SEDIMENTOLOGY AND HIGH-RESOLUTION SEQUENCE STRATIGRAPHY OF SHALLOW WATER DELTA SYSTEMS IN THE EARLY MARSDENIAN (NAMURIAN) PENNINE BASIN, NORTHERN ENGLAND.

Thesis submitted with the requirements of the University of Liverpool for the degree of Doctor in Philosophy by
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Abstract:

Sedimentology and high-resolution sequence stratigraphy of shallow water delta systems in the early Marsdenian (Namurian) Pennine Basin, Northern England.

The Pennine Basin formed during the early Carboniferous (Dinantian), and comprises a series of fault-defined sub-basins that were infilled by fluvial-deltaic systems throughout the Namurian. The basin-fill comprises turbidite-fronted deltas deposited in initially deep-water, which were overlain by deltas deposited in progressively shallow water throughout the middle and late Namurian. The Kinderscoutian-aged turbidite-fronted delta system prograded into the Pennine Basin, forming a palaeo-bathymetric high, over which the younger Marsdenian and Yeadoninan-aged mouthbar dominated deltas prograded. This shallowing trend continued into the Westphalian succession, where coal-rich delta top environments prevailed.

Throughout the Namurian of northern Europe, ammonoid-bearing marine bands form transgressive horizons that can be correlated across basins. The R2a1 *(Bilinguites gracilis)* and R2b5 *(Bilinguites metabilinguis)* marine bands mark the base and top of the early Marsdenian. Between R2a1 and R2b5 four marine bands (R2b1, R2b2, R2b3 and R2b4) separate a series deltas that contain basin-floor, turbidite, mouthbar, fluvial and bayhead depositional systems. The formation of new basin entry points and the use of pre-existing entry points allowed the Marsdenian deltas to prograde in a southerly direction over both the submerged Kinderscoutian delta and the underfilled Pennine Basin floor.

Palaeosols, transgressive and regressive key surfaces/ horizons are traced throughout the Marsdenian succession. The characteristics of these surfaces and horizons indicate strong alternations in water depth and depositional process occurred during deposition. These regressive-transgressive signatures can be correlated across the Pennine Basin, implying relative sea-level modulation occurred on a basinwide scale. Key surfaces and facies identified in field and borehole data are integrated with a depositional system model, allowing the analysis of the Marsdenian Stage by sequence stratigraphic techiques. This permits the refinement of the marine band correlation framework, allowing depositional environment distributions to be accurately predicted. Ammonoid-bearing marine bands represent basinwide surfaces that separate upward-coarsening successions. Within each upward-coarsening succession, key surfaces allow the definition of the upper and lower parasequence boundaries that define genetically related depositional units. Nineteen high-order sequences are identified between the R2a1 and R2b5 marine bands, and contain high-order forced-regressive/ lowstand and transgressive-highstand systems tracts suggesting that the polarity of sea-level fluctuated frequently during deposition. For example, the upper and lower bounding surface of high-order sequences is formed by a regressive surfaces. Low-order sequences comprise sets of stacked high-order sequences that lie between sequence boundary master surfaces. The presence of high- and low-order maximum flooding surfaces suggests sea-level modulation fluctuated on a at least two orders of magnitude, suggesting sea-level modulation curve was compound rather than simple.

Sequence boundaries are regionally erosive unconformities, representing a fall in relative sea-level, which are overlain by progradationally stacked depositional systems. Within the early Marsdenian, a series of high-order sequence boundaries coalesced to form a broader, low-order sequence boundary during low-order falling stage and lowstand systems tracts. The erosional sectors of the low-order sequence boundaries correlate to interfluve palaeosols on the incised valley margins, and bypassed lowstand turbidite deposits in southern parts of the Pennine Basin. Incised valleys formed during the R2b1 to R2b5 sequences are filled with stacked mouthbar, fluvial and bayhead dominated high-order sequences deposited during the low-order transgressive systems tract. Within the incised valley, high-order forced-regressive and lowstand systems tracts comprise progradationally stacked mouthbar and fluvial deposits, and are overlain by a high-order transgressive-highstand systems tract comprising retrogradationally stacked bayhead deltas. A regionally extensive marine band (the maximum flooding surface) caps incised valley fill, and defines the base of the highstand systems tract. On shelfal or platform areas, such as that formed by the submerged Kinderscout delta, accommodation is limited. The lack of accommodation implies that during relative sea-level fall, mouthbar systems often migrated rapidly in a basinward direction. Under these conditions mouthbar dominated delta systems generated characteristics suggesting deposition during the forced regressive

systems tract. Tidal and estuarine signatures are observed throughout the incised valley fill. These signatures are often cryptic due to the suppression of the tidal signature by the outflow-dominated nature of the delta system.

The broad high and low-order sequence architecture is apparent when palaeogeographic reconstructions are integrated with the sequence stratigraphic analysis. While the R2a1-R2b2 and R2b4 sequences are retrogradationally to weakly progradationally stacked, the R2b3 sequence represents a major drop in base level and an episode of marked basinward progradation.

Marsdenian depositional systems are similar to those of the Quaternary Gulf-Coast deltas, and comparisons are drawn between the magnitude and estimated periodicity of sea-level change during these periods. The observation that eustatic modulations in the Namurian have similar magnitudes and frequencies to the cycles of sea-level change in the Quaternary suggests that Milankovitch cyclicity may have been the controlling mechanism behind changes in Namurian relative sea-level.

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Figure Captions

Chapter 1

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Chapter 2

Figure 2.1 Photomosaic of Haigh Gutter (SE001121), illustrating the gradational transition from offshore mudstone basinal facies to offshore silty mudstone facies. These prograding pro-delta front sediments are overlain by erosive based, low-density turbidite facies, representing deposition from gravity driven flows, in the delta front area. Note the gradational colour change from the base towards the upper part of the section, due to the increase in coarser (siltstone) clastic sediment; and the disseminated thin turbidite beds in the offshore silty mudstone facies, representing deposition from waning laterally unconfined turbidite flows.

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Figure 2.12 Controls on mouthbar deposition. While supply system velocities are high, distributary channels have the capacity to entrain a greater amount of sediment. When this sediment enters the basin, it rapidly falls from suspension as it is deposited in the mouthbar (Coleman 1981). Wright (1977) suggested river mouth areas are influenced by two control parameters; i) the buoyancy of the effluent plume (which is dependant on the density contrast between plume and basinal waters) and, ii) the inertia associated with river discharge. To a certain extent these are mutually related, i.e. during times of high discharge, the fluvial inertia is greater, and the river water has a higher capacity to entrain and hold sediment in suspension. 2.12.1 When the effluent plume has a higher density than the surrounding basinal waters it will tend to sink on entry to the basin, producing a 'hyperpycnal flow' (Bates 1953; Wright 1977). This occurs when the fluvial discharge and sediment fluxes are high, i.e. the effluent has a high density due to the large bedload and clasts held in suspension by turbulence (Holdsworth & Collinson 1988; Collinson 1988; Orton & Reading 1993). Characteristically mouthbars deposited by bedload- friction processes are dominated by tractional transport mechanisms, comprising common cross bedded facies and shallow channelised features. 2.12.2 A decreased discharge in the fluvial supply system gives rise to a lowering in the amount of bedload and suspended sediment. This leads to the generation of a buoyant, or 'hypopycnal flow' (Bates 1953; Wright 1977). This situation need not occur wholly due to allocyclic modulations in fluvial discharge and sediment fluxes, but also in areas marginal to the axial region of the mouthbar plume. The facies generated by this type of mouthbar are dominated by suspension deposition through the basinal waters, with rare tractional or low-density turbidity currents transport sediment down the mouthbar.

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Figure 2.19 Image from Branshaw Quarry (SE032402). White line differentiates between facies above and below fair-weather wave base in the lower shoreface. Below the fair-weather wave base, the lower shoreface is characterized by interbedded fine-grained sandstone and silty-mudstone lithologies. In this facies, the upper surfaces of sandstone beds commonly have preserved symmetric wave-ripples (rule=1m). (see Figure 4.7)

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Figure 2.22 Tidally influenced cross-bedded sandstone
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Figure 2.23 Schematic representation of the fluvial (or tidally influenced) barform facies.

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Figure 2.26 Deformation structures at Bingley Road Quarry (SE053382). These are confined between the upper and lower bounding surfaces of individual planar giant cross bedding sets (barform set 1 & 2) of the fluvial barform facies. Individual forests are overturned in the direction of cross-bed dip and the base of the deformed zone gently dips sub-parallel, truncating the underlying giant cross-beds (upper part of barform 1). Some foresets have a flame like appearance, and are sub-vertical in orientation. Subsequent fluvial barform sets (barform set 2) migrate over the slumped feature and appear deformed, albeit to a lesser extent than the older set, implying that the deformation was syndepositional (photograph used with the permission of Paul Kabrna, Pendle & Craven Geological Society).

Figure 2.27 Laterally accreting channelized sandstone association in the upper part of Leicester Mills Quarry.
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Figure 2.28 Photographs of palaeosols (tape measure on figures 2.28.1 & 2.28.2 = 1m)
2.28.1 An extremely well preserved Lycophyte stump, with *Stigmaria* roots plunging at a shallow angle into the substrate. This type of root morphology probably represents the adaptation to growth in standing water, or poorly drained substrates when water was freely avaliable to plant roots (from Harper Delph Clough; SD716317).
2.28.2 Example of a Muddy palaeosol facies. This facies develops where the water-table was close to the substrate surface during palaeosol formation. These conditions were common on Marsdenian delta top environments, where shallow water bayhead deltas filled areas to emergence, allowing the development of pedogenic processes. 2.28.3 Graphic log from Harper Delph Clough (SD716317), illustrating stratigraphic position of pedogenic horizons. 2.28.4 Example of the sand-rich 'ganister' palaeosol facies from Fletcher Bank Quarry (SD805164). This facies comprises residual quartz and feldspars. 2.28.5 The sand-rich ganisteroid palaeosol is often highly indurated, and has either isolated voids after rootlets or carbonaceous streaks (from Warland Wood Quarry, SD946201), which are mainly sub-vertical in orientation. 2.28.6. Pedogenic 'ped' structures in waterlogged palaeosol beneath the *Bilinguites gracilis* (R2a1) marine band at Hard Head Clough (SE023129). Cubic 'crumb' like texture, and an abundance of hair-like carbonaceous rootlets suggests a fine mesh of rootlets developed during pedogenesis. 2.28.7 *Stigmaria* root, in muddy sandstone substrate with abundant hair-rootlets radiating perpendicular to main tap root. These roots are commonly sub-horizontal and close to the palaeosol top-surface. *Stigmaria* occur in both muddy and sand rich 'ganister' palaeosols.

Chapter 3

Figure 3.1 Part of section at Ponden Clough (SE981365). Interpretation of photomosaic, and log correlation, illustrating facies associations and key surfaces (dotted line= inferred position). This locality contains one of the few exposures of the Keighley Bluestone. Facies boundaries in the inter mouthbar association are gradational, inferring little influence by wave or current reworking. This association contains *Zoophycos*, *Hyalostelia* (sponge spicules) and *Lingula*, suggesting a brackish-marine to restricted marine environment (Waters et al. 1996). Transgressive horizons and marine indicators are common in the lower part of the section, while mouthbar facies increase in proportion up the section, suggesting proximity to a fluvial source. Facies boundaries become increasingly erosive up the section

Figure 3.2 Fourier analysis of heterolithic mouthbar facies lamina cyclicity
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Kvale (1997) 3.2.5 MEM Fourier analysis plot of data from figures 3.2.2 and 3.2.3 by Al Archer, University of Kansas, and corresponding ranges between this data set and that displayed in Figure 3.2.4.

Figure 3.3 Modulations in discharge generate variations in the amount of fluvial water entering the basin (Figure 3.3.1). In basins with marine waters a 'saline wedge' forms during periods of low-effluent velocity, as the density difference between plume and basinal waters allows underflow of the saline wedge (Nemec 1995). The 'buoyant plume' (*sensu.* Wright 1977) creates a strong density layering, which is elongated by the outward flowing effluent, and carries sand-sized clasts into the mouthbar area. Tidal modulations influence the inertial force of the plume, and the position of the saline wedge within the outlet channel (Figure 3.3.2). During the ebb tide, the inertia of the outflowing plume is enhanced, creating a mouthward shift in grainsize distribution. The grainsize distribution of a hypothetical tidally modulated low discharge plume (Figure 3.3.2) reveals a repeated thin-thick lamina set, deposited during the ebb and flood cycles and intercalated silty-mudstone lamina.

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Figure 3.6 Bayhead pro-delta, delta front association and bayhead delta top association at Fletcher Bank Quarry (SD805164). 3.6.1 Photomosaic of upper part of Fletcher Bank Quarry.
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Chapter 4

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The basin floor mudstone association is overlain by an amalgamated turbidite association. Candidate correlations to marine bands are illustrated, although no ammonoid bearing marine fauna are identified in the quarry section. Within the turbidite association, channel bodies are stacked and possess overbank low-density facies.

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PART 1: INTRODUCTION

Objectives of this study

The aims of this project are broadly two-fold; to improve the correlation framework within the Marsdenian deltaic interval of the Pennine Basin, and to provide interpretations on the variability and sedimentological characteristics of the delta systems. Within these general objectives, specific aims include;

1. The description of facies and facies associations, which provide solid building blocks for the analysis of variations in depositional environment and the interpretation of processes that controlled and/or influenced deposition. (see Facies and Facies Association Chapter).

2. The description of surfaces and horizons that can be used to correlate strata within the basin, the interpretation of the mechanisms behind their formation, and their limitations and accuracy as correlatable markers (see Key Surface Chapter)

3. To record the basin-wide distribution of facies associations and stratigraphical markers, thereby allowing the generation of palaeogeographical reconstructions that permit the realisation of depositional environment extent (see Palaeogeography Chapter).

4. To integrate the facies, facies associations and stratigraphical markers and produce a model that predicts their stratigraphic relationships and distribution across the basin, and to incorporate biostratigraphical and sequence stratigraphic techniques to produce an accurate correlation for the Marsdenian of the Pennine Basin (see Sequence Stratigraphy Chapter).

5. To delineate the sequence and timings of basin fill episodes, and to highlight the sedimentation transportation pathways which existed during the Marsdenian; particularly with respect to the deposition of delta systems through the individual sub-basins of the Pennines (see Palaeogeographic Chapter).

6. To ascertain the mechanisms controlling the distribution of facies associations and stratigraphical markers in Marsdenian delta systems, specifically with the aim to elucidate the control of tectonic processes and eustatic relative sea-level fluctuation (see Controls on Deposition Chapter).

Approach

This project integrates BGS map-data, borehole datasets and cored intervals held by the British Geological Survey with over one-hundred field localities. Field data was collected during the summer months of 1998 and 1999, and includes logged sections and interpreted photomosaics. Data from core and archived sources was collected throughout the duration of the project, and collated with the field data to provide an integrated, laterally extensive dataset which covers a significant part of the Pennine Basin and adjacent areas. The memoirs of the Geological Survey provided details for the majority of the field localities identified during this project (Green *et al.,* 1878; Wright *et al.,* 1927; Wray *et al.,* 1930; Bromehead *et al.,* 1933; Edwards *et al.,* 1950; Stephens *et al.,* 1953; Eden *et al.,* 1957; Earp *et al.,* 1961; Price *et al.,* 1963; Smith *et al.,* 1967; Aitkenhead *et al.,* 1985; Arthurton *et al.,* 1988; Cooper & Burgess, 1993; Cameron *et al.,* 1993; Brandon *et al.,* 1998; Waters, 1999).

Perhaps the most important approach used during this project was the integration with the current BGS mapping programme in northern England (Waters, 1999). This has proven beneficial to both the generation of a high-resolution correlation framework/ environmental interpretations and provided data of significance to the mapping of the lithostratigraphic units within Marsdenian of the Pennine Basin.

This thesis is laid out in four parts, which contain eight chapters. Part 1 comprises an introduction to the Namurian of the Pennine Basin. Part 2 (Chapters 2 to 5) comprises the basic 'building blocks' upon which the basin analysis is built, and comprises a rigid hierarchy of descriptions and interpretations. Part 3 (Chapters 6-8) comprise the basin analysis, with the majority of chapter six (in Section 6.2) written in paper format. Part 4 comprises a synthesis of all the previous interpretations.

The following sections begins with an historical overview of research carried out on the Pennine Basin, description of the structural elements that control the basin boundaries, and then focus on the lithostratigraphic nomenclature used in the Namurian Pennine Basin

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Chapter 1: The Pennine Basin & the Namurian: Background details

1.1. The tectonic origins of the Pennine Basin

The Pennine Basin lies between the Appalachian and Canadian Maritime Basins to the west and the North Sea and European Basins to the east (Figure 1.1.1 & 1.1.2) (Ziegler, 1982; Besly, 1988; Blakey, 2001). The margins of the Pennine Basin are defined by structural elements that developed during the Caledonian Orogeny and influenced the development of the basins of northern England throughout the late Palaeozoic. The Caledonian Orogeny occurred between the late Ordovician-early Devonian (approximately 470 -390 Ma) (Tanner & Bluck, 1999; Soper *et al.,* 1999). During this period the Midlands Microcraton progressed northwards, and indented the Laurasian continental plate (Fraser & Gawthorpe 1990; Figure 1.1.3). The collision generated a set of NE-SW faults (the Church Stretton, Bala, Pendle and Quernmore Faults) and NW-SE faults (the Hoton and Morley Campsell Fault systems). Subduction of the Rheic Ocean, in the Rheno-Hercynian back-arc generated an extensional province 500 km to the north, in the area now knowm as northern England (Leeder & McMahon 1988; Figure 1.2.1). Driven by the plate margin processes during the Hercynian Orogeny, rifting along these fault trends generated a series of half-grabens that developed into the Northumberland-Solway, Stainmore, Bowland, Cleveland, Edale, Gainsborough and Widmerpool sub-basins (Johnson, 1967; Gawthorpe, 1987; Fraser & Gawthorpe, 1990). The Craven Fault system delineates the northern margin of the Pennine Basin and separates the Askrigg structural high in the north (buoyed-up by granite) and the southern syn-rift basins. Rifting during the Dinantian extensional tectonic phase created a basinwide marine transgression, which generated significant water-depth, allowing platform and ramp carbonate systems to develop on syn-rift highs. The underfilled basins present at the end of the Dinantian produced depocentres that became the focus of delta deposition during the Namurian (Gawthorpe, 1987; Lee, 1988; Fraser & Gawthorpe, 1990). The Namurian marks the onset of the post-rift stage of basin infill, when passive thermal subsidence exceeded rifting as the main mechanism of basinwide accommodation development (Leeder, 1982; Fraser & Gawthorpe, 1990). This resulted in the decreased definition of the 'sub-basins' formed during the Dinantian, and the generation of a broad basin which has been termed the 'Central Province' (Ramsbottom, 1969; Collinson, 1988). During the Namurian the margins of the Central Province, or Pennine Basin as it is termed in this study, became less defined as it was infilled by clastic deposits. This is confirmed by the presence of a broader bullseye isopach pattern of Namurian and Westphalian deposits (Fraser & Gawthorpe, 1990).

Evidence for tectonic activity on basin bounding faults is scarce. However, possible seismically initiated dewatering structures are observed in the Marsdenian (i.e. Bingley Road Quarry, SE053382; see Figure 2.26) and syn-depositional fault movements are inferred on the Aire Valley Fault system during the Marsdenian (Waters *et al.,* 1996a) and the Morley-Campsall Fault during the Westphalian (Giles, 1989). The integration of fault trends identified by British Geological Survey and gravity-derived basin configuration models with tectonic-influenced facies identified during this study, enables the influence of basin boundary faults during the Namurian to be assessed (Lee 1988; Figure 1.2.2).

The cause for the initiation of such a dramatic influx of sediment is enigmatic, but probably involved the capture of major drainage basins in the Caledonide Mountain Belt or a major 'avulsion' in the fluvial feeder systems (Sims, 1988). The northward drift of Pangaea into the equatorial regions probably generated high denudation rates, due to the increased severity of mechanical and chemical weathering processes. It is also possible that climate change forced by the modification of continent configuration during the Caledonian Orogeny led to increased denudation in the hinterland (Rowley *et al.,* 1985).

1.2. History of sedimentological research in the Carboniferous of the Pennine and adjacent Basins

The Carboniferous deposits of the United Kingdom have provided economic deposits of coal, limestone, aggregates, building-stone, brick-clay, fireclay and a reliable water supply since the industrial revolution. Of all the counties within the United Kingdom, those of Lancashire, Yorkshire and northern Derbyshire have provided the largest amount of these resources (Figure 1.3.1). In this region, the area between the Peak District and the Yorkshire Dales comprises the most extensive outcrop of Upper Carboniferous aged deposits. This region has become a classic field area, where the concepts of sedimentology and stratigraphy have been taught and tested.

The term 'Millstone Grit' was initially used to describe the strata between the Carboniferous limestone (broadly Dinantian) and the Coal Measures (Westphalian) across northern England (Whitehurst, 1778). The interval that forms the focus of this study, was previously termed the 'Middle Grits', due to its stratigraphic position between the major gritstone intervals of the Kinderscoutian and Yeadonian (Stephens *et al.* 1953; see Figure 1.3.2 & 1.4.1). The dominantly deltaic succession of the Namurian series has been reclassified by the BGS as the Millstone Grit Group, and the Marsdenian Stage is the equivalent of the Middle Grits (Ramsbottom, 1981).

Data collection during the early 1900's provided a new insight into the variation in the rock types and the fossil faunas. Fastidious data collection by Geological Survey field geologists and individuals such as W.S. Bisat (Bisat, 1924; Bisat, 1928) initiated the detailed analysis of the Namurian deposits of the UK. The work of Bisat, the palaeontologist who first used ammonoid bearing marine bands to correlate stratal units across the Pennine Basin, was to prove crucial in the mapping of northern England. During the 1920's and 1930's the completion of the Rossendale, Huddersfield and Glossop map-sheets relied on the marine band framework set up by Bisat (Wright *et al.,* 1927). Investigation of these memoirs reveals that the geologists involved in their compilation had a sound understanding of the variability and processes that controlled deposition during the Namurian. Additionally, but slightly later, the investigation of Pennsylvanian sections in the Appalachian Basin inferred deposition was influenced by periods of cyclical sea-level rise and fall related to the waxing and waning of polar glaciers (Wanless & Shepard, 1936).

In the past 50 years, the discovery of oil and gas deposits, first in the onshore East Midlands Basin (Falcon & Kent, 1960), and later in the Ems Graben, western Germany (Dunay & Flint, 1999) led to the exploration and discovery of other Carboniferous reservoir intervals (Maynard *et al.,* 1997). It was the quest for better correlations between sand-rich delta cycles that led to the refinement of the marine band correlation framework and the identification of cyclicity on a basin fill scale (Ramsbottom *et al.,* 1962; Ramsbottom, 1977; Ramsbottom *et al.,* 1978).

The correlation of Namurian marine bands allowed the identification of 'Mesothems', which comprise major coarsening upward regressive cycles (Ramsbottom, 1977). The gritstone units of the lower Marsdenian (R2a1-R2b5) are positioned within Ramsbottom's 'N9' and 'N10' mesothems. Within these mesothems, minor coarsening-upwards sandstone cycles, or 'cyclothems', are identified. Within individual cyclothems, interbedded mudstone to siltstone parts are mapped separately from the sandstones, which often form topographic features on outcrop. Each sandstone is mapped as lithostratigraphical units and named after a type-section exposure. Although sandstones may have different names on adjacent map-sheets, the objectivity of the BGS mapping methodology, based on the identification of marine-band faunas, implies that time-equivalent lithostratigraphical units they may be compared across the basin if the marine bands or equivalent intervals above and below the unit are identified. While this correlation technique is proven in the generation of onshore geological maps, its use in the offshore has led to mixed results. Many sand-rich prospects prove uneconomic, due to a dependency on very subtle stratigraphical or structural traps.

The early to mid-1980's was a fairly quiet period in the field of Namurian research. Renewed interest in the science of stratigraphy was sparked with the publication of genetic (Galloway, 1989) and sequence stratigraphic concepts (Posamentier *et al.,* 1988; Posamentier & Vail, 1988; Van Wagoner *et al.,* 1988; Jervey, 1988). Sequence stratigraphic appraisal both onshore and offshore, along with the comparison between reservoir intervals and outcrop analogues has led a renewed interest in the analysis of the Pennine and adjacent Basins (Read, 1991; Collinson *et al.,* 1992; Hampson *et al.,* 1999; Davies *et al.,* 1999; Jones *et al.,* 2000). This is in part a response to the technological advances in the discovery and management of smaller petroleum reservoirs in the now mature Southern North Sea (Cameron *et al.,* 1993), and the drive to identify and model analogues for other deltaic reservoirs.

1.3. Lithostratigraphic nomenclature within the Pennine Basin

Within the Pennine Basin, the base of the Millstone Grit Group is highly diachronous (from the Late Pendleian E1c in the north to the Marsdenian R2b in the south of the basin). The Marsdenian is dominated by interbedded mudstones and siltstones with coarse-grained feldspathic sandstones and subordinate coals and seat-earths (Figure 1.4.2). Thick-shelled ammonoids (goniatites) occur in distinct, widely distributed, dark grey or black mudstone horizons (marine bands). The ammonoid faunas are often diagnostic of specific marine bands that can be correlated through the Pennine Basin and other Namurian basins within the British Isles (Bisat 1924; Ramsbottom *et al.* 1978, see Figure 1.3.2).

Marine bands also provide an excellent basis for the correlation of the interbedded sandstones when used in conjunction with sequence stratigraphic (Posamentier *et al.,* 1988; Posamentier & Vail, 1988; Van Wagoner *et al.,* 1988), and spectral gamma ray/ palynological techniques (Davies & McLean, 1996). Standardisation of Namurian marine-band nomenclature is based on that suggested by Ramsbottom *et al.* (1978), with the exception of the *Reticuloceras* which has been replaced by *Bilinguites* in the Marsdenian, and *Cancelloceras,* which has replaced *Gastrioceras*. The marine band index system used in this project (Figure 1.3.2) was introduced by (Ashton, 1974), and modified by (Holdsworth & Collinson, 1988; Riley *et al.,* 1993).

In the case of the Marsdenian Stage the lithostratigraphical nomenclature can be bewildering (Figure 1.5). Between *Bilinguites gracilis* (R2a1) and *Bilinguites metabilinguis* (R2b5) four to five major sandstone intervals can be differentiated across the Pennine Basin, depending on the number and correlation of the marine bands. In stratigraphical order these comprise, i) Alum Crag Grit, ii) Readycon Dean Flags, Scotland Flags and East Carlton Grit, iii) Woodhouse Flags, iv) Brandon Grit, Fletcher Bank Grit, Gorpley Grit, Heydon Rock, Midgley Grit, Pule Hill Grit, Rivelin Grit, Revidge Grit and Woodhouse Grit, and v) Helmshore Grit. Additionally, the Keighley Bluestone, a siliceous blue-grey siltstone with marine fauna, forms a lithostratigraphic unit in the area south-west of Keighley.

Lithostratigraphical mapping in areas of the Pennines relies on 'feature-mapping'; namely, the tracing in topographic, geomorphologic or vegetation changes. This methodology creates object-based maps, which allow the detailed elements of the stratigraphy to be taken into account. This study utilises both maps from recent and archived surveys to fit in the correlation of minor transgressive, regressive or emergent events within the current lithostratigraphic nomenclature. It is the identification, classification and correlation of surfaces and horizons within the broader marine band framework that provides the increased accuracy of correlation within the Marsdenian.

PART 2:
DEPOSITIONAL ELEMENTS OF THE MARSDENIAN PENNINE BASIN

Chapter 2: Facies Descriptions and Interpretation

2.1. Introduction and Methodology

Sedimentary facies are defined on the basis of physical and sedimentological characteristics; including grainsize, colour, composition, bedding, sedimentary structures, fossils and ichnofabrics. Such characteristics are produced by a number of parameters. These included the composition of the sediment, the processes involved in deposition and reworking, and the extent and type of organisms that interacted with the deposits. Most processes are non-unique, and occur in several environments of deposition. It is therefore the context of the facies that allows environmental interpretations to be made (see Section 6.2). Many facies schemes exist for Namurian (Walker, 1966a; Collinson *et al.,* 1977; McCabe, 1977) and Westphalian deposits (Guion & Fielding, 1988; Guion *et al.,* 1995; Jones *et al.,* 1995). Some facies in this study are similar to those described in previous Carboniferous research, while some new facies are described (e.g. those interpreted as tidally-influenced).

In order to be objective, each facies is dealt with individually. The facies are first described, and then interpreted in terms of processes and environment of deposition. Throughout the remaining text, the interpretative descriptions (i.e. offshore marine mudstone or interdistributary bay) are used when describing facies.

2.2. Facies identified in this study

2.2.1. Black mudstone with marine fauna

This facies represents approximately 1-3% of the sucession within the Marsdenian of the northern Pennine Basin, although it is highly distinctive. It comprises massive or fissile black to very dark grey mudstone, devoid of silt or coarser clastic material. Laminations are generally sub-millimetre in thickness, continuous and often fissile. Thick- and thin-shelled ammonoid faunas include *Anthracoceras*, *Homoceras*, *Bilinguites gracilis* and the various forms of *Bilinguites sp*. are commonly preserved as flattened impressions, along with bivalves *Dunbarella*, *Caneyella* and *Posidonia* (see Figure 4.1). In some examples, the pectenoid *Dunbarella* is observed, almost exclusively at the base of the facies in a high-density, low-diversity population. *Dunbarella*-rich horizons often contain abundant, well-preserved drifted plant debris. Elliptical to spherical concretionary masses of calcium carbonate containing a high-density bivalve and ammonoid shelly debris occur. These nodules commonly weather to a brown-to-buff unconsolidated mudstone. The mudstone surrounding these nodules is often indurated and commonly contains well preserved three-dimensional ammonoid debris and other shelly faunas. Isolated concretionary masses commonly deform surrounding laminations. Rare impressions of leaf and stem plant debris are observed throughout this facies.

Interpretation: Offshore marine mudstone

The muddy, parallel laminae suggests deposition from suspension in a standing body of water away from coarse-grained clastic input points. The high faunal diversity implies deposition during a period when marine conditions prevailed, while the high-faunal density suggests low sedimentation rates and the concentration of marine faunas. This facies always overlies sediments deposited within shallower delta-top/ front environments, and by inference, represents a regional transgressive event. The pervasive black colour of the mudstone implies that anoxic conditions prevailed in the substrate-water contact or within the sediment. The degree of anoxia may have been influenced by the amount of organic debris reworked during flooding of the underlying delta top. Mobilization of organic debris during transgressive reworking may have increased water column turbidity, while bacterial breakdown may have utilised free oxygen, therefore creating anaerobic conditions. The presence of epifaunal *Dunbarella* at the bottom of this facies suggests quiescent conditions following the initial flooding surface. The high density and low diversity of this fauna implies a restricted environment, where either a soft substrate or low oxygen conditions controlled colonization (Wignall, 1993a).

2.2.2. Black mudstone with no marine fauna

This facies is slightly more common than the offshore marine mudstone facies, and comprises black mudstone with no coarser grained clastic material, and no ammonoid or bivalve faunas.

Interpretation: Offshore black mudstone.

This clay-dominated facies appears to have been deposited by suspension from a standing body of water. During the deposition of this facies, the absence of ammonoid or bivalve faunas may imply that conditions were not conducive to preservation of shelly material. Alternatively, environmental conditions such as the presence of anoxic or nutrient-free water may not have allowed colonization by suspension-feeding marine faunas on the substrate or in the overlying water column.

2.2.3. *Parallel-laminated grey silty-mudstone with isolated sharp based sandstones*

This facies comprises grey to dark grey mudstone with rare centimetre-thick siltstone to fine-grained sandstone beds. It is fairly common throughout the Marsdenian Pennine Basin. A parallel lamination commonly defined by mica dominates the mud-rich parts of this facies. Some of the mud-rich intervals are massive with disseminated silt-sized carbonaceous debris and localized bioturbation. Body fossils are absent from this facies. Thinly laminated, light and dark grey mudstone are observed. The lighter laminae are commonly less than 2mm in thickness, and possess uneven basal surfaces with faint relief. Isolated laminae of light grey mudstone are also distributed within the darker grey mudstone. Rounded to elliptical pyritic concretionary structures up to 4mm in diameter are draped by the surrounding lighter grey mudstone.

Rare, very fine- to fine-grained sands, with a maximum bed thickness of 0.75 m are observed. These sandstones are either erosively based with sharp top surfaces, or have both a gradational basal and top surface (Figure 2.1). Rare broad D-shaped flute casts up to 1m in width and 0.05 m in depth, are associated with centimetre-scale flute casts and tool marks. Small-scale structures are superimposed onto those large flutes. The isolated, relatively thin character of these beds implies that the scouring at the base is shallow. Massive sandstone normally overlies the erosive-base, but in some cases, rare 0.1 m thick sets of cross-bedding and mud rip-up clasts overlie the erosive base. The majority of the middle of such beds are massive, while current ripple laminations are preserved in the upper part of beds.

Interpretation: Offshore silty-mudstone

These mudstones were deposited from suspension, with the rare sandstone and micaeous laminae representing periods of increased sediment input into the basin. Isolated bioturbated laminations may imply changes in nutrient and oxidization levels at the sediment-water interface. The paucity of body fossils and the grey colour may imply anoxic deposition, or non-marine conditions. The structureless, thin beds of fine-grained sandstone were deposited from suspension, following the initial passage of a turbulent gravity flow. The two scales of scouring may imply variation in flow velocity and density during deposition. Similar, if slightly larger, ‘mega-flutes’ are described by (Elliott, 2000), which form widespread erosion surfaces flanking shallow turbidite channels, and are attributed to differential erosion by turbidity currents.

The pyritic structures are interpreted as peloidal deposits of an infaunal organism. Assuming that conditions at or beneath the substrate surface were anoxic, it is possible that organic material was reduced to pyrite. Changes in anoxia at regional or local scales during deposition, or oxidisation of the sediments by bioturbation may account for the variablility in the colour of the mudstone.

2.2.4. Interbedded thin-bedded sandstone and silty mudstone.

This facies comprises interbedded fine- to medium-grained sandstone comprising dominantly of quartz grains and dark grey mica-rich silty mudstone. It is locally common in the southern parts of the study area, but absent from the northern area. The sandstone beds range from 0.02 to 0.2m in thickness, while the silty mudstone is commonly less than 0.1m in thickness (Figure 2.2). The base of the sandstone beds comprise planar discontinuous sub-horizontal erosive surfaces which are overlain by low angled cross-bedding. Widely spaced delicate flute- and tool-marks (flutes up to 0.04m across, 0.01m deep) are common on the base of beds, and are not as strongly developed as those in the massive thick-bedded turbidite sandstone facies (Figure 2.3). Fining upward trends are observed in some beds, and the transition from sandstone to silty-mudstone interbeds appears gradational over 0.05-0.1m. The base of the lower sandstone rich bed is seen to rest conformably on the silty-mudstone, but in some cases, erodes through to the underlying sandstone rich bed, therefore creating an amalgamated bed. Asymmetric ripple laminations are seen within the upper horizons of the sandstone beds, and flaser laminae are ubiquitous within the silty-mudstone lithology. In comparison with the massive thick-bedded turbidite sandstone facies, interstitial clay is more abundant within the matrix. The base of this facies is planar and parallel to bedding with in underlying facies, and no channel-like features are observed.

Interpretation: Low-density turbidite sandstone.

The juxtaposition of interbedded sharp-based sandstone and silty-mudstone suggests deposition from mixed tractional and suspension processes. The occurrence of flute- and tool-marks, the sharp and laterally continuous erosive bases of beds, along with the broad upward fining nature of the sandstones to the silty-mudstone facies, suggests that this facies was deposited from turbidity currents. In comparison with other classification schemes (Lowe 1982), the sandstone beds represents the Ta-Tb division, with localized thin development of Tc. The gradation between sandstone beds and the silty-mudstone partings (equivalent to Td-Te) implies that the transition between waning turbidite deposition and suspension deposition was gradual. The sheet geometry of the beds within this facies suggests deposition by an unconfined flow. This may have formed during the late stage of turbidity current deposition, in an turbidite overbank environment, or in the outer parts of a turbidite lobe.

2.2.5. *Thick-bedded sandstone, with mudstone conglomerate*

Fine- to coarse-grained sandstones dominate this facies. This facies is common in the southern and eastern Pennine Basin, where it makes up to 10-15% of the succession. Grains are commonly sub-angular to sub-rounded, implying a low textural maturity. At the base of individual beds granule sized clasts are commonly observed, and the grain size within beds appears to fine upwards. Little or no clay is present within the matrix. Beds generally appear massive, are between 0.2 and 0.75 m thick, and possess few obvious primary depositional sedimentary structures. However, faint sub-horizontal laminae, small mud clasts and asymmetric ripple laminations are observed towards the top of individual beds. The base of the beds are either planar (Figure 2.2, 2.3 and 2.4), or may have up to several metres of relief (Figure 2.5). Viewed on a broad-scale, basal erosive surfaces appear channelized, and cut underlying beds at low angles (Figure 2.5 & Figure 4.12). Flute and tool marks, with flutes up to 0.04 m across, 0.02 m deep) are ubiquitous on the base of beds, and are often locally concentrated in sets (Figure 2.3). In laterally continuous exposures, broad (>10m wide, ~5 m deep) concave-upwards channel features are seen. Large rafts of very dark grey mudstone (>30 cm width, >5 cm thick) are commonly observed near the base of channels. These rafts are commonly supported by the coarser sandstone matrix, and occur in concentrated horizons (Figure 2.6). The mudstone rafts are frequently rectangular in cross-section, contain internal parallel laminations and have rounded to sub-rounded corners.

Interpretation: High-density turbidite sandstone

The presence of the sharp bases, along with the fining-upwards grainsize trend of individual beds and the occurrence of sole-marks, suggests this facies was deposited by waning turbidite currents. The lack of mudstone partings between beds implies that they were either deposited in rapid succession, or that scouring associated with the passage of the turbidite flow may have eroded the underlying sediments. It seems likely that some of the thicker beds represent amalgamated multi-phase turbidites deposited during the waning stage of flow, with the base of individual events inferred by the presence of faint internal laminae and small mud-clasts. Similar facies are described in the Kinderscoutian and Marsdenian Stage of the Pennine Basin (Table 1).

	Comparable turbidite facies from the Pennine Basin		
	(Walker, 1966a; Walker, 1966b)	(Collinson, 1968a)	(Collinson *et al.,* 1977)
High Density turbidite	Facies 'B' and 'C'	Facies 7 – Massive bedded coarse sandstone	Medium bedded turbidite sandstone Thick-bedded turbidite sandstone Deformed conglomerate
Low Density turbidite	Facies 'A'		Thin bedded turbidites

Table 1 Comparisons between turbidite facies described in this chapter, and other studies of the Kinderscoutian and Marsdenian Stages.

2.2.6. Parallel-laminated siltstone/ mudstone with rare thin interbedded sandstone.

This facies comprises grey micaceous siltstone in a matrix of black to dark grey mudstone, and is common throughout the Marsdenian Pennine Basin, especially in the Huddersfield sub-basin where it may make up to 25-30% of the succession. Occasional isolated beds of very fine-grained sandstone, with both sharp and gradational bases are observed. Laminations are either laterally continuous or discontinuous, and are defined by alternations in grainsize and colour (Figure 2.7). Some outcrops show a gradual increase in the proportion of silt upwards, with laminations thickening and becoming increasingly continuous. In this lithology, faint cross laminations and rare symmetrical ripple laminations are observed. Sharp based, massive fine-grained sandstone (>0.01 m thick) with delicate tool marks, and silty sandstones (>0.04 m thick) with gradational tops occur within the silty parts of this facies, in association with silty flaser laminations. Horizons of *Planolites* bioturbation are laterally continuous across exposures for several metres (Figure 2.8), and rare *Lingula* are observed in some sections (see Section 4.2.2). Occasional indurated pale calcareous silty mudstone horizons are observed, which are laterally continuous, massive, and fracture with a concoidal cleavage.

Sub-vertical, concave-up listric surfaces occur in the upper, thicker laminated siltstones of some units. Laminations on either side of the surface appear to be offset by up to 10cm. Localised areas of increased iron-oxidisation are associated with the listric surfaces and within the surrounding sediments. Zones of loading appear below the silty horizons, and appear to form confined horizons, in which small (up to 10cm width, 3cm relief) load casts are common, along with concentric cuboidal or spherical structures. These form indurated features, and are confined to traceable sets of laminae.

Interpretation: Interdistributary bay

The dominant layer parallel bedding in this facies suggests that it was deposited predominantly from suspension, while the presence of both continuous and discontinuous laminae suggest that rates of deposition or reworking may have varied temporally. This is also evident from the rapid spatial and temporal variations in the proportion of mud to silt, and the deposition of sandy laminations. The alternate presence and absence of *Planolites* bioturbation may infer variations in oxygen and nutrient availability within the substrate. The occurrence of current- and wave-ripple laminations in silt-rich portions implies that the shallow margins of interdistributary bays were influenced by wind driven waves and current transport processes that reworked the substrate surface. *Planolites* occurs in a range of marine and non-marine environments throughout the Namurian (Eager *et al.,* 1985), and cannot therefore be used to infer open-bay conditions, while the presence of *Lingula* suggests slightly brackish marine basinal waters. The calcareous indurated horizons may represent zones of increased diagenetic cementation. The bedding-parallel orientation suggests a syn-depositional origin that may correlate laterally to flooding events.

Loading and dewatering structures are common within this facies. Compaction due to the deposition of overlying bay-fill sandy sediments deposited by bay-fill mouthbars, crevasse splays and fluvial sandstone, may account for these loading features. The presence of listric shear surfaces suggests small-scale slide displacement along discrete syn-depositional shear planes.

2.2.7. Carbonaceous silty mudstone, with interlaminated siltstone and fine- to medium-grained sandstone.

This facies comprises buff micaceous silty mudstone and dark grey mudstone, with thick isolated interbeds of siltstone to medium-grained sandstone, forming a heterolithic lithology that is banded on a millimetre to centimetre scale. It is locally common, but is generally rare, making up to 5-10% of the early Marsdenian succession. The black to dark grey mudstone forms the matrix throughout this facies. Asymmetric and rare symmetric ripple laminations and subtle flaser to lenticular laminae characterise the silt-rich components, while the sandstone laminae often contain a single set of current ripple laminations (Figure 2.9). Sub-spherical balls of siltstone, up to 5mm in diameter are present. These commonly overlie scour surfaces, which are draped by the black mudstone lithology. Carbonaceous debris is locally abundant, and in some cases, well preserved *Cordaities sp.* and *Neuropteris* plant debris is commonly associated with rafted coal lenses and logs. The sand-rich beds within this facies usually contain asymmetric and symmetric ripple laminations, which are commonly preserved as three-dimensional forms on the top of the beds. Palaeosols (see Key Surface Chapter) are associated with thin laterally non-persistent coal streaks. Ichnofaunas include isolated *Pelecypodichnus,* bivalve escape structures and rare *Cochlichnus* within the muddier components (Figure 2.10).

Interpretation: Shallow-water open marine bay

The heterolithic character and delicate sedimentary structures suggest this facies was deposited during the complex interaction of low velocity tractional currents and suspension deposition. Suspension fall-out was the dominant process of deposition, but current and wave reworking affected the substrate surface. The presence of well-defined laminations and current deposited sandstone laminae suggest tractional processes, resulting from reworking of the delta top substrate. The siltstone balls do not appear biogenic in origin, and are suggested as the product of winnowing currents during the deposition of this facies.

The occurrence of palaeosols and presence of wave ripples infers deposition in localised shallow to emergent conditions and the abundance of well-preserved plant debris suggests a local source of plant material. A high organic content and the lack of benthic fauna implies a poorly oxygenated, stagnant water column.

2.2.8. Zoophycos bioturbated silty mudstone.

This facies comprises black to very dark blue-grey siltstone interlaminated with thin discontinuous, very fine- to fine-grained sandstone laminae, with abundant carbonaceous debris forming diffuse laminations. It occurs in limited parts of the basin, and probably accounts for less than 0.5% of the early Marsdenian succession. Bedding and laminations are generally undulose, and weathered bedding surfaces are highly indurated, and contain abundant *Zoophycos* trace fossils (see Figures 4.5.5 and 4.5.8) throughout the facies (Eager *et al.,* 1985). These consists of positive epirelief feeding tubes/ lamella (up to 0.2 m wide) forming an overlapping network of concentric arcuate spreite. Each spreiten is approximately 3-4 millimetre in width, and forms a positive epirelief ridge. Lamella converge at the central apex, forming an area of positive relief approximately 10mm in width. Some outcrops have white acicular forms, which have been interpreted as *Hyalostelia* sponge spicules (Stephens *et al.,* 1953). Rare shell fragments are also observed, but are too fragmentary to allow identification.

This facies is more indurated in comparison with subjacent facies, and generally forms areas of positive topography.

Interpretation: Shallow-water restricted marine bay

The restricted grainsize, and lack of wave or current indicators suggests that this siltstone was deposited from suspension, while the presence of *Zoophycos* and *Hyalostelia* imply a shallow, restricted marine conditions (Cavaroc & Ferm, 1968). It is therefore possible this facies developed in restricted marine conditions, that may have been laterally adjacent to prograding delta complexes (Stephens *et al.,* 1953; Earp *et al.,* 1961; Wignall & Maynard, 1996).

2.2.9. Parallel-bedded sandstone

This facies comprises well-sorted fine- to medium-grained sandstone, with inter-laminations of mudstone and siltstone. It is fairly common and makes up to 20-25% of the Marsdenian succession in the Huddersfield sub-basin, but less in the southern and eastern Pennine Basin. Laminations and bedding are sub-parallel and laterally persistent across exposures. The beds contain faint sub-parallel internal primary depositional structures, or rare cross-bedding and cross-lamination. Bedding thickness is generally between 0.05-0.1m and beds have a variety of depositional dips, varying from sub-horizontal to ~20° (Figure 2.11.1). Rare slumping occurs within isolated beds, and flute and tool marks are seen on bedding planes. Minor scours, cutting individual bedding planes are observed, but are rare. Trace fossils in this facies include abundant *Olivellites, Thalassinoides* and rarer unlined tubes, cf. *Skolithos* (see Figure 4.5.10).

Interpretation: Bedload-dominated sand-rich mouthbar.

The sandstone in this facies was deposited predominantly from suspension, with a minor component of tractional transport. This facies is inferred as a sand-rich mouthbar environment, proximal to the fluvial point source, where sediment flux rates were high and deposition rapid. Other researchers suggest *Olivellites* is the trace of an epifaunal arthropod, inferring that although the water may have had a high turbidity, fauna could live on the substrate and nutrients were available (Yochelson & Schindel, 1978; Eager *et al.,* 1985).

The relatively high flow velocity of the fluvial input maintains a high density of suspended sediment. This sediment is rapidly deposited from suspension in the mouthbar (Coleman, 1981). River mouth areas are influenced by three control parameters; i) the buoyancy of the effluent plume (which is dependant on the density contrast between plume and basinal waters), ii) the inertia associated with river discharge, and iii) friction with the substrate (Wright, 1977). These parameters are mutually related to an extent, i.e. during times of high discharge, the fluvial inertia is greater, and the river water has a higher capacity to entrain and hold sediment in suspension.

The effluent plume will tend to sink on entry to the basin, producing a 'hyperpycnal flow' (Figure 2.12) if it has a higher density than the surrounding basinal waters (Bates, 1953; Wright, 1977). This situation would have been most typical of the Carboniferous Pennine Basin where the sediment flux and fluvial discharge was high, i.e. the effluent has a high density due to the large bedload, and clasts are held in suspension by turbulence (Holdsworth & Collinson, 1988; Collinson, 1988; Orton & Reading, 1993).

2.2.10. Parallel-laminated siltstone to parallel-bedded medium-grained sandstone.

This facies is characterized by variable proportions of grey micaceous siltstone and medium-grained sandstone in a matrix of black to dark grey mudstone; the alternations in grainsize and colour define laminations. Like the previous facies, it is fairly common, and makes up +25-30% of the early Marsdenian succession in the eastern Pennine Basin. Gradational upward coarsening packages; up to 3m in thickness comprise repeated siltstone to sandstone units that show an upward increase in bed thickness (Figure 2.13). These coarsening upwards packages commonly have interlaminated siltstone and mudstone at their base, and are overlain by fine-grained sandstone in beds between 0.005-0.05m in thickness. The upper part of the unit has rare asymmetric ripple laminations and broad, shallow load casts. Isolated and amalgamated horizons of *Planolites, Olivellites and Scolicia* bioturbation occur within this facies. These horizons are laterally continuous and can be traced across exposures for several meters. The base of this facies has a either a gradational contact with underlying offshore or interdistributary bay deposits, or low angled sharp erosive base. The top is often erosively truncated by proximal mouthbar or fluvial facies associations.

<u>Interpretation: Buoyancy-dominated fine-grained mouthbar</u>

The dominance of parallel laminations and bedding, and the paucity of cross-bedding or cross-lamination suggests that this facies was deposited from suspension. The presence of continuous and discontinuous laminations imply the rate of deposition and degrees of bioturbation may have varied laterally. Variations in sediment supply rate and sediment calibre are evident from the rapid spatial and temporal variations in the proportion of mud to silt, and the deposition of sandy laminations from suspension. The cyclic occurrence of layers with

Planolites bioturbation implies nutrient availability within the substrate and water column varied. The overall lateral continuity of bedding suggests that deposition was unconfined. This facies represents deposition from suspension in a mouthbar setting, where rare tractional or low-density turbidity currents transport sediment down the mouthbar.

These conditions are likely when the density of the supply plume is low. Decreased discharge in the fluvial supply system can give rise to a lowering in the amount of bedload and suspended sediment, leading to the generation of a buoyant, or 'hypopycnal flow' (Bates 1953; Wright 1977; see Figure 2.12).

2.2.11. Sheet sandstones with low angle cross bedding

This facies comprises medium- to coarse-grained sandstone, with a high micaceous content and rare muddy drapes within the toe-set. This facies makes up to 5% of the succession in the eastern Pennine Basin, and was not observed in the western Pennine Basin. Planar cross bedding with a low angle (5-10º) is the dominant sedimentary structure, and occurs in sets up to 0.5m thick. Toe-sets are commonly swept out, top-sets are not preserved, and reactivation surfaces are common. Excellent exposures of this facies can be seen in the upper part of the section at Pule Hill (SE032107; Figure 2.14). Ripple lamination is commonly observed within foresets, and often indicates a transport direction opposite to that inferred from the low angled cross-beds. The cross-bedded sets can be traced laterally over 10-30m, appear mutually erosive and possess rare shallow concave up channels. Upper phase plane lamination is common on bedding surfaces. Palaeoflow measured from the cross- beds and upper phase plane bedding appears generally concordant. *Pelecypodichnus* is commonly observed where bed boundaries have no upper phase plane lamination. Massive sandstone, up to 0.2m thick and 10m across appear as lens like features within the low angle cross-bedded sandstone. The upper surface of this sub-facies is often not eroded by the cross-bedded sandstone, symmetrical wave rippled surfaces are observed.

Interpretation: Shoaling mouthbar

The low angled toe and foresets, lack of top-set preservation, along with the presence of upper phase plane bedding suggests that this facies was partly transported by fluvial waters flowing at a high velocity, while the presence of wave ripple lamination also suggests deposition in very shallow water. The close association of this facies with fluvial and mouthbar sandstone suggests proximity to a near emergent fluvial mouth. Specifically, during high fluvial discharge, frictional and inertial processes dominate effluent plumes, and sediment is transported in shallow water flowing at high velocities. By inference, bedforms in this environment are transient features, that are subject to rapid changes in geometry and lateral extent. This facies represents the zone of deposition where fluvial waters spill over the top of a prograding mouthbar. In this environment such flows are laterally unconfined, and are subject to rapid velocity fluctuations due to modulation in distributary channel discharge. A similar facies is described by Bristow & Myers (1989) within the Rough Rock (Yeadonian) of the central Pennines, and is inferred as deposited in a shoaling mouthbar (their Facies 4).

2.2.12. Parallel-bedded sandstone interbedded with siltstone/ silty mudstone

This facies comprises fine- to medium-grained sandstone with a muddy matrix, with common sub-parallel inter-laminations of mudstone and siltstone. This facies is rare and probably accounts for less than 2% of the succession. Bedding thickness is generally between 0.05-0.1m and commonly sub-horizontal (Figure 2.15). Rare flute and tool marks are seen on bedding planes. Trace fossils present in this facies include *Olivellites* and *Pelecypodichnus*.

At Kebroyd Bridge (SE04452120) a sub-horizontally bedded example of this facies appears to show cyclical variation in bed thickness on a metre scale. On this scale bed thickness varies from 0.1m to 0.02m on an apparent ~25-28 bed cycle, and individual beds commonly possess a micaceous silty lamination (1-5mm thickness) on their upper surface (Figure 2.16). The base of this facies is commonly sharp or erosive onto the silty interdistributary bay or offshore facies. The upper surface is either truncated by shoaling mouthbar/ distributary channel facies, or overlain a flooding surface and offshore or interdistributary facies.

<u>Interpretation: Tidally-influenced sand-rich mouthbar.</u>

The apparent cyclical thickness variation in association with the paired beds and thin laminations seen at Kebroyd Bridge (SE044212) is suggestive of a quasi-rhythmic fluctuation in the flux of sediment supply. Either allocyclic modulations in fluvial discharge or the influence of a tidal range may create such rhythmic bedding. Similar facies have been described in the Carboniferous of the Pennines and Appalachian Basin (Broadhurst, 1988; Read, 1991; Aitkenhead & Riley, 1996).

2.2.13. Siltstone to fine-grained sandstone interlaminated with grey to black silty mudstone

This facies comprises thin, parallel-laminated, micaceous siltstone to fine-grained sandstone interlaminated with grey to black micaceous silty mudstone (Figure 2.17). Proportions of sandstone to siltstone can vary, and mica is abundant in both sandstone and siltstone rich components of this facies. It probably accounts for less than 4% of the total succession within the early Marsdenian.

The bases of sandstone laminae are commonly erosive, and have scour-enlarged *Pelecypodichnus* aligned in the direction of palaeoflow. Bivalve escape structures create a pervasive sub-vertical fabric that truncates the primary depositional structures within the sandstone laminae. Some sandstone beds fine upwards into the grey silty mudstone. Cross laminations are draped by mudstone lamina where beds comprise of interlaminated fine-grained sandstone with higher proportions of muddy-siltstone. Asymmetrical, commonly starved ripple cross-lamination with flow reversals and less common symmetrical ripples characterise these siltstones. Reactivation surfaces are common in the sand-rich portions (Figure 2.18). Ichnofaunas include *Planolites*, *Curvolithus*, *Rosselia* and rare *Chondrites*. Carbonaceous debris is locally common, but mostly rare, and where seen is often reworked and fragmentary.

Interpretation: Tidally-influenced silt-rich mouthbar

The layer-parallel, interlaminated fabric, high mica content and small size of the cross laminations imply that this facies was deposited dominantly by suspension, but was subject to frequent low velocity currents and reworking. The variable proportions of fine sandstone to siltstone may be dependent on the proximity of a clastic source. This facies represents deposition in a shallow-water mouthbar front, analogous to distributary mouthbars in terms of process, but where the substrate is affected by reworking in the shallow water depths. Current ripple laminations with bimodal flow indicators are interpreted as representing periods of flow reversal during deposition. Subtle-laterally discontinuous reactivation surfaces are sub-parallel to bedding, and commonly truncate and are overlain by ripple structures.

In a shallow water mouthbar, variations in the degree of wave re-working or palaeocurrent resulting from clastic input from a number of sources, providing conditions for apparent flow reversal. For example, sediment may be either supplied from the mouthbar feeder channel, or overbank flooding associated with high discharge events in the distributary channel. Alternatively, flow reversals may have been caused by ebb and flood tidal influences affecting the mouthbar front during deposition. The *Planolites*, *Curvolithus*, *Rosselia* and *Chondrites* ichnofauna present in this facies suggests a brackish to marine water column (Eager *et al.,* 1985).

2.2.14. Interbedded sandstone and silty mudstone with symmetrical wave ripples and hummocky cross-stratification.

This facies comprises well-sorted fine- to medium-grained sandstone comprising quartz-feldspar (70% quartz, 20% feldspar, 10% mudstone), in parallel decimetre to metre-scale beds, with interbedded discontinuously laminated silty-mudstone (Figure 2.19). It is rare throughout the basin, and probably accounts for less than 0.5% of the succession. Beds have laterally continuous upper and lower surfaces (see Figure 4.7), and possess decimetre scale tabular and trough cross bedding. Flaser laminae and thin sandstone beds are common and rare *Cochlichnus* and *?Arenicolites* are observed. The sand-rich beds within this facies have sharp planar erosive bases, and *Pelecypodichnus* is present in high-density clusters. Current structures dominate the lower part of the sandstone inter-beds, implying infrequent episodes of tractional transport. Symmetrical ripple laminations with crest-to-crest wavelengths between 0.05-0.1m are present on the top surface of sandstone beds. Ripple profiles are observed, with common crest bifurcation. Hummocky cross-stratification increases in density towards the upper part of this facies, and several beds comprise amalgamated hummocks up to 1.5m in wavelength.

Interpretation: Wave influenced shoreline between storm wave base and offshore transition zone

Silty mudstones were deposited from suspension, and punctuated by the influx of tractionally transported sandstone. Symmetrical ripples on the upper bedding surface suggest they were affected by wave processes during and after deposition. This pattern is consistent with the offshore-transition zone, where mud-silt suspension deposition during fair-weather periods is punctuated by the of storm-events. During storms, high amplitude waves rework the shoreface to the storm-wave base, and transport sand into the offshore-transition zone by tractional and suspension processes, therefore creating the interbedded nature of this facies. The apparent increase in wave generated structures from the base to the top of this facies, and the high density of hummocky cross-stratification is suggestive of shoreline progradation.

2.2.15. Well-sorted fine- to medium-grained sandstone with amalgamated hummocky cross-stratification.

This facies comprises well-rounded to sub-angular, fine- to medium-grained, grain-supported, quartz-rich sandstone, and like the previous facies, probably accounts for less than 0.5% of the early Marsdenian succession. The sandstone beds are up to 1m thick and have a quartz-rich cement, that provides this facies with a high induration. Beds are sharp based with centimetre-scale erosional relief. The predominant sedimentary structures are amalgamated hummocky cross-stratification, with wavelengths in the order or 1-3m and crest to trough heights of up to 0.4m (Figure 2.20). Uni-directional trough cross-bedding occurs as isolated sets, and discontinuous minor erosive scours with less than 0.05m of relief often feature. Upper surfaces are characterised by symmetric ripple laminations with crest-to-crest wavelengths in the order of centimetres, implying that low energy wave processes also affected this facies during deposition. Some beds are truncated by quartz-rich pebble horizons with 0.05-0.2m of erosive relief. A high density-high diversity ichnofauna including *Rhizocorallium (janeuse & irregular), Treptichnus, Chondrites, Taenidium, Phycodes, Curvolithus*, is observed on some bedding surfaces (see Figures 4.5.1 to 4.5.4). Steep-sided erosive channels (2m wide, 1.5m deep) are observed in the upper part of this facies, and cutting into the underlying amalgamated hummocky cross-stratification (see Figure 4.8). These channels are filled with massive, well-sorted quartz-rich sandstone.

Interpretation: Shoreface above fair-weather wave base

The textural and mineralogical maturity of this facies suggests that it has undergone some element of mechanical reworking. Amalgamated hummocky cross-stratification and symmetrical ripples suggests the influence of basinal wave processes (Dott & Bourgeois, 1982), while the high density and diversity of marine trace-fossils implies that this facies was deposited in shallow water with normal salinity (Pemberton *et al.,* 1992). The presence of gravel-lag deposits suggests that wave reworking occurred to the extent that much of the finer substrate was removed and re-deposited, leaving a coarse-grained gravel-lag deposit.

Uni-directional currents from the shoreface to offshore may explain the presence of steep sided cut and fill channels, suggesting rapid cut and fill. The channels are further evidence for strong wave energy associated with storm events. The short-term increase in the elevation of sea-level via banking storm-waves at the shoreline, creates a hydraulic head at the shoreface, and the sea-ward return flow of water.

2.2.16. Trough and planar cross-bedded sandstone, with log debris

This facies comprises coarse- to fine-grained sandstone with disseminated granules and pebbles. It is fairly common, and accounts for greater than 15-20% of the early Marsdenian succession. The base of the facies is commonly coarser than underlying facies with tree trunk impressions, carbonaceous debris, and mud intraclasts. The bases of bedsets commonly have concave upward erosive surfaces, suggesting channelisation of flow on a metre scale. Trough and planar tabular cross bedding are the dominant sedimentary structures, although millimetre thick planar lamina and massive beds centimetre thick beds are observed. Set thickness is variable, with a maximum observed thickness of 0.5m. Bedform topsets are rarely observed, but where preserved they consist of current rippled sands or surfaces with primary current lineation. Bottom sets are commonly fine-grained, with a greater proportion of mica in comparison with the rest of the facies (Figure 2.21). Very coarse to granular, grain-supported centimetre thick grain flows erode dune foresets. Where top-sets are preserved in dune-forms, crest heights are between 0.1m-2m, and palaeocurrents indicators suggest a dominantly uni-directional flow regime.

Interbeds of carbonaceous silty mudstone are disseminated within this facies, are rich in carbonaceous debris, and are current ripple laminated and with wave reworked ripple laminations. *Pelecypodichnus* are common above set boundaries and where the dune top-set surfaces are preserved. Escape structures are rarely longer than 0.05m, and often verge in the direction of palaeoflow. Scour pits, enhanced by the erosion of bivalve resting traces are observed on erosive surfaces with dimensions between 0.05m up to 0.03m in length. This facies is often laterally continuous over several tens to hundreds of metres.

Interpretation: Cross-bedded sandstone.

The transport of coarse calibre sediment and the presence of locally erosive surfaces suggests that high flow velocities were sustained during transport of this facies. The vertical stacking of cross bedding suggests a continual transport of sediment, with rare flow waning, although massive sands may be deposited from suspension during rapid flow waning. Flow waning may account for the localized preservation of dune topsets (Collinson & Thompson, 1989). In outcrop, the lateral continuity of this facies suggests that it was laterally unconstrained during deposition, while the presence of amalgamated bedforms implies vertical accretion. *Pelecypodichnus* are the resting trace of semi-infaunal freshwater bivalves (Eager *et al.,* 1985). When the bivalves become engulfed by migrating sediment, they attempt to escape sub-vertically through the sediment. The silty horizons are suggested to have been deposited from suspension during low discharge periods. The dominantly unidirectional sediment transport direction, and the absence of associated abandonment deposits suggests that this facies represents bedform migration in a fluvial system.

2.2.17. Cross-bedded sandstone with mud-drapes.

This facies comprises coarse- to fine-grained sandstone with disseminated granules and pebbles, and thin muddy drapes (Figure 2.22.1 and 2.22.2), and probably accounts for approximately 5% of the Marsdenian succession. This facies is similar to cross-bedded sandstone facies within which it is often interbedded, but is distinguished by the presence of mud drapes on toeset and foreset surfaces, and a greater abundance of re-activation surfaces within bed-sets. The drapes comprise micaeous-muddy laminae that often, but not always, occur in pairs, up to 5mm apart. When traced, mud drapes in some cross bed sets occur in bundles, which are rhythmically spaced.

Interpretation: Tidally-influenced cross-bedded sandstone

The presence of paired muddy drapes, associated apparent bundling, and reactivation surfaces suggests either tidal processes (Nio & Yang, 1991), or regular pulsing of fluvial discharge. In this instance the mudstone drapes were deposited during periods of tidal slack, while the sandstone foresets represent deposition during the ebb and reworking during the flood tides. The local absence of a tidal signature, and the variation between this and the cross-bedded sandstone without mud-drapes can be explained by erosion by enhanced fluvial ebb currents from fluvial dominated channels. In this case, the inertia associated with the basinward flow of a body of fluvial water could have suppressed any tidal signature or allowed erosion of sedimentary evidence for tidally influence deposits.

2.2.18. Giant cross-bedded sandstone

This facies comprises medium- to coarse-grained sandstone, with fine-grained, sub-horizontally laminated, micaceous sandstone common in the toe-sets, and probably accounts for up to 5-10% of the Marsdenian succession. Individual foresets are between 0.15-0.4m thick, and commonly massive, although some possess trough cross bedding with common reactivation surfaces (Figure 2.23). The average apparent dip of foresets is around 19º, with a maximum of 30º recorded on some foresets, and a minimum close to sub-horizontal in asymptotic toeset fines. Two geometric forms of giant cross-bedding are observed; those with planar foresets and those with asymptotic foresets. Both forms occur in single, or double sets, overlain by abandonment facies/ flooding surfaces, or are separated by an erosive surface with metre scale relief (Figure 2.24). Giant cross-bedded sandstone sets often have a maximum set height between toeset and upper set boundary of 7-8m, although sets average 4m in height. This facies is often associated with erosively based cross-bedded sandstone facies (Figure 2.25). The base of this facies is commonly sharp, and overlain by down-lapping cross-beds. Concave upwards erosive bases are often observed within this facies in laterally extensive exposures. Trackways interpreted as that of a sub-aqueous amphibian have been identified on giant cross bed foreset surfaces at Bingley Road Quarries (SE053382; Martin Whyte, *pers. comm.*).

Metre-scale deformation structures in giant cross-bedded sandstone facies are stratigraphically restricted to the sandstones between the R2b2-R2b3 marine bands

(Woodhouse-Scotland Flags), at Bingley Road Quarry (SE053382), to a lesser extent at Park Wood Quarries (SE068407) and on isolated outcrops on Rombalds Moor. The structures at Bingley Road Quarry are confined between the upper and lower bounding surfaces of individual planar giant cross bedding sets (Figure 2.26). Subsequent giant cross-beds prograde over the slumped feature and appear less deformed than the underlying set, implying that the deformation was syn-depositional. Individual forests are overturned in the direction of cross-bed dip and the base of the deformed zone gently dips sub-parallel to the dip of the giant cross-beds. This suggests that the deformation structures have a slumped component, with slumping in the direction of giant cross-bed dip. Some foresets have a flame like appearance, and are sub-vertical in orientation, or verge in an opposite to the cross-bed dip.

Interpretation: Fluvial barform

The dominantly massive nature of the giant foresets suggests accretion was predominantly by grainflow down the lee slope of the bedform. The association with cross-bedded sandstone facies suggests occasional tractional transport flow down the lee-slope of this feature; i.e. 'intrasets' (Collinson, 1968b). Intra-facies erosive surfaces have less relief than those described in the Kinderscoutian by Hampson (1997). Finer grained deposits in the toe-sets, imply that deposition occurred from suspension in the lee of this feature, suggesting that flow separation on the crest of this bedform was sufficient to allow a substantial flow waning within the trough area. Barform crests are straight (planar foresets), with rarer three dimensional dune forms (asymptotic foresets), suggesting the temporal dominance of turbulent flow fluvial discharge. The abundance of intra-facies erosive surfaces suggests a fluctuating flow-regime. Where the base of this facies is observed (Figure 2.24), suggests that the base of this facies is erosive, and by analogy with Kinderscoutian examples (Hampson, 1997), may be deposited within a channel. This study has not identified 'undulatory bedding' (*sensu.* McCabe 1977) in Marsdenian outcrops (see Case Study 3).

Various theories for the genesis of the scale and type of deformation structures (seen in barform set 1 and 2, Figure 2.26) may be invoked, and it seems likely that no individual mechanism was wholly responsible. The generation of planar giant cross beds in shallow-water prograding delta fronts could certainly have the potential to be effected by slumping, and loading on the delta top, or seismic triggering could have initiated slumping on this scale. There is no direct evidence for syn-depositional seismic activity during the Marsdenian, although this mechanism should not be ruled out. Additionally, rapid deposition may have inhibited syn-depositional dewatering via flow through the permeable substrate. In this event, an increase in the pore fluid pressure, would be associated with a decrease in the mechanical strength of the sediment, and led to rapid failure via dewatering and/ or slumping (see sequence stratigraphic paper for discussion of this).

2.2.19. Giant cross-bedded sandstone with muddy drapes

The lithology, geometry, scale, and sedimentary structures of the giant cross beds are similar to the giant cross-bedded sandstone facies. It probably accounts for 1-2% of the early Marsdenian succession. However, giant cross-bed toesets and foresets are commonly draped by micaeous-muddy laminae that often occur in pairs, up to 0.5cm apart, and in some examples repeated sets appear bundled. Re-activation surfaces are common throughout this facies, as are erosive surfaces. This facies is found in association with *giant cross-bedded sandstone*, which overlies a coset of this facies at Moselden Heights Quarry (SE044163; Figure 2.24). It is also associated with erosively based cross-bedded sandstone and tidally-influenced cross-bedded sandstone.

Interpretation: Tidally-influenced barform

The process generating the foresets in this facies is inferred as similar to that of the giant cross-bedded sandstone facies. However, muddy-drapes are suggestive of rhythmic flow waning and suspension deposition. The presence of paired muddy drapes, associated with apparent bundling, and reactivation surfaces suggests that tidal processes acted upon the sediments during deposition (Nio & Yang 1991). If tidal mechanisms are responsible for the deposition of this facies, it is likely that deposition occurred in the landward most part of the tidal reach, where the tidal influence is weak and fluvial processes dominate. Alternative processes may have caused these structures, such as the pulsing in fluvial discharge during sediment transport, allowing suspension deposition of silt and mudstone during low discharge stage.

2.2.20. Heterolithic interbedded sandstone-siltstone with Pelecypodichnus and thin palaeosols

This facies comprises interbedded siltstone and sandstone beds. Sandstone bed thickness is commonly less than 0.5m, while the siltstone can vary from 1-5m in thickness. It is rare and probably accounts for less than 0.5% of the Marsdenian succession. Both lithologies commonly form an interbedded facies where siltstone and sandstone beds are broadly equal thickness. The sandstone and siltstone beds have a sheet-like geometry and downlap onto underlying sediment of the same facies over 50-75m (see Figure 4.8.3). Common 2-5m wide, <0.4m deep concave-up scours are seen on the base of the sandstone beds, while the upper surfaces have commonly preserved three-dimensional symmetrical and asymmetrical ripple lamination, implying flow waning, and/ or reworking by wave ripples. Cross bedding sets within sandstone beds generally appear to decrease in thickness through individual beds. Climbing ripples are common within the upper part of these beds. Siltstones are commonly carbonaceous and highly micaceous, with lenticular to flaser sandstone laminations. Thin (>0.1m) palaeosol horizons with rooting and carbonaceous debris are common within this facies, but tend to occur toward the upper sections. *Pelecypodichnus* and bivalve escape structures are common on the bases of sandstone and siltstone beds.

Interpretation: Overbank flood deposits/ crevasse splay

The broad sheet like geometry of the sandstones suggests deposition from unconfined sheet-flows, while the concave upward erosive scours represent the local development of shallow channels. The sandstone and siltstone interbedded character suggests fluctuations in flow velocity, while the upwards decrease in bedform size within individual sandstone beds implies deposition by an individual fluvial event of waning flow strength or shallowing in water depth. The siltstones were deposited predominantly by suspension, with isolated low-velocity currents that transported or reworked coarser grained clastics. The association of preserved wave ripple lamination and thin palaeosols associated with both lithologies imply localized shallowing and emergence. This facies was deposited in an area of shallow standing water adjacent to levees and active fluvial channels. Similar environments are ascribed to the sub-aerial levee/ bayhead environment of the Mississippi River Delta (Coleman, 1988).

2.2.21. Parallel-bedded inclined heterolithic sandstones and carbonaceous silty mudstone.

Lithologies in this facies comprise interbedded fine- to medium-grained sandstone and carbonaceous silty mudstone. Sandstone beds are often less than 0.2m in thickness, have layer parallel bedding and are commonly massive, with some examples having weak internal erosive scours with relief in the order of 0.05 - 0.1m. The finer grained interbeds are often thinner (less than 0.1m thickness), and pinch out laterally where sandstone beds become amalgamated. Asymmetric ripple and flaser laminations are common in the carbonaceous silty mudstone lithologies. Carbonaceous debris is often scattered throughout this lithology, is often less than 0.03m diameter, and commonly reworked. This facies is rare, and probably accounts for less than 0.5% of the Marsdenian succession.

In extensive exposures (i.e. Fletcher Bank Quarries SD805164; Figure 4.9) the bedding within this facies inclines at low angles (less than 10°) for considerable distances (up to 350m), and the basal contact is represented by a series of the down-lapping beds defined by the interbedded lithologies (Figure 2.27). Sandstone beds often appear to thin towards the downlapping surface, and in some cases pinch out completely before they down-lap at the base of this facies.

Interpretation: Laterally accreting point bar

The planar and inter-bedded relationship of the lithologies in this facies, along with the lack of tractional sedimentary bedforms, suggests that the dominant mode of deposition was from low velocity tractional currents and suspension. The geometry and lateral continuity of the inclined bedding surfaces is suggestive of lateral accretion surfaces that are characteristic of point bar formation in low discharge and mixed bedload, sinuous fluvial channel systems (Allen, 1964; Orton & Reading, 1993). Individual beds of sandstone are inferred as erosively based where they are seen to truncate underlying lithologies. Where downlapping sandstone beds thin laterally, erosive scours at the base of the bed diminish in relief and bedding becomes concordant with the background carbonaceous silty mudstone. This suggests fluctuations in flow velocity along point bar bed surfaces, that may be dependent on proximity to the axis, or the distribution of flow in the fluvial channel.

On laterally accreting meander point bars, sediments are deposited primarily from suspension and low velocity tractional events, suggesting that the channels dominantly maintained low fluvial discharges and a mixed mud-sand bedload (Jackson, 1981; Smith, 1987; Jordan & Prior, 1992).

2.2.22. Rooted mudstone with carbonaceous debris

This facies comprises pale grey to cream mudstone, with common ferrous orange-red staining. Carbonaceous debris is abundant throughout, and appears as either silt to fine-grained sandstone, or in string-like forms. Rootlets are abundant, and range in length from a few millimetres to several decimetres. Common hair-like roots are preserved as carbonaceous streaks, along with sub-horizontal *Stigmaria* rootlets (Figure 2.28.6). Clean exposures often reveal spherical masses of fine rootlets, producing radiating (star-like) forms. The mud matrix is commonly organised into grain-supported mudstone aggregates that are up to 15mm in length, and either cubic, sub-rounded or lath-like in form (Figure 2.28.6). Between these aggregates, very little interstitial clay material is observed and slickensides are common. Common circular and elongate pores appear throughout, occurring at a similar density and diversity as the hair-like structures. Pores are commonly infilled with fine-grained debris. Numerous disseminated, matrix supported, angular, lath shaped fragments of lithified sandstone are present. These are often of the same lithology as the underlying facies. The rooted mudstone is commonly overlain by thick (both *in-situ* and drift/ swamp) coals (less than 10cm thick).

Interpretation: Muddy palaeosol

The mudstone aggregates comprising lath- and crumb-like ped structures are interpreted to be produced by the interaction of root hairs and substrate, whereas the carbonaceous streaks probably represent the *in-situ* remains of rootlets (Retallack 1988). Slickenside surfaces are probably formed by post-depositional loading and dewatering due to overlying sediment deposition or syn-depositional rootlet activity.

The high density and diversity of roots and the abundance of grain supported mud aggregates, inferred to be ped structures (Retallack, 1988), suggest that copious rooting occurred. The cemented, grain-supported texture of the peds, the paucity of interstitial clays, and the presence of rootlets suggests that fluid flow occurred dominantly in a sub-vertical orientation during formation. High density rooting such as this may have provided mechanical strength to the paleosol, making subsequent erosion by river channels less-likely (McCabe, 1984; Jones *et al.,* 1995). During palaeosol formation, erosion is inferred, from the presence of Krotovinas (washed in horizons). Such structures are formed by the removal and infilling of large roots during palaeosol reworking (Retallack, 1988). The mud-rich lithologies within this facies, along with the lack of desiccation indicators suggests that this type of palaeosol was formed in waterlogged, poorly drained conditions during periods of high water-table.

2.2.23. Rooted quartz-rich sandstone

This facies is characteristically a highly indurated, cream to white, medium- to coarse-quartz sandstone, with rare weathered feldspathic clasts. The sandstone is grain supported and often has a sugary texture (Figure 2.28.5). A matrix is absent in the greater part, although localized orange-brown interstitial clays are observed on some horizons. Well-preserved *Stigmaria* are abundant within this facies (Figure 2.28.8). Aggregates of individual sand grains form a variable fabric within this sub-facies, which in the some parts appears sub-rounded to lath-shaped and sub-horizontally orientated. The basal part of this facies often contains discontinuous convex up surfaces, which are intersected by an anastomosing network of sub-vertically orientated slickensided fractures. Above this, an iron oxidized sub-horizontally laminated medium-grained sandstone commonly separates the lowest part from a sub-vertically rooted medium sandstone.

Interpretation: Leached sand-rich 'ganister' palaeosol

The presence of rooting and the strongly leached nature of this facies implies palaeosol formation on a well-drained sandy substrate (either fluvial or mouthbar facies associations). By inference, the palaeosol may have formed while the water-table was relatively low, therefore allowing downward flow of ground water. The leaching of clays through the substrate forced the concentration of residual quartz and feldspar in an albic horizon (Percival, 1986). Deformation within the fabric appears to be concentrated along restricted, deep-seated, slikensided surfaces. This suggests water movement occurred along well-developed, possibly temporally persistent planes. The lack of thin hair-like rootlets forming a framework to support the palaeosol suggests the environment during formation may have restricted the growth of widespread vegetation.

2.2.24. Rooted interbedded sandstone and siltstone palaeosol.

This facies is commonly associated with interbedded carbonaceous mudstone, sandstone (maximum bed thickness 0.4m) and siltstone (maximum bed thickness 0.2m) overlain by thin (<5cm) dull coal facies, and is generally thinner than the sand-rich leached ganisteroid palaeosols (Figure 2.28.2). Large *Stigmaria* roots are common, as are hair-like roots. The muddier beds within this facies have similar characteristics to the rooted mudstone, but generally possess a greater proportion of carbonaceous debris. *Pelecypodichnus* are common at the base of the sandstone beds. Current ripple lamination is often preserved where rooting is limited. Harper Clough Delph (SD716317) has a spectacular example of this palaeosol type (Figure 2.28.1 and 2.28.3), with an extremely well preserved *Lycophyte* stump, similar to an example referred to by (Wagner, 1995) (his Figure 9). It possesses a characteristic four cornered shape, with *Stigmaria* roots plunging at a shallow angle into the substrate.

Interpretation: Waterlogged heterolithic palaeosol

The lack of evidence for leaching implies formation in waterlogged conditions, while the interbedded sheet-like character implies from suspension and tractional transport of sand, in an unconfined flow. This facies is considered to be similar to the crevasse splay facies in terms of characteristic geometry and primary depositional process. The generation of palaeosol horizons within this facies was probably coincident with substrate deposition, and implies periodic substrate emergence. Similar facies are described from the Pennsylvanian Cumberland Group (Nova Scotia), which are attributed to deposition in an anatomising fluvial overbank plain (Rust *et al.,* 1984). The occurrence of upright, woody plants implies inundation of the floodplain by sediment-bearing flood water, which rapidly bury the trees.

2.2.25. Black mudstone with abundant leaf impressions and plant debris and thin drift coal.

This facies comprises dark grey to black parallel-laminated, slightly micaceous mudstone, with discontinuous silty laminae. Well-preserved, scattered plant debris is abundant in the lower part of the facies. In several exposures, there appears to be a gradational contact between the black coaly mudstones and the underlying palaeosol facies. Rare rootlet horizons associated with orange weathered clay horizons are present throughout this facies. Laterally discontinuous lath-shaped to sheet-like drifted vitric coal debris is common. Rare disarticulated *Dunbarella* valves and *Caneyella* are found in the upper part of the facies.

Interpretation: Allochthonous/ swamp coal

The dominance of a black mudstone suggests deposition from suspension with rare silt input. The abundant well-preserved plant debris suggests a fairly local source; perhaps representing reworked organic matter associated with underlying paleosol facies, or washing in of material from surrounding areas. Standing water or heavily waterlogged conditions during the formation of this facies allowed the horizon of organic matter to be deposited from suspension and preserved over a large spatial area. The accumulation of large quantities of decaying plant debris probably increased ground water or water column acidity, restricting bacterial activity, and allowing preservation of organic material. The association of this facies, with an underlying, commonly mud-rich, waterlogged paleosol suggests a persistently high water-table during formation. Drift coals are common in sediment starved lacustrine environments (McCabe, 1984), where floating plant debris is allowed to accumulate. The rare occurrence of *Dunbarella* and *Caneyella* suggests such shallow water lacustrine systems were sometimes open to marine influences, or inundated by low magnitude flooding/ storm events.

2.2.26. Dull coal, with well preserved plant debris and micaceous silty partings

This facies comprises a thin, poorly developed coal, which is rarely thicker than 20cm. The coal has rare micaceous and silty partings, and is well laminated. Isolated laminae of millimetre thick vitric coal are observed. Well-preserved plant debris is abundant in both mesh like accumulations and isolated leaf and stem impressions. Common laminae of fusain (fossil charcoal) are abundant within some coals and exposed surfaces often possess a oily sheen.

Interpretation: In-situ coal

This coal has a minimal clastic component and was deposited on abandoned or clastic starved parts of the delta-top. This is corroborated by the presence of charcoal deposits suggesting the occurrence of fires, which may have occurred in dry or semi-dry environments. Rare silt partings suggest the presence of some suspension or wind-derived deposits.

Chapter 3: Facies associations

3.1. Introduction

In order to analyse the environments of deposition in the Marsdenian, and to provide a useful 'building-block' for palaeo-environmental analysis and correlation, facies are grouped into *facies associations*. Boundaries between facies delineate lateral and vertical changes in environment of deposition. This relationship conforms to *Walters Law* (Walther, 1894), which states that facies in vertical contact with each other must be the product of spatially neighbouring environments during deposition. A *facies association* comprises a set of facies that are found together in a vertical succession, without any major break in deposition. Facies are described in the Facies Chapter, and only their relationships and shared characteristics are described here.

3.2. The hierarchy and differentiation of facies associations from depositional systems

Thirteen facies associations are recognised in this study, and these are further grouped into four *depositional systems* (see Chapter 5). The grouping of sedimentary packages into facies associations and depositional systems permits flexible manipulation of facies data in palaeo-environmental analysis and correlation. This is useful when studying sedimentary packages on a variety of scales, as it allows the investigation of lateral variation of environment over tens of metres to tens of kilometres. Secondly, and most critically, the application of this system to both graphic correlation and palaeogeographic reconstruction increases the potential of identifying hierarchical stacking patterns on a variety of scales.

3.3. Present-day and ancient environmental analogues

Nine 'case studies' are presented in the next three chapters. These represent detailed analogues of depositional environments, which are critical in understanding of the processes and mechanisms that forced changes in environment, and ultimately the stacking pattern of facies. Modern analogues include examples from the US Gulf Coast (particularly areas including and surrounding the Mississippi River Delta), while ancient analogues concentrate on examples from the Upper Carboniferous basins of Europe, Canada and the US.

3.3.1. Basinal mudstone association

The offshore marine band mudstone facies, or in some cases, the offshore black mudstone, commonly forms the basal unit in this association. The presence or absence of marine conditions during deposition is inferred by the identification of marine faunas. Parallel laminations occur throughout all facies, with the exception of weak sole marks and flute casts of various scales in the offshore prodelta silty-mudstone facies. All facies contacts are gradational and non-erosive.

Interpretation of basinal mudstone association

This association represents the distal underfilled basinal areas of the delta, where suspension deposition predominates over tractionally derived input. It is not possible to ascertain whether the offshore prodelta silty mudstone facies was deposited in saline waters via analysis of lithology and grainsize alone. However, the abundance of suspension deposited siltstone suggests a distal clastic input source and a potential increase in fluvial discharge into the basin.

If the Pennine Basin possessed a compartmentalised form during the Marsdenian, the high volumes of fluvial discharge may have diluted and displaced saline waters. Marine incursions may therefore have been short-lived (Holdsworth & Collinson, 1988), implying that the majority of the basinal mudstone association was deposited in a low or non-saline basin. Although lithology and grainsize analysis will not determine basin-water salinity during deposition, the abundance of marine macro-faunas and ichnofacies in the basinal mudstone association may suggest saline conditions prevailed.

In the offshore prodelta the suspension deposition of siltstone may have increased the amount of oxygen arriving at the sea-bed, accounting for the degree of bioturbation. Changes in salinity, basin currents or nutrient level may also allow anoxia to occur in the water column or the sediment-water interface (Holdsworth & Collinson, 1988; Wignall & Maynard, 1993). Conditions favourable to the colonisation of bioturbating organisms may therefore have been dependant on a variety of regional and local factors including the firmness and oxygenation at the substrate-water interface.

The presence of parallel laminations, sole marks and flute casts of various scales in the offshore prodelta silty-mudstone facies, suggests that the flow regime varied throughout the deposition of this facies. Likely mechanisms for the initiation of flow in this setting are i) instability and slumping within adjacent delta mouthbar association due to loading, or ii) allocyclic changes in the flow regime in the mouthbar distributary channels, that may have increased the amount of suspended material entering the basin. Increased suspended load could have produced hyperpycnal flows that may have initiated low-density turbidity currents (Bates 1953).

Similar associations are described throughout the Namurian i.e. facies H and J in the Kinderscoutian Grindslow Shales of northern Derbyshire (Walker, 1966a); siltstone and silty sandstone facies between the Marsdenian Alum Crag Grit and R2b5 marine band near Blackburn (Collinson *et al.,* 1977); part of log facies B from the Marsdenian of the East Midlands (Church & Gawthorpe, 1994). The basal areas of the on the continental shelf fronting the Mississippi River Delta also comprises similar lithofacies (Coleman 1981). Mississippi prodelta lithofacies comprise a blanket of suspension deposited clays, that have a high lateral continuity and low lithological variation.

3.3.2. Turbidite facies association

Facies boundaries within this association are erosive, and the association commonly overlies the basinal mudstone association, with a sharp contact, which has up to 10m of erosive relief in some sections (see Figure 4.12). The relatively steep gradient of this contact over short distances and thickness variations delineates significant areas of localised scouring and infill, and the areas where channel complexes developed. Within the channel complexes, shallow concave-up channel geometries are common (between 10-30m in width and less than 5m in thickness). When traced laterally, individual channels often appear mutually erosive. Mudstone raft lithologies are common in the base of individual channels away from the margin of the channel complex, suggesting they are deposited close to the channel axis. Where the base of this association is planar and bedded is concordant with underlying facies, deposition is inferred as having occurred in laterally unconfined basin-floor turbidite lobes (see Figure 4.11). The sequence stratigraphic paper discusses the positioning of channelised and basin-floor turbidites, and suggests the relative positioning of these forms.

Interpretation of Turbidite facies association

High flow velocities during transportation are invoked because of the overall coarse grainsize of this association. Turbidite processes dominate transportation in this association, but high-density currents may have been short lived, and a decrease in flow strength probably forced the collapse of the turbidity flow (Lowe, 1982). Under these hydrodynamic conditions, flow waning led to the deposition of the lower density turbidite sandstone. The classification of horizons within this association according to Bouma's horizons (Bouma, 1962), or horizons defined by Lowe (Lowe, 1982) is often not possible. Erosion during the deposition of the overlying bed is common, implying that it is difficult to identify the base and top of an individual turbidite deposited bed (Walker, 1965). This is especially true in the case of channelised turbidites, where lateral confinement implies that successive beds are mutually erosive. Facies comparable to the high-density turbidite sandstone are described in the Marsdenian west of Blackburn (Collinson *et al.* 1977; medium, thick-bedded turbidites and deformed mudstone conglomerates) and the Kinderscoutian of northern Derbyshire (Walker 1966a; facies B, C and E).

The repetition of fining upward units with basal surfaces that are both erosive and non-erosive within the low-density turbidite sandstone facies implies that the process of deposition was episodic and underwent variable turbulence. Similar facies are described in the Marsdenian west of Blackburn (Collinson *et al.* 1977; thin bedded turbidite sandstone facies) and the Kinderscoutian of northern Derbyshire (Walker 1966a; facies A and B). It is likely that this facies represents levee, overbank or channel abandonment deposits, which are difficult to differentiate due to similarities in depositional process (Mutti & Normark, 1987; Clark & Pickering, 1996) and limited outcrop extent in the Pennines. During periods of high discharge, turbidity currents may have transported sediment along flow-pathways to adjacent overbank areas by avulsion processes or flow-stripping from the upper part of the flow. If this is the case, low-density turbidity flows may represent overbank deposits, which are deposited during and after the collapse of the turbidity flow (Lowe, 1982).

3.3.3. Inter mouthbar bay association

The interdistributary bay facies forms the main component of this association, while shallow water restricted marine bay facies are geographically restricted to the northern parts of the basin. The occurrence of the *Zoophycos* and sponge spiculite prone shallow water restricted marine bay facies at isolated localities (such as Ponden Clough SD981364) reveals a gradational upper and lower contact with the interdistributary bay facies (Figure 3.1). The base of this association is gradational where it is seen to overlie the basinal mudstone association and sharp where overlying proximal and distal mouthbar front association.

Interpretation of inter mouthbar bay association

The sharp basal contact suggests an abrupt cessation of mouthbar deposition, forced by either an allocyclic decrease in discharge from distributary channels, or flooding and retrogradation of the supply systems. Either of these events would have allowed the commencement of suspension deposited interdistributary bay sediments.

Zoophycos is generally regarded as an ichnofauna associated with the deep marine deposits (Seilacher, 1967). However *Zoophycos* also occurs in shallow marine environments where the sea-bed is not reworked by currents, waves or other bioturbation (Hallam 1975) and the occurrence of *Zoophycos* in close proximity to the delta front mouthbar association suggests a shallow water origin for this setting. Sponge spiculite horizons in the Marsdenian appear analogous to the spiculites of the upper Pottsville to lower Allegheny (Pennsylvanian) of southern Ohio (Cavaroc & Ferm, 1968). These are interpreted as the deposits of a transgressive event, when little or no clastic input was entering the basin.

Similar mudstone facies are described from the Barataria Bay, of the Mississippi River Delta (Tye & Kosters, 1986). In this area inter-mouthbar bay deposits comprise fine-grained bioturbated siltstone and deposited within an interdistributary bay. Their work uses the presence of brackish and marine bivalve fauna and flooding surfaces to differentiate between restricted and open-marine bay sequences, in a similar way to the observation of *Zoophycos* and *Lingula* in Namurian deposits.

Facies associations with similar characteristics to Marsdenian deposits have been interpreted by Coleman (1981) in the modern Mississippi River Delta interdistributary bay deposits. Modern analogues include open waters, which may, or may not, be connected to the open sea, and are surrounded by sub-aerial exposed environments including levees, marsh and infilled interdistributary bay. Modern Mississippi bays are often less than 4m in depth, and deepen in a basinward direction (Elliott, 1974). The silty component of interdistributary bays is suggested by Coleman (1981) to have been deposited during times of high fluvial discharge when crevasse splays and levee overbank flooding leads to the transport of fines into the bay areas. In the Marsdenian Pennine Basin silt-prone components are thin, suggesting limited clastic input into the bays. Fining-upward sand-rich beds suggest density undercurrents may have been responsible for deposition of these beds (Elliott, 1974), implying that sediment may have traveled some distance from its fluvial source.

3.3.4. Distal mouthbar association

The buoyancy-dominated fine-grained mouthbar facies forms the main component of this association, and it is often interbedded with thin, offshore prodelta silty mudstone facies, representing the deposition of background silty sediments. The base of this association is always gradational, overlying the basinal mudstone association or the inter mouthbar trough association, while the top is either gradational or sharp (and sometimes erosively) overlain by the proximal mouthbar association.

Interpretation of distal mouthbar association

Within the buoyancy-dominated fine-grained mouthbar facies, repeated coarsening up units may infer an autocyclic control on sediment input close to the point source axis. At periods of low fluvial discharge, the flow may have contained less suspended sediment and therefore depositing little sediment. When the effluent plume has a lower density than the surrounding basinal waters it will tend to float, producing a 'hypopycnal flow' (Bates 1953; Wright 1977; Orton & Reading 1993, see Figure 4.12). This case is more likely if basinal waters are either saline, or the density of the effluent is lower than the basinal waters (either by the dilution of sediment or the transport of finer/ less dense particles). The expansion of an effluent plume into relatively deep water also may allow flow separation from the sediment-water interface (Orton & Reading, 1993). Hypopycnal flows allow deposition from suspension over large areas, and could therefore produce laterally unconfined, suspension dominated mouthbars.

Similar deposits occur in the distal mouthbar of the delta front environment of the Mississippi River Delta (Coleman, 1981). In the examples from the Mississippi deposits in this setting comprise small-scale, cross-laminated bioturbated siltstones interbedded with mudstone and thin sandstone beds. The beds in these heterolithic units all dip seaward at angles no greater than one degree.

3.3.5. *Proximal mouthbar association*

The proximal mouthbar association comprises the bedload-dominated, sand-rich mouthbar and rarer tidally influenced sand-dominated mouthbar facies. The proximal mouthbar facies is commonly overlain by the shoaling mouthbar facies and/or cross-bedded sandstone facies. Facies boundaries within the association are often sharp and erosive.

The upper contact of this association is either i) a sharp-based contact with the inter mouthbar trough association representing a deepening event ii) a waterlogged clay-rich or heterolithic palaeosol facies of the delta top association, representing fill to base level and emergence, or iii) a sharp erosive contact with the base of the channelised fluvial sandstone facies of the fluvial plain association (with 0-10m erosive relief) representing a shallowing event. The base of this association is either gradational or sharp, overlying distal mouthbar associations or inter mouthbar trough associations (see Section 6.4.1). Bedding surfaces within this association reveal a range of geometries ranging from low angled sheet deposits to steeply dipping bedding surfaces.

Cyclically laminated mouthbar deposits

Rare exposures (e.g. Kebroyd Bridge SE045213) of the proximal mouthbar association contain the tidally-influenced sand-rich mouthbar facies (see Figure 4.16). Examination of this facies reveals striking patterns in lamina thickness.

A repeated centimetre-scale pairing of thin-thick lamina is clear in both photographs (see Figures 2.15 & 2.16), and on the lamina thickness bar charts (Figure 3.2.1 & 3.2.2). On this scale bed thickness varies from 0.1m to 0.02m on an apparent ~25-28 bed cycle, and individual beds commonly possess a micaceous silty lamination (1-5mm thickness) on their upper surface. This pairing of lamina thickness is clear in both photographs (Figure 3.2. & 3.3), and on the lamina thickness bar charts. On a broader scale, lamina occur in sets of thinner and thicker cycles over 1-2m (see Figure 2.16). Sections 1 and 3 from the Kebroyd Bridge section reveal the number of lamina between those of maximum thickness is 43 and 48 respectively.

Fourier time-series analysis of mouthbar lamina

Fourier time series analysis is used here to determine harmonic components of a complex wave. The 'wave' is formed by a string of data; the lamina thickness, which forms the sandstone lamina thickness bar-chart (Figure 3.2.2 & 3.2.3). The data was collected in the field, and measured to nearest half millimetre using Vernier forceps. In all three sections, the thickness of both sandstone and mudstone/ micaceous laminae were measured. Fourier time series analysis was run on Kebroyd Bridge sections 1 and 3 by Professor Alan Archer, University of Kansas, using the same Fourier analysis program used by Archer & Kvale (1997) to analyse the Kinderscoutian mouthbar sections. The Fourier transform is run on the data in MATLAB, and the data input and results from this analysis can be seen in Figures 3.2.2, 3.2.3 & 3.2.5. The short length of the input data string, along with the apparently 'noisy' nature of the data implies that the output of the Fourier transform is not as refined as that of (Archer & Kvale 1997). This is because periodicities greater than ten lamina cannot be resolved in the sections detailed here, due to the short length of the dataset. When compared to the Fourier transform results from lamina in the Kinderscoutian mouthbar deposits (Archer & Kvale 1997), both datasets have similar ranges of harmonic output, and the results of the analysis from section 1 and 3 share broadly similar peaks and troughs (Figure 3.2.5). The harmonic wave output of sections 1 and 3 falls into two frequency groups; between 1.9- 2.6 and 3.6- 4.4, suggesting an output equating approximately to 2 and 4 lamina. Although periodicities greater than ten lamina cannot be resolved due to the short length of the data set, periodicities between the thickest lamina in section 1 and 3 (43 and 48 lamina respectively) may be calculated (Figure 3.2.1 & 3.2.3).

Interpretation of proximal mouthbar association

Shoaling mouthbar and cross bedded sandstone facies represent mouthbar distributary channels, and differ from the stacked fluvial channel association, as they comprise laterally discontinuous, single storey channels that are less than 10m thick. These channels supply sediment to the advancing mouthbar and are therefore genetically linked to deposition taking place in the basinal direction; distributary channels described by Collinson & Banks (1975) and Collinson (1988).

Bedding surfaces represent true depositional dip, and their lateral extent reveals the geometry of individual mouthbars. Mouthbars with shallow dipping geometries represent deposition in shallow water, while those with steeper fronts are similar to ‘Gilbert type’ deltas, and were deposited in deeper water. When correlated laterally a proximal mouthbar association may vary in dip from steeply to shallow, suggesting localised changes in depositional gradient via changes in water depth. Deposition in the lower part of this association is dominated by hyperpycnal or ‘gravity underflows’ that transport sediment by a combination of tractional and suspension processes depositing sediment from suspension over a laterally unconfined area basinward of the sand-rich mouthbar (Wright, 1977).

Delta front mouthbar facies have been identified in the Marsdenian of western Ireland and (Pulham, 1988) and South Wales (Hampson, 1998). These share many characteristics with the examples from the Marsdenian of the Pennine Basin, and due to extensive nature of the coastal exposures provide a clear picture of the lateral continuity of mouthbar deposits. Similar facies are also described in the Pennine Basin by Collinson & Banks (1975), Miller (1986) Church & Gawthorpe (1994), Martinsen *et al.* (1995), Hampson *et al.* (1996). The combination of mouthbar and distributary channel associations in the Bideford Group (Namurian), North Devon (Elliott, 1976), is similar to the succession described here, and suggestive of a highly constructive, rapidly prograding mouthbar complex (see Figure 4.15).

The Haslingden Flags occur in the Yeadonian of the Pennine Basin, where they form an easterly sourced sandstone (Collinson & Banks, 1975; McLean & Chisholm, 1996), inter-bedded by the *Cancelloceras cumbriense* (G1b1) marine-band and overlain by the *Anthraconia bellula* marine-flooding horizon. They were previously inferred as coeval to the overlying

Rough Rock, but recent work has correlated the *Anthraconia bellula* transgressive horizon, which separates the Haslingden Flags from the Rough Rock, with an *Anthracoceras* marine band in the eastern Rossendale sub-basin (Hampson, 1995; McLean & Chisholm, 1996). This suggests that the Haslingden Flags represent a single elongate delta that prograded in an easterly direction into the underfilled Rossendale sub-basin, which was subsequently inundated by a basin wide marine transgression.

The Haslingden Flags may represent the initial, and brief (allocyclic driven?) input of sandstone from an eastern provenance, explaining why they form the only true 'birdsfoot' type delta system identified in the Pennine Basin. If the Haslingden Flags were allowed to develop uninterrupted by the *A. bellula* transgression, the subsequent progradation of many individual elongate delta systems, as is inferred in the Marsdenian, may have generated a diachronous based mouthbar complex, where many mouthbar bases laterally coalesced forming a mouthbar sheet. Additionally, exposures are often small and spatially separated in the Pennines, so the elements defining the margins of individual elongate delta geometries are difficult to distinguish. Therefore the amalgamated mouthbar systems formed of many individual elongate mouthbars appear sheet-like in geometry. An indication of the elongate form of individual deltas in the delta front mouthbar depositional system can be see in the palaeogeographic reconstructions (see Figures 8.2 & 8.3) and strike correlation panel sections (see Figure 7.2).

In the Mississippi River Delta, submarine ridges (sub-aqueous levees) develop in response to the broadening and shallowing of the supply channels (Coleman, 1981). This shallowing leads to channel bifurcation and the eventual waning in flow velocity allowing deposition of massive sandstone with a wave reworked upper surface. However, the shoaling mouthbars in the Marsdenian Pennine Basin is not influenced by basinal wave reworking to the same extent as those of the present day Mississippi River Delta. The combination of muddy drapes and reactivation surfaces also suggests fluctuation in flow velocity, while reversed flow ripple lamination in the foresets implies flow reversal. Flow separation and return fluid flow up the lee slope of the bed form may account for the reversed ripple lamination, however, a tidal influence could also account for the combination and orientation of these sedimentary structures.

Silesian tidally influenced mouthbars

Facies identified by Aitkenhead & Riley (1996) are interpreted as the deposits of a tidally influenced mouthbar, and these appear analogous to the facies described at Kebroyd Bridge. These facies comprise interlaminated mudstone-sandstone facies that lie in a stratigraphic position between the correlative equivalents of the Kinderscoutian-aged R1c3-R1c4 marine bands (Aitkenhead & Riley 1996, Figure 1.3.1). This interpretation is corroborated by harmonic analyses of the data, which infer periodicities that are interpreted as tidally driven (Archer & Kvale, 1997).

Previous studies have suggested similar tidally influenced mouthbars affected sediment deposition in the Silesian of the Pennine Basin and the Midland Valley of Scotland (Broadhurst, 1988; Read, 1992), and contemporaneous examples are seen in the Francis Creek Shale Member (equivalent to Bolsovian of the Pocahontas Basin, Illinois). The Appalachian Basin was connected to marine waters to a greater degree than those in the Pennine Basin, presumably because the basin geometry was entirely open to marine influences (see Section 9.2). The Silesian systems to the west of the Appalachians have very different palaeobotanic and geographic characteristics to those with European affinities, suggesting a faunal and geographic divide between the European-Maritime and Appalachian basins .

Cyclically laminated siltstone-sandstone/ mud rhythmites have been identified within the Upper Limestone Group (Arnsbergian) of the Kincardine Basin of the Midland Valley, Scotland (Read 1992). These rhythmites lie in the lower progradational part of the Fallin Sandstone, and are interpreted as the deposits of an ebb-dominated mouthbar. Aitkenhead & Riley 1996 studies of these facies are based wholly on cored boreholes, where the preservation of these sedimentary features is excellent (Read, 1992).

Rhythmite deposits from the Breathitt Group (equivalent to Langsettian) of the Eastern Kentucky Coal Field, are interpreted as annual, neap-spring and tidally influenced heterolithic channel fill successions (Greb & Archer, 1998). Annual cyclicity is inferred where cycles comprise more than 24 couplets and individual beds contain evidence for more than one depositional event. Lunar monthly cyclicity is inferred if a cycle ranges between 14-24 couplets (this number takes erosion of lamina into account), and each bed has evidence for only one depositional event per bed. The deposits Greb & Archer interpret as a tidally influenced heterolithic channel fill appear similar in lithology, contain similar sedimentary structures and lamina thicknesses to the Marsdenian tidally-influenced mouthbar facies. Additionally both the UK and US examples overlie an erosive unconformity similar to Key Surface R4 underlying the facies at Kebroyd Bridge. Re-examination of the US examples may reveal that they were deposited by an effluent plume in a tidally-influenced sharp-based mouthbar, rather than as a channel fill.

The influence of tidal processes on plume deposition

Tidal processes and dynamics within plume generated deposits are not as well understood as the role of tides in dune form deposition e.g. (Visser, 1980; Allen, 1981). Modulations in discharge generates variations in the amount of fluvial water entering the basin, and the amount of sediment transported (Figure 3.3.1). In basins with marine waters a 'saline wedge' forms during periods of low effluent velocity, when the density difference between plume and basinal waters allows underflow of the saline wedge (Nemec, 1995). During periods of low discharge the saline wedge is a significant feature, propagating up to 15-20 km from the outlet source (Wright & Coleman, 1974), only being displaced from the channel during periods of very high discharge. Within the distributary channel the relatively static nature of the saline wedge inhibits seaward bedload transport, while the mixing of fluvial and marine waters creates turbulence and forces the bedload into suspension (Nemec, 1995). As the plume passes over the shoaling mouthbar and saline wedge it is compressed, undergoes a hydraulic jump, and becomes non-turbulent. The resultant 'buoyant plume' (Wright, 1977) creates a strong density layering, which is elongated by the outward flowing effluent, and carries sand-sized clasts into the mouthbar. Tidal modulations influence the inertial force of the plume, and the position of the saline wedge within the outlet channel (Figure 3.3.2). Flood tides suppress the outflow and fluvial water is banked up within the channels, leading to an increased frequency of overbank bursts in the fluvial system (Andorfer, 1973). Decreasing fluvial outflow forces vertical mixing within the channel, and allows the saline wedge to migrate upstream, increasing the turbulence of the outflow, depositing sediment in the proximal mouthbar and creating a headward shift in grainsize distribution. During the ebb tide, the inertia of the outflowing plume is enhanced, creating a mouthward shift in grainsize distribution. During slack water, the mixing of the saline wedge with the fluvial waters from the distributary channel decreases the rate of effluent outflow, allowing the flocculation of clay particles and their deposition from suspension. The grainsize distribution of a hypothetical tidally modulated low discharge plume (Figure 3.3.2) reveals a repeated thin-thick lamina set, deposited during the ebb and flood cycles and intercalated silty-mudstone lamina.

The calibre of the sediment and type of sediment transport processes within the plume denote the characteristics of the sediment deposited. In Marsdenian facies, the sediment has a bimodal grainsize distribution; comprising very fine- to medium-grained sandstone and siltstone or silty-mudstone, which is often rich in mica-flakes. The muddier component of this sediment was carried in suspension, while the coarser sand-rich portion was transported by tractional or mixed saltation-suspension processes. Both mud and sand components were transported during periods of higher flow regime (i.e. the ebb tide, and to a lesser extent the flood tide). The sand-rich component was rapidly deposited during periods of waning flow, while the mud and silt fractions were deposited during periods when no flow or flow is very reduced (i.e. during tidal slack water).

Within the plume deposits, thick-thin pairs of sandstone lamina were formed by the ebb and flood cycles within a single semi-diurnal tidal cycle. The thicker sandstone lamina were deposited during the ebb stage. During the ebb period flow regime increased as fluvial inertia and out-going tide both flow in a downstream direction. The resultant higher flow regime increased the amount of entrained coarser clastic sediment (typically very-fine to medium grained sandstone), and creating a seaward shift in the distribution of fine to medium grained sandstone. In Marsdenian deposits, the tidal slack water interval is generally represented by a micaeous veneer or a thin mudstone laminae deposited from suspension. The micaeous laminae was reworked by the flood tide producing a thin, often mud-rich, sand lamina that is generally thinner than the sand lamina deposited by ebb currents. At the high-tide slack water a second mudstone or a micaeous lamina was deposited, completing the thick-thin or semi-diurnal tidal cycle. Therefore, any set of mouth-bar deposits influenced by tidal processes may possess sandstone lamina that occur on a repeated periodicity of two (Figure 3.4).

On any point of the earth's surface, two gravitational maxima pass in any 24-hour period, which create two semi-diurnal tides of different magnitudes in the open-ocean. These gravitational maximas are generated independently of one another. The stronger dominant tide is created by the gravitational pull of the moon on the surface of the ocean; whereas the weaker, or subordinate tide, is formed by centripetal forces as the earth rotates and pushes the ocean surface away from the earth (Figure 3.4). The dichotomy in tidal range during a 24-hour period is known as the diurnal inequality. When this is applied to the model of tidally influenced plume deposition, diurnal inequality should generate a periodicity of four, as it occurs over two semi-diurnal periods. Diurnal inequality should produce a thin-thick cyclicity in alternate semi-diurnal tidal units (Figure 3.4); as the dominant tide produces thicker sandstone lamina pairs, and the sub-ordinate tide produces thinner sandstone lamina pairs.

As the moon rotates around the earth, the sun and the moon fall into alignment (or syzygy) or lie perpendicular (or quadrature), on a 14-day periodicity. When the sun-moon are in alignment, the tidal bulge is amplified by up to 30%, increasing tidal ranges and creating stronger, spring, tides. When the sun-moon system are quadrature relative to the earth, the combined gravitational forces are not as strong, and the resultant tidal bulge is smaller, therefore creating weaker neap tides. It may therefore be possible to resolve a cyclical motif associated with lunar procession, when 28 or more semi-diurnal units appear in a stacked succession of tidally influenced plume deposits, and no hiatal or erosive surfaces are present. In theory, when deposition is influenced by a spring-neap cycle, periodicities of 56 lamina, representing 28 semi-diurnal cycles deposited over 14 days, may be expected. However, deviations from the expected cyclicity may relate to increased fluvial discharge. During periods of high discharge, erosion of the substrate removes lamina and generates erosional-breaks within tidal mouthbar units. In summary, the model of plume-influenced deposition should permit the recognition of tidal cyclicity in sandstone lamina on a semi-diurnal (2), a diurnal (4) and half-lunar monthly (56) cyclicity.

When compared to the Fourier transform results from lamina in the Kinderscoutian mouthbar deposits (Archer & Kvale, 1997) both data-sets have similar ranges of harmonic output, and the results of the analysis from section 1 and 3 share broadly similar peaks and troughs (Figure 3.2.5). Comparing the model for tidally influenced plume deposition to the results of the Fourier transform suggests that the periodicities of 2 and 4 signatures represent semi-diurnal tides and unequal (dominant and subordinate) semi-diurnal tides. The presence of thick-thin-thick cycles over 43 and 48 lamina for sections 1 and 3 is suggestive of a spring-neap-spring cyclicity. If this represents a 14 day lunar cycle, the absence of a full set of 56 lamina may be due to reworking and erosion by fluvial (grainflow) processes, which can be demonstrated by the presence of scour features with onlap fill in the section (see Figure 4.16). Identifying cryptic tidal influences within a sequence stratigraphic framework permits the isolation of mechanisms causing tidal amplification. In the Marsdenian Pennine Basin, tidally influenced deposits are located within incised valleys that form palaeo-coastal embayments, which are interpreted as incised valleys (see sequence stratigraphic paper and Section 6.6). Such coastline geometries provided zones where the passing tidal bulge was amplified, allowing a tidal motif to develop within the embayment sedimentary fill.

The integration of the model for a tidally-influenced mouthbar, and the Fourier time-series analysis provides a useful tool for the identification of cryptic tidal signatures within plume deposits. Such deposits have a higher preservation potential than sediments deposited within areas where higher flow regimes could rework or erode part of the sequence. The application of the approaches described to plume deposits in other basins may allow the identification of cryptic tidal signatures.

Other causes of cyclicity within plume deposits

The role of the 'saline-wedge' in controlling the buoyancy of the plume at the river mouth has been discussed above. Mixing between the saline water and the fluvial water caused by density contrasts between the water masses produces a series of internal waves (Wright & Coleman, 1974). The mixing of the water in the internal waves increases the turbulence of the water column (Ewing, 1950; Shand, 1953; Scrutton & Moore, 1953). Mouthbars deposited from plumes influenced by stable internal waves have the potential to deposit rhythmically laminated suspension deposits. The deposition of rhythmic laminated siltstone facies in Marsdenian-aged mouthbars of the Clare Basin is suggested to have deposited by this mechanism (Pulham, 1988). However, such processes are influenced but fluctuations in fluvial discharge, and do not obviously explain the periodicity of 2 and 4 inferred from the Time-series analysis.

3.3.6. Mouthbar top association

This association is rare, but where observed, it is commonly thin (less than 1m), and thins laterally over several kilometres. The association always overlies the proximal mouthbar association, which is commonly reworked by waterlogged muddy and heterolithic palaeosols. Where they occur, allochthonous drift or swamp coal facies are often thin (less than 0.1m), and are overlain by inter mouthbar trough, distal mouthbar or basinal mudstone associations.

Interpretation of mouthbar top association

The common occurrence of the waterlogged muddy and heterolithic palaeosols in this association suggests deposition close to sea/ lake level, implying infill to base-level, emergence and colonisation by land plants. Allochthonous drift or swamp coal facies are often thin (less than 0.1m), representing increased water-depths and reworking of the palaeosol.

3.3.7. Bayhead prodelta association

Offshore 'marine band' mudstone and offshore black mudstone facies are often absent or thin (less than 2m in thickness) in this association. Ammonoids are rarely preserved, and other marine faunas including *Lingula* and *Dunbarella,* commonly form high-density, low diversity populations. *Zoophycos* bioturbation commonly forms indurated horizons superimposed on the upper surface of palaeosol facies or the multi-storey channel facies association.

Interpretation of bayhead prodelta association

This facies association developed in restricted marine conditions, basinward of the main area of bayhead delta deposition. In this area connections to fully marine conditions were limited, preventing colonisation by nektonic species, including ammonoids, and marine fauna such as bivalve and brachiopod larvae stages.

3.3.8. Bayhead delta front association

The shallow water open marine-influenced bay facies in this association are overlain by, and laterally adjacent to, the tidally influenced sand- and silt-rich mouthbar facies. The base of the bayhead delta front association is either gradational with the underlying bayhead prodelta association, or may be sharp based if underlain by key surface R4 (see Key Surface Chapter, Enclosure 4) which represents the unconformity underlying bayhead delta front association. The top of the bayhead delta front association is often removed by erosion and/ or colonised by vegetation (see Figures 3.6 & 4.10.2).

Heterolithic lithologies, in the tidally influenced silt-rich mouthbar contain abundant flow reversal ripple cross-laminations with muddy drapes and subtle erosive reactivation surfaces, while marine ichnofaunas commonly include *Planolites*, *Curvolithus*, *Rosselia* and *Chondrites*.

At Middleton Towers (SD409588) a single bayhead delta cycle comprising a coarsening upwards wave reworked mouth-bar, with lower shoreface facies (above fairweather wave base; see Figure 4.6) overlies an offshore black marine mudstone facies. Within the upper part of the cycle wave-ravinement surfaces and gravel-lag horizons are common. The upper part of the bayhead association has a diverse marine ichnofauna; including *Treptichnus*, *Chondrites*, *Taenidium*, *Phycodes*, *Planolites* (*montanus* and *annulatus*) *Rhizocorallium* (*jenense* and *irregulare*) which are typical of lower shoreface ichnofabrics (Pemberton *et al.* 1992).

Interpretation of bayhead delta front association

The tidally influenced silt-rich mouthbar facies represents deposition in shallow-water affected by reworking by wave and tide processes, while the lateral continuity of lamina and bedding implies unconfined deposition. The tidally influenced silt-rich mouthbar facies suggests influence by marine waters (see Figure 2.22), while the ichnofaunas also imply a brackish-marine environment.

The bayhead delta front association is the most variable of all the associations described in this study. Deposition is influenced by fluvial, tidal and basinal wave processes, implying a wide variety of sub-environments exist within this association. The shallow water open marine bay facies is similar to the black mudstone and rhythmite mudstone (Haszeldine, 1984) from the Duckmantian of north-east England. In these Westphalian examples, organic-rich black mudstones are interpreted as deposits from a non-marine sediment-starved water column. Haszeldine (1984) describes alternating couplets of pale and dark mudstone, interpreted as deposited from fine-grained turbid underflows. Within this succession flaser to lenticular current ripple laminated siltstone laminae contain single sets of current ripple laminations, which are inferred as the deposits of waning low-density turbidite currents. Unlike the bayhead delta front association described here, Haszeldine suggested that the paucity of marine bivalve faunas, the presence of siderite and a lack of pyrite (suggesting non-reducing conditions) indicated a lacustrine environment.

Haszeldine also described an upward fining, very fine-grained sandstone with bivalve resting traces on the base of sandstone beds, and interpreted this as the deposits of turbid underflows sourced from a nearby distributary feeder channels. Comparison between the tidally influenced sand-rich mouthbar facies in this study, and the rhythmically inter-bedded lithologies, suggests clear differences in the styles of deposition. The facies described by Haszeldine suggest a tractionally deposited supply, which was influenced by tidal processes during deposition. Reviews of facies in the Westphalian Pennine Basin (Guion & Fielding, 1988; Guion *et al.,* 1995), describe similar facies that are attributed to deposition in freshwater lacustrine conditions. The buoyancy dominated heterolithic mouthbar facies is simiar to facies within the Bullion Rock (Langsettian) near Wigan (Broadhurst *et al.* 1980; Figure 3.5). Within the Bullion Rock, sets of regularly interbedded sandstone-siltstone lamina are interpreted as the infill of a shallow lacustrine bay (Figure 3.5.1). Like the tidally-influenced bayhead facies in this study, the Bullion Rock was probably deposited in a submerged to semi-emergent environment. Depending on how open marine connections were to this environment, tides may or may not have influenced deposition. Broadhurst *et al.* (1980) suggests the rhythmic bedding within the Bullion Rock is the product of variable river inputs generated by seasonal monsoons, although it is intriguing to compare the facies described by Broadhurst, with the similar tidally influenced facies described in this study (Figure 3.6).

The wave reworked bayhead delta front association at Middleton Towers (SD409588) shares characteristics with wave reworked facies in other intervals of the Namurian. For example, the lower shoreface facies described in this study is similar to facies within the

Harthope Ganister (Arnsbergian) of County Durham (Percival, 1992). Extensive quarried outcrop allowed Percival to interpret 'thinly bedded facies' as the deposits of a back-ridge barrier island environment. The limited lateral extent of exposures within the Marsdenian does not allow the lateral relationship of the shoreface deposits to be assessed. Facies in the Great Yoredale Cyclothem of the Pendleian, comprise interbedded heterolithic mudstone, siltstone and wave-rippled sandstones (Ainsworth & Crowley, 1994). These facies appear similar to those at Middleton Towers, and are inferred as the product of deposition within the lower shoreface transition zone. Additionally, interbedded wavy laminated sandstone facies from the Fossil Sandstone (Arnsbergian) on the Askrigg block are suggested as deposited between fair-weather and storm wave base on the coastal side of a transgressive barrier island complex (Brenner & Martinsen, 1990).

The gravel rich horizon at Middleton Towers appears similar to the 'Amroth pebble bed' described by Hampson (1998). Lying beneath the *Cancelloceras cancellatum* marine band in the Yeadonian of South-Wales, the pebble bed is interpreted as a storm lag deposit, produced by the winnowing of the sea floor substrate during periods of frequent wave activity. The proximity of the gravel-lag and bioturbated horizon at Middleton Towers (Figure 3.7.1) implies that sea-level modulated rapidly during deposition (Figure 3.7.2). During rising relative sea-level Plint (1988) suggests a gravel lag is generated by the transgressive winnowing of shoreface sediments. In the case of the example at Middleton Towers, this is reworked and colonized by infaunal ichnofaunas leading to the generation of a high-density, high diversity ichnohorizon (Key Horizon T3; Figure 3.7.1).

Bayhead delta top association

This association is dominated by coarse-grained sandstone, with ripple and dune bedforms in both channelized fluvial sandstone and tidally influenced cross-bedded sandstone facies. Where exposed in quarry sections, channel margins are linear, and individual channel fills do not contain evidence for inclined point bar deposits. Overbank deposits in the bayhead delta association are both located adjacent to, and overlying fluvial channel belts.

Waterlogged muddy and heterolithic palaeosol facies commonly overlie the delta front mouthbar association, and are themselves overlain by the drift coal/ swamp facies. The well-preserved example of a *Lycophyte* stump in heterolithic palaeosol facies at Harper Clough Delph, Blackburn (SD716317) provides a good insight into the form of the tree roots palaeobotanical species within the Marsdenian (see Figure 2.28.1).

Interpretation of bayhead delta top association

This facies association represents the emergent area of the bayhead delta lying behind the shoreline. The lack of point bar deposits suggests that channels within this association are straight and not sinuous. High discharge fluvial systems possess a high inertia and therefore have a greater potential to erode, avulse and transport material as bedload. Straight channels commonly feed present day bayhead delta front areas, suggesting that bayhead mouthbars received a high discharge.

Channel margins in Namurian bedload dominated fluvial systems are frequently straight (McCabe, 1977; Hampson, 1997), suggesting high discharge fluvial supply systems were dominant throughout the Namurian (Orton & Reading 1993; see Section 3.3.12). Map analysis from the present day channel complexes and sub-aerial exposed levees of the Atchafalaya and Cubits Gap deltas clearly illustrates the straight geometry of the bifurcating mouth-bar feeder distributary channels (Van Heerden & Roberts 1988; see Figure 5.4). The lack of in-channel fine-grained deposits, or lateral accretion surfaces suggests that channels were predominantly straight and did not meander.

These deposits are similar to those described in the lower part of the Yoredale Cyclothem (Upper Limestone, E1), Weardale, North Yorkshire (Elliott, 1975). The facies are the product of overbank flooding, levee and minor crevasse channels generation.

Intervals with closely spaced data-sets such as the Langsettian of the East Midlands Coalfield (Guion, 1984; Guion *et al.,* 1995), permit the three-dimensional reconstruction of crevasse splay and overbank (levee) deposits. These studies demonstrated that overbank environments lay parallel to channel margins, and that the overbank area was proportional to the width and depth of the channel. Crevasse splay deposits form lobate features up to 1 km in width, with individual splays sandstone beds up to 1m in thickness. Holocene-Recent examples from the Mississippi reveal natural levee and interdistributary trough facies have similar dimensions to Marsdenian overbank facies (Tye & Coleman, 1989a; Tye & Coleman, 1989b; Tye & Coleman, 1989c). Although the Langsettian and Recent examples are used as analogues for this facies in the Marsdenian, the lack of continuous outcrop makes it impossible to trace this facies to the extent of those studies highlighted above. Comparisons of spatial and stratigraphic parameters are made in Table 4.

Vegetation and roots in the palaeosol facies commonly destroy or disrupt sedimentary structures in the bayhead delta front sediments. As they develop at, or slightly above base level, wetland environments are important indicators of environmental change. Delta-top often subside, or are inundated by fresh or brackish water to produce a shallow low-lying swamp

(McCabe 1984). The study of such present day environments is critical in addressing issues such as land-management and conservation. This has produced a volume of data (Coleman, 1988; Roberts & Coleman, 1996; Coleman *et al.,* 1998a) that can be applied as analogues for ancient systems (Coleman, 1981).

The Marsdenian drift coal/ swamp facies is allochthonous, suggesting it is more likely to have been deposited by a floating swamp, rather than an insitu vegetated area. Present day environments comparable to those responsible for the deposition of drift coal/ swamp facies are located within the Mahakam Delta, Borneo, where interdistributary headland areas comprise shoreline parallel ridges of woody debris up to 2m thick. Similar floating swamps are also found in the Okefenokee Swamp of Florida, where following submergence, peat rafts within the lake are accreted to the lake margins (McCabe, 1984). In the Mississippi, extensive floating mats of organic material are reworked by increased water-table levels in swamp environments due to the storm surges associated with hurricane events (Coleman *et al.,* 1998b).

The *Lycophyte* tree-stump observed at Harper Delph Clough (SD716317; see Figure 2.28.1) appears to have a radial propagating root system, which may acted to increase structural support in water-logged or swamp conditions (see Section 2.2.24). *Lycophyte* species developing crowns of this dimension (up to 1.5m across) are likely to have grown in waterlogged soils, where the mechanical strength of the substrate was not great (DiMichele & Phillips, 1994). Similar *Lycophyte* stumps documented in the Bolsovian of Palencia, Spain appear to have been inundated by flooding events, suggesting the presence of water logged conditions (Wagner, 1995).

The geometry and form of the root stump at Harper Delph Clough, is similar to many modern day mangal root systems. Although no mangrove facies have recognised in deposits older than the Cretaceous, it is possible that other plant or tree-fern species may have inhabited this environmental niche previously. As the tree-stump at Harper Clough Delph formed at, or just above sea-level, within the bayhead delta depositional system, it is possible that this species may have been exposed to saline conditions. Such environments were common on bayhead delta environments, where either marine transgressions, or subsidence due to compaction may have led to the development of brackish or marine conditions.

Modern mangrove communities comprise many species that possess similar root morphologies and features that increase tolerance to salt (Plaziat, 1995). These features include supporting roots that protrude above the surface of the water, pneumataphores (aerial roots) that enable the plant to take in oxygen during low tide and salt excluding membranes which enable them to utilise sea-water without undue salt build-up.

Such morphological adaptations have not been recognised within Carboniferous palaeobotanical samples. Carboniferous species with mangrove-like characteristics, such as prop-roots in *Cordaites* (Cridland, 1964), have been interpreted as not necessarily indicative of water-logged environments (Plaziat, 1995).

3.3.10. Interfluve association

This associaton is generally between 1-5m thick, and overlies fluvial/ tidal channel and abandonment associations; or bayhead mouthbar front associations in rare examples (e.g. Middleton Towers SD409588). The leached sand rich 'ganister' palaeosol is common, although the waterlogged mud-rich palaeosol facies do occur. The base of the palaeosol facies completely reworks the underlying substrate. The upper contact of the association is often sharp and overlain by the marine offshore facies association. Overbank flood deposits/ Crevasse splay facies are rare, and where observed, reworked by pedogenic activity.

Interpretation of interfluve association

The leaching of clays through the substrate forces the concentration of residual quartz and feldspar in an albic horizon within the sand-rich 'ganister' facies (Percival, 1986). The presence of permeable sandy substrates is probably an important control in the formation of ganister horizons. Other mechanisms may increase the rate of leaching within the palaeosol, including a fall in base-level forcing a lowering in water-table levels, or (the rare) aggradation on the interfluve flood plain, leading to thickening of the substrate. It is likely that this facies developed in a perennially well-drained environment, as there is little accumulation of organic material within the palaeosol. However, leaching is significant in this facies, and therefore water must have been available via precipitation and/ or overbank flooding. Similar leached sand-rich facies are described in the Rough Rock of the Pennine Basin (Hampson *et al.,* 1996) and correlated equivalents in South Wales (Hampson, 1998). This facies is also described in the Namurian succession from the Clare Basin in Ireland, where in-situ *Stigmaria* and *Calamites* rooting, associated with rhizoconcretions and polygonal desiccation cracks suggested freely drained conditions developed in a humid-tropical climate (Davies & Elliott, 1996).

During transgression events, raised water-tables overprinted leached palaeosols with gleyed (waterlogged) characteristics (Aitken & Flint, 1994). Some Marsdenian palaeosols also appear to have formed under initially leached conditions, and are overprinted by later gleyed conditions (see Section 4.3.2).

The presence of semi-dry conditions in Namurian interfluve areas is corroborated by the presence of charcoal deposits, suggesting the occurrence of wild fires (Scott & Jones, 1994). Isolated horizons of fusain have been observed in some of the thicker Marsdenian coals, and these may represent charcoal beds. Coals are generally thin (less than 10cm thick) although the thickness of the original organic matter is estimated to be approximately thirteen times greater than that of the preserved coal (Elliott, 1965; McCabe, 1984), suggesting a significant build-up of organic material (estimated as a maximum of 1.5m). Comparisons between previously described facies (Guion & Fielding, 1988; Guion *et al.,* 1995; Jones *et al.,* 1995), and the characteristics and setting of this facies are suggestive of 'Raised Swamps' that are elevated above the adjacent main channel belts.

3.3.11. Channel abandonment association

This association overlies fluvial and tidal channel associations and is variable in both lithology and thickness. The basal contact of this association is sharp and non-erosive, suggesting an abandonment and/ or flooding surface. Palaeosol facies occur as thin horizons (less than 0.3m), and waterlogged muddy or heterolithic palaeosols often rework underlying sediments and are commonly overlain by thin allochthonous drift/ swamp coal facies (less than 0.2 m), that comprise a high proportion of suspension deposited black mudstone.

Interpretation of channel abandonment association

This association often overlies the fluvial and tidal channel associations, suggesting deposition after channel abandonment. The plugging of the channel by mudstones and siltstone deposited from suspension implies that this association represents a period of channel abandonment due to autocyclic decreases in flow velocity. The occurrence of thin palaeosols suggests formation over short periods of time, implying transient pedogenic activity occurred during only brief periods of emergence.

3.3.12. Channelized sandstone association

Erosive concave-upward channel bases often have up to 10m of demonstrable relief within this association. Each channel base often contains cosets of either channelized cross-bedded sandstone facies or fluvial barform facies, suggesting erosion and infill by sediments transported under similar flow regimes. The lateral continuity of these channels is difficult to observe, due to the lack of laterally continuous exposure, but laterally extensive quarries (such as Fletcher Bank Quarry SD805164) imply channel widths in excess of 800m (Okolo, 1983).

Interpretation of channelized sandstone association

The lack of evidence for lateral accretion surfaces at the channel base and margins, along with the dominantly unidirectional sediment transport direction combined with the paucity of abandonment deposits suggests this facies represents a fluvial system with straight margins. The system probably comprised a series of braided channels, where avulsion was the main mechanism of channel migration, (Hampson, 1997; Hampson *et al.,* 1999).

Muddy paired drapes, reactivation surfaces and bundling suggest a tidal influence during deposition of this association. Alternative processes could have also caused these structures, such as the pulsing in fluvial discharge during sediment transport, and suspension deposition fines during low discharge stage.

Carboniferous rivers discharged large volumes of fluvial water (Collinson, 1988), that would have suppressed marine influence within distributary channels (Holdsworth & Collinson, 1988) and therefore, the range of the tidal flood waters. Ebb dominated tidal systems potentially produced less re-activation surfaces than tidal systems with similar ebb-flood current strengths, as reworking by the flood current is suppressed by the fluvial dominated current (Visser, 1980). This produces an asymmetric tide, in which the flood tide is subordinate and incapable of eroding into the mud-drapes or produce reactivation surfaces on dune foresets or toesets (Demowbray & Visser, 1984).

In the in-channel bar facies at Moselden Heights Quarry (SE043164), the density of the mud-drapes appears to decrease from the base towards the top of barform set 1, while the base of barform set 4 also appears to possess common mica-mud drapes (see Figure 2.24). This may imply a decrease in tidal influence during down-channel progradation of individual bars, which could be associated with the increase in fluvial inertia as the fluvial regime approaches. During this event an increase in fluvial flow regime would have suppressed the tidal signature, and eroded tidally influenced sedimentary structures. The suppression of flood tides by high discharge fluvial currents explains why Marsdenian tidal deposits are subtle in character, and why tidally influenced cross-bedded sandstones have not been commonly identified in the Namurian of the UK.

Soft sediment deformation is observed in channel bar facies at several localities in the area around Bingley Road Quarry (SE053382; see Figure 2.26), Park Wood Quarries (SE068407) and isolated exposures on Rombalds Moor (see Figure 8.4). Strike-slip movement is documented on the Morley-Campsell Fault Zone during the Langsettian and Duckmantian (Giles, 1989). This fault zone was linked to the Craven Fault System, and a tectonic influence has been suggested for the distribution of the Keighley Bluestones during the Marsdenian (Waters *et al.* 1996a; see Section 9.2.2) . Minor fault-splays in this area could provide a mechanism for the initiation of soft sediment deformation (see Section 9.2.2.1). However, other mechanisms may be also responsible for these features. Load structures up to 3 meters thick are described in the delta front deposits underlying forced regressive mouth-bars in the Marsdenian of South-Wales (Hampson, 1998). These may have a similar mode of formation to the deformation structures observed in the Bingley Road Quarry, and suggested to have formed during a decrease in relative sea-level. Similar soft-sediment deformation in mouth-bar deposits of the Mesaverde Group from the Campanian of Wyoming formed during rapid base-level fall (Fitzsimmons & Johnson, 2000). Under these conditions rapid loading associated with the swift progradation of a clastic wedge may have created ideal conditions for massive dewatering/ collapse.

Carboniferous 'giant cross beds'

There has been much debate regarding the origin of 'giant cross beds' within the Kinderscoutian of the Pennine Basin (Collinson, 1968a; Collinson, 1968b; McCabe, 1977; Hampson, 1997). Examples from the Marsdenian are smaller in scale, but similar in geometry and lithology. The following brief review clarifies the arguments and theories that have previously been debated.

The original interpretation by Collinson (1968b) suggested that Kinderscoutian giant cross-bedding, is commonly associated with and overlain by cross-bedded sandstone facies, represents a prograding delta front, delta top and feeder system. He suggested that this morphology is similar to Gilbert's (1883) delta classification, the large scale cross bedding being equivalent to the 'foreset' and overlain by medium scale cross-bedded sandstone equivalent to the 'topset'. Collinson implied rapid sediment supply is required for progradation of this 'deltaic sedimentation unit', that the 'unit' prograded into standing water with depths slightly greater than the thickness of the set, and the subsequent deposition of the cross bedded topset beds requires a slowly rising base level.

The basal surface of giant cross bed sets often overlies an unconformable erosive surface (McCabe, 1975; McCabe, 1977), suggesting deposition occurred in channels. The thinning of giant cross beds onto channel margins and the occurrence of undulatory-bedded sandstone facies is suggested to represent the current lineations or 'sand-ribbon', formed by large obstructions to flow that created downstream corkscrew vortices (McCabe, 1977). In the case of three-dimensional dune-forms, the current can generate sand-ribbons downstream of the dune, modifying the dune or barform into a swept lunate to catenary form. McCabe also interpreted giant-cross beds as in-channel bank attached fluvial barforms. Recently, two distinct types of giant foreset have been recognized (Hampson, 1997); i) those with a simple clinoform geometry, and ii), giant-cross beds with erosive surfaces bounding individual cosets, creating a complex clinoform geometry.

i) Giant cross beds with simple geometries represent the progradation of a delta front (Collinson, 1968b). A fluvial origin for simple giant cross beds seems unlikely, as scouring is not observed at the base of giant foresets. Additionally, a downdip decrease in giant cross bed thickness would be expected, rather than the increase in cosets thickness noted by (Collinson, 1968b). Fluvial bars typically possess complex internal stratification, while internal structures are rarely preserved in the geometrically simple Kinderscoutian examples. However, erosive surfaces underlying the base of some giant foresets suggests an early stage of erosion during the onset of delta front progradation. In this case, autocyclic increases in basin supply pathways, or allocyclic fluctuations in supply system discharge may be responsible for such erosive features (Hampson, 1997).

ii) Hampson (1997) inferred that geometrically complex clinoforms with internally erosive set bounding surfaces are the product of in-channel fluvial bedform migration. Clinoform surfaces often reveal a change in orientation across erosive surfaces, implying rapid lateral changes in channel flow direction. This type of giant cross bedding has been inferred representative of a within-channel fluvial barform, or a braided river channel confluence (Hampson, 1997).

Two comparable facies have been identified in the Westphalian-aged Seaton Sluice Sandstone, near Newcastle-upon-Tyne (Haszeldine, 1983). Simple, tabular sets with planar foresets (facies S1) are inferred as deposited by in channel, straight crested sandsheets that are arranged tiers that descend in a down current direction. Haszeldine implied that sand was transported along the bedform top-sets by tractional processes, and was avalanched down the lee face. The top-sets are often preserved, and comprise cross-bedded sandstone suggesting ripple and dune-forms were present within the channel. The large trough cross-bedded sandstone facies (S2) described by Haszeldine, has concave-up forests that occur in association with facies S1. Facies S2 is interpreted as representing the coeval development of dunes generated during increased turbulence within the channel. Both facies possess concordant palaeocurrent flow directions, implying the down-stream migratation of lateral (bank-attached) or a medial (non-bank attached) bars in a straight sided channel.

In-channel bar forms with both simple and complex geometries occur within the Marsdenian. The association of the cross-bedded sandstone and barform facies in the Marsdenian is suggestive of the 'deltaic sedimentation unit' first described by (Collinson, 1968b). Subsequent analysis by Hampson (1997) recognised that these units were not the product of a prograding delta, but the deposits of a large-scale channel (McCabe, 1977). Hampson suggested that giant cross beds with simple geometries represent delta front deposition (Collinson's 'deltaic sedimentation unit') and those with complex geometries represent in channel bar forms. However, the distinction between the two types described by Hampson (1997) is not as clear in the Marsdenian. This is because outcrops are often of poor lateral extent, and barform facies may be confused with the bedload dominated mouthbar facies in the delta front mouthbar association overlain by a distributary channel (see Figure 5.7).

Where laterally continuous exposures are observed in the Marsdenian, bars occur within channels, and are overlain by cross-bedded sandstone facies and the channel abandonment association. This association suggests initial channel incision was followed by the progradation of a barform/ barforms in a basinward direction, which infilled the channel. Avulsion, forced by autocyclic processes is suggested as the mechanism of channel axis lateral migration. Incision and erosion of underlying channel systems forces the generation of new channels, which are subsequently infilled and produce the stacked fluvial channel association.

3.3.13. Laterally accreting channelized sandstone association

This association has been recognised at Fletcher Bank Quarry (SD805164) and Leicester Mills Quarry (SD619164). The basal contact erosively overlies channel abandonment associations, but the base is often difficult to define when in contact with the channel abandonment association. This is because the basal erosive contact is often bedding parallel, and overlain by the carbonaceous silty mudstone lithology of the downlapping low-angle lateral accretion surfaces. However, the geometry of the lateral accretion surfaces is well defined when observed from a distance (see Figure 2.27 & Figure 4.9). In the upper parts of this association, thin beds of the channelized fluvial sandstone facies are common. These often thicken to reveal channel geometries with estimated depths up to 4m and widths of 15m (see Figure 4.9). Thin channel abandonment facies or waterlogged muddy palaeosol facies commonly cap this association suggesting channels fill to emergence and point bar tops are colonised by land plants.

Interpretation of laterally accreting channelized sandstone association

Meandering fluvial channel sandstones with lateral accretion surfaces are not recognised as a common channel fill geometry in the Namurian of the Pennine Basin (Davies *et al.,* 1999). This is probably because the high fluvial discharges and the coarser bedload grainsize were not conducive to the development of meandering fluvial systems during the Namurian (Orton & Reading, 1993).

Interestingly, the succession at Fletcher Bank Quarry (SD805164), comprises two stacked channels with erosvie bases (Okolo, 1983). Okolo (1983) interpreted these as representing a period of channel 'side fill'. Okolo suggested that the channel containing this facies was filled during a period of channel abandonment, when tractional currents moved in a direction oblique to the channel margins, allowing the suspension deposition of interbedded silts and mudstone. Palaeocurrent data highlights the perpendicular orientation difference between in-channel cross-bedding (palaeoflow to the south-east) and inclined bedsets (dipping to the north-east). Comparison with Okolo's data implies that the inclined bedsets are equivalent to the lateral accretion surfaces identified in this study.

Meandering fluvial channels are well-documented in Westphalian-aged deposits. During the Westphalian extensive flat delta top conditions became established (Guion & Fielding, 1988; Guion *et al.,* 1995). The paucity of meandering channel in the Namurian suggests discharge was greater, possibly due to the fluvial channels in the delta front deposits of the Namurian possessing steeper river profile gradients than those on the Westphalian delta top. Westphalian meandering channels described in the literature are both mud and sand dominated, suggesting that both grainsize and flow velocity controlled the type and geometry of channel fill. For example, 'epsilon stratified silts' in the Langsettian/ Duckmantian of Northumberland are interpreted as the product of a laterally migrating channel that transported silty-muds (Haszeldine, 1984). A laterally accreting fluvial/ crevasse channel has also been identified in the Langsettian-aged Greenmoor Rock (Waters *et al.,* 1996a). Lateral accretion surfaces occur in sets up to 2m in thickness, suggesting channels are of smaller proportion than those observed in the Marsdenian.

Potential mis-interpretation of 'meandering fluvial systems'

Interbedded ripple cross-laminated sandstone-siltstones with large-scale inclined bedding are described in the mouthbar facies of the Yeadonian-aged upper Haslingden Flags of south-east Lancashire (Collinson & Banks, 1975). The Upper Haslingden Flags comprises a 27m thick mouthbar sandstone, with inclined bedding which dips up to 14°. Collinson & Banks suggest the inclined bedding represents the point bar accretion surfaces of a laterally migrating channel, implying the thickness of the epsilon cross bedding sets (up to 27 m) represents the approximate channel depth during deposition.

However, mouthbar facies are deposited close to the mouthbar front, and the gradient of the mouthbar front was probably a significant control on the velocity of the flow within the feeder channels. Proximity to the mouthbar front is likely to raise flow velocities within the channel, increasing the bedload and straightening the channel geometry. The facies and the sloping bed geometries described by Collinson & Banks in the Upper Haslingden Flags are similar to the steeper mouthbar fronts described within the delta front mouthbar association (see Figure 2.11). In light of interpretations from Marsdenian delta front mouthbars, reinterpretation of the Haslingden Flags may reveal that they are also the product of a prograding mouthbar, rather than a meandering channel complex.

Chapter 4: Key Stratal Surfaces and Horizons

4.1. Introduction

This study uses transgressive surfaces (representing deepening events), regressive surfaces (representing shallowing events) and palaeosol horizons (representing periods of emergence) to correlate depositional successions. Key surfaces represent laterally continuous horizons, which may overlie a variety of facies. Documentation and analysis of the variability, hierarchy and preservation potential ensures accurate correlation of the key surfaces and horizons. This chapter details the GENERIC characteristics of each key surface or horizon. The index figures T1-T5 and R1-R4 refer to a TYPE of either transgressive (T) surface or regressive surface (R), and although examples of each surface are shown, should not be confused with a specific surface.

4.2. Transgressive horizons (T1, T2, T3, T4 and T5)

Transgressive horizons form distinctive stratal units, which often contain faunas and ichnofacies that suggest a marine signature during deposition. Typically facies associated with the flooding surface are finer grained than underlying facies, and often comprise mudstones and siltstones. Five types of flooding surface are described; each is associated with predictable facies and has its own set of characteristics.

4.2.1. Type T1 key horizons: Marine band

In the Namurian Pennine Basin, ‘marine bands’ are laterally extensive, readily identified horizons that enable high confidence correlations and provide the main correlation framework (Bisat 1928; Ramsbottom 1977; Ramsbottom *et al.* 1978).

Marine bands are offshore marine mudstone facies, and comprise up to 3m of black, well-laminated mudstone with diverse marine faunas. The offshore marine mudstone facies commonly overlies the fluvial, bayhead delta and delta-front the delta-top facies (see Section 5.1.1). Three distinct types of horizon occur within each marine band; i) an ammonoid condensed horizon, ii) a *Dunbarella* condensed horizon and, iii) a faunal free condensed horizon.

i) T1a: Ammonoid condensed horizon; Ammonoids are often disseminated throughout the offshore marine mudstone facies, but are concentrated in one or more ammonoid rich horizons, which are typically less than 0.15m thick. In rare examples, similar horizons are observed in thicknesses up to 15m (Aitkenhead & Riley 1996). Marine bands contain an index thick-shelled ammonoid species (*Bilinguites gracilis, Bilinguites bilinguis, Bilinguites eometabilinguis, Bilinguites metabilinguis*; Figure 4.1), and other thin-shelled ammonoid species (*Anthracoceras, Dimorphoceras, Hudsonoceras*) which are preserved as flattened impressions or three-dimensional casts. Ammonoid-rich horizons often contain disseminated calcareous and sideritic concretions (up to 0.1m in diameter), that weather to an orange clay, which is readily identified in the field (Figure 4.1.1 to 4.1.4).

ii) T1b: Dunbarella condensed horizonS; The laminae above the basal contact of the offshore marine mudstone facies commonly contain the pecten-like bivalve *Dunbarella*, with abundant well-preserved *Cordaites* leaf and stem debris. *Dunbarella* commonly cover whole laminae, and are often preserved in life-position (Figure 4.1.7). Dunbarella-rich laminae are usually separated from overlying ammonoid-rich layers by several laminae containing rare ammonoids, *Dunbarella* and *Caneyella*.

iii) T1c: Non-fossiliferous black mudstones; Where the basal contact of the offshore marine mudstone does not contain *Dunbarella*, it often comprises a black or dark grey mudstone, which is often massive or has poorly defined laminae. This horizon is often less than 0.3m thick, and forms an indurated feature in weathered outcrops. A similar facies described and analysed by Wignall & Maynard (1996) is characterised by high authigenic uranium and total organic carbon values relative to adjacent mudstone (Figure 4.3).

Interpretation of type T1 horizons

The pervasive black colour of the offshore marine mudstone facies implies that anoxic conditions prevailed in the water column and/ or sediment during deposition. The presence of ammonoids, brachiopods and marine bivalves in high-density/ low-diversity faunas infers deposition in relatively hostile conditions. The T1a horizon contains the maximum abundance of ammonoids within the marine band, and is inferred as a semi-isochronous unit that can be correlated across the basin (Bisat, 1928; Ramsbottom, 1977; Ramsbottom *et al.,* 1978).

In deeper basin areas the substrate-water interface is often poorly oxygenated, due to the lack of bottom-currents, and stratification in the water-column which is enhanced by wind driven advection cells (Demaison & Moore, 1980; Wignall & Maynard, 1993). The paucity of oxygen flux to the basin floor may account for the lack of benthic fauna, and the black anoxic appearance of the mudstones.

The high-density/ low-diversity marine suggests that marine conditions influenced the environment in which this horizon was deposited (Calver, 1968). It seems likely that nutrients were freely available in the water column for the nektonic organisms, and are therefore unlikely to have restricted the development of macrofauna. Ammonoid-rich horizons are commonly thin, suggesting suitable conditions for the colonisation of marine fauna only last for brief periods. In a semi-enclosed basin, such as the Pennine Basin, the discharge of large amounts of fluvial water may have rapidly diluted the basinal salinity (Holdsworth & Collinson, 1988).

Profiles through Namurian marine bands reveals cyclic faunal changes in the areas of Ashover in Derbyshire (Figure 4.2), but these are not observed in proximal clastic rich areas (Ramsbottom *et al.,* 1962). These faunal changes are interpreted as representing periods of faunal change due to salinity fluctuation, and do not take into account the effect of clastic input or basinal desalination, which effects the thickness of the marine bands and the depositional environment (Holdsworth & Collinson, 1988).

The pecten-like morphology of *Dunbarella*, and their concentration at the base of the offshore marine mudstone facies suggests that these epifaunal bivalves lived in relatively quiet water conditions following initial marine flooding. The high-density fauna suggests a restricted/ stressed environment, where either soft substrate or low amounts of oxygen limit the faunal colonisation (Wignall, 1993b). Such conditions may be expected during the initial flooding of the delta top, when organic debris associated with deposition on the delta top is mixed with incoming marine water. Rotting and bacterial breakdown of the plant debris the water column may cause an increase in acidity during the initial flooding, and therefore decrease the amount of free oxygen.

The absence of fauna within the T1c horizon suggests that either the environment of deposition was unsuitable for the colonisation of marine fauna, or any fauna present during deposition was not preserved. Within the lower Marsdenian of the Pennine Basin (R2a1-R2b2) the 'Denshaw Horizon' (type locality; Haigh Gutter SD001121) appears to have formed under analogous anoxic conditions(Wignall & Maynard, 1996). The *Dunbarella* and non-fossiliferous black mudstone horizons (T1b and T1c) occur below the ammonoid-rich horizon (T1a) at a number of marine band outcrops, including a section at Pule Hill (Figure 4.3). This depositional motif is similar to the basal transgressive black shale (the Owd Betts marine band; Figure 4.4) that lies below the *Cancelloceras cumbriense* (G1b1) marine band (Wignall & Maynard, 1996).

Analogues of Marine Band Key horizons (T1); Namurian marine bands have been proven to be laterally continuous surfaces that can be traced within the British Isles, across the North Sea (Ramsbottom *et al.,* 1978), and into Poland and Germany (Patteisky, 1959; Korejwo, 1969). This was used by Ramsbottom *et al.* (1978) to systematically correlate marine bands and their sand-prone intercalated lithostratigraphic units across the British Isles. The initial faunal study of marine bands (Bisat, 1924) continues to be the standard reference for ammonoids identification in the Namurian of northern England. The ammonoid stratigraphy referred to here is based on the re-designated index scheme of Ashton (1974) , Holdsworth & Collinson (1988); and Riley *et al.* (1993), rather than the original designation of Bisat (1924).

The diversity of Namurian ammonoid species is fortuitous, and suggests rapid evolution at a rate of nearly one index species per cyclothem (Ramsbottom *et al.,* 1978; Riley *et al.,* 1993). The estimated time interval between marine bands is estimated to be approximately 65,000 years (Riley *et al.,* 1993). This figure was calculated by averaging the number of marine bands between dates attained from Pb-Ur radiometric dating of zircons in K-Bentontite horizons within the Arnsbergian of the Pennine Basin and Westphalian C tonsteins from Germany (Lippolt *et al.,* 1984).

There are problems with using the early Marsdenian R2b1, R2b2 and R2b3 *(Bilinguites bilinguis)* marine bands as correlatable surfaces. The classification of the Marsdenian marine bands (Holdsworth & Collinson, 1988; Riley *et al.,* 1993) highlights the repetition of the *Bilinguites bilinguis* species in marine bands R2b1, R2b2 and R2b3. This creates a problem when using ammonoids solely to classify marine bands as the repetition of the *Bilinguites bilinguis* species makes the identification of individual marine bands difficult. Differentiation between the R2b1 (*Bilinguites bilinguis* early mut. β) and R2b2 (*Bilinguites bilinguis* late mut. β) marine bands in previous studies relied either on detailed biostratigraphic analysis, or the identification of all three marine bands in a single section and the assignation of index codes (R2b1-R2b3) in stratigraphic order. Further complexity is created if the ammonoid-rich marine bands are traced into delta front and delta top areas, as they lose their ammonoid faunas.

4.2.2. Type T2 key horizons: Lingula-Planolites horizons

Lingula and *Planolites* are present in massive indurated black mudstone horizons less than 0.05m thick. *Lingula* are often preserved as carbonaceous impressions in life-position. *Planolites* horizons are best exposed in borehole core and water-scoured gullies through mudstone successions, which produce clean cross-sections. *Planolites* is characterised by dense traces with a rounded to elliptical from in cross section (up to 3mm wide, and up to 10mm in length). The trace is often sub-parallel to the laminations, and is infilled with silty mudstone that is lighter in colour than the surrounding matrix. Pyritic concretions are seen to infill the burrows in rare examples.

Interpretation of type T2 key horizons

Lingula brachiopods indicate brackish marine conditions in the Namurian (Ramsbottom *et al.,* 1978; Waters *et al.,* 1996a). Marine band horizons have often been correlated with *Lingula* horizons in delta top areas (Wray *et al.,* 1930; Bromehead *et al.,* 1933; Stephens *et al.,* 1953). In the lower Marsdenian *Lingula* horizons have been correlated to the ammonoid-bearing *Bilinguities bilinguis* (R2b2) marine band and the *Zoophycos* and sponge spicule rich Keighley Bluestone lithofacies unit (Waters *et al.,* 1996a). Similar brackish water facies from cores in the Upper Mannville (lower Cretaceous) of Alberta are represented by morphologically simple ichnofacies, including *Planolites* (Pemberton & Wightman, 1992). Pale indurated horizons associated with *Lingula* and *Planolites* horizons are interpreted as zones of early diagenetic cementation, and could also correlate to flooding events (see Figure 6.2). The problem of recognising and correlating similar cryptic surfaces in the Carboniferous has been recently addressed (Davies & McLean, 1996). Palynological and spectral gamma-ray data were used in conjunction with horizons comprising bioturbated siltstone and *Promytillus* faunas from the Grindslow Shales (Kinderscoutian). These horizons were correlated into basinal areas, where they are represented by ammonoid-bearing marine bands. By comparing these examples and applying a similar methodology to this study, *Lingula* and *Planolites* faunas are inferred as representing the correlative equivalent of a marine band unit in interdistributary bay and delta top areas.

4.2.3. Type T3 key horizons: Marine-bioturbated horizon

This type of horizon occurs in a variety of facies, and has various ichnofauna densities and diversities. However, two end member types occur; those that comprise a highly intensive ichnofabric (Figure 4.5.1 to 4.5.5) and those that are less intensive (Figure 4.5.6 to 4.5.10). All marine trace fossils described here are endogenic, i.e. made by infaunal animals (Seilacher, 1964). Apart from bedding-plane exposures that are weathered in the field, intensely bioturbated horizons are often difficult to study, and specimens often require slabbing and polishing perpendicular to bedding. Body fossils are absent from T3, with the exception of rare broken shell fragments.

High-density bioturbation; At Middleton Towers (SD409588), weathered bedding planes are intensely bioturbated. Traces found include; *Treptichnus*, *Chondrites*, *Taenidium*, *Phycodes*, *Planolites* (*montanus* and *annulatus*) *Rhizocorallium* (*jenense* and *irregulare*; Figure 4.5.1 to 4.5.4). These traces were formed at the same time as hummocky cross stratification and wave ravinement surfaces (Brandon *et al.,* 1998), which provide three dimensional relief to bedding planes, and locally truncate bioturbated beds (Figure 4.3). Field observation and polished specimen analysis of the upper most flooding horizon at Fletcher Bank Quarries (SD805164) reveals a well-cemented, quartz-rich sandstone with a muddy-micaceous matrix (Figure 4.5.1 and 4.5.4). Allochthonous shelly debris is scattered through this lithology. The bioturbated bed is less than 0.2m thick, and contains common *Chondrites, Rhizocorallium, Teichichnus, Planolites* and *Palaeophycus.*

Low-density bioturbation; Ichnofaunas that produce a less intense fabric are dominated by disseminated sub-horizontal infaunal grazing traces, which do not completely rework the substrate. Such ichnofaunas include up to 80% *Olivellites and Scolicia* (*Scolicia* differentiated by the lack of an epichnial crest and flatter cross section), but also includes *Pelecypodichnus* (and bivalve escape structures), rare *Curvolithus* and very rare *Skolithos* (Figure 4.5.10). In some outcrops (Ponden Clough SE981364; Branshaw Quarry; SE032401, Leicester Mills Quarry; SD619164, Bare Clough; SE018308) dark grey-blue chertified siltstone-mudstones are bioturbated by *Zoophycos* (see Section 2.2.8). This horizon is often thin (less than 0.1m), and forms a highly indurated 'pavement' when exposed as laterally continuous bedding surfaces. Sparse *Teichichnus* and *Palaeophycus* bioturbation is often found in the zone beneath *Zoophycos* horizons. Both *Teichichnus* and *Palaeophycus* form a high-density/ low-diversity ichnofauna in siltstone or fine-grained sandstone beds between 0.3-2m thick. In isolated examples (i.e. Smalley Delph Quarry; SD716317, 8 -10m from base of section), *Teichichnus* occurs without *Zoophycos* as a high-density/ low-diversity fauna in the bedload-dominated sand-rich mouthbar facies (Figure 4.5.6).

Facies in which low density ichnofaunas occur are mostly deposited by suspension and low velocity tractionally currents (i.e. overbank flood deposits/ crevasse splay facies, shallow water open marine bay facies, and mouthbar facies).

Interpretation of type T3 key horizons

All bioturbation associated with the T3 horizon is infaunal. The paucity of epifaunal (grazing) trace fossils may be due to reworking by bioturbation or by mechanical wave/ current processes. Horizons with low-diversity ichnofaunas are associated with suspension deposited substrates, suggesting quiet bottom conditions with little or no current or wave reworking. In this environment the absence of epifaunal traces suggests a lack of oxygen/ nutrients in the sediment, or overprinting by later infaunal bioturbation. Such horizons could be generated by several mechanisms within the mouthbar environment. These include: marine transgression; mouthbar abandonment forced by allocyclic alterations in the supply system; or switching in the depocentre forced by a decrease in availability of accommodation. The lack of diagnostic marine faunas (ammonoid, brachiopod or marine bivalve) associated with, or within the sediments overlying T3 suggests that marine fauna did not colonise this environment, or that they were not preserved. Transgressive events in shallow marine systems are often associated with the development of gravel lag deposits, interpreted as wave ravinement surfaces. The association of high-density ichnofaunas with the gravel lag wave-ravinement surface suggests that this type of horizon underlies a significant marine flooding event.

Horizons similar to the T3 are observed in the middle-lower shoreface facies in the Cretaceous on the Western Interior Basin of the US (Pemberton & Wightman, 1992). Similar lower-diversity T3 *Zoophycos* ichnohorizons are also described on mouthbar tops in the Namurian of County Clare (Pulham, 1988). The *Zoophycos* from County Clare appear connected to sub-vertical burrows, that pass upward through a metre or more of overlying sediment, and represents infaunal bioturbation associated with a marine transgression. Within the Marsdenian of the Pennine Basin *Zoophycos* bioturbation occurs in association with sponge spicules in the chertified Keighley Bluestone (Stephens *et al.,* 1953; Wignall & Maynard, 1996; Waters *et al.,* 1996a). The Keighley Bluestone is inferred as a delta top restricted marine lagoon environment, which formed during either during a marine transgression, or on a topographically high area of the seabed. However, the Keighley Bluestone formed in a restricted geographic area, and cannot be utilised in regional correlation.

In the case of low-density bioturbation, evidence for marine faunas or wave-reworking is not observed, and it seems likely that mouthbar abandonment is more likely to be a mechanism for the genesis of some lower density ichnohorizons. If low-density ichnohorizons represent the upper surface of the mouthbar, and the hiatal surface is formed by mouthbar abandonment, there is no need to invoke transgressive control on the formation of low-density ichnohorizons. If this is the case, it may be impossible to correlate low-density ichnohorizons within the basin, unless discharge fluctuations or delta abandonment effects several adjacent mouthbar complexes. Abandonment in delta systems is often followed by sediment compaction and mouthbar subsidence, the net effect of which is effectively a flooding/ a localised transgression. In this scenario, the transgressive surface is localised, and does not form a basinwide correlatable surface.

4.2.3. Type T4 key surfaces: Flooding surface separating fluvial plain/ bayhead delta depositional systems

This surface separates facies of the fluvial plain depositional system from those of the overlying bayhead delta depositional system. Grainsize generally decreases above the surface, as deposition becomes dominated by mudstone and silty mudstone. In examples where the overlying bayhead facies are silt-rich, the exact position of this surface may be difficult to locate due to the similarity of grainsize. The surface often demonstrates depositional relief in laterally continuous exposures (up to 2-3m vertical relief over ten's of metres).

Examples of this surface can be seen at Fosters Delph Quarry (SE021273), Branshaw Quarry (SE032402) and Fletcher Bank Quarry (SD805164; Figures 4.7 & 4.8). When individual fluvial plain/ bayhead delta flooding surfaces are correlated over several kilometres, the younger bayhead delta often demonstrates rapid variation in depositional water depths, from bayhead prodelta, delta front and delta top associations. At Branshaw Quarry, the deposits overlying the fluvial plain/ bayhead delta flooding surface are dominated by crevasse splay facies with thin rooted horizons and coals, common symmetric and asymmetric ripples and rare *Teichichnus*. The correlated equivalent of this surface is exposed at Fosters Delph Quarry (SE021273), 11.5km in a southerly (basinward) direction (Enclosure 2; localities 20 and 29). In this area it is overlain by the shallow open marine bay facies and interbedded with an upward coarsening succession of lower shoreface (between SWB-FWWB) facies.

Interpretation of type T4 key surfaces

In exposures of restricted stratigraphical extent, the fluvial plain/ bayhead delta flooding surface has the potential to be misidentified as the basal surface of the channel abandonment association. However, the channel abandonment association is thinner (typically less than a few metres), and contains common drifted plant debris. The delineation of the fluvial plain/ bayhead delta flooding surface is useful, as it is laterally extensive and separates a diverse set of depositional environments; namely fluvial plain facies from bayhead facies. The lateral change in facies overlying surface each T4 type surface suggests deposition in a variety of water depths, implying that the bayhead delta did not advance over the full extent of the transgressive surface. The flooding event associated with T4 forced a short-term retrogradation of the supply system that allowed rapid progradation of the geographically widespread bayhead mouthbars. It is unlikely that the T4 surface was formed by compaction related subsidence, as it forms a surface which can be correlated for several tens of kilometres through a variety of depositional systems (see Enclosure 1 and 2).

4.2.4. Type T5 key surfaces: Flooding surface separating bayhead delta top/ bayhead prodelta associations

The T5 flooding surface occurs at the base of the bayhead prodelta association delineates the base of the repeated bayhead delta sequences (see Figure 5.3). Common allochthonous carbonaceous debris and suspension deposited micaceous silt-sized clasts are found in the silty-mudstone deposits immediately overlying T5. The silty mudstones are typically thin (less than 3m), and are gradationally overlain by suspension deposited siltstone and fine-grained sandstone of the bayhead delta association. Lateral correlation of this surface is problematic, as the underlying bayhead delta depositional system thickens or condenses over several to ten's of kilometres.

Interpretation of type T5 key surfaces

The common occurrence of scattered fine-grained clastics within the mudstone overlying T5 implies suspended silt was present in the water column during flooding. This suggests that very little retrogradation of the clastic supply system is associated with the flooding, or that current or wave processes during flooding forced re-suspension of the unconsolidated bayhead mouthbar top substrate. Transgressive reworking of the palaeosol surface may also account for the abundance of disseminated reworked plant debris (see Section 4.3.3). At the time of transgression, the water column was highly turbid, and potentially anoxic/ dysaerobic due to the bacterial decay of organic matter and the utilisation of free oxygen within the water column. The lack of oxygen and nutrients during and shortly after transgression may account for the paucity of ichnofaunas associated with this surface.

The T5 transgression was of a low magnitude, and although a eustatic cause cannot be ruled out, it seems that subsidence on the delta top, caused by compaction or isostatic loading of the underlying sediments is a more likely mechanism for formation.

4.3 Emergent / Palaeosol horizons (Palaeosol types; P1, P2 and P3)

Palaeosols, underclays and seat-earths have been described in the Carboniferous since the commencement of geological mapping in northern England commenced (Green *et al.,* 1878). Palaeosols were known to be important markers, but in the early literature, descriptions often tend to concentrate on their overlying coals. Palaeosols and coals record delta-plain emergence, the colonisation of the substrate by land plants and the effect of rooting. Palaeosol motifs are controlled by substrate type (i.e. sand-rich, mud-rich or heterolithic lithologies), and interactions with environmental parameters, (e.g. substrate, water-table hydrodynamics and climate). Although palaeosols demonstrate considerable lateral heterogeneity (from mud- to sand-dominated) or can be laterally absent, they are useful indicators of regional emergence that are used to facilitate correlation in the Marsdenian Pennine Basin.

Three palaeosol facies are identified in this study: i) leached sand-rich 'ganister' palaeosols; ii) heterolithic palaeosols; and iii) mud-rich palaeosols. The characteristics of all three types are highly dependent on substrate type and water-table levels, which can vary during palaeosol formation. For example, changes in water-table during formation implies an initially leached palaeosol may be overprinted by a waterlogged motif. Pedogenic signatures are therefore diffuse and are consequently correlated as horizons, and not surfaces. Palaeosol horizons are considered as instantaneous geologic features, probably developing between 10's- to 100's of years. Variations in substrate type, climate, along with the hydrodynamics of the water-table, and the presence, absence, or combination of these factors explains their variability and distribution.

The palaeosols described here occur on widespread low-lying delta-top areas that lie slightly above sea-level. Such peneplain environments are susceptible to marine inundation, due to localised subsidence or regional eustatic flooding, and lowering of water-tables during marine regressions.

4.3.1. Type P1 key horizons: Leached palaeosols

The thickness of leached 'ganister' palaeosol horizons varies between 1- 10m. The matrix is often cream-pale grey and quartz-rich, with a sugary texture and carbonaceous sub-vertical rooting that is often well preserved. The fabric and structural characteristics of the leached palaeosol facies are described in the leached palaeosol Facies Description Chapter (see Section 2.2.23). This horizon occurs in the interfluve or bayhead delta top association and forms a caping layer to the channelised and laterally accreting channelised sandstone association, and the channel abandonment association. Lateral correlation of the leached palaeosol horizon can be difficult, as its composition is reliant on the underlying lithologies.

Interpretation of type P1 key horizons

The mechanisms behind ganister palaeosol formation are outlined in the facies description of the leached sand-rich 'ganister' palaeosol. Comparisons between characteristics and processes involved in the formation of both ancient and modern freely-drained soil horizons, suggest that kaolinisation and the removal of soluble bases leads to the development of a leached albic palaeosols or ganisters (Percival, 1986; Sigleo & Reinhardt, 1988). The rate of ganister formation also increases with temperature, due to the increased rate of feldspar degradation to kaolinite (Sigleo & Reinhardt, 1988). The climate of the Pennine Basin during the Namurian is suggested to have been warm and seasonally wet; conditions that are ideal for the development of leached palaeosols. Although the rate of development decreases rapidly after initial weathering, palaeosols that form over longer time-periods are deeper, and possess increasingly leached profiles (Sigleo & Reinhardt, 1988). Critically, the water-table must be low, and precipitation must exceed evaporation to allow drainage and leaching of the substrate. Where the leached palaeosol horizon develops on mudstone lithologies, the lack of free-drainage may make leached motifs difficult to identify. However, where leached pedogenic horizons are represented by thin leached palaeosols, evidence of rooting and colour change is often well preserved.

4.3.2. Type P2 key horizons: Waterlogged palaeosols

This horizon comprises the heterolithic palaeosol and the muddy palaeosol facies and occurs in horizons on average 1.5m, and up to a maximum thickness of 3m. Waterlogged palaeosol horizons are usually overlain by surface T5, and the upper surface of the P2 horizon is often reworked by tractional processes during flooding.

Interpretation of type P2 key horizons

The interbedded, suspension deposited sandstone-mudstone lithologies of the waterlogged heterolithic palaeosol facies formed in shallow standing water. This is supported by the occurrence of tree-stumps with radiating *Stigmaria* rooting systems (DiMichele & Phillips, 1994; Wagner, 1995). Horizon type P2 therefore represents a gradually accreting overbank soil that formed close to a clastic source. Where waterlogged palaeosols develop in muddy substrates, carbonaceous debris (comprising rootlets and leaf material) is often well-preserved. Rooting in this facies is often intense, and it is often impossible to observe any primary depositional fabric. This horizon is deposited at or very close to lake, mire or sea-level, and it is likely that this horizon development occurred in a shoaling and eventually emergent shallow interdistributary bay of a bayhead delta.

Palaeosol formation can be highly diachronous; i.e. they develop over an extended period (Retallack, 1990). This is because many of parameters affect their formation, such as substrate type, climate, and the hydrodynamics of the water-table. The relative influence of these factors may help to explain their spatial and temporal variability. The palaeosols described in the Marsdenian formed at, or slightly above sea-level. Such peneplain environments are susceptible to marine inundation, caused by subsidence or eustatic flooding, or declining water-tables, during periods of marine regression.

The climate of the Pennine Basin throughout the Namurian and Westphalian is thought to be seasonal (i.e. the occurrence of periodic annual modulations between wet and dry periods), due to the identification of episodic fluvial deposits inferred as the product of seasonal 'monsoon' floods (Broadhurst *et al.,* 1980; Vanstone, 1991). This signature is difficult to establish from the palaeosol record alone (Okolo, 1982; Collinson, 1988), although seasonally high discharges from fluvial channels may have led to overbank flooding, and regional wetting of the interfluve. The effect seasonal climate has on pedogenic processes is dependent on the amount of water available during palaeosol development. When the substrate is waterlogged (i.e. during a high water-table) plant growth is continuous, whereas during periods of low water-table, the substrate holds little or no available water. Palaeosols within the Pennine Basin do not show signs of periodic wetting and drying, but are more commonly, either completely leached or waterlogged (Percival, 1992). Because eustatic sea-level change controls the water-table beneath developing soil-horizons, it is difficult to extract a seasonal signature from Marsdenian palaeosols. It therefore more likely that eustatic sea-level change and its effect on water-table level controlled the degree of waterlogging/ leaching rather than seasonality.

A varied climate system in the Carboniferous of the UK is also suggested by the occurrence of fusain (fossil charcoal) in Dinantian to Westphalian successions (Nichols & Jones, 1992; Scott & Jones, 1994; Falcon-Lang, 1998; Falcon-Lang, 1999a; Falcon-Lang, 1999b; Falcon-Lang, 1999c). Beds with abundant fusain suggest that dry conditions allowed the proliferation of catastrophic wildfires, which destroyed vegetation and lead to weathering of the exposed soil. Such weathering potentially led to widespread sub-aerial erosion, and increased fluvial channel bedload. The paucity of fusain-beds in the Marsdenian suggests an absence of prolific wildfires, potentially due to the waterlogged conditions in the bayhead delta and delta top environments.

4.3.3. Type P3 key horizons: Leached palaeosol horizon with waterlogged overprint

The upper parts of some ganisteroid (leached) palaeosols often have a muddy matrix between ped structures. Rooting is common within type P3 horizons, and some examples have erosive scours that are filled with massive, reworked palaeosol deposits. P3 key horizons are often overlain by either horizon types T1,T2 or T3. In some cases marine bioturbation from horizon type T3 appears to rework palaeosol deposits (Figure 4.5.5).

Interpretation of type P3 key horizons

The association of horizon P3 with transgressive horizons T1-T3 suggests rising sea-level had the capacity to rework the palaeosol surface. An increased water-table would have been the precursor to marine inundation and would have generated the muddy overprint. It is not clear from these examples whether the mud within the palaeosol originated within the palaeosol and was reworked during raising of the water-table (i.e. a mud-rich hardpan), formed by weathering processes during waterlogged periods, or was imported by during the marine transgression.

Palaeosols with similar characteristics are described from the Breathitt Group (Pennsylvanian) of eastern Kentucky. Well-drained palaeosol horizons formed on incised-valley interfluves, and were often overprinted by clay-rich, gleyed soil horizons, inferred as the product of water-table elevation, during periods of sea-level rise. This interpretation is supported by other workers, who use clay (kaolinite/ illite) mineralogical variations to invoke the same two-stage pedogenic model for the genesis of other Pennsylvanian interfluve palaeosols (Gardener *et al.,* 1988).

4.4. Regressive surfaces (Unconformity types;R1, R2 and R3)

Regressive surfaces delineate a sharp- basinward step in facies, i.e. the facies underlying the surface represent relatively basinward deposits, and those overlying represent relatively proximal deposits. Although it is often difficult to identify erosional relief on surfaces in the field, correlation often reveals relief on a local to basinwide scale. Individual unconformities have a variable amount of erosive relief, and some erode through other key horizons/ surfaces.

4.4.1. R1 key surfaces: unconformities underlying the channelised sandstone association.

This type of surface separates the channelised sandstone association from underlying facies associations of the delta front. Correlation of surface R1 reveals that underlying facies associations differ across the basin, from inter distributary bays to distal mouthbars. The channelised sandstone association directly overlying this unconformity often comprises a granule/ pebble lag, which records a marked increase in grainsize (from fine- to medium-grained sandstone to a coarse-granule sandstone). Plant debris commonly overlies the unconformity, and often occurs as log-jams, with granules and small quartz and feldspar pebbles. Due to the grainsize and facies change, this surface is often easy to identify in both logged sections and boreholes. However, it is important to assess the variation in characteristics and lateral extent of R1 as it may be misidentified as surface R3 or the base of laterally discontinuous channelised sandstones, which commonly demonstrate less than 10m of relief.

There are few large exposures demonstrating the characteristics of this unconformity, and even in those that are laterally extensive, it is difficult to find direct evidence of a down-cutting relationship that may be associated with an angular unconformity. However, at Fletcher Bank Quarry (SD805164) the integration of logged sections using laterally continuous datums, suggests that 10.5m of relief is apparent over 200m (Figure 4.9 and 4.10.2).

Interpretation of R1 key surfaces

The increase in grainsize, and the dominance of tractional transported sandstones suggests that an increase in flow velocity is associated with the generation of surface R1. This may have been caused by either an allocyclic increase in fluvial input or a shallowing of depositional environment. Individual R1 surfaces can be traced for up to 60 km, and have an erosive relief up to 40m (Enclosure 1 and 2). This suggests either; i) the fluvial channels had the capacity to erode through considerable stratigraphy, ii) differences in the bathymetry existed across the basin, or iii) differential compaction may have modified the amount of relief. With the extent and depth of erosive relief on surface R1 being so widespread, it is likely that erosion occurred in response to a regionally significant shallowing event.

The most widely correlatable R1 surface in the early- to mid-Marsdenian has been previously identified at the base of the Midgley Grit in Yorkshire (Wignall & Maynard 1996). According to Wignall and Maynard (1996) this surface has up to 80m of regional erosive relief and can be correlated 25 km to the west, where it is equivalent to the base of channel-fill facies observed in the Fletcher Bank Grit at Fletcher Bank Quarry (SD805164) (Okolo 1983). At this locality, the R1 surface comprises a set of stacked concave-up channel bases that erode into a delta front mouthbar complex. This study has demonstrated that the estimated 80m of erosive relief is exaggerated, and that it is a function of their correlation of R2a1 (*Bilinguites gracilis*) and the R2b (*Bilinguites bilinguis*) marine bands (see Section 6.5).

Fluvial facies associations infilling channel features are described throughout the Namurian of the UK. Distributary channels within the Lower Kinderscout Grit in the Pennine Basin are described as having irregular and inclined basal surfaces (between 10° and vertical), which have strike directions parallel to palaeocurrent directions in the overlying fluvial facies (McCabe 1977). When traced regionally, these surfaces demonstrate relief of 20-40m, which may be the product of combined multi-phase erosion and/or depositional relief. The equivalent surface at the base of the Lower Kinderscout Grit and in the upper Grindslow Shales (LKG7-8) on the Kinderscout Plateau is inferred as underlying a regionally regressive, erosively-based deltaic sandstone (Hampson 1997). The erosive relief at the base of this surface is often greater than 40m, which is more than can be attributed to individual erosively based channels (up to 25m thick). The Farewell Rock (Yeadonian, South Wales), also has a similar erosive base with regional erosive relief (Hampson 1998).

4.4.2. Type R2 key surfaces: unconformities underlying turbidite association

This surface separates the basinal mudstone and turbidite associations from younger turbidite associations. In laterally-continuous exposures, (e.g. Mouselow Quarry; Figure 4.11) surface R2 has a concave-upwards geometry (up to 100m wide and 20m high), which truncates underlying bedding. Other outcrops reveal a planar base that lies parallel to bedding in underlying facies (e.g. Figure 4.12), but 'downsteps' sharply 5-8m over a similar lateral distance. Whether planar or channelised, the R2 surface has common flute marks and groove casts. High-density turbidite facies always overlie surface R2. They comprise fining upwards beds coarse- to medium-grained sandstones and mudstone rafts, which commonly overlie the unconformity at channel axes, with abundant flute and tool marks.

Exposures of this horizon are rare, but are observed at Mouselow Quarry (SK024952), Crowden Great Brook (SE063032) and Alum Crag (SD636280), while numerous examples are inferred from subsurface borehole data. The exposures at Mouselow Quarry and Alum Crag are continuous for over 200m, and provide the majority of field observations.

Interpretation of type R2 key surfaces

The 'stepped' profile observed at Alum Crag appears to be formed by a series of laterally coalesced channels. This implies flow confinement, suggesting that the surface developed as the product of several large turbidite events. This scenario suggests surface type R2 was produced by many individual episodes of turbidite erosion, implying that it is diachronous. The exposure at Mouselow Quarry (SK024952; Figure 4.11) reveals the relationship between surface R2 and the bedding in the underlying facies. Channels have steep margins (>20°), and are filled with high-density turbidite facies, which erode into, or conformably overlie low-density turbidites, that were deposited from thin sheet-like turbidity currents. The high-density of concave-up channels, and the stepped channel margin profile suggests that erosion by episodic turbidity currents was common. During periods of high-discharge and sediment bypass, the flow is confined within the channels had a high erosive potential, whereas deposition occurred during flow waning and channel abandonment. Similar turbidite associations overlie delta front mudstone in several studies of the Namurian Pennine Basin, notably (Walker 1966a; Walker 1966b; Collinson 1968a; Collinson *et al.* 1977; Jones 1980).

Channels from the Grindslow Shales (Kinderscoutian) of the Pennine Basin are between 3-45m thick. They are inferred as the product of successive flows that enlarge the channels, which were subsequently infilled by waning turbidity currents (Walker 1966a; Walker 1966b). Although the amount of relief on the channel complex base, which is equivalent to surface R2, is greater in the Kinderscoutian examples described by (Walker 1966b) they possess similar characteristics to those in the Marsdenian.

Channelised turbidite facies are described in the Roaches Grit (R2b4-R2c1) of Northern Derbyshire, where turbidite channel axes demonstrate a maximum basal relief of 3.5m deposits (Jones 1980; Jones & Chisholm 1997). The channel axes pass laterally into planar-bedded sandstone that is inferred as channel levee/ overbank facies, deposited in a deep-water turbidite fan complex. The surface R2 described in this study is equivalent to the basal contact of the turbidite association described by Jones and Chisholm (1997).

4.4.3. Type R3 key surfaces: unconformities underlying single storey fluvial channel association.

Unconformable surface R3 separates facies of the underlying proximal mouthbar association from the channelised sandstone association. The basal part of the overlying channel association comprises fine- to coarse-grained sandstone, but unlike facies overlying surface R1, log debris and granule and pebble lag deposits are rare. Above the unconformity, the dominant sedimentary structures are cross-stratification, while lateral accretion surfaces are absent. However, when the surface is correlated over several kilometres, between 5-10m of erosive relief can be demonstrated on this unconformity. The lateral extent of an individual R3 surface is between 1-10 km (Figure 4.13).

As with key surface R1, the lack of continuous exposure makes it difficult to measure the amount of relief associated with this surface. For example, the R3 surface at Leyzing Clough (SE068082) and Wessenden Head Bore (SE062087) is suggested as equivalent to the R3 surface in the Woodhead Tunnel Bore (SE135011; Figure 4.13). At all three localities key surface R3 overlies the proximal mouth-bar association. When datumed on the overlying flooding surface, R3 has less than 6m of erosive relief (Figure 4.13).

Interpretation of type R3 key surfaces

Where the proximal mouthbar association is overlain by locally incisive/ thinning channelised sandstone, the channelised sandstone is inferred as the product of isolated mouthbar distributary feeder channels. Within the distributary channel, the abundance of cross-stratified sandstone, and paucity of lateral accretion surfaces suggests the development of a low sinuosity channel. Similar channel geometries are observed in the Mississippi River Delta (Figure 4.14). Present day examples from the Mississippi illustrate the development of single, straight channels (e.g. Southwest Pass and Pass a Loutre; Figure 4.14) that feed prograding mouthbar complexes during periods of high relative sea-level (Coleman, 1981). In the example from the Southwest Pass, the distributary channel and mouthbar are genetically linked, i.e. the channel supplies sediment at the mouthbar allowing the distributary channel to prograde basinward. This analogue implies that mouthbar progradation, and the development of the surface R3 unconformity is diachronous and laterally restricted.

Similar straight channel geometries, infilled with thin tractionally transported sandstone form mouthbar-distributary channel complexes in the Haslingden Flags (Yeadonian) of Lancashire (Collinson & Banks, 1975), and the Bideford Group (Namurian), North Devon (Elliott, 1976) (Figure 4.15). The examples from the Haslingden Flags and the Bideford Group are described as 'elongate' deltas, on the basis of sandstone distribution. The relationship between a mouthbar and a low-sinuosity distributary feeder channel is similar to the reconstructions of the Marsdenian-aged Scotland Flags (Wignall & Maynard, 1996) which also suggest an elongate delta form. In the examples from the Marsdenian, the absence of the channel fill and basal unconformity suggests that the channel association represents only localised incision. Field observations suggest that the channel fill above surface R3 comprises a single channelised sandstone, implying the cut and fill of a single channel form.

4.4.4. Type R4 key surfaces: unconformities underlying proximal and distal mouth-bar or bayhead delta front association 'sharp-based mouthbar'

Surface R4 sharply separates either the basinal mudstone association from the overlying distal/ proximal mouthbar association or bayhead prodelta association from the overlying bayhead delta associations. The grainsize above the surface, and throughout the overlying distal/ proximal mouthbar and bayhead delta front associations comprises uniform siltstone to fine- to medium-grained sandstone. Coarse lags, similar to those associated with surface R1 are not observed.

Suspension processes dominate the deposition of facies overlying surface R4. These facies commonly comprise mouthbar, and rare tidally influenced mouthbar and cross-bedded facies. Bedding within the mouthbars is either layer-parallel or comprise subtle, downlapping lobes. The consistently layer parallel character and bedding of the key surface and under/overlying facies creates difficulties in identifying erosive relief and downlap onto the key surface. Where the distal mouthbar association overlies the unconformity, the grainsize contrast between mouthbar facies and basinal mudstone association is mudstone to siltstone and occasionally surface R4 is difficult to identify. In the few laterally-extensive outcrops of this surface, this unconformity appears to have very little, if any, erosive relief (Figure 4.16 and 4.17). Although it is often not possible to identify erosive relief at individual exposures, datuming on overlying flooding surfaces reveals that the unconformity has 2-10m of relief. At Branshaw Quarry (SE032402), silt-rich mouthbar facies grade into a bayhead prodelta and bayhead delta top association (Figure 4.8 and 4.17). Surface R4 is a sharply defined surface, and reveals approximately 1m of erosive relief over a 75m exposure.

Interpretation of type R4 key surfaces

Sediments immediately overlying any R4 key surface lack granule-pebble lag or log-jam deposits, have no bedforms, and are dominated by parallel-bedded suspension deposited mouthbar facies. These observations suggest that the surface is unlikely to be formed by erosion during bedload transport and dune migration. However, the occurrence of sand-dominated mouthbars with inclined bedding suggests localised deposition in almost emergent mouthbars, which were probably dominated by a high flow regime. Such high flow velocities would have the capacity to down-cut and generate the 2-10m erosive relief observed. However, it is not possible to determine solely from field evidence whether erosion or relict bathymetric relief is responsible for the topography on this surface.

Sections through prograding coastline successions (wave-reworked shoreface or shelf delta mouthbar) typically record the gradual passage from marine to coastal plain or delta-top environments. Planar erosive surfaces, similar to surface R4, are described in the Middle Shales (Marsdenian) of South Wales, and demonstrate erosive relief and a marked basinward shift in overlying facies (Hampson, 1998). These surfaces denote the lower boundaries of sharp-based mouthbars and have laterally-extensive planar bases, which are correlated for up to 1000m. Eustatic sea-level fall is inferred as the controlling mechanism behind the formation of these sharp-based mouthbars. Falling sea-level would have produced an erosive base by the combination of erosion during the lowering of wave-base and fluvial scouring down-current of the fluvial source (Hampson, 1998). The sharp-based mouthbars in the Middle Shales are the fluvial-dominated equivalent to sharp-based shorefaces documented in wave-dominate deltas. Sharp-based surfaces separating basinal and wave-reworked facies have been described in numerous shoreface settings (Plint, 1988; Ainsworth & Crowley, 1994; Martinsen *et al.,* 1995; Fitzsimmons & Johnson, 2000). These can be traced laterally for considerable distances and can be utilised as a correlatable surface. In the Marsdenian of the Pennine Basin, the facies above key surface R4 do not appear strongly wave-reworked, suggesting the dominance of a fluvial rather than wave reworking processes.

4.5 Summary of key surface chapter

Key Surface/ Horizon	Fauna	Characteristics	Interpretation	Correlation potential
Key horizon T1: *Marine band*	Goniatites, Dunbarella, *rare* (Anthracoceras, Dimorphoceras, Hudsonoceras), Caneyella.	Black, well-laminated or massive mudstone, ammonoid rich horizons are typically less than 0.15m thick.	Implies that fully marine conditions are established, and a high-magnitude base-level rise	Forms the primary correlation framework, and has a high correlation potential over several 100 km's. Laterally correlatable with T2, T3, T4, P2 and P3.
Key horizon T2: *Lingula-Planolites*	*Lingula* and *Planolites*	Isolated horizons, less than 0.05m thick that are often dominated by a massive black, indurated mudstone matrix. Occurs within the interdistributary bay facies	Suggests slightly restricted marine or brackish waters, associated with a rise in base-level close to the shoreline	Correlates laterally with T1, T3, T4, P2 and P3 key horizon.
Key horizon T3: *Marine ichnohorizon*	*Zoophycos, Palaeophycus, Trephichnus, Chondrites, Taenidium, Phycodes, Planolites* (*montanus* and *annulatus*) *Rhizocorallium* (*jenense, irregulare* and *Teichichnus, Olivelittes* and *Scolicia, Skolithos* and rare *Curvolithus*	Comprises bioturbated substrate, of varying degrees, with abundant infaunal trace fossil faunas. Occurs only within the delta front and bayhead mouthbar top association	Suggesting base-level rise, or a depositional hiatus, when bottom conditions are well suited for benthic organisms. Substrate underwent little or no mechanical reworking. Generated by sea-level rise/ mouthbar abandonment.	Correlatable equivalent to Key horizons T1 and T2.
Key surface T4: *Flooding surface separating fluvial plain/ bayhead delta depositional systems*	None	Overlies the fluvial plain association and separates it from the bayhead delta depositional system. Grainsize decreases above the surface and deposition is dominated by mudstone and silty mudstone.	Slight base-level rise forced short-term supply system retrogradation, allowing rapid progradation of thin bayhead mouthbars.	Forms a geographically widespread flooding surface.
Key surface T5: *Flooding surface between bayhead delta top and bayhead prodelta associations*	None	Separates bayhead associations from the base of the overlying bayhead delta sequence. Comprises common allochthonous carbonaceous debris and suspension deposited mica silt clasts.	Low magnitude base-level rise; although compaction of underlying sediments is also a likely cause. Carbonaceous debris suggests reworking of the palaeosol horizons, or input from fluvial sources.	Potentially forms regionally correlatable surface, although dependent on lateral continuity of underlying bayhead delta depositional system.
Key surface R1: *unconformity-underlying multi-storey channelised sandstone association.*	None	Separates multi-storey stacked channel association from underlying facies associations of the delta front mouthbar depositional system. Increase in grainsize above surface, between 15-40m erosive relief.	Base level fall occurring at two orders of magnitude; i) 15-20m, or, ii) 40m; (see Enclosure 2).	Comprises coalesced channel bases, that are mutually erosive, absolute correlation of individual channels difficult unless exposure is extensive. Laterally correlates with R2, R4 and P1.
Key surface R2: *unconformity underlying turbidite association*	None	Separates the basinal mudstone, and turbidite association from younger turbidite associations	Erosively based channels with localised planar bedded overbank deposits. Common flute marks and groove casts abundant.	Comprises coalesced channel bases, that are mutually erosive, absolute correlation of individual channels difficult unless exposure is extensive. Laterally correlates with R1, R4 and P1.
Key surface R3: *unconformity underlying single storey fluvial channel*	None	Separates facies of the proximal mouthbar association from the overlying stacked channel association. Increase in grainsize above surface, and less than 6m of erosive relief.	Distributary feeder channels genetically related to, and overlying the proximal mouthbar association. A localised incisive feature.	Localised distributary channel base, locally correlated, but does not form basinwide correlatable surface.
Key surface R4: *'Sharp-based mouthbar'*	None	Separates basinal mudstone association from the overlying distal and proximal mouthbar association or bayhead prodelta association from the overlying bayhead delta front. Tidal processes often influence some of the mouthbar deposits. Although it is not possible to identify erosive relief at individual exposures, datumed on the overlying flooding surfaces, the unconformity has 2-10m relief.	Formed by erosion at the head of a mouthbar during lower base-level due to increased bedload transport and dune migration.	Comprises coalesced channel bases, that are mutually erosive, absolute correlation of individual channels difficult unless exposure is extensive. Laterally correlates with R1, R2 and P1.
Key horizon P1: Leached palaeosol	*Stigmaria,* rootlets	The matrix is often cream-pale grey in colour and quartz rich, with a sugary texture and carbonaceous sub-vertical rooting that is often well preserved.	The leached nature implies formation on a well-drained sandy substrate. Formed while water table was relatively low, allowing a downward flow of ground water. Leaching of clays through substrate forced concentration of residual quartz/ feldspar in an albic horizon (Percival, 1986).	Absolute lateral correlation of the palaeosol horizons can be difficult, as substrate type is an important control on development. Where mud-rich lithologies dominate, waterlogged soils may develop, or the rate of leaching through the substrate may be decreased.
Key horizon P2: *Waterlogged palaeosol*	*Stigmaria,* rootlets	Waterlogged palaeosols occur within the bayhead delta top association, and infrequently in the mouthbar top association. This horizon comprises the waterlogged heterolithic palaeosol and the waterlogged muddy palaeosol facies, occurring in horizons up to a maximum thickness of 3m thick, but are on average 1.5m in thickness.	Deposited in shallow standing water, and overlain by T5, suggesting flooding of the palaeosol during/ after formation.	
Key horizon P3: *Leached palaeosol horizons with waterlogged overprint*	*Stigmaria,* rootlets	Upper parts of some leached 'ganister' palaeosols often have a muddier matrix, and are reworked by infaunal bioturbation and tractional processes. This horizon is overlain by T1-T5.	Rising sea-level reworks the palaeosol surface, and raises water table. As the water table is raised, the leached palaeosol is overprinted with a waterlogged motif.	

Table 2 Summary of key surfaces/ horizons utilised in this correlation, with notes regarding relative magnitudes of base-level modulation and correlation potential of each surface/ horizon. Also see Enclosure 4.

Chapter 5: Depositional systems

5.1. Introduction: The significance of depositional systems

Depositional systems comprise between two and four facies associations, and are classified as either a *basin floor*, *delta front mouthbar*, *bayhead delta* and *fluvial plain* depositional systems. Major breaks in the succession may separate neighbouring facies associations and depositional systems and therefore represent significant erosive or hiatal events (Middleton, 1973). Such events are important as they may delineate basinwide shifts in depositional environment belts, or changes in basinal conditions (see Chapter 4). Such depositional breaks therefore serve as useful indicators of major environmental change and may form correlatable event markers.

5.1.1. The basin floor depositional system

This system represents the most distal depositional environments of the Marsdenian deltas. Sediments were deposited in sufficient water depths where basinal wave or tidal activity cannot rework the substrate surface. Sediment was deposited either from suspension, or from waning turbidite flows (Figure 5.1). Basinal mudstone associations with and without correlatable ammonoid-rich offshore marine band facies commonly attain thicknesses in excess of 30m for tens of kilometres. The basinal mudstone association is commonly thinner (less than 10m in places) where overlain by a prograding delta front mouthbar depositional system and often interdigitates with turbidite and distal mouthbar front associations. Because of this, the turbidite association is often laterally discontinuous, and individual beds thin rapidly, as do whole turbidite lithostratigraphic units (such as the Alum Crag Grit, west of Blackburn) (Price *et al.,* 1963; Collinson *et al.,* 1977). As the basinal floor depositional system is dominated by mudstone and siltstone lithologies, it is often not exposed.

5.1.2. The delta front mouthbar depositional system

This system comprises a coarsening upward succession that varies in thickness from 2 to 30m (Figure 5.2). The base and top surfaces of this system are delineated by transgressive or regressive surfaces, that can be correlated locally, and in some cases can be traced laterally for tens of kilometres. A decrease in the thickness of the distal and proximal mouthbar front, and mouthbar top associations is often accompanied by an increase in the proliferation of the inter mouthbar trough and basinal mudstone association. The presence of one or more of these associations is dependent on whether the thinning in the mouthbar front associations is due to a transition towards a basinal or into an interdistributary trough area. Mouthbar top associations in the delta front mouthbar depositional system are often thin (less than 1m), and palaeosol facies are rare, occurring predominantly where channelized fluvial sandstone is developed.

CASE STUDY 1: THE MISSISSIPPI RIVER DELTA; ANALOGUE OF DELTA FRONT MOUTHBAR DEPOSITIONAL SYSTEMS

Modern mouthbars are difficult to study because they are largely sub-aqueous and associated with high fluvial discharge. Syntheses of mouthbar deposit styles are based on theoretical, laboratory and field investigations, and concluded the forces involved during deposition involve interactions between river discharge, effluent and basin water density and tide/ wave modification (Wright 1977, Figure 2.12).

Examples of fluvial dominated mouthbars include the Balize delta of the Mississippi River (Coleman, 1981; Coleman *et al.,* 1998a). Mississippi mouthbars comprise the coarsest clastic material present within the delta system, are well-sorted and contain abundant organic debris in areas away from active deposition. Since the start of the Holocene a 200m thick delta has prograded across the continental shelf. This progradational succession comprises up to 100m of clay-rich prodelta deposits, overlain by regressive sand-rich mouthbar sediments and the ‘birdsfoot’ style linear distributary channels. The dominance of a linear channel pattern, and absence of channels with sinuous geometries, may be controlled by high fluvial discharge, or the effect of increased flow velocities due to the vicinity of the sloping mouthbar front.

Shallow water, geographically less extensive bayhead deltas commonly occur within inter-distributary troughs (see Figure 4.14). The great thickness of clay beneath the dense, water saturated regressive mouthbar deposits, creates ideal conditions for the injection of mud and the genesis of mud-diapirs during compaction. The differential thickness of the Balize delta (200m) when compared to examples from the Marsdenian (maximum uncompacted thickness of 35m) may explain why deformation features of a similar scale do not occur in the Pennine Basin. The rate of mouthbar regression during periods of stable sea-level is considered to be rapid. For example, the South West Pass mouthbar of the Mississippi River has migrated 10 km seaward in 195 years (Gould, 1970) (see Figure 4.14).

Examples of present day tidally influenced mouthbar deposits include those of the Changjiang River estuary (Yellow River), China (Dingan & Jiemin, 1996; Jiu-fa *et al.,* 2000)(Dingan & Jiemin 1996; Jiu-fa *et al.* 2000), where tidal currents are responsible for the reworking and resuspension of the mouthbar deposits.

5.1.3. The bayhead delta depositional system

This depositional system is represented by repeated coarsening upward cycles between 2-17m in thickness (Figure 5.3). The depositional sequences commence with an initial deepening event, followed by a coarsening upward bayhead delta succession. Facies in the lower part of the depositional system possess gradational boundaries, while facies in the upper part commonly exhibit evidence of depositional hiatuses and erosion. The bayhead delta depositionsal system is generally represented by a thinner succession than the delta front mouth-bar association, and comprises a higher proportion of mud, with sedimentary structures commonly indicating a tidal influence.

Generally, bayhead deltas develop in shallow water, are geographically restricted and dominated by mouth-bar deposits. In the twenty-five years from the initiation of sub-aqueous growth, approximately 9km^2 of emergent land was created by the progradation of the Atchafalaya River delta (interpretation based on maps in Van Heerden & Roberts 1988). If similar rates of rapid delta progradation occurred in the Marsdenian, it is likely that Carboniferous bayhead delta deposits represent the development of laterally extensive sheet deltas over relatively short (i.e. 100-1000's of year) time periods.

CASE STUDY 2: ATCHAFALAYA RIVER DELTA; AN ANALOGUE FOR THE BAYHEAD DELTA DEPOSITIONAL SYSTEM

The Balize delta has been the main Mississippi River Delta depocentre for the past 600-800 years (Coleman, 1981). During this time the course of the Mississippi River has undergone a decrease in gradient as the Balize delta has prograded across the shelf. Due to the decrease in the river graded-profile, flow velocities have waned, and the Mississippi River has become less efficient at transporting sediment. This has led to an increase in sediment deposition on the delta-top, creating conditions where avulsion and crevasse events are common. For example, the initiation of a new trunk stream, the Atchafalaya River, and the subsequent formation of the Atchafalaya River delta has led to the formation of a new sub-delta, or bayhead delta, developing on the delta-top (Tye & Kosters, 1986; Van Heerden & Roberts, 1988; Tye & Coleman, 1989b).

The mouthbar dominated Atchafalaya River delta is currently building into the shallow-water (2-4m) of an old interdistributary bay, the bathymetry of which deepens in a seaward direction (Van Heerden & Roberts 1988; see Figure 5.4.3). The shallow depth of the Atchafalaya Bay allows reworking by tidal currents, wind and basinal waves (Van Heerden & Roberts 1988).

Similar depositional environments are interpreted in the Marsdenian Pennine Basin to those observed in the Atchafalaya Bay. The thickness of Carboniferous and modern Atchafalaya bayhead successions are equivalent, suggesting deposition occurred in analogous environments of comparable water-depth (Table 3). Delta progradation in Atchafalaya Bay is laterally unconfined, and the delta has a sheet-like geometry (Table 4), which is also similar to the Marsdenian bayhead delta depositional systems. The Atchafalaya bayhead delta has prograded rapidly into bay areas, suggesting emergent conditions can develop rapidly. Initiation of this bayhead delta commenced in the early 1950's. Analysis of cross-sections from Van Heerden & Roberts (1988) reveals that the Atchafalaya bayhhead delta comprisies a seaward thinning wedge (between 0.5-1.5m thick) of prodelta clays (Figure 5.4.3), which are overlain by distal-bar sandstone (between 1-1.5m thick) representing the sub-aqueous stage of delta growth. The development of levees, fed by overbank discharge from the advancing fluvial supply system, led to emergence and colonization by algal-flat sediments. Detailed comparisons between the Atchafalaya River delta and Marsdenian bayhead delta cycles are highlighted in Table 4.

Facies this study	Environments in Atchafalaya delta (Tye & Kosters, 1986; Tye & Coleman, 1989b; Tye & Coleman, 1989c)
Waterlogged muddy palaeosol, Waterlogged heterolithic palaeosol, Allochthonous/ swamp coal, interbedded with marine shallow water muddy bay facies / Interdistributary bay facies	Backswamp
Overbank flood deposits/ crevasse splay	Natural Levee
Channelized fluvial sandstone (and/ or) Tidally influenced cross-bedded sandstone	Distributary Channel
Sand dominated mouth-bar, Silt dominated mouth-bar, Tidally influenced sand dominated mouth-bar and Tidal influenced silt dominated mouth-bar, Shoaling mouth-bar	Distributary Mouthbar
Offshore 'marine band' mudstone, Offshore black mudstone and Offshore prodelta silty mudstone	Prodelta/ Delta Front
Marine shallow water muddy bay facies, Interdistributary bay, Shallow water restricted marine bay	Lacustrine/ Inter distributary trough
Lower Shoreface above fair-weather wave base	Not described

Table 3 Comparison of facies from the bayhead delta association in the Marsdenian Pennine Basin with environments in the Atchafalaya bayhead delta, Louisiana

	Reference/ Measured Data					Estimated data		
	Fluvial Channel Discharge	Sediment Calibre	Thickness of individual bayhead delta cycles	Area covered by individual bayhead delta	Tidal range	Area dimensions of whole basin	Number of bayhead delta cycles in stacked succession	Potential volume of sediment in individual bayhead delta
Marsdenian Pennine Basin (this study)	***Greater than 1 $\times$ 10^4 m³ second ?(Holdsworth & Collinson, 1988) (estimate from Kinderscoutian deltas)***	***Medium-grained sandston e to siltstone***	***0-15m***	***~1-100 km²***	***Low, but evident***	***Greater than 3750 km² (crude estimate based on the lateral extent of bayhead delta facies)***	***3, or greater than 3***	***5.63×10^4 m³***
Westphalian Pennine Basin (Proximal-distal lacustrine delta of (Guion et al., 1995)	***Greater than 1 $\times$ 10^4 m³ second? (Holdsworth & Collinson, 1988)***	***Medium-grained sandston e to siltstone***	***0-8m***	***~1-100 km²***	***?0 to low***	***Greater than 40000 km² (based on conservative estimate of Westphalian outcrop area of 200km by 200km)***	***Much greater than 3***	***3.2×10^5 m³***
Present day Atchafalaya Basin (Van Heerden & Roberts, 1988; Tye & Coleman, 1989b)	***3.6×10^3 m³ second***	***Fine-grained sandston e to mudstone***	***3-6m***	***Eventually Greater than 20 km² (still prograding (Van Heerden & Roberts, 1988)***	***~0.5m average***	***~4700 km²***	***1***	***2.82×10^4 m³***

Table 4 Comparison of parameters of bayhead deltas in the Marsdenian and Westphalian Pennine Basin, and the Atchafalaya Basin (Present-day), Louisiana

5.1.4. *The fluvial plain depositional system*

The fluvial plain depositional system comprises a sheet-form succession 10-35m thick that can be traced laterally across the basin for up to 20-100 kilometres. The base of this depositional system is always sharp, and erodes into a diverse range of facies associations, including proximal and distal mouthbar and basinal mudstone associations (Figure 5.5). The stacked channel association provides a consistently thick succession, and forms a volumetrically important part of the Marsdenian Basin infill. Channel abandonment and interfluve associations are thickest, and pedogenic processes dominate where the stacked channel association thins. Amalgamated palaeosol horizons form successions up to 5m thick towards the basin-margins, and on structural highs (upper part of Figure 5.6).

Meandering fluvial systems have not been previously identified in Marsdenian Pennine Basin, but this study recognises laterally accreting channelized sandstone in the top-most part of the fluvial plain depositional system at Fletcher Bank Quarry (SD805164) and Leicester Mills Quarry (SD619164). The laterally accreting channelized sandstone always occurs above 30+m of channelized cross-bedded sandstone, and just below the capping palaeosol facies.

These stacked channel associations should not be confused with distributary channels within the delta front mouthbar association. Mouthbar distributary channels comprise single storey channels less than 10m thick that pinch-out laterally over several kilometres. These smaller channels are not spatially or temporally continuous like channelized sandstones within the fluvial plain depositional system (Figure 5.7).

CASE STUDY 3: PRESENT DAY AND ANCIENT ANALOGUES FOR THE FLUVIAL PLAIN DEPOSITIONAL SYSTEM

Previous workers have likened the river systems that fed the Namurian Pennine Basin to modern day, bedload dominated, braided fluvial systems (Shackleton, 1962; Bristow, 1987; Bristow, 1993). Such systems have been intensively researched, and the processes of transport, architecture, controls and deposits of braided systems are well understood (Miall, 1977; Miall, 1978; Orton & Reading, 1993). Braided river systems form multi-storey sand bodies, and comprise broad, shallow channels, which generally transports sand-grade sediment as bedload. The present day South Saskatchewan River is a good analogue for the depositional environment of the fluvial plain depositional system. Within individual channels, bars possess a variety of forms, and individual channel fills fine-upwards; while floodplains develop on areas adjacent to channels (Miall, 1978). The fluvial plain depositional system forms laterally extensive, multi-storey sheet sandstones; individual units of which can be correlated across the Pennine Basin (the Midgley Grit forming the base example).

The volume of sediment within the fluvial plain depositional system of the Midgley Grit is large. If we take a conservative thickness of 20m for the fluvial plain component of the Midgley Grit, and assume that it extends lateral over 100 kilometres square; an estimated 2×10-11 m^3, or 20 km^3 of sediment is present. Considering that this is a conservative estimate of the volume of sediment within the fluvial plain depositional system, the actual volume of sediment within the fluvial part of the Midgley Grit may be much larger. Additionally, the Midgley Grit is one unit of several fluvial plains within the lower Marsdenian, suggesting the volume of sediment transported by this river system throughout the Marsdenian was very large. Comparisons between the area covered by the fluvial plain and delta-top environments in modern day rivers reveals the true scale of the Marsdenian fluvial systems (Table 5).

Similar braided river systems have been identified in the Kinderscoutian (Collinson, 1968a; Collinson, 1968b; McCabe, 1975; McCabe, 1977) and Yeadonian (Bristow, 1987). Estimates of the dimensions of the fluvial channel systems suggest that river systems feeding the Kinderscoutian delta complexes were slightly smaller than the Mississippi River (McCabe, 1977). This interpretation is based on the thickness of channels observed in both ancient and modern delta systems. However, the discharge and mean grainsize calibre of the Namurian and Mississippi Rivers is dissimilar. While the Kinderscoutian deltas are commonly comprise medium to coarse grained sediment; siltstone is the mean grainsize of the present day Mississippi River (Orton & Reading, 1993). The use of grainsize as an idicator of the rate of fluvial deposition as coarse-grained deltas may from in areas where the distance from hinterland to depocentre is short. This is because very little is understood about the effects of primitive Carboniferous vegetation on weathering in the hinterland (Bristow, 1988). It is possible that hinterland soils were poorly developed; which reduced the amount of chemical denudation, increased the amount of surface run-off, maintaining the mineralogical immaturity of sediments in the fluvial plain depositional system. It is fair to assume that the flow regime of Namurian Rivers was high, due to the large scale of the bedforms observed within channels (up to 40m thick, Collinson 1968b; McCabe 1977; Hampson 1997). The lack of channel abandonment facies also suggests that channels were rarely abandoned.

Within the Rough Rock (Yeadonian), the paucity of evidence for channel lateral accretion, and the dominance of medium-coarse grained cross-bedded strata suggest that the fluvial component of this delta was deposited in a braided river system (Bristow 1988; Bristow 1993). Channel fill sandstones in the fluvial plain depositional system of the Marsdenian partly comprises deposits transported by meandering fluvial channels (the laterally accreting channelized sandstone association), and these occur at the top of this assembledge, below the overlying transgressive surface. Such channel networks are formed as the river gradient, or graded profile, is lowered. This is likely during the later stages of fluvial belt genesis, when aggradation of fluvial deposits leads to the development of a flat fluvial plain (see Chapter 6).

Deltaic system	**Area covered by delta plain (fluvial plain depositional system**
Ganges/ Brahmaputra delta	105,641km²
Mississippi River delta	28,568km²
Niger River delta	19,135km²
Mackenzie River delta	13,000km²
Nile River delta	12,512km²
Midgley Grit (Marsdenian, Pennine Basin)	**10,000km²**

Table 5 Comparison between land area covered by the delta/ fluvial plain environment during the deposition of the Midgley Grit and during the deposition of the present day deltas (Orton & Reading, 1993).

5.2 Summary of depositional systems chapter

Facies in association	Facies association	Facies depositional system	Order of associations in depositional system
High density channelized turbidite sandstone (and/ or) Low-density turbidite sandstone	Turbidite association	Basinal floor depositional system	(youngest) Turbidite association Basinal mudstone association (and interdigitated with turbidite association) (oldest)
Offshore marine band mudstone (and/or) Offshore black mudstone (and/or) Offshore prodelta mudstone	Basinal Mudstone association		
Allochthonous drift coal/ swamp, Overlying; Waterlogged muddy palaeosol (or) Waterlogged heterolithic palaeosol.	Mouthbar top association	Delta front mouthbar depositional system	(youngest) Mouthbar top association. Inter mouthbar trough association Proximal mouthbar front association Distal mouthbar front association (thin) Basinal mudstone association (oldest)
Interdistributary bay (and/or) Shallow water restricted marine bay	Inter mouthbar trough association		
Bedload dominated sand rich mouthbar (common) Cross bedded sandstone laterally altering to; Shoaling mouthbar. (and/ or rare) Tidally influenced sand dominated mouthbar	Proximal mouthbar association		
Buoyancy dominated fine-grained mouthbar Laterally altering to; Offshore prodelta silty-mudstone	Distal mouthbar association		
In-situ coal (or) Allochthonous/ swamp coal; (common) Waterlogged muddy palaeosol (or) Waterlogged heterolithic palaeosol, Cross bedded sandstone (and/or) Tidally influenced cross-bedded sandstone laterally altering to: Overbank flood deposits/ crevasse splay	Bayhead delta top association	Bayhead delta depositional system	(youngest) Bayhead delta top association Bayhead delta front association Bayhead prodelta association (oldest)
Buoyancy dominated mouth-bar (and/or) Tidally influenced silt dominated mouthbar, laterally altering to; Shallow water open marine bay (or) Interdistributary bay (and rare) Lower Shoreface above fair-weather wave base	Bayhead delta front association		
Offshore marine band mudstone (and/or) Offshore black mudstone (and/or) Shallow water restricted marine bay (and/or) Offshore prodelta silty mudstone.	Bayhead prodelta association		
Allochthonous drift/ swamp coal, (rare) Waterlogged heterolithic palaeosol (or) Waterlogged muddy palaeosol	Channel abandonment association	Fluvial plain depositional system	(youngest) Interfluve association Channel abandonment association Laterally accreting channelized sandstone association Stacked channel association (oldest)
Cross bedded sandstone (and/or) Fluvial barform (and/or) Tidally influenced cross bedded sandstone (and/or) Tidally influenced barform	Stacked channelized sandstone association		
Laterally accreting point bar (and localised channels of) Cross bedded sandstone	Laterally accreting channelized sandstone association		
In situ coal (and/ or) drift/ swamp coal (and/or) Leached sand rich 'ganister' palaeosol (rare) Overbank flood deposits/ Crevasse Splay	Interfluve association		

Table 6 Facies, facies associations and depositional systems in the Marsdenian Pennine Basin.

PART 3: BASIN ANALYSIS

Chapter 6: Sequence stratigraphic interpretation

6.1. Introduction

As noted in the introduction, marine bands are regional correlatable transgressive horizons that form horizons bounding upward-coarsening cyclothems. Marine bands therefore provide the bounding surfaces that form a 'genetic' sequence framework (Galloway, 1989). The genetic sequence framework provided by marine band maximum flooding surfaces allows analysis of the sand-rich successions between marine bands by sequence stratigraphic techniques (*sensu.* Posamentier *et al.* 1988; Posamentier & Vail 1988; Van Wagoner *et al.* 1988; Van Wagoner *et al.* 1990). The combination of genetic and sequence stratigraphic interpretations produces an increasingly refined correlation framework. The remaining part of this study will use sequence stratigraphic concepts, with the aim of elucidating the mechanisms and controls that control deposition.

Correlation within the Silesian Pennine Basin prior to the early 1990's relied on the identification of goniatite bearing marine bands (Ramsbottom *et al.,* 1962; Calver, 1968; Ramsbottom *et al.,* 1978) and the lithostratigraphic correlation of the sandstone units that lay between marine bands (Ramsbottom *et al.,* 1978). Much discussion was initiated by the suggestion that sequence stratigraphic concepts may be used the correlate and elucidate controls on deposition within the Namurian Pennine Basin (Read, 1991; Collinson *et al.,* 1992). Sequence stratigraphic concepts are based on continental margins (Posamentier *et al.,* 1988; Posamentier & Vail, 1988) and different depositional responses to subsidence rate, sediment supply, palaeobathymetry and eustacy were expected in the 'intra-cratonic' Pennine Basin (Collinson *et al.,* 1992). Despite these complexities, the application of sequence stratigraphic concepts to the Pennine Basin has led to an increased understanding of the controls and products of Namurian delta deposition (Martinsen, 1993; Church & Gawthorpe, 1994; Wignall & Maynard, 1996; Hampson *et al.,* 1996; Hampson *et al.,* 1997; Hampson, 1997; Jones & Chisholm, 1997).

Surface analogues for major Carboniferous gas reservoirs in the Southern North Sea, have been identified within the sand-rich fills of incised valleys (Hampson *et al.,* 1997; Hampson *et al.,* 1999). The generation of incised valley fills in response to a falling and rising relative sea-level is now recognized as a fundamental element of deltas where deposition is influenced by sea-level fluctuations. Criteria for the identification of incised valleys include; i) the recognition of an unconformable surface with erosive relief, ii) extensive, laterally correlatable interfluves, iii) sediments above the sequence boundary show a basinward shift in facies, iv) the incised valley shows evidence for a composite fill, often encompassing more than one high-order sequence, v) presence of tidal facies within the incised valley fill, because during rising sea-level, incised valleys commonly become estuaries.

The identification of sequence boundaries is often difficult, especially when lateral change in depositional environment leads to variation in their characteristics; (e.g. the change between fluvial (R1) and sharp-based mouthbars (R4) within mouthbar systems). The following section forms a paper that focuses on the interval between R2a1 and R2b3. It describes the low- and high-order sequences identified in the Marsdenian, and suggests which mechanisms control deposition within the Pennine Basin. These are dealt with in greater detail in the Chapter 10. There is some re-iteration of the assumptions and background between the sequence stratigraphic paper and the remainder of this chapter, although it is kept to a minimum. The interval between the base of the Midgley Grit and R2b5 is not dealt with in this paper, but is summerised in Section 6.3. Additionally, Table 7 summarises the characteristics of all high-order sequences dealt with in this study.

6.2. PAPER: A sequence stratigraphic investigation of the fluvially-dominated, tidally-influenced incised valley fills and forced regressive mouthbars in the early Marsdenian (Namurian) of the Pennine Basin, U.K.

6.2.1. Abstract

This paper presents a detailed sequence stratigraphic model for the mouthbar-dominated systems of the early Marsdenian (Namurian). The Marsdenian forms a key interval of the Namurian Pennine Basin, lying between the turbidite-fronted delta systems of the Kinderscoutian, and the coal-bearing coastal plain environment of the Westphalian. The R2a1 marine band *(Bilinguites gracilis)* marks a low-order glacio-eustatic flooding event, which transgressed Kinderscoutian-aged deltas in the eastern Pennine Basin, forming a submerged platform over which early Marsdenian mouthbar systems prograded. Detailed analysis of field, bore and core data reveals higher-order transgressive and regressive surfaces exist within, and can be correlated through, these early Marsdenian mouthbar dominated deltas. Both low- and high-order eustatic regressive events combine to form shoreline perpendicular incised valleys on the submerged platform during the low-order forced regressive and lowstand systems tract. Sediment bypassing these valleys is deposited in outflow-dominated mouthbars, which propagate through the valley, and onto the submerged platform. Each high-order sequence forms a regressive surface of fluvial erosion (RSFE) in the proximal valley, and a regressive surface of marine erosion (RSME) in the distal valley and submerged platform. Bypassed sediment deposited on RSME's in the distal valley forms offlapping high-order forced regressive mouthbars, which progressively downcut and coalesces with the previous RSFE/ RSME to form a master-sequence boundary. In areas where the platform is entirely absent, sediment bypass across the master-sequence boundary reaches the platform/ slope break, where turbidite systems transport sediment into deeper-water areas. Rising low-order sea-level generates accommodation within the valley, forcing deltafront regression, and stemming bypass to the platform edge. The flooding of the valley creates a shallow coastline embayment, which is infilled by tidally influenced/ reworked mouthbars. The point when low-order base level rise attains its maximum rate coincides with the formation of a marine band on the submerged

platform. Dependant on the combined influence of clastic supply, water-depth and salinity, the low-order maximum flooding surface contains (in increasingly shallow water) ammonoid, marine bi-valve/ brachiopod or marine ichnofaunas.

However, the presence of syn-depositional faults, seismic dewatering structures, and condensed footwall-high deposits suggests non-eustatic parameters (subsidence, tectonic and inherited bathymetry) also have an influence on the distribution and architecture of both high- and low-order key surfaces, systems tracts and facies associations. These non-eustatic parameters are added to the sequence stratigraphic model, and provide a clear picture of the controls, architecture and depositional systems that affected deposited within the Marsdenian Pennine Basin.

6.2.2 Background

The Pennine Basin was formed by crustal extension along Caledonide fault trends during the late Devonian and early Carboniferous (Dinantian), in response to northward directed subduction during Rheno-Hercynian orogeny (Leeder, 1982; Gawthorpe, 1987; Fraser & Gawthorpe, 1990). The basin comprises a series of smaller fault-defined sub-basins (Paper Figure 1) that were infilled by fluvial-deltaic systems throughout the Namurian and Westphalian. During the Namurian, the post-rift stage of basin development was dominated by delta systems in which sedimentary provenance is inferred to have been dominantly sourced from the decaying Caledonian-Appalachian Mountains to the north/ north-east (Drewery *et al.,* 1987).

The early Namurian Pennine Basin fill comprises turbidite-fronted deltas deposited in initially deep-water, which were overlain by deltas deposited in progressively shallow water throughout the middle and late Namurian (Paper Figure 2). The Kinderscoutian-aged turbidite-fronted delta system subsequently entered the Pennine Basin, forming a palaeo-bathymetric high, over which the younger Marsdenian and Yeadoninan–aged mouth bar dominated deltas prograded. This shallowing trend continued into the Westphalian succession, where coal-rich delta top environments predominated.

High-frequency marine transgressions are represented by ammonoid-bearing horizons (marine bands), which can be used as correlative horizons within the Namurian stratigraphy of Northern Europe (Ramsbottom, 1977). The accuracy of this marine band chronostratigraphic framework provides a testing ground for the application of high-resolution sequence stratigraphic techniques. This paper integrates sedimentological data from the Marsdenian and a key surface/ horizon framework with the aim of creating a detailed correlation based on the sequence stratigraphic analysis of this fluvial-mouth bar dominated succession.

The application of sequence stratigraphy to the study of deltaic systems involves the identification of systems tracts, which are deposited in an identifiable water depth. Characteristic environmental indicators (facies, facies associations, key surfaces/ horizons) within these systems tracts allow the comparison of relative sea-level throughout the sequence. The utilisation of sequence stratigraphic techniques on outcrop data is well-documented in the Upper Carboniferous of northern England (Maynard, 1992; Church & Gawthorpe, 1994; Church, 1994; Zaitlin *et al.,* 1994; Wignall & Maynard, 1996; Hampson *et al.,* 1996; Hampson *et al.,* 1997). This paper describes a high-resolution correlation framework comprising marine band, regressive and palaeosols that are utilised in the correlation of Marsdenian deltaic cycles.

6.2.3 A traditional perspective of basin fill and correlation within the Namurian Pennine Basin

The initial episode of deposition within the Pennine Basin was dominated by turbidite-fronted deltas from the early-Namurian to Kinderscoutian Stage (Collinson *et al.,* 1977). Following the *Bilinguites gracilis* (R2a1) flooding event (which forms the base of the Marsdenian Stage), mouthbar fronted delta systems dominated the early Marsdenian clastic infill (Paper Figure 3). The five deltaic successions identified between R2a1 and R2b5 often form regionally correlatable cyclothems, the upper sand-rich part of each cyclothem corresponding to a mappable lithostratigraphic unit, which is often named after a type locality. These units are the Alum Crag Grit (between R2a1-R2b1), the Readycon Dean Flags, Scotland Flags and East Carlton Grit (between R2b1-R2b2), the Woodhouse Flags (between R2b2-R2b3), the Midgley Grit (between R2b3-R2b4) and the Helmshore Grit (between R2b4-R2b5). This uses field, core and borehole sections from across the Pennine Basin to create a sequence stratigraphic interpretation for the succession between the R2a1 marine band and the regional erosive surface lying above the R2b3 marine band.

Between R2a1 and R2b3 several extensive mouthbar dominated sandstone sheets prograded into the Pennine Basin from the north. Mouthbar cycles are typically sharp-based and contain isolated fluvial sandstones, with laterally correlatable to palaeosols horizons. Each mouthbar-fluvial sandstone unit comprises between 10-35m of stacked fluvial channel or proximal mouthbar deposits, possessing a basal unconformity with up to 45m erosive relief, which represents a regional regional regressive surface (Church 1994; Church & Gawthorpe 1994; Wignall & Maynard 1996). Transgressive and regressive surfaces occur within each sand-rich deltaic succession, implying that a high-order relative sea-level fluctuation also influenced delta system deposition. A palaeosol, overlain by a marine band often caps the each deltaic sheet, suggesting emergence prior to marine transgression.

SHRIMP analysis (Riley *et al.,* 1993) of zircon grains in bentonite (ash-fall tuff) bands provides absolute date horizons within the Namurian fill of the Pennine Basin, which allows the estimation of average marine band periodicities (65 000a). This data allows the extrapolation of dates for the Kinderscoutian-Marsdenian stage boundary (the R2a1 marine band) at 312.75Ma and the Marsdenian-Yeadonian stage boundary (the G1a1 marine band) at 312.23Ma. Prior to the early 1990's correlation in the Silesian Pennine Basin relied on the identification and correlation of ammonoid bearing marine bands (Ramsbottom *et al.,* 1962; Calver, 1968; Ramsbottom *et al.,* 1978). Between marine bands lithological units were used to sub-divide the stratigraphy into coarsening-upwards 'cyclothems' or 'minor cycles' (Ramsbottom, 1980; Holdsworth & Collinson, 1988) and groups of cyclothems termed 'mesothems' (Ramsbottom, 1977).

The suggestion that sequence stratigraphic concepts may be used the correlate stratal packages and elucidate controls on deposition within the Namurian of the Pennine Basin initiated much discussion (Read, 1991; Collinson *et al.,* 1992). The 'intra-cratonic' Pennine Basin (Collinson *et al.,* 1992) experienced very different responses to subsidence rate, sediment supply, palaeobathymetry and eustacy, than continental margins on which sequence stratigraphic concepts are based (Posamentier *et al.,* 1988) (Posamentier & Vail, 1988).

Despite these initial complexities, the application of sequence stratigraphic concepts to the Pennine Basin has led to an increased understanding of the controls and products of Namurian delta deposition (Martinsen, 1993; Church & Gawthorpe, 1994; Wignall & Maynard, 1996; Hampson *et al.,* 1996; Hampson *et al.,* 1997; Hampson, 1997; Jones & Chisholm, 1997). This paper aims to use sequence stratigraphic concepts to explain the relationships and controls between the distribution of facies associations and depositional sequences within the early Marsdenian Pennine Basin.

6.2.3.1 Depositional elements of the Marsdenian Pennine Basin

Several facies schemes exist for Namurian Stages (Walker, 1966a; Collinson *et al.,* 1977; McCabe, 1977) and Westphalian deposits within the Pennine Basin (Guion & Fielding, 1988; Guion *et al.,* 1995). Early Marsdenian facies are similar to those described by previous Carboniferous researchers, and encompass a range of depositional environments from interfluve to basin-floor environments (Paper Figure 4). Detailed sedimentological analysis of the Marsdenian has recently identified new facies, including those interpreted as tidally-influenced (Brettle *et al.,* 2001).

Stratigraphic surfaces are used in this study to delineate sedimentary packages that can be correlated across the Pennine Basin. These surfaces can are transgressive (representing deepening events), regressive (representing shallowing events) or pedogenic (representing emergence; Paper Figure 4). Such surfaces are interpreted as the product of fluctuating relative sea-level, and their characteristics reflect the water-depth and depositional environment in which they formed. For example, in an up-dip area a transgressive event may be represented by a waterlogged palaeosol horizon within interfluve facies, whereas it may be represented by a marine ichnohorizon, or a marine band down-dip. However, the correlation of time-equivalent horizons across facies boundaries requires a degree of care, especially in emergent environments, as leached pedogenic horizons are often overprinted by waterlogged signatures (see example in Paper Figure 9).

Although fluctuating relative sea-level had a strong influence, other mechanisms also altered the style of the basin fill. For example, changes in the size of drainage basin via tectonic or geomorphic mechanisms (Summerfield, 1991) variations in hinterland climate (Blum, 1990) and avulsion within the basin receiving sediment can potentially characteristics of the delta deposits (Roberts, 1998). The combination of these mechanisms suggests that deltaic systems are susceptible to rapidly fluctuating sediment supply, cannibalisation and period abandonment.

6.2.4 Accommodation and its control on facies and key surface variability

The waxing and waning of Gondwanan ice-masses has been the proposed cause of Namurian relative sea-level fluctuation since Wanless & Shepard (1936) first identified upward- coarsening cyclothem units in the Appalachian Basin of the US. Cyclothems and marine bands have been identified throughout the Euramerican province (Wanless & Shepard, 1936; Ramsbottom *et al.,* 1978; Heckel, 1986) and these are suggested as the product of fluctuating glacio-eustatic sea-level. Comparison suggests glacio-eustatic highstands and lowstands in the palaeo-equatorial Euramerican province are concomitant to periods of interpreted of low and high Carboniferous Gondwanan ice-mass volume (Veevers & Powell, 1987). Due to the tie between control (ice-cap) and product (cyclothem and marine band) it is very tempting to suggest that glacio-eustatic modulation was the primary control on Namurian relative sea-level.

Recent studies also suggest Namurian-aged deposits experienced periods of falling relative sea-level almost as regularly as that record in the marine bands (Wignall & Maynard, 1996; Hampson *et al.,* 1997; Hampson, 1997; Hampson, 1998). Regressive surfaces are identified throughout the Namurian-aged succession of the UK. Such regressive surfaces (often inferred as sequence boundaries) often sharply delineate distal delta-front facies from overlying proximal mouthbar and fluvial-tidal facies. These surfaces represent unconformities, and possess between 0-25m of basin-wide incision; which can be correlated for up to 60 km (Hampson *et al.,* 1997). In up-dip areas these regressive surfaces correlate with coeval interfluve facies, comprising leached palaeosol facies. Additionally, an individual regressive surface may be traced from shallow (mouthbar-dominated) to deeper-water (turbidite-dominated) environments (Paper Figure 5) Regressive surfaces therefore cross sub-basin boundaries, which have different supply points, fill-timings and subsidence rates, suggesting it is unlikely that they are produced by thermal or active basin subsidence.

Duration estimates for the Namurian have recently decreased from 26 Ma (Lippolt *et al.,* 1984) to 5.5 Ma (Riley *et al.,* 1993). Estimated rates of thermal subsidence for the Pennine Basin have therefore increased from 0.015 mm per year (Leeder & McMahon, 1988) to 0.07 mm per year. Such subsidence rates imply that the deposition of an averaged 30m thick deltaic lithostratigraphic unit would have taken approximately 0.4 Ma if thermal subsidence alone was responsible for accommodation generation. With the Namurian duration estimated at 5.5 Ma in, a maximum number of

13 lithostratigraphic units and 14 marine bands would have formed within the Pennine Basin. However, Riley *et al.* (1993) identified 52 marine bands within the Namurian, implying that thermal subsidence alone cannot account for the amount of accommodation generated during this period. Regional thermal subsidence therefore appears to produce broad areas of background accommodation, which accounts for an estimated maximum of 26% of the accommodation generated during the Namurian. Subsidence driven by sediment compaction could also have generated accommodation. However, compaction driven subsidence, driven by dewatering, generates patchy, rather than regional areas of accommodation (Coleman, 1981; Van Heerden & Roberts, 1988; Tye & Coleman, 1989b; Coleman *et al.,* 1998a).

Evidence for contemporaneous activity along fault-zones appears to be localised and subtle (Collinson, 1988). Recent fieldwork has interpreted massive de-watering structures in the Woodhouse Flags (see description of the R2b3 sequence) as tectonically triggered (Brettle, 2001). These de-watering structures occur in an area of fault-splays off the NW-SE trending Craven fault-system. Restricted marine facies (containing *Zoophycos* and sponge spicules) appear to outcrop on tectonically uplifted footwall highs (Waters *et al.,* 1996a) which run parallel to the Craven fault-system. Both de-watering structures and restricted marine facies therefore suggest syn-depositional tectonic activity was restricted in areas adjacent to major basin-bounding faults.

The dominance of shallow-water mouthbar environments during the Marsdenian relies on the regional shoaling influence inherited from the Kinderscoutian-aged delta system. This shallow-water area was between 30-50 km in width, and lay within the Huddersfield sub-basin. During the Marsdenian this area maintained water-depths up to an estimated 40m (varying from emergent to just below storm wave base) and became a locus of shallow-water deposition. The submerged Kinderscoutian delta therefore produced a platform upon which extensive shoreline transgression and regression occurred during periods of relative sea-level modulation. There is no doubt the Kinderscoutian represented a period of high discharge and sediment input into the Pennine Basin, providing ample capacity to rapidly fill accommodation generated by subsidence or glacio-eustatic flooding events (Collinson, 1988). Some discharge fluctuations into the Pennine Basin had the potential to modulate deposition rates and the degree of avulsion (Broadhurst *et al.,* 1980; Broadhurst, 1988), but are unlikely to deposit units that can be traced basinwide. The repetition and highly-correlatable nature of each transgressive-regressive surface, at both low and high order, suggests it is unlikely that allocyclic-driven sediment supply variations or delta-lobe abandonment is responsible for the formation of transgressive-regressive surfaces.

Marine-bands mark severe and regional trangressive events, that cannot be accounted for by the low-rates of thermal subsidence or the localised influence of active tectonics. It is therefore intuitive to suggest glacio-eustatic sea-level fluctuations provided the most significant short-term mechanism for generating and destroying accommodation during the Namurian. In comparison, thermal subsidence probably produced a broad background increase in accommodation, whereas activity along fault-zones only had a local influenced on accommodation.

6.2.5 The application of sequence stratigraphic concepts to the early Marsdenian

Within the Pennine Basin cyclothems form broadly coarsening-upwards cycles bounded by marine band flooding surfaces or their cryptic equivalents (Paper Figure 6) (Waters *et al.,* 1996b). Marine bands form a 'genetic stratigraphic' framework (Galloway, 1989) within which cyclothems can be traced. The genetic stratal framework provided by marine bands allows the subsequent analysis of each cyclothem by sequence stratigraphic techniques, thereby producing an increasingly refined stratal framework. Parasequences (*sensu.* Van Wagoner *et al.* 1990) comprise genetically related and relatively conformable bedsets bound by flooding surfaces or their correlated equivalents (mud-rich palaeosols or hiatal surfaces). The application of the parasequence definition is not clear-cut within the Marsdenian, as extensive regressive surfaces (the regressive surfaces of marine/ fluvial erosion) occur between what would otherwise be considered parasequence-bounding flooding events (Paper Figure 6). This necessitates the modification of the sequence stratigraphic framework to take account for the absence of readily identifiable parasequences.

6.2.5.1 High-order sequences

The base of each high-order sequence comprises a regressive surface with between 2-10 m of erosive relief, which is overlain by mouthbar and/ or distributary channel facies, suggesting accommodation generation occurred after the formation of the regressive surface. A transgressive surface caps the mouthbar/ distributary channel facies, and forms an upper bounding surface to each sand-rich sequence (Paper Figure 6). Regressive surfaces can erode through the underlying transgressive surface and coalesce with an older regressive surface, complicating the identification of entire high-order sequences.

In order to utilise high-order sequences it is necessary to explain the characteristics within and mechanisms effecting shallow-water mouthbar dominated settings. Two distinct types of mouthbar are recognised the early Marsdenian. They either comprise ii) mud-sand lithologies with broadly layer-parallel bedding deposited by buoyant plume-processes (see in core in Paper Figure 7), or ii) sand-rich lithologies, with clinoform bedding with depositional dips up to 21° up to 20m in height, (see Paper Figure 7 and 17 for example) and deposited dominantly by bedload processes. Analysis of mouthbar laminae thickness data from mouthbar buoyant plume deposits using Fourier time-series techniques suggests a semi-diurnal and diurnal tidal influence exists within the early Marsdenian (Brettle *et al.,* 2001). Both mouthbar types typically occur in each high-order sequence; bedload-dominated mouthbar deposits commonly lie at the base of each unit, and buoyancy-dominated mouthbars cap the bedload-dominated mouthbar (Paper Figure 6). The relationship between fluctuating relative sea-level and the influence this has on the architecture of high-order sequences is best illustrated by a model illustrated in Paper Figure 8.

During periods of rising and high relative sea-level (Paper Figure 8.1 & 8.2), accommodation exceeded the amount of sediment input, and high-order sequences developed a weakly progradational or aggradational stacking pattern. Under such circumstances shoreline regression is said to be 'normal' (*sensu.* Posamentier & Morris, 2000). As relative sea-level fell, the platform became increasingly shallow; sediment input exceeded accommodation and delta progradation was 'forced' across the platform. 'Normal' mouthbar progradation therefore ceased at the onset of falling relative sea-level (Paper Figure 8.3) and accommodation destruction resulted in the generation of a regional regressive surface.

Under these circumstances, relative sea-level fall erodes into the underlying highstand platform forming the regressive surface of marine erosion (RSME) basinward of the mouthbar, and the regressive surface of fluvial erosion (RSFE) within the fluvial setting (*sensu.* Plint & Nummedal, 2000). The RSME forms a broad, shallow incised valley that bypasses sediment across the platform and deposits it basinward of the incised valley mouth. As relative sea-level fell, and the graded-profile was raised, the RSFE migrated headward and cannibalised up-dip parts of the delta. Reworking of the delta increased the amount of sediment bypassing the incised valley, forcing the deposition of an offlapping *bedload-dominated mouthbar* (Paper Figure 8.4). The deposition of bedload dominated mouthbars during relative sea-level fall resulted in an abrupt basinward shift in facies and an offlapping stacking pattern similar to the 'forced regressive' deposits observed in many shoreface and deltaic successions (Plint, 1988; Hadley & Elliott, 1993; Hunt & Tucker, 1993; Ainsworth & Crowley, 1994; Plint & Nummedal, 2000). During the period when relative sea-level was at its lowest, the potential existed for bypassed sediment to overshoot the mouthbar system, and the platform/slope break, to form a submarine turbidite fan.

The subsequent relative sea-level rise forced the point of basin fluvial input landward and lowered the graded-profile; while synchronously generating a thin sheet of accommodation within the incised valley. This flooding event (*sensu.* a transgressive tidal /marine ravinement surface *sensu.* Zaitlin *et al.* 1994) is rarely marked by indicators of marine conditions in the proximal mouthbar, but it is often represented by a marine ichnofauna horizon (i.e. Paper Figure 9) in the distal mouth bar. The restoration of the graded-profile decreased sediment cannibalisation on the up-dip delta, lessening the amount of coarse-grained sediment entering the basin. The combination of lowered sediment input, decreased supply system inertia force and the sheet-like geometry of the accommodation allowed the progradation of the thin *plume-deposited mouthbars*. The identification of subtle tidal indictors within *plume-deposited mouthbars* corroborates the notion that the incised valley allowed the amplification of a tidal-range, and the preservation of tidally influenced deposits (Allen, 1981; Allen & Posamentier, 1993; Zaitlin *et al.,* 1994). Each mouthbar dominated high-order sequence therefore lies within a distal incised valley, and comprises an initial bedload-dominated forced-regressive component in the basinward part of the valley, and a later transgressive-highstand plume-dominated mouthbar deposited in the proximal incised valley.

The predictable shift in facies distributions and key stratal surfaces in each high-order sequence is similar to those described by larger-scale (3^{rd} order) sequence-scale relative sea-level modulations (*sensu.* Van Wagoner *et al.* 1988). Such similarities permit the definition of systems tracts within each high-order sequence. In this classification, the basal regressive surface (RSME/ RSFE) is equivalent to the sequence boundary. Overlying the regressive surface of erosion, bedload-dominated mouthbars generated during falling and low relative sea-level are equivalent to the forced-regressive and lowstand systems tract. In cases where it is not eroded by the next regressive surface of erosion, the transgressive surface within each high-order sequence corresponds to a maximum flooding surface. Overlying the high-order maximum flooding surface, buoyancy-dominated mouthbars were deposited as relative sea-level rose during the late transgressive to highstand systems tract. These repetitive high-order sequences share architectural characteristics to those described by Plint & Nummedal (2000; their figure 4), and a similar separation of each high-order maximum flooding surfaces by a RSME/ RSFE. Because of this, high-order sequences are therefore used rather than parasequences as the fundamental stratal unit in the early Marsdenian.

6.2.5.2. Low-order sequences

The section detailing the history of correlation within the Namurian noted regional-scale unconformites have been correlated throughout various intervals of the Carboniferous Pennine Basin. Such unconformites are the product of several coalesced regressive sufaces, which formed during a period of low-order falling and lowstanding relative sea-level (Paper Figure 10).

As illustrated in Paper Figure 10 the superposition of a low- and high-order base level curve implies that high-order sequence boundaries coalesce on the low-order falling limb, forming regional unconformity. Each regional unconformity is therefore a low-order or 'master' sequence boundary (*sensu.* Posamentier & Allen, 1999), which forms a diachronous base to the sand-rich upper unit of each cyclothem. Each sequence boundary has the potential to develop at differing rates, due to the variations in discharge of the fluvial system forming individual high-order regressive surfaces. This notion suggests allocyclic mechanisms may control the rate of high-order regressive surface formation and ultimately the rate of sequence boundary genesis.

During low-order falling stage and lowstand systems tract, bypass of the platform areas allowed the deposition of turbidite-dominated high-order sequences basinward of the incised valley mouth. Rising low-order relative sea-level generated accommodation within the incised valley, allowing the deposition of the low-order transgressive system tract. During the low-order transgressive system tract, basinward stepping mouthbar and fluvial dominated high-order sequences were deposited on the master sequence boundary. Marsdenian low-order incised valley fills often contain several transgressive and regressive surfaces (i.e. 'a composite fill' *sensu.* Zaitlin *et al.* 1994), implying high-order relative sea-level fluctuated throughout the low-order transgressive systems tract (Paper Figure 11). The low-order transgressive system tract culminated in the formation a marine band, which corresponds to a low-order maximum flooding surface, and the highest magnitude high-order maximum flooding surface.

Incised valleys are formed by low-order relative sea-level falls and should possess correlatable interfluve horizons, the but dominance of sub-aqueous deposition, and the sporadic nature of the outcrop implies candidate interfluve areas are often not exposed. However, the stacking of successive high-order sequences during the formation of the low-order incised valley does allow the definition of systems tracts; which will be used in the remainder of the paper to describe the fill within each low-order incised valley.

6.2.6. Sequence stratigraphic analysis of the early Marsdenian

Sequence stratigraphic concepts shift the emphasis of the correlation scheme away from marine bands (low-order maximum flooding surfaces) to master sequence boundaries; although marine bands are still acknowledged as providing the most explicit correlation framework in the Namurian. When combined with the marine band framework, sequence stratigraphy provides an accurate means of interpreting the heterogeneity within the sand-rich parts of each low-order sequence. As marine bands remain unique biostratigraphic marker horizons, they are therefore used in the naming of each low-order sequence.

In this section, the sequence stratigraphic concepts described above are integrated with facies schemes and palaeocurrent data (Paper Figure 12), allowing the assessment of basin-fill architecture during each sequence

6.2.7. The upper R2a1 sequence

Forming the base of the study interval, the R2a1 (*Bilinguites gracilis*) marine band represents the Kinderscoutian-Marsdenian Stage boundary (Ramsbottom, 1981; Cleal & Thomas, 1996) and can be correlated throughout Namurian deposits of the Pennine Basin, Southern North Sea, South Wales and south-western Ireland (Ramsbottom *et al.,* 1978). R2a1 forms the maximum flooding surface within high-order sequence 1 (HS1). HS1 is the only high-order sequence of the R2a1 sequence that is Marsdenian-aged, and comprises up to 10m of basinal mudstone facies. It is underlain by fluvial-mouthbar facies of the Upper Kinderscout Grit (Davies & McLean, 1996; Waters *et al.,* 1996a; Hampson, 1997; Waters, 1999), which form the lower part of HS1 (not illustrated on Figure 5 or 11).

6.2.7.1. Interpretation of the R2a1 sequence

The R2a1 marine band is a pan-European flooding event that is traced across all sub-basin bounding faults within the Pennine Basin. This suggests that R2a1 represents a major eustatic-driven flooding event, during which the generation of accommodation is independent of tectonic subsidence. The omnipresence of basinal mudstone facies within HS1, suggests the flooding event represents the drowning of the underlying Kinderscoutian deltas, and the establishment of relatively deeper-water conditions across the basin (Holdsworth & Collinson, 1988).

The strong retrogradational stacking pattern of the upper part of HS1 suggests the R2a1 marine band represents the low-order maximum flooding surface within the R2a1 sequence, where HS1 comprising the low-order highstand systems tract.

6.2.8. The R2b1 sequence

The regressive surface at the base of HS2 represents the initial regressive surface marking the base of the R2b1 sequence. In the Rossendale and Bowland sub-basins HS2 is represented by up to 30m of lowstand turbidites, which entered the basin from the northwest (Paper Figure 12.1). These turbidites are thickest between the footwall high of the Quernmore Fault to the west and Kinderscoutian deltas to the east (Paper Figure 1), where they form the *Alum Crag Grit* (Paper Figure 5.1 & 13). In the Huddersfield sub-basin, HS2 comprises a RSME overlain by discrete sharp-based mouthbars, which according to palaeocurrent analysis, were fed by a sediment supply point entering the basin from the north-east. The upper high-order transgressive-highstand unit of HS2 comprises an ammonoid-bearing high-order flooding surface overlain by basinal mudstone facies in the Rossendale and Bowland sub-basins, and a marine ichnofauna horizon overlain by distal mouthbar/ basinal mudstone facies in the Huddersfield sub-basin.

A second RSME marks the base of HS3, eroding into, and coalescing with, the regressive surface at the base of HS2. Like HS2, HS3 also reveals a marked dichotomy in depositional processes; turbidites were deposited in the Rossendale and Bowland sub-basins, whereas mouthbar were deposited in the Huddersfield sub-basin. Isolated exposures of HS3 also contain buoyancy-dominated mouthbar facies, revealing laminae that occur in cyclic bundles of two and four (described by Brettle *et al.* 2001). Forming the maximum flooding surface within HS3, the R2b1 *(Bilinguites bilinguis)* marine band is of reduced lateral extent and contains a lower density ammonoid fauna when compared to R2a1. The high-order transgressive-highstand unit overlying R2b1 comprises a thin succession of basinal mudstone facies, which are less than 5m in thickness.

6.2.8.1. Interpretation of the R2b1 sequence

During HS2 and HS3, mouthbar parasequences dominated the Huddersfield and Harrogate sub-basin fill, whereas turbidite-dominated parasequences dominated the Rossendale and Bowland sub-basins. Analysis of the palaeogeographic and tectonic map suggest the turbidites deposits in the Rossendale and Bowland sub-basins were funnelled through the deeper, under-filled parts of the basin. The variation between mouthbar facies in the east of the basin, and turbidite facies in the west reflects the dichotomy in water-depth that existed at the onset of the R2b1 sequence.

The tying of HS2-3 mouthbar deposits in the Huddersfield sub-basin to age equivalent turbidite-deposits in the Rossendale and Bowland sub-basins is difficult; due to the discontinuity of outcrop and the paucity of exposures covering significantly thick stratal sections. At Alum Crag (SD636280) HS2 is represented by a stacked succession of turbidite channels overlain by planar-bedded lower-density turbidites (Paper Figure 13). This exposure is the only real insight into the deeper-water, turbidite dominated deposits of the Rossendale sub-basin. The presence of channelised turbidite deposits overlain by bedding parallel deposits probably represents an initial period of continuous bypass during the low-order forced-regressive to lowstand system tract, and gradual accumulation and backstepping of the turbidite system during the low-order late-lowstand and early transgressive systems tract.

The high-order maximum flooding surface within HS2 represents a period of accommodation generation and a landward shift of the supply input point. The subsequent parasequence comprises basinal mudstone facies in the Rossendale sub-basin and interdistributary bay and thin plume-dominated mouthbar deposits in the Huddersfield sub-basin. In the Huddersfield sub-basin, these are located above the axis of the HS2 high-order forced-regressive/ lowstand mouthbar unit, and are interpreted as distal low-order transgressive systems tract mouthbars that re-used supply systems developed during the high-order transgressive-highstand (Paper Figure 11.1). The high-order maximum flooding surface within HS2 is therefore inferred as the low-order initial flooding surface (*sensu.* Zaitlin *et al.* 1994) within the incised valley system.

The RSME at the base of HS3 represents a second, more extensive regression, which is extensively overlain by low-order forced regressive systems tract mouthbar deposits in the Huddersfield sub-basin, and low-order lowstand systems tract turbidite deposits in the Rossendale and Bowland sub-basin. This regressive surface coalesces with the regressive surface at the base of HS2, forming the first significant low-order master sequence boundary (Posamentier & Allen, 1999; Posamentier & Morris, 2000). The sequence boundary defining the base of the incised valley/ turbidite system within the R2b1 sequence is therefore diachronous, as it comprises regressive surfaces formed during the relative sea-level fall of HS2 and HS3.

The mouthbars of HS2 and HS3 comprise bedload-dominated and fining-upwards plume-dominated types (Paper Figure 5.2, 7 & 14). Such variations could be the product of distance from a distributary channel source, but the regionally correlative nature of each high-order sequence makes it increasingly plausible that such variations are the product of deposition during varying stages of each high-order sequence (i.e. bedload dominated during forced-regression, plume-dominated during transgression). It is within the buoyancy-dominated mouthbar succession that the cyclic mouthbar laminae described by Brettle *et al.* (2001) are interpreted as the product of a subtle tidal influence on the mouthbar plume during deposition.

The restricted geographic extent and the low-density of marine fauna contained within R2b1 suggests that it represents a less significant transgressive surface than R2a1. However, it forms the low-order maximum flooding surface within the R2b1 sequence and separates the low-order transgressive system tract incised valley fill from the low-order highstand systems tract represented by the upper part of HS3.

6.2.9. The R2b2 sequence

A RSME marks lower bounding surface of HS4, which incises the R2b1 marine band in the southern Huddersfield sub-basin. HS4 reveals a marked fining-upwards trend; comprising sand-rich high-order forced-regressive-lowstand systems bedload dominated mouthbar facies, which are overlain by mud-rich interdistributary bay and plume-deposited mouthbar facies deposited above the high-order maximum flooding surface. In the southern Huddersfield sub-basin, the mouthbar-dominated succession between HS5-7 is collectively known as the *Readycon Dean Flags*. In parts of the northern Huddersfield sub-basin the high-order forced-regressive/ lowstand system of HS4 comprise bedload-dominated mouthbars (known as the *Scotland Flags*), which contain sets of conjugate syn-depositional faultlets in isolated outcrops (i.e. Great Mount Quarry SE01002750; Paper Figure 15.1 & 16). Turbidite facies dominate HS4 in the Rossendale and southern Huddersfield sub-basins (Collinson *et al.*, 1977; an unnamed lithostratigraphic unit, Paper Figure 5.1). A cryptic flooding surface, separates the high-order forced regressive-lowstand system of HS4 from the overlying thin and retrogradationally stacked high-order transgressive-highstand system in the Blackburn Tunnel section. However, the high-order transgressive-highstand system of HS4 is removed by the base of HS5 across most of Pennine Basin, which forms the second RSME in this sequence. In the northern Huddersfield and central Rossendale sub-basins HS5 comprises fining-upwards bedload- (forced regressive-lowstand system) to plume-dominated (transgressive-highstand system) mouthbars and fluvial facies (known as the *East Carlton Grit*), while bedload-dominated mouthbars continued to dominate the southern Huddersfield sub-basin. Isolated down-dip orientated outcrops of the high-order forced-regressive/ lowstand system of HS5 in the southern Huddersfield sub-basin reveal clinoform surfaces, representing a prograding mouthbar succession (Paper Figure 17). The outcrop at Digley Quarry reveals at least two sets of shoaling mouthbar delta-front clinoforms bounded by an erosive surface. The lower set of clinoforms represent the deposits of a basinward encroaching bedload-dominated mouthbar, with isolated channels of fluvial transported sandstone, whereas the upper clinoforms represent shoaling deposits and the passage of the mouthbar crest.

The RSME at the base of HS6 erodes into HS5, and is overlain by high-order forced-regressive/ lowstand systems comprising fluvial deposits in the Rossendale and northern Huddersfield sub-basins and bedload-dominated mouthbar deposits in the southern Huddersfield sub-basin. The high-order maximum flooding surface within HS5 is an easily correlated surface, as it is overlain by distal bayhead rather that buoyancy dominated mouthbar facies. HS6 attains its maximum thickness of +20m in the Rossendale sub-basin, but is less than 5m thick across the majority of the basin. The base of HS7 comprises a RSFE overlain by up to 10m of fluvial sandstone in the Huddersfield sub-basin, where it incises HS6, locally eroding through the high-order maximum flooding surface in HS6. In the Rossendale sub-basin HS7 thickens to 10m, and is represented by a fining-upwards bayhead mouthbar in the Heywood Bore (SD8385108976). The R2b2 *(Bilinguites bilinguis)* marine band is of reduced lateral extent and contains a lower density ammonoid fauna in comparison with R2a1, but outcrops more extensive than R2b1. The R2b2 marine band is succeeded by the high-order transgressive-highstand systems basinal mudstones of HS7, outcropping northward to a greater extent than the high-order transgressive-highstand of HS3 (overlying R2b1).

The 'Keighley Bluestone' lithofacies in the northern Huddersfield sub-basin is a condensed equivalent of the entire R2b2 sequence. It comprises a thin (30cm to 5m thick) chertified mudstone and containing *Hyalostelia* sponge spicules and *Zoophycos* ichnofaunas (see Paper Figure 18 for outcrop example). Analysis of the structural maps reveals the Keighley Bluestone is restricted to a faulted footwall high, formed by faults that splay off the South Craven Fault System (Paper Figure 16).

6.2.9.1. Interpretation R2b2 sequence

The base of the R2b2 sequence is formed by the RSME underlying HS4 in the northern Huddersfield sub-basin. This coalesces with the RSFE/ RSME at the base of HS5 in the southern Huddersfield sub-basin, forming a low-order master sequence boundary. Across the Pennine Basin, the delta-front created by the R2b2 sequence regressed to a greater extent than during the R2b1 sequence. Such a basinwide delta progradation implies the amount of sediment supply exceed the amount of accommodation generated by the combination of Dinantian rifting and transgressive events between R2a1 and R2b1.

The bedload-dominated mouthbar high-order forced regressive/ lowstand systems of HS4 in the Huddersfield sub-basin form the low-order forced-regressive to lowstand incised valley fill. The coeval turbidite facies in the Rossendale sub-basin are suggested as the high-order forced-regressive to lowstand deposits, which entered the basin from the north-west.

The thin fining-upward trend to plume-dominated mouthbar deposits in the upper part of HS4 represent the onset of the high-order transgressive to highstand systems tract. High-order maximum flooding surfaces are generally marked by a fining-upward trend more often than a distinct transgressive event indicator of the within the high-order sequences of the R2b2 sequence (see discussion of the high-order sequences). The high-order maximum flooding surface within HS4 represents the weak low-order initial flooding surface (*sensu.* Zaitlin *et al.* 1994) within the R2b2 sequence, and marks the onset of the low-order transgressive systems tract. This is eroded by the base of HS5, which coalesces with the base of HS4, forming a master sequence boundary. HS5 marks the onset of mouthbar deposition across the Huddersfield and Bowland sub-basin boundary as delta progradation filled accommodation, allowing the development of significant areas of shallow-water northwards of the east-west line of section.

Paper Figure 17 reveals that the forced-regressive-lowstand system of HS5 possesses a set of shoaling clinoform surfaces that dip in a southern (basinward) direction, which decrease in gradient through the section. Similar shoreline geometries are observed in forced regressive deposits, and are inferred as the product of deposition in an increasingly shallow water (Posamentier & Morris, 2000).

The RSME/ RSFE underlying the high-order forced-regressive-lowstand systems of HS6 erode through HS5 and into HS4. This is overlain by fluvial deposits in the Rossendale sub-basin and the bedload-dominated mouthbar deposits of the southern Huddersfield sub-basin. The high-order maximum flooding surface within HS6 generated sufficient accommodation to establishment a laterally extensive shallow-water setting into which bayhead delta system bayhead deltas prograded. This high-order sequence marks a significant event in the early Marsdenian, when flooding of the incised valley forced a marked backstepping of the supply system and the generation of accommodation within the incised valley. A RSME forms the base of HS7 across the majority of the basin, but is replaced by a RSFE in the central Huddersfield sub-basin. The dominance of an RSFE in this area is probably the product of proximity to the main axis of clastic input at the time (Paper Figure 12.2).

Interestingly, the condensed marine conditions that prevailed during the deposition of the Keighley Bluestone occur in an area relatively close to the basin-entry point, which was protected from deltaic influx throughout the R2b2 sequence. The lack of clastic input was probably the result of bypass around the structural high on which the Keighley Bluestone outcrops, suggesting that the basin floor bathymetry part controlled water-depth and facies distribution. The syn-sedimentary faulting observed in Great Mount Quarry (Paper Figure 15) suggests some tectonic activity in the basin, which may have been related to movement on faults synthetic to the Morley-Campsell or Denholme Clough Fault zones (Paper Figure 16). The extent of outcrop and the density of ammonoid faunas suggests that the low-order transgression associated with the formation of R2b2 is of greater the that of R2b1, but less than R2a1.

6.2.10. The R2b3 sequence

The regressive surface underlying HS8 comprises a RSME across almost the entire basin, and marks the base of the R2b3 sequence. This surface was subsequently coalesced with regressive surfaces at the base of HS9, HS10 and HS11. HS8 comprises a fining-upwards turbidite succession in the Rossendale sub-basin (Paper Figure 5.1) and fining upwards bedload to buoyancy-dominated mouthbars in the northern and mid-Huddersfield sub-basin (Paper Figure 18). In the southern Huddersfield sub-basin (Paper Figure 19; Mouselow Quarry; SK02459515), the high-order lowstand system comprises thin channelised turbidites with mudstone rafts (maximum dimension up to 1m), formed from reworked turbidite-offshore mudstone facies. The equivalent high-order forced-regressive/ lowstand system in the eastern Huddersfield sub-basin comprises a thicker (up to 20m) bedload-dominated mouthbar succession (Paper Figure 5.2; locality 15, Ponden Clough). The high-order maximum flooding surface within HS8 is formed by a flooding surface, and is overlain by retrogradationally-stacked basin floor facies in the Rossendale sub-basin and inter-mouthbar bay deposits in the southern

Huddersfield sub-basin (see Brettle *et. al.* 2001, his Figure 3.2). The base of HS9 is marked by the second RSME in the R2b3 sequence; which is overlain by bedload-dominated mouthbars and tidally influenced distributary channel facies in the northern and mid-Huddersfield sub-basin. In the northern Huddersfield sub-basin several localities exposing HS9 lie adjacent to the Aire-Valley and Morley-Campsell faults, and contain large-scale dewatering structures (Paper Figure 15.2). The exposure at Bingley Road Quarry (SE05353810) comprises metre-scale bar mouthbar foresets, with each coset appearing to have experienced differing degrees of sub-vertical oriented deformation.

In the southern-most Huddersfield and eastern Rossendale sub-basins HS9 comprises a series of turbidite sandstones interleaved with silty- and a channelised turbidite facies. The base of HS10 is marked by a third regressive surface; and comprises a RSME overlain by bedload-dominated mouthbar facies in the Rossendale sub-basin, and a RSFE overlain by fluvial facies in the northern Huddersfield sub-basin. The high-order forced regressive system of HS10 represents the most basinward extent of an RSME prior to the R2b3 marine band.

In the northern Huddersfield sub-basin the mouthbar- and fluvial-dominated succession lying between the basal regressive surface of HS8 and R2b3 are known as the *Woodhouse Flags*. Within the middle Huddersfield sub-basin, the R2b3 sequence is retrogradationally stacked and forms a thinner, backstepping unit in comparison to the R2b2 sequence. In this area the high-order maximum flooding surface within HS10 separates the underlying mouthbar-dominated *Woodhouse Flags*, from the bayhead succession of HS10 and HS11. In this area HS11 comprises mouthbar facies, and a prograding shoreface succession at Fosters Delph (SE02152717; Paper Figure 20.1). On average, wave ripple crests in the lower-shoreface trend 75-255º, an orientation that should lie concordant to a flow oscillation trending ENE-WSW. The equivalent high-order forced-regressive/ lowstand system of HS11 in the northern Huddersfield sub-basin is dominated by bayhead deltatop facies comprises waterlogged palaeosols and overbank facies (Paper Figure 20.2; Branshaw Quarry, SE03284005).

The R2b3 *(Bilinguites bilinguis)* marine band rarely contains ammonoid faunas, but is typically marked by marine ichnofaunal horizons (typically *Zoophycos, Teichichnus* and *Rhizocorallium)* in mouthbar or bayhead facies (Paper Figure 20.2;

Branshaw Quarry) or a calcareous horizon in interdistributary bay facies (Deanhead Clough, SE02601445; Paper Figure 14). The high-order transgressive-highstand system of HS11 overlies the R2b3 marine band, and comprises basin floor mudstones in the Rossendale and Bowland sub-basins, and distal bayhead delta mudstones in the Huddersfield sub-basin. HS11 delineates the upper limit of the succession discussed in this paper, and is incised by a regressive surface at the base of the next sequence, marking the base of the *Midgley Grit*.

6.2.10.1 Interpretation R2b3 sequence

The incised valley created during the R2b3 sequence is aggradationally to retrogradationally stacked on the R2b2 sequence, because R2b2 forced a more extensive shoreline regression than the R2b1 flooding event. However, the outcrop available in the deep-water southern Huddersfield sub-basin (Paper Figure 19; Mouselow quarry) reveals a thicker R2b3 sequence turbidite succession, suggesting R2b2 did not reduce bypass through the incised valley. The presence of large mudstone rafts within the R2b3 sequence suggests that up-dip slope and turbidite systems were cannibalized during this period.

The master sequence boundary forming the base of the R2b3 sequence comprises the coalesced regressive surfaces of HS8, HS9 and HS10, implying that it forms the most diachronous of all the early Marsdenian low-order sequence boundaries. The increase from two to three coalesced regressive surfaces explains why the R2b3 sequence boundary has greater erosive-relief in the Huddersfield sub-basin when compared to the R2b1 and R2b2 sequence boundaries. In comparison, the R2b3 incised valley fill is also more heterogeneous; reflecting the greater fluctuations in accommodation that occurred within this incised valley.

The variations in basin bathymetry inherited from Dinantian extension and Namurian basin fill continued to influence deposition during the R2b3 sequence. Turbidite facies suggest deeper-water conditions prevailed in the underfilled Rossendale and Bowland sub-basin between HS8 and HS9, until accommodation infill forced mouthbar deposition during HS10 and HS11. In the shallower waters of the Huddersfield sub-basin co-indices with the increased occurrence of fluvial facies in HS10, reflecting the increased proximity of the basin-input point and high sediment input rates in this area (Paper Figure 12.3). In this area, HS9 and HS10 comprise basal regressive surfaces overlain by high-order forced regressive bedload-dominated mouthbar deposits. Unlike previous mouthbar-dominated high-order sequences, HS9 does not contain a high-order transgressive system, because as noted above, the proximity of the basin-input point probably resulted in an increased clastic input into the Huddersfield sub-basin, and the decreased likelihood that the high-order transgressive system tract would develop.

The large-scale dewatering structures of HS9 only occur in the northern-most part of the Huddersfield sub-basin, in proximity to the South Craven Fault Zone. The apparent severity of the dewatering event suggests such structures were probably triggered by frequent syn-depositional seismic activity along the boundary of the Askrigg block-Pennine Basin. Dewatering structures similar to those seen in Paper Figure 7 are observed in the Northumberland basin to the north of the Askrigg-Alston Block (Leeder, 1987), and are also inferred as the product of seismic activity along active fault-zones. The area in where seismic-dewatering occurred lies close to where the Keighley Bluestones outcrop (Paper Figure 16), further corroborating the notion that the northern margin of the Huddersfield sub-basin was actively influenced by tectonic processes. The juxtaposition of basin-bounding fault trends (Paper Figure 1) and Kinderscoutian-aged palaeogeographic reconstructions (Collinson, 1968a) reveals that a sediment-supply conduit lay in the area bounding the north-eastern Huddersfield sub-basin and the Askrigg Block during the late Kinderscoutian. This suggests that fluvial supply pathways initiated during the Kinderscoutian were reused and modified, creating an antecedent drainage pathway during the Marsdenian. Entrenching of the fluvial systems through the faulted transfer zone suggests the Craven Fault zone formed a *synthetic interbasin transfer zone* (Gawthorpe & Hurst, 1993), linking the Askrigg block to the north with the Pennine basin to the south. This basin-input point in the north-eastern Huddersfield sub-basin supplied sediment since the Kinderscoutian; gradually infilling syn-rift accommodation and ultimately leading to the dominance of bayhead, overbank and waterlogged palaeosol facies during HS11.

At the beginning of the Marsdenian the western Pennine Basin was relatively underfilled due to of absence of a Kinderscoutian-aged input. Due to the greater amount of accommodation in the Rossendale and Bowland sub-basins, the Marsdenian-aged deltas therefore experienced a protracted infill and lower rates of shoreline regression. Differential rates of shoreline regression between the eastern and western Pennine Basin forced the formation of a broad coastal embayment by the time the R2b3 sequence was deposited (Paper Figure 12.3). This notion is corroborated by the ENE-WSW oriented shoreface at Fosters Delph Quarry (Paper Figure 20.1), which with the absence of open-marine connections, is interpreted as the product of storm and wind-driven waves amplified by the shoreline geometry (Paper Figure 12.3).

The R2b3 marine band forms the high-order maximum flooding surface in HS11. This is difficult to correlate in the Huddersfield sub-basin, and identification relies on the used of calcareous and marine ichnofaunal horizons. The high-order transgressive-highstand system of HS11 consists of basinfloor mudstones in the Rossendale and Bowland sub-basins, suggesting R2b3 forced a transgression of the delta shoreline. The delta-shoreline probably prograded rapidly after R2b3, as it contains abundant suspension deposited micaceous-siltstone and carbonaceous debris in the Huddersfield sub-basin.

The regressive surface separating HS11 from the progradationally stacked mouthbar facies of the overlying sequence marks the base of the Midgley Grit, and the lowest high-order regressive surface at the base of the R2b4 sequence.

6.2.11 Discrepancies with previous analyses and sequence stratigraphic interpretations

Wignall and Maynard (1996) presented the most recent interpretation of the early Marsdenian fill of the Pennine basin. Several findings of this study contradict those of Wignall and Maynard, and these need to be addressed. The application of sequence stratigraphic concepts relies on the interpretation of detailed sedimentological data, and several discrepancies have come to light when particular sections and sequence stratigraphic interpretations are compared. Importantly, Wignall and Maynard classified marine bands as parasequence bounding surfaces, used parasequences as the highest-order sequence stratigraphic unit and could therefore not appreciate the significance of the high-order regressive surfaces lying between marine bands.

Marine bands form the primary correlative horizon framework, and mis-identification or failure to identify them can lead to miscorrelation. At the classic Pule

Hill section (SE03151015; Cleal & Thomas 1996) Wignall & Maynard identify the basal marine band as R2b1. This interpretation is contrary to other interpretations that suggest this marine band is the R2a1 marine band (Bisat, 1924), as it contains *Bilinguites gracilis* (N.Riley, pers. comm.) and lies less than 5m above the top of the upper Kinderscout Grit (inferred from the dip-slope to the west). The mis-identification of R2a1 suggests Wignall and Maynard's classification of the overlying silty horizon as the Readycon Dean Flags, and the R2b2 marine band, is incorrect. Unfortunately, Wignall & Maynard (1996) also missed both R2b2 (represented by a marine band) and R2b3 (represented by a calcareous siltstone with *Planolites*) in the Deanhead Clough section (between SE02751452 and SE02601445). They therefore inferred that the 62m of exposed mouthbar facies in Deanhead Clough corresponds to the Readycon Dean Flags, when it actually represents both Readycon Dean and Woodhouse Flags (Paper Figure 14).

In the southern-Huddersfield sub-basin Wignall & Maynard (1996) inferred the existance of a NW-SE trending incised-valley in their equivalent to the R2b2 sequence, which contained the turbidite-dominated 'Howels Head Flags'. They suggested the 'Howels Head Flags' representing the feeder-system of the Alum Crag Grit, which lies in the eastern Rossendale sub-basin. Several lines of evidence suggest this connection does not exist. At several localities, including Crowden Great Brook (SE06300320), flute cast and tool marks in the 'Howells Head Flags' constantly suggest turbidity palaeocurrents trended in a south to southeasterly direction. This study suggests the cryptic equivalent of the R2b1 marine band occurs within the Alum Crag Grit lithostratigraphic unit (Paper Figure 5.1) defined by Price *et al.* (1963) and Collinson *et al.* (1977). The 'Alum Crag Grit' therefore comprises two turbidite sandstones; a lower unit (HS2-HS4), which is part exposed at Alum Crag; SD63632804), and an upper unit (HS6-HS8) recorded in a tunnel section examined by Collinson *et al.* (1977). If the 'Howells Head Flags' did form part of a turbidite feeder system bypassing the southern Huddersfield sub-basin and depositing the Alum Crag Grit, one would expect at least some evidence of turbidite systems between HS2-HS4 in this area. This is not the case, and to the contrary, the R2b1 sequence consists of a thin condensed section throughout in this area.

Palaeocurrents at the Alum Crag exposure suggest that the Alum Crag Grit was fed from a northerly, and not a southeastern source, as suggested by Wignall and Maynard. A southeastern source appears improbable, as turbidite currents would have had to traverse the delta-front slope, against the force of gravity (Paper Figure 12.3). In addition, the proposed south-eastern source area remained an underfilled basinal area until the after the R2b5 sequence (Jones, 1980).

6.2.12. Summary; sequence stratigraphic model for a mouthbar-dominated incised valley on a submerged platform

The three sequences in the early Marsdenian contain, and are delineated by, key surfaces and system tracts. This section consolidates these features and present a generic model describing the processes that may have influenced the formation (Paper Figure 21) and the typical features of an early Marsdenian mouthbar dominated incised valley fill.

6.2.12.1. The sequence boundary

Early Marsdenian master-sequence boundaries comprise more than one high-order regressive surfaces; each the product of a high-order relative sea-level fall (Paper Figure 21). The positioning of each high-order sequence on the falling low-order limb is critical, as this dictates the extent to which each high-order regressive event is either suppressed or extended. This superposition controls the rate, amplitude and extent of each regressive surface of fluvial erosion; with the initial regressive events producing a broad but shallow incised valley, into which later RSFE's become entrenched. Similar entrenchment processes are suggested to control the spatial distribution of the active Colorado River alluvial system during the Pleistocene to Recent (Blum, 1993; Blum & Törnqvist, 2000). To this extent, incised valley development is fixed during the initial high-order regressive event, and becomes increasingly confined. This pattern coincides with a period when the source of bypassed sediment altered, from a dominantly hinterland source (*sensu.* the conveyor belt model, Blum & Törnqvist 2000) to include that produced by headward incision (*sensu.* the vacuum cleaner model, Blum & Törnqvist 2000). The continuous outflow-dominated nature of deposition within the Pennine Basin suggests that even though headward had the potential to increase the amount of sediment bypassing the incised valley, hinterland drainage remained the dominant sediment source throughout the throughout the Namurian.

Perhaps more significant than reworking parts of the up-dip delta, headward erosion confined flow within the incised valley, focused bypass down the valley axis, raised discharge rates, and potentially increased the rate of RSME propagation and mouthbar deposition on the submerged Kinderscoutian platform. The RSFE therefore initially forms a broad feature (Paper Figure 21, point 1) that becomes increasingly confined, whereas the RSME is initially confined and broadens (Paper Figure 21, points 2-4). In the Huddersfield sub-basin, where deposition predominantly occurs on the submerged Kinderscoutian platform, this suggest the rate of emergence and deltafront progradation may have rapidly increased during the later low-order forced regression, as the fluvial feeder system incised headward.

While the master sequence boundary formed in areas of bypass, leached palaeosols developed synchronously in interfluve areas. Within the basin, these have a fairly low-preservation potential, as subsequent high-order transgressions have the potential to either rework or overprint interfluve areas with gleyed signatures during periods of raised watertable (i.e during a low-order transgressive systems tract). Similar overprinted interfluve facies identified in the Pennsylvanian of eastern Kentucky, are interpreted as sequence boundaries that underwent wetting during a rising water-table associated with a rising in sea-level (Aitken & Flint, 1994; Aitken & Flint, 1996). Comparisons may also be drawn with floodplain deposits in the Dunvegan Formation of Alberta, where ‘compound’ palaeosols are attributed to an episode of falling, then rising base-level (McCarthy *et al.,* 1999). Indeed, the classification by McCarthy *et al.* (1999) of floodplains as formed under episodes of either high, low or zero to negative accommodation corresponds to interpretations made in the Pennine Basin.

Additionally, the formation of waterlogged palaeosols (Paper Figure 21, point 3) upon mouthbar emergence may be overprinted by leached palaeosols (Paper Figure 21, point 4) formed on exposed mouthbars within the late low-order forced regressive/ early lowstand systems tract. Such palaeosols are have not been identified within the Pennine Basin as i) they have been overlooked, ii) the subsequent base-level fall occurs rapidly so waterlogged palaeosols are not generated, or, iii) the severity of leaching associated with high-order base-level fall completely removes any gleyed signature. Either way, the sporadic distribution of the waterlogged palaeosols in the Huddersfield sub-basin suggests the presence of low-accommodation conditions (*sensu.* McCarthy *et al.* 1999), whereas up-dip areas experienced zero to negative accommodation.

6.2.12.2.The maximum flooding surface

Marine bands are condensed intervals that represent the point where low-order relative sea-level reached its maximum rate of rise. At the onset of the Marsdenian, the R2a1 transgressive event flooded the Kinderscoutian delta in the north-eastern Pennine basin. The extent of the R2a1 transgression was such that it formed a wide submerged platform across the north-eastern Pennine Basin. The R2b1, R2b2 and R2b3 low-order maximum flooding surfaces were less significant in comparison; only raising relative sea-level to the extent of forcing weak retrogradational stacking of highstand systems tracts of HS3, HS7 and HS11. Assuming each high-order relative sea-level oscillation was constant in amplitude and wavelength, the marine band must signify the point at which the maximum rate of high-order relative sea-level rise lies closest to the maximum rate of low-order relative sea-level rise on the rising limb. Each marine band therefore represents the high-order flooding event of greatest amplitude within the low-order sequence, and the point when the combined rate of low- and high order relative sea-level rise reached its maximum. Like each transgressive surface, the marine band varies characteristics within the incised valley. Within the distal valley, on the submerged shelf, or on the basin floor, it is represented by the typical ammonoid bearing marine-band. With progressively increased proximity to the valley head, decreased water depth and salinity and increased water turbidity imply the density of ammonoid and *Dunbarella* decreased, whereas *Lingula* and marine ichnofauna increased. At the point where sea-level rise forces shoreline regression over previously emergent areas, transgressive reworking forms a transgressive surface of marine erosion (TSME *sensu.* Plint, green book). Although this feature is not currently identified within the early Marsdenian incised valley it is a significant in basins that experience basin wave-reworking; i.e. parts of the Cretaceous Western Interior basin Plint etc), and may be a significant key surface in the proximal areas of the palaeo-valley.

6.2.12.3. Systems tract distribution

In the distal incised valley and on the submerged platform areas, sediment bypassing the valley during the low-order regressive systems tract was deposited in offlapping forced mouthbars (Paper Figure 21). The forced-regressive mouthbar illustrated in Paper Figure 17 reveals at least two sets of shoaling mouthbar clinoforms bounded by erosive surfaces. The lower clinoform set represents the deposits of a basinward encroaching bedload-dominated mouthbar, with isolated channels of fluvial transported sandstone, which is eroded and overlain by the upper clinoforms that represent the shoaling deposits of a mouthbar crest. This set of shoaling mouthbar front clinoforms dip in a southerly (basinward) direction, and decrease in gradient through the section in a similar manner to those described by Posamentier & Morris (2000), that are inferred as the product of deposition in an increasingly shallowing waterdepths during a high-order forced-regressive system tract.

Where sediment bypassed the submerged platform, and passed down the platform/ shelf break, it deposited thick forced-regressive to lowstand turbidite lobes. These form the Alum Crag Grit in the deep-water areas of the western Pennine Basin, but the lack of outcrop in the southern Pennine Basin means that these are not observed in their most distal extent. The turbidite feeder channels representing the R2b2 and R2b3 sequence in the southern-most part of the Pennine Basin are exposed at Mouselow Quarry (Paper Figure 19), implying that this area comprised a delta-front slope environment, with bypass continuing in a southerly direction, where the deposition of lowstand turbidite lobes is suggested occurred.

Valley fill commenced as low-order relative sea-level rose, creating an initial flooding surface and generating accommodation within the valley. This forced the rate of accommodation generation to outpace sediment supply; resulting in delta shoreline regression, decreased bypass, and the cessation of basin-floor turbidite deposition. The accommodation created by the low-order transgression also allowed the onset of shallow-water bayhead delta deposition within the incised valley. These transgressive systems tract bayhead mouthbars are typically thinner than those formed during the forced regressive and lowstand systems tract, and often occupy areas away from the axis of forced regressive systems tract deposition. Deposition of the forced regressive systems tract occurred below sea-level, and created a topographic high on the sea-bed. The combination of this topography and raised sea-level may have forced mouthbar deposition off the central incised valley axis during the highstand and transgressive systems tract, explaining why distal bayhead deltas expanded rapidly across the platform.

Flooding of the incised valley led to the generation of a wide coastal embayment, in which the tidal range was amplified. As predicted by incised valley fill (Zaitlin *et al.,* 1994) and drowned river valley/ estuarine models (Allen, 1991; Dalrymple *et al.,* 1992; Allen & Posamentier, 1993) the intensity of tidal reworking processes increases during the transgressive systems tract. During the low-order transgressive systems tract, the valley remained shallow, allowing the tidally-influenced deposition and reworking, and explaining the presence of the cryptic tidal signatures identified within the buoyancy-dominated mouthbar and distributary channel facies in HS3 and HS8-10 (Brettle *et al.,* 2001). Fluvial-dominated valleys such as those of the Marsdenian were influenced by tidal processes, but were also susceptible to overprinting and reworking by fluvial currents. Fluctuations in river discharge or fluvial system retrogradation during high-order flooding events may have potentially suppressed the fluvial regime and allowed tidal facies development. However, the low-order maximum flooding surface caused extensive deepening in the distal and mid-valley, forcing the cessation of tidal processes in these areas. In the areas of the valley where the transgressive systems bayhead deltas became shoaling or emergent, crevasse-splay and waterlogged palaeosols developed. These reached their most basinward in the Huddersfield basin during the HS11 (Paper Figure 20.2).

After the maximum flooding surface the platform and distal valley experienced a protracted fill by the distal highstand system tract deltafront mouthbars, whereas the proximal and mid-valley succumbed rapidly to input. For example, parts of HS11 lying in the middle section of the valley (Paper Figure 12.3) comprise basin floor and interdistributary facies containing a large proportion of suspension deposited siltstone and carbonaceous debris (including those on Paper Figure 14). This area lies fairly close to an input point, implying the R2b3 flooding event only forced weak deltafront and supply point regression.

6.2.12.4. Modifications to the model: The effect of antecedent drainage, bathymetry, subsidence and tectonics

The sequence stratigraphic model above represents an ideal scenario, where glacio-eustatic fluctuations are the sole mechanism behind sequence formation. As noted throughout this paper, other mechanisms including inherited bathymetry, antecedent supply systems, subsidence and tectonics modify the extent and amount of accommodation within Marsdenian incised valley fills. While the effect of non-glacio-eustatic mechanisms on accommodation are subtle, they nevertheless explains some incised valley fill characteristics.

6.2.12.5. Antecedent drainage and bathymetry

When viewed in the broadest sense, the Marsdenian forms part of a continuum of basin-infill which commenced with Dinantian rifting and carbonate deposition, was effected by the input of Namurian deltas, and culminated in syn-rift accommodation infill with the development of a coal-forming delta plain during the Langsettian. The Marsdenian lies at a key interval in the history of the Namurian Pennine Basin, when the initially deeper-water turbidite fronted deltas were replaced by shallower-water mouthbar conditions (Paper Figure 2). This trend was initiated during the early Namurian, when fluvial supply systems began to fill the Huddersfield sub-basin, and culminated in the formation of a broadly emergent platform by end of the Kinderscoutian stage (Paper Figure 1). Therefore the amount of accommodation available at the start of the Marsdenian was distinctly dichotomous; with the north-western (Brandon *et al.,* 1998) and southern Pennine Basin remaining largely underfilled, and the north-eastern area predominantly filled. Whereas shoaling mouthbar systems were deposited on the platform created by older deltas, turbidite systems gradually prograded into the underfilled parts of the basin.

It is obvious that the Huddersfield sub-basin received much sediment during the Kinderscoutian and Marsdenian. The north-eastern most areas of Huddersfield sub-basin became part of an antecedent drainage pathway during the Marsdenian, through which sediment input continued. This clastic entry point is situated close to an area where several major fault lineaments converge with the Craven Fault system, suggesting the fault zone provided a long-lived supply pathway through which the Huddersfield sub-basin was supplied with sediment. It also appears possible that the north-western sediment input point was also formed close to an intersection of the Quernmore and

Craven Fault Zones (Paper Figure 12.1). However, the paucity of constant outcrop adjacent to the Craven Fault system does not allow the identification of footwall or hangingwall fans. In such settings either subaerial or submarine fans might be expected to represent the product of locally derived debris, or bypass over fault-controlled topographic relief. However, the observation of antecedent drainage conduits occurring in conjunction with fault-zones does suggest tectonic and inherited topography did have a significant influence on the spatial distribution of Marsdenian depocentres. In the deeper-water Rossendale and Bowland sub-basins the pre-existing basinfloor topography provided by the footwall high of the Quernmore Fault and the tilted footwall high of the Rossendale block appeared to control turbidite distribution (Gawthorpe, 1987; Lee, 1988). It seems likely that turbidites were funnelled through the deeper parts of the basin, which lay between the Kinderscoutian platform to the east and the individual structural highs in the east. This may account for the limited outcrop extent of the Alum Crag Grit and the overlying temporal equivalents of the Readycon Dean and Woodhouse Flags (Paper Figure 5).

6.2.12.6. Subsidence

Although thermal subsidence produces a basinwide increase in accommodation, it only accounts for one-quarter of accommodation generated during the entire Namurian. It seems unlikely that a incised valley fill would be significantly influenced by such a broad accommodation increase to any significant extent. Subsidence due to compaction probably had a greater effect on the amount of accommodation. The dominance of sand-rich lithologies implies de-watering probably occurred rapidly after deposition and loading. Perhaps more significant might be compaction associated with the high-pressure loading of the Kinderscoutian delta during delta progradation during the Marsdenian.

6.2.12.7. Tectonics

Direct evidence suggesting active tectonics affected the Marsdenian is restricted to isolated outcrops containing seismic dewatering structures and syn-sedimentary faultlets in the northern Huddersfield sub-basin. The restricted occurrence of dewatering structures/ syn-depositional faultlets in zones adjacent to fault splays off the Craven fault-system infers active tectonics are specific to this area. While earthquakeactivity is likely to have little effect on accommodation within the incised valley, it may have the potential to trigger submarine turbidity currents in unstable areas of the submerged deltaslope.

The restricted marine faunas of the Keighley Bluestone provide indirect evidence for active tectonic uplift, which like the dewatering and syn-depositional faults, lie within the zone of fault splays in the northern Huddersfield sub-basin. This lithostratigraphic unit sits on a structural high, and is inferred as the condensed, restricted marine equivalent of the R2b2 sequence. Whereas restricted marine conditions prevailed on the structural high area, the zone surrounding the Keighley Bluestone outcrop was subject to mouthbar and fluvial deposition. Although active tectonism may therefore have affected the amount of accommodation in this part of the basin, it seems unlikely to influence accommodation across the incised valley to any great extent.

6.2.13. Conclusion

This paper has implies that eustatic rather than and non-eustatic (susbsidence, active tectionics and inherited bathymetry) controlled deposition to a greater extent within the early Marsdenian. Eustatically driven regressive surfaces of marine and fluvial erosion often have less than 5m erosive relief, and are inferred as the product of sea-level fall and incision during periods of falling and low sea-level. Such regressive surfaces define high-order sequences, which coalesce to form master sequence boundaries during periods of falling low-order sea-level. Master sequence boundaries separate the lower mudstone/ siltstone-rich part of each cyclothem from the sandstone-rich upper part, implying that each cyclothem is not a repetitively deposited cyclic unit, but is punctuated by a series of depositional hiatuses. Within each low-order sequence, a marine band forms a low-order maximum flooding surface, which also represents the highest magnitude high-order transgressive event within the low-order sequence.

Facies associations and systems tracts occur in a fairly predictable order within each low-order sequence. Each sequence commences with a master sequence boundary, which is overlain by a low-order lowstand to transgressive systems tract that comprises fluvial, mouthbar and bayhead deposits. Significant in the Pennine Basin, transgressive systems tract deposits often contain previously overlooked tidally influenced mouthbar and fluvial facies that were formed as the incised valley formed a coastal embayment during the low-order transgressive systems tract. In the upper transgressive systems tract, high-order sequences become increasingly dominated by bayhead conditions, until the low-order maximum flooding surface forced the highest magnitude flooding event, and the deposition of a marine band. Each marine band is overlain by offshore or interdistributary bay mudstones representing the highstand systems tract, and is incised by a subsequent sequence boundary that forms the upper bounding surface of the sequence.

Both high- and low-order transgressive and regressive surfaces can be reliably traced across sub-basin boundaries, which probably have different subsidence and fill histories. In combination with the presence of marine faunas associated with each flooding surface, this suggests each relative sea-level fluctuation represents a eustatic sea-level change, formed independently of variations in sediment input and basin tectonics. The delineation of active tectonic trends that have a predictable influence on facies distributions, corroborates the notion that active tectonism had only a minor influence on controlling deposition.

Whereas correlation within the Namurian-aged basins of the UK has traditionally relied on cyclothem and marine band identification (Ramsbottom *et al.,* 1978; Waters *et al.,* 1996a), this paper has demonstrated that high-resolution sequence stratigraphic concepts can be used in conjunction with the existing marine band and lithostratigraphic frameworks to define high- and low-order sequences. Analysis of these sequences has enabled refinement of the marine band-lithostratigraphic unit based correlation framework, allowing the correlation of the detailed depositional elements and architectures within the early Marsdenian fill of the Pennine Basin.

End of paper (figures and figure captions follow on unnumbered pages)

6.3. Sequences not in the previous paper; The R2b4 and R2b5 sequences

In this section the R2b4 and R2b5 sequences are dealt with in the same manner as sequences R2b1 to R2b3 in the sequence stratigraphic paper. Figure 6.1 illustrates the general stacking pattern all the early Marsdenian low-order sequences.

6.3.1. R2b4 sequence

An R4 surface separates the high-order transgressive/ highstand systems tract of HS11 from the base of the progradationally stacked delta front mouthbar for HS12. HS13 comprises a wedge of progradationally stacked mouthbar and basin floor turbidites in the Gainsborough Trough, which are absent in the Rossendale sub-basin. The high-order sequence boundary at the base of HS14 has up to 20m demonstrable erosive relief across the basin (Enclosure 1; localities 8-10/ 11). This represents an episode of sufficient bypass that incised the high-order maximum flooding surface with HS13. Both HS12 and HS14 contain high-order maximum flooding surfaces, which force the aggradational to retrogradational stacking of the overlying high-order transgressive-highstand systems tract. The high-order transgressive-highstand systems tract of HS14 is significant, as it comprises bayhead delta facies in the central Rossendale sub-basin. In the northern Huddersfield sub-basin HS 15 is absent (Enclosure 1; locality 5-6), and in the south-eastern Rossendale sub-basin the R2b4 marine band overlies HS13-14. The high-order transgressive/ highstand systems tract of ?HS14 (Enclosure 1; locality 7) comprises a retrogradationally stacked thin distal bayhead delta that is eroded by the R1 surface that forms the base of HS15.

The interfluves present on the basin margins (Enclosure 1; locality 1 and 15) developed synchronously during HS12-15, although a definitive correlation is not made. Interfluves experienced leaching during the deposition of regressive events but became waterlogging during the periods of rising sea-level. The change in palaeosol signature is probably driven by the influence of falling and rising sea-level on the water-table. The interval between HS12 and the high-order maximum flooding surface in HS15 (R2b4) comprises the **Midgley Grit** lithostratigraphic unit, which forms a basinwide correlatable sandstone succession.

The R2b4 marine band represents the high-order maximum flooding surface within HS15. In the northern Huddersfield and Rossendale sub-basins the R2b4 marine band has a sparse ammonoid fauna, and is represented by a T4 horizon, while in deeper parts of the basin an ammonoid-rich marine band (T1 horizon) is fully developed. Despite the paucity of fauna, the T4 horizon representing the R2b4 marine band forms a correlatable surface that delineates the fluvial plain depositional system from the overlying tidally influenced bayhead deltas (HS16-19).

6.3.2. R2b5 sequence

Bayhead deltas with the R2b5 low-order sequence are commonly capped by muddy palaeosol horizons, and these are overlain and reworked by T5 flooding horizons. Successive bayhead dominated high-order sequences decrease in thickness up-section, from a maximum of 18m in HS16 (Enclosure 1; locality 8) to 2m in HS19 (Enclosure 1; locality 2-9). HS19 is higher in the bayhead delta succession and are strongly estuarine in character as bimodal current-ripples and back-flow ripples with muddy drapes are common (see Figure 2.18). All five bayhead deltas between R2b4 and R2b5 are defined by sharp mouthbars with R4 basal surfaces. Together, they form the **Helmshore Grit** lithostratigraphic unit.

In the area east of Lancaster, recent mapping and interpretation of seismic data (Brandon *et al.,* 1998) suggests that locality 1 (Enclousure 1; Middleton Towers) lay on a structural high, generated by late Dinantian-early Namurian inversion on the Quernmore Fault. To the west of the Quernmore fault, the high-order forced-regressive/ lowstand systems tract of HS16 comprises a wave-reworked mouthbar facies overlain by a T3 transgressive surface of marine erosion (a wave-ravinement surface and a high-density bioturbated horizon) representing a shoreface environment (Enclosure 1; locality 1).

6.3.3. The high-order transgressive-highstand system above R2b5

The R2b5 marine band (horizon T1) separates the bayhead delta succession from a deep water, basinal mudstone depositional system across the entire width of the Pennine Basin. The high-order transgressive-highstand systems tract of HS19 overlies R2b5 and represents a major basin-wide retrogradational event, which like R2a1, can be correlated into the southern North Sea, south Wales and western Ireland (Ramsbottom *et al.,* 1978).

Table 7 Description of high-order sequences; including descriptions of high-order systems tracts with lower bounding surface variability, the depositional systems observed across the basin, and high-order sequence stacking pattern in relation to the underlying high-order sequence.

High-order sequence	Systems tract	Lower bounding surface	Depositional system dominant in each basin	Stacking Pattern
HS 1	High-order Transgressive-Highstand systems tract	**R2a1 marine band** (T1). Can be traced into Southern North Sea, South Wales & south-west Ireland (Ramsbottom *et al.,* 1978)	Basin floor mudstone across all sub-basins.	Very Strongly Retrogradational
HS 2	High-order Forced-Regressive/ Lowstand systems tract	R2 in Rossendale & Bowland sub-basins, ?R4 in Huddersfield sub-basin	Basin floor turbidites in Rossendale & Bowland sub-basins, delta front mouthbars in Huddersfield sub-basin	Progradational
	High-order Transgressive-Highstand systems tract	T1 in Rossendale sub- basins,(Collinson *et al.,* 1977), non deposition/ eroded by PS4 & PS6 in eastern sub-basins	Basin floor mudstone, in Rossendale & Bowland sub-basins	Retrogradational
HS 3	High-order Forced-Regressive/ Lowstand systems tract	R2 in Rossendale sub- basins, R4 in Huddersfield sub-basin	Basin floor turbidites in western basin, delta front mouthbars in Huddersfield sub- basin; eroded by P6 in eastern basin	Strongly Progradational
	High-order Transgressive-Highstand systems tract	**R2b1 marine band** T1 Rossendale sub- basins, T1 & T2 in Huddersfield sub- basin & northern Gainsborough Trough	Basin floor mudstone in Rossendale & Bowland sub-basins, eroded in part by P6 in Huddersfield sub- basin. The Keighley Bluestone forms in the northern Huddersfield sub-basin	Retrogradational
HS 4	High-order Forced-Regressive/ Lowstand systems tract	R2 in Rossendale sub- basins & Gainsborough Trough, R4 in Huddersfield sub-basin	Basin floor turbidites in Rossendale & Bowland sub-basins & Gainsborough Trough, delta front mouthbars in Huddersfield sub-basin	Strongly Progradational
	High-order Transgressive-Highstand systems tract	T3? in Rossendale sub- basin	Basin floor mudstone in Rossendale & Bowland sub-basins	Weakly Retrogradational

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HS 5	High-order Forced-Regressive/ Lowstand systems tract	R4 in Rossendale sub- basin & northern Gainsborough Trough; R1and leached palaeosol in the Huddersfield sub-basin (locality 25; Enclosure 1)	Delta front mouthbars in Rossendale & Bowland sub-basins & northern Gainsborough Trough & fluvial sandstones, with leached interfluves in the Huddersfield sub-basin	Strongly Progradational
	High-order Transgressive-Highstand systems tract	T3 in the northern Gainsborough Trough & ?Rossendale Basin, / & mud overprinted palaeosol in the Huddersfield basin (locality 34; Enclosure 2).	Basin floor mudstones in the northern Gainsborough Trough	Weakly Retrogradational
HS 6	High-order Forced-Regressive/ Lowstand systems tract	R1 & mud rich palaeosol (locality 34; Enclosure 2) in southern Huddersfield Basin	Bayhead delta in all basins except the Gainsborough Trough. Possible fluvial sandstone (continuation of supply system initiated during allomember 7 (locality 31; Enclosure 2)	Retrogradational
	High-order Transgressive-Highstand systems tract	**R2b2 marine band** T1 in all basins	Basin floor mudstones in all sub-basins	Strongly Retrogradational
HS 7	High-order Forced-Regressive/ Lowstand systems tract	R2 in Rossendale sub- basins, R4 in Huddersfield basin	Delta front mouthbars in Huddersfield sub-basin, Basin floor turbidites in Rossendale & Bowland sub-basins	Progradational
	High-order Transgressive-Highstand systems tract	?T1 in Rossendale sub-basin, rare T3 in Huddersfield sub-basin	Basin floor mudstones in Rossendale sub-basin, eroded/ thin distal delta front mouthbar in Huddersfield sub-basin (locality 23; Enclosure 1)	Retrogradational
HS 8	High-order Forced-Regressive/ Lowstand systems tract	R2 in Rossendale sub-basin, localised R1 & R4 in Huddersfield sub-basin	Basin floor turbidites in Rossendale sub-basin, delta front mouthbar & rare fluvial systems in Huddersfield Basin (locality 23; Enclosure 1)	Progradational
	High-order Transgressive-Highstand systems tract	T1 in Rossendale sub-basin	Basin floor mudstones in Rossendale sub-basin (only seen in Blackburn Tunnel Section).	Weak Retrogradational

Continued overleaf

HS 9	High-order Forced-Regressive/ Lowstand systems tract	R2 in Rossendale sub-basin, R1 in Huddersfield Basin & R4 in northern Gainsborough Trough	Basin floor turbidites in Rossendale sub-basin, fluvial systems & shoreface on bay margins in Huddersfield Basin (see Figure 9.3), & thin delta front mouthbar in the northern Gainsborough Trough.	Strong progradational
	High-order Transgressive-Highstand systems tract	T4 in the Rossendale & Huddersfield sub-basin, & mud overprinted palaeosol in parts of the Huddersfield Basin (locality 20 & 29; Enclosure 2)	Bayhead delta in all sub-basins	Retrogradational
HS 10	High-order Forced-Regressive/ Lowstand systems tract	R4 in Rossendale & Bowland sub-basins, R1 in Huddersfield sub-basin	Delta front mouthbar in Rossendale & Bowland sub- basins and fluvial sandstone in Huddersfield sub-basin	Progradational
	High-order Transgressive-Highstand systems tract	**R2b3 marine band** T1 in the Rossendale sub-basin, T2 & T3 in the Huddersfield sub-basin & Gainsborough Trough	Bayhead delta in the Huddersfield sub-basin/ Basin floor mudstones in other sub-basins	Retrogradational
HS 11	High-order Forced-Regressive/ Lowstand systems tract	R4 in all basins, with the exception of the north west Bowland sub-basin where the development of leached interfluves initiated (locality 1; Enclosure 1)	Delta front mouthbar in all sub-basins	Progradational
	High-order Transgressive-Highstand systems tract	T3 in the Rossendale Basin, ?R2 in the northern Gainsborough trough & non deposition/ eroded by P19 & younger allomembers in Huddersfield sub-basins	Delta front mouthbar in the Rossendale Basin & ?Basin floor turbidites in the Gainsborough Trough	Aggradational-weak retrogradational
HS 12	High-order Forced-Regressive/ Lowstand systems tract	R1 across the Rossendale & Huddersfield sub-basins & R4 in the Gainsborough Trough	Fluvial sandstone in the Rossendale & Huddersfield sub-basins, & delta front mouthbar & ? Turbidite sandstones in the Gainsborough Trough	Progradational
	High-order Transgressive-Highstand systems tract	T2-T3 in central Huddersfield sub-basin and T3 in Rossendale and Bowland sub-basin	Interdistributary trough/ Delta front mouthbar association in Huddersfield and Rossendale sub-basin	Retrogradational
HS 13	High-order Forced-Regressive/ Lowstand systems tract	R1 in western parts of the Rossendale sub-basin, allomember eroded in the Huddersfield sub-basin, no deposition in the Gainsborough Tough	Fluvial sandstone in the Rossendale & Huddersfield sub-basins	Strongly Progradational

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	High-order Transgressive-Highstand systems tract	T4 in the western Rossendale & eastern Huddersfield sub-basin (localities 17-19; Enclosure 1), eroded by younger allomembers in the eastern Rossendale & western Huddersfield sub-basins	Bayhead delta in the western Rossendale & eastern Huddersfield sub-basin	Retrogradational
HS 14	High-order Forced-Regressive/ Lowstand systems tract	R1 throughout all sub-basins	Fluvial sandstone in all sub-basins	Strongly Progradational
	High-order Transgressive-Highstand systems tract	?suggestion of T4 (locality 7; Enclosure 2) Rossendale Basin	?thin (less than 5m) Bayhead delta in Rossendale sub-basin	Weakly Retrogradational
HS 15	High-order Forced-Regressive/ Lowstand systems tract	R1 across the Rossendale & Huddersfield sub-basins & ?T2 in the Gainsborough/ Goyt Trough	Fluvial sandstones in the Rossendale & Huddersfield sub-basins, with Turbidite sandstones in the Gainsborough/ Goyt Trough	Progradational
	High-order Transgressive-Highstand systems tract	**R2b4 marine band** T1 in the Rossendale sub-basin & Gainsborough/ Goyt Trough, T4 in the northern Huddersfield sub-basin with the exception of T1 around Keighley	Basin floor mudstones in all basins, with the exception of distal bayhead delta mudstones in the north-east Huddersfield Basin	Strong Retrogradational
HS 16	High-order Forced-Regressive/ Lowstand systems tract	R4 in Rossendale, Huddersfield & Bowland sub-basins	Bayhead deltas overlain by waterlogged palaeosols in the Rossendale & Huddersfield sub-basins	Weak progradational-
	High-order Transgressive-Highstand systems tract	T5 across the Rossendale, Huddersfield & Bowland sub-basins	Distal bayhead delta mudstones	Weak retrogradational

Continued overleaf

HS 17	High-order Forced-Regressive/ Lowstand systems tract	R4 in the central Rossendale, Huddersfield & Bowland sub-basins	Geographically restricted bayhead deltas, with distal bayhead delta mudstones in adjacent areas	Weak progradational
	High-order Transgressive-Highstand systems tract	T5 in only in the central Rossendale, Huddersfield & Bowland sub-basins	Geographically restricted distal bayhead delta mudstones	Weak retrogradational
HS 18	High-order Forced-Regressive/ Lowstand systems tract	R1 & R4 in Rossendale, Huddersfield & Bowland sub-basins	Bayhead deltas, with fluvial feeder systems in the Rossendale & Huddersfield sub-basins	Weak progradational
	High-order Transgressive-Highstand systems tract	T5 across the Rossendale, Huddersfield & Bowland sub-basins	Distal bayhead delta mudstones	Weak retrogradational
HS 19	High-order Forced-Regressive/ Lowstand systems tract	?R4 across the Rossendale, Huddersfield & Bowland sub-basins	Thin bayhead delta mudstones	Weak progradational
	High-order Transgressive-Highstand systems tract	**R2b5 marine band** T1 in the Rossendale & Bowland sub-basin & Gainsborough/ Goyt Trough, T4 and T1 in the northern Huddersfield sub-basin	Basin floor mudstone across basin.	Very Strongly Retrogradational

6.4. Interpretation of R2b4 and R2b5 sequences

6.4.1. Interpretation of the R2b4 sequence

A 'stepped' geometry is apparent on the low-order sequence boundary at the base of the sandstone unit above R2b3 when it is correlated through successive sections across the Pennine Basin. The regressive surface at the base of the sand-rich unit is overlain HS12 to HS15 in different parts of the basin, implying that it is diachronous. The resultant R1,R2 and R4 regressive surfaces forms a 'master-sequence boundary' (see sequence stratigraphic paper and Chapter 7 for discussion) that has up to 40 metres erosive relief and is overlain by a wedge of basinward-stepping fluvial and mouthbar facies (comprising the Midgley Grit). In deeper parts of the basin, the R2 basal surface in HS13 represents the erosive base of a turbidite channel complex or the basal surface of a proximal turbidite lobe. This surface and the overlying turbidite deposits resulted from sediment bypass of the proximal delta during falling and relatively low sea-level.

A thick ganister palaeosol (P1 horizon) at locality 1 and an overprinted (P3) palaeosol at locality 15 (Enclosure 1; & Figure 6.2) are the correlated interfluve equivalents of the R1, R2 and R4 regressive surfaces. The palaesols were formed on interfluves throughout the period of falling and relatively low sea-level. The presence of interfluves on both sides of the regressive sandstone permits an accurate correlation of width (up to 90 kilometres). A thick P1 horizon developed on the structural high west of the Quernmore Fault between HS12 to HS15.

Within the regressive sandstone, stacked high-order sequences of delta front mouthbar, fluvial and bayhead delta depositional systems prograde over the R1, R2 and R4 surfaces. This implies a relatively high modulating sea-level existed in comparison to that during the formation of the regressive surface. At its thickest, the regressive sandstone attains a maximum thickness of 60m (Enclosure 1; between the base of HS12 and R2b4; localities 7, 8 and 9). The high-order maximum flooding surface within HS12, is interpreted as the initial flooding surface within the regressive sandstone. Tidally influenced fluvial facies occur throughout HS13-HS15, and although their distribution is sporadic, tidal facies are often eroded by fluvial dominated, cross-bedded sandstone facies (Enclosure 1; localities 7 and 12).

The R2b4 marine band separates the underlying fluvial plain depositional systems from the younger bayhead delta depositional system of HS16-HS19. The absence of a fully marine signature across the whole basin suggests that the magnitude of flooding associated with the R2b4 marine band was only sufficient to partly force the retrogradation of the supply system.

6.4.2. Interpretation of the R2b5 sequence

The dominance of bayhead deltas above R2b4 implies that individual transgressions created thin (*circa.* 15m) but extensive areas of accommodation. Tidal and estuarine motifs in HS16-HS19 indicates the continued presence of tidal currents. The successive thinning of individual bayhead deltas within the R2b5 sequence suggests a decreased amount of accommodation was available, while the sharp-based mouthbar provides evidence implying base-level modulation occurred.

6.5. Low-order sequence stacking patterns

This section integrates the interpretations of low-order sequences and describes their stacking patterns and therefore the broad baasin-fill architecture. Assuming constant sediment input and no lateral shift in the basin depocentre, comparisons between stacked low-order sequences (sequence *sensu.* Posamentier *et al.* 1988; Posamentier & Vail 1988), allows the relative magnitude of base level rise associated with each marine band (Figure 6.1, 6.2 and 6.3). The R2a1 marine band can be correlated with confidence beyond the Pennine Basin, and separates the Kinderscoutian delta from the retrogradationally stacked R2a1 sequence (Figure 6.1). HS1 and HS3 are retrogradationally stacked on the Kinderscoutian delta. The spatial distribution of the facies associations reveals the marked dichotomy in depositional environment. The underfilled Bowland and Rossendale sub-basins possess slope/ basin floor turbidites, while and Huddersfield sub-basin contains mouthbars on the submerged Kinderscoutian delta top. The submerged Kinderscoutian delta system effectively forms a platform, over which the R2b1-R2b4 low-order sequences prograde (Figure 6.1). These sequences combine to form a *shoal water / inner shelf delta* that protrudes onto the Kinderscoutian delta. The turbidites deposited in the Rossendale and Bowland sub-basins represent deposits that bypassed the delta front during periods of low-order forced-regressive to lowstand relative sea-level.

The R2b1 marine band is not as significant as R2a1 and only modulates relative sea-level to the extent of forcing the retrogradational stacking of the transgressive/ highstand sytems tract of HS3 (Figure 6.2 & 6.3). HS4 and HS6 are progradationally stacked, and possess the similar facies association distribution as the R2a1 sequence. The base of HS4 erodes through the R2b1 marine band in the southern part of the Huddersfield sub-basin (Figure 6.3). The surfaces underlying the regressive sandstone units in low-order sequences R2a1-R2b2 comprises amalgamated R1-R2-R4 regressive 'master surfaces' (which form a low-order sequence boundary; see paper and section 6.5 for disscussion). For example, HS2 and HS3 in the Bowland sub-basin both possess correlatable regressive surfaces, which amalgamate to form a single low-order sequence boundary in the Huddersfield sub-basin (Figure 6.2). Therefore each low-order sequence boundary is the product of at least two high-order relative sea-level falls, and not a single high magnitude sea-level fall. This implies that the regressive surface and its correlative interfluves are composite and diachronous, and absolute correlation is only possible if the key surface sequence and environment distributions are fully understood.

The high-order sequences overlying the sequence boundary in the R2a1 sequence (HS2 & HS3), R2b1 sequence (HS4 to HS7) and R2b2 sequence (HS 8 to HS11) fluctuate between progradation and retrogradational stacking patterns, implying that sea-level modulated rapidly during deposition. Cryptic tidal signatures are located in HS3 and HS8 (in the initial low-order incised valley fill; see Figures 2.15, 2.16 & 4.16). The incised valley fill within the R2b3 sequence culminated in emergence and the formation of the palaeosol horizons (Figure 6.2; HS10 and HS11). The increased thickness of the incised valley fill throughout the deposition of the R2b1-R2b2 sequences is probably due to the encroachment of the fluvial supply system during this period (Figure 6.1).

The R2b4 sequence is progradationally-stacked on the R2b3 sequence. The basinal mudstone depositional system of the high-order transgressive-highstand systems tract of HS11 separates R2b3 from the regressive sandstones of HS12 to HS15. The lithologies within the high-order transgressive-highstand systems tract of HS11 contain a large proportion of suspension deposited siltstone deposited as suspended clastic debris, suggesting the close proximity of a clastic source (Figure 6.2). This implies that the thickness of the high-order transgressive-highstand systems tract cannot alone be used as evidence of a higher magnitude sea-level rise in comparison between R2b1, R2b2 and R2b3. Variations in the magnitude of transgressive events and the proximity of the supply system explain why marine bands in some areas have distinctive ammonoid faunas, and why some have cryptic, non-ammonoid motifs.

The base of HS11 forms the first regressive surface of the master sequence boundary delineating the base of the R2b4 low-order sequence. This sequence boundary is diachronous, and occurs at the base of HS11 in the valley axis, and HS14 or HS15 towards its margins. Unlike the successions overlying the sequence boundaries in sequences R2b1, R2b2 and R2b3, high-order forced regressive/ lowstand systems tract of mouthbar associations are overlain by transgressive horizons, formed by high-order sea-level modulations. The initial stage of mouthbar progradation is coeval with the continued formation of the regressive surface (R4) within the R2b3 sequence. This suggests that the mouthbar associations within HS12 and HS13 represent forced regressive deposits, which are preserved by the flooding of the system by the high-order transgression.

The incised valley fill overlying the sequence boundary formed during the R2b4 sequence forms a succession of greater width (up to 90km) and depth (up to 40m of erosive relief) than the previous sequences. The sand-rich incised valley fill is influenced throughout deposition by relative sea-level modulations generated by high-order transgressions-regressions. The extent of progradation is such that the delta system reaches the shelf-slope break formed by the submerged Kinderscoutian delta system in the southern Huddersfield sub-basin by the end of the R2b3 sequence. The R2b3 sequence is therefore similar in form to a lowstand *shelf edge delta*. However it seems better describe this phenomenon as a *platform edge delta* because the R2b4 sequence does not develop at a continental-shelf edge.

The marine transgression associated with R2b4 is of sufficient magnitude to partly inundate the incised valley, creating accommodation in which thick bayhead delta systems prograded (the Helmshore Grit; Figure 6.2). The incised valley fill in the R2b5 sequence has a similar density of transgressive surfaces and retrogradationally to aggradationally stacked high-order sequences in comparison to the R2b4 sequence, implying the continued modulation of high-order relative sea-level. Towards the margins of the incised valley, shoreface associations replace bayhead delta systems (Figure 6.2; HS16). These are also influenced by modulations in base level, and the spectacular section at Middleton Towers (SD409588) reveals a shoreface with a basal regressive contact of the basin floor mudstones of HS16, comprising a transgressive surface of marine erosion overlain by a gravel lag, and a marine bioturbated (T3) horizon (see Figure 3.7).

As the amount of accommodation decreased within the incised valley during the final stages of R2b5 sequence, minor base-level modulations had an increasingly important effect on forcing the backstepping of bayhead delta systems (Figures 6.1 & 6.3). Bayhead deltas become emergent and developed both leached and waterlogged palaeosol horizons due to the influence of modulating sea-level on the water-table.

The R2b5 marine band represents a high-amplitude flooding surface, which completely caps the R2b4-R2b5 sequence incised valley fill (Midgley and Helmshore Grits) and interfluves, separating them from the overlying basin floor mudstones systems. The overlying basin floor mudstones are up to 10m thick in the north-east Huddersfield sub-basin (Enclosure 2; locality 8), and contain a coarsening upwards mouthbar, implying the re-establishment of a prograding delta front.

6.6. A generic sequence stratigraphic model for Marsdenian

This section details the generic characteristics of a low-order sequence in order to allow the examples from the Marsdenian to be utilized for other intervals of the Namurian. Marine bands undoubtedly provide a first order framework, on which other correlation frameworks can be applied. Application of the sequence stratigraphic framework focuses the interpretation on the surfaces bounding the sand-rich part of each sequence, and increases the resolution of the correlation between marine bands. The polarity and magnitude of relative sea-level modulation associated with each transgressive horizon and regressive surface is assessed by comparing the characteristics and positioning of low-order sequences and high-order sequence (Figure 6.1, 6.2 and 6.3). The observed variations between the retrogradational-progradational stacking of systems tracts within high-order sequences and the stacking patterns of low-order sequences suggests that relative sea-level fluctuation occurs on at least two orders of magnitude.

High-order sequence stacking patterns are produced by high frequency and low magnitude (high-order) sea-level fluctuations. High-order sequences often alternate between progradational and retrogradational stacking patterns, and high-order sequence thicknesses vary between 7-20m.

Low-order sequence stacking patterns are generated by low frequency/ high magnitude sea-level fluctuations. Periods of marine band deposition are associated with the point of the maximum rate of sea-level rise (the maximum flooding surface; *sensu* Hampson *et. al.,* 1997). Conversely, the maximum rate of sea-level fall corresponds to the generation of the master sequence boundary within each low-order sequence. Low-order sequences are generally between 10-50m thick within the incised valley, and up to 130m thick in basinal turbidite fans.

The combination of two relative sea-level modulation curves with contrasting frequencies and magnitudes produces a compound curve. This compound curve is used to explain the origins and interrelationships between depositional systems, key surfaces/ horizons. This allows the interpretation and delineation of low-order systems tracts (*sensu.* Van Wagoner *et al.* 1988, Posamentier *et al.* 1988, Posamentier & Vail 1988).

6.6.1. Low order forced regressive and falling-stage systems tracts

During lower order falling stage systems tract, the low- and high-order curves combine to produce a stepped sea-level fall, with extended high-order falling limbs and suppressed high-order rising limbs (Figure 6.4). R1, R2 and R4 surfaces are formed during these high-order regressions.

The R4 surface often has uniform erosive relief (usually less than 1m) and is always associated with overlying coarser-grained mouthbar deposits. This suggests the generation of this surface is closely associated with the process of mouthbar deposition (see Figure 4.16). In the mouthbar environment, scour surfaces lie basinward of the river-mouth and form linear or bifurcating scour features (see Figure 2.12.1). These scour surfaces only form adjacent to the river-mouth due to the rapid dispersion of the river-effluent and the loss of flow regime. The Marsdenian mouthbars prograded rapidly in a basinward direction and the scour surfaces were transient features that are susceptible to sediment inundation and preservation by the prograding mouthbar. During periods of falling sea-level, the velocity of the effluent plume entering the basin increased as the graded profile of the river was raised. As effluent discharge intensifies, the flow became increasingly friction-dominated (see Figure 2.12.1; Wright 1977). The extent and severity of scouring and mouthbar migration increased with increasing friction, and the scour feature (or regressive surface of marine erosion; see paper for discussion) migrated rapidly in a basinward direction. The mouthbar succession is only preserved during subsequently increasing sea-level and high-order sequence progradation. Minor relative sea-level falls have the capacity to locally form R4 surfaces but several generations of R4 surface during a low order sea-level fall may coalesce to form laterally continuous sequence boundaries (Figure 6.3).

On platform areas (e.g. that formed by the submerged Kinderscoutian delta) a laterally extensive forced regressive systems tract formed as the result of extended periods of sea-level fall (Figure 6.3). During periods of falling low-order sea-level accommodation on the platform was produced either by high-order sea-level modulation, or was inherited from the previous low-order transgression of the platform. The thin sheet-like accommodation was infilled by mouthbar deposits, implying that subsequent mouthbars had to avulse laterally into areas that had not previously received sediment. These Marsdenian mouthbar deposits are similar to those described from shoreface environments (Plint 1988; Posamentier & Chamberlain 1993; Hadley & Elliott 1993). Although such forced regressive systems tracts are well documented in coastal deposits but few examples are documented in areas with limited wave-reworking (Posamentier & Allen 1999; Kolla *et al.* 2000). However, forced regression deposits from both deltaic and wave-reworked coastlines and deltaic forced regressive deposits share characteristics in addition to the sharp regressive basal surface (Posamentier & Allen, 1999; Plint & Nummedal, 2000; Posamentier & Morris, 2000). These similarities include the formation of progressively lower-relief clinoforms as the forced regressive wedge progrades in a basinward direction (as seen rather dramatically in Figure 2.12.1 and 2.12.2).

In the mouthbar-fluvial depositional system, the high-order sequence boundaries coalesced during the falling stage and early lowstand systems tract to effectively form a diachronous sequence boundary 'master surfaces' (Figure 6.4.2). Master surface sequence boundaries have been described by Posamentier & Allen (1999). In cases where the early high-order sea-level falls produce prograding wedges that do not bypass sediment to the basin floor, forced regressive high-order systems tract deposits may be preserved. Forced regressive high-order systems tracts are most clearly identified where they are capped by subsequent higher-order transgressive horizons. However, transgressive surfaces have a low preservation potential due to erosion during the later stages of incised valley formation. Bypass of sediment through the incised valley eventually leads to the deposition of downlapping progradational turbidite deposits (overlying an R2 surface). These progradational successions alternate with retrogradationally stacked high-order sequence formed during higher order relative sea-level rise.

6.6.2. Low order late falling stage and early lowstand systems tract

The maximum amplification rate of the falling limbs of the high-order fluctuations occurs during the late falling stage and early lowstand systems tract as the maximum rate of relative sea-level fall is approached (Figure 6.3 & 6.4.1). The maximum rate of relative sea-level fall is associated with higher-order fluctuations, and lies close to the point of the maximum rate of lower-order relative sea-level fall. As the rate of lower-order relative sea-level fall decreases (during the early lowstand systems tract), both falling and rising limbs of the higher order curve become equal in magnitude. Therefore during the period of low-order sea-level fall the rate of sea-level fall associated with higher-order sea-level fluctuations increases, reaches a maximum, and then decreases. This implies that the magnitude of sea-level fall associated with sequence boundary generation will change through time (Figure 6.4.1). Early higher-order relative sea-level falls are of low magnitude and produce shallow, broad incised valleys. As the rate and magnitudes of relative sea-level fall increases, the incised valleys produced are narrower and deeper (see sequence stratigraphic paper for discussion). The fluvial system became entrenched within this developing incised valley (Figure 6.4.2), which explains the 'terraced' appearance of the depositional strike section through the 'master' sequence boundary (see Enclosure 1; localities 3-6). Continued bypass of sediment through the incised valley keeps supplying sediment to the lowstand turbidite-dominated delta front, which progrades in a basinward direction.

High-order sea-level rises are superimposed on the falling low-order relative sea-level curve. Transgressive surfaces located within the incised valley fill correspond to the maximum rate of sea-level rise during these periods. These high-order maximum flooding surfaces produce thin retrogradationally stacked high-order transgressive-highstand systems tracts that have a very low preservation potential in shelfal areas, but may be intercalculated within the prograding turbidite system. These retrogradationally stacked systems tracts are preserved due to their position on a rising part of the low-order sea-level curve, where there is no significant eroison during periods of high-order sea-level fall. The relative position of the distributary feeder channels controls the spatial distribution of erosion surfaces developed during high-order sea-level fall. The first transgressive horizon within the incised valley is only preserved when accommodation generation outstrips sediment supply. This implies that the initial flooding surface is diachronous, and it cannot be used as an absolute correlation datum (Figure 6.3). The combination of the incised valley geometry and the marine transgression may create tidal amplification (*sensu.* Zaitlen *et al.* 1994), producing the tidally-influenced structures present in these high-order sequences (Figure 6.5; see sequence stratigraphic paper for discussion).

6.6.3. Low order early transgressive stage systems tract

During periods of low-order sea-level rise, high-order sea-level rise is amplified, and falls are suppressed. This generates sufficient accommodation in the distal incised valley, allowing the storage of sediment, and the short-term cessation of sediment supply to the turbidite-dominated succession in the deeper parts of the basin. In the shelfal area, high-order relative sea-level falls are still of sufficient magnitude to generate incision and high-order sequence boundaries (either regressive surfaces of marine erosion, or regressive surfaces of fluvial erosion; see sequence stratigraphic paper). High-order transgressive and highstand systems tracts formed during the periods of low-order rising relative sea-level are aggradationally stacked during the rising high-order limb, and progradationally stacked during periods of high-order high relative sea-level.

As the rate of low order sea-level rise increases, high-order flooding events increase in magnitude. The high-order sequence stacking between these flooding events is controlled by the amount of sediment entering the basin. Progradation occurs if sediment supply exceeds the rate of accommodation generation during the flooding event, whilst retrogradation occurs if sediment supply is surpassed by the rate that accommodation is generated.

Prior to the point of the maximum rate of relative sea-level rise (the low-order maximum flooding surface); high-order flooding surfaces are cryptic, and inferred by the presence of thin retrogradationally stacked high-order transgressive and highstand systems tracts in the incised valley fill. In deeper parts of the basin these transgressive surfaces are correlated to increasingly marine-dominated flooding horizons (T3), and eventually to condensed mudstone intervals (the high-order flooding surfaces HS2 and 4 are examples of this in the Rossendale sub-basin, see Figure 6.2). At the maximum rate of relative sea-level rise, transgression allows condensed deposition in the distal incised valley. The R2b4 marine band is an example of this, and is characterised by an ammonoid (T1) horizon in the distal incised valley, and *Lingula-Planolites* (T2) in proximal areas. Maximum flooding surfaces, with their identifiable ammonoid index species form the 'genetic' stratigraphic framework (Galloway 1989), which in past have provided the regionally correlatable marine bands, that have proven so valuable in previous correlations (Ramsbottom *et al.* 1978; Waters *et al.* 1996a).

6.6.4. Low order late transgressive stage to highstand systems tract

The maximum flooding surface is overlain by a laterally extensive bayhead delta depositional system in up-dip areas, where high-order sequence boundaries and maximum flooding surfaces are defined by extensively correlatable transgressive and regressive surfaces. Evidence for tidal influence is more common within the bayhead delta depositional systems than in the underlying fluvial-dominated incised valley fill.

During the low-order transgressive systems tract, the rate of sea-level rise began to outpace the amount of sediment input, and the bayhead delta system transgressed landward. Palaeosols become less common within the incised valley fill, with the exception of areas where sediment supply keeps pace with accommodation generation near to fluvial input points. By the low-order highstand systems tract, the shoreline has retreated to its maximum landward position, and the marine transgression led to the deposition of basin floor mudstones across the incised valley.

Following the maximum flooding surface, the rising limbs of the high-order sea-level curve were extended and the falling limbs shortened. This difference is expressed by the retrogradational to aggradational high-order sequence stacking patterns developed during high-order sea-level rise, compared to weakly progradational stacking during periods of high-order falling sea-level. The difference between the magnitude of the high-order fall and rise decreases as the rate of low-order sea-level rise decreases and the high-order fall and rise equilabrates.

6.7. Summary: An improved correlation based on sequence stratigraphic concepts.

Transgressive and regressive horizons/ surfaces refine the correlation framework provided by the marine bands and delineate retrogradationally and progradationally stacked units (*sensu.* high-order forced regressive/ lowstand and transgressive and highstand systems tracts). These key surface/ horizon signatures and high-order systems tract stacking patterns allow the polarity of relative sea-level modulation to be defined throughout a stacked succession of low-order sequences. Changes in relative sea-level have a known effect on the distribution of depositional systems, so when eustatic sea-level signatures are correlated across the basin, the distribution of depositional environments can be predicted. This forms the basis of the sequence stratigraphic correlation within the lower Marsdenian, which augments the marine band correlation framework, increasing the number of correlative ties and the accuracy of the correlation. The development of a systems tract model combining both high and low order eustatic sea-level oscillations explains the lateral distribution and the variability of depositional environments and key surface/ horizon characteristics within sequences.

The stacking pattern of the sequences across the basin allows the broad architecture of the basin fill to be inferred. The retrogradational to weakly progradational stacking of sequences R2b1 to R2b2 and R2b3 suggests deposition occurred during a background transgressive-highstand systems tract. However, when individual mouthbar-dominated high-order sequences are traced laterally they often have a regressive surface of marine erosion at their base. These are interpreted as high-order forced regressive systems tract mouthbars that are restricted to shallow-water platform areas, where minor sea-level changes have a major impact on the positioning of the depositional environments. The depositional systems overlying the R2b4 sequence boundary step noticeably into the basin, and represent a major eustatic lowstand and transgressive incised valley fill. Several smaller incised valleys of smaller dimensions occur in all of the sequences. Comparison of small and large scale incised valleys reveals that they both have similar fill and key surface/ horizon characteristics. Both scales of incised valley have basal regressive surfaces and are filled with retrogradationally and progradationally stacked high-order sequences. When traced across the basin, sequence boundaries and their correlated interfluves comprise a master-sequence boundary that is formed by a series of coalesced minor regressive surfaces. Such surfaces are produced by the combined erosive effect of low and high-order base level fall.

High-order sequences within the low-order transgressive systems tract fill of the incised valley are delineated by a series of transgressive horizons and regressive surfaces that where formed by the combined effect of a rising low-order and fluctuating high-order eustatic sea-level signature. The R2b5 marine band is found across the study area as is identified outside the Pennine Basin (Ramsbottom *et al.,* 1978). The widespread flooding associated with the R2b5 marine band, re-establishes basinal conditions across a large geographic area and suggests a high-amplitude eustatic sea-level rise. The R2a1 marine band can also be traced outside the Pennine Basin, and seperates delta-top from basin-floor depositional systems. Both R2a1 and R2b5 are therefore regarded as the products of similar, large magnitude, transgressive events that can be correlated across Europe, suggesting regional tectonics or local compaction was not responsible for their formation. High- and low-order sequence architecture, and the geographic extent of the transgressive-regressive surfaces are controlled by relative sea-level modulations produced by the combination of eustatic sea-level fluctuations on at least three frequencies (high-order sequence, low-order sequence, sequence sets).

Chapter 7: The magnitude and frequency of relative sea-level change in the early Marsdenian

7.1 Introduction; The frequency and magnitude of relative sea-level change during the Carboniferous.

The waxing and waning of the Gondwanland ice-cap has been the proposed cause of Carboniferous relative sea-level fluctuations since the initial recognition of transgressive-regressive 'cyclothems' (Wanless & Shepard, 1936). Interpretations suggest the main regression correlates to the period of ice cap growth, lending weight to this hypothesis (Figure 7.1; Veevers & Powell 1987). In the late Namurian of northern Europe, the sharp rise in eustatic sea-level roughly corresponds to the inferred onset of ice-cap waning at the start of the Westphalian (Figure 7.1). The eustatic signature in the Appalachian Basin broadly mirrors that of Europe, with the exception of the late Namurian transgression in Europe.

The duration of the Carboniferous has been intensely debated, and new dating techniques and sample points provide an increasingly accurate absolute-dating framework (Harland et al., 1982; Klein, 1990; Waters *et al.,* 1996b). Absolute dating techniques are required if the frequency of relative sea-level fluctuation is to be assessed. Absolute dates from the Namurian succession of the Pennines are derived from bentonite (volcanic ash) bands, which are sporadically distributed. The interval between between transgressive-regressive surfaces must be calculated as an average between absolute date datapoints. Absolute time-picks from the 40Ar-40Ar dating of tonsteins (Lippolt *et al.,* 1984) led to the assertion that the averaged periodicity of marine bands during the Namurian was 160 ? 64ka (Leeder & McMahon, 1988; Maynard & Leeder, 1992). Absolute dates from the Appalachian Basin suggested that the length of the Namurian was shorter, and the averaged periodicity between marine bands was 40-120ka (Heckel, 1986). Accurate dates are required to determine subsidence rates, and this initiated an interest the in re-calculation of the length of the Namurian series within the Pennine Basin (Leeder & McMahon, 1988; Riley *et al.,* 1993). Power spectral analysis of transgressive-regressive cycles from the Yeadonian of the Pennine Basin used the pre-SHRIMP absolute dates to calculate periodicities of 120 000 years with magnitudes of sea-level change up to 42m (Maynard & Leeder, 1992).

Maynard & Leeder (1992) assessed the magnitude of relative sea-level change, by noting critical indicators of water depth change relative to base level, i.e. palaeosols/shoreface deposits. Such depth markers are common in some parts of the lower Marsdenian of the Pennine Basin, where successions with multiple flooding surfaces and palaeosols occur in the bayhead delta depositional system. The most recent absolute dates are provided by the SHRIMP analysis of Namurian bentonite clay bands in the Arnsbergian and Westphalian. Using the dates from these bands suggests a much shorter average duration between Namurian marine bands of approximately 65 000 years (Riley *et al.,* 1993).

7.1.1. Estimating relative sea-level rise in Carboniferous successions with multiple palaeosol-flooding surface horizons

In examples where the high-order forced-regressive/ lowstand systems tract bayhead deltas are capped by palaeosols, all available accommodation is inferred as filled, allowing emergent conditions to prevail. Eustatic sea-level rise and flooding (during the high-order transgressive and highstand systems tract) is interpreted to have followed this period of emergence and the palaeosol is overlain by a regionally flooding surface (normally a T5 surface in bayhead environments). A succession of stacked high-order sequences exhibiting these characteristics suggests multiple cycles of transgression and emergence. In the early Marsdenian this type of succession only occurs in the bayhead delta depositional system of the R2b5 sequence (the Helmshore Grit).

The magnitude of relative sea-level change is estimated by using up to four such successions in the Helmshore Grit (Enclosures 1 and 2). Estimates of the magnitude of sea-level change do not account for compaction within the underlying delta system, or the effect of relative sea-level fall on the generation of regressive surfaces at the base of the high-order regressive-lowstand systems tract. Compaction due to post-depositional de-watering may cause subsidence on the delta top, but this would probably have a negligible effect on the thickness of individual high-order sequences (see Section 9.2.2.2). Regressive surfaces erode into the bayhead prodelta/ offshore facies of the underlying high-order transgressive-highstand systems tract, but do not incise the high-order maximum flooding surface. Therefore, the thickness difference between the retrogradationally stacked high-order transgressive-highstand systems tract and the progradationally stacked high-order forced-regressive/ lowstand systems tract allows an estimate of the net amount of sea-level rise during the deposition of high-order sequences. The data from the R2b5 sequence suggests an estimated 7-20m of relative sea-level modulation occurred during the deposition of the paired progradationally-retrogradationally stacked high-order systems tracts.

The Helmshore Grit represents the last stage of fill prior to the R2b5 low-order maximum flooding surface that caps the incised valley. Deposition during this interval occurred on a predominantly flat delta top, with low relief topographic highs and interfluves on the east and west margins (see Figure 8.6). In this environment, even minor transgressive events created regional flooding, which are observed as laterally continuous transgressive surfaces within the Helmshore Grit lithostratigraphic unit. In areas of the basin where waterlogged palaeosols (or other depth markers such as shoreface sequences) are not present, calculation of the magnitude of relative sea-level rise is not possible. This implies that estimates of the magnitude of relative sea-level modulation can not be extracted from the delta front mouthbar, or basinal mudstone depositional systems.

The integration of the correlation framework within the R2b5 sequence, along with an understanding of tectonic and subsidence effects on accommodation generation would provide a testing ground for the technique used by Maynard & Leeder (1992), and a measurement of the magnitude of relative sea-level change.

7.1.2. The frequency and magnitude of Quaternary relative sea-level change

Eustatic sea-level change during the Quaternary is characterised by asymmetric, high-amplitude sea-level cycles with superimposed cycles of shorter frequencies and magnitude. These cycles are classified as fourth and fifth order relative sea-level modulations (Vail *et al.,* 1991) and are inferred as responsible for the deposition of transgressive-regressive cycles on continental shelves (Suter & Berryhill, 1985). Fifth-order sea-level rise is characterised by a low-magnitude (± 20-30 m) sea-level fluctuations occurring over a short periods (20 000 years); while fourth-order sea-level fall occurs over high-magnitudes (± 130m) and longer periods (100 000 years; Figure 7.1.2). Fifth-order modulation is superimposed on the fourth-order fluctuation and generates a stepped appearance to fourth-order falling limbs, while rising fourth-order limbs are uninterrupted by the fifth-order modulation. The asymmetry of the Quaternary sea-level curve is explained by continental ice-cap dynamics. As continental ice-masses develop they generate a prolonged period of eustatic sea-level fall, wheras the ice caps melt over a much shorter time period.

7.1.3. Comparisons between the frequency and magnitude of Namurian and Quaternary relative sea-level change

Absolute rates of sea-level change cannot be calculated for the Marsdenian, due to the lack of an absolute dates tied to the correlation-framework. However, the polarity of sea-level modulation can be inferred from high- and low-order sequence stacking patterns (see sequence stratigraphy paper). Due to the absence of absolute dating points, this chapter outlines the differences between the Quaternary and Namurian by comparing; i) the relative sea-level modulation patterns between the Marsdenian and Quaternary ii) the estimated magnitude of relative sea-level change, and iii) the estimated duration of Namurian marine bands and major Quaternary flooding events. These will be addressed individually.

7.1.3.1. Comparisons of Modulation pattern:

Comparing the estimated 65ka between Namurian marine bands and the time-elapsed between the maximum rates of fourth-order Quaternary fourth-order sea-level rise reveals interesting similarities. Up to 118ka is inferred to have elapsed between the maximum rate of sea-level rise on alternate Quaternary rising sea-level limbs (120- 140 and 10- 20 ka). This period approximately corresponds with the periodicity of two Namurian marine bands (130 ka). The R2b4 marine band is preserved within an incised valley, suggesting that other marine bands may also be present within lowstand to transgressive systems tract incised valley fills. The R2c4 and R2c5 (*Reticuloceras coreticulatum* and Butterly) marine bands are candidates for such intra-incised valley marine bands (Hampson, 1997), and may explain why some Namurian marine bands are cryptic or have rare ammonoid faunas. The periodicities of Namurian marine bands that lie below and above the incised valley fill (i.e. not including intra-incised valley fill marine bands) are therefore interpretted as similar to the periodicity of Quaternary fourth order flooding events. This has implications for the generation of a relative sea-level modulation curve (see Section 7.2).

7.1.3.2. Comparisons of Magnitude:

The minimum thickness of accommodation generated by valley incision can be estimated from the dimensions of the incised valley fill, i.e. in the case of the R2b4 and R2b5 sequences, between the low-order sequence boundary in the R2b4 sequence and the R2b5 maximum flooding surface. Between the sequence boundary within the R2b3 sequence and the R2b5 maximum flooding surface, a maximum of 115m of sediment was deposited within the incised valley of the southern Rossendale sub-basin (Enclosure 1; localities 17 & 18). This part of the basin remained underfilled during the Kinderscoutian, and was not influenced by the Kinderscoutian inherited bathymetry during the Marsdenian.

The R2b4 low-order maximum flooding surface represents a lower magnitude sea-level rise compared to that generated during the R2b5 maximum flooding surface, although the associated sea-level rise was greater than those generating marine bands R2b1-R2b3 (see Figure 7.3.2). Assuming that the valley was incised to base level during sequence boundary formation, the 115m of strata within the incised valley therefore represents the minimum amount of accommodation generated between the lowest point of relative sea-level and maximum relative sea-level during the R2b4 and R2b5 sequences. This magnitude of sea-level change is very similar to the 130m difference in eustatic sea-level between Quaternary lowstand-highstand periods.

7.1.3.3. Comparisons of Frequency

The Quaternary asymmetric falling fourth-order relative sea-level limb with a superimposed fifth order cycle (Figure 7.2) between 20ka and 140ka could have potentially produced a similar scale of erosion to that observed during sequence boundary genesis in the R2b3 sequence (see Figure 7.1).

The low-order transgressive-highstand systems tract of the R2b1 and R2b2 sequence has many transgressive-regressive surfaces, implying frequent base level modulation. During the Quaternary the rising limb between 120ka and 140ka has superimposed perturbations caused by the higher-order modulation, wheras such perturbations are absent on the rising limb at the onset of the Holocene.

As mentioned in the summary of the previous chapter, both the R2a1 and R2b5 maximum flooding surfaces are more regionally extensive than R2b1, R2b2, R2b3 and

R2b4. Therefore R2a1 and R2b5 are placed on the points of a fluctuating sea-level curve where the maximum rate of sea-level rise is at a maximum. The periodicity of this curve is estimated to be ~325ka (65ka × 5 marine bands).

Nineteen high-order sequences are identified between the six marine bands used in this study (R2a1 to R2b5). Using the calculated 65ka marine band frequency from Riley *et.al.* (1994), and assuming the ratio of 4 high-order sequences per low-order sequence is constant throughout the Namurian, individual high-order sequences are estimated to occur at approximately 16ka intervals of time (65ka ÷ 4; duration of Namurian marine band divided by number of high-order sequences between marine bands). This figure is similar to the duration of the Quaternary 19-23ka (fifth-order) cycles (Figure 7.2). However, Quaternary fourth-order cycles occur at periodicities of ~122 ka, approximately half that of marine bands (Figure 7.2). The contrast between Marsdenian and Quaternary perturbations may be due to the characteristics of the background subsidence or relative sea-level curve during the late Namurian (Figure 7.1.1 & 7.2). The Marsdenian and Yeadonian represent a period of background rising relative sea level (Veevers & Powell, 1987) corresponding to a period of decreased ice-cap volume in Gondwanaland (Figure 7.1.1, point B). A background rising sea-level would have extended rising higher order relative sea-level limbs. Such conditions would have created aggradational to retrogradation sequence stacking patterns, such as those responsible for the aggradational stacking of sequences R2a1-R2b3 (see Figure 6.1).

7.2 Sea-level curve for the early Marsdenian

As noted in the sections above three relative sea-level fluctuations are observed in the Marsdenian (Figure 7.3.1 and Table 8). When high- and low-order sequence and 'grouped-sequence' sea-level modulation curves are combined, a composite sea-level curve is generated (Figure 7.3.2). The amalgamated sea-level curve reveals attenuation and amplification of sea-level, in a similar manner to the combination of the hypothetical high- and low-order limbs used in the analysis of systems tracts (see Figure 6.3). The composite sea-level curve reveals the positions of sequence boundaries and marine band maximum flooding surfaces (Figure 7.3.2), but also alludes to their relative significance. The combined sea-level curve implies variations in curve gradient vary

	Estimated Frequency (using estimated **65 ka** marine band periodicity from Riley *et al.* 1994).	**Estimated Magnitude** (taken from thickness of uncompacted sedimentary sequences)	**Comparable Order of sea-level change**
High-order sequence	~16 ka	6-20m	Fifth-order
Low-order sequence	**65 ka** (Riley *et al.* 1994)	115m (between sequence boundary and maximum flooding surface)	Two marine bands = fourth-order
Groups of low-order sequences	325 ka (between R2a1-R2b5	?	?

Table 8 Parameters used in the generation of a composite sea-level modulation curve for the early Marsdenian (see Figure 7.3.1).

greatly between individual maximum flooding surfaces and sequence boundaries. Therefore each low-order sequence boundary and low-order maximum flooding surface formed at different rates of falling and rising relative sea-level. This notion is corroborated by the observation that marine bands R2a1 and R2b5, along with the low-order sequence boundary at the base of the R2b4 sequence form markers that are more extensive than other marine bands and sequence boundaries.

The interpretation in Figure 7.3.2 focuses on a detailed part of the curve illustrated by Church & Gawthorpe (1994), and reveals the complex history of sea-level modulation during this interval. The sea-level curve of Church & Gawthorpe (their Figure 11), corroborates the notion that the interval between *R. gracilis* (R2a1) and *R. metabilinguis* (R2b5) was a period of rising relative sea-level that was punctuated by minor falls in relative sea-level. Church & Gawthorpe inferred that the R2a1 sequence represents a fourth-order sea-level rise, which deposited a 'transgressive sequence set', which corresponds with the retrogradationally-stacked R2b1-R2b3 sequence patterns observed in this study (see Figure 6.1). Church & Gawthorpe applied estimates of marine band periodicity (130-200 ka) (Holdsworth & Collinson, 1988; Leeder & McMahon, 1988; Maynard, 1992; Maynard & Leeder, 1992) to designate a fourth-order sea-level cyclicity. After the submission of Church & Gawthorpe's paper, reappraisal of absolute dates of ash-fall tuffs using SHRIMP techniques implied that the estimated duration of the Namurian was 65ka, rather than 130-200ka (Riley *et al.,* 1993). This suggested that two, rather than one marine band cycle approximated to fourth-order (~122 ka) sea-level change during the Quaternary. Therefore during a Namurian fourth-order cycle three marine bands should be deposited (Figure 7.2). This hypothesis is supported by interpretations suggesting R2b4 forms a flooding event of lesser magnitude between R2b3 and R2b5 (see Section 7.1.3).

Apart from the re-estimation of marine band periodicity, the sequence stacking patterns inferred by Church & Gawthorpe are similar those inferred in this study. Church & Gawthorpe illustrated that the R2b1, R2b2 and R2b3 sequences are aggradationally stacked and possess limited evidence for low-order lowstand incision. The observation of aggradational sequence stacking patterns was used to infer that these low-order sequences were deposited during a period of high relative sea-level. This compares with the generally high sea-level that occurs between R2b1 and R2b3, and the aggradational to retrogradational stacking of the low-order sequences during this interval (see Figures 6.1 and 7.3.2).

7.3 Summary of Magnitude and Frequency analysis

The modeled sea-level curve presented here (Figure 7.3.2) fits with the broad Marsdenian- Yeadonian sea-level curve of Church & Gawthorpe (1994). This study has combined the accurate delineation of low- and high-order sequences within the Pennine Basin, along with reduced estimated marine band periodicity to allow re-estimation of the order of sea-level change during the Marsdenian.

Comparisons between Quaternary and Marsdenian sea-level curves suggest that eustatic modulation at magnitudes and periodicities similar to those during the Quaternary controlled deposition during the Namurian. Eustatic modulations in both the Quaternary (Suter & Berryhill, 1985; Pillans *et al.,* 1998) and the Namurian (Veevers & Powell, 1987) are interpreted as the product of glacio-eustatic fluctuations. These are suggested as controlled by long-term quasi-periodic variations in the orbital parameters of the earth. The Milankovitch theory of climate change (Milankovitch, 1941; de Boer & Smith, 1994) has been used to explain the changes in relative sea-level inferred from deep-sea δ18O isotope core data (Pillans *et al.,* 1998). The similarities between Quaternary and Namurian relative sea-level modulations suggest that Milankovitch orbital parameters may have controlled the extent of Namurian polar ice, and therefore eustatic sea-level.

Chapter 8: Palaeogeographic reconstruction

8.1. Introduction: Palaeogeographic reconstructions

This chapter utilises data from field, borehole, core and archived datasets to create isopach maps and delineates the palaeogeography within each sequence (Figures 8.2.1, 8.3.1, 8.5.1, 8.6.1 & 8.7.1). When the results of the sequence stratigraphic analysis are added, the geometries of the incised valleys and incised valley fills are apparent throughout the Marsdenian of the Pennine Basin (Figures 8.2.2, 8.3.2, 8.5.2, 8.6.2 & 8.7.2).

The reconstructions represent the pre-Marsdenian basin geometry (Figure 8.1), the basin geometry at the end of the R2b1 sequence (Figure 8.2), the R2b2 sequence (Figure 8.3), the R2b3 sequence (Figure 8.5), the R2b4 sequence (Figure 8.6) and R2b5 sequence (Figure 8.7). Isopach maps of the sandstone-rich incised valley fill underlying the marine bands are measured in metres, and the geographic extents of sedimentary systems are highlighted. The maps also illustrate palaeocurrent data, which is represented by an arrow and the direction of flow/ number of measurements. Over 200 data points were used to constrain the isopachs and system boundaries. Figures 8.2 to 8.5 are also represent in the sequence stratigrphic paper (Paper Figures 12.1 to 12.3).

8.1.1. Palaeogeography at the end of (R1c) Kinderscoutain Stage

Between the Pendleian and Kinderscoutian, turbidite-fronted deltas were overlain by younger shoaling-deltas that became increasingly fluvial-dominated as they prograded south through the Pennine Basin (Jones, 1980; Collinson, 1988). Individual delta sequences underwent stages of transgressive-regressive development that were modulated by high-frequency/ low magnitude fluctuations in relative sea-level, generated by the waxing and waning of ice-masses on the Gondwanan super-continent (see Section 7.1).

By the onset of the Kinderscoutian, deltas entering the basin from the north-east had part-filled the Huddersfield and Rossendale sub-basins and the Gainsborough Trough (Collinson, 1968a; McCabe, 1975; Collinson *et al.,* 1977; Jones, 1980). The southern margin of the turbidite-fronted Kinderscoutian delta onlapped the Derbyshire Dome in the south, and entered the north-south trending Goyt Trough. The northernmargin of the main Kinderscoutian depocentre lies slightly basinward of the Craven-Quernmore Fault system (Figure 8.1).

The integration of thickness and distance estimates from the base of delta-slope to delta-top allows a crude estimation of the angle of slope on the submerged delta-front at the end of the Kinderscoutian. Between the basin floor and delta top, the Kinderscoutian delta has a compacted thickness of 100-350m (Bromehead *et al.,* 1933; Walker, 1966a; Hampson, 1997). The distance between the delta top and condensed basinal sections in the area west of the Lask Edge Fault (Collinson *et al.,* 1977; Jones, 1980) is approximately 10 kilometres, implying the gradient of the Kinderscoutian delta-front was between 0.6-1.7°. This angle is probably underestimated, as the effect of compaction is not taken into account. Additionally, the precise definition of the shelf-slope break for the Kinderscoutian delta is enigmatic, due to the paucity of outcrop data in this area.

By the end of the Kinderscoutian underfilled sub-basins lay to the west, south and north-west. It is these sub-basins that became the focal areas of delta deposition during the Marsdenian.

8.1.2. Observations: the R2b1 sequence

The R2b1 sequence isopach-palaeogeographic map (Figure 8.2.1) reveals the thickest areas of sediment accumulation are centered in the Rossendale sub-basin (up to +40m) and the Harrogate sub-basin (up to 20m). Data in the western part of the study area is restricted to localities along the inverted Pendle Monocline structure (Alum Crag; SD636280), and borehole data sets (Roddlesworth, SD656123 and Heywood SD830090). Data-points are therefore scarce, and the isopachytes are poorly constrained (Figure 8.2.1). In the Harrogate sub-basin, a high-density of borehole and field localities allows the delineation of a depocentre, that although bifurcating, trends in a linear NE-SW direction. Towards the south, the isopachytes continue this trend, with a southerly-trending thickened lineament developing into the Gainsborough sub-basin, and south-westerly into the Rossendale sub-basin. Turbidite deposits (the Alum Crag Grit) are underlain by a R2 surface in the western areas, while basinal deposition prevailed in the areas of non-deposition to the south (see Figure 6.2.2). An underfilled slope sector that coincides with the submerged Kinderscoutian delta-slope separates the turbidite-dominated western sub-basins from the mouthbar-dominated eastern sub-basins (see Figure 8.2).

The distribution of depositional systems reveals that fluvial deposits are restricted to the north-east of the basin, while mouthbar systems lie to the south and west. The base of the fluvial deposits is marked by a R1 type surface, while an R4 surface delineates the base of the mouthbar succession. These deposits correlate with the Alum Crag Grit in the Rossendale sub-basin.

R2b1 sequence interpretations

Fluvial systems that entered the north-eastern part of the Pennine Basin during the Kinderscoutian persisted supplying sediment throughout the Marsdenian. On the submerged Kinderscoutian delta-top in the north-east, a thickened succession, as denoted by the isopachytes (up to 40m +), and the dominance of fluvial facies suggests the presence of a fluvial channel belt. The area with the thicker succession coincides with the junction of the Morley-Campsall and Craven Fault system. It appears that this fault junction was a long-lived locus for fluvial transport, and a tectonic control is therefore suggested as responsible for the positioning of this basin entry point. A comparison of the relative positions of the Kinderscoutian and Marsdenian delta system suggests the R2b1 sequence lies aggradationally on the Kinderscoutian delta. It is therefore likely that the basin entry points used during the Kinderscoutian (Figure 8.1) were reused during the R2b1 sequence (Figure 8.1 & 8.2.1).

During deposition of the R2b1 sequence, the entry point in the north-east supplied mouthbars in the Huddersfield sub-basin. The R4 surface that underlies the tidally-influenced mouthbar succession forms part of the sequence boundary, which correlates with an R1 surface in the northern part of the Huddersfield sub-basin (see Enclosure 2). Tidally influenced deposits lie within the incised valley, and represent the infill of the valley during the transgressive systems tract (Figure 8.2.2). The isopachyte distribution and the SW and SSE palaeocurrent trend implies that two delta-lobes prograded from the mouth of the incised valley towards the south and west of Huddersfield (Figure 8.2.2). This suggests that during the period of lowered relative sea-level two mouthward migrating shoaling delta systems developed in the basinward part of the incised valley, in an area overlying the submerged Kinderscoutian delta top.

Although there is a lack of evidence for deformation or mass-movement on the submerged Kinderscoutian delta-front, the gradient the submerged slope possesses (0.6-1.7°) would have been susceptible to gravitational mass-transport processes. Progradation of R2b1 sequence may have led to excessive loading in the Kinderscoutian delta system, and the initiation of mass-creep and slides or slumps (Figure 8.2.2).

An additional clastic entry point appears to have commenced supplying sediment in the underfilled, north-western part of the basin. A lobate delta-front prograded across the Bowland and Rossendale sub-basins during the R2b1 sequence, and comprises delta-front and basin-floor turbidite fans. Sections provided by the Blackburn Sewer Tunnel infer up to 40m of turbidite deposits in this part of the basin (Figure 8.2.1; Collinson *et al.* 1977). These overlie an R2 surface, which represents the correlated equivalent of the R1 and R4 surface in the proximal incised valley. The presence of erosive or concordant contacts and the identification of downlap surfaces allows the relative position of the turbidite deposits to be placed in relation to the delta front slope (Posamentier & Allen, 1999). On delta front slopes, the falling or lowstand systems tract turbidites have an increased potential to erode due to the increased amount of bypassed sediment. As turbidity currents are driven by the increased potential energy associated with the presence of a slope, they generate channel systems, down which the turbidites are funneled. The outcrop at Alum Crag (SD636280; see Figure 4.12 and Paper Figure 13) reveals an erosive R2 surface, which is overlain by several coalesced channelized turbidites. This suggests that turbidite deposition in this area had a capacity to erode. This is inferred as the product of either increased flow density because of proximity to a sediment-rich effluent plume, or increased potential energy due to its positioning on a delta front slope. The lack of laterally extensive turbidite exposures in the area to the south of the Alum Crag locality does not allow the identification of downlap basin floor surfaces, so the true extent of the channelised turbidite system cannot be assessed. However, the limited lateral extent of the Alum Crag Grit is inferred by feature mapping (Price *et al.,* 1963). The presence of a basal R2 surface with erosive relief in the order of tens of metres (see Figure 4.12), along with the map outcrop pattern suggests that the Alum Crag Grit forms a broad lens shaped lithostratigraphic unit. The correlation panel (Enclosure 1) identifies an intra-sequence marine band flooding event within the thickest part of the Alum Crag Grit in the

Rossendale sub-basin (Collinson *et al.,* 1977). This flooding event does not appear to have an obvious motif within the mouthbar and fluvial-dominated incised valley fill to the east of the basin. The lack of a marine signature in the northern part of the basin suggests that fluvial systems may have filled most of the proximal incised valley, reducing the potential for minor marine transgressions to flood the valley. This is confirmed by the observation of palaeosols in the north-eastern part of the basin. Palaeosol motifs are rare during the R2b1 sequence, but the mud-rich palaeosols that cap mouthbar deposits and suggest emergence in the extreme north east of the basin.

8.1.3. Observations: the R2b2 sequence

As with the R2b1 sequence isopach distribution pattern, the R2b2 sequence dataset suggests that an east-west divide existed between a thicker western zone (up to 40m), and a thinner eastern zone (10-20m). Isopachyte distribution patterns imply that the underfilled zone separating the east-west depocentres during the R2b1 delta cycle was less defined during this period (Figure 8.3.1).

A high-density borehole and field section dataset in the south-west Huddersfield sub-basin permits the accurate definition of isopachytes in this area. The spatial and temporal facies distribution on the map and correlation panel suggests shallow-water deltas overlying R1 and R4 regressive surfaces existed in eastern parts of the basin (equivalent to the East Carlton Grit and Readycon Dean Flags). In western areas R2 regressive surfaces are overlain by turbidite-fronted deltas, then a R4 surface, and a thin mouth-bar system in western areas (equivalent to the unnamed sandstone of Collinson et al. 1977). Fluvial systems are broader, and located in slightly more basinward areas than those during the R2b1 sequence. The NE-SW trending fluvial system that persisted during the R2b1 sequence appears to have widened, and developed an E-W trend in the mid-Huddersfield sub-basin. Mud-rich palaeosols commonly overlie mouthbar fronted deltas in the north-east of the Pennine Basin, but are absent in the area to the west of the submerged Kinderscoutian slope-shelf break. The geographically restricted 'Keighley Bluestones' were deposited in the northern Huddersfield sub-basin during this period.

R2b2 sequence interpretations

Comparison of the R2b1 sequence and R2b2 sequence isopach data reveals that the later system formed a thinner aggradationally stacked delta. After the R2b1 marine incursion, mouthbar deltas, some appearing tidally influenced, prograded from the clastic entry points in the north-west and north-east. These are overlain by thin bayhead deltas in the west of the basin. Synchronous with the deposition of shallow-water deltas on-top of the submerged Kinderscoutian delta in eastern part of the basin, deeper water turbidite-fronted deltas in western areas prograded across the basin that had remained underfilled since the early Namurian. The presence of an east-west divide between shallow water deposition in the Huddersfield sub-basin and deep-water deposition in the Rossendale and Bowland sub-basin during the R2b2 sequence is inferred by the presence of mud-rich palaeosols in the bayhead deltas in the eastern area. Age equivalent intervals towards the west comprise delta-front mouthbars, implying deposition in deeper water environments. In this area, an isopach thickness of 40m + towards the hanging wall trough of the Rossendale sub-basin represents the thick turbidites underlying the shallower delta-front deposits.

Incised valleys are delineated by R1,R2 and R4 surfaces with the similar characteristics as those in the R2b1 sequence. Unlike the R2b1 sequence, the R2b2 sequence has not one, but four basinwide regressive surfaces. The lowest of these surfaces, at the high-order sequence boundary at the base of HS4 partly erodes the R2b1 marine band in the central part of the Pennine Basin. This suggests that the incised valley formed during the R2b1 sequence was generated during HS5 and HS6 , and is therefore diachronous. Additionally, the lateral extent of the R2b2 sequence incised valley is greater than the R2b1 incised valley (Figure 8.3.2). This is because the R2b2 sequence reused and widened the incised valley generated during the R2b1 sequence. The R2b1 transgression forced retrogradational stacking of the R2b2 sequence, providing a wide but shallow area of accommodation basinward of the fluvial entry point. The partial infilling of the basin by the R2b1 sequence implies that relative sea-level falls in this area influenced a wide geographic area of relatively shallow water (estimated at less than 30m deep from the thickness of the R2b2 delta system). This explains why the incised valley system formed and infilled during the R2b2 sequence has a broader geographic extent than the incised valley during the R2b1 sequence.

Comparison with the basin configuration model (see Figure 1.2.2; Lee 1988) suggest an area of thinner isopachytes occurs close to the northern margins of the Huddersfield sub-basin (Figure 8.3 & 8.4). The base of the Huddersfield sub-basin comprises the southerly dipping hanging wall of the Craven Fault system. The southerly directed slope of the hanging wall might have increased the potential for bypass during the incised valley formation; explaining why the deltaic deposits are less than 10m thick in this area. The thinning of the R2b2 sequence was also noted by Waters *et al.* (1996). Waters *et al.* noted that the Keighley Bluestone outcrops in this area, and that it is bounded by the Craven Fault system to the north and the antithetic Denholme Clough Fault to the south. Interestingly, the 'Keighley Bluestones' contain restricted marine faunal indicators (*Zoophycos*, *Lingula*, and sponge spicules) suggesting deposition below sea-level. The integration of the structural model implies that the Keighley Bluestone developed on an area of raised sea-bed topography. It is suggested that this formed as a consequence of its location on the south-easterly sloping high of a graben between the Denholme Clough Fault and the Morley-Campsell Fault. Evidence for a structural control is also inferred twenty kilometres to the south, at an intriguing locality (Great Mount Quarry SE009275; Figure 8.4). This locality comprises the mouthbar deposits of the Readycon Dean Flags. Occurring within the mouthbar succession are a set of NE-SW trending syn-depositional minor faults (see Figure 2.11). Field-mapping to the north of Hebden Vale reveals that Great Mount Quarry lies in the horst block formed by two small tectonic faults, striking NW-SE with throws between 5 to 15m (Waters, *pers comm.*). Modelling of gravity data (Lee, 1988) and recent BGS field mapping has not highlighted any major tectonic-faults in this area, although a NE-SW extensional regime may correspond with that generated by faults antithetic to the Denholme Clough-Morley Campsall Fault (Figure 8.4; also see Paper Figure 16). Similar syn-depositional faults are described in the Rough Rocks Flags in outcrops adjacent to the Denholme Clough Fault (Waters *et al.,* 1996a). A seismic trigger for syn-depositional extension seems probable, but the presence of a sloping front to the mouthbar, and gravitational collapse of the oversteepened and overloaded N-S oriented delta front may also have been a major driving force. Deposition in the eastern part of the basin occurred in mouthbar-dominated delta-front deposits that were interbedded with rare turbidites (? instability, seismic or fluvial underflow triggered). These prograded in a southerly direction, crossing the Holme Disturbance, and entered the Gainsborough Trough.

The significance of the forced regressive systems tract and R4 (the regressive surface of marine erosion) in respect to its palaeogeographic position

Rapid lateral variations in isopachyte thicknesses in the southern Huddersfield sub-basin suggests that the area basinward of the fluvial-mouthbar incised valley fill comprises a series of mouthbars of non-uniform thickness. These mouthbar deposits are underlain by R4 surfaces. The mechanisms behind the formation of the R4 surface are enigmatic until it is placed in its palaeogeographic and sequence stratigraphic context.

The R4 surface is restricted to wide areas of relatively shallow-water mouthbar deposition. The modulation of relative sea-level in these areas provides and destroys large areas of accommodation. Such rapid elimination of accommodation allows the formation of forced regressive systems tract mouthbar deposits. This partly explains how the R4 surface propagates across the submerged Kinderscoutian delta platform (Figure 8.2.2 and 8.3.2). During the R2b2 sequence, mouthbar-dominated low-order forced-regressive systems tract are deposited on shelfal or platform areas, and R1 and R2 regressive surfaces move basinward in northern parts of the basin. The southern Huddersfield sub-basin therefore represents an area where sediments bypassed down the incised valley are deposited on the platform, and not transported directly down the delta front slope. During the subsequent transgressive systems tract, sediment was stored within the incised valley preserving the forced regressive deposits on the flooding platform.

8.1.4. Observations: the R2b3 sequence

The low-order sequence underlying the R2b3 marine band (the Woodhouse Flags) occupies a slightly less basinward position than the R2b2 sequence, and isopachyte distributions in the western zone suggest this succession is slightly thinner (maximum of 30m; compared to 40m in the R2b2 sequence; Figure 8.5.1). In the Goyt Trough / western Gainsborough Trough deposition of basinal mudstone occurred throughout the R2b3 sequence. Facies distributions also suggest that the R2b3 sequence has shifted landward in relation to the previous interval.

In the eastern part of the Pennine Basin, fluvial systems are concentrated in N-S and NE-SW bands, which are concordant with a 20m thick NE-SW trending mouthbar system in the southern Gainsborough Trough. While the fluvial systems are underlain by R1 surfaces, the mouthbar depositional system is underlain by an R4 surface. The Bowland and Rossendale sub-basins are dominated by turbidite and mouthbar systems, which are underlain by R2 and R4 surfaces and attain a maximum thickness of 30m. Palaeosol horizons are also present, but restricted to eastern and northern areas where mouthbar areas became emergent (Figure 8.5.1).

R2b3 sequence interpretations

In the south-east of the basin, the R2b3 sequence is thinner and does not prograde basinward to the extent of the R2b2 delta front. Fluvial feeder systems erode the top of the mouthbar-dominated delta front, and palaeosol horizons commonly overlie individual delta-top facies associations in the east of the basin. The presence of palaeosols implies an increased extent of emergence; while mouthbar and turbidite facies infer the prevalence of deeper water conditions in the western basin.

As with the R2b2 sequence, the dichotomy of emergence and non-emergence represents the bathymetry inherited from the Kinderscoutian delta systems (Figure 8.1). The incised valley that formed during the R2b3 sequence is retrogradationally stacked on the R2b2 sequence, suggesting that R2b2 was of a magnitude significant to force the delta system to back-step out of the basin (Figure 8.5.2). While a bathymetric step existed between the western turbidite and the eastern mouthbar-dominated lowstand/ incised valley fill during the R2b2 sequence, the Woodhouse Flags incised valley fill contains mouthbar deposits almost across the full width of the basin (Figure 8.5.2).

The Goyt Trough and western Gainsborough Trough area remained underfilled during the R2b3 sequence. Overlying the area of mouthbars formed and abandoned during the R2b2 sequence, mouthbars of laterally varied thickness prograded into the eastern margin of the underfilled zone. Comparison with the structural element map reveals that the northern margin of the underfilled zone approximately coincides with the Holme Fault/ Disturbance. The fault block configurations suggest the Gainsborough Trough and Huddersfield sub-basin share similar orientations (Lee 1988). These fault blocks possess northerly highs, which shallowly slope in a southerly direction. It is plausible that bypass down the sloping footwall accounts for the aggregation of mouthbar deposits in the southern Gainsborough Trough. The elongated geometry of the isopach map in this area also suggests a south-west directed progradation, rather than a linear basinward shoreline migration. This coincides with the south-westerly flowing fluvial feeder system.

Further evidence of tectonic activity in the northern Huddersfield sub-basin is inferred from the massive dewatering-slump structures observed in the Woodhouse Flags at Bingley Road Quarry (SE053382; see Figure 2.26). The area in which the dewatering-slump structures occur lies between the South Craven-Denholme Clough Fault. It is suggested that this area experienced seismic events of a significant magnitude that triggered massive de-watering events. Similar de-watering-slump features in the Stanningley Rock (Langsettian), which outcrops 15km to the east adjacent to the Craven Fault system (Waters *et al.,* 1996a) and in the lower Carboniferous deposits of the Northumberland Basin (Leeder, 1987).

The underfilled zone of the Goyt Trough / western Gainsborough Trough formed a wide, shallow bay, bordered by a shoreline to the east and a shallower area of active mouthbar deposition to the west and north-west. Amplification of basin swell waves within this semi-enclosed, but basinward-facing bay explains the presence of the late transgressive systems tract wave-dominated shoreline observed at Foster Delph Quarry (SE022273; see Figure 4.7). The wave ripple crests should lie approximately concordant with the shoreline orientation, suggesting an averaged orientation of 75-255º (from seven readings), implying that the shoreline lay at approximately ENE-WSW orientation.

In the west of the Pennine Basin, turbidite fronted mouthbar delta systems were deposited in an increasingly shallowing basin. The absence of palaeosols in the Bowland and Rossendale sub-basins suggests that this area did not fill to emergence, except in the northern areas adjacent to the Craven Fault system.

8.1.5. Observations: the R2b4 sequence

The isopach map of the R2b4 sequence reveals an increased thickness of the delta system compared with the previous delta systems (Figure 8.6.1). The isopach data in the south is constrained by few data points, but isopachytes appear to be E-W oriented, while northerly protuberances appear in line with the axes of fluvial systems. A dichotomy still exists between the thinner delta (10-20m) system in the east, and the thicker (30-50m) delta system in the west, where turbidite systems occupy the basinal areas. Unlike the older delta systems, areas of fluvial and mouthbar deposition appear to thin in a proximal direction. The incised valley fill between the low-order sequence boundary and the R2b4 marine band correlates to the Midgley Grit lithostratigraphic unit. The base of the Midgley Grit is diachronous, between HS12 and HS15, and the base is marked by both R1 and R4 regressive surfaces.

In the east of the Pennine Basin, fluvial systems and palaeosols occur in a more basinward position than those in the west, following the trend observed in the previous delta systems. Analysis of the correlation panels reveals three laterally correlatable high-order maximum flooding surfaces, and several palaeosol horizons occur within the Midgley Grit (see Enclosures 1 & 2). Leached palaeosols rework underlying fluvial sediments, but only exist on the basin margins away from the sub-basin depocentres. Palaeocurrent data suggests a generally southern direction of transport, although some measurements suggest up ninety degrees variance to the east or west.

R2b4 sequence interpretations

The Midgley Grit is dominated by mouthbar and fluvial systems, which are up to 50m thick in western zones of the Pennine Basin. The diachronous base of the Midgley Grit suggests the R1,R2 and R4 surfaces that define the sequence boundary are produced during separate periods of low order base level fall (see Chapter 6). The western incised valley fill is dominated by multi-storey fluvial sandstones that are separated from mouthbar sandstones by R1 surfaces. Enclosure 1 reveals that the mouthbar system thins from the east to the west, following a general shallowing trend generated by the inherited bathymetry of the Kinderscoutian and earlier Marsdenian delta systems.

The R2b4 sequence is thinner in parts of the basin away from the main incised valley axes (Figure 8.6.1). During the low-order lowstand systems tract sediment bypassed the delta top through fluvial systems. This is carried to the submerged Kinderscoutian delta platform edge, where the effluent plume is influenced by gravitational forces and transported deeper into the basin by turbidity currents. This produced a basinward-stacking facies distribution and delta-front progradation. Bypass of the delta-top via the incised valley accounts for lateral thinning in areas marginal to fluvial supply pathways. The thinner isopach area also corresponds to interfluve zones possessing freely-drained 'ganisteroid' palaeosols (Figure 8.6.2). The combination of the fluvial system, with the development ganister palaeosols, and the basinward-stacking of the delta system suggests that an enlargement of the incised valley may have generated the interfluve areas, a wide bypass zone and the pronounced delta front progradation (Hampson, 1997).

In the southern part of the Pennine Basin, basinward of the incised valley mouth, turbidite systems form an extensive fringe on the E-W trending delta front slope. These lowstand systems tract turbidites form a band across the south of the basin; suggesting the merger of delta systems which previously developed from separate incised valley conduits in the north-west and north-east.

The fluvial-mouthbar depositional system contact within the incised valley lies in a basinward position and forms a more linear feature in comparison with previous incised valley fills. The R2b4 incised valley is wider than those in the previous sequences, suggesting that either the R2b4 sequence incised valley underwent a protracted period of fill, or rates of deposition were greater (see Section 9.2.3.1). Rare meandering fluvial channels are observed in the upper parts of the R2b4 incised valley fill suggesting a decreased graded profile occurred within the incised valley during the late R2b4 sequence (Fletcher Bank Quarry; SD805164, Leicester Mills Quarry; SD619164; see Figures 2.27 and 4.9). Isolated tidally-influenced fluvial facies occur in the eastern part of the incised valley during the R2b4 sequence (see Figure 2.22.1 from Moselden Heights Quarry, Enclosure 1, locality 12). This suggests that the incised valley was open to marine waters, allowing the amplification of tidal processes.

Palaeocurrent data from the R2b4 sequence suggests palaeoflow within the incised valley is concordant with the pathway trend of the incised valleys from sequences R2b1 to R2b3. Throughout the R2b4 sequence, clastic material was supplied from a north-westerly and north-easterly direction. In the Rossendale and Huddersfield sub-basins palaeocurrent data suggests both palaeocurrent trends occur in a relatively small area. Detailed re-examination of palaeoflow from individual high-order sequences may elucidate directions suggesting clastic input from both sources.

8.1.6. Observations: the R2b5 sequence

Distribution of the isopachytes reveals that the R2b5 sequence is more uniform in thickness than the underlying sequences (Figure 8.7.1). Thicker isopachytes exist in the Rossendale sub-basin, while the R2b5 sequence thins to completely to the north-east of the Pennine Basin. In this area the R2b5 marine band directly overlies the correlated T2 equivalent of the R2b4 marine band, separated by as little as one metre of mudstone (Enclosure 1, localities 3 and 5).

In the eastern and southern parts of the Pennine Basin the correlation panels reveal that the R2b5 sequence comprises bayhead mouthbars, with several flooding surfaces (T4 and T5 surfaces), which can be correlated across the Pennine Basin. Individual bayhead delta systems have sharp correlatable bases (R4 surfaces) and are commonly capped by mud-rich palaeosols that suggest emergence. To the north-west of the Quernmore Fault, shoreface deposits dominate the bayhead system. The wide lateral spread of data makes the estimation of the width of the fluvial supply systems difficult, and although sparse, the palaeocurrent data suggests a wide variation in palaeocurrents (between south-east and west).

R2b5 sequence interpretations

The uniform distribution of the R2b5 sequence implies that deposition occurred over a broad area, in which the underlying bathymetric topography (depositional or tectonic) had less control than during the formation of the underlying systems. It is likely that the large volumes of sediment deposited in the R2b4 incised valley may have filled sub-basins, creating a broad, flat delta-plain by the time of the R2b4 transgression. The R2b4 transgression formed a laterally extensive area of accommodation in the incised valley, within which the bayhead deltas of the Helmshore Grit were deposited. The presence of a generally flat delta plain at the onset of the R2b5 sequence is corroborated by the high correlation potential of the minor flooding events that occur within the bayhead system (see Enclosures 1 and 2). Progradation of the R2b5 sequence high-order sequences was not inhibited to same the extent as the older Marsdenian deltas, where differential bathymetry decreased the degree to which flooding events could be correlated.

The R2b5 sequence is also retrogradationally stacked within the incised valley to the extent that it onlaps the palaeogeographic high formed by the combination of the delta depocentre in the east and the Askrigg Block in the north. The bayhead delta system as a whole appears to fill the incised valley formed during the R2b4 sequence (Figure 8.7.2).

Analysis of the palaeocurrent data suggests a range of flow directions within the incised valley fill. The lack of pronounced fluvial supply pathways suggests either a decrease in the amount of fluvial input into the basin or more likely, the divergence of supply systems across the broad and flat delta plain. Such a divergent set palaeoflow would be expected on a flat delta top. Under these conditions the lowered graded profile of the supply system during the late transgressive systems tract may allow the formation of meandering river-systems, or increase the potential for rivers to avulse. Tidally influenced fluvial and mouthbar systems occur throughout the basin, suggesting the incised valley was still partly submerged, providing an area where tidal amplification processes operated (Figure 8.7.2).

The presence of the shoreface deposits to the north of the Quernmore Fault block (Middleton Towers; SD409588) suggests localised, but significant basinal wave reworking during this period (Figure 8.7.2). Unlike the R2b3 sequence, where shoreline deposits are identified, the paucity of field localities implies that the position and extent of the coastline cannot be delineated. However, the position of this locality is on the eastern margin of the incised valley, where shoreline environments may have formed marginal to areas of high sediment input. Mud-rich palaeosols occur throughout the R2b5 sequence, and are present as far south as the Goyt Trough (Quick Edge Quarry; SD967036). This implies that by the end of the sequence, progradation had infilled the incised valley to at least as far as this position.

8.2. Summary and comparisons with previous palaeogeographic reconstructions

The integration of interpretations from field, borehole and well-log data has allowed the generation of a series of palaeogeographic and sequence stratigraphic time-slices through the Marsdenian of the Pennine Basin. This extends previous reconstructions (Collinson *et al.,* 1977; Jones, 1980; Church, 1994) and provides a insight into the large-scale depositional architecture of Pennine Basin. The lateral change from the deeper water deposits to the west, and shallow water deposits to the east is a striking example of how shelf-edged deltas may develop synchronously to deep-water turbidite lobes in other parts of the basin. Such interpretations imply that relative sea-level modulation was not the only control mechanism, but the timing and amount of sediment input was also important in dictating basin infill distribution (Church & Gawthorpe, 1997). Similar conclusions drawn by Church & Gawthorpe (1997) in the Marsdenian- Yeadonian of the east-Midlands, where both autocyclic and eustatic mechanisms are suggested to control deposition during the Namurian.

Regional palaeogeographic reconstructions (Collinson *et al.,* 1977; Jones, 1980) illustrate the palaeogeographic configuration of the Pennine Basin during the Namurian (Figure 8.1). They suggest the main Marsdenian depocentre lay in the Bowland and Rossendale sub-basins, where the thickest Marsdenian succession occurs (see Enclosure 1). The palaeogeographic maps of Church (1994) (his page 253 & 254) illustrate reconstructions of the system equivalent to the R2b3 and R2b4 sequences. These maps contain fairly crude interpretations of the facies distributions, but share several similar and dissimilar elements with the palaeogeographic interpretations presented here. Up to the end of the R2b3 sequence the fluvial supply pathways according to Church's model comprised a NNE-SSW trending fluvial system in the southern Huddersfield and Harrogate sub-basins, and a ENE-WSW trending system in the north-west Huddersfield sub-basin. By comparison, these fluvial supply pathways are similar in position and orientation to those illustrated on Figure 8.2.2, 8.3.2, 8.5.2, 8.6.2 and 8.7.2.

The R2b4 sequence reconstruction for the late lowstand systems tract (Church 1994; page 254) suggests the presence of a main E-W trending incised valley, with a minor N-S valley running parallel to the Lask Edge Fault lineament. This study suggests that palaeocurrents trend in N-S direction, rather than E-W. Unlike the interpretation presented here (Figure 8.6.2), no reference is made to a north-westerly clastic basin entry point in the Rossendale and Bowland sub-basins. With the inherited bathymetry from the platform edge of the Kinderscoutian delta it is possible that some flow may have been directed in a westerly direction. However, with sediment supply in the east of the basin directed towards the south down the main incised valley, it seems unlikely that such a major incised valley with this orientation existed in this part of the basin.

Only one previous paper has published palaeogeographic reconstructions of the Marsdenian Stage in the Pennine Basin (Wignall & Maynard, 1996). The isopach maps produced by Wignall & Maynard (1996) are similar in some respects to those presented in this study. Of particular note is their delineation of a NE-SW trending distributary feeder channel within the interval they termed the Scotland Flags (Figure 8.2.2). However, the correlations made during this study disagree with those applied by Wignall & Maynard. The unit Wignall & Maynard term as the Scotland Flags is inferred as the lateral equivalent to the Alum Crag Grit (see precise of Wignall and Maynard in the sequence stratigraphic paper). The reconstruction they made for the Alum Crag Grit interval also suggests that an incised valley feeder system fed the Alum Crag Grit from the south-east. The palaeogeographic reconstructions made during this study reveal that not only are palaeoflow indicators trending in a southerly direction, but such an incised valley would have to had formed parallel to the shoreline (Figure 8.2.2).

PART 4:
SYNTHESIS

Chapter 9: Controls on deltaic deposition within the Pennine Basin

9.1. Introduction

This chapter integrates all the interpretations made in the previous eight chapters, with the aim of focusing on the mechanisms that controlled deposition within the Marsdenian Pennine Basin. The Marsdenian succession lies between the turbidite-fronted deltas of the Kinderscoutian and the fluvial plain/ coal-rich Westphalian strata, and it shares characteristics of both intervals. As a result, sedimentological and stratigraphic elements of both the Kinderscoutian and Westphalian can be studied and compared in what appears to be a key period during the evolution of the Pennine Basin.

9.2 Mechanisms controlling deposition in the Marsdenian Pennine Basin

9.2.1. Basin Geometry

Topography associated with the Dinantian syn-rift stage of basin formation was an important control on the geometry of the Marsdenian Pennine Basin. On the uplifted Askrigg and Alston Blocks to the north of the Pennine Basin, deposition was condensed (Ramsbottom *et al.,* 1978; Church, 1994). Additionally, a structural high west of the Quernmore Fault, created by re-activiation during the late Dinantian/ early Namurian probably inhibited the lateral migration of the depositional systems (Brandon *et al.,* 1998). This structural high explains the condensed/ thin Marsdenian deposits in the western part of the study area.

9.2.1.1. Geometry of basin fill controlled by the topography of early Namurian deltas

Topography formed by older Namurian deltas influenced the distribution and thickness of depositional systems during the Marsdenian. During the interval between R2a1 and R2b3, the topography of the submerged Kinderscoutian delta led to the creation of deeper basinal waters in the Bowland and Rossendale sub-basins, and shallower waters in the Huddersfield sub-basin. Early Marsdenian deltas were therefore deposited on a platform provided by the submerged Kinderscoutian delta system, and in deeper waters where the basin remained underfilled during the Kinderscoutian.

Although turbidite systems in the Marsdenian are thinner than those of the Kinderscoutian (10-50 m; compared with up to 250m in the Kinderscoutian), they share similar sedimentary structures, facies and facies architecture (Collinson, 1988). Throughout the Namurian, gradual infilling of accommodation produced by Dinantian rifting probably restricted the lateral extent of available accommodation. The decreased accommodation available in the Marsdenian compared with that during the Kinderscoutian probably controlled the difference between the scale of these successions. During the Kinderscoutian deltas prograded from the north, filling a significant proportion of the basin. A gradual reduction in the number of turbidite systems throughout the Namurian coincidenced with the infill of the initial syn-rift basinwide accommodation. Additionally, topography on the basin floor had the potential to control turbidite systems transportation pathways, implying that as turbidites pass down-slope, they bypassed areas of topographic relief on the basin floor (older lobes, tectonic high areas).

9.2.1.2 Topographic controls on mouthbar distribution

Mouthbars form seabed topography in areas basinward of the river mouth. Compensational stacking of subsequent deposits would be expected if the mouthbars created significant topography on the seabed. The deposition of bayhead or mouthbar-dominated high-order sequences leads to emergenceand shoreline progradation. Autocyclic switching during the deposition of bayhead or delta front mouthbar high-order sequences occurs in response to decreased accommodation. High-order accommodation is the product of high-frequency, low-amplitude eustatic increases in relative sea-level (Section 7.2). Decreasing accommodation adjacent to the sediment supply point leads to an autocyclic shift in the depocentre. Autocyclicity within high-order sequences is therefore driven by the capacity of sediment supply system to fill accommodation. This mechanism drives delta progradation within the high-order sequence, generating extensive mouthbars. It is unlikely that autocyclic shifts in accommodation during high-order sequence deposition influenced the distribution of entire delta sequences.

Eustatic sea-level modulates the stacking pattern of low-order sequences during the Marsdenian (Chapter 7). Low-order sequences dominated by mouthbars form extensive sheets that were predominantly formed during periods of low-order forced regression (Section 6.6.1). Accommodation destruction during the low-order forced regressive systems tract led to increased bypass and the increased likelihood of seabed reworking at mouth of the river. Erosion basinward of the river mouth may have eroded sea floor topography, diminishing the influence compensational stacking had on the distribution of mouthbar deposits.

9.2.1.3. Position of clastic basin entry point

Palaeocurrent data suggests that fluvial supply pathways initiated during the Kinderscoutian were reused and modified during the Marsdenian (see Palaeogeographic maps in Chapter 8). Entrenching of the fluvial systems may have led to the ‘fixing’ of fluvial supply pathways during the Kinderscoutian, which developed into an antecedent or *inherited drainage* system (Summerfield, 1991) during the Marsdenian. Fixing of the entry point north of the Huddersfield sub-basin during the Kinderscoutian, and its re-use throughout the Marsdenian, may explain why Marsdenian deltas are stacked on those deposited during the Kinderscoutian. The Kinderscoutian/ Marsdenian clastic entry point is situated close to an area where several major fault lineaments converge with the Craven Fault system (see Figure 8.1). This fault-zone provided a supply pathway through which the Huddersfield sub-basin was supplied with sediment during the Marsdenian and Kinderscoutian. The paucity of outcrop in the area adjacent to the Craven Fault system does not allow the delination of locally sourced footwall or hangingwall fans, such as those seen in Pliocene-present day rift zones (Eliett & Gawthorpe, 1995). The dominance of a southerly directed palaeoflow suggests a transfer zone formed in the faulted area and focused antecedant drainage. The Craven Fault zone therefore forms a *synthetic interbasin transfer zone* (Gawthorpe & Hurst, 1993) that links the block area to the north with the basin to the south.

9.2.2. Subsidence processes and basin tectonics

Thermal subsidence was the principle tectonic control on sedimentation in the Pennine Basin during the late Namurian (Leeder, 1982; Leeder & McMahon, 1988). Averaged thermal subsidence rates of 0.015mm per year are inferred for the Northumberland, Stainmore and Bowland Basins during the Namurian (Leeder & McMahon, 1988). These rates are low, suggesting that 2 Ma of thermal subsidence would be necessary to create the average 30m of accommodation required for the deposition of a deltaic sequence. In order to calculate subsidence rates, the duration of the Namurian was estimated at 26 Ma by Leeder & McMahon (1998). This implies that subsidence alone would have had little control on the total amount of accommodation generated during the Namurian. The increased accuracy of dating by SHRIMP techniques has decreased the estimated duration of the Namurian to approximately 5.5 Ma (Riley *et al.,* 1993). The estimated duration of 5.5 Ma suggests the rate of thermal subsidence therefore occurred at approximately 0.07mm per year. The average low-order sequence of 30m thickness would have therefore taken 0.43 Ma to form if thermal subsidence (at a rate of 0.07mm per year) alone was wholly responsible for accommodation production.

9.2.2.1 Tectonically driven subsidence

Recent research has identified localised areas of active tectonic subsidence in the Silesian Pennine Basin. Analysis of the Westphalian succession in the South Staffordshire Basin has highlighted the presence of local syn-sedimentary growth faults (Waters *et al.* 1994). Additionally, studies of the Rough Rock (Yeadonian) have revealed the widespread influence of structures formed by syn-depositional tectonic processes. Such syn-sedimentary faults and dewatering structures appear closely associated with basin-scale fault trends (Bristow, 1987), suggesting synchronous tectonic and depositional processes. The localised extent of the Keighley Bluestones (see sequence stratigraphic paper) appears to correlate with the footwall-high of the Huddersfield sub-basin (Lee, 1988). Waters *et al.* (1998) recognised the outcrop limit of the Keighley Bluestones as lying in the area between the Aire Valley and the Denholme Clough Fault. The area between these faults probably formed a topographically high horst at the time of deposition (see Figure 8.4, and Paper Figure 16), and the restricted marine faunas within the bluestone suggest deposition away from the actively prograding deltas (Stephens *et al.,* 1953). De-watering and soft-sediment deformation structures identified in this study have been interpreted as seismically triggered de-watering structures (see Figure 2.26). Lying antithetic to the Aire Valley Fault, syn-depositional growth faulting associated with the Denholme Clough-Morley Campsall Fault appears to enforce the notion of active tectonic subsidence in this area (Great Mount Quarry, SE009275; see Figures 2.11.4 and 8.4).

The area described above is of particular interest, as it is lies in the pathway of the major basin entry point that crossed the Craven Fault System during the Kinderscoutian and Marsdenian (Section 8.1.1). The positioning of this fluvial pathway may have occurred in response to presence of a conduit, controlled by the presence of a topographic low provide by fault intersections on the Craven Fault system.

9.2.2.2. Compaction driven subsidence

Subsidence due to sediment compaction leads to the generation of localised areas of accommodation in delta top environments. Bayhead environments possess low-gradient profiles implying that compaction allows rapid marine inundation. This process occurs in the present day bayhead deltas, where subsidence is the primary cause of landloss; e.g. as in the Mississippi River Delta area (Coleman, 1981; Roberts & Coleman, 1996; Roberts, 1997). Bayhead deltas from the modern Mississippi River delta are analogues for the bayhead deltas of the Helmshore Grit and the coal-bearing delta top system of the Westphalian (Jones, 1980; Guion & Fielding, 1988; Fulton & Williams, 1988; Guion *et al.,* 1995). While facies variability in the Namurian is strongly controlled by sea-level fluctuations, the influence of the subtler autocyclic controls is more obvious in the Westphalian (Fulton & Williams, 1988). It has been suggested that peat-mire thickness and local syn-depositional topography controlled the thickness of Westphalian coal seams (Fielding, 1987). Similarly, differential compaction between mud- and sand-rich facies of the Elland and Greenmoor cycles (Langsettian) of Bradford, is suggested to form topographic lows leading to the lateral stepping of the Greenmoor Sandstone relative to the Elland Flags (Waters *et al.,* 1996a). Depocentre positioning due to subtle topographic changes is also suggested to have influenced the thickness of deposits in other parts of the Westphalian (Guion *et al.,* 1995; Jones *et al.,* 1995). However, assessment of the controls topography or local compaction had on the occurrence of high-order maximum flooding surfaces is not possible due to the lesser extent of marine bands and their correlative equivalents (Calver, 1968).

It is possible to ascertain that compaction-driven subsidence has a minimal influence on the variation of high-order sequence thickness within the bayhead delta system. This is because the bayhead environment has a series of closely spaced and correlatable depositional water-depth indicators. The amount of accommodation available on the delta top is dependent on the magnitude of eustatic sea-level fluctuation and subsidence in the underlying delta. The widespread extent of individual depth indicators (palaeosols and shoreface facies) suggests that fluctuations in relative sea-level were widespread in extent. It therefore seems likely that eustatic modulation had a greater control on the distribution of bayhead deltas than compaction in the underlying bayhead deltas.

On a broader scale, long-term subsidence due to compaction of whole delta systems, such as the deltas of the Kinderscoutian Stage, could have potentially led to geographically widespread subsidence. The effect of this is subsidence is probably insignificant when compared to the topographic high provided by the Kinderscoutian delta, which has a pronounced influence on the facies association distributions throughout the Marsdenian (see Chapter 8).

9.2.3 Eustatic sea-level modulation.

Ammonoid-bearing marine bands can be correlated across the Namurian of Europe (Ramsbottom *et al.,* 1978). The regional extent of marine bands suggests the mechanism controlling their distribution is independent of regional tectonic subsidence or local compaction. Marine bands are therefore represent low-order maximum flooding surfaces (*sensu.* Van Wagoner *et al.* 1988), which are the products of flooding associated with eustatic sea-level rise (Ramsbottom, 1977). Additional evidence suggesting the significance of eustatic sea-level change comes from the high-order sequence boundaries and maximum flooding surfaces that occur between marine bands in the Pennine Basin. These surfaces are also inferred as the product of fluctuations in eustatic base level.

As early as 1936, waxing and waning the of Gondwanaland ice-sheets was inferred as the mechanism controlling fluctuations in facies and cyclothem distributions in the Upper Carboniferous of the Appalachian Basin (Wanless & Shepard, 1936). Wanless & Shepard inferred that periods of glacial advance and low relative sea-level led to the development of large scale ‘erosive channels’, whereas periods of glacial retreat led to high relative sea-level and the extensive deposition of deeper water mudstones. The problems associated with such correlations become apparent when comparison are made between the estimated extent of Carboniferous ice-sheets and the degree of eustatic fluctuation (Caputo & Crowell, 1985; Veevers & Powell, 1987; Smith & Read, 2000). The geographic distance between polar ice-sheets and palaeoequatorial regions, combined with the difficulty in correlating between continental and marine sequences, make such absolute correlation impossible. The characteristics of transgressive horizons in the Euramerican Province vary along the palaeoequatorial region. While ammonoid-bearing marine bands prevail in the Namurian of the UK, thin marine limestone beds mark the transgressive events in the Appalachian Basin (Wanless & Shepard, 1936; Heckel, 1984; Heckel, 1986).

9.2.3.1. Geometry of the incised valley

The Midgley-Helmshore Grit incised valley is wider (up to 90 km between interfluves) than those in the Kinderscoutian, which are up to several tens of kilometres in width (Hampson, 1995; Hampson *et al.,* 1999). This mirrors a trend suggesting incised valleys in the early Silesian are narrower than those in those in the late Silesian (Hampson *et al.,* 1999).

The initiation of a new basin entry point in the north-east Bowland sub-basin, and the merger of supply pathways may be responsible for the enlargement of the incised valleys created during the subsequent lowstand (see Figures 8.1 to 8.7, Collinson *et al.* 1977). The increased width of the incised valley suggests more accommodation was avaliable within the Marsdenian incised valleys than those formed during the Kinderscoutian. While the narrow width of the Kinderscoutian incised valleys implies their shoreline experienced rapid retrogradation and progradation during the transgressive systems tract, the greater width of the Marsdenian incised valleys implies they underwent a protracted infill. The lateral dispersion of sediment within the incised valley increased the differentiation between the high-order forced-regressive/ lowstand and transgressive/ highstand systems. This explains why evidence for several base-level fluctuations and tidal signatures are observed within the Marsdenian incised valley fill and not in those of the Kinderscoutian. The delta systems of the Marsdenian were deposited in a platform position above the Kinderscout delta top (see Section 6.5), while Kinderscoutian deltas are 'turbidite-fronted', steeper in cross-sectional profile, and less spatially extensive than those of the Marsdenian (Collinson *et al.,* 1977). Marsdenian deltas were therefore formed at, or just below sea-level, and were liable to extensive shoreline retrogradation valley during marine transgressions.

9.2.3.2. Influence of tidal reworking within Marsdenian incised valleys

Unlike other Namurian incised valley fills which contain 'simple' stacked-multistorey fluvial sandstones formed during one regressive-transgressive cycle (Hampson *et al.,* 1997), those in the Marsdenian comprise 'composite' fills (Zaitlin *et al.,* 1994), suggesting multiple transgression-regression events occurred during fill of the incised valley. Comparisons drawn between incised valley fills in UK and US suggest those within the Appalachian Basin contain evidence for a tidal-influence, while those in the Pennine basin do not (Davies *et al.,* 1999). However, this study has identified evidence for tidal currents within incised valley fills containing delta front mouthbars, bayhead mouthbars and fluvial facies. Tidal signatures are predicted in incised valleys, as amplification of tides occurs within the embayed coastline (Dalrymple *et al.,* 1992; Zaitlin *et al.,* 1994). It has also been suggested by previous studies that the Pennine Basin was not influenced by oceanic water-masses, implying basinal processes had a negligible influence on the deposition of Namurian deltas (Collinson, 1988). Tidally influenced deposits occur in plume (Section 3.3.5.), channel (Section 3.3.12.) and rare sub-tidal flats (Section 3.3.8.). The subtle character tidal of indicators within the lower incised valley suggests tidal processes had a limited capacity to completely rework the early incised valley fill. An increased abundance of tidal facies in the upper incised valley fill (e.g. the Helmshore Grit) implies tidal reworking had a greater influence in the late stage of valley fill.

The identification of high-order maximum flooding surfaces and sequence boundaries and tidally-influenced deposits implies the extent that tidal processes influence deposition has to be reassessed for the Marsdenian, and potentially other intervals of the Namurian.

9.2.3.3. The frequency of eustatic sea-level change

Rising relative sea-level forces the transgression of the supply system, decreasing the capacity of supply system to transport clastic material into the basin, producing a landward shift in facies depositional belts (Posamentier & Vail, 1988; Van Wagoner *et al.,* 1988). The opposite is true for falls in relative sea-level, where the basinward migration of the supply system, forces the regression of the depositional system and a basinward shift in facies depositional belts (see Paper Figure 8).

The frequency of the flooding or regressive events may also effect the characteristics of depositional systems. The timing of falling relative sea-level has a significant influence on facies distribution. If several high-order sequence boundaries coalesce to form a low-order sequence boundary, the basinward extent of the regression is likely to be greater. This is demonstrated in the early Marsdenian where the sequence boundary at the base of the R2b1 and R2b3 sequences comprise 2-4 regressive surfaces (Section 6.6.1). When incision is restricted to the platform areas created by the submerged Kinderscoutian delta, high-order sequence boundaries are overlain by thin forced regressive systems tract mouthbars.

Four or five high-order sequence boundaries coalesce to form the low-order sequence boundary at the base of the R2b3 sequence. The incised valley fill overlying the low-order sequence boundary comprises thicker mouthbars and multi-storey fluvial sandstones (see Enclosure 1 and 2; Figure 6.2). The palaeogeographic interpretation implies that the R2b4 sequence incised valley fill prograded further into the basin than those of the preceding sequences. The coalescence of high-order sequence boundaries during sequence boundary genesis is inferred as the product of a combined falling low-order limb, and falling high-order limbs at the onset of HS12 (see Figure 6.3).

9.2.3 4. Comparison between Quaternary and present day relative sea-level modulations

Similar patterns of sea-level modulation exist between the Namurian and the Quaternary (see Section 7.1.3). The magnitude of relative sea-level rise during a fourth-order Quaternary sea-level cycle (130m; Section 7.1.2), is comparable to the sea-level change between the formation of a Marsdenian low-order sequence boundary and overlying low-order maximum flooding surfaces (115m; Section 7.1.3.2). 'Stepped' profiles are present during low order falling relative sea-level of both the Namurian and Quaternary. During the Namurian, periods of rising relative sea-level are stepped (Section 6.6.3), while Quaternary sea-level rise is abrupt and non-punctuated (see Figure 7.1.2). The presence of a late Namurian background rising sea-level is suggested as the cause the stepped sea-level rise (Section 7.2). Comparisons with the estimated Gonwanaland ice-cap extent corroborate this suggestion (Veevers & Powell, 1987).

This evidence reveals a decrease in the amount of ice during the late Namurian, which corresponds with a period of inferred high relative sea-level during the Marsdenian and Yeadonian (see Figure 7.1.1). The R2a1 marine band appears to mark the onset of this interval, suggesting that the increased relative sea-level during the late Namurian of Europe may coincide with waning of the Gonwanaland ice-cap. The deposition of cyclical delta cycles in the Europe and US also correspond to the high eustatic sea-level of the late Namurian (see Figure 7.1.1). A global increase in eustatic sea-level therefore seems the most plausible explanation for the amplification of high-order rising sea-level limbs (see Figure 7.3.2).

The identification of sea-level modulation signatures in the Namurian sections of European and North American basins may allow the qualitative assessment of eustatic-driven accommodation generation. Comparisons of sea-level modulation signature and the estimated amount of accommodation between basins may improve the current correlations that are based on biostratigraphic (Ramsbottom *et al.,* 1978) and chronostratigraphic frameworks (Davies *et al.,* 1999). This method could be integrated with current palaeobotantical and faunal data to provide a framework for the global correlation of Namurian sections. This notion is corroborated by a tantalising insight from Veevers & Powell (1987). They inferred that the relatively low sea-level across Europe during the mid-Namurian corresponds to the Mid-Carboniferous Hiatus in the US. This interval also appears to coincide with the period of maximum Gonwanan ice-cap extent.

9.2.4. Climate in the basin and hinterland

Fluctuations in Namurian river discharge have been related to seasonal variations in river discharge (Collinson, 1968a; Jones, 1980; Hampson, 1997), and evidence suggesting episodic flooding in fluvial overbank areas has also been recognised (Okolo, 1983). The transportation of enormous amounts of sediment must have involved very high discharge rivers, which may have related to seasonally wet periods in the hinterland (Collinson, 1988). Modulations in river discharge had a direct control on facies distributions in the proximal delta. Basinal mudstones and turbidite sandstones are deposited in the relatively deep water of the delta-slope or basin floor. It is difficult to assess if position relative to clastic input points, or seasonal variations in discharge dictate their extent.

Throughout the Carboniferous, climate progressively changed due to the influence of the uplifted Caledonian mountain belt, and the effect it had on global atmospheric circulation (Raymond *et al.,* 1989; Ottobliesner, 1993). Deposition was influenced by both tropical conditions, and the longer-term periodic wetting and drying associated with flooding caused by the waxing and waning high-latitude glaciations. Sediment input to the Westphalian-aged bayhead deltas is interpreted as modulated by monsoon-like conditions (Broadhurst *et al.,* 1980; Broadhurst, 1988), and analysis of palaeobotanical data from the lower Carboniferous suggests the influence of a seasonal climate (Falcon-Lang, 1999a).

Attempts to use palaeosol horizons to unravel the climatic signature within the basin are probably too simplistic (see Section 3.3.10). Palaeosols in the Marsdenian are always either waterlogged, leached or overprinted, and true variegated examples are rare. Additionally, little is known about the climatic signature of the hinterland area, where denudation of the Caledonian mountain belt generated sediment that was transported to the basins. Therefore, the climate in hinterland areas may have had a greater effect on the type or characteristics of sediments than the climate within the basin.

9.3 Summary

Accommodation generation and sedimentation were controlled by a combination of factors during the Marsdenian. These factors dictated the position of depositional environments, and the characteristics of transgressive and regressive surfaces.

The most significant control is eustatic sea-level fluctuation. Eustatic sea-level fluctuation modulates the amount of accommodation generated by compaction, thermal sag, sea floor topography and localised tectonics. The delta system deposited in the Kinderscoutian provided a broad topographic feature that influenced the extent depositional environments during the early Marsdenian. The deposition of Marsdenian delta systems also shifted depocentre positioning, forcing compensational stacking of subsequent low-order sequences.The positioning of basin entry points also had and influence on the geometry and timing of basin fill. Fluvial entry points generated during the Kinderscoutian dictated the pathway of early Marsdenian fluvial systems. The modification and the interaction of these systems with topography created by

Marsdenian deltas influenced the autocyclic switching of depocentres. While the control of climate is difficult to distinguish within the basin, variations in fluvial discharge due to allocyclic climate variations may have influenced the amount of sediment entering the basin. The effect of tidal reworking is limited to incised valley fills. Tidally influenced facies are subtle, suggesting they had little influence on delta deposition.

Chapter 10 Conclusions

This study has integrated data from the field, boreholes, cored-intervals and the current mapping programme of the Geological Survey to allow a new interpretation of the environmental facies, key surfaces/ horizons and palaeogeography of the Marsdenian Pennine Basin. The correlation framework combines the current marine band framework with sequence stratigraphic concepts to increase the accuracy of the correlation. Specifically, the application of sequence stratigraphic concepts has allowed the delineation of systems tracts, incised valley fills, and their internal architecture. For the first time, this has allowed the isolation and identification of a tidal signature within the Namurian of the Pennine Basin. The accuracy of the correlation has delineated areas where tectonic and inherited topography controlled deposition, and allowed estimates of sea-level magnitude and frequency. The interpretation introduces new and interesting deposition elements of Namurian deposition, while the correlation surpasses that of the previously used cyclothem model.

The Marsdenian Stage; stratigraphic position & depositional environments

The Pennine Basin is proposed to have formed by back-arc extension during Dinantian times and infilled by fluvial-deltaic systems during the Namurian. The Marsdenian lies between the turbidite-fronted deltas of the Kinderscoutian and the fluvial plain/ coal-rich Westphalian succession, and shares characteristics of both intervals.

The R2a1 (*Bilinguites gracilis*) marine band marks the base of the Marsdenian Stage. The interval between R2a1 and R2b5 (*Bilinguites metabilinguis*) comprises a series of shoaling and platform edge deltas that contain basin floor, turbidite, delta front mouthbar, fluvial and bayhead depositional systems. The delta system that prograded into the Pennine Basin during the Kinderscoutian formed a palaeo-bathymetric high over which subsequent deltas developed. The formation of new basin entry points and the use of pre-existing entry points allowed the Marsdenian deltas to pass over palaeoslope formed by the Craven Fault system. The Marsdenian delta system therefore prograded in a southerly direction over both the submerged Kinderscoutian delta and the underfilled basin floor.

Within the basin floor depositional system, the ammonoid-bearing marine bands provide a genetic key surface framework. Marine bands allow the broad correlation of the intervening stratal units across the basin. Within these stratal units, palaeosols, transgressive and regressive surfaces are traced through depositional environments. The motif of these surfaces and horizons reflects changing water depths and the depositional processes that occurred during deposition. The strong regressive-transgressive signature associated with these markers, and their wide lateral extent implies that relative sea-level modulation occurred on a basinwide scale.

Sequence Stratigraphic interpretations

Integration of the marine band and key surface/horizon framework with the depositional system model permits the application of sequence stratigraphic techniques to the Marsdenian Stage. Transgressive and regressive surfaces form correlatable boundaries that define high-order sequences. This allows the accurate prediction of facies distributions, and refinment of the correlation framework.

The alternation between shoreline regression and transgression between high-order forced-regressive/ lowstand systems tract and the high-order transgressive-highstand systems tract suggests that the polarity of sea-level fluctuated within each high-order sequence. The lower and upper bounding surfaces of alternate high-order sequences possesses comprise a sequence boundary, and are separated by a high-order maximum flooding surface. Rapid modulations in relative sea-level, along with the formation of low-order sequences over longer time periods suggests that sea-level modulation occurred on a composite curve comprising fluctuating high- and low-order sea-level modulations.

During the low-order forced regressive and lowstand systems tract, several high-order sequence boundaries colaesse to form a regionally correlatable erosive unconformity (a master sequence boundary). The most extensive sequence boundary in the early Marsdenian lies at the base of the Midgley Grit. This sequence boundary correlates to interfluve palaeosols on the western (exposed at Middleton Towers) and eastern margins (in the Colne Rad Mills Borehole) of an incised valley, and lowstand turbidite deposits in southern parts of the Pennine Basin.

The submerged Kinderscout delta formed a platform on which low-order forced regressive systems tract mouthbars formed during the early Marsdenian. Accommodation was limited in these areas, and decreased relative sea-level led to the basinward migration of mouthbar deposits. Low-order forced regressive systems tract deposits were preserved when high-order sea-level rise was superimposed on intervals of low-order sea-level fall. Accommodation was generated within the incised valleys during the high-order rising sea-level, allowing the preservation of forced-regressive system tract mouthbars.

Tectonic or compaction driven subsidence does not seem to account for the magnitude and lateral extent of the transgressive and regressive key horizons/ surfaces. Rates of thermal subsidence occurred at a maximum rate of 0.03mm per year, while compaction driven subsidence cannot account for the broad geographic extent of the transgressive horizons. Eustatic sea-level change is therefore inferred as the mechanism most likely to be responsible for the formation of transgressive horizons.

Characteristics of Marsdenian incised valley fills

Incised valleys formed during the low-order R2b1 to R2b5 sequences are filled with stacked mouthbar, fluvial and bayhead-dominated high-order sequences deposited during the low-order transgressive systems tract. Within these incised valleys, the initial fill deposits are progradationally stacked, and become increasingly retrogradationally stacked towards the later stage of fill. Lower parts of the incised valley fill contain fluvial facies and delta front mouthbars, whereas shallow water bayhead delta deposits dominate the retrogradationally stacked high-order sequences. A regionally extensive marine band (the low-order maximum flooding surface) caps incised valley fill, and defines the base of the highstand systems tract.

Tidal and estuarine signatures are observed throughout the low-order incised valley fill, but are often cryptic due to the suppression of the tidal signature by the dominance of outflow from the delta system. Analysis of laminae cyclicity in plume-deposited mouthbars reveals diurnal and semi-diurnal periodic currents influenced deposition. Mud-couplets, drape bundling, reactivation surfaces and flow reversal indictors suggest tidal currents influenced deposition within fluvial channels.

Comparison of sequence stacking patterns allows the analysis of basin architecture. While the R2b1-R2b3 and R2b3 sequences are retrogradationally to weakly progradationally stacked, the sequence boundary at the base of the R2b4 sequence represents a major drop in base-level and an episode of marked basinward progradation. Combining fluctuating sea-level curves of vaying magnitudes and frequencies allows the construction of a eustatic sea-level curve for the early Marsdenian.

Analogues with present day depositional systems

The facies exposed in the Marsdenian are analogous to those deposited in the Quaternary outflow-dominated Mississippi delta system. Comparison of Carboniferous and Quaternary deltaic environments therefore allows the influence of varying magnitudes and frequencies of sea-level change to be tested.

Marsdenian eustatic sea-level modulation appears to occur at a similar frequency to that during the Quaternary. High-order maximum flooding surfaces occurring at an estimated averaged periodicity of 16 ka, a periodicity similar to the fifth-order eustatic sea-level modulation observed in the Quaternary (19-23 ka). While sharing similar magnitudes and periodicities of eustatic sea-level change, rising intervals of the sea-level curve have an increased magnitude in the Quaternary, and are suppressed in the Namurian. The opposite is also true for the high-order falling limbs of the relative sea-level curve, where the falling parts of the sea-level curve are extended and the rising parts suppressed. Suppression or extension of rising/ falling sea-level change occurs when a compound relative sea-level curve, comprising fluctuations of varying magnitude and frequency, is superimposed on a background rising/ falling sea-level. In the case of the early Marsdenian, this suggests that background sea-level was rising, and the late Namurian was a period of decreasing polar ice cap extent and increasing eustatic sea-level. This is a notion corroborated by current interpretations of global Carboniferous ice-cap extent and eustatic sea-level, which suggests the late Namurian was a period of rising eustatic sea-level.

Appendixes

Appendix 1: Lithostratigraphic figure

Our abbreviated, current understanding of the lithostratigraphy -marine band relations in the Huddersfield sub-basin. Might be useful to keep a copy of this handy little figure out, for quick reference.

Appendix 2: Key for sedimentary logs

As used in all sedimentary logs throughout this thesis.

Enclosures

The two fold-out sheets (Enclosure 1; the East-West Correlation Panel, and Enclosure 2) are present in the back of the thesis. Marine bands R2b1, R2b2, R2b3, R2b4 and R2b5 form datum surfaces, on which the underlying strata are hung. This preserves the geometry of each low-order sequence, permitting a clear interpretation of facies association distribution. The graphic logs on these enclosures are simplified versions of those on the enclosed CD-ROM. Note that both panels are split, with the sequences datumed on R2b1 and R2b3 lie on slightly different basin cross-sections to those datumed on R2b4 and R2b5. Locality reference numbers on the enclosures refer to the following localities;

Enclosure 1: West-East correlation panel

The logged sections are; 1. Middleton Towers, Heysham (SD409588) 2. Leicester Mills Quarry, Anglezarke (SD621161) 3. Roddlesworth Bore, Belmont (SD656123) 4. River Roddlesworth Section, Belmont. (SD659212) 5. Harper Clough Delph & Smalley Delph Quarry (SD716317) 6. Hodge Clough, Ramsbottom (SD787194) 7. Fletcher Bank Quarries & Bore, Ramsbottom (SD805164) 8. Gin Hall Bore, Bury (SD805144) 9. Heywood Bore (SD83090) 10. (upper) Warland Wood Quarry, Todmorden (SD947202) (lower) Bottomley Clough, Todmorden (SD944214) 11. Noah Dale Core, Rishworth (SE019217) 12. Moselden Heights Quarry, Scammonden (SE043164) 13. M62 Road Cut, Scammonden (SE044168) (data from Church 1994) 14. Dean Head Clough, Scammonden (SE026144) 15. Colne Road Mills, Huddersfield (SE145160) 16. Blackburn Tunnel Section (from Collinson et al. 1977) (SD632280 to SD642271) 17. Roddlesworth Bore Belmont (SD656123) 18. Fletcher Bank Bore, Ramsbottom (SD805164) 19. Heywood Bore (SD83090) 20. Bottomley Clough, Todmorden (SD944213) 21. Manshead Moor Tunnel, Cragg Vale (SD993207 to SE004193) 22. Horse Hey Clough, Rishworth (SE003189) 23. Noah

Dale core, Rishworth (SE019217) 24. Scammonden Dam Borings (SE053166) 25. Colne Road Mills, Huddersfield (SE145160).

Enclosure 2: North-South correlation panel

The logged sections are; 1. Cowloughton Clough, Cowling (SD981364) 2. Ponden Clough, Stanbury (SD981364) 3. Holme Chapel Bore (SD880290) 4. Paul Clough, Cliviger (SD912280) 5. Gorpley Clough, Todmorden (SD918237) 6.(upper) Warland Wood Quarry, Todmorden (SD947202) (lower) Bottomley Clough, Todmorden (SD944214) 7. Heywood Bore (SD83090) 8. Culvert Clough, Denshaw (SD983128) 9. Great Clough, Scammonden (SE031147) 10. Wessenden Head Bore (SE062087) 11. Hincliffe Mill Quarry, Holmebridge (SE132072) 12. Woodhouse Quarry, Holmebridge (SE128064) 13. Rake Dike, Holme Moss (SE101053) 14. Woodhead Tunnel Boreholes (SE13601) 15. Loftshaw Clough, Langsett Moor (SK170994) 16. Quickedge Quarry, Oldham (SD9670367) 17. Mouselow Quarry, Glossop (SK02459515) 18. Cowloughton Clough, Cowling (SD965420) 19. Clough Beck, Keighley (SE063433) 20. Branshaw Quarry, Oakworth (SE032401) 21. Parkwood Quarries & Parkwood Brickpit, Keighley (SE065407) 22. Woodhouse Quarry, Haworth (SE062396) 23. Ponden Clough, Stanbury (SE981364) 24. Wickering Crag, Haworth (SE048372) 25. Nan Scar Clough, Oxenhope (SE039336) 26. Rag Clough, Oxenhope (SE015336) 27. Middle Moor Clough, Crimsworth Dean (SE993336) 28. Nook Quarry, Hebden Bridge (SE010275) 29. Fosters Delph Quarry, Mytholmroyd (SE022273) 30. Bare Clough, Luddenden Dean (SE018308) 31. Fulshaw Clough, Luddenden Dean (SE028301) 32. Cat-i-th well Clough, Luddenden Dean (SE042282) 33. Triangle Rail Section, Ripponden (SE045212 34. Noah Dale Core, Rishworth (SE019218) 35. Great Clough, Scammonden (SE031147) 36. Pule Hill Section, Marsden (SE032100) 37. Leyzing Clough, Wessenden Head (SE068082) 38. Wessenden Head Bore (SE062087) 39. Crowden Great Brook, Black Hill (SE063032) 40. Rake Dike, Holme Moss (SE101053) 41. Loftshaw Clough, Langsett Moor (SK170994) 42. Mouselow Quarry, Glossop (SK022962).

Enclosure 3: Maps of borehole, core and field data localities

All localities mentioned in the text are illustrated on this fold-out panel.

Enclosure 4: Figure showing summary of facies, facies associations and key surfaces

Taken from sequence stratigraphic paper and key surface chapter.

Enclosure 5: Marsdenian Pennine Basin dataset CD-ROM

The CD-ROM contains detailed field data collected during the 1998-9 fields seasons. To open the browser select **Open_Me_Locality.html**. and use the forward/ backward buttons and hyperlinks to navigate. All data is in .PDF format;

Graphic logs with written descriptions for all the localities mentioned in this project, along with several that were included in the fieldwork programme but not utilised in the analysis. I have endeavored to keep this data objective i.e. free from facies, key surface or sequence stratigraphic interpretations.

A series of correlation panels produced shortly after the field season. These panels use the marine bands as correlation horizons, and do not include the detailed key surface and facies association data shown on Enclosure 1 and 2.

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