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THE SEDIMENTARY EVOLUTION OF THE 'EXMOOR BASIN' DURING THE LATE EMSIAN - EARLY EIFELIAN: THE LYNTON FORMATION

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**THE SEDIMENTARY EVOLUTION OF THE 'EXMOOR BASIN' DURING THE
LATE EMSIAN - EARLY EIFELIAN: THE LYNTON FORMATION**

by

CHRISTOPHER JOHN POUND

A thesis submitted to the University of Plymouth in partial fulfillment for the degree of

DOCTOR OF PHILOSOPHY

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Christopher John Pound
24th September 1995

ABSTRACT

THE SEDIMENTARY EVOLUTION OF THE 'EXMOOR BASIN' DURING THE LATE EMSIAN - EARLY EIFELIAN: THE LYNTON FORMATION

by

CHRISTOPHER JOHN POUND

An integrated investigation of the sedimentological, ichnological and syn-sedimentary tectonic aspect of the late Emsian to early Eifelian Lynton Formation (Lynton Beds of previous studies) has revealed a varied mudstone-dominated shallow marine succession which accumulated in a rapidly subsiding rifted basin. The 'Exmoor Basin' developed in response to a Devonian phase of transtension associated with dextral shear along a fundamental east-west lineament to the north (the Bristol Channel Fault Zone). The Lynmouth - East Lyn Fault (a splay off the BCFZ?) was active throughout the deposition of the Lynton Formation and strongly influenced the depositional styles developed along its length. A re-evaluation of the β /subsidence curve for the 'Exmoor Basin' using the latest biostratigraphic and lithological thickness data indicates a pattern consistent with a strike-slip basin; Carboniferous thermal phase β /subsidence values suggest only 10% crustal thinning compared to values c. 50 % claimed by previous authors (Dewey 1982, Sanderson 1984).

The base of the exposed Lynton Formation is characterised by extensive intraformational slide deposits and the presence of phosphatic material which represents a highstand deposit that correlates with the eustatic transgressive T-R event Ic of Johnson *et al.* (1985). Following a period of gradual reduction in accommodation space the sequence was punctuated by a massive influx of sand and granule grade material deposited at the base of the Lynmouth - East Lyn Fault scarp. This material was swept together into a series of offshore sand ridges and a shoreface deposit adjacent to the fault scarp. A new process-response model has been developed to describe the offshore sand ridges that were moulded by a combination of semi-permanent trade wind induced geostrophic flow, oscillatory currents and (possibly) weak tides.

The central part of the Lynton Formation records a gradual upwards increase in relative accommodation space and decrease in the influence of semi-permanent currents; dysaerobic substrates became widespread and a localised anoxic mud developed offshore.

The transition into the overlying Hangman Sandstone Group was marked by the southward progradation of a sandy shoreline in the face of a period of world-wide eustatic sea-level rise. The older, more northerly shoreline was dominated by longshore currents whilst the younger shoreline preserved a mixed (lower energy) storm- and wave-dominated sequence. The rate of shoreline progradation was relatively slow and the Lynton Formation - Hangman Sandstone Group boundary is markedly diachronous; the thickness of the exposed Lynton Formation varies from 200m adjacent to the Lynmouth - East Lyn Fault, where previously unrecognised outliers of the Hangman Sandstone Group occur, to 250m some 5km down-palaeoslope.

Although the ichnofauna was locally diverse, with 27 distinct ichnotaxa recognised within the Lynton Formation, the succession was dominated by a gradation between straight *Palaeophycus tubularis* burrows and branching *Chondrites* systems reflecting the response of an organism tolerant to dysaerobic conditions.

The study demonstrates the value of integrating sedimentological, ichnological and structural techniques when studying Devonian marine shelf successions which accumulated in a tectonically unstable setting.

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At no time during the registration for the degree of Doctor of Philosophy has the author been registered for any other University award.

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Relevant scientific seminars and conferences were regularly attended at which work was often presented; several papers were prepared and published.

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Pound, C. J. 1983. The sedimentology of the Lower-Middle Devonian Staddon Grits and Jennycliff Slates on the east side of Plymouth Sound, Devon. *Proc. Ussher Soc.*, **5**, 465-472.

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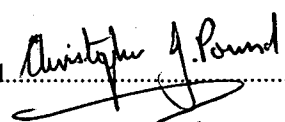
Ussher Society Annual Conferences:

4th - 6th January 1981 - Fowey Hotel, Fowey

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1. INTRODUCTION

1.1 PROLOGUE

The Lynton Beds (Simpson 1964, Tunbridge 1978, Tunbridge & Whittaker in: Goldring *et al.* 1978, Evans 1980, 1983, Edmunds *et al.* 1985, Bluck *et al.* 1989) are the oldest rocks in north Devon and crop out in the core of the Exmoor Anticline whose axis trends NNE through Lynton (enclosure 1). The succession is exposed along 11.4km of the north Devon coastline, between Ramsey Beach (6465 4940) to the west, and Ninney Well (7353 4957) to the east. The exposure continues inland as a narrow belt, to a fold-closure in the proximity of Oare (802 473). The contact between the Lynton Formation and the overlying Hangman Sandstone Group is conformable along the southern margin of the Lynton Formation outcrop, locally being offset by NW-SE and NE-SW trending late normal faults. The northern contact with the overlying Hangman Sandstone Group is delineated by the ESE - WNW trending Lynmouth - East Lyn Fault, a major reverse fault with a quoted downthrow in excess of 1800 to 2000m to the NNE (Edmunds *et al.* 1985). The exposed sequence, nowhere is the base of the Lynton Beds visible, has been estimated as being between 300 and 400m thick. The age of the Lynton Beds, based on conodonts and miospores recovered by Knight (1990a, b), is latest Emsian to early Eifelian. This contrasts with a late Emsian age indicated by the Brachiopod fauna (Evans 1980, 1983). The base of the Lynton Beds may be equivalent in age to the top of the Dartmouth Group of south Devon (Dineley 1966, Allen 1979a).

Lithologically the Lynton Beds predominantly comprise a millimetre-scale interlamination of lenticular bedded, heterolithic dark grey cleaved mudstones with siltstones and very fine to medium grained sandstones, locally containing thicker beds of wave-rippled and hummocky cross-stratified sandstones (Bluck *et al.* 1989). The sequence is intensely disturbed by bioturbation, so much so that Simpson (1957) coined the term "tunnel sandstone" for beds in the Woody Bay area penetrated by branching tubes of the trace fossil *Chondrites*. A macro-fauna comprising brachiopods, bivalves and crinoid ossicles is associated with lesser numbers of plant, bryozoan, stromatoporoid? and coral fragments, gastropods, tentaculitoids and, rarely, fish remains, orthocones and trilobites (Edmunds *et al.* 1985).

The fauna, repeated evidence of penecontemporaneous erosion and concentration of the shelly fauna into lumachelles, led Simpson (1964) to ascribe a shallow marine origin to the Lynton Beds, with evidence preserved that implied deposition during periods of calm alternated with storm-wave stirring of the substrate.

It is now generally accepted that the Lynton Beds accumulated on a wave-dominated shelf between storm and fair-weather wave-base and exhibits a gradual deepening from its lower horizons to the middle part of the succession (Bluck *et al.* 1989). The upper part of the succession and the transition into the overlying basal Hollowbrook Formation of the Hangman Sandstone Group represents the approach to a sandy, low wave-energy shoreline (Tunbridge 1983a).

Localised occurrences of a matrix-supported intraformational conglomerate have been reported immediately adjacent to the Lynmouth - East Lyn Fault (Tunbridge & Whittaker in: Goldring *et al.* 1978, Edmunds *et al.* 1985). This was named the 'Lyn Conglomerate' by Tunbridge (1986) who interpreted the conglomerate to be the product of debris flows sliding down a submarine fault scarp formed by syn-sedimentary tectonic movements on the Lynmouth - East Lyn Fault, a splay off the Bristol Channel Fault Zone.

The geographic/geomorphologic setting of the Lynton area is described in appendix A.

NOTE: *From this point onwards the sequence previously referred to as the 'Lynton Beds' will be referred to as the 'Lynton Formation' in accordance with the new stratigraphical designation proposed in section 1.8.3 of this thesis.*

1.2 OBJECTIVES

Goldring and Langenstrassen (1979) and Tsien (1989) have emphasised the influence of syn-sedimentary tectonic movements on facies distribution in the Devonian of the Rhenish Schiefergebirge and Ardenne, with siliciclastic rocks the product of physical processes of rapid sedimentation whilst carbonate rocks are mainly the product of biological and chemical processes during tectonically calm periods. The equivalent thick, presumed continuous Devonian sequence in north Devon similarly developed on the tectonically unstable margin of the 'Old Red Sandstone Continent' and records the interdigitation of periods of continental and shallow marine deposition. Although detailed sedimentological studies and accompanying palaeoenvironmental reconstructions have been documented for some of the more arenaceous intervals (Baggy Sandstones - Goldring 1971; Hangman Sandstone Group - Tunbridge 1978), the argillaceous intervals have received scant attention. For example, recent studies in north Devon (Tunbridge 1978, Evans

1980, 1983, Edmonds *et al.* 1985) have drawn attention to the lack of a detailed sedimentological analysis of the Lynton Formation and the need for a formal lithostratigraphic scheme.

The initial objectives of this study, therefore, were to:

1. Describe the Lynton Formation sequence using modern sedimentological techniques.
2. Relate the prolific ichnofauna to lithofacies, and the physical environment that they accumulated in, to enable ethological models to be constructed/refined for the preserved ichnotaxa.
3. Place the sedimentary sequences into a regional context to elucidate the depositional history of north Devon during late Emsian - early Eifelian times.
4. The Lynton Formation contains several distinctive sandstone-body types which provide new insights into the deposition of sand-bodies on muddy shelves. Facies sequence models will be developed to illustrate the sandstone-body types.
5. Place the strata previously referred to as the *Lynton Beds* into a formal lithostratigraphic scheme.

N.B. Much of the Lynton Formation comprises a monotonous muddy heterolithic facies. In order to avoid repetition, this facies type is discussed in some detail in chapter 4 where some of the best examples of the facies are described. For this reason chapters 3 and 5, where muddy heterolithic facies from other levels in the Lynton Formation are described, are included for the sake of completeness. *The sedimentological models presented in chapters 3 & 5 are consequently somewhat more superficial than those presented elsewhere in this thesis.*

1.3 METHODS

An initial reconnaissance of all the available Lynton Formation exposures revealed that no complete section through the Formation is available; the most complete sections were observed on the coast. Due to lichen-cover and poor exposure inland (see Appendix A), and the complex nature of the finely interlaminated heterolithic strata, it was decided that a limited number of representative coastal sections and more favourable inland exposures would be logged in fine detail. Although the coastal sections are well exposed, access problems abound due to the majority of the coastline being composed of steep cliffs rising directly

from boulder-strewn intertidal zones swept twice daily by the strong tidal currents of the Bristol Channel. Rope techniques were used to reach the more inaccessible exposures, although an inflatable dinghy was employed to reach exposures too dangerous to approach by rope. Other exposures, particularly between Woody Bay and Heddon's Mouth, could only be approached by traversing along the base of the cliffs. The latter technique, which could often only be employed on favourable spring tides, allowed a reconnaissance of exposures in order to ascertain whether any unusual facies variations were exposed; repeated return visits were impossible. Fortunately, all the exposures finally chosen for study in detail were *relatively* easy of access using the techniques described above, coastal exposures generally being accessible during the low tide period of all tides. Nevertheless: *the reader is cautioned that many of the coastal exposures are significantly more craggy than those typically encountered during normal geological field work - a fall in these circumstances, particularly below high-water mark on this inaccessible coastline, could have potentially serious consequences.*

In order that subtle variations could be resolved within the heterolithic strata which dominate the succession logging was carried out on a centimetre-by-centimetre scale. Logs were initially drawn up at a 1:10 scale. Sample blocks were collected from horizons where sedimentary structures were not revealed by weathering and were slabbed, polished and matt varnished in the laboratory. This technique was only successful with heterolithic rocks, the sandstones generally lacked sufficient granulometric contrast to reveal structures. Due to the frequently steep dip, with strata running up precipitous and inaccessible cliffs, and also abundant faulting, it was usually impossible to trace facies variations laterally. Where exposure did allow lateral facies variations to be observed, 3-D fence diagrams were constructed or photomontages were taken. A full set of 20 X 20cm overlapping photographic plates were taken during a Wessex helicopter flight kindly arranged by the R.A.F. Air-Sea Rescue base at Chivenor.

Although prolific, the ichnofauna was difficult to collect, particularly in coastal exposures where the compact nature of the wave-washed slab resisted collection with a hammer and chisel. Therefore, photographs were obtained and, where weathering accentuated relief in ichnofossils, latex rubber impressions were taken. Positive replicas were then cast in the laboratory, using Strand Glass polyester resin coloured with Strand Glass slate grey. All specimens are now housed in the Department of Geological Sciences, University of Plymouth.

1.4 THESIS STRUCTURE

During the course of the present study a wide range of interrelated themes were explored: process sedimentology (primary sedimentary structures and their hydrodynamic significance), ichnology, facies sequence analysis and palaeo-environmental reconstruction, structural geology (particularly syn-sedimentary deformation), basin formation processes and lithostratigraphy. It was decided that the thesis would be structured around the central theme of the sedimentary evolution of the 'Exmoor Basin' during the period that the Lynton Formation was deposited.

The lack of distinctive marker horizons, and uncertainty regarding the lateral equivalence of sections due to structural complications, meant that it was not possible to erect a formal lithostratigraphic hierarchy for the Lynton Formation. Nevertheless, a number of distinctive lithofacies and facies associations were recognised and these are defined in section 1.8.

There is strong sedimentological and structural evidence that the Lynton Formation was deposited in a tectonically unstable setting. These lines of evidence are drawn together in chapter 2 where they are described in the context of the structural geological evolution of the Lynton area.

The body of the thesis is concerned with the detailed description of the suite of physical sedimentary and biogenic structures, their grouping into facies associations and palaeoenvironmental reconstruction based on analogous sequences documented from both the geological record and Recent environments. The detailed description, taxonomy and ethological interpretation of the associated ichnotaxa has been separated out into appendix B. For the purpose of sub-division into chapters a series of 'mega-facies' were defined based on the grouping of a common broad set of characteristics recognised at particular levels within the Lynton Formation. The mega-facies scheme is described in section 1.7.

Finally, chapter 8 discusses the processes that formed the 'Exmoor Basin' and controlled its subsidence history prior to describing the detailed basin-fill history during late Emsian - early Eifelian times.

1.5 HISTORY OF RESEARCH RELATING TO THE LYNTON FORMATION

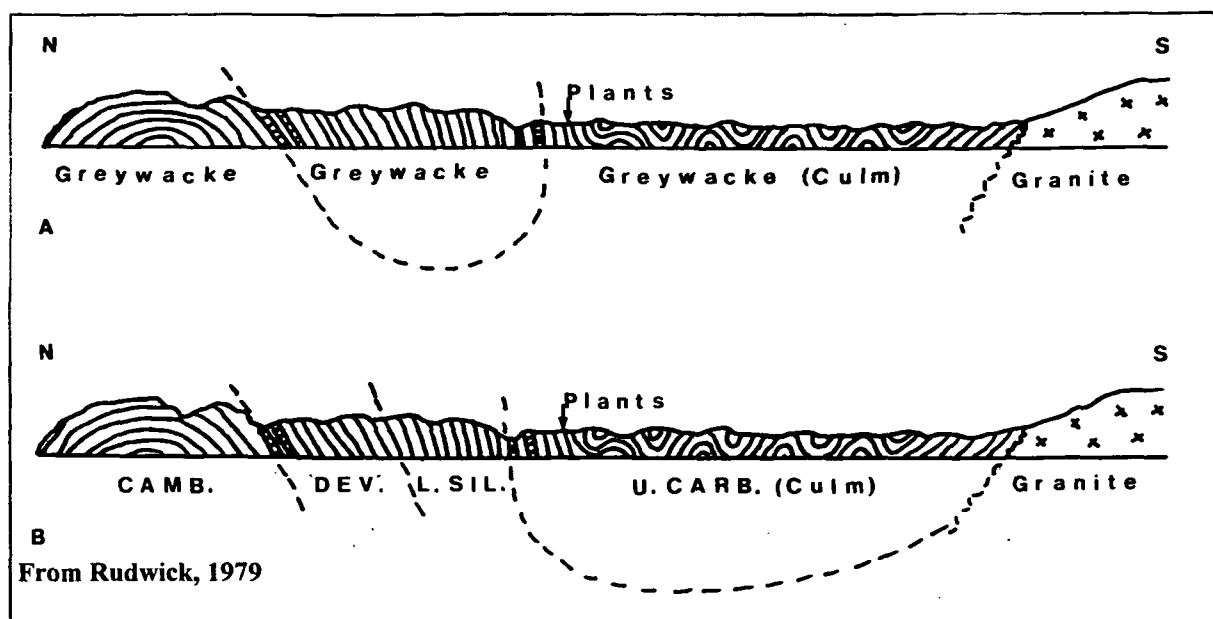
The early history of research into the Lynton Formation is directly related to the unravelling of the stratigraphic succession in SW England, which led to the erection of the Devonian System by Sedgwick and Murchison (1839). The history and controversy that surrounded the foundation of this system is more fully documented by Rudwick (1979, 1985). Early research was primarily directed towards lithological and faunal subdivision and correlation, 'way-up' criteria being unavailable in the mid-Nineteenth century to allow a structural analysis of the succession. The faunal lists compiled during this period are discussed by Evans (1980) and Edmunds *et al.* (1985), particularly the brachiopod faunas.

The earliest attempt to subdivide the rocks of SW England was that of Conybeare (1823), who described four divisions: Granite, "Metalliferous Slates", Slate and "Stratified Rock". By 1835, when the geological maps prepared by de la Beche were published, the only boundary shown within the Palaeozoic succession was that between the Culm series of central Devon and the "Grauwacke" to the north and south.

1837 saw the commencement of a series of papers providing a more detailed subdivision of the strata in north Devon. Williams (1837) divided the carbonaceous strata into nine groups, his "Linton slates and limestones" occurring above the "Cannington Park limestone" (base) and "Foreland and Dunkery sandstones". The previous summer, Sedgwick and Murchison had undertaken a geological traverse of Devonshire, reporting their results (Sedgwick & Murchison 1837a) to the British Association meeting. They concluded that a broadly synclinal structure containing Culm occurred in central Devon, with "Lower Silurian" and "Cambrian" strata rising to the surface in north Devon, the term "Devonian" being used to denote "Upper Cambrian" strata. They subdivided the succession in north Devon on lithological grounds, recognising that the sequence was folded into a major anticline with the oldest rocks, the "calcareous group of Linton" (*sic*) exposed in the core. Sedgwick and Murchison's views on the sequence in Devonshire contrasted with de la Beche's interpretation of a synclinal structure in north Devon, which repeated the greywacke (Culm) of central Devon in north Devon (compare A and B in text-figure 1.1).

Weaver (1838), in describing the structure of the area between Bideford and the Foreland, placed the "Linton calcareous slates" above the "Foreland sandstone", the lowest of his eight divisions, adopting the term "transition group" for the conformable sequence below the Culm. The "Foreland sandstones" were also

recognised by Sedgwick (1838), placing “a series of coarse arenaceous slates” (p.680) below the “calcareous slates of the river Lyn”.



Text-Fig. 1.1 The structure of central and north Devonshire: two alternative interpretations from the 1830's

Redrawn and slightly simplified for direct comparison. **A.** De la Beche's interpretation in November 1834 (from a letter to Sedgwick). The correlation of two limestone bands in north Devon led him to infer a synclinal structure there, so that the fossil plants from the Culm series near Bideford appeared to be well down in the total 'Greywacke' succession, which de la Beche at this time believed to be entirely pre-Silurian. **B.** Sedgwick and Murchison's interpretation in August 1836 (from a newspaper report of the British Association meeting). Their conviction that the fossil plants from the Culm series must be of Coal Measures age led them to infer a broadly synclinal structure for central Devon, with 'Lower Silurian' and 'Cambrian' strata rising to the surface in north Devon. The term 'Devonian' was used to denote the 'Upper Cambrian' strata. Note that no explicit criteria for 'way-up' in vertical or overturned strata were available to geologists in the 1830's (From: Rudwick 1979).

In 1839, under pressure to publish in order to vindicate his interpretation of the sequence in Devonshire, de la Beche produced his: "Report on the Geology of Cornwall, Devon and West Somerset". Detailed lithological descriptions and faunal lists were presented, along with sketch sections. The "Linton grey beds" were placed in his "grauwacke group" above the "sandstones of Foreland Point". 1839 also witnessed the publication of a paper by Sedgwick and Murchison which further amended their 1837 paper. Sedgwick's fieldwork in Devon the previous summer had shown the central Culm to grade conformably into older strata, as de la Beche had previously maintained. A date younger than Lower Silurian seemed probable for the strata and the term "Devonian System" was introduced for the greywacke rocks that lay to the north and the south of the Culm and equivalent to the Old Red Sandstone. In defining their new system, they acknowledged the work of the palaeontologist William Lonsdale (later published in 1840) who recognised that corals from the limestones of south Devon were intermediate between Silurian and Carboniferous forms.

In 1840 Sedgwick and Murchison published their *magnum opus*: "On the physical structure of Devonshire, and on the subdivisions and geological relations of its older stratified deposits." They now recognised that the oldest rocks in north Devon were to be found in the East Lyn valley, in the core of the 'Exmoor Anticline', the rocks of Foreland Point being equivalent in age to those found between Woody Bay and Combe Martin. They also predicted a "downcast fault to the north" (p.644) in the area east of Lynmouth. Their "Group 1" (i.e. Lynton Formation) were described as slightly micaceous, siliceous sandstones contained in cleaved chloritic beds, the base of the group not being exposed. The top of the group, exposed in the Valley of Rocks, passed under a succeeding group of ". . . red micaceous and siliceous flagstones." Sedgwick and Murchison's interpretation of the position of the Foreland Grits differed from their earlier views, and those of Weaver (*op. cit.*) Indeed, the precise stratigraphical position of the Foreland Grits was not satisfactorily resolved until the study of Tunbridge (1978) who recognised that they formed part of the Hangman Sandstone Group.

After the intense period of research in the 1830's the pace of activity slowed in north Devon. In 1865 Jukes produced a correlation of the rocks of north Devon with the Old Red Sandstone of Ireland. The great thickness of rock exposed in north Devon was explained by a fault postulated to occur between Morteohoe and Wiveliscombe, which repeated the sequence. The existence of this fault was refuted by Etheridge (1867) who produced palaeontological evidence that demonstrated that the sequence in north Devon was conformable. Additionally, the latter paper contained a sketch map showing "Lynton sandstones" (Foreland Grits) below the "Lynton Slates", the latter group occurring as an outlier in the Quantocks. In the same year, Hall (1867) also produced a sketch map. It showed the Devonian of north Devon divided into seven groups, the "Lynton zone" occurring above the "Foreland Group". At the beginning of this century, Ussher (1906) combined the maps of the latter two workers with his own of north Devon and west Somerset. This map showed the Lynton Beds occurring above the Foreland Grits and below the Hangman Grits.

In 1910 Hamling prepared a map for the Geologists Association field meeting in north Devon (Hamling & Rogers 1910). As discussed by Goldring (1952), this map combined all the existing information and therefore closely resembled Ussher's map, the only structural addition being the faulted boundary between the Lynton Beds and Hangman Grits on the coast east of Lynmouth. This map has served as the definitive geological map of north Devon until relatively recently.

The next major phase of research into the Lynton Formation was undertaken by Simpson. In 1951, in a paper discussing problems in the stratigraphy of the marine Devonian of Britain, he pointed out that most workers had accepted Weaver's (1838) succession in which the Foreland Grits underlay the Lynton Formation. Simpson preferred the interpretation of Sedgwick and Murchison (1840) in which the Lynton Formation crops out in the core of the 'Exmoor Anticline', the Foreland Grits being the downfolded equivalents of the Hangman Sandstone Group on the northern limb of the anticline.

In 1953 the Geologist Association undertook a second excursion to the Lynton area (Simpson & Kidson 1954). Although the main objective of the excursion was to discuss the effects of the 1952 flood upon morphology, exposures in the Lynton Formation were examined to the east of Lynmouth Beach, where tight recumbent folding was recognised, and in the East Lyn Valley. In the latter exposures, large U-shaped burrows were observed which they compared to *Arenicolites subcompressus* (Eichwald) of the German Rhaetic.

Further trace fossil material from the Lynton Formation was discussed by Simpson (1957) when he documented occurrences of *Chondrites* on the eastern side of Woody Bay, the term "tunnel sandstone" being applied to particularly dense burrow populations. Simpson's unpublished notes indicate that he intended to revise the nomenclature of the Lynton Formation fauna. However, ill-health precluded completion of the work and the notes have been subsequently lost, the only published contribution being a short paper in 1964. This paper gave a faunal list which suggested a late Emsian to early Eifelian age for the Lynton Formation, a date subsequently used in House *et al.* (1977). Simpson perceptively attributed the occurrence of thick sandstone beds and lumachelles in the Lynton Formation to storm deposition on a normally quiet sea-bed.

More recently a resurgence in interest in north Devon has taken place. Tunbridge (1978, 1980, 1981b, 1983a, b, 1986) examined the transition from the uppermost beds of the Lynton Formation into the succeeding Hangman Sandstone Group, concluding that the sequence represents a series of regressive phases in an approach to a low energy, wave-influenced sandy shoreline, the coast being supplied with sediment from the north by ephemeral streams. Evans (1980), in a review of the Early Devonian brachiopod faunas of Britain, concluded that the Lynton Formation fauna revealed a late Emsian age, although an early Eifelian age could not be discounted for the uppermost beds (Evans 1983), and was deposited in a near-shore, "sub-tidal" setting.

Hamling's 1910 map was finally superseded in 1981 when the Ilfracombe (Sheet 277) geological map was published by the Institute of Geological Sciences. This sheet shows the Lynton Formation, termed "Lynton Slates", as a 300 to 400m thick group of sediments occurring in the core of the 'Exmoor Anticline', bounded to the north by the Lynmouth-East Lyn reverse fault. The accompanying sheet memoir was published in 1985 (Edmonds *et al.*) The British Geological Survey have now mapped the adjacent Minehead (sheet 278) area and are in the process of publishing the results (Dr. R. A. Edwards, B.G.S. Exeter Office - pers. comm.)

Most recently Knight (1990a, b) undertook a detailed integrated micropalaeontological investigation of the Lynton Formation, Hangman Sandstone Group, Ilfracombe Slates and Morte Slates. A date no older than latest Emsian age was established for the Lynton Formation, with the uppermost horizons of possible earliest Eifelian age.

1.6 REGIONAL PALAEOGEOGRAPHIC, TECTONIC AND STRUCTURAL SETTING

1.6.1 Palaeomagnetism and the Palaeo-latitudinal Setting

Although: "Devonian paleomagnetic data are not yet sufficient to resolve many paleogeographic problems, and much of the existing data appears to be unreliable" (Witzke & Heckel 1989, p.51), the majority of published work points to a palaeo-latitude of between 15°S and 30°S for southern Britain during the late Early Devonian - see table 1.1. This palaeo-latitudinal range is consistent with sedimentological evidence which shows that the Lower Old Red Sandstone of southern Britain was deposited in an arid to semi-arid environment (Allen 1974a, b, 1979a) and that the Devonian limestones of SW England were deposited in a warm, tropical sea (House 1975).

| Year | Author(s) | Palaeo-latitude | Time | Notes |
|------|-------------------------------------------------|-----------------|------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| 1973 | Smith, Briden & Drewery | 25°S | Early Devonian | |
| 1979 | Scotese, Bambach, Barton, Van der Voo & Ziegler | 0° | Emsian | Noted that no reliable Early Devonian poles were available from North America for the reorientation of Laurussia |
| 1981 | Briden & Duff | 20° - 30°S | Early Devonian | |
| 1984 | Briden, Turnell & Watts | 20° - 30°S | Early Devonian | |
| 1984 | Scotese | 15°S | Late Emsian interpolation | Early Devonian ("Gedinnian-Siegenian" i.e. Lochkovian-Pragian of revised Devonian timescale - Ziegler & Klapper 1985) = 20°S; Late Devonian (Fammenian) = 10°S |
| 1985 | Livermore, Smith & Briden | 15°S | Late Emsian interpolation | Ludlow - Emsian = 20° - 30°S; Eifelian - Tournaisian = 5°S |
| 1985 | Scotese, Van der Voo & Barrett | 15°S | Late Emsian interpolation | Early Devonian ("late Gedinnian - early Siegenian" i.e. late Lochkovian - early Pragian of revised Devonian timescale - Ziegler & Klapper 1985) = 20°S; Late Devonian (late Frasnian - early Fammenian) = 10°S |
| 1989 | Witzke & Heckel | 30°S | Mid Devonian | Palaeolatitude based on a combination of palaeomagnetic, palaeoclimatic, palaeobiogeographic and tectonic evidence. |
| 1990 | Scotese & McKerrow | 20°S | Late Early Devonian (Emsian) | |
| 1990 | Torsvik, Smethurst, Briden & Sturt | 30°S | Early Devonian (380Ma) | 380Ma equates approximately with the Eifelian/Givetian boundary (Harland <i>et al.</i> 1990) |
| 1990 | Kent & Van der Voo | 5°S | Early Devonian | |

Table 1.1 - Published palaeo-latitude interpretations for southern Britain during the late Early Devonian

1.6.2 Palaeogeographical Setting

Tsien (1989) presented a series of palaeogeographic maps spanning the Devonian Period for north-western and Central Europe. The Emsian map showed a NE-SW trending Variscan geosyncline (cf. 'Rheic Ocean') on the southern flank of the 'Old Red Continent'. The geosyncline was divided by a 'Normannian-Mid German High' (discontinuous) separating the 'Cornwall-Ardenno-Rhenish Basins' to the north from the 'Saxothuringian-Barandian and Central Armorican Basins' to the south. The southern flank of the geosyncline was bounded by the 'Ligerian-Arverno-Vosgian-Moldanubian Cordillera' and was connected at its eastern end with the 'Proto Tethys' ocean and 'Uralian Seaway' to the south and south-east respectively.

More locally, the reader is directed to a paper by Bluck *et al.* (1989) which provides a comprehensive review of the palaeogeographic and tectonic setting of the Devonian of England, Wales and Scotland. The following sections, therefore, focus on any area known (or suspected) to contain late Emsian to early Eifelian strata in

southern Britain and published work that supplements the Bluck *et al.* account. For reference, text-figure 1.2 reproduces the Early Devonian palaeogeographic reconstruction of Britain, and columns indicating the possible relationships between Devonian rocks of south Wales, Devon and Cornwall from Bluck *et al.*

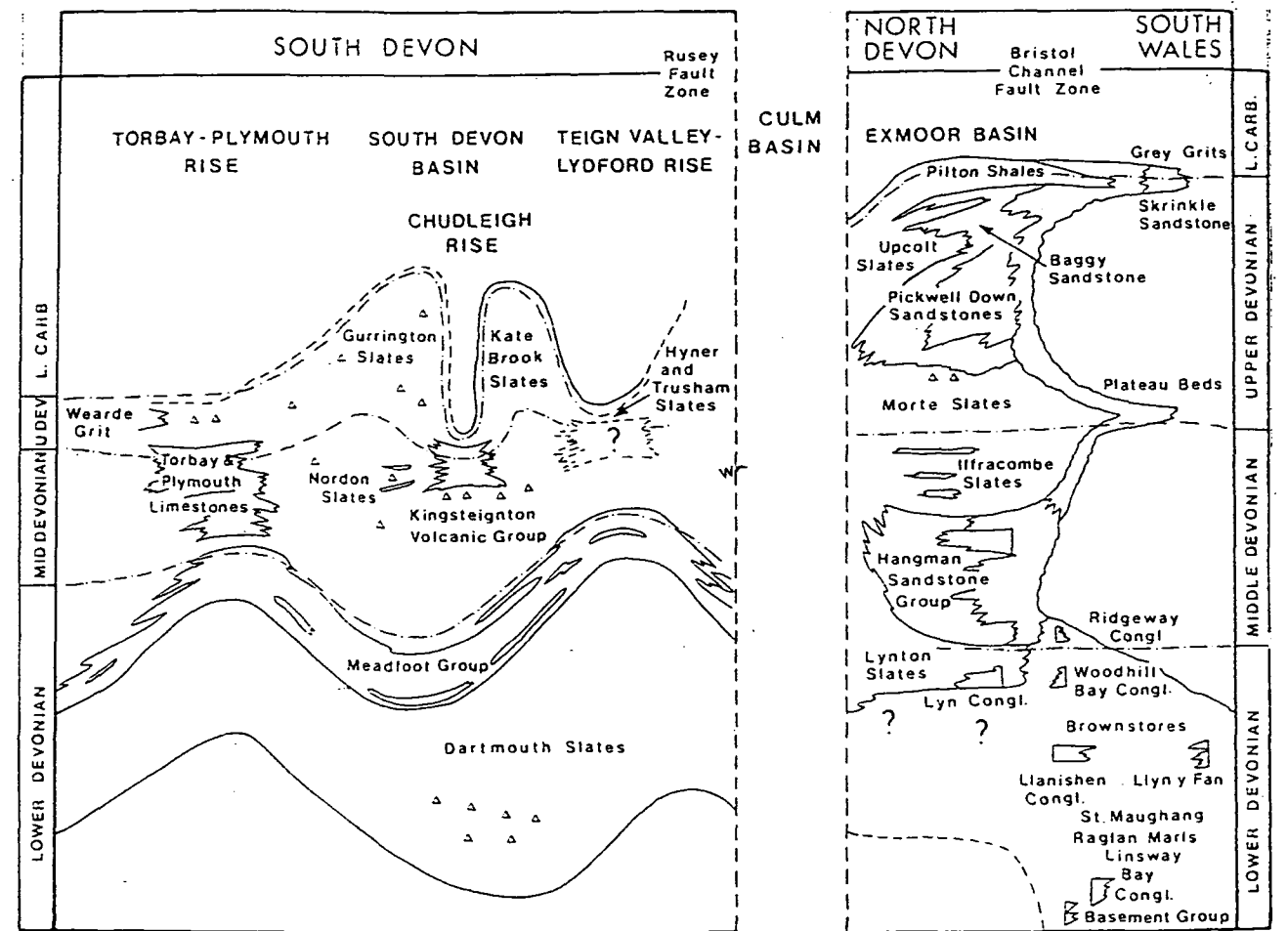
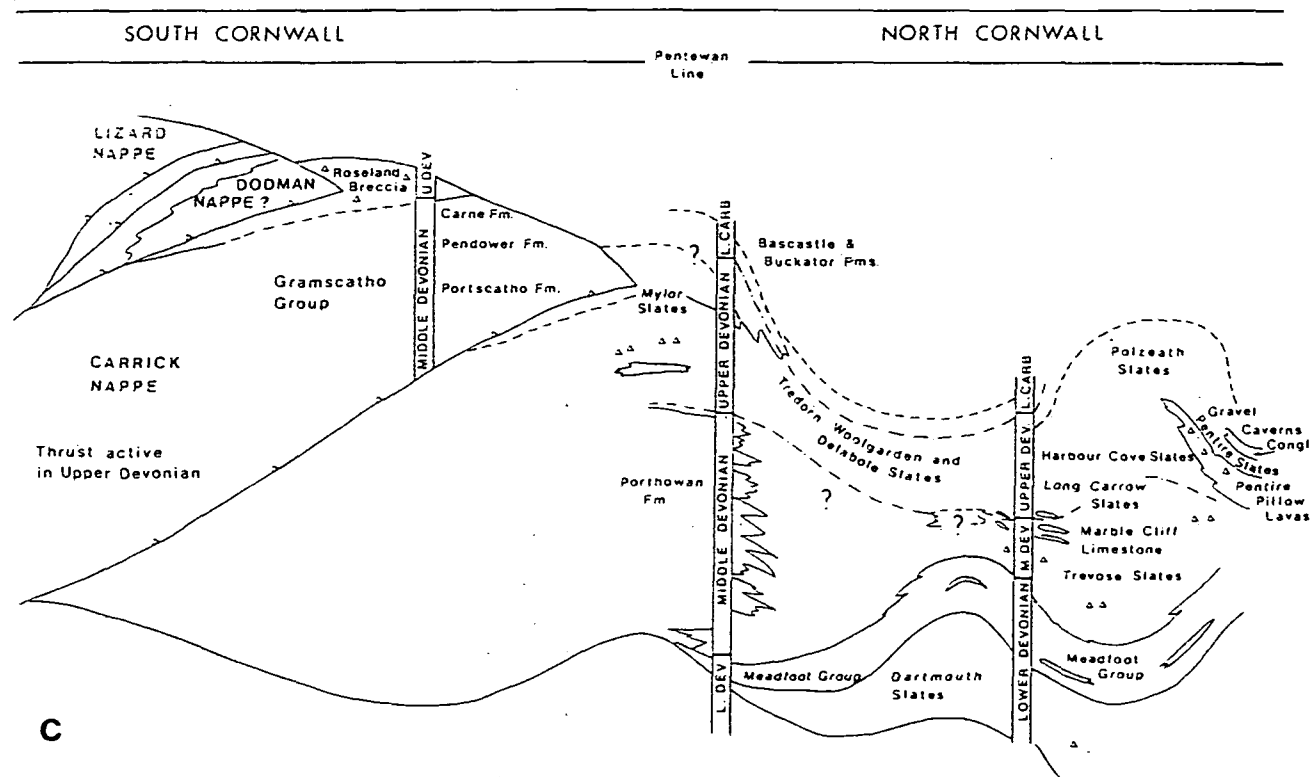
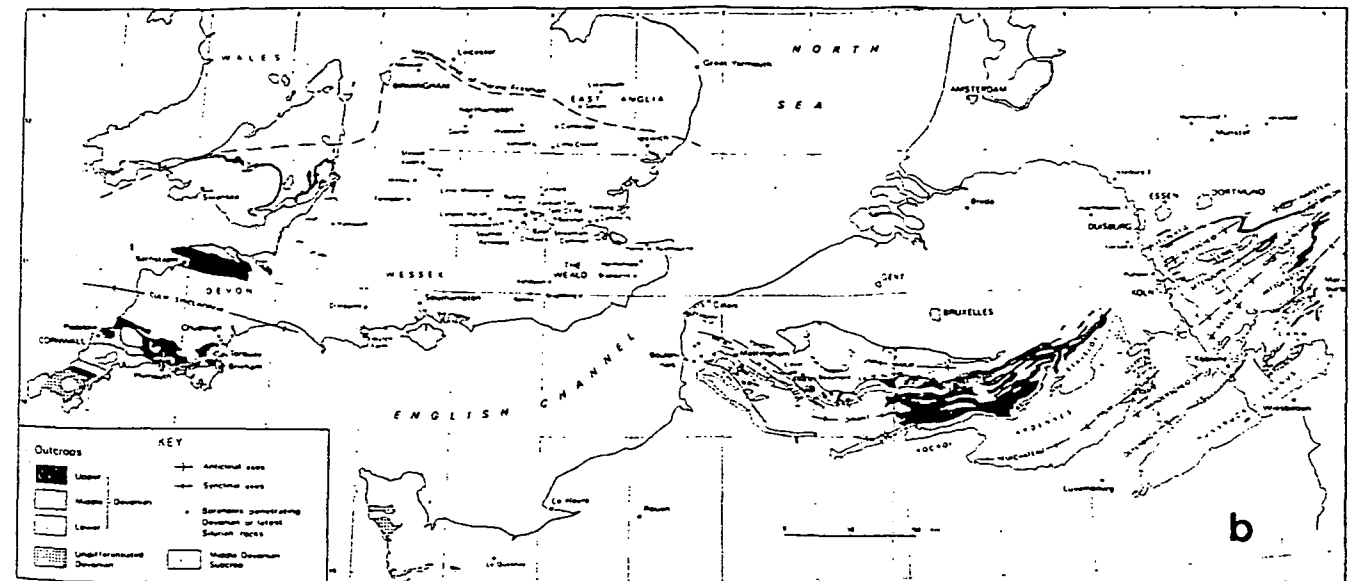
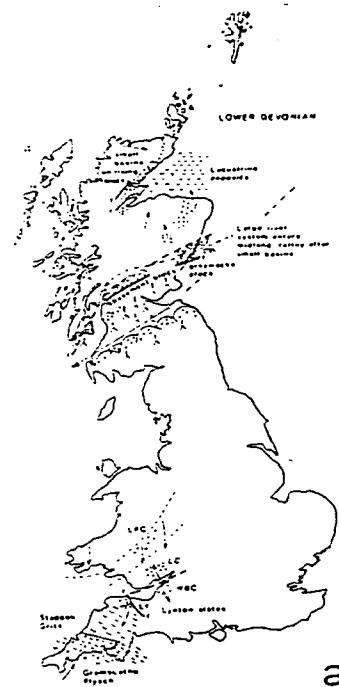
1.6.2.1 Old Red Sandstone Magnafacies

1.6.2.1.1 Southern Ireland

The only possible strata of late Emsian to early Eifelian were reported by Allen (1979a) who placed the fluvial and aeolian Caherbla Group of the Dingle peninsula in the Emsian. More recent biostratigraphic studies, however, now place the Caherbla Group in the Late Devonian (Graham & Clayton 1989).

1.6.2.1.2 South Wales

During the Early Devonian, thick developments of Old Red Sandstone accumulated in southern Britain, derived dominantly from upland regions to the north and north-west (Allen 1974a, Bluck *et al.* 1989). The Brownstones are the youngest Lower Old Red Sandstone deposits preserved in Wales and the Welsh Borderland, forming a broad belt stretching from Carmarthen Bay to the north and east of the South Wales coalfield and extending eastwards to the Forest of Dean. Isolated outcrops occur near Portishead on the southern margin of the Bristol Channel (Pick 1964). Tunbridge (1981a) has carried out a detailed sedimentological investigation of the Brownstones of central south Wales, where a series of red siltstones and red to brown sandstones display a broad coarsening-upwards trend. Tunbridge attributed the coarsening-upwards trend to a southerly migration of the Early Devonian 'fall-line' (Allen 1974a), with thin-bedded 'distal' sheet flood sandstones, enclosed in floodbasin silts, passing upwards into 'medial' low sinuosity channels interbedded with floodbasin silts, finally passing upwards into 'proximal' low sinuosity channels forming multi-storey sandstone bodies interbedded with 0.3 to 1m thick siltstones. Geographically isolated coarse clastic episodes interrupt the alluvial plain sequences (Williams 1964, Allen 1975, Tunbridge 1986) and are interpreted as being shed from localised uplift along faults within the basin with movement associated with late Caledonian transpression (Tunbridge 1986, Woodcock 1987).



Text-fig.1.2 Late Early Devonian Palaeogeography

a. Generalised reconstruction of Devonian palaeogeography during the 'Lower' Devonian. b. Map showing the relations of the Devonian rocks of southern Britain to the Ardennes and Rhenish Schiefergebirge. c. Diagram illustrating possible relationships between the Devonian rocks of south Wales, north Devon, south Devon, north Cornwall and south Cornwall. From Bluck *et al.* (1989) figure 6a, 7 plus 8 & 9 combined.

In SE Wales Lower Old Red Sandstone sedimentation ceased during the mid to late Emsian i.e. palynological data presented by Richardson and Rasul (1979) from the Lower Old Red Sandstone indicated a mid-Emsian age for the youngest recovered samples. Active uplift of the south Wales area was underway by the Eifelian, with Lower Old Red Sandstone deposits being reworked and deposited to the south (Bluck *et al.* 1989). Figure 8 (reproduced as part of text-figure 1.2c) of Bluck *et al.*, however, shows the Old Red Sandstone magnafacies extending up into the Mid Devonian and punctuated by two conglomeratic episodes (Tunbridge 1986): the Woodhill Bay Conglomerate (shown as late Early Devonian) of the Bristol area and the Ridgeway Conglomerate (shown as early Mid Devonian) in Dyfed. There are no published biostratigraphic indicators, however, to support these dates and, based on palynological data from the enclosing laterally equivalent Brownstones, an older (i.e. pre- late Emsian) date is more likely for the two conglomerates e.g. Powell (1989 figure 6) suggests a mid Pragian age for the Ridgeway Conglomerate.

SW Dyfed is divided into a series of fault-bounded blocks, each with its own Devonian stratigraphy directly related to pre-Variscan tectonic control on basin bounding faults in an overall extensional environment (Powell 1989). The southerly downthrowing basin-bounding faults were interpreted as a series of synthetic faults to the faults delineating the 'Bristol Channel Landmass' palaeohigh. Figure 6 of Powell showed Brownstones deposition extending into the late Emsian in the Broad Haven basin. To the south, in the Winsle Block, deposition of the fluviatile Cosheston Group also extended into the late Emsian, with a coarse alluvial fan deposit (the New Shipping Formation, derived from the north) developing locally in the hanging-wall of the Benton Fault. Following the period of active extension during the Early and Late Devonian, the Lower Carboniferous was marked by more passive subsidence prior to reactivation of the extensional fault system during Variscan compression.

In southern Britain, therefore, the Lower Old Red Sandstone magnafacies appears to be restricted to SW Dyfed and concealed strata in SE England reported from borehole studies:

1.6.2.1.3 Concealed Devonian of SE England

Faringdon No. 1 Borehole: Falcon (1955) and Falcon and Kent (1960) record 299m of red and purple indurated sandstone, shales and siltstone including occasional pebble bands and limestone horizons. Allen (1979a) attributed the beds to a proximal and medial alluvial origin. Spores, described by Mortimer and

Chaloner (1972), indicate a date no older than late Emsian, whilst the occurrence of *Ancyrospora* (Mortimer 1967), if proved, would give a date no younger than late Emsian.

Canvey Island Borehole: Smart *et al.* (1964) described red and brown siltstones and mudstones, arenaceous and rudaceous rocks and some intraformational conglomerate, recording: plant debris, spores, acritarchs and ostracods from this borehole. Palynological evidence yielded a late Emsian age (Mortimer 1967). Allen (1979a) noted that the facies described are similar to fining-upwards sequences observed in the Apply Barn borehole where mainly grey and green sandstones with conglomerates of mixed intraformational and exotic type lie with grey, purple, and red mudstones in fining-upwards sequences 5 to 10m in thickness (Poole 1969). Allen (*op. cit.*) noted that the occurrence of the ostracod *Leperditia* sp. in the Canvey Island borehole suggests a marine influence, assigning the sequence to a distal alluvial setting. It is apparent, therefore, that only in concealed late Emsian deposits in SE England (Faringdon No.1 & Canvey Island) are deposits typical of Old Red Sandstone alluvial facies preserved.

1.6.2.2 Marine Deposits - SE England

It has long been understood that Devonian shelf and near-shore clastic facies successions in NW Europe developed in a tectonically unstable setting (Goldring & Langenstrassen 1979). During the past 15 years there has been a upsurge in published accounts relating the previously enigmatic Devonian and Carboniferous succession in SW England to syn-sedimentary basin development and tectonics, although much work still remains to be done in unravelling this tectonically complex region. For more detailed reviews of basin evolution in SW England during the Devonian and Carboniferous the reader is referred to: Selwood and Durrance (1982), Dineley (1986), Bluck *et al.* (1989) and Selwood (1990).

Devonian deposition in SW England north of the Lizard was initiated on continental crust that formed the southern margin of the 'Old Red Sandstone Continent' (Selwood 1990). During the Lochkovian and early Pragian thick continental deposits of the Dartmouth Group extended southwards, but late Pragian transgression resulted in the progressive northwards onlap of marine deposits which accumulated in extensional basins, each of which had a different sequence fill and subsidence history. Matthews (1977), in a seminal review of the upper Palaeozoic successions in SW England, recognised a series of east-west trending basins which were developed and filled sequentially from south to north in front of the advancing Variscan

deformation front: 'Roseland Basin' ('Gramscato Basin' of later accounts), 'Trevone Basin', 'Bude Basin', 'Exmoor Sink'; Selwood & Thomas (1986) subsequently recognised an additional basin with its own discrete history of deposition: the 'South Devon Basin'.

1.6.2.2.1 'Exmoor Basin'

The term 'Exmoor Basin' follows Webby (1966) in preference to Matthews (1977) name of 'Exmoor Sink'.

Quantock Hills:

Webby (1965b) described a series of light brown and brownish-grey siltstones and silty slate containing *Chonetes* sp., *Tentaculites* sp. and crinoid ossicles, which crop out in the core of the eastward plunging Courtway Anticline (ST 148 366 to ST 156 356). Webby named these the Little Quantock Beds and postulated that they may be of Lynton Formation age or younger. This interpretation differed from that of Ussher (1908), who believed the strata represented downfaulted Ilfracombe Slates. The outcrop of the Little Quantock Beds is now restricted to a few doubtful blocks in banks and ditches. Evans (1980) only retrieved crinoid fragments and the significance, therefore, of the Little Quantock Beds must remain equivocal.

North Devon Coast:

The shallow marine Lynton Formation (Simpson 1964, Tunbridge 1978, Tunbridge & Whittaker in: Goldring *et al.* 1978, Evans 1980, 1983) crops out in the core of the Exmoor Anticline (enclosure 1) and the base of the Lynton Formation is therefore not exposed. The base of the Lynton Formation may be equivalent in age to the top of the Dartmouth Group of south Devon (Dineley 1966, Allen 1979a). The possibility of concealed Early Devonian strata occurring beneath Exmoor is discussed in section 1.6.4.4 where the interpretation of the negative gravity anomaly in the Exmoor region is examined.

The regressive transition from the Lynton Formation into the overlying Eifelian (Knight 1990a) Hangman Sandstone Group (Tunbridge 1978, 1983a,) records an approach to a wave-dominated shoreline. Tunbridge (1978) divided the Hangman Sandstone Group into five formations: the basal Hollowbrook Formation (Tunbridge 1983a), a unit 70m in thickness, composed of shoreface sandstones and offshore heterolithic beds. The Trentishoe Formation, c.1 250m of continental sheetflood sandstones, thin floodbasin siltstones,

laminated and desiccated ephemeral lake mudstones (Tunbridge 1981b) with a top member, the Yes Tor Member (Tunbridge 1980), composed of a sequence of thick purple siltstones with pedogenic carbonate nodules, reaching a thickness of 20m. The overlying coarse-grained sandstones and conglomerates of the Rawns Formation (148m) represent a short-lived episode of high-energy alluvial sedimentation with angular clasts derived from a nearby northerly source. A marine transgression marks the base of the Sherrycombe Formation, a succession composed of a series of upwards-coarsening sequences representing estuarine or fan-delta deposits. The Little Hangman Formation (c. 100m) records more open marine conditions in a series of interbedded, heterolithic grey mudstones and sandstones which pass upwards into the earliest Givetian to Frasnian Ilfracombe Slates (Knight 1990a).

1.6.2.2.2 South Devon and Cornwall North of the Lizard

During the upper Palaeozoic the area covering south Devon and Cornwall north of the Lizard was dominated by two basins: the 'South Devon Basin' and 'Trevone Basin' (see review in Selwood 1990).

'South Devon Basin'

The 3-4km thick Dartmouth Group (Smith & Humphreys 1989, 1991) are thought to represent the southernmost extension of the Old Red Sandstone magnafacies (Dineley 1966), although Selwood (1990 p. 204-205) has recently suggested that:

"If it is accepted that the flysch in the South Devon and Trevone Basins is a product of a northward advancing deformation front, then there is little evidence for newly deformed Upper Palaeozoic sediments belonging to such a basin. Rather an intermittently uplifted basement land mass or land masses is indicated. In this setting the Middle Devonian carbonates might have developed as fringing reefs to such land masses; and the Lower Devonian continental and shallow marine clastics might have had a southern provenance".

The base of the Dartmouth Group is not exposed and an age range for the Group implies deposition extending from the late Lochkovian - early Pragian (Dineley 1986 - date based on traquaraspid fish remains; Davis 1990 - date based on palynomorphs) into the Pragian, and possibly into the lowest Emsian (Blicek 1982). The Dartmouth Group records deposition of muddy fluvial deposits in a distal terminal fan setting

within a very large basin which was subsiding rapidly along syn-sedimentary fault lines. During periods of particularly rapid subsidence shallow perennial lakes developed. The intercalation of several volcanic layers suggests that the rapid subsidence of the basin was driven by high-heat flow generated by crustal thinning (Smith & Humphreys 1991).

During the widespread late Pragian transgression (House 1975, 1983, Johnson *et al.* 1985) the continental Dartmouth Group was replaced by the shallow marine 'Meadfoot facies' (Evans 1980) of the Meadfoot Group deposited on a marine shelf (Richter 1967, Harwood 1976, Pound 1983 - enclosure 12, Holder & Leveridge 1986a). The base of the Meadfoot Group is diachronous, being late Pragian in age in eastern Cornwall (based on brachiopods recorded from a marine incursion near the top of the Dartmouth Group - Evans 1981) and the base of the Meadfoot Group (Evans 1980), and mid Pragian to Emsian in age in south Devon (Evans 1980). To the east of Plymouth Sound the 'Meadfoot facies' is locally referred to as the Bovisand Formation (Harwood 1976). The lower levels of the Bovisand Formation were deposited on an outer shelf below storm wave-base, the occurrence of pyrite throughout the sequence indicating that anoxic conditions developed below the sediment water interface; the upper Bovisand Formation was deposited within storm wave-base (Humphreys & Smith 1989).

The onset of the overlying 'Staddon facies' (Evans 1980) of the Meadfoot Group was also diachronous based on dates for the succession obtained from brachiopod assemblages (Evans 1980) and the Staddon Grits thicken westwards into Cornwall (Dineley 1961). Evans noted that only a sparse fauna was obtained from the 'Staddon facies', therefore hindering precise correlations, and more importantly, observed that the 'Staddon facies' and 'Meadfoot facies' are lithologically similar in the critical Looe and Fowey exposures. Thus, the paucity of the fauna and the difficulty of differentiating the 'Staddon facies' from thick sandstones in the 'Meadfoot facies' makes the precise positioning of the Meadfoot-Staddon boundary hazardous, and the 'Staddon facies' may be restricted to the late Emsian (Dr. K. M. Evans, pers. comm.) Dean (1989) recovered a palynomorph assemblage from the Bovisand Formation and lower Staddon Grits from the east side of Plymouth Sound and reported a probable late Emsian (possibly younger) miospore assemblage. Knight (1990a) re-examined this assemblage and considered the assemblage to be ". . . slightly older than those of the Lynton Formation" (p.359).

The 'Staddon facies' in the Plymouth area represents the development of a fluvial-dominated, low wave-energy delta and is thought to represent a regressive sediment pulse supplied by a local fault-block to the north (Pound 1983 - enclosure 12). The supply of sediment to the delta ceased in late Emsian times and uppermost Staddon facies sediments were locally reworked by waves into a series of bars. The succeeding Jennycliff Slates record an upwards transition from a muddy nearshore facies (restricted bay, lagoonal and estuarine environments), developed as the sea transgressed and drowned the fluvial upper 'Staddon facies' (Humphreys & Smith 1989), to deposition on a more open storm-dominated shelf (Pound *op. cit.*)

Humphreys and Smith utilised phosphatic horizons in the Plymouth area for the purpose of stratigraphic correlation and assigned the Bovisand Formation to T-R Cycle 1b of Johnson *et al.* (1985) i.e. late Pragian, whilst the lower Jennycliff Slates were assigned to T-R Cycle 1c i.e. mid Late Emsian. However, these assignments conflict with published data for the region and Humphreys and Smith recognised that "... palynological work currently being undertaken on the south Devon sequence suggests that a downward shift of the stage boundaries is required (A. Dean, pers. comm.)" (p.123). When Dean's subsequently published dates (1989) were applied, the phosphatic horizons in the Bovisand Formation equate to T-R Cycle 1c of Johnson *et al.* (1985) i.e. mid late Emsian, whilst the lower Jennycliff Slates equate to T-R Cycle 1d i.e. late Early Eifelian. Thus, the Staddon Grits in the Plymouth area are equivalent in age to the Lynton Formation. Further east at South Bay (south of Brixham) Smythe (1973) reported early Eifelian icriodid conodonts in transitional beds from the Staddon Grit into overlying crinoidal limestones. In the Kingsteignton area a thick pile of volcanic rocks generated an intrabasinal rise in the 'South Devon Basin' during the early Mid Devonian (Selwood 1990).

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Between the 'South Devon Basin' and the 'Trevone Basin', in the Liskeard area, Burton and Tanner (1986) demonstrated the existence of a persistent shelf area from Emsian times until the onset of basinal sedimentation in the Late Devonian.

'Trevone Basin'

In north Cornwall Beese (1982) assigned a late Emsian age to the Bedruthan Slates which are found at the base of a thick succession of basinal deposits developed during Mid Devonian progradation and retreat of turbidite fans. Holder and Leveridge (1986a) suggested that these deposits resulted from onlap of clastic sediments from the 'Gramscatho Basin' which supplied distal turbidites to the Trevone Basin which must, therefore, have already been developing in late Emsian times.

1.6.2.2.3 'Gramscatho Basin'

The 1.5km Roseland Breccia Formation (Frasnian or younger) contains olistoliths of Ordovician, Silurian and Early Devonian age shelf sediments, together with igneous and metamorphic rocks (Holder & Leveridge 1986a). This putative shelf sequence was cut by Early Devonian extensional basin development. Although late Early Devonian basin-fill deposits are suspected there is no positive evidence that the basin-fill deposits range lower than the middle Eifelian. The end of extensional basin development was marked by the onlap of the basinal deposits of the Porthowan Formation during the Mid Devonian over Dartmouth Group and Meadfoot Group deposits to the north (Bluck *et al.* 1989).

1.6.3 Plate Tectonics and the Palaeo-stress Field in the Exmoor Region During the Late Emsian - Early Eifelian

The purpose of this section is to provide a brief overview of the tectonic climate that prevailed during this period in order to place the development of the Lynton Formation sequence and 'Exmoor Basin' in context. In order to ascertain the nature of the late Emsian - early Eifelian stress field in the Exmoor region it is first necessary to examine the plate tectonic setting of SW England at that time.

The overall tectonic setting of SW England during the Devonian and Carboniferous has been the subject of intense and continuing debate. To the north, the closure of the Lower Palaeozoic Iapetus Ocean (McKerrow & Ziegler 1972) and the collision of North America (Laurentia), Acadia, Great Britain and northern Europe (Baltica) created an orogenic belt extending from New York State to Spitsbergen referred to as the 'Old Red Sandstone Continent' (Laurussia). Palaeomagnetic data reviewed by Torsvik *et al.* (1990) showed that by Early Devonian times the apparent polar wandering paths for northern and southern Britain converged,

implying effective closure of the Iapetus ocean in Britain. The line of the resulting Iapetus suture passes through the Solway Firth and thence through the paratectonic Caledonides of Ireland (Phillips *et al.* 1976). Baltica and Laurentia most likely collided in earliest Devonian time to form Euramerica, followed by docking of Armorica against Laurentia/Baltica to form the 'Old Red Sandstone Continent' during the Early Devonian.

The closure of Iapetus was diachronous, occurring during the late Llandovery - Wenlock Stages in Greenland and Scandinavia, Wenlock - Pragian Stages in Britain, and final early Devonian closure in Acadia and Laurentia (Scotese *et al.* 1985). Palaeomagnetic data for Laurentia and Gondwana suggests that the putative 'Rheic Ocean' (McKerrow & Ziegler 1972) separating Laurussia from northern margin of Gondwana was relatively narrow during the early Devonian, and may have been closed by the late Devonian. This narrowness is consistent with biogeographic data presented by Barrett (1984).

To the south, however, a far more complex picture is presented. Broadly, four models have been proposed to explain the tectonic evolution of the Rhenohercynian Zone of the Variscides. Each of the models is briefly outlined below.

(i) **Palaeozoic 'Rheic Ocean'** Various authors have proposed a large mid-European ocean, termed the 'Rheic Ocean' (e.g. Burrett 1972, McKerrow & Ziegler 1972, Johnson 1973 & Burne 1973), based on palaeomagnetic evidence suggesting a large separation between Pangea and Gondwanaland during the mid-Palaeozoic (Smith *et al.* 1973 and Tarling 1979) and a faunal province distribution suggesting a wide ocean (Burrett 1972, Cocks & Fortey 1982). It was suggested that the ocean opened during the Lower Palaeozoic and closed during the Upper Palaeozoic.

(ii) **Narrow, Rift Basin** McKerrow and Ziegler (1972), Dewey and Kidd (1974) and Sawkins and Burke (1980) suggested that a major ocean never became established in mid-Europe. Instead, a small rift basin, possibly of 'Red Sea' type, opened and closed during the Devonian and early Carboniferous. This interpretation is consistent with a palaeomagnetic reconstruction offered by Morel and Irving (1978) and biogeographic data presented by Barratt (1985).

(iii) **A Narrow, Partly Oceanized, Marginal Back Arc Basin Reading** (1973) was the first author to suggest that the Variscan geology of SW England can be explained in a 'Japan Sea' type marginal back arc basin context, a view expanded upon by several subsequent authors (e.g. Bromley 1976, Leeder 1976, 1982b, Anderton *et al.* 1979, Dewey 1982, Floyd 1982 and Holder & Leveridge 1986a), possibly on the margin of a 'Proto-Tethys Ocean' (Floyd 1972, Nicolas 1972). This model was proposed to explain an extensional phase (cf. Matthews 1977) of deformation during the Devonian in SW England, the extension being a response to northward subduction of a 'Massif Central' ocean with the Bretonic arc being situated above the subducting oceanic crust (note, however, that Holder & Leveridge viewed the subduction as being southward directed). Dewey (1982) and Leeder (1982b) stressed that the subduction was oblique.

(iv) **Intracontinental Strike-slip Régime** Badham (1976, 1982), Arthaud and Matte (1977), Sanderson (1984), Barnes and Andrews (1986), Franke (1989) rejected an active subducting plate margin setting to propose an alternative model of intracontinental strike-slip deformation, within a zone of dextral shear, for south-west England. Riding (1974), Badham and Halls (1975), Badham (1982), and Franke (1989) suggested that the major strike-slip faults bounded microplates. Franke described a number of terranes from western and central Europe that originated as rifted fragments off northern Africa which were accreted on to the European margin during NW-directed Variscan convergence. Late Palaeozoic extensional basins generated during a period of crustal thinning were inverted during Variscan convergence and foreland basins developed in front of the advancing deformation front.

It should be stressed that the above four models are not mutually exclusive, as may be expected for models based upon the same geological evidence! For example, Barnes and Andrews' (1986) model for an intracontinental strike-slip régime shares many features with the narrow, partly oceanized marginal back arc basin model (iii). In particular, Barnes and Andrews envisaged closure of the 'Massif Central Ocean' at the end of the Devonian accompanied by the formation of the Ibero-Armorican Arc. Following crustal convergence, northerly-directed tectonic transport stacked up thrusts and produced foreland basins which migrated northwards throughout the Carboniferous. The Barnes and Andrews model, however, differs from the marginal back arc basin model in attributing Devonian extension to an intracontinental dextral strike-slip régime rather than the back arc extension proposed in the latter model.

Although a wide 'Rheic Ocean' [model (i)] is now generally recognised as untenable, the Variscan geosyncline now being interpreted as developing as a narrow arm off the 'Proto-Tethys Ocean' (Tsien 1989), the presence and nature of a subduction zone south of Cornwall is still a matter of debate (Bluck *et al.* 1989). There is, however, increasing evidence that strike-slip motions played a significant part in basin development. Badham (1982), in arguing for the importance of strike-slip movements presented four points which must be taken into account in any model to explain the plate tectonic evolution of the Rhenohercynian zone:

- a) The sedimentary, tectonic, igneous and metamorphic history varies in both timing and type of process along the orogen.
- b) There is a lack of laterally extensive evidence of initial rifting and later subduction along the length of the orogen.
- c) There is no evidence for a sustained phase of compression. Instead, evidence points to episodic and localised extension and compression.
- d) Ophiolites, although present, have differing timing and emplacement histories.

In summary, basins developed in SW England during the Devonian through the interplay of crustal **extension** with strike-slip movements ('transtension' *sensu* Harland 1971). It is now widely accepted that during the Devonian extensional strike-slip shear across SW Britain was dextral (Barnes and Andrews 1986, Holdsworth 1989, Bluck *et al.* 1989, Selwood 1990); Devonian dextral shear has been attributed to late Caledonide oblique movements along the Iapetus Suture (Tunbridge 1986). By the time the Variscan deformation front had reached north Devon in the Late Carboniferous a switch to dextral shear in a **contractional** régime ('transpression' *sensu* Harland *op. cit.*) developed in response to NW-directed Variscan compression deflected around the London - Brabant Massif (Gayer & Nemcok 1994) and was accompanied by large-scale dextral transcurrent movements occurring along the Bristol Channel - Bray Fault system (Holder & Leveridge 1986b). Local evidence for dextral Variscan transpression in north Devon has been described by Andrews (1993) from the Pilton Shales in Croyde Bay.

1.6.4 Structural Setting of the Lynton Formation

The tectonic history of the major structural elements within the Lynton Formation may be conveniently ascribed to three separate phases of deformation: syn-sedimentary tectonics, the Variscan orogeny and Tertiary reactivation. These are described in turn below.

1.6.4.1 Syn-sedimentary Deformation (D0)

Tunbridge & Whittaker (in: Goldring *et al.* 1978) recorded that Lynton Formation rocks occurring immediately adjacent to the Lynmouth - East Lyn Fault at Ninney Well Bay (enclosure 1) contained “. . . intraclasts in the slates (*that*) have been derived from stratigraphically lower Lynton Beds” and that: “Inland, similar occurrences have been recorded, only in close proximity to the fault, and may indicate intra-Devonian movement along the same line” (p.9). Similarly, Whittaker and Edmonds (in: Edmonds *et al.* 1985) in describing the rocks at Ninney Well Bay noted that: “The Lynmouth East - Lyn Fault shows features . . . consistent with a fracture active locally in Lower Devonian times. Clasts of older Lynton Slates appear in the younger Lynton Slates, the source area lying close to the north” (p.63). The conglomerate was described as comprising “. . . clasts of dark greenish grey silty and sandy shale and siltstone . . . in dark grey slates . . . The clasts are up to 0.3 m long and 0.15 m wide, generally non-angular, and lithologically similar to rocks lower in the Lynton Slates sequence” and that: “The clasts of the coastal exposures appear to die out laterally away from the fault within about 5m, and yet persist vertically through an unknown thickness of not very intensely disturbed slates” (p.27). Whittaker and Edmonds also made reference to unpublished notes made by Prof. S. Simpson which recorded occurrences of the conglomerate associated with the Lynmouth - East Lyn Fault at Myrtleberry Cleave (7417 4901) and north of the old limestone quarry at Watersmeet (7473 4872). The intraformational conglomerate described by Whittaker and Edmonds was named the “Lyn Conglomerate” by Tunbridge (1986) who documented the intraformational conglomerate to be the product of submarine debris flows sliding down a fault scarp developed along the Lynmouth - East Lyn Fault, interpreted to be a syn-sedimentary splay from the Bristol Channel fault system in a strike-slip setting.

1.6.4.2 The Variscan Orogeny (D1)

Sanderson and Dearman (1973) divided SW England into twelve structural zones. The Lynton Formation occurs in their ‘zone 1’, which is characterised by upward-facing folds overturned towards the north, which

have sub-horizontal axes and an east-west trend. The fold attitude and cleavage steepens to the south, folds becoming upright south of Barnstaple.

The main structural aspect of the north Devon region is probably the product of a single orogenic phase: the late Carboniferous paroxysmal phase of the Variscan orogeny (Whittaker & Edmonds 1981; Edmonds *et al.* 1985). Structures preserved mark the passage of a zone of deformation which migrated northwards, affecting southern Cornwall and southern Devon during the Late Devonian and Early Carboniferous, the Bude - Bideford area in post mid-Westphalian times (Hecht 1992), reaching the Mendips by the latest Carboniferous (Shackleton *et al.* 1982; Coward & Smallwood 1984). Potassium-argon dating (Dodson & Rex 1971) supports the timing of the northward migration of Variscan deformation from the Late Devonian to the end of the Carboniferous. Anomalous dates from north Devon were attributed to varying amounts of detrital mica. Variscan structures in north Devon formed in response to NW-directed compression in a zone of thin-skinned deformation with small (c. 11° at Wild Pear Beach, Combe Martin) amounts of rotation in the local structural transport direction (Gayer & Nemcok 1994).

The dominant structural feature of north Devon is the 'Exmoor Anticline' ("Lynton Anticline" of Whittaker & Edmonds 1981; Edmonds *et al.* 1985) which verges northwards and has an axis inclined to the south, folding the Early Devonian to Early Carboniferous succession along an ESE-WNW axis with a small plunge to the ESE. The anticline has a gently dipping southern limb and a close to vertical (e.g. SW of Foreland Point, 753 510), sometimes inverted (e.g. north of Great Red Gully, 748 507), northern limb (Whittaker *in*: Whittaker & Edmonds 1981). The Lynton Formation crops out in the core of this feature (enclosure 1). The anticline has been traced offshore into the Bristol Channel (Lloyd *et al.* 1973) where geophysical evidence suggests that the fold axis trends more nearly east-west (Brooks *et al.* 1977). Simpson (1951) noted that slickensides are common in north Devon, trending towards the north. Whittaker (*op. cit.*) observed truncated vertical tension gashes that indicate displacements of up to 3m along slickensided bedding surfaces associated with bedding plane slip related to the folding of the 'Exmoor Anticline'.

Small- and medium-scale folds in north Devon, parasitic on the 'Exmoor Anticline', are overturned towards the north and have an associated southerly-dipping axial planar cleavage (Simpson 1971). This cleavage is common with the axial planar cleavage that fans around the 'Exmoor Anticline' and is steeply upright in the south of Exmoor, shallowing to dip 30° - 40° southwards in the Lynton Formation.

On the coast east of Lynmouth, tight overturned folds have been recognised (Simpson & Kidson 1954, Simpson 1971). These are similar to other small-scale folds overturned towards north, recognised in the Ilfracombe area (Holwill *et al.* 1969) and to the west of Little Hangman (Whittaker in: Whittaker & Edmonds 1981).

Unpublished illite crystallinity determinations were obtained for a series of pilot samples from the Lynton Formation by Dr. T. J. Primmer (ex University of Bristol). They reveal that the Lynton Formation is of a greenschist metamorphic grade (subsequently corroborated by illite crystallinity results published by Kelm & Robinson 1989), with the exception of samples taken in close proximity to the Lynmouth - East Lyn Fault, which indicated a retrogression to anchizone grade. Primmer (pers. comm.) notes that retrogression is common along fault zones, other examples having been recorded from the Port Wrinkle and Port Nadler faults of south Cornwall.

The presence of a major fault intersecting the coast east of Lynmouth was first alluded to by Sedgwick and Murchison (1840), who recognised that the rocks of the Foreland were equivalent to the rocks cropping out in Woody Bay and Combe Martin (now referred to as the Hangman Sandstone Group). In order to account for the sharp juxtaposition of the Lynton Formation and the Hangman Sandstone Group on the coast east of Lynmouth, Sedgwick and Murchison predicted a ". . . downcast fault to the north . . ." (p.664 - see section 1.5). Early maps showing the contact between the Lynton Formation and the Foreland Point sequence, however, showed the contact as a conformable junction, many early authors believing the rocks of Foreland Point to be **older** than the Lynton Formation (Etheridge 1867, Hall 1867, Ussher 1906 - see section 1.5). The first map to show the contact between the Lynton Formation and the rocks of Foreland Point as being a faulted one, was that of Hamling (in: Hamling & Rogers 1910) who concurred with the earlier view of Sedgwick and Murchison (1840) that the Foreland sequence was equivalent in age to the Hangman Sandstone Group sequence which crops out along the coast between Woody Bay and Combe Martin, an interpretation confirmed by Tunbridge (1978).

The precise position of the Lynmouth-East Lyn Fault was mapped by Whittaker between 1972 and 1975 (B.G.S. Sheet 277 - Ilfracombe - see enclosure 1) who recognised the fault to be a reverse fault dipping southwards at about 75° (see cross-section E-F on B.G.S. Sheet 277, incorrectly printed as 45° in the accompanying sheet notes). This figure was derived by detailed mapping of the fault plane in the deeply

incised East Lyn valley in the proximity of Watersmeet (Dr. A. Whittaker, pers. comm.) The figure of 75° agrees with observations made by the author at Ninney Well Bay (on the coast east of Lynmouth - see enclosure 1) where the extent of the fault plane can be fairly accurately delineated by an examination of the lithologies exposed in outcrops adjacent to the covered fault line (plate 3.1A).

Whittaker (*op. cit.*) estimated the fault to have a throw of about 2 000m near Lynmouth, the throw diminishing somewhat to the east (see notes accompanying B.G.S. Sheet 277), this figure having been derived by simple trigonometric methods (Dr. A. Whittaker, pers. comm.) Whittaker adjudged the progressive 'pinching-out' of the Lynmouth Formation to indicate an eastwards decrease in the throw of the fault.

Ussher (1889) traced the course of the Lynmouth-East Lyn Fault as far east as Brockwell (929 431) near Wooton Courtenay on the western margin of Porlock vale. The known extent of the fault, therefore, is in excess of 20km. By continuing the trend of the Lynmouth - East Lyn Fault from the easternmost point mapped by the B.G.S. (sheet 277 - Ilfracombe 1981) at Malmsmead (792 477), the fault trends 109° for a further 14km. Within the area of sheet 277 the fault trends 101° for 3.5km between the East Lyn valley west of Wilsham (753 485) and Malmsmead; from the East Lyn valley west of Wilsham to the northern flank of Wind Hill, where the trend is offset by a NNE-SSW trending late normal fault (see section 2.3), the fault trend is 114° . A marine survey by Lloyd *et al.* (1973) indicated that offshore, the core of the 'Exmoor Anticline' continues its trend uninterrupted in a WNW direction; presumably the trend of the Lynmouth - East Lyn Fault also continues uninterrupted offshore.

Two sets of minor faults occur in the Lynton area; the most prominent set trend NW-SE, the second set trending SW-NE (Whittaker & Edmonds in: Edmonds *et al.* 1985). These faults were observed at intervals in the order of one-hundred metres along the coast, although the offset of beds indicated that displacements are of only a few metres. Hobson and Sanderson (1983) attribute a Variscan origin to the conjugate fault sets.

1.6.4.3 Tertiary Reactivation (D2)

Shearman (1967) observed that conjugate fault sets of Variscan origin in the Lundy Island and Lee Bay areas displace Tertiary dykes, indicating that reactivation occurred along these features during the Tertiary.

1.6.4.4 Regional Geophysics

Much of south Wales, the Bristol Channel and north Devon is underlain by a seismic basal refractor interpreted as being Precambrian crystalline basement (Brooks *et al.* 1993 - who provided a review of the published data base for the regional seismic network). Brooks *et al.* noted that the basal refractor lies at a depth of c. 8km near the north Devon coast.

The major scale of the Lynmouth - East Lyn Fault is connected to the wider issue of whether Exmoor is underlain by a major thrust. Falcon (in: discussion of Cook & Thirlaway 1952) proposed a thrust to explain the gravity gradient across the Quantock Hills, citing evidence from the Cannington Park inlier where Devonian rocks are apparently thrust over Carboniferous rocks (Ussher 1908). The western extension of this gravity anomaly under Exmoor was discussed by Bott *et al.* (1958) who also favoured a thrust to explain the anomaly. Bott and Scott (1964) reviewed the earlier interpretation and suggested that the anomaly could be explained by either a northward thickening of low density Lower Old Red Sandstone, or concealed Devonian and Carboniferous beneath a thrust.

The extension of the gravity gradient into the southern Bristol Channel region was examined by Brooks and Thompson (1973) who proposed an Early Carboniferous basin, partially overthrust by the Devonian of north Devon, to explain the observed anomaly. They noted that an interpretation could also be arrived at without recourse to a major thrust. Matthews (1975) preferred to explain the anomaly in terms of a thick Early Devonian sequence thinning to the south and terminating in an east-west fault zone running across southern Exmoor.

Whittaker (1975) produced evidence showing that the Rodway Beds, which had previously been interpreted as Devonian, were of Namurian age and therefore overlie the Carboniferous Limestone in the Cannington Park area. This interpretation left Namurian strata closely adjacent to Mid Devonian in the Halseycross Farm

inlier to the SW of Cannington and required an intervening concealed fault with a downthrow of several thousand metres to the north. Whittaker suggested that this may represent the near surface trace of the 'Exmoor thrust'.

In 1977, Brooks *et al.* presented new seismic refraction evidence which indicated that a high velocity layer lay at a depth of only 2.3km under Exmoor. This layer could not be explained by the regional aeromagnetic field which implies a deep magnetic basement in this area. Brooks *et al.* offered two explanations: either north Devon is underlain by shallow non-magnetic basement which deepens rapidly to the north, or, a distinctive Lower Palaeozoic or Early Devonian layer (e.g. quartzite or limestone) is present. The latter interpretation only required a thin layer overlying a much thicker sequence of low velocity rocks which gave rise to the main gravity anomaly. Whittaker (1978) published a short note, commenting on the interpretation of Brooks *et al.*, suggesting that north Devon may be underlain by Lower Palaeozoic quartzite lithologies, citing quartzite clasts obtained from Old Red Sandstone 'exotic' conglomerates of the Bristol Channel region as evidence. Whittaker envisaged deformed slaty rocks of north Devon being separated from 'basement' of Gres Armoricaire - Gres de May type, by a plane of décollement. Matthews (1981) preferred to interpret the high velocity layer as an analogue of the Early Devonian Emsquartzit of Germany.

Mechie and Brooks (1984) presented a Generalised velocity-depth model for north Devon. The model featured a high velocity zone between a depth of 2 and 4.5km separated by a thrust (linked to the Bristol Channel fault Zone) from a thick layer of lower velocity interpreted as 'autochthonous' Palaeozoic strata. It was suggested that the Palaeozoic 'autochthon' was underlain by Precambrian basement at a depth of 8km.

The southward dipping major seismic reflection event at the base of the high-velocity zone underlying Exmoor at a relatively shallow depth was again interpreted by Brooks *et al.* (1988) as a Variscan thrust plane. Beneath the seismic event ". . . discrete packages of parallel or subparallel reflectors . . . (which) can be tied to Carboniferous Limestone sequences outcropping onshore" (p.439) were imaged, and were interpreted as being consistent with the gently folded and little faulted Old Red Sandstone and Carboniferous of south Wales. The sub-Mesozoic basement above the southward dipping event, however, was considered to be consistent with the severely folded and faulted Devonian and Carboniferous sequence observed in north Devon. E-W seismic profiles across northern Exmoor indicated that the southward dipping seismic event lies at a depth of 4 to 5 km below northern Exmoor. Brooks *et al.* suggested that the thrust was reactivated by

Mesozoic extension, cutting up through the Mesozoic succession to be expressed at the surface as the Bristol Channel Fault Zone, a zone of normal faults throwing down to the south. Mesozoic reactivation was confirmed by subsequent detailed seismic lines off the Devon coast presented in Brooks *et al.* (1993).

The shallow high-velocity zone overlying the Variscan thrust in the north Devon region was discussed in detail by Brooks *et al.* (1993). When the effects of Mesozoic refolding were subtracted the high-velocity zone was found to dip towards the SSE, at an average dip of 15 - 20°, from a depth of c. 2km in the Lynton area to 3km near Cannington Park. The layer was interpreted as lying below, and in normal stratigraphic contact with, the Devonian succession in north Devon. The seismic character of the few kilometres thick high velocity layer suggested to Brooks *et al.* a uniform lithology, possibly of schistose and/or gneissose rocks cf. the Start schists and Eddystone gneisses of south Devon.

Moving eastwards, the major Variscan thrust and associated negative Bouguer anomaly extends for at least 150km, passing beneath the Somerton Anticline in Somerset (Donato 1988), where the fault plane dips SSW, and on to the Vale of Pewsey where the fault dips SSE (Kenolty *et al.* 1981, Chadwick *et al.* 1983). Evidence was presented for these two areas to suggest that the major thrust underwent Mesozoic extensional reactivation, in common with the Bristol Channel Fault Zone to the west. The seismic character of the major south dipping reflector corresponds to similar events imaged in the Celtic Sea (BIRPS & ECORS 1986) and the Rhenish Massif of Germany (Meissner *et al.* 1981).

Donato (1988) recorded a zone of slightly steeper gravity gradient, superimposed on the wider gradient, which appeared to correlate with a shallow seismic event sub-parallel to the major Variscan reflection event. Extrapolated westwards, the steeper gradient correlates with the fault at Cannington Park, lending further weight to the hypothesis of a 'Cannington Park thrust'.

It should be noted, however, that the concept of an 'Exmoor - Cannington thrust' (i.e. linking the Lynmouth - East Lyn Fault with the Cannington Park Fault) in accounting for the observed Bouguer anomaly is wholly inadequate. As was discussed above, the regional Bouguer anomaly is a product of the deeper southerly dipping major Variscan thrust which reaches the surface at the Bristol Channel Fault Zone. Furthermore, the lateral continuity of the surface trace implied by the term 'Exmoor-Cannington Thrust' is only apparent and ignores the major dextral strike-slip NW-SE trending Cothelstone Fault which would be expected to offset

the fault trend. The onshore trace of the Lynmouth - East Lyn Fault is 13 to 17 km south of the Bristol Channel Fault Zone in the Lynton area and increases eastwards. The 'Cannington Park Thrust' trace, however, is only 7km south of the Bristol Channel Fault Zone. Furthermore, Whittaker (1975) notes that the E-W to WNW-ESE fault trend at Cannington Park must swing to the NW before reaching the Quantock Hills to the west, suggesting a curved fault trend. It is concluded that the 'Cannington Park Thrust' lies sub-parallel to, and north of, the Lynmouth - East Lyn Fault, and would display a similar (putative) westwards increasing closing angle with the Bristol Channel Fault Zone.

In conclusion, the idea of a plane of décollement beneath Exmoor is compelling. N.B. The term 'décollement' is preferred to 'thrust' as there is no evidence that *older* rocks lie above *younger* rocks on a regional basis. A regional low-lying cleavage dipping towards the south (also occurring in the Brendon area - Webby 1965a), and tight, north-facing overturned folds imply significant tectonic shortening and movements towards the north. Indeed, Shackleton *et al.* (1982) recorded high flattening strains by a pressure-solution cleavage in the Ilfracombe area, indicating a shear zone dipping towards the south at approximately 45°, possibly representing shortening by as much as 20km locally. In their model to explain the tectonic evolution of SW England they extended their basal plane of décollement beneath Exmoor.

Based on the above interpretation the Lynmouth - East Lyn Fault would have been reactivated as a thrust during the Variscan and probably had a listric geometry¹, soling-out onto the major Variscan thrust/density inversion interface imaged by Brooks *et al.* (1988). This fits the model of Gayer and Jones (1989) which shows a period of extension in SW England and south Wales during the Devonian and Early Carboniferous prior to inversion during post-Westphalian Variscan contraction. In this model the Bristol Channel Fault Zone is interpreted as a major Variscan thrust forming the floor thrust of an extensive thrust sheet.

1.7 MEGA-FACIES AND THICKNESS OF THE LYNTON FORMATION

Published estimates of the thickness of the Lynton Formation appear to have been based upon width of outcrop considerations. Simpson (1959, p.68) suggested a: "Thickness probably greater than 1500 ft.", although a revised estimate of "about 1,300 feet" was proposed in a subsequent paper (Simpson 1964, p.122). More recently, Goldring *et al.* (1978) suggested a figure of c. 300 - 400m, figures consistent with

¹ The pre-Variscan fault geometry would probably have been planar or convex-up - see section 2.2.4.

those published by Whittaker and Edmonds (1981). The recognition of outliers of the Hangman Sandstone Group during the present study (see section 1.8.3), on ground previously mapped as Lynton Formation, however, invalidates the above estimates based on an outcrop width considerations. A revised estimate of the thickness of the Lynton Formation is presented below, and is based upon the measurement of sections, and the intervening strata between sections, along with the lateral correlation of sections on the basis of facies and lithology.

Enclosure 2 is an attempt to show the relative positions of locations described, and sections logged, during the course of the present study. In the absence of distinctive chronostratigraphic marker horizons or a biostratigraphical subdivision of the Lynton Formation, the correlation of sections has been based upon facies and lithological characteristics. Thus, it should be borne in mind that the majority of correlations shown in enclosure 2 are diachronous; the vertical thickness axis should not be thought of as representing an absolute chronostratigraphic scale. The horizontal axis of enclosure 2 attempts to relate sections in terms of their relative position in a section drawn normal to palaeoslope (see section 2.2.1.2.2 for the method used to determine the regional palaeoslope trend). In practice this was achieved by considering the distance of individual locations/sections from the line of the Lynmouth - East Lyn Fault and projecting their position onto a two-dimensional plane. This faultline was a major syn-sedimentary lineament which had a profound influence on the position of facies during the deposition of the Lynton Formation (chapters 2 and 8). *It should be stressed that enclosure 2 may well contain inaccuracies due to the inherent problems to be expected when attempting to correlate sequences which lack distinctive marker horizons and occur in a structurally complex region. Nevertheless, the scheme proved to give geologically reasonable results in practice and it is believed to show relationships which are broadly correct.*

As noted in section 1.4 a series of 'mega-facies', based on the grouping of a common broad set of characteristics recognised at particular levels within the Lynton Formation, have been defined for the purpose of sub-division into chapters. None of the divisions between mega-facies is sharp i.e. a gradual transition is observed between the set of characteristics that defines one mega-facies into the set of characteristics that define the succeeding mega-facies. The following paragraphs describe, in chronological order, these mega-facies. For detailed descriptions of localities and sections, their grid references, ease of access etc. the reader is referred to the introductory section of each mega-facies chapter.

1.7.1 The 'Basal Mega-facies' of the Exposed Lynton Formation

The lowest exposed beds of the Lynton Formation crop out at Yellow Stone and east Lynmouth Beach at Black Rocks (north of Point Perilous) and are characterised by numerous decimetre scale intraformational slides and slump scars. The first sequence accessible for the purpose of logging, however, occurs 400m to the SW of Yellow Stone at Wringcliff Bay (see chapter 3) at a level above the basal zone of intraformational sliding and slumping. The Wringcliff Bay section occurs c. 8m above the oldest strata exposed at Yellow Stone, although the precise position of the section is unclear as the two sequences are separated by late normal faulting, across which bedding cannot be matched. The transition into the succeeding 'lower-middle mega-facies' is visible at Ruddy Ball, some 40m above the base of the exposed Lynton Formation.

1.7.2 The 'Lower-middle Mega-facies'

The basal zone of the 'lower-middle mega-facies' is characterised by a series of laterally discontinuous sandstone-bodies. The 'Lee Stone facies association' (defined in section 1.8.3.1.1), is a distinctive coarsening-upwards sequence which has been logged at Lee Stone and Lynmouth Beach and was also observed at Ruddy Ball. The 'Watersmeet lithotype' (defined in section 1.8.3.1.2) occurs throughout the Lynton Formation (see section 4.1.1) but attains its maximum thickness at Watersmeet at this level in the sequence; thinner developments at this level occur in the lower parts of the A39 road and Lee Stone sections.

At Duty Point (6947 4968) an unbroken succession, some 48m in length, is present in the vertical cliff face above the 'Lee Stone facies association' and contains several metre-scale units of sandstone. Representative sections of this sequence were logged in the outcrop above Lee Stone and outcrops immediately above the cliff top (described in chapter 4), the only accessible exposures of this part of the succession. The top of the Duty Point succession was correlated with the base of the West Crock Point succession on the basis of facies and lithology. This correlation was corroborated by trigonometric calculations.

1.7.3 The 'Upper-middle Mega-facies'

Along the western side of Crock Point a 124m thick succession, the longest continuous section logged within the Lynton Formation, of monotonous heterolithic bedding occurs. The sequence is characterised by the absence of sandstone-body developments of appreciable thickness, the frequent occurrence of the trace fossil

Chondrites and a 9m thick development of largely unbioturbated mudstones containing thin beds of graded, un-rippled sandstone at 4/5th height in the succession.

1.7.4 The Transition to the Hangman Sandstone Group Adjacent to the Lynmouth - East Lyn Fault: The 'Upper Proximal Mega-facies'

As noted above, outliers of the Hangman Sandstone Group occur at Hollerday Hill, The Tors west of Wind Hill and Summerhouse Hill. Sequences through the boundary are exposed on the WNW flank of Hollerday Hill above the Valley of Rocks, the north spur off the eastern end of South Cleave, The Tors on the western flank of Wind Hill, the crag above the Glen Lyn Gorge and Oxen Tor. In the field the junction between the Lynton Formation and the Hangman Sandstone Group was taken at the point where thick sandstones become the dominant lithology (see section 1.8.3).

Oblique aerial photographs of the inaccessible sea-cliffs below the Valley of Rocks indicate that a complete succession, offset locally by minor late normal faulting, may be traced from Yellow Stone (base of the exposed Lynton Formation) to the Lynton Formation - Hangman Sandstone Group junction exposed on the WNW flank of Hollerday Hill. Measurements taken from the oblique aerial photographs indicate a thickness of approximately 200m for the Lynton Formation in the vicinity of the Valley of Rocks i.e. immediately south of the offshore projection of the Lynmouth-East Lyn fault. A similar thickness was calculated for the boundary in the vicinity of Wind Hill. Moving southwards away from the line of the Lynmouth - East Lyn Fault the (trigonometrically) calculated height of the boundary indicates that the Lynton Formation is somewhat thicker than at Wind Hill and Hollerday Hill. Comparison with correlation of the sections noted in the preceding mega-facies suggests the Lynton Formation - Hangman Sandstone Group boundary in the region immediately south of the Lynmouth-East Lyn fault is somewhat younger than the thick mudstone unit developed at 4/5th height in the West Crock Point section.

1.7.5 The Transition to the Hangman Sandstone Group Distal to the Lynmouth - East Lyn Fault: The 'Upper Distal Mega-facies'

Further south of the Lynmouth - East Lyn Fault, and its projected offshore continuation, the Lynton Formation - Hangman Sandstone Group boundary is exposed at Little Burland, Great Burland and Hollow Brook Combe waterfall (see section 1.8.3). The boundary forms a prominent break-of-slope which may be mapped eastwards and westwards (Edmonds *et al.* 1985). The top of the west Crock Point succession is

separated from the boundary mapped to the south by a fault zone which downthrows to the north. The actual fault zone is seen in the cliffs at 686 491 where the coastline of west Crock Point swings east-west into Woody Bay. These faults are shown on B.G.S. Sheet 277 (Ilfracombe) and have a NW-SE trend. Reconnaissance by the author indicated that the faults can be traced inland in a SE direction and are responsible for forming a prominent col at Bonhill Top (689 488) that divides the spur of land that extends NE from Martinhoe Common. The amount of displacement across this fault zone is unknown and it did not prove possible to match units across the fault zone.

The lithological characteristics of the upper 1/5th of the west Crock Point succession resemble a similar sequence observed in the inaccessible (and therefore unloggable) coastal cliffs between Wringapeak (6716 4955) and The Cow and Calf (6643 4966) which occur immediately below the Little Burland succession. The top of the west Crock Point succession and the base of the Little Burland sequence are, therefore, tentatively regarded as being approximately equivalent in age. This correlation is supported by the fact that there is no overlap in lithology/facies between the top of the west Crock Point succession and the base of the Little Burland succession. Field evidence indicated that amalgamated hummocky cross-stratified sandstone-bodies ('Woody Bay facies association' - defined in section 1.8.3.1.3) occur in the Little Burland sequence (chapter 7), whilst only single hummocky cross-stratified beds occur in the top of the West Crock Point succession (chapter 5). The correlation shown on enclosure 2 indicates that the Lynton Formation has a **minimum** thickness of 255m south of this fault zone.

The sections logged WNW of Barbrook and at Woody Bay west occur close to the mapped boundary between the Lynton Formation and the Hangman Sandstone Group. These sections are interpreted as occurring at a similar stratigraphic level to the lower part of the Little Burland sequence below the Lynton Formation - Hangman Sandstone Group boundary.

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In conclusion, the Lynton Formation - Hangman Sandstone Group boundary is believed to be diachronous, the thickness of the **exposed** Lynton Formation varying from 200m immediately south of the Lynmouth-East Lyn Fault, and its projected offshore extension, to a figure in excess of 255m some 2km south of the Lynmouth - East Lyn Fault.

1.8 STRATIGRAPHY OF THE LYNTON FORMATION

1.8.1 Introduction

Devon is, nominally, the type area for the Devonian System of Sedgwick and Murchison (1839), and as such, has a long history of research (see section 1.5). The principle Devonian divisions - epochs and stages - were established by decision of the International Commission of Stratigraphy in May 1984 (Ziegler & Klapper 1985 - see table 1.2).

| Period | Epoch | Stage | Ma | |
|----------|----------------|------------|-----|-----|
| Devonian | Late Devonian | Famennian | 363 | |
| | | Frasnian | 367 | |
| | Mid Devonian | Givetian | 377 | |
| | | Eifelian | 381 | |
| | Early Devonian | Emsian | 386 | |
| | | Pragian | 390 | |
| | | Lochkovian | 396 | |
| | | | | 409 |

Table 1.2 Devonian epoch and stage names along with chronometric scale

After Harland *et al.* (1990)

The B.G.S. map covering the Lynton area (Whittaker & Edmonds 1981), and the accompanying sheet memoir (Edmonds *et al.* 1985), re-designated the strata previously known as the *Lynton Beds* to the new name of *Lynton Slates* in apparent contravention of the conventions set out in Holland *et al.* (1978). The purpose of the following sections is to formally assign the strata previously known as the *Lynton Beds* to the appropriate lithostratigraphic division, applying the conventions established by Holland *et al.*, and to define a series of distinctive facies sequence associations and lithotypes.

1.8.2 Stratigraphical Nomenclature

A full synonymy of the various names given to the Lynton Formation is given in table 1.3. The name most commonly applied to the formation is the *Lynton Beds*, the name which appears in the Devonian part of the

Lexique Stratigraphique International (Simpson 1959) and the Geological Society of London Special Report on Devonian correlation in the British Isles (House *et al.* 1977). Recent mapping undertaken by the British Geological Survey in north Devon has resulted in the term *Lynton Slates* being applied to the formation, a name that first appeared on a map figured by Scrivener and Bennett (1980, p.55) and which subsequently was used on the Survey's revision of Sheet 277 - Ilfracombe (Whittaker & Edmonds 1981) and in the accompanying sheet memoir (Edmonds *et al.* 1985).

| Date | Author | Page | Name |
|-------|------------------------------|----------|------------------------------|
| 1837 | Williams | 95 | Linton slates and limestones |
| 1837b | Sedgwick & Murchison | 560 | Calcareous group of Linton |
| 1838 | Weaver | 589 | Linton calcareous slates |
| 1839 | de la Beche | 47 | Linton grey beds |
| 1839 | Williams | 116 | Calcareous slates of Linton |
| 1840 | Sedgwick & Murchison | 696 | Group No 1 |
| 1841 | Phillips | 183 | Linton Group |
| 1853 | Sharpe | 28 | Linton beds |
| 1867 | Etheridge | 580 | Lynton slates |
| 1867 | Etheridge | 590 | Lynton beds |
| 1867 | Hall | 373 | Lynton zone |
| 1906 | Ussher | 11 | Lynton beds |
| 1910 | Hamling in: Hamling & Rogers | Plate 13 | Lynton slates and Grits |
| 1980 | Scrivener & Bennett | 55 | Lynton Slates |
| 1990b | Knight | 306 | Lynton Formation |

Table 1.3 - The synonymy of the names given to the Lynton Formation.

In reference to the strata previously known as the Lynton Beds, Edmonds *et al.* (1985) designated the name Lynton Slates stating that: "They (*i.e. the Lynton Beds*) are not referred to a lithostratigraphic hierarchy either as a Group or as a Formation, and it has not proved possible to subdivide them" (p.5).

Evans (1983, p.303) noted that: "The name Lynton Slates does not assign the succession to the formal lithostratigraphical hierarchy and they (*sic*) is no improvement over the old name, moreover, this term is particularly misleading as slates form a very small proportion of the total succession which is dominated by finely-laminated sandstones and mudstones". Evans (p.303) concluded that: "Examination of the Lynton Beds in the light of the guide to stratigraphical procedure indicates that probably 'Formation' would be the most appropriate designation. Until such time as Formation is formally proposed, . . . it is proposed here that the name Lynton Beds is retained and that the inappropriate designation Lynton Slates be avoided". The first published account where the name *Lynton Formation* was used was Knight (1990b), although no formal designation/justification was given.

During the present study it has not proven possible, using currently available techniques, to subdivide the formation into mapable units. For this reason it is proposed in the following section that the strata previously referred to as the *Lynton Beds* should be accorded the status of 'Formation' within the lithostratigraphic hierarchy.

1.8.3 Redesignation to Formation Status of the Strata Previously Known as the 'Lynton Beds'

Evans (1983, p.303) noted that: "The guide to stratigraphical procedure (Holland *et al.* 1978) recommends that 'Bed', with a capital initial letter, be restricted to the smallest lithostratigraphical unit recognised in formal classification. In the same work (Holland *et al.* 1978, p.11), however, it is recommended in the interests of stability of nomenclature that names which include 'Beds' (with an initial capital letter) be retained except in the case of a major definitive revision". The detailed sedimentological analysis and recognition of distinctive facies sequence associations and lithotypes herein, of the strata previously referred to as the *Lynton Beds*, is considered to constitute a definitive revision.

With reference to the recommendations of the International Subcommittee on Stratigraphic Classification, set down in Hedberg (1976), the applicability of the term 'Formation' to the strata previously referred to as the *Lynton Beds* is justified by the basic concept of a "formation as the primary unit of lithostratigraphy" (p.32). Although Hedberg notes that: "The thickness of units of formation rank follows no standard" . . . "Practicability of mapping and of delineation on cross sections is an important consideration in the establishment of formations" (p.32).

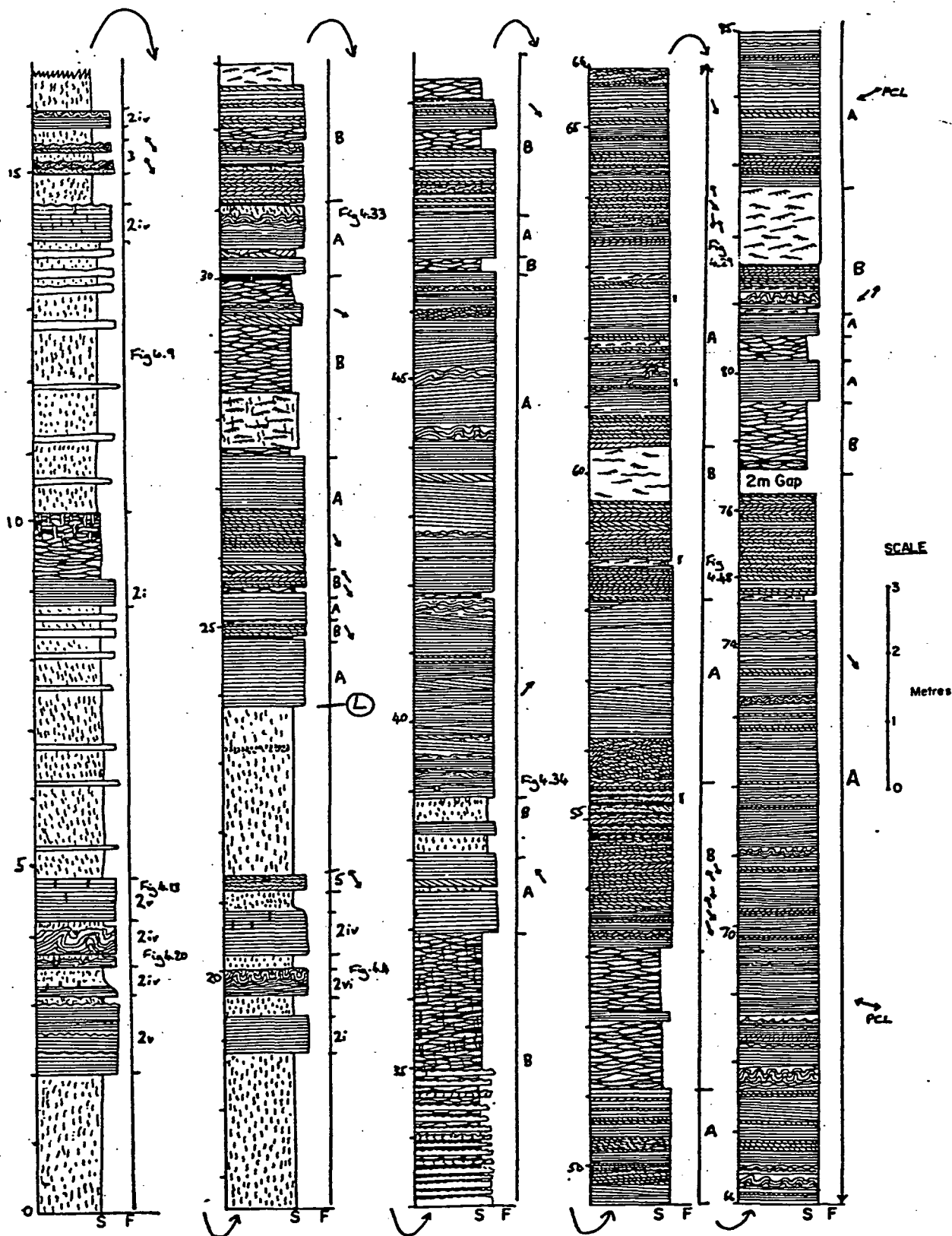
When naming a lithostratigraphic unit, Hedberg (p.40) recommends that: "The name of a lithostratigraphic unit should be formed from the name of an appropriate local geographic feature, combined with the appropriate term for its rank, . . . or with the name of the dominant rock type of which the unit is composed, or with both". The geographic feature Lynton, the main town upon the outcrop of the Formation, is preserved in conformity with recommendation G (chapter 5) of Hedberg (*op. cit.*) which suggests that where there is a change in the lithological designation of a unit or a change in rank, it is not necessary to propose a new geographic term. The justification for referring the strata to formational status has previously been stated. Ideally, a lithological term should also be included. Although the term "Lynton Slates" has recently appeared in print (see above), in accordance with the observation of Evans (1983) that slates form a relatively minor

proportion of the sequence, and in the absence of a suitable alternative lithological term, no lithological term is included in the name.

In conclusion, it is recommended that the strata previously referred to as the *Lynton Beds* be replaced by the term *Lynton Formation*. It is recognised that this recommendation will need to appear in press, with suitable descriptions and definitions, before being formally accepted. Nevertheless, the name *Lynton Formation* has been adopted throughout this thesis.

No previous attempt has been made to formally define, in stratigraphic terms, the strata previously referred to as *Lynton Beds*. In view of the recommended redesignation of the Lynton Beds to 'formation' status it is considered desirable that the Lynton Formation be defined using the procedures for establishing lithostratigraphic units set down in Hedberg (*op. cit.*)

The top of the Lynton Formation was defined by Tunbridge (1978), who described the conformable transition from the Lynton Formation to the Hollowbrook Formation (the basal formation of the Hangman Sandstone Group). Tunbridge defined the base of the Hollowbrook Formation as ". . . the point where sandstone becomes dominant over mudstone. This point coincides approximately with the disappearance of *Chondrites*" (p.97). The type section of the Hollowbrook Formation was rather vaguely defined as lying ". . . on the cliffs west of Woody Bay, where almost complete sequences can be found at Hollowbrook (SS669495) and nearby Great Burland (SS664497)". The only point at which the contact between the Lynton Formation and the Hollowbrook Formation was exactly defined was at 23.85m on the sedimentological log of Great Burland (figure 4.22, reproduced herein as text-figure 1.3).



Hollowbrook Formation
Great Burland

Text-fig. 1.3 Section from the Lynton Formation into the Hollowbrook Formation exposed at Great Burland.

Grid Ref: SS 663 497. The Hollowbrook Formation follows the thick bioturbated grey mudstone at the point "L". This bed lies just above the coastal footpath at this locality. Figure 4.22 of Tunbridge (1978).

The section at Great Burland clearly exposes the transition from the Lynton Formation to the Hollowbrook Formation, the latter formation being approximately 80m thick at the type locality (Tunbridge *op. cit.*) Unfortunately the uppermost beds of the Hollowbrook Formation are not exposed and for this reason the section at Great Burland cannot be considered to be a unit-stratotype of the Hollowbrook Formation. However, the section at Great Burland serves well as a boundary-stratotype for the Lynton Formation - Hangman Sandstone Group contact, although the section was not specifically designated as such by Tunbridge. For this reason, the section at Great Burland (SS 665 495) is designated as the boundary-lectostratotype for the Lynton Formation - Hangman Sandstone Group contact herein. The precise contact occurs in the outcrop immediately beneath the lower footpath at Great Burland (SS 6648 4953) and is shown at 23.85m on the sedimentological log of Great Burland (text-figure 1.3) at the point where a 2.45m thick unit of lenticular bedding is sharply overlain by a 1.2m thick unit of parallel-laminated fine-grade sandstone.

It was noted above that Tunbridge (*op. cit.*) defined the boundary between the Lynton Formation and the Hangman Sandstone Group as the point at which sandstone becomes dominant over mudstone, Tunbridge (p.97) observing that: "This point coincides approximately with the disappearance of *Chondrites*." However, an examination of the section at Great Burland by the author revealed that the sandstone-filled burrows referred to the ichnogenus *Chondrites* by Tunbridge lack the unequal dichotomous branching characteristic of *Chondrites* and are considered by the author to be more correctly referable to the ichnospecies *Palaeophycus tubularis* (Hall). Furthermore, sandstone-filled *P. tubularis* burrows occur in many of the lenticular-bedded units within the Hollowbrook Formation. For these reasons, it is considered herein that the ". . .disappearance of *Chondrites*" should not be used as a criterion for the distinction of the boundary between the Lynton Formation and the Hangman Sandstone Group. However, Tunbridge's criterion of ". . . the point where sandstone becomes dominant over mudstone" (p.97) to define the Lynton Formation - Hangman Sandstone Group boundary, has considerable utility as ". . . the change from predominantly slaty to predominantly sandy lithologies produces a strong feature which is traceable inland" (Edmonds *et al.* 1985, p.6). This characteristic was used to map the boundary between the Lynton Formation and the Hangman Sandstone Group shown on Sheet 277 (Ilfracombe - Whittaker & Edmonds 1981.)

It is desirable at this point that a series of auxiliary reference sections (boundary - hypostratotypes *sensu* Hedberg *op. cit.*) are designated in order to elucidate the nature of the Lynton Formation - Hangman

Sandstone Group boundary. In addition to the sequence at Great Burland, Tunbridge (*op. cit.*) also logged a sequence through the Hollowbrook Formation at Hollowbrook waterfall and cliff (figure 4.23; SS 670 497). Unfortunately, the base of the sequence exposed at Hollowbrook occurs several metres above the boundary between the Lynton Formation and the Hollowbrook Formation (see figure 4.24 of Tunbridge, which shows a correlation between the logged sections at Hollowbrook and Great Burland). For this reason, the sequence at Hollowbrook obviously cannot serve as a boundary-hypostratotype.

During the course of the present study, two sequences were logged through the transition from the Lynton Formation into the Hangman Sandstone Group: one at Oxen Tor, the other at Little Burland. These two sequences may usefully serve as boundary-hypostratotypes for the Lynton Formation - Hangman Sandstone Group boundary.

The sequence exposed at Oxen Tor is shown on Enclosure 9; a sedimentological analysis of this section being given in chapter 6. At Oxen Tor (SS 729 490) 25.2m of the Lynton Formation - Hangman Sandstone Group transition crops out. The precise contact between the Lynton Formation and the Hangman Sandstone Group is taken at 9.85m on the log at the point where lenticular bedding containing thin-bedded sandstones is sharply overlain by a sandstone sequence comprising units of cross-bedding and wave-ripple cross-laminations. The boundary is taken at the point where a 17cm thick unit of strongly bioturbated lenticular bedding is overlain by a 3cm thick parallel-laminated sandstone with a wave-rippled top, the contact being exposed in the outcrop at SS 7288 4899. The boundary is taken at this point as it marks the disappearance of muddy units of appreciable thickness.

A sedimentological log of the sequence at Little Burland is shown on Enclosures 11A and 11B. A sedimentological analysis of the section is given in chapter 7. At Little Burland (SS 663 495) 71.8m of the transition from the Lynton Formation to the Hangman Sandstone Group is exposed. The boundary between Lynton Formation and the Hangman Sandstone Group is taken at 31.2m on the log at the point where hummocky cross-stratified sandstone-bodies set within bioturbated lenticular bedding, containing thin-bedded sandstones, are sharply overlain by a thick sandstone sequence containing only thin, decimetre-scale units of lenticular bedding. The boundary occurs in the exposure at SS 6629 4955 at the point where a 10cm thick unit of bioturbated sandstone is erosively overlain by a 22cm thick unit of bioturbated sandstone. The boundary is taken at this point as it marks the disappearance of muddy units of appreciable thickness.

Transitions from the Lynton Formation to the Hangman Sandstone Group, of the type exposed at Oxen Tor, are also exposed on the WNW flank of Hollerday Hill, north spur off the eastern end of South Cleave (7095 4958), a crag above the Glen Lyn Gorge (7228 4905) and the summit of the WNW spur of Wind Hill (7304 4942). Lichen cover and grass-covered breaks in exposure, however, precluded the construction of sedimentological logs. The presence of these exposures indicates that the outcrop of the Lynton Formation shown on Sheet 277 (Ilfracombe) is incorrect. Outliers of the Hangman Sandstone Group should be shown on Hollerday Hill (714 498), the hill between Valley of Rocks and Dean (707 487), Summer House Hill (725 499) and the summit of the WNW spur of Wind Hill (731 494). Unfortunately, the mapping of the position of these outliers was beyond the scope of the present study and must be left for future workers to determine.

1.8.3.1 Designation of Facies Sequence Associations and Lithotypes Within the Lynton Formation

Although it has not proven possible during the course of the present study to subdivide the Lynton Formation on the basis of mapable units, several distinctive facies sequence associations and lithotypes were recognised. It is not appropriate to assign 'Member' status to these units as, with the possible exception of the 'Watersmeet lithotype', their characteristics in small, isolated inland outcrops would be insufficiently unique to guarantee accurate identification.

1.8.3.1.1 'Lee Stone Facies Association'

The type sequence for the 'Lee Stone facies association' is defined at the western end of Lee Stone (6948 4970) and occurs between 7.35m and 9.63m in the logged sequence (text-figure 4.2) where the facies association is 2.28m in thickness. An auxiliary section through the 'Lee Stone facies association' at the eastern end of Lee Stone (6952 4969) is defined as occurring between 7.16m and 10.58m on the log (text-figure 4.3), where the facies association attains a thickness of 3.42m. The 'Lee Stone facies association' comprises a prominent series of sandstone-dominated heterolithic units which are flaser and cross-bedded, with occasional thin units of lenticular bedding. A full account of the sedimentology of the 'Lee Stone facies association' is given in chapter 4.

A further auxiliary reference section of the 'Lee Stone facies association' is designated at east Lynmouth Beach (7293 4962) and is defined as occurring between 1.46 and 3.26m as shown on the log (text-figure 3.1b) where the facies association is 1.8m thick. A further exposure of the 'Lee Stone facies assoc.' occurs

at Ruddy Ball (7137 5005) where the Member is approximately 2m thick. Unfortunately the exposure was too inaccessible to undertake detailed logging.

1.8.3.1.2 'Watersmeet lithotype'

The type sequence for the 'Watersmeet lithotype' is defined at the river-cliff exposures on the northern bank of the East Lyn River, 250m east of Watersmeet (7464 4865) and is shown in columns B and C of enclosure 7, where it occurs between 1.02m (column B) / 1.08m (column C) and 5.1m (column B) / 5.17m (column C), the lithotype varying laterally in thickness from 4.08m (column B) to 4.09m (column C). The 'Watersmeet lithotype', at the type locality, comprises trough and planer cross-bedded granule-grade conglomeratic cosets separated by thin mudstone-draped erosion surfaces. A full account of the sedimentology of the 'Watersmeet lithotype' is given in chapter 4.

An auxiliary reference section of the 'Watersmeet lithotype' is designated at Lee Stone and occurs between 5.64m and 6.01m on the western logged section (text-figure 4.2) and between 4.52m and 5.41m on the eastern logged section (text-figure 4.3). The lithotype varies in thickness from 0.37m at the western end of Lee Stone, to 0.89m at the eastern end of Lee Stone.

1.8.3.1.3 'Woody Bay Facies Association'

The type sequence of the 'Woody Bay facies association' is defined at the western side of Woody Bay (6871 4905) and is shown on Enclosure 10, where it is defined as occurring between 12.78m and 16.47m on the log, where the Member is 3.69m in thickness. The 'Woody Bay facies association' is a prominent sandstone body comprising amalgamated sets of hummocky cross-stratification and occasional thin silty-mudstone units. A full account of the 'Woody Bay facies association' is given in chapter 7.

An auxiliary reference section of the 'Woody Bay facies association' is designated at the road cutting 0.7km WNW of Barbrook (7077 4787) and is defined as occurring between 20.99m and 23.0m on the log (text-figure 7.1), the facies association being 2.01m in thickness.

1.8.4 Age of the Lynton Formation

Simpson (1964) assigned a ". . . late Emsian or early Eifelian age" (p.121) to the Lynton Formation, based on an examination of their faunal content. Subsequently many authors (House & Sellwood 1966, House *et al.* 1977, Edmonds *et al.* 1975) have placed the Lynton Formation - Hangman Sandstone Group boundary well within the Eifelian, whilst Goldring *et al.* (1978) and Edmonds *et al.* (1985) restricted the Lynton Formation to the upper Emsian, although Edmonds *et al.* (1985) noted that fossil localities sampled for biostratigraphical purposes were well scattered and the possibility of Eifelian strata being present could not be discounted.

Two detailed biostratigraphic studies have been undertaken in recent years which have included samples taken from the Lynton Formation: Evans (1980, 1983) examined the brachiopod fauna whilst Knight (1990a & b) studied the microflora and microfauna.

The stratigraphic ranges of the taxa obtained by Evans indicated a late Emsian age for the Lynton Formation, although Evans noted that the lack of stratigraphically useful taxa in the uppermost horizons of the Lynton Formation meant that ". . . an early Eifelian age for the uppermost horizons cannot be discounted" (1983, p.301).

Knight (1990a) stated that: "Based on miospore data, the Lynton Formation is interpreted as no older than latest Emsian in age, with the uppermost horizons and overlying Hollowbrook Formation of possibly earliest Eifelian age. Conodont faunas recovered from the isolated, inland section at Watersmeet . . . are of early Eifelian aspect" (p. i). The recovery of *Icriodus retrodepressus* from the Watersmeet locality is of critical importance as the inception of this species coincides with the base of the *partitus* Conodont Zone which is now formally recognised as marking the base of the Eifelian Stage (Ziegler & Klapper 1985). The Watersmeet exposure was the stratigraphically oldest locality that yielded a conodont fauna during Knight's study. Consideration of the relative stratigraphic position of the Watersmeet section, some 50m above the base of the exposed Lynton Formation (enclosure 2), indicates that the bulk of the Lynton Formation is Eifelian in age. Knight (1990a, p.348) noted that "the Eifelian age interpreted (based on the conodont faunas) is compatible with the speculative age assessments provided by the palynomorph assemblages also recovered from this sequence" (i.e. the Lynton Formation).

2. THE TECTONIC EVOLUTION OF THE LYNTON FORMATION

2.1 INTRODUCTION

The tectonic history of the Lynton Formation was divided to three separate phases of deformation in section 1.6.4 for the purpose of reviewing material published to date: syn-sedimentary deformation, the Variscan orogeny and Tertiary reactivation. These 3 phases are used below as a basis for describing new information revealed during the course of the present study. N.B. Section 2.2.1 describes a major D0 structure, but due to its close relationship with D1 structures the two deformation phases are discussed in the same section.

2.2 SYN-SEDIMENTARY DEFORMATION (D0)

Several features interrupting the 'normal' Lynton Formation sequence display characteristics indicative of a syn-sedimentary tectonic origin:

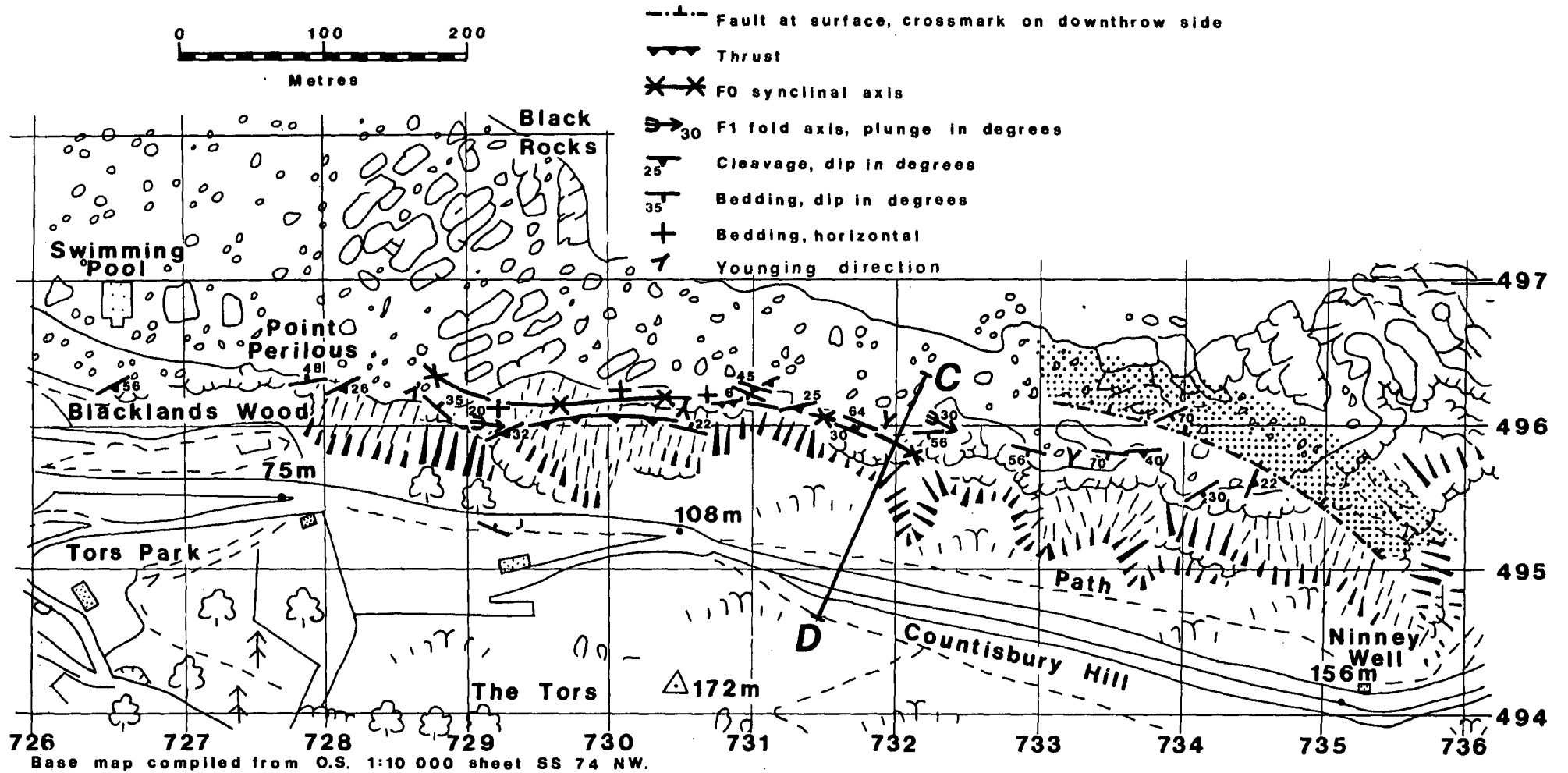
2.2.1 Slide Exposed in the Cliffs East of Lynmouth

Grid Reference: 733 496.

A survey of the 1km long coastal cliff section between Lynmouth and Ninney Well (see Enclosure 1), exposing some of the lowest horizons within the Lynton Formation, has revealed the presence of a major pre-tectonic fold. This fold lies within the northern (overturned) limb of the 'Exmoor Anticline'. The local development of extensive soft sediment deformation features in the rocks exposed between Lynmouth and Ninney Well, and the fact that the fold predates the first phase of Variscan deformation, suggests that the fold should be interpreted as a large-scale syn-sedimentary fold. The general sedimentology of the sequence exposed between Lynmouth and Ninney Well is described separately in chapter 3.

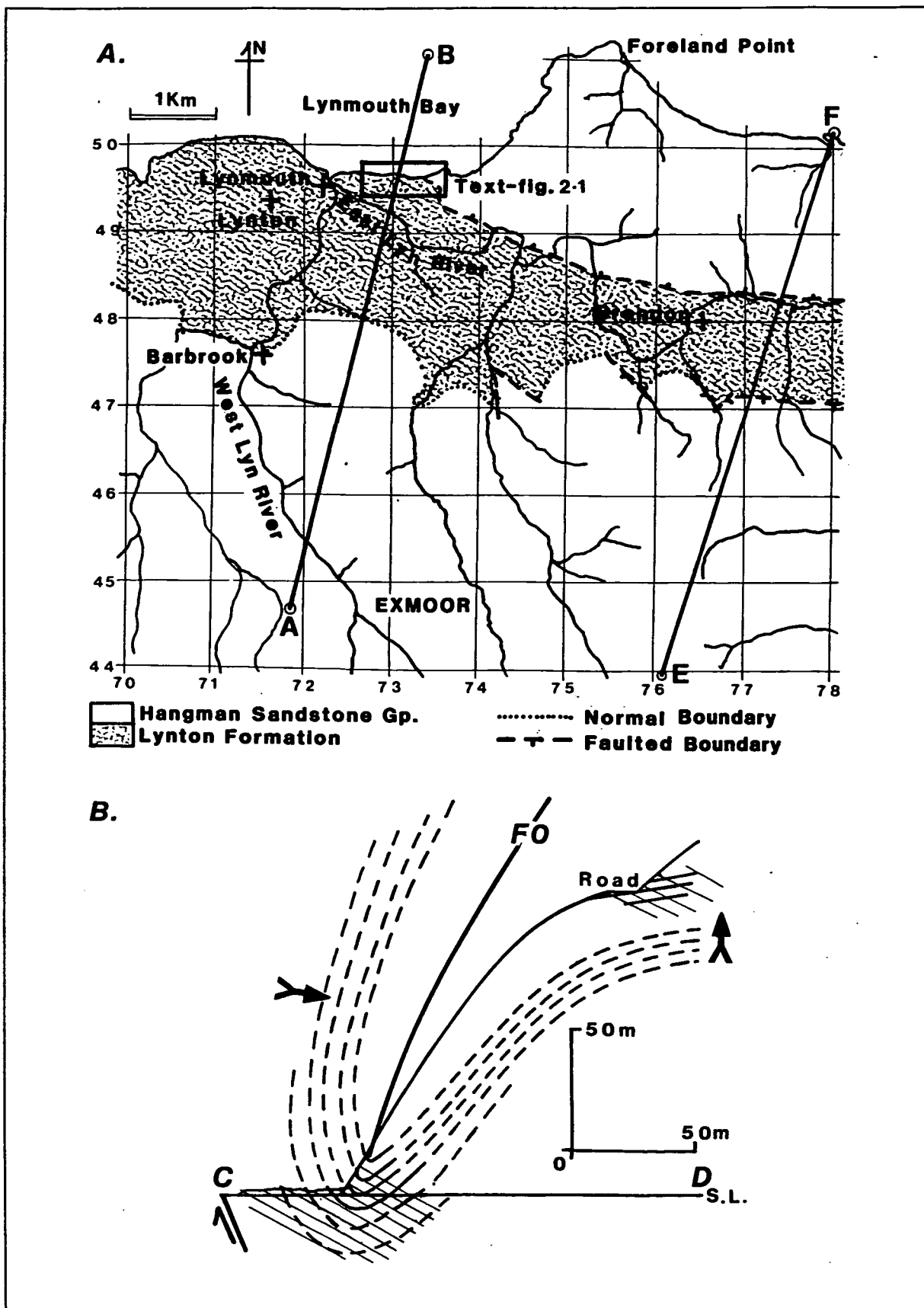
The field mapping of the structural geology of the Lynmouth to Ninney Well section was carried out as a joint study with Dr. T. J. Chapman (Murphy Eastern Oil Company - ex Plymouth Polytechnic).

2



Text-fig. 2.1 Geological structure east of Lynmouth.

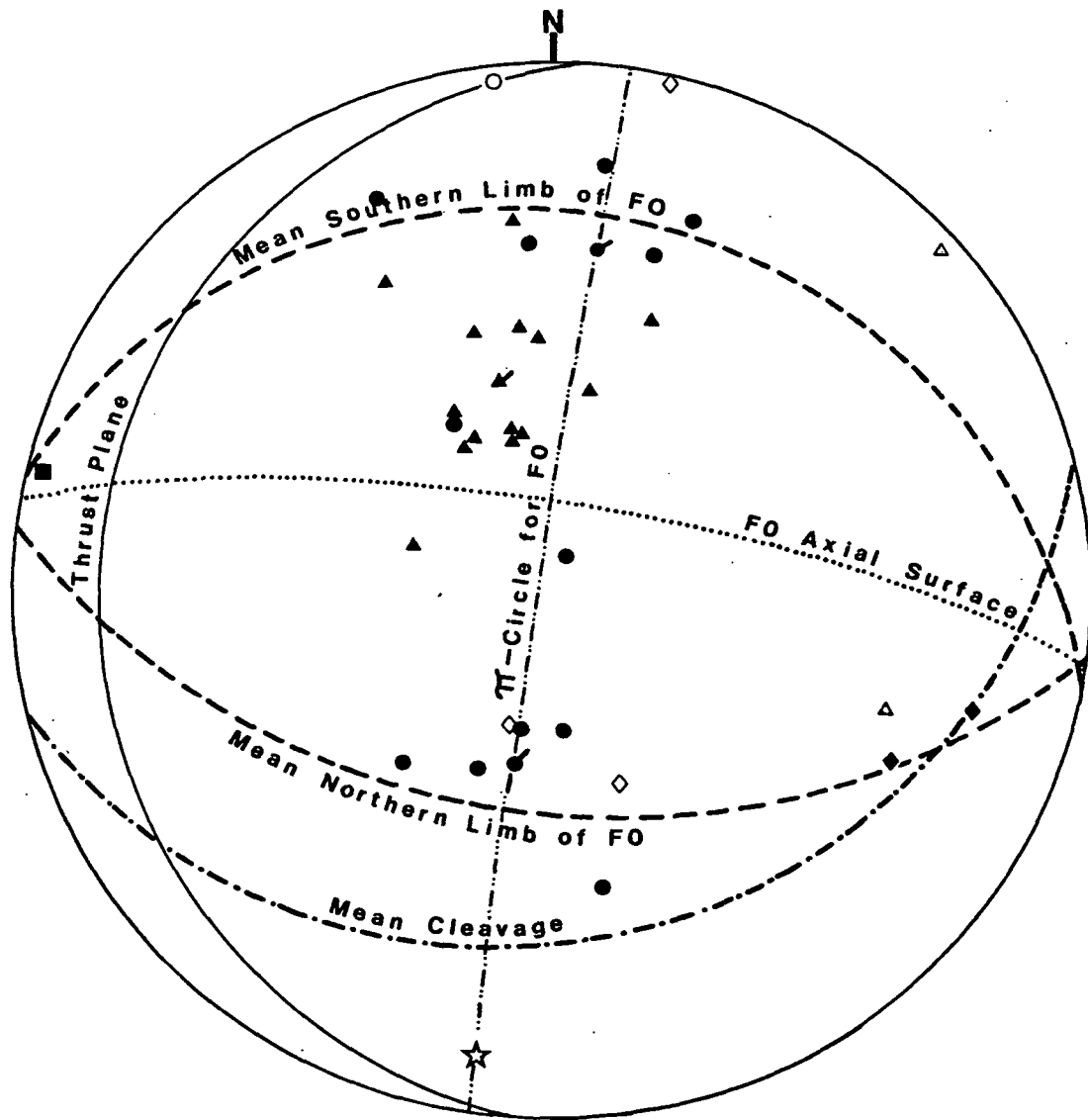
Section C - D is shown on text-figure 2.2B.



Text-Fig 2.2 Location of north Exmoor sections & section through cliffs east of Lynmouth.

A. Location map of north Exmoor, showing the position of the sections shown in text-figure 2.4.

B. Diagrammatic cross-section through the cliffs east of Lynmouth. Position of section is marked on text-figure 2.1. Bedding shown by thick solid (actual) and dashed (interpreted) lines, cleavage by thin lines.



- ▲, ✎ Pole to cleavage, mean.
- , ✎ Pole to bedding, mean.
- F0 fold hinge line, measured.
- ◆ F1 fold hinge line.
- △ Cleavage/bedding intersection lineation.
- ◇ Bedding-plane slip slickenside.
- Thrust-plane slickenside.
- ☆ Pole to F0 axial surface

Text-fig. 2.3 Stereogram of eastern Lynmouth beach structural section.

Text-figure 2.1 shows a structural map of the foreshore and cliffs east of Lynmouth, whilst figure 2.2B shows a true-scale cross-section through the foreshore and cliffs (marked C-D on text-figure 2.1). Text-figure 2.2A shows the location of the map and sections described in this section. In the cliffs on the western half of the map (text-figure 2.1) the bedding dips steeply northwards and youngs **northwards**, whilst near the base of the cliff the bedding becomes flat lying (7292 4962). On the eastern half of the map, the beds dip vertically to steeply northwards, but young **southwards** high in the cliffs, becoming flat lying near the base of the cliffs (plate 2.1A). Further eastwards, the southerly younging beds low in the cliffs dip southwards. In a small cave at grid reference 7317 4960 the southerly dipping beds pass around a fold hinge and become northerly dipping. This well exposed fold axis trends 098°-278° with virtually no plunge.

It is clear that a major synform is present in the cliffs east of Lynmouth which has an east-west trending axis as shown on the map (text-figure 2.1) and profile through the structure (text-figure 2.2B). An equal-area projection of the bedding (text-figure 2.3) demonstrates that the synform has a slight plunge towards 098° and has a maximum interlimb angle of 97°. Way-up evidence in the form of trochoidal wave-ripples, cross-bedding and graded rhythmites indicate that both limbs are the right way up. The northerly limb dips and youngs southwards, whilst the southerly limb dips and youngs northwards. Thus, the synform is a syncline.

Cutting across and transecting the syncline, and apparently unrelated to it, is a well developed slaty pressure solution cleavage which typically dips at 25° to 35° southwards (plate 2.1C), regardless of the two limbs of the synform (text-figure 2.2B, 2.4C - inset). This is the only tectonic/metamorphic foliation in these rocks and is clearly unrelated to the main syncline, as shown in text-figures 2.2B and 2.4C. Any axial planar foliation associated with the main syncline would dip steeply northwards, but in fact none is developed. A well developed pressure solution cleavage is common in north Devon and is associated with small- and medium-scale folds overturned towards the north (e.g. at Combe Martin - Whittaker & Edmonds 1981). This southerly dipping cleavage is axial planar, although fanning, in relation to the main 'Exmoor Anticline' (text-figure 2.4B). The major syncline in the cliffs east of Lynmouth is designated F0 as it is pre-tectonic and probably sedimentary in origin. The 'Exmoor Anticline' and associated small- and medium-scale folds with an axial planar cleavage are designated F1.

Text-figure 2.4C (inset) shows the interesting relationship of the cleavage and F1 tectonic folds with the F0 syncline. On the northern limb, the bedding is steeper than the cleavage, yet it is not inverted. The minor F1

tectonic folds verge south and are **downward** facing (*sensu* Bell 1982). On the southern limb, the cleavage and bedding dip at about the same angle, but in opposite directions. The minor F1 tectonic folds still verge south, but are upward facing. In summary, the minor fold vergence and cleavage does not change across the F0 fold axis, but the facing direction does change.

In conclusion, the F0 syncline is believed to be of syn-sedimentary origin. The cleavage and F1 minor tectonic folds were produced later, on the steep northern limb of the 'Exmoor Anticline', during the primary Variscan deformation in north Devon. Prior to overturning in the north limb of the 'Exmoor Anticline' the F0 syncline would have had a low-lying axial plane and would have verged and faced towards the south. The syncline must have had a corresponding anticline to the north of it; it appears that this anticline has been faulted out by the Lynmouth - East Lyn Fault.

In addition to folding, the sequence east of Lynmouth is also disrupted by thrusting (see text-figure 2.1). The most extensive thrust may be traced horizontally westwards from grid reference 7305 4961 to 7294 4960 (marked on text-figure 2.1) and creates a major step-like feature in the cliff-face. Several small thrusts are visible in the cliff-face figured in plate 2.3A. The upper thrust in this figure cuts through a small fold developed in the northern limb of the F0 syncline. This feature is interpreted as a fold-thrust structure (*sensu* Williams & Chapman 1983), the fold developing in front of a propagating thrust tip line as a ductile response to strain; continued thrust propagation resulted in the thrust cutting through the fold. At grid reference 7308 4962 a westerly dipping bedding-parallel thrust surface can be seen at beach level (plate 2.3B; shown on stereogram - text-figure 2.3). Although the thrust surface dips westwards, a well developed thrust-plane slickenside indicates a north-south sense of movement on the thrust. In summary, the thrusts developed in the sequence east of Lynmouth represent north-south shortening interpreted to have developed during the primary Variscan deformation of north Devon.

2.2.1.1 Origin of F0 Syncline

Apart from the pre-tectonic nature of the F0 syncline already described, a further line of evidence suggests a sedimentary origin for the fold. The sedimentary sequence cropping out in the cliffs east of Lynmouth differs from Lynton Formation sediments developed elsewhere, with the exception of beds of the same age at Yellow Stone, in that it contains extensive developments of soft sediment deformation. Much of the

sequence contains 'disturbed' sediments displaying evidence of lateral flowage, and in cases where heterolithic lithologies are developed, loading (plate 2.4A); discrete décollement surfaces are only rarely recognised (e.g. plate 2.4B). The tops of many of the units that have been disturbed are truncated by penecontemporaneous erosion surfaces, overlain by regularly bedded sediments. In some outcrops, multiple layers of erosionally-truncated disturbed units can be recognised (plate 2.4C).

In conclusion, the wide development of soft sediment deformation structures in the sequence east of Lynmouth suggests slope instability of the seafloor immediately adjacent to the Lynmouth - East Lyn Fault. In conjunction with the demonstrably pre-tectonic nature of the F0 syncline, suggesting that the F0 syncline is syn-sedimentary in origin and probably represents a response to contractional strain in the leading edge of a major slide unit.

2.2.1.1.1 Nomenclature

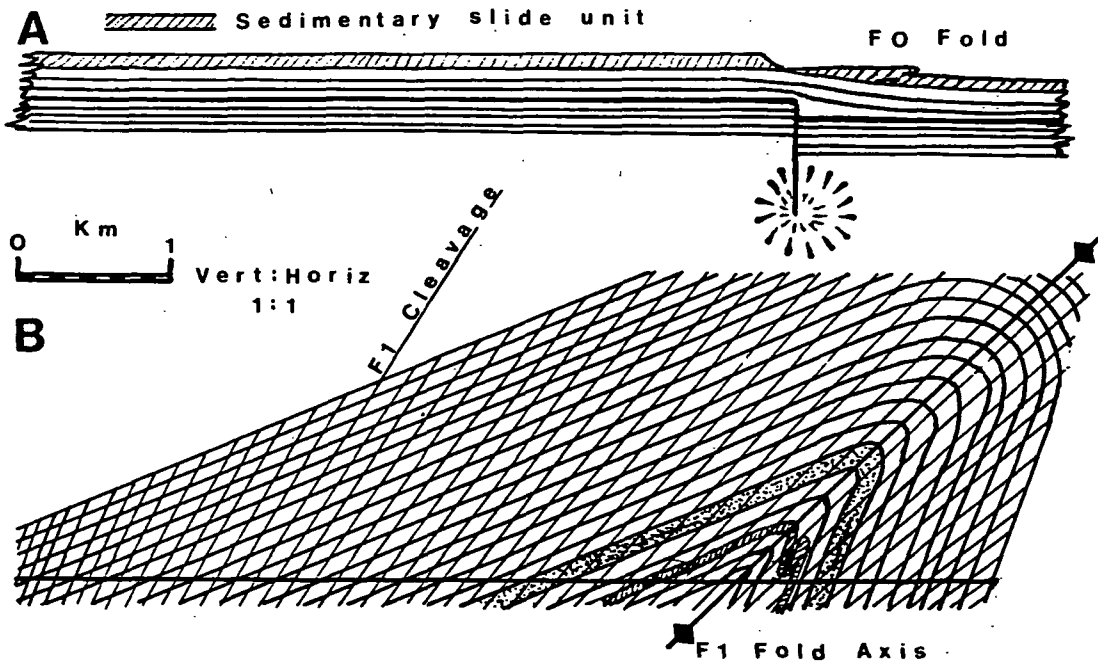
In conformity with the recommendations of Woodcock (1979a), the term slump is reserved for specific types of slide where backward rotation on a more or less horizontal axis parallel to the slope is recognised (see text-figure 2.5; e.g. Ruddy Ball slump - see section 2.2.2). No such backward rotation has been recognised within the sequence involved within the F0 syncline. For this reason, the F0 syncline is assigned to the broader category: *slide unit*. The initial triggering of sliding may have been related to either:

- (i) Seismic shocks.
- (ii) A build up of pore-fluid pressure.
- (iii) Oversteepening of slopes.
- (iv) Cyclic loading by oscillating currents associated with a tsunami.

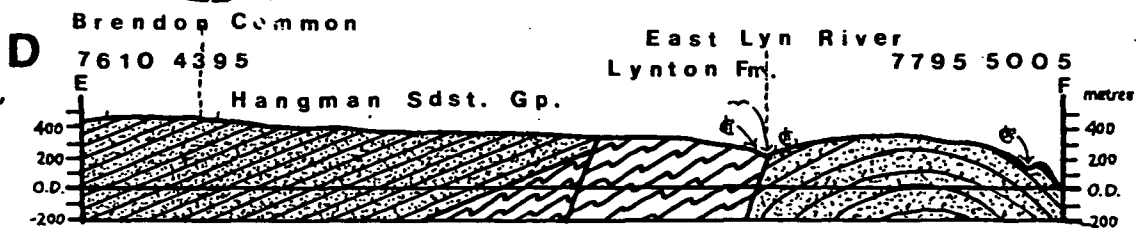
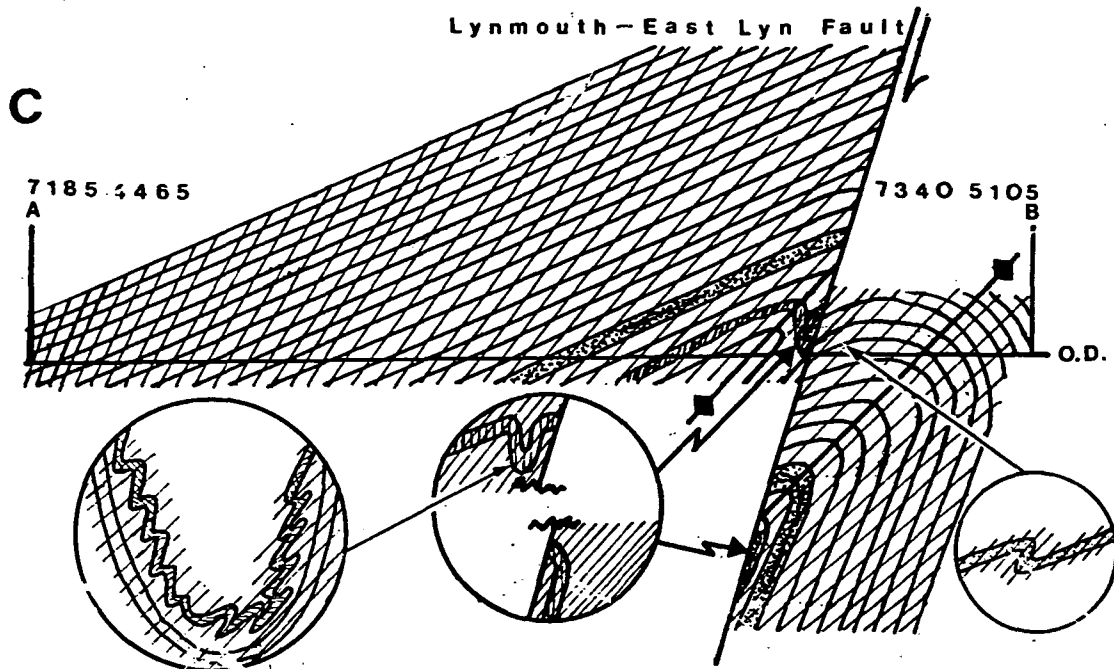
Any of the above processes may have been aided by bioturbation leading to slope destabilisation (see 4.2.2). Although a seismic shock is shown as the trigger in Text-fig. 4.9A, any of the above four mechanisms could have triggered the slide. Text-figure 2.4 shows the conjectured evolution of the sedimentary sequence in north Exmoor.

Text-Fig. 2.4 Structure of the Lynmouth - East Lyn Fault

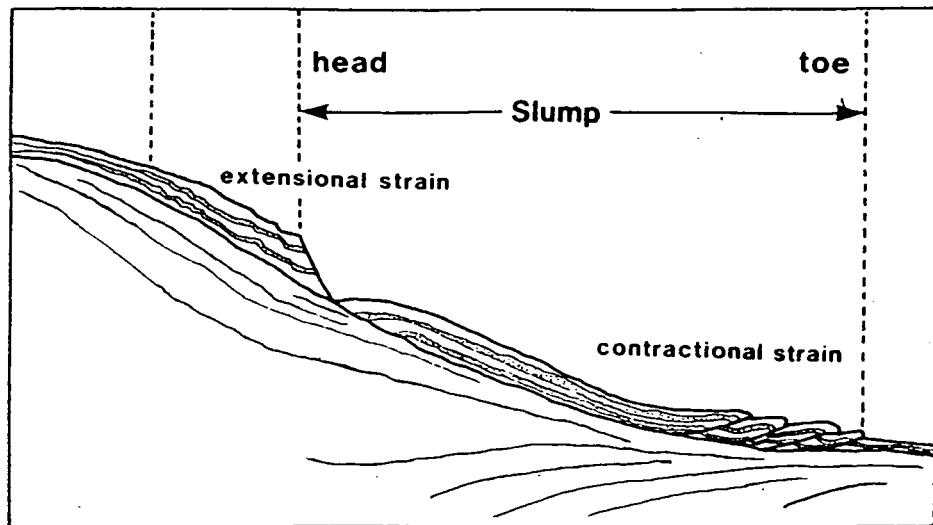
- A. Generation of the F0 syncline in the compressional leading edge of a sedimentary slide unit. The rotational slump scar in the extensional trailing edge of the slide sheet is conjectural. Although a seismic shock on the Lynmouth - East Lyn Fault is shown as the trigger for slide movement, a tsunami, a build up in pore-fluid pressure or oversteepening of slopes could equally as well have triggered the slide. The latter mechanism is particularly appealing; any movement on the Lynmouth - East Lyn Fault would have resulted in slope oversteepening. Compare with text-figure 2.5. N.B. The Lynmouth - East Lyn Fault plane is shown as vertical, although it could equally well have been listric.
- B. Formation of F1 major overturned fold during the primary Variscan deformation of north Devon.. Note the fanning cleavage
- C. Reverse movement on the Lynmouth - East Lyn Fault resulting in the cutting out of the complimentary anticline to the F0 syncline. Line of section shown on text-figure 2.1. Simple trigonometric calculations indicate a movement in the order of 1600m on the Lynmouth - East Lyn fault, compared to the 2000m proposed by Whittaker (notes accompanying B.G.S. Sheet 277 - Ilfracombe). The inset showing the F0 syncline shows the interesting relationship of the cleavage and the F1 tectonic folds with the F0 syncline - see text for discussion.
- D. Section from B.G.S. Sheet 277 (Ilfracombe). Compare with C above.



Note: Scale of FO fold is exaggerated



Section from Geol. Surv.G.B., Sheet 277 (Ilfracombe)



Text-fig. 2.5 Idealized cross-section of a submarine failure as interpreted from seismic sections.

From Farrell (1984 Fig 6c); after Lewis (1971).

2.2.1.1.2 Palaeoslope

The movement direction of the slide sheet is assumed to have been perpendicular to the mean axis of the slide sheet folding (F0) (cf. Woodcock 1979b). Thus, it is necessary to find the original orientation of the F0 slide sheet fold. Its present orientation and direction as measured directly or as determined from poles to bedding (π -axis - see text-figure 2.3) is a few degrees towards 098° i.e. slightly south of east. The slide sheet fold is now in a near vertical plane on the northern limb of the 'Exmoor Anticline'. If these beds are unfolded, it is clear that the slide sheet slid in a direction approximately towards the south or south-west.

The accurate determination of the original orientation of the F0 fold was found by removing the plunge of the regional fold axis (F1) and then rotating the steep limb back to horizontal. The regional fold axis (F1) of the 'Exmoor Anticline' plunges approximately 22° towards the ESE as determined from a stereographic projection of bedding in the north Exmoor region. This ESE plunge is corroborated by the outcrop pattern and by the intersection of the cleavage and bedding east of Lynmouth. When the regional (F1) fold axis is unplunged and the gross bedding east of Lynmouth is rotated about the F1 axis back to horizontal, it is apparent that the F0 axis plunges a few degrees towards the ESE. Thus, the F0 fold was formed during the translation of a slide sheet down a SSW dipping palaeoslope (see text-figure 2.8).

2.2.1.2 The Significance of the Size of the F0 Slide Unit

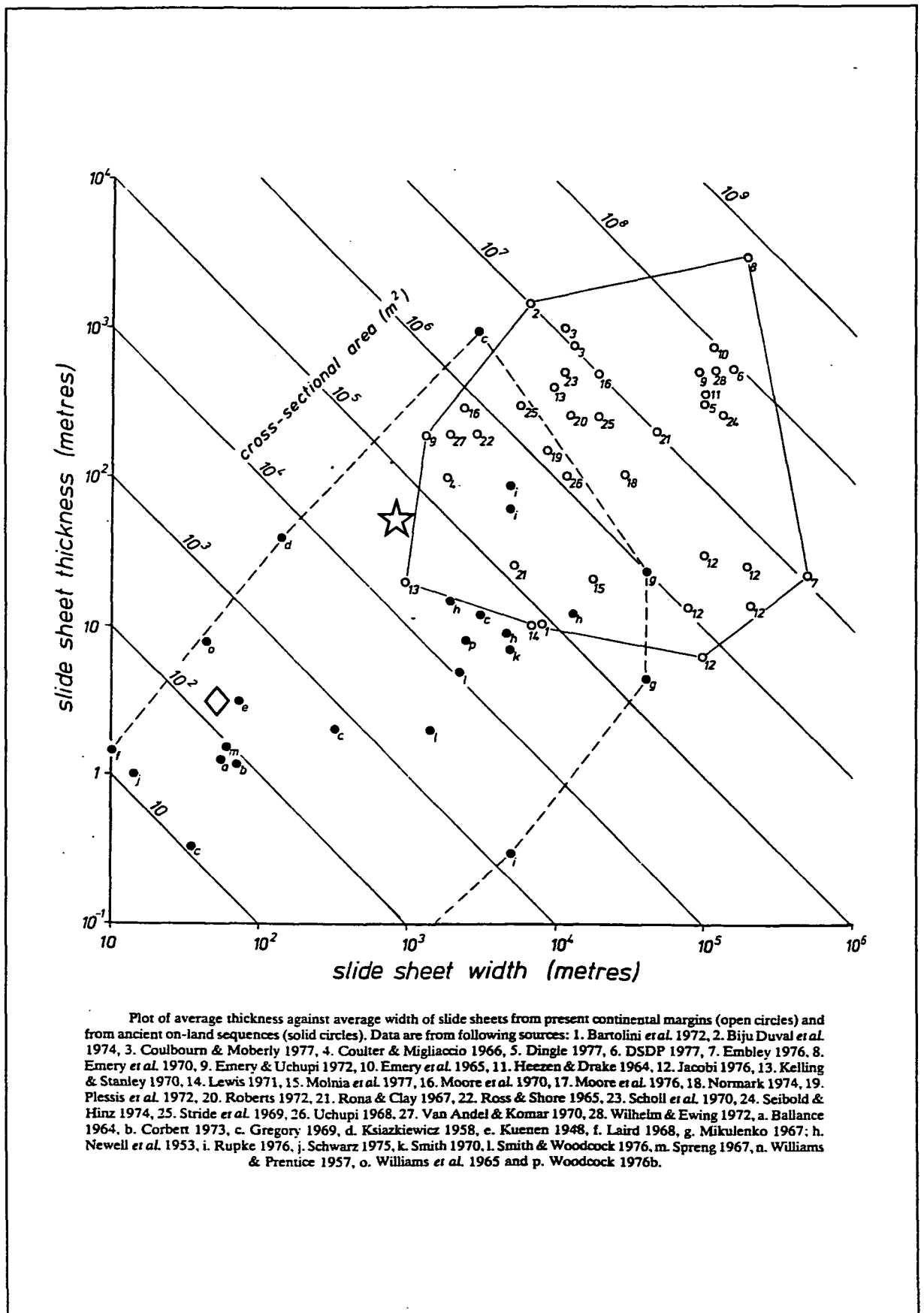
Woodcock (1979a, p.137) has observed that “. . . the submarine slides now commonly revealed by seismic reflection profiling on present-day continental margins are, on average, several orders of magnitude larger than the slumps and slides inferred from the on-land ancient geological record.” Woodcock showed this diagrammatically by plotting slide sheet thickness against slide sheet width (reproduced in text-figure 2.6). Woodcock attributed this discrepancy to the fact that:

- (i) Small-scale slides on present-day continental margins cannot be imaged by seismic reflection.
- (ii) Ancient examples of large slide units are being misinterpreted and attributed to some other tectonic mechanism.

When the Lynton Formation slide unit is plotted on Woodcock's graph (☆ in text-figure 2.2) it is clear that the unit plots near the overlap in fields of modern continental margin slides and ancient slides. Furthermore, the Lynton Formation plot is based on the **minimum** size of the slide, due to limited exposure.

In conclusion, the Lynton Formation slide sheet is one of the largest of such sheets recognised in the geological record.

An example of a slide sheet from the Eocene Hecho Group (slope deposits) is shown in plate 2.5A. This sheet is demonstrably syn-sedimentary in origin. It is suggested that the Eocene example is analogous in origin to the F0 fold **before** it was refolded during the primary Variscan deformation of north Devon.



Text-fig. 2.6 Plot showing the dimensions of slide units in the geological record and from present day continental margins.

The slump unit at Ruddy Ball (◇) and F0 slide unit east of Lynmouth (☆) have been plotted on the diagram. References are given in Woodcock (1979a) - Diagram = figure 1 of Woodcock.

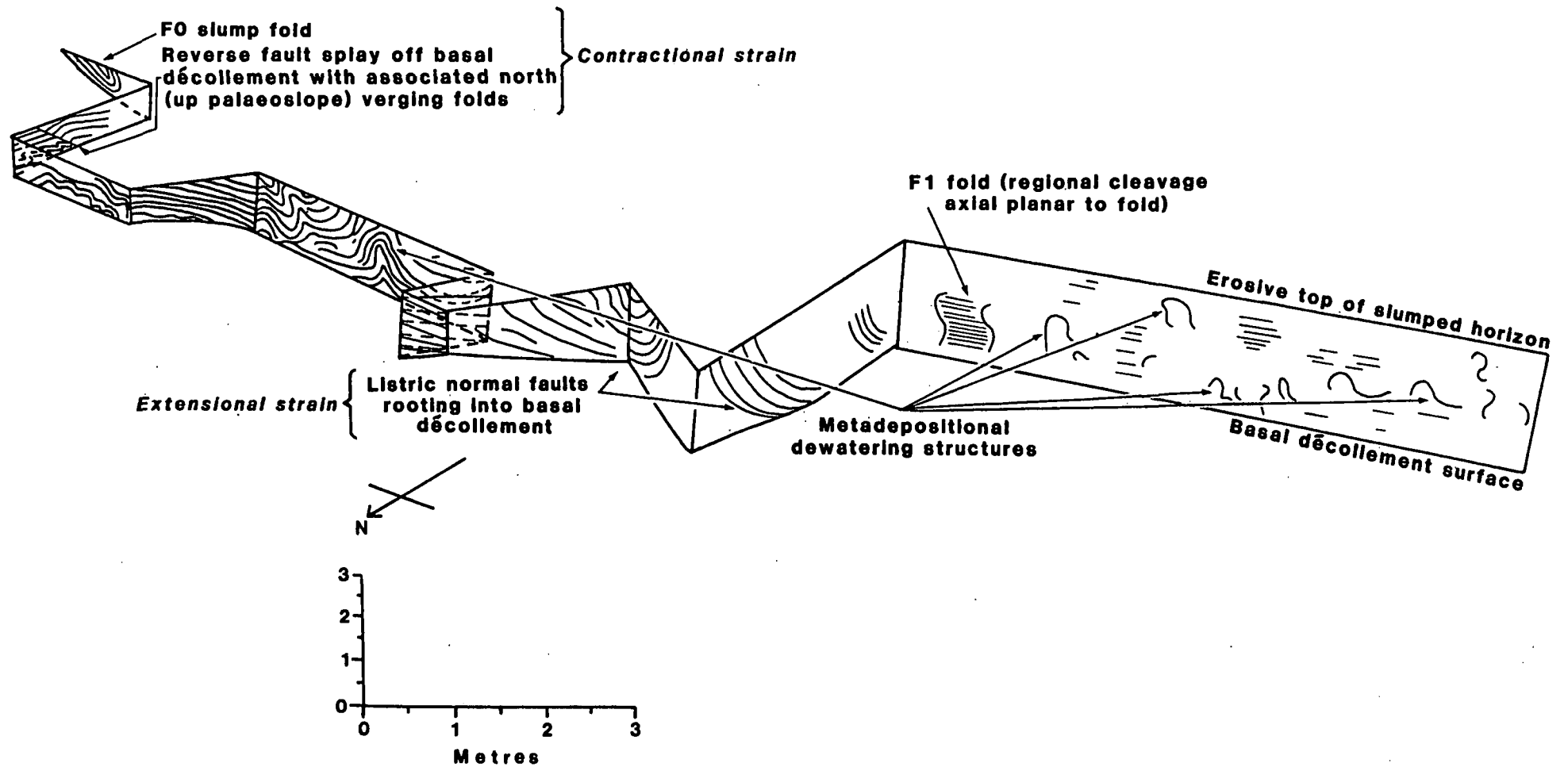
2.2.2 Ruddy Ball Slump

Grid Reference: 7137 5005

At Ruddy Ball, a location immediately south of the projected line of the offshore continuation of the Lynmouth - East Lyn Fault, a slump scar is preserved. The concave-up basal décollement surface is several tens of metres wide and is overlain by an infill of disturbed sediments reaching a thickness of several metres. The infill displays large-scale metadepositional dewatering structures and deformational features indicating overall shortening, although in detail, rapid transitions are observed from features indicating extensional strain to features indicating contractional strain.

The horizon figured in text-figure 2.7 is termed a *slump* i.e. a specific type of *slide* where rotational motion occurs on a concave-upward shear plane (Coates 1977, who defined *slides* as downslope mass movements “. . . with displacement along recognised shear surfaces where the ruptured mass moves with some semblance of unitary motion”). Farrell (1984) has described features commonly associated with slumped sediment masses. Several of the features described by Farrell were observed within the Ruddy Ball slumps (see text-figure 2.7):

- (i) A basal décollement / shear plane overlain by sediments displaying phenomena indicative of overall shortening.
- (ii) Rapid spatial transitions from features indicative of extensional strain (listric normal faults rooted into the basal décollement) to features indicative of contractional strain (reverse faulting with associated upslope-verging folds; slump folds).
- (iii) The erosive truncation of slump folds before the deposition of the overlying beds.



Text-Fig. 2.7 Ruddy Ball (7137 5005) slump deposit.

View facing SE, of the eastern 25m of a 50m wide slump above a convex-upwards décollement / shear plane. Note the rapid change laterally from structures indicating extension to structures indicating contraction (cf. Farrell 1984). The slump has been refolded locally by north-facing minor folds during the Variscan deformation of north Devon (D1). The tectonic cleavage at Ruddy Ball is axial planar to the F1 folds.

The movement direction of the slump sheet is assumed to have been perpendicular to the mean axis of the slump folds (Woodcock 1979b), although not all fold vergence directions and facing directions give a downslope direction, opposing vergences occasionally occurring (Farrell 1984). In addition, the majority of listric normal faults are assumed to dip down-palaeoslope (Farrell *op. cit.*) The measurement of the limited number of accessible slump folds and listric normal faults at Ruddy Ball suggests a palaeoslope dipping towards the SSW, a direction perpendicular to the Lynmouth - East Lyn Fault, a feature believed to have had a major influence on sedimentation patterns during the deposition of the Lynton Formation and consistent with the palaeoslope dip direction indicated by the F0 slide sheet described in section 2.2.1.

The initial triggering of slump movement may be related to:

- (i) Seismic shocks.
- (ii) A build up of pore-fluid pressure due to migration of pore water.
- (iii) Oversteepening of slopes.
- (iv) Cyclic loading by oscillating currents accompanying a tsunami, resulting in liquefaction

{(i) to (iii) - Allen 1982 vol. B, Chap. 9; (iv) K. A. Kastens in Cita *et al.* 1982}. Processes (i) to (iv) may have been aided by slope destabilisation resulting from bioturbation (Stanley 1971, Hecker 1982).

An examination of recent submarine slumps (Lewis 1971) has revealed a characteristic pattern of basal failure above which the strain pattern is predominantly contractional at the downslope end and extensional at the upslope end (text-figure 2.5). However, whilst the failed mass is moving downslope (translational phase), sequential velocity changes initiate strain waves which will propagate through the moving body and superimpose a new strain pattern on that already present in the sediment mass (Farrell *op. cit.*) Further strain overprinting will occur as the shear strength of the basal décollement surface exceeds the shear stress acting on it, and the sediment mass comes to rest, a contractional strain wave propagating from the front to the back of the unit where a change in slope has initiated termination of movement, an extensional strain wave propagating from the back to the front of the unit where dewatering from the back of the unit has initiated termination of movement (Farrell *op. cit.*)

In conclusion, the lack of a clear sequence of strain overprinting in the Ruddy Ball slump precludes an interpretation relating the observed strain pattern in the Ruddy Ball slump to the pattern of strain predicted by the dislocation model of Farrell (*op. cit.*) Nevertheless, the presence of a reverse fault dipping down-palaeoslope and associated northward (up-palaeoslope) verging minor folds (see text-figure 2.7; cf. Farrell *op. cit.* figure 3b) must have formed by a contractional strain wave propagating up-palaeoslope, perhaps as a result of movement termination originating at the toe of the slump due to a change in slope. However, large-scale metadepositional (*sensu* Allen 1982 vol. 2, p.352: arising “. . . either just before or immediately after deposition ceases”) dewatering structures (see text-figure 2.7) suggest that termination of movement may have been initiated by a drop in pore-fluid pressure due to dewatering. Dewatering structures of the scale observed in the Ruddy Ball slump have been recorded by M. A. Chan (cover photograph, *Bull. Am. Assoc. Petrol. Geol.*, Jan. 1984, vol. 68, no. 1) in delta-front distributary deposits.

Finally, the size of the slump at Ruddy Ball is similar to that of many slide units in the geological record - see text-figure 2.6.

2.2.3 Intraformational Conglomerate

In the sediments adjacent to the Lynmouth - East Lyn fault at Ninney Well Bay (7347 4956), an unusual conglomeratic lithology is developed. This deposit was first recognised by Prof. S. Simpson (unpublished notes - referenced in Edmonds *et al.* 1985 - see section 1.6.4.1) who noted that the conglomerate only occurs in isolated outcrops immediately adjacent to the Lynmouth - East Lyn Fault on Lynmouth beach and in the East Lyn valley. More recently the conglomerate was described during the remapping of the Ilfracombe sheet (Edmonds *et al. op. cit.*). The name 'Lyn Conglomerate' was applied by Tunbridge (1986) who documented the intraformational conglomerate to be the product of submarine debris flows down a fault scarp developed along the Lynmouth - East Lyn Fault, a syn-sedimentary splay from the Bristol Channel fault system.

2.2.3.1 Description

During the present study, the conglomerate was observed in the cliffs east of Lynmouth, occurring between the Lynmouth - East Lyn Fault at Ninney Well Bay (see plate 3.1A) and the Black Rocks (7295 4965). The conglomerate is particularly well exposed immediately adjacent to the fault at Ninney Well Bay and in the

walls of the gullies at 7333 4968 and 7322 4959. Exposures of the conglomerate have also been recorded at Myrtleberry (7417 4901) and north of the old limestone quarry at Watersmeet (7473 4872) by Prof. S. Simpson (unpublished notes - referenced in Edmonds *et al.* 1985), although an extensive reconnaissance during the present study failed to reveal these exposures.

Immediately adjacent to the Lynmouth - East Lyn Fault at Ninney Well Bay (7347 4956), the conglomerate occurs in units up to 3m in thickness, separated by thin (<0.4m) units of strongly cleaved lenticular bedding. Although inaccessible in its higher reaches, the conglomerate appears to persist for some considerable thickness immediately adjacent to the fault. The conglomerate thins rapidly westwards, reaching thickness of as little as 9cm at the Black Rocks (7295 4965), where the conglomerate is interbedded with thick sequences of heterolithic lithologies. The conglomerate appears to occur in laterally extensive sheets which have undulating, non-erosive bases, extending in excess of 50m away from the line of the fault to Black Rocks, contrary to the observation of Whittaker & Edmonds (in: Edmonds *et al.* 1985, p.27) indicating that clasts in the conglomerate: “. . . die out laterally away from the fault within about 5m”. The tops of the sheets are irregular, with clasts protruding from the tops of the deposits, these irregular surfaces being concordantly draped by the overlying deposits (plate 2.5B). The conglomerate comprises intraformational clasts ‘floating’ in a massive, structureless matrix of dark grey mudstone with a low silt content. The ratio of clasts to matrix ranges from approximately 1:10 in the thick conglomerate units adjacent to the fault (plate 2.5C) to 1:4 in the more distal, thinner units (plate 2.6A). The conglomerate is poorly sorted, polymodal, and has a fairly even distribution of clasts within any one unit. The clasts range from 0.3cm in diameter to 60cm in length (plate 2.5C), reaching maximum thickness of 16cm (plate 2.6B); clasts range from well rounded to tabular in shape (plate 2.5C).

A wide range of clast lithologies are present, although all clast types have been observed to occur elsewhere in the lower Lynton Formation i.e. the clasts are intraformational in origin. The following clast types have been recorded:

- (i) Light grey silty-mudstone, clasts ranging from rounded to tabular in shape (plate 2.5C).
- (ii) Limestone; calcareous silty-mudstone with >50% carbonate by weight, clasts tend to be rounded.
- (iii) Siltstone (plate 2.6C), clasts tend to be sub-rounded.
- (iv) Rounded to sub-rounded clasts of sandstone with a calcite cement (plate 2.6B).

- (v) Sub-rounded to tabular clasts of heterolithic lithology (plate 2.7A).
- (vi) Well rounded vein quartz of granule to pebble grade (plates 2.5C & 2.7B).
- (vii) Disarticulated and fragmented bivalve and brachiopod valves (plate 2.7B).
- (viii) Crinoid ossicles.
- (ix) Bone fragments.

Many of the above range of clast-types show evidence suggesting that they were transported in an unlithified form. Many of the clasts with a large mudstone component show buckling (plates 2.5C & 2.7A) and some clasts were fractured but retained their original outline (plate 2.7B). Other clasts have a diffuse boundary with the surrounding matrix. The rounded to subrounded clasts of sandstone, however, have a calcite cement (plate 2.6B) and appear to have been transported in a lithified state. The calcite is interpreted as early diagenetic in origin. Similar intraformational sandstones with an early diagenetic origin were recorded by Goldring (1971) within his Rough Facies.

In terms of a clast fabric within conglomeratic units, the measurement of the orientation of the axes of clasts (where lengths : $A > B > C$) reveals a preferential alignment of the plane within which the A and B axes lie, parallel to bedding i.e. tabular (plate 2.5C) and oblate (plate 2.6B & C) clasts lie parallel to bedding. In rare circumstances, the examination of vertical sections through the deposit reveals an upwards increase in clast percentage and clast size, developed near the base of individual units i.e. inverse grading (plate 2.5C). Bedding plane views of the conglomerate (plate 2.6A) do not reveal any preferred clast orientation.

No evidence of colonisation by a hard-bodied fauna (*in situ* shells), or soft-bodied fauna (bioturbation) has been recorded within conglomeratic units.

2.2.3.2 Discussion

Fisher (1971) described the characteristics of debris flows, noting that they are poorly sorted and commonly contain large fragments resting unsupported in a finer grained, internally structureless matrix. Fisher also noted that tabular clasts often lie parallel to flow surfaces, and inverse graded intervals may occasionally be developed. It is apparent that the Lynton Formation conglomerate closely fits Fisher's description.

Middleton and Hampton (1976) defined debris flows as sediment gravity flows in which clasts (sand, gravel and boulders) are primarily supported by the yield strength of the matrix in which they are dispersed, with a contribution from clast buoyancy. The yield strength of the matrix (competence), composed of clay-laden water, was again stressed as the primary factor for clast entrainment by Hampton (1975, 1979). Debris flows behave as non-Newtonian fluids, Bingham plastic behaviour best describing their behaviour (Johnson 1970), and flow in a laminar manner (Kurtz & Anderson 1979).

Given the above physical basis, debris flows have been recognised widely in both the geological record and the Recent. Many of the features described from Recent and fossilised debris flows may be observed in the Lynton Formation debris flow.

Cossey and Ehrlich (1979) recorded debris flows thinning away from active faults, the debris flows having a similar thickness and lateral persistence to the Lynton Formation examples. The Lynton Formation debris flow geometry's fit into the 'sheet-like' category of Hill *et al.* (1982). The non-erosive nature of the base of many debris flows is attributed to their laminar nature of flow by Johnson (1970) who notes that bases are only erosional in cases where the rigid plug migrates through to the base of the deposit during flow. Clasts protruding from the top of the deposit are common in debris flows (Enos 1977), a poorly sorted, massive deposit with a polymodal clast population characterising debris flows (Hill *et al.* 1982).

Debris flows commonly contain only intraformational clasts. For example James *et al.* (1980) noted that clasts in the Cow Head Breccia of Western Newfoundland, which they interpreted as a debris flow, were similar to surrounding lithologies. The presence of contorted clasts (cf. Peterson 1965) suggests that many of the clasts were transported in an unlithified state.

The presence of tabular clasts aligned parallel to flow surfaces is believed to reflect the laminar nature of flow (Hampton 1975, Enos 1977, Hill *et al.* 1982), as is the preservation of fractured clasts (Enos *op. cit.*) Inverse grading in debris flows is believed to be related to the internal distribution of dispersive pressures; a full review of the origin of inverse grading is given by Naylor (1980). The lack of a fauna in debris flows has been noted by both Hecker (1982) and Hill *et al.* (*op. cit.*)

In conclusion, the conglomeratic deposits adjacent to the Lynmouth - East Lyn Fault display many of the features characteristic of debris flows. The intimate association of the debris flows with the Lynmouth - East Lyn Fault suggests that debris flow transport may have been triggered by movement along the fault line resulting from either a seismic shock or, alternatively, localised steepening of the palaeoslope. Enos (1977) has noted that debris flow initiation need not be catastrophic, a build up in pore-fluid pressure being sufficient to initiate movement; such movement would be aided by a localised steepening of the palaeoslope over a fault line. However, it is also possible that the debris flows may have been triggered by cyclic loading by oscillatory currents accompanying a tsunami.

In conformation with the recommendations of Hedberg (1976), the Lynton Formation debris flows are not referred to a formal lithostratigraphic hierarchy as slides, slumps, mud (debris) flows were regarded by Hedberg to emphasise mode of origin, and thus should only be given informal names. Furthermore, the debris flow deposits have not been assigned to a named lithotype (see section 1.8.1.3) as the Lynmouth Beach section adjacent to the Lynmouth - East Lyn Fault is structurally complex and a suitable type section could not be found.

2.2.4 Palaeo-shoreline Trend and Fault Styles

Evidence presented in the previous sections indicates that the Lynton Formation was deposited in a tectonically active region and that the Lynmouth - East Lyn Fault was subject to periods of syn-sedimentary movement during the period that the Lynton Formation was deposited. Several other lines of evidence presented in later chapters support the conjecture that the Lynton Formation shelf was tectonically active and that the Lynmouth - East Lyn Fault exerted an influence on depositional patterns preserved within the Lynton Formation. These additional lines of evidence are summarised below:

- (i) The thickest development of the 'Watersmeet lithotype' occurs near Watersmeet and is interpreted as having been catastrophically emplaced and subsequently reworked (see section 4.4). This locality occurs immediately adjacent to the Lynmouth - East Lyn Fault and contains palaeocurrent indicators which suggest that the fault plane presented a positive scarp feature at the time of deposition of the 'Watersmeet lithotype' at this stratigraphic level (see section 4.2.4).

(ii) A coarsening-upwards sequence occurring along the A39 road in the East Lyn Valley (described and interpreted in section 4.3.3) contains evidence for rapid shallowing resulting from catastrophic emplacement of large volumes of sand which were subsequently reworked. The location of the sandstone-body proximal to the Lynmouth - East Lyn Fault is interpreted to be linked to movement on the fault triggering the catastrophic input of the sand preserved as a sandstone-body in the upper part of the A39 sequence (see section 4.4).

(iii) SE of Wringapeak (6730 4944) a 3.4m thick sandstone unit, exposed in a coastal gully, displays large-scale soft sediment deformation features (see plate 7.7; described in section 7.6.1). The sandstone unit is several metres thick and is referable to the 'Woody Bay facies association'. It comprises cosets of hummocky cross-stratification that have been extensively perturbed by soft sediment deformation in the form of 'ball-and-pillow' structures. Although the generation of soft sediment deformation features can be triggered by a variety of mechanisms, loading on the scale observed at this locality strongly suggests a seismic-shock origin. Although this locality does not lie adjacent to the offshore extension of the Lynmouth - East Lyn Fault, it is tempting to suggest that movement on this or associated faults may have supplied the requisite seismic pulse to trigger the generation of the load structures at this locality.

Movement on the Lynmouth - East Lyn Fault at the close of the Early Devonian may have been linked to movements in the Bristol Channel Fault Zone to the north which Tunbridge (1986) suggested was active during the Devonian. The Bristol Channel Fault Zone in the outer Bristol Channel trends towards 099° (see compilation figure 7 of Tunbridge *op. cit.*), whilst the Lynmouth - East Lyn Fault has an average onshore trend towards 108° . Extrapolation of this average trend offshore indicates that the Lynmouth - East Lyn Fault would converge on the Bristol Channel Fault Zone after some 80km, at a point 13km south of St Govan's Head in south Dyfed. The figure for the average trend of 108° for the Lynmouth - East Lyn Fault, however, is significantly influenced by a 3.5 km anomalous section between Wilsham and Malmsmead. If this short anomalous section is ignored, the trend of the fault swings from 109° to 114° in a westerly direction. Thus, the fault appears to have an overall arcuate trend, the closing angle increasing in a westerly direction. If the trend of 114° is extrapolated offshore, the Lynmouth - East Lyn Fault would meet the Bristol Channel Fault Zone after 50km, at a point 20km SW of Worms Head in SW Gower.

Although highly speculative, there is no direct evidence to support the conjecture that the Lynmouth -East Lyn Fault does continue for an appreciable distance offshore to converge on the Bristol Channel Fault Zone (the seismic survey reported in Lloyd *et al.* 1973 only traced the fault a short distance offshore of the north Devon coast), the hypothesis of the faults converging is consistent with expected fault patterns for regions dominated by strike-slip motion (see section 1.6.3). Furthermore, arcuate faults converging on major fault zones with an increasing closing angle are characteristic of strike-slip dominated zones (cf. Crowell 1974). If the Lynmouth - East Lyn Fault does continue offshore to meet the Bristol Channel Fault Zone, the figure of 114° (possibly even progressively increasing further westwards), suggesting an arcuate fault trend, is more likely than an average 108° linear fault trend.

If we do consider the Lynmouth - East Lyn Fault as converging on the Bristol Channel Fault Zone (a hypothesis documented in Tunbridge 1986), a westward-thinning wedge geometry results (text-figure 2.8) typical of anastomosing fault systems occurring in strike-slip fault zones (cf. Crowell 1974). By applying the results of studies on present day fault systems, an understanding of the fault system at the close of the Early Devonian may be gained.

A discussion of published material relating to the palaeo-stress field in the Exmoor region during the late Emsian - early Eifelian presented in section 1.6.3 concluded that during that period the region experienced dextral strike-slip shear along E-W trending fault zones and overall crustal extension/thinning (i.e. 'transtension' *sensu* Harland 1971) and localised uplift. Sanderson and Marchini (1984) mathematically modelled the effects of shear on a theoretical strain ellipsoid and considered the geological structures that could be expected to result; the conclusions of this study are diagrammatically shown in text-figure 2.9. For the special case where there is neither extension or compression across the shear zone ($\alpha^{-1} = 1$ in text-figure 2.9) classic Riedel shears will develop and 45° obliquity of structures will result. Where there is overall shortening across the shear zone ('transpression' *sensu* Harland *op. cit.*; $\alpha^{-1} < 1$ in text-figure 2.9) folds and thrusts initiate at much lower angles to the zone, whereas extensional features such as veins, dykes and normal faults will initiate at higher angles. Where there is overall extension across the shear zone ('transtension' *sensu* Harland *op. cit.*; $\alpha^{-1} > 1$ in text-figure 2.9) folds and thrusts initiate at much higher angles to the zone, whereas extensional features such as veins, dykes and normal faults will initiate at lower angles. Note in particular that in text-figure 2.9 the diagram for transtension ($\alpha^{-1} > 1$) shows the dextral

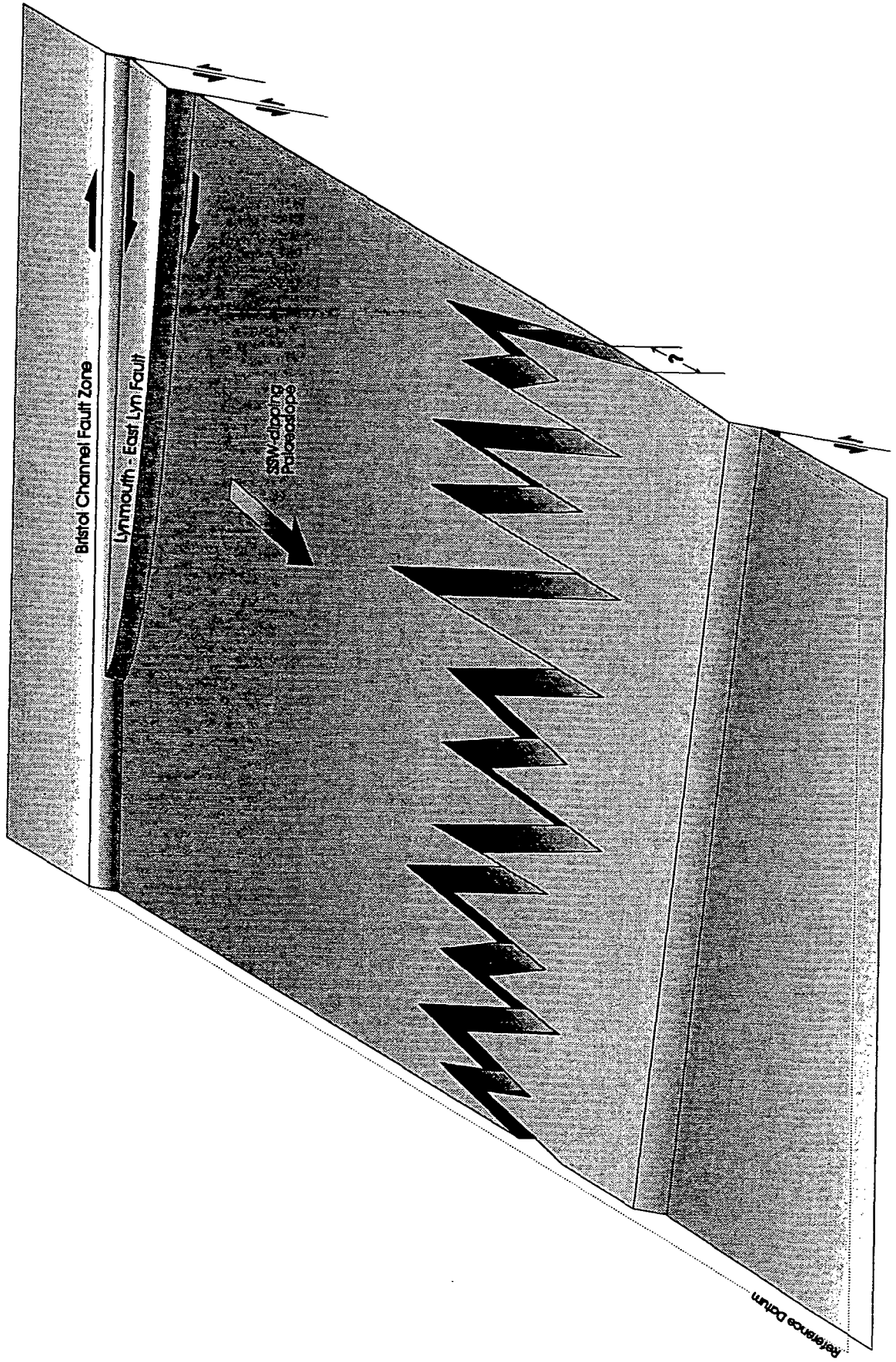
synthetic Riedel shear (labelled 'R') as being parallel to the shear zone, whilst extensional normal faults develop at a low angle to the shear zone.

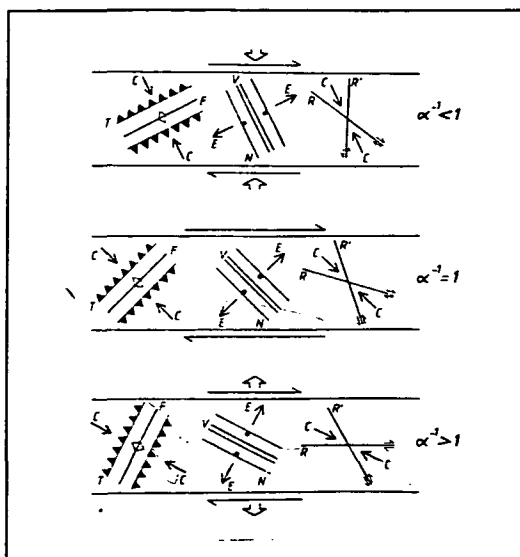
By applying the transtensional model to the north Devon region at the close of the Devonian we could predict that dextral shear along the E-W trending Bristol Channel Fault Zone would result in extensional normal faults developing at a low angle to the shear zone and trending ESE-WNW; the Lynmouth - East Lyn Fault is interpreted to be one such extensional normal fault. There is insufficient geological evidence preserved to ascertain whether the Lynmouth - East Lyn Fault was *created* in response to late Early Devonian dextral transtension or whether the fault had a more ancient origin i.e. Tunbridge (1986) proposed that the Bristol Channel Fault Zone was sited along a long-lived zone of structural weakness with a history spanning back to at least the Lower Palaeozoic. A further prediction of the Sanderson & Marchini model is that transtension will result in: "crustal thinning, subsidence and basin development" (p.458) - this theme is explored in chapter 8.

Evidence presented in this chapter indicates that the Lynmouth - East Lyn Fault had a significant influence during certain depositional episodes preserved within the Lynton Formation. Furthermore, slump and slide sheet deposits indicate a SSW dipping palaeoslope, a direction normal to the faultline, suggesting that the faultline ran parallel to the palaeo-slope and therefore the palaeo-shoreline trend. This is consistent with the observation of Heward (1981, p.246) that: "The majority of modern non-deltaic shorelines are oriented alongshore perpendicular to palaeoslope".

Text-fig. 2.8 The Lynmouth - East Lyn Fault and its relationship to the Bristol Channel Fault Zone
Diagram = over page

The block diagram shows the interpreted geometrical relationships of syn-sedimentary fault blocks developed in the Bristol Channel and north Devon area during the late Emsian in a dextral transtensional setting. Compare with text-figure 2.9, $\alpha^{-1} > 1$ case: the Bristol Channel Fault Zone corresponds to the major dextral wrench fault trend whilst the Lynmouth - East Lyn Fault corresponds to the expected normal fault trend. This resulted in a half-graben type basin developing with a hanging-wall basin floor that dipped towards the WSW, a direction normal to the trend of the Lynmouth - East Lyn Fault. A normal fault parallel to the trend of the Lynmouth - East Lyn Fault, delineating the southern margin of the basin, is predicted. The faults are extensional and dip to the south - no attempt has been made to show whether the faults were planar or listric at depth. Note that hanging-wall downthrow would have been accompanied by (a smaller amount of) foot-wall uplift. N.B. The palaeo-horizontal reference datum does not imply sea-level for any particular point in the late Emsian.





Text-fig. 2.9 The Sanderson - Marchini transpression model

Diagrams to show orientations of fractures in the transpression model (for the dextral overall shear case):

$\alpha^{-1} < 1$: transpression i.e. horizontal shortening across the shear zone

$\alpha^{-1} = 1$: classic wrench tectonics where horizontal distance across the shear zone is maintained

$\alpha^{-1} > 1$: transtension i.e. horizontal extension across the shear zone

C - compression axis; **E** - extension axis; **N** - normal faults; **T** - thrust faults; **R** - synthetic (dextral) Riedel shears or wrench faults; **R'** - antithetic (sinistral) Riedel shears or wrench faults; **V** - veins, dykes or extension fractures; **F** - fold axes.

See text for discussion.

The WNW-ESE shoreline trend predicted for the Lynton Formation appears to have persisted through the Devonian period in the north Devon region. Webby (1966) discussed the palaeogeography of the Givetian-Frasnian deposits of north Devon and west Somerset (i.e. the marine uppermost Hangman Sandstone Group and the Ilfracombe Slates), in terms of the oscillation of a WNW-ESE trending strandline which retained its orientation over a considerable period (synthesised in figure 2 of Webby). Goldring (1971), in a study of the Baggy Beds (Famennian), noted that "... the shoreline remained relatively stable during Baggy times along a WNW-ESE line" (p.38). It appears, therefore, that a WNW-ESE shoreline trend persisted from the late Emsian through to Famennian times in the north Devon area, a period spanning more than 20 million years. Goldring & Langenstrassen (1979 p.81) noted that for Devonian shelves "though the rate of sediment input was high, subsidence ensured that the shoreline was relatively stable over long periods". Harms *et al.* (1982) have observed that in many geological sequences subsidence or loading compaction closely match deposition

rate giving rise to thick paralic sequences e.g. the 3000 to 5000m thick Pliocene and Pleistocene succession along the storm- and wave-dominated Makran coast, western Pakistan. The Makran overlies an active subduction zone with a coastal mountain range subject to continuous uplift on the margin of the Indian Ocean in an area subject to a monsoonal climate with strong summer storms with large amounts of sediment supplied to the coast by widely spaced but ephemeral streams. The preserved sequences showed a series of regressive pulses (1000 - 3000m thick comprising many individual cycles in places as many as 20 to 60 representing shoaling sequences) slope to near shore facies transitions; an average sedimentation rate of 1 to 3m per thousand years was reported. The major regressive pulses showed lobate forms in plan view up to 100km broad, each lobe is thought to be a focus of fluvial drainage supply where shelf progradation was more rapid and coarser grained near the stream mouths and finer grained and less rapid between the lobes. Sand was moved offshore rapidly by wave and storm currents.

It is possible that the influence of the Lynmouth - East Lyn Fault on the shoreline trend continued along the projected eastward and westward extension of the observed faultline. Once this shoreline trend had become established, it is likely that the trend would have been sustained over a considerable period of time, albeit fluctuating in response to transgression/regression and basin subsidence, in the absence of any further change in regional syn-sedimentary tectonic style. An alternative possibility is that the Lynmouth - East Lyn Fault is but one of a set of parallel faults, any of which could have influenced the position of the Devonian shoreline in the region.

Geomorphologists, in classifying shoreline types, generally include the category of 'fault controlled coasts' e.g. Johnson (1919), Cotton (1952), Shepard (1963). Examples of present day coastlines controlled by faults include the North Island coast of New Zealand and the NE coast of San Clemente Island, California (Johnson 1919); Gawthorpe *et al.* (1994) described an active fault controlling the shoreline trend ('South Alkyonides fault segment') along the edge of the Gulf of Alkyonides in central Greece. Surlyk (1978) gave an example of ancient fault controlled coastlines from the Mesozoic basin of East Greenland where faults ". . . were active during sedimentation and formed, or at least controlled the position of the actual coastlines" (p.130). The preceding discussion indicates that the Lynton Formation shoreline should be assigned to the geomorphological category of a 'fault controlled coast'

It is appropriate at this point to discuss the characteristics of the fault pattern presented in figure 2.8 as it provides a potentially predictive framework for discussing the pattern of depositional styles presented in the subsequent mega-facies chapters of this thesis. Gibbs (1987) has shown that the majority of sedimentary basins that have developed around the British Isles are not the product of simple extension. Rather, they are of mixed mode (transpressional and transtensional). The faults that define these basins will either be listric or planar. The latter structures detach at around 10 to 15km in a ductile zone in the lithosphere. The planar faults must act as a domino fault array in order to resolve the geometric 'space problem' that would otherwise result at the end of an array. It is not possible to state whether the Lynmouth - East Lyn Fault would have had a planar or listric geometry during Devonian extension. Jackson *et al.* (1982), however, examined seismically active domino planar fault arrays in the Aegean and reported dips of $\cong 40^\circ$; these would probably be too steep to act as ramps during compression (Gibbs 1987). The reactivation of the Lynmouth - East Lyn Fault with a reverse sense of movement during Variscan transpression suggests, therefore, that the fault possibly had a listric geometry during the Devonian transtensional phase.

Yielding and Roberts (1992) presented an extensive analysis of normal faulting in the North Sea and the sedimentation patterns resulting from foot-wall uplift as fault blocks regained isostatic equilibrium following a seismic event. The ratio of uplift to subsidence is approximately 1:4 - a figure that is applicable to most normal fault zones. Thus, uplift of the foot-wall and subsidence of the hanging-wall will generate positive and negative loads respectively. Erosion of the uplifted foot-wall block results in deposition of sediment in the hanging-wall graben thus reducing these initial loads and restoring isostatic equilibrium. The implication of this model is that extension/transension across normal faults will result in uplift of the foot-wall to the basin margin and possibly emergence (?with localised fringing littoral sandstone prisms). This emergence and ensuing erosion can result in localised deposits of coarser foot-wall material being incorporated into the hanging-wall depocentre (?Watersmeet lithotype'). In addition, extensive slumping can be expected along the active fault scarp. Where fault throws are large, transport of material eroded from the area of maximum uplift may be predominantly down the dip-slope of the foot-wall rather than directly over the steep fault-scarp. This can lead to hanging-wall basins being starved of coarser grade material and becoming dominated by argillaceous successions. The effect of hanging-wall subsidence and foot-wall uplift will result in half-graben basins which may have a foot-wall sill / emergent barrier (see text-figure 2.8) which can result in reduced basin circulation and possibly anoxic bottom conditions.

2.2.5 Conclusion

A picture has been presented in the preceding sections of movement, in a dextral transtensional setting, on the ESE-WNW trending Lynmouth - East Lyn Fault providing a trigger to initiate the translation of slide sheets, slump deposits and debris flows down a palaeoslope dipping towards the SSW. The fault-line itself appears to have provided a major influence on the trend of the contemporary shoreline.

2.3 THE VARISCAN OROGENY (D1 DEFORMATION)

Transtensional forces during the Devonian and Carboniferous in the Exmoor region were replaced by compressional forces during the late Carboniferous marking the onset of Variscan deformation in the area. The structural weakness presented by the Lynmouth - East Lyn Fault was exploited by Variscan compressive forces and resulted in major reverse displacement on the fault - see section 1.6.4 for a review of published material relating to the structural geology of the Lynton Formation.

A slaty cleavage of pressure solution origin is developed in argillaceous lithologies where it dips on average 25° to 35° towards the south. Cleavage 'augening' is frequently developed around biogenically-produced sandstone-filled tubes. In sandstones a less well developed spaced pressure solution cleavage is present. Cleavage refraction is frequently observed at the junction of different lithologies. The cleavage intensity is greater adjacent to the Lynmouth-East Lyn Fault, where a lensoid cleavage fabric develops in heterolithic lithologies due to pressure-solution, similar to that described by Hobson and Sanderson (1983, Fig. 6.9b). The cleavage is axial planar to the main 'Exmoor Anticline' and small- and medium-scale folds which are overturned towards the north and verge southwards.

At Duty Point upper cliff (6943 4967) a number of populations of 'mantled' tubes, produced by an organism similar to Recent cerianthid anemones, occurs (see appendix B). The tubes have been tilted to plunge at a mean angle of 22° 36' towards 173° 46' (see text-figure B.5 for stereogram plot of tube populations) from their original vertical orientation. Strain calculations assuming simple shear (appendix C) yielded a value of 6.5 : 2.5 : 1 for the X:Y:Z strain ellipsoid at this locality, indicating that the tubes have been tectonically extended by approximately 260%. The tubes lie at an angle close to the direction of maximum extension, some of the tubes actually lying within the plane of cleavage. The Lynton Formation X:Y:Z strain ellipsoid

compares with an average value of 7.5 : 6.5 : 1 for the Ilfracombe Group recorded by Jeffery (1971) based on detailed measurement of ooids.

In section 2.2.1 D1 deformation structures were described from the east Lynmouth Beach area. Small-and medium scale folds were described that refold the F0 synform. Minor thrusts and fold-thrust structures are developed which exhibit a N-S sense of movement. At Ruddy Ball soft sediment deformation structures are refolded by F1 medium-scale folds (text-figure 2.7). During the course of this study similar fold-thrust structures to those occurring at east Lynmouth Beach were also observed in the Trentishoe Formation of the Foreland Point section. Thrusting is believed to be a response to shortening during the primary Variscan deformation. Elsewhere on the north Devon coast Hollwill *et al.* (1969) described similar thrusting within the Ilfracombe Slates.

At this point it is appropriate to discuss the amount of movement on the Lynmouth - East Lyn Fault resulting from D1 reactivation. Whittaker and Edmonds (in: Edmonds *et al.* 1985, p.63) suggest “. . . a throw of 1800m to 2000m or possibly more” for the Lynmouth - East Lyn Fault in the Lynmouth - Countisbury area. The estimate was based on the fact that: “Hangman Grits resembling the Sherrycombe Formation or parts of the Rawn’s Formation have been brought into contact with strata low in the Lynton Slates”. No indication was given for the precise method of calculation of this figure. Observations by the author and Dr. I. P. Tunbridge (University of Plymouth, pers. comm.) indicate, however, that the Hangman Sandstone Group exposed in the Lynmouth - Countisbury area is solely of Trentishoe Formation type. The following paragraphs attempt to give a more detailed reasoning to enable an estimate of the throw of the fault in the Lynmouth region to be derived.

The Lynmouth - East Lyn Fault trends ESE-WNW across the Lynmouth beach foreshore and into the gully at Ninney Well (see text-figure 2.1) where the fault plane dips southwards at 75°. The line of the fault is then diverted to a NNE-SSW trend from the head of the gully 150m inland onto the flank of Wind Hill, before resuming a ESE-WNW trend (see enclosure 1). Edmonds *et al.* (1985) attribute this diversion to a branch fault origin. A more plausible explanation based on field evidence is that the ESE-WNW trend of the fault has been offset by a late normal fault.

North of the fault in the Lynmouth - Foreland Point area, the Trentishoe Formation adopts a broadly anticlinal structure representing the downfaulted axial zone of the 'Exmoor Anticline'. Field observations, and the use of a photomontage taken during a helicopter reconnaissance of the coastline reveal that the oldest Trentishoe Formation crops out at beach level in the core of the anticline where it intersects the coastline at a point in the region of the zig-zag path down to south Sillery Sands (739 496) - not at Blackhead (745 502) as suggested by Whittaker and Edmonds (*op. cit.*).

Observations and trigonometric considerations indicate that approximately 1 400m of Trentishoe Formation is exposed between the zig-zag path and Foreland point (754 511) on the northern limb of the anticlinal structure. This figure compares with a thickness of approximately 1 250m of Trentishoe Formation given by Tunbridge (1978) based on evidence from the coastal section between Ramsey Beach and Combe Martin (south of the Lynmouth - East Lyn Fault.)

It is possible that the greater thickness of Trentishoe Formation north of the fault may be due to facies variations across the fault. Nevertheless, the rocks exposed in the core of the anticline at Sillery Sands are probably near the base of the Trentishoe Formation. Observations and trigonometric calculations suggest approximately 400m of Trentishoe Formation is exposed between the core of the anticline and beach level adjacent to the fault at Ninney Well.

To the south of the Lynmouth - East Lyn Fault on Lynmouth beach the Lynton Formation has been folded into a broad synclinal structure (F0 syncline - see text-figure 2.1). Enclosure 2 indicates that the 'Lee Stone facies association' exposed at 7292 4962, in the core of the F0 syncline to be some 50m above the base of the exposed Lynton Formation sequence. The strata exposed on Lynmouth beach young towards the core of the F0 syncline (text-figure 2.1) suggesting that the Lynton Formation exposed at beach level immediately adjacent to the Lynmouth - East Lyn Fault must be near the base of the exposed Lynton Formation, confirmed by the large amount of intraformational slumping and sliding similar to that exposed at Yellow Stone where the oldest exposed Lynton Formation crops out.

We are now in a position to estimate the thickness of strata offset by the Lynmouth - East Lyn Fault at Ninney Well:

400m Trentishoe Formation (see discussion earlier in this section)

c. 70m Hollowbrook Formation (Tunbridge 1978)

200m Lynmouth Formation (thickness adjacent to fault - see enclosure 2)

The above indicates a figure of 670m total offset. This figure, however, should be qualified by the following two points. Firstly, it has been suggested in section 2.2.4 that the fault controlled the orientation / position of the Early Devonian shoreline. For this reason, it is possible that the Hollowbrook Formation may not have been developed to the north of the fault. Secondly, it was noted earlier in this section that the thickness of the Trentishoe Formation differs on either side of the fault, but comparisons suggested that the Trentishoe Formation exposed in the core of the anticline at Sillery Sands was probably near the base of the Formation. There is no stratigraphic control, unfortunately, on the total thickness of the Trentishoe Formation north of the fault and it is possible that a significant thickness of Trentishoe Formation is present below the exposed core of the anticline.

Given a figure of 670m stratal offset, it is now possible to calculate the throw on the fault in the Lynmouth area by trigonometric calculation based on the following assumptions:

- (i) The fault plane dips 75° southwards.
- (ii) Bedding strike is parallel to the fault. This appears to be true i.e. the fault is a true strike fault.
- (iii) Uniform dip of bedding either side of the fault. Although complicated by the F0 syncline south of the fault it appears that bedding dips southwards across the fault at a mean of 45° (B.G.S. Sheet 277 - Ilfracombe).

The calculated minimum throw on the Lynmouth - East Lyn Fault is 1 340m. It should be noted, however, that this figure is very sensitive to minor variations in the above assumptions. By altering the angle of the fault plane and bedding dip across the fault both by 5° , and the estimate of stratal difference across the fault by 100m, either side of the estimated values, a range of values from 887m to 2251m throw are obtained (see table 2.1). These deviation from the estimate are within the range of expected observational and natural variation. It can be concluded that the throw on the Lynmouth - East Lyn Fault should be given as being in

the order of a minimum of 1 000m to a maximum of 2 000m, rather than the 1 800m to 2 000m given by Whittaker and Edmonds (*op. cit.*)

| Estimate of Stratal Difference ↓ | Fault Plane Dip ⇒ | 70° | 75° | 80° |
|-------------------------------------|----------------------------|-------|-------|-------|
| | Bedding Dip Across Fault ↓ | | | |
| 570m | 40° | 1 140 | 994 | 887 |
| | 45° | 1 349 | 1 140 | 994 |
| | 50° | 1 667 | 1 349 | 1 140 |
| 670m | 40° | 1 340 | 1 168 | 1 042 |
| | 45° | 1 585 | 1 340 | 1 168 |
| | 50° | 1 959 | 1 585 | 1 340 |
| 770m | 40° | 1 540 | 1 342 | 1 198 |
| | 45° | 1 822 | 1 540 | 1 342 |
| | 50° | 2 251 | 1 822 | 1 546 |

Table 2.1 Range of values for the throw of the Lynmouth - East Lyn Fault in the Lynmouth area - given in metres.

Whittaker and Edmonds (*op. cit.*) suggested that the throw decreases to about ¹⁵⁰⁰~~1 500~~m on the edge of the district (near Oare). This figure was based on the observation that in the east of the district rocks higher in the Lynton Formation are faulted against the Trentishoe Formation. It was noted above, however, that the figure in the Lynmouth area was based on the incorrect observation of Sherrycombe and Rawn's Formation rocks faulted against the Lynton Formation. It should be concluded, therefore, that there is no field evidence to suggest that the throw on the Lynmouth - East Lyn Fault decreases to the east.

Apart from the Lynmouth - East Lyn Fault, the only other major strike fault affecting the Lynton Formation is the Tippacott Fault which "... trends E-W between Tippacott (76855 4713) and the Badgworthy Water valley (7890 4715) just east of the district" (Whittaker & Edmonds *op. cit.*, p.63). The fault marks the southern boundary of the Lynton Formation in this area (see Enclosure 1) and has a downthrow to the south which is unlikely to exceed 50m.

NW-SE trending normal faults are numerous in the district and are observed to intersect the coast at distances in the order of every 100m. The throw of these faults is generally only of several metres. An antithetic set of NE-SW faults are occasionally observed. Hobson and Sanderson (1983) attribute a Variscan origin to the two fault sets which probably have a late normal fault origin.

The only evidence of igneous activity within the Lynton Formation is a single dyke cutting the sequence at Ramsey Beach (6463 4936). Thin sections examined by the author reveal that the dyke has a doleritic composition. Randomly oriented laths of sericitized plagioclase feldspar (indeterminate) sit in a pyroxene groundmass that has been chloritized. Vessicles have either a quartz and chlorite or clinocllore infill. Late stage fractures are infilled with dolomite. Lloyd *et al.* (1973) described a similar dolerite from the Horseshoe Rocks which crop out beneath the Bristol Channel, 6.4km NW of Ilfracombe. The outcrop of the intrusion along the strike of the Devonian rocks and the sheared nature of the dykes led the authors to ascribe a Variscan date to the intrusion. The dyke at Ramsay Beach trends 150° - 330° (dipping 75° to the SW) and is offset by bedding-plane slip associated with Variscan folding. This indicates that the dyke is of an early Variscan age or older. When the effects of folding are subtracted the original orientation of the dyke is indicated as dipping 62° towards the WSW. Andrews (1993) presented evidence from the Pilton Shales at Croyde Bay indicating that Variscan deformation in north Devon was the product of dextral transpression. By applying the transpression model of Sanderson and Marchini (1984) to a dextral shear zone trending slightly south of east in north Devon (Gayer & Nemcok 1994), if dykes are developed a NNW - SSE trend is predicted (see text-figure 2.9, diagram for $\alpha^{-1} < 1$ i.e. transpression). This direction is coincident with the trend observed for the Ramsey Beach dyke. It is concluded, therefore, that the Ramsey Beach dyke developed in response to early Variscan transpression.

In summary, the structures within the Lynton Formation described above as being attributable to Variscan deformation are broadly consistent with "Zone 1" of Sanderson and Dearman (1973). Their Zone 1, characterising the Exmoor area, displays folds with axes which are sub-horizontal and trend east-west and are overturned towards the north and face upwards to the north.

2.4 TERTIARY DEFORMATION (D2)

Horizontal components of motion on the NW-SE trending (synthetic) set of late normal faults have been attributed by Shearman (1967) to Tertiary reactivation; no evidence to either confirm or refute this evidence was found during the course of this study.

3. THE 'BASAL MEGA-FACIES' OF THE EXPOSED LYNTON FORMATION

3.1 INTRODUCTION

The lowest sequences within the 'basal mega-facies' crop at Yellow Stone (7060 4998) and on east Lynmouth beach near Point Perilous (7280 4965) - see enclosure 2. The base of the cliffs at Yellow Stone is permanently wave-washed, even at lowest spring tides, and for this reason the sequence could not be logged. Nevertheless, visual observations confirmed that the Yellow Stone sequence exhibits characteristics closely similar to those observed in the sequence near Point Perilous. The beach section east of Lynmouth comprises vertical cliffs passing upwards into grassed slopes. This section is probably the most accessible of the coastal exposures of the Lynton Formation and the sequence has been described by Tunbridge & Whittaker (in: Goldring *et al.* 1978); the sequence is, however, structurally complex - e.g. sandstone layers are frequently 'augened' in the cleavage (plate 2.1B) - see section 2.2.1. The cliff base is protected by large boulders and *in situ* wave-washed slabs; small patches of sand occasionally lie between the boulders. The coast east of the small headland at 7311 4963 is accessible for 1½ hours either side of low tide.

The oldest part of the sequence east of Lynmouth crops out in foreshore exposures, named the Black Rocks, to the north of Point Perilous. Unfortunately these exposures are wave-washed and, for the most part, barnacle-covered and it did not prove possible to construct a log through the sequence. Nevertheless, notes and photographs were taken and have been included in the facies descriptions presented below. A short representative log through a sequence of slightly younger 'typical' 'basal mega-facies' was measured in the base of the cliff face at Point Perilous (7288 4961) - see text-figure 3.1A. Point Perilous is accessible for approximately 2 hours either side of low tide.

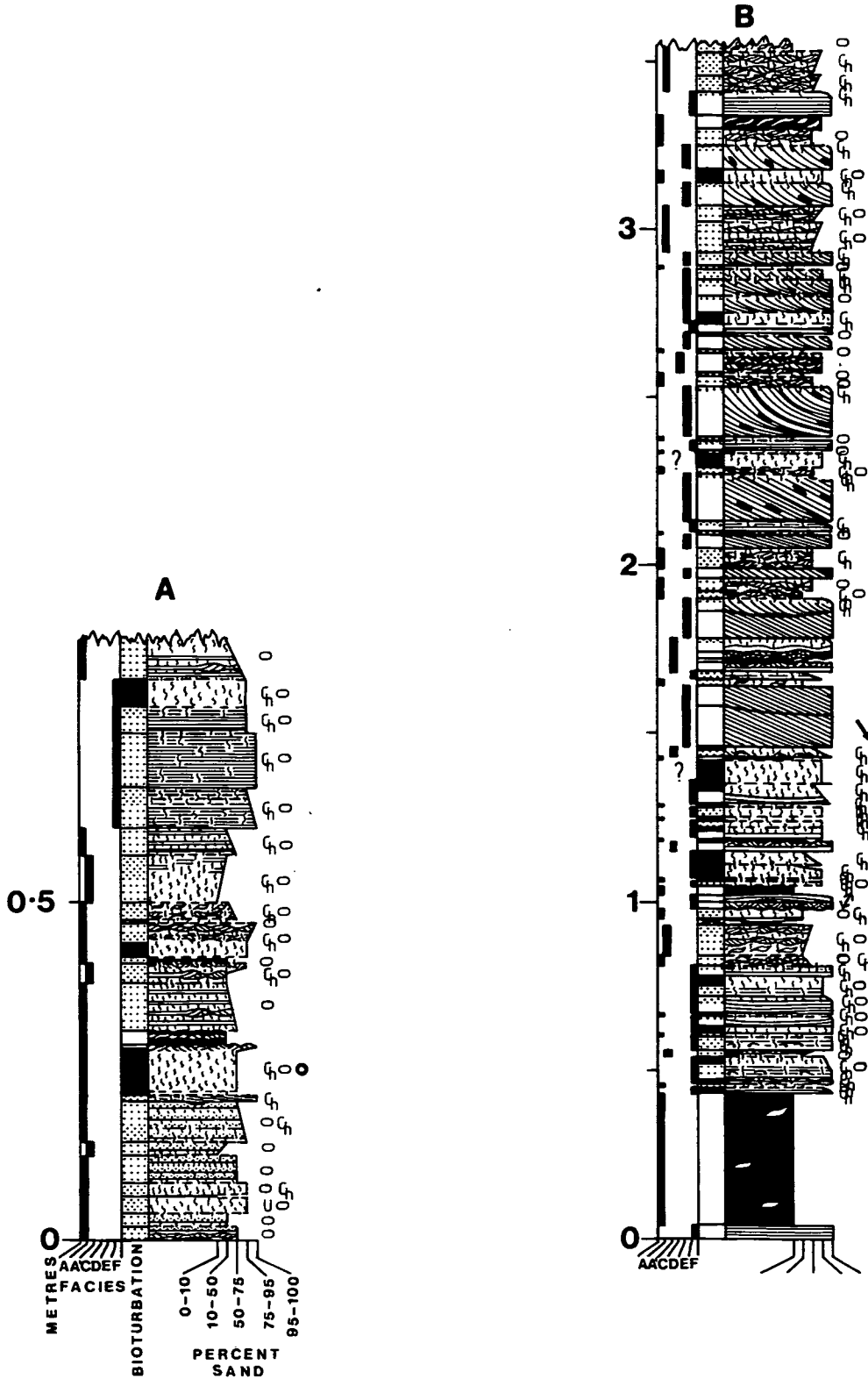
A section through the middle part of the 'basal mega-facies' occurs in foreshore slabs exposed 400m to the SW of Yellow Stone at Wringcliff Bay (7029 4978 to 7022 4967) c. 8m above the oldest strata exposed at Yellow Stone, although the precise position of the section is unclear as the two sequences are separated by late normal faulting, across which bedding could not be matched. The slabs dip at 10° towards the SSW. This section was logged (text-figure 3.2) and is accessible approximately between mid and low tide.

Text-fig. 3.1 Logs through sections at east Lynmouth beach

- A. Cliffs east of Lynmouth (7288 4961). Log through 'typical' basal mega-facies sediments exposed on Lynmouth beach.
- B. Cliffs east of Lynmouth (7292 4962). Log through a coarsening-upwards - fining-upwards sandstone body attributable to the 'Lee Stone facies association' (which is defined as occurring between 1.46 and 3.26m on the log) within 'lower-middle mega-facies' deposits - described in chapter 4.

Facies for both logs: **A** = thinly interlayered sandstone/mudstone bedding; **A'** = coarsening-upwards microsequences; **B** = not present in these localities; **C** = wavy bedding; **D** = flaser bedding; **E** = cross-bedded sandstone; **F** = thin-bedded horizontally-laminated sandstone.

Lynmouth Beach-East.



Text-fig. 3.2 Wringcliff Bay - log through slabs exposed on beach below Castle Rock

The logged sequence runs from 7029 4978 to 7022 4967. Facies present in this section: **1** = thinly interlayered sandstone/mudstone bedding; **2** = thinly-bedded sheet sandstones; **4** = flaser bedding. Facies on log can be ascertained by examination of the symbol types.

The 1.5km stretch of coast between Wringcliff Bay and Lynmouth (see enclosure 1) comprises north-facing, vertical cliffs exposing some of the lowest visible strata of the Lynton Formation. The base of the cliffs is protected by large boulders and wave-washed slabs. At Ruddy Ball, a 50m wide plano-convex lens of convoluted and folded sediments, representing a slump scar and its infill in the 'lower-middle mega-facies', is exposed high in the cliff face (see section 2.2.2; can only be approached with the use of a rope). Ruddy Ball can be reached from Lynmouth Beach, to the east, via an arduous scramble, accessible for 2 hours either side of a low tide. The transition into the succeeding 'lower-middle mega-facies' is visible at Ruddy Ball (7137 5005), some 40m above the base of the exposed Lynton Formation (see enclosure 2).

The following sections provide descriptions and interpretations of the facies found within the 'basal mega-facies', followed by a discussion on the palaeoenvironmental significance of the facies sequence in the 'basal mega-facies'.

3.2 FACIES

3.2.1 Facies 1 - Thinly Interlayered Sandstone/Mudstone Bedding

3.2.1.1 Description

Shown as facies A and A' on text-figure 3.1. Facies 1 comprises 79.8% of the Wringcliff Bay section, the only section of significant length measured through the 'basal mega-facies', where units have a mean thickness of 6.1cm ($n = 214$, $\sigma_{n-1} = 8.0\text{cm}$, min. thickness = 0.5cm, max. thickness = 90cm, where: n = sample size and σ_{n-1} = standard sample deviation - this notation is used throughout the thesis).

Although the mudstone-dominated heterolithic sediments preserved in the 'basal mega-facies' are broadly similar to those found elsewhere in the Lynton Formation, the oldest deposits (Yellow Stone and Black Rocks) are unusual in containing a higher proportion of graded rhythmites and horizontally-laminated sandstone streaks than elsewhere in the Lynton Formation, with the exception of the west Crock Point section through the 'upper-middle mega-facies' (described in chapter 5). Two other features are also characteristic of the heterolithic sediments of the Yellow Stone and Black Rocks sequences: scour-forms and intraformational slide horizons. Although both of these features are present elsewhere in the Lynton Formation, they are significantly more frequent in the Yellow Stone and Black Rocks section than elsewhere.

Individual facies 1 units generally exhibit a tabular geometry and the boundaries between facies 1 units are generally arbitrary, corresponding to abrupt changes in the gross sandstone content between individual units. Laterally extensive erosion surfaces, with up to 2cm relief, commonly separate facies 1 units. Rarely, these erosion surfaces have isolated ripples of clean sandstone resting upon them. These sandstones gradationally overlie bioturbated heterolithic deposits. Internally facies 1 units have a sandstone content varying between 0 and 95%, variations within individual units defining fining-upwards (e.g. 5.5m on the Wringcliff Bay log - text-figure 3.2), coarsening-upwards (e.g. 1.0m on the Wringcliff Bay log) and coarsening-fining-upwards microsequences; the most frequently developed microsequences coarsen-upwards but are not as common as in the 'lower-middle mega-facies' (see section 4.2.2). 1 to 3m thick intervals of groups of units with an overall higher sandstone content than that in the surrounding sequence are occasionally observed - this phenomenon is particularly apparent in the cliff face below Castle Rock which rises above the Wringcliff Bay section. The grain-size of the sand in facies 1 varies between coarse silt and fine sand, units occasionally containing fine, comminuted shell debris. Individual sandstone layer characteristics allow four sub-facies to be defined:

Graded Rhythmites:

These occur as very thin beds of coarse silt to medium sand grade and are of rather uniform thickness (2 to 10mm). Beds are laterally persistent in the order of metres, beds being separate by mudstones of varying silt content, which are up to 25mm in thickness. The graded rhythmites comprise between 25 and 75% of this type of heterolithic unit in the east Lynmouth beach section. Graded rhythmites are illustrated in plate 2.7C.

The bases of the graded rhythmites are sharp, normally planar, although they locally fill irregularities and biogenic structures. The tops of the units are planar and diffuse. Internally the beds fine upwards, although grain-size variation within single beds rarely exceeds one phi unit. Occasionally, faint horizontal laminations may be discerned within the sandstone bed.

Horizontally-laminated Sandstone Streaks:

This sandstone layer type is more frequent in the 'basal mega-facies' than elsewhere within the Lynton Formation, with the possible exception of the west Crock Point sequence. Individual streaks have planar

bases and tops and range in thickness from 1 to 20mm; some streaks pinch-out laterally. Units containing horizontally-laminated sandstone streaks are particularly common above 16.3m on the Wringcliff Bay log where they are associated with a trace fossil assemblage containing *Chondrites* sp. a, *Teichichnus rectus* and *Bergaueria* sp.

Unconnected Lenticular Bedding:

This bedding type comprises isolated sandstone lenses, 2 to 15mm in height and 7 to 30mm in length, set within mudstone. Individual lenses exhibit a lamination style characteristic of wave and wave-current (combined flow) origin. The paucity of bedding plane surfaces meant that crestline trends were difficult to measure.

Connected Lenticular Bedding:

Connected lenses generally have a pinch-and-swell morphology, although planar-based sandstone layers also occur. Sandstone layers range from 2 to 20mm in thickness; rippled upper surfaces have wavelengths between 5 and 10cm and amplitudes of 5 to 10mm. Individual layers exhibit a lamination style characteristic of wave and wave-current (combined flow) origin e.g. at 14.2m on the Wringcliff Bay log ripples trending 030-210° had a set trending 070-250° superimposed on them. Again, the lack of bedding plane surfaces made crestline trend measurement difficult, although a trend of 092-272° was measured at 4.35m on the Wringcliff Bay log. More occasionally, ripples are of a unidirectional current origin (plate 2.7C).

~

As noted above and in discussed in section 2.2.1, the oldest 'basal mega-facies' deposits (near Black Rocks and at Yellow Stone) are characterised by extensive soft sediment deformation features, particularly intraformational slide sheets. The slide sheets are from 2 to 200cm in thickness and comprise a basal planar décollement overlain by folded, frequently recumbent, horizontally shortened layers of heterolithic bedding (plate 2.4B). The top of the deformed zone is generally planar as a result of penecontemporaneous erosion (plate 2.4B & C). In some cases rotational slump scars and intraformational microfaults are visible representing syn-sedimentary zones of extension created during the sliding events. Thus, the sequence

contains closely adjacent zones of both intraformational compression and extension conformably bounded by facies 1 heterolithic units. The slide sheets are associated with an unusually large number of intraformational scours that have been conformably filled with facies 1 heterolithic deposits (plate 3.2A, B); the scours and their fills are frequently truncated by planar, laterally extensive erosion surfaces (plate 2.4C). Finally, there is much evidence to indicate that the oldest 'basal mega-facies' deposits were subjected to more frequent episodes of soft sediment deformation than was present in other levels within the Lynton Formation. For example, lateral flowage of sediments with a high pore-water content, flame structures where an underlying mud was injected into overlying sandstone layers due to density loading (plate 2.4A) are common. Soft sediment deformation structures and scours are much less frequent in the Wringcliff Bay section, e.g. a 23cm wide, 3cm deep scour concordantly filled with sandstone laminae occurs at 6.05m on the Wringcliff Bay log.

3.2.1.2 Biofacies

Facies 1 contains a diverse ichnofauna dominated by a gradation between *Palaeophycus tubularis* (plate B.14D) and *Chondrites* sp. a (see appendix B for a discussion of this phenomenon). In summary, straight *P. tubularis* burrows grade into *P. tubularis* burrows with occasional unequal-dichotomous branching in some sequences. The *P. tubularis* burrows, both branched and unbranched, are aligned parallel to the mean palaeo-current direction i.e. NNW-SSE (sub-oblique to the dip direction of the SSW-dipping palaeoslope) e.g. 53 and 83cm above the base of the Wringcliff Bay log. The burrow entrances faced into a semi-permanent, obliquely-offshore flowing, geostrophic current that has been interpreted to have existed during the deposition of the lower 2/3rds of the Lynton Formation (see section 4.2.2.3 and chapter 5). In turn, the obliquely-offshore-dipping branched burrows grade into radially disposed branching systems attributable to *Chondrites* sp. a (plate B.3A - located 2.25m above the base of the Wringcliff Bay log). Occasionally, *Chondrites* sp. a burrows are associated with the much smaller diameter burrows of *Chondrites* sp. b (plate B.5B). Finally, *Chondrites* sp. a burrows become excluded and only the small *Chondrites* sp. b burrows remain. Significantly, occurrences of *Chondrites* are associated with the horizontally-laminated sandstone streak sub-facies of facies 1 (the top metre of the Wringcliff Bay section) in an environment where the substrate was slightly below storm wave-base. This gradation from straight *P. tubularis* burrows to *Chondrites* sp. a and then a solitary assemblage of *Chondrites* sp. b burrows has been interpreted (see appendix B - discussion of *Chondrites*) as indicating a progressive decrease in the oxygen content of the sediment below a surficial oxygenated layer (see model shown in text-figure 5.2). Thus, much of the 'basal

mega-facies' (the portion containing *Chondrites*) had a dysaerobic zone of sediment below the sediment/water interface.

Teichichnus rectus occurs in more distal facies 1 deposits (graded rhythmite and horizontally-laminated sandstone streak sub-facies) where it is associated with *Chondrites* sp. a and b, along with *Bergaueria* sp. burrows (plate B.2A, B & C) in the upper part of the section. Occasional specimens of *Aulichnites* sp. on the top of sandstone layers, *Rosselia socialis* (plate B.21D) and 'mantled' tubes also occur. A single specimen of *Megagraption aequale* was observed in the Wringcliff Bay section whilst an isolated specimen of *Arenicolites* sp. b was observed in a muddy sandstone in the east Lynmouth beach section.

The proportion of biogenic 'churning' (fossitextura deformativa), resulting in the loss of primary physical sedimentary structures, can vary within facies 1 units. In particular, the proportion of biogenic 'churning' frequently increases upwards within a unit e.g. 1.3m on the Wringcliff Bay log. Some units have been totally destratified by biogenic 'churning' e.g. 1.1m on the Wringcliff Bay log.

The sequence conforms to Schäfer's (1972) vital-lipostrate biofacies, and vital-pantostrate in the more distal parts of the section indicated by a higher proportion of graded rhythmites and horizontally-laminated sandstone streaks. Occasional horizons of laminated mudstone contain no evidence of biogenic reworking and have been assigned to Schäfer's letal-pantostrate biofacies.

3.2.1.3 Interpretation

Facies 1 closely matches facies 1 in the 'lower-middle mega-facies' and the following interpretations are based on the detailed discussion of the 'lower-middle mega-facies' facies 1 presented in section 4.2.1. In summary, the graded rhythmites are believed to represent silt and sand settling-out beneath storm wave-base from storm-generated suspension clouds and is regarded as the most distal sub-facies within facies 1 (see text-figure 4.8). The occasional presence of indistinct horizontal laminae within the graded rhythmite layers probably reflects the influence of wave orbitals on settling events from a turbulent suspension cloud. The horizontally-laminated sandstone streak facies was deposited in a slightly shallower environment immediately below storm wave-base, the laminae being the product of 'pulsating' fall-out from suspension under oscillatory flow. The unconnected and connected lenticular sub-facies represent the migration of

ripples below fair-weather wave-base during periods of increased wave / combined flow activity, interrupting appreciable periods of mud deposition below fair-weather wave-base. The unconnected sub-facies was deposited at similar depths to the connected sub-facies but was starved of sand, hence the discontinuous nature of the lenses. The laterally extensive planar erosion surfaces preserve the passage of high-energy events creating a zone of net erosion, whilst units exhibiting a thin coarsening-upwards cap of wave / combined flow interference ripples represent the winnowing of fines by slightly lower energy events which were still of sufficient energy to winnow and ripple the substrate.

The trace fossil assemblages are interpreted as being attributable to Seilacher's (1967a) *Cruziana* ichnofacies, but based on comparisons with the west Crock Point succession which has a very similar set of trace fossils (see enclosure 4), may contain some zones which are transitional to Seilacher's *Zoophycos* ichnofacies.

The significance of the intraformational slide units and soft sediment deformation features in the oldest 'basal mega-facies' deposits is discussed in section 3.3.

3.2.2 Facies 2 - Thinly-bedded Sheet Sandstones

3.2.2.1 Description

Facies 2 comprises 10.7% of the Wringcliff Bay section where units have a mean thickness of 1.8cm ($n = 72$, $\sigma_{n-1} = 1.1$ cm, min. thickness = 0.5cm, max. thickness = 7cm). The thin-bedded sheet sandstones, very fine to medium sand in grade, are generally enclosed in connected and unconnected lenticular bedding (facies 1). Some beds, e.g. 21cm above the base of the Wringcliff bay log, contain finely comminuted shell debris, crinoid ossicles and bryozoan fragments. Beds have planar to irregular bases which occasionally scour to a depth of several centimetres into the top of the underlying unit e.g. the bed figured in plate 3.2B infills a scour with laminae that lie sub-parallel to the scour surface and thicken into the scour. Elsewhere, bed bases overlying burrowed mudstones indicate that the sand was 'piped' down into the underlying open burrow systems.

The bulk of the beds comprise a thick zone of parallel to gently undulatory laminae internally, the undulations having wavelengths of several decimetres and an amplitude of several centimetres; undulatory

laminae can meet both the bed base and bed top asymptotically (plate 5.3A). Beds are frequently parallel-laminated throughout their thickness e.g. the beds occurring between 7.1 and 7.3m on the Wringcliff Bay log. The uppermost part of some beds comprises a complexly interwoven pattern of cross-lamination diagnostic of an oscillatory/combined-flow origin; the cross-lamination is generally form-discordant, although some sets of form concordant laminae were observed (8.43m on the Wringcliff Bay log). In some cases the entire unit is wave-ripple cross laminated (e.g. 9.4m on the Wringcliff Bay log), with up to 3 sets of cross-lamination occurring in a single bed.

Bed tops are generally planar (e.g. 1.07m above the base of the Wringcliff Bay log) or 'hummocky'; in some cases a hummocky upper surface passes laterally into a wave-rippled top e.g. 3.8m on the Wringcliff Bay log. The hummocks have wavelengths of between 15 and 120cm and amplitudes of up to 4cm. Occasionally the bed top is rippled by either wave, combined-flow or interference ripples and in some cases an isolated sigmoidal sandstone lens offshoot occurs above the ripple crests (see text-figure 4.7 and accompanying discussion in chapter 4 for a discussion on the origin of these features). Ripple crests were difficult to measure due to the nature of the outcrop, but a set trending 000 - 180° was observed 3.92m above the base of the Wringcliff Bay log. Beds are generally laterally persistent over the width of the outcrop, although where the beds are 'hummocky' the bed can pinch-out laterally resulting in a series of isolated plano-convex lenses.

3.2.2.2 Biofacies

Bed tops are frequently penetrated by indistinct biogenic structures, although distinct burrows of *Bergaueria* sp. are common in the upper part of the Wringcliff Bay sequence. Epirelief trails attributable to *Aulichnites* sp. occur on some bed tops when viewed in plan. More occasionally, beds are extensively disturbed by biogenic 'churning', which in some cases extended down to the base of the unit and resulted in the total loss of primary physical lamination. Facies 2 is assigned to Schäfer's vital-lipostrate biofacies.

3.2.2.3 Interpretation

Facies 2 matches facies 3 described in chapter 7 (section 7.5), to which the reader is directed for a more detailed discussion and interpretation. In summary, facies 2 is the product of the combined action of storm waves and associated obliquely-offshore-directed currents generated by on-shore directed winds. Initial storm conditions resulted in erosion, followed by net deposition during steady flow conditions at the storm

peak. Finally, the top of the bed may have been wave-current rippled as the storm waned. The top of the bed was subsequently reworked by bioturbation and/or oscillatory currents. The large number of planar bed tops suggests that the beds were deposited by storm-generated bottom currents in a zone below storm-wave base.

3.2.3 Facies 3 - Coquinas

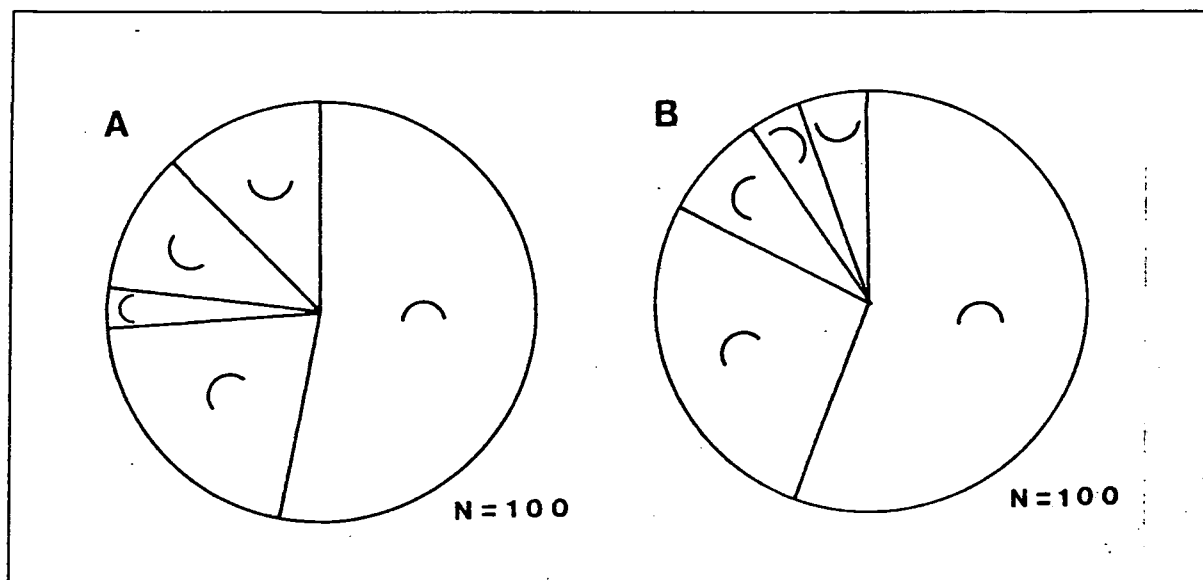
3.2.3.1 Description

Crinoid coquinas, although infrequent elsewhere in the Lynton Formation, are relatively abundant in the oldest deposits preserved in the Lynmouth beach section. The crinoidal coquinas are best observed in the walls of the sea caves at 729 496. The coquinas occur in thick mudstone units of variable silt content where rippled and laminated sandstones are only preserved rarely. Individual crinoidal coquinoid units have a thickness ranging from 1 to 7 cm and are composed almost entirely of crinoid ossicles. Units are either laterally persistent with a tendency to pinch-and-swell in thickness, or form isolated lenses 20 to 50cm in width. Although crinoidal debris is generally concentrated in the coquinas, crinoid ossicles are also liberally disseminated through units of the host mudstone in the oldest 'basal mega-facies' deposits. The crinoid coquinas have a grain supported fabric, comprising individual ossicles of an average 3 to 8mm. diameter, the ossicles being completely disarticulated and poorly sorted. Many of the ossicles have been partially dissolved by the strong pressure solution cleavage which commonly affects these deposits (plate 3.2C).

In addition to crinoid coquinas, shelly coquinas are also unusually common in the 'basal mega-facies' (plate 3.3A). The coquinas comprise concentrated zones of mainly disarticulated shells set within predominantly argillaceous units containing occasional single valves. The lowest unit exposed at Ruddy Ball, immediately underlying an 8m thick sequence of lenticular bedding, exhibits a 2m thick (minimum; base not exposed) unit of silty-mudstone containing disarticulated bivalves, rare brachiopods and occasional crinoid columnals (plate 3.1B). This horizon was sampled during the B.G.S. mapping of the Ilfracombe sheet; the resulting faunal assemblage is shown in column K of table 1 of the Ilfracombe sheet memoir (Edmonds *et al.* 1985). The modes of life of the column K bivalve fauna (see table B.1), indicates that the Ruddy Ball assemblage comprises epi-byssate and shallow infaunal suspension-feeders and some infaunal deposit-feeders.

The vertical, wave-polished surface figured in plate 3.1B was selected for the detailed measurement of the percentage of articulated versus disarticulated bivalve shells and their orientations. The results are presented

in text-figure 3.3B, where the sample is shown alongside a sample from a similar facies observed in the west Crock Point section (text-figure 3.3A). In the Ruddy Ball sample only 10% of the shells were articulated, the articulated forms occurring in less disturbed zones where the shell density was lower than elsewhere, indicating that the shells had not been concentrated by penecontemporaneous winnowing to the extent of the more concentrated zones. Of the articulated forms, none appeared to be in life position.



Text-fig. 3.3 Pie chart showing the orientation of bivalve and brachiopod shells.

Measurements made on exposed vertical faces through the deposits. **Location A:** West Crock Point (6876 4941) - shown at 31m on West Crock Point Log (enclosure 8A). **Location B:** Ruddy Ball (7137 5005). See text for discussion.

The two samples figured above were measured in shell accumulations exposed in units with similar sedimentological context. The resultant orientation classes for the two samples (five orientation classes following those proposed by Salazar-Jimenez *et al.* 1982) are closely similar, and for this reason, are interpreted together.

In the 2m thick shell accumulation at Ruddy Ball, calcareous concretions, ranging from 2cm to 10cm in diameter and having a rounded outline, occur occasionally (plate 3.1C). The calcite concretions have diffuse margins and valves can be observed to pass uninterrupted from the host rock into the concretion. Similar concretions have been observed in muddy deposits in Woody Bay (6794 4891). The concretions are interpreted as early diagenetic in origin. Similar concretions were recorded in the shell accumulations described by Bouma *et al.* (1982).

3.2.3.2 Biofacies

Although the shells and crinoids have been disarticulated the shell valves are generally intact and were probably not transported a great distance before reaching their final resting place i.e. they are para-autochthonous; the crinoid ossicles, however, were probably swept in from another environment (see below) i.e. they are allochthonous. The para-autochthonous nature of the shelly material, when taken together with evidence for winnowing causing a concentration of valves at certain levels, suggests that facies 3 should be assigned to Schäfer's vital-lipostrate biofacies.

3.2.3.3 Interpretation

Although the coquinas occurring within the 'basal mega-facies' may be sub-divided on the basis of their biogenic content, i.e. crinoidal and shelly, their similarities in terms of structure and aspect suggest a common physical origin. The coquinas are interpreted as representing the post-mortem accumulation of biogenic material, zones of biogenic clast concentration swept together during periods of increased winnowing (storms?)

Although bivalve and brachiopod valves are very occasionally articulated, all the crinoid ossicles are disarticulated. This reflects rapid post-mortem disarticulation that is characteristic of crinoids in areas of persistent wave and current activity (Anderson 1968, Brower 1973, Brower & Veinus 1974). There is an apparent absence of suitable areas for crinoid colonisation in the environments preserved by the lower Lynton Formation, a situation analogous to the presence of crinoidal debris in the *Diplocraterion yoyo* facies of the Devonian Baggy Sandstones of north Devon (Goldring 1971). Goldring suggested that the crinoidal debris of the *D. yoyo* facies may have been derived from stratigraphic 'highs'; it is possible that the foot-wall fault block of the Lynmouth - East Lyn Fault (see text figure 2.8), may have provided the requisite 'high'.

The thick shelly developments shown in text-figure 3.3 are interpreted together as the bivalve orientation are similar. In both samples over half of the valves are horizontal convex-up, a further quarter being inclined convex-up. Flume and field observations by many authors (see references in: Salazar-Jimenez *et al.* 1982) have shown that a convex-up attitude is hydrodynamically stable. It is hereby suggested that the Lynton Formation samples represent a post-mortem accumulation of valves on a muddy substrate, the valves resting in a hydrodynamically stable position. At certain horizons within the shell accumulations, concentrated zones

of valves occur with virtually all the valves convex-up. These concentrated horizons are interpreted as representing periods of increased winnowing of the substrate (storms?). The remaining classes on the pie charts represent vertical and concave-up shells. In muddy environments, valves may be overturned by predators, scavengers and bioturbation, in conjunction with the life and death history of the valves (Clifton 1971). In detail, the pie charts shown in text-fig. 3.3 closely correspond to the histograms for valve orientation in lightly bioturbated sediments figured in Salazar-Jimenez *et al.* (1982, figure 5). Salazar-Jimenez *op. cit.* found that the proportion of horizontal and inclined convex-up valves decreases, and the proportion of vertical and inclined concave-up shells increases, with increasing bioturbation, the proportion of horizontal concave-up valves first increasing, then decreasing. The Lynton Formation accumulations are, therefore, interpreted to be a post-mortem accumulation of shells, with zones of valve concentration representing periods of increased winnowing (storms?), that have subsequently been bioturbated. The presence of articulated valves suggests that the accumulations have not suffered a significant degree of post-mortem transport i.e. they represent a para-autochthonous assemblage.

Bouma *et al.* (1982) described and figured (figures 45 - 47) shell accumulations similar to those observed in the Lynton Formation, from the muddy shelf deposits of the Miocene Narrow Cape Formation of Kodiak Island, Alaska. Bouma *et al.* interpreted the shell accumulations, consisting of mainly convex-up valves dispersed throughout units or concentrated in bands, to be the product of storm winnowing.

3.2.4 Facies 4 - Flaser Bedding

3.2.4.1 Description

Facies 5 comprises 9.5% of the Wringcliff Bay section where units have a mean thickness of 9.4cm ($n = 11$, $\sigma_{n-1} = 5.9$ cm, min. thickness = 3.5cm, max. thickness = 23cm); the majority of occurrences of this facies are restricted to a zone between 14.75 and 16.3m above the base of the Wringcliff Bay log. Facies 4 has been observed throughout the thickness of the 'basal mega-facies', but tends to be more common in the upper half of the mega-facies.

Units of facies 5 comprise a regular alternation of 1 to 3cm thick zones of very fine to fine grade sandstone with a lamination style characteristic of wave / combined flow / interference rippling separated by thin mudstone layers. The majority of the pinch-and-swell sandstone layers are continuous, being separated by 1

to 5mm thick mudstone laminae which either maintain a constant thickness over both ripple crests and troughs or thicken into troughs ('wavy flaser' bedding referred to in section 4.2.6.1); isolated mudstone flasers occur only rarely. The sandstone:mudstone ratio tends to remain relatively constant through the thickness of individual beds. Individual sandstone layers have bases which are either planar or preserve the rippled profile at the top of the underlying bed i.e. a massive/structureless sandstone basal layer; bed tops are either rippled or erosively truncated by the overlying unit.

3.2.4.2 Biofacies

This facies is occasionally disturbed by the effects of biogenic 'churning' but distinct burrows (fossitextura figurativa) are more common. *Bergaueria* sp. and *Palaeophycus tubularis* burrows commonly penetrate facies 4 in the upper part of the Wringcliff Bay sequence and epirelief trails attributable to *Aulichnites* sp. occur on some rippled bed tops when viewed in plan. Facies 4 is assigned to Schäfer's vital-lipostrate biofacies.

3.2.4.3 Interpretation

The genesis and hydrodynamic significance of flaser bedding is discussed in section 4.2.7.3, to which the reader is referred for a detailed discussion. In summary, facies 4 represents an environment where conditions allowing the preservation of sand were more favourable than those allowing the preservation of mud. Relatively frequent periods of increased wave energy would have resulted in the mobilisation of the rippled substrate and the winnowing of mud. The lack of intraformational mudstone chips in the sandstone layers suggests that a significant period elapsed between periods of ripple mobilisation, allowing consolidation of the basal mud layer which would have effectively 'armoured' the underlying rippled surface (Hawley 1982).

3.2.5 Facies 5 - Bone-Bed

3.2.5.1 Description

It has long been recognised that, in contrast with the remainder of the Lynton Formation sequence, bone fragments are present in the Lynmouth beach sequence. For example, Hamling (1908) observed that although identifiable specimens were scarce, *Pteraspis* remains could be identified.

During the course of the present study bone fragments were observed in several exposures on east Lynmouth beach including a single, thin 'bone-bed' (*sensu* Reif 1982) i.e. “. . . a sediment which is enriched in highly fractured and abraded vertebrate bones. Very often the bone fraction is well-sorted with grain sizes of fine to coarse gravel” (p.299). The bone-bed occurs in an intertidal platform, the Black Rocks (7291 4963), forming a shallow recess in the smooth wave-washed slabs. The unit has a sharp, planar erosive base with occasional shallow scours into the underlying silty-mudstone. Although the upper surface of the unit passes gradationally into the overlying silty-mudstone, the transition is relatively abrupt. The unit is laterally persistent across the exposure, varying in thickness from 6 to 8 cm.

Internally, the bone-bed comprises a very poorly sorted mixture of granule to pebble grade clasts which are matrix supported in a silty-mudstone; a weak normal grading is discernible locally. Clasts comprise the following: disarticulated bivalves and crinoid ossicles, well rounded vein quartz granules, intraclasts of mudstone and limestone of pebble grade, phosphatic nodules and abraded and rounded bone fragments, some porous (plate 3.3B).

In addition to the bone-bed, isolated bone fragments are occasionally observed within muddy heterolithic units (facies 1). In particular, a 12cm by 10cm 'arrow-head' shaped fragment with a faint internally-laminated structure was observed in a large boulder at 7291 4963 (plate 3.3C). This specimen was tentatively identified as being part of a large Arthrodire - possibly part of the nuchal plate - by Prof. D. Dineley (University of Bristol). The size of the fragment suggests a fish in the order of 1m in length, a size unknown for this order of placoderm fishes in the Early Devonian; placoderms did not attain similar dimensions until the Mid and Late Devonian in Europe and North America (Prof. D. Dineley - pers. comm.)

3.2.5.2 Biofacies

There is an apparent absence of ichnofossils penetrating facies 5. This factor, taken together with the scoured base of the bed, indicates that facies 5 should be referred to Schäfer's letal-lipostrate biofacies.

3.2.5.3 Interpretation

The association between phosphatic nodules and the accumulation of fish remains is common at sites of apatite formation at present (SW African continental shelf) and in the past e.g. Dartmouth Group (Humphreys & Smith 1989). Phosphatic bone fragments (apatite) of Devonian age tend to be restricted to thin pebbly horizons and bone-beds, frequently associated with disconformity surfaces (Humphreys & Smith

op. cit.) and frequently result from winnowing occurring at the peak of basin-wide transgressive events which supply fresh phosphorous which is then utilised by organisms. Examples include: the Ludlow Bone Bed where Old Red Sandstone deposition commenced with shallow marine, wave influenced shelf sand shoals during of the deposition of the Downton Castle Sandstone and intertidal Temeside Shales during Pridoli times (Allen 1985); the 'Coprolite Bed' of the Lower Cretaceous Speeton Clay in Yorkshire (Scott *et al.* 1987). Humphreys and Smith (*op. cit.*) recorded examples from the Dartmouth Group, Bovisand Formation ('Meadfoot facies' of the Meadfoot Group) and Jennycliff Slate Formation (Plymouth Group) along the eastern side of Plymouth Sound - see section 1.6.2.2.2.

The laterally extensive, 6 to 8 cm thick, erosively based bone-bed with a diffuse top fits very closely with similar deposits described by Reif (1982) from the Middle Triassic of SW Germany. Reif assigned the Middle Triassic bone-beds to the 'condensation bone-bed' category of Aepler and Reif (1971) and further suggested that the beds had a storm ('tempestite') origin. The well rounded nature of the phosphatic fragments was attributed to the maturity of the deposits, the erosive bases and diffuse bed tops suggesting a storm origin. Reif proposed that the bone-beds represented condensation with a slow sedimentation rate of fine-grained material that was subsequently repeatedly winnowed during storms. The Lynton Formation bone-bed corresponds to this general model.

Reif noted that 'condensation bone-beds' are found predominantly in shallow marine deposits of early transgressive or late regressive origin. When the overall aspect of Lynton bone-bed is considered, an early transgressive 'highstand' origin appears to be appropriate.

Given a transgressive 'highstand' interpretation for the bone-bed, and the relatively high proportion of distal facies 1 graded rhythmites and horizontally-laminated sandstone streaks in the units that enclose the bone bed, it is probable that facies 5 occurs within a zone equating to Seilacher's (1967a) *Zoophycos* ichnofacies. In the absence of well developed trace fossil assemblages in the units adjacent to the bone-bed, however, this interpretation cannot be unequivocally substantiated.

3.3 INTERPRETATION OF THE 'BASAL MEGA-FACIES' SEQUENCE

The suite of primary sedimentary structures and biogenic traces preserved within the 'basal mega-facies' succession preserves a sequence that is transitional between open shelf conditions, with well oxygenated bottom waters, and more restricted conditions, where a dysaerobic environment became widely established. From time-to-time the basin circulation was sufficiently restricted to enable unbioturbated anoxic black muds to develop locally in facies 1.

The above facies pattern is typical of much of the Lynton Formation and that developed on many modern and ancient marine shelves subject to an alternation of open and restricted circulation (Johnson & Baldwin 1986, e.g. northern California shelf: Leithold 1989). The basal mega-facies, however, contains several features uncharacteristic of the remainder of the Lynton Formation:

- the presence of phosphatic material, including fish remains
- a relatively low proportion of arenaceous sediment
- a large amount of syn-sedimentary intraformational slide and slump horizons , along with soft sediment deformation features
- a high volume of crinoid ossicles and shelly material

The exposed base of the Lynton Formation is characterised by a range of features which suggest that the beds were deposited during a transgressive episode: (i) Large amounts of intraformational sliding and slump scars which are typical of shelf deposits on relatively steep continental shelves generated during rapid onlap over continental coastal plain facies (Galloway 1989). (ii) The presence of phosphatic fragments and a thin bone-bed is also typical of transgressions which bring a fresh supply of phosphorous which can subsequently be used by organisms. (iii) The abundance of crinoidal debris at this level, relative to the remainder of the Lynton Formation, suggests that crinoid 'meadows' developed nearby in a zone of relatively low sedimentation rates, possibly on a topographic high formed by the foot-wall to the Lynmouth - East Lyn Fault (see text-figure 2.8). Goldring (1971 p.13) observed that: "The absence of crinoidal debris in the lower Baggy Beds may be related to the regional palaeogeography, the incoming of crinoids possibly coinciding

with a marine transgression northwards. The entry of crinoid debris coincides with a sharp change in the overall facies-pattern.”

In summary, the basal deposits of the Lynton Formation are consistent with highstand deposits which accumulated during the peak of a transgression. Biostratigraphic indicators (see section 1.8.4) are consistent with a late Emsian age for the base of the exposed Lynton Formation and the interpretation of Johnson *et al.* (1985) suggesting that Lynton Formation deposition was initiated by the 1c transgression of their nomenclature (see text-figure 8.4) is tentatively confirmed. The bone-bed can be considered to be coincident with the maximum flooding surface (*sensu* Van Wagoner *et al.* 1988) of the transgressive event.

The above transgressive interpretation is supported by the relative paucity of sand-grade sediment in the oldest deposits of the Lynmouth Beach section, within which the bone-bed occurs, compared to the remainder of the Lynton Formation. Trapping of sand at the shoreline is characteristic of transgressive episodes (Swift 1976, Goldring & Langenstrassen 1979, Cant 1984).

The upper half of the ‘basal mega-facies’, exposed in the Wringcliff Bay succession, indicates a decrease in relative accommodation space over time representing a gradual increase in the sediment accumulation rate on the shelf as sediment built out over the maximum flooding surface.

The occurrence of debris flow deposits interdigitating with the east Lynmouth beach succession (see section 2.2.3) indicates that throughout the period that the ‘basal mega-facies’ was deposited debris flows slid down a fault scarp along the line of the Lynmouth - East Lyn Fault.

4. THE 'LOWER-MIDDLE MEGA-FACIES' OF THE LYNTON FORMATION: SAND-BODY PRESERVATION ON A MUDDY SHELF

4.1 INTRODUCTION

The 'lower-middle mega-facies' of the Lynton Formation evidences a gradual shallowing of relative accommodation space and an increase in sand supply to the palaeo-shelf as compared with the underlying 'basal mega-facies' described in the previous chapter. As with all other contacts between mega-facies described from the Lynton Formation, the contact with the underlying and overlying mega-facies is gradual. The 'lower-middle mega-facies' is considered to commence some 20 to 30m above the base of the exposed Lynton Formation and passes upwards into the 'upper-middle mega-facies' approximately 100m above the base of the exposed Lynton Formation - see enclosure 2.

4.1.1 Localities & Logged Sections

The 'lower-middle mega-facies' is dominated by a thick sequence of finely interbedded mudstones and sandstones punctuated by occasional sandstone horizons, each several metres in thickness. The sandstone horizons are of three broad types:

- (i) 'Lee Stone facies association' - lensoid sandstone-bodies with coarsening-upwards (CU) centres and coarsening-upwards - fining-upwards (CUFU) margins. The sandstones are of fine sand grade and are generally well sorted; individual beds exhibit wavy bedding, flaser bedding, planar and trough cross-bedding plus parallel-lamination. The sandstones within the 'Lee Stone facies association' are notably more calcareous than the beds of very fine to fine grade well sorted sandstone occurring elsewhere within the Lynton Formation. The interpretation of this facies association is given in section 4.3.1.
- (ii) 'Watersmeet lithotype' - Moderately to poorly sorted units arranged in decimetre-scale cross-bedded sets; the sets are frequently bounded by mud draped erosion surfaces. Lithologically, granule-grade clasts are set in a matrix varying between sublithic arenites with a ferroan calcite cement (Lee Stone) to a finely crystalline sandy dolomitized extrasparite at Watersmeet (classification of Folk 1959). Although thick (\leq 4m) developments of the 'Watersmeet lithotype' are restricted to the 'lower-middle mega-facies,' thinner beds have also been observed at several levels in the 'upper proximal mega-facies' at Rugged Jack (707 498) and Castle Rock (7041 4973) in the Valley of Rocks - see section 6.2.8 and enclosure 2 - as well as

the 'lower-middle mega-facies' in a track-side exposure east of Lynmouth (7268 4923 - see also enclosure 2). The interpretation of this lithotype is given in section 4.3.2.

(iii) 'Thick coarsening-upwards parallel-laminated sandstones' - road-side exposures along the A39 in the East Lyn valley expose a thick (>6m) unit of thick-bedded parallel-laminated sandstones within which mud content decreases upwards (winnowed-upwards trend) whilst preservation of biogenic activity decreases upwards. The interpretation of the A39 road-side section is given in section 4.3.3.

In total, nine sections were logged in detail (see enclosure 2) within the five localities (see enclosure 1) considered in this chapter:

4.1.1.1 Watersmeet

At Watersmeet the NW-flowing East Lyn River meets a N-flowing stretch of river formed by the confluence of Hoarok Water and Farley Water 1km to the south. One quarter of a kilometre up-stream of Watersmeet a 4m thick development of the 'Watersmeet lithotype' is exposed on the banks of the East Lyn River where bedding dips c. 10° towards WNW. On the southern side of the river a section was logged through the old quarry face above the lime kiln (7464 4862) - log A on enclosure 7. On the opposite bank of the river (7462 4865) two sections were logged through the river-side slabs and short natural cliffs above - log B on enclosure 7 was taken through the NE end of the exposure, whilst log C was taken through the SW end (down-stream) of the exposure.

In addition to the exposures one quarter of a kilometre up-stream of Watersmeet, palaeo-current measurements from limited exposures through parts of the 'Watersmeet lithotype' were also taken from: the path-side exposures on the ^-shaped promontory of land formed between the confluence of the two rivers at Watersmeet i.e. on the opposite side of the river, directly facing Watersmeet House (7440 4864), where bedding dips 10° towards 186°; river-side bluffs on the west bank of East Lyn River 100m north (down-stream) of Watersmeet (7444 4878) where bedding dips c. 10° towards the north.

Exposures of intraformational conglomerate (described in section 2.2.3 where they were attributed to a submarine debris flow origin) have been recorded from near the line of the Lynmouth - East Lyn Fault at

Myrtleberry (7417 4901) and north of the old limestone quarry at Watersmeet (7473 4872) by Prof. S. Simpson (unpublished notes - referenced in Edmonds *et al.* 1985), although an extensive reconnaissance during the present study failed to reveal these exposures. These occurrences suggest that debris flows continued to be shed from the line of the fault until at least the close of deposition of the 'lower-middle mega-facies' hereabouts - see enclosure 2.

4.1.1.2 East Lynmouth Beach

The section exposed between Lynmouth and Ninney Well (7261 4962 to 7348 4957), where the Lynmouth - East Lyn Fault has juxtaposed the Lynton Formation against the younger Hangman Sandstone Group, comprises vertical cliffs passing upwards into grassed slopes. This section is probably the most accessible of the coastal exposures of the Lynton Formation. The cliff base is protected by large boulders and *in situ* wave-washed slabs; small patches of sand occasionally lie between the boulders. The coast east of the small headland at 7311 4963 is accessible for 1½ hours either side of low tide. The sequence has been described by Tunbridge & Whittaker (in: Goldring *et al.* 1978).

As described in section 2.2.1, the Lynton Formation sequence that crops out adjacent to the Lynmouth - East Lyn Fault along the coastal section east of Lynmouth is deformed by a major syncline of syn-sedimentary origin (D0), with associated soft sediment deformation structures (see chapter 3). Superimposed upon the sedimentary deformation are structures of D1 tectonic origin relating to the Variscan deformation of north Devon. The tectonic deformation comprises minor folds, overturned towards the north, and an axial planar pressure solution cleavage (intensively developed locally) dipping southwards, associated with the Exmoor anticline, along with fold-thrust structures (see section 2.3).

The complex syn-sedimentary and Variscan deformation of the section exposed between Lynmouth and Ninney Well has resulted in deposits of both 'basal' and 'lower-middle mega-facies' being preserved in relatively close proximity. A log through a sequence typical of the 'basal mega-facies' in this section is shown in text-figure 3.1A and described in section 3.2. At grid reference 7292 4962 a coarsening-upwards - fining-upwards sandstone body of 'Lee Stone facies association' type is preserved at the cliff base in a part of the sequence attributable to the 'lower-middle mega-facies' - shown in text-figure 3.1B and discussed within this chapter.

It should be noted that intraformational conglomerate deposits, described in section 2.2.3 where they were attributed to a submarine debris flow origin, are preserved throughout the exposed thickness of the Lynton Formation sequence immediately adjacent to the Lynmouth- East Lyn Fault at Ninney Well. This indicates that debris flows were being shed from the line of the fault throughout the time that the basal and 'lower-middle mega-facies' were deposited hereabouts.

4.1.1.3 Ruddy Ball

The 1.5km stretch of coast between Wringcliff Bay and Lynmouth (see enclosure 1) comprises north-facing, vertical cliffs exposing (at the level of the cliff base) a section from the lowest visible strata of the Lynton Formation at Yellow Stone (grid reference 7060 4999) through to the 'lower-middle mega-facies' at Ruddy Ball (grid reference 7137 5005). At Ruddy Ball a 50m wide plano-convex lens (0 - 2.5m thick) of convoluted and folded sediments, representing a slump scar and its infill, is exposed high in the cliff face (can only be approached with the use of a rope). The base of the cliffs is protected by large boulders and wave-washed slabs. Ruddy Ball may be reached from Lynmouth Beach, to the east, via an arduous scramble, accessible for 2 hours either side of a low tide

The sequence at Ruddy Ball comprises a 2m thick silty-mudstone containing disarticulated bivalves and rare brachiopod valves (described in section 3.2.3) at the top of the 'basal mega-facies.' The shell bed passes upwards, via 8m of lenticular bedding, into the plano-convex slump scar lens (described in section 2.2.2) near the base of the 'lower-middle mega-facies;' the slump horizon is overlain in turn by: 30 - 40cm of lenticular bedding, 20cm pure mudstone, a 3 - 4m thick coarsening-upwards sequence attributable to the 'Lee Stone facies association.'

Although the 'lower-middle mega-facies' deposits at Ruddy Ball are too inaccessible to be logged, the section is discussed in section 4.3.1 where the interpretation of the 'Lee Stone facies association' is presented.

4.1.1.4 A39 Road Section - East Lyn Valley

A series of road-side exposures along the A39 in the East Lyn Valley, between the quarry at 7307 4908 and 7318 4879, reveal an 11-12m thick coarsening-upwards sequence capped by thick units of parallel-laminated

sandstone (log shown in text-figure 4.1). Bedding dips 06° towards 340° , cleavage dips 65° towards 190° . These exposures have previously been described by Simpson and Kidson (1954) and by Tunbridge & Whittaker (in: Goldring *et al.* 1978); both sets of authors drew attention to the presence of U-shaped tubes, attributed to the trace fossil *Arenicolites cf. subcompressus*, in these exposures.

4.1.1.5 Duty Point

The type locality of the 'Lee Stone facies association' crops out at Lee Stone, Duty Point. The exposure trends east-west and is north-facing. Two sections were logged through the 'Lee Stone facies association' - one at the western end of Lee Stone (6948 4970, text-figure 4.2) and the other at the eastern end of the exposure (6952 4969, text-figure 4.3). The log bases equate approximately with high tide mark, the rocks below being barnacle covered and unsuitable for logging. The logs terminate at the point where the exposure passes into steep, grassy slopes above.

Bedding at Lee Stone dips 14° towards 190° the tectonic cleavage dipping 26° towards 190° . A large amount of bedding-plane slip has occurred, marked by thin horizons of clay gauge. More rarely, slickensides occur which have a lineation which plunges 13° towards 164° . The measurement of extension of 'mantled' tubes (see Appendix C), occurring in the upper Duty Point cliff section, indicates a ratio of 6.5:2.5:1 for the X:Y:Z strain ellipsoid hereabouts.

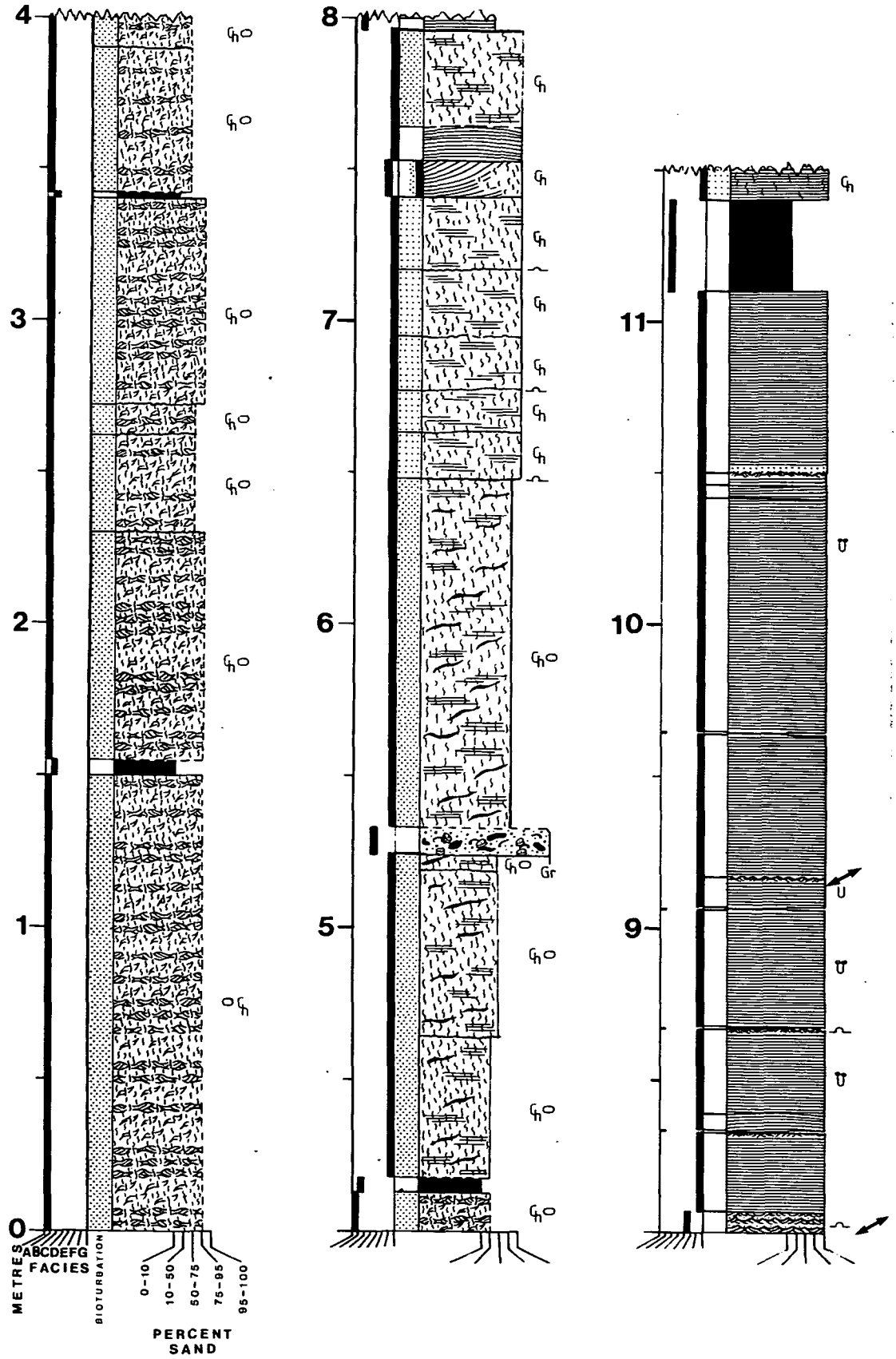
Two additional sections were logged at Duty Point: the exposure above Lee Stone (6953 4963, text-figure 4.4) and Duty Point cliff top (6944 4967, text-figure 4.5), in order that an understanding of the lithologies surrounding the 'Lee Stone facies association' could be gained.

Text-fig. 4.1 Log through road-side exposures along the A39 in the east Lyn Valley.

The log base is located in the quarry at 7307 4908; the log top is located in the face exposed at 7318 4879.

Facies: **A** = thinly interlayered sandstone/mudstone bedding; **B** = graded mudstone; **C** = wavy bedding - not present at this locality; **D** = bioclastic sandstone; **E** = wave cross-laminated sandstone; **F** = isolated trough cross-bedded sandstone set; **G** = thick-bedded parallel-laminated sandstone.

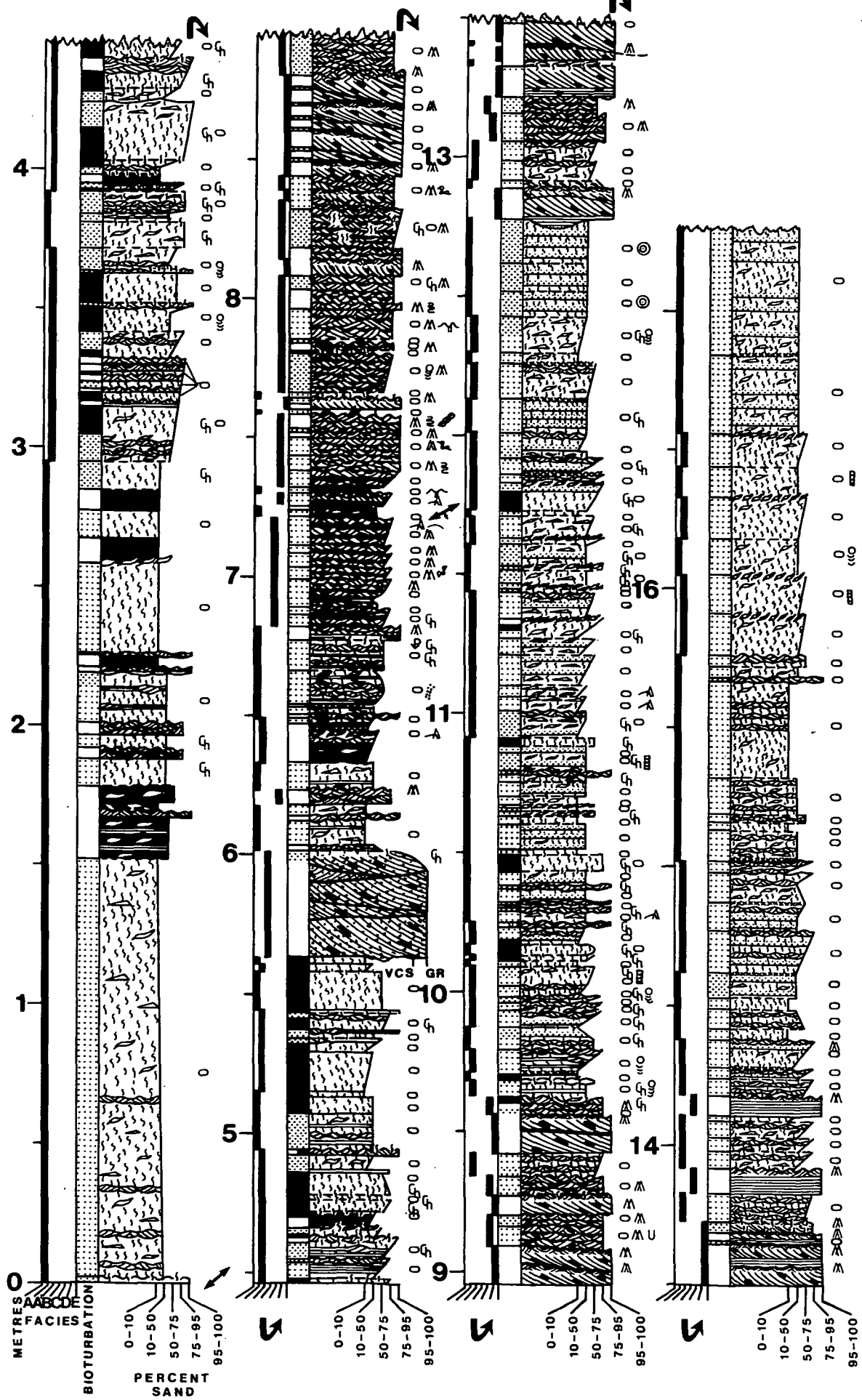
A39 Road Section : East Lyn Valley



Text-fig. 4.2 Western Lee Stone. Log through the 'Lee Stone facies association' and surrounding sequence.

Grid reference = 6948 4970. The 'Lee Stone facies association' is defined as occurring between 7.35m and 9.63m on the log. The 'Watersmeet lithotype' is defined as occurring between 5.64m and 6.01m on the log. Facies: A = thinly interlayered sandstone/mudstone bedding; A' coarsening-upward microsequences; B = cross-bedded granule conglomerate; C = wavy bedding; D = flaser bedding; E = cross-bedded sandstone.

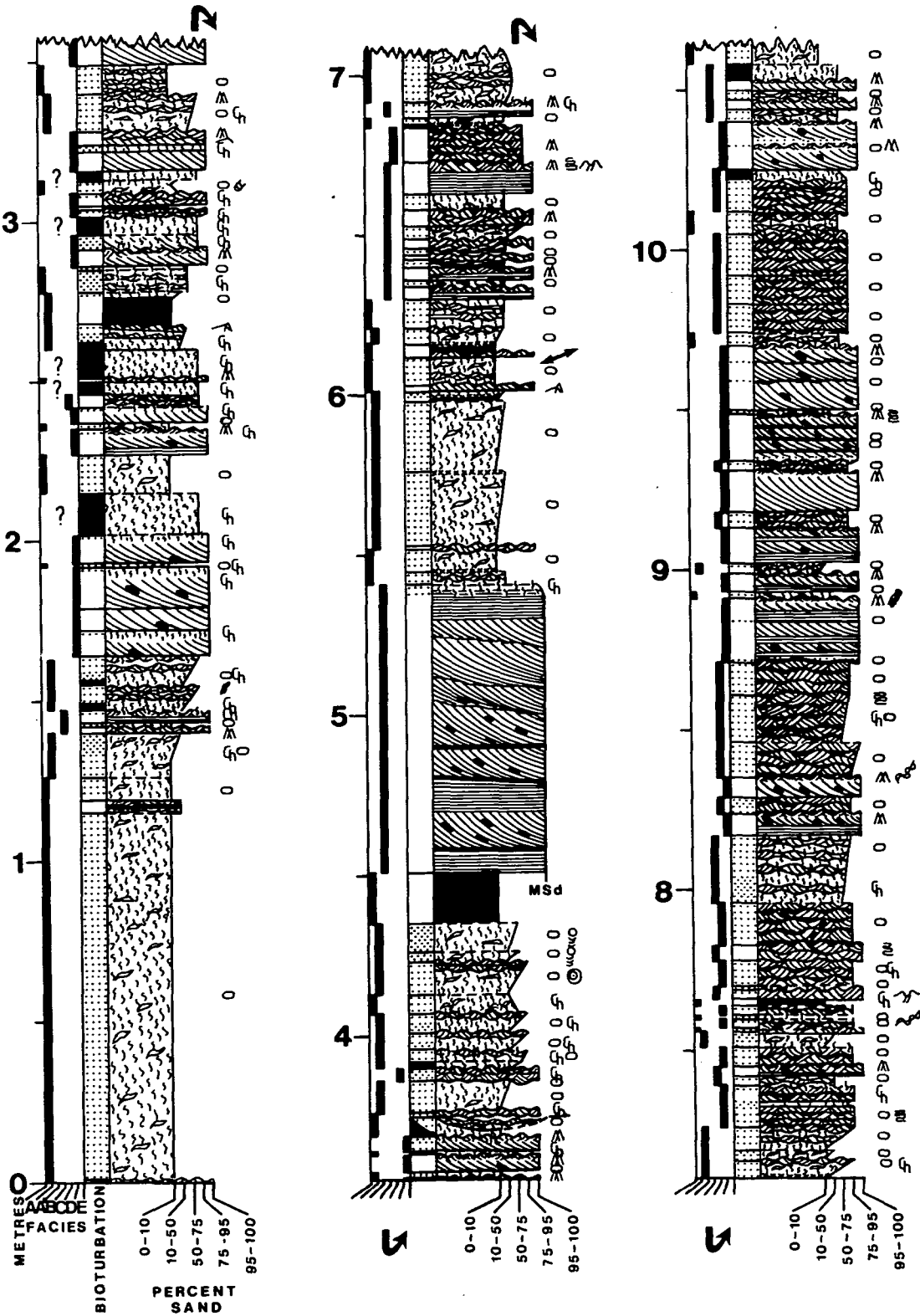
Western Lee Stone



Text-fig. 4.3 Eastern Lee Stone. Log through the 'Lee Stone facies association' and surrounding sequence.

Grid reference = 6952 4969. The 'Lee Stone facies association' is defined as occurring between 7.16m and 10.58m on the log. The 'Watersmeet lithotype' is defined as occurring between 4.52m and 5.41m on the log. Facies: A = thinly interlayered sandstone/mudstone bedding; A' coarsening-upward microsequences; B = cross-bedded granule conglomerate; C = wavy bedding; D = flaser bedding; E = cross-bedded sandstone.

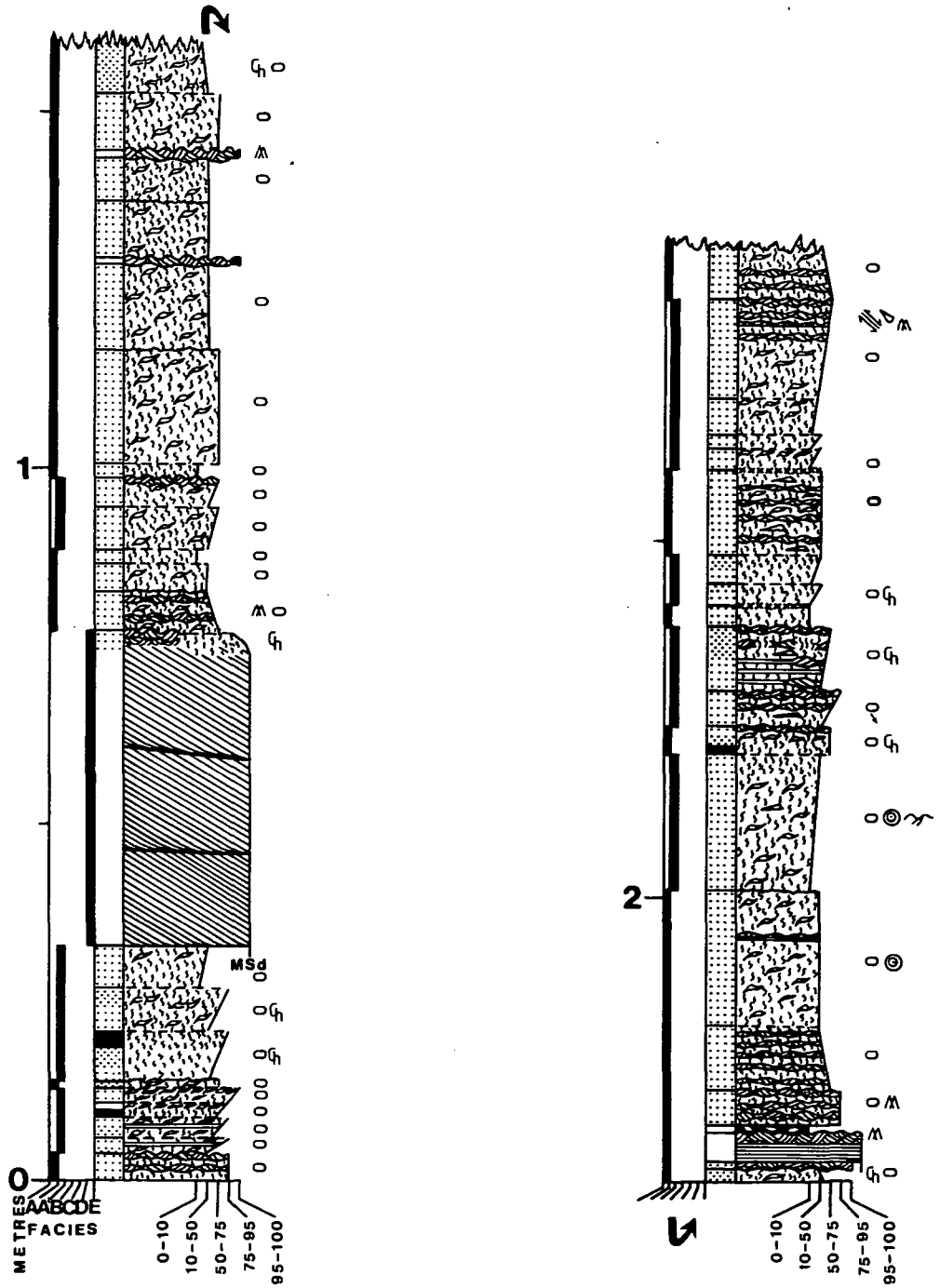
Eastern Lee Stone.



Text-fig. 4.4 Outcrop above Lee Stone. Log through typical heterolithic units within the 'lower-middle mega-facies.'

Grid reference = 6944 4967. Facies: A = thinly interlayered sandstone/mudstone bedding; A' = coarsening-upward microsequences; B = cross-bedded granule conglomerate; C = wavy bedding; D = flaser bedding; E = cross-bedded sandstone.

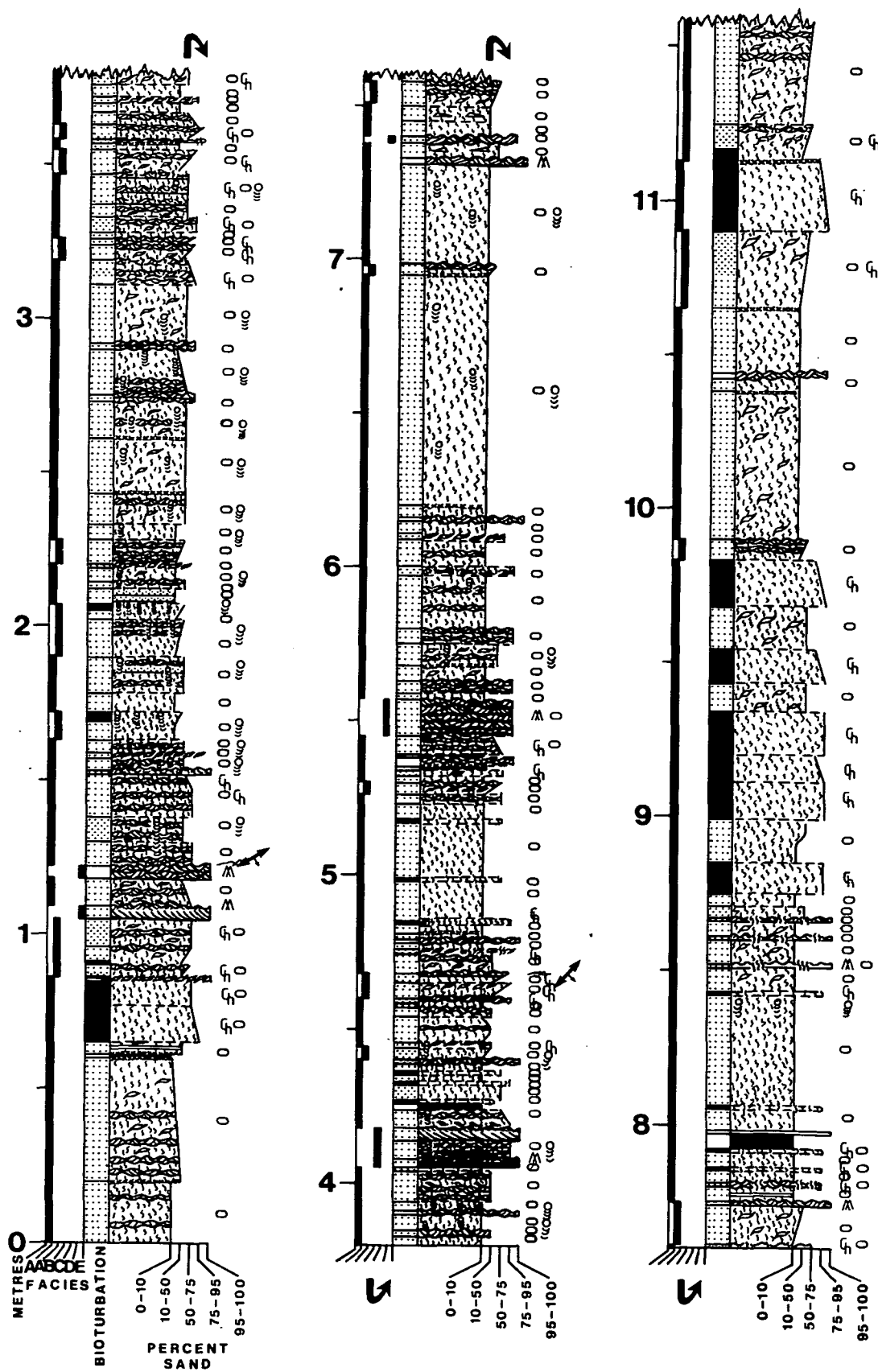
Outcrop Above Lee Stone.



Text-fig. 4.5 Duty Point cliff top. Log through typical heterolithic units within the 'lower-middle mega-facies.'

Grid reference = 6944 4967. Facies: A = thinly interlayered sandstone/mudstone bedding; A' = coarsening-upward microsequences; B = cross-bedded granule conglomerate; C = wavy bedding; D = flaser bedding; E = cross-bedded sandstone.

Duty Point—Cliff Top.



4.1.2 Approach

The past three decades have witnessed a significant rise in the application of facies sequence analysis to palaeoenvironmental reconstruction, a period heralded by the landmark study of cyclic sequences in the Lower Westphalian of north Devon by de Raaf *et al.* (1965). This rise has resulted from a greater understanding of the hydrodynamic significance of bedforms, the growing availability of studies of modern environments and the ever increasing number of studies of ancient sequences. The above factors have allowed the erection, and refinement, of a series of a widely applicable depositional models. Unfortunately, there is an increasing tendency for the description of individual facies to be kept to a minimum, perhaps due to the increasing pressure on journal space, and the interpretation of individual facies to be based upon the ultimate interpretation of the facies sequence as a whole. This approach is exemplified by Walker's (1979) statement in a review article of *Facies and Facies Models*, that: "... many, if not most facies identified in the field have ambiguous interpretations - a cross-bedded sandstone facies, for example, could be formed in a meandering or braided river, a tidal channel, an offshore area dominated by alongshore currents or an open shelf dominated by tidal currents. The key to interpretation is to analyse all of the facies communally, in context. The sequence in which they occur thus contributes as much information as the facies themselves" (p.1).

Two, common approaches have stemmed from the over-zealous application of facies sequence analysis. Firstly, many authors have striven to fit their sequence to a previously described modern and/or ancient analogue. However, such an approach hinders the generation of new, radical environmental models which may be necessary to adequately explain a new or unusual facies sequence. For example, there is a paucity of documented examples of modern shelves where sedimentation is influenced by permanent and semi-permanent current systems. As a result, sedimentologists have tended to ignore the possibility of such processes in moulding ancient sand-bodies. This situation prompted the convenors of a research symposium on the: *Sedimentology of Shelf Sands and Sandstones*, in a series of suggested topics for discussion, to ask the question: "What is the role of permanent and semipermanent current systems on shelves? Are they the forgotten process?" (Knight & McLean 1986, p.x). Although several papers were submitted detailing the rôle of semi-permanent geostrophic currents on modern shelves (Martin & Flemming 1986, Nelson 1986), no paper was received which considered the rôle of semi-permanent currents in the interpretation of an ancient sequence.

The second approach has been to present a hydrodynamic interpretation for a given sedimentary structure, based upon its position within a facies sequence. The application of this approach is well illustrated by the history of interpretation of hummocky cross-stratification. The majority of authors have ascribed a storm-wave origin to hummocky cross-stratification, usually with scant regard for hydrodynamic considerations, this interpretation being based on the frequent occurrence of hummocky cross-stratification below sediments attributed to generation within fair-weather waves, and above sediments deposited below storm wave-base, within a regressive facies sequence. As will be seen in Chapter 7, however, hydrodynamic considerations preclude a purely oscillatory origin for hummocky cross-stratification and, furthermore, it is probable that hummocky cross-stratification is polygenetic in origin.

In particular, the detailed study of the 'Lee Stone facies association' within the following chapter is an attempt to redress the imbalance presented by the above approaches to facies-sequence analysis. The emphasis throughout the chapter will be on a detailed description of each facies and an interpretation of individual facies in terms of the hydrodynamic and ethological significance of its constituent physical and biogenic structures (section 4.2). Only then will an attempt be made to reconcile individual facies within overall facies models (section 4.3). Although, ultimately, the exact environmental position of many structures remains uncertain, the application of techniques such as the recognition of tidal bundles (cf. Boersma 1969), provides a less ambiguous interpretation of individual facies. This in turn has allowed, it is believed, a new, radical and more realistic model to be generated for the 'Lee Stone facies association' than would otherwise have been possible with a less rigorous examination of individual facies.

4.2 FACIES

The following sub-sections provide a detailed description and interpretation of the physical and biogenic structures preserved within the facies recorded on the logs described in section 4.1.1. As a different facies nomenclature was used for each of the logged sections it has been necessary to develop a common scheme for the purpose of description; this is presented in table 4.1.

| Log ↓ | Watersmeet | East Lynmouth Beach | A39 Road Section | Duty Point |
|----------|----------------|------------------------|------------------|------------|
| A | 1 | 1 | 1 | 1 |
| A' | 1 - calcareous | 1' | - | 1' |
| B | 3 | - | 2 | 3 |
| C | 5 | 5 | - | 5 |
| D | - | 6 | 4 | 6 |
| E | - | 8 | 7 | 8 |
| F | - | 10 | 9 | - |
| G | - | - | 11 | - |

Table 4.1 Chapter 4 facies scheme.

Key: **1** = thinly interlayered sandstone/mudstone bedding; **1'** = coarsening-upwards microsequences within facies 1; **2** = graded mudstone; **3** = cross-bedded granule conglomerate; **4** = bioclastic sandstone; **5** = wavy bedding; **6** = flaser bedding; **7** = wave cross-laminated sandstone; **8** = cross-bedded sandstones; **9** = isolated trough cross-bedded sandstone set; **10** = thin-bedded horizontally-laminated sandstone; **11** = thick-bedded parallel-laminated sandstone. **Watersmeet** = logs shown in enclosure 7; **East Lynmouth Beach** = log shown in text-figure 3.1B; **A39 Road Section** = log shown in text-figure 4.1; **Duty Point** = logs shown in text-figures 4.2 - 4.5.

4.2.1 Facies 1 - Thinly Interlayered Sandstone/Mudstone Bedding

Facies 1 is shown as facies A on all nine logs described within this chapter; calcareous horizons of facies 1 at Watersmeet that occur below the 'Watersmeet lithotype' are denoted A' on enclosure 7.

Although present in all of the sections logged and described within this chapter, this facies is best exposed in the sections logged at Duty Point, where abundant wave-washed slabs and clean vertical faces have enabled detailed investigations of the physical and biogenic structures to be recorded. In contrast, at east Lynmouth beach facies 1 is strongly disturbed by tectonic cleavage whilst the exposures at Watersmeet and along the A39 road section are strongly weathered and lichen-covered. Occasional clean surfaces at the latter localities, however, indicate that the physical and biogenic structures preserved in facies 1 in the Watersmeet, east Lynmouth beach and A39 road sections are similar to those exposed in the Duty point sections. *Unless stated otherwise, all observations within section 4.2.1 refer to the sections exposed at Duty Point.*

Facies 1 records the background sedimentation style within which the sandstone-body types of the 'lower-middle mega-facies' are preserved, forming units from 1 to over 150cm in thickness. Unit boundaries on the logs are generally arbitrary, tending to mark abrupt changes in sandstone percentage. More occasionally, units may be separated by planar, laterally extensive (see correlation of facies 1 units at Watersmeet - shown at base of columns B & C on enclosure 7), erosion surfaces with up to 3cm relief; no biogenic structures

cross these surfaces. A clean, rippled sandstone is frequently present beneath the erosion surfaces, representing the winnowing of fines during the erosional event that created the surface.

In essence, this facies is a mudstone facies which contains thin sandstone (well sorted, grade ranges from coarse silt to fine sand, very rarely medium sand) layers in varying amounts. The sandstone content of an individual unit may range from 0% to 95%. Although pure mudstones are rare, e.g. a 40cm thick unbioturbated silty-mudstone near the base of the east Lynmouth beach log (text-figure 3.1B) which only contains unconnected sandstone lenses at certain levels, individual mudstone layers reached a maximum thickness of only 10cm.

At Watersmeet several thicker (10-30cm) units of silty-mudstone, homogenised by biogenic 'churning', contain disarticulated bivalve or brachiopod valves and quartz granules (plate 4.6B). The valves are 5 - 15mm in length and randomly distributed and oriented throughout individual units. Occasionally the valves are concentrated into 2 - 3cm thick matrix-supported zones of uniform thickness (e.g. 39 and 46cm above the base of log B on enclosure 7; plate 4.6C), extending up to 5m laterally. The tops and bottoms of these shell concentrations are gradational, merely marking a rapid increase/decrease in shell content. No evidence was observed of the shells being in life position. The units denoted as facies A' on the Watersmeet logs reacted to dilute hydrochloric acid in the field; acid digestion in the laboratory revealed that these units contain between 35% and 45% carbonate by weight. Thin section examination revealed poorly defined graded laminae consisting of calcite microspar with poor to moderately sorted sub-angular detrital quartz clasts of fine to very fine sand-grade, although larger clasts sporadically occur together with very coarse sand-grade chert, sandstone and quartzite clasts 1 to 2mm in diameter. Articulated crinoid stems, coarse crinoid fragments and coarse brachiopod shell debris are also present. Mica minerals (?biotite) are rare and no feldspar grains were observed. Not all laminae are calcareous - several laminae are of siliceous mudstone / siltstone, some with a cherty appearance. The dominant carbonate microspar is interpreted to be recrystallized primary micrite.

Facies 1 sandstone layers are randomly disposed within individual units, with the exception of sequentially arranged microsequences, particularly coarsening-upwards micro-sequences described separately in section 4.2.2 (facies 1'). Sandstone intercalations are of four broad types and are described individually in the sections below.

4.2.1.1 Facies 1 - Sub-type A - Graded Rhythmites

4.2.1.1.1 Description

The least common of the four sandstone layer types - this sub-type was not observed in the Watersmeet or A39 road sections. Sub-type A consists of 2 to 7mm thick laminae, laterally persistent in the order of decimetres and rather uniform in thickness, composed of coarse siltstone to very fine sandstone. Individual sandstone layers are randomly disposed within units, intervening mudstone layers never exceeding 3cm in thickness. Sandstone:mudstone ratios range from 1:5 to 1:20 within individual units. Multiple layers of bioturbation occur within single units.

The sandstone layer bases are sharp and planar, although they may locally infill irregularities and biogenic structures. Bed tops are diffuse, grading upwards into mudstone. The sandstones are well sorted with the grain size range never exceeding one phi unit within an individual sandstone layer. Faint horizontal-laminations were occasionally observed.

4.2.1.1.2 Interpretation

Ashley (1975) noted that two types of thin-bedded rhythmites may be distinguished: seasonal couplets ('varves') and those produced during single sedimentation events. The random disposition of the sandstone member in the Lynton Formation couplets and the large mudstone:sandstone ratio precludes a seasonal origin. Furthermore, deposition during a single tidal cycle ('tidal bedding' of Reineck & Wunderlich 1968a) may also be ruled out, the muds being too thick ($\geq 3\text{cm}$) to have been deposited during the slack water phase of a single tidal cycle. Additionally, multiple levels of bioturbation within individual mudstone units indicates an appreciable time span for the deposition of the mudstone units.

Examples of graded rhythmites produced by repeated non-cyclic, single sediment pulses have been recorded from a diverse range of environments: lacustrine deltas (Forstner *et al.* 1968); shelf seas: tide-dominated (Reineck & Singh 1972), wave-dominated (de Raaf *et al.* 1977), storm-dominated (Gadow & Reineck 1969); turbidite fans (Piper 1970).

Of the shallow marine occurrences, Reineck and Singh (1972) attributed deposition of graded sand layers to fall-out from turbulent suspension clouds created during storms in still, or slowly moving, waters below

wave-base. De Raaf *et al.* (1977) described graded siltstone and sandstone layers ('facies M1a, silt- and sand-streaked muds'), noting that the siltstone streaks were usually graded, having sharp bases and diffuse tops, whilst the sandstone layers usually did not display grading. A similar relationship was observed in the Lynton Formation examples. De Raaf *et al.* (*op. cit.*) also attributed the graded layers to fall-out from suspension clouds generated during storms.

The Lynton Formation graded rhythmites represent silt and sand settling-out beneath storm wave-base from storm-generated suspension clouds with a sufficient grain size range to produce grading. The presence of an indistinct lamination probably reflects the pulsing influence of wave orbitals in the overlying water column on settling rates from the turbulent suspension cloud.

4.2.1.2 Facies 1 - Sub-type B - Horizontally-laminated Sandstone Streaks

4.2.1.2.1 Description

Of the localities considered within this chapter, this sandstone layer type has only been observed in the Duty Point sections (plate 4.1A). Sub-type B is characterised by sandstone horizons of coarse siltstone to fine sandstone grade which are randomly disposed through each unit, the intervening mudstones never exceeding 3cm in thickness (plate 4.1A). The sandstone:mudstone ratio ranges from 1:5 to 1:20 within individual units. The sandstones have sharp, planar bases and tops, although the base may infill local irregularities and biogenic structures. Horizontal-laminations, although present, are faint due to the low granulometric contrast. Individual sandstones are laterally persistent in the order of metres and are of a rather uniform thickness. Laminae occasionally swing-up into incipient cross-lamination giving a pinch-and-swell appearance to the sandstone layers.

4.2.1.2.2 Interpretation

Thin, horizontally-laminated sand streaks are common in a wide range of sedimentary environments. Marine examples include: storm-sand layers in shelf muds (Gadow & Reineck 1969), delta front sequences (Coleman 1976), continental slopes (Dott & Bird 1979).

Reineck and Singh (1972) described examples of horizontally-laminated sand streaks from the North Sea and experimentally produced similar examples through the sedimentation of suspension clouds in current

velocities below those required to generate a rippled bed. Graded rhythmites were treated as a special case of horizontally-laminated sand in which a sufficient granulometric contrast was available to produce grading. De Raaf *et al.* (1977) described examples from a Lower Carboniferous, wave-dominated sequence in County Cork, Ireland. They noted that the lamination was poorly defined due to a low granulometric contrast, indicating the absence of sand transport by traction, leaving fall-out from suspension as the dominant mechanism of sedimentation.

The Lynton Formation examples are believed to represent fall-out from storm-generated suspension clouds beneath storm wave-base. The presence of incipient cross-lamination indicates the proximity of storm wave-base in some examples.

4.2.1.3 Facies 1 - Sub-type C - Unconnected Lenses

4.2.1.3.1 Description

This bedding style forms one of the two most common sub-types of thinly interlayered sandstone/mudstone bedding, connected lenses (see 4.2.1.4 below) being the other common bedding style; both sub-types are present in all of the logged sections considered within this chapter. Individual units comprise isolated lenses of coarse siltstone to fine sandstone grade and are set within mudstone. The sandstone lenses range from 3 to 10mm in height, 15 to 50mm in length and are randomly disposed within mudstone units, the sandstone lenses forming 5% to 90% of any individual unit. Internally, the sandstone lenses are cross-laminated. A broad range of genetic types are preserved, defined by two end members:

- (i) **Unidirectional Current-Ripples** These are strongly asymmetric; form-sets have planar bases, individual form-sets ranging from 15 to 40mm in length and 3 to 7mm in height. The form-sets never exceed one ripple cross-lamination set in thickness. The crestlines are sharp and straight when viewed in plan, lee-side 'spur-like' protrusions are occasionally visible in plan view. The form-sets are form-concordant and display planar foreset laminae. Palaeocurrents were predominantly directed obliquely palaeo-offshore i.e. towards the SE (plate 4.1A) down a SSW-dipping palaeoslope (see section 2.2.1), although onshore directed ripples may occasionally be observed. The above features are indicative of straight-crested current-ripples.

(ii) **Oscillation Ripples** The lenses are symmetric, or near symmetric in cross-section, ranging from 15 to 50mm in length and 3 to 10mm in height. Individual form-sets are commonly greater than one set in thickness, the set boundaries being curved and disposed at a low-angle. The form-sets are usually form-discordant, although a single form-concordant draping lamina is occasionally present. Form-set bases are irregular or arcuate, the crests being rounded, or very rarely sharp. In plan, the crestlines are straight and commonly bifurcate (plate 4.1B). The foreset laminae are concave-up, frequently dipping at less than the angle of repose and commonly thicken laterally to produce a swollen-lens geometry. Observed ripple asymmetry is not necessarily consistent with the internal structure. Adjacent ripple form-sets within a ripple train may display opposed cross-lamination directions. The crestlines generally have a NE-SW trend. The above features are consistent with those defined by Boersma (in: de Raaf *et al.* 1977) as being diagnostic of wave-generated ripples.

The above represent the two end-members of a broad spectrum, the majority of ripples in facies 1 unconnected lenses appearing to be of a combined-flow origin. At the time of writing, no detailed study is available of the lamination styles diagnostic of ripple generation by combined-flows. The problem is further compounded by the fact that the outcrops examined in this study generally trend east-west, a direction perpendicular to the palaeocurrent and wave-oscillation directions, thus producing a paucity of true ripple cross-sections on the often relatively smooth vertical rock faces. On the logs, ripples of wave or combined-flow origin are shown by the sigmoidal symbol, whilst pure current-ripples have a separate, distinct symbol (see Log Legend - enclosure 3A).

Tanner (1967) empirically derived a series of ripple indices to distinguish between wave- and current-generated ripples. Several large bedding planes are available at Lee Stone which allow the application of Tanner's indices. The results are presented in text-figure 4.6.

Text-fig. 4.6 Characteristics of lenticular bedded combined-flow ripples at Lee Stone.

Location A: Bedding plane at 6946 4970 - connected lenticular bedding (Facies 1, sub-type D).
Location B: Bedding plane at 6948 4970 - isolated lenticular bedding (Facies 1, sub-type C). The upper half of the figure shows data for both localities plotted on Figure 35 of Reineck and Singh (1980 - *pars*). Note that for each of the three indices of Tanner (1967), the mean value for each data set falls nearer the current-produced field in the case of the ripples from locality B. The **lower half of the figure** shows ripple profiles obtained from both localities. Note that the ripples from locality B are more strongly asymmetrical i.e. the unidirectional component of the flow was larger for the ripples at locality B than for the ripples at locality A. This phenomenon is consistent with the ripple indices recorded for each locality. This fact is further evidenced by the ripples at locality B having small lee-side spurs, a feature characteristic of current-ripples. Also note the wave-modified profiles to the ripples at locality B (arrowed).

PI₁ : Parallelism index

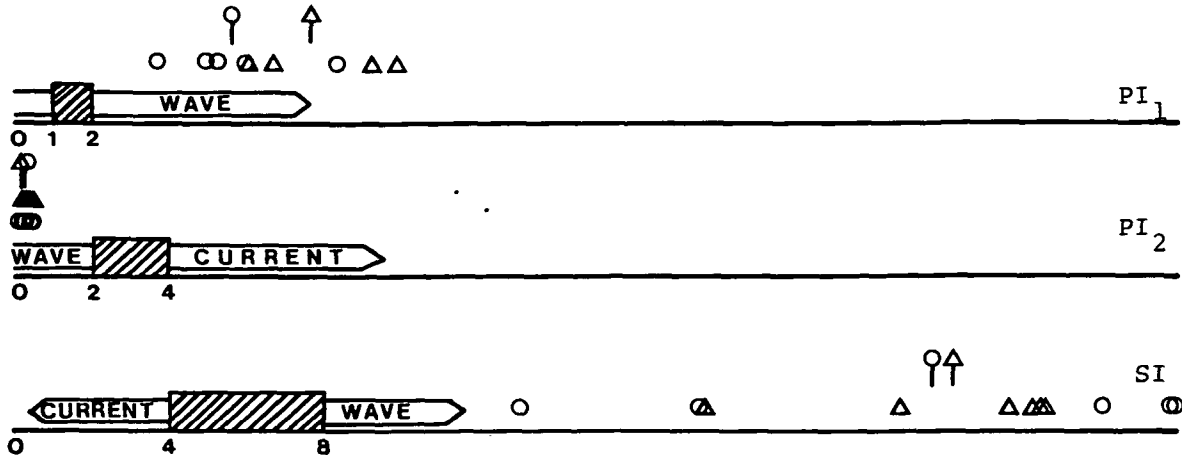
$$= \frac{\text{length of curved part of crestline} \times \text{min. ripple length}}{\text{mean ripple length} \times \text{max. ripple length}}$$

PI₂ : Parallelism index

$$= \frac{\text{max. ripple length} - \text{min. ripple length}}{\text{mean ripple length}}$$

SI : Straightness index

$$= \frac{\text{length of curved part of crestline}}{\text{departure of curvature from straight (crest) line}}$$



Hatched areas indicate where values of both wave and current ripples overlap.

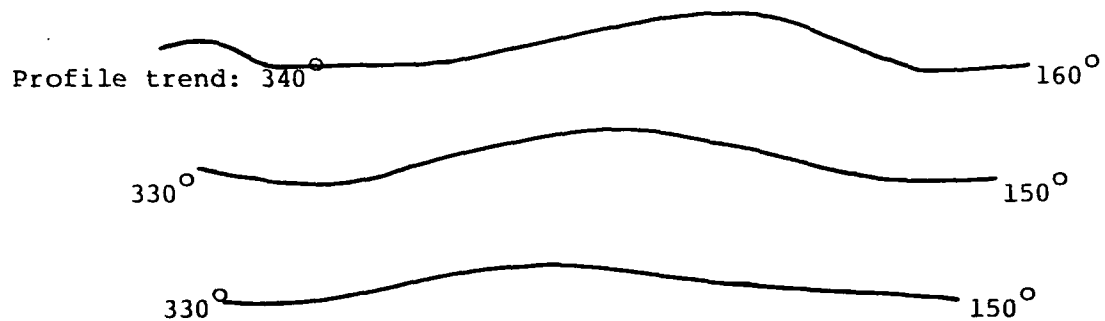
Δ Location A Existence fields based on data of Tanner (1967).

○ B

| Mean

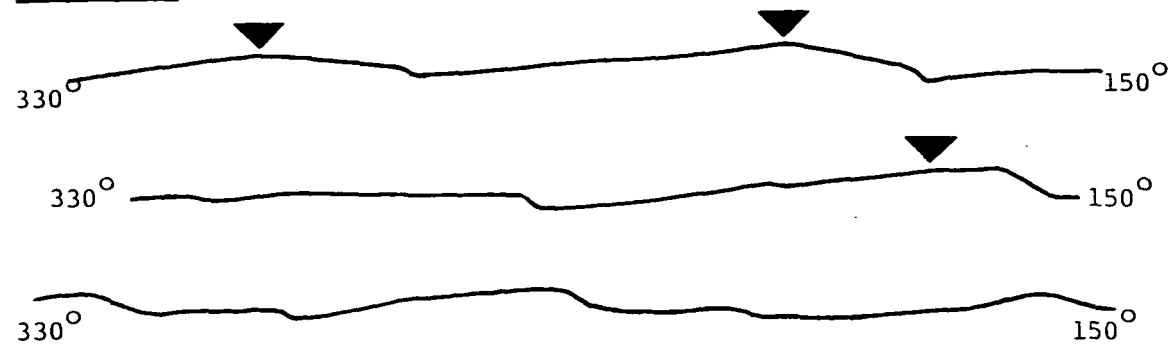
RIPPLE PROFILES.

Location A:



Profiles drawn at 1:1 scale.

Location B:



▼ = Wave-rounded crest, modifying asymmetric current ripple.

4.2.1.3.2 Interpretation

The form-sets described above fall into the category of lenticular bedding with single lenses ('linsen bedding') of Reineck and Wunderlich (1968a). The ripples almost universally fall into the thick lens sub-division, although some of the purely current-generated lenses approach the thin lens sub-division. Isolated lenticular bedding has been reported from a wide range of environments, including: lacustrine (P. A. Allen 1981), delta front (Coleman 1976), intertidal and sub-tidal (Reineck & Wunderlich 1968a) and wave-dominated shelf (de Raaf *et al.* 1977), and is therefore not environmentally diagnostic. This type of bedding represents the alternation of current and/or wave action depositing sand, alternating with slack water when mud was deposited, conditions being more favourable for the deposition and preservation of mud than for sand (Reineck & Singh 1980). Sand supply was meagre, producing ('starved') isolated ripple lenses.

The most common conditions for the generation of lenticular bedding are tidal. However, the more regular alternation of sand and mud produced by either diurnal or spring-neap tidal cycles is lacking in the Lynton Formation examples. Moreover, it is clear that the majority of ripples are of a combined-flow origin, with the unidirectional component being obliquely palaeo-offshore-directed. Lenticular bedding produced by purely tidal processes would be expected to mainly display pure current ripples (Reineck & Singh 1980), with wave-ripples being only locally important (Reineck 1960a).

An insight into the nature of the currents involved in the generation of the combined-flow ripples may be gained by considering the results of the combined-flow ripples exposed on the two extensive bedding planes at Lee Stone (text-figure 4.6). It may be seen that the ripples for both localities plot in the wave-generated field. This is consistent with observations of bifurcating crestlines (plate 4.1B), a distinguishing feature of wave-generated ripples (Reineck & Wunderlich 1968b). Further examination of text-figure 4.6 reveals that for each of the indices, the ripples at the locality shown as 'B' fall closer to the current-ripple field. This feature accords with the ripple profiles (text-figure 4.6 - 'Ripple Profiles') being more strongly asymmetric for locality B. Furthermore, small incipient spurs, a feature indicative of current-ripples (Reineck & Singh 1980), are observed on the lee-side of locality B crests. Wave-rounded crests (arrowed) modify these profiles.

In summary, the ripples are of combined-flow origin, the ratio of the component of oscillatory : unidirectional flow component varying from locality to locality. The ripples shown in text-figure 4.6 from locality B record a larger unidirectional current component than locality A ripples. Wave modification of previously produced current-ripples must be ruled out, as bifurcations in crestlines from both localities could only be produced by the simultaneous action of waves and currents i.e. the modification of current-ripples by later oscillatory currents (to produce bifurcations) would involve sand transport which would have destroyed originally current-produced structures such as spurs.

Examination of the lamination style of the combined-flow ripples confirms Reineck & Singh's (1980) observations of combined-flow ripples with a lamination style almost identical to straight crested current-ripples, being form-concordant, but also form-discordant.

The lenticular bedding with unconnected lenses bedding sub-type is interpreted as representing the obliquely palaeo-offshore migration of ripples with a limited sand supply (i.e. 'starved' ripples), below fair-weather wave-base, due to an increase in environmental energy due to a combination two or more of: (i) a lowering of wave-base during a period of increased wave energy, (ii) flow of a semi-permanent geostrophic current (see 4.2.2.3), (iii) weak tidal currents. For example, currents strong enough to transport sand as bed-load may only have occurred when a semi-permanent geostrophic flow was enhanced by: a) peak diurnal tidal flow, and/or b) spring tides and/or c) lowering of wave-base (an essentially random event). The key point is that individual flow components due to waves, geostrophic flow or tides were unable in isolation to transport sand as bed load, but the combination of two or more of these factors could support the transport of sand as bed load. The events that were responsible for the transport of sand as bed load interrupted appreciable periods of mud deposition from suspension, below fair-weather wave-base.

4.2.1.4 Facies 1 - Sub-type D - Connected Lenses

4.2.1.4.1 Description

Sandstone laminae in this bedding sub-type are usually 4 to 6mm in thickness, although they may range up to 2cm thick. They are composed of well-sorted coarse siltstone to fine sandstone and occasionally medium grade sandstone. Rare units of moderately-sorted sand occur which contain up to 30% of comminuted, convex-up shell fragments which range in size from 1 to 5mm, and up to 5% of well rounded vein quartz

granules in a matrix supported texture. The shell fragments have usually been decalcified and units containing such fragments weather to a brown colour.

The lateral persistence of sandstone units in this bedding sub-type is less than for the graded rhythmite and horizontally-laminated sandstone streak bedding types, occasionally forming isolated lenses which persist laterally for up to 2m, representing mobile sand patches on a muddy substrate. In common with the unconnected lenticular bedding type, it is apparent that two end members are present:

(i) **Unidirectional Current-Ripples.** These are strongly asymmetric in profile and never exceed more than one set in thickness. The foreset laminae are planar and lee side spurs occur occasionally (plate 4.1C). Unit bases are planar and frequently followed by a thin horizontally-laminated zone which is overlain by a unidirectional current-rippled zone. Although pure unidirectional current-generated ripples are rare, the majority appear to have been produced by the migration of straight-crested current-ripples. A few lunate current ripples, with concave-up foresets, have been recorded (plate 4.2A).

(ii) **Oscillation Ripples.** These are often two sets thick, but are rarely three sets thick. Both the set bases and the unit bases are scooped. The crests are usually rounded, although they are occasionally sharp (trochoidal), but troughs are always rounded. The above features produce a pinch-and-swell geometry to the sandstone lenses. The ripple wavelength is usually in the range of 7 to 12cm, rarely reaching 20cm; the amplitude ranges between 8 and 10mm. The internal lamination style of the sandstone layers consist of low-angled, undulating laminae, or more rarely, horizontal-lamination swinging-up into cross-lamination. The ripples are form-discordant and have scooped lower set boundaries which impart a lensoid geometry to the sand units. Rarely, a final stage, form-concordant laminae may be seen to drape the ripple profile. The dip direction of the foreset laminae is bipolar, although accurate measurements could not normally be achieved in the field. Incipient form-concordant lenses have been observed in some of the thinner units. Plan views are rare, but the few crests that have been observed trend ENE-SSW, a direction normal to the SSW-dipping palaeoslope.

In many cases a 'hummocky' plan surface is seen, and whatever section is cut through the hummock, the profile remains near-symmetric. The hummocks are approximately circular in plan, having gently curving convex-up sides. Both the apices and the swales between the 'hummocks' are rounded. The

swale-to-swale distance, via the 'hummock' apex, ranges from 7 to 15cm, the 'hummocks' reaching heights of between 8 and 10mm. Whatever direction a cross-section is cut through the 'hummocks' the internal lamination style remains the same, consisting of features diagnostic of generation by waves, including: scooped lower set boundaries, low-angled and undulating laminae, form-discordancy, a lensoid set geometry, an occasional final stage draping lamina, bundled up-building of sets and multi-directional foreset dip directions.

Again, it must be emphasised that the above types are end members of a broad spectrum. It is apparent that the majority of the examples seen in the Lynton Formation represent combined-flow or wave-modified current forms (e.g. text-figure 4.6 - location A; plate 4.2B).

A commonly occurring connected lens type consists of a planar based unit with a parallel-laminated basal zone passing upwards into an erosively based set of wave or combined-flow origin. The basal parallel-laminated zone comprises horizontal laminae that are parallel to the planar base of the unit, the base occasionally displaying drag marks (plate 4.3B). Occasionally, surfaces are exposed parallel to bedding, through the parallel-laminated zone. These surfaces display primary current lineation in some cases. The parallel-laminated base occasionally passes upwards into a current-rippled zone which may have a wave-modified symmetrical profile. However, the most common type of cross-lamination overlying the basal parallel-laminated zone displays an asymmetric ripple profile with bundled lenses and a unidirectional cross-lamination style diagnostic of production by wave orbitals with a velocity asymmetry (cf. de Raaf *et al.* 1977 - Figure 8, 6). Set bases are scooped and **erosionally** overlie the parallel-laminated zone. This is in contrast to undulating parallel-lamination swinging-up into cross-lamination, a feature diagnostic of wave-generation. It should also be noted that the lamination is bold in comparison to the faint lamination produced by sedimentation from suspension clouds. This infers a traction origin for the lamination (cf. Reineck & Singh 1972). The above features suggest that the parallel-lamination was generated under a unidirectional current.

Within the connected lenticular bedding type, isolated sigmoidal sandstone lenses frequently occur in the mudstones immediately above the rippled tops of the connected sandstone layers. These also occur above ripples in facies 5, 6 and 8 (described in sections 4.2.6, 4.2.7 and 4.2.9 respectively). Examples of sigmoidal sandstone lenses occurring above ripple profiles are shown in the field sketches reproduced in text-figure 4.7. The sigmoidal sandstone lenses always lie on the palaeo-landward side of ripple crests and are completely

surrounded by mudstone. Where observed, the cross-lamination within the sigmoidal sand lenses records a obliquely palaeo-offshore transport direction, the sigmoidal lenses climbing up the backs of the underlying ripples.

It was previously noted that connected lenticular bedding can occur as isolated, lensoid sand patches on a muddy substrate. Plate 4.2C shows evidence that these sand sheets were periodically reworked/transported, interrupting appreciable periods of low energy conditions when the sand patches were bioturbated. The existence of sand patches which were periodically transported during periods of increased environmental energy is central to the explanation of the coarsening-upwards microsequences described in section 4.2.2, a particular type of thinly interlayered sandstone/mudstone unit.

4.2.1.4.2 Interpretation

The bedding style described above falls into the category of lenticular bedding with thick, connected lenses of Reineck and Wunderlich (1968a). It should be noted, however, that the bedding style described above does not strictly conform to Reineck and Wunderlich's classification which states that only up to 75% continuous wavy sands should be classified as connected lenticular bedding; where there are more than 75% continuous wavy sands, the bedding should be termed 'wavy bedding'. In the present study all millimetre-scale, continuous, wavy sands are included in the bedding type 'lenticular bedding'. The connected lenticular bedding of this study corresponds to the lithotype 'M2 (variation a)' of de Raaf *et al.* (1977). This type of bedding has been recorded from identical, and as diverse, environments as isolated lenticular bedding, differing from the latter in that an abundant sand supply was available. Thus, connected lenticular bedding is not diagnostic of any single environment.

In summary, the connected lenticular bedding is interpreted as representing the introduction of sand via traction processes into an environment where conditions were more favourable to the deposition and preservation of mud than sand, the supply of sand being sufficient to ensure that the rippled sand patch was continuous, as opposed to a field of isolated ripples starved of sand (i.e. isolated lenticular bedding).

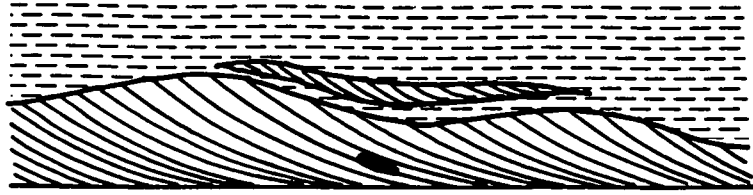
Text-fig. 4.7 Isolated, sigmoidal sandstone lenses in mudstones immediately overlying rippled sandstone horizons; drawn from field sketches.

A. Combined-flow ripples from facies 1', bedding sub-type D (connected lenticular bedding near the top of a coarsening-upwards microsequence), Lee Stone, 6949 4969. **B.** Wave-generated rippled sandstone from facies 1', bedding sub-type D (connected lenticular bedding near the top of a coarsening-upwards microsequence), Western Woody Bay, 6779 4904. **C.** Wave-generated cross-lamination erosionally overlying unidirectional current-generated parallel-lamination. Facies 5 (wavy bedding), Lee Stone, 6949 4969. **D.** Wave-generated cross-lamination with continuous mudstone lamina separating sets, facies 6 (flaser bedding), Lee Stone, 6957 4969. Note that in every case the isolated sigmoidal lens is preserved on the landward side of the underlying ripple profile, and the isolated sandstones were migrating in a palaeo-off-shore direction.

PALAEO -
OFFSHORE

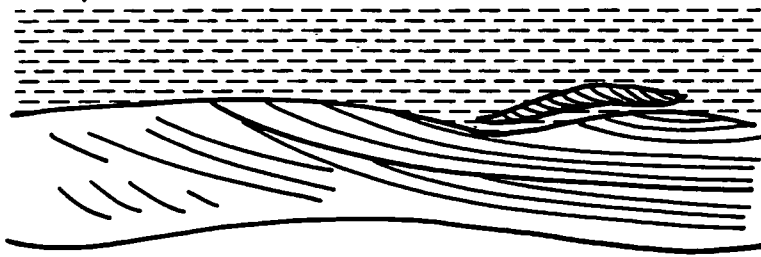


A



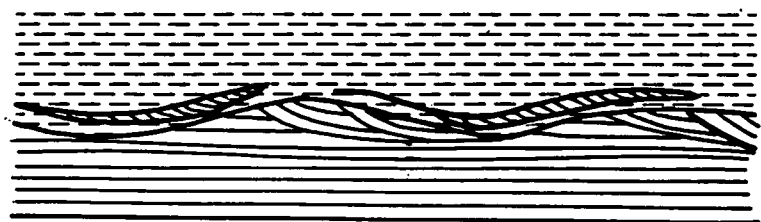
3 cm

B



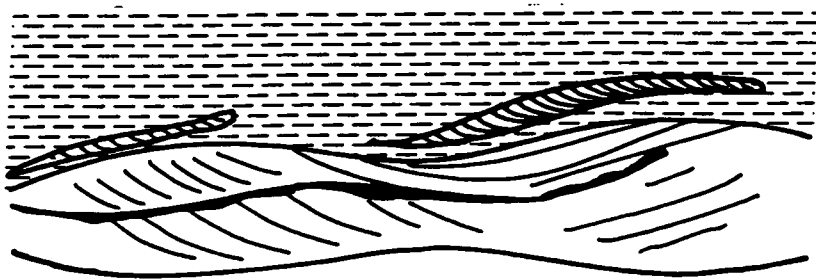
3.5 cm

C



10 cm

D



3 cm



MUD

At this point, it is necessary to discuss the nature of the hydrodynamic régime that prevailed during the emplacement of the continuous sand lenses. Firstly, the question may be posed: was the parallel-lamination at the base of many of the sand units produced under a unidirectional or oscillatory flow regime? The following characteristics are pertinent:

(i) The lamination is horizontal and parallel to a planar base, as opposed to subparallel and undulating laminations characteristic of wave-generated laminae.

(ii) Primary current lineation has been observed in some of the parallel-laminated zones (plate 4.3C). Primary current lineation arises from the bursting cycle of viscous sublayer streaks under conditions of natural turbulent flow (Allen 1982). Thus, in order to generate primary current lineation, a fully developed turbulent boundary layer is necessary. Unfortunately: "There is no adequate theory for the mechanics of turbulent flow at this time. For oscillating flows, where turbulence generated at a given location may return as a modified disturbance, a universal velocity distribution has not been established, partly because of disagreement on Reynolds criteria characterising laminar, transitional and turbulent régimes, partly because intermittent turbulence prevails over a wide Reynolds number range, and partly due to the dependence of the similarity profiles on the characteristic frequency of oscillation. Steady flow principles and assumptions, especially for the velocity distribution and the critical shear stress are not applicable to wave-induced flows, except where quasi-steady conditions prevail, i.e. where the frequency of oscillation is low and the wave amplitude is much smaller than the wavelength of oscillation" (Teleki 1972 p. 52-53). Taking the above considerations into account, it appears unlikely that primary current lineation could have developed under the short period waves that could be expected to have developed in the basin of limited fetch in which the Lynton Formation was deposited (see chapters 1, 2 & 8 for a discussion of the regional palaeogeography) except where waves shoaled on a shoreface and a strong flow asymmetry developed.

(iii) The parallel-laminated base occasionally passes upwards into a current-rippled zone which may have a wave-modified symmetrical profile.

(iv) The most common type of cross-lamination overlying the basal parallel-laminated zone displays an asymmetric ripple profile with bundled lenses and a unidirectional cross-lamination style which is

diagnostic of generation under wave orbitals with a velocity asymmetry (cf. de Raaf *et al.* 1977 - figure 8,6).

(v) Set bases are scooped and erosionally overlie the parallel-laminated zone. This is in contrast to undulatory lamination swinging-up into cross-lamination, a feature diagnostic of wave-generation (cf. de Raaf *et al.*, *op. cit.* - figure 8, 3).

(vi) The lamination is bold in comparison to the faint lamination produced by sedimentation from suspension clouds. This infers a traction origin for the lamination (cf. Reineck & Singh 1972).

The above features point to the genesis of the parallel-lamination under a unidirectional current.

In the description section above, it was noted that the tops to many of the connected sandstones consisted of a series of approximately circular 'hummocks.' The 'hummocks' appear to be equivalent to the 'three-dimensional vortex ripples' produced experimentally in oscillatory flow tunnels at the transition from two-dimensional wave-ripples to upper oscillatory plane bed (Carstens *et al.* 1969, Lofquist 1978). The significance of this ripple type is discussed in the 'Interpretation' section of facies 6 (section 4.2.7.3), a facies in which the 'hummocky' ripple type attains its maximum development. At this point it is sufficient to note that by analogy with wave tunnel experiments, the ripples are developed in sands finer than about 0.3mm at moderately high orbital velocities after a long period of evolution from initially small and straight-crested vortex ripples.

It was noted in the 'Description' section above that many of the rippled tops within the connected lenticular sandstone bedding type display isolated, sigmoidal sandstone lenses in the mud immediately overlying the rippled top of a continuous sandstone layer; the isolated sigmoidal sandstone lenses also occurring above ripples within facies 5, 6 and 8, examples being shown in text-figure 4.7. The sequence of events described below are suggested to account for the production of the sigmoidal sandstone lenses.

Following the emplacement of the connected lenses, the ripple profiles were blanketed with mud, either by sedimentation of mud from suspension clouds formed during the period of increased environmental energy, or fair-weather mud deposition. Owen (1970) and Einsele *et al.* (1974) noted that bed density increases rapidly near the base of a mud bed, and that the shear strength of the bed increases with its density,

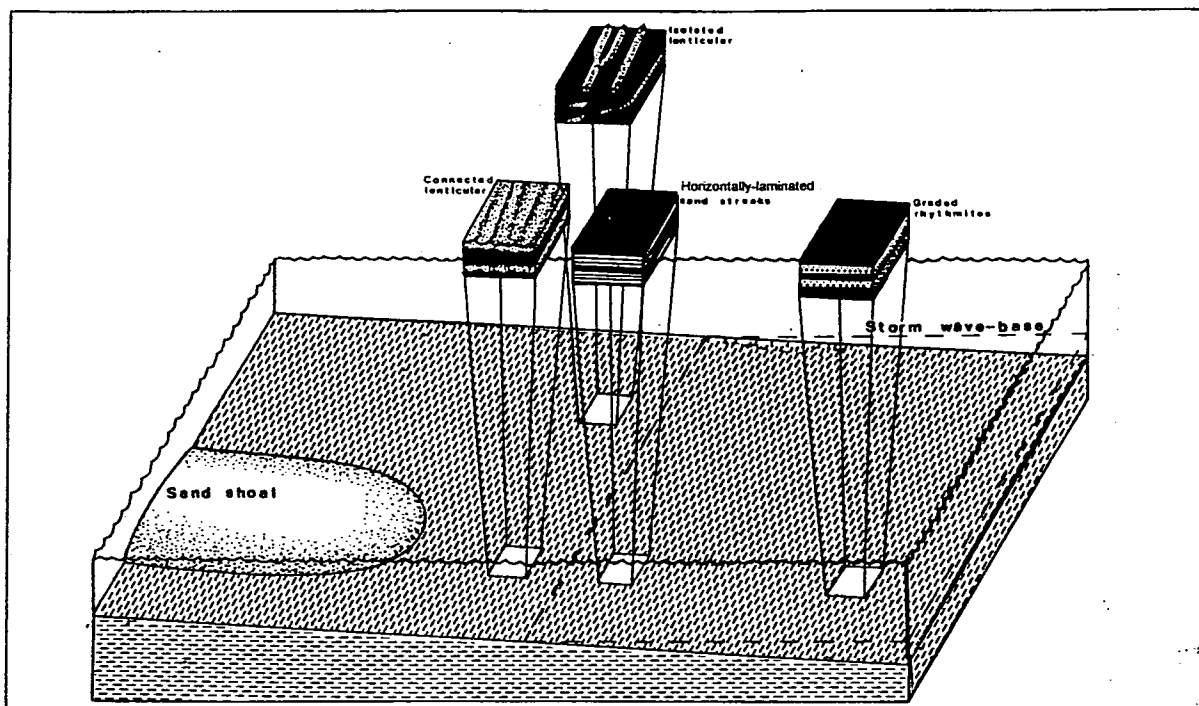
producing a lower, relatively resistant mud layer. An ensuing increase in environmental energy would remove the unconsolidated mud, leaving a thin, continuous, resistant mud lamina protecting the underlying sand layer. As environmental energy waned, sand-grade material would have been transported obliquely palaeo-offshore by the unidirectional flow component of a combined-flow over the muddy, rippled substrate, as a traction deposit in the form of isolated lenses where the sand supply was restricted. As the high energy event waned further, the sand would be preferentially deposited on the landward side of the underlying ripple in the form of a sigmoidal lens, the sand requiring greater energy to climb up the underlying ripple back as opposed to climbing down a ripple, for any given flow velocity.

4.2.1.5 Facies Summary

The thinly interlayered sandstone/mudstone facies may be subdivided according to the sandstone member type. All sub-types were deposited below fair-weather wave-base. The various bedding types may be interpreted in terms of their proximity to peak flow wave-base and the availability of a sand supply. The graded rhythmites were deposited well below peak flow wave-base whilst the horizontally-laminated sand streaks were deposited close to peak flow wave-base under the influence of storm-wave orbitals. The isolated and connected lenticular sub-types were deposited within peak flow wave-base, the isolated and connected lenses differing only in the availability of a sufficient sand supply. The relationships of the various bedding types is shown diagrammatically below in text-figure 4.8.

In the introductory section to section 4.2.1 it was noted that the facies 1 units underlying the 'Watersmeet lithotype' (facies 3) at Watersmeet contain a high proportion of carbonate mud and bioclastic debris interlaminated with siliciclastic mudstones and siltstones. The presence of recrystallized primary carbonate mud and common bioclasts suggests deposition took place on, or in proximity to, a shallow shelf with high carbonate production, probably from the erosion of calcareous shelled biota e.g. crinoid meadows. This may be the result of an episodic increase in carbonate production since most of the Lynton Formation is relatively carbonate poor e.g. it may relate to a transgressive episode (that initiating the marine Lynton Formation above a 'Dartmouth Group' type alluvial succession? - see section 1.6.3) which would have temporarily pushed back the source of siliclastic supply from the shelf allowing biota to flourish. Alternatively, there may have been a climatic aberration allowing carbonate production to increase. The lithological content and sorting of the facies 1 calcareous units certainly indicates that the bulk of the material is allochthonous,

comprising an assemblage similar to the overlying 'Watersmeet lithotype'. This, taken together with the localisation of calcareous facies I units to a thin zone immediately underlying the 'Watersmeet lithotype' at Watersmeet, suggests that the material was derived from winnowing and down-palaeocurrent transport of the 'Watersmeet lithotype' sand-body which then subsequently prograded over the calcareous facies I units.



Text-fig. 4.8 Distribution of bedding types in facies 1 (thinly interlayered sandstone/mudstone bedding) in relation to wave-base and sand supply.

4.2.1.6 Biogenic Structures

In the following sections, an account is given of the ichnofauna contained within facies I and the biofacies that they characterise. The ethology (behaviour) and ecology (response of an organism or population to surroundings) of each ichnotaxon is only briefly summarised; the reader is referred to appendix B for a full account of the taxonomy, ethology and ecology of each ichnotaxon. A summary of the distribution of ichnotaxa through the facies considered in this chapter is given in enclosure 4.

Using the scheme documented in Frey and Pemberton (1985) the ichnofauna of facies 1 can be subdivided into several types based on the ethological and preservational characteristics of individual ichnotaxa:

4.2.1.6.1 *Repichnia* (Crawling Traces)

Although many animals may have moved over, or within, the facies 1 substrate, traces of their passage will only be preserved if, in the case of open burrows, they are subsequently infilled with sediment of a contrasting lithology (see section 4.2.1.6.4), or where the traces were constructed at the junction between two contrasting lithologies. Of the latter type of preservation, *Aulichnites* sp. and 'paired, parallel grooves' are preserved as epireliefs on sandstones in facies 1 and represent the locomotory traces of animals moving over a sandy substrate. By reference to appendix B, *Aulichnites* sp. represents the ploughing furrow of a ?gastropod, whilst the 'paired, parallel grooves' represent the impressions of podia produced by the locomotion of a vermiform animal.

The preservation of delicate 'paired, parallel groove' locomotary traces, on the surface of several coarse siltstone to fine sandstone grade laminae overlain by a mudstone cap, provides evidence for quiescent periods of non-transportation / non-deposition of both sand and mud grade sediment during the deposition of facies 1.

4.2.1.6.2 *Pascichnia* (Grazing Traces)

Infaunal traces left by the *Phyllocytes* sp. animal are preserved as epireliefs on sandstones within facies 1. By reference to appendix B, *Phyllocytes* sp. represents the collapse of an originally open burrow of a deposit-feeding vermiform animal, the original burrow being kept open by the animal pressing its faecal material into the burrow wall.

The occurrence of *Helminthorhapse japonica* in facies 1 (see appendix B for a detailed account of the ethology of this ichnospecies), preserved as concave epireliefs on rippled lenticular sands, is at first sight rather disconcerting. To date, *H. japonica* and its synonyms (synonymy given in Seilacher 1977a) have only been recorded from flysch sequences, occurring in trace fossil suites which are referable to Seilacher's *Nereites* ichnofacies - an ichnofacies in which organisms have evolved highly efficient feeding strategies in response to a restricted food supply. Indeed, Seilacher (1977a, p.300) has commented that *H. japonica*: ". . .

is one of the graphoglyptid patterns that come closest to the paradigm for sediment feeding.” Furthermore, the mode of preservation of *H. japonica* in facies 1 is atypical of graphoglyptid preservation. Seilacher (1962) considered graphoglyptids to be open mud burrows that became uncovered and immediately sand-cast by low velocity turbidity currents, occurring, therefore, as convex hyporeliefs on turbidite soles. Seilacher termed these traces “pre-depositional furrows.” The explanation in the following paragraph, provides two alternative possibilities to account for the unusual mode of occurrence of *H. japonica* in facies 1.

(i) **Open Burrow:** The emplacement of a connected train of sand ripples during a period of increased environmental energy would be immediately succeeded by the settling of fine-grained material which would have been suspended during increased particle flux during the period of raised environmental energy. The differing settling velocities of individual components of the fine-grained material would be reflected in a density grading of the post- environmental energy increase mud layer, organic detritus having a lower settling velocity than clay mineral grains. Thus, the basal layer of the post- environmental energy increase mud would have had a low organic content. It is hereby proposed that the *H. japonica*-animal, with its highly efficient infaunal grazing pattern, represents an opportunistic response in exploiting the basal part of the post- environmental energy increase mud layer which had a low organic content. Due to collapse of the open tunnel during sediment compaction, tunnels constructed entirely within mud would not be preserved. However, where the tunnels were constructed along a sand/mud interface, collapse of the mud roof of the tunnel onto the concave tunnel floor would result in concave epirelief preservation on sand surfaces.

(ii) **Surface Feeding:** The *H. japonica* animal was a surface feeder which grazed the surface of sand layers, which would have had a low organic content due to winnowing, and would have needed to use an efficient grazing pattern to conserve energy.

It is possible that the *H. japonica* animal burrowed or grazed muds within the Lynton Formation deposits. The post-burial metamorphic history of the Lynton Formation, however, makes it unlikely that any specimens would be preserved and this possibility will probably remain unresolved.

4.2.1.6.3 *Fodinichnia* (Feeding Traces)

Infaunal traces left by the *Gyrochorte cosmosa* animal are preserved as epireliefs upon sandstones within facies 1. By reference to appendix B, *G. cosmosa* represents the oblique passage through the substrate of a vermiform tube. 'Longitudinally striated tubes,' only observed in the exposure immediately above Lee Stone (6953 4963), are interpreted as representing stowing structures produced by an infaunal deposit feeder.

4.2.1.6.4 *Fodinichnia* / *Domichnia* (Feeding Traces / Dwelling Structures)

The most ubiquitous and cosmopolitan ichnotaxon within the Lynton Formation is *Palaeophycus tubularis*, a form which grades into *Chondrites* sp. a in the lower energy deposits of Wringcliff Bay (see chapter 3) and west Crock Point (see chapter 5). *P. tubularis* is interpreted herein to be the semi-permanent dwelling burrow of an omnivorous vermiform animal whose diet was maintained by predation and/or suspension-feeding, supplemented by infaunal deposit-feeding. The vacated burrow was passively infilled with sand and is usually preserved in either a bed-junction or burial mode of preservation (cf. Simpson 1957).

Where the host sediment was muddy, the burrows are generally disposed at a low angle to bedding (10° - 25°), a phenomenon which probably reflects a shallow redox potential discontinuity or a shallow nutrient-rich zone within the muds. It is probable that the above two factors were interrelated; in either case, the burrow would have been restricted to the sediment immediately below the sediment-water interface. Where the host sediment was of a sandy composition, the burrows maintain a near vertical attitude. The near vertical attitude is interpreted as reflecting a redox potential discontinuity that was at a much greater depth below the sediment-water interface in sandy sediments, the burrows therefore not being restricted to a thin, oxidised surface layer of sediment, as was the case in the muddy sediments. Furthermore, the sandy sediments would have probably contained very little organic detritus and the *P. tubularis*-animal would have had to have burrowed vertically downwards until it encountered an organically rich muddy layer which it could exploit horizontally (see plate B.14D). It should be noted, however, that differential compaction of the mud component, when compared with the sand component, plus the fact that the mudstones layers would have accommodated more shear strain than the more competent sandstone layers, would both have resulted in tubes preserved in sandstone layers being steeper than those preserved in mudstone layers.

The presence of penecontemporaneously exhumed burrows, partial collapse of unfilled sections of burrow and diagenetic haloes around burrow walls, suggests that the *P. tubularis*-animal secreted a mucus-type

substance to bind its burrow wall, thus imparting strength to the burrow structure. The manner of preservation of the coherent burrow structure, therefore, yields an insight into the environmental conditions that prevailed during the deposition of the sediments enclosing the *P. tubularis* burrows.

The rose-diagrams shown in text-figures B.9 and B.10A, B, C give the alignment of all the *P. tubularis* burrows on a given area of a single bedding plane. Thus, the rose diagrams show the alignment of both *in situ* tubes and penecontemporaneously exhumed tubes. If the unabraded, demonstrably *in situ* burrows from each sample are plotted separately (text-figure B.10A', B', C'), a strong NNW-SSE primary alignment can be detected, a trend which is parallel to the palaeocurrent direction. If the rose-diagrams for the whole samples are re-examined in the light of the alignment of the demonstrably *in situ* burrows, it may be seen that the alignment of the abraded, exhumed burrows has two modes: a major mode parallel to the palaeocurrent trend (NNW-SSE), which has a wider spread of values than the mode for the *in situ* tubes, and a minor mode perpendicular to the major mode. The major mode is interpreted as representing tubes that have been exhumed and aligned within the current, the minor mode representing burrows that have been rolled upon the substrate by the current. Using the terminology of Simpson (1957), the exhumed tubes are preserved in a 'burial mode' of preservation. The sediment that infills the *in situ* burrows can usually be seen to have been 'piped down' from an overlying bed ('bed-junction preservation'), although in rare cases the overlying bed may have been penecontemporaneously removed ('concealed bed-junction preservation'). The frequent occurrence of *P. tubularis* burrows in a burial mode of preservation, and occasionally in a concealed bed-junction mode of preservation, demonstrates that penecontemporaneous erosion commonly disturbed the sediments of facies 1.

It was noted above that in addition to the secondary alignment and rolling of exhumed *P. tubularis* burrows within the current, the *in situ* burrows within muddy sediments are also aligned, the *in situ* burrows being aligned parallel to the obliquely palaeo-offshore-flowing wind-induced geostrophic flow (text-figure B.10A', B', C'), the burrows dipping at an angle of 10° to 25° obliquely palaeo-offshore (SE). Thus, the burrow entrance faced into the prevailing current. Several hypotheses are advanced in to explain this phenomenon; the following paragraphs are a brief summary of the conclusions reached in Appendix B.

- (i) Where suspension-feeding forms part of the animals feeding activity, it would be advantageous to the animal to align its feeding-apparatus into the prevailing current (cf. Reidl 1971).

(ii) In cases where part of the animals diet was obtained by predation, it would be advantageous for the animal to align its feeding-apparatus into the prevailing current in order for the animals pressure and chemical sensors to detect approaching prey (cf. Ocklemann & Vahl 1970).

(iii) If the *P. tubularis*-animal returned to the surface to replenish its oxygen reserves, it would be of benefit to the animal if its respiratory-apparatus faced into the oxygen-rich prevailing current.

(iv) For a dense burrow population, the most efficient burrow configuration is for burrows to lie parallel to each other. However, to achieve such a burrow configuration, a fixed common 'reference point' is required. The obliquely palaeo-offshore-flowing semi-permanent geostrophic flow may have served as the requisite 'reference point.'

Individual factors suggested above to account for the primary alignment of the *in situ* *P. tubularis* burrows may have acted alone or alternatively, several of the factors may have acted in combination. In either case, the occurrence of aligned *P. tubularis* burrows in facies 1 indicates that the obliquely offshore-flow was semi-permanent in nature.

The size-frequency distribution of *P. tubularis* burrow diameters, shown in the histograms in text-figure B.10, also reveals details of the physical environment that prevailed during the tunnelling and preservation of the burrows. The histograms for the *in situ* burrows (A', B', C') show a strong negative skew, and no tubes under 3mm in diameter occur. The histograms for the *in situ* burrows are interpreted to be the product of an animal which had a planktonic larval stage which, on settling to the substrate, had a high initial growth rate, a mortality which increased with age or an accumulation of generations in the larger growth sizes. Unfortunately, no independent evidence is available to ascertain the growth rate and it is, therefore, impossible to attribute the negative skew to a particular factor.

The size-frequency histograms for the total sample of *P. tubularis* burrows (text-figure B.10A, B, C - combined sample of *in situ* & penecontemporaneously exhumed burrows) has an approximately Gaussian distribution, burrow diameters ranging from 1 to 6mm. This burrow size-frequency distribution is attributed to the following factors:

- (i) During exhumation, the burrows became abraded, resulting in a reduction in burrow diameter. Thus, the mean burrow diameter for the combined sample of exhumed and in situ tubes (A, B, C) is, in each sample, less than the mean burrow diameter of the in situ tubes alone (A', B', C').
- (ii) The presence of burrows with a diameter less than 3mm is the product of a reduction in burrow diameter caused by abrasion during exhumation.
- (iii) The presence of abraded tubes of 6mm in diameter infers that some of the abraded burrows were derived from a population of burrows which contained burrows in excess of 6mm in diameter. The larger burrows were presumably the product of burrow development in sediments with an increased nutrient content, a higher oxygen content or, deposition in a more turbulent environment.

Burrows referable to the ichnogenus *Rosselia socialis* are common in facies 1. In suitably exposed specimens, the stacked cone structure of *Rosselia socialis* may be observed to pass distally into simple, thinly-lined tubes referable to the ichnogenus *Palaeophycus tubularis* (where the distal portion of the burrow is a *fodinichnial*, i.e. feeding, trace). *Rosselia socialis* is interpreted (see appendix B) to have been produced by the upward migration of a cone-shaped depression representing the surface opening of a semi-permanent dwelling burrow i.e. *fugichnial* (escape structure) modification of a *domichnial* trace in response to substrate aggradation or degradation. The cone-shaped depression is believed to have served as a collecting pit for the organic detritus up on which the *Rosselia socialis* animal fed. In common with *P. tubularis* burrows in facies 1, the *Rosselia socialis* structures predominantly dip in an obliquely palaeo-offshore direction (SE); the conical collecting pit therefore faced into the obliquely palaeo-offshore-flowing geostrophic flow, presumably aiding the collection of organic detritus in the conical pit. The consistent obliquely palaeo-offshore dip of the *Rosselia socialis* structure again evidences the semi-permanent nature of the geostrophic flow. The ubiquitous truncation of the stacked cone structures at a particular horizon, representing a former sediment surface, demonstrates the frequent occurrence of penecontemporaneous erosion in facies 1.

The ichnospecies *Teichichnus rectus* has only been observed in sediments interpreted as having been deposited distally to sand bodies. The *T. rectus*-animal, therefore, appears to have preferred a low energy environment. *T. rectus* is believed to represent the combined feeding / dwelling burrow of a vermiform animal, the spreite structures being produced in response to sediment influx i.e. *fugichnial* (escape structure) modification of a *fodinichnial* / *domichnial* trace in response to substrate aggradation. Significantly, only retrusive spreite have been observed, inferring that the substrate within which the burrow was produced was

aggrading upwards and was not subject to periodic erosional events which would have resulted in a net degradation of the substrate and the production of protrusive spreite. The low environmental energy of the substrate is reflected by the nature of the sediments within which *T. rectus* is preserved. Graded rhythmites and horizontally-laminated sandstone streaks are more common within the sediments enclosing *T. rectus* than elsewhere in facies 1, indicating that the substrate rarely fell within the zone of periodically increased environmental energy. Furthermore, erosion surfaces are virtually absent, reflecting the absence of penecontemporaneous erosion.

Fürsich (1975), in a study of the distribution of trace fossils in the Corallian rocks of Yorkshire, Dorset and Normandy, has also found *T. rectus* to be confined to sediments deposited in a low-energy environment. Fürsich proposed three recurrent trace fossil associations which were primarily related to depth. Fürsich's *Teichichnus* association, characterised by *T. rectus*, was found to occur in argillaceous sediments containing shallow burrows, mainly of deposit feeders. These sediments were interpreted as having been deposited in a low energy régime with low rates of deposition and a fairly stable substrate i.e. offshore subtidal regions or lagoons. A similar low energy régime, with low rates of deposition and a fairly stable substrate, is envisaged for the facies 1 sediments which contain *T. rectus*.

4.2.1.6.5 *Domichnia* (Dwelling Structures)

In facies 1, passively infilled cylindrical burrows occasionally occur with burrow walls that are more thickly lined than those of *P. tubularis*; these burrows are assigned to the ichnospecies *P. heberti*. The rare examples of *P. heberti* in facies 1 are related to the occurrence of thicker than normal connected lenticular sandstone horizons. *P. heberti* is interpreted herein (appendix B) as the domicile of a vermiform animal, the distal part of the burrow displaying unlined/thinly lined foraging offshoots. Endobionts living in semi-permanent dwelling burrows require a burrow with a permanent wall to facilitate the circulation of respiratory currents within the burrow (Schäfer 1972). Due to the greater porosity of sand in comparison with mud, burrow walls in sandy sediments are more thickly lined than those constructed in muddy sediments. The rarity of *P. heberti* in facies 1 reflects the relative paucity of thick sandy horizons in facies 1.

'Mantled' tubes, proposed to represent the dwelling burrows of an animal with a similar ethology to cerianthid anemones (see appendix B), do not occur frequently in facies 1. When they do occur, however,

they can be found in great numbers e.g. cliff-top exposure at Duty Point (6940 4966) - see plate B.9A and B. Schäfer (1972) recorded a similar distribution of Holocene cerianthids in North Sea sediments.

Holocene cerianthids have been observed to live for 10 to 40 years in aquaria (Hyman 1940). The result of such longevity is that burrows attain large lengths in response to an aggradation of the substrate, Hyman recording burrows up to 1m in length. After a correction for tectonic extension (see appendix C), 'mantled' tubes in facies 1 reach maximum observed lengths of 16cm. The great length of the Lynton Formation 'mantled' tubes is attributed, by analogy with Holocene cerianthid burrows, to a longevity in the animal that built the 'mantled' tubes, the animal migrating upwards within its burrow in response to an aggradation of the substrate i.e. *fugichnial* (escape structure) modification of a *domichnial* trace in response to substrate aggradation.

The mode of preservation and infill of the 'mantled' tubes yields much valuable data relating to the physical sedimentary conditions that prevailed during the period of burrow construction. Three modes of tube infill occur; these are schematically shown in text-figure B.6. All three modes of infill may occur within any individual specimen. It should be remembered that the 'mantled' tubes pierce mudstones containing thin sandstone horizons, the thin sandstone horizons representing the incursion of coarser sediment, during periods of increased environmental energy, into an environment favouring the deposition and preservation of mud.

- (i) Successive Dwelling Structures. These consist of a series of U- to V-shaped meniscate structures, usually defined by mudstone laminae, with sandy sediments intervening between the mudstone laminae. A distinct laminated wall structure is usually visible. The U- to V-shaped mudstone laminae represent the position of the base of the dwelling cavity at a particular point in time. The sandstone infill intervening between the mudstone laminae represents the influx, during periods of increased environmental energy, of coarse material into the dwelling cavity, necessitating an upward migration of the dwelling cavity.
- (ii) Cone-in-Cone Structure. These consists of a series of stacked, steep-sided, V-shaped cones of sandstone in which the inverted apex of the cone may be either open or closed. No wall structure is visible. The cone-in-cone structure is interpreted to represent the escape trail produced by an upward, rapid migration of the 'mantled' tube animal in response to a large increment of sediment during a storm.

(iii) Passive Infill. Parallel-sided tubes with a thinly laminated wall structure. Interpreted as a burrow that has been passively infilled with laminated mudstones and normally graded sandstones after the 'mantled' tube animal had abandoned its burrow.

The above three modes of tube infill are consistent with the construction of a burrow in an environment that favours the deposition and preservation of mud, but periodically experiences an influx of coarser sediment during periods of increased environmental energy of variable magnitude.

The character of the surface the 'mantled' tubes are preserved beneath also reflects the nature of the physical sedimentary environment that prevailed during the construction and preservation of the 'mantled' tubes. It was noted previously that 'mantled' tubes do not occur frequently, but where they do occur, they can be found in great numbers. Where these dense populations do occur, the burrows within the population are normally truncated at a single, common horizon, e.g. population shown in plate B.9B. These horizons of common truncation are interpreted as representing the passage of a high energy event of unusually large magnitude, the burrow population preserved beneath such horizons thus being interpreted as a mass mortality assemblage. Using the terminology of Simpson (1957), two modes of preservation occur beneath the horizons of common truncation:

(i) Bed-Junction Preservation. This type of preservation results when a change in the type of sediment deposited occurs immediately upon vacation of the burrow. The new sediment type of the overlying bed infills the underlying burrows.

(ii) Concealed Bed-Junction Preservation. This type of preservation results when a change in the type of sediment deposited occurs immediately on vacation of the burrow. However, the new sediment type has been subsequently removed, the different lithology of the burrow infill beneath an erosion surface being the only evidence of a change in the sediment type. Simpson (1957) and Hallam (1975) have taken the occurrence of concealed bed-junction preservation to be good evidence of penecontemporaneous erosion. In some cases the sand that infilled the burrows may be located by tracing the erosion surface laterally, the occurrence of an isolated, rippled sandstone lens evidencing the passage of a discrete sand-patch over the substrate.

An example of concealed bed-junction preservation is shown in plate B.9B. In this photograph, the 'mantled' tube population occurs beneath an erosion surface with a thin mudstone-drape (arrowed); there is no trace of the sandstone that infilled the tubes. Furthermore, the diameters of the tubes occurring beneath the erosion surface in this example are unusually narrow, presumably indicating that the erosion event caused a mass mortality of the animals that inhabited the tubes, the growth of the tubes, therefore, being halted prematurely.

4.2.1.6.6 Fossitextura Deformativa

Schäfer (1972) defined a simple classification of biogenic traces based on biostratinomic criteria - see Table 4.2.

1. Burrowing textures (fossitextura deformativa)
2. Complete structures (fossitextura figurativa)
 - (a) sunk or internal structures
 - (b) elevated structures
3. Half-structures
 - (a) sculptured bedding planes
 - (b) sculptured boundary planes
4. Mobile structures

Table 4.2 Bioturbate Textures - a biostratinomical classification

From Schäfer (1972 - table 12)

Adopting the above classification, the dwelling burrows described in the previous section would be assigned to fossitextura figurativa - permanent cavities with hardened walls, where: "The boundary against the surrounding sediment is distinct" (Schäfer *op. cit.*, p.405). In the present section, the textures generated by the locomotion of burrowing organisms within the substrate, resulting in the deformation of the original fabric (perturbed burrowing textures, or fossitextura deformativa) will be examined.

Fossitextura deformativa are biogenic structures without any definite form appearing as mottled structures or irregular flecks of different grain size or colour. Howard (1975, p.134) noted that: "In the event of continuous, slow, uninterrupted sedimentation, the record is one of complete biogenic reworking, generally by a variety of organisms." However, it should be borne in mind that a densely bioturbated unit does not necessarily imply that the texture records an environment in which a myriad of organisms were crawling in, on, and through the substrate i.e. using the phraseology of Howard (*op. cit.*, p.135): "Typically, the high degree of bioturbation has more to do with the amount of time available for biogenic activity per unit accumulation of sediment than with animal density or "frenzy" of activity." By reference to the bioturbation column on the logged sections, it may be seen that, to a greater or lesser degree, the majority of facies 1 units are bioturbated. It may also be seen that *Palaeophycus tubularis* burrows are the predominant biogenic

structure. However, *P. tubularis* has a distinct, thinly lined burrow wall (see Appendix B) and is therefore referable to *fossitextura figurativa*. In units where bioturbation is intense, bioturbation by *P. tubularis* is supplemented by indistinct biogenic churning (see enclosure 3B for symbol) which may genuinely be referred to *fossitextura deformativa*. In some units the biogenic churning is so intense that no distinct biogenic structures may be defined, the biogenic churning imparting a gnarled texture to the sediment (see plate 4.2C).

In the units composed almost entirely of mudstone, the lack of granulometric contrast makes it difficult to determine whether the unit has been completely homogenised by bioturbation, or has not been bioturbated at all. This problem was resolved by examining the fissility of individual lithified mud units. Polishing of sawn surfaces of specimens collected from units with a poor fissility and a gnarled appearance revealed a 'swirled' texture, defined by faint differences in coloration on polished surfaces; the 'swirled' texture is interpreted to be biogenic in origin. In contrast, the polishing of sawn surfaces of specimens collected from units with a good fissility revealed a uniform coloration, such units being interpreted as non-bioturbated. A similar relationship between shale fissility and bioturbation was noted by Byers (1974) in a study of the Upper Devonian Sonyea Group of New York State and the Upper Cretaceous Pierre Shale of the Western Interior Seaway of North America. Byers concluded that shales which have not been biogenically disturbed retain their original horizontal fabric, imparting a good fissility to the shale. The horizontal fabric may either be the product of a lamination resulting from variations in sediment supply and/or sediment type, or the horizontal alignment of platy clay grains and carbonaceous detritus. Bioturbation, however, resulted in a randomisation of the horizontal fabric with a concomitant loss of shale fissility. Interestingly, illite crystallinity studies (Kelm & Robinson 1989) indicate that Lynton Formation mudrocks are of low greenschist metamorphic grade where: "The Devonian rocks of the region have undergone major recrystallization as shown by the strong alignment of the phases into the closely spaced cleavage" (p.150). Nevertheless, the metamorphic overprint was insufficient to completely eradicate biogenically randomised fabrics.

In the first half of the present century, *fossitextura deformativa* received little attention compared to that received by *fossitextura figurativa* (Schäfer 1972). However, since the classic study of Moore and Scruton (1957), who used the intensity of bioturbation to characterise delta and delta-influenced shelf environments of the Mississippi Delta, the proportion of biogenically reworked sediments, compared to the proportion of physically produced sedimentary structures, has been used to interpret the nature of depositional

environments. The precise ratio of biogenically reworked sediments to preserved physically produced sedimentary structures is dependent on physical energy (waves and currents), the rate of sedimentation and the density, adaptation and variety of organisms (Howard 1975). For example, Moore and Scruton (*op. cit.*) found that physically-produced layering was preserved where sediment rates were high, and waves and currents reworked the substrate. Expressed quantitatively, preserved layering exceeded destruction by burrowing when sedimentation rates exceeded 4cm per year. It was noted at the beginning of the present section that continuous, slow, uninterrupted sedimentation would leave a record of complete biogenic reworking of a deposit. However, intense bioturbation, with the complete erasure of physical-sedimentary structures is rare in facies 1 suggesting high deposition rates and/or high physical energy levels (see above).

A third factor is the density, adaptation and variety, of organisms. However, this factor has varied during the Phanerozoic, bioturbation increasing during the Phanerozoic (Bambach 1977, Thayer 1979). For example, the rapid diversification of holothurians at the end of the Devonian resulted in a marked increase in bioturbation of sediments (Thayer *op. cit.*) The lack of surface deposit- feeders ('biological bulldozers' of Thayer *op. cit.*) is reflected by the fact that the ichnofauna of facies 1 consists predominantly of a suspension-feeding infauna (enclosure 4). Rhoads and Young (1970) have shown that suspension-feeders will be displaced if surface deposit-feeders are present as the deposit-feeders suspend particles which will clog the filtering mechanism of suspension-feeders.

In conclusion, the paucity of completely bioturbated deposits in facies 1 suggests high deposition rates and/or high physical energy levels, although direct analogies cannot be drawn with Holocene sediments due to the decrease in bioturbation backwards in time through the Phanerozoic. It should be remembered that the fact that facies 1 is a predominantly muddy facies does not necessarily infer that physical energy levels were low during the deposition of facies 1. As McCave and co-workers (McCave 1970, 1971, 1972, McCave & Swift 1976) have shown, mud may be deposited and preserved at relatively high levels of tidal/wave/current energy.

In addition to the proportion of biogenic structures to physical sedimentary structures yielding details of relative rates of deposition and physical energy during deposition, biogenic structures can also provide direct evidence of the physical properties of the original substrate, particularly the original fluid content of the sediment. The value of such information is of great importance in palaeoenvironmental reconstruction as:

“... substrate coherence is probably one of the main factors controlling the distribution of trace-making organisms - even more so than sediment composition or grain size” (Howard 1975, p.141). Research in the field of the relationship of biogenic structures to substrate consistency has been pioneered by Rhoads (Rhoads 1967, 1970, 1975, Rhoads and Young 1970).

Substrate firmness of marine sediments (quantified in terms of sediment cohesion or water content) ranges from lithified rock to ‘soupy’ muds. The initial state of sediments may be ascertained in ancient rocks by an examination of deformation structures associated with biogenic traces. “Shallow-burrowing organisms (grazing or foraging traces) moving through an uncompacted sediment, high in water content (≥ 50 percent by weight), ordinarily produce a narrow zone of deformation around their burrows. Water-lubricated grains slide past one another in the loosely packed matrix, to accommodate volume displacement by the burrower. In contrast, an organism burrowing through a firm bottom, low in water content (≤ 50 per cent), deforms the bottom plastically; i.e. each burrow is surrounded by a relatively larger zone of deformation, extending several grain diameters away from the burrow wall” (Rhoads 1975, p.153-154).

Examination of the contacts between sandstones and mudstones in facies 1 fossitextura deformativa reveals that contacts are sharp, frequently displaying a zone of plastic deformation around biogenic structures (see plate 4.2C). This suggests that the sediments preserved in facies 1 had a low water content. In terms of rheology, the sediments would have behaved in a plastic manner during biogenic deformation. This deduction is corroborated by the fact that fossitextura figurativa in facies 1 have sharp contacts between the burrow and the host sediment. For example, plate 4.1A shows sharp contacts between *P. tubularis* burrows and the surrounding mudstone. Furthermore, there is frequently a zone of plastic deformation surrounding burrows in facies 1. For example, plate B.9C shows a wide zone of plastically deformed lenticular bedding surrounding a ‘mantled’ tube, although a component of the deformation was almost certainly imposed during sediment compaction and metamorphism. Rhoads (1970) suggested that the degree of compression of biogenic structures may also be used as a guide to determining the original water content of a substrate. Burrows in facies 1 show relatively minor compression (e.g. *P. tubularis* burrows shown in plate 4.1A); what little compression there is may be explained by sediment compaction and tectonic compression, the latter being evidenced by cleavage ‘augening’ burrows. Thus, the minimal amount of burrow compaction in facies 1 also suggests that the sediments had a low initial water content.

It was stressed above that the preserved sediments of facies 1 appear to have had a low original water content and deformation was plastic in nature. However, a study by Rhoads (1970) of Holocene substrates in a subtidal setting off the Massachusetts coast showed that the vertical distribution of infaunal deposit-feeders within a substrate controls the water content of the substrate: "The vertical distribution of deposit-feeders may extend several centimetres into the sediment but the greatest density of macrofaunal invertebrates in subtidal muds is found within the upper 10cm of the bottom" (p.394). Rhoads observed that sediments that had been highly biogenically reworked, indicated by a granular surface faecal zone, had water contents exceeding 60%, occasionally reaching as much as 80%. The rheological properties of muds with high water contents are thixotropic and burrows would show a diffuse wall structure where water-lubricated grains slid over one another; no zone of plastic deformation of surrounding sediment would occur. Postma (1967) has shown that the water content of fine-grained sediments has a major influence on critical erosion velocity; the higher the water content, the lower the initial velocity required for erosion for any given grain size. Thus, burrowed substrates are more easily resuspended than non-burrowed substrates (Rhoads & Young 1970). It is possible, therefore, that the original facies 1 substrate consisted of a bioturbated, thixotropic surface layer with a high water content, sediment with a lower water content and having plastic deformation characteristics underlying the surface thixotropic layer. The surface thixotropic layer would have been prone to resuspension and would therefore have had a lower preservation potential than the underlying plastic muds. However, work by Rhoads and Young (1970) has also shown that: "Reworking of muds by deposit-feeders (primarily errant polychaetes and protobranch bivalves) can produce a fluid sediment surface that is easily resuspended, even by weak tidal currents. High turbidity at the sediment surface effectively excludes suspension feeding-organisms that are sensitive to clogging" (Rhoads 1975, p.151). Rhoads and Young concluded that a spatial separation may be expected to occur between areas where fluidization of the muddy bottom by deposit-feeders results in the exclusion of suspension-feeders, and areas where the activity of deposit-feeders is minor, allowing suspension-feeders to survive. The presence in facies 1 of traces attributed, at least in part, to suspension-feeders - namely: *Rosselia socialis*, 'mantled' tubes, *P. tubularis* and *Teichichnus rectus* (see appendix B, for ethological reconstructions) - suggests that although semi-permanent currents swept the facies 1 substrate, the mud was firm enough to resist resuspension, inferring that a fluid surface of mud did not develop in facies 1.

In conclusion, the sharp sandstone/mudstone contacts present in fossitextura deformativa observed within facies 1 and the presence of deformation structures surrounding biogenic textures, suggests that the preserved

sediments of facies 1 had a low original water content i.e. it would have exhibited plastic rheological properties. The presence of suspension-feeding forms in facies 1 implies that a surface layer of mud with a high original water content (thixotropic) did not develop. The paucity of fossil texture deformative in facies 1 indicates that deposition rates were high, although, as stated above, care must be taken in making direct comparisons with Holocene shelf sediments as bioturbation rates have increased with time during the Phanerozoic.

4.2.1.7 Biofacies

Schäfer (1972) erected five biofacies which could be recognised, in both Holocene and ancient sequences, on the basis of the character of their organic and inorganic constituents, their shape, and spatial relationships. Schäfer's scheme was an adaptation of the six-fold classification of Schmidt (1958) who recognised that turbulence is the most decisive factor in delimiting biotopes i.e. turbulence determines the distribution and texture of grains and the oxygen supply available for organisms. However, Schäfer found that Schmidt's scheme could not be applied with any degree of accuracy to ancient sequences and therefore modified Schmidt's scheme accordingly.

Schäfer proposed three substrate types based on the structure, shape and spatial relationships of their constituent organic and inorganic grains: *astrate*, characterised by non-bedded organogenic material of permanent biocoenoses; *lipostrate*, characterised by many minor bedding disconformities, and *pantostrate*, characterised by conformable bedding. To each of these types, the modifying terms *vital* or *lethal* were added. *Vital* refers to biofacies comprising biocoenoses and taphocoenoses, whilst *lethal* refers to substrates comprising solely of taphocoenoses. It should be noted that, by definition, *astrate* cannot be composed of taphocoenoses (i.e. *lethal*) and must therefore only have *vital* as a prefix. Detailed schematic sketches, with accompanying explanatory notes, of each of Schäfer's five biofacies are reproduced in enclosure 5.

Schäfer proposed the term *vital lipostrate* to describe biofacies characterised by multiple bottom-biocoenoses that are destroyed at an early stage, taphocoenoses, and by disconformities (understood herein as including bedding disconformities in which the bedding planes above and below a break are essentially parallel, the break being marked by an erosion surface). It is apparent that the majority of facies 1 is referable to Schäfer's *vital lipostrate* biofacies.

In terms of physical sedimentary structures, facies 1 is dominated by connected and unconnected lenticular bedding which, together with erosion surfaces overlying burrows preserved in a concealed bed-junction mode of preservation (e.g. plate B.9B), indicates that the substrate periodically suffered agitation by waves and currents. Furthermore, evidence above (section 4.2.1.6.4) and discussed in detail below (section 4.2.2.3) suggests that semi-permanent geostrophic currents operated during the deposition of facies 1.

The bottom-biocoenoses of facies 1 also suggest that the substrate was periodically subject to sudden influxes of sediment of variable grain size. This is particularly apparent in the morphology of the 'mantled' tubes, which show evidence of an upward migration of the burrow in response to sedimentation (see text-figure B.6). However, storm events of unusually high energy occasionally resulted in the truncation of a population at a particular horizon (plate B.9B). Nevertheless, following such events the substrate appears to have been rapidly recolonised, as is evidenced by the paucity of unburrowed horizons in facies 1.

Taphocoenoses in facies 1 are of two generic types: allochthonous and para-autochthonous. The allochthonous taphocoenoses comprise abraded and disarticulated biogenic debris, chiefly crinoid ossicles, bivalve and brachiopod valves, concentrated either within sandstone layers e.g. unit at 1.51m on the log of the upper cliff at Duty Point (text-figure 4.5) or as thin zones of convex-up valves (i.e. in a hydrodynamically stable position - Clifton 1971) within silty-mudstones (plate 4.6B). Valves disseminated through biogenically homogenised silty-mudstone facies 1 units below the 'Watersmeet lithotype' (facies 3) at Watersmeet (enclosure 7, columns B & C) are interpreted as having a random orientation due to disturbance by biogenic 'churning.' The para-autochthonous taphocoenoses comprise burrows that have been penecontemporaneously exhumed and locally rolled on the substrate e.g. *P. tubularis* (see section 4.2.1.6.4).

In summary, the majority of facies 1 is classified as vital lipostrate. It is important to remember that: "The material record furnished by the biofacies (i.e. vital lipostrate) documents only a fraction of the events that have happened during the period of its existence because beds are lost and organogenic remnants are carried away" (Schäfer *op. cit.*, p.476).

Although the majority of facies 1 falls into the category of vital lipostrate, there are certain exceptions. The thinly interlayered sandstone/mudstone units occurring in the west Lee Stone section (text-figure 4.2),

between 9.6m and 13.8m and also between 14.2m and the top of the log, although having both unconnected and connected lenticular bedding, they also contain many units with a high proportion of graded rhythmites and horizontally-laminated sandstone streaks. Thus, many of the sandstone units that were deposited on the facies 1 substrate, preserved within the portions of the logged sequence noted above, accumulated below the zone of periodically increased environmental energy (see text-figure 4.8). Within the intervals of sequence mentioned above, decimetre-scale horizons exist where graded rhythmites and horizontally-laminated sandstone streaks were the only sand layers that were deposited and no erosional discordances interrupt the sequence suggesting that for appreciable periods of time deposition occurred on a substrate unaffected by currents capable of transporting sand-grade material as bed load. Nevertheless, the oxygen content of the water was sufficient to sustain permanent benthonic biocoenoses, as evidenced by the profuse ichnofauna. These sediments may be assigned to the vital-pantostrate biofacies of Schäfer (*op. cit.*) who proposed that the biofacies was: "Characterised by permanent benthonic biocoenoses, by taphocoenoses of nektonic and planktonic animals, and by complete, conformable bedding" (p. 480). The occurrence of *Teichichnus rectus* within units deposited below the zone of periodically increased environmental energy, was discussed above (section 4.2.1.6.4).

Rare units of unbioturbated facies 1 deposits, e.g. a 40cm thick unbioturbated silty-mudstone with occasional unconnected sandstone near the base of the east Lynmouth beach log (text-figure 3.1B). In addition, the base of several coarsening-upwards microsequences also show no trace of biogenic activity, e.g. unit at 3.95m on the western Lee Stone log (text-figure 4.2), and are uninterrupted by erosional disconformities. These unbioturbated horizons are interpreted as being referable to Schäfer's lethal-pantostrate facies.

In summary, facies 1 is predominantly composed of sediments referable to Schäfer's vital-lipostrate biofacies intercalated with occasional decimetre-scale referable to his vital-pantostrate and lethal-pantostrate biofacies.

4.2.2 Coarsening-Upwards Microsequences Within Facies 1

4.2.2.1 Description

Coarsening-upwards (CU) microsequences are considered as a special case of thinly interlayered sandstone/mudstone bedding, with cyclical increases in sandstone content defining decimetre-scale CU units. This microsequence type is shown in a separate column, denoted A', on the left-hand margin of the logs for east

Lynmouth beach (text-figure 3.1A) and Duty Point (text-figures 4.2 - 4.5). Weathering and lichen cover precluded the possibility of distinguishing any microsequences within the facies 1 units exposed at Watersmeet (enclosure 7) or the A39 road section (text-figure 4.1). Individual microsequence thickness varies from 2 to 38cm, averaging approximately 9cm. Thickness data for each of the four sections at Duty Point are presented in Table 4.3.

| SECTION | N | Max.(cm) | Min.(cm) | \bar{X} | σ_{n-1} |
|-------------------------|------------|-----------|----------|-------------|----------------|
| Lee Stone - west | 75 | 35 | 2 | 7.98 | 4.79 |
| Lee Stone - east | 22 | 23 | 3 | 5.35 | 5.35 |
| Outcrop above Lee Stone | 16 | 38 | 4 | 12.44 | 10.2 |
| Duty Point cliff top | 11 | 34 | 2 | 9.45 | 7.76 |
| Aggregate | 133 | 38 | 2 | 8.84 | 6.41 |

Table 4.3 CU microsequence thickness data for microsequences associated with the 'Lee Stone facies association,' Duty Point

\bar{X} = mean, σ_{n-1} = standard population deviation.

The variation in sandstone content between the base and top of a single CU microsequence can range from 5% to 100%. As a broad generalisation, units occurring well below, or above (i.e., deposited distally), a thick sandstone body (i.e. on the scale of the 'Lee Stone facies association' - see section 4.3.1) have sandstone contents increasing upwards from 0% to 30%, whereas CU microsequences immediately below sandstone bodies (i.e., deposited proximally) have sandstone contents which increase upwards from 80% to 95% sandstone. The sandstone content of an individual microsequence may increase either gradationally or by step-wise increments. Microsequences may commence, particularly in units well below or above a sandstone body, with a thick pure silty-mudstone which may exceed 5cm in thickness.

The sandstone content of a unit is contributed by thin primary sandstone intercalations of the sub-types described in section 4.2.1 and by the infill of biogenic structures. Rarely, one may occur to the exclusion of the other. Graded rhythmites and horizontally-laminated sandstone streaks are comparatively rare, predominantly occurring in microsequences well below or above sandstone bodies, where they generally occur in the lower part of a microsequence. The majority of sandstone intercalations are of the lenticular-bedded type. Within a single unit there is a very strong trend for isolated lenses to become connected upwards. Furthermore, there is also a tendency for the connected lenses to thicken upwards, e.g. plate 4.5.

When observations are compiled for all the CU microsequences that have been examined the following **idealised** sequence emerges. From the base upwards: (i) silty-mudstone; (ii) graded rhythmites and horizontally-laminated sandstone streaks (no preferred distribution); (iii) isolated lenticular sandstone lenses;

(iv) connected lenticular sandstone lenses which become thicker towards the top of the microsequence; (v) flaser bedding, mudstone flasers becoming less connected upwards. Member (v) is rare, only occurring in microsequences immediately below sandstone bodies.

Unit bases are of three types:

(i) A pure, generally relatively thick ($\leq 12\text{cm}$), mudstone which either drapes the preceding unit conformably (plate 4.5D) or erosively truncates final sandstone of the preceding unit. The mudstones are evenly cleaved, exhibiting no evidence of internal disconformities or bioturbation.

(ii) Relatively thin ($\leq 5\text{cm}$) homogenous silty-mudstone drape on the final sandstone of the preceding unit. The drape is sharp-based in the case of an underlying rippled or laminated sandstone, but is gradational where the previous unit top has been bioturbated by biogenic structures that cross the boundary of the two units. Individual units tend to have an uneven cleavage, displaying a 'gnarled' appearance on both weathered (plate 4.5A) and fresh surfaces - where grain size differences are sufficient a 'swirled' texture can occasionally be observed.

N.B. It is recognised for both (i) and (ii) that the lowermost mudstone laminae of a CU microsequence may well have been deposited from a sediment cloud suspended during the event that emplaced the uppermost sandstone of the preceding CU microsequence, and as such, should strictly be assigned to the preceding CU microsequence. In practice, however, it is not possible to distinguish between a mudstone drape associated with the final high-energy event that emplaced the preceding unit, and the true basal mudstone laminae of the succeeding unit. For this reason, the base of CU microsequences are taken at the contact between the final sandstone of the preceding unit and the thick mudstone deposits at the base of the overlying microsequence.

(iii) A planar, laterally persistent erosion surface, with up to 2cm relief, truncating sedimentary and/or biogenic structures of the preceding unit (plate 4.5A). The erosion surface may be locally obscured by bioturbation. Type (iii) CU microsequence bases can be overlain by type (i) or (ii) (silty-) mudstone units.

The top of a CU microsequence commonly consists of a thin, clean, rippled sandstone (facies 1, sandstone sub-type D), often reworking bioturbated sediments. Ripples are usually of a three-dimensional vortex type with isolated sigmoidal sandstone lenses above them. In other examples, mud was introduced into clean

sands at the top of a microsequence by bioturbation. This has the effect of producing a thin (<1cm) fining-upwards top to the CU microsequence.

CU microsequences attain their maximum development below, and immediately above, sandstone bodies i.e., deposited proximally to the sandstone body where an abundant sand supply was available. Plate 4.4B shows particularly striking examples of CU microsequences from the lower part of the Lee Stone eastern section. CU microsequences may also be observed within thinly interlaminated sandstone/mudstone microsequences within the 'Lee Stone facies association' proper. Examples displaying the variability of CU microsequences are shown in plate 4.5.

4.2.2.2 Biofacies

Although the biofacies that occur in facies 1 have been discussed previously (section 4.2.1.7), it is appropriate at this point to examine the specific distribution of biofacies within individual CU microsequences.

The majority of CU microsequences are bioturbated. The most prominent biogenic structures are sandstone-filled *Palaeophycus tubularis* burrows in mudstone, and general biogenic churning. *P. tubularis* burrows occur in mudstones at the base of microsequences and between sandstone intercalations. The biogenic churning generally affects the more sandy upper parts of units. Occasionally, biogenic churning is severe enough to destroy all of the primary sedimentary structures within a unit. Multiple episodes of bioturbation may be discerned within an individual microsequence e.g. plate 4.5A.

Within a microsequence bioturbation generally increases upwards, although a few examples of bioturbation decreasing upwards have been recorded (e.g. plate 4.4A). In some cases the total sandstone content of a microsequence is contributed by biogenic structures, predominantly *P. tubularis*, the burrow density increasing upwards within a microsequence.

The generally upward increase in burrow density has two possible explanations:

- (i) The upper surface of the mudstone represents the palaeo- sediment/water interface. Burrow density would naturally decrease downwards beneath the interface, the upper 10 to 15cm of substrate being the most biologically active (Webb *et al.* 1976).
- (ii) Increasing sand supply with time would result in the younger burrows (i.e. in the upper part of a microsequence) being sand filled. Earlier burrows, at the base of a microsequence, would become infilled with mud or collapse and therefore become undetectable when the sediment becomes lithified, due to the lack of granulometric contrast.

Differing lithologies of sandstone infilling the burrows at different levels within a microsequence provides evidence to support an interpretation based on process (ii) described above, although this process was probably supplemented by process (i). The source of sand in process (ii) was, presumably, isolated patches of connected and unconnected lenticular sands (facies 1, sandstone sub-types C and D) which migrated over the substrate during increased flow energy, the only record of their passage being the infilling of the burrows with sand cf. 'concealed bed-junction preservation' of Simpson (1957). Burrows towards the upper part of a CU microsequence may be disturbed by biogenic 'churning'.

Examining the specific distribution of biofacies within individual CU microsequences is best approached by describing the biofacies that are present within a complete, idealised CU microsequence.

Of the two basic types of (silty-) mudstones devoid of sandstone horizons occurring at the base of a CU microsequence:

- (i) The relatively thick, pure mudstones or homogenous silty-mudstones which are evenly cleaved (suggesting that the muds were not disturbed by bioturbation cf. Byers 1974; see section 4.2.1.6.6), and exhibiting no evidence of internal disconformities, are assigned to Schäfer's (1972) lethal-pantostrate biofacies.

(ii) The relatively thin zones of heterogeneous silty-mudstones displaying an uneven cleavage and a 'gnarled' appearance on both weathered and fresh surfaces and contain occasional sandstone-infilled *Palaeophycus tubularis* burrows. The limited thickness of these zones of silty-mudstone is due to either a restricted supply of mud or, penecontemporaneous erosion of the muds. In either case, the muds were subsequently bioturbated, as is evidenced by the 'gnarled' appearance of both fresh and weathered. These mudstones are therefore referable to Schäfer's vital-pantostrate biofacies or, where there is evidence of penecontemporaneous erosion, to the vital-lipostrate biofacies.

Moving upwards within our idealised complete CU microsequence into conformably bedded mudstones with graded rhythmites and horizontally-laminated sandstone-streaks, sandstones frequently infill *P. tubularis* burrows and the beds are therefore referable to the vital-pantostrate biofacies..

Resting gradationally above the mudstones with graded rhythmites and horizontally-laminated sandstone streaks are lenticular bedded sandstones and mudstones more typical of facies 1. As was discussed in detail in section 4.2.1.2.5 the lenticular bedded sandstones and mudstones with abundant biogenic traces are referable to Schäfer's vital-lipostrate biofacies.

The uppermost part of our idealised CU microsequence comprises flaser bedding interpreted to be the product of increased environmental energy emplacing sands into an environment where muds were deposited during periods of lower environmental energy, the majority of the mud being subsequently eroded during the next phase of increased environmental energy. The rippled upper surfaces of the sands were colonised by *the Palaeophycus heberti*-animal and the intervening mud flasers were extensively bioturbated, predominantly by *P. tubularis*. The flaser bedding is also referable, therefore, to the vital-lipostrate biofacies of Schäfer.

4.2.2.3 Interpretation

The interpretation of the specific distribution of mudstone/sandstone layers within individual CU will be presented by considering an idealised CU microsequence:

At the base of the idealised CU microsequence, (silty-) mudstones devoid of sandstone horizons occasionally occur. Two contrasting mechanisms for the emplacement of these units can be invoked. Firstly, the mudstone may represent a single event deposit (*sensu* Seilacher 1982a), perhaps resulting by deposition from a plume

of fine sediment generated during a river flood event, deposited and preserved at mid-shelf depths cf. Drake *et al.* (1972) or alternatively, deposition from a cloud of fine-grained sediment generated by the suspension of shelf muds during a period of increased environmental energy. Graded mudstones recorded from the Upper Devonian Pilton Beds of north Devon by Goldring and Langenstrassen (1979) are a possible analogue for the Lynton Formation pure mudstones and may be of an 'event deposit' origin - see also facies 2 (section 4.2.3). An alternative, second mechanism for the deposition of the thick, pure mudstones may be envisaged in which a mud unit accumulated by slow deposition, possibly during the monsoon season (see below). It should be noted that the deposition and preservation of mud does not necessarily infer a low environmental energy, but may instead reflect the possibility that the mud was the only sediment type supplied during the monsoon season. As McCave and Swift (1976) have shown, where the concentration of fine-grained sediment in suspension is sufficiently low, mud may be deposited and preserved at relatively high velocities as a result of net entrapment of fine suspended sediment in the viscous sublayer of a turbulent boundary layer.

Moving upwards within our idealised complete CU microsequence, mudstones with graded rhythmite and horizontally-laminated sandstone-streaks occur above the basal mudstones. The sandstones are interpreted as having been deposited below the zone of periodically increased environmental energy (see sections 4.2.1.1 & 4.2.1.2).

Resting gradationally above the mudstones with graded rhythmite and horizontally-laminated sandstone streaks in our idealised CU microsequence are lenticular bedded sandstones and mudstones typical of facies 1. The sandstones are interpreted as having been deposited within the zone of periodically increased environmental energy but below fair-weather wave base (see sections 4.2.1.1 & 4.2.1.2).

The uppermost part of our idealised CU microsequence comprises flaser bedding, although it should be noted that flaser bedding only occurs occasionally within CU microsequences. The flaser bedding is interpreted to be the product of periods of increased environmental energy emplacing sands into an environment below fair-weather wave base where muds were deposited during a period of decreased environmental energy, the majority of the mud being subsequently removed during the erosional phase of the next phase of increased environmental energy (section 4.2.7.3).

“In shallow water the chance of periodic alternating beds to be preserved is minimal. Here, repeated reworking and deposition of sediments by physical and biological agents or the influence of local sediment sources etc. will usually obliterate any primary rhythm” (Einsele 1982, p.7). The above quote serves to emphasise the importance of the Lynton Formation cycles and demands that a widespread, fundamental process was responsible for the Lynton Formation examples.

The deposition and preservation of shallow marine cycles is more common in limestone-marl sequences, where the primary rhythms become enhanced by diagenetic processes, the rhythms usually representing deposition on a relatively stable and sufficiently deep platform; or in a basin, with slow deposition rates (Einsele *op. cit.*)

Examples of shallow marine small-scale coarsening-upwards microsequences of tidal origin have been recorded from Holocene inshore mesotidal deposits of the North Sea by Terwindt (1971, 1981) and Jurassic tidal flat deposits by Sellwood (1975). Terwindt (1981) described decimetre-scale microsequences displaying a gradational transition from sand-clay alternating bedding and/or lenticular bedding and/or wavy and flaser bedding into cross-lamination (i.e. ‘coarsening-upwards WEAK CUR-MED CUR microsequences’ where ‘WEAK CUR’ = weak currents, and ‘MED CUR’ = medium currents). Bioturbation was minimal and primary sedimentary structures were produced by unidirectional currents. Terwindt (1981) calculated the frequency of occurrence for the various types of microsequence and found that the coarsening-upwards WEAK CUR - MED CUR microsequences had a 20% frequency compared with 40% fining-upward MED CUR - WEAK CUR microsequences and 20% for fining-coarsening MED CUR - WEAK CUR - MED CUR microsequences. Terwindt did not attribute the microsequences to any specific process, although he did note that the number and thickness of the mud layers recorded from flaser bedding within fining-upward MED CUR-WEAK CUR microsequences was not uniform; this variation was interpreted as possibly representing variations in current strength during the neap-spring tidal cycle.

Examples of coarsening-upward microsequences from a group interpreted as having been deposited in a non-tidal setting have been recorded by Sellwood (1970) from the British Lias. Three types of coarsening-upward

microsequence were distinguished. In each case, the sediment type, trace and body fossils indicated increasing environmental energy towards the top of the cycle. Cycle type I (1.4 to 3.6m in thickness) comprised clays, often with siltstone laminae that thickened and became more continuous upwards, with a pectinid-nuculid faunal assemblage and small pyrite filled burrows, passing upwards into siltstones and lenticular bedded fine sandstones with abundant suspension-feeding bivalves. The trace fossil assemblage was dominated by *Rhizocorallium* and *Diplocraterion*; *Teichichnus* was also recorded. In every case the burrow spreite were retrusive. Scours, circular in plan, 1m in width and 10cm in depth were recorded in the top of some cycles. Type II cycles (0.15 to 0.9m in thickness) had a similar trace fossil assemblage to type I cycles, but occurred in a more calcareous facies, limestone units representing the coarser portion of each cycle. Cycle type III (0.3 to 1.5-m in thickness) comprised silty-clays, which become more silty upwards, differing from type I cycles in the occurrence of *Pinna*, a more argillaceous lithology and fewer trace fossils.

Where exposure was favourable, each of the cycle types could be traced over 5km laterally, suggesting a broad causative mechanism. Sellwood related the cycles to the deposition of clay below wave-base, sedimentation being more intermittently rapid in the silty cycle top where waves and currents were stronger. Each cycle was thought to represent a shallowing-upwards event related to eustatic instability combined with tectonic subsidence. Calcareous nodules, rich in ammonites at the cycle tops, were taken to represent: “. . . a probable still-stand in sedimentation and transgressive phase” (p.496).

A purely tidal origin for the Lynton Formation CU microsequences, cf. Terwindt (1981), is precluded by the fact that the cross-lamination within the Lynton Formation cycles is of a combined-flow origin (with oscillatory currents as the principle component), whereas the cross-laminations described by Terwindt were of a unidirectional current origin. Furthermore, only 20% of Terwindt's microsequences were of a CU variety. Examination of the logs reveals that the vast proportion of thinly interlayered sandstone/mudstone units showing a preferred internal organisation are of a CU type, although rare fining-upwards and coarsening-fining-upward microsequences have been recorded (see logs). Finally, Lynton Formation deposits display few features that can be considered to be diagnostic of generation solely by tides (see section 4.2.4) - none provide unequivocal evidence of tides.

The Lynton Formation cycles more closely parallel those described by Sellwood (1970), particularly his type I cycles, where environmental energy was higher at the top of a unit. However, Sellwood invoked eustatic

instability combined with tectonic subsidence to account for his cycles. This mechanism seems unlikely in the case of the vast number of Lynton Formation microsequences which have a relatively uniform thickness and occur throughout the complete thickness of the Lynton Formation sequence, although they are much more frequent in the deposits immediately above and below the 'Lee Stone facies association.' In the search for a more suitable mechanism, the following characteristics are pertinent: relatively uniform thickness, multiple levels of bioturbation within a microsequence and a rhythmic change in environmental energy.

In order to evaluate the controlling factors responsible for the genesis of the Lynton Formation CU microsequences it will be posited that the pertinent features listed in the previous paragraph could be explained by a seasonal climatic control and/or proximity of a sand source during deposition.

Before any discussion proceeds on the way in which the Early Devonian palaeoclimate of southern Britain could influence the formation of the CU microsequences, a brief review of the late Early Devonian palaeoclimate affecting palaeo- NW Europe is apposite. Using palaeomagnetic data the majority of authors place southern Britain between 15° and 30° south of the equator during the late Early Devonian (see section 1.6.1), within an expected belt of SE trade winds that currently stretch from a latitude 5/10° to 35° N/S and at the margin of a zone stretching between 5° and 25° where occasional (several times a season) winter storms and occasional (c. once per 3 000 years) hurricanes can be expected (Marsaglia & Klein 1983). It is possible, however, that southern Britain lay within a zone stretching between 25° and 45° where winter storms are common and hurricanes occur occasionally (Marsaglia & Klein *op. cit.*)

By analogy with the present distribution of climatic belts (Riehl 1979) a monsoonal climate, characterised by seasonal wind reversals, may be expected. It should be noted, however, that the different configuration of continents, and the barriers presented by mountain belts in particular, would have had a major influence on climatic belt distribution during the Devonian, limiting direct analogies with present day climatic belt distribution. Indeed, Witzke & Heckel (1989) noted that:

“Monsoonal circulation is developed in response to seasonal temperature contrasts over large mid-latitude continental masses, producing strong seasonal variations in wind direction and rainfall that disrupt or eliminate the zonal pattern. Depending on geographic factors, monsoonal circulation potentially can establish seasonally humid climatic conditions at latitudes generally characterised by

the zonal dry belts” (p.51). But went on to note that: “. . . three lines of evidence suggest that monsoonal circulation may not have been a significant feature for Devonian Euramerica: 1) land-sea temperature contrasts would have been ameliorated by the widespread epicontinental seas, particularly during the Middle and Late Devonian; 2) virtually all Devonian global palaeogeographic models show a decided asymmetry of land about the equator, whereas monsoonal systems apparently are best developed when the distribution is more symmetrical . . . ; and 3) equatorial mountains may counter the effects of the developing monsoon . . . , and one or more Euramerican mountain ranges . . . transect or straddle the equator on virtually all published Devonian paleogeographic maps.”

Parish (1982) also discussed Devonian palaeoclimate. Although Parish's palaeoclimatic reconstruction for the Emsian is rejected as data were plotted on the base maps of Scotese *et al.* (1979) which place southern Britain on the equator, a position disputed by subsequent authors (see section 1.6.1), a useful review of factors influencing the formation of a monsoonal climate was presented. On a uniform globe, high pressure zones centre 30° N and S of the equator and are best developed in winters i.e. in summers low pressure develops over continents which breaks up cells. Using the example of contemporary Asia, the central continental region is markedly hotter in summer and cooler in winter because it is isolated from the ameliorating influence of the oceans. Furthermore, a high mountain range (Himalayas) presents an E - W barrier which shields the centre of the continent from oceanic influence during winter and further enhances the likelihood of high pressure developing. In summer, the Himalayas channel and enhance low pressure over Afghanistan and Pakistan. The result is a marked monsoonal seasonality. Parish discounted a monsoonal climate for the Devonian of the region that is now NW Europe on the basis that the Caledonide mountain belt had a N-S trend.

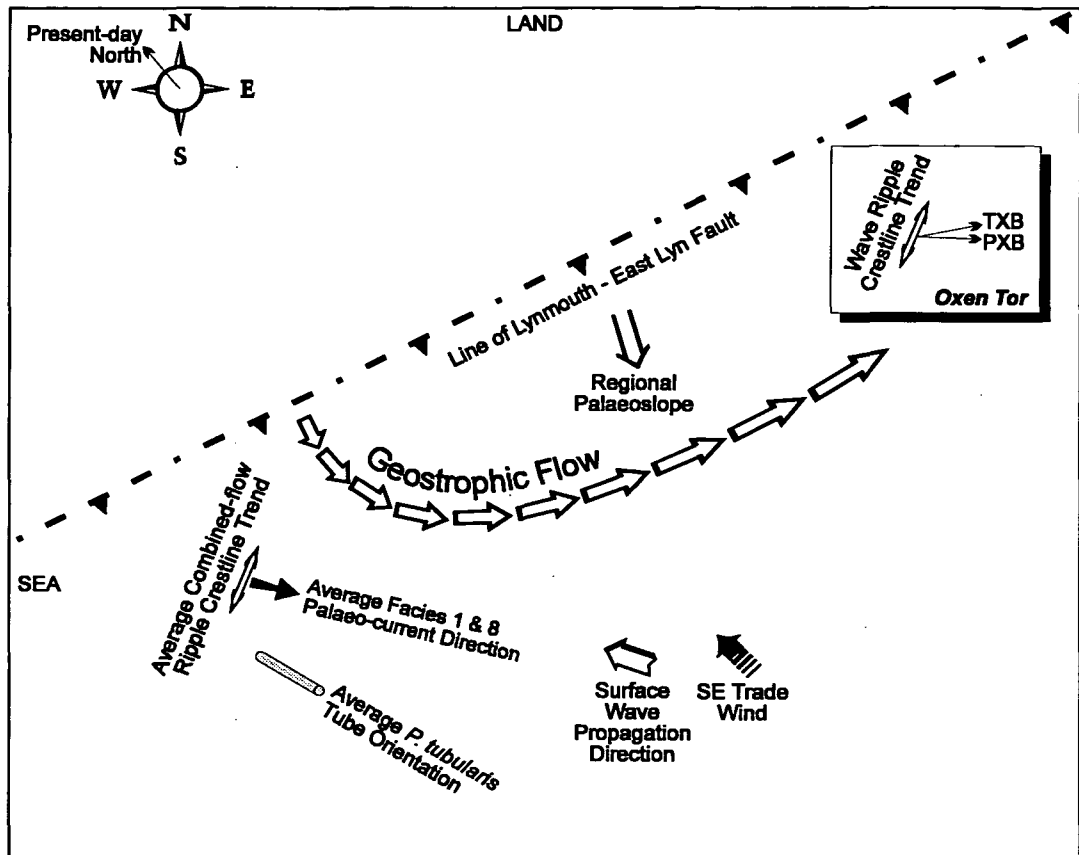
Nevertheless, there is an independent line of evidence support a monsoonal climatic setting may have prevailed (locally?) in southern Britain during the Early Devonian. Allen (1974b) discussed the abundance of pedogenic carbonate horizons (calcretes) in the Lower Old Red Sandstone of the Anglo-Welsh outcrop. Allen, following the survey of Goudie (1973) noted that calcretes require a warm to hot climate for their formation, with a mean annual temperature range of 16° to 20° C and a marked seasonal rainfall, 100 to 150mm in range. Pseudoanticlines developed in calcretized horizons were attributed to alternate, seasonally

controlled, wetting and drying of the deep clayey soil and subsoil (cf. 'gilgai'). The above features are typically developed in a monsoonal climate.

For the purpose of initial discussion a monsoonal climate will be assumed for Early Devonian times in southern Britain. The following scenario can be envisaged for the genesis of the Lynton Formation CU microsequences.

An anticyclonic system centred over the Variscan geosyncline sea during the southern winter would have generated constant SE trade winds blowing onshore, calms being practically non-existent over the sea. The resultant wind-shear over the sea surface would produce a palaeo-NW-flowing surface wind-drift current in the water column. When this current met the land barrier of the eastern side of the Old Red Sandstone Continent, water would have 'piled-up' against the shoreline. This 'piling' would have resulted in an opposing counter-current flowing obliquely offshore at depth, deflected (to the left in the palaeo- southern hemisphere) by the Coriolis force to set up an along-shore (at mid-depth) geostrophic flow (Strahler 1963 - see text-figure 4.9). During the winter trade wind season a semi-permanent geostrophic current would have flowed, progressively moving sand patches obliquely offshore. For a given fixed point on the substrate, with a sand patch lying in a shore-ward direction of the fixed point at the commencement of the trade wind season, the sand patch would become progressively more proximal as the winter advanced.

With the onset of summer monsoon season and the accompanying southward shift of the intertropical convergence zone, the characteristic monsoonal wind reversal would occur, with moist unstable westerly winds of a lower velocity and greater variability, blowing offshore from the Old Red Sandstone Continent. Any surface wind-drift current produced on the western side of the Variscan geosyncline sea would have flowed offshore. As a combination of no barrier being presented to this offshore-flowing surface current and the limited fetch reducing effective wave energy at depth, a geostrophic flow would not have been generated and sand transport due to meteorological currents would cease during fair-weather periods affecting the shelf. This factor, in part, could account for the asymmetric development of the CU microsequences.



Text-fig. 4.9 Palaeo-shelf configuration showing inferred palaeo-wind direction and resulting palaeocurrent indicators recorded at Lee Stone

The compass rose shows palaeo-North (palaeomagnetic rotation taken from Torsvik *et al.* 1990); palaeoslope dip calculated from palaeo-translation direction of the east Lynmouth beach slide sheet (documented in section 2.2.1.1.2). SE trade winds, with the propagation of surface waves deflected to the left (in the Southern hemisphere) by the Coriolis force, would have 'piled' water against a palaeo-scarp / palaeo-shoreline (taken as line of Lynmouth - East Lyn Fault, with interpreted late Early Devonian downthrow to the palaeo-SSE during the rifting stage of basin formation - see section 2.2.4). This 'piling' would have resulted in a counter geostrophic flow, also have been deflected progressively towards the left by the Coriolis force (cf. Strahler 1963). The average combined-flow ripple crestline trend, taken from measurements of facies 1, 5, 6 & 8 at Lee Stone, is shown along with the average direction of cross-lamination and cross-bedding dip within facies 1 & 8 at Lee Stone and average *Palaeophycus tubularis* tube orientation from Lee Stone (see table B.2). The inset shows rotated palaeocurrent indicators from Oxen Tor (see text-figure 6.1).

Starting at the base of an idealised CU microsequence, cycle bases normally consist of a mudstone drape on the rippled top of the previous CU microsequence. Occasionally, however, planar laterally extensive erosive events occur at, or just above, the cycle base. During the southern summer, moist unstable air over the western sector of tropical seas results in hurricane formation (Riehl 1979). The passage of a hurricane would be expected to produce an erosion event. Sand would only be deposited by such an event if the storm track moved onshore to produce a storm-surge with its resultant obliquely offshore-flowing sediment-laden geostrophic flow. A large proportion of the storm tracks would not have crossed the coastline and no sand deposit would result, the passage of the hurricane being marked only by an erosion event. Furthermore, the passage of a hurricane is a relatively transient meteorological event, 'coupling' with the water surface being weak and resulting in only weak geostrophic flow patterns being generated. However, stirring of the substrate

during the hurricane would winnow-out and suspend fine-grained sediment and ripple the remaining sandy lag deposit. Clean, rippled sandstones are common below erosion surfaces in the Lynton Formation. The low frequency of these erosional events reflects the rarity of a hurricane affecting a given coastal area. For example, Ager (1973) reported that for any particular site on the Gulf Coast of Mexico, there is 95% probability that the site will be affected by a hurricane at least once every 3000 years. Alternatively, the erosion surface could merely represent the passage of a migrating sand patch (that would have sourced the sandstone lenses at the top of the microsequences) which left no depositional record.

The thick mudstones seen at the base of many cycles can be interpreted as representing summer season deposits generated by monsoonal rains affecting continental areas and producing river flow (river beds would have been dry in the arid winter) transporting sediment to the coast, which would be predominantly muddy at low flow stage (Dr. I. P. Tunbridge, pers. comm.) The resulting increase in suspended fine-grained sediment in shelf waters alone could result in mud being deposited where previously sand had been transported (cf. McCave 1970). In addition, lower wind velocities in the summer would result in lower values of wave effectiveness at the bed (McCave 1971) which would also allow fine-grained sediments to accumulate where sand had been transported previously.

Occasional summer season cyclonic fronts (Riehl 1979) would have increased wave and current energy, entraining and transporting sand in suspension clouds, resulting in the deposition of sand-grade material in zones (below fair-weather wave-base) where mud would have normally been deposited during summer fair-weather periods. The resulting sand deposit in the basal part of our idealised CU would, therefore, be of a graded rhythmite and horizontally-laminated sandstone-streak type, the type being dependent on the degree to which wave-base was deepened during the cyclonic storm event, not the proximity of the sand supply. The absence/weakness of summer geostrophic flow, even where it may have been enhanced by a weak tidal flow, would have been insufficient to rework the graded rhythmite and horizontally-laminated sandstone-streaks when they were deposited. These sand layers would be buried by mud during the summer and, therefore, protected from erosion during the subsequent winter trade wind season when semi-permanent bottom-currents capable of transporting sand-grade material as bed load would have prevailed.

Early in the winter trade wind season, with sand patches situated distally (for example) to a fixed point, sand migration due to tractive processes operating during higher energy periods would be insufficient to reach the

fixed point. As winter progressed, and the sand patch became more proximal, traction deposits would begin to reach the fixed point in the form of 'starved' ripple trains, preserved as isolated lenticular bedding. As sand supply increased with time, the lenses would become less 'starved' and, therefore, more connected and thicken upwards. Note that the upward change in lenticular bedding type and thickness is **not** related to an increase in energy but instead reflects a relative increase in sand supply over time.

The explanation above, however, contains a paradox: summer trade winds blow at a relatively constant rate (Riehl 1979) - why, therefore, do lenticular bedded mudstone and sandstone layers alternate towards in the upper part of CU microsequences? Firstly, McCave (1971) showed that mud can be deposited at relatively high velocities if the density of suspended mud is adequate. The alternation of mudstone and sandstone layers, therefore, may be more a response to the sediment source type rather than due to large variations in bottom-current velocity. Secondly, evidence from facies 3 and 8 suggests that tidal currents may have enhanced geostrophic currents, although the velocity of the tidal currents acting in isolation would have been insufficient to transport sand-grade material as bed load. It is possible, therefore, that the lenticular sand lenses in facies 1 only actively migrated when any geostrophic current was enhanced by either peak diurnal or spring tidal flow. The combination of: (i) variation in the power of wave orbitals touching the bottom and enhancing shear stress, thus lowering the threshold for sand transport, (ii) unidirectional geostrophic currents, (iii) tidal currents, would suppress any tendency for rhythmicity in the mudstone/sandstone alternation that may be expected if tidal currents were the dominant force.

A question can be raised as to the ability of onshore-blowing trade winds to set up a geostrophic flow of sufficient velocity to transport sand. The following calculations, although only approximate, suggest that trade wind generated geostrophic flow is sufficient to transport sand. Using the analogue of the present day Brazilian coast, an area affected by summer monsoons and winter trade winds, Ratisbona (1976) records steady trade winds of 5 to 10m s⁻¹ over the southern Atlantic Ocean during the winter season. Allen (1982, vol. B, ch.12), in a comprehensive review of shelf meteorological currents noted that the velocity of a current moving offshore is approximately 3.3% of the onshore surface wind velocity measured 10m above the air/ocean interface. Thus, current velocities in the order of 0.165 to 0.33m s⁻¹ may be expected for winds of 5 to 10m s⁻¹. Using the depth-velocity-grain size diagram of Rubin and McCulloch (1980) for fine sand in water depths of 4 to 20m, a current velocity in the order of 0.35m s⁻¹ would be required in order to transport sand in the form of current-ripples. Given that the superimposition of oscillatory currents will lower the

threshold for sediment movement (Komar 1976), and the fact that much of the sand is of very fine sand and coarse silt grade, it is apparent that the trade wind induced current would be sufficient to transport sand during periods of higher wind velocities, resulting in the production of lenticular bedding. Furthermore, any tidal current component that may have been present would also have served to enhance sand transport thresholds, possibly accounting for the rare fining-upward and coarsening-fining-upward cycles recorded from the Lynton Formation (cf. Terwindt 1981).

The palaeogeographic reconstruction of Tsien (1989) shows that Variscan geosyncline was connected at its eastern end to the 'Proto Tethys Ocean' and 'Uralian Seaway'. Authors such as Klein & Ryer (1978) and Slater (1986) have shown that tides could have been generated in ancient epeiric and mioclinal shelf seas, e.g. Cretaceous Western Interior Basin of the USA where Elliott (1986b) has presented unequivocal evidence for the operation of a diurnal tidal régime, whether or not the seaway was connected to the open ocean. It is reasonable, therefore, to suggest that a weak tidal current may have been present during the deposition of the 'lower-middle mega-facies' of the Lynton Formation.

CU microsequence cyclicity could alternatively be explained as being entirely the result of seasonal variability in the strength of trade winds in the absence of any monsoonal seasonal influence. Trade winds are strongest during the winter season when high pressure cells centred 30° N/S of the equator are best developed (Parish 1982). The base of the CU sequence could represent deposition during the Devonian southern summer when the trade winds would have not been sufficiently strong to generate a geostrophic flow capable of transporting sand-grade material; flow velocity would have been sufficient, however, to transport and deposit mud grade material cf. the mechanism described by McCave (1971). With the onset of the winter increase in trade wind strength, geostrophic flow would have increased to the point where sand-grade material would have been transported as bed load, producing the sandstone lenses characteristic of the upper part of CU microsequences, the lenses becoming more connected upwards with increasing proximity of a source of sand-grade material (as described above).

Although a detailed interpretation for the sequence of physical and biogenic structures preserved in the CU microsequences, based on the monsoonal / trade wind climatic variation influencing the transport of sand grade material, is alluring in a palaeoclimatic sense, a note of caution is necessary. It is possible (probable?) that the CU sequences are solely the product of increasing sand body proximity for a given point, where the

sand body was transported under a predominantly unidirectional geostrophic flow régime induced by trade winds in a climate lacking any seasonal monsoonal influence. Although tidal current path separation can result in unidirectional tidal flow patterns locally (e.g. Stride 1982), tides are ruled out as being the sole source of unidirectional transport in the Lynton Formation deposits due to the relative paucity of features diagnostic of generation by tides (see section 4.2.4) - although tidal currents may have enhanced meteorologically-induced geostrophic flow. In summary, although the rôle of trade wind induced unidirectional geostrophic currents in dictating the proximity of sand patches for a given point appears to be critical to the genesis and preservation of the CU microsequences (whilst the presence of monsoonal seasonal variation is not), the precise interpretation of the meteorologic and tidal factors influencing the generation of the unidirectional currents must remain equivocal. **No monsoonal seasonal variation is proven by Lynton Formation CU microsequence cyclicality.**

4.2.3 Facies 2 - Graded Mudstone

This facies only occurs in the A39 road section, where it is shown as facies B on the log (text-figure 4.1).

4.2.3.1 Description

Two to 30 cm. thick units of silt-poor, normally graded mudstone constitute facies 2. Individual units display sharp, planar bases which are locally erosive; bed tops are planar and show a rapid gradation into the succeeding unit. The mudstone exhibits a good tectonic cleavage, bioturbate textures being very rare and, if present are restricted to a thin, laterally restricted zone at the top of beds. Facies 2 occurs with both lenticular bedding of facies 1 and the parallel-laminated sandstones of facies 11 (plate 4.13B).

4.2.3.2 Biofacies

The erosive nature of unit bases and absence of bioturbation, with the exception of rare laterally restricted zones at the top of some beds, indicates that the majority of facies 2 should be assigned to Schäfer's (1972) lethal-lipostrate biofacies, although bottom conditions became sufficiently oxygenated locally to support occasional bioturbation of bed tops in zones which are assigned to Schäfer's vital-lipostrate biofacies.

4.2.3.3 Interpretation

Several authors have recorded graded mudstones, of the character described above, from shallow marine sequences. Goldring and Langenstrassen (1979, p.84) described: “. . . sharp-based, graded mudstones and siltstones in which a fauna is virtually absent (letal pantostrate)” as common components of Devonian open shelf and shoreface clastic facies, although they noted that the mode of genesis was unclear.

Wignall (1989) described graded mudstone horizons from the Jurassic shallow marine Kimmeridge Clay of Dorset. The graded mudstones are common constituents of the overall Kimmeridge Clay sequence and have sharp planar or slightly erosive bases and flat occasionally bioturbated tops, organic debris increasing upwards within the beds. Wignall ascribed a “tempestite” origin to these beds, indicating that the beds were the product of rapid deposition from a waning flow of storm origin.

Examples of graded mudstones have also been recorded from modern shallow marine sequences. Morton (1981) described graded mud horizons within cores recovered from the Gulf of Mexico and the North Sea. Morton suggested that the graded muds were a more distal equivalent of proximal sandy storm units, stating that the: “. . . rapid deposition of storm beds is commonly indicated by homogeneity of mud and lack of burrowing” (p. 387).

Aigner and Reineck (1982, p.193) described: “Sequences of unbioturbated mud that sharply cut through underlying bioturbation, and just show minor post-event bioturbation, towards the top, are interpreted as pure mud tempestites” further suggesting that: “They seem to represent an extremely distal end member of storm sedimentation.”

The similarity in characteristics of the graded mudstone units of facies 2 and graded mud units recorded from both ancient and modern shelf sequences attributed to storm event deposit origin (i.e. a class of mud “event deposit” cf. Seilacher 1982a) suggests a similar mode of genesis - the storm being a tropical cyclone in the case of the Lynton Formation examples. The occurrence of facies 2 throughout the A39 road section, however, indicates that the graded mudstone should not be regarded as a distal (in the sense of proximity to shoreline) end member of storm sedimentation cf. Aigner and Reineck, and Morton (*op. cit.*) A more

plausible explanation is that facies 2 is the product of a storm event sourced from erosion of a zone of silty-mud deposits rather than sands.

A second hypothesis is offered: the graded mudstone is an event deposit triggered by syn-sedimentary tectonic movement initiating slumping and sliding on the shelf which resulted in the suspension of large amounts of fine grained material that would have settled out to form a mud blanket of the type preserved by facies. The local scouring at the base of the mud deposit is attributable to localised currents generated in response to the downslope translation of large amounts of sediment.

N.B. The possibility that facies 2 represents a mud blanket deposited solely from a plume of turbid river flood water overriding denser (saltier) water (i.e. "hypopycnal flow" of Bates 1953) is discounted. Although a hypopycnal plume could account for the supply of mud, a hypopycnal plume origin is not consistent with the laterally extensive planar, locally erosive, base of facies 2.

4.2.4 Facies 3 - Cross-Bedded Granule Conglomerate

Facies 3 is shown as facies C on both the Watersmeet (enclosure 7) and Duty Point (text-figures 4.2 - 4.5) logs. In addition to these two localities a 75cm thick unit was observed within the 'lower-middle megafacies' in a track-side exposure east of Lynmouth (7268 4923). Thin horizons of the 'Watersmeet lithotype' have been recorded elsewhere within the Lynton Formation [see section 4.1.1(i) and enclosure 2].

4.2.4.1 Description

This facies comprises granule-grade clasts set in a matrix of medium-grade sandstone arranged in decimetre-scale cross-bedded sets. The sets are frequently bounded by mudstone-draped 'pause planes' (*sensu* Terwindt 1981). The following account is based on observations made of the occurrence of this facies in the Duty Point area i.e. a laterally extensive unit low in the Lee Stone section (5.64m on the Lee Stone west log, text-figure 4.2 and 4.5m on the Lee Stone east log, text-figure. 4.3) and the 4m thick sandstone-body that crops out on either side of the East Lyn River 250m east of Watersmeet (enclosure 7).

Lithologically, facies 3 at Lee Stone comprises moderately to poorly sorted sublithic arenites with a ferroan calcite cement. Medium to fine grade, sub-angular to sub-rounded quartz grains are set in a silty (partly

recrystallised) matrix. Lithic fragments are rare in comparison to facies 3 at Watersmeet; some ?pyrite is also present.

At Watersmeet the lithofacies comprises a moderately to poorly sorted (although it is *on average* better sorted than at Lee Stone), locally well sorted where foresets are well organised, finely crystalline sandy dolomitized microsparite (classification of Folk 1959) containing abundant bioclasts and some fine, subangular, detrital quartz. The spar is localised and may represent secondary infilling of fractures and small solution features, the microspar presumably originating as a recrystallised primary micrite. Bioclasts are dominated by bryozoans and echinoderm/crinoid debris, much being relatively coarse. Occasional pyrite and opaque heavy minerals are present.

In both cases fresh surfaces are medium grey in colour, weathering to dark brown due to decalcification. Grains range in size up to granule-grade and are set in a moderately sorted matrix of medium-grade sandstone (fine sandstone grade locally), always with a matrix-supported texture. Sandstones can be locally well sorted where either foresets are well developed or thin zones of parallel-lamination occur. Digestion in hydrochloric acid of three samples each from Lee Stone and Watersmeet yielded a mean value of 44.6% ($\sigma_{n-1} = 3.9$) and 78.7% ($\sigma_{n-1} = 6.8$) carbonate by weight respectively. A mudstone draping an erosion surface within facies 3 at Watersmeet yielded 27.7% carbonate by weight.

Visual examination of facies 3 revealed:

- Silty-mudstone intraclasts (plate 4.12C).
- Well rounded quartz (coarse sand to granule grade).
- Crinoid ossicles.
- Disarticulated brachiopod and bivalve valves (particularly abundant at the eastern side of Lee Stone at the top of the facies - see plate 4.7A).
- Phosphatic bone fragments.

Thin section examination of facies 3 revealed:

- Quartz grains containing vacuoles, epidote inclusions and etched surfaces.
- Polycrystalline quartz grains.

- ❑ Micaceous quartzite that has been pressure welded.
- ❑ Crystal lithic tuff containing microlites of feldspar.
- ❑ Volcanic rock fragments with an authigenic dolomite cement.
- ❑ Phyllite with quartz, epidote and muscovite micas.
- ❑ Siltstone clasts containing polycrystalline quartz grains and muscovite.
- ❑ Crinoid ossicles which, at Watersmeet, frequently display partial dolomitisation.
- ❑ Phosphatic bone fragments, some with concentric growth rings visible.
- ❑ Coral/?bryozoan fragments.
- ❑ Authigenic dolomite rhombs.

Of particular significance is the preservation of articulated sections of crinoid stems within cross-bedded units at both Lee Stone (plate 4.7B) and Watersmeet where a 2cm long section comprising 15 articulated ossicles was recorded (text-figure 4.13). Mudstone intraclasts, up to 2.5cm in length, frequently rest on foreset surfaces (text-figures 4.12 & 4.13), particularly in cross-bedded sets immediately overlying mudstone-draped pause planes e.g. 4.5m on the log of eastern Lee Stone (text-figure 4.3). There is also evidence that mud intraclasts caused a localised obstruction to flow (plates 4.11B, 4.12C and text-figure 4.12).

The facies has a sharp planar base, lying above beds of facies 1 at both Lee Stone and Watersmeet, and locally infills biogenic tubes, particularly of the 'mantled' tube type (see appendix B) in the top of facies 1 at Lee Stone. Small scours, 2cm to 30cm in diameter and up to 4cm in depth (e.g. plate 4.7A & text-figures 4.10 & 4.11) occur occasionally at the base of the facies. At Lee Stone the top of the facies is generally gradational, with biogenic churning introducing mud to a depth of 5 to 8cm below the top of the facies (plate 4.7C). Small concave-up symmetrical scours cut the top of the facies locally, individual scours ranging from 10 to 25cm in diameter and 4 to 7cm in depth. Elsewhere, the top of the unit has been reworked by waves and currents to give a zone of well sorted, ripple cross-laminated fine sandstone (plate 4.7C). Towards the eastern side of Lee Stone, the top of the facies is cut by two major features, a channel-like form and a scour, which are described in detail below. At Watersmeet the top of the facies is sharp and planar and draped by a 6 - 8cm thick mudstone which has only been bioturbated locally where sand-grade material was introduced.

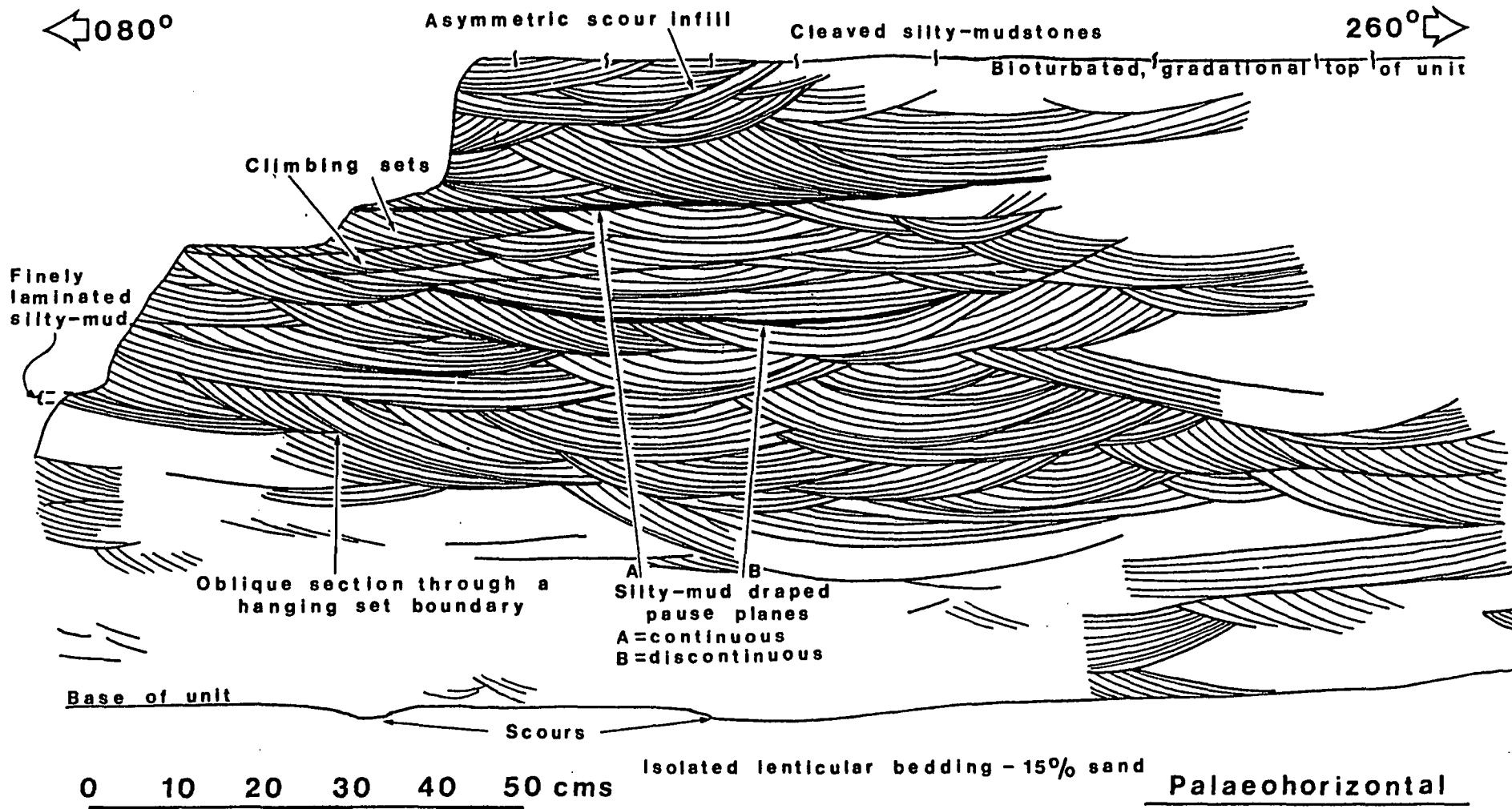
At Lee Stone facies 3 attains its thickest development (approx. 90cm) at the eastern side of the exposure and thins westwards, the unit pinching out completely at the western side of Lee Stone whilst at Watersmeet the unit retains a reasonably constant thickness of 4m over the width of the outcrop. Three stratification types characterise facies 3; these are described below:

4.2.4.1.1 Trough Cross-Bedding

Trough cross-bedding is the most common bedding type within facies 3. Sets range from 2cm to 11cm in thickness and have sharp, erosive, concave-up lower bounding surfaces in sections viewed normal to palaeocurrent direction ('festoon bedding' - see text-fig. 4.10). Set bases are scooped to sub-planar when viewed in sections parallel to the palaeocurrent. The bounding surfaces to sets are usually sub-parallel, but may also diverge down-palaeocurrent (rarely converging) to give wedge shaped sets (cf. de Raaf & Boersma 1971, Anderton 1976, Levell 1976b) in the order of 20 to 80cm in length, with persistence ratios (visible lateral extent / mean thickness, Anderton 1976) of 5 to greater than 20. Hanging set boundaries (cf. Allen 1973, Levell 1980b - see text-figure 4.12) and reactivation surfaces (cf. Collinson 1970) were observed only rarely.

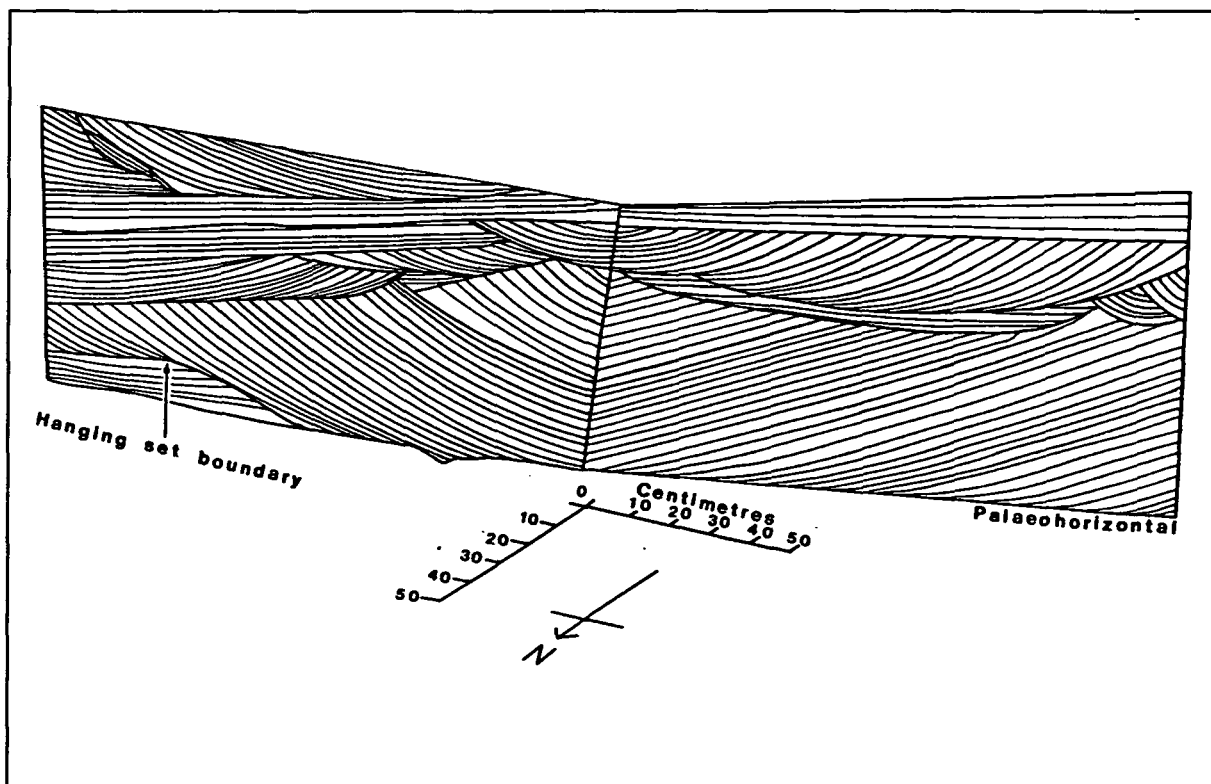
Measurement of foreset thicknesses at the lime kiln quarry east of Watersmeet revealed that sets range from 2 to 11cm in thickness ($\bar{x} = 5.29\text{cm}$, $\sigma_{n-1} = 2.36\text{cm}$, $n = 31$) and average foreset spacing range from 2 to 6mm ($\bar{x} = 3.8\text{mm}$, $\sigma_{n-1} = 1.12\text{mm}$, $n = 37$), the thicker foresets corresponding with the coarser sediment grades. Sorting within an individual set is generally poor. Foresets are concave-up (plate 4.8A), occasionally passing into planar bottomsets at their base, but more commonly infill scour troughs. The infill to these scours, when viewed in sections normal to palaeocurrent, is normally concordant, although asymmetric infills have been recorded (text-figure 4.10). Individual laminae are picked out by differential weathering, giving a ribbed appearance, indicating that each laminae is graded. No grading down foreset surfaces towards the foreset toe was observed.

Occasionally, foresets within a single set are observed to be oversteepened and convex-up (plate 4.11C, text-figure 4.12), resembling the metadepositional deformation of cross-bedding by down-slope translation described by Allen (1982, vol. B, ch. 9).



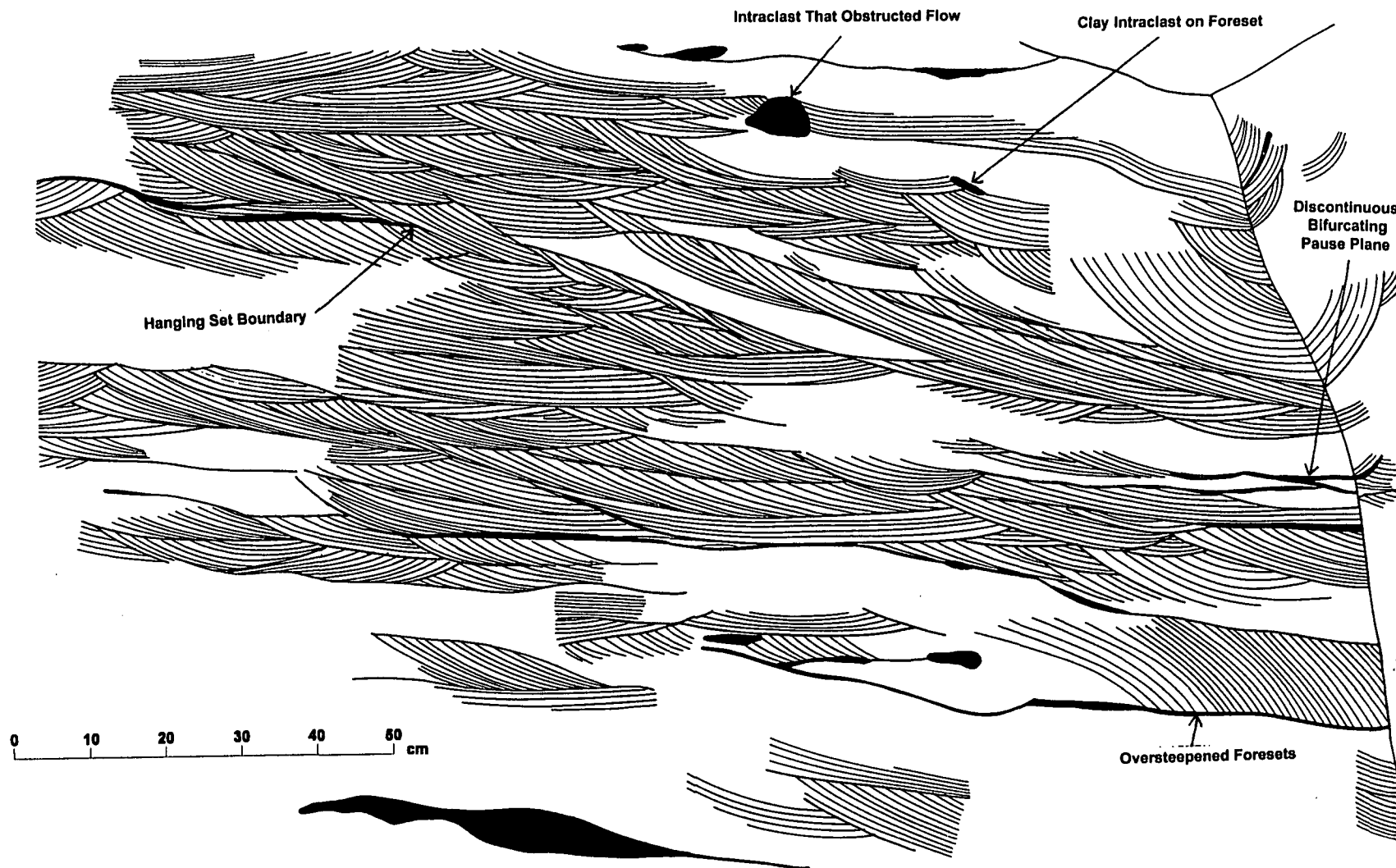
Text-fig. 4.10 Cross-bedding in facies 3 at Lee Stone

Diagram taken from field sketch of the entire thickness of facies 3 at eastern Lee Stone (grid reference 6956 4969). The face is normal to the palaeocurrent trend. Note the 'festoon' nature of the cross-bedded sets.



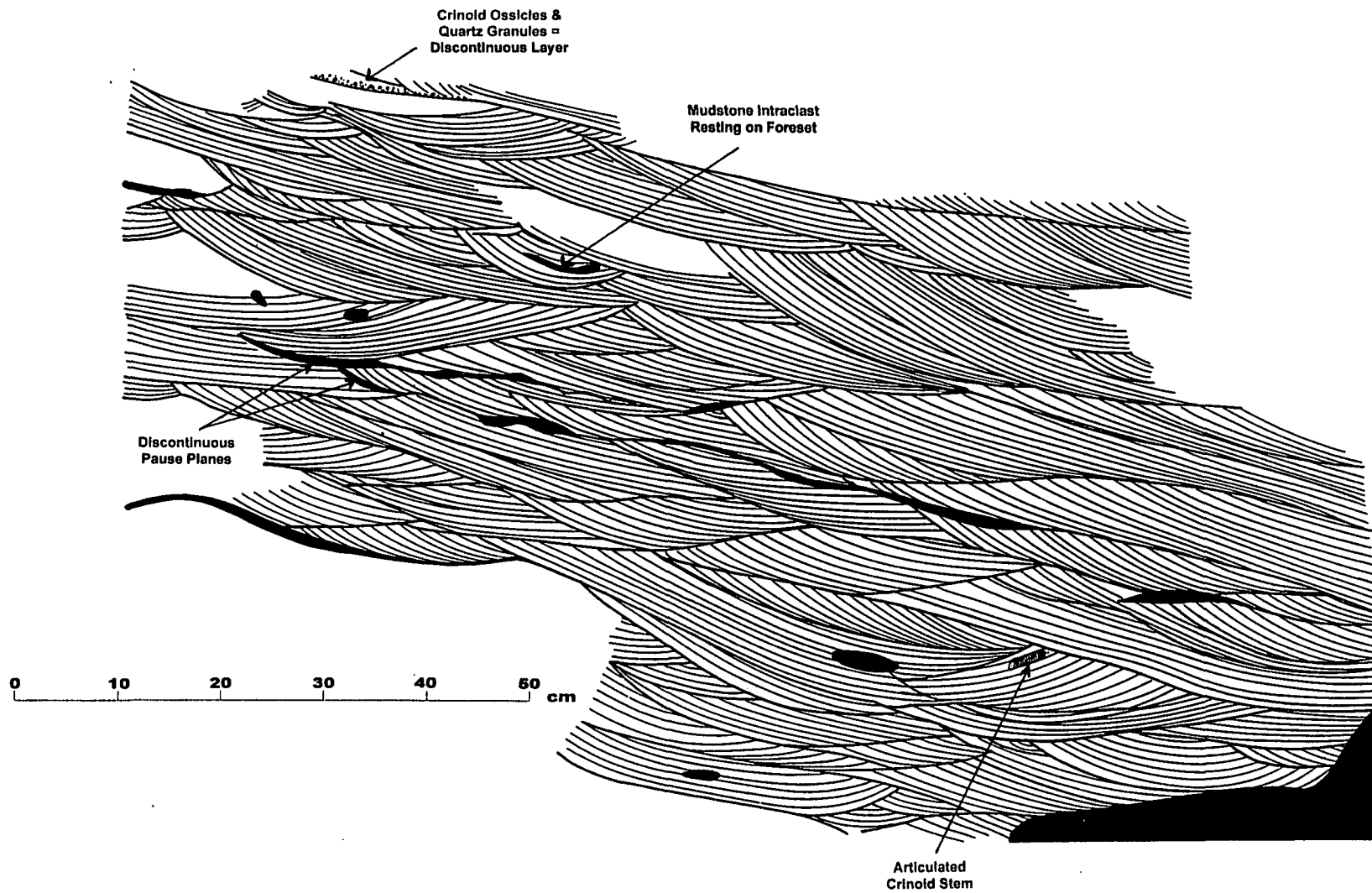
Text-fig. 4.11 Cross-bedding in facies 3 at Lee Stone

Diagram taken from a field sketch of the complete thickness of facies 3 (grid reference 6947 4970). The lower two-thirds of the unit comprises a planar cross-bedded set which migrated towards the SE (obliquely palaeo-offshore). Note the asymptotic toes to the foresets as they meet the planar base of the unit which only shows local, minor scouring. The cross-bedded sets in the upper part of the unit display the 'festoon' lamination style characteristic of trough cross-bedding - note the erosional, scooped bases to the trough cross-bedded sets. Also note the presence of a hanging set boundary and the parallel-laminated sets.



Text-fig. 4.12 Cross-bedding in facies 3 near Watersmeet

Diagram taken from a field sketch of facies 3 (grid reference 7465 4881). Detail of frame A on text-figure 4.16 - for orientation of face, see text-figure 4.16.



Text-fig. 4.13 Cross-bedding in facies 3 near Watersmeet
 Diagram taken from a field sketch of facies 3 (grid reference 7465 4881). Detail of frame B on text-figure 4.16 - for orientation of face, see text-figure 4.16.

Trough cross-bedding palaeocurrents at Lee Stone have a low variance (see text-fig. 4.14) and are unimodal, dipping towards the SSE. Reversed sets are rare, although 'herringbone' cross-stratification (cf. Reineck 1963) is occasionally observed. The NNW directed sets are always thinner than adjacent SSE directed sets. Palaeocurrent variation between vertically adjacent sets is frequently greater than 60°.

At Watersmeet, the vast majority of cross-bedding is of 'trough' type and no attempt was made to distinguish between 'trough' and 'planar' bedding when palaeocurrents were measured. The results are shown in text-figure 4.15, where the vector strengths are shown on a map on which the line of the Lynmouth - East Lyn Fault has been plotted. It can be seen that the cross-bedding at the two localities closest to the fault have a lower variance (i.e. stronger vectors) than those further away. Furthermore, the mean palaeocurrent for the two localities closest to the fault flowed SSE down a SSW-dipping palaeoslope, whereas the mean palaeocurrent for the two localities furthest from the fault flowed ESE along the strike of the palaeoslope.

4.2.4.1.2 Planar Cross-Bedding

Planar cross-bedding is rare at Watersmeet; at Lee Stone it is generally observed at the base, the top and the lateral margins of the facies (see text-figure 4.11). Sets range from 4cm to 12cm in thickness and are bounded by sharp, sub-parallel surfaces which occasionally converge down-palaeocurrent (rarely up-palaeocurrent) to give wedge-shaped sets (plate 4.7A) which are 30cm to 80cm in length, with persistence ratios of 5 to greater than 25. Foreset angle may decrease down-palaeocurrent as the set thickens (cf. Boersma & Terwindt 1981; plate 4.7A). Lower bounding surfaces of sets are planar and non-erosional, rarely erosional (text-figure 4.11). Occasionally, broad shallow scours, up to 10cm wide and 1cm deep, occur at the set base. Reactivation surfaces and hanging set boundaries (text-figure 4.11) occur infrequently.

Sorting within these sets is generally moderate to good i.e. better sorted than in the trough cross-bedded sets. Individual foreset laminae range in thickness from 1mm to 5mm and are planar with an average dip of 19° (31°, max, N = 51, $\sigma_{n-1} = 6.4$) at Lee Stone; foreset toes are asymptotic. Differential weathering picks out foreset laminae and attests to grading within individual laminae. However, it has not been possible to ascertain whether the grading is normal or reverse. Silty-mudstone intraclasts are less frequent than in the trough cross-bedded sets and tend to occur in sets immediately above silty-mudstone draped pause planes.

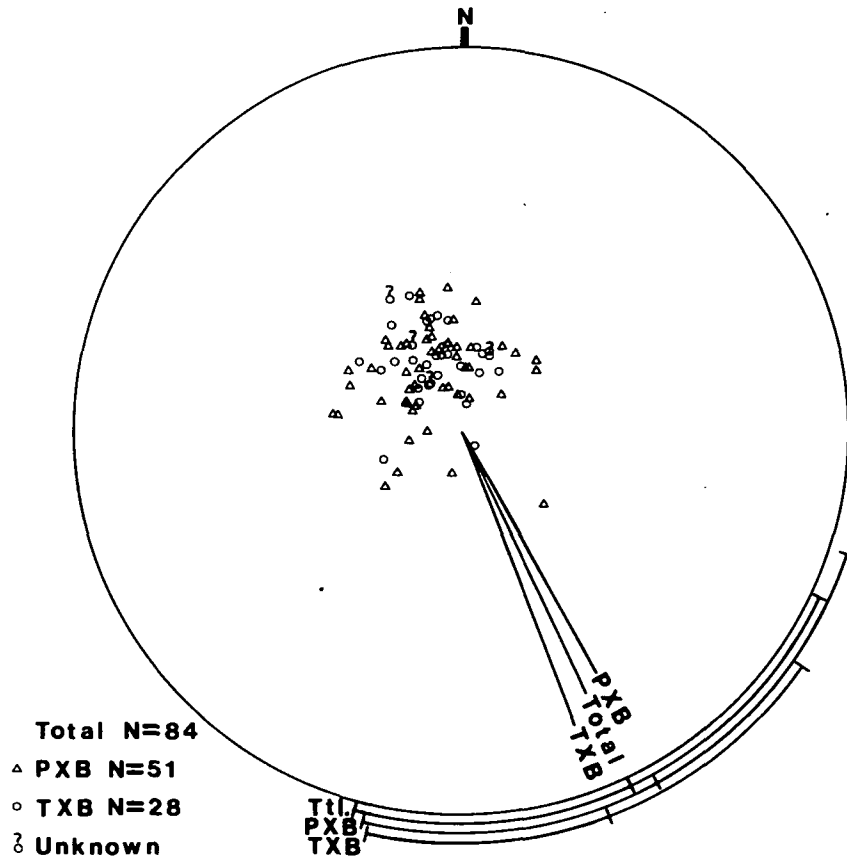
Soft sediment deformation structures have been recorded from the tops of some planar cross-bedded sets, with forest tops becoming slightly oversteepened in the highest 1cm to 2cm (plate 4.7A).

At Lee Stone palaeocurrents again have a low variance (text-figure 4.14), exhibiting an essentially unimodal distribution directed towards the SSE. Of the measured foreset dip directions, only one foreset records migration that was in a NNW direction. Furthermore, on examination of the exposed portion of facies 3, 'herringbone' cross-stratification is only rarely observed, NNW-directed sets always being thinner than adjacent SSE-directed sets.

4.2.4.1.3 Parallel-lamination

The least frequent of the three primary sedimentary structure types comprising facies 3 is parallel-lamination (text-figure 4.11) which has only been observed at Lee Stone. The laminae occur in sets up to 10cm in thickness, with individual laminae displaying normal grading and ranging from 5mm to 10mm in thickness. Lamina sets are occasionally wedge-shaped (text-figure 4.11). Convex-up shells occur within some lamina sets. Parallel-laminated sets generally occur where facies 3 attains its greatest thickness, being concentrated in zones near the base and near the top of the facies.

~



| | |
|-------------------------------------|-------------|
| Mean Vector Direction: Total | 155° |
| PXB (Planar Cross-Bedding) | 151° |
| TXB (Trough Cross-Bedding) | 160° |

| | |
|----------------------------------------------------|--------------|
| Estimate of Spread of Angular Values: Total | 75.5% |
| PXB | 72.0% |
| TXB | 82.3% |

| | |
|--------------------------------------|-------------|
| Mean Angular Deviation: Total | ±40° |
| PXB | ±43° |
| TXB | ±34° |

| | | |
|---------------------------------------------|-----------------------------|---------------------|
| Rayleigh Test of Significance: Total | <10⁻²⁰ | ∴ Non-random |
| PXB | <10⁻¹⁰ | |
| TXB | <10⁻⁵ | |

Text-fig. 4.14 Lee Stone: Facies 3 - Foreset Vector Means

Text-fig. 4.15 Watersmeet: Facies 3 - Foreset vectors in relation to the Lynmouth - East Lyn Fault

Palaeocurrent vectors (width of line proportional to strength of vector) plotted on a field map showing the line of the Lynmouth - East Lyn Fault - tip of arrow-heads marks the point at which palaeocurrents were measured.

Lime kiln quarry (grid reference 7464 4862):

$$\mu_0 = \text{Mean vector direction} = 90^\circ$$

$$L = \text{Estimate of spread of angular values} = 26.2\%$$

$$S = \text{Mean angular deviation} = \pm 70^\circ$$

$$\text{Rayleigh test of significance} = < 0.05 \therefore \text{non-random}$$

N side East Lyn River, opposite lime kiln (grid reference 7462 4855):

$$\mu_0 = 168^\circ$$

$$L = 70.8\%$$

$$S = \pm 44^\circ$$

$$\text{Rayleigh test of significance} = < 10^{-10} \therefore \text{non-random}$$

Watersmeet (grid reference 7440 4864):

$$\mu_0 = 101^\circ$$

$$L = 56.7\%$$

$$S = \pm 53^\circ$$

$$\text{Rayleigh test of significance} = < 10^{-4} \therefore \text{non-random}$$

Old bridge, S of Chislecombe Bridge (grid reference 7444 4878):

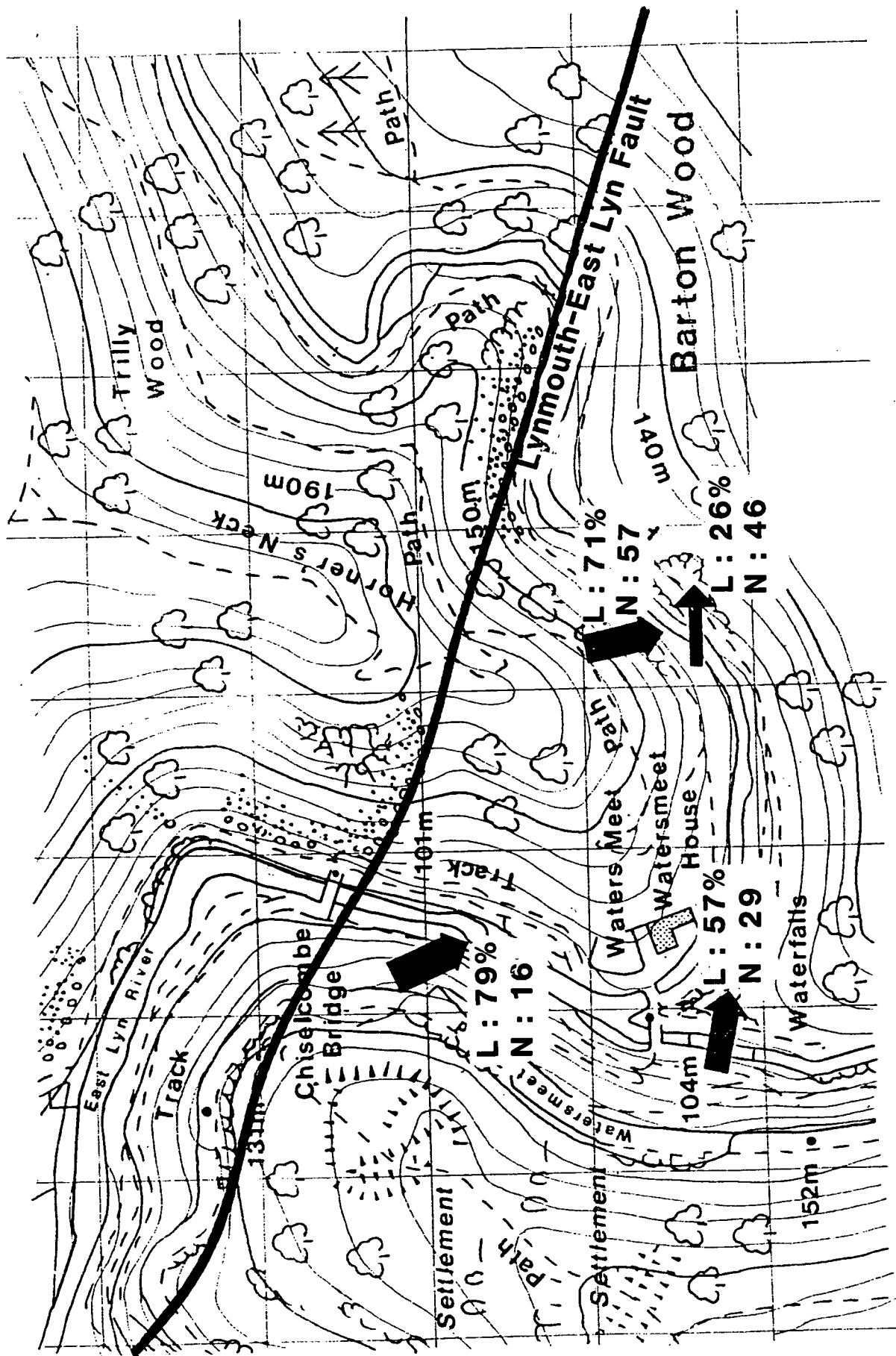
$$\mu_0 = 156^\circ$$

$$L = 79.3\%$$

$$S = \pm 37^\circ$$

$$\text{Rayleigh test of significance} = < 10^{-4} \therefore \text{non-random}$$

The preserved palaeocurrent pattern strongly suggests that the Lynmouth - East Lyn Fault presented a SSW-facing scarp feature at the time of deposition and that a geostrophic flow influence was present - see text-figure 4.9 for comparison. Palaeo-on-shore trade winds would have 'piled' water against the fault scarp resulting in a palaeo-off-shore counter-flow at depth which would have been progressively deflected by the Coriolis force towards the left in the palaeo- Southern hemisphere. Thus, expected palaeocurrents would be obliquely off-shore adjacent to the scarp, becoming scarp-parallel further away from the scarp.



Facies 3 is internally subdivided by a series of planar 'pause planes' (*sensu* Terwindt 1981, 'structural diastems' using the terminology of Boersma 1969) which may be either erosional or non-erosional. These planes are usually, but not universally, draped with finely laminated silty-mudstone 1mm to 30mm in thickness. Silty-mudstone intraclasts frequently lie on pause plane surfaces (plates 4.11B & 4.12C). Three types of pause plane are recognised (cf. Terwindt *op. cit.*):

(i) Concordant, non-erosive drapes of mudstone, 1mm to 4mm in thickness, on foreset surfaces.

(ii) Discontinuous planar surfaces which are either erosive or non-erosive. These surfaces have a lateral continuity of up to 70cm and are frequently draped with silty-mudstone laminae (text-figure 4.10, 4.12, 4.13 & plate 4.12B). Where surfaces of this type terminate, the overlying cross-stratified set 'steps-down' to produce larger scale forests (i.e. the hanging set boundary type noted above - see text-figure 4.12). Occasionally, erosion surfaces can pass laterally into zones of wave-rippled fine sandstone (plate 4.14A) or, where non-erosive, mudstones can drape and preserve wave-ripple profiles at the top of the underlying cross-bedded set. It should be noted at this point that the discontinuous pause planes are always the result of subsequent scour/truncation of originally continuous pause planes. A granule lag comprising well rounded vein quartz and crinoid ossicles was observed to rest on a pause plane at Watersmeet (text-figure 4.13).

(iii) Laterally extensive pause planes (plate 4.11A) which may be traced for over 30m (text-figure 4.16, correlation between the three logged sections shown on enclosure 7) and are always erosive, although relief never exceeds 6cm. A silty-mudstone drape, up to 30mm in thickness, is sometimes present (text-figure 4.10). No pebble lags have been recorded beneath these surfaces. The surfaces are generally sub-parallel and lie close to palaeo-horizontal, although several surfaces in the lime kiln quarry east of Watersmeet dip at up to 12° towards SSW (see text-figure 4.16). These laterally extensive surfaces can laterally intergrade between three types: erosive breaks between cross-bedded sandstone sets / mudstone-draped erosion surfaces / wave or combined-flow rippled surfaces.

Text-fig. 4.16 Fence diagram of the old quarry above the lime kiln showing development of pause planes in facies 3.

250m east of Watersmeet - grid reference 7465 4881. The unmarked areas on the lower part of each face are covered in soil, loose stone and vegetation, as is the zone towards the top left of face 5 where the broken lines represent the extrapolation of pause planes.

Face trends:

Face 1 (furthest left) = 080° - 260°

Face 2 = 030° - 210°

Face 3 = 100° - 280°

Face 4 = 150° - 330°

Face 5 = 090° - 270°

Face 6 = 010° - 190°

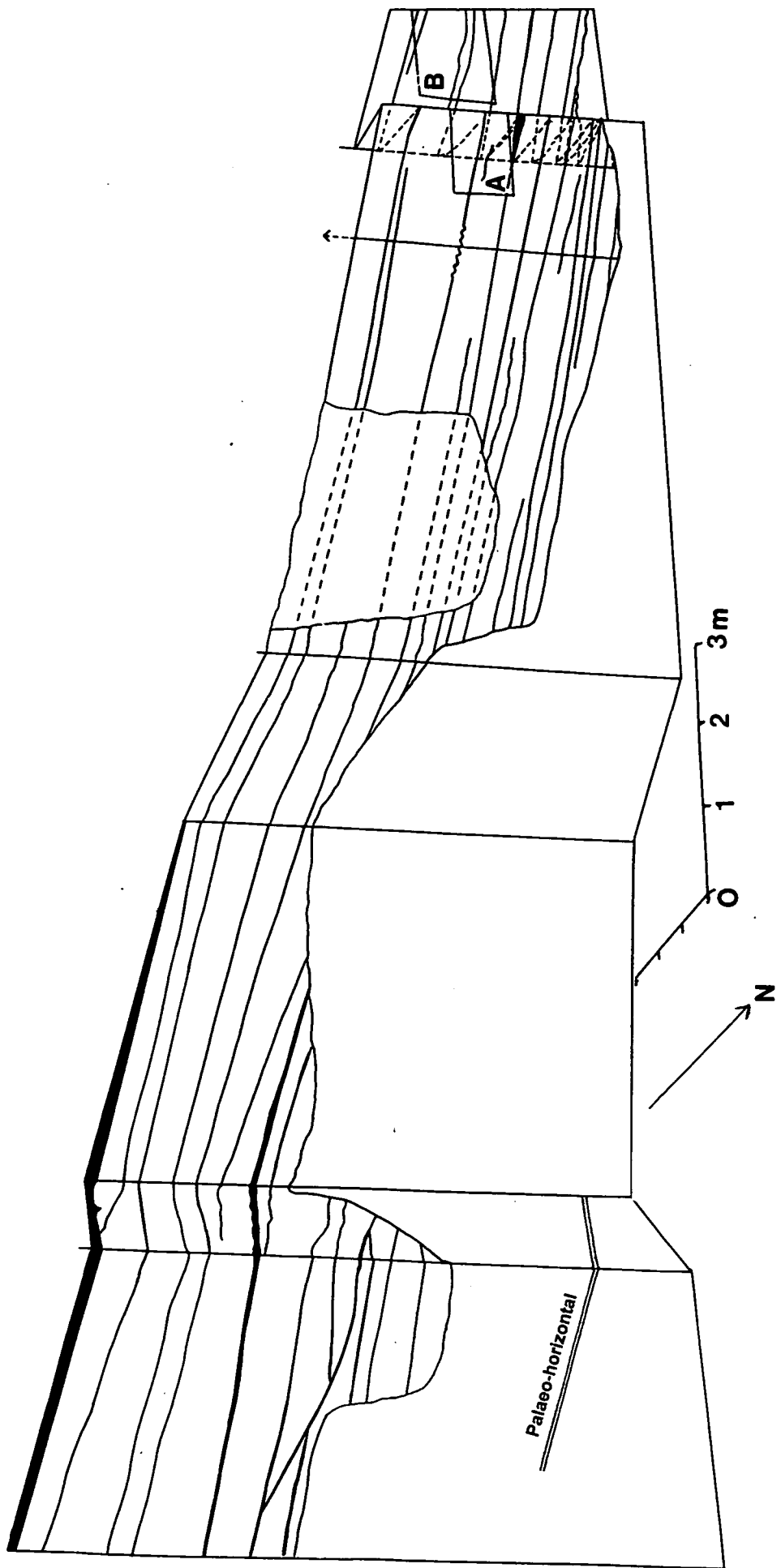
Face 7 (furthest right) = 080° - 260°

The fence diagram has not been corrected for regional tectonic dip which is c. 10° towards the WNW. The dip of the sub-parallel pause planes is so close to the regional tectonic dip as to be indistinguishable in most cases i.e. the pause planes were near palaeo-horizontal. A few pause planes located near the intersection of faces 1 & 2, however, indicate pause plane dips of up to 12° towards the SSW.

The solid vertical line, and broken continuation, towards the right centre of face 5 marks the line of the logged section shown in column A on enclosure 7. The 3rd pause plane down shows a crinkled line where it crosses the line of the logged section - the crinkling represents a zone of wave-ripples preserved beneath a muddy-siltstone drape.

The thick black band at the top of faces 1, 2 & 3 is a mudstone which is laterally equivalent to the thick mudstone unit above the 'Watersmeet lithotype' on the two logged sections north of the East Lyn River (logs B & C on enclosure 7) and can be seen to 'pinch out' in a westerly direction.

Field sketches of the cross-beds enclosed in frames A & B are shown in text-figures 4.12 & 4.13 respectively.



At the eastern end of Lee Stone, the top of facies 3 is cut by a major channel-like scour surface (plate 4.9). The surface is erosive and cuts through the whole unit, which has a local thickness of 45cm to 60cm, penetrating to lenticular bedding (facies 1) beneath. The resultant scour is 10m in width, symmetrical in shape, and is floored by lenticular bedding of the unit immediately underlying facies 3. The depth to width ratio of the scour is approximately 1:23. Unfortunately, exposures do not permit the determination of the three-dimensional extent of the scour and it cannot be unequivocally described as channel-like. If the scour was of a channel origin, the channel would have had an **approximately** N-S orientation i.e. normal to the palaeoshoreline.

The scour has a concordant fill of lenticular bedding (facies 1) which thickens towards the centre of the scour i.e. individual beds thicken when traced from the margin towards the centre of the scour. However, the top of the scour is erosional and there is an angular discordance between the lenticular bedding (plate 4.9). At the western side of the scour, the top of the deposit has been locally reworked, a thin stringer of conglomerate having locally migrated over the top of the fill (plate 4.9).

At the extreme eastern side of Lee Stone, a section parallel to palaeocurrent direction is available through facies 3. At this locality a 4.5m wide asymmetrical scour with a concave-up base has been cut to a depth of 25cm into the top of facies 3 (plate 4.10). On the up-palaeocurrent side of the scour, a sharp knick-point was formed at the point of scour initiation, whilst the downstream margin of the scour forms a more gentle angle with the top of the unit. The initial 15cm of the scour infill comprises a concordant drape of lenticular bedding which may be traced without break from the scour to a wedge of cross-bedded sandstone resting on top of the unit. No change in thickness is observed in the lenticular bedded drape, and therefore the shape of the scour-form surface has been replicated on the upper surface of the lenticular bedded drape. The topmost part of the scour infill consists of up to 25cm of well sorted, fine- to medium-grade sandstone which weathers to give a flaggy appearance. The flaggy sandstones are interpreted to be the toesets of a major dune which must have been several metres in height judging from the scale of toesets, the majority of the dune having been subsequently removed by penecontemporaneous erosion; the scour itself would represent the scour trough of the major dune. The toesets are concave-up and dip 10° to 15° towards the south, the toesets on the up-palaeocurrent side of the scour almost paralleling the scour surface and passing down-palaeocurrent into ripple cross-laminated sandstones with mudstone drapes. At the down-palaeocurrent side

of the major scour the foreset toes fill smaller shallow scours cut into the underlying heterolithic fill. A series of silty-mudstone-draped pause planes occur at 3cm, to 10cm intervals within the foreset cosets, subdividing the fill into a series of form-concordant wedge-shaped sets which thin down-palaeocurrent. At the toe of some of the toesets the sands immediately beneath the 'pause planes' display structures similar to the "crinkly bedding" described by Terwindt (1981). The scour fill is completed at the down-palaeocurrent side of the scour by a final wedge of toesets passing into lenticular bedding. The top of the scour fill is marked by a laterally extensive erosion surface. The scour infill closely fits the descriptions of tidally-generated foreset and toeset bundles given by Terwindt (1981).

4.2.4.2 Biofacies

At first sight, facies 3 conforms well with Schäfer's description of his lethal-lipostrate biofacies: "Characterised by abundant taphocoenoses and by disconformities due to considerable erosion; water is well aerated but continuous shifting of coarse material prevents establishment of a benthonic fauna and biocoenosis. What biogenic remains (heavy valves) can be found come from, neighbouring benthonic biocoenoses" (1972, p.478). The cross-bedding preserved within the granule conglomerate attests to the fact that for much of the time, the facies 3 substrate was mobile and consequently, well aerated. Major erosional discordances bound cross-bedded cosets and indicate that facies 3 was occasionally subjected to periods of very high energy which resulted in nett erosion. Taphocoenoses comprise disarticulated brachiopod and bivalve valves (plate 4.7A), crinoid ossicles and phosphatic bone fragments. The occurrence of an articulated crinoid stalk (plate 4.7B) suggests that in many cases, after initial burial, allochthonous biogenic material was not subsequently reworked. However, the presence of mudstone draped pause planes in facies 3 indicates that periods of active bedform migration were interspersed with periods of relative quiescence when mud was deposited on pause planes.

It was noted in a previous section (4.2.4.1.1) that three types of pause plane occur:

- (i) Concordant, non-erosive mud drapes on foreset surfaces.
- (ii) Discontinuous planar surfaces which are either erosive or non-erosive.
- (iii) Laterally extensive, erosive pause planes.

Bioturbation of the mudstones on type (i) pause planes is very rare, being more frequent in the mudstones on type (ii) pause planes and common in the mudstones on type (iii) pause planes. The bioturbation consists almost exclusively of sandstone-infilled *Palaeophycus tubularis* burrows and occasional zones of fossitextura deformativa.

The above distribution in the degree of biogenic reworking of the mud on the various types of pause planes is interpreted as being related to the temporal significance of the individual pause plane types. The type (i) non-erosional, concordant foreset drapes represent only minor breaks in bedform migration, such as those experienced during e.g. the slack water phase between tides. The length of exposure of the mud drape before burial during the subsequent phase of bedform migration would have been insufficient to allow biotic exploitation of the mud. The type (ii) laterally discontinuous erosive and non-erosive pause plane drapes represent a more major break in bedform migration than the type (i) pause planes and reflects events of a lower order of magnitude. Burrowing would be restricted to localised patches exploited by opportunistic forms, time being insufficient for the establishment of a major infaunal population. The type (iii) laterally extensive, erosional pause plane mudstone drapes represent the most major type of break in active bedform migration and reflect events of a similar order of magnitude to the seasonal variation in environmental energy discussed in section 4.2.2.3, the mudstones on the extensive pause planes resulting from deposition during the low energy (monsoon?) season. The major break in active bedform migration during the low energy season is reflected by the fact that the mud drape on type (iii) pause planes is frequently well burrowed.

In conclusion, the physical and biogenic structures preserved in the granule conglomerate of facies 3 suggest that facies 3 should be assigned to Schäfer's lethal-lipostrate biofacies. However, the presence of bioturbated mudstone drapes on pause planes in facies 3 indicates that vital conditions prevailed at certain periods during the deposition of facies 3. Thus, facies 3 pause planes are considered to be attributable to Schäfer's vital-lipostrate biofacies.

Knight (1990a) recovered a diverse icriodid conodont fauna from facies 3 at Watersmeet, but noted that the fauna lacks presumed associated acodinan elements. Of the fauna recorded, *Icriodus culicellus*, a form commonly associated with crinoidal or coralliferous limestones, was interpreted as reflecting shallow, possibly agitated water conditions; *I. cf. wernerii* and *I. retrodepressus* reflect quieter, shallower conditions. Taken as a whole, Knight interpreted the fauna as representing quiet, shallow water conditions. The

imbalance in forms recovered was attributed to current sorting suggesting that the fauna was: "probably derived from a somewhat quieter palaeoenvironmental setting, and transported a ?short distance (based on preservation)" (p.399). The conclusions of Knight are consistent with the interpretations based on sedimentology and ichnofauna presented herein.

4.2.4.3 Interpretation

At Lee Stone, facies 3 attains its maximum thickness near the eastern end of the exposure. When viewed from a direction normal to palaeocurrent the unit thins both to the east and west, pinching-out completely over a distance of 90m to the west. The base of the unit is planar. Thus, facies 3 has a plano-convex geometry in sections normal to palaeocurrent; sections parallel to palaeocurrent through the facies are too limited to make any geometric assessment. At Watersmeet the lithotype retains an average thickness of 4m over the width of available exposures.

Palaeocurrent data (text-figures 4.15 & 4.15) indicate a dominant transport direction towards the SSE (obliquely palaeo-offshore), although exposures at Watersmeet away from the line of the Lynmouth - East Lyn Fault exhibit a wider variation in palaeocurrent direction and have a mean which is towards the ESE i.e. sub-parallel to palaeoslope strike.

An examination of the suite of stratification types in facies 3 suggests that tidal processes were, in part, responsible for the erosion and transport of the facies 3 substrate. If considered individually, the following stratification features observed in facies 3 are not unequivocally diagnostic of influence by tidal processes. However, when the suite of structures is considered in combination, the influence of tidal processes is suggested: wedge-shaped sets, pause planes with mudstone drapes, hanging set boundaries, reactivation surfaces and occasional occurrences of 'herringbone' cross-stratification. The strongest evidence for tidal processes are the toset bundles infilling the scour at the top of facies 3 at Lee Stone (plate 4.10); these bundles are strongly reminiscent of the tidally-generated bundles described by Terwindt (1981).

In summary, it is apparent that facies 3 represents sand-bodies migrating in a down palae-slope direction. The sand-bodies appear to have been active for an appreciable period of time. This is evidenced by the thin stringer of facies 3 material migrating over the top of the lenticular bedded fill of the channel-like scour at

the top of facies 3 at Lee Stone (plate 4.10). The fill comprises thin sandstones, representing an interlude where environmental energy increased, set in bioturbated mudstones, the fill reaching a maximum thickness of 60cm. The many sandstones preserved within the scour suggest that the scour fill must have a significant period of time to accumulate.

In order that a mechanism for the transport of the sand-bodies may be deduced, the nature of the internal stratification is examined below.

The dominant stratification types of facies 3 consist of trough cross-stratification and planar cross-stratification, representing the migration of sinuous-crested and straight-crested dunes respectively. The former match the internal form of 'megaripples' of Daboll (1969), 'type 2 megaripples' of Dalrymple *et al.* (1978) and 'simple dunes', 'dunes' or 'sand waves' of Klein (1970), Harms *et al.* (1974) and Jackson (1976). The latter have internal structures similar to bedforms referred to as: 'sand waves' (Daboll 1969), 'type 1 megaripples' (Dalrymple *et al.* 1978), 'simple sand waves' (Klein 1970), 'diminished dunes' (Smith 1971) and 'transverse bars' (Jackson 1976). Allen (1982, vol. A, ch. 8), however, considers the separation into two distinct bedform types erroneous. A review by Allen of experimental data on dune existence fields revealed a substantial overlap between the two types, indicating that only one hydrodynamic class of bedform was present. Allen referred to the two end members as 'two-dimensional dunes' and 'three-dimensional dunes' (this nomenclature is followed herein), noting that three-dimensional dunes are created under more 'central' conditions, with two-dimensional dunes arising near to either the lower (probably) or upper (unlikely) limit of the dune existence field. This indivisibility of dune bedforms was encountered in facies 3, planar cross-bedded sets frequently displaying incipient scours, a gradation, albeit abrupt, being recognised between planar cross-bedding and trough cross-bedding.

A vast amount of data (summarised in Harms *et al.* 1982) has allowed existence fields for dunes to be plotted. In terms of geometry, dunes formed by relatively weak flows tend to be of the straight-crested two-dimensional type, and those formed by relatively strong flows tend to be highly three-dimensional and irregular in plan view. In facies 3 the two-dimensional dunes are generally observed near the lateral margins of the unit, inferring an increase in environmental energy towards the unit crest at Lee Stone. Two-dimensional dunes are also observed at the base of the deposit at Lee Stone, probably reflecting the inability of lee vortices to erode scour pits in the cohesive mud which underlay the deposit.

Many of the features displayed in facies 3 dunes have been described from tidal settings by previous authors.

Diagnostic features are noted below, along with the interpretation of the original author.

(i) **Sharp Set Boundaries** Klein (1970) recognised that set bounding surfaces in tidally-generated cross-beds are frequently sharp. This was attributed to turbulent scour during periods of strong tidal action.

(ii) **Wedge-Shaped Sets** De Raaf and Boersma (1971), Anderton (1976) and Levell (1980b) have reported the common occurrence of wedge-shaped sets within tidally-generated sequences. De Raaf and Boersma (*op. cit.*) noted that the frequent incidence of sets of a short length, or if they were long, containing pause planes, reflected the bi-directional and intermittent character of tidal currents. The frequent occurrence of silty-mudstone draped pause planes resting on foreset surfaces [type (i) pause planes of section 4.2.4.1.1] was attributed by Boersma and Terwindt (1981) to sedimentation from suspension during tidal slack-water phases.

(iii) **'Herringbone' Cross-Stratification** Reineck (1963) was the first worker to propose 'herringbone' cross-stratification as diagnostic of sediment working by tidal currents. Subsequently, 'herringbone' cross-stratification has become accepted as one of the most diagnostic characteristics of tidal sedimentation. However, the occurrence of 'herringbone' cross-stratification in facies 3 is rare. This reflects the view of Klein (1970) that 'herringbone' cross-stratification tends to form only during the period of transition from spring to neap tides as the depth of reworking with each successive (ebb or flood) tide becomes less. Thus, sets of flood-oriented cross-strata may develop on an ebb-dominated region of a shoal during a spring phase of the tidal cycle and become preserved as the subsequent tides move progressively towards neaps. However, long-term preservation of flood-oriented sets of cross-stratification is probably low because as the shoal gradually accretes, each successive phase of deep-level reworking accompanying the spring phase of the tidal cycle will tend to destroy the flood-oriented set formed during the previous spring phase of the tidal cycle (Bridges 1982).

(iv) **Hanging Set Boundaries** (*sensu* Allen 1973) are common in ancient tidal deposits (e.g. Anderton 1976, Levell 1980b). They represent the downstream coalescence of two or more small-scale sets into a single medium-scale set (Levell *op. cit.*). Hanging set boundaries are only observed rarely in facies 3.

(v) **Reactivation Surfaces** were first described from a fluvial sequence (Collinson 1970), but are common in tidal deposits (De Raaf & Boersma 1971, Banks 1973b, Sellwood 1975, Anderton 1976, Boersma & Terwindt 1981). Klein (1970) attributed the occurrence of reactivation surfaces in tidal deposits to the modification of bedform morphology by the weaker, reversing tidal current. Levell (1980b) recognised that their genesis in tidal sequences is probably polygenetic, some representing convex-upwards set boundaries formed by the eroding lee-side eddy of advancing superimposed dunes, others formed by current reversals, whilst yet others may be due to wave reworking. Again, this type of structure is uncommon in facies 3.

(vi) **Sets With a Low Angle of Climb** Anderton (1976) recorded dune sets, 3cm to 25cm in thickness, in medium- to coarse-grade sandstone displaying planar, sub-parallel, erosional set boundaries. This description accords well with examples observed in facies 3 (text-figure 4.10). Anderton interpreted the climbing dunes as the product of dune migration during a storm-reinforced tide. Nio (1976) has also recorded climbing dunes within a tidal sand-body.

Oversteepened cross-sets have been described from fluvial environments by Allen and Banks (1972) who interpreted them as the product of current drag on liquefied sand, with liquefaction being produced by seismic shaking. The oversteepened foresets at Lee Stone are of a much more restricted occurrence than the examples of Allen and Banks, being confined to the top part of a cross-set. Furthermore, there is no evidence for vertical water escape features which might be expected from seismic shocks. Although seismic shaking could be invoked from activity on the Lynmouth - East Lyn Fault (chapter 2), perhaps a more plausible mechanism is that offered by Dalrymple (1979) who proposed that soft-sediment deformation could be initiated by wave action.

In addition to the nature of cross-stratification inferring a tidal influence in the origin of facies 3 cross-beds [discussed in (i) to (vi) above], the character of the internal discordances within facies 3 also resembles, at least in part, similar structures described from tidal deposits elsewhere. The type of discordance observed in facies 3 have been variously termed 'structural diastems' (Boersma 1969); 'discontinuity planes' (de Raaf & Boersma 1971) and 'pause planes' (Terwindt 1981). Terwindt (*op. cit.*) stated that these planes were the product of ". . . erosional or non-erosional surfaces representing a stand-still phase of . . . megaripple migration during the subordinate tide" (p.11-12). Terwindt recognised three types of pause plane: foreset

drapes [described in (ii) above], limited and extensive, the limited planes representing minor or truncated planes of the extensive type. All three types have been recognised in facies 3.

Low angle, laterally extensive discontinuity surfaces are common in ancient subtidal sand-bodies and have been described by Anderton (1976), Johnson (1977), Hobday and Reading (1972) and Nio (1976). Of particular relevance are the surfaces described by Hobday and Reading ('facies B') and Johnson ('facies 4') in the shallow marine Upper Quartzitic Sandstone Member of the Varanger Fjord area of Finmark, which are late Precambrian in age (referred to as the Skalmeset Sandstones by Hobday & Reading), as these surfaces were not purely the product of erosion but were generated by a complex combination of erosion and deposition. A similarly complex combination of erosion and deposition appears to have operated during the moulding of the facies 3 surfaces.

The discontinuity surfaces described by Hobday and Reading dip at between 6° to 17° and exhibit a slightly sigmoidal shape, enclosing lenses of trough cross-bedded sandstone with a migration direction running obliquely down the major discontinuity surfaces. The surfaces are complex, zones of wave ripples with crestlines oriented oblique to the strike of the major surfaces passing laterally into erosion surfaces which are frequently scoured into by a succeeding set of trough cross-bedding. Hobday and Reading attributed the production of this facies to fair-weather progradation of a sand-body by the migration of three-dimensional dunes, interspersed with periods of storm activity which generated the discontinuity surfaces, wave-ripples forming on these surfaces as the storm abated. The even spacing of the surfaces was attributed to the regular alternation of storm and fair-weather seasons.

Johnson (*op. cit.*) undertook a more detailed study of the Upper Quartzitic Sandstone Member sequences, noting that the regularly spaced surfaces were planar to concave-up, attaining dips of 6° to 15°; wave-ripple crestline trends were either oblique or parallel to the strike of the discontinuity surfaces. In contrast to Hobday and Reading, Johnson proposed that the three-dimensional dunes represented migration during the **highest** energy conditions, whilst the major discontinuity surfaces were the product of waves when unidirectional currents were inactive. Johnson suggested that the concave-up geometry of the surfaces may have been a response to shoaling waves cf. modern shoreface profiles.

The presence of major discontinuities in the dominant coset bounding surfaces within facies 3, produced as a result of scour into the surfaces by three-dimensional dunes migrating obliquely down the surfaces

('hanging-set boundaries'), along with horizons of wave and combined-flow rippling bounding cosets, accords with the interpretation for the Upper Quartzitic Sandstone Member surfaces given by Johnson. The coset boundaries within facies 3 are, therefore, interpreted to represent periods of inactivity in the unidirectional currents which drove bedform migration preserved in facies 3 cross-bedding. Conversely, the cross-bedding represents high-energy conditions, a unidirectional, SSE-flowing current reinforced by a weak (?spring) tidal current (see above for evidence of tidal activity) resulting in thick packets of cross-bedding building obliquely across a SSW-wards accreting bar face. Evidence presented within previous facies interpretations within this chapter indicates that the SSE-flowing current was semi-permanent in nature i.e. burrow entrances facing into the trade wind induced geostrophic current for at least part of the year as trade winds blew onshore. Levell (1980b), in a study of the Lower Sandfjord Formation (late Precambrian) of north Norway, a sub-tidal sand ridge deposit, invoked a similar mechanism to account for the migration of cross-bedded sets, suggesting that semi-permanent currents were necessary to reinforce weak tidal currents.

In conclusion, facies 3 dunes migrated in response to semi-permanent wind-induced currents during the trade wind season, reinforcing a weak tidal régime. Peak energy would have been attained during the conjunction of spring tides and/or periods of strong trade winds, when current velocities would have been sufficient to initiate dune migration. The highest values of bed shear stress would have been achieved at the bar crest, where the relatively shallow conditions would result in the greatest 'wave-effectiveness' at the sea-bed, sufficient to sustain the migration of three-dimensional dunes. These dunes migrated down major SSW-dipping bar face surfaces, locally scouring into the surfaces; SSW-wards accretion of the bar would have resulted. Exceptionally, currents were sufficiently strong to allow upper phase plane bed conditions to become established, resulting in dunes becoming 'washed-out' to give localised parallel-laminated deposits.

A question can be posed as to the extent of tidal current influence in moulding facies 3. Firstly, no wedge-shaped set has been observed that displays a full 'tidal bundle slackening sequence' of the type described by Boersma and Terwindt (1981, figure 6) for a single tide - the evidence that is preserved in facies 3 for tides is more equivocal. The occurrence of successive foresets decreasing in angle down-current (plate 4.7A) does not necessarily infer slackening over a single tide, it could also represent a waning storm event. Furthermore, 'herringbone' cross-stratification, hanging set boundaries and reactivation surfaces are relatively scarce in facies 3. These facts suggest that tidal currents, alone, were not responsible for the transport of facies 3 sediments and probably played a subordinate rôle in bedform transport.

Several positive lines of evidence are present in facies 3 that support the hypothesis that sediment transport was not solely the product of tidal currents. These are listed below:

4.2.4.3.1 Palaeocurrents Were Dominantly Unimodal

It has been shown (text-figure 4.14) that both trough cross-bedding and planar cross-bedding at Lee Stone indicate that palaeocurrents flowed dominantly towards the SSE, the variance between the two cross-stratification types (9°) falling well within one circular standard deviation for the total ($\pm 40^\circ$). At Watersmeet the two localities closest to the line of the Lynmouth - East Lyn Fault preserve palaeocurrents which flowed towards 156° and 168° (with a variance of $\pm 37^\circ$ and 44° respectively), whilst the two localities furthest from the line of the fault preserve evidence for palaeocurrents flowing towards 90° and 101° (with a variance of $\pm 70^\circ$ and 53° respectively) i.e. the palaeocurrents further away from the fault flowed sub-parallel to palaeoslope strike and had a greater variance than the obliquely palaeo-offshore flowing palaeocurrents preserved in the localities closer to the line of the fault. The occurrence of predominantly unidirectional cross-bedding in tidally-generated sand waves is common and has been discussed by Johnson and Stride (1969), Klein (1977), Johnson 1978, Levell (1980b), Terwindt (1981) and Johnson *et al.* (1982). Levell (*op. cit.*) suggested three explanations for this phenomenon:

- (i) Asymmetry of the tidal ellipse resulting in a regional transport dominance over large areas of the shelf.
- (ii) Mutually evasive ebb and flood channels deforming the tidal ellipse by bottom friction, reflection and shielding, thus generating a time velocity asymmetry and, therefore, a local transport dominance.
- (iii) Preservation, which is three-fold:
 - a) Reversed tops to 'cat-back' bars are not preserved, reactivation surfaces being the only evidence of tidal current reversal.
 - b) The lateral migration of sub-tidal bars, which results in the preferred preservation of either flood or ebb members.
 - c) In the majority of tidal seas, sediment transport rates are greatest when tidal currents are enhanced by wind-drift, which may be semi-permanent, reinforcing one tidal phase.

Levell (*op. cit.*) concluded that in the geological record the most likely explanation for unidirectional cross-beds in tidal sand-bodies is the reinforcement of tidal currents by another current system.

This conclusion is in agreement with the work of Johnson and Stride (1969) who indicated that wind-drift and semi-permanent currents, enhancing tidal currents, are important processes in the production of tidal sand-bodies. This phenomenon was also considered important by Johnson *et al.* (1982) who noted that: “Seasonal winds in consistent directions cause relatively long term strong unidirectional currents, which can determine the nett sand transport direction of the continental shelf during a particular season, even if strong tidal currents are present. This effect becomes more important for strong winds, shallow water and where there is only a small ebb-flood tidal current asymmetry” (p.86).

Johnson *et al.* (*op. cit.*) proposed the Torres Strait, between Australia and New Guinea, as an example of a modern sea where semi-permanent wind-drift currents produce a nett current effect. This example is particularly pertinent to the present study as the nett current effect was the product of a seasonal wind reversal of SE trade winds and the NW monsoon. The relevant current velocity data are presented in table 3.4.

| | | |
|-------------------------------|-------------------------------------------------------|----------------------|
| NW MONSOON (Summer) | W-going mean spring tidal current | 2m s ⁻¹ |
| | E-going mean spring tidal current + non-tidal current | 2.7m s ⁻¹ |
| SE TRADE WINDS (Winter) | W-going mean spring tidal current + non tidal current | 3m s ⁻¹ |
| | E-going mean spring tidal current | 2m s ⁻¹ |

Table 4.4 Mean spring tidal current velocities for the Torres Strait.

After: Hydrographic Department, Admiralty (1945).

Table 3.4 shows that the seasonal wind reversal results in currents in each direction varying by as much as 1m s⁻¹. Furthermore, the west-going current reaches an annual mean spring peak which is 0.3m s⁻¹ greater than the east-going current. This suggests a nett westerly transport of sand. This suggestion is substantiated by the occurrence of numerous sand banks extending westward in the lee of islands and reefs (shown in plate 3.20 of Stride 1982).

4.2.4.3.2 Immature Sediment Texture

The grain texture of tidal sand-bodies has been discussed by Reineck (1963), Houbolt (1968), Klein (1970), Balazs and Klein (1972), Vischer and Howard (1974) and Klein (1977). Grain-size distributions were found to be dominantly unimodal (Reineck 1963, Houbolt 1968, Klein 1970), although bimodal populations have occasionally been recorded (Klein 1975a, 1977). Klein (1977) attributed bimodal populations to different modal classes being derived from different source materials, noting that this phenomenon could also produce

multimodal populations. The quartz grains in Klein's examples were characterised by supermature grains (cf. Balazs & Klein 1972) which were ascribed to the long distance of grain transport in a tidal sand-body.

Facies 3 is relatively poorly sorted, although a crude bimodal grain-size distribution is present. It seems that the bimodality is related to differing source materials (cf. Klein 1977). Vein quartz and quartzite grains are well rounded, but volcanic rock fragments, biogenic debris, metamorphic rock fragments and bone debris are rounded to subangular. It appears that the well rounded nature of the quartz grains is related to a previous cycle of erosion. The textural immaturity of facies 3 suggests that the sediments were not reworked *in situ* by tidal currents to any great extent.

4.2.4.3.3 Presence of an Articulated Crinoid Stem

Post-mortem disarticulation of crinoids is rapid, accelerated disarticulation occurring in areas of persistent wave and current activity (Anderson 1968, Brower & Veinus 1974, Liddell 1975, Miller 1979). Rapid sedimentation ensures the preservation of complete crinoids and arms (Liddell 1975, Brower & Veinus 1974), whereas crinoids often become entirely disarticulated in areas of reduced sedimentation, even when velocities are low (Lane & Matthews 1965, Anderson 1968, Brower 1973, Brower & Veinus 1978). The occurrence of a 7cm section of articulated crinoid stem resting on a foreset surface in facies 3 at Lee Stone (plate 4.7B) and the 2cm section at Watersmeet (text-figure 4.13) indicates rapid burial of the specimens. This suggests that dune migration was unidirectional and/or spasmodic, as bi-directional tidal migration would have resulted in reworking and disarticulation of the crinoid stem, whereas unidirectional (?spasmodic) dune migration could bury and preserve the specimen.

The crinoid stems must have been transported to these localities from another environment as no surface suitable for crinoid colonisation has been recorded in the Duty Point sections. Similarly, Goldring (1971) recorded crinoid debris in his *Diplocraterion yoyo* facies and noted that no surface was present in the exposed Baggy Beds that was suitable for crinoid colonisation. Goldring suggested that the crinoid debris may have been derived from nearby stratigraphic 'highs'. Crinoid ossicles are rare in facies 3 at Lee Stone, but are relatively more abundant at the Watersmeet localities adjacent to the Lynmouth-East Lyn Fault. During Lynton Formation times extensional movement on this fault would have formed a positive area to the NNE (chapter 2). That this fault may have formed a positive scarp feature during the time of deposition of the 'Watersmeet lithotype' at Watersmeet appears to be corroborated by the preserved palaeocurrent pattern

(see text-figure 4.15). It is suggested that this positive area may have been suitable for crinoid colonisation, providing crinoid debris which would have been abundant at the base of the fault scarp, but more scarce distally i.e. Lee Stone.

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In conclusion, dune migration in facies 3 was achieved by trade wind generated geostrophic currents which in isolation were insufficient to induce dune migration, but, when reinforced by weak tidal currents, were capable of transporting dunes.

The origin of parallel-lamination in facies 3 is problematical. Klein (1977) noted that the preservation of lower flow régime plane beds is common in tidal sand-bodies, but warns that a lower phase plane bed origin should only be considered if associated with interbedded micro-cross-lamination produced by migrating current-ripples. However, the grain-size in facies 3 is greater than 0.6mm. Southard and Boguchwal (1973) have shown that current-ripples do not occur in grain-sizes greater than 0.6mm and, therefore, Klein's criterion cannot be applied to the granule conglomerate of facies 3.

Fine-grained sandstone filling the scour at the eastern end of Lee Stone has permitted the preservation of foreset bundles displaying structures reminiscent of tidally-generated bundles (cf. Boersma 1969). The absence from the rest of the facies of the structure seen in the fine-grained sediments of the scour fill is related to grain-size. As Allen (1982) has discussed, current-ripples are only produced in sands finer than medium-grade. Thus, the fine sand grade scour infill has allowed the development of many of the ripple-scale features diagnostic of tidal sedimentation which could not develop in the coarse granule-sandstone characteristic of much of facies 3. An interpretation of the events figured in plate 4.10 is given below; bracketed numerals and letters refer to the notation used on the plate.

The top of facies 3 at Lee Stone (A) is cut by a 4m wide scour (surface A/B). The up-palaeoslope end of the scour (right-hand side of the plate) is initiated at the toe of an obliquely palaeo-offshore directed wedge-shaped set of cross-strata (off-plate, right) that has migrated, non-erosively, over the planar top of the facies 3 unit. The scour surface is deepest at its up-palaeoslope end, gradually shallowing down palaeoslope. The shape of the scour is similar to the lower set boundary types created in neap to spring tidal sequences

reported by Terwindt (1981, figure 8). In the latter, a transition occurs from a flat lower set boundary at the onset of a spring tide, to a steeper lower set boundary towards full spring tide. As the approach to the subsequent neap tide occurs, the lower set boundary rises. Terwindt attributed the change in the level of the lower set boundary, to an increase in vortex action during spring tides lowering the surface, the surface then rising as the subsequent neap tide approaches. No sediment was deposited in the trough due to the strength of the vortex, which produced an area of net erosion. It is likely that tidal current velocity was enhanced by geostrophic currents, the presence of which is evidenced by the abrupt end of dune migration and the lack of deposition in the scour trough. The surface is therefore interpreted as a dune-toe scour trough cut by a lee vortex enhanced by increased currents during either a storm or a spring tide. The dune dimensions must have been appreciable (in the order of metres) judging from the 4m wide diameter of the scour. The only remnant of the dune is the cross-bedded wedge (off-plate, right) described above, the remainder of the dune being truncated by event C/D, described below. N.B. Several sandstone horizons set within mudstones are visible in the inaccessible Duty Point cliff above Lee Stone which comprise single cross bedded sets 70-150cm thick (e.g. plate 4.19C) - sand-waves preserved as sandstone-bodies of this type are proposed to be analogous to those responsible for the formation of the scour into facies 3 under discussion here.

Following the scouring event, a relatively quiescent period ensued, allowing the deposition of a 10 to 20cm thick unit of lenticular bedding (B) which concordantly draped the scour surface and cross-bedded wedge (off-plate, right). The thickness of this unit infers a period of dune inactivity. This was perhaps related to a period of lower energy during the ?monsoon season (see section 4.2.2.3).

Unit C records a further phase of dune migration which has resulted in the deposition of toesets displaying structures characteristic of deposition from tidal currents (cf. Terwindt 1981): neap tidal bundles of form-concordant toesets and ripple profiles preserved beneath slack water mud drapes and, spring tidal bundles of form-concordant toesets with 'crinkly bedding' preserved beneath slack water mud drapes and ripple foresetting on pause planes, the foresetting opposing the major dune trend.

Boersma and Terwindt (1981) have shown that characteristic structural sequences are generated in response to tidal variations over: (i) a single tide and, (ii) the neap-spring tidal cycle. During the dominant tide, a lateral succession of cross-strata is generated, termed a tidal bundle (Boersma 1969), which is laterally enclosed by pause planes representing the stand-still phase of dune migration during the subordinate tide.

Within an ideal tidal bundle, three structural intervals may be distinguished and are arranged in a lateral sequence:

- (i) Pre-vortex acceleration structures produced by flow acceleration following the turn of the tide.
- (ii) Full vortex structures with intermittent avalanching occurring over the entire lee front of the dune under full vortex activity.
- (iii) Slackening structures representing deceleration of the tidal current with settling of the suspension load resulting in progressive deposition in the toe area at a gradually lower angle. Finally, only ripples move over the lee front and trough.

During the succeeding subordinate tide, flow reversal is marked by:

- (i) Accentuation of the pause plane by erosion at the top of the dune as the lee side becomes the stoss side.
- (ii) Generation of up-slope and obliquely climbing ripples on top of the pause planes to give opposite ripple foresets and 'crinkly bedding'.
- (iii) Formation of an ebb or flood cap at the dune top.

Terwindt (1981) notes that the basic tidal sequence described above is not always developed completely and is dependent on bedform characteristics (particularly height), current strength, available material, etc. For dunes over 0.5m in height, as was almost certainly the case for the dune under discussion here judging by the toeset dimensions, reactivation structures are frequently absent from lower parts of the dune and slackening structures are characterised by form-concordant toesets. In the scour under discussion here only the toesets have been preserved. Thus, any structures that were generated at the top of the dune by the subsequent tide would not have been preserved. In addition, the effects of the diurnal inflow or outflow of water may be obscured by wind-generated inflow or outflow. The nature of the tidal bundle also depends on whether deposition-occurred during the spring or neap tidal phase of the tidal cycle. The bundles observed in the facies 3 scour are interpreted below:

1. **Neap Tidal Bundles** During neap tides sedimentation is dominated by fall-out from suspension, to give form-concordant toesets (plate 4.10, **x**). Further into the trough cosets of ripple cross-lamination (plate 4.10, **y**) pass into more isolated sets (plate 4.10, **z**) with increasing distance. The ripple-forms are draped with mud from the dune front during slack water.

II. **Spring Tidal Bundles** During the acceleration phase, suspension fall-out generates form-concordant toesets (plate 4.10, **a**) with 'crinkly bedding'. However, during the full vortex stage, deposition in the trough is minimal, full vortex structures generally wedging-out and having an angular or subangular relationship to the toesets. The slackening phase results in the production of 'crinkly bedding' (plate 4.10, **b**) preserved beneath slack water mud drapes (**c**). The velocity of the subordinate tide will also be greater during spring tides, allowing the generation of ripple foresetting on pause planes which oppose the major dune trend (**d**).

Terwindt (1981) recognised that a transition from neap to spring tides would generate a coarsening-upwards megaripple (dune) microsequence of tidal bundles. Terwindt's extensive examination of cores and peels from Holocene and sub-Holocene deposits revealed that in the majority of cases, the boundary between individual coarsening-upwards megaripple microsequences was erosional. It may be seen that the facies 3 scour infill comprises two (incomplete) coarsening-upwards dune microsequences cf. Terwindt (1981).

Boundary **C/D** is a laterally extensive erosion surface which dips obliquely palaeo-offshore at 3° to 5° truncating the major dune which produced the unit **C** toesets. The erosion event that produced this surface must have been of a large magnitude in order to truncate and remove the dune alluded to above. It is probable that the erosion surface is analogous in origin to the laterally extensive erosional pause planes described previously. The erosion surface **C/D** is draped with a 6cm thick mud (plate 4.9 **D**) representing post- high energy event deposition of mud suspended during the **C/D** erosion event. Unit (**D**) grades upwards into lenticular bedding (**E**) (see facies 1).

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The origin of the channel-like scour figured in plate 4.9 is problematical, as only one section, parallel to depositional strike, is available through the scour. The lack of shear surfaces and soft sediment deformation structures in the deposit immediately adjacent to the scour, precludes a rotational slump-scar origin for the scour. Instead, the scour appears to be more analogous with the 'furrows' (cf. Dyer 1970) described from tidal shelf deposits and erosion surfaces/channels recorded from ancient sub-tidal deposits (Goldring 1971, Banks 1973a, Anderton 1976). Alternatively the scour may be the scour pit formed at the toe of a major sand-

wave cf. the example described above. The scour under discussion at this point, however, lacks any evidence of having been a sand-wave scour pit e.g. toe-set sand preservation.

Goldring (1971) noted that the most common type of erosion surface within the Upper Devonian Baggy Beds of North Devon are “. . . irregularly parallel or slightly sinuous grooves cut in shale, siltstone or sandstone . . . up to 1.2m broad and 20cm deep, with semi-elliptical to subrectangular cross-section, and are generally flat-floored and occur singly or more commonly in parallel groups” (p.33). Although larger in scale, the facies 3 example displays certain features seen in the Baggy Beds examples. Goldring observed that “. . . many grooves cut through sandstone and siltstone are floored by the underlying shale, which probably offered greater resistance to erosion” (p.33). Furthermore, an example from the Rough Wall Member of the Baggy Beds displays a subconcordant fill of laminated very fine grade sandstone and silty shales (*Arenicolites curvatus* facies) similar to that present in the facies 3 scour. The consistent trend of the Baggy Beds grooves, constant breadth and depth, average width to spacing ratio of 1:6, gently sloping to vertical margins and up- and down-dip facing division of grooves strongly suggests that the grooves are analogous to flow-parallel furrows (cf. Dyer 1970) observed on present-day tidally-dominated continental shelves, although they could also presumably form on a shelf with a mixed meteorological and tidal current pattern.

On present-day continental shelves, furrows form in sediment of mud to gravel-grade, furrows in gravel-grade material corresponding to peak near-surface currents in the order of 1.5m s^{-1} . Furrows have been documented to reach maximum lengths of 8km, widths of 30m and depths of 1m. In plan the furrows are straight to slightly sinuous and join in a preferred direction, generally joining in the direction of the stronger current (Belderson *et al.* 1982).

Flood (1983) undertook a literature review of Holocene furrows and presented a classification and model for the initiation and evolution of furrows. Their parallelism with flow and regular width to spacing ratios (1:5 to 1:15), suggested to Flood that their origin is related to secondary helical flow-cell circulations within the fluid boundary layer, a mechanism that had first been suggested by Dyer (1970). Flood envisaged cells being initiated by spatial variations in bed shear stress and localised abrasion by, or scour around, particles on the sea bed. Thus, zones of flow convergence will sweep together coarse material with the ability to abrade furrows. Furrows may be observed to pass down-current into sand ribbons (cf. Kenyon 1970) along tidal

current transport paths (Belderson *et al.* 1982). Flood's (1983) model, however, was mainly derived from observations of furrows developed in fine-grained cohesive sediments, Flood noted that furrow-like bedforms develop in shelf sands related to secondary flow circulation may be due to a different mechanism, given the different physical process of erosion, transportation and deposition. Indeed, Karl (1980) suggested that shore-normal furrows ('mesoscale current lineations'), developed in fine- to medium-grade sand on the Californian shelf, formed in areas of secondary helical flow cell **divergence**, sand ribbons forming in areas of flow convergence.

In conclusion, the scour of plate 4.9 is interpreted as being of furrow origin, probably related to a secondary helical flow cell circulation superimposed on a mean palaeo-offshore flow during recurring episodic periods of directionally stable strong currents (cf. Flood 1983) e.g. trade wind season geostrophic flow with a superimposed (?spring) tidal flow.

Goldring and Aigner (1982) noted that scours with a 'banded' (sand/mud) infill are common in shallow marine sequences, the 'banded' infill suggesting that infill occurred over a period of time. Goldring and Aigner observed that the scour surfaces acted as preferential sites for colonisation by benthic organisms, suggesting that the firmness of the partially lithified, exhumed substrate and the protected aspect of the scours flavoured settlement. No evidence of preferential colonisation was observed on the facies 3 scour surface.

The genesis and predicted geometry of facies 3 sandstone-bodies is discussed in section 4.3.2.

4.2.5 Facies 4 - Bioclastic Sandstone

This facies only occurs in the A39 road section, where it is shown as facies D on the log (text-figure 4.1).

4.2.5.1 Description

A single 10cm thick unit of bioclastic sandstone occurs in the A39 road section, set within a sequence of bioturbated, parallel-laminated muddy sandstones (at 5.3m on the log). The unit has an average thickness of 10 cm, appearing as a decalcified brown and deeply weathered horizon at outcrop (plate 4.13A). The unit can be traced over the available width of the exposure with little or no thickness variation. The base of the unit is

a laterally extensive, planar erosion surface with a local relief not exceeding 1cm. The top of the unit exhibits a fining-upwards, gradational contact with the overlying parallel-laminated muddy sandstone.

Internally, the unit comprises a matrix of fine to medium grade sandstone containing clasts of: disarticulated crinoid ossicles and brachiopod valves, intraformational mudstone and well rounded vein quartz granules and lithologically resembles the 'Watersmeet lithotype' (facies 3). The unit is poorly sorted and no primary bedding structures were observed.

4.2.5.2 Biofacies

The facies 4 horizon was sampled by Evans (1980 - recorded as locality ND31A) who recovered a fauna containing 84.5% *Subcuspidella lateincisa*, a brachiopod which was free-lying on the sediment surface. Evans noted that although locality ND31A yielded a 'death assemblage' (*sensu* Boucot 1953), if post-mortem faunal mixing of the assemblage is assumed to have been minimal, then the fauna would have been derived from 'Benthic Assemblage 2' of Boucot (1975) i.e. low intertidal zone down to 6-9m.

Knight (1990a) recovered a solitary specimen of the conodont *Icriodus culicellus*, a form commonly associated with crinoidal or coralliferous limestones and weakly agitated to turbulent shallow waters.

There is no evidence within facies 4, even at the top of the bed, of bioturbation.

Facies 4 conforms to Schäfer's description of his lethal-lipostrate biofacies: "Characterised by abundant taphocoenoses and by disconformities due to considerable erosion; water is well aerated but continuous shifting of coarse material prevents establishment of a benthonic fauna and biocoenosis. What biogenic remains (heavy valves) can be found come from, neighbouring benthonic biocoenoses" (1972, p.478) - although the emplacement of facies 4 is interpreted to be the result of a single high-energy event (see Interpretation below).

4.2.5.3 Interpretation

The sharp planar base and gradational top of facies 4 suggests the passage of a short-lived, high-energy event. Similar laterally extensive, 3-30 cm. thick sheet-like beds containing shell debris and having sharp

erosive bases and planar, gradational tops have been described by Brenner and Davies (1973) from the Upper Jurassic of the Western Interior, U.S.A. Brenner and Davies termed such beds 'storm lags' and attributed them to reworking of the sea floor over extensive areas by storm waves and associated currents. It is proposed that facies 4 was emplaced by similar processes to those postulated by Brenner and Davies.

The disarticulated crinoid ossicles and brachiopod valves, along with well rounded vein quartz granules found within facies 4 do not lithologically match any found within the A39 road section. This suggests that facies 4 is allochthonous, rather than para-autochthonous (*sensu* Seilacher 1982a). The lithology of facies 4 is more characteristic of that found within the 'Watersmeet lithotype' (see sections 4.2.4.) It is suggested, therefore facies 4 was derived from a source similar to that for the 'Watersmeet lithotype' and was emplaced by an 'exceptional' storm event.

4.2.6 Facies 5 - Wavy Bedding

Facies 5 is shown as facies C on the Watersmeet (enclosure 7), east Lynmouth beach (text-figure 4.1) and Duty Point (text-figures 4.2 - 4.5) logs. This facies does not occur in the A39 road section.

4.2.6.1 Description

Wavy bedding is the least common facies type (by volume) within the Watersmeet, east Lynmouth beach and Duty Point sections. The thickness of this facies ranges from 2 to 42cm, individual units comprising a regular alternation of 1 to 17cm thick wavy beds of fine to very fine sandstone (up to medium-sand grade at Watersmeet) interbedded with 1 to 19cm thick lenticular bedded units (facies 1, connected & unconnected lens types - plate 4.8B). At east Lynmouth beach the sandstone beds frequently contain finely comminuted shell debris.

The sandstone beds are generally laterally persistent in the order of tens of metres, although the facies 5 bed above the 'Watersmeet lithotype' north of the East Lyn River (columns B & C on enclosure 7) could not be traced to the lime kiln exposure (column A on enclosure 7) 40 metres away; beds display no major change in thickness laterally. The bases of the sandstone beds are of two types: planar or undulatory, the latter having a relief of up to 1cm and a wavelength of 6 to 9cm. Bed bases frequently display SE-NW aligned impersistent groove casts, prod marks and hypichnial ridges infilling exhumed *Palaeophycus tubularis* burrows (plate

4.8C) at Lee Stone, but bed bases at the remaining localities are poorly exposed. Indistinct, sinuous, post-depositional hypichnial grooves formed by interstratal ?deposit-feeders are also occasionally observed on bed bases at Lee Stone.

Where sandstone bases are planar, the lower portion of the bed comprises a zone of parallel-laminations up to 6cm in thickness, which in some cases are broadly undulatory with a wavelength of up to 70cm and an amplitude of up to 2cm; no primary current lineations have been recorded. These parallel-laminae are generally overlain by 1 to 2 sets of cross-lamination, although one set at Watersmeet has a planar top overlain by lenticular bedding (facies 1). The cross-laminated sets have scooped bases, erosionally truncating the underlying parallel-laminated zone. Laminae are low-angled, occasionally swinging-up from horizontal-lamination, laminae frequently thinning laterally to give a swollen-lens morphology. Individual sets usually display a unidirectional foreset dip direction, although both laterally and vertically adjacent sets may display opposing foreset dip directions. Ripple forms are usually form-discordant, although a late-stage draping lamina was occasionally observed. The cross-lamination style is diagnostic of generation by waves (Boersma in: de Raaf *et al.* 1977). Where the sandy base is undulatory, a parallel-laminated zone is not developed. Instead, the basal zone of the bed is composed of a thin (less than 1.5cm thick) zone of massive sand. The massive sands are overlain by cross-laminations identical to those following the parallel-laminations described above.

Sandy unit tops display symmetric to near-symmetric ripple profiles with a wavelength of 6 to 9cm and an amplitude of 7 to 11mm. Ripple troughs are always rounded, whilst crests are usually rounded, rarely trochoidal. Interestingly, the ripple profile remains near-symmetrical regardless of the trend of section through any given ripple, inferring that the ripples are dome-shaped. The few available bedding plane surfaces available at Lee Stone confirm this observation, displaying a smooth topography of circular to near circular hummocks separated by depressions. No preferred direction of hummock elongation is visible. The sides of the hummocks slope in the order of 15° to 30° . These ripples appear to be equivalent to three-dimensional vortex ripples observed in oscillatory flow tunnels at the transition from two-dimensional wave-ripples to upper oscillatory plane bed (Carstens *et al.* 1969, Lofquist 1978). The significance of this ripple type is discussed in section 4.2.7 in relation to facies 6, within which three-dimensional vortex ripples are particularly well developed. At this point it is sufficient to note that, by analogy with the oscillatory flow tunnel runs, three-dimensional vortex ripples develop after a long period of evolution from initially small and

straight crested ripples in sands finer than 0.3mm. Sigmoidal offshoots, of the type described in section 4.2.1.4, frequently occur in muds immediately overlying the rippled upper surface of the sandstone beds, always occurring on the onshore-facing side of the ripple profile (text-figure 4.7C).

The fine grained sediments intervening between the wavy sands are of two types:

- (i) Thin mudstones, in the order of 1 to 2cm in thickness, draping the rippled upper surface of the underlying wavy sandstone. The mudstones occasionally contain sigmoidal sandstone offshoots associated with the underlying wavy sandstone and sandstone filled *Palaeophycus tubularis* burrows. The mudstones maintain an approximately constant thickness over the underlying rippled upper surface of the wavy sandstone bed. Thus, the upper surface of the mudstone has a rippled surface in-phase with the underlying rippled wavy sandstone. The drape of this type of wavy topped mudstone by the succeeding wavy sandstone results in the production of a wavy sandstone with an undulating base of the type described above.
- (ii) Two to nine centimetre thick beds of thinly interbedded sandstone/mudstone (facies 1, see 4.2.1), generally of unconnected and connected lenticular type bedding. The bedding may be organised into a series of CU microsequences (facies 1' - see section 4.2.2).

4.2.6.2 Biofacies

Biogenic structures occasionally descend from the upper surface of the wavy sandstones in the form of *Palaeophycus heberti* burrows, frequently occurring in dense clusters (plate B.13A, B). More occasionally, the uppermost few centimetres have been biogenically destratified (i.e. fossitextura deformativa). Rare bedding plane views of the upper surface of wavy sands reveal occasional epichnial traces of *Helminthorhapha japonica*, *Phyllocytes* sp. and *Gyrochorte cosmosa*. *Palaeophycus tubularis* burrows occasionally intersect the wavy sandstone layers. The biogenic component of the lenticular bedding intervening between the wavy sandstones does not differ from that observed in facies 1 (see section 4.2.1.6).

The wavy sands are interpreted to be the product of major storm events (see Interpretation below), the *P. heberti* burrows representing post-storm colonisation of the sandy substrate by tube-dwelling animals. The thickly-lined burrow wall of *P. heberti* (see appendix B) reflects the response of tube-dwelling organisms to

a sandy substrate i.e. the production of a thick mucus-bound wall in porous sediments (cf. Brafield 1978). The presence of *P. heberti* burrows standing proud of the rippled sandstone surface, in a 'chimney-like' fashion, attests to penecontemporaneous scour of unbound sediment from around the resistant mucus-bound burrow wall.

The presence of *H. japonica* on rippled sandstone surfaces is interpreted to be the response of an opportunistic foraging animal to muds with a low organic content, the muds immediately overlying the rippled surface being the product of post-storm settling of fine-grained material suspended during the storm. As organic detritus would have had a lower settling velocity than mineral grains, the muds immediately overlying the rippled surface would have had a low organic content (see section 4.2.1.6).

The presence of *Phyllodoites* sp. and *G. cosmosa* preserved epirelief on the rippled upper surface of wavy sandstones is interpreted, in common with examples of *Phyllodoites* sp. and *G. cosmosa* in facies 1, to be the result of preservation due to the granulometric contrast between sandstone and the overlying mudstone. Both *Phyllodoites* sp. and *G. cosmosa* are interpreted to be the traces left by an infaunal vermiform animal migrating through the substrate. Their preservation was dependent upon the presence of a lithological contrast; the junction between the rippled sand and the overlying mud would have provided such a contrast.

In relation to biofacies, if the wavy sandstones are considered in isolation their geologically 'instantaneous' deposition suggests that they should be assigned to Schäfer's (1972) lethal biofacies i.e. non-life. However, if we take facies to be ". . . the whole set of attributes possessed by the deposited sediment laid down in a particular environment" (Leeder 1982a, p.119), we should consider whether the wavy sandstone was conducive or hostile to colonisation **subsequent** to deposition. Thus the occurrence of *P. heberti* and other biogenic traces within the wavy sandstone layers dictates that the wavy sandstones should be assigned to Schäfer's vital biofacies. Furthermore, the bioturbated lenticular bedding intervening between the wavy sandstones must be considered as part of the wavy bedded facies if Leeder's definition of facies is adhered to (see above), and thus reinforces the assignation of the wavy bedded facies to Schäfer's vital biofacies. The presence of penecontemporaneous scour around *P. heberti* burrows and erosional discordances in the lenticular bedding intervening between the wavy sandstones indicates that the wavy bedding should be assigned to Schäfer's lipostrate biofacies. In conclusion, facies 5 is assigned to Schäfer's vital-lipostrate biofacies (see enclosure 4).

4.2.6.3 Interpretation

Wavy bedding is a common constituent within tidally-produced sequences, including: intertidal flats (Reineck 1967), estuaries (Howard & Frey 1975), subtidal sand-bodies (Berg 1975) and on tidally-influenced deep-sea rises (Klein 1975b). However, the production of wavy bedding is not restricted to tidal environments; in shallow marine sequences, wavy bedding has been recorded from both modern and ancient wave-dominated shorefaces (Howard & Reineck 1972, Tunbridge 1983a).

Reineck and Wunderlich (1968a) have produced a classification of finely interbedded rippled sands and muds, based on the lateral continuity of the minor member and the relative proportion of sand and mud. In their scheme wavy beds were defined as the alternation of continuous mud and ripple bedded sand layers. Wavy bedding differs from connected lenticular bedding in that **no more** than 75% of the rippled sands are continuous in lenticular bedding. In the field, however, the application of Reineck and Wunderlich's classification proves impractical as the 75% continuity criterion forms an arbitrary boundary between the two bedding types, with no specific sedimentological/hydrodynamic significance. The lateral continuity of rippled sands frequently changes over a short distance, as does the proportion of isolated lenses interbedded with connected ripples. The distinction between wavy flaser and wavy bedding is also unsatisfactory, as thin, laterally continuous muds frequently become discontinuous where two vertically adjacent wavy sands become amalgamated to produce wavy flaser bedding.

For the above reasons the following modifications to Reineck and Wunderlich's classification are applied throughout this thesis. Connected lenticular bedding refers to millimetre-scale connected wavy sandstones. The term wavy bedding is restricted to centimetre-scale connected rippled sandstones. Finally, the wavy flaser category includes **continuous** wavy muds interbedded with centimetre-scale rippled sandstones. Thus, a composite unit of centimetre-scale rippled sandstones interbedded with lenticular bedding is termed wavy bedding (cf. Tunbridge 1983a). Two wavy bedding types have been distinguished in facies 5, distinguished from each other on the nature of the base of the wavy sandstone:

4.2.6.3.1 Units comprising wavy-based and -topped sandstones separated by thick wavy mudstones

This bedding type conforms to Reineck and Wunderlich's (1968a) original definition of wavy bedding. Reineck and Wunderlich proposed that wavy bedding was a type of 'tidal bedding', produced by the

alternation of current or wave action and slack water during a single tide. Subsequently, considerable doubt has been expressed as to whether mud layers of the required thickness can be deposited and **preserved** during the slack water phase of a single tide. Although millimetre thick muds (Reineck & Wunderlich 1969) and even 2cm thick muds (Wunderlich 1978) have been observed to deposit during the slack water phase of a single tide, McCave (1970) has noted that subtidal heterolithic bedding types “. . . are not directly comparable with similar bedding types on tidal flats deposited during slack tide, because in the latter environment tidal slack water may be abnormally long (5 hours) and bottom concentrations (of fine-grained sediment) may be very high (>200mg l⁻¹)” (p.4157). McCave concluded that the deposition and preservation of mud layers during a single tidal slack water phase was extremely unlikely; the introduction of sand into an area of fair-weather mud accumulation during periodic storm activity being a more plausible mechanism.

An alternative mechanism for mud preservation was proposed by Terwindt and Breusers (1972). Experimental results suggested to them that centimetre-scale mud layers could form from the accumulation of several thinner (\cong 3mm) mud layers deposited during successive slack water intervals during the neap phase of the neap-spring tidal cycle. The deposition of wavy sand layers would occur during increased current velocities during the spring phase of the spring-neap tidal cycle. However, extensive flume experimentation examining the potential for preservation of mud layers in sands under a tidal regime led Hawley (1981) to conclude that Terwindt and Breusers' mechanism was “. . . extremely unlikely due to the low velocities required to erode such beds” (p.711). Furthermore: “The formation of even one sand-mud couplet every two weeks implies a total accumulation of approximately 0.5m yr⁻¹, far higher than that found in nearshore areas” (p.710). Instead, Hawley proposed that the preservation of thick mud layers in the subtidal environment occurs when the range of current velocities over a tidal cycle (both diurnal and spring-neap) does not exceed the mud erosion threshold, sands only being deposited by higher than ambient velocities i.e. normally storms (cf. McCave 1970).

In facies 5, massive sandstones draping underlying wavy mudstone surfaces (following the interpretation of Collinson & Thompson 1982, p.101): “. . . imply rapid deposition from suspension, most probably by the deceleration of a heavily sediment-laden current”. In the shallow marine environment the generation of a sediment-laden current is most likely to be related to storm activity. Thus, evidence from facies 5 type (i) wavy sands supports McCave's (1970) hypothesis i.e. wavy sandstones set within mudstones are the result of periodic storm activity and not tidal current velocity variability. Although the storm events that emplaced the

wavy-based sands were obviously of insufficient energy to erode the underlying wavy muds, it is probable that the wavy mudstones represent the preservation of a lower consolidated dense mud layer underlying a thicker unconsolidated mud column (cf. Owen 1970), the overlying unconsolidated muds having been suspended by storm action due to their lower shear resistance.

The wave cross-laminated top to the wavy sandstones represents storm-wave reworking of the storm deposited sand. When fair-weather conditions returned, the upper surface of the sand was colonised locally by the *P. heberti*-animal and the sandy substrate was slowly buried by muds accumulating below fair-weather wave-base. Interstratal deposit-feeders subsequently exploited the newly created sand/mud-interfaces at the base and top of the wavy sands.

4.2.6.3.2 Units comprising planar based wavy sandstones, separated by relatively thick units of thinly interlayered sandstone/mudstone bedding (facies 1)

Planar based, wavy topped, centimetre-scale sandstones set in finer lithologies are common constituents within shallow marine siliciclastic sequences (Goldring & Bridges 1973, Banks 1973a, Anderton 1976, Vos 1977, de Raaf *et al.* 1977, Tunbridge 1983a). These examples equate with the 'type B' storm layers of Allen (1982, vol. B, Fig. 12-12) and parallel the wave-ripple laminated sublittoral sheet sandstones in the scheme of Johnson (1978, Fig. 9.50). The latter author proposed that this type of sandstone was deposited from a decelerating current, followed by progressive oscillatory wave action and represents the deposit of a single storm.

The planar base of the facies 5, type (ii) wavy sandstones was clearly erosive, having eroded the underlying mud substrate to exhume the tunnels of infaunal deposit-feeders which were subsequently cast by the storm-sand to give hypichnial ridges on the planar base of the wavy sandstones. This type of preservation of 'graphoglyptid' tunnels is common on erosive turbidite bases (Seilacher 1977a). The parallel-laminated sandstone overlying the planar base appears to be of unidirectional upper phase plane bed origin, the well defined laminae being truly parallel and flat-lying. In contrast, the undulatory, pinch-and-swell type of lamination, commonly swinging-up into low-angled cross-lamination, characteristic of upper phase oscillatory plane bed deposits (de Raaf *et al.* 1977) is absent from facies 5. The parallel-laminated zone is erosively overlain by wave-ripple cross-laminations produced by the migration of three-dimensional vortex ripples (cf. Harms *et al.* 1982) which represent storm-wave reworking of the parallel-lamination. Thus, the

sequence of erosion to parallel-lamination to wave cross-laminations represents a waning flow sequence, the former two members produced by storm-generated unidirectional obliquely offshore-flowing geostrophic currents, whilst the upper member represents reworking of the sandy substrate by oscillatory currents as the storm abated. Following the storm, the upper surface of the sand was colonised by *P. heberti* dwelling-tubes and the upper part of the wavy sand was locally biogenically destratified. Biogenic structures descending from upper surfaces are typical of storm-generated sublittoral sheet sandstones (Goldring & Bridges 1973). The planar based wavy sandstones are separated by relatively thick sequences of muds deposited beneath fair-weather wave-base. The processes described in section 4.2.1 introduced thin connected and unconnected sandstone lenses. Thus, the planar based wavy sandstones, representing relatively major storms, are separated by lenticular bedding representing sand incursions into a muddy environment during relatively minor peaks in environmental energy.

In conclusion, the wavy based and wavy topped sandstones, separated by thin wavy mudstones, represent relatively high frequency minor storm events proximal to an abundant sand supply (sublittoral shoal?). The massive, non-erosive bases to the wavy based sandstones attest to the relatively low energy of some of the storm events. In contrast, the planar-based wavy sands separated by lenticular bedding represent relatively low frequency, high energy storm events, the connected and unconnected sandstones in the intervening lenticular bedding representing lower-energy peaks in environmental energy. The thick units of lenticular bedding separating the planar-based wavy sandstones record the 'normal' sedimentation style at that particular site, the planar-based wavy sandstones representing 'rare events'. Thus, type (i) wavy bedding represents deposition proximal to a sand source, whilst type (ii) wavy bedding represents deposition more distal to the sand source.

4.2.7 Facies 6 - Flaser Bedding

Facies 6 is shown as facies D on the east Lynmouth beach (text-figure 3.1B) and Duty Point (text-figures 4.2 - 4.5) logs. In comparison with Lee Stone, facies 6 is relatively rare in the east Lynmouth beach section. This facies does not occur in the Watersmeet or A39 road sections.

4.2.7.1 Description

Flaser bedded units range from 3cm to 42cm in thickness and comprise a regular alternation of centimetre-scale very fine to fine-grade sandstone and millimetre-scale mudstone laminae. The sandstones have an average thickness of 2cm to 3cm, the mudstone laminae reaching a maximum thickness of 1cm in ripple troughs and 5mm over ripple crests. The sandstone content of individual units ranges between 70% and 95%. Flaser bedding is the dominant bedding type within the major sandstone-body at Lee Stone i.e. the 'Lee Stone facies association.' The flaser bedded units in the upper part of the 'Lee Stone facies association' are locally cut by NW-SE trending micro-channels which have a maximum width of 1m (plate 4.14A) and depths ranging from 4 to 8cm.

The sandstone beds have a very variable lateral persistence, averaging 1m to 10m, frequently displaying lateral wedging-out of the unit due to penecontemporaneous erosion. Vertically adjacent sandstone beds are usually separated by a continuous wavy mudstone lamina. More rarely, vertically adjacent sandstone beds amalgamate locally, the crests of the underlying rippled sandstone being intersected by the base of the overlying sandstone. This results in the preservation of isolated mudstone flasers in the troughs of the lower ripple set. Thus, two types of flaser bedding are defined:

- (i) Flaser bedding with continuous wavy mudstone laminae separating the sandstone beds. The mudstone laminae usually thicken into the ripple trough. This type of flaser bedding was referred to as wavy flaser bedding in section 4.2.6.
- (ii) Flaser bedding with discontinuous mudstone layers of a convex-upwards lens shape preserved in ripple troughs ('simple flaser bedding' *sensu* Reineck & Wunderlich 1968a).

Both type (i) and (ii) flaser bedding may locally contain bifurcations of mudstone laminae. When units are traced laterally, type (i) flaser bedding and type (ii) flaser bedding commonly pass into each other within a particular mudstone horizon. Both type (i) and type (ii) flaser bedding are shown in plate 4.14B.

In the vast majority of cases, the base of the sandstones are sharp, non-erosive and wavy (plate 4.14B, C), with a relief of up to 1.5cm and a wavelength of 5 to 12cm. The paucity of exposed basal surfaces precluded

the observation of any sole marks. Where the basal surface of the sandstone is wavy, it is followed by a thin (5 to 10mm thick) zone of massive sandstone (plate 4.14C). This basal massive zone is identical to those seen in wavy based sandstones in facies 5 (section 4.2.6). Although intraformational mudstone chips occur throughout the complete thickness of many of the wavy sandstones, they are particularly common in the basal massive zone. The mudstone chips range in size from 5 to 15mm and are frequently contorted. Very rarely, the sandstone base is sharp, planar and erosive, being followed by a thin (<8mm) zone of the flat-lying parallel-laminations identical to those seen at the base of the planar-based wavy sandstones discussed in section 4.2.6.

The upper, cross-laminated zone within the sandstone beds is the product of the interplay, to varying degrees, between unidirectional and oscillatory currents. Pure current-ripples appear to be absent from facies 6, although combined-flow ripples in which the dominant formative current was unidirectional occasionally occur (plate 4.15A). In vertical section, the combined-flow ripples generated by predominantly unidirectional currents have a strongly asymmetric, rounded ripple profile, ripple symmetry indices (length of horizontal projection of stoss side ÷ length of horizontal projection of lee side) exceeding 2.2 in every case, reaching a maximum of 3.8. The vertical form-index (ripple length ÷ ripple height) ranges from 11 to 15. In plan, the ripple crests are straight to slightly sinuous, the ripples lacking spurs and stoss-side ridges. Set thickness ranges from 5mm to 15mm, foresets generally being planar. Set bases are also planar. In summary, the characteristics described above of the combined-flow ripples, match well with the characteristics of combined-flow ripples described by Allen (1982, vol. 1, chap. 11; referred to as 'wave-current ripples').

At the opposite end of the combined-flow ripple spectrum, ripples generated by purely oscillatory flow are rare. Combined-flow ripples generated by predominantly oscillatory currents are symmetrical, or nearly symmetrical, in profile with the crests being rounded, or in rare cases trochoidal; ripple troughs are always rounded. The ripple symmetry generally falls below a value of 1.7. The ripple wavelength ranges from 5cm to 13cm, the ripple height ranging from 5mm to 10mm. Sets attain a maximum thickness of 2cm, the set bases always being scooped and erosional (plate 4.16), resulting in a 'swollen lens' set geometry.

Combined-flow ripples generated by predominantly oscillatory flow display a cross-lamination style comprising sets of concave-up low-angle undulating laminae, inclined laminae occasionally swinging-up

from horizontal-laminations. Offshooting and draping foresets occur occasionally, as do single form-concordant laminae draping ripple profiles. Sets tend to display a bundled pattern of up-building. Foresets in adjacent sets frequently display an opposed direction of migration, adjacent sets usually showing a structural dissimilarity. The majority of ripples are form-discordant. In summary, the lamination style is identical to that generated by purely oscillatory flow (cf. Boersma in: de Raaf *et al.* 1977). Many of the characteristic features described above are figured in plate 4.16.

At Lee Stone an abundance of rippled bedding planes (e.g. plate 4.15) permits the examination of facies 6 in plan view. Straight-crested, 'two-dimensional wave-ripples' are virtually absent, the vast majority of bedding planes consisting of a complex pattern of decimetre-scale three-dimensional hummocks, combined-flow and interference ripples. The hummocks are approximately circular in plan view, with gently curving, convex-upwards flanks. Whatever section is cut through the hummocks, the profiles remain near-symmetrical. Apex-to-apex distances range from 5cm to 13cm, the hummocks attaining heights in the range of 5mm to 10mm. The internal lamination style is characteristic of generation by waves (plate 4.16).

It must be stressed that combined-flow ripples generated by dominantly unidirectional currents, combined-flow ripples generated by dominantly oscillatory currents and 'three-dimensional wave-ripples' are all members of a broad spectrum of wave and combined-flow ripples, all intermediate types having been recognised in facies 6. Furthermore, it is only when facies 6 is observed in plan view that the true complexity of the ripple morphology is fully visible. In addition to three-dimensional 'hummocky' wave-ripples and combined-flow ripples, a complex pattern of interference ripples is present. The pattern is usually the product of wave-ripples superimposed on wave-ripples, although a ripple asymmetry in many of the interference ripple profiles implies a velocity asymmetry in the formative currents or the superimposition of a unidirectional current upon the oscillatory current.

Wave-wave interference ripples have been recorded by many authors from the natural environment (Kindle 1917, Bucher 1919, Inman 1957, Bajard 1966, Komar 1973, Roberts 1974, Martinez 1977). In common with interference ripples formed experimentally (Bagnold 1946, Manohar 1955, Mogridge 1973), naturally occurring interference ripples comprise two, or occasionally three, sets of trochoidal ripple profiles arranged at a steep angle forming a 'brick', 'tile' or hexagonal pattern. Allen (1982, vol. 1, p.434) noted that "... the available evidence strongly suggests that two sets of crests are invariably formed simultaneously". In

contrast, interference ripples in facies 6 lack the regularity of previously described interference ripples. The facies 6 interference ripples comprise ripple crests that have become **subsequently** reworked by currents, generating ripple crests at an angle to the initial set. The currents that produced the modifying set, however, did not operate for a sufficient period of time to completely rework the underlying ripple set. This resulted in a thin veneer of rippled sand, with crestlines at an angle to partially preserved underlying ripple crests, giving a 'hummocky' topography to facies 6 bedding planes (plate 4.15B). Both the modifying and modified ripple sets may be of either wave or combined-flow origin. At Lee Stone, up to three generations of interference ripples have been recorded on individual bedding planes. Plate 4.15C shows a close-up photograph of an interference rippled bedding plane.

It is worth noting at this point that vertical sections through both three-dimensional vortex (wave) ripples and interference ripples are approximately similar i.e. ripple profiles are symmetrical or near-symmetrical in whatever section is taken through the ripple, and the ripple cross-lamination is diagnostic of generation by waves. Thus, the distinction between three-dimensional vortex ripples and interference ripples can only be made if a bedding plane view of the ripples is available. For this reason, the double inverted 'V' symbol on logs within this thesis (i.e. '3-D wave-ripples' on the legend, enclosure 3A) covers both three-dimensional vortex ripples and interference ripples.

In many cases reworking of ripples can be demonstrated to have occurred after ripple troughs had become partially infilled with mud. Ripple crests became modified by wave and current activity after mud had settled in ripple troughs and resulted in 'spill-over aprons' (*sensu* Seilacher 1982b) interdigitating with the mud infilling ripple troughs (plates 4.14B & 4.15D). The "spill-over aprons" were later tilted down by mudstone compaction; tilting down of this type having also been recorded by Seilacher (1982b, figure 3) and Aigner (1982, figure 3A). The 'spill-over aprons' figured in plate 4.15D only occur on the north-facing flanks of the ripples, suggesting that the 'spill-over aprons' were generated by southerly-directed unidirectional currents rather than by oscillatory currents.

In common with facies 1 and 5 at Lee Stone, isolated sigmoidal sandstone lenses frequently occur in the mudstones immediately above rippled sandstones. As with facies 1 and 5, the sigmoidal lenses generally occur above the landward-facing side of the underlying ripples (text-figure 4.7D) although sigmoidal lenses

occasionally occur above the seaward-facing flanks of the underlying ripples (plate 4.14C). The genesis of the isolated sigmoidal lenses is discussed in section 4.2.1.4.

4.2.7.2 Biofacies

The assemblage of biogenic structures in facies 6 closely resembles that observed in facies 5 (see enclosure 4). *Palaeophycus heberti* and *P. tubularis* burrows descend from the upper surface of the wavy sandstones, bedding plane views revealing that both *P. heberti* (plate B.13A, B) and *P. tubularis* (plate B.15C) tend to occur in dense clusters. A single occurrence of *Arenicolites* sp. descending from the upper surface of a wavy sandstone has also been observed (plate B.1G). The upper few centimetres of wavy sandstones have occasionally been biogenically destratified (fossitextura deformativa).

The multitude of well exposed rippled bedding planes within facies 6 reveal a large number of well preserved biogenic epireliefs of *Helminthorhappe japonica*, *Phyllodocites* sp. and *Gyrochorte cosmosa*. Rarely, vertically disposed 'mantled' tubes cut facies 6. The mudstone flasers intervening between the wavy sandstone layers are normally bioturbated, frequently densely, by sandstone filled *P. tubularis* burrows. More rarely, the mudstone flasers have a 'swirled' appearance i.e. fossitextura deformativa.

The wavy sandstone layers in facies 6 are interpreted to be event deposits generated by an increase in environmental energy (see section 4.2.7.3). The burrows of *Arenicolites* sp. a, *P. heberti* and *P. tubularis* represent a post- high-energy event colonisation of the sandy substrate by tube-dwelling vermiform organisms. As was noted in the biofacies description of facies 5 (section 4.2.6.2) the thickly-lined burrow wall of *P. heberti* reflects the response of tube-dwelling organisms to sandy, porous sediments i.e. the production of a thick mucus-bound wall. The presence of both *P. heberti* and *P. tubularis* burrows standing proud of rippled sandstone surfaces in a 'chimney-like' fashion, demonstrates that the mucus-bound burrow walls were resistant to penecontemporaneous scouring.

As with facies 1 and 5, the occurrence of concave epireliefs of *H. japonica* on rippled sandstone bedding planes is interpreted to be the response of an opportunistic foraging animal to muds with a low organic content in the zone immediately above the rippled surface. Furthermore, the presence of epireliefs of

Phyllodoce sp. and *G. cosmosa* is believed to be, as with facies 1 and 5, the result of preservation due to the granulometric contrast between sand and the overlying mud.

In conclusion, the rippled sands of facies 6 were colonised by vermiform organisms which subsisted in whole, or in part, on a diet gathered by suspension-feeding (*P. herberti*, *P. tubularis*, *Arenicolites* sp. a and 'mantled' tubes). This suspension-feeding activity infers the presence of semi-permanent currents acting on facies 6. This is consistent with the presence of multiple phases of 'spill-over rippling' (plate 4.15D). The mudstone component of facies 6 is normally bioturbated by *P. tubularis* which gathered food by a combination of suspension-feeding and/or predation, combined with deposit-feeding (appendix B). More rarely, the mudstones have been biogenically destratified (*fossitextura deformativa*).

In summary, facies 6 is characterised by the burrows of organisms that lived in a well aerated, agitated environment which was frequently subjected to high energy events which rippled the sand component of the facies. It is apparent, therefore, that facies 6 is referable to Schäfer's (1972) vital-lipostrate facies.

4.2.7.3 Interpretation

Flaser bedding has been recorded from a diverse range of tidally-generated sequences, including: intertidal flats (Reineck 1967), estuaries (Howard & Frey 1975), sub-tidal sand-bodies (Berg 1975) and on tidally-influenced deep sea rises (Klein 1975b). The occurrence of flaser bedding, however, is not restricted to tidal environments, Howard and Reineck (1972) and Tunbridge (1983a) having recorded flaser bedding from modern and ancient wave-dominated shorefaces respectively. It is clear, therefore, that flaser bedding is not diagnostic of a particular environment.

As was noted in section 4.2.6.3, the application of Reineck and Wunderlich's (1968a) classification of finely interbedded sands and muds in the field proved impracticable. The scheme of Reineck and Wunderlich was, therefore, modified. The term 'wavy bedding' is restricted to centimetre-scale rippled sandstone interbedded with centimetre-scale units of mudstone and lenticular bedding (see 4.2.6.3); the 'wavy flaser' category includes continuous millimetre-scale wavy mudstone laminae interbedded with centimetre-scale rippled sandstones (with wavy tops and bases). Flaser bedding observed within the Lynton Formation displays

examples of all Reineck and Wunderlich's flaser bedding types i.e. 'bifurcated wavy', 'wavy', 'bifurcated' and 'simple' (see figure 1 of Reineck & Wunderlich *op. cit.*)

The term flaser bedding ('flaserschichten') was coined by Reineck (1960a) who suggested that the rippled sand layers represented strong mid-tide currents, the intervening mud layers representing deposition during slack water. This view was reiterated in a later paper (Reineck & Wunderlich 1968a) and accepted by Klein (1977). McCave (1970), however, argued that ". . . mud layers found in flaser bedding cannot be deposited during a single tidal slack water . . . because the rate of mud deposition is just not high enough" (p.4157), further noting that: "Mud flasers from a marine environment are not directly comparable with apparently similar bedding types on tidal flats deposited during slack tide . . . because in the latter environment tidal slack water may be abnormally long (\cong 5 hours), and bottom concentrations may be very high ($>200 \text{ mg l}^{-1}$)" (p.4157). McCave concluded that the deposition and preservation of mud layers during a single tidal slack water phase was extremely unlikely; the introduction of sand into an area of fair-weather mud accumulation during periodic storm activity being a more plausible mechanism.

Despite McCave's objections, experiments undertaken by Reineck and Wunderlich (1967, 1969) and Wunderlich (1969) showed that flaser bedding could be produced by a diurnal tidal velocity variation mechanism.

An alternative mechanism for mud preservation was proposed by Terwindt and Breusers (1972). Experimental results suggested to them that centimetre-scale mud layers could form from the accumulation of several thinner (\cong 3mm) mud layers deposited during successive slack water intervals during the neap phase of the neap-spring tidal cycle. The deposition of wavy sand layers would occur during increased current velocities during the spring phase of the spring-neap tidal cycle.

More recently, Hawley (1981) carried out extensive set of flume experiments in order to examine the deposition and preservation of mud layers in sands. It was concluded in section 4.2.6.3 that the wavy based and topped sandstones separated by thick mudstone layers [type (i) wavy bedding] could not be produced and preserved by a diurnal tidal velocity variation mechanism, this conclusion being based on the findings of Hawley (*op. cit.*) However, Hawley has also concluded that flaser bedding, with its much thinner discontinuous mud layers, could not be produced by a diurnal tidal velocity variation mechanism, preferring:

“A mechanism which more readily permits the deposition of thicker beds and longer consolidation times, such as storm action” (p.699). Hawley’s conclusions were challenged by Terwindt and Breusers (1982) who suggested that Hawley had used an inappropriate equation for critical shear velocity above a rippled bed. Essentially, Hawley used an equation for a hydraulically rough bed; Terwindt and Breusers suggested that a partial mud cover of the rippled bed would, in part, react as a hydraulically smooth surface, thus raising the critical shear velocity that would be required to erode the mud. In reply to Terwindt and Breusers’ discussion, Hawley (1982) accepted that Terwindt and Breusers were correct, stating: “I believe that my original suggestion - that storm activity is the main cause of flaser beds in tidal areas - is incorrect and that my alternative mechanism - that they form as the result of basal consolidation of thick deposits formed in a single slack water period - is more likely to be the main cause of flaser beds” (p.907).

In conclusion, flaser bedding can be produced and preserved by a diurnal tidal velocity variation mechanism and by longer term velocity variations such as those generated during the spring-neap tidal cycle or storms. Three lines of evidence suggest that the flaser bedding in the Lynton Formation was **not** produced by a diurnal tidal velocity variation mechanism. Firstly, flaser bedding in tidally-dominated areas shows some kind of regular alternation in the number and thickness of the clay layers in a vertical sense (Terwindt 1981, Terwindt & Breusers 1982). No such systematic variations have been observed in facies 6. Secondly, the rippled sandstones of facies 6 were colonised by vermiform organisms which subsisted in whole, or in part, on a diet gathered by suspension-feeding (see 4.2.7.2 above). The dwelling-tubes of the suspension-feeders indicate that the facies 6 substrate was **not** subject to diurnal transport of sand and large influx of mud, but suggests that the layering in facies 6 was the product of much longer term velocity variations such as those experienced during the spring-neap tidal cycle and/or seasonal variation in current velocity. Thirdly, multiple layers of ‘spill-over aprons’ on the flanks of ripples in facies 6 (plate 4.15D) indicate that the introduction of mud into the ripple troughs must have been incremental and punctuated by periods of sand reworking over the muds that infilled troughs. In summary, the layering of facies 6 is interpreted to be the product of long-term variations in current velocity, probably representing winter trade wind season geostrophic flow, perhaps during the spring phase of the spring-neap tidal cycle.

In facies 6 massive sandstones draping underlying wavy mudstone laminae suggests (following the conclusions of Collinson & Thompson 1982, p.101): “. . . rapid deposition from suspension, most probably by the deceleration of a heavily sediment-laden current”. The preservation of wavy mudstone laminae

beneath a non-erosive, massive sandstone drape is thought to be the function of the current which deposited the sandstone having attained its sediment load capacity c.f: "As a storm current accelerates, the bottom is eroded. During the period of peak flow, however, the sediment load in the bottom boundary layer of the flow rapidly adjusts to the available fluid power. As long as velocity remains constant and the sediment load is in equilibrium with it, capacity has been attained and further erosion cannot occur" (Swift and Rice 1984, p.48). It is probable, however, that the wavy mudstone laminae represent the preservation of a lower consolidated dense mud layer that underlay a thicker unconsolidated mud column (c. Owen 1970), the overlying mud having been suspended by an increase in environmental energy due to their lower shear resistance.

In rare cases the base of the sandstone layer is planar and demonstrably erosive. In such cases the planar base is overlain by a thin zone of parallel-lamination. Where the base is planar, the underlying mudstone has usually been eroded where it overlies ripple crests. Thus, the mudstone underlying planar-based sandstones usually occur as isolated flasers.

As was described in section 4.2.7.1, facies 6 displays a wide range of combined-flow ripples, interference ripples and 'hummocky' three-dimensional vortex ripples. Pure current-ripples and straight-crested wave-ripples appear to be virtually absent.

The combined-flow ripples in facies 6 are asymmetric, with their steep sides facing towards the SSE. Combined-flow ripples are the product of ". . . one or both of: (1) a significant wave-generated mass-transport (at the bed normally in the direction of wave-propagation) superimposed on the grain-mobilizing oscillatory flow component, and (2) the presence of a unidirectional current unrelated to the presence of the waves (e.g. wind drift, tidal current, thermohaline circulation) and not necessarily acting in the direction of wave-propagation" (Allen 1982, Vol. 1, p.450). It is not possible, using preserved ripple morphology, to distinguish whether the ripple asymmetry was the product of wave-generated mass-transport or the superimposition of an related unidirectional component (Allen *op. cit.*) However, the combined flow ripple crestlines are orientated ENE-WSW (palaeo- NE-SW) a direction normal to presumed palaeo-SE trade winds, the resultant waves, therefore, propagating towards the palaeo-NW; the steep sides of the combined-flow ripples face towards the SSE, opposite to the direction which would be expected if the asymmetry was the product of waves propagating towards the NNW. Furthermore, palaeocurrents in facies 3 and 8 indicate that SSE-directed unidirectional currents were responsible for the production of cross-bedding. It is

concluded, therefore, that the asymmetry of the combined-flow ripples was the product of a superimposed SSE-directed unidirectional current.

As was discussed in section 4.2.7.1, interference ripples in facies 6 lack the regularity of previously described interference ripples. As opposed to previously described interference ripples, in which the two or more sets of crests formed simultaneously (Allen 1982 vol. 1, chap. 11), the interference ripples in facies 6 comprise ripple crests that have become **subsequently** reworked by waves generating crests at an angle to the initial set. The waves that produced the modifying set, however, did not operate for a sufficient period of time to completely rework, and therefore erase, the underlying set of ripple crests. Such ripples suggest the operation of short-lived wave-trains of variable direction, presumably in response to reflection and refraction due to a complex bottom topography.

An important ripple type in facies 6, and also in facies 1, 5 and 8, are the rounded 'hummocks' assigned to the 'three-dimensional vortex ripples' described by Harms *et al.* (1982). Three-dimensional vortex ripples have only been observed in oscillatory-flow tunnels (Carstens *et al.* 1969 - termed 'three-dimensional dunes'; Lofquist 1978) in sands finer than about 0.3mm at moderately high maximum orbital velocities. These forms are ". . . strongly three-dimensional at equilibrium, after a long period of evolution from initially small and straight-crested rolling-grain ripples" (Harms *et al.* 1982, p.2-37). The bed comprises a series of rounded 'hummocks' separated by rounded swales. During low-velocity runs the side slopes are step, but become increasingly low-angled as velocity is increased, reaching as little as 10°. There is little tenancy toward elongation of the hummocks in the direction normal to oscillation. Carstens *et al.* (*op. cit.*) found that three-dimensional vortex ripples formed at a ratio of ripple amplitude to mean grain diameter of 775 to 1 700, dune amplitude decreasing almost linearly with increasing water-motion amplitude.

As three-dimensional vortex ripples have only been produced experimentally, it is possible that they are the consequence of flume wall effects. However, it is probable that they do occur naturally, the appropriate observations having yet to be made (Harms *et al.*, *op. cit.*)

In conclusion, the 'hummocky' wave ripples in the Lynton Formation bear a striking resemblance to experimentally produced three-dimensional vortex ripples and are, therefore, interpreted to be the product of moderately high maximum orbital velocities acting on a bed of very fine to fine-grade sand.

The interface between the rippled sandstone and overlying mudstone laminae displays epireliefs of *Helminthorhapha japonica* which is interpreted to be the response of an opportunistic foraging animal to muds with a low organic content which were deposited during the waning phases of the storm responsible for the emplacement of the underlying rippled sand.

The presence of *Palaeophycus heberti*, *P. tubularis* and *Arenicolites* sp., the dwelling burrows of animals which subsisted in whole, or in part, on a diet gathered by suspension-feeding, colonising the rippled sands suggests the presence of permanent or semi-permanent currents flowing in a consistent direction (towards SSE). The presence of such currents is also suggested by the occurrence of multiple levels of 'spill-over aprons' on ripple flanks and isolated sigmoidal sandstone lenses along with the presence of *P. heberti* and *P. tubularis* tubes standing proud of rippled sandstone surfaces in a 'chimney-like' manner, suggesting penecontemporaneous scour around resistant burrows. The mudstone laminae are interpreted to be the product of fall-out from suspension of material suspended during high-energy events and accumulation during fair-weather conditions. That mud can be deposited during fair-weather, in the presence of permanent or semi-permanent currents, has been proven by the study of McCave and Swift (1976) who suggested a mechanism of viscous sub-layer diffusion from a water column with a low suspended mud concentration.

In summary, facies 6 represents deposition in an environment favouring the preservation of sand, and with an abundant sediment supply. Individual wavy sandstones are thought to represent peaks in environmental energy, perhaps due to trade wind season geostrophic flow with a superimposed peak (?spring) tidal flow, rather than diurnal tidal velocity variations. Muds were deposited over rippled sand surfaces during fair-weather in an environment influenced by permanent or semi-permanent currents.

4.2.8 Facies 7 - Wave-rippled Sandstone

This facies only occurs in the A39 road section where the single unit of this facies it is shown as facies E on the log (text-figure 4.1).

4.2.8.1 Description

A single 7 cm. thick coset of very fine to fine grade sandstone, displaying 1 to 3 cm. thick sets of cross-laminations with scooped set bases and symmetrically rippled tops constitutes facies 7. Ripple crests are

frequently truncated by planar erosion surfaces. Nevertheless, symmetrical ripple forms are occasionally preserved under a thin blanket of massive sandstone forming the base of the succeeding set. Cross-laminations are low-angled, occasionally swinging-up from near-horizontal laminations, laminae frequently thinning laterally to give a swollen-lens morphology. Individual sets usually display a unidirectional foreset dip direction, although both laterally and vertically adjacent sets may display opposing dip directions; lamination below ripple forms is generally form-discordant. Observed ripple crests trend ENE-WSW, a direction oblique to the SSW dipping regional palaeoslope and normal to palaeo-SE trade winds.

4.2.8.2 Biofacies

Several sinuous convex epirelief trails (?*Aulichnites* sp.) have been observed on rippled surfaces. The rippling of this facies suggests that the environment was well aerated and it is therefore assigned to Schäfer's (1972) vital lipostrate biofacies (see enclosure 4).

4.2.8.3 Interpretation

The above characteristics are diagnostic of wave-generated ripples (cf. de Raaf *et al.* 1977). Facies 7 is interpreted as the product of relatively sustained wave activity at, or above, fair-weather wave-base.

4.2.9 Facies 8 - Cross-bedded Sandstones

This facies only occurs in the east Lynmouth beach (text-figure 3.1B) and Duty Point (text-figure 4.2 - 4.5) of the logged sections where it is shown as facies E on the logs.

4.2.9.1 Description

Cross-bedded cosets range from 11cm to 88cm in thickness at Lee Stone but only reach a maximum thickness of 15cm at east Lynmouth beach; individual sets range from 3cm to 31cm in thickness at Lee Stone and between 4cm and 15cm at east Lynmouth beach, although the thinner sets appear to represent penecontemporaneous erosion of originally thicker sets (plate 4.18C). Individual foreset laminae range from 1mm to 15mm in thickness, laminae frequently thinning in a down-foreset direction. The thicker laminae tend to occur in sets of a coarser grain-size ranging from very fine to medium sand in grade. The finer sandstones tend to be well sorted and lithologically mature (sub-litharenites), the medium-grade sandstones

being only moderately to well sorted and containing more bioclastic debris (comminuted shell debris and crinoid ossicles) and rock fragments, although they are still classified as sub-litharenites. Cross-bedded sandstones are a major component of the 'Lee Stone facies association.'

Cross-bedded sets at the top of the 'Lee Stone facies association' at Lee Stone are locally cut by NW-SE trending micro-channels which have a maximum width of 30cm and depths of up to 7cm (plate 4.19B).

The cross-bedding in facies 8 may be divided into two broad types: planar cross-bedding and trough cross-bedding. These two types, however, are end-members of a broad, indivisible spectrum, all intermediate types having been observed in facies 8.

Sets with predominantly planar foresets tend to occur directly above muddy heterolithic units, in isolated cosets within muddy heterolithic facies (e.g. plate 4.18B, 0.3 - 0.8m on the log of the outcrop above Lee Stone - text-figure 4.3), or at the margins of the 'Lee Stone facies association' (plates 4.17A & 4.18A) and the lower half of the 'Lee Stone facies association' exposed at east Lynmouth beach (text-figure 3.1B). Sets attain a thickness of up to 35cm (e.g. 0.7m on log shown in text-figure 4.4). Set bases are planar and/or erosive, shallow scours up to 10cm in diameter and 1cm in depth occasionally being observed on faces parallel to the palaeocurrent direction. Bounding surfaces to sets tend to be planar and parallel, a small angle of down-climbing occasionally being observed e.g. coset shown in plate 4.18B, resulting in wedge-shaped sets (plate 4.18B) cf. de Raaf and Boersma (1971), Anderton (1976) and Levell (1980b). Individual foresets meet the set base asymptotically, although the contact tends to be more angular in sets immediately above muddy facies e.g. plate 4.18B. Occasionally foreset surfaces have a mudstone drape. Intraformational mudstone clasts are scarce and tend to rest near to the toes of foresets.

Trough cross-bedded sets, in which the foresets are concave-up, tend to occur within the core of the 'Lee Stone facies association' e.g. plate 4.17C. Sets range up to 24cm in thickness (plate 4.17C). In vertical sections parallel to the palaeocurrent direction the set bases are generally scooped (plate 4.17C), although sets lying above muddy units tend to be more planar (plate 4.18C). In vertical sections normal to the palaeocurrent direction, the set bases comprise a series of intersecting scour troughs i.e. 'festoons'. The foreset laminae infilling the scour troughs rest concordantly above the scour trough base, the infill being symmetric. Rarely, the laminae infilling the scour trough 'onlap' against the scour trough base, the infill

being asymmetric. Individual foresets are concave-up and 'spoon-shaped', approaching the scour trough base asymptotically or, more rarely, passing into planar toesets. Individual sets have lateral 'persistence ratios' (i.e. visible lateral extent ÷ mean thickness - Anderton 1976) values ranging from 5 to 20. Some sets are visibly lens-shaped e.g. set at 13.5m on log shown in text-figure 4.2.

In trough cross-bedded cosets the set bounding surfaces are virtually never parallel, either converging (plate 4.17B) or diverging in a down-palaeocurrent direction giving wedge-shaped sets (plate 4.17B) cf. de Raaf and Boersma (1971), Anderton (1976) and Levell (1980b). Sets frequently down-climb the bounding surfaces. Occasionally, the upper surface to sets is convex-upwards (plate 4.17B & C) cf. Anderton (1976) and Levell (1980b). Reactivation surfaces (cf. Collinson 1970) are rare. Foreset surfaces frequently have a mudstone draping lamina which usually thins up the foreset surface and often pinches-out completely (plate 4.17C). Mudstone intraclasts are common in trough cross-bedded sets, particularly in sets which overlie mudstone-draped pause planes e.g. set at 13.4m on log shown in text-figure 4.2. In trough cross-bedded sets mudstone intraclasts tend to lie on the upper parts of individual foresets (plate 4.19A).

Cosets within facies 8, comprising sharply bounded sets of cross-bedding, are bounded by three types of surface:

4.2.9.1.1 Erosion surfaces with or without a mudstone drape

The erosion surfaces are divisible into two types:

(i) Discontinuous planar surfaces which may be traced laterally for distances of up to 1.2m. Mudstone drapes, if present, reach thicknesses of up to 5mm. Where the laterally discontinuous erosion surfaces terminate, the cross-bedded set overlying the erosion surface 'steps-down' and a larger cross-bedded set results cf. 'hanging set-boundary' of Allen (1973) and Levell (1980b).

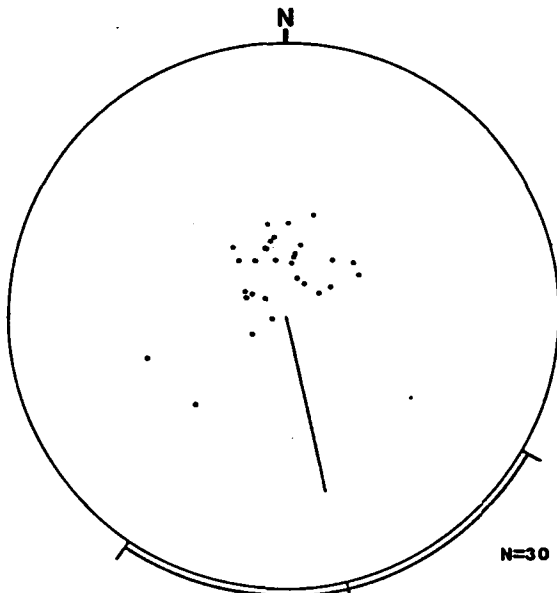
(ii) Laterally extensive erosion surfaces which may be traced laterally for distances of up to 7m, having a relief of up to 4cm. Silty-mudstone drapes, reaching thicknesses of 10mm, are frequently present. The laterally extensive planar, or occasionally convex-up, erosion surfaces dip at angles between 5° and 8° towards the SSW, the cross-bedded sets down-climbing these surfaces obliquely, the cross-beds migrated

towards the SSE (Lee Stone = text-figure 4.17, single planar cross-bedded set at east Lynmouth beach = text-figure 3.1B, observations of facies 8 exposed at Ruddy Ball also indicate palaeocurrents directed towards SSE), the palaeocurrent vector magnitude being high for facies 8 and unimodal.

4.2.9.1.2 Rippled Coset Tops

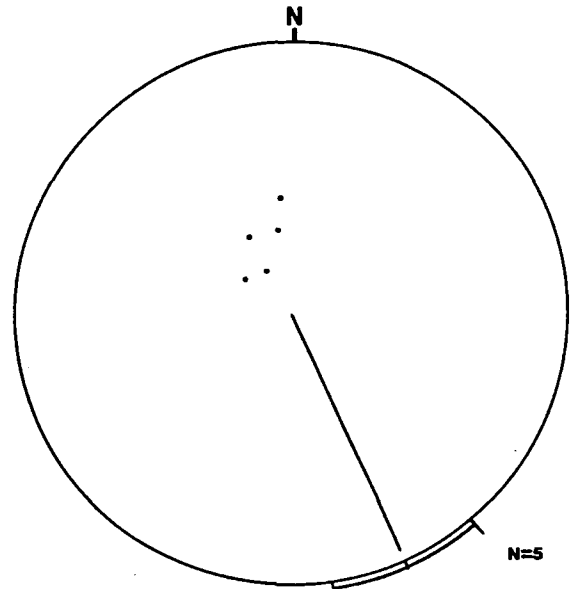
Frequently the upper boundaries of cosets are ripple cross-laminated, the cross-laminae resulting from the interplay, to varying degrees, of unidirectional and oscillatory currents. In detail, the cross-lamination styles and ripple forms do not differ from those observed in facies 6, described in section 4.2.7.1, to which the reader is directed for a detailed account. A summary of the characteristics of the ripple cross-lamination styles and ripple forms in facies 8 is given below.

Pure current-ripples appear to be absent from facies 8, although a wide spectrum of combined-flow ripples generated by predominantly unidirectional currents have a rounded, strongly asymmetric ripple profile, ripple symmetry indices (i.e. length of horizontal projection of stoss side \div length of horizontal projection of lee side) ranging from 2.6 to 3.6, the vertical form index (i.e. ripple length \div ripple height) ranging from 8 to 14. Combined-flow ripples generated by predominantly oscillatory currents are symmetrical, or nearly symmetrical, in profile with ripple troughs and the majority of ripple crests being rounded, although trochoidal crests were observed occasionally. The ripple symmetry index reaches a maximum value of 1.9, the vertical form index ranging from 5 to 12. The internal lamination styles of the ripples in facies 8 are identical to those in facies 6 (described in section 4.2.7.1).



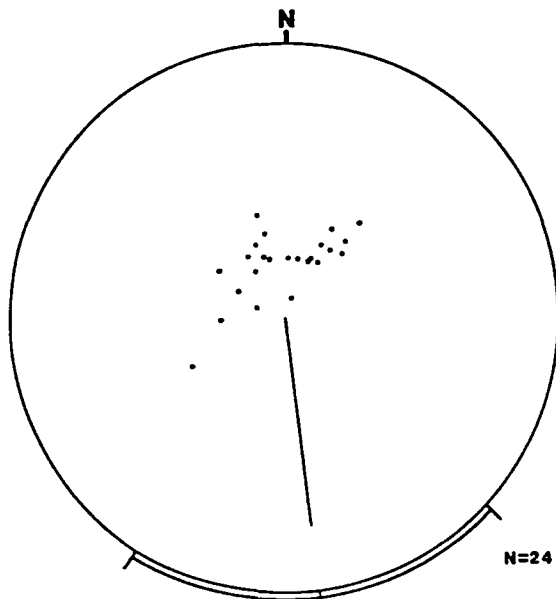
A. WESTERN LEE STONE

μ_0 = Mean vector direction = 167°
 L = Estimate of spread of angular values = 66.3%
 S = Mean angular deviation = $\pm 47^\circ$
 Rayleigh test of significance = $< 10^{-5}$ ∴ non-random



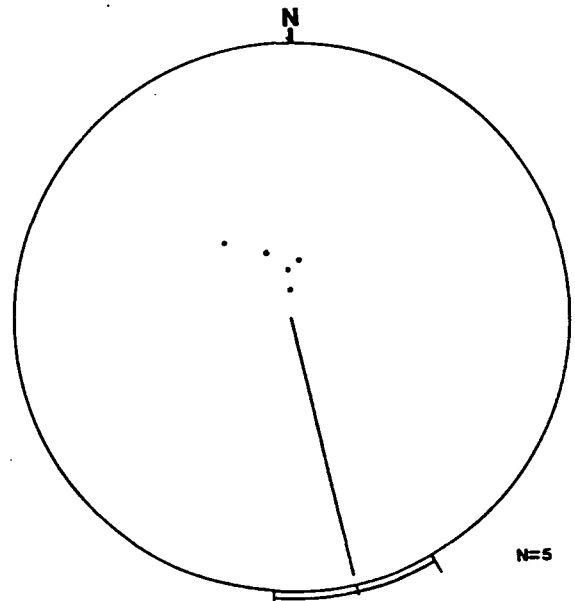
B. WESTERN LEE STONE - UPPER SHOAL

μ_0 = 155°
 L = 95.9%
 S = $\pm 16^\circ$



C. EASTERN LEE STONE

μ_0 = 174°
 L = 75.2%
 S = $\pm 40^\circ$
 Rayleigh test of significance = $< 10^{-5}$ ∴ non-random



D. DUTY POINT CLIFF TOP

μ_0 = 167°
 L = 95.6%
 S = $\pm 17^\circ$

Text-fig. 4.17 Lee Stone: Facies 8 - Foreset Vector Means

In plan view it is apparent that 'two-dimensional' straight to slightly sinuous crested combined-flow ripples are rare in facies 8, the majority of combined-flow ripple crestlines intersecting in a complex interference pattern. As in facies 6, the interference ripple crestlines do not conform to a classic 'brick', 'tile' or hexagonal pattern. The facies 8 interference ripples comprise ripple crests that have become subsequently reworked by currents generating ripple crests at an angle to the initial set. The currents that generated the modifying set, however, did not operate for a sufficient period of time to completely rework the underlying ripple set. This resulted in a thin veneer of rippled sandstone, with crestlines at an angle to partially preserved underlying ripple crests, giving a 'hummocky' topography to the ripple.

More rarely, 'three-dimensional vortex (wave) ripples' may be distinguished on facies 8 bedding planes, although such ripples can only be differentiated from combined-flow ripples if bedding plane views are available as the two ripple types display identical characteristics when viewed in vertical section (see 4.2.7.1). Straight-crested ripples formed by oscillatory currents with a velocity symmetry have not been observed in facies 8.

Occasionally, ripple profiles in facies 8 are overlain by a thin mudstone layer, which may be continuous or discontinuous depending on the degree of erosion during the emplacement of the overlying cross-bedded set. As in facies 6, the deposition of the mud appears to have been incremental, as is evidenced by the occurrence of 'spill-over aprons' on ripple flanks, interdigitating with the mudstone infilling the ripple trough.

4.2.9.1.3 Biogenic Structures Descending from the Upper Bounding Surface of Cross-bedded Cosets

Biogenic structures frequently descend from the upper bounding surface of cross-bedded cosets, or exploit the mudstone laminae that overlie erosion surfaces. A detailed account of the biogenic structures in facies 8 is given in section 4.2.9.2.

In summary, facies 8 is internally subdivided by a hierarchy of 'pause planes' (*sensu* Terwindt 1981 - structural diastems' of Boersma 1969):

(i) Concordant, non-erosive drapes of mudstone, 1 to 3mm in thickness, resting on foreset surfaces.

Frequently, the mudstone lamina thins up the foreset surface, occasionally pinching-out completely.

(ii) Discontinuous, planar or convex-upwards surfaces that are either erosional or non-erosional. These surfaces are frequently overlain by a mudstone lamina.

(iii) Laterally extensive, planar or convex-upwards erosion surfaces. A silty-mudstone lamina, ranging up to 10mm in thickness, is normally present. These surfaces dip at 5° to 8° towards the SSW, the cross bedding of facies 8 down-climbing these surfaces obliquely.

Rarely, units of parallel-laminae are intercalated with the cross-bedded sets of facies 8 at Lee Stone. The sets of parallel-laminae reach thicknesses of up to 7cm and are laterally impersistent, being erosively truncated by cross-bedded units.

4.2.9.2 Biofacies

As is shown in enclosure 4, the ichnofaunal assemblage preserved within facies 8 closely resembles those observed in facies 5 and 6. *Palaeophycus tubularis* and *P. heberti* descend from the upper bounding surfaces of cosets, bedding plane views indicating that both ichnospecies tend to occur in dense clusters. In addition, the upper few centimetres of cosets have frequently been biogenically destratified (fossitextura deformativa). More rarely, *P. tubularis* burrows exploit the mudstone drapes resting on foreset surfaces and intraformational mudstone clasts that are disseminated throughout cosets. The mudstone drapes resting on coset bounding surfaces have almost universally been exploited by *P. tubularis* burrows, many of the drapes locally exhibiting a 'swirled' appearance i.e. fossitextura deformativa.

Examination of the many, well exposed rippled bedding plane views through facies 8 reveals a large number of well preserved epireliefs of *Gyrochorte cosmosa*, *Helminthorhapha japonica* and *Phyllocites* sp.

The cross-bedded sandstones of facies 8 are interpreted to be the product of a semi-permanent trade wind generated geostrophic current system with a superimposed weak tidal régime (see section 4.2.9.3). The *P. tubularis* and *P. heberti* burrows represent the colonisation of a sandy substrate by tube-dwelling vermiform organisms during relatively long periods of bedform inactivity. Many of the *P. heberti* and *P. tubularis* burrows stand proud of the rippled surfaces in a 'chimney-like' fashion, attesting to the resistance of the mucus-bound burrow walls to penecontemporaneous erosion.

In common with facies 1, 5 and 6, epireliefs of *H. japonica* on rippled sandstone bedding planes are believed to represent the response of an opportunistic foraging animal to muds with a low organic content in the zone immediately above the rippled surface. Epireliefs of *G. cosmosa* and *Phyllodoites* sp., in common with facies 1, 5 and 6, are preserved due to the granulometric contrast between sand and overlying muds.

The ichnofauna of facies 8 comprises the burrows (*P. tubularis* & *P. herberti*) of vermiform organisms which subsisted in whole, or in part, on a diet gathered by suspension feeding. The presence of burrows of suspension-feeders suggests that semi-permanent currents were active during the deposition of facies 8, a conclusion which is consistent with the presence of multiple phases of 'spill-over aprons'. The interface between rippled coset upper bounding surfaces and overlying mudstones has been exploited by the burrows of infaunal deposit-feeders (*G. cosmosa*, *H. japonica* & *Phyllodoites* sp.) In summary, the ichnofauna of facies 8 represents the burrows of organisms that lived in a well-aerated environment which was subjected to periods of higher environmental energy driving bedform migration. Facies 8 is, therefore, referred to Schäfer's (1972) vital-lipstrate biofacies.

4.2.9.3 Interpretation

Facies 8 is a major component bedding type of the 'Lee Stone facies association', although it is volumetrically subordinate to facies 6. In addition, isolated cosets of cross-bedding, referred to facies 8, locally occur in heterolithic lithologies of facies 1 type e.g. outcrop above Lee Stone (0.3m to 0.8m on text-figure 4.4).

The dominant stratification types of facies 8 consist of trough cross-bedding and planar cross-bedding, representing the migration of sinuous-crested and straight crested dunes respectively, termed 'three-dimensional dunes' and 'two-dimensional dunes' by Allen (1982, vol. A, chap. 8). A full list of synonyms of Allen's dune types is given in section 4.2.4.3. Allen noted that a substantial overlap occurs between the two dune types, which he took as indicating that only one hydrodynamic class of bedform was present. Allen further noted that three-dimensional dunes are created under more 'central' conditions, two-dimensional dunes arising near to either the lower (probably) or upper (unlikely) limit of the dune existence field. As with dunes in facies 3, this indivisibility of dune bedforms was encountered in facies 8, planar cross-bedded sets

frequently displaying incipient scours, a gradation, albeit abrupt, being recognised between planar and trough cross-bedding.

As was noted in section 4.2.4.3, a vast amount of data (summarised in Harms *et al.* 1982) has enabled existence fields for dunes to be plotted. In summary, two-dimensional, straight-crested dunes tend to be formed by relatively weak flows, whilst three-dimensional, sinuous-crested dunes tend to be formed by comparatively stronger flows. In the 'Lee Stone facies association' planar cross-bedding tends to occur at the margins of the facies association, trough cross-bedding occurring in the core of the facies association. The 'Lee Stone facies association' is interpreted as a sand ridge (see section 4.3.1.3). Thus, the distribution of cross-bedding types within the 'Lee Stone facies association' suggests that environmental energy increased towards the sand ridge crest, as would be expected. In addition to the 'Lee Stone facies association' sequential setting, planar cross-bedded sets also tend to occur within sequences of more muddy lithology, probably reflecting the inability of lee vortices to erode scour pits in the cohesive mud that underlay the planar cross-bedded sets. Furthermore, the isolated cosets of cross-bedding set within facies 1 type lithologies, interpreted as representing duned sand patches migrating across a muddy substrate (see below), are generally comprised of planar cross-bedded sets. This suggests that the currents which induced the migration of the duned sand patches were weaker than those that moulded the sand ridge.

Although the cross-bedding of facies 8 exhibits a unimodal palaeocurrent distribution (text-figure 4.17), many of the characteristic features of the cross-bedding indicate that tidal currents were, at least in part, responsible for enhancing semi-permanent unidirectional currents and thus enhancing the ability of currents to support bedload transport of the facies 8 substrate. The following features are diagnostic of a tidal influence; a detailed discussion of each feature is given in section 4.2.4.3, the subsection being indicated by the Roman numeral in parentheses after each feature, along with the interpretation of previous authors: Sharp set boundaries (i), wedge-shaped sets (ii), hanging set boundaries (iv), reactivation surfaces (v). It should be noted that no single characteristic provides unequivocal evidence for tidal activity; rather, it is the complete suite of characteristics, when considered together, that suggests a tidal influence.

Thin, laterally impersistent sets of parallel-laminae are rare in facies 8, tending only to occur towards the upper centre of the 'Lee Stone facies association' at Lee Stone when considered in a three-dimensional sense. The cross-bedding in facies 8 indicates that the environmental energy increased towards the central

crestal region of the 'Lee Stone facies association' (see above). It is suggested, therefore, that the parallel-laminae are of upper plane bed origin, representing short periods of increased environmental energy (?storms), during which dune bedforms were 'washed out'. With a return to lower environmental energy dune bedforms became re-established and would have laterally truncated the parallel-laminae sets, resulting in the preservation of thin, laterally impersistent sets of parallel-laminae. Sets of parallel-laminae interpreted to have been the product of upper plane bed conditions have been recorded from ancient sandstone-bodies interpreted to have been deposited by tides (Banks 1973b - 'facies 4'; Levell 1980a - 'Plsd facies'; 1980b - 'facies 3') and storms (Tillman & Martinsen 1984 - 'central bar [planar laminated] facies').

Subdividing facies 8, and thus the 'Lee Stone facies association,' are a series of internal discordances termed 'pause planes' (*sensu* Terwindt 1981). The three types of 'pause plane' recognised by Terwindt have also been recognised in facies 8 (see also facies 3 - section 4.2.4.3):

(i) Foreset drapes comprising a concordant, non-erosive drape of mudstone resting on foreset surfaces.

Drapes of this type were attributed by Terwindt and Boersma (1981) to sedimentation from suspension during tidal slack-water phases.

(ii) Limited, generally erosional, pause planes, frequently with a mudstone drape, represent truncation planes of the extensive type.

(iii) Extensive erosional pause planes, normally having a mudstone drape and/or wave / combined-flow rippling.

Low angle, laterally extensive discontinuity surfaces are common in ancient subtidal sand-bodies and have been described by Anderton (1976), Johnson (1977), Hobday and Reading (1972) and Nio (1976). A full discussion of these examples has previously been given in section 4.2.4.3. In summary, facies 8 dunes are interpreted as having migrated in response to semi-permanent wind-induced currents during the trade wind season, reinforcing a weak tidal régime. Peak energy would have been attained during the conjunction of spring tides and/or periods of strong trade winds, when current velocities would have been sufficient to initiate dune migration. The highest values of bed shear stress would have been achieved at the bar crest, where the relatively shallow conditions would result in the greatest 'wave-effectiveness' at the sea-bed, sufficient to sustain the migration of three-dimensional dunes. These dunes migrated down major SSW-dipping bar face surfaces, locally scouring into the surfaces; SSW-wards accretion of the bar would have

resulted. Exceptionally, currents were sufficiently strong to allow upper phase plane bed conditions to become established, resulting in dunes becoming 'washed-out' to give localised parallel-laminated deposits.

The ridge flanks would have experienced lower energy conditions due to lower 'wave-effectiveness' at the bed as a function of greater water depth; two-dimensional dunes would have migrated in these lower energy conditions. In interbar zones two-dimensional duned, isolated sand patches (e.g. 0.3 - 0.8m on log shown in Text-fig. 4.3) would have been mobilised (cf. Johnson 1977; Levell 1980a). These sand patches would have been out-of-phase with the large-scale hydrodynamic processes responsible for sweeping the major sand ridges together on the muddy substrate. as such, the duned sand patches would have acted as a major source of sand, supplying sand in an offshore direction to the sand ridges.

During lower energy periods (neap tides and low / moderate energy trade winds), migration of the dunes would have been relatively slow, possibly being restricted to the bar crest. When migration periodically ceased completely, dune foreset surfaces became draped with mud. Wave action on the ridge flanks would have generated the major SSW-dipping pause planes. Turbulence at the sea bed would have been sufficient to allow colonies of suspension-feeding organisms to establish. Bar height may have been restricted by fair-weather waves (cf. Johnson 1977).

The lowest energies encountered during the formation of facies 8 would have occurred during the (?monsoon) summer season, a period when dune migration would have ceased. Mud derived from low-flow stage river outflow would have been deposited on foreset surfaces and on the major pause planes. Periods of high energy would occasionally occur during the summer (?monsoon) season due to the passage of tropical cyclones. Tropical cyclones are relatively short-lived events and do not effectively couple with the sea surface to give a major shelf flow circulation pattern (Swift *et al.* 1983). As a result the product of occasional tropical cyclonic events would be dominantly erosional, possibly reducing the bar crest height and generating extensive wave-rippled surfaces.

4.2.10 Facies 9 - Isolated Trough Cross-bedded Sandstone Set

This facies only occurs in the A39 road section, where it is shown as facies F on the log (text-figure 4.1).

4.2.10.1 Description

A single unit of trough cross-bedded, medium grade sandstone is exposed within the A39 road section (at 7.5m on the log). The trough cross-bedded set is 10 to 12 cm. in thickness, individual foreset laminae displaying an asymptotic relationship to the lower set boundary. Although the direction of foreset dip could not be measured accurately, an approximately northward dip could be discerned i.e. palaeo-onshore. The nature of the exposure precludes a more detailed description being given.

4.2.10.2 Biofacies

The unit has been extensively biogenically 'churned', resulting in the total localised obliteration of primary structures, and is therefore referred to Schäfer's (1972) vital-lipstrate biofacies.

4.2.10.3 Interpretation

The interpretation of facies 9 is enigmatic due to the absence of detailed data. Nevertheless, the lack of structures diagnostic of tidal activity, and the presence of predominantly wave generated structures in the remaining facies within the A39 road section, suggest that facies 8 is the product of sinuous-crested megaripple migration in response to wave generated currents. (see also facies 11 interpretation).

4.2.11 Facies 10 - Thin-bedded Horizontally-laminated Sandstone

This facies only occurs in the east Lynmouth beach section, where it is shown as facies F on the log (text-figure 3.1B).

4.2.11.1 Description

Facies 10 comprises laterally persistent beds of faint, horizontally-laminated fine sandstone which occasionally fine upward to coarse siltstone. Individual beds are 2 to 8 cm. in thickness and have sharp,

planar bases. Bed tops are planar but gradational, frequently displaying a rapid upward-fining trend;. Although internal lamination is generally horizontal, laminae occasionally display a broadly undulatory form.

4.2.11.2 Biofacies

Localised bioturbation descending from bed tops is common and occasionally develops into an intensely 'churned' texture resulting in the obliteration of primary bedding structures. This facies is referred, therefore, to Schäfer's (1972) vital-lipstrate biofacies.

4.2.11.3 Interpretation

Thin beds of sandstone comprising horizontal-laminations as the only internal structure are comparatively rare in shallow marine sequences preserved in the geological record (e.g. Goldring & Bridges 1973, Anderton 1975). Such beds are however, commonly observed within cores recovered from modern shelf sequences (e.g. Reineck *et al.* 1967, 1968, Gadow & Reineck 1969, Reineck & Singh 1972, Howard & Reineck 1981). The sharp bed bases, fining-upwards, gradational bed tops and bioturbation descending from the upper surface of beds suggested an 'event deposit' attributable to storm action to the above authors.

Allen (1982) proposed a physical model for thin shelf sandstone units of storm origin. Allen suggested that in zones proximal to a sand source, an idealised waning flow sequence would display parallel-lamination of upper phase plane bed origin with an overlying zone of wave-current cross-lamination. The applicability of this model to the facies 10 units is brought into question by the consistent absence of an upper zone of wave-current cross-lamination. It is suggested therefore, that the horizontal-lamination of facies 10 is not the product of an upper phase plane bed flow regime - a contention supported by the absence of primary current lineation and generally faint lamination.

The model presented in Reineck and Singh (1972), whereby sand entrained in suspension clouds settles during the waning phase of a storm to give horizontal-lamination, appears more appropriate in explaining the generation of facies 10. Reineck and Singh suggested that: "As generally only slow bottom currents are present, the sediment deposited is not reformed into ripples, but sedimented in the form of laminated sand" (p.123). The facies 10 beds are interpreted to be the product of sand fall-out from suspension under the

influence of wave-orbitals just below storm wave-base, i.e. wave orbitals did not touch the bottom, in a zone proximal to sand supply.

4.2.12 Facies 11 - Thick-bedded Parallel-laminated Sandstone

This facies only occurs in the A39 road section, where it is shown as facies G on the log (text-figure 4.1).

4.2.12.1 Description

A series of 4 to 114cm. thick sets of parallel-laminated coarse silt to fine sand-grade sandstones, referred to as facies 11, dominate the upper 2/3rds of the A39 road section. The sandstone is well sorted and comprises an average of 95% sub-angular quartz grains with rare mica and ?pyrite; silty material is present in samples from samples of this facies collected from lower parts of the section. Individual laminae are parallel and display an average spacing of 1 to 3mm. Laminae are generally 'flat', although a gentle undulation is occasionally visible e.g. 7.6m on the log. No primary current lineation has been observed.

Set bases generally display a low-angled, laterally extensive erosion surface which is planar with up to 1cm. relief; this gives sets a very gentle 'wedge' geometry. Occasionally, set bases display a massive sandstone draping, and thereby preserving, underlying wave-ripple crests of the preceding unit. Set tops are either planar, frequently displaying truncation of laminae by the succeeding parallel-laminated set, or wave-rippled (plate 4.21A). Where observed, wave-ripple crestlines trend ENE-WSW, a direction oblique to the SSW dipping palaeoslope; ripple profiles display an asymmetry towards the SSE indicating a palaeo-offshore flow component.

Cosets are generally separated by 1 to 2cm. thick units of lenticular bedding (facies 1), although single units of graded mudstone (facies 2), bioclastic sandstone (facies 4), wave-ripple cross-lamination (facies 7) and trough cross-bedded sandstone (facies 9) also separate cosets.

4.2.12.2 Biofacies

In the logged sequence there is an overall coarsening-upwards trend within facies 11, accompanied by an upwards decrease in biogenic disruption of primary structures. At the base of the facies 11 sequence (plate

4.13A) the parallel-laminated units are silty and have a high mud content, muddy wisps being commonly preserved. Biogenic 'churning' is ubiquitous, resulting in the total loss of primary bedding structures locally. Sandstone filled *Palaeophycus tubularis* burrows are common locally. Sinuous convex epirelief trails (?*Aulichnites* sp.) are occasionally observed on combine-flow ripple surfaces.

The upper reaches of the facies 11 sequence (Plate 4.13B) comprise well-sorted, clean, fine grade sandstone, the mud and biogenic 'churning' component visible in the lower part of the sequence having been lost. The upper part of the sequence is also characterised by the presence of U-shaped tubes, referable to the ichnogenus *Arenicolites*. Two ichnospecies can be distinguished: tubes penetrating to a depth of up to 5 cm. are referred to *A. sp. a* ; tubes penetrating to depths of up to 30cm. are referred to *A. curvatus* (see appendix B for detailed descriptions and discussion). The tubes of both ichnospecies in the A39 road section are always truncated by penecontemporaneous erosion. *Arenicolites* tubes are characteristically found within littoral to shallow sublittoral sequences (Howard 1978).

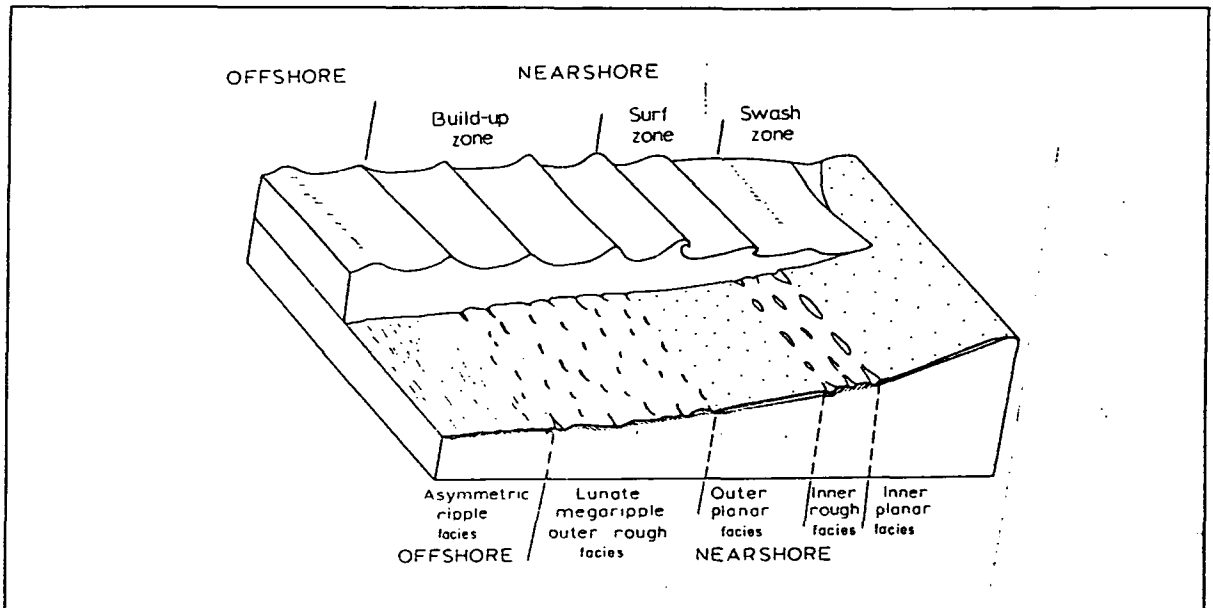
Measurements of the *A. Curvatus* tubes preserved within the A39 road section (text-figure B1) indicate that the plane of the U-tubes have a mean dip of 80° to the SSE (offshore), the mouths of the tubes opening into the offshore flowing mean palaeo-flow. This phenomenon is attributed to the response of a vermiform suspension feeder, dwelling in the U-tubes, directing its feeding apparatus into the prevailing current.

In summary, facies 11 is referred to Schäfer's (1972) vital-lipostrate biofacies.

4.2.12.3 Interpretation

The absence of both a well developed wedge-shaped set geometry and inversely graded laminae (cf. Clifton 1969) suggests that the parallel-lamination of facies 11 was not generated in the swash zone. Furthermore, none of the following characteristic features of sub-aerial exposure have been observed in facies 11: 'ladder ripples' (Bajard 1966), swash marks, antidune lenses, rill marks, flat topped or double-crested wave-ripples, wave-ripples with rounded crests and pointed troughs (Reineck & Singh 1980) or run-off channels (Clifton *et al.* 1973).

Clifton *et al.* (1971) observed flat to broadly undulating laminated sands in the transition between the outer portion of the surf zone and the inner portion of the zone of wave build up. This observation was made during a study of part of the Oregon coast, a non-barred high energy coastline environment. Clifton *et al.* termed the above parallel-laminated zone the 'outer planar facies' (see text-figure 4.18). The parallel-lamination was ascribed as to an upper flow régime origin under conditions of pseudo-unidirectional flow generated by flow asymmetry under shoaling waves. During periods of lower wave energy, wave-ripples become established in this zone.



Text-fig. 4.18 Zonation of wave activity and facies of sedimentary structures within and adjacent to the high-energy nearshore.

After Clifton *et al.* (1971).

Immediately offshore of the 'outer planar facies' an 'outer rough facies' of lunate megaripples with landward dipping foresets, produced by shoaling waves with a strong flow asymmetry, is developed (see text-figure 4.18); this zone is analogous to facies 9 cross-bedding. Cores recovered by Clifton *et al.* commonly contained a mixture of structures characteristic of the 'outer planar' and 'outer rough' facies cf. A39 road section sequence.

Howard and Reineck (1981) recorded a parallel-laminated sand zone extending from the backshore to -9m water depth in nearshore transects reported from the modern high-wave-energy Californian shoreline which also strongly resembles facies 11. Extensive vibracoring revealed that in many areas the bulk of the shoreface comprised laminated sand, with interbedded cross-bedded sands only occurring occasionally.

Similarly, Howard and Reineck (1972) in a study of the low-energy, mixed wave- and tide-dominated Sapelo Island (Georgia, U.S.A.) coastline also recorded thick parallel-laminated sands extending across a large portion of the upper shoreface and foreshore.

In conclusion, facies 11 and the associated facies 9 are proposed as having developed in the zone of wave build-up and outer surf zone in a nearshore, wave-dominated environment cf. Clifton *et al.* (1971); flow is directed onshore in this zone. Several asymmetric wave-ripple profiles and the dip of the *Arenicolites curvatus* burrows preserved within facies 11, however, indicate that during certain periods, flow was directed offshore. This phenomenon may reflect periods of decreased wave energy when the offshore zone of 'asymmetric ripple facies' of Clifton *et al.* would have migrated onshore. Extended periods of lower energy resulted in the preservation of lenticular bedding (facies 1).

4.3 FACIES SEQUENCE ANALYSIS

4.3.1 'Lee Stone Facies Association'

The logged sections described from east Lynmouth beach (text-figure 3.1B), Lee Stone (text-figures 4.2 & 4.3), along with visual observations recorded at Ruddy Ball (see section 4.1.1.3), comprise a muddy shallow marine sequence enclosing a major sandstone-body type - the 'Lee Stone facies association' [1.46m - 3.26m, 6.83m - 9.63m & 7.20m - 10.58m on the respective logs - see also section 4.1.1 (i)]. Two sections, 40m apart, were logged through the 'Lee Stone facies association' and its enclosing deposits at Lee Stone; a summary and correlation of the two sections is shown in text-figure 4.19.

Physical and biogenic evidence discussed during the consideration of individual facies in the preceding sections indicate that the 'Lee Stone facies association' was deposited in a more turbid, slightly shallower environment than the conformable preceding and succeeding muddy deposits, inferring a sand-body that stood in positive relief relative to the surrounding muddy substrate. Furthermore, the presence of SSW-dipping 'pause planes' within facies 8 indicates palaeo-offshore lateral accretion of the sand-body. These features strongly suggest that the 'Lee Stone facies association' represents a discrete sand ridge - as opposed to an extensive sand sheet deposit representing the culmination of a regional shallowing event preceding a deepening episode.

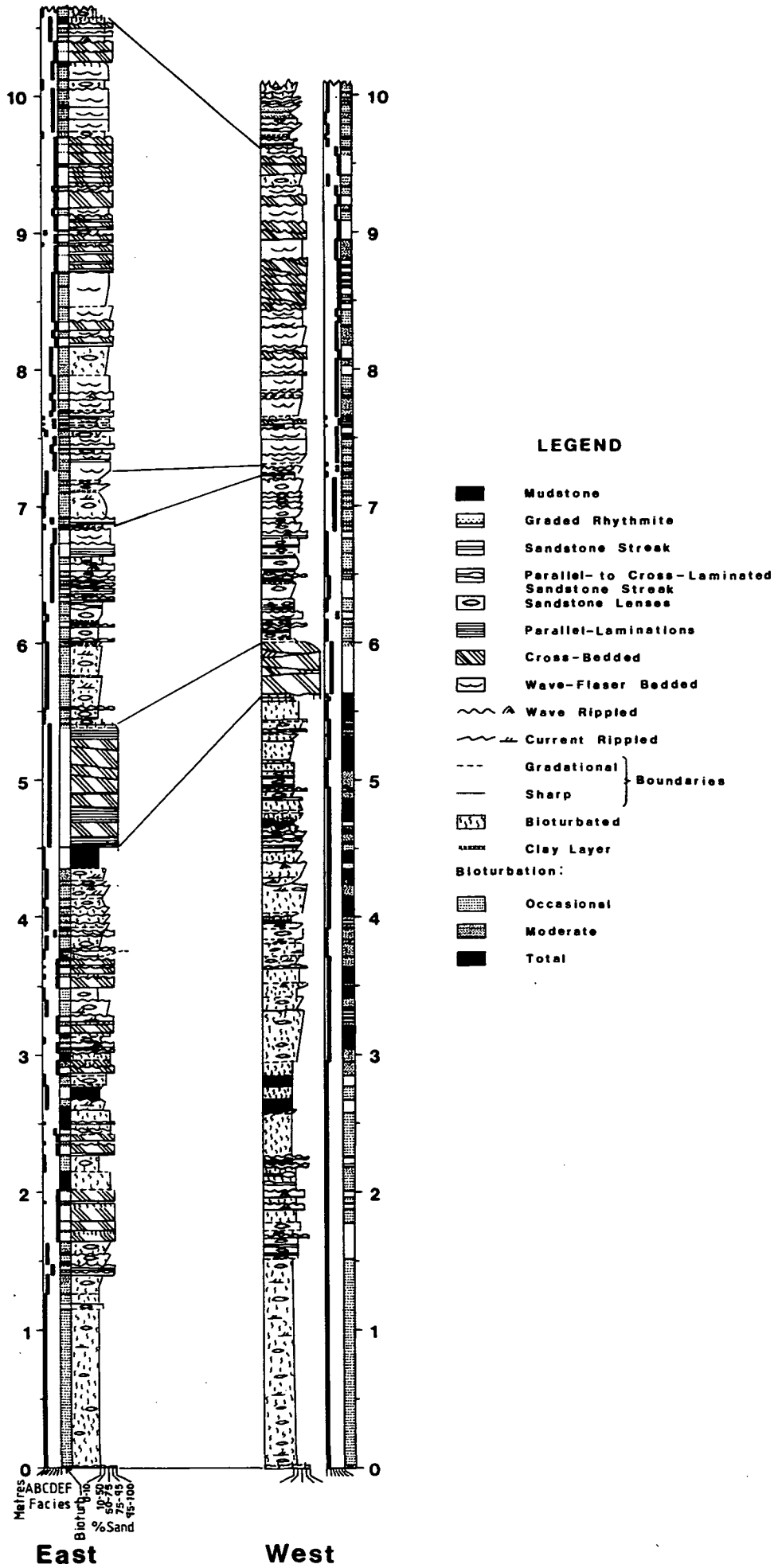
A qualitative assessment of the facies sequences at Lee Stone shown in the logs is shown in text-figure 4.19 and reveals a coarsening-upwards (CU) trend in the western sequence i.e. facies: 1⇒1'⇒5⇒6⇒8⇒1. The eastern sequence exhibits a coarsening-upwards followed by a fining-upwards (CUFU) trend i.e. facies: 1⇒1'⇒5⇒6⇒8⇒6⇒5⇒1. In both sequences the single facies 3 unit is enclosed within facies 1 units, the facies 3 unit appearing to not upset the general facies sequence trend of the two sections. A CUFU trend can also be discerned in the east Lynmouth beach section (text-figure 3.1B). The logs, however, reveal that the interdigitation of facies is complex, indicating that a quantitative approach to facies sequence analysis is justified. For this reason observed-minus-random facies transition analysis was undertaken for the three sequences, using the method described in Cant and Walker (1976); the results are tabulated in appendix E and shown graphically in text-figure 4.20, where only transitions with a greater than random probability of occurring are shown.

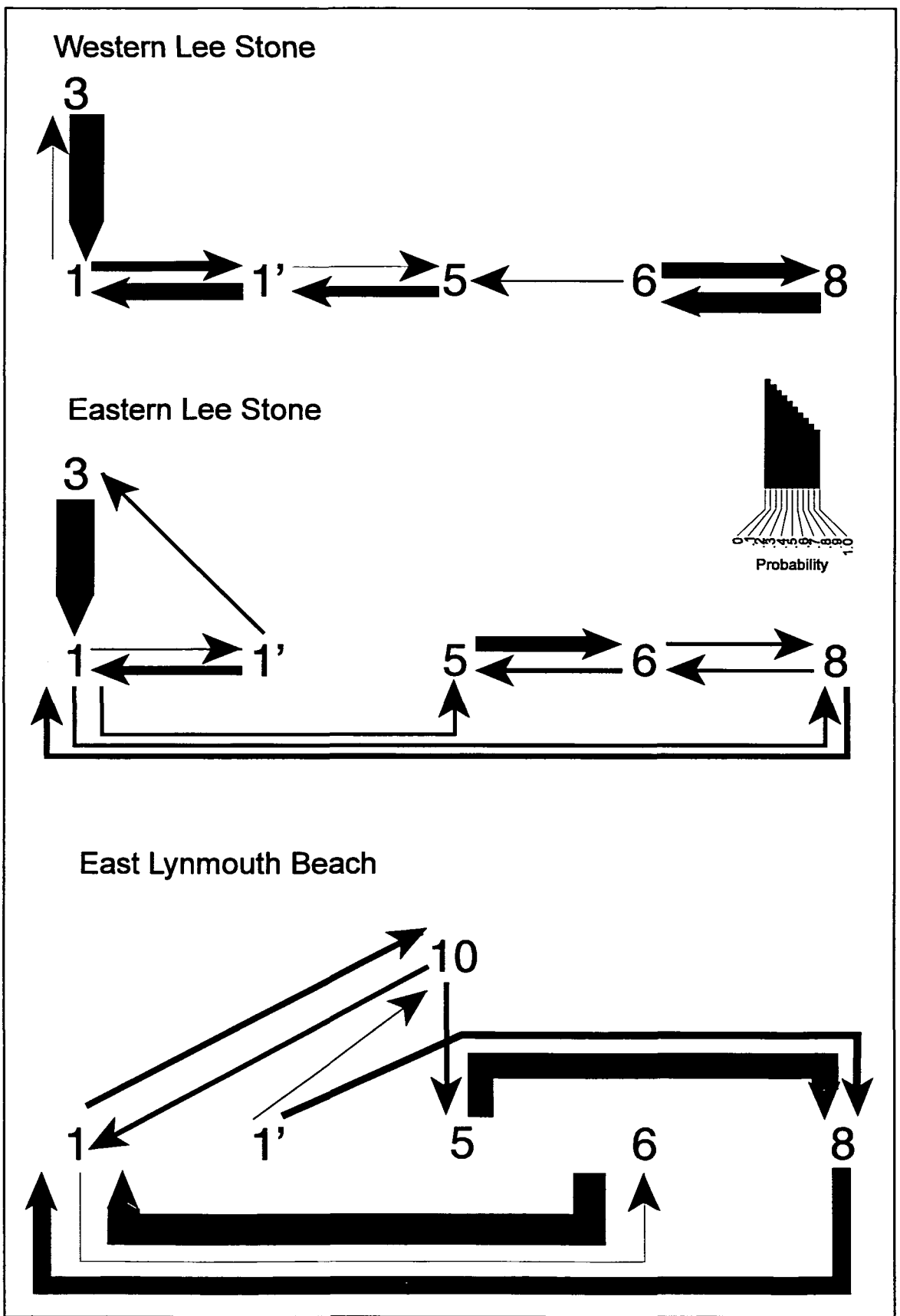
Prior to discussing the sedimentological significance of the consequent facies transitions consideration must be given to the cautionary warnings applying to this type of analysis contained in de Raaf *et al.* (1965) and Reading (1978). These authors stressed the significance of the nature of the contacts between facies, the presence of major erosional contacts implying the possibility of non-deposition, or perhaps the beginning of a new depositional cycle. An examination of the Lee Stone sequence serves to confirm that there are no major erosion surfaces for which a significant depositional hiatus / removal of facies is implied. Even the major SSW-dipping 'pause planes' of facies 8 are interpreted as fair-weather bar surfaces intervening between bar accretion deposits **within** the sandstone-body. Furthermore, there is no evidence of the development of biogenic/early diagenetic hardgrounds within the logged sequences. It appears, therefore, that the use of the facies transition analysis technique for the Lee Stone sequences is justified.

Text-fig. 4.19 Summary logs of sequences at Lee Stone along with correlation based upon prominent, laterally extensive horizons.

The lower two sets of tie-lines enclose the 'Watersmeet lithotype,' the middle tie-line connects a prominent slickensided bedding plane slip horizon, whilst the upper two sets of tie-lines enclose the 'Lee Stone facies association.' Facies: **A** = thinly interlayered sandstone/mudstone bedding; **A'** = coarsening-upwards microsequences; **B** = cross-bedded granule conglomerate; **C** = wavy bedding; **D** = flaser bedding; **E** = cross-bedded sandstones.

Lee Stone Correlation.





Text-fig. 4.20 Observed-minus-random probability diagrams, for sequences logged at Lee Stone and east Lynmouth beach, showing transitions with a greater than random probability of occurring.

Calculations shown at Appendix E. Facies: **1** = thinly interlayered sandstone/mudstone bedding; **1'** = coarsening-upwards microsequences; **3** = cross-bedded granule conglomerate; **5** = wavy bedding; **6** = flaser bedding; **8** = cross-bedded sandstone; **10** = thin-bedded horizontally-laminated sandstone. See text for discussion.

Examination of text-figure 4.20 reveals clear facies transition sequences for the two Lee Stone sections; the east Lynmouth beach section is a little less clear. Thinly interbedded sandstone/mudstone bedding (facies 1) exhibits a tendency (weaker in the eastern Lee Stone CUFU sequence) to pass upwards into coarsening-upwards microsequences (facies 1') reflecting an upwards increase in proximity of the offshore migrating sand ridge which provided the source of sand necessary to pick out the microsequences. This tendency is not seen in the less well organised CUFU east Lynmouth beach section where the CU microsequences are not so well developed and the sequence is interpreted as being more lateral in relation to a ridge centre when compared with the eastern Lee Stone CUFU sequence. At Lee Stone the reverse tendency for facies 1' to pass upwards into facies 1 is stronger; this is due to facies 1' usually being overlain by facies 1 within a sequence that is dominated by facies 1.

The single occurrence of a unit of cross-bedded granule conglomerate in the Lee Stone sections (facies 3 - the 'Watersmeet lithotype') set within units of facies 1 / 1' is reflected in the weak probability for a facies 1 / 1' to facies 3 transition; the probability for a transition from facies 3 to facies 1 is, unsurprisingly, very high!

The east Lynmouth beach section shows an approximately equal probability for transition from facies 1 to thin-bedded horizontally-laminated sandstone layers (facies 10) and the corresponding reverse transition. This is interpreted to be the result of relatively minor (random) storm events suspending sand at the ridge crest, the sand then falling out of suspension below storm wave-base on the ridge flank. At east Lynmouth beach there is a strong facies 10 to wavy bedding (facies 5) transition probability; the reverse trend is not developed. This represents a tendency for thin-bedded horizontally-laminated sandstones (deposited just below storm wave-base) to be replaced upwards by wavy bedded sandstones (deposited within storm wave-base) within an upwards-shallowing sequence.

Also at east Lynmouth beach there is a tendency for CU microsequences (facies 1') to be succeeded by thin-bedded horizontally-laminated sandstone (facies 10) and cross-bedded sandstone (facies 8); the reverse tendencies are not developed. This trend reflects interdigitation at the ridge margin, where the sand supply was sufficient to enable CU microsequences to be picked out, of fair-weather deposits and the products of minor storms (facies 10) or longer periods when unidirectional currents became established (facies 8).

The western Lee Stone section shows a weak probability for CU microsequences (facies 1') to pass upwards into wavy bedding (facies 5), a function of the upwards increase in the proximity of the palaeo-offshore-migrating sand ridge, reflected in the storm sandstones of facies 5 being thicker than those of facies 1'. The reverse trend is stronger due to the interdigitating nature of facies 1' and 5 i.e. units of facies 5 are predominantly overlain by facies 1' - although there is a general upwards tendency for facies 1' to be replaced by facies 5. The eastern Lee Stone section shows a weak probability for transition from facies 1 to 5. This phenomenon is interpreted to be a function of the western Lee Stone section representing a progradational sequence through a ridge centre, whilst the eastern Lee Stone section represents a progradational sequence through the ridge flank. As discussed previously (see section 4.2.9.3), the sand ridge migrated in a uniformly offshore direction (SSW). Thus, a point immediately down-current of the ridge centre would be subject to influx of sand to pick out CU microsequences, whereas a point off the ridge flank would only receive sands when currents deviated from its normal direction i.e. an essentially random event - in summary, a point down-current of the ridge centre would have a strong probability of receiving sands, whereas a point off the ridge flanks would only experience random events.

The eastern Lee Stone sequence exhibits a strong probability for wavy bedding (facies 5) to pass upwards into flaser bedding (facies 6), the reverse tendency being much weaker. The western Lee Stone section displays a weak facies 6 to 5 transition probability, but no reverse tendency. The lack of a facies 5 to facies 6 tendency in the western Lee Stone section is attributed to the western Lee Stone sequence having a strong coarsening-upwards trend, reflecting an upwards increase in environmental energy at the ridge centre, whereas the eastern Lee Stone section interdigitates with the surrounding muddy lithologies (CUFU trend) reflecting fluctuating environmental energy. At east Lynmouth beach facies 1 shows a weak tendency to pass directly into facies 6, the reverse trend being much stronger, a pattern attributed to the laterally distant position of the east Lynmouth beach section in relation to a ridge centre, reflected in a resistance of facies 8 to become established.

Considered individually, both Lee Stone sections display an approximately equal probability for transition between flaser bedding (facies 6) and cross-bedded sandstone (facies 8) and the converse; the probabilities for the eastern Lee Stone section are, however, weaker. The parity of the facies 6/8 transitions reflects the alternation of intermittent wave/geostrophic-dominated processes (facies 6) with obliquely palaeo-offshore-flowing geostrophic currents initiating facies 8 dune migration. The variation in the magnitude of the

probabilities between the two sections is interpreted to be due to dune migration being more active in the central ridge zone (western Lee Stone section) than at the ridge flank (eastern Lee Stone section).

At east Lynmouth beach facies 8 develops directly from either CU microsequences (facies 1') or wavy bedding (facies 5). This reflects a position laterally distant to a ridge centre, which would have resulted in both limited sand supply and turbulence, and thus flaser bedding (facies 6) was only poorly developed at this locality.

The eastern Lee Stone and east Lynmouth beach sections alone exhibit a propensity for transition from facies 8 to facies 1 (accompanied by a weak reverse tendency at eastern Lee Stone). This transition type is attributed to deposition on the ridge flank, where facies 8 dune migration would have been more sporadic in contrast to the higher energy ridge crest zone.

In summary, the western Lee Stone section exhibits a coarsening-upwards sequence interpreted as a section through the core of an offshore-migrating sand ridge. Thick inter-ridge deposits of thinly interlayered sandstone/mudstone bedding, with the development of coarsening-upwards microsequences proximal to the ridge facies association, pass upwards into wavy bedding deposited adjacent to the ridge margin. The ridge margin itself comprises flaser bedding; cross-bedded sandstones becoming increasingly dominant towards the ridge core i.e. palaeo-ridge crest. A thin development of flaser bedding caps the ridge deposits prior to a return to inter-ridge deposits of thinly interlayered sandstone/mudstone bedding.

The eastern Lee Stone section exhibits a coarsening-upwards followed by fining-upwards trend, interpreted as a section through the flank of an offshore-prograding sand ridge. This section differs from the western Lee Stone section in that the ridge and adjacent deposits exhibit a symmetrical development of wavy bedding passing upwards into an interdigitating sequence of flaser bedding and cross-bedded sandstones, succeeded by a wavy bedded cap. Both the eastern Lee Stone and western Lee Stone sequences contain a single unit of cross-bedded granule conglomerate set within the lower inter-ridge mudstone-dominated deposits.

The east Lynmouth beach section also exhibits a coarsening-upwards followed by fining-upwards trend, interpreted as a section through the flank of an offshore-migrating sand ridge. The sequence is less well organised than the CUFU sequence at eastern Lee Stone, however, a phenomenon interpreted to be the result

of the east Lynmouth beach section being in a more lateral position in relation to a ridge core than the east Lee Stone section. This is reflected by the fact that facies 1 deposits frequently interdigitate with facies 8 cross-beds in the east Lynmouth beach section, whilst facies 6 flaser bedding takes the corresponding interdigitating position at eastern Lee Stone (see logs: text-figures 3.1B and 4.3).

A similar relationship of a CU sequence through the core of bar deposits and a CUFU sequence at the margin was recognised by de Raaf *et al.* (1977) within Lower Carboniferous deposits in County Cork, southern Ireland - interpreted as representing wave-generated sublittoral bars.

A further difference between the eastern Lee Stone and western Lee Stone sequences is the extent to which the ridge deposits have been bioturbated. It can be seen from text-figure 4.19 that the ridge deposits in the western Lee Stone section are more heavily bioturbated. This phenomenon is interpreted to be a function of the ridge centre sequence being deposited in a more turbulent, well aerated environment when compared to the ridge flank deposits.

The facies sequence information for the Lee Stone and east Lynmouth beach sequences are summarised in two idealised vertical sequences - a ridge centre and ridge flank sequence - see text-figure 4.21.

The log for western Lee Stone (text-figure 4.2) shows a second sandstone-body development of a CUFU (ridge flank) 'Lee Stone facies association' type (12.78m to 13.14.18m on the log) i.e. above the main development of the CU (ridge centre) 'Lee Stone facies association' developed at western Lee Stone (6.83m - 9.63m on the log). This suggests that 'Lee Stone facies association' sand ridges may have exhibited a tendency to vertically stack with a small vertical offset.

Johnson & Baldwin (1986) presented a useful scheme for discriminating between the main shallow marine siliciclastic facies (see their figure 9.33, p.259). The scheme can be used to place examples of both modern and ancient environments within a spectrum defined on the basis of inferred depositional processes (storm-dominated / tidal and current-dominated / wave-dominated) and sand/mud content (mud-dominated / mixed sand-mud / sand-dominated). The 'Lee Stone facies association' clearly falls between the 'mixed sand/mud' and 'sand-dominated' classes, whilst there is evidence for the presence of the following depositional processes (see section 4.2): semi-permanent wind-driven currents (geostrophic flow - important), tides

(minor; equivocal evidence), storms (significant in heterolithic 'inter ridge' and 'ridge flank' facies) and fair-weather waves (only significant where enhanced by unidirectional geostrophic and tidal currents i.e. the 'Lee Stone facies association' which accumulated below fair-weather wave-base).

The purpose of the following sections is to compare the 'Lee Stone facies association' with both ancient (section 4.3.1.1) and modern (section 4.3.1.2) analogues falling within a similar part of the spectrum presented in the scheme of Johnson and Baldwin, particularly in order to elucidate the relative importance of the range of depositional processes responsible for moulding the 'Lee Stone facies association' shelf sand ridge. Although there is now a substantial and growing body of published material documenting coarsening-upwards sandstone-bodies with characteristics close to those preserved within the 'Lee Stone facies association,' effort will be concentrated on a detailed discussion of a few particularly well documented examples. This discussion is followed by the presentation of a process-response depositional model for the 'Lee Stone facies association' and surrounding sequence (section 4.3.1.3). Finally, possible causes for sand-body preservation on a muddy shelf are discussed for the 'Lee Stone facies association' (section 4.3.1.4).

4.3.1.1 Ancient Analogues of the 'Lee Stone Facies Association'

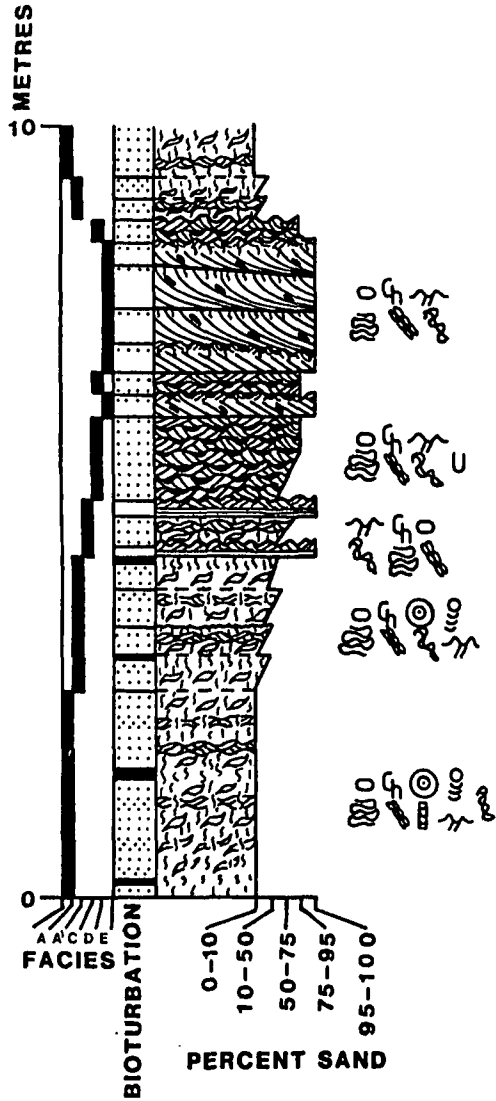
Johnson and Baldwin (1986) observed that: "Ancient linear sand ridges are recognised in two ways: (1) where an elongate sand body has a trend approximately parallel with the dominant internal palaeocurrent direction or known regional transport path . . . and (2) where large-scale, low-angle ($\leq 6^\circ$) surfaces are preserved in which the dominant internal palaeocurrent direction is parallel to, or slightly oblique to the strike of these surfaces" (p.263). Although the limited exposure of sandstone-bodies of the 'Lee Stone facies association' type precludes evaluation against the first criterion of Johnson and Baldwin, measurement of cross-bedding dip direction in relation to facies 8 'pause planes' (see section 4.2.9.1) indicates that the 'Lee Stone facies association' fits the second criterion of Johnson and Baldwin. This conclusion is consistent with the facies sequence model described for the 'Lee Stone facies association' in the preceding section.

Ancient sand ridge analogues for the 'Lee Stone facies association' fall into three broad categories within the tidal- to storm-dominated spectrum described by Johnson and Baldwin (*op. cit.*): 'tidal,' 'tidal/storm' and 'storm/current interactive' systems. Representative examples from each of these types, and comparisons drawn with the 'Lee Stone facies association', are discussed in turn below.

Text-fig. 4.21 Idealized sections through the 'Lee Stone facies association' derived from facies transition analysis.

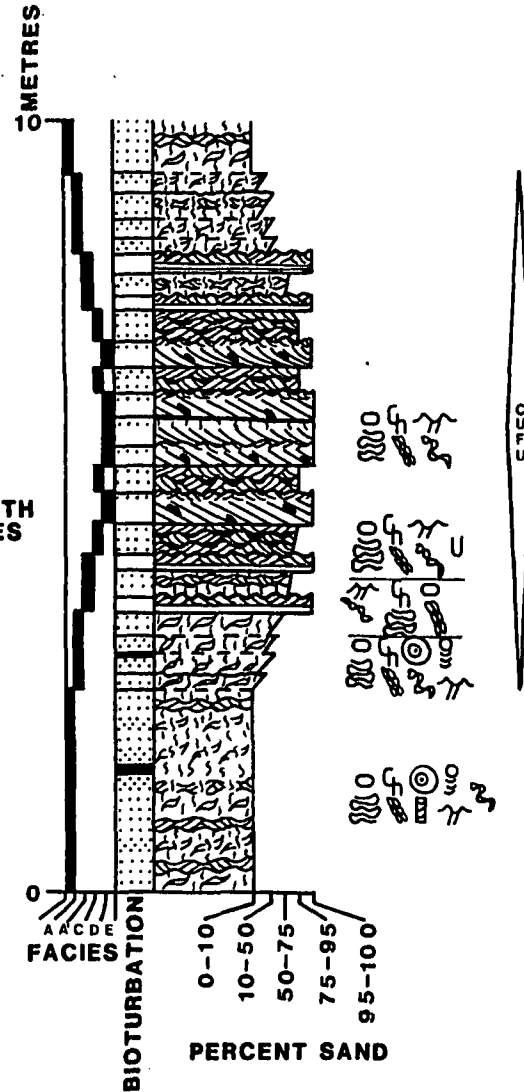
Facies: **A** = thinly interlayered sandstone/mudstone bedding; **A'** = coarsening-upwards microsequences; **C** = wavy bedding; **D** = flaser bedding; **E** = cross-bedded sandstones. See text for discussion.

RIDGE CENTRE



INTER-RIDGE DISTAL
 INTER-RIDGE PROXIMAL
 RIDGE MARGIN
 RIDGE CORE
 RIDGE MARGIN
 INTER-RIDGE PROXIMAL WITH THICK STORM SANDSTONES
 INTER-RIDGE PROXIMAL
 INTER-RIDGE DISTAL

RIDGE FLANK



INTER-RIDGE DISTAL
 INTER-RIDGE PROXIMAL
 INTER-RIDGE PROXIMAL WITH THICK STORM SANDSTONES
 RIDGE MARGIN
 RIDGE CORE
 RIDGE MARGIN
 INTER-RIDGE PROXIMAL WITH THICK STORM SANDSTONES
 INTER-RIDGE PROXIMAL
 INTER-RIDGE DISTAL

4.3.1.1.1 Ancient Tidal Sand Ridges

Tides are likely where narrow, elongate seaways are connected with major ocean basins (Bridges 1982). As discussed in section 1.6.3.2, during the Emsian the NE-SW trending Variscan geosyncline on the southern flank of the 'Old Red Continent' was connected at its eastern end with the 'Proto Tethys' ocean and the 'Uralian Seaway' to the south and south-east respectively. Thus, any tidal influence preserved in the Lynton Formation would be consistent with the palaeogeographic setting. Equivocal evidence of a tidal influence in the 'lower-middle mega-facies' was discussed in section 4.2.4.3.

Examples of tidal ridges preserved in the geological record have been given by a number of authors e.g. Anderton (1976), Nio (1976), Levell (1980b). Internal structures reported from these deposits ranged from simple avalanche foresets to complex, compound sets comprising large, low-angle surfaces which are separated by smaller-scale cross-bedding, which dipped mainly down-slope, but occasionally also dipped up-slope. Allen (1980) presented a model of tidal sand wave formation which showed that as the time-velocity pattern becomes more symmetric there is a change from avalanche foresets to increasingly complex compound sets with 'herringbone' patterns. Allen (1982b) applied this model to the Lower Greensand of southern England which was dominated by southerly flow believed to be related to strongly asymmetrical tidal currents (ebb-dominated) which were preferentially enhanced by additional unidirectional currents, either fluvial, thermohaline (oceanic) or meteorological (wind-driven) in origin.

Although there is equivocal evidence of tidal influence in the deposits encasing the 'Lee Stone facies association' the absence of features such as well developed tidal bundles (cf. Boersma & Terwindt 1981) etc. suggests that tidal currents played a relatively minor rôle in moulding the 'Lee Stone facies association' sand ridge. Furthermore, shoreface deposits preserved in deposits of a similar age in the A39 road section suggest that the basin was wave-dominated during 'lower-middle mega-facies' times e.g. there is no evidence of features such as tidal inlet migration channel lags (Elliott 1986).

4.3.1.1.2 Ancient Tidal/Storm Sand Ridges

The late Precambrian to early Cambrian seas bordering the Iapetus Ocean have provided several documented examples of tide/storm interactive sand ridges (Banks 1973a, Anderton 1976, Johnson 1977, Levell 1980b)

whilst other examples have been reported from ancient epicontinental seas (Hereford 1977, Tankard & Hobday 1977, Hobday & Tankard 1978, Cotter 1983). Internal features within these sandstone-bodies show evidence for tidal currents being enhanced by storm-generated flow e.g. Anderton (1976 - late Precambrian Jura Quartzite) reported that bedform migration appears to have reached a peak during spring tides and a minimum during neap tides, but that bedform migration was significantly enhanced by storms e.g. thick ($\leq 4.5\text{m}$) tabular sets of cross-bedding were interpreted to be the product of sand wave migration in response to wind-drift currents generated by storm-waves.

Of the above examples, the study by Johnson (1977) of the late Precambrian Upper Dakkavarre Formation, north Norway (referred to as the Skalmeset Sandstones by Hobday & Reading 1972, who first suggested that the deposit represented a westerly prograding sub-tidal sand bar) is worth describing in detail as it provides a useful analogue from which comparisons can be drawn with the 'Lee Stone facies association'. Johnson recognised 5 coarsening-upwards sequences, each between 2 and 10m thick. No bioturbation was developed in the sequence i.e. it of Precambrian age. The following facies were described (*comparisons with the 'lower-middle mega-facies' italicised*):

- Facies 1 - fine wave rippled sandstones (c. 1cm) interbedded with 2cm siltstones, both deposited from suspension. *Distal facies 1.*
- Facies 2 - moderate-energy wavy-laminated sandstones with waves more intense than facies 1 and high-energy thicker storm parallel- to wavy-laminated waning sequences; wave-ripple crestlines trended SE-NW. *Facies 5 and 6.*
- Facies 3 - 15 to 30cm sets of trough cross-bedding exhibiting a unidirectional palaeocurrent trend towards the NW and occasionally the NE. *Facies 8.*
- Facies 4 - westwards-inclined, convex-up surfaces dipping between 6° and 15° , regularly spaced and separated by trough cross-bedding (migrated towards NW, occasionally NE) identical to facies 3 but migrating sub-parallel to the strike of the inclined surfaces. The trough cross-bedding was interpreted to represent periods of relatively high-energy, whilst the inclined surfaces were interpreted as having been moulded by wave scour under shoaling wave conditions in periods of relatively low-energy when unidirectional currents were not flowing. *'Pause planes' that dissect the 'Lee Stone facies association'.*

□ Facies 5 - coarse tabular cross-bedding, some foreset toes exhibiting backflow ripples, displaying reactivation surfaces, hanging set boundaries and a downcurrent decrease in set thickness. Units were up to 4m thick and were erosively bounded, overlying an erosively channelled base with a trend that cut obliquely across the migration direction of the facies 4 inclined surfaces. Four channels, up to 4m in depth, were deeply dissected and laterally connected. The channels were interpreted as representing mutually opposing flow in tidal channel systems. *Although a possible channel does occur in the sequence at the eastern side of lee stone it has a passive infill of facies 1 thinly interbedded sandstone/mudstone. Thus there appears to be no 'lower-middle mega-facies equivalent' to Facies 5 of Johnson.*

The angular relationship between wave-ripple crestlines, cross-bedding migration directions and the inclination of the major surfaces is similar to those recorded in the 'Lee Stone facies association'. Johnson discounted the rôle of semi-permanent currents in moulding the sand-body, stating that ". . . there is no positive evidence for their presence in this instance" (p.257).

The presence of unidirectional palaeocurrent patterns in sandstone-bodies attributed to a tidal origin is common, e.g. Narayan (1971), Banks (1973a) and Anderton (1976), and has been attributed to preferential bedform migration during higher energy periods induced by wind-drift and storm-surge (Johnson & Stride 1969, Swift 1974). Also, tides flowing around sand-banks with an asymmetric cross-section results in the preservation of foresets on the steeper face rather than gentle face e.g. Well Bank in North Sea reported by Houbolt (1968). Johnson ruled out 'storm-surge ebb' currents in accounting for the observed palaeocurrent pattern on the basis that the currents would have flowed offshore whereas the Upper Quartzitic Sandstone Member preserves an alongshore palaeocurrent trend. However, subsequent literature has now shown that the effect of the Coriolis force in generating alongshore geostrophic flow is significant (see review in Swift & Rice 1984). Thus, storm-generated flows may have played a more important rôle in moulding the Upper Quartzitic Sandstone Member than Johnson recognised. Johnson concluded that tides and direct storm-generated wind-drift oscillatory currents enhancing bottom shear stress were the primary currents responsible for moulding the Upper Quartzitic Sandstone Member; a pure storm-formed ridge origin was discounted because of the lack of waning flow features in facies 3 & 4.

Johnson developed a process-response model for the Upper Quartzitic Sandstone Member based on long-term relative energy, mainly due to tides and waves. 'Phase 1' was the period of highest energy, when tidal

currents were enhanced by storm induced currents, and resulted in rapidly migrating dunes draping the bar crest and flanks and extending into troughs and offshore. Sand suspended by high energy currents and wave agitation moved offshore in seaward thinning and fining sheets generating bar face accretion. 'Phase 2' was a period of fair-weather when spring tides and waves were dominant; the resulting current patterns were more variable. The bar face was reworked by tides and waves and fines were deposited from suspension offshore (facies 1). Bars built vertically, but their height was limited by fair-weather wave-base. During 'Phase 3' fair-weather neap tides and wave were dominant but fluctuated in intensity. Dunes at the bar crest were probably inactive during this phase, whilst bar flanks modified by waves (facies 2) and fines were winnowed from the bar crest and deposited in the inter-bar zone. Thus, 'Phase 3' was primarily a period of bar degradation.

Whilst the Upper Quartzitic Sandstone Member shares many features in common with the 'Lee Stone facies association' there are several significant differences which suggest that the rôle of tidal channels was less significant in moulding the 'Lee Stone facies association' relative to the Upper Quartzitic Sandstone Member:

- ❑ The most persuasive argument in favour of tidal currents for the Upper Quartzitic Sandstone Member is the mutually exclusive bi-polar palaeocurrents in the facies 5 channels. No analogues were observed in the Lynton Formation.
- ❑ The consistent alignment of *Palaeophycus tubularis* burrows facing obliquely *onshore* (see appendix B) throughout the 'lower-middle mega-facies' is interpreted to be a response to an offshore-flowing semi-permanent current. If the current responsible for the burrow alignment had been tidal in origin, burrows facing obliquely *offshore* would be expected occasionally, representing an area of opposed flow resulting from ebb-flood tide separation paths.

In summary, the 'Lee Stone facies association' exhibits many features in common with tide/storm generated sand ridges but evidence for tidal currents playing a significant rôle in relation to obliquely-offshore directed semi-permanent currents is lacking.

4.3.1.1.3 Ancient Storm/Current Interactive Sand Ridges

The Western Interior Seaway of North America was a winter-storm-dominated seaway with predominantly shoreline-parallel currents. During the Upper Cretaceous the seaway was mainly muddy but did contain a number of widely distributed lenticular, elongate offshore sand bars which formed at distances of up to 100km from the palaeo-shoreline. Each sandstone-body is 5 to 20m thick and up to 160km long with long axes parallel to the palaeocurrent direction. The sandstone-bodies characteristically coarsen-upwards from bioturbated mudstones, through ripple-laminated fine sandstones into coarser cross-bedded sandstones. The sandstone-bodies contain significant petroleum reserves which have been extracted since the end of the last century and there is therefore an extensive published record integrating surface outcrop observations with sub-surface seismic profiles and well log data.

The Shannon and Sussex Sandstones (Berg 1975, Spearing 1976, Brenner 1978, Boyles & Scott 1982, Hobson *et al.* 1982, Shurr 1984, Tillman & Martinson 1984) comprises 5 to 65cm thick sets of planar and trough cross-bedded sandstone, common containing mudstone drapes and mud rip-up intraclasts. The sandstones represent the southerly progradation of elongate sand sheets of very low relief with surfaces dipping $<0.5^\circ$ (Seeling 1978). Thin-bedded wave-rippled sandstones found in the lower part of the sandstone lentils are typical of storm sandstones (Spearing 1976) and there was an orthogonal relationship between wave-ripple crests and the dominant palaeocurrent mode (Spearing 1976), supporting the inference that storms and/or storm-enhanced fair-weather basinal currents (tidal or oceanic) dominated sedimentation patterns. Boyles & Scott (1982) described abundant landward-dipping (to the NW), low-angle accretion surfaces in the central bar facies with associated westerly migrating cross-bedding. The sandstone-body was interpreted as having developed in response to southerly-flowing fair-weather oceanic currents cf. (SE African shelf model) combined with westerly-directed storm-generated currents. La Fon (1981 p.720) preferred an interpretation involving "... storms possibly augmented by weak tidal currents".

Taking one of the studies in detail, Shurr (1984) documented the Shannon Sandstone from the Upper Cretaceous of Montana where he recognised a hierarchy of sandstone-bodies: lithosome > sheet > lentil > elongate lens > small-scale facies packages. Lentils comprised 12-18m thick elongate lenses of 50km² lateral extent. In turn, each elongate lens comprised a series of coarsening-upwards cycle which individually ranged from 3 to 6m in thickness and 0.12km² in lateral extent; the small-scale facies were arranged in an imbricate pattern. The lithosome and sand sheets were attributed to regional-scale long-term tectonic events, whilst the

lentic - elongate lens - small-scale facies package hierarchy was interpreted as analogous to morphologic elements recognised on modern shelves i.e. complex sand ridge fields - individual sand ridges - sand waves, respectively; n.b. the term 'sand waves' was used in reference to units of dunes/megaripples with a relief of up to 4m described from modern shelves, not sand waves as described from flumes.

Perhaps the most detailed study of 'Western Interior Seaway' coarsening-upwards sandstone-bodies was that of Tillman & Martinsen (1984) who documented two vertically stacked shelf-ridge complexes in the Shannon Sandstone Member of the Cody Shale, Powder River Basin, Wyoming. These sandstone-bodies were deposited 70 miles from the contemporary shoreline at inner to mid shelf depths (below fair-weather wave-base in 20 to 100m of water) where they were moulded by S- to SW-flowing shore-parallel currents periodically intensified by storms. The ridges had a N-S trend i.e. slightly oblique to palaeocurrent direction. The following facies were described (*comparisons with the 'lower-middle mega-facies' italicised*):

- ❑ Central Bar Facies - clean trough (80%), planar-tangential (20%) and planar (trace) cross-bedded sandstones which commonly had a truncated horizontal to sub-horizontal upper surface; sets frequently thinned in a down-palaeocurrent direction and shale rip-up clasts were frequent, although burrowing was rare. Palaeocurrents were occasionally bi-directional towards the very top of the facies. The cosets between erosional boundaries occasionally thinned in a direction oblique to the mean palaeocurrent direction. *Facies 8 cross-beds.*
- ❑ Central Bar (Planar Laminated) Facies - planar-laminated beds of sub-horizontal laminae with occasional burrowed zones; this facies was rare. *Parallel-laminated zones within facies 8.*
- ❑ Bar Margin Facies (Type 1) - cross-bedded sandstone which contained a higher proportion of mud rip-up clasts than the Central Bar Facies cross-beds; the facies also contained interbedded shale. Beds were on average thinner than the Central Bar Facies. Burrowing was generally rare, but became common locally. *Facies 8 in the CUFU ridge margin sequence logged at the eastern end of Lee Stone. The facies 8 cross-bedded units contained a higher proportions of mudstone rip-up clasts, and thin heterolithic beds separating the cross-bedded units, at eastern Lee Stone when compared with the equivalent cross-bedded units in the ridge core sequence logged at the western end of Lee Stone.*

- ❑ Bar Margin Facies (Type 2) - cross-bedded to rippled sandstone in which burrowing was rare. *Facies 6 (flaser bedding) interbedded with facies 8.*

- ❑ Interbar Facies - Thinly interbedded sandstone and shale in which the sandstones comprised mainly current ripples; wave-ripples were less frequent. Burrowing was moderate, locally high. *Facies 1 and facies 5 (wavy bedding).*

- ❑ Interbar Sandstone Facies - Rippled sandstones comprising mainly current ripples; wave-ripples were less frequent. Burrowing was rare to moderate. This facies was interpreted as having been deposited lateral to the higher energy portions of the ridges as well as at the bases of the ridges during their initial development. *No equivalent facies developed in the 'lower-middle mega-facies' i.e. the Lynton Formation shelf appears to have been muddier than the sequences described by Tillman and Martinsen.*

- ❑ Shelf Sandstone Facies - sub-horizontally-laminated sandstone containing rare trough cross-bedded sets; bioturbation was rare. *No equivalent facies developed in the 'lower-middle mega-facies' i.e. the Lynton Formation shelf appears to have been muddier than the sequences described by Tillman and Martinsen.*

- ❑ Bioturbated Shelf Sandstone Facies - 75 to 95% burrowed (compared with 5 to 27% in other Shannon Sandstone facies); physical structures were restricted to rare traces of cross-lamination and parallel-lamination. This facies developed at the base of the ridge complex and between the two stacked ridge complexes and represents periods of slow deposition. *Cf. Facies 1, although the 'lower-middle mega-facies' does not appear to have had as high a sand-grade input as the Shannon Sandstone shelf.*

- ❑ Shelf Siltstone Facies - sub-horizontally laminated siltstones with a moderate degree of burrowing. *No equivalent facies developed in the 'lower-middle mega-facies' i.e. the Lynton Formation shelf does not appear to have had as high a silt-grade input as the Shannon Sandstone shelf.*

- ❑ Bioturbated Shelf Siltstone Facies - 75 to 95% (5 to 27% in other facies) burrowed with rare low-angle bedding and small ripples. *Facies 1.*

- ❑ Shelf Silty Shale Facies - current ripples and sub-horizontal silty laminae which were lightly to moderately burrowed. *Facies 1, although the 'lower-middle mega-facies' examples appear to reflect an environment in which oscillatory currents were more dominant than their Shannon Sandstone counterparts.*

N.B. Tillman and Martinsen reported a 5m wide erosion channel which cut through the rippled interbar sandstone facies and was filled concordantly with rippled sandstone of the same facies (figure 24C of Tillman and Martinsen 1984). This channel and its fill is analogous to the putative channel at the eastern side of Lee Stone that cut through facies 3 and was filled with facies 1 inter-ridge heterolithic deposits.

Tillman & Martinsen observed that facies association sequences varied between locations e.g. the Interbar Sandstone Facies was overlain by the Central Bar Facies in some areas, whereas elsewhere the Interbar Facies passed upwards into the Bar Margin Facies, and was in turn overlain by the Central Bar Facies. Furthermore, in some cases the sandstone was only locally developed, suggesting that the ridge spacing and thickness was smaller than elsewhere; more usually the sandstone formed an extensive sheet comprising imbricate lentils. The ridge system was extensive: subsurface geophysical studies revealed that the Lower Shannon Sandstone represented the migration of an elongate ridge 18 by 17 miles in lateral extent and up to 75 feet thick. The major ridge comprised several coalesced smaller ridges and was multi-crested; the smaller ridges trended N-S, a direction oblique to the SW-directed palaeocurrents (the distant shoreline trended NE-SW). Wave ripple crest orientations were variable, trending between ENE-WSW to N-S, many interference wave ripples (e.g. Fig 23D). The obliquity of the ridge in relation to palaeocurrent was regarded by Tillman and Martinsen to be analogous to the storm-moulded sand ridges oriented at c. 30° to the current on the Atlantic shelf of the U.S.A. (as described in Stubblefield *et al.* 1984). The ridges appeared to stack obliquely upwards in a direction parallel to the palaeoflow. In contrast, the Upper Shannon Sandstone had an oblate geometry, compared with the strongly linear developments of the Lower Shannon Sandstone. Tillman and Martinson attributed this difference to a syn-sedimentary, actively growing, palaeo-high which localised sand deposition in the Upper Shannon Sandstone.

The Shannon Sandstone was deposited during a time of falling sea-level or possibly during a minor stillstand; the end of the 'Shannon regression' was marked by a phase of rapid transgression. The sand was probably supplied from a large contemporaneous deltaic system and the sand ridges built upwards (the ridges grew vertically rather than laterally) from a gently sloping mud surface below. Prior to, and possibly concurrent with, ridge formation, less competent currents flowed between the bars to give the rippled deposits; short- and long-term fluctuations in the dominantly unidirectional currents gave rise to the interfingering of facies. Shale deposition and bioturbation occurred during low energy periods, whilst the

introduction of strong currents after relatively long periods of quiescence resulted in the introduction of abundant clay rip-up clasts.

In comparison with the Shannon Sandstone, the 'Lee Stone facies association' was deposited on a shelf where the silt- and sand-grade input was lower. Thus, the 'Lee Stone facies association' did not form into a laterally extensive sand sheet, but was restricted to a series of laterally discontinuous lenses several hundred metres in width. The Shannon Sandstone was extensively burrowed by a deposit-feeding infauna that left a mottled / 'churned' texture, whereas the 'lower-middle mega-facies' contained a more established / permanent infauna that contained both suspension- and deposit-feeders. This suggests that the 'Lee Stone facies association' ridge may have had a lower sediment supply (less suspended sediment) and/or was subject to periods of non-migration (allowing a suspension-feeding infauna to establish) as compared to the Shannon Sandstone. Finally, oscillatory currents were more pronounced on the 'lower-middle mega-facies' shelf than the Shannon Sandstone shelf.

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In summary, the 'Lee Stone facies association' contains many features in common with sand ridge sequences reported from elsewhere in the geological record. In detail, evidence preserved within the 'Lee Stone facies association' more closely matches features reported from ancient sand ridges interpreted as having had a storm/current interactive origin, although there is some equivocal evidence to suggest that tidal currents may have enhanced fair-weather and storm currents. The 'Lee Stone facies association' lacks the well-organised tidal bundles etc., that characterise ancient sand ridges moulded predominantly by tidal currents.

4.3.1.2 Modern Analogues of the 'Lee Stone Facies Association'

Modern sand ridges are predominantly tidal (sub-section 4.3.1.2.1) or storm-maintained (sub-section 4.3.1.2.2) - a comparison of the characteristics of tidal and non-tidal ridges is discussed in sub-section 4.3.1.2.4, along with the significance of these differing characteristics in relation to the interpretation of the 'Lee Stone facies association'.

4.3.1.2.1 Modern Tidal Sand Ridges

The southern North Sea is probably the most intensively studied area of modern tidal sand ridge formation e.g. Off (1963), Houbolt (1968), and Caston (1972). The region is swept by diurnal and semi-diurnal currents which flow parallel to sand ridge crests; ebb and flood currents frequently separate along well defined paths (Caston & Stride 1970). The sand ridges migrate in the direction facing the steeper slope, which dips between 4° and 7° , resulting in large internal bedding planes (Houbolt 1968). In contrast with storm-formed ridges, numerous dunes and sand-waves develop; these features are most common on the gentler faces (where there is a lower potential for preservation).

Tidal sand ridges (tidal sand banks) are linear bedforms whose long axes are oriented as much as 20° obliquely to the direction of the strongest tidal current (Kenyon *et al.* 1981). Transport processes on the faces of these ridges are dominated by either ebb or flood currents; tidal inequality, therefore, leads to asymmetric cross-sections resulting in major ($\cong 3^\circ$ to 7°) internal bedding planes separated by small-scale cross-stratification.

4.3.1.2.2 Modern Storm-maintained Sand Ridges

The intensively studied NW Atlantic shelf off the east coast of the U.S.A. has provided a wide range of studies detailing the mechanics of storm-maintained sand ridge generation e.g. Swift *et al.* (1972, 1973) and Stubblefield *et al.* (1975). During fair-weather periods the shelf is dominated by oscillatory and wave-surge currents, with wave-ripples oriented parallel to the long-axes of ridges. Waves maintain the ridges in a configuration that is sub-parallel to the shoreline (Duane *et al.* 1972) but the long-term effect of waves is mainly destructional, through winnowing at the bar crest. Large, low-angle surfaces have been reported from seismic profiles of the Bethel Shoal, Florida from where the configuration of the shoals runs normal to the seaward-dip of the low-angle progradation surfaces (Duane *et al.* 1972). The storm-current path is sub-parallel to the ridge crests (Swift *et al.* 1973), but importantly there are smaller-scale flow-transverse bedforms with lee slopes dipping in the direction of the storm-generated currents (Swift 1972, Swift *et al.* 1972, 1973). The storm-currents are responsible for the constructional phases of ridges. Cross-ridge channels have not been reported from Atlantic shelf (Duane *et al.* 1972).

Swift & Field (1981) documented a storm-maintained sand ridge field from the Maryland Sector of the NW Atlantic shelf which provided a useful overview of the features that characterise storm-maintained ridges. The Maryland Sector ridges are spaced between 4.5 and 6.5km apart and make a 10 to 35° opening angle with shoreline. Individual ridges have a relief of between 3 and 12m and side slopes are generally less than one degree, although offshore ridges can reach 7° locally. A systematic change in morphology takes place from shoreface ridges to nearshore ridges, thence to offshore ridges (in >16m water). The slope angle was observed to decrease as the asymmetry increased, i.e. the seaward flanks of offshore ridges are up to five times as steep as landward flanks, and the cross-sectional area is increased. The grain size is 90° out of phase i.e. the grain-size is coarsest on landward flank of ridges, not on the ridge crest - as the landward flanks erode, the seaward flanks aggrade. The ridges are maintained by a storm-related geostrophic flow régime which peaks during 'north-easter' storms; nett transport is towards the SE i.e. slightly seaward of the ridge crest. Sand waves were only recorded from inter-ridge swales where they had a \approx 100m spacing and faced southwards. This feature was attributed to be due to waves suppressing sand wave formation on ridge crests. The intermittent nature of the storm-generated currents is reflected in the coexistence of megaripples on ridges and sand waves in swales on the one hand and mud lenses occurring in troughs elsewhere. Mud accumulates during the quieter summer months, but during winter storms the ridges are reactivated, storm-generated currents building megaripples and sand waves and scouring-out mud patches. Interference ripples are common on ridges and appear to be the result of current ripples superimposed on wave ripples (e.g. figure 10C of Swift & Field).

The association of sand waves, migrating under wind-drift generated currents, with storm-maintained sand ridges has been recorded by Boggs (1974), Hunt *et al.* (1978), Swift *et al.* (1978), Stride & Chesterman (1973).

4.3.1.2.3 A Comparison of Modern Tidal and Non-tidal Sand ridges: Applicability to the 'Lee Stone Facies Association'

Swift (1975) observed that there is a strong structural similarity between storm-generated and tidal ridges. Nevertheless, Belderson (1986) published a set of diagnostic criteria to distinguish between tidal and storm-maintained ridges - these are summarised in table 4.5.

Belderson (*op. cit.*) reported that tidal sand banks form where there are relatively strong tidal currents (i.e. 90cm s^{-1} near-surface, 55cm s^{-1} at 30m depth) and there is an abundant sand supply, whilst tidal sand sheets tend to form where currents are weaker. Belderson disputed the continuum from tidal sand banks to storm-generated sand ridges proposed by Swift (1975) on the basis that in areas of intermediate tidal current strength there is a zone occupied by fields of sand waves (i.e. the 'sand sheet zone').

Permanent or semi-permanent unidirectional currents can override peak ebb-flood tidal asymmetry, even where the unidirectional current is weaker Belderson (1986). This can result in sand ridges/banks forming outside their normal tidal limits e.g. off the east Scottish coast due to the Fair Isle Current ($\cong 50\text{cm s}^{-1}$ - Belderson 1986). This is particularly significant off NW Denmark where tidal currents only reach 25cm s^{-1} but storm flow can enhance currents to 200cm s^{-1} (Stride & Chesterman 1973). Similar storm-flow enhancement of weak tidal currents has been reported from sand ridges forming under the geostrophic Agulhas Current off SE Africa (Martin & Flemming 1986).

| | Tidal Sand Banks | Storm-maintained sand ridges |
|----------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------|
| <i>Height</i> | 20-30m high | 3-12m (7m average) |
| <i>Length</i> | $\leq 70\text{km}$ | $\leq 20\text{km}$ |
| <i>Shape</i> | Offshore they are linear, inshore (where abundant sand supply) where they are parabolic. | Elongated |
| <i>Spacing</i> | 2-30km | 0.5 - 7km |
| <i>Crests</i> | sharp (except where near surface planed by waves) | smooth |
| <i>Angle to flow</i> | $0-20^\circ$, but generally $7 - 15^\circ$ | more oblique than tidal $\pm 28^\circ$ is in better accordance with Huthnance theory |
| <i>Angle with coast</i> | Related to peak tidal direction | $35-40^\circ$, but can be up to 60° |
| <i>Slope angles</i> | $\leq 6^\circ$ | 1° or less, with steepest slope of 2° (rarely up to 7°), even shallower slope for offshore ridges which are c. 0.5° |
| <i>Internal structures</i> | Internally cross-stratified units occur between $\cong 5^\circ$ master-bedding surfaces. But can contain larger angle-of-repose cross-sets created by superimposed sand waves | Absence of superimposed sand waves results in pervasive small-scale cross-beds bounded by master-bedding surfaces dipping at c. 5° |
| <i>Sand waves</i> | Abundant and semi-permanent | generally rare and inter-ridge |

Table 4.5 A comparison between the characteristics of tidal and storm-maintained sand ridges

Based on data presented in Belderson (1986)

In comparing the 'Lee Stone facies association' with modern tidal and storm-maintained sand ridges a storm-maintained origin appears to be more consistent with the published characteristics summarised in table 4.5:

- Height \cong 4m
- Length - unknown, but probably < 1km
- Shape - unknown
- Spacing - in the order of several hundred metres
- Crests - not preserved
- Angle to flow \cong 45° i.e. facies 8 cross-bedding indicates a mean SSE palaeocurrent, the 'Lee Stone facies association' 'pause planes' indicate that the sand ridge prograded towards the SSW.
- Angle with coast - ?parallel i.e. the 'Lee Stone facies association' 'pause planes' indicate that the sand ridge prograded towards the SSW (down-palaeoslope)
- Slope angles - 5 to 8° - this angle is more consistent with tidal sand ridges, although Swift & Field have reported offshore storm-maintained sand ridges from the Maryland Sector of the NW Atlantic shelf with slopes of up to 7°.
- Internal structures - no evidence of large-scale angle-of-repose cross-beds **within** the 'Lee Stone facies association'
- Sand waves - generally rare and inter-ridge

4.3.1.3 Depositional Model for the 'Lee Stone Facies Association' and Surrounding Sequence

The purpose of this section is to draw together the combined lines of evidence of processes preserved within the 'Lee Stone facies association' and the surrounding sequence, and their comparison with ancient and modern analogues presented in the previous two sections, in order to build a process-response model to account for the preserved facies association.

In section 4.3.1.1 a representative sample of coarsening-upwards shallow marine sandstone-bodies closely analogous to the 'Lee Stone facies association' was described from a variety of geological settings. It is clear that this type of coarsening-upwards sequence is a pervasive phenomenon in Jurassic to Cretaceous siliciclastic sequences deposited in the epicontinental 'Western Interior Seaway' of North America and they provide a good analogue for the 'Lee Stone facies association'. Swift and Rice (1984) described a set of models to account for the generation of the 'Western Interior Seaway' coarsening-upwards sandstone-bodies, based on an understanding of fluid and sediment dynamics gained from the study of modern shelves, particularly the present day eastern seaboard of the U.S.A. Swift and Rice augmented their model with circumstantial evidence contained in the stratigraphic and tectonic setting of the 'Western Interior Seaway'.

The core of Swift and Rice's (1984) model emerged from the application of theoretical studies by Smith (1970) and Huthnance (1982) which used numerical flow-modelling to understand bedform stability for a range of typical shelf conditions. Figueredo *et al.* (1981) demonstrated that these models could be used to understand the generation of modern sand ridges forming on the present-day middle Atlantic continental shelf of North America. Swift and Rice applied these models to the ancient coarsening-upwards sandstone-bodies preserved in muddy shelf deposits of the Cretaceous 'Western Interior Seaway'. Because the resulting process-response model generated by Swift and Rice was firmly based in a consideration of fundamental shelf fluid and sediment dynamics it has a wide applicability to the understanding of both modern and ancient shelf sandstone-bodies. The model of Swift and Rice will, therefore, be used as the basis for interpreting the conditions that gave rise to the 'Lee Stone facies association' sandstone-bodies and their surrounding deposits, with only minor modifications being made to take into account the differing settings of the 'Western Interior Seaway' and the 'Exmoor Basin' e.g. palaeo-latitude and palaeo-climate. The ensuing paragraphs provide a précis of the model of Swift and Rice and their counterparts interpreted for the 'Lee Stone facies association' and surrounding sequence, followed by the construction of a process-response model for the 'Lee Stone facies association' and surrounding sequence.

Swift and Rice used the storm-dominated mesotidal middle Atlantic continental shelf of North America as an analogue for the storm-dominated micro- to meso-tidal Western Interior Cretaceous Seaway. Tidal currents are insufficient to drive observed sediment transport on the middle Atlantic shelf and sediment transport is dominated by geostrophic flow generated during storms. The key features of geostrophic flow relating to

sediment transport on the shelf are described below (see section 4.2.2.3 for a discussion of geostrophic flow on the Lynton Formation shelf):

- (i) Where wind stress moves surface water landwards, water is piled against the shoreline ('coastal set-up') and a pressure gradient is established; a compensatory bottom return current will flow seawards for the duration of the storm event.
- (ii) The bottom flow is deflected by the Coriolis force (deflection to the right in the Northern Hemisphere, to the left in the Southern Hemisphere). When the Coriolis term balances with the pressure term in the equation of motion a shore-parallel bottom flow will result.
- (iii) Mid-latitude storms (generally of 2 to 4 days duration) more effectively 'couple' with the sea surface than hurricane events (which are more transient) and are thus more effective in generating geostrophic flow; the return flow generated by water piled against the shoreline falling in level as the storm relaxes is insignificant. N.B. Although not discussed by Swift and Rice, the semi-permanent nature of trade winds would be particularly effective in generating geostrophic flow.
- (iv) Storm waves accompanying the geostrophic flow will superimpose a high frequency oscillatory component on the unidirectional geostrophic component at the sea-floor. The resultant shear stress at the bed is greater than the sum of the two components considered individually (Grant & Madsen 1979, Hammond & Collins 1979) - geostrophic flows are, therefore, especially efficient in entraining bottom sediment.

Swift and Rice discussed the application of the Rouse equation (governing expected sediment concentration profiles above non-consolidated beds) when parameterised for typical shelf hydraulic and sediment dynamic régimes. They concluded that as storm flow accelerates the bottom will be eroded, but at peak flow the sediment load in the bottom boundary layer must rapidly adjust to the available flow power. For a constant velocity, and equilibrium sediment load, further erosion cannot take place as maximum capacity has already been attained i.e. the substrate is effectively 'armoured'. As the storm abates and flow decelerates, a graded bed will be deposited. Of particular significance is the conclusion that each high-energy event can only erode a few centimetres down into the bed before capacity is reached. On a relatively flat portion of shelf (i.e. in an area not subject to spatial acceleration over topographic highs) the product of this cyclic acceleration and

deceleration of storm flows will be thin-bedded heterolithic deposits (see text-figure 4.22A - termed 'temporal acceleration model' by Swift & Rice). Very broad, gentle topographic and peak flow gradients will exist, even on a relatively flat muddy shelf, which will result in localised trends towards long-term deposition.

Instabilities arising over an initially plane bed, or due to slight initial irregularities in the bed, will result in a regularly spaced pattern of velocity variation superimposed on the mean flow component. Numerical stability analysis was used to explore whether these velocity variations will be amplified or damped for a given set of parameters. Where the velocity variation is amplified, the analysis sought the limiting condition of amplification set by other variables at which a new state of quasi-equilibrium was attained. In summary, water flowing over a sand bed with an undulating surface will be out of phase with the topography, the greatest values of shear occurring on the up-current surface as flow accelerates and converges with the bottom (see text-figure 4.22B); bottom shear stress and sediment discharge reach a maximum at this point, then decrease from that point over the crest resulting in an ordered array of regularly spaced flow-transverse bedforms. However, although this analysis holds true for bedforms of ripple, megaripple and sand wave scale, Huthnance (1982) showed that sand ridges, i.e. bedforms with spacings of up to several kilometres, tend to be aligned at small angles to flow irrespective of whether they are storm-built or tide-built. This was attributed to the phase lag for bedforms of sand ridge scale being insufficient to cause crestral aggradation i.e. flow accelerating over the ridge crest will be too rapid to allow deposition to occur. For sand ridges, flow was considered as having a cross-ridge component which accelerates to clear the topographic high, and an along-ridge component that loses energy due to frictional drag with the bottom and results in deposition (see text-figure 4.22C). Three significant conclusions can be drawn from this analysis:

- (i) A sand bed is inherently unstable in response to flow in excess of threshold velocity - any initial bottom irregularity will be amplified.
- (ii) Sand ridges will extend in a direction slightly oblique to the cross-flow direction and will have a spacing determined by the separation and attachment of turbulent wakes; given a few millennia, the sand ridges evolve into an orderly array of regularly spaced bedforms. Palaeocurrents will record a mean trend that will be slightly oblique when compared to the direction of sand ridge aggradation.

Text-fig. 4.22 Key elements of Swift & Rice (1984) model for sand-bodies preserved on muddy shelves

A. Model for the production of 'normal' shelf facies by flows whose velocities vary in time rather than in space. S, Si, Cl refer to sand, silt and clay. (From Swift & Rice 1984 - figure 10).

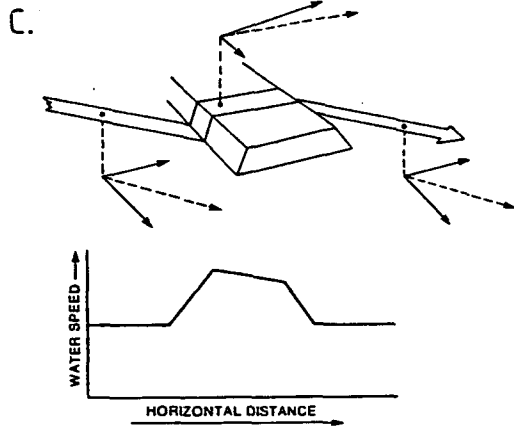
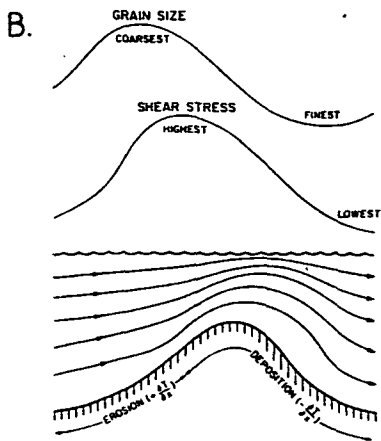
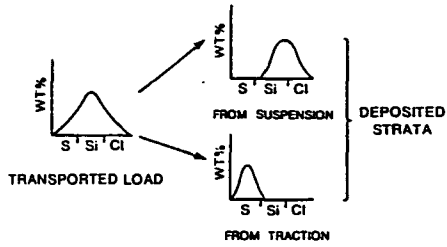
B. Model for the growth of a bedform from a slight initial topographic high, due to phase lag between topography and bottom shear stress. (from Swift & Field 1981, based on Smith 1970 : reproduced from Swift & Rice 1984 - figure 11).

C. Model for the growth of a large-scale (2km spacing) sand ridge, based on the analysis of Huthnance (1982). (From Swift & Rice 1984 - figure 12).

D. Process model for sand-body formation on a muddy shelf due to the occurrence of spatial as well as temporal velocity variations. S, Si, Cl refer to sand, silt and clay. (From Swift & Rice 1984 - figure 13).

E. Model depicting the generation of an upward-coarsening sequence. Upbuilding of shelf surface leads to greater fluid power expenditure, and a shift of the partition between deposited fraction and bypassed fraction towards the coarse end. (From Swift & Rice 1984 - figure 16).

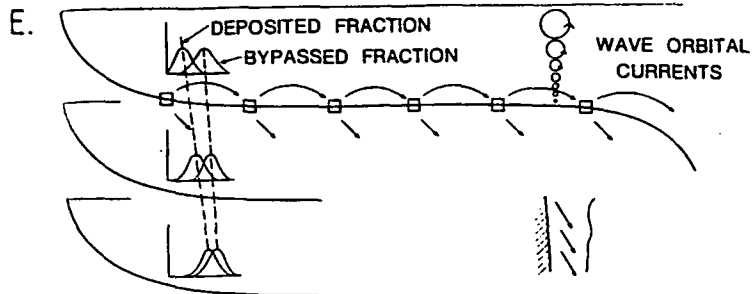
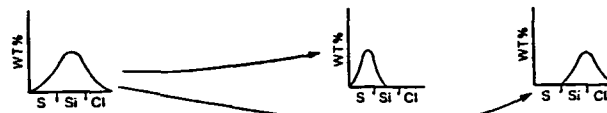
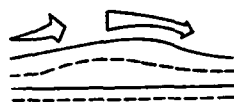
A. NORMAL SHELF SEDIMENTATION (TEMPORAL ACCELERATION MODEL)



D. SAND BODY FORMATION (SPATIAL ACCELERATION MODEL)

BOTTOM CURRENT ACCELERATES UP FORWARD SLOPE OF HIGH. DECELERATES OVER CREST

SAND LENS DEPOSITED OVER HIGH: FINES DEPOSITED DOWN CURRENT



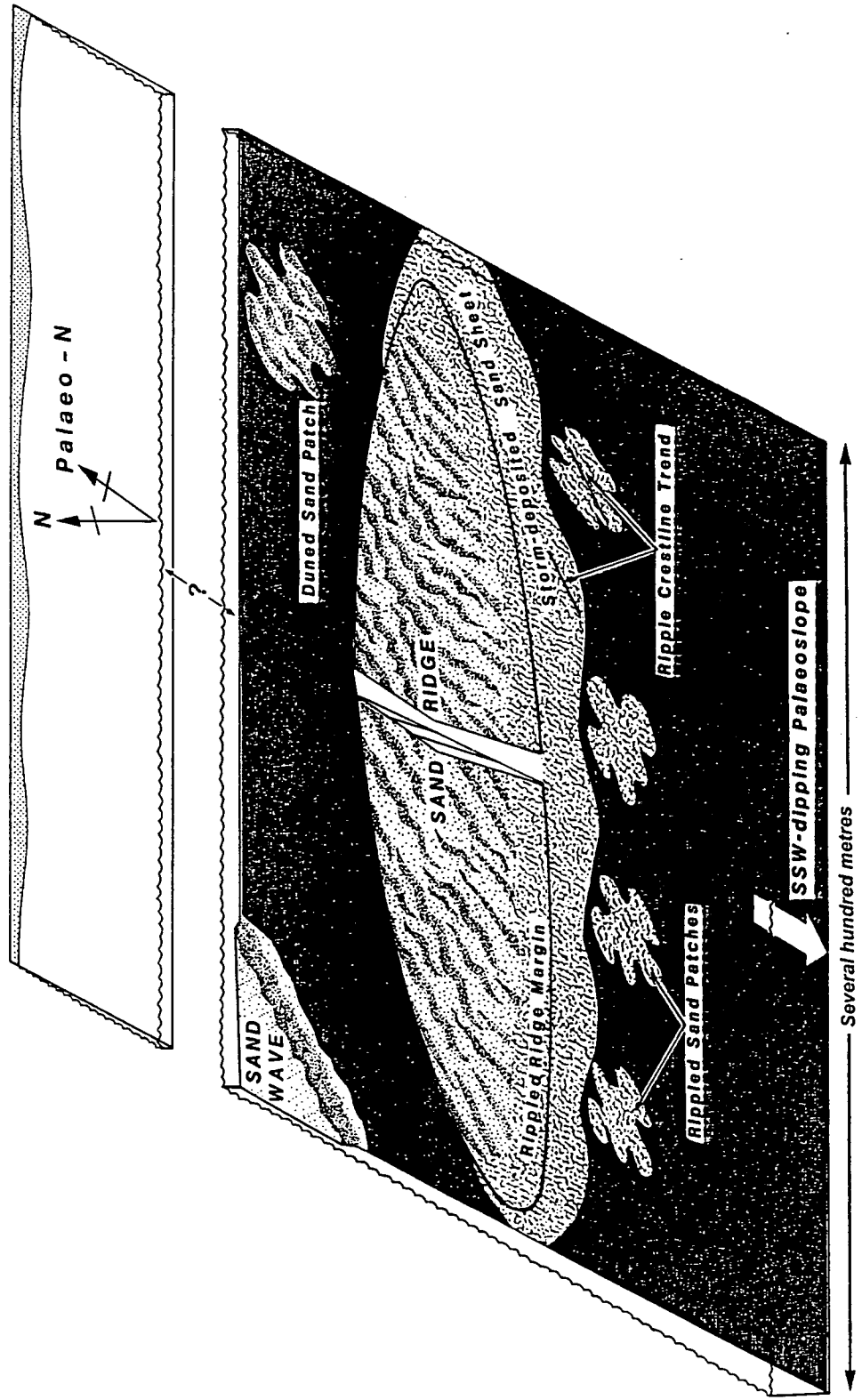
RELATION OF STORM CURRENTS TO SHORELINE AND SHELF EDGE

(iii) Deposition on topographic highs can occur throughout a period of increased environmental energy in response to spatial deceleration. Furthermore, spatial deceleration will result in the fractionation of the deposited load, which will become finer-grained downstream i.e. the topographic highs will be topped with fine sand whilst muds will be deposited in the down-stream shadow of the topographic highs (see text-figure 4.22D - termed 'spatial acceleration model' by Swift and Rice). Most importantly, particles are not deposited because velocity drops below a certain critical velocity, but because the flow power available to suspend sediment *per unit area of bottom* drops.

It is now possible to integrate the various processes described in the chapter so far into an overall process-response model for the 'Lee Stone facies association' and enclosing deposits. The model is based upon both spatial and temporal variations in the relative energy of the geostrophic, wave and tidal currents that have been invoked as the prime factors responsible for moulding the 'Lee Stone facies association' sand-body - see text-figure 4.23. The model incorporates processes discussed in section 4.3.1.2 from modern shallow marine sand ridges.

The 'lower-middle mega-facies' shelf had a mixed clastic input that was dominantly argillaceous. Fair-weather biogenic processes served to mix the substrate, particularly during the summer when semi-permanent currents that were active in the winter trade wind season would have been less active or inactive. During the winter trade wind season wave energy was both stronger and considerably more sustained giving rise to a geostrophic current that flowed obliquely offshore at depth; this unidirectional current may have been superimposed on a weak tidal régime.

Occasional summer storm events would have eroded and entrained the top few centimetres of inter-ridge silty & sandy mud deposits. The depth of erosion would be limited by the equilibrium sediment capacity of the bottom boundary layer at peak flow. In the distal inter-ridge zone sand would have mainly been deposited from suspension below (graded rhythmites) or near (horizontally-laminated sand streaks) storm wave-base. In the proximal inter-ridge zone silt- and sand-grade material would have been transported as bedload. Where the silt/sand supply was starved, isolated ripples would have migrated across the substrate (unconnected lenticular bedding); where the silt/sand supply was adequate, rippled sand patches would have developed (connected lenticular bedding). Areas of net erosion resulted in laterally extensive planar erosion surfaces being preserved in the inter-ridge muddy heterolithic facies.



Text-fig. 4.23 Model for the 'Lee Stone facies association' and surrounding deposits

During the winter trade wind season sand patches created by fractionation of mixed silty- and sandy-muds during summer storms would have been mobilised by the semi-permanent geostrophic current and stronger waves (?and tides); nett migration was towards the SSE (obliquely-offshore). In areas proximal to sand supply (sand ridges and sand waves) increasing proximity with time, relative to a fixed point, to an obliquely-offshore aggrading sand-body would have resulted in the development of coarsening-upwards microsequences in inter-ridge heterolithic deposits. The establishment of helical flow cells during the trade wind season in inter-ridge areas led to the scouring of furrows. There is no evidence to indicate that tidal flow either created, or became preferentially directed along, these furrows.

The influx of a large volume of sand onto the 'lower-middle mega-facies' shelf 'primed the pump' for the development of a sand ridge field (see section 4.4.2). Acceleration of geostrophic flow over initial irregularities on the shelf would have been amplified, causing the sand to be swept into discrete ridges. The ridges would have either been located over topographic highs or became spaced (c. several hundreds of metres), over a period of millennia, according to Huthnance (1982) bedform stability theory. In the case of the 'Lee Stone facies association' at Lee Stone the sandstone-body stratigraphically overlies a plano-convex lens of the 'Watersmeet lithotype' (vertically separated by several metres of muddy heterolithic deposits). It appears, therefore, that the ridge formed over a small muddy topographic high that caused flow to accelerate locally. Furthermore, evidence at the western end of Lee Stone suggests that another ridge complex was vertically stacked above the logged 'Lee Stone facies association'.

In detail, the sand ridges were dominated by three-dimensional dunes (occasionally two-dimensional in the slightly lower velocity currents that would have been present around the ridge margins) obliquely down-climbing (towards the SSE) major surfaces that reflected the SSW-directed (down-palaeoslope) aggradation of the sand ridge. The ridges long axes were approximately parallel to the strike of the palaeoslope. The dunes migrated under the influence of geostrophic flow during the trade wind season, but migration was intermittent i.e. as evidenced by mud drapes and mud rip-up clasts. Migration may have been restricted to periods where a spring tide and/or increased wave activity enhanced the geostrophic flow above the threshold to support active bedform migration. During periods of increased wave activity alone (no active mega-ripple migration) the mega-rippled ridge surface would have been rippled by combined-flows. Interference rippling predominated due to the geostrophic unidirectional current not being orthogonal to the

wave oscillation direction - Tillman & Martinsen (1984) reported a similar pattern in the Shannon Sandstone. The more turbulent, and therefore better oxygenated, sand ridge supported an active soft-bodied fauna comprising: infaunal deposit feeders that maintained open burrow systems and benefited from the semi-permanent currents that would have irrigated the burrow system with oxygenated water, infaunal suspension feeders and epifaunal browsers.

Large isolated zones of sand in inter-ridge areas were formed into duned sand patches, or even sand waves with angle-of-repose foresets where the sand supply was sufficient. The sand waves (plate 4.19C) are not inconsistent with the interpreted geostrophic origin for the sand ridge fields (Boggs 1974, Hunt *et al.* 1978, Swift *et al.* 1978, Stride & Chesterman 1973).

During the summer (?monsoon season) mud deposition would have dominated, partially blanketing the sand ridge complex; dune migration would have ceased. Short duration intense summer storms would have rippled (combined flow type) the duned surface and laterally extensive 'pause planes' would have developed during major storms which resulted in net erosion of the ridge. A storm sand apron would have been deposited around the ridge flank, interdigitating with inter-ridge muddy heterolithic deposits..

4.3.1.4 The Size of the 'Lee Stone Facies Association'

Although the lateral extent of the exposures within which the 'Lee Stone facies association' was observed were limited it is clear that both the width and length of individual sandstone lenses is in the order of hundreds of metres; the thickness of the sand lenses is c. 2m. When these dimensions are compared with those reported for similar coarsening-upwards sandstone-bodies preserved in Western Interior Seaway deposits it is apparent that the 'Lee Stone facies association' is appreciably smaller - see table 4.6. In the scheme of Shurr (1984) which details a dimension hierarchy for ancient shelf sandstone-bodies the 'Lee Stone facies association' sandstone-body falls into the smallest class (V) i.e. 'facies package'. In comparison, the majority of the sandstone-bodies shown in table 4.6 fall within the 'elongate lens' (IV), 'regional lenticle' (III) and 'sheet' (II) classes, corresponding to sand ridges, sand-ridge fields and areas of sandstone deposition confined to lineament-bound blocks, respectively. Shurr's class V 'facies packages' were interpreted as representing 'sand waves'. This nomenclature is confusing as Shurr's 'sand waves' do not conform to the 'sand waves' comprising high-angle avalanche foresets described by other workers and used

herein for inter-ridge sandstones with large-scale angle-of-repose foresets (see text-figure 4.23). For this reason the 'Lee Stone facies association' is referred to as a small-scale sand-ridge.

| Lithostratigraphic Unit | Author | Sandstone-body Height / Thickness (m) | Length (km) | Width (Km) |
|--------------------------------|-----------------------------------|---------------------------------------|----------------------------|------------|
| Duffy Mountain Sandstone | Boyles & Scott (1982) | 27 | >50 | 8 - 16 |
| Gallup Sandstone | Campbell (1971) | 6 | 6 - 64 | 3 |
| Semilla Sandstone Member | La Fon (1981) | 12 - 21 | >20 | 8 - 15 |
| Shannon Sandstone Member | Spearing (1976) | 15 / 22 | 50 / 100 | 30 / 50 |
| | Seeling (1978) | 20 | 17.6 | 2.4 |
| | Shurr (1984) | 12 - 18 | Area c. 52 km ² | |
| | Tillman & Martinson (1979 & 1984) | 20 - 23 | >29 | >11 |
| Sussex Sandstone Member | Berg (1975) | 12 | 32 - 160 | 8 - 48 |
| | Brenner (1978) | 30 | 50 | >10 |
| | Hobson <i>et al.</i> (1982) | 12.2 | 45 | 1.6 |
| Viking Formation | Evans (1970) | 3 - 10 | 113 | 11 - 22 |
| | Koldijk (1976) | 2.5 | 30 | 3.3 |
| | Reinson <i>et al.</i> (1983) | 9.5 | 40 | 1.6 - 3 |
| | Beaumont (1984) | 18 - 36 | Not given | |
| 'Lee Stone facies association' | | 2 | <1 | <1 |

Table 4.6 Dimensions of some coarsening-upwards sandstone-bodies interpreted as linear sand ridge deposits reported from the Western Interior Seaway of the U.S.A. - comparison with the 'Lee Stone facies association'

The limited size of the 'Lee Stone facies association' sand ridge compared with coarsening-upwards sequences described from shelf sandstone-bodies in the Western Interior Seaway is interpreted to be due to: (i) the 'lower-middle mega-facies' shelf being poor in sand-grade material; (ii) the 'Exmoor Basin' had relatively small dimensions (in terms of fetch and limited water depth) which would have resulted in smaller sand ridge spacings and heights according to Huthnance (1982) bedform stability theory.

Shurr (*op. cit.*) observed that lensoid 'small-scale facies packages' were commonly imbricately stacked (e.g. figure 14 of Shurr). A similar imbricate stacking is observed at Lee Stone where a sandstone body above the western end of the exposure was vertically stacked above the logged 'Lee Stone facies association'. This stacking is attributed to amplification of an initial shelf irregularity (the plano-convex lens of 'Watersmeet lithotype' in this particular instance) causing flow acceleration accompanied by down-current deceleration and deposition which would have been self sustaining over a period of millennia, resulting in a long-lived site of preferential ridge aggradation.

4.3.2 The Geometry of the 'Watersmeet Lithotype' at Watersmeet & Lee Stone

A comparison of the 'Watersmeet lithotype' (facies 3) with the cross-bedded facies 8 of the 'Lee Stone facies association' shows many characteristics held in common:

- ❑ Cross-bedding style i.e. both facies exhibit: hanging set boundaries, reactivation surfaces, climbing sets, mudstone intraclasts resting on foresets, wedge-shaped sets, similar set thicknesses, similar palaeocurrent distributions, rare sets of 'herringbone' cross-stratification, a similar style of set bounding surfaces.
- ❑ Both facies are internally sub-divided by the same hierarchy of surfaces: concordant mudstone drapes on foreset surfaces - discontinuous planar surfaces (overlying erosional or non-erosional surfaces; with or without a mudstone drape) - laterally extensive erosion surfaces (with or without a mudstone drape).

The prime difference between facies 3 and facies 8, therefore, is the textural maturity of the latter in comparison to the poorly sorted character of the 'Watersmeet lithotype' and immaturity in clast types. A consequence of this textural difference is the different biogenic assemblages preserved by the two facies. Although facies 3 could support an infauna it is restricted to a few examples of *Palaeophycus tubularis* burrows in mudstone clasts and drapes and indistinct biogenic 'churning' penetrating a few bed tops. In contrast, facies 8 supported a diverse assemblage of infaunal deposit feeders, infaunal suspension feeders and epifaunal browsers. This difference in assemblage diversity is attributed to the textural differences in the substrate (which would also have affected the preservation potential of biogenic traces between the two facies) rather than to any difference in the hydrodynamic environment in which the two facies were deposited.

In summary, the similarity between the suite of physical sedimentary structures in facies 3 and facies 8, and their close stratigraphic relationship at Lee Stone, implies that both facies were moulded in the same hydrodynamic environment i.e. geostrophic flow enhanced by oscillatory currents (possibly) superimposed on a weak tidal régime. Furthermore, the presence of low-angle laterally extensive 'pause planes' in both facies is suggestive of sand-bodies that were subject to periods of accretion interspersed with periods of inactivity.

Despite the above similarities, however, the 'Watersmeet lithotype' and the 'Lee Stone facies association' have a strikingly different set of facies associations. The 'Lee Stone facies association' coarsening-upwards sequence represents the aggradation of a sand ridge over a period of millennia, during which time the deposit achieved a relatively high degree of textural maturity when compared with the 'Watersmeet lithotype'. In contrast, the 'Watersmeet lithotype' abruptly interrupts a sequence of muddy heterolithic marine shelf deposits. The only evidence for the presence of a major sandstone-body in the sequences that immediately encase the 'Watersmeet lithotype' at Lee Stone and Watersmeet is a higher proportion of storm-generated thinly-bedded sheet sandstones, presumably derived from the nearby sand-body.

In section 4.4.3 a case is put for a rapid (?catastrophic) influx of 'Watersmeet lithotype' sediment to account for the developments of this facies in the 'lower-middle mega-facies'. Due to its relatively coarse grade the facies 3 substrate would have required higher flow powers to initiate mega-ripple migration when compared to the facies 8 substrate. This in part accounts for the textural immaturity of facies 3. The presence of the laterally extensive erosional 'pause planes' in facies 3 indicates that the sand-body was structurally organised, although only to a limited degree, and subject to alternating periods of aggradation and quiescence over a number of seasons. Certainly the high carbonate content, including primary micrite, in the facies 1 muddy heterolithic deposits that immediately underlie the 'Watersmeet lithotype' at Watersmeet suggests winnowing of argillaceous material from the 'Watersmeet lithotype'. The 'Watersmeet lithotype' then subsequently prograded over the carbonate-enriched muds. Nevertheless, it appears that the 'Watersmeet lithotype' sand-bodies were buried relatively rapidly under a muddy heterolithic blanket long before they could be swept together into (?near) equilibrium sand-bodies of a 'Lee Stone facies association type'. The burial process may have been rapid if a large amount of argillaceous material was input to the basin by the event that introduced the 'Watersmeet lithotype' sediment at this level.

In summary, the geometry of the 'Watersmeet lithotype' is difficult to predict as the sand-bodies did not attain hydrodynamic equilibrium. The deposit at Lee Stone is clearly plano-convex, suggesting at least a degree of moulding took place prior to premature burial preserving the observed geometry. The deposit at Watersmeet would have initially had a 'fan-like' geometry following rapid emplacement at the foot of the Lynmouth - East Lyn Fault scarp (see section 4.4.3). Evidence presented in the preceding paragraph, however, indicates that the deposit migrated along the base of the fault scarp under the influence of normal

shelf currents. The Watersmeet deposit may, therefore, have an incipient plano-convex ridge geometry but exposure at Watersmeet is limited and a tabular geometry is all that is visible at outcrop.

4.3.3 Thick, Coarsening-upwards Parallel-laminated Sandstone Sequence

The sequence preserved within the A39 road section exhibits a coarsening- and winnowed-upwards trend. The lower part of the sequence comprises strongly bioturbated lenticular bedding (facies 1) with intercalated graded mudstone units (facies 2). The lenticular bedded sequence is replaced upwards by bioturbated, parallel-laminated muddy sandstone (facies 11) containing a unit of bioclastic sandstone (facies 4). Moving upwards through the parallel-laminated sequence, units become coarser and cleaner, also becoming less bioturbated, the top of the exposed sequence only displaying isolated *Arenicolites* burrows. These features indicate that turbulence increased upwards through the sequence. The upper part of the parallel-laminated sequence contains units of onshore-directed trough cross-bedding (facies 9), wave-ripple cross-laminated sandstone (facies 7). Unfortunately, in the absence of further exposure, the continued vertical extent of the sandstone-body is unknown, as is the nature of the critical contact with the overlying lenticular bedded sequence.

In the discussion of facies 11, above, it was alluded that the sequence preserved within the A39 road section closely matches the predicted vertical sequence for progradation of a non-barred high energy, wave-dominated nearshore environment cf. Clifton *et al.* (1971). With reference to the model presented by Clifton *et al.*, reproduced herein as text-figure 4.18, the lower lenticular bedded portion of the sequence represents an offshore sequence deposited below fair-weather wave-base in the absence of flow asymmetry generated by shoaling waves. The base of the parallel-laminated sequence equates with the base of the nearshore sequence. The absence of an extensive sequence representing a zone of lunate megaripples, cf. the 'outer rough facies', reflects the paucity of sediment coarser than fine sand grade. Clifton (1976) demonstrated that the generation of megaripples by asymmetric wave orbitals is a function of grain size, megaripples only being generated where the sand is of medium to coarse grade.

The parallel-laminated sequence recorded in the upper part of the A39 road section corresponds to the 'outer planar facies'. The lower part of the parallel-laminated sequence represents deposition in the inner portion of the zone of wave build up. Fluctuations in fair-weather wave-base are represented by the extensive

bioturbation which marks periods of relative quiescence. High wave energy was not persistent enough to allow mud to be significantly winnowed from the sand/mud admixture. The single set of onshore-directed trough cross-bedding developed in conditions analogous to the 'outer rough facies', its existence permitted by the localised presence of medium grade sand. Clifton *et al.* observed a similar intercalation of 'outer rough facies' in can cores, attributing the intercalation to fluctuations in wave-base/energy. A major storm event was responsible for sweeping in the allochthonous debris of the bioclastic sandstone unit.

The upper part of the parallel-laminated sequence represents the inner zone of the 'outer planar facies' and was deposited under the influence of the outer part of the surf zone. Flow asymmetry would have been persistent enough to generate conditions of near constant pseudo-unidirectional flow allowing an upper phase plane bed flow regime to become established; the product of this flow régime is the thick sequence of parallel-laminated clean sandstone. The relatively persistent turbidity of this zone allowed the *Arenicolites* organism to become established. During periods of decreased wave-energy, wave-ripple fields developed.

In conclusion, the A39 road section represents the progradation of a wave-dominated shoreface; the observed sequence recording offshore through to middle shoreface environments. However, the transition from offshore to lower shoreface deposits is unusually abrupt i.e. there are no thicker storm-emplaced sheet sandstones intercalated with lenticular bedding in the part of the sequence interpreted as marking the upper offshore (immediately below 4.14m on the log). An apron of storm-emplaced sheet sandstones in the upper offshore normally presages an overlying sandy shoreface sequence in a vertical progradational succession e.g. Oxen Tor (section 6.3) and Little Burland (section 7.11). The transition from offshore to shoreface deposits is marked by a graded mudstone (facies 2) at 4.14m on the log through the A39 section; there is no evidence of an unconformity or depositional hiatus at this level. The A39 offshore to shoreface succession is, therefore, rather enigmatic in that a rapid shallowing and influx of sand-grade material is indicated at the point that the shoreface part of the succession occurs.

As noted above, the absence of further exposure precludes analysis of the critical 'inner rough facies' and 'inner planar facies', representing the inner surf zone and swash zone environments, that would be predicted to overlie the exposed sequence.

Several other studies of modern wave-dominated shelf to shoreface profiles have revealed predicted vertical sequences similar to that preserved in the A39 road section. Bernard *et al.* (1962), in a now classic study of a progradational barrier island profile from the Gulf of Mexico, described a transition from heavily bioturbated shelf muds containing occasional storm-generated sand layers, through thinly interlayered graded sand units within silty clays with abundant bioturbation of the lower shoreface. The middle shoreface comprised laminated sand and cross-bedding, along with shell layers which frequently graded upwards into fine sand; bioturbation was abundant. Finally, the upper shoreface displayed parallel-laminated sand with low-angled discordances and occasional ripple-bedded layers; bioturbation in this zone was virtually absent. Although longshore bars developed parallel to the coast, the bars were levelled during storms and were, therefore, considered to have a low preservation potential.

Reineck and Singh (1971) described a shelf to beach sequence on the non-barrier and non-tidal Tyrrhenian Sea coast. Strongly bioturbated shelf muds containing laminated and weakly graded storm silt layers were succeeded by lower shoreface, strongly bioturbated laminated sands via a transition zone of almost completely bioturbated fine sand. The middle shoreface displayed moderately bioturbated, laminated sands containing gentle internal discordances and minor amounts of ripple cross-lamination. Weakly bioturbated, cross-bedded and cross-laminated medium grade sands of the upper shoreface were succeeded by a swash cross-laminated foreshore sequence and backshore dune deposits. In contrast to the Galveston Island sequence described above, a longshore bar was observed and was noted to only be active during storms, resulting in a middle shoreface parallel-laminated sand-body which migrated onshore.

The additional two studies cited above contain important implications for the model proposed for the A39 road section. Firstly, it is apparent that the internal sequence does not contain characteristics allowing a distinction to be made between a barrier and a non-barrier coastline to be made. The critical major erosion surface and back barrier washover fan and lagoonal facies necessary to distinguish between a barrier and non-barrier sequence are not exposed in the A39 road section; the barrier/non-barrier origin of the A39 road section must therefore, remain equivocal. Secondly, the parallel-laminated sequence may, or may not, represent a middle shoreface longshore bar; the studies cited above suggest that differentiation between a sequence containing a longshore and a non-barred sequence is very difficult. Nevertheless, the absence of a major erosion surface, overlain by a lag deposit, at the base of the parallel-laminated sequence suggests that no longshore bar was developed; the onshore migration of a longshore bar would be characterised by the

presence of such an erosion surface and overlying lag deposit. The possibility that the A39 road section represents a well developed multiple-barred environment can certainly be discounted. Studies of multiple-barred environments by Davidson-Arnott and Greenwood (1976) and Hunter *et al.* (1979) indicate that preserved deposits for a barred coastline would be dominated by rip channel deposits overlying a strong erosion surface and bar trough facies; there is no evidence of analogous deposits preserved in the A39 road section.

The A39 road section also closely resembles ancient beach/barrier sequences attributed to progradation of a wave-dominated shoreline. Howard (1972) described an Upper Cretaceous sequence from Utah which exhibits a passage upwards from highly bioturbated mudstones, with thin bedded sandstones and isolated current-generated structures (offshore) into bioturbated, parallel-laminated sandstones containing occasional cross-bedded units (lower shoreface). These were succeeded by weakly bioturbated, low-angle laminae containing minor amounts of ripple cross-lamination and trough cross-bedding (upper shoreface). Finally, a sequence of trough cross-bedding overlain by swash cross-lamination, with no bioturbation, was exhibited (upper shoreface to foreshore).

Campbell (1971) described a similar Upper Cretaceous sequence from New Mexico. Bioturbated mudstones containing laminated and ripple siltstones (offshore) were overlain by bioturbated parallel-laminated shoreface sandstones with occasional cross-laminated intervals. The top of the sequence displayed a sequence of swash cross-lamination and wave-ripples.

4.4 SAND-BODY PRESERVATION ON A MUDDY SHELF

On enclosure 2 the A39 coarsening-upwards shoreface sequence, occurrences of the 'Lee Stone facies association' at Duty Point, Ruddy Ball and east Lynmouth Beach, and the 'Watersmeet lithotype' at Watersmeet are shown as being laterally equivalent. The purpose of this section is to explore the reasons for these sandstone-bodies being preserved in an otherwise thick, muddy shelf succession. In the cliff-face below Castle Rock (704 497) a section through the 'lower-middle mega-facies' above the Wringcliff Bay log (see chapter 3), reveals no evidence of any significant change in base-level / sandstone-body development / major erosion surface at the presumed level of the 'lower-middle mega-facies' sandstone-body developments seen elsewhere, or indeed at any other level in the Castle Rock cliff-face. This indicates that the 'lower-middle

mega-facies' sandstone-body developments are not the result of a rapid increase/decrease in relative sea-level / relative accommodation space. With this factor in mind, combined with strong evidence that the Lynton Formation was deposited in a tectonically active basin, it is posited that the 'lower-middle mega-facies' sandstone-body developments are a response to a tectonically-induced episode. It is in this context that the possible origin of each of the three sandstone-body types is discussed in turn in the following sub-sections.

The preservation of sand-bodies in muddy shelf sequences has perplexed many researchers, particularly those working on the Jurassic - Cretaceous Western Interior Seaway of North America where sandstone lentils are frequently encased in thick shelf mudstone sequences. The problem is essentially one of sand supply to the shelf i.e. how is the 'littoral energy fence' (*sensu* Allen 1970b), where wave-surge transports sediment onshore, breached? The following mechanisms have been proposed:

(i) Storm erosion of coastal sands (Hayes 1967, Reineck & Singh 1972, Morton 1981, Johnson & Belderson 1969, Caston 1976).

(ii) Circulatory tidal sands connected to beaches and offshore bars (Clayton *et al.* 1983).

(iii) Erosional transgression & shoreface retreat (Swift *et al.* 1978, Field 1980, Stubblefield *et al.* 1984).

McCave (1985) stated that: "offshore linear banks, abandoned by rising sea level, are thought to form closed sand circulations in many cases and thus not be supplied from the coastal zone" (p.60). Examples of sandstone-bodies in the Western Interior Seaway interpreted as offshore sands related to periods of still-stand or transgression include Hobson *et al.* (1982), Palmer and Scott (1984). Beaumont 1984 suggested that the Viking Formation represented a transgressed delta reworked into coarsening-upwards shelf sand ridges.

(iv) Offshore transport of sand through tide-dominated delta mouths moving sand beyond the zone of onshore wave-surge (Meckel 1975, Levell 1980b). Swift (1976) referred to this mechanism as the 'ebb tidal jet' and 'flood stage jet'. Levell (*op. cit.*) described the 1.5km thick Lower Sandfjord Formation (N Norway) and attributed sand supply to repeated transgression and regression of tide-dominated deltas flanked by sandy coastal plains with extensive braided stream systems. Levell suggested that syn-depositional faults may have caused stacking of the various facies belts.

(v) River mouth flooding (Drake *et al.* 1972)

(vi) Reworking of ephemeral fan deltas / terminal fans cf. Tunbridge (1981a). I.E. Sediment is deposited at the coast as ephemeral fan deltas (cf. Glennie 1970) but would have had a low preservation potential due to being reworked by coastal and shelf currents.

In discussing the preservation of shelf sand-bodies Cant and Hein (1986) suggested that external factors interplay in varying ratios: sea-level variation / sediment input / subsidence (a combination of tectonic and sediment loading). Cant and Hein proposed that the interplay of these mechanisms allows three styles of sandstone-body to be recognised:

(i) *Cyclic coarsening-upward* - Cf. Western Interior Seaway foreland basin of U.S.A., particularly hummocky cross-stratified sandstone-bodies. Each progradational coarsening-upwards cycle was related to thrust events.

(ii) *Noncyclic shelf deposits* - Comprises sequences that lack bed thickness or grain-size trends. e.g. the Lower Cambrian Gog Quartzite of western Canada which consists of 1500m of mainly sandstone with thin interbedded mudstone. The Gog Quartzite was interpreted as the product of rapid tectonic subsidence with high sediment input which kept pace with subsidence. Thus, no significant facies trends developed due to strong semi-permanent wind-driven and/or tidal currents which redistributed sediment evenly over shelf; no major shoreline progradation took place.

(iii) *Stratigraphically isolated patchy sandstones* - This type consists of coarse deposits which are isolated at a single stratigraphic level. Examples included the Cretaceous Cardium and Viking Formations of western Canada, a conglomeratic unit preserved within muddy shelf deposits. The Cardium and Viking Formations were interpreted to be the product of a tectonic pulse supplying coarse material. In contrast, Weimer (1984) suggested a eustatic low-stand origin for the Cardium and Viking Formations i.e. a rapid fall in sea-level resulted in fluvial incision and rapid erosion due to steepening of gradients resulting in rapid deposition on the shelf. Lowstand channels and disconformities were not preserved because they were drowned by the subsequent transgression and winnowed. The Cardium & Viking Formation sandstone-bodies original morphologies (e.g. barriers and deltas) were preserved, however, although the suite of physical and biogenic structures was altered to reflect the new environmental setting.

With these mechanisms in mind we may now examine the 'lower-middle mega-facies' sandstone-bodies:

4.4.1 A39 Coarsening-upwards Shoreface Sequence

In section 4.3.3 the A39 road section succession was interpreted as representing the progradation of a wave-dominated shoreface. In the absence of a change in relative sea-level, or a geologically 'instantaneous' decrease in the relative accommodation space available, progradation must be the product of the rate of deposition exceeding the rate of accommodation space change (Van Wagoner *et al.* 1988). Relatively few modern shorefaces are actively prograding systems. Those that are result from unusually high rates of sand input, generally from well established perennial fluvial systems e.g. the coast of Nayarit (Mexico) supplied by the Rio Grande de Santiago (Curry *et al.* 1969). Heward (1981) published an extensive review of both ancient and modern prograding shoreline deposits and stressed that observed and interpreted rates of progradation require *direct* fluvial sediment supply. Sediment supply to the Lynton Formation shoreline, however, occurred at rather ill-defined input zones fed by ephemeral streams (Tunbridge 1981a), there being no evidence for established channels reaching the semi-arid shoreline to produce distinct prograding deltaic lobes (Tunbridge 1983a). It seems unlikely therefore that progradation of the A39 shoreface could have resulted from high rates of fluvial sand supply, particularly given the rapid shallowing implied by the sharp upper offshore to lower shoreface transition.

It is suggested that the enigmatic A39 sandstone-body was the product of catastrophic emplacement of sand at the foot of the Lynmouth - East Lyn Fault scarp, a hypothesis supported by the localised nature of the A39 sandstone-body. This rapid input would have resulted in localised shallowing above the offshore deposits that accumulated at the foot of the fault scarp prior to the catastrophic event. Initially the deposit would presumably have had a lobate geometry, but over time the sands would have been reworked by waves and longshore currents to give a linear wedge-shaped geometry parallel to the line of the Lynmouth - East Lyn Fault. Cant (1984) recorded a similar phenomenon in the Lower Cretaceous Spirit River Formation of Alberta (Canada) which exhibited a series of regressive cycles within an overall transgressive situation. Each cycle comprised parallel-laminated shoreface sands abruptly overlying shelf mudstones and siltstones. Cant attributed this phenomenon to: ". . .the influx of vast amounts of sediment to overwhelm sea-level rise

and cause rapid progradation” (p551), although the mechanism to explain the vast influx was not given. Regardless of the interpretation for the mode of sediment input, sand supply to this stretch of the Lynmouth - East Lyn Fault was short-lived as exposures in the slopes above the A39 road section indicate a return to offshore muddy heterolithic deposition.

There is a further line of evidence to support the catastrophic sand supply hypothesis. It was noted in section 4.3.3 that the point of rapid shallowing of the sequence, identified by the onset of shoreface deposits, is marked by a graded mudstone event deposit (facies 2) with a sharp planar to erosive base. Of the several possible interpretations offered for the origin of this event deposit in section 4.2.3.3 one stands out: the graded mudstone is an event deposit triggered by syn-sedimentary tectonic movement initiating slumping and sliding on the shelf which resulted in the suspension of large amounts of fine grained material that would have settled out to form a mud blanket. The local scouring at the base of the mud deposit was attributed to localised currents generated in response to the downslope translation of large amounts of sediment. Thus, if this interpretation for the origin of the graded mudstone is correct, it supports the interpretation of the catastrophic emplacement of a sand that resulted in a localised progradation of a shoreface over offshore muds capped by a graded mudstone event deposit. Additionally, a metre above the graded mudstone event deposit a bed of poorly sorted bioclastic sandstone (facies 4) occurs. Although the bed was not assigned to the ‘Watersmeet lithotype’ (because it lacks the characteristic cross-bedded structure), it does have a similar lithology. The bed was interpreted as an allochthonous storm-generated event layer - it is possible that the material was derived from a sand ridge comprising material catastrophically deposited at the foot of the fault-scarp at the same time as the ‘Watersmeet lithotype’ at Watersmeet was emplaced.

N.B. It is possible that the top (not exposed) of the shoreface sequence preserved by A39 sandstone-body was never sub-aerially exposed i.e. the rapid input of sand at the base of the Lynmouth - East Lyn Fault scarp, and its subsequent reworking, may have taken place below low-water mark

4.4.2 'Lee Stone Facies Association' Sandstone-body Origin

Modern sand ridges are moulded into the upper surface of the Holocene transgressive sand sheet that covers modern shelves (Swift 1976). Even where mud is blanketing sand ridge fields on modern shelves, the sand ridges protrude upwards through the mud from the underlying Holocene transgressive sand sheet (e.g. Twichell *et al.* 1981). Consequently there are no documented examples of sand ridge lentils preserved within muddy shelf deposits.

In the absence of modern analogues, it is of benefit at this point to examine the mechanisms proposed by Swift & Rice (1984) to account for Western Interior Cretaceous Seaway coarsening-upwards sandstone-bodies. Swift and Rice recognised that the application in isolation of the stability analysis model of Huthnance (1982 - see section 4.3.1.3) to the Western Interior Cretaceous Seaway coarsening-upwards sandstone-bodies would require sand-bodies to have occurred at every horizon from the shoreline to the shelf edge spaced at distances determined by the Huthnance theory; the preserved record, however, clearly does not support such a simplistic picture. Swift and Rice attributed the concentration of the Western Interior Cretaceous Seaway coarsening-upwards sandstone-bodies into narrow (<30m thick) zones, laterally restricted into 'lentils', to be the product of the sand content of the storm-transported load exceeding a critical threshold so that the 'temporal acceleration model' (see text-figure 4.22A) is no longer valid and the 'spatial acceleration model' (see text-figure 4.22D) becomes applicable. Swift and Rice concluded that: "the pump must be primed by increasing the sandiness of the horizon on a regional basis before sand bodies can be triggered and grow" (p.55). The source of geologically sudden sand input was interpreted by Swift and Rice to be tectonic rather than eustatic in origin given the foreland basin setting of the Western Interior Cretaceous Seaway. Abrupt movements uplifting thrust plates would have both supplied a rapid pulse of sediment plus loaded the crust and induced subsidence. As the up-thrust block was denuded the rate of sediment supplied would have eventually exceeded the rate of load-induced subsidence and the shelf deposits would have recorded a shallowing- and coarsening-upwards style. Swift and Rice concluded, therefore, that the regional coarsening-upwards cycles fossilised the thrusting history of the Sevier Overthrust Belt that bordered the Western Interior Cretaceous Seaway. Each cycle coarsens upwards due to shallowing upwards as accommodation space diminishes with sediment input i.e. wave agitation of the sea floor will increase with time at any given point. At the beginning of each coarsening-upwards cycle mud-

grade material would generally be trapped on the shelf, whilst towards the end of the cycle mud-grade material would largely be bypassed over the shelf edge (see text-figure 4.22E).

Swift and Rice proposed a second, and more direct, tectonic mechanism to account for the Western Interior Cretaceous Seaway coarsening-upwards sequences by suggesting that reactivation of basement highs caused localised high points on the shelf surface. The relief of each high point was sufficient to cause spatial acceleration which in turn triggered the growth of localised sand ridges via the Huthnance (1982) mechanism described in section 4.3.1.3. Swift and Rice quoted the Virginia Beach Massif, Platt Shoals Massif and the Hudson Divide from the modern North Atlantic shelf off the U.S.A. as analogues where basement relief causes regional variations in storm flow intensity reflected in the pattern of sediment deposition around them. Ancient examples given the Shannon Sandstone Member at Salt Creek which developed over an actively growing palaeo-high (Tillman & Martinsen 1984).

The application of the Huthnance model by Swift and Rice to the Western Interior Seaway coarsening-upwards sandstone-bodies, sand ridge growth being triggered by a tectonically-induced influx of sand-grade material, provides an attractive mechanism to explain the 'Lee Stone facies association' preservation. In the absence of any sea-level change at the time that the 'Lee Stone facies association' accumulated, however, a mechanism is still required to explain how a tectonically-induced influx of sand made its way onto the Lynton Formation shelf. Of the non-eustatic mechanisms mentioned above, none provide a particularly plausible mechanism for providing the large volumes of sediment to the shelf during a particular period in time. Again, it is suggested that a catastrophic influx of sand, triggered by movement on the Lynmouth - East Lyn Fault, may have supplied the requisite volume of sand-grade material cf. the A39 coarsening-upwards shoreface sequence. Whereas the sand preserved in the A39 sequence was 'trapped' by normal wave-dominated shoreface maintenance processes, any sand that by-passed the 'littoral energy fence' during catastrophic emplacement was free to be redistributed across the shelf by fair and foul weather processes of the type that led to the generation of the 'Lee Stone facies association' sand ridge.

There is some indirect evidence to support the putative catastrophic mechanism invoked to account for the 'Lee Stone facies association' sand influx. At Ruddy Ball the 'Lee Stone facies association' is underlain by 30 to 40cm of connected and unconnected lenticular bedding resting on the infill to a large slump scar; the

slump scar is interpreted as representing down-slope sediment movement triggered by a seismic event (see section 2.2.2). 1.8km to the WSW at Duty Point the 'Lee Stone facies association' is underlain by 1.3 to 1.5m of lenticular bedding resting on a unit attributable to the 'Watersmeet lithotype'. The following set of events are proposed to account for these two sequences. A major movement on the Lynmouth - East Lyn Fault triggered the catastrophic emplacement of sand (e.g. A39 road section sand-body) and 'Watersmeet lithotype' material (e.g. at Watersmeet) at the foot of the fault scarp. The seismic event also triggered slumping at Ruddy Ball. 'Watersmeet lithotype' sediment deposited on the shelf, beyond the 'littoral energy fence', migrated for a period under peak normal shelf currents (geostrophic / wave / tide combined in varying proportions) as a series of irregular granule patches. The patches were not particularly mobile due to their coarse grain-size and were reasonably rapidly buried resulting in the preservation of an immature, poorly sorted grain fabric (e.g. at Duty Point below the 'Lee Stone facies association'). The sand-grade material, however, catastrophically emplaced on the shelf was moulded by normal shelf currents, migrating over a period of millennia, to form a series of sand ridges. Insufficient evidence is preserved to either state whether the sand ridges had attained regular spacings conforming to Huthnance bedform stability predictions before being encased in shelf muds, or whether the ridges were preferentially sited over palaeo-highs on the shelf.

4.4.3 'Watersmeet Lithotype' Origin

In section 4.3.2 the hydrodynamic origin of the 'Watersmeet lithotype' was attributed to a similar set of processes to those that moulded the 'Lee Stone facies association' i.e. semi-permanent geostrophic flow enhanced by a weak tidal current and storm-wave reworking. However, whereas 'Lee Stone facies association' type sand-bodies moved towards becoming a set of regularly spaced sand ridges conforming to the Huthnance (1982) stability theory over a period of millennia, 'Watersmeet lithotype' sand-body migration would have been more intermittent due to the coarser grain size i.e. as is evidenced by the immature sediment texture of the 'Watersmeet lithotype' in relation to the 'Lee Stone facies association'; the 'Watersmeet lithotype' sand-bodies would not have become equilibrium bedforms (i.e. attaining Huthnance stability spacings) prior to being buried by mud and preserved. It can be predicted, therefore, that the 'Watersmeet lithotype' sand-bodies would have had an irregular sand patch distribution with an elongated, broadly plano-convex geometry (cf. the Lee Stone occurrence) representing incipient ridges. The elongation

direction would have been slightly oblique to a line drawn normal to the mean palaeocurrent direction according to Huthnance stability theory.

As described in the preceding sub-section, it is suggested that the 'Watersmeet lithotype' was catastrophically emplaced at the foot of a submarine fault scarp, where it reached its maximum observed thickness at Watersmeet, and was subsequently reworked by strong longshore and shelf currents; palaeocurrent indicators in the 'Watersmeet lithotype' at Watersmeet indicate that the Lynmouth - East Lyn scarp was a positive feature at the time of emplacement. The deposit sharply overlies shelf deposits of thinly interlayered sandstone/mudstone bedding with a high carbonate content and has an abrupt top. Evidence for the catastrophic event that introduced the 'Watersmeet lithotype' at this stratigraphic level was described in the preceding sub-section.

5. THE 'UPPER-MIDDLE MEGA-FACIES' OF THE LYNTON FORMATION

5.1 INTRODUCTION

The western side of Crock Point provides the longest unbroken succession through the Lynton Formation that is suitable for logging. The logged section (enclosures 8A, B, C, D, location of section shown on enclosure 6) comprises a 123m long sequence through the 'upper-middle mega-facies', a monotonous succession of muddy heterolithic deposits. The sequence is occasionally cut by minor NW-SE trending faults which have a small throw and bedding can easily be traced across these faults. The base of the section occurs in an isolated 'keel' of rock at the extreme NW of Crock Point (4875 4945), some 2m above the barnacle cover on the north face of the 'keel'. The top of the section occurs in foreshore slabs (6857 4918) where the sequence is disturbed by a major zone of NW-SE trending faults and extensive quartz veining. The slabs along the west side of Crock Point dip 25 to 30° towards SSW. The tidal access window for the section is only 1½ hours either side of low tide at the northern end. It is possible to walk the 1.2km across the beach from the vicinity of Woody Bay lime kiln (6769 4898 - see enclosure 6), although this is not recommended as a scramble over treacherous algae-covered boulders marks much of the route and there is a danger of being cut-off by tides at a small promontory (683 491). For the purpose of logging, access was gained to the section by abseiling down the slabs at 6876 4943.

Although the west Crock Point section is superficially a monotonous succession of muddy heterolithic deposits there are several features of palaeoenvironmental interest:

- Channel-forms between: 12-13, 18-21, 29, 39-41m on the log.
- Slumped bed at 46m on the log
- Thick, unbioturbated mudstone units, with occasional thin graded sandstone layers, between 94 and 103m on the log
- An interesting ichnofacies distribution, particularly an upward gradation from *Palaeophycus tubularis* through to *Chondrites* sp. a burrows (see appendix B) through the lower $\frac{4}{5}$ ^{ths} of the section.

The purpose of the following sections is to provide a relatively brief description and interpretation of each facies, followed by the discussion of a model explaining trends in primary physical sedimentary and biogenic structures through the succession and their palaeoenvironmental significance.

5.2 FACIES

5.2.1 Facies 1 - Thinly Interlayered Sandstone/Mudstone Bedding

5.2.1.1 Description

Facies 1 comprises 86.3% of the west Crock Point succession where units have a mean thickness of 13.1cm ($n = 792$, $\sigma_{n-1} = 14.0\text{cm}$, min. thickness = 0.5cm, max. thickness = 100cm). Individual facies 1 units generally exhibit a tabular geometry (plate 5.1A), although there are examples of scours and channels in the lower half of the section. In many cases the boundaries between facies 1 units are arbitrary, corresponding to abrupt changes in the gross sandstone content of units. Many of the boundaries between facies 1 units have been exploited by bedding-plane slip, presumably reflecting the contrast in competency between adjacent units, and exhibit a clay gouge up to 2cm in thickness, occasionally accompanied by 0.5 to 2cm thick zones of quartz slickenside surfaces e.g. 26.75 and 38.6m on the log.

Laterally extensive erosion surfaces, with up to 2cm relief, commonly separate facies 1 units. Rarely, these erosion surfaces have isolated ripples of clean sandstone resting upon them or a thin (0.5 to 2cm thick) drape of unbioturbated mudstone (facies 3) e.g. 12.2m on the log. Elsewhere, horizons of clean rippled sandstone, overlain by a thin, generally unbioturbated, mudstone drape occur. The sandstones gradationally overlie bioturbated heterolithic deposits. The draping mudstone is frequently exploited by present-day erosion (plate 5.1B).

Internally facies 1 units have a sandstone content varying between 0 and 95%, variations within individual units defining fining-upwards, coarsening-upwards and coarsening-fining-upwards (e.g. 8.1m on the log) microsequences; the most frequently developed microsequences coarsen-upwards but are not as common as in the 'lower-middle mega-facies' (see section 4.2.2). In addition, there is a tendency for sandstone layers to thin upwards within fining-upwards microsequences and to thicken upwards on coarsening-upwards

microsequences. 1 to 3m thick intervals of groups of units with an overall higher sandstone content than that in the surrounding sequence are occasionally observed e.g. 6.9 to 8m on the log.

The grain-size of the sand in facies 1 varies between coarse silt and fine sand, units occasionally containing fine, comminuted shell debris. Individual sandstone layer characteristics allow four sub-facies to be defined:

Graded Rhythmites:

Thin, sharp-based coarse silt to fine sandstone layers with gradational tops and thicknesses ranging from 1 to 10mm (plate 5.1C). The sharp bases of individual layers is frequently irregular, infilling local irregularities in the underlying substrate, although overall the bases are planar. Although the granulometric contrast is frequently low within individual layers a weak normal grading is frequently visible. In addition, a faint internal lamination is occasionally present. The tops of layers are generally diffuse, grading into the overlying mudstone, although the contact is occasionally relatively abrupt. The graded rhythmites are more common in the upper part of the section than lower down and are frequently unbioturbated e.g. 118.7m on the log.

Horizontally-laminated Sandstone Streaks:

This sandstone layer type is more frequent in the west Crock Point section than elsewhere within the Lynton Formation, with the possible exception of the 'basal mega-facies'. Individual streaks have planar bases and tops and range in thickness from 1 to 15mm; some streaks pinch-out laterally. Occasionally the sandstone layers contain intraformational mudstone clasts up to 3mm in length e.g. 48.6m on the log. Horizontally-laminated sandstone streaks are commonly associated with *Chondrites* sp. a e.g. 9.5m on the log.

Unconnected Lenticular Bedding:

This bedding type comprises isolated sandstone lenses, 2 to 12mm in height and 10 to 30mm in length, set within mudstone. Individual lenses exhibit a lamination style characteristic of wave and wave-current (combined flow) origin. The paucity of bedding plane surfaces makes crestline trends difficult to measure.

Connected Lenticular Bedding:

Connected lenses generally have a pinch-and-swell morphology, although planar-based sandstone layers are occasionally visible. Sandstone layers range from 2 to 20mm in thickness; rippled upper surfaces have wavelengths between 5 and 10cm and amplitudes of 5 to 10mm. Individual layers exhibit a lamination style characteristic of wave and wave-current (combined flow) origin. Again, the lack of bedding plane surfaces made crestline trend measurement difficult. Where crestlines were visible it was apparent that wave-wave interference ripples are common e.g. at 37.5m on the log ripples trending 074-254° modify a previous set of indeterminate trend.

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Facies 1 is occasionally disturbed by soft sediment deformation which locally disturbed bedding intensively. Near the base of the section (2m on the log) a 50cm thick zone of convolutions and load pillows are visible. The convolutions have an approximately vertical axis of symmetry and decrease in amplitude both upwards and downwards within the unit. Load pillows of sandstone are common and frequently large, attaining lengths of up to 80cm. There is no evidence for lateral flowage having occurred when this unit was deformed. The absence of bioturbation in this and the preceding unit suggests that the substrate remained 'soupy' following the deformation event that produced the convolutions. This contention is supported by the presence of loaded bases to the thin sheet sandstone (facies 2) and isolated sandstone lenses in the succeeding unit.

Higher in the section (46m on the log) a 33cm thick slide unit is preserved which exhibits bedding dipping at approximately 40° to the south within a recumbent fold which faces towards the south; the top of the slide unit is planar and has been bioturbated. When the effect of Variscan folding was subtracted a palaeoslope dipping towards the south was indicated (see section 2.2.1 for a discussion of the use of slide units for ascertaining palaeoslope direction).

In the zone between 12 and 41m on the log a series of channel/scour-forms occur. The first group of channels/scours appear 12.2m above the base of the logged sequence and comprise 4 channels/scours, with depths of 23, 27, 19 and 9cm, which cut into underlying facies 1 units; the latest channel/scour incises into

the infill of the penultimate channel/scour. The channel/scour bases are erosive and have a dip of up to 10° at their margins. The channels/scours were infilled with a concordant drape of unbioturbated connected (rarely unconnected) lenticular bedding which thickens towards the centre of the channels/scours - this group of channels is figured in plate 5.2 which shows the cross-cutting nature of the channels/scours and the planar erosion surfaces that truncate the channels/scours. The second and third channel/scour scour complexes occur at 18.5 to 19.1 (59cm deep) and 19.6 to 20.3m (70cm deep) above the base of the logged sequence respectively. In both cases the channels/scours have been filled concordantly with bioturbated lenticular bedding which thickens towards the centre of the channels/scours and is internally sub-divided by channel/scour erosion surfaces that follow that are concordant with the basal channel/scour erosion surface. The fourth channel/scour complex occurs between 28.9 and 29.6m above the base of the log and is 70cm deep and 4m in width. The channel/scour is filled with a sequence of unbioturbated silty-mudstone broken only by a 5cm thick thinly-bedded sheet sandstone (facies 2) near the top of the fill which exhibits a basal zone of load pillows; the top of the channel/scour scour is truncated by a planar erosion surface. The fifth channel complex occurs between 39.3 and 39.9m in the logged sequence and is 60cm deep and 3.6m wide. A bedding plane view of the upper surface of this feature enabled the feature to be unambiguously identified as a channel-form with an axis that trends 156-336° i.e. oblique to the S-dipping palaeoslope hereabouts (see above) and parallel to the mean SSE-palaeocurrent direction in the west Crock Point succession. The sixth and final channel/scour structure occurs between 40.1 and 40.9m in the logged sequence and is 76cm deep and 4.6m wide. The channel/scour was filled by a single unit of bioturbated lenticular bedding.

Between 30.2 and 31.2m on the log a 1m thick zone of bioturbated lenticular bedding occurs which contains a high proportion of disarticulated shells (only \cong 1% are articulated). There are several zones, each up to 8cm thick, where the shells have been concentrated by penecontemporaneous winnowing, although the shells are still matrix supported. Bioturbation appears to have reoriented valves in less concentrated zones. Although the shells are disarticulated the valves appear to have suffered relatively little damage which suggests they were not transported far i.e. the assemblage is para-autochthonous. This unit is figured in plate 5.5A and is shown on the pie chart in text-figure 3.3A. Elsewhere there are pockets of comminuted shell debris with a grain-supported fabric in a fine to medium sandstone matrix which fill gutter cast within facies 1 (e.g. plate 5.5B) cf. Goldring and Aigner (1982).

5.2.1.2 Biofacies

Facies 1 contains a diverse ichnofauna dominated by a gradation between the open burrow systems of *Palaeophycus tubularis* (as evidenced by occasional specimens with collapsed burrow walls e.g. plate B.14B) and *Chondrites* sp. a (plate B.3.C & D) - see appendix B for a discussion of this phenomenon. In summary, straight *P. tubularis* burrows grade into *P. tubularis* burrows with occasional unequal-dichotomous branching upwards through the west Crock Point sequence (B4A to C). The *P. tubularis* burrows, both branched and unbranched, are aligned parallel to the mean palaeo-current direction i.e. NNW-SSE (sub-oblique to the dip direction of the southerly-dipping palaeoslope). The burrow entrances faced into a semi-permanent, obliquely-offshore flowing, geostrophic current that has been interpreted to have existed during the deposition of the lower 2/3^{ths} of the Lynton Formation (see section 4.2.2.3 and chapter 3). In turn, the obliquely-offshore-dipping branched burrows (e.g. plate B.3E) grade into radially disposed branching systems attributable to *Chondrites* sp. a. *Chondrites* sp. a burrows are frequently associated with the much smaller diameter burrows of *Chondrites* sp. b (plate B.5A & C) before, finally, *Chondrites* sp. a burrows become excluded and only the small *Chondrites* sp. b burrows remain. Significantly, occurrences of *Chondrites* are generally associated with the graded rhythmite and horizontally-laminated sandstone streak sub-facies of facies 1 in an environment where the substrate was below storm wave-base. This gradation from straight *P. tubularis* burrows to *Chondrites* sp. a and then a solitary assemblage of *Chondrites* sp. b burrows has been interpreted (see appendix B - discussion of *Chondrites*) as indicating a progressive decrease in the oxygen content of the sediment below a surficial oxygenated layer. Thus, much of the 'upper-middle mega-facies' (the portion containing *Chondrites*) had a sub-surface dysaerobic zone of sediment. *Muensteria* sp. (plate B11A-D) is occasionally found, generally in association with *Chondrites* sp. a. A similar association has been recognised by Wetzel (1983) in modern cores recovered from the Sulu Sea basin (Phillipines).

N.B. Simpson (1957) reported that the majority of *Chondrites* specimens in the Lynton Formation occurred in a 'concealed bed junction' mode of preservation. Observations recorded during the present study, however, indicate that the 'concealed bed junction' mode of preservation is relatively rare, the majority of *Chondrites* specimens occurring in a 'bed junction' mode of preservation e.g. plate B.3F. The observations regarding the mode of preservation of *Chondrites* reported in this study conform with observations reported by Hallam (1975).

An extensive development of *Teichichnus stellatus* (plate B.23D-F) occurs near the base of the section (between 5 and 7m on the log) where dense networks of burrows can be distinguished at three particular levels within the succession. In contrast, isolated specimens of *Teichichnus rectus* more commonly occur in the top 10m of the section, as do 'mantled' tubes. Seilacher (1963) observed that *Teichichnus*, *Chondrites* and *Zoophycos* were frequently associated in the Devshish Sandstone (Sinat, Iraq). Throughout the west Crock Point succession specimens of *Rosselia socialis* were only occasionally observed. A single specimen of *Zoophycos* sp. a (plate B.25A) was found at 39.9m in the logged succession, whilst a single specimen of *Zoophycos* sp. b (plate B.25B) was found at 23.6m in the logged succession. Biogenic 'churning' (*fossitextura deformativa*) is extensively developed throughout the succession, completely destratifying physical sedimentary lamination locally.

The sequence conforms to Schäfer's (1972) vital-lipostrate biofacies, and vital-pantostrate in the more distal parts of the section indicated by a higher proportion of graded rhythmities and horizontally-laminated sandstones within facies 1. Occasional horizons of laminated mudstone contain no evidence of biogenic reworking, e.g. the 'soupy' muds at 3m, the channels at 12 to 13m and the channel fill at 34m on the log; these have been assigned to Schäfer's letal-pantostrate biofacies.

Evidence for a nektonic biota was restricted to a single specimen of an orthocone (plate 5.5C) which has a post-mortem alignment parallel to the SSE-NNW palaeocurrent trend where it came to rest on the substrate; Miall (1984, figure 5.53) has reported a similar post-mortem orthocone alignment parallel to the palaeocurrent trend.

5.2.1.3 Interpretation

Facies 1 is closely analogous with facies 1 in the 'lower-middle mega-facies' and the following interpretations are based on the detailed discussion of the 'lower-middle mega-facies' facies 1 presented in section 4.2.1. In summary, the graded rhythmities are believed to represent silt and sand settling-out beneath storm wave-base from storm-generated suspension clouds and is regarded as the most distal sub-facies within facies 1 (see text-figure 4.8). The occasional presence of indistinct horizontal laminae within the graded rhythmite layers probably reflects the influence of wave orbitals on settling events from a turbulent suspension cloud. The horizontally-laminated sandstone streak facies was deposited in a slightly shallower

environment immediately below storm wave-base, the laminae being the product of 'pulsating' fall-out from suspension under oscillatory flow. The unconnected and connected lenticular sub-facies represent the migration of ripples below fair-weather wave-base during periods of increased wave / combined flow activity, interrupting appreciable periods of mud deposition below fair-weather wave-base. The unconnected sub-facies was deposited at similar depths to the connected sub-facies but was starved of sand, hence to discontinuous nature of the lenses. The laterally extensive planar erosion surfaces preserve the passage of high-energy events creating a zone of net erosion, whilst units exhibiting a thin coarsening-upwards cap of wave / combined flow interference ripples represent the winnowing of fines by slightly lower energy events which were still of sufficient energy to ripple the substrate.

The lack of shear surfaces and soft sediment deformation structures in the deposit immediately adjacent to the channel/scour structures in the lower part of the sequence precludes a rotational slump-scar origin for these features. Goldring (1971) recorded similar structures to the west Crock Point channels/scours in the Late Devonian Baggy Beds of North Devon where "... irregularly parallel or slightly sinuous grooves cut in shale, siltstone or sandstone . . . up to 1.2m broad and 20cm deep, with semi-elliptical to subrectangular cross-section, and are generally flat-floored and occur singly or more commonly in parallel groups" (p.33). Examples in the Rough Wall Member of the Baggy Beds displayed a subconcordant fill of laminated very fine grade sandstone and silty shales ('*Arenicolites curvatus* facies') similar to that present in the facies 1 structures. The consistent trend of the Baggy Beds grooves, constant breadth and depth, average width to spacing ratio of 1:6, gently sloping to vertical margins and up- and down-dip facing division of grooves strongly suggests that the grooves are analogous to flow-parallel furrows (cf. Dyer 1970) observed on present-day tidally-dominated continental shelves, although they could also presumably form on a shelf with a mixed meteorological and tidal current pattern.

A review of a similar channel/scour in the 'lower-middle mega-facies' (see section 4.2.4.3) concluded that the channel/scour structures in the Lynton Formation are of furrow origin, probably related to a secondary helical flow cell circulation superimposed on a mean palaeo-offshore flow during recurring episodic periods of directionally stable strong currents (cf. Flood 1983) e.g. trade wind season geostrophic flow with a

superimposed (?spring) tidal flow - this interpretation, it is suggested, is equally applicable to the west Crock Point channels/furrows.

Goldring and Aigner (1982) noted that scours with a 'banded' (sand/mud) infill are common in shallow marine sequences, the 'banded' infill suggesting that infill occurred over a period of time. Goldring and Aigner observed that the scour surfaces acted as preferential sites for colonisation by benthic organisms, suggesting that the firmness of the partially lithified, exhumed substrate and the protected aspect of the scours favoured settlement. No evidence of preferential colonisation was observed on the facies 1 channel/scour surfaces.

Although the facies 1 in the 'upper-middle mega-facies' is characterised by a strikingly similar suite of primary physical sedimentary structures and trace fossil assemblages to facies 1 in the 'basal mega-facies', the west Crock Point section exhibits a much lower proportion of soft sediment deformation structures and intraformational slide units. This suggests that by 'upper-middle mega-facies' time the Lynton Formation shelf was more stable.

The trace fossil assemblages are interpreted as being transitional between Seilacher's (1967a) *Cruziana* and *Zoophycos* ichnofacies. This assignment is based on the fact that the trace fossil assemblages contain many of the forms present in the 'lower-middle mega-facies', which was clearly represented a *Cruziana* assemblage (see chapter 4), and rare specimens of *Zoophycos*. N.B. It is possible that the west Crock Point succession contains a higher proportion of specimens of *Zoophycos* than is suggested by the collection gathered during the course of this study i.e. relatively few bedding-plane surfaces are available in the west Crock Point section and *Zoophycos* is notoriously difficult to distinguish from sandstone streaks in vertical cross-section.

5.2.2 Facies 2 - Thinly-Bedded Sheet Sandstones

5.2.2.1 Description

Facies 2 comprises 4.0% of the west Crock Point succession and individual beds have a mean thickness of 3.3cm ($n = 145$, $\sigma_{n-1} = 3.5\text{cm}$, min. thickness = 1cm, max. thickness = 21cm). The thin-bedded sheet sandstones, very fine to fine sand in grade, are generally enclosed in connected and unconnected lenticular

bedding (facies 1), as figured in plate 5.1A. Beds have planar bases, occasionally planar tops, and parallel or low-angle laminae internally which meet both the bed base and bed top asymptotically (plate 5.3A); beds are laterally persistent over the width of the outcrop. Some of the beds have a 'hummocky' top e.g. the bed figured in plate 5.6A has a hummocky top with a wavelength of 150cm and an amplitude of 4.5cm, pinching-out laterally. This has resulted in a set of discontinuous plano-convex lenses where the core of each hummock comprises shell debris in isolated plano-convex lenses 50cm in wavelength and 1cm in amplitude. Elsewhere, laterally discontinuous plano-convex lenses of facies 2 sheet sandstones were observed to overlie a laterally continuous planar erosion surface. Bed tops are frequently penetrated by indistinct biogenic structures. The bulk of the beds comprise a thick zone of parallel to gently undulatory laminae, the undulations having a wavelength of several decimetres and an amplitude of several centimetres. The uppermost part of the majority of beds is characterised by a lamination style diagnostic of an oscillatory/combined-flow origin. Soft sediment deformation structures are common e.g. 102.9m on log a bed which is parallel-laminated at the base passes upwards into convolute laminae. Elsewhere beds are occasionally broken up into load pillows (cf. Dzulynski and Kotlarczyk 1962) within the lenticular bedding of facies 1 (plate 5.3B & C). No proximity trends (cf. Aigner & Reineck 1982) for the sheet sandstones could be detected within the west Crock Point sequence.

5.2.2.2 Biofacies

Facies 2 bed tops are frequently penetrated by indistinct biogenic structures and, rarely, penetrated by 'mantled' tubes. More occasionally, beds are extensively disturbed by biogenic 'churning', which in some cases extended down to the base of the unit and resulted in the total loss of primary physical lamination. Facies 2 is assigned to Schäfer's vital-lipstrate biofacies.

5.2.2.3 Interpretation

Facies 2 matches facies 3 described in chapter 7 (section 7.5), to which the reader is directed for a more detailed discussion and interpretation. In summary, facies 2 is the product of the combined action of storm waves and associated obliquely-offshore-directed currents generated by on-shore directed winds. Initial storm conditions resulted in erosion, followed by net deposition during steady flow conditions at the storm peak. Finally, the top of the bed is wave-current rippled as the storm waned. The top of the bed was subsequently reworked by bioturbation or oscillatory currents. The planar tops to some of the hummocky

beds is reminiscent of HFM variants of HCS described in section 7.6 where beds of this type were interpreted to have been deposited by storm-generated flow with high sediment fluxes in an environment below storm wave-base. It is of interest to note that the Jenyncliff Slates (Eifelian), cropping out along the eastern side of Plymouth Sound (see Pound 1983 - enclosure 10), frequently exhibit HCS units of the HFM variety (see plate 7.18B).

5.2.3 Facies 3 - Thin Mudstone Beds

5.2.3.1 Description

Facies 3 comprises 1.4% of the west Crock Point succession and individual beds have a mean thickness of 7.4cm ($n = 22$, $\sigma_{n-1} = 6.3$ cm, min. thickness = 2cm, max. thickness = 26 cm). Units are generally silt-poor and normally graded with sharp, planar bases which are locally erosive; bed tops are planar and show a rapid gradation into the succeeding unit. The mudstone exhibits a good tectonic cleavage, bioturbate textures being rare and, if present are restricted to a thin, laterally restricted zone at the top of beds where they are picked out by sand piped down from the overlying bed (plate 5.4A & B).

5.2.3.2 Biofacies

The erosive nature of unit bases and absence of bioturbation, with the exception of rare laterally restricted zones at the top of some beds, indicates that the majority of facies 3 should be assigned to Schäfer's (1972) lethal-lipostrate biofacies, although bottom conditions became sufficiently oxygenated locally to support occasional bioturbation of bed tops in zones which are assigned to Schäfer's vital-lipostrate biofacies. At 23.6m on the logged section a specimen of *Zoophycos* sp. b (plate B.25B) was observed along with a single occurrence of *Megagraption aequale* (plate B.20A-C) which had been reburrowed by *Planolites montanus*. Biogenic 'churning' (fossitextura deformativa) disturbs tops of some units.

5.2.3.3 Interpretation

Facies 3 strongly resembles facies 2 of the 'lower-middle mega-facies' and the reader is directed to section 4.2.3 for a detailed discussion of the origin this bedding type. In summary, two possible origins were described. Firstly, a 'mud tempestite' where the beds were viewed as the product of rapid deposition from a waning flow of storm origin in a distal setting was proposed. An alternative explanation was offered: the

graded mudstones are event deposits triggered by syn-sedimentary tectonic movement initiating slumping and sliding on the shelf which resulted in the suspension of large amounts of fine grained material that would have settled out to form a mud blanket of the type preserved by this facies. The local scouring at the base of the mud deposit was attributed to localised currents generated in response to the downslope translation of large amounts of sediment.

N.B. The possibility that facies 3 represents a mud blanket deposited solely from a plume of turbid river flood water overriding denser (saltier) water (i.e. 'hypopycnal flow' of Bates 1953) is discounted. Although a hypopycnal plume could account for the supply of mud, a hypopycnal plume origin is not consistent with the laterally extensive planar, locally erosive, bases observed in some facies 3 units.

5.2.4 Facies 4 - Flaser Bedding

5.2.4.1 Description

Facies 4 comprises 1.3% of the west Crock Point succession where individual beds have a mean thickness of 7.8cm ($n = 20$, $\sigma_{n-1} = 4.7\text{cm}$, min. thickness = 1.5cm, max. thickness = 19cm). Units of this facies comprise a regular alternation of 1 to 3cm thick zones of very fine to fine grade sandstone with a lamination style characteristic of wave / combined flow / interference rippling separated by thin mudstone layers. An example of wave-wave interference rippling is figured in plate 5.5C. The majority of the pinch-and-swell sandstone layers are continuous being separated by 1 to 5mm thick mudstone laminae which either maintain a constant thickness over both ripple crests and troughs or thicken into troughs ('wavy flaser' bedding referred to in section 4.2.6.1); isolated mudstone flasers occur only rarely. The sandstone:mudstone ratio tends to remain relatively constant through the thickness of individual beds. Individual sandstone layers have bases which are either planar or preserve the rippled profile at the top of the underlying bed i.e. a massive/structureless sandstone basal layer; bed tops are either rippled or erosively truncated by the overlying unit.

5.2.4.2 Biofacies

This facies is occasionally disturbed by the effects of biogenic 'churning' and distinct burrows (*fossitextura figurativa*) are restricted to *Palaeophycus tubularis* and rare occurrences of *Rosselia socialis*. Facies 4 is assigned to Schäfer's vital-lipstrate biofacies.

5.2.4.3 Interpretation

The genesis and hydrodynamic significance of flaser bedding is discussed in section 4.2.7.3, to which the reader is referred for detail. In summary, facies 4 represents an environment where conditions allowing the preservation of sand were more favourable than those allowing the preservation of mud. Relatively frequent periods of increased wave energy would have resulted in the mobilisation of the rippled substrate and the winnowing of mud. The lack of intraformational mudstone chips in the sandstone layers suggests either that a significant period elapsed between periods of ripple mobilisation, allowing consolidation of the basal mud layer which would have effectively 'armoured' the underlying rippled surface (Hawley 1982).

5.2.5 Facies 5 - Thick Mudstone Units

5.2.5.1 Description

Between 93.9 and 103.1m above the base of the logged section a thick sequence of silty-mudstones containing thin graded rhythmites and horizontally-laminated sandstone streaks and punctuated by thicker sandstone layers (cf. facies 2), is developed. This facies comprises 7.0% of the west Crock Point succession and individual beds have a mean thickness of 38.9cm ($n = 22$, $\sigma_{n-1} = 45.0\text{cm}$, min. thickness = 4cm, max. thickness = 160cm); the intervening thicker sandstone layers (cf. facies 2) have a mean thickness of 2.75cm ($n = 12$, $\sigma_{n-1} = 3.8\text{cm}$, min. thickness = 0.5cm, max. thickness = 15cm).

The mudstones are medium dark grey in colour and are well-cleaved and in some cases separated by laterally extensive planar surfaces e.g. 96.7m on the log. The mudstones contain graded rhythmites and horizontally-laminated sandstone streaks which reach a maximum thickness of 1cm. In addition there are some thicker sandstone layers (cf. facies 2) which generally have planar tops and bottoms, although the layer at 94.6m has a wave-rippled top locally and the layer at 102.2m is wave-ripple cross-laminated throughout its thickness. Three of the layers (at 98.9, 100.6 and 102.4m on the log) contain finely comminuted shell debris and rounded vein quartz granules, the first two layers exhibiting erosive bases. Near the top of facies 5 a thicker (15cm) sheet sandstone occurs (plate 5.6C) which contains a layer of symmetric convolute lamination which dies out both upwards and downwards within the unit.

5.2.5.2 Biofacies

The presence of a good fissility in the facies 5 mudstones is interpreted to reflect the absence of bioturbation. A similar relationship between shale fissility and bioturbation was noted by Byers (1974) in a study of the Late Devonian Sonyea Group of New York State and the Upper Cretaceous Pierre Shale of the Western Interior Seaway of North America. Byers concluded that shales which have not been biogenically disturbed retain their original horizontal fabric, imparting a good fissility to the shale. The horizontal fabric may either be the product of a lamination resulting from variations in sediment supply and/or sediment type, or the horizontal alignment of platy clay grains and carbonaceous detritus. Bioturbation, however, resulted in a randomisation of the horizontal fabric with a concomitant loss of shale fissility.

The absence of biogenic structures and the horizontally-laminated character of this facies suggests that facies 5 should be assigned to Schäfer's letal-pantostrate biofacies.

5.2.5.3 Interpretation

By analogy with facies 1, the predominance of graded rhythmites and horizontally-laminated sandstone streaks indicates that facies 5 was deposited below, and at some levels close to, storm-wave base; the local development of a wave-ripples (at 94.6 and 102.2m on the log) indicates that exceptional storm waves touched the facies 5 substrate. Although the bulk of the sand content of facies 5 was deposited from suspension the thicker sandstone sheets, and the scoured bases to the layers containing comminuted shell debris, indicate that the substrate was occasionally swept by currents strong enough to transport sand and granule-grade material as bedload. The most striking characteristic of facies 5, however, is the absence of bioturbation. Both Seilacher (1964) and Rhoads (1975) have noted that the absence of trace fossils in shallow marine mudstones is a useful indicator that the muds were deposited in anoxic conditions. Taken in conjunction with the relatively large number of *Chondrites* burrows found in the upper part of the west Crock Point succession that encloses facies 5 (suggesting that dysaerobic conditions existed below the sediment/water interface - see appendix B) it is proposed that conditions below the facies 5 substrate were anoxic. An explanation invoking a 'soupy' substrate which could not support an infauna may be ruled out on the grounds that the sandstone layers have sharp bases and do not exhibit load structures.

It should be noted at this point that Knight (1990a) sampled facies 5 and reported a paucity of palynomorphs at this and stratigraphically adjacent levels. In reference to the interpretation of facies 5 as representing anoxic bottom conditions Knight observed that “. . . from a palynological view point, the lack of amorphous organic matter within this section does not substantiate such a claim” (P.406).

Anoxic black laminated shales / mudstones have been recorded by many authors from the Devonian (House 1983, e.g. Gutschick and Wuellner 1983). The most intensively studied successions are the black shale tongues in classic Devonian succession in New York State. House (*op. cit.*) and Johnson *et al.* (1985) have used the black shale incursions as indicators of rising sea-level which caused basinal anoxic mud conditions to onlap across continental shelf sequences. The correlation of the New York State black shales with similar sequences in Belgium, Germany and European Russia has enabled qualitative eustatic sea-level curves to be constructed for the Devonian (see figure 7 of House 1983 and figure 12 of Johnson *et al.* 1985 - the latter is reproduced in text-figure 8.4 herein). During periods of transgressive onlap sediment became trapped at the contemporary Devonian shoreline (Goldring & Langenstrassen) which enabled the anoxic muddy sequences to become established on continental shelves (Brett & Baird 1985). McGhee & Bayer (1985) produced simple quantitative models which demonstrated that transgressive-regressive episodes due to changes in sea-level, when superimposed on rapid tectonic subsidence, will tend to enhance the preserved record produced by stratified anoxic basin waters moving up on to shelf during onlap. The major Devonian black shale cycles were attributed to a eustatic sea-level origin, whilst the smaller-scale cycles were attributed to localised basin subsidence which had a tectonically-driven origin.

Wetzel (1982) reviewed the origin of laminated black shale / mudstone sequences and proposed a four-level hierarchical scheme with suggested mechanisms for black shale formation at each scale. Facies 5 appears to correspond to Wetzel's 'macro-scale' where: "One unit may represent hundred (*sic*) to about a million years" (p.433). Wetzel attributed macro-scale variations to: (i) Thermohaline stratification, normally requiring a restricted/silled basin. With this mechanism: "An anoxic period is commonly terminated by increasing water circulation or tectonic movements changing the configuration of the basin and thus increasing water circulation" (p.434). (ii) High volumes of organic matter being supplied to a basin (upwelling). (iii) Periods of high sea-level resulting in high organic productivity in coastal waters and anoxia at depth. (iv) Sediment-starved lakes. Mechanism (ii) can be ruled out based on the observation of Knight (1990a) that facies 5 does not contain amorphous organic matter. There is no evidence in coeval deposits in the Valley of Rocks (see

enclosure 2) to support mechanism (iii) where high organic activity in coastal waters would be expected. Clearly the sediment starved lake mechanism (iv) does not apply to the Lynton Formation.

In summary, facies 5 is interpreted as preserving an anoxic event which formed during a period of restricted circulation in the 'Exmoor Basin'. Although the presence of a few erosively based sandstone layers and wave-ripples may appear to not be consistent with an anoxic interpretation for facies 5 Wetzel (*op. cit.*) noted that bottom current events giving rise to shell stringers, ripples etc. are more frequent than was previously assumed in anoxic shale deposits.

5.3 INTERPRETATION OF THE 'UPPER-MIDDLE MEGA-FACIES SEQUENCE'

The suite of primary sedimentary structures and biogenic traces preserved within the west Crock Point sequence records a transition from open shelf conditions with a *Cruziana* ichnofacies, similar to the suite of structures and traces found in the underlying 'lower-middle mega-facies', through to more distal enclosed shelf conditions with a *Zoophycos* ichnofacies. The trace fossil suite indicates that the transition to more distal conditions was accompanied by a decrease in the pore water oxygen content below the sediment/water interface, with dysaerobic conditions being widely developed. The decrease in the pore water oxygen content reached a peak 4/5^{ths} of the way up the west Crock Point sequence where the thick facies 5 mudstone development indicates that anoxic bottom conditions persisted for a considerable period of time.

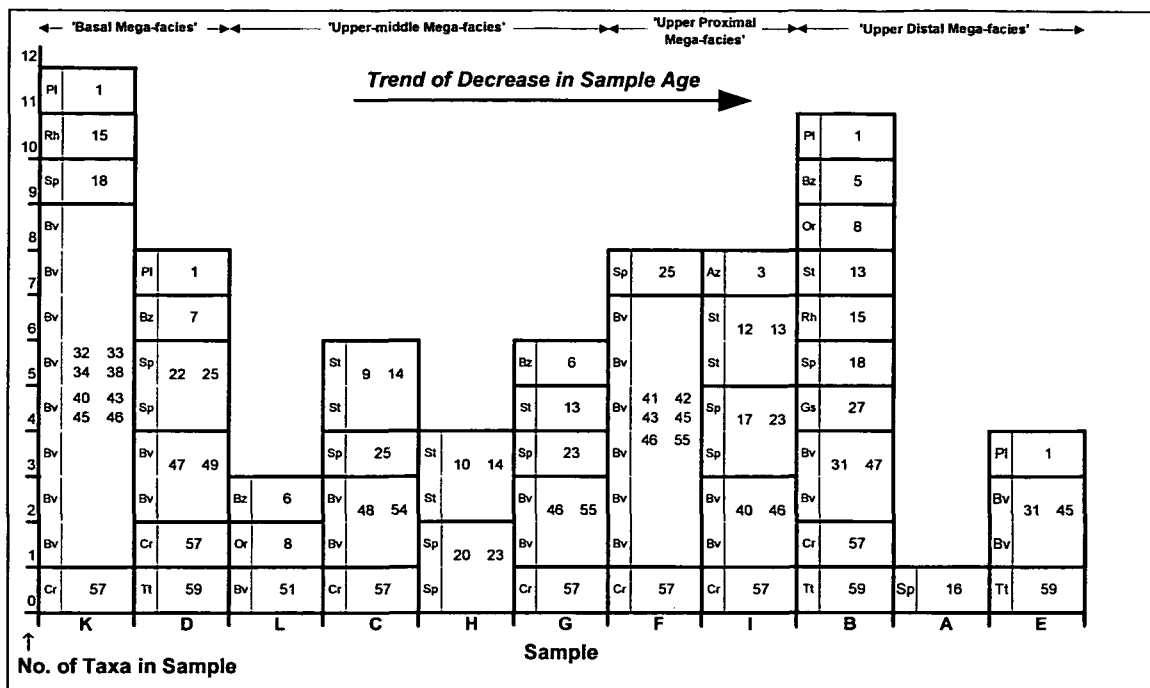
A range of models are now available to test the above interpretation of the 'upper-middle mega-facies' sequence, in particular, the interpreted changes in pore water oxygen content with time. Early models for oxygen deficient marine biofacies (Rhoads & Morse 1971, Byers 1977, Savrda *et al.* 1984, Thompson *et al.* 1985) explored the influence of dissolved oxygen concentration on the calcareous shelly fauna. Rhoads and Morse initially defined three oxygen-related biofacies: *aerobic* (dissolved oxygen > 1.0ml per litre), *dysaerobic* (dissolved oxygen 0.1 - 1.0ml per litre¹), *anaerobic* (dissolved oxygen < 0.1ml per litre). The oxygen-related facies were established on the basis that heavily calcified organisms are only found in waters where dissolved oxygen \geq 1.0ml per litre, whilst polychaetes, nematodes and crustaceans can tolerate dissolved oxygen in the range 0.1 - 1.0ml per litre. Rhoads & Morse concluded that: "... low oxygen marine

¹ Thompson *et al.* (1985) preferred a narrower range, 0.1 - 0.3ml per litre, to define the dysaerobic zone based on extensive box core analyses of the central Californian continental slope. Savrda & Bottjer (1991) observed that the data of Thompson *et al.* was flawed and recommended that the 0.1 - 1.0 ml per litre range should be retained.

basins are dominated by a low diversity assemblage of small soft-bodied organisms with a high surface area to volume ratio. Many of these organisms are deposit-feeders, this trophic group being well adapted to low oxygen sulphide-rich environments" (p.419) - this description closely matches much of the 'upper-middle mega-facies'.

In order to test the 'upper-middle mega-facies' sequence against the Rhoads and Morse model the diversity of the macro-fossil assemblages collected during the B.G.S. re-mapping of the Ilfracombe sheet was plotted as a histogram - the result is shown in text-figure 5.1. The histogram demonstrates that the macro-fossil content of the 'upper-middle mega-facies' is somewhat less diverse than those collected from elsewhere in the Lynton Formation. Although the number of samples are limited the result is consistent with an interpretation for the 'upper-middle mega-facies' as having had a lower pore-water dissolved oxygen content relative to other mega-facies in the Lynton Formation. N.B. Samples A and E were collected from near the Lynton Formation - Hangman Sandstone Group boundary where the encroaching littoral conditions (see chapters 6 and 7) would be expected to have resulted in low-diversity macro-fossil assemblages.

More recently, research has turned towards trace fossil assemblages in reconstructing the oxygenation histories of ancient marine bottom waters as they appear to provide more sensitive indices than the calcareous fauna used by earlier workers (Savrda & Bottjer 1986, 1989, 1991, Easthouse & Driese 1988, Wignall 1991). In particular, the presence of the trace fossil *Chondrites* appears to provide a useful pointer to dysaerobic bottom conditions (Seilacher 1978, Bromley & Ekdale 1984, Ekdale 1985, Vossler & Pemberton 1988, Wignall 1991).



Text-fig. 5.1 Macro-fauna & -flora age distribution and diversity through the Lynton Formation

The histogram shows the relative ages of the fossil localities sampled by Edmonds *et al.* (1985) and the faunal/floral diversity for each sample. Data taken from table 1 of Edmonds *et al.* Note that the horizontal axis does not imply a scale divided into equal units of time.

Key:

Taxa numbered according to the scheme used in table 1 of Edmonds *et al.* (1985). N.B. Only taxa recorded from the localities listed below are included in this key. For this reason the numbering is not contiguous. Characters in parantheses after each phylum / class correspond to the abbreviation used on the graph. **Plantae (Pl):** 1 - Fragments indet. **Anthozoa (Az):** 3 - solitary coral indet. **Bryozoa (Bz):** 4 - *Fenestella*, 5 - *Polypora*, 6 - fenestellid indet., 7 - bryozoan indet. **Brachiopoda:** ORTHIDA (Or): 8 - *Platyorthis longisulcata*, STROPHOMENIDA (St): 9 - *?Chonetes sarcinulatus*, 10 - *C. sordidus*, 12 - "*orthotetes hipponyx*" of Whidborne (1910), 13 - chonetacean indet., 14 - strophomenoid indet. RHYNCHONELLIDA (Rh): 15 - rynchonellacean indet. SPIRIFERIDA (Sp): 16 - cf. *Alatiformia alatiformis*, 17 - *Brachyspirifer?*, 18 - cf. *Eurispirifer paradoxus*, 20 - *Mauispirifer?*, 22 - *Retzia*, 23 - *Subcspidella lateinisa*, 25 - spiriferacean indet. **Gastropoda (Gs):** 27 - *Bucanella?* **Bivalvia (Bv):** 31 - *Carydium sociale*, 32 - *Cimitaria cf. acutirostris*, 33 - *Cypricardella cf. rhomboidalis*, 34 - *Cypricardinia*, 38 - '*Leptodomus*' cf. *arcuata*, 39 - *Modiomorpha modiola*, 41 - '*Myophoria*', 42 - *Myilarca?* 43 - '*Nuculana*', 44 - *Nuculoidea ?trigona*, 45 - *Palaeoneilo maureri*, 46 - *P. cf. primaeva*, 47 - *Paracyclas rugosa*, 48 - *?Prothyris plicata*, 49 - *P.?* 51 - *Ptychopteria (Actinopteria) fasciculata*, 52 - *P. (A.) sp. ?nov.*, 53 - *P. (A.) cf. musicata*, 54 - *?P. (A.)*, 55 - *Tancrediopsis aff. subcontracta* **Crinoidea (Cr):** 57 - columnals **Tentaculitoidea (Tt):** 59 - *Tentaculites*.

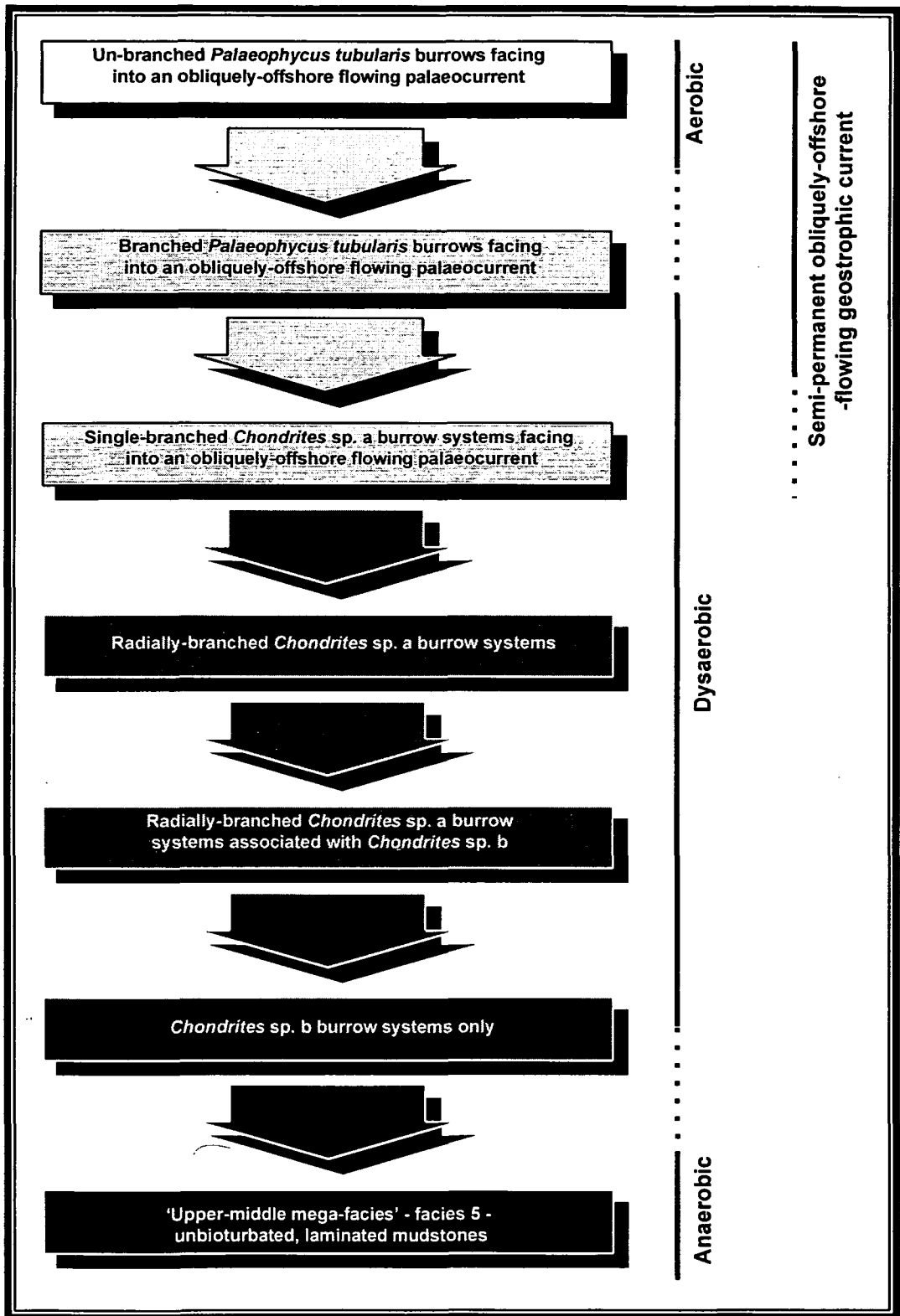
Sites:

Sites and grid references listed in table 1 of Edmonds *et al.* (1985). In order to plot the relative age sequence of the localities only sites that could be assigned to an approximate level (and therefore mega-facies) in the Lynton Formation are included - for this reason the lettering is not contiguous. **A** - Heddon's Mouth, Trentishoe; cliff face (6533 4961); **B** - exposures by and in floor of track to Woody Bay, Martinhoe (6764 4907 to 6767 4904); **C** - Lee Bay, Martinhoe, reef (6937 4919); **D** - Wringcliff Bay, Lynton, reef (7013 4975); **E** - road cutting WNW of Barbrook (7060 4794), **F** - the Valley of Rocks, Lynton, SW side (7029 4939), **G** - the Valley of Rocks, Lynton, south side (7065 4955), **H** - the Valley of Rocks, Lynton, north side (7097 4993), **I** - Hollerday Hill, Lynton, NW flank (7118 4988), **K** - reef NW of Hollerday Hill, Lynton (7148 5005), **L** - road cutting, Myrtleberry Cleave, Lynton (7375 4871).

Bioturbation in modern substrates may be divided into three levels: a surface mixed layer (not normally preserved due to low shear strength of the homogenised sediment and subsequent overprinting by the transitional layer), a transitional layer produced by deeper burrowers and a historical layer below the level where new bioturbation occurs (Ekdale *et al.* 1984). Much of the new work relies on the recognition of trace fossil tiering which can be established by observing cross-cutting relationships in the preserved ichnofauna of the historical layer (Wetzel 1983, Ekdale 1985, Savrda & Bottjer 1986). Unfortunately the 'upper-middle mega-facies' is dominated by the *Palaeophycus tubularis* - *Chondrites* sp. a continuum - other traces are relatively infrequent. As a result it did not prove possible to establish cross-cutting relationships or, therefore, unambiguous tiering within the ichnofossil assemblages. Thus, the construction of palaeo-oxygenation curves using the 'oxygen-related ichnocoenosis' model for 'distinctly burrowed beds' could not be carried out for the 'upper-middle mega-facies'. The dominance of the *Palaeophycus tubularis* - *Chondrites* sp. a continuum, however, may of itself be taken to be an indicator of reduced levels of dissolved oxygen in the 'upper-middle mega-facies' substrate. The most likely candidates for tiering is the association of *Chondrites* sp. a and b. As discussed in appendix B, the association of *Chondrites* sp. b with *Chondrites* sp. a tends to occur at the radially branching *Chondrites* sp. a end (i.e. most dysaerobic) of the *Palaeophycus tubularis* - *Chondrites* sp. a continuum. The smaller *Chondrites* sp. b burrows persisted after the *Chondrites* sp. a burrows had been excluded, indicating a reduction in dissolved oxygen levels within the bottom waters, although no examples of *Chondrites* sp. b burrows cutting across *Chondrites* sp. a burrows were observed. This continuum has allowed a palaeo-oxygenation model for the Lynton Formation to be constructed - see text-figure 5.2.

Savrda and Bottjer (1991) proposed a scheme for sub-dividing the anaerobic biofacies:

- Anaerobic*: containing no micro-bioturbation or autochthonous microbenthic body fossils.
- Quasi-anaerobic*: containing micro-bioturbation and/or autochthonous microbenthic body fossils (e.g. benthic foraminifera and scolecodonts).
- Exaerobic*: containing *in situ* epibenthic macroinvertebrate body fossils (e.g. bivalves, molluscs and gastropods).



Text-fig. 5.2 Palaeo-oxygenation model for the Lynton Formation - esp. the 'upper-middle mega-facies'.

Of the 238 palynological samples collected by Knight (1990a) from the Lynton Formation through to Morte Slates succession 43% contained scolecodonts. Knight observed that: "Strata considered to be more distal in aspect (e.g. the Morte Slates, or the most basinwards facies of the Lynton Formation - exposed west of Crock Point, Pound *pers. comm.*) proved notably devoid of scolecodonts" (p.406). This observation strongly supports the assignment of facies 5 to the revised *anaerobic biofacies* of Savrda and Bottjer (1991).

The lack of cross-cutting relationships, taken together with the 'bed-junction' ('piped') mode of preservation of *Palaeophycus tubularis* - *Chondrites* sp. a burrows in the heterolithic deposits of facies 1, closely matches Savrda & Bottjer's (1991) *short-term event* type within their 'indistinctly burrowed bed' model. This suggests that the dissolved oxygen content of the burrowed horizons within the 'upper-middle mega-facies' fluctuated over relatively short time spans; the dissolved oxygen content of the substrate may have been close to the dysaerobic / anaerobic boundary for much of the time and an infauna could only become established when a storm event (of the type that introduced the sandstone layers into the facies 1 environment) re-oxygenated the bottom waters and allowed a burrowing infauna to become established. A subsequent storm event introducing sand would have enabled the burrows to be preserved as sand was 'pumped' into the open burrow systems.

Evans (1980) observed that the Lynton Formation brachiopod fauna was impoverished when compared with the Meadfoot Group which had a much stronger resemblance to that of the 'typical' Rhenish magnafacies. Evans attributed this phenomenon to be a reflection of high rates of terrigenous influx or fresh water from fluvial input. Given the ephemeral nature of the fluvial systems developed on the contemporaneous coastal plain (Tunbidge 1978, 1981b, 1983a, 1984) the latter factor can be ruled out. Although the Lynton Formation accumulated during a period of high argillaceous input (see section 8.3) the Lynton Formation was relatively starved in respect to arenaceous input; there is no reason to suggest that brachiopod diversity would have been limited by large volumes of argillaceous input. It is proposed, therefore, that the impoverished brachiopod fauna of the Lynton Formation was caused by widespread dysaerobic bottom conditions.

Although the west Crock Point sequence exhibits a general upwards trend towards substrate deoxygenation, reaching a peak in the facies 5 anoxic mudstones 4/5th of the way up the sequence, the observation made in

section 5.2.2.1 should be reiterated at this point: namely, that that the facies 2 thinly-bedded sheet sandstones do not indicate any proximality trends similar to those described by Aigner and Reineck (1982).

The long-term deoxygenation trend through the west Crock Point succession occurred during a period of eustatic sea-level rise that intervened between transgressive events Ic and Id of Johnson *et al.* (see text-figure 8.4). A high-frequency pattern of re-oxygenation events was superimposed on this long-term trend as storm-generated currents re-introduced oxygenated bottom waters. Both the onset and termination of the conditions that gave rise to the facies 5 anoxic mudstone was, however, relatively abrupt. In section 5.2.5.3 it was concluded that facies 5 conditions were probably initiated by thermohaline stratification generated by tectonic movements creating a restricted circulation in the 'Exmoor Basin'. Rhoads and Morse (1971) suggested two reasons for stagnant basin conditions developing:

- surface mixing by waves becoming shallower due to smaller wave heights which results from the fetch of a basin being decreased due to e.g. the emergence of a barrier.

- movement of bottom waters becoming impeded by sills suppressing lateral mixing with oceanic waters

Given the tectonic setting of the 'Exmoor Basin' (see section 2.2.4) the following scenario can be envisaged to account for the restricted circulation that gave rise to the deposition of facies 5. Foot-wall uplift of the basin floor at the southern margin of the Lynton Formation resulted in the formation of a sill/lip that served to either (i) restrict circulation in the basin, or (ii) caused deep water waves to shoal and reform, thus reducing the effective fetch of the 'Exmoor Basin' (see text-figure 2.8). Later fault movements resulted in a further change to the basin configuration re-established a deep basin circulation, accounting for the relatively abrupt cessation of facies 5 conditions.

6. THE TRANSITION TO THE HANGMAN SANDSTONE GROUP ADJACENT TO THE LYNMOUTH - EAST LYN FAULT: THE UPPER PROXIMAL MEGA-FACIES

6.1 INTRODUCTION

As was described in section 1.7, the top of the Lynton Formation appears to be diachronous, implying a southerly / basinward progradation of the palaeo-shoreline over time (see enclosure 2). The oldest Lynton Formation - Hangman Sandstone Group transitional sequences occur adjacent to the Lynmouth-East Lyn Fault below the crest of outliers, unrecognised on BGS Sheet 277 (Ilfracombe), of the Hangman Sandstone Group at: Hollerday Hill, The Tors west of Wind Hill and Summerhouse Hill. Sequences through the boundary are exposed on the WNW flank of Hollerday Hill above the Valley of Rocks (7118 4988), north spur off the eastern end of South Cleave (7095 4958), The Tors on the western flank of Wind Hill (7304 4942), the crag above the Glen Lyn Gorge (7228 4905) and Oxen Tor (7288 4899). The latter sequence was logged owing to the relatively good quality of exposure in comparison to the other, generally heavily lichen-covered, outcrops of equivalent age. The southerly-thickening wedge of younger Lynton Formation deposits and the transition to the Hangman Sandstone Group, distal to the Lynmouth - East Lyn Fault, is described in chapter 7. A comparison of the 'upper proximal mega-facies' and 'upper distal mega-facies', and the succeeding basal Hangman Sandstone Group sequences, is discussed in section 8.2.

6.2 OXEN TOR SEQUENCE

At Oxen Tor a 25.2m thick nearly complete coarsening-upwards section through the Lynton Formation - Hangman Sandstone Group transition is exposed in a rib that runs down the steep hill-side immediately below a sharp kink in the lower path which contours the sharp break in slope below the crest of Oxen Tor. Cosets of well sorted cross-bedded quartz sub-litharenites, weathering white / light grey, mark the top of the logged sequence (enclosure 9) which occurs at path-level. This locality serves as a boundary-hypostratotype for the Lynton Formation - Hangman Sandstone Group transition (see section 1.8.3), the precise boundary being taken at 9.85m on the log at the point where thinly interlayered sandstone/mudstone bedding with thin sandstone beds is sharply overlain by a sandstone sequence comprising cosets of cross-bedded sandstone and wave-ripple cross-laminations i.e. where a 17cm thick unit of strongly bioturbated thinly interlayered sandstone/mudstone unit is overlain by a 3cm thick parallel-laminated sandstone with a wave-rippled top.

Seven facies are defined at Oxen Tor; they are described in the following paragraphs, along with a contrasting parallel-laminated sandstone facies observed on the WNW flank of Hollerday Hill above the Valley of Rocks. A facies sequence analysis of the Oxen Tor sequence is then discussed, prior to the definition of a depositional model to describe the Lynton Formation - Hangman Sandstone Group transition proximal to the Lynmouth - East Lyn Fault and a comparison with both Recent and geological examples published for similar sequences.

6.2.1 Facies A - Thinly Interlayered Sandstone/Mudstone Bedding

Thinly interlayered sandstone/mudstone bedding dominates the lower 10m of the logged sequence, occasionally being interrupted by thin-bedded sandstones (facies B) or isolated units of cross-bedding (facies C). Within facies A primary sedimentary structures are difficult to discern due to heavy lichen cover, extensive biogenic 'churning' and penetration by *Palaeophycus tubularis* burrows, the latter generally occurring in units with a higher sandstone content. Bioturbation increases upwards within the sequence and can result in the total overprinting of primary sedimentary structures locally. The facies comprises almost exclusively of unconnected and connected sandstone lenses (sub-types C & D of facies 1 described in section 4.2.1) of very fine to fine, occasionally medium, grade sandstone. Isolated sigmoid-shape unconnected sandstone lenses off-shooting from an underlying horizon of connected lenses, of the type figured in text-figure 4.7, occur occasionally. Individual units vary from 2cm to 1.2m in thickness; unit boundaries are either taken at points of abrupt change in gross sandstone content or are marked by laterally extensive (across the available outcrop width c. 5 - 10m) planar erosion surfaces with up to 2 cm relief e.g. 0.45m on the log. Units occasionally exhibit a thin coarsening-upwards cap of wave / combined flow interference ripples e.g. 4.1m on the log. No coarsening-upwards sequences of the type described at e.g. Duty Point (facies 1') were observed.

Facies A represents the migration of ripples below fair-weather wave-base during periods of increased wave / combined flow activity, interrupting appreciable periods of mud deposition below fair-weather wave-base. The laterally extensive planar erosion surfaces preserve the passage of high-energy events creating a zone of nett erosion, whilst units exhibiting a thin coarsening-upwards cap of wave / combined flow interference ripples represent the winnowing of fines by slightly lower energy events which were still of sufficient energy

to ripple the substrate. The upward increase in the combined feeding/dwelling *P. tubularis* burrows may represent an upward increase in environmental energy / agitation of the substrate.

6.2.2 Facies B - Thinly-bedded Sheet Sandstones

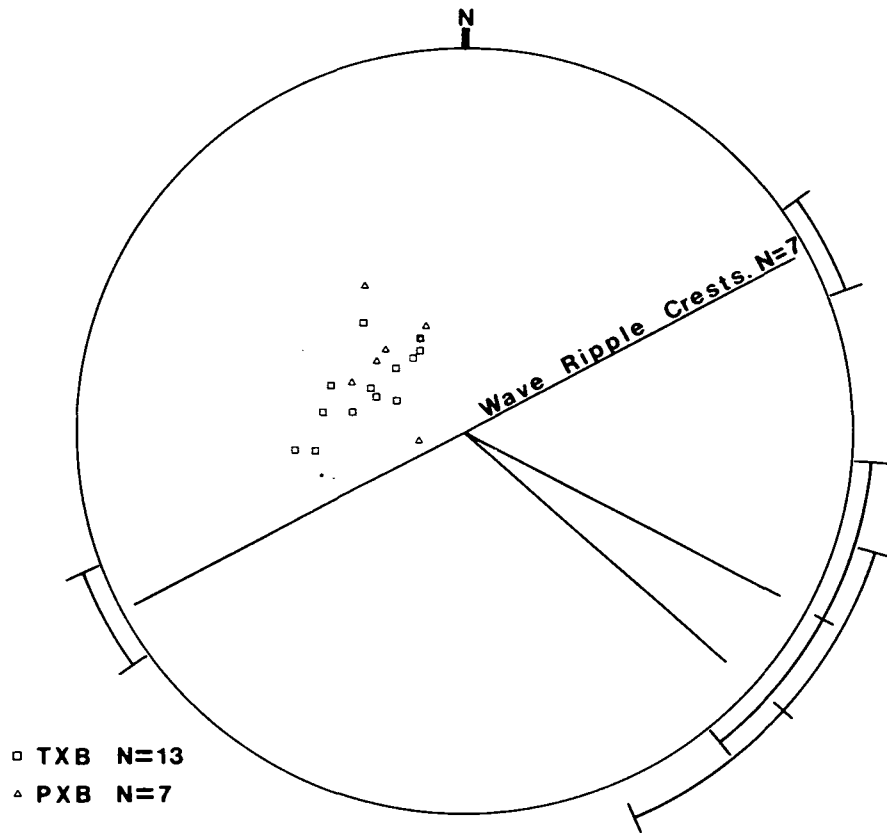
Facies B comprises 2.5 to 10cm thick (4.5cm average) beds of sheet sandstones, very fine to fine in grade, which are interbedded with facies 1 (connected sandstone lens sub-type); beds are laterally persistent over the width of the outcrop. Beds of facies 2 are rare in the lower part of the sequence but become abundant in the 1m or so beneath the Lynton Formation / Hangman Sandstone Group boundary (8.68 - 9.85m on the log), at which point the sequence becomes significantly sandier. Bed bases are planar and generally erosive, but scours up to several centimetres in depth are occasionally observed (e.g. 8.68cm on the log), or gradational; in the latter case the bed base is recognised by a very abrupt increase in sand content. Bed tops are either gradational or marked by ripples of a wave or combined flow / interference ripple origin, the latter corresponding to the three-dimensional hummocks described in section 4.7.1; bed tops are frequently penetrated by indistinct biogenic structures. Internally, the lower part of a bed can occasionally comprise a thin zone of massive sandstone draping and preserving the wave rippled top of the underlying bed (9.18 & 9.86m on the log) to give an amalgamated facies B unit. The bulk of the beds comprise a thick zone of parallel to gently undulatory laminae, the undulations having a wavelength of several decimetres and an amplitude of several centimetres. The uppermost part of the majority of beds is characterised by a lamination style diagnostic of an oscillatory/combined-flow origin. Biogenic 'churning' can obliterate primary lamination locally.

Facies B matches facies 3 described in chapter 6 (section 6.5), to which the reader is directed for a detailed discussion and interpretation. In summary, facies B is the product of the combined action of storm waves and associated obliquely-offshore-directed currents generated by on-shore directed winds. Initial storm conditions resulted in erosion, followed by net deposition during steady flow conditions at the storm peak. Massive sandstones at the bed base, preserving ripple profiles at the top of the underlying unit, represent fall-out from suspension from a sediment cloud that had reached capacity to entrain sediment and therefore did not have the flow-power available to erode the underlying substrate, thereby effectively 'armouring' the initial bed topography. Finally, the top of the bed is wave-current rippled as the storm waned. The top of the bed was subsequently reworked by bioturbation or oscillatory currents.

6.2.3 Facies C - Isolated Beds of Trough Cross-bedding

Facies C beds comprise single sets or cosets (maximum of 5 sets per coset) of trough cross-bedding between 8 and 45cm in thickness; individual sets are 9.6cm thick on average ($n = 28$, $\sigma_{n-1} = 2.3$, min. = 5cm, max. = 16cm). In the lower part of the sequence facies C is interbedded with facies A, whilst higher in the sequence there is a tendency for facies C to be interbedded with facies F or, more occasionally, facies E. Beds are well sorted and are of medium quartz-rich sandstone grade, weathering white/grey, although the mud and lithic content is higher than facies G. Set boundaries are planar, frequently exhibiting localised shallow irregular scours, or drape and preserve wave-ripple profiles at the top of the preceding set. Facies C set boundaries contrast with the undulatory/scooped surfaces that separate facies G cross-bedded sets. Internally, foreset laminae are 1 to 3mm thick and approach the lower set boundary asymptotically. Set tops are either erosionally truncated by the succeeding cross-bedded set, rippled (wave or combined flow / interference ripple origin) or biogenically 'churned', bioturbation always descending from the upper set surface.

Text-figure 6.1 shows cross-bedding dip directions, split into trough and planar cross-bedding, summed for occurrences in facies C, D and G at Oxen Tor, along with the crest trend of wave-ripples. The summing of dip directions across facies is reasonable in that there is no significant change in cross-bedding dip direction between the facies containing cross-bedding - see palaeocurrent dip arrows shown on the log column for Oxen Tor (enclosure 9). Foreset dip directions indicate bedform migration towards the ESE, a direction sub-parallel to the strike of the palaeoslope and trend of the Lynmouth - East Lyn Fault. The palaeocurrents, corrected in relation to palaeomagnetic rotation, are shown on text-figure 4.9, along with the trend of the Lynmouth - East Lyn Fault, palaeoslope dip and inferred palaeo-wind directions.



| | |
|-------------------------------------------------|----------|
| Mean Vector Direction: TXB Trough Cross-Bedding | 119° |
| PXB Planar Cross-Bedding | 132° |
| Wave Ripple Crests | 062-242° |

| | |
|-------------------------------------------|-------|
| Estimate of Spread of Angular Values: TXB | 91.9% |
| PXB | 90.3% |
| Wave Ripple Crests | 96.6% |

| | |
|-----------------------------|------|
| Mean Angular Deviation: TXB | ±23° |
| PXB | ±25° |
| Wave Ripple Crests | ±7° |

| | |
|------------------------------------|------------------------------|
| Rayleigh Test of Significance: TXB | <10 ⁻³ non-random |
|------------------------------------|------------------------------|

Text-fig. 6.1 Oxen Tor Palaeocurrents

The palaeocurrents, corrected in relation to palaeomagnetic rotation, are shown on text-figure 4.9

The trough cross-bedding represents the migration of sinuous-crested dunes (see section 4.2.4.3 for discussion of this bedform type) by either unidirectional or pseudo-unidirectional (i.e. strongly asymmetrical oscillatory flow produced by shoaling waves cf. Clifton 1976) currents; there is no evidence for tidal activity in the upper Lynton Formation (see section 7.7). The style of facies C cross-bedding is strongly similar to facies 5 at Woody Bay west (see section 7.7) and a similar hydrodynamic origin is invoked here i.e. facies 5 is the product of the obliquely-longshore migration of a duned sand patch under the influence of storm-generated geostrophic flow, an interpretation consistent with observed palaeocurrent indicators. As the storm waned oscillatory currents would become dominant relative to the ebbing geostrophic flow; the duned sea-floor would have been moulded by oscillatory currents into a series of wave / combined-flow / interference ripples. The interaction of the ebbing geostrophic flow with oscillatory currents may account for the frequent occurrence of combined-flow / interference ripples. The duned and rippled sand patch would only have been active during periods of storm activity and would have been subject to biogenic colonisation and mud deposition during inter-storm periods.

6.2.4 Facies D - Planar Cross-bedding

Facies D beds are restricted to the lower part of the upper third of the logged sequence where they are interbedded with facies E and F. Beds comprise single sets or cosets (maximum of 3 sets per coset) of planar cross-bedding between 8 and 38cm in thickness; individual sets are 12.1cm thick on average ($n = 8$, $\sigma_{n-1} = 4.8$, min. = 8cm, max. = 22cm). Beds are well sorted and are of medium quartz-rich sandstone grade, weathering white/grey. Set boundaries are either planar or drape and preserve wave-ripple profiles at the top of the preceding set. Internally, foreset laminae are 1 to 3mm thick and approach the lower set boundary asymptotically, or have a thin massive zone at the base where they drape and preserve ripple profiles at the top of the preceding set. Foreset dip directions indicate bedform migration towards the SE, a direction sub-parallel to the strike of the palaeoslope and trend of the Lynmouth - East Lyn Fault, but closer to the mean palaeoslope (SSW-dip) than the ESE-dipping trough cross-bedding. Set tops are either planar, rippled (wave or combined flow / interference ripple origin) or biogenically 'churned', bioturbation always descending from the upper set surface. Bioturbation is less pervasive than in facies C.

Planar cross-bedding is the product of the migration of straight-crested dunes formed in weak, when compared to the flow power required to generate sinuous-crested dunes, unidirectional or pseudo-

unidirectional flows (see section 4.2.4.3 for a detailed discussion of the hydrodynamic significance of these structures). Facies D is set within sandier deposits than facies C, in a zone where wave activity was the dominant force in moulding the substrate (facies E and F). The presence of a structure produced by unidirectional currents in an environment dominated by an oscillatory flow regime suggests that the zone was periodically subjected to either strongly asymmetrical oscillatory flow produced by shoaling waves cf. Clifton (1976) or is the product of the obliquely longshore migration of a duned sand patch under the influence of storm-generated geostrophic flow. The latter explanation is more consistent with observed palaeocurrent indicators. The fact that the palaeocurrent direction of facies D (SE) is closer to the mean palaeoslope dip (SSW) than the trough cross-bedding (ESE) recorded within other facies (see text-figure 4.9) is attributed to be a function of the geostrophic flow origin. In the southern hemisphere the seaward-directed bottom flow set up by the shoreward movement of the upper water column in response to storm wind shear is progressively deflected to the left by the Coriolis force as it accelerates offshore, until the Coriolis force and the pressure gradient force terms balance in the equations of motion. This results in the generation of shore-parallel geostrophic flow (Strahler 1963). The flow power required to form planar cross-bedding is less than that required to form trough cross-bedding i.e. the currents responsible for the generation of the planar cross-bedding would have been less accelerated and would not have been deflected as much as the flow that created the trough cross-bedding. As the storm waned oscillatory currents would have become dominant relative to the ebbing geostrophic flow; the duned sea-floor would have been moulded by oscillatory currents into a series of wave / combined-flow / interference ripples. The interaction of the ebbing geostrophic flow with oscillatory currents may account for the frequent occurrence of combined-flow / interference ripples. The duned and rippled sand patch would only have been active during periods of storm activity and would have been subject to wave reworking, biogenic colonisation or mud deposition during inter-storm periods.

6.2.5 Facies E - Flaser Bedding

Units of this facies comprise a regular alternation of 2 - 3cm thick zones of very fine to fine grade sandstone with a lamination style characteristic of wave / combined flow / interference rippling and thin mudstone layers. The majority of the pinch-and-swell sandstone layers are continuous being separated by 1 to 7mm thick mudstone laminae which either maintain a constant thickness over both ripple crests and troughs or thicken into troughs ("wavy flaser" bedding referred to in section 4.2.6.1); isolated mudstone flasers occur only rarely. This facies is frequently disturbed by the effects of biogenic 'churning', particularly the

mudstone component. The sandstone:mudstone ratio tends to remain relatively constant through the thickness of individual beds, although in one case (c. 17m on the log) the sandstone content increases upwards within a unit. Facies E beds are restricted to the lower part of the upper half of the logged sequence where they are interbedded predominantly with facies D and F and more occasionally with facies C and G. Beds range between 2 and 34cm in thickness, averaging 14.4cm ($n = 9$, $\sigma_{n-1} = 10.5$) and have bases which are either planar or preserve the rippled profile at the top of the underlying bed; bed tops are either rippled or erosively truncated by the overlying unit.

The genesis and hydrodynamic significance of flaser bedding is discussed in section 4.2.7.3, to which the reader is referred for detail. In summary, facies E represents an environment where conditions allowing the preservation of sand were more favourable than those allowing the preservation of mud. Relatively frequent periods of increased wave energy would have resulted in the mobilisation of the rippled substrate and the winnowing of mud. The lack of intraformational mudstone chips in the sandstone layers suggests either that a significant period elapsed between periods of ripple mobilisation, allowing consolidation of the basal mud layer which would have effectively 'armoured' the underlying rippled surface (Hawley 1982).

6.2.6 Facies F - Bipolar Ripple Cross-lamination

Facies F displays many of the characteristics present in facies E, but lacks mudstone laminae and flasers. Cosets of well sorted very fine to fine grade quartzose sandstone, weathering white to light grey, range between 4 and 39cm in thickness, averaging 13.0cm ($n = 20$, $\sigma_{n-1} = 7.7$). Boundaries between cosets are either planar and erosive or a thin zone of massive sandstone drapes and preserves the underlying wave / combined-flow / interference rippled topography. The few exposed plan views of wave-ripples indicate an ENE-WSW crestline trend (text-figure 6.1) i.e. a direction slightly oblique in relation to palaeoslope strike, but one that would have been normal (allowing for palaeomagnetic rotation) to the propagation direction of surface waves in respect to a SE trade wind (see text-figure 4.9). Internally, facies F comprises 0.5 to 2cm thick sets of complexly interwoven cross-laminae with opposing dip directions and a suite of characteristics diagnostic of generation by waves (Boersma in: de Raaf *et al.* 1977). In some cases wave-ripple crests are repeatedly truncated within a coset e.g. unit at 18.6m on the log. Towards the upper part of the sequence containing facies F thin (3 to 14cm thick) zones of clean, well sorted parallel-laminated sandstone occur, erosionally truncating underlying ripple crests; individual laminae are 1 to 2mm in thickness. Hand-lens

examination of fresh surfaces indicated that heavy minerals are concentrated within laminae, although thin sections were not taken to establish the precise mineral assemblage. No evidence of primary current lineation was observed, although there is a paucity of bedding plane views of this facies F variant. Primary lamination within facies F is occasionally locally disrupted by biogenic 'churning'.

Facies F bears a strong resemblance to facies 7 at Little Burland (see section 7.9) - the facies interpretation presented below follows that for facies 7 at Little Burland. The basic lamination style is the product of prolonged periods of sustained oscillatory action at or above fair-weather wave-base. Localised patches of biogenic 'churning' reflect a well oxygenated substrate that could support an infauna, although the predominance of primary physical structures over biogenic traces suggests that the substrate was frequently mobilised by current activity. Taken in their overall context the thin parallel-laminated zones are interpreted to be the product of oscillatory sheet flow formed during periods of increased storm-wave activity cf. Roep *et al.* (1979).

6.2.7 Facies G - Trough Cross-bedded Cosets

The upper quarter of the Oxen Tor sequence is dominated by thick cosets of well sorted, trough cross-bedded cosets of quartzose medium grade sandstone. Cosets range from 15 to 100cm in thickness, averaging 61.3cm ($n = 8$, $\sigma_{n-1} = 30.3$) and are bounded by planar erosion surfaces which can plane-off the wave-rippled upper surface of the preceding unit. Individual sets average 6 to 7cm in thickness, reaching a maximum of 12cm; sets tend to maintain a constant thickness up/down palaeocurrent direction. Individual foresets are concave-up and 'spoon-shaped' approaching the scour trough base asymptotically or, more rarely, passing into planar toesets (plate 4.21B). In vertical sections normal to the palaeocurrent direction the set bases cross-cut to give a series of intersecting scour troughs ('festoon bedding' - plate 4.21B & C). The foreset laminae infilling the scour troughs rest concordantly above the scour trough base, the infill being symmetric. The mean dip direction of facies G foresets indicates a migration direction towards the ESE (text-figure 6.1) i.e. a direction slightly offshore of being parallel to palaeoslope strike (text-figure 4.9). Very occasionally, adjacent sets show opposing foreset dip directions ('herringbone' cross-bedding - plate 6.1A), in which case the WNW-directed (subordinate palaeoflow) sets tend to be thinner (c. 50%) than the adjacent ESE-directed (dominant palaeoflow) sets. Set boundaries are scooped to broadly undulatory, except where 'herringbone' cross-bedding is present and the set boundary tends to be planar (plate 6.1). Set tops very occasionally display a

thin zone of intricately bundled cross-lamination, up to 2 sets in thickness, indicative of wave reworking. There is no evidence of biogenic traces preserved within facies G.

Facies G preserves the migration of sinuous-crested dunes ("three-dimensional dunes" of Allen 1982, vol. A, chap. 8) under faster flow conditions than the straight-crested dunes ("two-dimensional dunes" of Allen *op. cit.*) preserved in facies D planar cross-bedding lower in the Oxen Tor sequence - see section 4.2.4.3 for a full discussion of the hydrodynamic significance of these structures. This suggests that mean environmental energy increased upwards in the Oxen Tor sequence, a feature consistent with the overall coarsening-upwards trend and the sandstones becoming progressively better sorted and less muddy upwards within the sequence. The well sorted, quartzose sandstone lithology of facies G and the absence of mudstone layers / intraclasts and biogenic structures indicates that this facies was deposited in a high-energy environment subjected to relatively constant agitation within fair-weather wave-base; there is no evidence to suggest that facies G was aeri ally exposed at any point during deposition e.g. intertidal indicators such as: 'ladder ripples' (cf. Bajard 1966), swash marks, antidune lenses, rill marks, symmetrical ripples with planed-off crests, flat topped or double-crested wave-ripples, wave-ripples with rounded crests and pointed troughs (Reineck & Singh 1980) or small run-off channels cut at low water on an intertidal surface (cf. Clifton *et al.* 1973). Although true 'herringbone' cross-stratification occasionally occurs in facies G there is a lack of convincing evidence that this facies was deposited by tidal currents e.g. wedge shaped sets, hanging set boundaries, reactivation surfaces, rhythmic tidal foreset bundles (see section 4.2.4.3 for a discussion of the diagnostic characteristics of cross-bedding deposited by tidal currents). The sinuous-crested dunes migrated in an alongshore direction, with a small offshore component. When palaeogeographic and palaeoclimatic considerations are taken into account a shore-parallel pseudo-unidirectional geostrophic flow would have been expected to develop in response to onshore 'piling' of water by trade wind induced wind shear (text-figure 4.9). Brief periods of decreased environmental energy would have allowed oscillatory currents to dominate, resulting in the reworking of dune tops to give thin wave-rippled zones. The planar erosion surfaces bounding cosets evidence periodic, major high-energy events which would have resulted in the duned substrate being 'washed out' and the zone being subject to nett erosion cf. the beach face accretion surfaces produced by storm planation described by van Straaten (1965).

6.2.8 Facies H - Parallel-laminated Sequence - Valley of Rocks

Although not present at Oxen Tor, a sequence of parallel-laminated sandstones, of similar lithology to facies G at Oxen Tor, is present on the WNW flank of Hollerday Hill above the Valley of Rocks (7118 4988). The heavy lichen cover on the rocky rib running up the WNW flank of Hollerday Hill precluded detailed logging of the sequence. Nevertheless, an upwards transition from thinly interlayered sandstone/mudstone bedding (facies A), with an upwards increase in thinly-bedded sheet sandstones (facies B), to facies H parallel-lamination can be discerned where it is interbedded with rare units of facies G. Unfortunately a break in the upper third of the sequence prevents the nature of the transition to facies G and H being observed.

N.B. Several units attributable to the 'Watersmeet lithotype' (see section 4.2.4) occur nearby on the crags below Rugged Jack and Castle Rock (see enclosure 2):

- 1.35m thick unit c. 55m below the top of the Lynton Formation (7063 4982)
- 5 to 10cm thick unit c. 35m below the top of the Lynton Formation (7071 4988)
- 5 to 10cm thick unit c. 25m below the top of the Lynton Formation (7084 4990)
- 5 to 10cm thick unit c. 25m below the top of the Lynton Formation (7041 4973).

The 1.35m thick unit was sampled by Knight (1990a) and yielded a monospecific conodont fauna containing *Icriodus culicellus*, a form commonly associated with crinoidal or coralliferous limestones and weakly agitated to turbulent shallow waters. A similar monospecific fauna was recovered from facies 4 in the A39 road section (see section 4.2.5).

Facies H comprises well sorted fine grade quartzose sandstones organised into 8 to 15cm thick lamina sets of 1 to 2mm thick laminae; lamina sets intersect at a low-angle (<5°) and dip in a palaeo-offshore direction (plate 6.1C). Hand-lens examination of fresh surfaces indicates heavy minerals are concentrated within laminae, although thin sections have not been taken to establish the precise mineral assemblage. No evidence of primary current lineation was observed. Several shallow scours, reaching a maximum of 45cm in width and 6cm in depth (plate 6.1B) are preserved. The scours have a concordant infill, with laminae thickening towards the centre of the scour; several discrete phases of scour infill can occasionally be detected (plate

6.1B). These features resemble examples of 'swaley cross-stratification' (Leckie & Walker 1982a) figured in Brenchley (1985) figure 11. Occasional sinuous convex epirelief traces c. 6mm wide and 8mm wide, 40mm long near-vertical ?*Skolithos* sp. tubes occur in bed tops at certain levels.

Facies H closely resembles facies 8 at Little Burland (see section 7.10), but lacks the thin ripple cross-laminated zones, interpreted to be the product of deposition under oscillatory sheet-flow conditions (Clifton 1976) on the upper shoreface of a storm- and wave-dominated shoreline. As at Little Burland, there is no evidence of intertidal exposure preserved in facies H e.g. intertidal indicators such as: 'ladder ripples' (cf. Bajard 1966), swash marks, antidune lenses, rill marks, symmetrical ripples with planed-off crests, flat topped or double-crested wave-ripples, wave-ripples with rounded crests and pointed troughs (Reineck & Singh 1980) or small run-off channels cut at low water on an intertidal surface (cf. Clifton *et al.* 1973).

6.3 FACIES SEQUENCE ANALYSIS

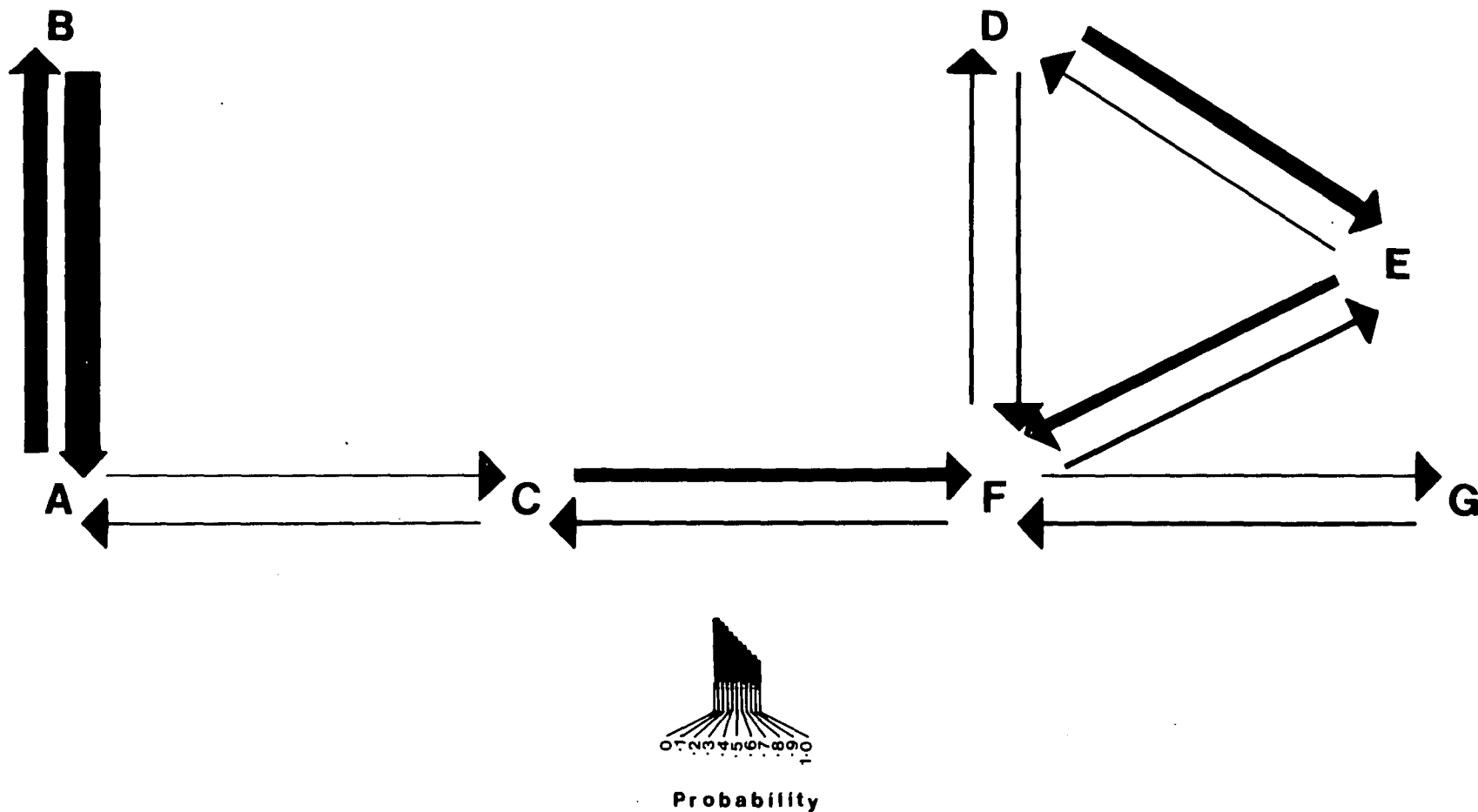
The sequence at Oxen Tor (enclosure 9) has an overall coarsening- and shallowing-upward trend. A complex upward facies transition is observed from thinly interlayered sandstone/mudstone bedding and thinly-bedded sheet sandstones units (offshore deposits) with isolated trough cross-bedded sets becoming more frequent upwards (offshore - shoreface transition), into wave cross-laminated sets and parallel-laminated sandstones (lower to mid shoreface), occasionally displaying planar cross-bedded and flaser bedded units, finally passing into trough cross-bedded cosets (upper shoreface). In order to elucidate the significance of this sequence an observed-minus-random probability analysis was undertaken; the results are presented in text-figure 6.2 where only transitions with a greater than random probability are shown.

Thinly interlayered sandstone/mudstone bedding (facies A) shows a relatively strong probability of transition to thinly-bedded sheet sandstones (facies B) with an even stronger probability of transition back to facies A. This reflects the frequent introduction of beds of storm origin, in an essentially random pattern, into an offshore zone of deposition. Facies A shows a weak tendency of transition to isolated beds of trough cross-bedding and a slightly higher probability of return to facies A type deposits reflecting the rare introduction of facies C beds into the upper parts of the mudstone-dominated sequence at Oxen Tor. When facies C did become established, however, there is a relatively strong probability of transition to bipolar ripple cross-lamination (facies F) and only a weak probability of return to facies C - this phenomenon indicates that once

the coarsening- and shallowing-upwards sequence was initiated there was little possibility of deposition reverting to a deeper, muddier environment. Facies F shows a weak probability of transition to trough cross-bedded cosets (facies G) and a slightly higher probability of transition back to facies F. This feature is probably due to the absence of the continuation of the Oxen Tor sequence and therefore the transition to the facies that would have succeeded facies G not being measured i.e. the logged portion of facies G represents the initial establishment of this facies, in a sequence sense, which would have shown a high tendency to return to the preceding facies F deposits.

Text-figure 6.2 exhibits an interesting variant sequence: bipolar ripple cross-lamination (facies F) \Rightarrow planar cross-bedding (facies D) \Rightarrow flaser bedding (facies E) \Rightarrow return to facies F. There is a weak probability for the initiation of the variant sequence (facies F \Rightarrow D) and an approximately equal probability of return to facies F (although the facies F \Rightarrow D transition is more probable than the facies F \Rightarrow G transition). Once planar cross-bedding had become established, however, there was a strong probability of transition to flaser bedding and thence back to bipolar ripple cross-lamination; the probabilities for the corresponding return F \Rightarrow E and E \Rightarrow D transitions were much weaker. The main transition from bipolar ripple cross-lamination to trough cross-bedding represents the progradation of a relatively high-energy upper shoreface sequence. The variant sequence is attributed to lower energy conditions which allowed the planar cross-bedded event deposit units to be preserved in a wave-influenced environment which favoured the preservation of mud in calmer periods until a return to higher energy conditions is heralded by the bipolar ripple cross-laminated facies.

Facies H is not present at Oxen Tor and could not, therefore, be included in the quantitative observed-minus-random probability analysis. Nevertheless, the stratigraphic position and inferred hydrodynamic position infer an upper shoreface setting that occurred above the facies G environment.



Text-fig. 6.2 Observed-minus-random probability diagrams, for sequences logged at Oxen Tor, showing transitions with a greater than random probability of occurring. Calculations shown at Appendix F. Facies: **A** = thinly interlayered sandstone/mudstone bedding; **B** = thinly-bedded sheet sandstones; **C** = isolated beds of trough cross-bedding; **D** = planar cross-bedding; **E** = flaser bedding; **F** = bipolar ripple cross-lamination; **G** = trough cross-bedded cosets.

The profile of bioturbation intensity shows a gradual upwards increase in the lower mudstone-dominated part of the sequence (offshore to shoreface toe) before decreasing and finally almost disappearing in the overlying sandstone-dominated sequence, representing the progradation of a shoreface over offshore deposits i.e. a transition from Seilacher's (1967a) *Cruziana* to (in facies H) *Skolithos* ichnofacies. The profile is somewhat unusual in that most modern and ancient sequences record an increase in bioturbation away from the toe of the shoreface in offshore deposits (Howard & Reineck 1981 and referenced cited therein). This anomaly may be a feature of the upper proximal facies fauna requiring a relatively low tolerance to oxygen deficiency, preferring the more agitated bottom conditions near the toe of the shoreface than the less oxygenated offshore environment. More usually, the shoreface sequence represents an upward increase in the proportion of physical versus biogenic structures.

In summary, the above facies sequence analysis of the 'upper proximal mega-facies' and succeeding basal Hangman Sandstone Group is interpreted as representing the progradation of a storm- and wave-dominated shoreline. The following sections compare this sequence with analogous sequences recorded from both the Recent and geological record.

6.3.1 Modern Analogues

On wave-dominated shorelines, sediment eroded from the upper shoreface during storms is deposited as washover fans in lagoons or storm-event layers on the lower shoreface and offshore. Thus, the lower shoreface tends to be dominated by the alternation of fair-weather and storm-deposited layers, whilst the upper shoreface is dominated by fair-weather deposits which are repeatedly interrupted by erosion surfaces produced by storm wave planation. (Sonu & van Beek 1971). In environments of relatively stable sea-level, low to moderate subsidence and a continuous supply of sediment, beach or barrier island shorelines will prograde in a seaward direction (Bernard *et al.* 1962).

The most pertinent analogue for the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition sequences is the landmark study of the depositional structures and processes in a non-barred, high-energy wave-dominated shoreline setting off the Oregon (U.S.A.) shoreline by Clifton *et al.* (1971 - see text-figure 4.18 for a reproduction of Clifton *et al.*'s facies zonation scheme). The following paragraphs describe

the facies recognised by Clifton *et al.* and compare them to the facies recorded in the 'upper proximal megafacies' - basal Hangman Sandstone Group transition sequences. Moving in a landward direction:

Clifton *et al.*'s 'inner offshore facies' comprised asymmetric (no preferred facing direction was observed) wave ripples, frequently exhibiting cross-cutting sets of interference ripples. This facies is comparable to the thinly interlayered sandstone/mudstone bedding and thinly-bedded sheet sandstones facies. The presence of frequent mudstone layers in facies A is attributed to the Lynton Formation shelf being more muddy than the Oregon shelf and the fact that facies A probably represents a more offshore environment below the depth studied by Clifton *et al.* which also explains the lower degree of ripple asymmetry recorded in the Lynton Formation wave-ripples.

On the Oregon shoreline the outer portion of the zone of wave build-up (the 'outer rough facies') comprised a zone of landward facing megaripples 30 to 100cm in height. On first consideration this facies does not appear to have an equivalent in the Oxen Tor sequence. However, Clifton *et al.* noted that 'outer rough facies' megaripples become re-oriented under longshore currents. During the study period longshore currents were not significant, in relation to oscillatory currents, on the Oregon shoreline. Furthermore, the 'outer rough facies' ". . . may, under small short-period waves, become so modified as to be nearly unrecognisable. Under the smallest waves encountered . . . average wave period of 6 seconds . . . the facies . . . contained a combination of straight symmetric sand ripples 10 to 15 cm high and irregular megaripples which mostly lacked a lunate form." (p. 665). Given the strong longshore currents that would be expected on the trade wind dominated Lynton Formation shoreline (see section 4.2.2.3), and that the limited fetch of the putative 'Exmoor Basin' would have given rise to waves with a period of ≤ 4 seconds (see chapter 8), then it is unlikely that an 'outer rough facies' type would have developed off the Lynton Formation shoreline. The equivalent to the Oregon 'outer rough facies' appears to be the isolated sets of trough cross-bedding (facies C), that are found interbedded with facies A and predominantly near the base of the bipolar ripple cross-laminated facies F, representing the migration of sinuous-crested dunes under faster than ambient longshore currents. In common with the Oregon 'outer rough facies', the tops of facies C cross-beds show evidence of reworking by symmetric oscillatory flow, representing lower energy periods that intervened between periods of active dune migration.

The inner portion of the wave build-up zone to outer surf zone ('outer planar facies' of Clifton *et al.*) was characterised by parallel-laminated sands developed under oscillatory sheet flow with occasional zones of wave-ripples (c. 10cm wavelength). This facies is thought to be an analogue for the wave-ripple cross-laminated facies F which contains thin zones of parallel-lamination interpreted as having been generated under oscillatory sheet flow conditions. The fact that facies F has only thin zones of parallel-lamination compared with the Oregon 'outer planar facies' is attributed to the Lynton Formation being a lower-energy wave climate than the open ocean Oregon coast, an interpretation supported by the preservation of the lower-energy variant facies sequence: bipolar ripple cross-lamination (facies F) \Rightarrow planar cross-bedding (facies D) \Rightarrow flaser bedding (facies E) \Rightarrow return to facies F.

The upper shoreface on the Oregon coastline was dominated by onshore-dipping cross-bedding created under shoaling wave conditions ('inner rough facies') developed between the inner surf and the swash zones. Off steep beaches, large (15-20 cm high) long-crested symmetrical ripples developed, whilst off gently sloping beaches a series of 1 to 2m wide, 10 to 50cm deep, troughs developed which were elongated parallel to the shoreline and migrated seawards over time. The latter structure was preserved as predominantly seaward-dipping trough cross-bedded sets. Clifton *et al.* noted that: "Longshore currents, which commonly reach their maximum velocities over the general area occupied by the inner rough facies, modify the shape of the depressions and the direction of their migration. Under the influence of longshore currents, the steepest side shifts toward the up-current end of the depressions and cross-stratification shows a pronounced longshore component" (p.657) cf. the facies G trough cross-beds. Furthermore, Clifton *et al.* recorded that: "Rarely, shoreward-facing lunate megaripples about 10 cm high and a meter in span were present in the seaward portion of the inner rough facies" (p.657). This latter phenomenon is analogous to the thin sets of palaeo-shoreline-facing cross-beds that are interbedded between thicker sets of ESE-dipping cross-bedding to produce 'herringbone' cross-bedding in facies G.

Finally, the uppermost Oregon shoreface and foreshore were dominated by seaward-dipping sets of parallel-lamination ('inner planar facies') developed under shoaling wave conditions during fair-weather. Facies H is interpreted as representing the lower part of the 'inner planar facies' i.e. there is no evidence of exposure / foreshore deposits preserved.

In summary, the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition is interpreted as preserving the progradation of a wave-dominated shoreline with a relatively gently sloping shoreface. The environment of deposition was of lower energy than the Oregon coastline, allowing mud to be deposited on the shoreface in quieter periods (flaser bedded facies E). Importantly, longshore currents played a significant role in bedform migration when compared to their minimal influence on the Oregon shoreline deposits studied by Clifton *et al.*

A further valuable study of relevance to interpreting the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition is that of Howard and Reineck (1981) who compared the depositional facies recorded from modern examples of both low-energy (wave- and tide-dominated: Sapelo Island, Georgia, U.S.A. - see also Howard & Reineck 1972) and high-energy (California shelf) beach-to-offshore sequences. The following zones were observed off the Californian coast:

Offshore (>-19m water depth i.e. near to below storm wave-base): The transects extended further out into the offshore than the study by Clifton *et al.* (1971) and the offshore zone was found to be thoroughly bioturbated (90 - 100%) with only remnants of parallel-laminated thin storm sands preserved reflecting the alternation of storm and fair-weather conditions. The facies A / B alternation in the 'upper proximal mega-facies' equates to these offshore zones, although bioturbation is less intense in facies A / B, presumably a function of the fact that bioturbation has increased during the Phanerozoic (Bambach 1977, Thayer 1979).

Transition (-9 to -19m water depth i.e. approximately between fair-weather and storm wave-base): The offshore transition and lower shoreface comprised erosive-based parallel- and undulatory-laminated sand layers with wave-rippled tops, alternating with bioturbated silty muds similar to the upper part of the facies A / B sequence at Oxen Tor. Cross-bedding was absent.

Nearshore: The following zones were reported:

- Small-scale (oscillation) ripple lamination (below low water to c. -9m, maximum depth of -29m): Equates to facies F.
- Cross-bedded sand (mean high water to c. -9m): 10 to 30cm thick sets with erosional set boundaries with a dominant longshore component and subordinate onshore and offshore modes. Equates to facies G.
- Parallel-laminated sand (backshore to -9m): Wedge-shape sets of seaward-dipping parallel-laminated sand with heavy mineral layers. Equates to facies H.

Bioturbation was practically non-existent above a depth of c. 6m. Below 6m bioturbation was present but the bedding was dominated by primary physical sedimentary structures

In comparison the low-energy, mixed wave- and tide-dominated Sapelo Island sequence was broadly similar but exhibited the following differences when compared with the higher-energy Californian sequence:

- the water depths at which the boundaries of the facies occur
- facies thicknesses
- thin mud lenses were occasionally preserved in the shoreface sequence
- cross-bedding is not typical of the Sapelo Island shoreface, although this could be due to the absence of sand of a suitable grain size to allow dunes to form.

The significance of the comparison between the Californian and Georgia sequences to the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition is that the main discriminating factor between low- and high-wave-energy sequences is the facies thickness. Given the complex interplay of subsidence rate, sea-level changes and sediment supply and their impact on the thickness of ancient sequences it is not possible to say unequivocally whether the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition was generated on a low- or high-wave-energy shoreline. Nevertheless, the presence of mudstone flasers in facies E and the bipolar ripple cross-lamination (facies F) ⇒ planar cross-bedding (facies D) ⇒ flaser bedding (facies E) ⇒ return to facies F lower energy facies variant indicates that the 'upper proximal

mega-facies' - basal Hangman Sandstone Group transition was generated on a lower wave-energy shoreline than the ocean-facing Californian sequence.

In comparison with sequences described from wave-dominated barred-shorelines (Davidson-Arnott & Greenwood 1974, Hunter *et al.* 1979), where oblique and shore-parallel bars complicate the upper shoreface and are dissected by rip-current channels with seaward-directed planar cross-bedding, the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition lacks the following diagnostic characteristics:

- repetition of vertically stacked repetitive facies belts preserved by the migration of bars across the upper shoreface
- major erosion surfaces at the base of units representing the migration of a rip-channel
- significant amounts of onshore-directed cross-bedding formed by the shoaling waves responsible for the construction and maintenance of the bars

Finally, as was noted for the A39 coarsening-upwards sequence (section 4.3.3), the sequence at Oxen Tor does not contain characteristics allowing a distinction to be made between a barrier and a non-barrier coastline to be made. The critical major basal erosion surface and back barrier washover fan and lagoonal facies necessary to distinguish between a barrier and non-barrier sequence are not exposed in the section and the barrier/non-barrier origin of the sequence must, therefore, remain equivocal.

6.3.2 Ancient Analogues

Elliott (1986) noted that in ancient sequences interpreted as representing the progradation of wave-dominated shorelines evidence for storm-dominated deposits is generally only present in the lower parts of sequences, being absent in the upper part due to extensive reworking by fair-weather shoaling waves.

Roep *et al.* (1979) described a Messinian (Miocene) sequence near Alemeria in southern Spain where mudstones with thin storm sandstones, interpreted as representing the offshore to shoreface transition zone (cf. facies A & B at Oxen Tor), passed upwards into wave-ripple cross-lamination and parallel-lamination (cf. facies F) interpreted as representing fair-weather and storm-weather shoaling waves respectively.

Deposits representing the upper shoreface shoaling wave environment comprised trough cross-bedding (cf. facies G), which was locally conglomeratic, which in turn passed upwards into parallel-laminated sandstones (cf. facies H) and beach rock erosion surfaces and breccia. The Messinian sequence differs from the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition in that the upper shoreface trough cross-bedding had a dominantly onshore palaeocurrent pattern (shoaling-wave-dominated), compared to the predominantly longshore pattern recorded in facies G (longshore-current-dominated).

Sequences representing prograding wave-dominated shorelines where longshore currents were significant have been documented in the Cretaceous Gallup Sandstone of New Mexico by McCubbin (1972, 1982) and Harms *et al.* (1982). The transition from offshore burrowed laminated siltstone and thin beds of hummocky cross-stratified sandstone (HCS) of storm origin into the lower shoreface is marked by an increase in the thickness and proportion of beds of HCS. The upper shoreface sequence was dominated by trough cross-bedded sandstone indicating longshore palaeocurrents directed towards the SSE with some opposed dips to the NNW cf. facies G at Oxen Tor. Finally, low-angled sets of parallel-lamination with heavy mineral placers preserved a foreshore environment. (cf. facies H). The Gallup Sandstone differs from the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition in that HCS is well developed in the storm-dominated lower shoreface - offshore transition zone and a lower shoreface wave-ripple cross-laminated zone (cf. facies F at Oxen Tor) was not preserved in the Gallup Sandstone. These two differences may be interconnected. The Gallup Sandstone was deposited in the Cretaceous Western Interior Seaway and at the time was located at a latitude of 33°N (Duke 1985) within a zone stretching between 25° and 45° where winter storms are common and hurricanes occur occasionally (Marsaglia & Klein 1983). This location would have been on the margin of the trade wind zone that currently stretches from a latitude 5/10° to 35° N/S, a zone within which southern Britain was located at the close of the Lower Devonian (see section 4.2.2.3). In the Lynton Formation semi-permanent longshore currents would have prevailed during the trade wind season, maintaining the shoreface profile which would have been rippled by oscillatory currents in its lower part and would have obliterated the bulk of evidence for storm activity on the lower shoreface. In contrast, the Gallup Sandstone was deposited near the boundary between a trade wind and mixed hurricane / winter storm environment where the absence / negligible presence of semi-permanent wind-forced currents would have resulted in a storm-maintained lower shoreface profile, within the lower regions of which storm beds would have a higher chance of preservation than their Lynton Formation counterparts.

Clifton (1981) described a series of Miocene high-wave-energy shoreline deposits from the SE Caliente Range, California, which cropped out as c. 50 coarsening-upwards progradational sequences separated by thin units of conglomeratic transgressive lag deposits overlying major erosion surfaces. Within an individual sequence the offshore transition to lower shoreface comprised units of horizontally- and cross-laminated siltstone and fine sandstone containing coarser lenses of cross-bedded coarse sandstone and granules exhibiting obliquely-onshore palaeocurrents. The lower shoreface deposits were truncated by a major erosion surface overlain by offshore-directed sets of cross-bedded coarse grade sandstone interpreted as rip channel deposits in a barred shoreline i.e. the rip channels would have reworked bar deposits as they migrated alongshore. The rip channel deposits were overlain in turn by offshore-dipping parallel-laminated foreshore deposits and capped by red or green fluvial mudstones. The 'upper proximal mega-facies' - basal Hangman Sandstone Group transition lacks the mid shoreface rip channel erosion surface overlain by offshore-directed sets of cross-bedding of the Miocene sequences, suggesting that the Lynton Formation shoreline was not subject to the migration of rip channels along a barred shoreface.

In summary, a comparison of the 'upper proximal mega-facies' - basal Hangman Sandstone Group transition with similar sequences reported from the geological record suggests that the sequence is closer to prograding wave-dominated shoreline examples interpreted as being dominated by longshore currents than those either interpreted as being dominated by shoaling wave processes or rip channels migrating along a barred shoreline. Rip currents would be expected to develop along a non-barred, longshore current dominated shoreline but they would be more ephemeral than the channelised deposits associated with a barred shoreline and would not necessarily be expected to leave a record.

7. THE TRANSITION TO THE HANGMAN SANDSTONE GROUP DISTAL TO THE LYNMOUTH - EAST LYN FAULT: THE 'UPPER DISTAL MEGA-FACIES'

7.1 INTRODUCTION

The boundary between the Lynton Formation and the overlying Hangman Sandstone Group, as defined by Tunbridge (1978; see also section 1.8.3 herein), is diachronous, the Lynton Formation thickening by approximately 55m in a basinward (southerly) direction (see enclosure 2). This chapter describes and interprets the sedimentology of the basinward-thickening wedge at the top of the Lynton Formation and the transition to the succeeding Hangman Sandstone Group distal to the Lynmouth - East Lyn Fault. The transition to the Hangman Sandstone Group proximal to the Lynmouth - East Lyn Fault is described in chapter 6. A comparison of the changing character of the transition between the proximal and distal settings is discussed in section 8.3.

The basinward-thickening wedge is differentiated from the remaining Lynton Formation by the ubiquitous presence of 'hummocky cross-stratification' (*sensu* Harms 1975), individual beds frequently being amalgamated to form 1 to 5m thick sandstone-bodies referable to the 'Woody Bay facies association' (see section 1.8.3.1.3 for definition). Examples of hummocky cross-stratification are visible along the coastline between western Woody Bay (675 492) and the most westerly exposure of the Lynton Formation at Ramsey Beach (6465 4938), continuing inland along a broad strip adjacent to the southern boundary of the Lynton Formation.

For the purpose of elucidating the sedimentology of the basinward-thickening wedge at the top of the Lynton Formation, a detailed analysis and discussion of the facies present in three representative sections is presented for: Woody Bay west, WNW Barbrook and the Lynton Formation - Hangman Sandstone Group transition at Little Burland.

The section at Woody Bay west (Enclosure 10) occurs in the low, eastward facing cliff section running NNE from the lime kiln towards the old pier (6770 4903 - see enclosures 1 & 6, plate 7.4C); the section is accessible for 3 hours either side of low tide. Bedding dips at 32° towards the SSW. The type section of the 'Woody Bay facies association' is defined at this locality as occurring between 12.78 and 16.47m on the log. The precise stratigraphic position of this locality, in relation to the overall Lynton Formation succession, is

unclear due to the absence of horizons suitable for correlation. Evidence from mapping, however, suggests that the section occurs towards the base of the basinward-thickening wedge at the top of the Lynton Formation (see enclosure 2).

The section WNW of Barbrook (text-figure 7.1) lies in the road cutting to the east of the lay-by 0.7km, WNW of Barbrook (7077 4787 - see enclosure 1); bedding dips 35° towards the SSW. This section contains the auxiliary reference section of the 'Woody Bay facies association' (see section 1.8.3.1.3 for definition), which occurs between 20.99m and 23.0m on the log. Again, the precise stratigraphical position of this locality in relation to the overall Lynton Formation sequence is unclear. Mapping considerations, however, indicate that the section lies towards the top of the Lynton Formation (see enclosure 2).

The Little Burland section (enclosures 11A & 11B) comprises a series of small outcrops forming a rocky rib immediately below the summit of the lower of the two coastal footpaths that link Heddon's Mouth with Woody Bay (6629 4957 to 6626 4949 - see enclosure 1); bedding dips 27° towards the SW. The section displays the coarsening-upwards transition from the Lynton Formation to the Hollowbrook Formation, the basal Formation of the Hangman Sandstone Group. A boundary-hypostratotype for the Lynton Formation - Hangman Sandstone Group boundary was defined at this locality (see section 1.8.3 for details), occurring at 31.2 m on the log. This section has previously been logged by Tunbridge (1978, figure 4.4). The boundary-lectostratotype of the Lynton Formation - Hangman Sandstone Group was defined at Great Burland (663 497- see section 1.8.3 for detail), some 200m ESE of Little Burland. The sequence at Great Burland has been logged and described by Tunbridge (1978 figure 4.22 & 1983a, Figure 11) and is sedimentologically very similar to the Little Burland sequence.

As noted above, the facies occurring in the three detailed logged sections will be considered together for the purpose of description and analysis. As a different facies nomenclature was used for each of the three logged sections it has been necessary to develop a common scheme for the purpose of description; this is presented in table 7.1.

Text-fig. 7.1 Road cutting 0.7km WNW of Barbrook - log through the auxiliary reference section of the 'Woody Bay facies association'.

Grid reference 7077 4787. The 'Woody Bay facies association' is defined as occurring between 20.99m and 23.0m on the log. Facies: **A** = thinly interlayered sand/mud bedding; **B** = does not occur at this locality; **C** = thin bedded sheet sandstones; **D** = hummocky cross-stratification; **E** = bipolar ripple cross-lamination.

| Facies given on log ↓ | Woody Bay West (Enclosure 10) | WNW Barbrook (text-figure 7.1) | Little Burland (enclosures 11A & 11B) |
|-----------------------|-------------------------------|--------------------------------|---------------------------------------|
| A | 1 | 1 | 1 |
| A' | 1' | - | - |
| B | 2 | - | 3 |
| C | 3 | 3 | 4 |
| D | 4 | 4 | 6 |
| E | 5 | 7 | 7 |
| F | - | - | 8 |

Table 7.1 Chapter 7 facies scheme.

Key: **1** = thinly interlayered sandstone/mudstone bedding; **1'** = coarsening-upwards microsequences; **2** = pure mudstone; **3** = thinly bedded sheet sandstones; **4** = hummocky cross-stratification; **5** = cross-bedding; **6** = wavy bedding; **7** = bipolar ripple cross-lamination; **8** = parallel-laminated sandstone with thin ripple cross-laminated zones

7.2 FACIES 1 -THINLY INTERLAYERED SANDSTONE/MUDSTONE BEDDING

Facies 1 is shown as facies A on all three logs described within this chapter.

Units of facies 1 range between 1 and 100cm in thickness, the thicker units displaying a laterally extensive tabular geometry. Many of the boundaries depicted on the logs between units of facies 1 are arbitrary, corresponding with abrupt changes in gross sand content of units. Towards the base of the WNW Barbrook section (text-figure 7.1) many of the boundaries have been exploited by bedding plane slip and display a deeply weathered clay gouge. Elsewhere, units of facies 1 are frequently separated by laterally extensive erosion surfaces (plate 7.1A) with a relief of up to 2cm. In the lower part of the Little Burland sequence a 1.5m wide scour-form (channel?) cuts down 6cm into the underlying lenticular bedding, the scour-form itself being infilled with lenticular bedding (plate 7.1B).

Although facies 1 generally overlies other facies non-erosively, thereby draping and preserving the morphology of the upper bedding surface of the preceding facies, several erosive contacts between hummocky cross-stratification (facies 4) and overlying lenticular bedding have been observed (plate 7.11B). Thinner units of facies 1 tend to be laterally inpersistent, particularly when intervening between hummocky cross-stratified cosets (facies 4) or parallel-laminated sandstone cosets containing thin ripple cross-laminated zones (facies 8). Rarely, isolated concave-up 'pods' of lenticular bedding are preserved in the swales of hummocky cross-stratified cosets (plate 7.11C).

Internally, facies 1 units have sandstone:mudstone ratios varying between 5% and 95%, variations within units defining fining-upwards or coarsening-upwards microsequences. The latter type of microsequence is more common in the Woody Bay west section than elsewhere and is discussed below as a specific subset of facies 1 i.e. facies 1' (see section 7.3). The grain size of the sandstone in facies 1 varies from coarse siltstone to fine sandstone grade. Individual sandstone layer characteristics allow four sub-facies to be defined; these are described below.

Thin beds of sharp, planar based sandstone with gradational tops and a thickness of 5 to 10mm are termed graded rhythmites. Although granulometric contrast is generally low, weak normal grading can frequently be discerned. Graded rhythmites are uncommon in the sections under consideration in this chapter, mainly being restricted to the Woody Bay west section e.g. at 1m on the log (enclosure 10).

Horizontally-laminated sandstone streaks are also uncommon in the sections considered within this chapter, again being principally restricted to the Woody Bay west section e.g. 4.2m on the log. Individual streaks have planar bases and tops, ranging in thickness from 2 to 5 mm. Internal lamination is faint and horizontal to gently undulatory.

The bulk of facies 1 comprises unconnected and connected lenticular bedding. The unconnected lenticular bedding takes the form of isolated sandstone lenses, 2 to 10mm in height and attaining widths of 15 to 40mm, set within silty mudstone. The internal lamination style is characteristic of a wave-current combined flow origin (plate 7.1C). The connected lenses have a pinch-and-swell morphology, the thickness of individual lenses varying between 3 and 20mm, wavelengths varying between 5 and 8cm and amplitudes between 5 and 10mm. Again, the lamination style is characteristic of a wave-current combined flow origin. Ripple crestlines are only visible rarely and were hard to measure; observations suggest that the majority of crestlines trend between WNW-ESE and WSW-ENE i.e. parallel to oblique in relation to the WNW-ESE striking palaeoslope. Several of the upper surfaces of the connected lenses display 'micro-hummocks' referable to '3-D vortex ripples' of wave origin cf. section 4.2.1.4. Thinly-bedded sheet sandstones (facies 3) are frequently interbedded with connected lenticular bedding (plate 7.15B).

The deposits preserved in facies 1 were frequently disturbed by soft sediment deformation, lateral flowage/sliding (plate 7.11B) and loading of sand lenses within silty muds (plate 7.13C) having been observed.

Facies 1 has been extensively disturbed by biogenic activity, although the extent of this activity is hard to assess on well weathered, lichen covered units, particularly in the WNW Barbrook section where much of the 'bioturbation' column on the log has been marked '?' as a consequence. Biogenic 'churning' is locally intense enough to completely homogenise units, resulting in mottled/gnarled silty-mudstones which exhibit a poor cleavage in contrast to the pure, well cleaved mudstones of facies 2. Five to seven millimetre diameter tubes attributable to *Palaeophycus tubularis* are ubiquitous, while single examples of *Teichichnus rectus* and *Roselia socialis* have been recorded at Woody Bay west and Little Burland respectively. N.B. Tunbridge (1978 and 1983a) referred to the presence of *Chondrites* at Great Burland (see section 1.8.3). The absence of *Chondrites* in the sections under consideration in this chapter and observations by the author at Great Burland indicate that Tunbridge's references to *Chondrites* should be referred to weakly branching forms of *P. tubularis*. Measurement of *P. tubularis* burrow orientation at Woody Bay west indicates a strong north-south tube alignment (see text-figure B9), a direction normal to the palaeocurrent at this locality - see appendix B for discussion.

The interpretation of facies 1 is based upon that for the closely analogous facies 1 at Lee Stone - see section 4.2.1 for a detailed discussion and references. The graded rhythmites are the product of fall-out from storm-generated suspension clouds of sediment of a sufficient grain size range to give a normally graded bed. Deposition occurred well below wave-base, away from the influence of wave orbital currents. The horizontally-laminated sandstone streaks were deposited nearer to storm wave-base, the horizontal-lamination reflecting energy pulses of wave orbital origin which influenced fall-out from a storm-generated suspension cloud. In contrast, the unconnected and connected lenticular bedding were deposited within wave-base where storm generated currents were sufficient for bedload traction, the unconnected lenses reflecting a limited sediment supply i.e. distal to sand-bodies.

In conclusion, facies 1 was deposited in an environment favouring the preservation of mud. The presence and character of sandstone layers was dependent on the proximity of wave-base which would have fluctuated in response to storm-events. Occasionally, the major storm events had a nett erosive affect, resulting in the development of laterally extensive erosion surfaces. The biofacies is generally characteristic of Schäfer's (1972) vital-lipstrate (see enclosure 4) i.e. agitated, well-aerated water with intermittent sediment movement.

7.3 FACIES 1' -COARSENING-UPWARDS MICROSEQUENCES

Facies 1' is shown as facies A' on the Woody Bay west log (enclosure 10), the only sequence within which it has proven possible to differentiate this facies 1 sub-type.

Individual facies 1 microsequences range in thickness from 3 to 32cm and display an upwards increase in sandstone content within the range 5% to 95%. Microsequence boundaries are gradational or sharp, never erosional, frequently preserving the morphology of the upper surface of the underlying unit. The majority of microsequences exhibit an upwards transition from unconnected to connected lenticular bedding; graded rhythmites and horizontally-laminated sandstone streaks have not been observed. Bioturbation in the form of *Palaeophycus tubularis* burrows is common, burrows occasionally crossing unit boundaries.

By analogy with similar coarsening-upwards microsequences preserved within the Lee Stone sequence (see section 4.2.2), facies 1 microsequences are interpreted as the product of seasonal deposition in a mixed trade wind / monsoonal climate and /or the increasing proximity of a sand patch with a unidirectional migration path in response to geostrophic flow inducing semi-permanent currents..

7.4 FACIES 2 - PURE MUDSTONE

Facies 2 is shown as facies B on the Woody Bay west log (enclosure 10), the only sequence within which this facies occurs within the sections considered within the current chapter.

Facies 2 comprises 4 to 13cm thick pure mudstones with a silt content not exceeding 10%. Mudstones of this facies may be distinguished from thick silty mudstones of facies 1 by the presence of a strong tectonic cleavage, absence of mottling and absence of a gnarled texture. Bed bases are sharp, draping and preserving the morphology of the upper surface of preceding units e.g. hummocky surfaces at 15.0 and 15.4m on the Woody Bay west log. Unit tops are sharp or gradational and are planar. The majority of units lack sufficient granulometric contrast to develop a graded profile. The unit at 15.8m on the Woody Bay west log, however, shows a good normal grading, the silt content decreasing upwards within the unit (plate 7.14A). Facies 2 is found interbedded with cosets of hummocky cross-stratification (facies 4). The absence of bioturbation indicates that facies 2 belongs to Schäfer's (1972) letal-pantostrat biofacies (see enclosure 4).

The interpretation of facies 2 is based upon analogous deposits in the A39 road section (see section 4.2.3 - facies 2). Individual mudstone beds are interpreted as event deposits generated either by (i) storm action sourced by erosion of a zone of silty-mud deposits, or (ii) settling from a cloud of fine-grained material suspended by the down-slope translation of slides and slumps triggered by syn-sedimentary tectonic movements. This type of bed is a common component of Devonian open shelf and shoreface clastic facies (Goldring and Langenstrassen 1979).

7.5 FACIES 3 - THINLY-BEDDED SHEET SANDSTONES

Facies 3 is shown as facies B on the Little Burland log and facies C on the Woody Bay west and WNW Barbrook logs.

Single, 2 to 14cm thick beds of sheet sandstones are frequently interbedded with facies 1, particularly with connected lenticular bedding (plate 7.15B; see also section 7.2). Individual beds are generally laterally persistent over the width of the exposure, although isolated bi-convex lenses up to 25cm in length (e.g. plate 7.3A) and infilling shallow scours (0.2 & 25.7m on Little Burland log) are occasionally visible. Grain size ranges from coarse siltstone to medium sandstones, clean quartz sub-litharenites occasionally occurring (plate 7.15B).

Bed bases are planar (plate 7.15B) to undulatory (plate 7.2A), with up to 15mm relief; erosive contacts are common, frequently with small scours visible (plates 7.2B & C, 7.3B). Soft sediment deformation structures are rare, although loading (8.0m on Woody Bay west log) and flame structures (plate 7.6B) have been observed. Indistinct tool marks are visible on the lower surface of some beds.

Symmetric to near-symmetric wave- / combined-flow ripple profiles are visible at the top of many beds (plate 7.2B), wavelengths ranging from 7 to 12cm with amplitudes of up to 1cm. Crestlines trend between ENE to WSW and NNW to SSE, the majority tending toward the former trend. Bedding plane views of ripple surfaces indicate that crestlines are generally straight, crest profiles being either trochoidal (plate 7.12A) or, more commonly, rounded (plate 7.12B); interference ripple patterns are common (plate 7.12C). More rarely, bed tops are planar (plate 7.2C).

Internally, individual beds occasionally exhibit a normally graded profile (0.7m on Woody Bay west log), although the majority have a uniform size distribution. Rarely, a lag of shell debris is present at the base of a bed (48.2m on Little Burland log) or a massive sand (1.2m on Woody Bay west log) draping and preserving the ripple profile at top of the preceding unit to give an amalgamated facies 3 unit (8.4 and 11.0m on WNW Barbrook log). Parallel- to undulatory-lamination is the dominant internal lamination style of facies 3, occasionally to the exclusion of all other lamination styles. Beds comprising solely parallel-lamination (i.e. horizontal laminae), are rare, a basal zone of parallel-lamination swinging-up into undulatory-lamination being more common (plate 7.3C). In many cases the entire bed may comprise undulatory-lamination, the laminae frequently displaying a tangential relationship with lower surface of the bed (plate 7.3B). Undulatory-laminae have a wavelength of 10 to 15cm and an amplitude not exceeding 1cm; up to 3 sets of laminae may be present in any one unit/coset (plate 7.2C). Within a coset, set boundaries are sharp. undulatory and truncate laminae of the preceding set, sometimes at a high angle (plate 7.2C). Primary current lineation has not been observed.

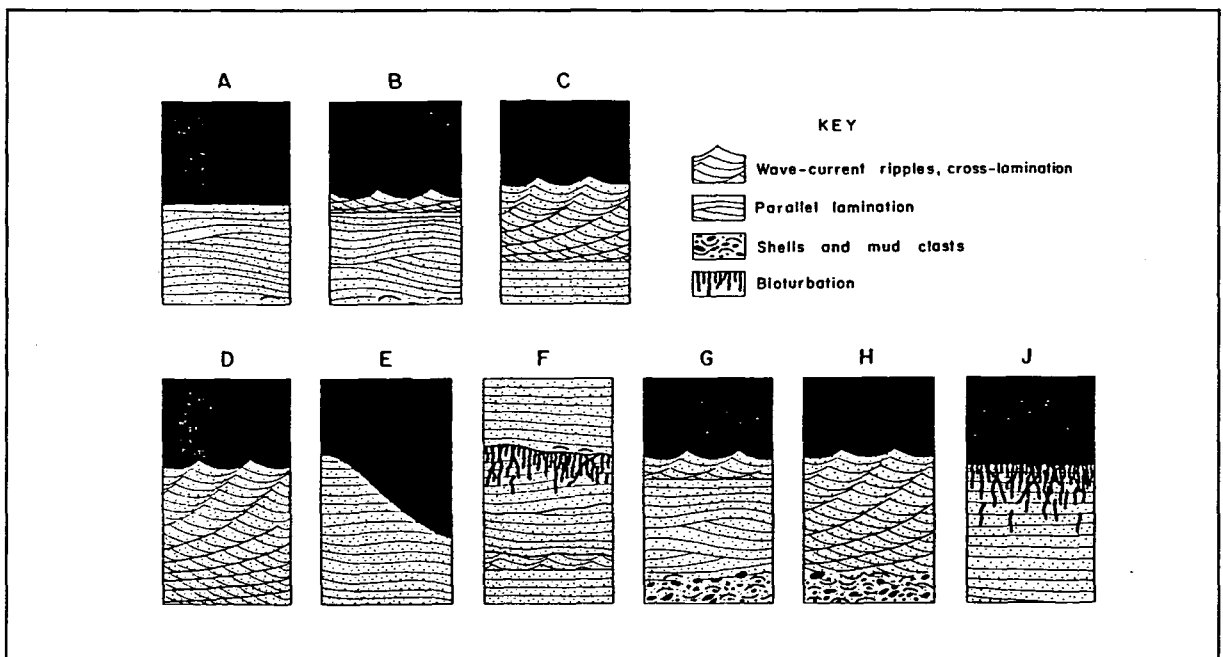
The uppermost zone of many beds is characterised by a lamination style diagnostic of an oscillatory/combined-flow origin. Laminae are generally form-discordant in relation to the ripple profile at the top of the bed (plate 7.15B), although they may be form-concordant (plate 7.3C); adjacent ripples frequently display a dissimilar lamination style (plate 7.3C). Several beds are composed entirely of wave-ripple cross-lamination, a maximum of three sets having been observed in a single bed (11.5m on WNW Barbrook log). Very rarely the uppermost zone of a bed may display an alternation of 1 to 2mm thick parallel streaks of sandstone and mudstone e.g. 0.7m on Woody Bay west log.

Facies 3 is frequently disturbed by biogenic structures. Convex hyporelief open-looped trails and concave hyporelief broad, sinuous grooves have been observed on the base of one bed (Plate 7.6C) and are interpreted as pre-depositional tunnels that were exhumed and cast during the high energy event that emplaced the sheet sandstone cf. Seilacher (1962). More commonly, biogenic 'churning' descends from the upper surface of beds, resulting in the complete loss of primary lamination locally. Bedding plane views of the upper surfaces of beds frequently exhibit concave and convex epirelief trails (plate 7.12C & 7.13A).

Goldring (1966) was the first author to recognise and describe sheet sandstones as a distinct facies in ancient shallow marine sequences, attributing their origin to the combined action of waves and unidirectional

currents; specific reference to storm action was not made. Subsequently, shallow marine sheet sandstones have been reported in a large number of studies, the beds being attributed to a product of storm action. Beds of this type were termed 'sublittoral sheet sandstones' by Goldring and Bridges (1973) and "storm lag" deposits by Brenner and Davies (1973). Tunbridge (1983a) assigned similar beds in the Great Burland section to his Facies 3A - 'parallel laminated sandstone (single beds)' - proposing that the beds originated by fall-out from sand-laden suspension clouds entrained by storm waves.

Allen (1982) provided an extensive literature review of shallow marine sheet sandstones, citing approximately fifty examples from the Phanerozoic record. The majority of examples were clastic, although several carbonate examples were quoted. Allen synthesised the findings of his literature review into a scheme based upon the variation of sedimentary structures within reported vertical sequences of beds ascribed to a storm origin (figure 12-12 of Allen, reproduced below as text-figure 7.2).



Text-fig. 7.2 Schematic representation of vertical sequences reported from beds thought to be storm sand-layers.

See text for discussion. From: Allen (1982).

A consideration of the suite of vertical sequences preserved within facies 3 indicates that the majority of members of Allen's scheme are present within the Lynton Formation; these are described below with reference to Allen's scheme shown in text-figure 7.2. Type A beds are exclusively composed of parallel-lamination, cf. Goldring & Bridges (1973), Anderton (1975) and Brenchley (1985), and are similar to many storm sands reported from modern shelf sequences (see discussion of facies 10, Lynmouth Beach in section

4.2.11). In type B beds, parallel-lamination is capped by a wave-rippled bed top, cf. Goldring and Bridges (1973) and Hamblin and Walker (1979), whilst in type C beds parallel-lamination is capped by a wave-current or current-rippled bed top, cf. Banks (1973a), Anderton (1976) and Vos (1977); type D beds are solely composed of either wave-current or current cross-lamination, cf. de Raaf *et al.* (1965), Anderton (1976) and Vos (1977). No examples of type E beds, where the bed top is erosively truncated and overlain by mudstone, have been observed in facies 3.

Type G and H beds exhibit a basal lag comprising a mainly suspension-feeding fauna, cf. Brenner & Davies (1973), Bridges (1975) and Vos (1977). Goldring and Bridges (1973) noted that the basal lag fauna may be swept together from over a wide area, whereas the fauna in the surrounding mudstones frequently appears to have been smothered *in situ*. Type J beds have bioturbated tops, cf. Howard (1972), Howard and Reineck (1972), Goldring and Bridges (1973) and Reineck and Singh (1980), representing fair-weather colonisation following the storm event that emplaced the bed.

Allen (1982) noted that the above sequences were present in beds that were generally laterally extensive, overlying a planar to gently undulose or irregular erosion surface. Allen also observed an inverse relationship between sandstone bed thickness and the thickness of surrounding mudstone units i.e. the thinner sandstones tend to be set within thicker mudstone units; this phenomenon is present in facies 3. In summary, with the exception of Allen's type E bed sequence (erosive top overlain by mudstone), all of Allen's bed sequences have been observed within facies 3.

Johnson (1978) synthesised the geological literature relating to mechanisms of emplacement of storm sandstones. The beds are generally attributed to waning flow of sediment-laden currents, individual beds recording single storm events comprising: initial storm erosion followed by deposition and finally, post-storm reworking. Oscillatory and associated currents winnow the substrate leaving a basal lag deposit, the overlying parallel-laminated sand representing fall-out from suspension, high bed shear stress sweeping the sand into flat sheets with primary current lineated surfaces. As flow wanes further, the bed top is rippled by waves and currents. Post-storm reworking of the bed top may either take the form of wave-rippling or bioturbation. Storm generated sheet sandstones are found in two broad settings: 'shoreface-shoreline association' and the 'open shelf association' (Goldring and Bridges 1973). Facies sequence analysis of the sections described within this chapter (see section 7.11) suggests that facies 3 at Woody Bay west and WNW

Barbrook is referable to the 'open shelf association', whilst at Little Burland the 'shoreface-shoreline association' is appropriate.

Storm sands have been described from many modern shelf settings, their characteristics being broadly similar to those described from ancient sequences. Hayes (1967) described a sharp based, normally graded sand layer from the Gulf of Mexico which attained a maximum thickness of 9cm. Hayes proposed a turbidite origin for the bed, suggesting that a density current was triggered by a storm surge ebb flow. Reineck and co-workers have reported examples of modern storm sands from: the North Sea (Reineck *et al.* 1967); German Bight (Gadow & Reineck 1969) where sand layers of between 2cm and less than 1cm thickness exhibit solely parallel-lamination, solely cross-lamination and parallel-lamination overlain by cross-lamination; the Gulf of Gaeta (Reineck & Singh 1972); Georgia coastline (Howard & Reineck 1972). A storm surge ebb mechanism (cf. Hayes 1967) augmenting wave generated currents was invoked for the Gulf of Gaeta study, the suggestion being advanced that sand only accumulated during the waning phases of storms. Finally, Howard and Reineck (1981), in a study of a high-energy coastline in southern California, envisaged purely wave-induced currents to be responsible for the emplacement of storm sands.

Nelson (1982), in a study of cores retrieved from the Yukon shelf, proposed an idealised vertical sequence of sedimentary structures for storm sands, mimicking the Bouma (1962) sequence. A basal zone of parallel-lamination was succeeded by cross-lamination of a mainly trough-type ascribed to an oscillatory origin along with convolute bedding, succeeded in turn by 'plane-parallel lamination' and finally capped by mud; beds thinned from 20cm proximally to 1cm distally. The upper plane parallel-laminated zone comprised an alternation of sand and plant debris and may be analogous to the uppermost zone of parallel sandstone and mudstone streaks observed at the top of several facies 3 beds. Nelson proposed that the sand units were deposited by rapidly waning flow of wind generated currents resulting in an ebb flow (cf. Hayes 1967), sand suspension being aided by storm-wave liquidation of the substrate. Nelson noted that turbidity currents were neither necessary or hydrodynamically appropriate for the emplacement of shelf storm sands.

Aigner and Reineck (1982), in a study of proximity trends of storm sands in the Helgoland Bight, invoked the following mechanism to explain the emplacement of storm sands. A wind-induced gradient current (cf. Allen 1982) flows offshore at depth to compensate for onshore surface flow during a storm. Flow is concentrated in tidal channels before expanding onto the shelf in a 'jet-like' pattern.

A consideration of modern and ancient examples of storm sands, combined with knowledge of shelf hydrodynamic processes, has led to the erection of several broad classes of model to explain the emplacement of shelf storm sands. The seminal study of Hayes (1967) has been cited by many others, either to suggest that storm-wave generated suspension clouds moved offshore as a density / turbidity current (e.g. Goldring & Bridges 1973; Kelling & Mullin 1975; Vos 1977; Hamblin & Walker 1979; Wright & Walker 1981) or to emphasise the importance of storm surge ebb currents alone (e.g. Goldring & Bridges 1973; Brenner & Davies 1973; Brenchley *et al.* 1979), or a combination of wave-induced currents with storm surge ebb currents (e.g. Reineck *et al.* 1967, 1968; Gadow & Reineck 1969).

The study by Hayes, however, has been critically evaluated by Morton (1981) and McCave (1985), who noted that water levels in the lagoon behind the coastline studied by Hayes actually **fell** during Hurricane Carla. Thus, the requisite pre-conditions to generate Hayes' hypothesised storm surge ebb flow, i.e. a positive water surface 'set-up' at the coastline, would have resulted in an **increase** in water levels in the lagoon. Furthermore, flow measurements taken during the storm are opposite to those implicit in the model of Hayes. In addition, Hayes suggestion of turbidity/density current generation is hydrodynamically unreasonable for typical shelf conditions/morphology, a factor also noted by Aigner and Reineck (1982).

Several authors have suggested that the combined action of storm waves and an ebb tide current may be important in the generation of storm sands (e.g. Goldring & Bridges 1973; Anderton 1976; Reineck *et al.* 1967, 1968; Gadow & Reineck 1969). The universal applicability of this model, however, has been questioned by Allen (1982) on the grounds that evidence for large-tidal flows is relatively rare and the resultant currents are insufficient to account for the volume of sand transport over a wide area necessary to account for sheet storm sand emplacement.

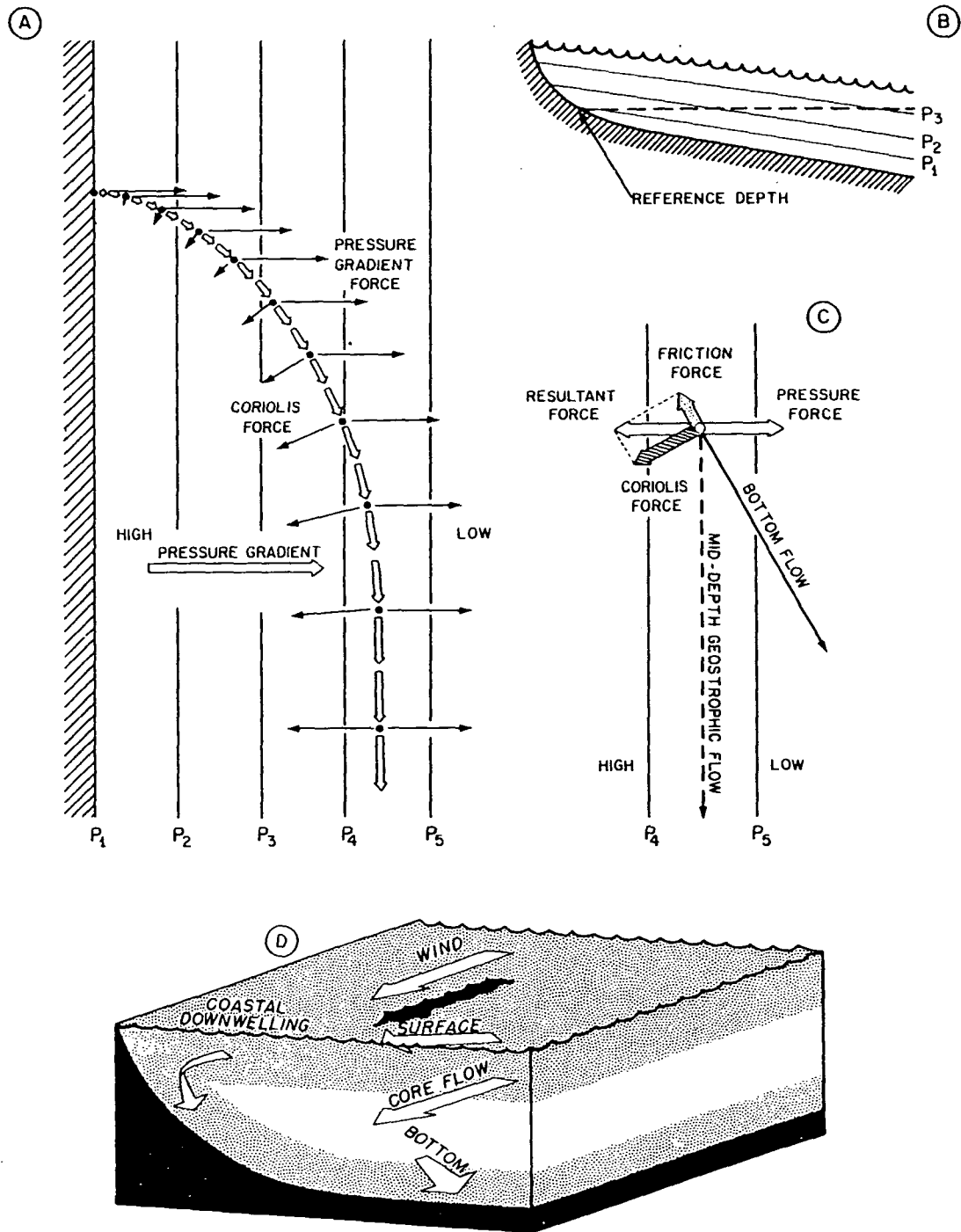
Goldring and Bridges (1973), Vos (1977) and Howard and Reineck (1981) considered the possibility of storm waves alone being responsible for storm sand generation. This model is flawed, however, in that it does not account for the requisite nett unidirectional flow over a wide shelf area, necessary for the generation of storm sheet sandstones.

Allen (1982) proposed a semi-quantitative physical model for storm sand emplacement based upon consideration of the hydrodynamic régime of modern shelf storm events and observations of ancient storm

sandstones. Allen stressed the importance of wave interaction with unidirectional wind-induced currents (cf. Goldring 1966; Goldring & Bridges 1973) during storm events (as opposed to the waning phase of storm events), demonstrating that the resulting hydrodynamic régime adequately explains the lateral extent of both ancient and modern storm sheet sands and the sedimentary structure they exhibit. Allen emphasised that optimum conditions are achieved with sustained onshore-directed winds, which generate a nett surface onshore-directed flow, resulting in 'piling' of water against the shoreline i.e. positive set-up. The positive set-up is compensated by an offshore-directed current immediately above the sea bed; Allen termed this offshore-flowing current a 'gradient current' as it flows parallel to the dip of the mean shelf surface.

In Allen's model, the outer shelf receives laminated muds, the majority of which would be obliterated by post-storm bioturbation cf. facies 2. The mid-shelf is characterised by thin allochthonous sands, interbedded with thicker muds, displaying erosional bases, normal grading and wave-current rippled tops. The inner shelf is dominated by thick allochthonous sands interbedded with thinner autochthonous muds. The bed soles of sand sheets exhibit channelling, rilling and fluting; bed interiors displaying a relatively stable full-storm stage flow, bed tops becoming normally graded. With increasing storm energy, para-autochthonous muds are winnowed to give shell lags with a locally derived faunal aspect. Beds are dominated by parallel-lamination and bed tops exhibit asymmetric wave-current ripples reflecting the wave-induced offshore-directed bottom current. Parallel- to cross-laminated sands are generated proximally; more distally, complete units are cross-laminated (see text-figure 7.2).

Although Allen's model is relatively simple it is effective in explaining structures and trends preserved in ancient shallow marine sheet sandstone sequences. The essentially two-dimensional nature of the model, however, is limiting in that it does not take into account the Coriolis force in developing the characteristic three-dimensional flow typical of shelf storm systems (Strahler 1963, Csanady 1982; Swift *et al.* 1983; Vincent 1986). The resulting three-dimensional model is shown and described in text-figure 7.3. This model is now generally accepted and widely applies to shelf storm systems - see also chapter 4.



Geostrophic flow on the continental shelf. *A*, In this plan view, a parcel of water at a reference depth moves seaward (to the right) in response to pressure-gradient force. As it accelerates, it experiences a Coriolis force impelling it to the right of its trajectory. *B*, Cross-section of hypothetical shelf experiencing geostrophic flow illustrating relation of sea surface slope, isobaric surfaces, and reference depth. *C*, Relation between geostrophic flow and flow in bottom boundary layer. *D*, Block diagram of geostrophic flow, showing landward moving upper boundary layer (stippled), fluid interior (clear), and seaward veering boundary layer (stippled). Modified from Strahler (1963) and Swift (1976a).

Text-fig. 7.3 Geostrophic flow on the continental shelf.

From: Swift & Rice (1984).

In conclusion, facies 3 is attributed to the combined action of storm waves and associated offshore-directed currents generated by onshore directed winds cf. model of Allen (1982) modified to take account of the three-dimensional dynamic system resulting from the Coriolis force (text-figure 7.3). Initial storm conditions produced erosion, followed by net deposition during steady flow conditions at the storm peak. Finally, the top of the bed is wave-current rippled as the storm wanes. The top of the bed is reworked by bioturbation or oscillatory currents. The assemblage of sedimentary structural sequences preserved by facies 3 is very diverse, the majority of the sequences proposed by Allen (1982 - see text-figure 7.2) being preserved in the Lynton Formation. Facies 3 at Woody Bay west and WNW Barbrook occurs in an 'open shelf association', whilst the Little Burland examples are preserved in a 'shoreface-shoreline association'.

7.6 FACIES 4 - HUMMOCKY CROSS-STRATIFICATION

Facies 4 is shown as facies D on the Woody Bay west and WNW Barbrook logs and as facies C on the Little Burland log.

7.6.1 Description

Facies 4 occurs in either single beds set within units of facies 1, or as units of amalgamated beds forming sandstone-bodies referable to the 'Woody Bay facies association' (see section 1.8.3.1.3) i.e. 2 to 4m thick sequences predominantly comprising sets of hummocky cross-stratification (*sensu* Harms 1975). The sandstone-bodies also contain occasional thin partings of lenticular bedding (facies 1), either as isolated pods (plate 7.11C) or as thin (1 to 2cm thick) continuous drapes, e.g. two examples at 16.3m on Woody Bay west log, or pure mudstone drapes (facies 2), which preserve the form of the underlying hummocky topography (plate 7.14A, text-figure 7.5).

Lithologically, facies 4 sandstones are well to very well sorted and of coarse siltstone to fine sandstone grade. This grain size is typical of hummocky cross-stratification (HCS) e.g. Harms (1975) and Bose (1983) reporting fine to very fine sandstone for HCS, Campbell (1966) noting coarse siltstone to very fine sandstone and Cant (1980) recording coarse siltstone to fine sandstone. Petrographically, facies 4 comprised sub-mature sub-litharenites with a matrix of clay and detrital mica, although well sorted examples high in the Little Burland sequence are matrix poor and contain heavy minerals, mainly haematite and tourmaline. Fresh

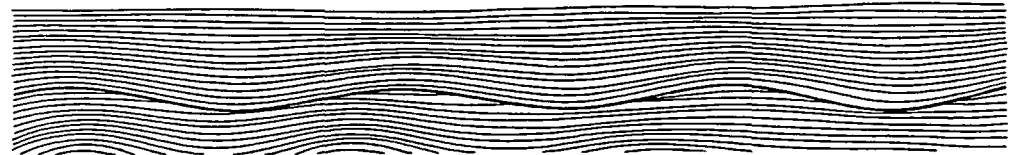
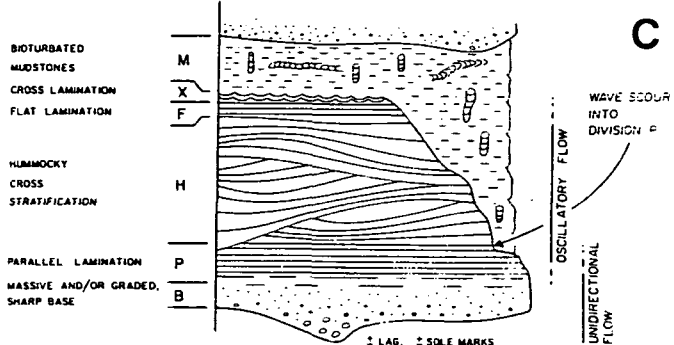
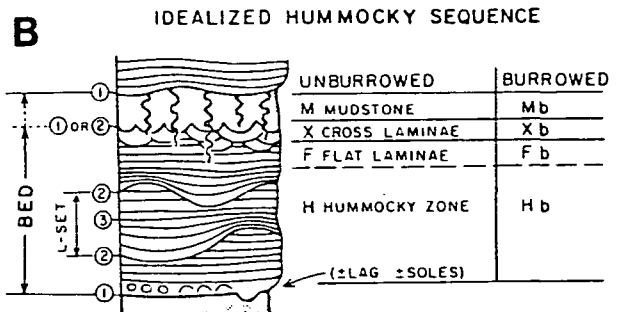
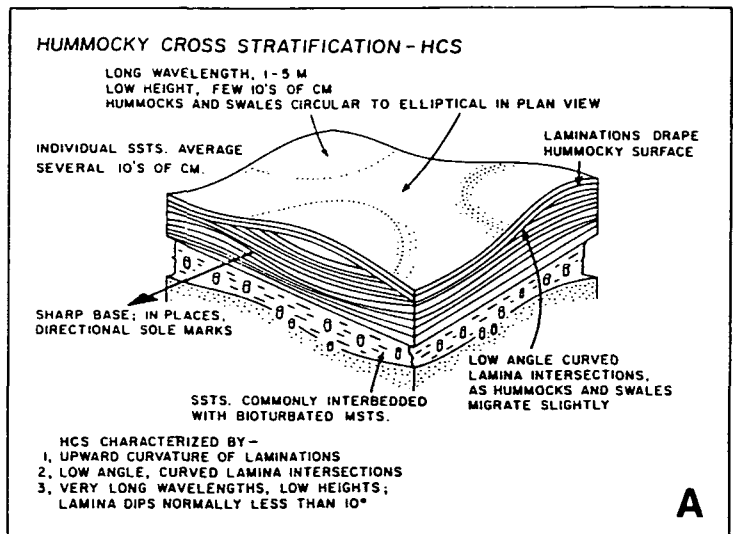
surfaces are medium grey in colour, weathering to a light grey or a pale yellow/green colour; lichen cover is frequent in the WNW Barbrook and Little Burland sections.

The diagnostic features characterising HCS are shown in text-figure 7.4A. In facies 4, individual sets range between 3 and 30cm in thickness, measurements taken from the Little Burland sequence revealing a mean thickness of 10.6cm (n= 135, max.= 27cm, min. = 3cm). Cosets attain thicknesses of up to 265cm, the thicker cosets tending to occur within sandstone-bodies of the 'Woody Bay facies association'. Measurements taken from the Little Burland sequence reveal a mean of 3 HCS sets per bed (n = 54, ave. = 2.9, max. = 10, min. = 1) and a mean bed thickness of 27.5cm (n = 54, =28.6, max. =120cm, min = 3cm). Individual beds are laterally persistent over the width of the exposure. Measurements from facies 4 compare with reported set thicknesses of HCS: 15 to 50cm, average set thickness becoming thinner towards the base of progradational sequences (Harms 1975); 30cm maximum (Campbell 1966); 2 to 20cm (Hamblin & Walker 1979); 15 to 20 cm at the base of a progradational sequence, reaching 90cm at the top of the sequence (Leckie & Walker 1982a); "several decimetres" (Walker & Hunter 1982a).

The lower bounding surfaces of sets are sharp, either non-erosively draping the underlying topography, or incising into the underlying unit (e.g. 11.5m on the Little Burland log) cf. Harms (1975). Erosion surfaces are either planar, with up to 2cm relief, being laterally extensive over the width of the exposure, or curved/undulatory to give a hummocky topography, erosion surfaces dipping in a random manner around the points of the compass at up to 15°, and rarely up to 30°. The resulting hummocky topography comprises broad antiforms with a crestal spacing of 60 to 120cm. These dips compare with 10°, rarely 15° (Harms 1975) and 3 to 6° (Walker 1979), the hummock spacing comparing with 50 to 500cm (Hamblin & Walker 1979) and up to 3m recorded by Cant (1980). Laminae frequently thicken into these scours cf. Cant (1980).

Text-fig. 7.4 Characteristics of hummocky cross-stratification.

- A. Characteristics of hummocky cross-stratification. **From:** Walker (1982).
- B. Idealized vertical sequence for hummocky cross-stratified beds. **From:** Dott & Bourgeois (1979, 1981).
- C. Modified idealized vertical sequence for hummocky cross-stratification. **From:** Walker *et al.* (1983).
- D. Classification of hummocky cross-stratification lamination styles - see text for discussion and Pound (1986 - enclosure 14).



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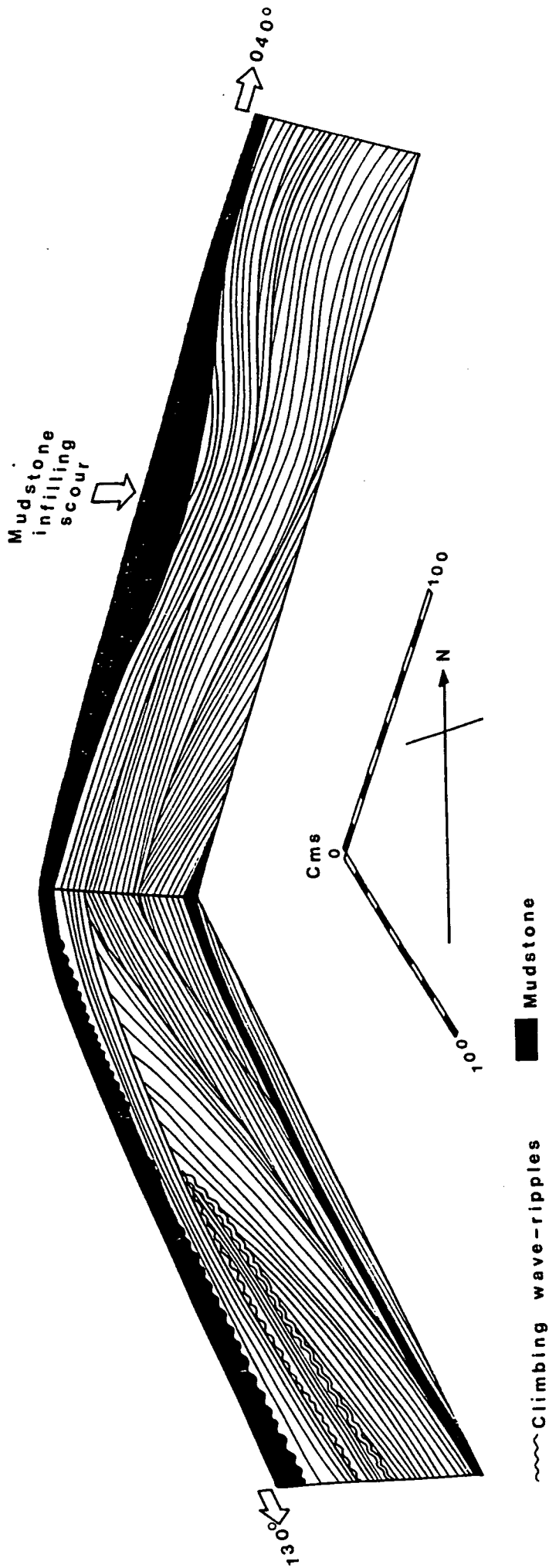


Hummocky Cross-Stratification : Lamination Styles

D

Text-fig. 7.5 Fence diagram through hummocky cross-stratified coset at Woody Bay west.

Grid reference: 6769 4902. The following features should be noted: thin mudstone draping and preserving hummock topography at base of left-hand face; hummocky topography is predominantly of 'scour-and-drape' type, although there is evidence of tangential type development towards the right-hand side of the right-hand face; climbing wave-ripples developed at the top of a hummocky cross-stratified set; the top of the hummocky cross-stratified set is locally wave-rippled, but has also been scoured, the scour having been subsequently infilled by pure mudstone (facies 2).



Occasionally, erosion surfaces pass laterally into zones where hummocky laminae are conformably amalgamated (plate 7.4A). More normally, laminae of preceding HCS sets are truncated at an angle of 5° to 12° (plate 7.9D). Zones of biogenic 'churning' (fossitextura deformativa - see plate 7.11A) or discrete sandstone-filled *Palaeophycus tubularis* burrows frequently penetrate downwards from amalgamation surfaces cf. Hamblin and Walker (1979), Walker and Hunter (1982a), Cant (1980) and Goldring (1984). Troughs between hummocks ('swales') frequently intersect, giving rise to very low angle (3° to 6°) curved intersections of laminae cf. Hamblin and Walker (1979). Isolated pods of lenticular bedding (facies 1) are occasionally preserved in these swales (plate 7.11C).

Dott and Bourgeois (1982) proposed a hierarchical scheme for surfaces within HCS sequences (cf. Campbell 1966). Sharp, erosional basal surfaces were termed 'first-order boundaries', internal truncations of laminae were defined by 'third-order boundaries' - see text-figure 7.4.B in which the numerals enclosed in circles denote the order of the surface. Dott and Bourgeois postulated that where second-order (i.e. 'amalgamation') surfaces merge laterally into zones where laminae are concordant, formation during a continuous depositional (intra-storm?) event was likely, the hummocky surface being generated by a combination of scour and deposition within a brief time span. The presence of pebble, shell or mudstone intraclast lags above second-order surfaces, however, suggested to Dott and Bourgeois that the second-order surfaces can also represent pulses within a single storm event or season i.e. ". . . separated by hours, days, or weeks" (p.666). Goldring (1984, p.9) suggested the following: "Criteria for distinguishing between first-order surfaces of amalgamation from second-order discordances (due to factors inherent to the flow régime)": shell lags; irregular erosion surfaces, often with appreciable relief, and possibly undercutting relationships (indicating considerable erosion and, at least, partial cementation); bioturbation, occasionally terminating against the surface; discontinuous muddy horizons; mica and macerated plant debris concentrations. The frequent presence of biogenic structures descending from second-order surfaces within facies 4 indicates that many of these surfaces represent an appreciable hiatus in primary deposition (inter-storm?), particularly where isolated pods of lenticular bedding are preserved within swales (termed 'M-cut-out type' amalgamation by Dott & Bourgeois, 1982, i.e. M-zone of idealised HCS vertical sequence is cut-out -see text-figure 7.4.B).

Rarely, faint indistinct tool marks are visible on the basal surface of HCS units e.g. 0.5 and 7.3m on the Little Burland log. Although various authors have discerned a directional significance in HCS sole marks i.e. the sole marks are normally parallel to palaeoflow and /or inferred palaeoslope dip direction (e.g. Cant 1980, Leckie & Walker 1982a, Bose 1983), no directional significance could be ascertained from the few facies 4 examples. More frequently at the base of a unit sandstone from the overlying bed is 'piped-down' into burrows at the top of the preceding unit cf. Walker and Hunter (1982a). In addition, several examples of open-looped trails, preserved in convex-hyporelief, have been observed where the HCS unit overlies pure mudstone (facies 2). These open-looped trails are interpreted as originating from pre-depositional tunnels which have been exhumed and cast during the high energy event that emplaced the HCS unit cf. Seilacher (1962).

Bed bases occasionally exhibit load structures which are frequently associated with flame structures between the load casts. The load casts range from 10 to 20cm in width and detrude between 1 and 5cm into underlying muddy lithologies, the laminae within the load cast displaying a simple down-warping. In a rather inaccessible gully on the coast SE of Wringapeak (6730 4944) a 3.4m thick sandstone body of 'Woody Bay facies association' type exhibits extensive, large-scale load structures (plate 7.7). One particular HCS unit overlying a pure mudstone lithology (facies 2) exhibits loading to an extent where isolated load 'pillows' have developed. Similar near-total load deformation of a thick amalgamated HCS sandstone body has been described by Hunter and Walker (1982a).

Internally, the HCS units of facies 4 exhibit a wide range of sedimentary structures. For the purpose of discussion the idealised vertical sequence for HCS proposed by Dott and Bourgeois (1982) and modified by Walker *et al.* (1983) - see text-figure 7.4.B and C respectively - will be used. Proceeding from the base and working upwards, a basal 'B-zone' is occasionally present in the form of a shell lag e.g. 40.2 and 41.0m on the Little Burland log. The shell lags, rather than forming continuous layers of consistent thickness, tend to occur as concentrated lenses concordantly overlain by HCS laminae. The concentrations take the form of either plano-convex lenses at the core of the hummocks (cf. Brechley & Newall 1982), the lenses attaining widths of up to 50cm and heights of 5cm (e.g. plate 7.5B & text-figure 7.8 - see also plate 5.6A from the west Crock Point section described in chapter 5) or as lenses below the apex of intersecting HCS set bounding surfaces (plate 7.5C).

The shell lags contain bivalves, mudstone intraclasts (cf. Bose 1983) and tentaculitids. The valves, although thin, are generally intact but are disarticulated, the majority being convex-upwards i.e. in a hydrodynamically stable position (Clifton 1971). The presence of occasional concave-upwards valves tends to correspond with mud clinging to the valves, presumably affecting their hydrodynamic stability (Plate 7.5A). Goldring and Langenstrassen (1979) also recorded mud clinging to valves within storm layers, suggesting that this phenomenon resulted from the valves being derived from erosion and winnowing of an underlying muddy substrate. The occurrence of this phenomenon within an amalgamated HCS sequence of facies 4 suggests the loss of fair-weather muddy sediments cf. 'time-compensated' amalgamated sheet sandstones *sensu* Goldring (1984).

A fauna was recovered from two localities in the Little Burland sequence and was examined by Dr. David Butler (ex British Geological Survey). The first locality - 40.2m on the log, sample registered as BGS ZO 4846-4865 - yielded a fauna comprising valves of the nuculoid bivalve *Carydium* aff. *rugosum* (Haffer), *Palaeonucula* cf. *krachtae* (Roemer), bryozoans?, a crinoid columnal and a tentaculitoidean? The second locality - 41.0m on the log, sample registered as BGS ZO 4829-4845 - yielded a fauna comprising moulds of bryozoans and *Carydium* aff. *rugosum*. Butler ascribed a "latest Emsian or Middle Devonian" age to these faunas. Butler noted that the *C.* aff. *rugosum* specimens appear to be intermediate in form between *C. sociale* and the early Givetian *C. rugosum* and exhibit concentric folds used to strengthen the valves against stress. The concentric folds are better developed in comparison to specimens of *C. sociale* recovered by Butler from Dean Steep (200m west of the WNW Barbrook section). Butler suggested that the Little Burland specimens were probably derived from a higher energy environment than the Dean Steep specimens. This hypothesis is reinforced by the dentition of the Little Burland specimens which indicates a better internal purchase of one valve against the other, so that there is less chance of the valves rotating in relation to each other.

Occasionally the base of an HCS bed is massive, draping the topography of the preceding unit e.g. at 21.4m on the WNW Barbrook log wave-ripple profiles are preserved beneath a thin drape of massive sandstone. Equally rarely, where the base of a unit is planar a thin zone of parallel (i.e. horizontal) lamination occasionally develops cf. Hamblin & Walker (1979), Walker *et al.* (1983) and Brenchley (1985). Developments of this type were termed 'P-zone' by Walker *et al.* (*op. cit.* - see text-figure 7.4C). Hamblin & Walker (*op. cit.*) attributed the parallel-lamination to an upper-phase plane bed origin analogous to Bouma division B, an interpretation followed by Walker *et al.* (*op. cit.*). Brenchley (*op. cit.*), however, noted that the

parallel-lamination nearly always lacks primary current lineation and suggested that the laminae are generated by rapid deposition from suspension.

Hummocky laminae directly overlying the basal surface of a set frequently approach the basal surface asymptotically; this lamination style is herein termed 'tangential type' (see text-figure 7.4D) and infers both lateral and vertical accretion of the hummock cf. Dott and Bourgeois (1982). In situations where the basal surface defines the flat top of a muddy unit and the sand supply was limited, isolated hummocks are preserved - see plate 5.6A and B for examples preserved within the more distal west Crock Point section (discussed in chapter 5). More commonly, the sand supply was sufficient to allow the hummocks to become connected (see text-figure 7.4D and plate 7.9D). Where the lower set bounding surface is itself hummocky as a result of differential scour, laminae also commonly approach this surface tangentially cf. Campbell (1966), Hamblin and Walker (1979) - see plate 7.4B. The draping of differentially scoured surfaces is the most common form of lamination style observed within HCS both within the Lynton Formation and from elsewhere e.g. Dott and Bourgeois (1982), Bourgeois (1983) and Brenchley (1985); this lamination style is herein termed 'scour-and-drape' - see text-figure 7.4D.

Laminae within scour-and-drape type HCS may either approach the differentially scoured underlying surface tangentially (as described above) or parallel the surface cf. Harms (1975). Differentially scoured surfaces almost universally overlie a sandy lithology (text-figure 7.5), although differential scour into muddy lithologies is occasionally visible (plate 7.4.B) cf. Brenchley (1985, figure 9). Laminae are frequently parallel (i.e. everywhere equidistant), but also frequently thicken laterally into swales (plate 7.4A) cf. Bose (1983). Thickening into swales as the top of the hummocky unit is approached leads to a gradual flattening of the hummocky topography and was termed 'F-zone' by Dott and Bourgeois (1982) - see text-figure 7.4B.

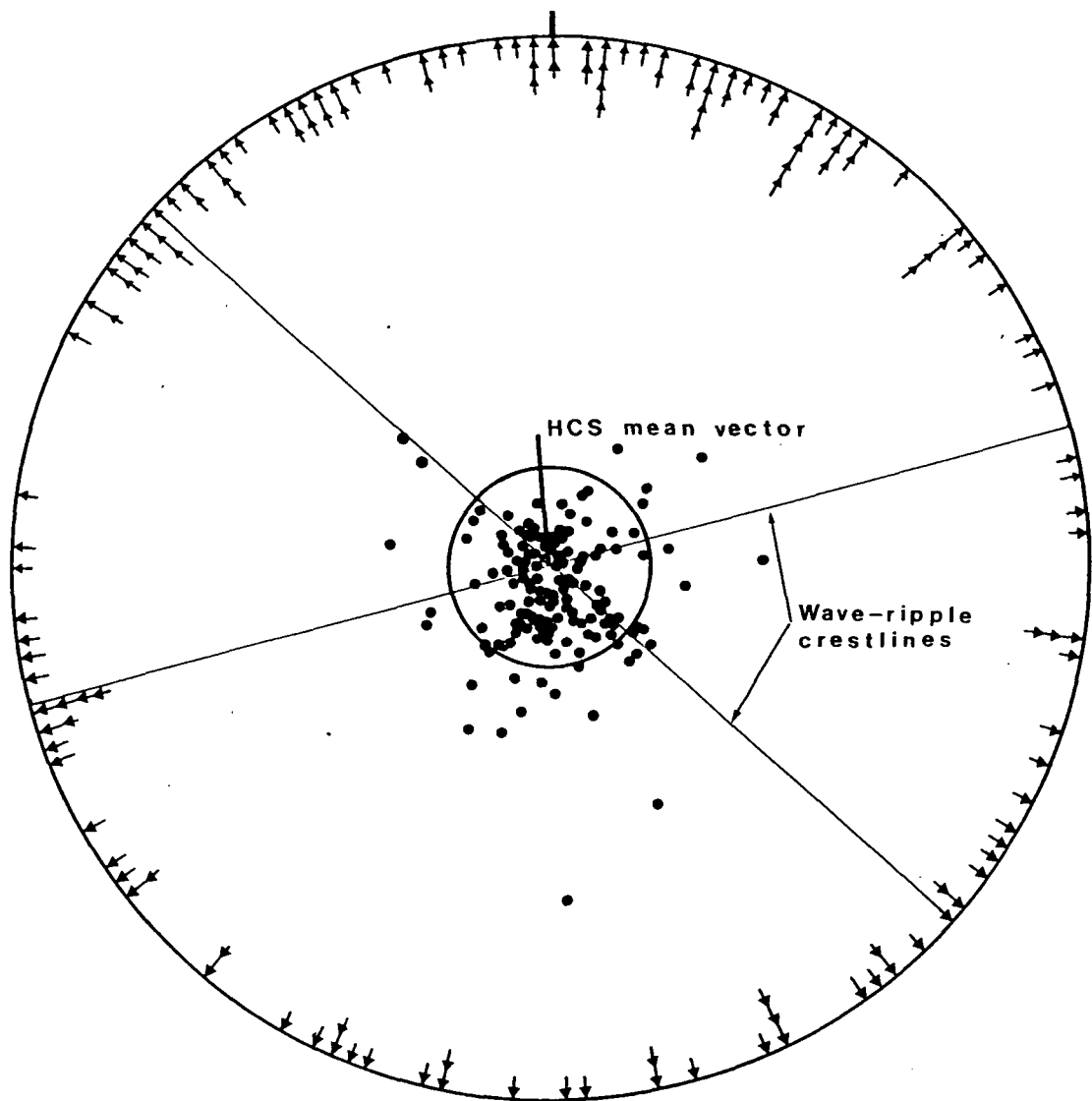
Very rarely, laminae thicken over hummock crests to give a third type of HCS lamination style termed 'vertical accretion type' herein - see text-figure 7.4D. Campbell (1966), Brenchley and Newall (1982), Hunter and Clifton (1982), Brenchley (1985) and Allen and Underwood (1985) have described laminae thickening over hummock crests, the latter three authors along with Bourgeois (1983) all noting that this is the rarest HCS lamination style. Furthermore, Dott and Bourgeois (1982) caution that vertical accretion type HCS may not be the product of true 'fanning' (cf. Harms 1975) of laminae, but the result of many minor laminae discordances. Careful examination of facies 4 examples confirms that the laminae thickening over

hummock crests is genuine (plate 7.9A collected from locality featured in plate 7.4A). Some sets exhibit a vertical transition from laminae thickening over crests to laminae thinning over crests (plate 7.4A) i.e. a transition from vertical accretion type hummocks to F-zone laminae.

Considering all facies 4 lamination styles together, laminae range from 1 to 4mm in thickness (average 2mm) and have a 'flaggy' fissility. Lamina grading is not generally apparent due to the well sorted nature of the sandstones. Occasionally, however, a density grading is visible, with micaceous grains concentrated at the top of individual laminae (plate 7.9B). Density grading of hummocky laminae has also been described by Levell (1980b) and Dott and Bourgeois (1982). Rarely, 3 to 8mm long mudstone intraclasts are visible (plate 7.9C) cf. Bose (1983).

Measurements of the dip of hummocky laminae at Woody Bay west (text-figure 7.6) and Little Burland (text-figure 7.7) reveal a random disposition around the points of the compass, a phenomenon characteristic of HCS (Harms 1975; Hamblin & Walker 1979; Bourgeois 1980; Dott & Bourgeois 1982; Brenchley 1985). Laminae generally dip at up to 15° (cf. Harms 1975; Hamblin & Walker 1979; Dott & Bourgeois 1982; Bose 1983; Brenchley 1985), although some laminae attain dips of up to 35°. The steeper laminae tend either to be the product of scour infill (plate 7.8D) or soft sediment deformation. Dips of up to 35° for HCS laminae have also been recorded by Hunter and Clifton (1982) and Brenchley (1985).

The Woody Bay west exposure is valuable in that hummocks are visible in plan. The majority appear to be circular (plate 7.8A), although some are elongated in an approximately E-W direction i.e. sub-parallel to palaeoslope strike (plate 7.8B). Hamblin and Walker (1979), Leckie and Walker (1982a) and Brenchley (1985) have recorded circular hummocks, whilst elongated hummocks have been recorded by Campbell (1966, 1971), who records elongation parallel to the shoreline, Hamblin and Walker (1979), who record elongation at 45° to shore-normal sole marks, and Cant (1980).



HCS:

Arrows= dip directions

Dots= poles to hummock surfaces

L= Estimate of spread of angular values=24.3%

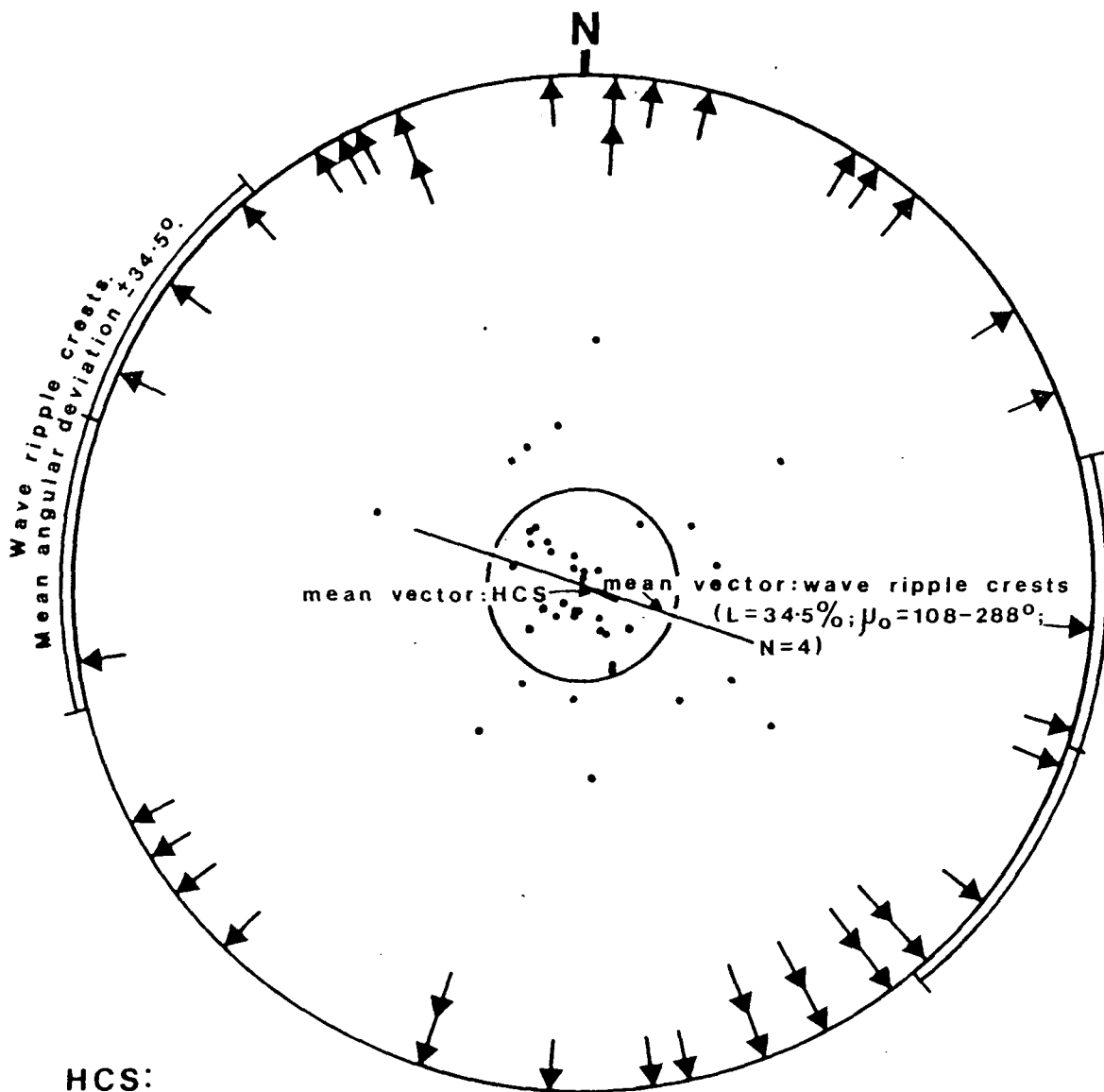
μ_0 =Mean vector direction=356°

s= Mean angular deviation=±71°

Rayleigh test of significance=10^{-3}:within non-random field

N=151

Text-fig. 7.6 Hummocky cross-stratification lamina dip - Woody Bay west



HCS:

Arrows = dip directions.

Dots = poles to hummock surfaces.

Dots within small circle represent dips $< 15^\circ$.

L = Estimate of spread of angular values = 7.6%.

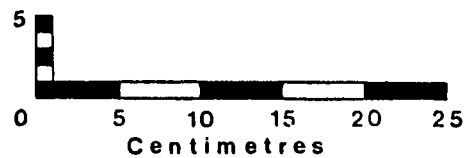
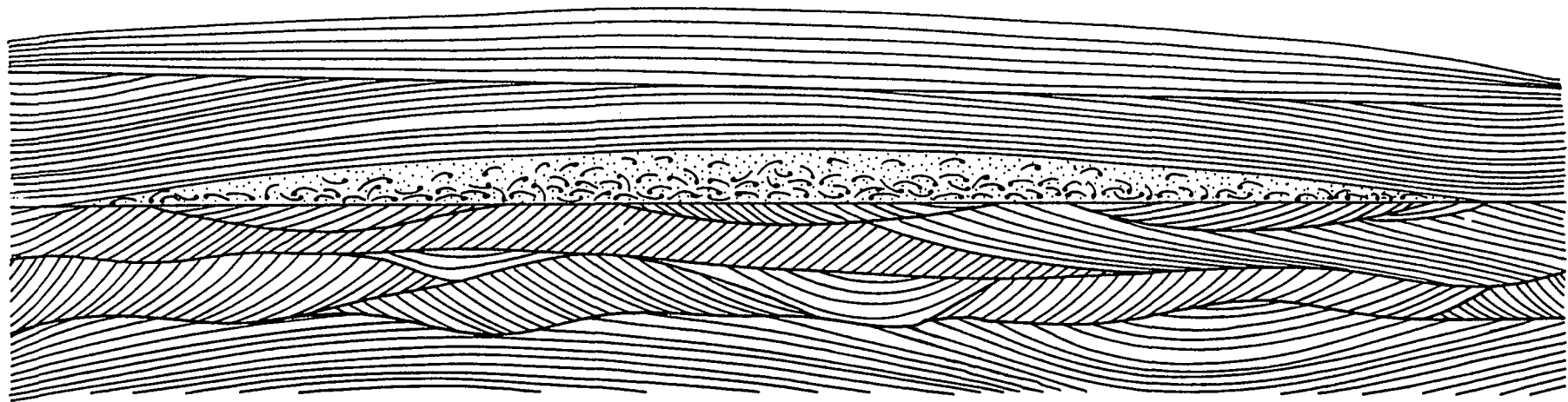
μ_0 = Mean vector direction = 112° .

s = mean angular deviation = $\pm 78^\circ$.

Rayleigh test of significance = $> 0.7 \therefore$ random.

N=40.

Text-fig. 7.7 Hummocky cross-stratification lamina dip - Little Burland



Text-fig 7.8 Field sketch made at Little Burland showing a plano-convex 'swell lag' concentration at the core of a hummock in facies 4.

Grid reference: 6629 4949. Note that the majority of valves are convex-upwards and are supported in a fine sandstone matrix. The shell lag is conformably draped by hummocky laminae. The shell lag succeeds a zone of wave-ripple cross-lamination at the top of the preceding HCS unit.

Plan views of hummocky surfaces occasionally reveal a grain lineation (plate 7.9C), the lineation direction paralleling the wave oscillation direction recorded by wave-ripple crestlines preserved at the top of the unit e.g. 21.8m on the WNW Barbrook log. Similar grain lineations over hummock surfaces have been recorded by Cant (1980), Levell (1980b) and Brenchley (1985). Goldring and Bridges (1973) undertook a particle orientation analysis of a sublittoral sheet sandstone and found that the primary elongation mode was parallel to the palaeocurrent, whilst a secondary mode was observed perpendicular to the palaeocurrent. The coincidence of primary current lineation and wave oscillation direction supports, at least in part, a wave origin for the laminae, a conclusion also reached by Brenchley (1985). It should be noted, however, that at the time the lineated laminae were deposited, the oscillatory currents may have been either strongly asymmetric or have had a strong unidirectional current superimposed. Nevertheless, the occurrence of grain lineation on foreshores is clearly oscillatory in origin; it seems probable that these lineations could also form subaqueously (Dr. H. E. Clifton, U.S. Geol. Survey, Menlo Park - pers. comm.) Oscillatory flume observations made by Dr. T. Astin (Reading University - pers. comm.), indicated a strong grain lineation develops in vortex ripples of medium sand at depths of 10 to 20 cm and wave periods of between 0.9 and 1.1 seconds. The grain lineation was partly the product of grain long-axes parallel to the water motion and partly due to size sorting into coarser and finer bands cf. primary current lineation of unidirectional origin

Although the above observations suggest that the grain lineation in facies 4 may result from oscillatory flow (which was possibly either strongly asymmetric or had a strong unidirectional current superimposed), caution must be exercised as the lineation may be of a purely unidirectional current origin. Indeed, Collinson and Thomson (1982) figure a hummocky bedform closely resembling HCS and exhibiting a strong grain lineation. Collinson (pers. comm.) stated that the photograph was taken at the type locality of the Moraenesø Formation in the Wandel Valley of NE Greenland (Upper Proterozoic). The sequence within which the hummocky bedform was preserved represents the sandy fill of a palaeo-valley containing a wide, low-sinuosity fluvial system which experienced a very 'flashy' discharge. Collinson also reported that he had observed similar lineations on hummocky bedforms in a late Precambrian fluvial succession on the north side of Verangerfjord, N. Norway. Similarly, Rust and Gibling (1990) reported antidunes from a Westphalian fluvial sequence in Nova Scotia with primary current lineation of unidirectional flow origin. Cheel (1991) undertook a detailed analysis of grain fabrics preserved within HCS and found that within individual beds the basal laminae grain fabric was the product of unidirectional flow with a superimposed

oscillatory component. Higher in the bed the grain fabric was of oscillatory origin. Cheel concluded that HCS was the product of waning unidirectional flow superimposed on storm waves.

Measurement of hummock dimensions within the Little Burland sequence revealed a mean wavelength of 99.3cm ($n = 6$, $\sigma_{n-1} = 15.3$, max. = 120cm, min = 76cm) and a mean amplitude of 6.8cm ($n = 5$, $\sigma_{n-1} = 3.3$, max. = 12cm, min. = 5cm). Measurement of isolated tangential type hummocks (cf. text-figure 7.4D) in the more distal west Crock Point section (see Chapter 5) revealed wavelengths in excess of 150cm (plate 5.6A). The preceding measurements compare with measurements quoted in the geological literature of (wavelength and amplitude respectively): 10cm to 5m and 1 to 50cm, ripple indices of 4 to 17 (Campbell 1966); 5 to 15ft. and an average 10 inches (Campbell 1971); one to a few metres and 10 to 50 cm (Harms 1975); 1 to 5m and normally 10 to 20cm, seldom exceeding 30 to 40cm (Hamblin and Walker 1979); up to 5m and 10 to 15cm (Cant 1980); 1 to 5m and a few 10's of cms. (Leckie & Walker 1982a); 0.7 to 1.7m and 5 to 15cm (Bose 1983); up to 4m and up to 1m (Brenchley 1985). Structures described by Gilbert (1899) that resemble HCS had wavelengths of 1 to 10m and amplitudes of 15 to 100cm.

It should be noted however that the preservation potential of hummocks is lower than the intervening swales, hummocks frequently being planed-off by erosion. This is particularly true of hummocks in amalgamated sandstone bodies cf. Brenchley (1985). It was this phenomenon that led Campbell (1966) to coin the original name for HCS as 'truncated wave-ripple lamination'.

Caution must be exercised when perusing the geological literature of HCS as it appears that the term has been applied to structures which do not conform to the generally accepted diagnostic features for the structure, particularly dimensions. For example, Reineck (1976 - plate 1, figure 2), Vos (1977 - figure 9D) and Brenchley (1985 - figure 5) all figure structures with amplitudes and wavelengths of approximately 2cm and 10cm respectively, attributing the structures to HCS. It appears that the structures should more correctly be assigned to a 3-dimensional wave-ripple origin.

HCS laminae in facies 4 are frequently contorted by convolute lamination, a common feature of shelf sheet sandstones (Goldring & Langenstrassen 1979). The convolutions occur in zones attaining maximum thicknesses of 22cm, occasionally disturbing the complete bed e.g. 42.7m on the Little Burland log. The amplitude of the convolutions tends to first increase and then decrease upwards within a set cf. Allen (1982).

The convolutions, when viewed in three-dimensions, are dome-shaped (plate 7.10A), although zones of more disorganised convolutions are occasionally visible (plate 7.10B). Convolutions attain wavelengths of 30cm, and amplitudes of 5cm. The axis of symmetry of convolutions is generally near vertical, although some examples have axes inclined towards the SE (e.g. plate 7.10A) i.e. in the same direction as the mean palaeo-current. Laminae are occasionally punctured at their crest, presumably to form a pathway for water escape (e.g. 44.1m on the Little Burland log). Convolutions are frequently truncated by penecontemporaneous erosion (plate 7.10A) cf. Dott & Bourgeois (1982), with truncation occasionally by a succeeding set within a coset.

Using the nomenclature of Allen (1982), the majority of convolutions are metadepositional i.e. generated just before or immediately after deposition ceases (distinguished by penecontemporaneous erosion of the top of the convolution.) Where no truncation is exhibited, it is possible that the convolution was post-depositional in origin. Conversely, the occurrence of convolutions truncated by second-order surfaces (*sensu* Dott and Bourgeois 1982) within cosets may indicate a syndepositional origin for some of the convolutions

Although seismic activity (cf. Allen & Banks 1972; Brenchley & Newall 1977; Dalrymple 1979; Allen 1982) may have provided the requisite 'trigger' for liquefaction resulting in the development of convolute lamination, there being evidence for syn-sedimentary movement on the Lynmouth - East Lyn Fault (see section 2.2), the widespread occurrence of metadepositional and syndepositional convolutions, and their frequency of occurrence in facies 4, suggests an origin associated with bed deposition. A review of convolute laminae in shallow water sequences by Dalrymple (1970) indicated that liquefaction can be generated by **wave-induced stress** acting upon the bed. Although Allen and Banks (*op. cit.*) have shown that **shear-stress** of the bed alone cannot produce liquefaction, the shear-stress produced by an offshore-flowing current may have been sufficient to tilt the vertical axis of the convolution figured in plate 7.10A in an offshore direction, the convolution itself being generated by wave-induced stress acting vertically on the bed.

As noted previously, several examples of hummocky laminae diminishing in amplitude towards the top of sets, resulting in a flattening of the hummocky topography, have been observed e.g. plate 7.4A and 16.4m on the Woody Bay west log. This lamination style, termed F-zone laminae, does not exhibit any primary current lineation, a phenomenon suggesting an origin resulting from fall-out from suspension cf. Bose (1983) and Brenchley (1985). Very rarely the F-zone laminae display an alternation of 1 to 3mm thick fine and coarse

grained laminae, the finer grained laminae ranging from mudstone to coarse siltstone in grade - this alternation occupying a zone up to 5cm in thickness at the top of a bed e.g. 7.3m on the Little Burland log. This alternation of coarse and fine laminae at the top of the bed is analogous to the top of several facies 3 beds - see section 7.5. The F-zone is frequently overlain directly by pure mudstone of facies 2 in the Woody Bay west section e.g. 16.5m, on the log.

The top of HCS beds very commonly exhibits a zone of cross-lamination diagnostic of generation by an oscillatory or combined-flow. Laminae are generally form-discordant with respect to ripple profiles preserved at bed tops, although form-concordant laminae are occasionally visible. Adjacent ripples frequently display dissimilar lamination patterns; bundled up-building of lamina sets is common.

At Woody Bay west several examples of undulatory laminae with wavelengths of 5 to 7cm and amplitudes of 4 to 8mm, rounded troughs and less rounded or trochoidal crests, were observed close to the top of HCS units. The laminae are parallel (i.e. equidistant) and the axis drawn through successive lamina ripple crests is sub-vertical (75° to 80°) - see plate 7.13B. This lamination style closely resembles that described for supercritical climbing wave-ripples by Allen (1982). Climbing wave-ripples within the 'x-zone' (see text-figure 7.4B) of HCS have previously been reported by Dott and Bourgeois (1982); Bose (1983) has recorded climbing current-ripples in the X-zone. The presence of climbing wave-ripples is attributed to high rates of fall-out from suspension during the waning phases of the event that emplaced the HCS.

The majority of plan-views available at the top of rippled facies 4 HCS beds exhibit a complex arrangement of interference ripples (cf. Cant 1980; Brenchley 1985), indicative of complex and involved formative wave patterns, and "micro-hummocks" (cf. Brenchley 1985) attributed to a '3-D vortex wave-ripple' origin (see section 4.2.7.3 for discussion of this type of wave-ripple). Where linear wave-ripple crestlines are observed they trend between ENE-WSW and SE-NW at both Woody Bay west (text-figure 7.6) and Little Burland (text-figure 7.7). The linear wave-ripple profiles are generally symmetric (cf. Hamblin & Walker 1979; Leckie & Walker 1982a), but may also be asymmetric occasionally (cf. Hamblin & Walker 1979) indicating a mean SE -directed (offshore) flow component.

Several examples of differential scour into the top of HCS, to produce a hummocky topography, have been observed (cf. Brenchley 1985). These scours are either infilled with pure mudstone of facies 2 (e.g. text-

figure 6.5) or lenticular bedding of facies 1 (e.g. plate 7.11B). In other cases planar erosion surfaces, with up to 2cm relief, have been observed at the top of HCS sets (cf. Bose 1983) and are overlain by lenticular bedding of facies 1 (plate 7.4A).

Very rarely, 'gutter casts' (*sensu* Whitaker 1973), ranging up to 12cm in width and 2cm in depth and having steep sides of up to 45° (cf. Brenchley 1985) mud rounded bases, are observed. The gutter casts incise into the top of HCS beds and are infilled with either mudstone or sandstone of the succeeding unit. In one example (plate 7.6A) the gutter cast has been filled preferentially from one side resulting in a laminated sandstone infill dipping at an angle near to the 'angle of repose' for sand avalanching cf. Aigner (1982 - figure 9A).

Scours at the top of sublittoral sheet sandstones are common (Goldring & Aigner 1982) and frequently exhibit evidence that scour formation is unrelated to the formation of the underlying HCS (Brenchley 1985).

In facies 4, biogenic structures commonly descend from the top of HCS beds (cf. Hamblin & Walker 1979; Cant 1980; Walker & Hunter 1982a), generally resulting in reworking of the originally sharp bed top to give a gradational interface with the succeeding unit (cf. Brenchley 1985). Bioturbation generally either occurs in the form of discrete 5 to 7mm diameter, sandstone-filled *Palaeophycus tubularis* descending at angles of 45° to 80° from the horizontal, or biogenic 'churning' (*fossitextura deformativa*) - see plate 7.11A. Bioturbation tends to be at a maximum at bed tops and frequently results in a complete localised loss of primary structures (e.g. 45.0 and 44.7m in the Little Burland log). The presence of biogenic structures immediately below set boundaries (amalgamation surfaces) within HCS cosets demonstrate that an appreciable period of time must have lapsed prior to deposition of the overlying set (cf. Goldring 1984) i.e. sufficient time for the preceding set to be colonised and bioturbated; the converse does not hold true i.e. the absence of bioturbation does not necessarily indicate the absence of a significant gap in time between deposition of two sets - the absence of bioturbation may merely indicate that the surface was not colonised.

Occasionally, structures resembling biogenic vertical escape trails (cf. Goldring 1971, plate 6f; Schäfer 1972, fig. 220; Howard 1978) are visible within HCS beds. These structures indicate that deposition of the HCS bed was rapid. Where bedding plane views of wave-rippled tops of HCS. units are available, sinuous convex and concave epirelief trails up to 4mm in width are occasionally visible e.g. 23.0m on WNW Barbrook log.

The Lynton Formation HCS units lack a distinctive trace fossil suite; this is in contrast to the assemblages reported by other workers e.g. Frey (1990), Pemberton *et al.* (1992).

In comparison with the schemes for idealised HCS bed sequences proposed by Dott and Bourgeois (1982) and modified by Walker *et al.* (1983) - see text-figure 7.4 B and C respectively - facies 4 exhibits a wide range of variability. Amalgamated H (e.g. 8.4m on the Little Burland log) and Hb (e.g. 31.6m on the Little Burland log) units are very common, as are Hmb (e.g. 21.2m on the Little Burland log), HM (e.g. 15.0m on the Woody Bay west log) and HXM units (e.g. 14.1m on the Woody Bay west log), the latter being particularly prevalent in the Woody Bay west section; HX units (e.g. 7.7m on the Little Burland log) are somewhat less common. The following variants only occur rarely: HF (e.g. plate 7.4A and 16.4m on the Woody Bay west log), BHX (e.g. plate 7.5B), BH (e.g. plate 7.5C), PHXM and PH (e.g. 13.8m on the Woody Bay west log); these variants occasionally exhibit biogenic structures descending from the upper part of the unit. It is of interest to note that the Jenyncliff Slates (Eifelian), cropping out along the eastern side of Plymouth Sound (see Pound 1983 - enclosure 10), frequently exhibit HCS units of the HFM variety (see plate 7.18B).

With reference to Figure 8 of Dott and Bourgeois (1982), amalgamated HCS sequences are occasionally of the 'M-cut-out type' (e.g. plate 7.11C) or 'lag type' (e.g. plate 7.5B & C), whilst 'truncated contortion' (e.g. plate 7.10A) and 'bioturbated type' (e.g. plate 7.11A) units are common.

7.6.2 Interpretation

Hummocky cross-stratification was first described as a distinct primary sedimentary structure by Campbell (1966 & 1971) who proposed the name of 'truncated wave-ripple lamination' for the structure. The term 'hummocky cross-stratification' was first coined by Harms (1975), this now being the term by which the structure is generally known, although other authors have referred to 'hummocky stratification' (e.g. Dott & Bourgeois 1980); HCS is a sub-type of Goldring and Bridges (1973) 'sublittoral sheet sandstone' class. The term 'swaley cross-stratification' has been applied to amalgamated units of HCS where only the swales are preserved, swaley cross-stratification tending to occur above HCS and below beach deposits in progradational sequences e.g. Leckie and Walker (1982b) and Hein (1982). HCS occurs in sequences ranging from Lower Cambrian (Hein 1982) to Pleistocene (Duke 1984) in age - see Dott and Bourgeois

(1982) for a list of examples - and has been recorded from the following environments: prograding shorelines (Campbell 1971; Harms *et al.* 1975; Hamblin & Walker 1979; Hein 1982; Leckie & Walker 1982a; Rosenthal 1982; Walker & Hunter 1982a, b, c) including muddy shorelines (Cant 1980), transgressive shorelines (Bourgeois 1980; Bose 1983), tidal channels (Campbell & Oaks 1973), estuaries (Campbell & Oaks 1973), deltas (Vos 1977) and lakes (Duke 1984, Eyles & Clark 1986). Additionally, structures identical to HCS in morphology, although not referred to as HCS by the original author, have been reported from environments as diverse as fluvial (Collinson pers. comm. - see earlier discussion on grain lamination in HCS for references; Rust & Gibling 1990, Cotter & Graham 1991) and a flysch sequence deposited below the calcite compensation depth (Campanian to early Maastrichtian Flysch, Dranses Nappe, Plagersfluh Switzerland - pers. comm. Dr. P. W. Homewood, Institut de Geologie, Fribourg, Switzerland). Possible examples of HCS in an intertidal setting are preserved within a Palaeozoic sequence in Wisconsin, U.S.A. (Dott & Byers 1981) led the authors to comment that: "HCS is still a newly known feature, and perhaps it is not so environmentally circumscribed as assumed" (p.333). Furthermore, the occurrence of HCS is not restricted to siliciclastic sequences, examples of HCS in carbonate sequences having been recorded by several authors e.g. Kreisa (1981), Aigner (1982) and Wu (1982).

During the past two decades recorded occurrences of HCS have increased dramatically. The presence of sharp, frequently erosional bases to HCS, internal organisation into an idealised, apparently waning flow, sequence cf. Dott and Bourgeois (1982) and Walker *et al.* (1983), and bioturbation descending from bed tops cf. Howard (1972) and Goldring and Bridges (1973), has led to the almost universally accepted interpretation of HCS beds as 'event deposits' i.e. representing a single depositional event. Furthermore, the shallow water, mainly marine, setting of sequences containing HCS has led to the widely accepted hypothesis that HCS was emplaced during storm events i.e. 'tempestites' (*sensu* Ager 1973). The absence, however, of unequivocal contemporary examples of HCS, from either natural settings or flume studies, has resulted in a vigorous debate regarding the nature of the hydrodynamic régime that prevails during the deposition of HCS.

At the time of writing there is no universally accepted model relating the genesis of HCS to a specific hydrodynamic régime. The search for such a model has been hampered by the absence of unequivocal examples of HCS in studies of contemporary environments. Campbell (1971) reported a structure resembling HCS within a trench through a longshore bar, commenting: "that this structure may commonly form in the surf zone bordering beaches" (p.80). It is difficult, however, to extrapolate Campbell's observations to

examples of HCS in the geological record where deposition took place distal to a shoreline/surf zone (e.g. Walker 1983) in an environment where the hydrodynamic régime would have been significantly different. Campbell's observation does serve, however, to significantly emphasise that structures matching HCS in geometry are almost certainly polygenetic in origin - a conclusion necessary to explain the diverse range of environments from which HCS has been recorded in the geological record (see above).

Harms (1979) reported scuba diving observations made during storm activity at depths of 5 to 15m. Although HCS was not observed, fine sand was suspended during the peak current of each wave surge which then fell from suspension to form a thin lamina draping the underlying topography. Harms suggested that HCS may form in a similar manner.

Geehan (1979 - oral comm. reported in Dott & Bourgeois 1982) and Clifton (in: Hunter & Clifton 1982) both reported diving observations made immediately following winter storms which revealed the presence of hummocky surfaces. The latter author observed hummocky bedforms in fine to medium grained sand, the hummocks having a relief of up to 10cm in water depths 4 to 5m. In the absence of more detailed observations and systematic box coring for the above two studies, interpretation of these observations as HCS must remain equivocal.

Howard and Reineck (1981) have suggested that much of the parallel-lamination observed in cores recovered from many offshore-shoreface transition environments probably represents HCS. Howard and Reineck recognised, however, that it is not presently possible to recover cores of a sufficient size in relation to HCS to confidently diagnose the presence of HCS.

Finally, Swift *et al.* (1983) reported HCS-like structures from side-scan sonar and scuba observations made during studies of the storm-dominated inner Atlantic continental shelf off the eastern coast of the U.S.A. The hummocky structures were only observed in very fine to fine sand and did not exhibit any apparent slip-face, the steepest slopes not exceeding 12°. The hummocky pattern was short lived, only being visible immediately post-storm, the structure becoming rapidly erased or buried. Vibracores taken through the structure revealed sub-horizontal laminae, whilst box cores exhibited low-angled reversals of dip and laminae truncation. Duke (1985), however, noted that tracings taken from box cores and shown in figure 5 of Swift *et al.* display a maximum apparent dip of 31°, much greater than the 10 to 15° typical of HCS. Duke

suggested that the structures described by Swift *et al.* were washed-out megaripples. Swift *et al.* observed that the hummocks occurred in zones which were mutually exclusive to zones of straight-crested megaripples, tending to occur in zones which experienced intense flow acceleration during storms, occasionally attaining near-surf conditions. Again, in the absence of more detailed observations, including systematic grid box coring, these structures cannot be unequivocally presented as modern examples of HCS.

Several authors have attempted to calculate the frequency of storm sandstone layers in specific stratigraphical intervals in an effort to gain an insight into the nature of the processes that emplace thick storm sand layers. Reported recurrence intervals (see table 7.2) range between 400 and 15 000 years with an average of approximately 5 000 years. Great care, however, must be exercised when attempting to place an interpretation on these figures. For example, Reineck (1960b) estimated that only between ten-thousandth and one-hundred-thousandth of sediment deposited in modern tidal seas is actually fossilised. Although these figures will vary widely, dependent on factors such as subsidence rate, base level changes, sediment supply etc., it is clear that preservation potential must be considered before attributing any particular significance to the storm sand frequency quoted above. Nevertheless, several authors have used the above type of figures to suggest that the apparent frequency of HCS layers indicate 'rare' events. The influence of studies in the Gulf of Mexico (e.g. Hayes 1967) has led to the proposal that HCS is the product of hurricane-force winds (e.g. Kreisa 1981, Mount 1982). Palaeogeographical analysis of ancient inferred storm depositional systems (Marsaglia & Klein 1983; Duke 1985) indicates that only 70% or 73%, respectively, of these systems were deposited in latitudes subject to hurricanes whilst the remainder occurred in latitudes subject to winter storms. It is clear, therefore, that hurricanes are not a prerequisite for the formation of HCS.

| Stratigraphic Unit | Author | Calculated Recurrence Interval, Years |
|--------------------------------------|--------------------------------|---------------------------------------|
| "Passage Beds" (U. Jurassic) Alberta | Hamblin & Walker 1979 | 3 200 |
| Devonian, Germany | Goldring & Langenstrassen 1979 | 400 to 2 000 |
| Ordovician, Norway | Brenchley <i>et al.</i> , 1979 | 10 000 to 15 000 |
| Triassic, Germany | Aigner, 1982 | 2 500 to 5 000 or 5 000 to 10 000 |
| Ordovician, Virginia, USA | Kreisa, 1981 | 1 200 to 3 100 |
| Approximate average 5 000 years | | |

Table 7.2 Recurrence intervals for preserved storm emplaced sharp based sandstones.

After: Walker (1984).

Duke (1985) cited examples of HCS within sediments preserved in the ancient Lake Bonneville which must have been emplaced by storms much less intense than hurricanes. This occurrence led Duke to conclude that

HCS can be produced by minor storms but its preservation potential in shallow marine sequences would be much lower than that for the more exceptional hurricane events. This explanation accounted, Duke believed, for the fact that nearly three-quarters of examples of HCS are preserved in sequences deposited at palaeolatitudes subject to hurricanes.

During deposition of the Lynton Formation southern Britain was situated 25° south of the equator (see sections 1.6.1) which is within the 10° to 45° latitude range generally accepted as being subject to hurricane events (Duke 1985). The occurrence of HCS in the Lynton Formation, therefore conforms to the latitudinal constraints of the majority of ancient examples.

The preceding discussion has led us to a position where the depositional mechanisms proposed for HCS can be critically evaluated. As outlined above, structures with a geometric style matching that for HCS have been observed in a very diverse range of sedimentary environments. This factor suggests that HCS is polygenetic in origin. For the purpose of this discussion, only mechanisms applicable to the shallow shelf environment below fair-weather wave-base will be discussed.

The possibility that tsunamis may be a causative agent for sublittoral sheet sandstone emplacement has been considered by Goldring and Bridges (1973) and Dott and Bourgeois (1982). Duke's (1985) study of the latitudinal distribution of HCS, however, indicated a preferential distribution dependent on climatic belts. Duke concluded that tsunamis are not a viable mechanism for HCS generation as the tectonic events triggering tsunamis would be essentially random i.e. latitude independent.

Although there is widespread evidence for syn-sedimentary tectonic activity preserved within the Lynton Formation (see chapter 2), the widespread occurrence of HCS within the upper Lynton Formation would be accompanied by widespread soft sediment deformation triggered by seismic and/or tsunami-induced liquefaction - if tsunamis were responsible for the emplacement of HCS - this is not the case however. Furthermore, soft sediment deformation is predominantly found in the lower part of the Lynton Formation (see chapter 2), whereas HCS is restricted to the upper part of the Lynton Formation. It must be concluded, therefore, that tsunamis do not offer a plausible mechanism for the emplacement of HCS. This conclusion is supported by Dott and Bourgeois' (1982) observation that soft sediment deformation structures are not commonly developed in HCS.

Following Hayes' (1967) study of hurricanes Carla and Cindy, several authors have suggested storm surge ebb currents as the causative agent for HCS (e.g. Brenchley *et al.* 1979; Walker 1979). Studies of modern shelf storm systems, however indicate that storm surge ebb currents have a negligible effect in comparison to geostrophic flow during storm events (see section 7.5 for discussion).

Several described sequences exhibit HCS intimately related to turbidites (e.g. Hamblin & Walker 1979; Leckie & Walker 1982a; Rosenthal 1982; Walker & Hunter 1982a) which has led to the suggestion that HCS is the product of density/turbidity currents modified by storm waves. Again, studies of modern shelf storm systems indicate that the generation of density/turbidity currents by shelf storms is hydrodynamically unreasonable (see section 7.5 for discussion). Indeed, Swift *et al.* (1983) noted that for typical shelf conditions the criteria for autosuspension of a turbidity current are in order of magnitude less than that required to sustain a turbidity current, further noting that the geostrophic flow component during a typical shelf storm event would be many times greater than any acceleration due to turbidity currents, making the latter barely detectable.

Several authors have proposed that HCS is the product of purely oscillatory flow generated by storm waves e.g. Campbell 1966, 1971; Harms 1975; Bourgeois 1980; Hunter & Clifton 1981; Dott & Bourgeois 1982). A theoretical study by Allen (1985, Allen & Pound 1986), however, indicated that generation of HCS under purely progressive storm waves is incompatible with published hummock spacing values when considered against predicted values for the ratio of the amplitude of oscillation near the bed and bedform spacing values. Allen concluded that the predicted crest spacings of low-steepness bedforms produced by surface gravity waves are much smaller than observed spacing values for HCS. Allen conceded, however, that surface gravity waves may play an important part in enhancing bed shear stress relative to a unidirectional current acting alone and, furthermore, introduce a strongly three-dimensional structure to the near-bed water motion, thereby preventing substantial bedform migration.

With our current understanding of modern shelf storm systems it is clear that a combined-flow régime is present near the bed during shelf storm events. Where shelf storm winds generate an approximately onshore surface flow a compensating offshore bottom flow results during the storm. As the flow accelerates it reacts to the Coriolis force, establishing a strongly three-dimensional geostrophic flow pattern (see text-figure 7.3). Given the above norm for shelf storm systems, suggestions that HCS is generated under a combined-flow

régime are particularly attractive (e.g. Cant 1980; Swift *et al.* 1983; Nottvedt & Kreisa 1987, Allen & Underhill 1989, Arnott & Southard 1990).

As noted above, a study of the storm-dominated inner Atlantic continental shelf off the eastern U.S.A. by Swift *et al.* (1983) revealed bedforms resembling HCS during side-scan sonar and diving observation. The hummocky bedforms, interpreted as HCS by Swift *et al.*, occurred in zones mutually exclusive from straight-crested megaripped zones during periods of intense flow acceleration due to storm generated geostrophic flow. A spacio-temporal acceleration model was advanced to explain the occurrence of the hummocky bedforms whereby the seafloor is eroded during the intensifying period of a storm. Scour is deeper than that for waves alone due to net sediment transport out of the area. As the curve for acceleration flattens, scour becomes less intense; interaction between bottom irregularities and the velocity boundary later results in scour of pits into the sea floor. Soon the sea floor aggrades and mound-like bedforms rise around the pits, continued scour keeping the pits open.

Although the bedforms described by Swift *et al.* cannot be unequivocally classified as HCS - owing to the lack of detailed internal observation of the bedforms - the study does present a valuable insight into the flow régime within which HCS is most likely to develop. In summary, the study presents a hydrodynamically plausible explanation for HCS by reconciling large- and small-scale hydrodynamic components of modern storm systems with the morphology of HCS. The interpretation of HCS advanced by Swift *et al.* can be extended further by considering the study reported by Allen and Underwood (1989).

Allen and Underhill described an exposure of amalgamated HCS (cf. 'swaley cross-stratification' *sensu* Duke 1980) from the Bencliff Grit (Upper Jurassic) of Dorset. Swales were observed to be filled asymmetrically, laminae being concordant on the 'upstream' side and slightly discordant on the 'downstream' side of the swale. This phenomenon accounted for a slight bimodality in measurements of palaeocurrent direction. Several 'growing' hummock-forms (cf. 'vertical accretion' type HCS -text-figure 7.4D) were observed within the section, as were examples of a vertical transition through low-angled laminae, sub-critical climbing ripples, supercritical climbing ripples to, finally, symmetrical wave-ripples. Allen and Underhill interpreted the bedforms as the product of a combined-flow régime. The majority of structures exhibited growth via 'downstream' and vertical aggradation, which was attributed to high sediment fluxes generated by oscillatory flow enhancing the shear stress at the bed produced by a

predominantly unidirectional current. In summary, the bedforms were believed to be transitional between HCS and trough cross-bedding, the diminished 'foreset' angle in relation to fully developed trough cross-bedding, being attributed to heightened sediment fluxes cf. Jopling (1965) and Bagnold (1966). Allen and Underhill further emphasised that their conclusions indicated that the structures are not environmentally specific.

The preceding discussion, albeit brief in relation to published accounts now existing for HCS, serves to demonstrate that although the origin of hummocky cross-stratification is not yet clearly understood in terms of the hydrodynamic régime prevailing during deposition, consideration of small-and large-scale processes operating during modern storm events does give valuable insights into the probable mode of formation of HCS. In summary, the evidence from the study of modern storm events combined with conclusions drawn from occurrences of HCS in the geological record strongly suggests a combined -flow origin for the structure. HCS appears to represent a high sediment flux variant of trough cross-bedding deposited by a unidirectional current. In shallow marine sequences the high sediment flux is generated by enhancement of the unidirectional flow by oscillatory currents.

At this point it is expedient to discuss the above hypothesis in relation to Duke's (1985) conclusions regarding the palaeolatitudinal distribution of HCS. As described above, the majority (73%) of examples of HCS were deposited at latitudes dominated by hurricane activity, the remainder having been deposited at latitudes dominated by intense winter storms. Swift *et al.* (1983) have shown that intense winter storms couple with the water column more effectively than the more transient, faster moving hurricane-type event. For this reason, geostrophic flow becomes better developed during intense winter storms than during the passage of a hurricane. Duke reasoned that as a result of this phenomenon, bedforms moulded by the well established geostrophic flow accompanying intense winter storms would resemble bedforms generated under unidirectional flow e.g. trough cross-bedding. In contrast, bedforms moulded by the less well established geostrophic flow accompanying hurricane events would be subject to less intense unidirectional currents but stronger oscillatory currents resulting from the very large surface gravity waves generated by hurricanes. Duke concluded that oscillatory or multidirectional currents were the dominant generating force for HCS.

Developing the conclusions reached by Duke further, the more powerful oscillatory flow superimposed a weaker unidirectional flow during hurricane events, in comparison to those generated during intense winter

storms, would result in an enhancement of bed shear stress relative to that produced by a unidirectional current alone, with accompanying high sediment fluxes near the bed. As noted previously, the higher sediment fluxes will result in diminished foreset angles. Thus, HCS with its characteristic low-angle lamination and evidence for deposition under high suspended sediment loads would be expected to develop more easily during flow conditions generated by hurricane events as opposed to intense winter storm events. Furthermore, the preservation potential of HCS would be higher in areas dominated by exceptional events greatly increasing the depth of wave-effectiveness (i.e. hurricanes) relative to areas experiencing moderate increases in wave-effectiveness during less exceptional events (i.e. intense winter storms).

Returning to the study of Swift *et al.* (1983) which examined hummocky bedforms on the storm-dominated inner Atlantic shelf off the eastern coast of the U.S.A., Duke (1985) noted that figured box core tracings exhibit apparent foreset dips of up to 31° (see above), much steeper than the 10° to 15° typical of HCS. Thus, if fossilised, the bedforms ascribed to HCS by Swift *et al.* would not be interpreted as HCS but would be described as trough cross-bedding degraded by storm waves. Nevertheless, the bedforms described by Swift *et al.* do exhibit certain features characteristic of HCS, hummocky topography of a similar scale for example. In summary, the bedforms described by Swift *et al.* are interpreted herein as representing structures towards one end of a spectrum of bedforms generated by combined-flows i.e. the end at which unidirectional currents dominate relative to oscillatory currents. At the opposite, oscillatory dominated end of the spectrum, bedforms conforming to 'classic' HCS are generated under high sediment fluxes.

To conclude, although much of the above is speculative and largely inferential, when modern storm systems, the palaeolatitudinal distribution of HCS and the lamination style characterising HCS are considered together the following are suggested:

- (i) The wide environmental distribution of structures exhibiting a lamination style characteristic of HCS indicates that the structure is polygenetic in origin.
- (ii) Shallow marine examples of HCS can be plausibly accounted for by high sediment fluxes near the bed, resulting from oscillatory currents enhancing bed shear stress values, suppressing the lateral development of trough cross-bedding relative to its vertical development - both in terms of diminished foreset angles and modification of streamlines for the unidirectional flow component by oscillatory currents.

(iii) Although hurricanes generate greater increases in bed shear stress via oscillatory currents reinforcing the unidirectional flow component, relative to intense winter storms, hurricane-type events are not a prerequisite for the formation of HCS; HCS generated by hurricanes, however, will have a higher preservation potential. Rather, it is the increase in sediment flux, near the bed, which will be relatively greater during hurricanes, that is apparently a prerequisite for HCS formation. Where HCS is produced by non-hurricane events the necessary increase in sediment flux near the bed could be the product of factors such as storm-wave attack of a shoreface suspending large amounts of sediment which will be transported offshore by geostrophic flow during the storm, for example.

The hypothesis above is also consistent with the occurrence of structures with a lamination style resembling HCS. For example, occurrences of HCS-type structures were given above for flysch and fluvial sequences. In the flysch occurrences, the presence of HCS-type lamination can be accounted for by the suppression of foreset angles within dune cross-bedding (C_1 division of the modified Bouma sequence *sensu* Allen 1970a), in response to high suspended sediment load (common in turbidity currents). The occurrence of HCS type structures in the fluvial examples can be attributed to sinuous crested dunes in a flow with high suspended sediment concentration, perhaps due to a river flood event.

In the two cases above, the enhanced sediment flux near the bed would be due to processes other than oscillatory flow; in this sense HCS-type bedforms are polygenetic in origin. A distinction must be drawn at this point between:

- (i) Polygenetic in the sense that differing process types can generate high sediment fluxes near the bed which will modify megaripples forming in response to the unidirectional component of flow (which may itself be polygenetic in origin e.g. fluvial, geostrophic or turbidite flows).
- (ii) Polygenetic in the sense that HCS-style lamination can result from processes dissimilar to those resulting in megaripple formation under high sediment fluxes e.g. Astin (1984) gives examples of beach structures which resemble amalgamated HCS ('swaley cross-stratification' of Duke 1980).

To conclude, HCS and morphologically similar structures are polygenetic in origin. For this reason, great care must be exercised before drawing any inferences with respect to the nature of the hydrodynamic régime

within which the structure was deposited. In particular, the widely held view that the structure is indicative of deposition between storm and fair-weather wave-base (Walker 1984) is invalid. Before useful hydrodynamic conclusions can be drawn for the occurrence of HCS the hypothesis presented above must be rigorously tested theoretically and by flume and natural environmental observations - in particular, relating the lamination style and bedform dimensions to the hydrodynamic régime. Although such rigorous testing is beyond the scope of this study, it is of value to briefly evaluate observations of HCS within the Lynton Formation (section 7.6.1) with respect to the above hypothesis.

Firstly, the palaeolatitude of 25°S at which the Lynton Formation was deposited falls within the 10° to 45° latitude range within which hurricanes are significant. Thus, Duke's (1985) analysis of the palaeolatitudinal distribution of HCS, which concluded that the majority of examples were deposited at hurricane-dominated palaeolatitudes, indicates that the Lynton Formation examples were generated at a statistically optimal palaeolatitude. Secondly, the wide diversity of microsequences exhibited by individual HCS beds within the Lynton Formation corroborate the idealised B-P-H-F-X-M microsequence, with bioturbation descending from bed tops, proposed by Walker *et al.* (1983 - see text-figure 7.4C), a sequence consistent with a waning-flow 'event deposit' interpretation. Walker *et al.* interpreted the lower B and P intervals of the microsequence to be generated by unidirectional currents, the succeeding intervals being interpreted as oscillatory in origin. Although it is theoretically possible that the lower two intervals could be oscillatory in origin, i.e. the graded base could be produced by winnowing induced by waves and the parallel-lamination could be oscillatory in origin (see above), the presence of sole marks consistently indicating an offshore flow (Hamblin & Walker 1979; Leckie & Walker 1982a) strongly suggests that the lower two intervals are unidirectional in origin. This unidirectional interpretation is extended to the Lynton Formation examples.

Walker *et al.* interpreted the H-zone, succeeding the B- and P- zones, as oscillatory in origin. This interpretation conflicts with the unidirectional flow under high sediment flux model presented herein, albeit that the high sediment flux in shallow marine examples is attributed to oscillatory currents enhancing bed shear stress. As discussed above, a purely oscillatory origin for H-zone laminae is considered untenable on hydrodynamic grounds; a dominantly unidirectional interpretation for H zone laminae is, therefore, posited herein.

Rare plan views of hummocks within the Lynton Formation reveal that although the majority of hummocks are circular in plan, several are elliptical with their long axes sub-parallel to palaeoslope (i.e. flow-transverse bedforms), a phenomenon in accord with similar observations by Campbell (1966, 1971), Hamblin and Walker (1979) and Cant (1980). The presence of this flow-transverse tendency is compatible with a trough-cross bedding deposited under a high sediment flux interpretation, although obviously does not provide conclusive proof for the hypothesis.

It should be noted at this point that the grain lineation visible on many hummock surfaces could be either unidirectional, oscillatory or combined-flow in origin (see above) and, therefore, neither contradicts or directly supports the hypothesis. i.e. Allen (1968) has shown that foreset laminae produced from a bed-stock of fine sand in a sub-aqueous environment will be normally graded. Thus, the normal grading in HCS laminae could be the product of either 'foreset' laminae developing in trough cross-bedding deposited under a high sediment flux or fall-out from a widespread suspension cloud un-associated with processes giving rise to trough cross-bedding.

The observation that HCS within the Lynton Formation is predominantly of the 'scour-and-drape' type, in common with the majority of reported examples of HCS, is again consistent with the hypothesis for HCS formation presented herein. It is posited that 'scour-and-drape' type HCS is generated by an initial phase of scour by predominantly unidirectional flow acting upon a non-uniform granular substrate. The scours will be evenly spaced and comparatively uniform in an effort to become stable relative to the prevailing flow - which is equivalent to that normally giving rise to sinuous-crested dunes (megaripples). As the flow wanes and/or the suspended sediment load attains a maxima beyond which no further sediment can be entrained, nett deposition (drape) will ensue. The draping laminae will be of equal thickness over the hummocky surface, or even thicken over hummock crests, where the flow régime remains equivalent to that normally giving rise to sinuous-crested dunes. As the flow wanes further, however, laminae will thicken into swales to give F-zone laminae. Where the unidirectional flow component waxes and wanes relative to the oscillatory component, or as temporal and spatial fluctuations in suspended sediment concentrations occur during the storm, multiple sets of hummocky laminae will be developed within individual beds, separated by undulose erosion surfaces.

The more rare development of 'vertical accretion' and 'tangential' type HCS is indicative of growing bedforms and suggests that the hummocky laminae were deposited when the flow régime was equivalent to that normally giving rise to sinuous-crested dunes; the resulting bedforms are best regarded as climbing dunes.

Further evidence for a generic link between HCS and dune bedforms is contained within Dott and Bourgeois (1982) who noted that "... there is even evidence that hummocky stratification may grade to ordinary cross-stratification" (p.664). This quote was based on an observation by Dott in the Cambrian of Wisconsin (Bourgeois pers. comm.) This phenomenon may be explained by a spatial variation in suspended sediment concentration during the emplacing storm event; where the concentration was high foreset lamination would be suppressed to give HCS, whereas cross-bedding would develop in areas where the suspended sediment concentration was lower.

Moving upwards through the idealised HCS microsequence the X-zone within Lynton Formation examples of HCS predominantly comprises combined-flow and, more rarely, purely oscillatory types. This indicates that in the majority of cases unidirectional currents are still important during the waning phases of storm events. This is attributed to the fact that the geostrophic flow component (unidirectional) will wane more rapidly than the oscillatory currents directly produced by storm waves. As the storm waves persist after the geostrophic flow ceases it is clear that the oscillatory flow component will increase in importance relative to the unidirectional flow component. It follows, therefore, that if the unidirectional flow component remains significant during the generation of the majority of x-zone laminae (i.e. combined-flow ripples), the unidirectional flow component must have been even more important during the generation of the preceding H-zone laminae. This hypothesis is supported by the observations of Dott and Bourgeois (1982) who gave examples of X -zone laminae comprising solely current ripples and climbing current ripples; these observations suggest that H-zone laminae are predominantly the product of unidirectional currents.

Dott and Bourgeois (1982) noted that HFM microsequences are a common variant of the HCS idealised microsequence (see also Pound 1986 - enclosure 14). The HFM variant is common in the Jennycliff Slates, deposited on shelf fed by deltaic outflow (Pound 1983 - see enclosure 12). Other authors e.g. Dott and Bourgeois (*op. cit.*) - Eocene, Coos Bay, Oregon and S. C. Nolan (ex E.S.R.I., U. C. Swansea, pers. comm.) - Ballustree Member, Lower Carboniferous, Dublin Basin, Ireland). The Coos Bay examples were associated

with much evidence for solely unidirectional flow. The absence of X-zone laminae in the preceding examples from deltaic sequences suggests that oscillatory flow was unimportant in the generation of these occurrences of HCS. It is posited that these HCS beds were generated by unidirectional flow generated by either storm or river flood processes in the absence of significant oscillatory currents, the attendant high suspended sediment flux inhibiting the generation of foreset laminae, hummocky, laminae resulting. This hypothesis is supported by evidence that the Jennycliff Slates examples were deposited in an environment near or below storm wave-base (Pound 1983).

In conclusion, observations from the Lynton Formation and Jennycliff Slates are consistent with the hypothesis that HCS is equivalent to sinuous-crested dunes deposited under high sediment fluxes, but do not provide conclusive proof for the hypothesis. Nevertheless, the hypothesis does provide a consistent explanation for HCS, reconciling knowledge of modern shelf storm systems, hydrodynamic theory, observed HCS lamination styles and the palaeolatitudinal distribution of HCS.

7.6.3 Geometry of the 'Woody Bay Facies Association'

Although many authors have observed proximality trends in sequences containing storm sands, both within modern shelves (e.g. Gadow & Reineck 1969; Reineck & Singh 1971; Morton 1981; Aigner & Reineck 1982; Nelson 1982) and ancient sequences (e.g. Goldring & Bridges 1973; Brenchley *et al.* 1979; Kreisa 1981), no trend in proximality could be discerned within facies 4. Throughout the upper part of the Lynton Formation., however, HCS beds are frequently amalgamated to give 2 to 4 m, thick sandstone units referred to the 'Woody Bay facies association'.

Although a large body of literature relating to HCS now exists, comparatively little has been written in respect to the geometry and origin of amalgamated HCS sandstone units. This is largely due to the limited lateral extent of exposure available in many of the studies, the difficulty of confidently identifying HCS in cores (Allen & Underhill 1989) where the geometry of the sandstone unit is otherwise well defined by seismic evidence and, perhaps most significantly, the absence of unequivocal modern examples of HCS within shelf sand-bodies. The steep dip of exposures in the upper Lynton Formation (25° to 30°S.) relative to the more gently inclined lower Lynton Formation, and the frequency of late normal faulting along with the limited lateral extent and inaccessibility of exposures combine to make prediction of the geometry of the

'Woody Bay facies association' impracticable. During the present study, no lateral thickness variations in 'Woody Bay facies association' type sandstone units could be discerned, units assuming a tabular geometry across the width of available exposure.

Brenchley and Newall (1982), in a study of the Ordovician Cheney Longville Flags of Shropshire, suggested that occurrences of amalgamated HCS sandstone units represented sand lobes developed on the inner shelf. The absence of lateral exposure control, however, precluded the accurate assessment of sandstone unit geometry, but Brenchley and Newall envisaged a "mode of formation . . . analogous to rip currents but an order of magnitude larger" (p.1268), possibly located at the mouths of rivers or channels between barrier islands. No modern counterparts exist for the large-scale rip cells invoked by Brenchley and Newall and this mode of formation is, therefore discounted herein.

Tunbridge (1983) interpreted amalgamated sandstone units, containing HCS within the Great Burland sequence (200m to the E of the Little Burland sequence - see enclosure 1) as outer shoreface shoals analogous to modern examples described by Howard and Reineck (1972) from the upper shoreface of Sapelo Island, Georgia, U.S.A. and by Kumar and Sanders (1976) from shoreface sands along Long Island. Tunbridge observed that the amalgamated sandstone units only occurred in certain sequences and attributed this phenomenon to the growth of nearshore sandy shoals generated by storms, the shoals being ephemeral features. Tunbridge's interpretation was based on the position of the amalgamated sandstone units within a facies sequence, the study lacking the lateral exposure control to directly observe the sandstone unit geometry.

Atkinson *et al.* (1986) presented results of an integrated study of the depositional geometry of Miocene storm-generated sandstones from the north coast of Borneo. A series of 1.5 to 3m thick amalgamated sandstone beds interpreted as hummocky/swaley cross-stratification were correlated and mapped and found to take the form of offshore bars, up to 2km in length, approximately parallel to the palaeo-shoreline. It was suggested that the bar-like geometry is a result of the modification of distal storm sands by semi-permanent coastal currents.

Rice (1984) described a series of amalgamated HCS units within the Upper Cretaceous Mosby Sandstone of central Montana. Individual units averaged 1.5m in thickness and were elongate sub-parallel to the palaeo-

shoreline, attaining lengths of several kilometres; the units tended to correlate with the position of palaeo-‘highs’. Rice attributed the formation of the sandstone lenses to the temporal and spatial flow acceleration model for geostrophic flow presented in Swift and Rice (1984 - see section 4.3 for discussion on the application of this model to the ‘Lee Stone facies association’).

In addition to the lensoid geometry attributed to amalgamated HCS sandstone units by the above authors, several authors have recognised a sheet-like, tabular geometry e.g. Moore and Hocking (1983) and Brenchley (1985). The latter author noted that sheet-like tabular geometries tend to occur on nearly flat shelves and can form belts of over 100km in width.

The above examples serve to emphasise that amalgamated HCS sandstone units, within a vertical sequence lacking lateral continuity, could represent any one of either tabular or lensoid geometries and be either associated with a shoreline/shoreface or have formed on an open shelf. Furthermore, caution must be exercised when attributing amalgamated HCS units to a specific sandstone-body geometry in the absence of a clear understanding of the hydrodynamics of HCS emplacement - only an understanding of the hydrodynamics of HCS emplacement will allow the structure to be related to the processes generating sandstone-bodies of a specific geometry.

For the above reasons the geometry of ‘Woody Bay facies association’ sandstone units must remain equivocal.

7.7 FACIES 5 - CROSS BEDDING

Facies 5 is shown as facies E on the Woody Bay west log, the only sequence within which this facies occurs in the sections under consideration within the present chapter. The cross-bedded units occur at the top of the sequence (25.6 to 27.5m on the log).

Facies 5 comprises 4 to 22cm thick sets of fine to medium grained cross-bedded sandstones. Set bases are either erosional, exhibiting localised scours, planar or drape and preserve wave-ripple profiles at the top of the preceding set. Set tops are either erosionally truncated by the succeeding cross-bedded set, wave-rippled or biogenically ‘churned’, the bioturbation descending from the upper set surface, frequently mixing-in mud

to give a gradational contact with the overlying unit. Examination of vertical sections at varying angles through the wave-ripple profiles indicates that the ripples are either of 3-D ('micro-hummocks') or of interference origin; unfortunately, no plan views were available to enable a distinction to be drawn between the two types.

The cross-bedding is primarily of the trough type, although planar cross-bedding is also present e.g. 26.4m on the log. Foreset laminae range between 2 and 4mm in thickness and approach the lower set boundary asymptotically in trough cross-bedded sets. Measurement of foreset dip directions indicated formations by SE directed palaeocurrents ($138^{\circ}29'$, where $N = 5$, the mean spread of angular values = 94.2% and the mean angular deviation is $\pm 19.5\%$), a direction oblique to the SSW-dipping regional palaeoslope. Occasionally, sets are separated by thin zones of wave-generated bi-directional cross-lamination up to 2 sets in thickness.

Cosets are 19 to 86cm in thickness, a maximum of 9 sets having been observed within a single coset. Cosets are separated by thin units of unbioturbated silty mudstone (facies 1). Complete cosets have been observed to be disturbed throughout their thickness by biogenic 'churning' which can render internal set boundaries diffuse.

In the absence of more detailed data, interpretation of facies 5 is enigmatic. The trough cross-bedding and planar cross-bedding are of the same hydrodynamic class, representing the migration of sinuous- and straight-crested dunes respectively; the sinuous-crested dunes formed at higher environmental energies than the straight-crested dunes (see section 4.2.4.3 for discussion). The flow that generated these bedforms would have been either unidirectional or pseudo-unidirectional (i.e. strongly asymmetric oscillatory flow produced by strongly shoaling waves cf. Clifton 1976). There is no evidence for tidal activity in the upper Lynton Formation, e.g. 'Woody Bay facies association' sands would have been reworked during inter-storm periods if tidal currents of sufficient magnitude were active. Tidal currents are, therefore, discounted as possible agents for the transport of facies 5 bedforms.

The association of cross-bedding with HCS (facies 4) is unusual in a regressive setting, other examples include LaFon (1981), Dott and Bourgeois (1982). Bourgeois (1980) cited examples from an Upper Cretaceous transgressive sequence in SW Oregon and Dott and Bourgeois (*op. cit.*) gave an example from the Miocene of Cape Blanco, Oregon. Both groups of authors attributed the cross-bedding to the migration of

lunate-megaripples under shoaling waves in a wave-dominated nearshore environment cf. lunate-megaripples recorded in modern shoreface environments by Clifton *et al.* (1971). The lunate-megaripples described by Clifton *et al.*, however, migrated in an offshore direction, whereas facies 5 bedforms migrated obliquely offshore (SE). For this reason the following alternative explanation for the association of cross-bedding (facies 5) and HCS (facies 4) is advanced.

It is posited that the dunes of facies 5 migrated under a storm-induced, obliquely-offshore-flowing (SE) geostrophic current of the type invoked for the generation of HCS in facies 4. Dunes developed in preference to HCS for one, or a combination of, the following reasons:

- (i) The sands of facies 5 are coarser than those of facies 4 and would, therefore, have required a higher flow power to suspend them.

- (ii) There may have been less sediment suspended at source, e.g. due to a change in the shoreline configuration, resulting in a suspended sediment concentration which was insufficient to suppress the formation of angle-of-repose foresets, resulting in dunes forming as opposed to climbing dunes (HCS). As the storm waned the storm-waves oscillatory currents would become dominant relative to the ebbing geostrophic flow; the duned seafloor would be modified by the oscillatory currents, giving the thin cross-laminated cap to cross-bedded sets. The interaction of the ebbing geostrophic flow with oscillatory currents may account for the preponderance of interference ripples in facies 5.

In conclusion, facies 5 represents the offshore migration of a duned sand patch, of a coarser grade than the HCS sands that occur at this level, under the influence of storm-generated geostrophic flow; the patch would only have been active during periods of storm activity and would have been subject to biogenic colonisation and mud deposition during inter-storm periods.

7.8 FACIES 6 - WAVY BEDDING

Facies 6 is shown as facies D on the Little Burland log, the only sequence containing this facies in the sections under consideration within this chapter, where its occurrence is restricted to a zone between 48.8m

and 51.4m on the log interbedded with bipolar ripple cross-laminated units (facies 7), Units of facies 6 attain a maximum thickness of 1.6m.

Individual wavy bedded units comprise a regular alternation of 2 to 9cm, thick beds of fine to very fine grade sheet sandstones set within 1 to 29cm thick sequences of unconnected and connected lenticular bedding (cf. facies 1). The sheet sandstones are laterally persistent over the width of the exposure and may be divided into two types on the basis of the nature of the lower bounding surface: planar or undulatory. Planar based beds (plate 7.14B) frequently exhibit a basal zone of parallel-lamination up to 6cm in thickness; no primary current lineation has been observed. The parallel-lamination is overlain by a zone of bimodal trough cross-lamination occurring as 1 to 2cm thick sets with scooped bases incising erosively into the underlying parallel-laminated zone. The cross-lamination exhibits low-angle lamina sets which pinch-and swell to give a swollen-lens morphology. Laterally adjacent sets frequently exhibit opposing foreset dip directions. These features are diagnostic of wave-generated ripples cf. Boersma in: de Raaf *et al.* (1977).

Beds with undulatory bases (plate 7.15A) have a relief not exceeding 1cm, undulations exhibiting wavelengths of up to 8cm. The complete bed comprises bimodal cross-laminations identical to those described for the planar based sheet sandstones.

Sheet sandstone tops comprise symmetric to near-symmetric wave-ripple profiles of 6 to 8cm wavelength and up to 1cm amplitude; both ripple troughs and ripple crests are rounded. Lamina-sets beneath the rippled surfaces are generally form-discordant, although a single form-concordant draping lamina is occasionally visible. Vertical sections taken at different orientations through ripple profiles, coupled with rare plan views of bedding planes reveal that the ripples are dome-shaped ('micro-hummocks'). The domes are circular to slightly elliptical but do not exhibit a preferred direction of elongation; dome flanks have a dip of 15° to 25°. The dome-shaped ripples are reminiscent of 'three-dimensional vortex ripples' produced in oscillatory flow tunnels at the transition from two-dimensional wave-ripples to upper oscillatory plane bed (Carstens *et al.* 1969; Lofquist 1978). The three-dimensional vortex ripples evolve over a long period of time from straight-crested ripples in sand of less than 0.3mm diameter.

The sheet sandstones within facies 6 are separated by unconnected and connected lenticular bedding identical to that preserved in facies 1. The lenticular bedding is bioturbated by sand-filled *Palaeophycus tubularis* burrows of 5 to 7mm diameter.

The interpretation of facies 6 is based upon that for the closely analogous facies 5 at Lee Stone - see section 4.2.6 for a detailed discussion and references. Facies 6 is regarded as an amalgam of facies 1 (lenticular bedding) and facies 3 (thinly-bedded sheet sandstones). The two are grouped as one facies due to their regular alternation, the significantly smaller ratio of lenticular bedding to sheet sandstones distinguishing this facies from facies 1 sequences containing random occurrences of facies 3.

The undulose-based sheet sandstones represent moderate frequency minor storm events proximal to an abundant sand supply. In contrast the planar-based sheet sandstones represent higher energy storm events, as evidenced by the basal parallel-laminated zone, also of a moderate frequency and proximal to a sand source. The intervening lenticular bedding represents the 'normal' depositional mode at that particular site, with the unconnected and connected thin sandstones representing high frequency, relatively low energy storm events. The *Palaeophycus tubularis* burrows represent inter-storm colonisation of a muddy substrate.

Although wavy bedding has generally recorded from tidal environments, examples have also been described from modern and ancient wave-dominated shoreface environments (Howard & Reineck 1982; Tunbridge 1983a); no evidence of tidal activity has been observed within the sections under consideration within this chapter. The time span required to deposit the bioturbated lenticular bedded units intervening between the sheet sandstones indicates a frequency typical of major storms for the emplacement of the sheet sandstones as opposed to emplacement by processes with a tidal periodicity.

7.9 FACIES 7 - BIPOLAR RIPPLE CROSS-LAMINATION

Facies 7 is shown as facies E on the WNW Barbrook and Little Burland logs and comprises 3 to 34cm, thick cosets of very fine to fine grade sandstone. The sandstone is well sorted, weathered surfaces being white in colour; thin sections reveal a quartz sub-litharenite composition. Units have a very low mud content, some sets exhibiting an upwards decrease in mud content suggesting an upward increase in winnowing. Boundaries between cosets in facies 7 are generally planar and erosive, although rippled surfaces are

occasionally preserved. Facies 7 is found interbedded with thinly interlayered sandstone/mudstone bedding (facies 1), hummocky cross-stratification (facies 4), wavy bedding (facies 6) and parallel-laminated sandstone with thin ripple cross-laminated zones (facies 8); a single thinly bedded sheet sandstone (facies 3) has been observed to intervene between two facies 7 cosets.

Internally, facies 7 (plate 7.16A, B & C) is characterised by 2 to 3cm thick sets of intricately interwoven cross-laminations. Lower set boundaries are generally erosive, undulating in an irregular manner. Individual sets exhibit a swollen-lens morphology displaying unidirectional cross-lamination, adjacent sets frequently exhibiting opposing foreset dip directions. Individual ripple lenses arise from a bundled up-building of complexly interwoven sets. Single offshooting laminae occasionally drape foreset surfaces. Adjacent sets are generally structurally dissimilar.

The above features are characteristic of wave-ripple cross-lamination cf. Boersma (in: de Raaf *et al.* 1977). Ripple profiles are only preserved occasionally and comprise 7 to 10cm wavelength symmetric to slightly asymmetric ripples of up to 1cm amplitude. Ripple crests are normally rounded, although trochoidal crests have been observed (plate 7.16A); ripple troughs are always rounded (plate 7.16A). The paucity of plan views of bedding planes within facies 7 make ripple crestline trends difficult to measure; it is apparent, however, that the majority trend between NNW-SSE and WNW-ESE. Laminae are generally form-discordant beneath the rippled surfaces, although form-concordant offshooting and draping laminae are occasionally visible.

Millimetre-scale mud laminae occasionally separate sets of cross-laminae within facies 7 and give rise to localised zones of flaser bedding (*sensu* Reineck & Wunderlich 1968a). The mud laminae take the form of either isolated convex-up flasers preserved within ripple troughs or continuous drapes overlying and preserving rippled surfaces, mud laminae tending to thin over ripple crests. Isolated flasers occasionally pass laterally into a zone of continuous drape of ripple profiles. Rarely, vertical sections reveal mud laminae bifurcating to pass either side of cross-laminated lenses. Tunbridge (1983a) considered flaser bedding as a separate facies, distinct from bi-directional ripple cross-laminated sandstones within the Great Burland sequence. Flaser bedded sandstones are not as well developed in the Little Burland sequence, however, flaser bedding generally dying out both laterally and vertically within bipolar ripple cross-laminated cosets. For

this reason flaser bedding is not considered as a separate facies within the sections under consideration in this chapter.

Towards the top of the Little Burland sequence facies 7 contains several thin zones of spherical to sub-spherical sandstone pellets, 7 to 10mm in diameter. The surface of the pellets is generally smooth, although some exhibit a concentric ring structure (plate 7.17D). Polished sawn sections through samples of the pellet horizon indicate that many of the pellets had a very thin mud coating. The pellets are interpreted as faecal in origin, Reineck and Singh (1980) noted that faecal pellets are generally transported away from their place of origin through bioturbation or inorganic agencies. Evidence preserved within the faecal pellet zones at the top of the Little Burland sequence suggests that they are secondary layers i.e. "The main difference between . . . 'primary biogenic layers' of fecal pellets and 'secondary fecal-pellet layers' (produced by concentration of transported fecal pellets into layers) is that the primary fecal pellet layers overlie a sediment layer with a dense population, the sediment thus being strongly bioturbated. The secondary fecal pellet layers may overlie a completely unbioturbated horizon" (Reineck & Singh 1980, p.174). Of the three faecal pellet horizons at Little Burland, two overlie non-bioturbated sediments, the other overlies a zone of biogenic 'churning' which shows no evidence of biogenic activity by an organism large enough to produce faecal pellets of the size observed.

Facies 7 is only lightly bioturbated in comparison with other facies described within this chapter. Isolated 5 to 7mm, diameter sandstone-filled *Palaeophycus tubularis* burrows are observed occasionally, normally descending from the upper surface of cosets. More commonly, isolated patches of biogenic 'churning' (fossitextura deformativa) give rise to a loss of primary sedimentary structures locally (plate 7.16C).

Facies 7 is the product of prolonged periods of sustained oscillatory current action. Waves approached the ESE-WNW trending palaeoshoreline (see section 2.2.4) in a direction which varied between normal to the palaeoshoreline (ESE-WNW trending ripple crestlines) and strongly oblique to the palaeoshoreline (SSE.-NNW trending ripple crestlines). When these directions are corrected to remove palaeomagnetic rotation a palaeowind direction varying between WSW and WNW is indicated for generation of the waves that produced the respective sets of ripple crestlines. The slight asymmetry of several profiles indicates a unidirectional flow component to the predominantly oscillatory current. No preferred direction of asymmetry could be discerned and the origin of the unidirectional component must remain equivocal.

It is clear that the predominantly oscillatory currents responsible for the generation of facies 7 were not constant in respect of energy, as evidenced by the fluctuating mud and biogenic content at differing levels within facies 7. The highest energy levels are reflected by the preservation of clean, ripple cross-laminated sandstones devoid of biogenic traces. The absence of biogenic traces, when considered with the preservation of biogenic traces elsewhere in facies 7, suggest that wave energy was relatively constant at these levels, reworking any biogenic traces before they could be fossilised.

Conversely, flaser bedding and patches of biogenically churned sediments reflect periods of diminished wave activity. As was discussed in section 4.2.7.3, flaser bedding has been recorded in both tidal- and wave-dominated shallow marine sequences. The mud flasers are interpreted herein as the product of inter-storm deposition (McCave 1970, Hawley 1981), rather than following the original interpretation of mud deposition during tidal slack water (Reineck 1960, Reineck & Wunderlich 1968a), Tunbridge (1983, p.150) also attributed flaser bedding, in the Great Burland sequence, to: “. . . wave generation, forming under slightly less constant energy conditions that the cross laminated facies, permitting mud films to drape the wave ripple sands during quiet periods.”

Similarly, the zones of biogenic churning are also interpreted as the product of periods of diminished wave activity which permitted the preservation of the bioturbate texture. Aigner and Reineck (1983) presented a study of the present day shoreface off the Norderney barrier island (North Sea) in which they used the degree of bioturbation as an indicator of the extent of wave reworking. The first appearance of significantly bioturbated sediments, when following a profile in a seaward direction, was taken as marking the approximate depth of wave base. The depth of the effective wave base was noted as changing cyclically in response to an alternation between the winter storm season (deeper) and the temperate summer season (shallower). Following Aigner and Reineck's observations the biogenic churned zones within facies 7 are interpreted as representing periods of diminished wave action which intervened between storms/storm seasons.

It is clear that the facies association context of facies 7 differs between the WNW Barbrook and Little Burland sections. In the WNW Barbrook section facies 7 is restricted to thin horizons, never exceeding 10cm in thickness, and is found in association with lenticular bedding (facies 1), thinly-bedded sheet sandstones (facies 3) and hummocky cross-stratification (facies 4). These latter three facies were interpreted as the

product of deposition below fair-weather wave-base in a storm-dominated environment. Given this interpretation, the cleaner ripple cross-laminated component of facies 7 in the WNW Barbrook section is interpreted as the product of periods of prolonged wave activity immediately following storm activity, whilst the flaser bedding reflects an alternation between storm and fair-weather deposition.

Within the Little Burland sequence (enclosures 11A & 11B) facies 7 predominantly occurs in thick cosets up to 34cm in thickness towards the top of the section. Lower in the section thinner cosets of facies 7 were found in a similar facies association to those at WNW Barbrook and are ascribed to the same process of formation. The facies association of the thicker facies 7 cosets at the top of the section is, however, quite different, interdigitating with wavy bedding (facies 6) and parallel-laminated sandstone with thin ripple cross-laminated zones (facies 8), individual facies 7 cosets frequently being separated by thin zones of lenticular bedding (facies 1). Furthermore, mud flasers are less frequent in the thicker cosets of the upper part of the Little Burland section, suggesting deposition during periods of prolonged, relatively constant wave action.

The facies association of facies 7 within the upper part of the Little Burland sequence closely matches the shoreface facies sequences described by Tunbridge (1983) within the Great Burland section. Tunbridge ascribed the thick cosets of ripple cross-laminated sandstones within the Great Burland sequence to deposition in an inner shoreface environment intervening between wavy bedded deposits of the outer shoreface, frequently containing multi-storey parallel-laminated sandstones and hummocky beds representing outer shoreface shoals, and parallel-laminated sandstones with heavy mineral layers and thin ripple cross-laminated horizons of the foreshore. Tunbridge's interpretation for the thick cosets of ripple cross-laminated sandstone at Great Burland is followed herein for the thick facies 7 cosets in the upper part of the Little Burland sequence.

Tunbridge noted that the wave-generated ripple cross-laminated cosets bear a close resemblance to ripple cross-laminated sands described from the modern low wave energy, tide dominated shoreline of Sapelo Island, Georgia, U.S.A. (Howard & Reineck 1972) in a 'lower shoreface' environment (1 to 2m water depth). Howard and Reineck (1981) later compared the Sapelo Island profile with a modern high wave energy shoreline between Ventura and Port Hueneme, California, U.S.A. They observed that wave ripple cross-lamination is restricted to 'nearshore facies' (low water to 9.3m water depth) and commonly contains

concentrations of heavy minerals suggesting wave reworking of the substrate under conditions of minor sediment addition. Howard and Reineck concluded “. . . the similarity between the overall types of physical sedimentary structures are remarkable” (p.829) when comparing the Sapelo Island and Ventura-Port Hueneme shorelines. When considering the likely stratigraphic sequences to preserved for each sequence, the only significant difference would be that the sequence would comprise thicker facies units reflecting the increased depth of wave-base. The difficulty in accurately quantifying sediment compaction, basin subsidence rates and palaeotidal range precludes any attempt being made, however, to assign the sequence at Little Burland to a particular point on the spectrum between high and low energy shorelines.

Finally, it is of interest to note that the differentiation between the thin facies 7 coset facies association (WNW Barbrook and lower part of Little Burland sections), interpreted as being deposited in an offshore/lower shoreface position below fair-weather wave-base and the thick facies 7 coset facies association (upper part of Little Burland sequence), interpreted as being deposited in an upper shoreface position predominantly within fair-weather wave-base, is mirrored in the Ventura-Port Hueneme coastal profile. Howard and Reineck (1981) observed that wave ripple cross-lamination in their ‘offshore’ and ‘transition’ zones was restricted to thin cosets, frequently found in the upper part of storm-generated parallel-laminated and ‘hummocky’ sand beds. In comparison, upper shoreface deposits are dominated by thick cosets of wave-ripple cross-laminated sands.

7.10 FACIES 8 - PARALLEL-LAMINATED SANDSTONE WITH THIN RIPPLE CROSS-LAMINATED ZONES

Facies 8 is shown as facies F on the Little Burland log (enclosures 11A & 11B) where the facies occurs towards the top of the section. Of the sections described within this chapter, facies 8 is restricted to Little Burland where it is interbedded with thin units of lenticular bedding (facies 1) and bipolar ripple cross-lamination (facies 7). Facies 8 laminae sets range between 4 and 40cm in thickness, cosets attaining a maximum thickness of 2.3m. Individual sets assume a tabular to gently wedge-shaped geometry, set boundaries being planar to gently undulose and erosive, truncating laminae of the previous set at a low angle (maximum of 5°). Although the nature of the exposure precludes accurate measurements of the set boundary dip direction being made, the majority appear to dip in a southerly direction (offshore). Facies 8 is composed of a well sorted quartz sub-litharenite of coarse siltstone to fine sandstone grade which weathers a very light grey. Individual laminae are frequently picked out by dark bands comprising heavy mineral concentrations.

Tunbridge (1983a) described similar heavy mineral concentrations in the nearby Great Burland section in an identical facies, noting that the layers were dominated by haematite and tourmaline.

Internally, facies 8 comprises parallel-lamination, laminae typically ranging from 1 to 3mm in thickness and are interspersed with infrequent thin zones of ripple cross-lamination up to 3cm in thickness (plate 7.17A). No primary current lineation has been observed, although it should be noted that there is a paucity of bedding plane surfaces of facies 8 in the Little Burland section. The cross-laminated zones exhibit two distinct lamination styles. The first style comprises one set thick units with planar bases and tops with individual foresets being planar and generally dipping towards the south (offshore). This lamination style is ascribed to an origin under offshore-directed unidirectional flow. The second cross-lamination style comprises bi-directional trough cross-lamination, set bases exhibiting erosive 'scoops' into underlying parallel-lamination. Observed ripple profiles are symmetric to near -symmetric and are preserved beneath a cap of massive sandstone at the base of the succeeding unit. The second cross-lamination style is attributed to the effects of pure oscillatory and combined flows.

Several zones of soft sediment deformation, in the form of convolute lamination, have been observed. Individual zones attain a maximum thickness of 26cm and exhibit convolutions which steadily increase in amplitude upwards within a set, then decrease in amplitude towards the top of a set. The convolutions are always truncated by penecontemporaneous erosion. This style of convolute lamination was termed 'metadepositional' by Allen (1982) i.e. generated just before, or immediately after, deposition ceases. Amalgamated, multiple sets of convolute lamination have been observed, separated by undulose erosion surfaces.

Occasional shallow scours cut facies 8, reaching widths of 50cm and depths of 5cm (plate 7.17B & C). These features resemble examples of swaley cross-stratification (Leckie & Walker 1982a) figured in Brenchley (1985 figure 11). Individual scours display broadly symmetric, concave lower surfaces and contain a concordant infill, laminae thickening towards the centre of the scour.

Bioturbation within facies 8 is primarily restricted to dense colonies of vertical, sandstone filled *Skolithos verticalis* burrows. Individual burrows are approximately 1mm in diameter, gently sinuous and penetrate to depths of up to 35mm. Tube walls are picked out by a thin zone of muddy, organic debris and occasionally

display an annulated structure. The tube population is dense, tubes generally being separated by distances of only 3 to 10mm, tubes always being truncated by penecontemporaneous erosion. Multiple levels of burrowing are present within individual beds. *S. verticalis* is interpreted as the domicile of a suspension-feeding vermiform organism (see appendix B for a detailed discussion, also plates B24A, B & C). The more silty horizons of facies 8, where the mud content tends to be higher, tend to be burrowed by 5-7mm diameter sandstone-filled *Palaeophycus tubularis* tubes.

Facies 8 closely matches Tunbridge's (1983) 'parallel and cross laminated sandstone' facies described from the adjacent Great Burland section. Tunbridge assigned this facies to deposition in a beach environment, basing his interpretation on the presence of intertidal indicators such as symmetrical ripples with planed off crests (cf. Reineck & Singh 1980), 'ladder ripples' (cf. Bajard 1966), small channels possibly cut at low water on an intertidal surface (cf. Clifton *et al.* 1973) and heavy mineral layers possibly concentrated by wave swash (cf. McKee 1957, Clifton 1969, Vos 1977). Tunbridge concluded that the interbedding of cosets of wave-generated cross-lamination with parallel-lamination suggested fluctuating wave energy or a change in beach topography producing a complex of rippled and planar surfaces which could represent a low relief ridge and runnel system (cf. Parker 1975).

Thick sequences of parallel-laminated sandstones of the type described above are relatively common in the fossil record within sequences attributed to deposition in prograding shoreline sequences (Harms 1975, Elliott 1978, Heward 1981). The physical process responsible for the production of the parallel-laminae can either be swash-backwash (Clifton 1969) or oscillatory sheet-flow (Clifton 1976). Parallel-lamination formed by swash-backwash, as distinct from upper-phase plane bed deposits (Allen 1984), frequently exhibit inversely graded laminae (Clifton 1969) and crescentic swash marks (Reineck & Singh 1980) on bedding plane surfaces; such laminae are generally associated with indicators of intertidal exposure. None of these features were observed in facies 8 at Little Burland. Parallel-lamination produced by oscillatory sheet-flow forms only under waves of short or intermediate wave period as would be expected for the Lynton Formation (see chapter 2) as a response to intense wave activity close to the shoreline (Clifton 1976). The resulting laminae would be grouped into wedge-shaped sets and would lack evidence of intertidal exposure. Based upon the evidence preserved in the Little Burland sequence, facies 8 parallel-lamination is ascribed an origin under oscillatory sheet flow. The thin zones of oscillatory and current -ripple lamination are thought to reflect periods of less intense wave action under strongly shoaling waves which would have generated

markedly asymmetric ('pseudo-unidirectional') flow. It is possible, however, that some of the current ripples could represent the last stage of emergence with water running across an intertidal surface (cf. Reineck & Singh 1980).

Although Tunbridge (1983a) suggested that the alternation of parallel-lamination and cross-laminated intervals could be produced by the migration of a low relief ridge and runnel system (see above), this interpretation is not favoured for the Little Burland sequence where the cross-laminated intervals are thinner and less well organised than those in the Great Burland sequence. Furthermore, the preservation potential of ridge and runnel systems, representing a fair-weather equilibrium topography, is low (Harms 1975), whereas the preservation potential of shoreline deposits generated by periods of storm activity with high wave energy is high (Elliott 1978). Facies 8 at Little Burland is interpreted, therefore as the product of periods of sustained high wave energy (parallel-lamination), punctuated by periods of relative quiescence (cross-lamination - fair-weather), on the upper shoreface of a storm-and wave-dominated shoreline.

Upper shoreface parallel-laminated sands have been reported from both low energy (Howard & Reineck 1972) and high energy (Clifton *et al.* 1971; Howard & Reineck 1981) modern shorelines as well as ancient shoreline sequences (e.g. Harms 1975; Hamblin & Walker 1979). Interestingly, Hamblin and Walker described a series of broad shallow scours, 45-80cm across and 5-10cm deep, in the lower part of their 'planar bedded sandstone' facies, interpreted as beach deposits, which closely resemble the scours observed in facies 8. Walker and Hamblin (1982) reinterpreted this sequence and assigned the scours to swaley cross-stratification ('SCS', *sensu* Leckie & Walker 1982a), stating that they felt that the lower part of the sequence containing SCS represented a storm-dominated upper shoreface environment, whilst the overlying SCS-free parallel-laminated zone represented a true beach environment. Astin (1984), however, has pointed out that structures resembling SCS are visible in many shoreline deposits, ranging from backshore through to shoreface.

The frequent occurrence of convolute lamination in facies 8, relative to other facies in the Little Burland section, suggests that a wave-beat origin for this structure (cf. Dalrymple 1970) is more likely than a seismic shock origin.

The isolated dense colonies of *Skolithos verticalis* in facies 8 are typical of Seilacher's (1967a) *Skolithos* ichnofacies which characterises lower littoral to infralittoral, moderate to relatively high energy environments with a well sorted shifting particulate substrate. Physical sedimentary structures predominate over biogenic structures in this ichnofacies (Howard 1975) due to intense physical reworking. See Appendix B for a discussion on the ethology of *S. verticalis*. The assignment of facies 8 to the *Skolithos* ichnofacies is consistent with the upper shoreface interpretation based upon primary sedimentary structures.

7.11 FACIES SEQUENCE ANALYSIS

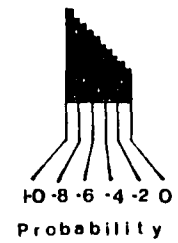
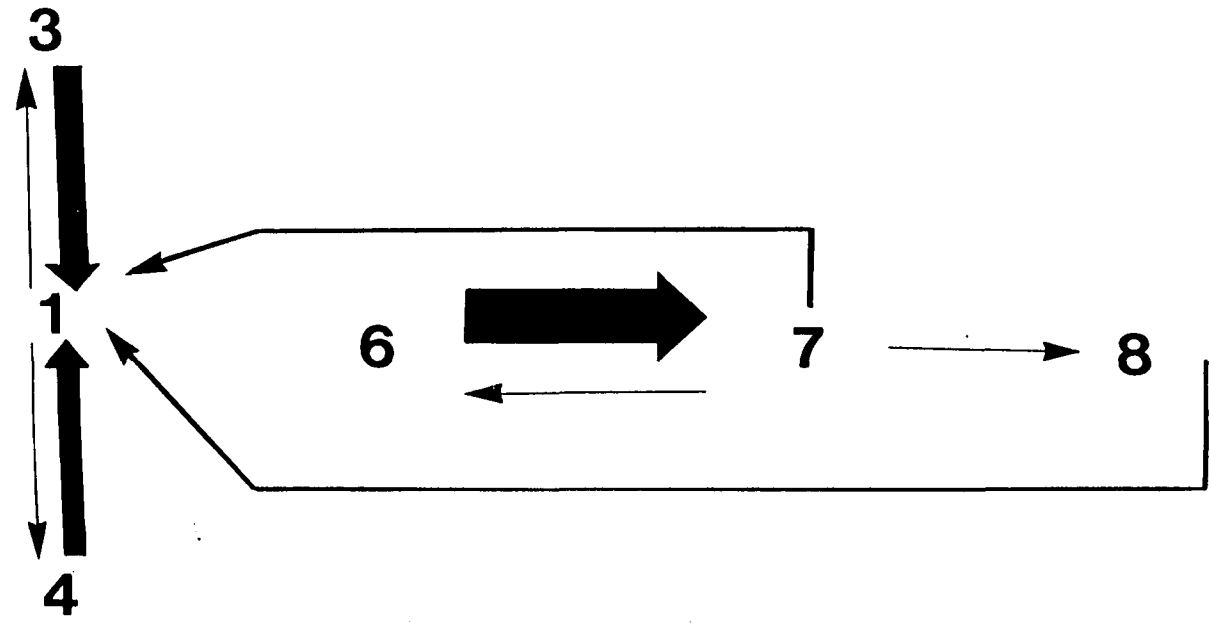
Of the three sections described in this chapter, the most variable sequence of facies occurs at Little Burland where a qualitative assessment of the section reveals a coarsening-upwards (CU) trend, itself comprising minor CU cycles. In comparison, the sections at Woody Bay west and WNW Barbrook comprise more muddy, shallow marine deposits which enclose an isolated sandstone body lithotype: the 'Woody Bay facies association'. The latter two sections resemble the lower part of the Little Burland section in terms of lithology and facies. As the Little Burland sequence comprised sequences representative of the major facies associations under consideration in this chapter, this section will be analysed in detail, the resulting interpretations being applicable to the other two sections.

The physical and biogenic structures within the Little Burland sequence indicate a shallowing upwards trend with a commensurate upwards increase in water turbulence. The log of this section, however, reveals that the interdigitation of facies is complex. It has been found necessary, therefore, to apply quantitative techniques to reveal the facies sequences present within this section. Observed-minus-random facies transition probabilities were calculated using the method outlined in Cant and Walker (1976); the results are tabulated in Appendix G and are shown graphically in text-figure 7.9, where only transitions with a greater than random probability of occurring are shown. The shortcomings of this technique, discussed in section 4.3, should be borne in mind during the ensuing discussion i.e. major erosional contacts between facies imply the possibility of non-depositional and perhaps, the beginning of a new depositional cycle. It will be shown that depositional cycles are present within the Little Burland sequence which implies that major depositional breaks must be present. Although facies 2 (pure mudstone) and facies 5 (cross-bedding) are not present within the Little Burland sequence, their relationship to other facies will be examined at the appropriate point.

Text-figure 7.9 reveals two distinct facies groupings within the Little Burland sequence. The first grouping comprises thinly interlayered sandstone/mudstone bedding (facies 1), thinly bedded sheet sandstones (facies 3) and hummocky cross-stratification (facies 4). The second grouping comprises wavy bedding (facies 6), bipolar ripple cross-lamination (facies 7) and parallel-laminated sandstone with thin cross-laminated zones (facies 8). Tunbridge (1983) undertook a similar observed-minus-random facies sequence analysis for the adjacent Great Burland sequence, where a more varied sequence is exhibited, and established similar facies sequence groupings to those at Little Burland, namely 'shelf sequences' and 'shoreface-foreshore sequences' respectively. At Little Burland the two sequences are related by both groups exhibiting a tendency for transition to facies 1. Each grouping will be considered in turn below.

In the lower part of the Little Burland section the facies sequence exhibits a weak tendency for transition from thinly interlayered sandstone/mudstone bedding (facies 1) to thinly bedded sheet sandstones (facies 3) or hummocky cross-stratification (HCS - facies 4). Both facies 3 and 4 are interpreted as the product of random storm events interrupting muddy shelf sedimentation (facies 1) below fair-weather wave base. There is a much stronger reverse tendency for facies 3 and 4 to return to facies 1, reflecting the tendency for random event deposits to revert to 'background' sediments.

The thin sheet sandstones almost invariably occur in single beds interspersed with thick sequences of thinly interlayered sandstone/mudstone bedding, whereas HCS, although occasionally occurring in single beds, tends to form 1 to 4m thick amalgamated units referable to the 'Woody Bay facies association'. These amalgamated sandstone bodies equate to Tunbridge's (1983) 'multi-storey beds' of parallel-laminated sandstone and HCS interpreted as outer shoreface shoals. Discussion of the geometry of this sandstone body type (7.6.3) indicated, however, that amalgamated HCS units can represent either tabular or lensoid geometries and can be associated with either shoreface or open shelf deposits. Indeed, amalgamated HCS units are associated with deposits interpreted as both open shelf (Woody Bay west and WNW Barbrook) and shoreface (uppermost Little Burland) in origin -see interpretations of individual facies.



Text-fig. 7.9 Observed-minus-random probability diagram, for sequence at Little Burland, showing transitions with a greater than random probability of occurring.

Calculations shown at Appendix G. Facies: **1** = thinly interlayered sandstone/mudstone bedding; **3** = thinly-bedded sheet sandstones; **4** = hummocky cross-stratification; **6** = wavy bedding; **7** = bipolar ripple cross-lamination; **8** = parallel-laminated sandstone with thin ripple cross-laminated zones.

The open shelf and shoreface associated interpretations are corroborated by the fact that the Woody Bay west and WNW Barbrook sections contain a significantly higher proportion of graded rhythmites and parallel-laminated sand-streaks in facies 1 when compared to Little Burland. Of the sandstone layer types found in facies 1, these two types represent the most distal environment, being deposited below and near storm wave-base respectively.

The facies grouping described above matches Tunbridge's (*op. cit.*) 'shelf sequences' comprising burrowed muddy sandstones containing erosion surfaces, lumachelles, sharp-based parallel-laminated sandstones and ripple cross-laminated beds - reflecting a distal to proximal trend of storm emplaced deposits respectively. Similar deposits interpreted as representing storm events occur associated with the facies grouping described above - i.e. storm generated erosion surfaces are present in facies 1 whilst lumachelles, although not considered as a separate facies herein, occur as lag deposits at the base of facies 3 and 4 beds; sharp-based parallel-laminated sandstones equate to facies 3 and the thin ripple cross-laminated beds match the thin beds of facies 7 described from WNW Barbrook (e.g. 5.1m on log) and the lower part of the Little Burland sequence. In addition to the storm event deposit types matching those described by Tunbridge, three further types are present in the sections under consideration in this chapter: pure mudstone units with sharp bases and which fine-upwards (facies 2), cross-bedded sandstone beds (facies 5) and HCS (facies 4). Facies 2 represents a storm event deposit sourced from an area of silts and muds whilst facies 5 is interpreted as the product of dune migration under a storm-generated geostrophic flow with low suspended sediment concentrations. Facies 4, however, is interpreted as originating under storm-generated geostrophic flow with high suspended sediment concentrations.

The proximal to distal classification of the storm units described by Tunbridge (*op. cit.*) is not followed herein as the character of the storm deposits is believed to reflect storm magnitude (random) and source area sediment type and availability, rather than direct proximity. For example, Tunbridge proposed that the erosion surfaces cutting muddy sandstones represented the most distal storm event deposits. This type of erosion surface, however, is visible in deposits high in the Little Burland sequence associated with proximal, shoreface sequences suggesting that the erosion surfaces represent an area of net erosion in a proximal storm environment in some cases. The only storm deposits believed to reliably indicate proximity are the graded rhythmites and parallel-laminated sandstone streaks representing a distal, offshore environment near storm

wave-base (see above) and the fauna within lumachelles at the base of HCS beds towards the top of the Little Burland sequence which indicate a turbid, proximal environment (see 7.6.3).

The alternation of bioturbated muddy deposits with storm sand layers is common in many modern shelf seas (Reineck & Singh 1971; Howard & Reineck 1972; Clifton 1976; Kumar & Sanders 1976; Aigner & Reineck 1982) and reflects storm events transporting sand from a sandy shoreface to a muddy shelf environment below fair-weather wave-base.

The second facies grouping revealed in text-figure 7.9 only occurs in the upper part of the Little Burland section. In this facies grouping there is a strong tendency for a transition from wavy bedding (facies 6) to bipolar ripple cross-lamination (facies 7), with a weak contrary trend, and a weak tendency for transition from bipolar ripple cross-lamination to parallel-laminated sandstone with thin cross-laminated zones (facies 8). Both facies 7 and facies 8 display a weak tendency for transition to thinly interlayered sandstone/mudstone bedding (facies 1). This facies grouping equates to Tunbridge's (*op. cit.*) 'shoreface-foreshore sequences'. The strong tendency for transition from wavy bedding to bipolar ripple cross-lamination reflects an upwards transition from the emplacement of sheet sandstones by minor storm events in a muddy outer shoreface/inner shelf setting proximal to an abundant sand supply (see 7.8 interpretation) to ripple cross-laminated sands accumulating in a mid-shoreface environment where sustained periods of oscillatory current action prevailed (see 7.9 interpretation). The weak contrary tendency suggests that the upwards transition from facies 6 to 7 was not uniform, the two facies interdigitate (e.g. 47.5 to 51.4m on the log) in response to a variation in wave-base depth, possibly seasonal, or a change in shoreline/basin configuration affecting the wave climate. Furthermore, facies 7 itself comprises a combination of clean, well sorted bipolar ripple cross-laminated sands reflecting prolonged periods of oscillatory current activity within fair-weather wave-base; wave-flaser bedding reflecting quiescent periods allowing mud deposition between periods of active sand bedload transport; and bioturbated zones representing the preservation of biogenic structures in an environment temporarily below fair-weather wave-base where sands were not actively reworked. The variation of physical and biogenic structures and sediment types preserved by facies 7 reflects fluctuations in wave-base in response to seasonal and/or basin configuration influences on wave climate in a middle shoreface setting.

Tunbridge (*op. cit.*) observed that several sequences in the Great Burland sequence show a transition from lenticular bedded inner shelf deposits to multi-storey parallel-laminated sands and hummocky cross-stratified outer shoreface shoal deposits, omitting wavy and flaser bedded facies. As noted above, HCS shoals are present in both open shelf and shoreface associations within the sequences under consideration in this chapter. Although no HCS shoal sub-sequence is revealed by text-figure 7.9 as a variant to the wavy bedded and flaser/cross-laminated lower to middle shoreface sequence, examination of the Little Burland log reveals examples of amalgamated HCS units passing upwards into thick zones of ripple cross-laminated sandstone (e.g. 47.5m on log) which equates to the outer shoreface shoal to middle shoreface deposits catalogued by Tunbridge.

The transition from facies 7 to parallel-laminated sandstone with thin ripple cross-laminated zones (facies 8) has a low probability, due in part to the limited number of occurrences of this facies at Little Burland. Although the contrary transition is present in the Little Burland sequence (e.g. 64.1m on log), the transition has a random probability of occurring (text-figure 7.9). This suggests that once a prograding shoreface sequence is established with wavy bedding passing upwards into flaser bedding / bimodal ripple cross-laminations, it will inevitably pass upwards into parallel-laminated sands with thin ripple cross-laminated zones attributable to an upper shoreface - foreshore environment. As noted in the discussion of facies 8, there is no unequivocal evidence of intertidal exposure preserved in the Little Burland sequence. Thus, the shallowest deposits preserved within the Little Burland sequence equate to an upper shoreface setting.

In summary, the character of individual facies and the observed-minus-random facies transition analysis of the Little Burland sequence indicate that the Woody Bay west and WNW Barbrook exposures preserve a storm-dominated open shelf environment whilst the Lynton Formation - Hollowbrook Formation transition exposed at Little Burland preserves the progradation of a wave-dominated shoreface over a storm-dominated open shelf environment.

The Little Burland sequence does not range as high into the Hollowbrook Formation as the adjacent Great Burland sequence described by Tunbridge (*op. cit.*) and there is no evidence of foreshore and alluvial plain facies preserved at the former locality. Nevertheless, the two sections are similar in that they both comprise an overall coarsening-upwards (CU) progradational shoreline sequence composed of smaller scale CU sequences. Examination of the Little Burland log shows individual CU sequences separated at heights of

35.4, 59.6 and 67.2m on the log giving four CU sequences within the overall CU sequence (see arrows on enclosures 11A & 11B indicating the bases of the CU cycles); this compares with seven CU cycles within the Great Burland sequence (see figure 11 of Tunbridge *op. cit.*) where a more complete sequence of the Hollowbrook Formation is preserved. The origin of these cycles is discussed in section 8.3.

7.12 DISCUSSION

The preceding section documents a model for the upper Lynton Formation 'wedge' and the transition to the succeeding Hollowbrook Formation. Deposition on a muddy storm-dominated open shelf attributable to Seilacher's (1967a) *Cruziana* ichnofacies (Woody Bay west, WNW Barbrook and the lower part of the Little Burland section) was succeeded by a wave-dominated shoreface sequence exhibiting a transition from Seilacher's *Cruziana* ichnofacies to a *Skolithos* ichnofacies (upper part of the Little Burland section). A comparison of a bivalve fauna recovered from a locality 200m west of the WNW Barbrook section with a fauna recovered from the Little Burland section provides corroborative evidence for an increase in environmental energy when comparing the latter with the former section (see section 7.6.1). This model closely accords with that presented by Tunbridge (*op. cit.*) for the adjacent Great Burland sequence, with minor divergences outlined above. Tunbridge proposed the low wave energy, coastal environment of Sapelo Island, Georgia (Howard & Reineck 1972) as a modern analogue for the Great Burland sequence. Both the Great Burland and the Little Burland sequences differ from modern high wave energy shoreline sequences (Clifton *et al.* 1973 and Davidson-Arnott & Greenwood 1976) in that they lack a zone of megarippled sand formed in the nearshore zone by the asymmetric orbital motions of shoaling waves. Tunbridge, however, recognised that the production of megaripples by asymmetric wave orbitals is dependent on a medium to coarse sand grade (Clifton 1976), a grain size fraction absent in the Great Burland and Little Burland sequences and their modern analogue off Sapelo Island. Tunbridge went on to note that the modern low and high wave energy shorelines sequences cited above also differ in the degree to which primary sedimentary structures are biogenically reworked in the offshore and lower shoreface environments, the low wave energy shoreline sequences exhibiting a much greater degree of reworking.

Ancient shoreline sequences attributed to low wave energy settings (Davies *et al.* 1971; Howard 1972; Elliott 1975) also appear to contain a far higher degree of biogenically reworked structures than those deposited in sequences proposed as having originated in a high wave energy environment. (Vos & Hobday 1977;

Bourgeois 1980). On the basis of the high degree of biogenic reworking of deposits interpreted as representing offshore and lower shoreface environments in the Great Burland sequence, Tunbridge proposed that the highest Lynton Formation and succeeding Hollowbrook Formation record a low energy wave-dominated sandy marine regressive sequence. Thayer (1979) has shown, however, that the relative proportion of biogenic activity has increased through the geological record. Thus, consideration of the proportion of bioturbation in a stratigraphically isolated section without reference to a contemporary section is invalid. Fortunately, a contemporary section is available to allow a comparison. The Oxen Tor section, described in chapter 6, records the progradation of a relatively high-wave energy shoreline. Comparison of the bioturbation columns on the logs of Oxen Tor (enclosure 9) and Little Burland sequences, for facies interpreted as offshore and lower shoreface in origin, reveals that the degree of biogenic reworking is greater in the Little Burland sequence. This suggests that the Little Burland section represents a relatively lower energy sequence than the Oxen Tor sequence. Comparison of the Little Burland and Oxen Tor sequences, therefore, broadly confirm the assertion of Tunbridge that the Great Burland section records the progradation of a low energy wave-dominated sandy shoreline, with the caveat that the label 'low energy' is taken as relative to the demonstrably high wave energy sequence preserved at Oxen Tor.

Progradation of the Lynton Formation -Hollowbrook Formation shoreline was achieved by wave reworking of sediment supplied to the shelf at rather ill-defined input zones by ephemeral streams (Tunbridge 1981a), there being no evidence for established channels reaching the semi-arid shoreline to produce distinct prograding deltaic lobes (Tunbridge 1983a). Tunbridge suggested that occasional temporary stream deltas may have developed, but their lack of preservation suggested to Tunbridge that the deltas would have been reworked soon after development (cf. Glennie 1970 p.124). It should be borne in mind, however, that present day semi-arid shorelines can exhibit deltaic sequences in which channel facies are restricted to a narrow band normal to the shoreline. Johnson (1982) described the facies of the Gascoyne delta of Western Australia which is developing on a passive cratonic margin with a semi-arid climate with coastal processes dominated by waves. Johnson noted that the semi-arid climate is only reflected in the subaerial delta environments, where red-brown desiccated muds are deposited; the marine sphere is not affected. The Gascoyne River flows intermittently, the river channel being straight, confined and displaying little braiding; the river does not migrate laterally across the deltaic plain. Any preserved stratigraphic sequence would comprise a narrow band of channel and levee deposits surrounded by extensive marine delta front deposits. Applying the analogue of the Gascoyne delta to the Lynton Formation - Hollowbrook Formation shoreline it is possible

that a narrow band of channel and levee deposits are preserved but are not visible due to the limited outcrop at this stratigraphic level. The delta front deposits would be indistinguishable from deposits preserved for a linear wave-dominated shoreline on the scale of outcrop available for this study.

8. CONCLUSION: EVOLUTION OF THE 'EXMOOR BASIN' AND THE LYNTON FORMATION SEQUENCE

8.1 INTRODUCTION

Webby (1966) coined the term 'Exmoor Basin' for the thick Devonian-Carboniferous sequence deposited to the north of an E-W line at the approximate latitude of Bideford, within which the Lynton Formation was deposited. Discussion in the preceding chapters has revealed that during the period that the Lynton Formation was deposited the basin was tectonically active. In particular, syn-sedimentary activity along the Lynmouth - East Lyn Fault appears to have left a conspicuous signature within the Lynton Formation sequence. Dewey (1982) and Sanderson (1984) have suggested that the 'lithospheric attenuation' model of McKenzie (1978) adequately explains the evolution of the 'Exmoor Basin', Dewey proposing a β stretching factor of 2 (50% lithospheric thinning) to explain the pattern of sediment accumulation rates over time, whilst Sanderson favoured "... a slightly lower value" (p.160) with a Moho rise of \cong 6km. On a wider scale, a growing body of literature from southern mainland Britain (see review in section 1.6.3) and southern Ireland (Graham & Clayton 1989) documents the development of localised sedimentary basins across the southern British Isles during the Upper Palaeozoic. The following section (8.2) is an attempt to use more recent stratigraphic thickness and biostratigraphic data to provide a more detailed subsidence plot against time to test the models developed by Dewey and Sanderson. The revised model is then discussed in relation to the sedimentary pile that developed within the 'Exmoor Basin'. This is followed by a section (8.3) that examines the sequence stratigraphy of the Lynton Formation and the relationship of the depositional record to global sea-level change, subsidence rate and sediment supply and their influence on local accommodation space over time. Finally, suggestions for future research are made (section 8.4).

8.2 EVOLUTION OF THE 'EXMOOR BASIN'

Before commencing a discussion on the β /subsidence plot for the 'Exmoor Basin' a cautionary note should be given. Brooks and Kiriakidis (1986), discussing the subsidence of the North Aegean Trough, warned that an overestimate of the amount of lithospheric thinning may result from considering basin subsidence rates in isolation. A more reliable method was suggested involving summing the displacement on listric normal faults, which flatten out into a décollement horizon at some depth in the crust. The total displacement across the faults gives the amount of lateral extension, from which the β stretching factor for the lithosphere can be

calculated if it is assumed that the deeper crust has undergone an equivalent amount of ductile stretching. Unfortunately, Variscan thrust reactivation of extensional faults (e.g. Lynmouth- East Lyn Fault) and later Mesozoic extensional reactivation (e.g. Bristol Channel Fault Zone) render such a technique impossible for the 'Exmoor Basin'. Nevertheless, the re-calculated β /subsidence plot for the 'Exmoor Basin' accords with independent geological evidence, as will be shown below.

It should also be noted that no attempt has been made to plot eustatic sea-level changes on the curve. Although qualitative curves for the Devonian (House 1983, Johnson *et al.* 1985) and Carboniferous (Ramsbottom 1979), and semi-qualitative curves based around measurements of onlap for the Late Palaeozoic (Ross & Ross 1987) are available, no accurate quantitative curves are available. The magnitude of any eustatic events, however, is not likely to grossly affect the overall subsidence curve.

The first feature of note on the β /subsidence plot (text-figure 8.1) is that the curve does not show simple exponential subsidence typical of pure extensional basins (cf. McKenzie 1978). Instead, rapid subsidence during the Devonian rapidly lessened in the early Tournaisian and was followed by a more gentle phase of subsidence through the Carboniferous. This pattern is characteristic of extensional basin development in strike-slip settings ('transtension' - Harland 1971) where lateral heat conduction to the basin walls results in rapid cooling and subsidence is greater than the rate predicted for the rift stage in the uniform extension model; the critical threshold below which lateral heat loss to the sides of basins becomes important is between 100 and 250km (Allen & Allen 1990). Thus, the simple extensional models of Dewey (1982) and Sanderson to account for the 'Exmoor Basin' are rejected. Furthermore, the model presented in text-figure 8.1 indicates a β stretching factor of 1.1 during the thermal subsidence phase; this compares with a value of c. 2 given by Dewey and Sanderson. This suggests crustal thinning of only c. 10% compared to 50% implied by the earlier models.

Miall (1984) presented a number of features which characterise the depositional style of strike-slip basins. These are described below (italicised) with observations on their applicability to the 'Exmoor Basin':

- (i) *Basin geometry is deep and narrow - evidence of syn-depositional relief may include conglomerates and breccias localised along faults delineating basin margins. Sediment accumulation rates are rapid.* The geometry of the 'Exmoor Basin' is unclear but: "The pronounced straight northern margin of the Culm

Basin presumably reflects a steep, early extensional fault that may have controlled basin formation and became inverted during foreland deformation” (Hecht 1992 p.37) - this line may mark the southern margin of the ‘Exmoor Basin’ and coincides approximately with the position of the southern margin of the ‘Exmoor Sink’ described by Matthews (1977). The northern margin of the ‘Exmoor Basin’ would have extended no further north than the Bristol Channel Fault Zone. The occurrence of the localised presence of an intraformational conglomerate, attributed to a debris flow origin (section 2.2.3), along the line of the Lynmouth - East Lyn Fault and palaeocurrent patterns preserved within the ‘Watersmeet lithotype’ at Watersmeet (section 4.2.4.1) indicate that the faultline was a positive feature during the late Emsian. As discussed above, from the late Emsian through to the close of the Devonian sediment accumulation rates were high.

(ii) *Lateral facies changes are rapid e.g. breccias along faulted margins pass laterally into lacustrine muds very rapidly.* Both the intraformational conglomerate noted in (i) and the ‘Watersmeet lithotype’ thin rapidly away from the line of the Lynmouth - East Lyn Fault.

(iii) *Syn-depositional unconformities within basins, and differing stratigraphies in adjacent basin fills, reflect syn-depositional tectonic activity on basin-margin fault lines.* Many authors have remarked on the variability of fills within adjacent basins across SW England (e.g. Matthews 1977, Goldring & Langenstrassen 1979, Bluck *et al.* 1989, Selwood 1990).

(iv) *The basin sediments frequently exhibit features, e.g. sediment petrology and hinterland geology, indicating offset from their source.* No examples have been documented from the ‘Exmoor Basin’.

(v) *Basins frequently exhibit offset of geomorphological features e.g. fluvial channels and alluvial fans.* No examples have been documented from the ‘Exmoor Basin’.

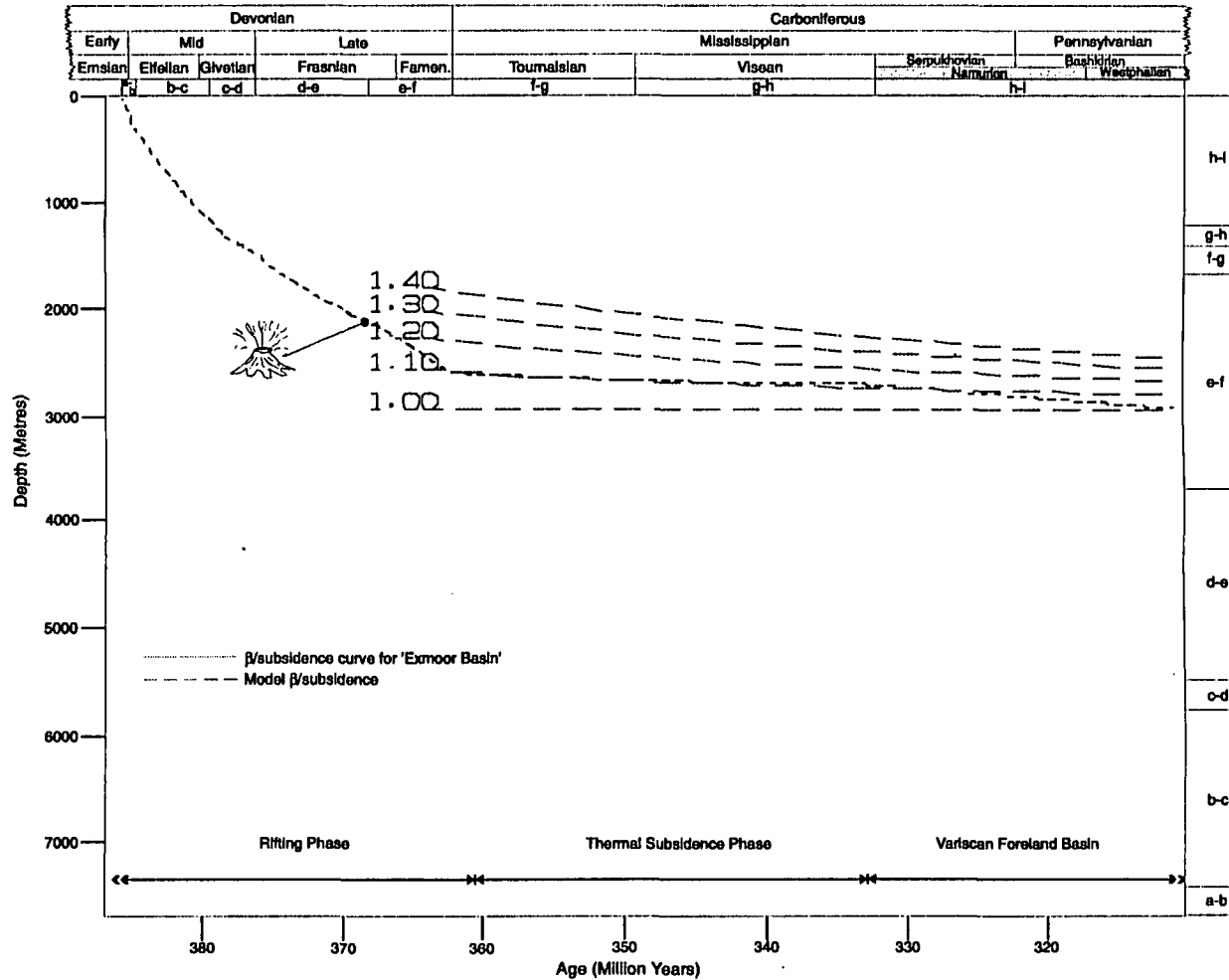
The above lines of evidence strongly suggest that the ‘Exmoor Basin’ had a strike-slip character. Gibbs (1987) classified basins into: extensional on listric faults, extensional on domino faults (pure shear), strike-slip (simple shear) and mixed-mode (general strain) and noted that: “In the UK, on and offshore there is increasing evidence from commercial seismic data that most basins are not simply extensional but are of mixed mode types (transpressional and transtensional basins cf. Sanderson & Marchini 1984; Harland 1971)” (p.19). It is concluded that the new transtensional basin model, based on evidence presented in text-

figure 8.1) is more consistent with the preserved geological evidence (see section 2.2.4) than the simple extensional models proposed by Dewey and Sanderson.

Text-figure 8.2 shows that during the Devonian 6000m (compacted figure) of sediment was deposited during a period of 23.5my i.e. $\cong 0.25\text{m} / 1000\text{yrs}$. This figure compares with other Devonian basin successions in NW Europe where the following deposition rates have been published: 0.4m / 1000yrs. (not stated whether the figure is for compacted or decompact rate) for the Mid to Late Devonian Munster Basin of southern Ireland (Graham & Clayton 1989), 0.3m / 1000yrs. (compacted) for the Early to Mid Devonian Lenne and Siegen troughs of the Rhenish Schiefergebirge (Goldring & Langenstrassen 1979). Powell (1989) calculated **uncompacted** deposition rates for the Pragian to Emsian Cosheston Group in SW Dyfed on either side of the syn-sedimentary Benton Fault. The deposition rate for the hanging-wall sequence was given as 0.2m / 1000yrs., whilst the foot-wall sequence was only 0.023m / 1000yrs. Deposition rates (uncompacted) of up to 2m / 1000yrs. have been recorded for some strike-slip basins (Miocene Ridge Basin of California where 10km of sediment accumulated in 5 million years - Link & Osborne 1978).

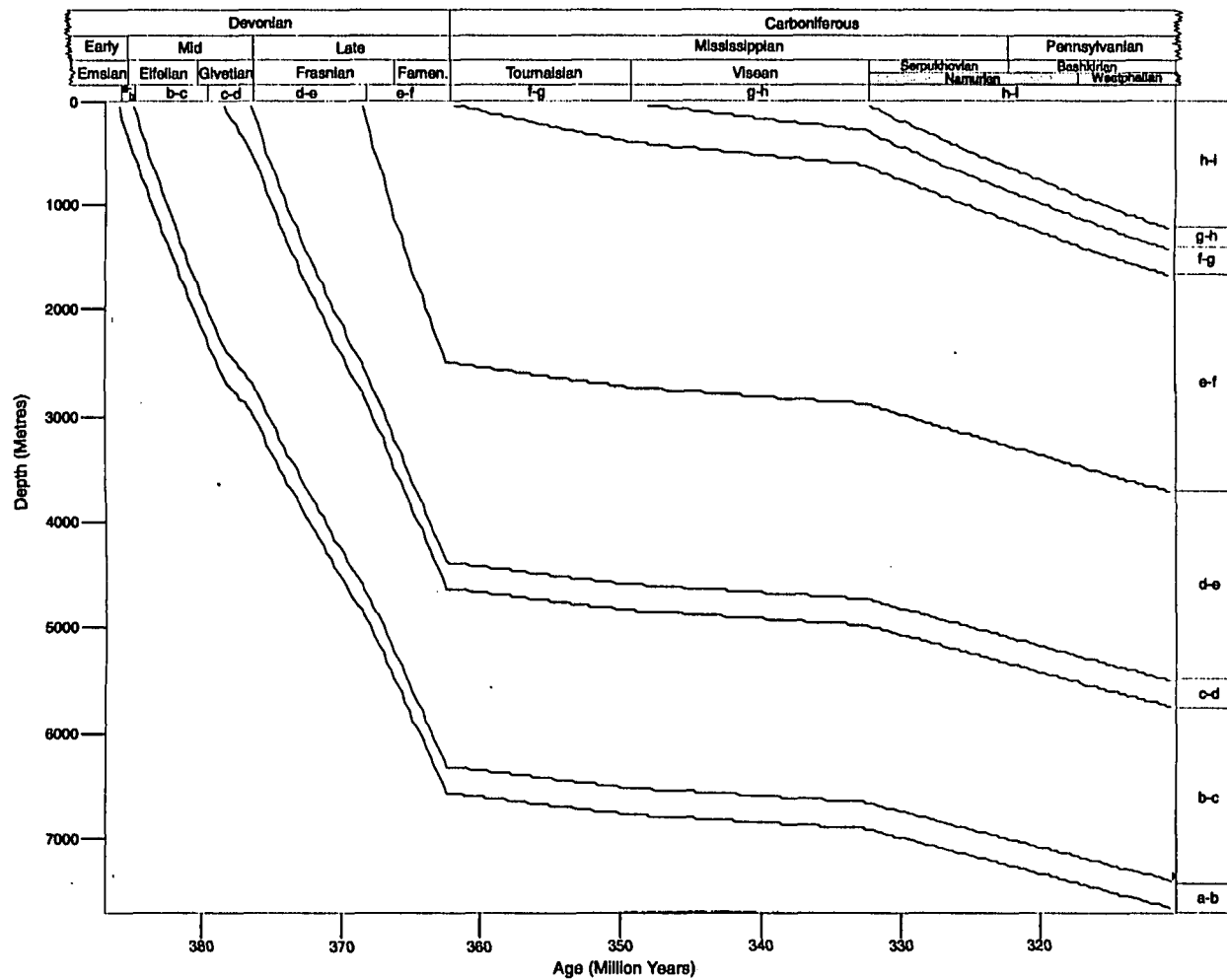
Thus, in contrast to many strike-slip basins, the 'Exmoor Basin' was not starved of sediment. A sizeable hinterland is suggested i.e. Allen and Allen (1990 p.132) observed that ". . . the size of drainage basin required to continuously feed the sedimentary basin and keep pace with tectonic subsidence is extremely large (several hundred times the size of the depositional area)".

The timing of basin initiation could not be ascertained due to the concealment of the succession below the exposed Lynton Formation. If the exposed Lynton Formation is underlain by Dartmouth Group alluvial facies (see section 8.3) then Emsian basin initiation may correlate with: "Extensional development of the Gramscatho basin (which) probably began in the upper part of the Lower Devonian" (Barnes & Andrews 1986 p.120) and the Mid Devonian onset of the Trevone and South Devon Basins (Selwood (1990), suggesting a widespread crustal stretching event. The 'rift phase' (see text-figure 8.3) following this event would have been dominated by a series of localised 'half-graben' type basins along the marginal areas of stretched crust (see text-figure 2.9), whilst 'pull-apart' basins, locally floored by oceanic material (e.g. the 'Gramscatho Basin'), would have developed further south where crustal thinning was greater.

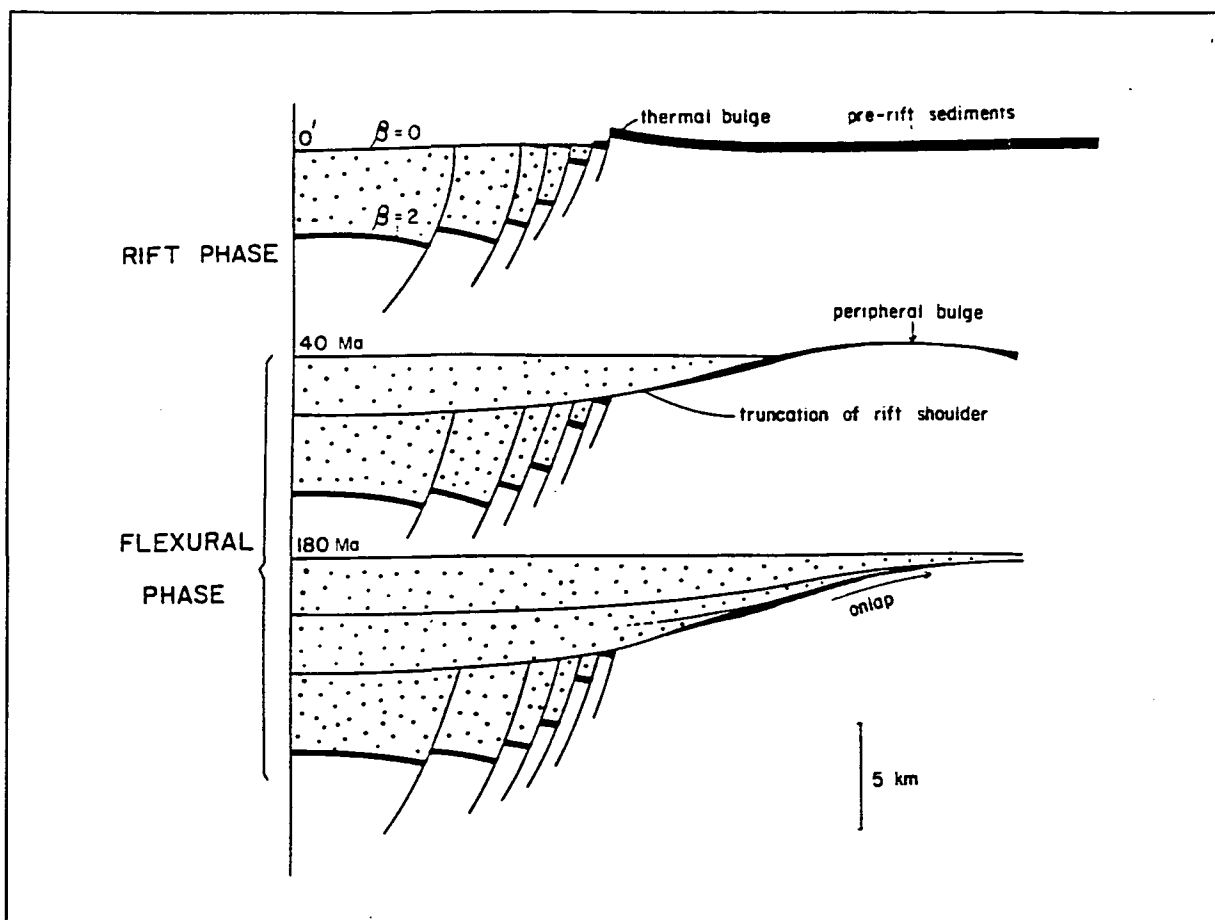


Text-fig. 8.1 β /subsidence plot for the 'Exmoor Basin' with thick (1500m) Morte Slates

The volcano symbol represents the 'Bittadon felsite' - a keratophyric tuff of contemporaneous volcanic origin found locally at the base of the Pickwell Down Sandstone. The data used to derive this plot are documented in appendix D. 'a-b', 'b-c' etc. represent the stratigraphic intervals described in appendix D. See text for discussion.



Text-fig. 8.2 Stratigraphic burial history for the 'Exmoor Basin' with thick (1500m) Morte Slates



Text-fig. 8.3 Evolution of a sedimentary basin during rift and flexural (thermal) phases assuming a lithospheric stretching model.

(From Dewey 1982 after McKenzie 1978)

There are several points of note relating to the deposits that accumulated during the Devonian rifting phase of the evolution of the 'Exmoor Basin':

Firstly, the thick continental deposits of the Trentishoe Formation within the Hangman Sandstone Group represent a major influx of sediment that stemmed an Eifelian eustatic transgressive phase and was derived from reworking of highest Old Red Sandstone deposits from the south Wales area (Tunbridge 1986). This period of erosion gave rise to the regional Mid Devonian unconformity that separates the Lower from the Upper Old Red Sandstone and beneath which the Lower Old Red Sandstone is gently folded (Jones 1956). The Mid Devonian unconformity across south Wales has been attributed to: early proto-Variscan movement (Dunne 1983), a late pulse of Caledonian activity (Tunbridge 1986), the Mid Devonian Acadian orogeny of Canada (Soper *et al.* 1987) and to foot-wall uplift on a series of southward-dipping E-W trending fault zones e.g. the Vale of Glamorgan Axis (Wilson *et al.* 1988). The lithospheric attenuation model of McKenzie (1978), however, predicts the development of a 'thermal bulge' at the margins of extensional basins (simple and strike-slip) during the rift phase (see text-figure 8.3). The development of a thermal bulge beyond the

edge of the 'Exmoor Basin' alone would have been of insufficient lateral extent to account for the Mid Devonian unconformity across south Wales. But if the 'Exmoor Basin' was one of many basins developing across southern Britain during the Mid Devonian, as was argued for above, then the predicted thermal bulge likely to result from such laterally extensive crustal extension would have been adequate to account for the observed Mid Devonian unconformity. The thermal bulge may have provided the driving force for the foot-wall uplift mechanism proposed by Wilson *et al.*

Secondly, Dewey (1982) notes that volcanics with an alkalic mafic chemistry become increasingly likely towards the end of the rift phase. A volcanic event of this type was preserved in the deposition of the 'Bittadon Felsite' (see text-figure 8.1), a keratophyric tuff which is locally found at the base of the Pickwell Down Sandstone and is interpreted as being of contemporaneous volcanic origin (Edmonds *et al.* 1985).

Thirdly, Edmonds *et al.* (1985) gave a thickness range of 600 - 1500m for the Morte Slates, favouring the lower figure on the basis that structural repetition of the sequence was suspected. Both the 1500m (text-figures 8.1 & 8.2) and 600m values (text-figures D1 & D2) were input to the '*Theta*' package. It can be seen that the 1500m value results in a smooth β /subsidence curve, whilst the 600m value results in a pronounced convex-up segment departing from the overall exponential curve. Dewey (1982) has shown that such departures can be expected where deposition rate is less than the subsidence rate; measurement of the departure along the vertical axis giving the depth of the water column above the basin floor. Using this method a maximum water depth of 250m is indicated for the highest Morte slates (late Frasnian). Webby (1966), however, suggested that the Morte Slates represented delta platform muds with a deeper prodelta phase near the middle of the sequence. More recently, Knight (1990a) found marine palynofloras in the Morte Slates that ". . . are conceivably contemporaneous with the encasing sediment" (p.404) and: "The recovery of a *Maranhites* spp. complex from the Morte Slates provides the most convincing evidence that at least a few of the microplankton assemblages recovered from this sequence are truly contemporaneous" (p.406). The Morte Slates palynoflora was regarded by Knight to be compatible with a shallow marine environment. Taken together, the observations of Webby and Knight do not support a water depth of c. 250m for the Morte Slates. Thus, it is tentatively suggested that the higher 1500m for the thickness of the Morte Slates is more likely.

The lower part of the Pilton Shales (Devonian part i.e. last portion of interval e-f on text-figure 8.1) accumulated in a shallow marine environment, passing upwards into the more offshore, deeper water deposits of the upper Pilton Shales (Carboniferous part is initial part of interval f-g on text-figure 8.1). The Carboniferous portion of the Pilton Shales contain a fauna of corals, bryozoa, brachiopods, gastropods, goniatites, bivalves and trilobites (Edmonds *et al.* 1985) indicative of non-euxinic conditions. At the base of the Viséan, however, a transition occurs into the predominantly shaly Codden Hill Chert, a relatively condensed sequence containing dark pyritic shales with impure limestones and cherts; a nektonic fauna of predominantly goniatites and free-swimming bivalves is preserved. Goldring (1962b) referred to this period of slow deposition rates during the early Carboniferous as the 'Bathyal Lull'. This period of reduced sedimentation corresponds to a sharp shallowing of the β /subsidence curve in text-figure 8.1 and is interpreted as the onset of the thermal subsidence / flexural stage of basin development.

The lithospheric attenuation model of McKenzie (1978), predicts the development of a 'peripheral bulge' at the margins of extensional basins (simple and strike-slip) during the flexural / thermal subsidence phase (see text-figure 8.3). The peripheral bulge moves progressively away from the basin margin with time, its influence being more widespread than the thermal bulge of the rift phase. It is possible that the peripheral bulge to the 'Exmoor Basin' and generically associated basins of SW England is represented by the 'St Georges Land' / East Midlands platform massif - a Carboniferous massif that extended over a large part of Wales and central England (Anderton *et al.* 1979). As predicted by the McKenzie model, Mississippian sedimentation in south Wales is characterised by progressive onlap (see text-figure 8.3), although dextral strike-slip movement along E-W and NE-SW trending fault zones resulted in syn-sedimentary tectonics that considerably complicate the picture (Wilson *et al.* 1988).

During the Viséan the β /subsidence curve in text-figure 8.1 departs from the model subsidence curve. Using the vertical axis departure distance technique described above a water depth of $\cong 50\text{m}$ is indicated for the basin in which the upper Codden Hill Chert was deposited. It should be noted that the relative deepening shown by the curve g-h may, in reality, underestimate the true amount of deepening during this interval as the effects of the Viséan eustatic transgression (Ramsbottom 1978) are not included on the β /subsidence plot.

At the beginning of the Namurian a small, but significant, inflection in the β /subsidence curve in text-figure 8.1 suggests a sudden increase in deposition rates. By mid-Namurian times, and persisting into the Westphalian, the β /subsidence curve indicates that the sediment accumulation rate *exceeded* the model thermal subsidence rate, which at first sight appears not to be geologically reasonable. Recent work published by Hecht (1992) describing the Culm basin of north Devon, however, provides an explanation for this paradox. Tournaisian and Visean neritic black limestones, cherts and black shales (the Codden Hill Cherts) were replaced by foreland basin deposits during the Namurian. Initially, Crackington Formation distal turbidites were deposited during a period when the subsidence rate was low when compared with other Variscan foreland basins. During the late Namurian slumping occurred, the first slump was interpreted to be coincident with abrupt acceleration of the subsidence path of the Culm basin due to tectonically-induced slope steepening. This abrupt increase in subsidence gave rise to thick Westphalian A, B & C deposits of a shallow storm-dominated lake, with a few marine incursions, preserved as the Bude Formation (Higgs 1991) and succeeded by the southward prograding fluvio-deltaic Bideford Formation (de Raaf *et al.* 1965, Elliott 1976). In detail, the Culm basin was a 'piggy-back, passive roof type basin' developed on top of the northward-prograding Variscan orogenic wedge (see figure 2 of Gayer & Jones 1989); the thrust tip rode up and over the 'Bristol Channel Landmass' during the Westphalian to produce a palaeo-high that provided the northerly source for the Bideford Formation deltaic succession.

In summary, the β /subsidence plot shown in text-figure 8.1 supports the model presented by Hecht for an early Namurian onset to a foreland basin development associated with the northward advance of the Variscan orogenic wedge. The Culm foreland basin subsidence history overprinted the post-Visean thermal subsidence phase of the earlier 'Exmoor Basin'. It should be noted that the backstripped β /subsidence plot presented by Dewey (1982, figure 30) for north Devon indicated a mid-Carboniferous water depth of 500m. This figure is now clearly incompatible with the shallow lake interpretation for the mid-Carboniferous Culm basin and provides further evidence against the earlier simplistic pure extensional basin subsidence models proposed by Dewey (1982) and Sanderson (1984). It is also worthy of mention that Matthews (1977) plot of northward basin migration in SW England was essentially based on the initiation of greywacke deposition (i.e. flysch). Thus, Matthews was plotting the onset of foreland basin initiation (e.g. Bude Formation foreland basin infill) overprinting the major extensional basin infills e.g. 'Gramscatho' and 'Exmoor Basins'.

In an attempt to corroborate the proposed burial history for the 'Exmoor Basin' infill, presented in text-figure 8.2, published geothermometry data for north Devon were reviewed in order to establish the thermal profile that would have developed in response to burial and an associated increase in overburden pressure. Kelm and Robinson (1989) studied illite crystallinity profiles across north Devon and concluded that: "There is no definable variation in grade either in IC values or mineralogy . . . across the stratigraphic divisions of the Devonian" (p.150). Knight (1990a) constructed geothermal profiles based on conodont alteration indices and vitrinite reflectance data and also found no definable variation in values across stratigraphic divisions. Knight interpreted the lack of dependency of uniformly high vitrinite reflectivities (>300°C) on stratigraphic position as indicating that coalification was syn-orogenic. Thus, the available geothermometry data indicates that the geothermal gradient that would have been expected to develop during Devonian extension and subsequent thermal subsidence has been overprinted by regional metamorphism associated with the Variscan orogenic phase.

Structural and seismic evidence (discussed in section 1.6.4) indicated that a 'thin-skinned' structural model (cf. Shackleton *et al.* 1982, Cooper *et al.* 1986, Hanna in: Chapman 1986, Williams & Chapman 1986) is applicable to the Variscan deformation of southern Britain. The 'thick-skinned' model of Sanderson (1984) is rejected, although Sanderson's transtensional model to explain the extensional origin of the Rhenohercynian basins of southern Britain in an extensional strike-slip setting appears to be valid.

During Variscan compression, faults that previously had an extensional geometry during Devonian dextral transtension were reactivated to accommodate crustal thickening by a reverse (thrust) sense of movement in a dextral transpressional setting (Gayer & Nemcok 1994) e.g. Lynmouth - East Lyn Fault, Bristol Channel Fault Zone (Brooks *et al.* 1988, Gayer & Jones 1989), Severn Estuary Fault Zone and Vale of Glamorgan Axis (Wilson *et al.* 1988) and SW Dyfed (Dunne 1983, Powell 1989).

Thickening of previously stretched lithosphere by thrust reactivation of extensional faults (thrust flaking of an elastic lid to a viscoelastic lithosphere) has been discussed by Jackson (1980) and was given the term 'elastic lid model' by Dewey (1982, figure 3.1). Modern examples of thrust reactivation of extensional listric faults have been discussed from the Zagros Mountains of Greece by Jackson (1980); ancient examples have been described from the Aegean Sea area (Mercier *et al.* 1976, Sorel 1976). It is proposed that the elastic lid model adequately describes the geological evolution of the Rhenohercynian zone of southern Britain.

8.3 SEQUENCE STRATIGRAPHY AND RELATIVE CHANGES IN ACCOMMODATION SPACE

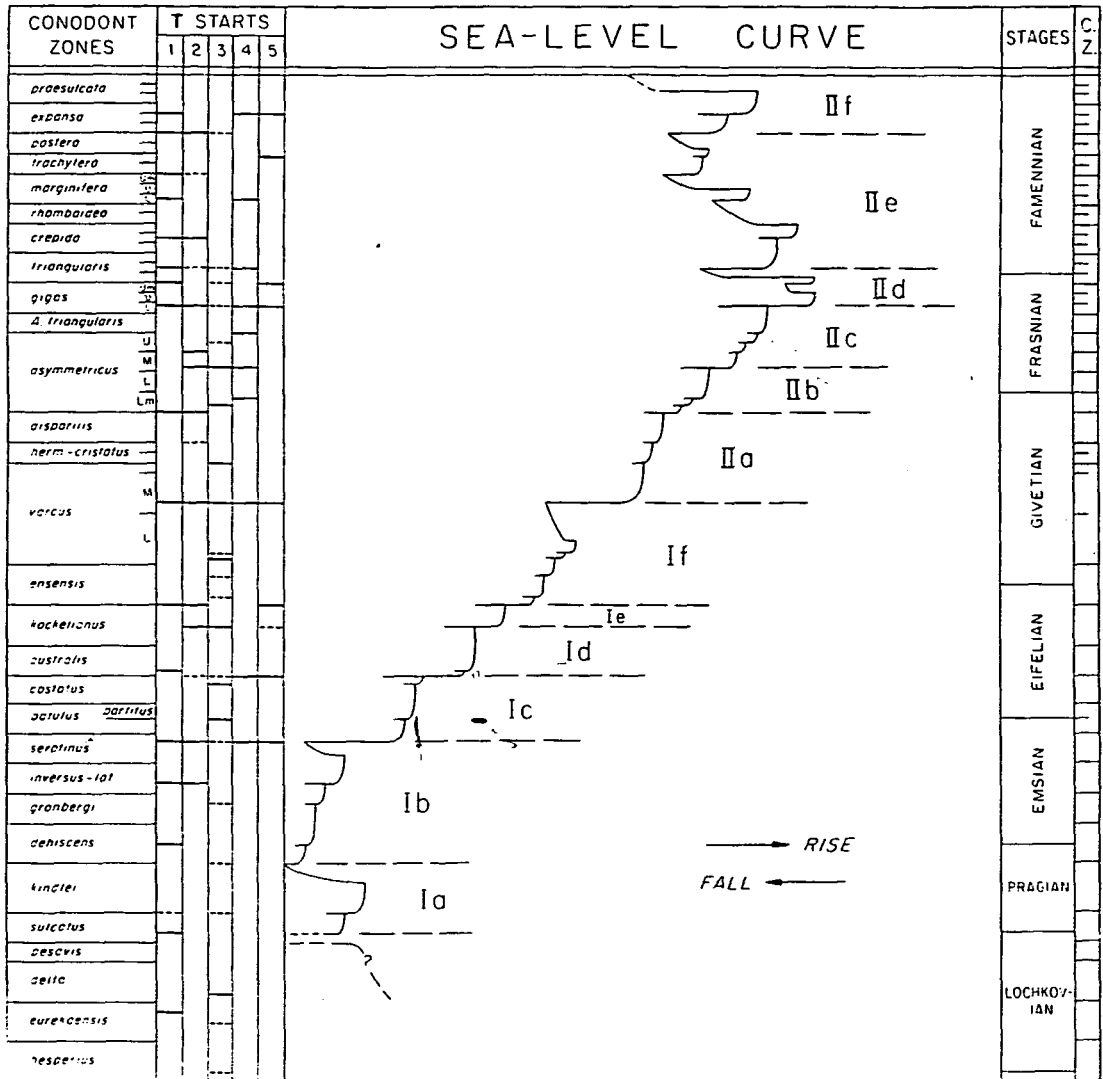
'Sequence stratigraphy', and its predecessor 'seismic stratigraphy', as developed by Vail and the 'Exxon school' has become widely accepted as a powerful tool in achieving high-resolution chronostratigraphic frameworks in which to place lithostratigraphic units (Wilson 1990). Recent studies have achieved a considerable degree of refinement in unravelling the complex interaction of eustatic sea-level changes, sediment supply, tectonic movements and subsidence. For example, Gawthorpe *et al.* (1994) described the Pliocene - Holocene of central Greece and demonstrated that sequence stratigraphy techniques can be applied to active extensional basins subject to frequent (i.e. glacio-eustatic) global sea-level changes to reveal characteristic sequence stacking patterns. The purpose of this section is to discuss the Lynton Formation succession in a sequence stratigraphic framework.

Enclosure 2 provides a conceptual section through the Lynton Formation normal to the line of the Lynmouth - East Lyn Fault with locations discussed in this thesis projected onto a two-dimensional plane. It is important to reiterate the point made in section 1.7 that, given the lack of lateral control and marker horizons within the Lynton Formation, the diagram can only ever be expected to be an approximation of the relative disposition of the various localities and logged sections. Furthermore, the diagram is based on lithostratigraphic thicknesses - no 'time-lines' are implied. Nevertheless, the diagram does serve a useful purpose in that it provides a framework for discussing the sequence stratigraphy of the Lynton Formation.

Johnson *et al.* (1985) produced a qualitative eustatic curve for the Devonian (reproduced in text-figure 8.4) and equated the onset of Lynton Formation deposition with the late Emsian transgressive event Ic of their scheme. Further south, the diachronous onset of the Meadfoot Group was related by Johnson *et al.* to onlap within cycle Ib. The exposed base of the Lynton Formation is characterised by large amounts of intraformational sliding and slump scars (chapter 3) which are typical of shelf deposits on relatively steep continental shelves generated during rapid onlap over continental coastal plain facies (Galloway 1989). The presence of phosphatic fragments and a thin bone-bed is also typical of transgressions which bring a fresh supply of phosphorous which can subsequently be used by organisms (chapter 3); the abundance of crinoidal debris at this level, relative to the remainder of the Lynton Formation, suggests that crinoid 'meadows' developed nearby in a zone of relatively low sedimentation rates, possibly on a topographic high formed by the foot-wall to the Lynmouth - East Lyn Fault (see text-figure 2.8). In summary, the basal deposits of the

Lynton Formation are consistent with highstand deposits which accumulated during the peak of a transgression. Biostratigraphic indicators (see section 1.8.4) are consistent with a late Emsian age for the base of the exposed Lynton Formation and the interpretation of Johnson *et al.* (1985) suggesting that Lynton Formation deposition was initiated by the Ic transgression is tentatively confirmed i.e. coincident with the onset of Staddon Grits deposition in the Plymouth area (see section 1.6.2.2.2). Thus, an inflection is shown near the base of the 'relative accommodation space' curve (enclosure 2), coincident with the approximate level of the bone-bed, to represent the peak of transgression. As the sedimentation rate subsequently increased, sediment building out into the basin would have presumably resulted in progressive downlap developing a maximum flooding surface. No evidence of this was observed, which is to be expected given the structural complexity and lack of lateral control at this level.

Moving upwards through the 'basal mega-facies', the Wingcliff Bay succession reflects a gradual shallowing sequence i.e. a decrease in relative accommodation space over time representing a gradual increase in sediment accumulation rate on the shelf as sediment built out over the maximum flooding surface.



Text-fig. 8.4 Qualitative eustatic curve for the Devonian

The diagram shows facies progression and T-R cycles and their relationship to Devonian conodont zones. Deepening events, plotted inside the eustatic curve relate to principle deepening events across the Devonian Euramerican platform. From Johnson *et al.* 1985.

The basal part of the succeeding 'lower middle mega-facies' is characterised by a series of sandstone-body developments: the 'Lee Stone facies association' (section 4.3.1), various developments attributable to the 'Watersmeet lithotype' (section 4.3.2) and a coarsening-upwards shoreface sequence developed in the A39 road section (section 4.3.3). The preservation of these sandstone-body types within a muddy shelf sequence is discussed in detail in section 4.4. In summary, movement on the Lynmouth - East Lyn Fault resulted in large amounts of sediment being catastrophically introduced into a muddy shelf environment. The effect of the fault movement on relative accommodation space is difficult to ascertain. Ignoring the effects of sediment influx for a moment, an increase in accommodation space would be expected to result from transtensional oblique slip movement on the Lynmouth - East Lyn Fault causing downward movement of the hanging-wall basin within which the Lynton Formation was deposited (section 8.2). However, each individual seismic episode on the faultline could only be expected to have resulted in < 1m subsidence of the hanging-wall floor to the basin (Yielding & Roberts 1992); the effect on accommodation space would have been negligible. Examination of the lower part of the cliff face below Castle Rock, which provides a section through the 'lower-middle mega-facies' above the Wringcliff Bay log, reveals no evidence of any significant change in base-level / sandstone-body development / major erosion surface at this, or indeed any other, level. Thus, the influx of sediment on to the floor of the hanging wall basin would have led to a geologically instantaneous relative decrease in *local* accommodation space. It is this phenomenon that is shown as the curves denoted by broken lines on the 'relative accommodation space' curve on enclosure 2. It is important to recognise that this phenomenon would have been localised. In areas where sediment was not introduced during the catastrophic episode, relative accommodation space would have remained unchanged (unbroken curve on enclosure 2 e.g. cliffs below Castle Rock).

Three separate broken line curves are shown on enclosure 2 as each location had a different local relative accommodation space history:

- (i) **'Watersmeet Lithotype'** This curve is drawn for the lime kiln exposures at Watersmeet where the 'Watersmeet lithotype' reached its thickest development (c. 4m thick - section 4.2.4). The 'Watersmeet lithotype' at this locality is interpreted as having been catastrophically emplaced at the foot of a submarine fault scarp and reworked by strong longshore currents. The deposit sharply overlies thinly

interlayered sandstone/mudstone bedding with a high carbonate content and has an abrupt top. For this reason the relative accommodation space is shown as a 'square-wave' curve on enclosure 2¹.

There is evidence for the catastrophic event that introduced the 'Watersmeet' lithotype at this stratigraphic level preserved further afield. At Ruddy Bull a slump scar occurs some 30 to 40cm below a coarsening-upwards sequence assigned to the 'Lee Stone facies association', whilst 2km to the WSW at Lee Stone a 0.4 to 1m thick plano-convex lens of the 'Watersmeet lithotype' occurs a metre or so beneath the 'Lee Stone facies association' (see text-figures 4.2 & 4.3). In the East Lyn valley a bed of the 'Watersmeet lithotype' occurs in a track-side exposure east of Lynmouth (see enclosure 2) and material of a similar composition to the 'Watersmeet lithotype' occurs in a storm-generated bed a metre above the base of the parallel-laminated shoreface deposits in the A39 road section (see text-figure 4.1) N.B. In chronological terms the slump scar was formed at the time of the tectonic movement, whilst the 'Watersmeet lithotype' sediment that was introduced by the tectonic event was subsequently reworked by shelf currents prior to being preserved in its current positions at Lee Stone and in the East Lyn valley and is consequently slightly younger than the slump scar at Ruddy Ball.

(ii) **A39 Coarsening-upwards Shoreface Sequence** The coarsening-upwards wave-dominated shoreface sequence is also interpreted as having been catastrophically emplaced at the foot of a submarine fault scarp. It is assumed that the geometry immediately following emplacement would have been lobate, but subsequent reworking by waves and longshore currents would have redistributed the sand to form a linear prism of shoreface sands running parallel to the strike of the Lynmouth - East Lyn Fault. The deposit abruptly overlies thinly interlayered sandstone/mudstone offshore deposits, suggesting a rapid shallowing occurred. Thus, the 'relative accommodation space' curve is drawn in part as a 'square wave', although the vertical component has been drawn as a steep curve to represent progradation of the shoreface following catastrophic emplacement of the sand. The point of catastrophic emplacement is marked by a graded mudstone event deposit interpreted to be the product of fallout of fine-grain material suspended by the seismic event that triggered the input of the sand. A thin bed of 'Watersmeet lithotype' material,

¹ The top of the broken line curve is not shown i.e. the time that elapsed before the sequence returned to 'normal' water depth (i.e. re-joined the solid main curve) is beyond the resolution allowed by the available preserved palaeo-water-depth indicators. Furthermore, the broken line curve is schematic; the broken line curve departure from the main solid curve would have been less pronounced if drawn to true scale i.e. the decrease in relative accommodation space that came about as a result of the catastrophic introduction of additional sediment would have been proportionately less in relation to the storm to fair-weather wave base depth difference than is shown by the curves.

one metre above the base of the shoreface sequence, is interpreted to be a storm-generated event deposit reworking material from a 'Watersmeet lithotype' sand ridge that originated from catastrophic emplacement of coarser-grained material at the base of the fault scarp further along strike. The top of the sequence is not exposed and it is not known whether foreshore deposits developed or whether the shoreface was backed by a lagoon. Small exposures in the slopes above the A39 road section indicate a return to offshore deposition with the preservation of thinly interbedded sandstone/mudstone - the curve for the top of the section has therefore been marked: ?????

(iii) **'Lee Stone facies association'** On enclosure 2 the 'relative accommodation space' curve¹ for the 'Lee Stone facies association' represents minor localised shallowing above a sand ridge development. The ridge developed in response to a massive influx of sand which has been linked to the event that emplaced the A39 shoreface sand prism (see section 4.4). There is supporting evidence for this hypothesis in that the 'Lee Stone facies association' comprises sandstones with a higher carbonate content than is typical for Lynton Formation sandstones. The petrography of the catastrophically emplaced sandstones at the base of the Lynmouth - East Lyn Fault scarp suggests a provenance from an area of high carbonate production (a palaeohigh formed by the foot-wall to the Lynmouth - East Lyn Fault?)

Sand that by-passed the 'littoral energy fence' during/following catastrophic emplacement was swept together over a period of millennia into a series of sand ridges spaced according to Huthnance (1982) bedform stability theory and/or situated over palaeo-highs; there is insufficient evidence preserved to state whether the bedforms had attained an equilibrium spacing conforming to Huthnance theory. Because the 'Lee Stone facies association' sand-bodies would have migrated back and forth across the shelf surface over a period of millennia they are preserved stratigraphically above deposits more directly linked to the major catastrophic event (the slump scar at Ruddy Ball and a bed of the 'Watersmeet lithotype' at Duty Point).

In summary, there is good geological evidence to support the correlations shown on enclosure 2, originally based on trigonometric considerations, for the sandstone-bodies preserved towards the base of the 'lower-middle mega-facies'.

At Duty Point the 40m high cliff face exposes a continuous section through thick mudstone-dominated heterolithic facies within the 'lower-middle mega-facies'. On closer inspection distinct metre-scale 'packets' can be distinguished which are marked by abrupt changes in the sandstone:mudstone ratio; the 'packet' boundaries are conformable and there is no evidence of any significant depositional hiatus. Internally the suite of primary physical and biogenic sedimentary structures remains the same across all packets, indicating that the 'packets' are not controlled by a significant change in water depth. Although the 'packets' are discussed at this point the phenomenon is seen throughout the mudstone-dominated components of mega-facies throughout the Lynton Formation e.g. cliffs below Castle Rock and the west Crock Point section.

It has not proven possible to laterally trace units defined on the basis of sandstone:mudstone ratios further than the available outcrop width at individual localities. This is best exemplified by the comparison of two laterally equivalent sections. The north-facing cliffs below Castle Rock (704 497) expose a near continuous section through the middle of the Lynton Formation above the logged section at Wringcliff Bay i.e. from between c. 25m to 180m above the base of the exposed Lynton Formation (see enclosure 2). Further west, the north-facing cliffs at Duty Point (695 497) expose a continuous section through the middle of the Lynton Formation above the logged section at Lee Stone i.e. from between c. 50m to 100m above the base of the exposed Lynton Formation (see enclosure 2). Although the exposures at these localities are very good, they are rather inaccessible due to the near-vertical nature of the cliff faces. It was possible, however, to construct vertical sequence profiles of sandstone:mudstone ratios by the use of binoculars, judicious use of abseiling and rock climbing techniques combined with the use of a rope, marked at 5m intervals, suspended down the cliff face to provide a scale. The resulting profiles revealed no correlation at any level, despite the fact that the two sequences are separated by only 900m laterally.

In order to better understand the supply of sediment to the shoreline, its subsequent dispersal and the resulting vertical changes in sandstone:mudstone ratios to be expected, it is first necessary to summarise the likely coastal setting. Tunbridge 1983a, discussing the Lynton Formation to Hollowbrook Formation shelf to alluvial plain transition at Great Burland and Hollowbrook, envisaged a distal coastal plain environment where sands were deposited by ephemeral sheet floods (Tunbridge 1981b) and where occasional high-sinuosity channels were interbedded with a mud-dominated clay playa facies (Tunbridge 1984). Flood sands reached the coastline occasionally, but no established channels systems entered the sea i.e. sediment reached

the shoreline via ill-defined zones from ephemeral streams. The flood sands reaching the coast would have been rapidly reworked and redistributed along the shoreline cf. modern coastal wadi sediments.

It is believed, therefore, that the heterolithic 'packets' preserve the changing pattern of sand and mud supply to the shoreline and subsequent dispersal across the shelf. It is not necessary to invoke eustatic sea-level changes to account for the presence of the 'packets'. The apparent lack of lateral continuity of the 'packets' is interpreted to be a function of sediment supply to the shoreline (point sources) and its subsequent distribution. With this interpretation, the expected 'packet' geometry would be lobate with very high thickness : lateral extent ratios. It is recognised that there may be a tectonic-control, at least in part, on the fluvial supply of sediment to the shoreline e.g. hinterland (foot-wall?) uplift; there is insufficient geological control, however, to assess the relative contribution of tectonic influences versus normal fluvial processes in controlling the type of sediment supplied to the shoreline.

As described in chapter 5, the anoxic thick mudstone towards the top of the west Crock Point succession marks the culmination of a gradual decrease in environmental energy (deepening?) and oxygen content of the substrate recorded over the significant period of time that it must have taken for the lower +90m of the west Crock Point to accumulate; the period of deposition coincides with the period of global eustatic transgression that intervened between transgressive events Ic and Id of Johnson *et al.* (see text-figure 8.4). Nevertheless, the onset of the conditions that gave rise to the anoxic mudstone was relatively abrupt. This was attributed in chapter 5 to tectonic uplift of the foot-wall block that defined the southern margin of the Lynton Formation resulting in the formation of a sill/lip that served to restrict circulation in the basin. As was described in section 2.2.4, if the fault blocks that floored the basins across Devon behaved in a 'domino' pattern, rotation on the blocks would have been synchronous and foot-wall uplift histories would have been similar for each basin. Thus, uplift of the 'Exmoor Basin' foot-wall to the south, that resulted in the formation of the sill, would have been accompanied by uplift of the foot-wall of the Lynmouth - East Lyn Fault (and Bristol Channel Fault Zone foot-wall blocks?) to the north (see text-figure 2.9). This uplift to the north would have resulted in sediment being shed to the south and may have initiated the southward (basinward) progradation of the shoreline preserved in the diachronous Lynton Formation - Hollowbrook Formation transition.

Before describing the Lynton Formation - Hollowbrook Formation transition, the expected effects on accommodation space as a result of the above set of events needs to be described i.e. the accommodation space curve for the upper part of the west Crock Point section. Accommodation space is shown as gradually increasing through the lower $2/3^{\text{rds}}$ of the west Crock Point section in response to gradual eustatic sea-level rise. A steepening in the curve, somewhat below the level of the 'thick mudstone', is drawn to show an increase in the rate of expansion of accommodation space to reflect the superimposition of tectonically-induced hanging-wall subsidence onto eustatically controlled sea-level rise. Note that this increase in accommodation space due to hanging-wall subsidence is not shown as a sharp inflection in the curve i.e. increase in accommodation space due to hanging-wall subsidence would have been the cumulative result of many individual fault movement episodes - each individual seismic episode on the faultline could only be expected to have resulted in < 1m subsidence of the hanging-wall floor to the basin (Yielding & Roberts 1992).

Following the period of anoxia, represented by the 'thick mudstone', basin circulation improved and the bottom waters were once again stirred-up and oxygenated and storm-waves could 'feel' the basin floor and transport the coarser sediment fraction as bedload. The substrate would have aggraded due to the influx of sediment produced by foot-wall uplift to the north, resulting in a decrease in accommodation space. The fact that this shallowing is superimposed on eustatic sea-level rise attests to the large volumes of sediment that must have been input to the basin at this time. This was accompanied to the north at this time by southward progradation of the shoreline i.e. recorded in the sequences proximal to the Lynmouth - East Lyn Fault.

Although the muddy heterolithic deposits in the upper Lynton Formation 'upper-distal mega-facies' wedge are broadly similar to those preserved lower in the Formation, i.e. deposited on a storm-and wave-dominated muddy shelf, the preserved sandstone-body type ('Woody Bay facies association') has a strikingly different character. The 'Woody Bay facies association' is characterised by hummocky cross-stratified (HCS) sandstones, whilst in comparison the sandstone-bodies in the lower Lynton Formation are characterised by cross-bedded units and a coarsening-upwards (CU) with a thin fining-upwards (FU) cap at their centre and CUFU margins ('Lee Stone facies association'). Both the 'Lee Stone facies association' and the 'Woody Bay facies association' are believed to have developed in response to obliquely-offshore directed geostrophic flow - yet the structures preserved internally differ significantly between the two facies associations. This divergence is believed to be due to one or more of three interrelated conceivable factors.

Firstly, possible evidence of tidal activity is preserved in the lower Lynton Formation (see section 4.2.4.3), whereas there is no evidence of tidal activity preserved in the upper Lynton Formation 'wedge'. The absence of tides in the upper Lynton Formation 'wedge' is thought to be due to these deposits forming in a basin which had become more restricted following the west Crock Point 'anoxia'. Erosion of the foot-wall scarp south of the exposed Lynton Formation would have resulted in the breaching of the sill that caused the restricted circulation at the time the anoxic muds were deposited. There would still have been a sufficient barrier, however, to prohibit the tidal wave from propagating into the upper Lynton Formation basin. It is suggested that the tidal currents in the lower Lynton Formation reinforced the semi-permanent (seasonal) geostrophic flow generated in response to onshore-directed trade winds - the resulting flow would have been sufficient to induce the migration of sandy bedforms during fair-weather conditions (see section 4.2.9). In the absence of tidal currents of measurable effect in the upper Lynton Formation 'wedge', however, trade wind induced geostrophic flow alone may not have been sufficient to initiate the migration of sandy bedforms allowing storm-produced HCS to be preserved.

Secondly, the presence of a barrier, even if submerged to a shallow depth, to the south of the area in which the upper Lynton Formation was deposited would have caused deep water waves to break and reform. This would have had the effect of reducing the apparent fetch of the 'Exmoor Basin' and thus, wave effectiveness at the seabed. This would have had a particularly profound effect on any semi-permanent geostrophic currents in the basin which, following the rise of the foot-wall barrier to the south, would have been significantly reduced. There is evidence that the semi-permanent geostrophic flow, an important component in bedload transport in the 'lower-middle mega-facies' (see chapter 4), was of diminished strength in the upper parts of the Lynton Formation. Coarsening-upwards microsequences (see section 4.2.2.3), which required semi-permanent geostrophic flow to establish, were only developed occasionally in the upper Lynton Formation e.g. Woody Bay west, and not at all in the Little Burland section.

Several workers are now starting to suggest that HCS forms in combined flows where the unidirectional component is relatively *weak* and suspended sediment fluxes are high (Nøttvedt & Kreisa 1987, Arnott & Southard 1990). If this interpretation is correct then the restriction of occurrences of hummocky cross-stratification to the upper Lynton Formation may be due to the unidirectional geostrophic flow component being weak relative to the oscillatory component. In the lower Lynton Formation, where strong (relative to oscillatory flow) geostrophic currents (enhanced by tidal currents) did apparently develop, the dominant

bedform in fine sand was cross-bedding, resulting in sandstone-bodies of the 'Lee Stone facies association' type being preserved.

The third factor to influence the internal structure of Lynton Formation sandstone-bodies was the proximity of the palaeoshoreline and sand supply. During lower Lynton Formation times the shoreline was more distal than the upper Lynton Formation 'wedge' sandy shoreline. Furthermore, sand may have been 'trapped' in the sub-basin to the north of the Lynmouth - East Lyn Fault whilst suspended mud by-passed the sub-basin and passed over the foot-wall barrier and into the hanging-wall basin that the Lynton Formation was deposited in. Certainly, the lower part of the Lynton Formation is muddier than the upper Lynton Formation; where sand-bodies did develop in the lower Lynton Formation the sand appears to have been catastrophically emplaced and reworked, over a period measured in millennia, into sand-bodies of the 'Lee Stone facies association' type (see above). Conversely, during storms affecting the proximal upper Lynton Formation sandy shoreline large volumes of sand would have been suspended in the upper shoreface region and swept offshore by geostrophic currents to be deposited under oscillatory flow conditions with a relatively weak unidirectional flow superimposed. As described above, high suspended sediment fluxes appear to play an important rôle in the formation of HCS.

The third factor does, however, raise a potential paradox: why was HCS formed off the 'upper-distal mega-facies' shoreline (Little Burland) but not off the 'upper-proximal mega-facies' shoreline (Oxen Tor, Valley of Rocks etc.)? The answer may be due to the 'upper-proximal mega-facies' shoreline being of a relatively higher wave-energy than the 'upper-distal mega-facies' shoreline (see above). High wave-energy shorelines develop a characteristic suite of pseudo-unidirectional bedforms (Clifton *et al.* 1971 - see discussion in chapter 6) under highly asymmetric oscillatory flow, whereas the pseudo-unidirectional bedform suite does not develop off lower wave-energy shorelines (see chapter 7). As noted above, HCS appears to only form in combined flows with a relatively small asymmetry / unidirectional component.

Knight [1990a, p.(i)] reported dates of "earliest Eifelian" for the uppermost Lynton Formation at Oxen Tor, Little Burland and Great Burland and the palynologically rich overlying Hollowbrook Formation. The base of the exposed Lynton Formation appears to equate with the late Emsian transgression Ic (text-figure 8.4) - see above. Transgressive event Id of Johnson *et al.* did not occur until well into Eifelian times (correlated with the *costatus* → *australis* conodont zone boundary - see text-figure 8.4), a date that is significantly

younger than the Hollowbrook Formation dates reported by Knight (*op. cit.*) Thus, the progradation of the upper Lynton Formation - Hollowbrook Formation shoreline took place during the period of eustatic marine transgression (House 1975, 1983, Johnson *et al.* 1985) that intervened between transgressive events Ic and Id of Johnson *et al.* (see text-figure 8.4). The progradational sequence implies, therefore, a high rate of sediment influx to stem the eustatic rise in sea level (cf. Tunbridge 1983a), the sediment being supplied from a rapidly uplifted source region in south Wales that gave rise to the Mid Devonian unconformity that separates Lower from Upper Old Red Sandstone.

The nature of the coarsening-upwards (CU) cycles within the upper Lynton Formation - Hollowbrook Formation progradational sequences at Little Burland, and those reported by Tunbridge (*op. cit.*) from nearby Great Burland and Hollow Brook Combe, corresponds to the 'asymmetric shoaling upward shelf cycles' of thick regressive deposits with only thin or non-existent basal transgressive records (Wilson 1975) now commonly referred to as *parasequences* (Vail *et al.* 1977). The parasequences are separated by *marine flooding surfaces* (arrows marking base of CU cycles on enclosures 11A & 11B) i.e. "a surface separating older from younger strata across which there is evidence of an abrupt increase in water depth" (Van Wagoner *et al.* 1988, p.39). Due to the lack of lateral facies control it is not possible to ascertain whether the cycles in the Lynton Formation - Hollowbrook Formation are the product of local shoaling or a regional scale mechanism where a rapid rise or fall in relative sea-level would have been followed by a phase of progradational seaward outbuilding. Furthermore, the lack of good lateral exposure has meant that it was not possible to directly establish the *stacking pattern* for the *parasequence set* i.e. whether the CU sequences are progradational / retrogradational / aggradational in origin. Although the Lynton Formation - Hollowbrook Formation transition CU cycles appear to conform to a classic *highstand systems tract* (Van Wagoner *et al.* 1988, p.44) no *maximum flooding surface* flooring the tract, or unconformity / *bounding discontinuity* terminating the tract was observed in the sections logged by the author. Therefore, the term *highstand systems tract* has not been applied.

The minor CU regressive cycles set within the overall regressive sequence could be analogous to Goodwin and Anderson's (1980, 1985) concept of 'punctuated aggradational cycles' (P.A.C.'s) wherein most stratigraphic deposits were believed to accumulate episodically in the form of thin shallowing-upwards cycles separated by surfaces of non-deposition. Goodwin and Anderson proposed that P.A.C.'s are controlled by phases of geologically instantaneous relative base-level *rises*, most P.A.C.'s having a basin-wide extent.

P.A.C.'s, however, are a special case where the rate of deposition is exactly balanced by the rate of accommodation increase (Van Wagoner *et al.* 1988, figure 1) i.e. they are interpreted to result from vertical *aggradation*, rather than *progradation*. Given the overall diachronous nature of the Lynton Formation - Hangman Sandstone Group boundary (enclosure 2) an *aggradational parasequence set* geometry can be ruled out. Furthermore, given that the younger 'upper-distal mega-facies' built out further in a basinward direction than the older 'upper-proximal mega-facies' (enclosure 2) the overall rate of deposition must have been greater than the increase in accommodation space. Thus, we can predict that the stacking pattern of the Lynton Formation - Hollowbrook Formation CU cycles would equate to a *progradational parasequence set* (Van Wagoner *et al.* 1988, figure 1).

Plint (1977) described a series of sharp-based shoreface sequences resting erosionally on offshore muds. The sequences were interpreted to represent erosion of offshore muds in advance of a shoreface moving seawards due to a rapid fall in base-level. As sea-level stabilised the shoreface continued to prograde in a seawards direction over the erosion surface cut into the offshore muds. Examination of the Little Burland sequence (enclosures 111A & 11B), however, indicates that the CU cycles consist of a basal zone of offshore lenticular bedding overlying an erosion surface cut into the top of the preceding shoreface sequence; there is no evidence of major erosion surfaces in the transition from offshore to shoreface deposits. In summary, Plint's rapid sea-level *fall* mechanism does not appear to be applicable in the case of the Little Burland sequence.

Care must be exercised with the parasequence interpretation, however, as local CU cycles representing weak transgressive/regressive pulses that did not induce a sedimentary response over the entire basin can occur, and cycles can amalgamate in nearshore zones where reduced subsidence and/or subsequent erosion can inhibit the preservation of the entire transgressive /regressive cycle. This phenomenon is well shown in Aigner's (1984) figure 11 where discrete cycles offshore become amalgamated nearshore. The cycles described by Aigner, from a storm-dominated carbonate ramp system of Upper Muschelkalk age in the intracratonic German Basin, are analogous to the cycles from the Lynton Formation - Hollowbrook Formation in that Aigner suggested that the minor cycles were not eustatically controlled. Aigner listed changes in the rate of sea level fluctuations, tectonic mechanisms (subsidence), changes in the earth's geoid and climatic-astronomic controls as possible causal agents for the smaller scale cycles and concluded that

tectonic controls were most likely since the minor shallowing-upward cycles appeared to be related to Variscan structural zones.

Given that the Lynton Formation - Hollowbrook Formation was deposited in a tectonically active setting (see chapter 2), and that the overall CU cycle was the product of a high rate of sediment influx from a tectonically uplifted source region in south Wales stemming a eustatic transgression at this time, it is reasonable to suggest that the smaller-scale CU cycles were tectonically controlled cf. Aigner (*op. cit.*) Also, see section 2.2.4 where a series of regressive cycles from the semi-arid Makran coast of Pakistan are described - the Makran provides a good analogue of the Lynton Formation progradational cycles.

It can be seen from the 'relative accommodation space' curves on enclosure 2 that the Little Burland sequence comprises several cycles of shoreface progradation, whilst the offshore sequences (Wringcliff Bay, West Crock Point, Duty Point) have smoother curves. This is interpreted to be a function of the degree of precision in assessing relative water-depth, and therefore 'relative accommodation space', preserved by the available geological indicators, as opposed to an increase in tectonic activity at the level of the Lynton Formation - Hangman Sandstone Group boundary. Shoreface sequences have a diverse range of facies belts related to water depth, whereas the relative water depth in offshore sequences must be defined in relation to the broad zone between storm and fair-weather wave-base i.e. a high proportion of graded rhythmites and horizontally-laminated sandstones within thinly interbedded sandstone/mudstone beds suggests a position close to storm wave-base, whereas a high proportion of connected and unconnected lenticular bedding suggests a position well within storm wave-base (section 4.2.1.3). This situation is exacerbated by the fact that storm wave-base was obviously not fixed i.e. storm wave-base would vary depending on the magnitude of each storm. The phenomenon of relative accommodation space resolution being finer in shoreface sequences than in offshore sequences is well known (Loutit *et al.* 1988). It is believed that in the Lynton Formation the offshore equivalents of the Little Burland CU cycles are the muddy heterolithic 'packets' described earlier in this section. It should be stressed, however, that the 'packets' are interpreted as having formed in response to a combination of tectonic and fluvial controls; the 'packet' frequency is likely to be higher than the CU cycle frequency.

In conclusion, the eustatic component of the 'relative accommodation space' curve shown on enclosure 2 is relatively simple i.e. from the 'bone-bed' near the base of the exposed Lynton Formation, interpreted as a

maximum flooding surface correlating with transgressive event Ic of Johnson *et al.* (1985), there is an increase in accommodation space as a result of eustatic transgression between events Ic and Id of Johnson *et al.* The complex pattern of changes in relative accommodation space superimposed on the eustatic component of the curve cannot be interpreted as glacio-eustatic in origin as the late Early Devonian to early Mid Devonian period was not subject to glaciation (Johnson *et al.* - *op. cit.*) The superimposed pattern is interpreted to be the result of tectonically-induced changes in accommodation space, primarily through controlling the rate of sediment supply. Thus, the statement by Johnson *et al.* (*op. cit.*, p.585) in discussing Devonian sea-level changes that: "The simultaneous occurrence of well-dated transgressive events in disjunct regions in a variety of depositional settings seems to require control by eustatic sea-level fluctuations great enough to overprint most ordinary subsidence histories" only applies in part to the observed 'relative accommodation space' curve for the Lynton Formation.

8.4 SUGGESTED FUTURE RESEARCH

Without a doubt the most perplexing task during the course of this study was the lateral correlation of locations and sections. In order to corroborate, or refute!, and refine the correlations presented in enclosure 2 a detailed mapping exercise, concentrating on the following aspects, would be of benefit:

- Confirm or disprove the lateral equivalence of the sandstone-bodies near the base of the 'lower-middle' mega-facies' i.e. 'Lee Stone facies association', the A39 coarsening-upwards shoreface sequence and 'Watersmeet lithotype' occurrences at this level.
- Trace the extent of outliers of the Hangman Sandstone Group at Hollerday Hill, The Tors west of Wind Hill and Summerhouse Hill. During the present study only upper shoreface deposits assigned to the basal Hollowbrook Formation of the Hangman Sandstone Group were found; an effort should be made to establish whether any continental Trentishoe Formation type beds occur within these outliers.
- Examine the structural fabric of veins and extension fractures, the orientation of minor- and medium-scale folds and determine thrust propagation directions etc. to confirm, or otherwise, that evidence from the dyke trend at Ramsey Beach (see section 2.3) for Variscan deformation being the product of dextral transpression is applicable to the Lynton area as a whole, and on a wider basis, the north Devon area cf. Andrews (1993), Gayer and Nemcok (1994).

Although the biostratigraphic resolution *within* the Lynton Formation is inadequate for the purpose of lateral correlation (Knight 1990a), several novel techniques are now available that may provide the required degree of resolution:

- ❑ The strontium isotope curve during the Devonian was particularly steep (Kovach 1980). Strontium isotope analysis of conodonts cf. Kovach (*op. cit.*) and Smalley *et al.* (1994), particularly from occurrences of the 'Watersmeet lithotype' throughout the Lynton Formation which are known to yield conodonts (Knight 1990a), may provide a finer scale of resolution than the biostratigraphic range of the conodonts themselves.
- ❑ The correlation of heavy mineral assemblages cf. Morton & Hurst (1995). This technique could be particularly useful in correlating the heavy mineral rich upper shoreface quartzose sandstones occurring around the Lynton Formation - Hangman Sandstone Group boundary, particularly in correlating the CU cycles between the Little Burland - Great Burland - Hollow Brook Combe sequences.
- ❑ Whole rock geochemistry cf. Racey *et al.* (1995).

Detailed petrographic and heavy mineral analyses of localised 'exotic' conglomerates around the Bristol Channel region have yielded a wealth of information relating to the tectonic evolution of the region (Tunbridge 1986). Analysis of the 'Watersmeet lithotype' to establish its provenance, and possibly the composition of the Lynmouth - East Lyn Fault foot-wall block, would provide a useful extension to the data base of these 'exotic' conglomerates. Additionally, in section 8.3 the provenance of the calcareous sandstone within the 'Lee Stone facies association' was linked to the catastrophic event that emplaced the calcareous 'Watersmeet lithotype' during the time that the basal part of the 'lower-middle mega-facies' was deposited. Petrographic, heavy mineral and whole rock geochemistry comparisons between the 'Watersmeet lithotype' deposited at this level and the 'Lee Stone facies association' may establish whether the common provenance hypothesis is geologically reasonable.

X-ray diffraction and cathodoluminescence analysis of the bone-bed found at east Lynmouth Beach would reveal the presence/proportions of apatite, francolite, pyrite etc. and thus facilitate a better understanding of the seawater geochemistry and oxygen content of the substrate at the time of deposition cf. the study of Humphreys and Smith (1989) on deposits of a similar age in south Devon.

Numeric tidal modelling has enabled the tidal régime of the Western Interior Seaway of north America to be better understood e.g. Slater (1986). As the Devonian palaeogeography of NW and central Europe becomes increasingly refined (e.g. Tsien 1989), modelling of this type should be applied to the 'Variscan geosyncline' which was periodically connected to the 'Proto Tethys' ocean. The results of such a study may reveal whether the interpretation of a tidal influence in the 'lower-middle mega-facies' is sound.

The question of whether the Lynmouth - East Lyn Fault soles out on a major Variscan plane of décollement, above which the Devonian-Carboniferous succession of north Devon has been shortened, was discussed in section 1.6.4.4. A detailed set of parallel seismic profiles along N-S transects covering the southern Bristol Channel and north Exmoor region would help to resolve the precise geometrical relationships between structures in this region. Brooks *et al.* (1994) suggested that a deep borehole would be needed to resolve the nature of the shallow (c. 2km) high-velocity layer that underlies north Exmoor in the sequence above the major plane of décollement. If such a borehole were ever to be sunk it would provide a useful seismic velocity control on the proposed set of N-S detailed seismic surveys.

Evidence preserved within the Lynton Formation supports the hypothesis that HCS forms under oscillatory flow with a relatively weak unidirectional current superimposed and high sediment fluxes. Workers seeking to establish the hydrodynamic conditions under which HCS forms may benefit from observing relatively low wave-energy shorelines lacking other unidirectional currents (oceanic, tidal etc.) where storm-generated suspended sediment fluxes are known to be high e.g. large lakes.

Finally, since the earliest days of research on the Devonian of north Devon attempts have been made to compare the sequence with equivalent outcrops in the Ardenne - Rhineland region e.g. Sharpe (1853). More recently, Goldring and Langenstrassen (1979) contrasted the styles of open shelf and near-shore clastic facies during the Devonian and recognised many similarities between north Devon and the Ardenne - Rhineland sequences, particularly the strong tectonic control on sedimentation patterns. It is hoped that the refinement of the environments of deposition of the Lynton Formation and their syn-sedimentary tectonic setting described herein will facilitate future workers in making more detailed comparisons with contemporaneous deposits in the Ardenne - Rhineland region.