



This work is protected by copyright and other intellectual property rights and duplication or sale of all or part is not permitted, except that material may be duplicated by you for research, private study, criticism/review or educational purposes. Electronic or print copies are for your own personal, non-commercial use and shall not be passed to any other individual. No quotation may be published without proper acknowledgement. For any other use, or to quote extensively from the work, permission must be obtained from the copyright holder/s.

THE SEDIMENTOLOGY OF CARBONIFEROUS FLUVIAL AND DELTAIC SEQUENCES;
THE ROACHES GRIT GROUP OF THE SOUTH-WEST PENNINES
AND THE PENNANT SANDSTONE OF THE RHONDDA VALLEYS

Colin Michael Jones, BA

Thesis for the degree of Doctor of Philosophy
of the University of Keele

1977

Volume I : Text

ABSTRACT

This thesis describes two Carboniferous fluvio-deltaic sequences in which river channel fills are dominant, the Roaches Grit Group (Namurian R2b) in the south-west Pennines and Lower Pennant Measures (Westphalian C) in the Rhondda Valleys. In addition, sedimentation in non-meandering sandy rivers is reviewed and discussed.

Non-meandering rivers can be classified into straight and two braided types. In the former, alternate bars attached to the channel sides give rise to solitary sets of cross-bedding which are eroded during falling stage. In braided rivers, channel fill characteristics vary with discharge regime. These range from regimes with pronounced short lived flood peaks with rapid rising and falling stages, probably producing relatively uniform cross-bedded fills, to those which show limited variation between high and low stage and where high stage sediments are mainly reworked. Multiple channel braided rivers should have channel abandonment sequences showing alternating periods of bedform movement and ponded water.

Palaeocurrents vary with regime and some braided rivers may show a variance similar to that of meandering rivers.

The Roaches Grit Group is an overall coarsening upwards sequence 375m thick at maximum, with Deep Water, Delta Slope, Delta Top and Delta Margin Associations. Within these, 20 lithofacies are described and discussed. The lowest Deep Water Association is dominated by turbidites which generally thicken up the sequence. The Delta Slope shows evidence of syn-sedimentary faulting. Density currents deposited much of the sediment and turbidity currents cut channels into the slope. Large fluvial channel fills dominate the Delta Top Association and large solitary sets of cross-bedding within these were probably produced by alternate bars. Sand-waves occurring in shallower parts of the channel produced cosets of tabular cross-bedding, and smaller forms superimposed during falling stage

produced convex-up erosion surfaces within individual sets. There was little low stage sedimentation and a discharge regime with a pronounced flood peak plus a prominent falling stage period is envisaged.

Palaeocurrents are towards the north-west and the delta prograded into a nw-se trending deep water trough, filling this part of the basin for the first time. The petrography and sedimentology however suggest a northerly sediment source. Tectonic movement along the southern margin of the basin diverted the river and at the Roaches led to channel fills being successively offset towards the north.

The Lower Pennant Measures, 550m thick at maximum, contain Fluvial Channel and Delta or Crevasse-delta Associations. 13 lithofacies are described and discussed. The Delta Association, largely restricted to the lower Lynfi Beds, consist of small, laterally variable coarsening upwards and fining upwards sequences, recording the filling of shallow, bays. In most of the succession this early progradational phase is reworked by the succeeding large braided river. This produced thick, laterally extensive sandstones composed of stacked channel fills with low order vertical facies sequences.

Key sections in the association show channel fills uninterrupted by major erosion. These contain major channel base, conglomerate, main sandstone, alternating beds, siltstone and seat-earth/coal components. The conglomerate records concentration in the thalweg, whilst the cleaner, main sandstone component was deposited in topographically higher parts of the channel. Sequences in this show little order and no upwards fining. They were largely determined by discharge changes with larger bedforms being commonly washed out during falling stage. The overlying alternating beds fine upwards and formed during progressive channel abandonment. Widespread coals record periodic diversions of the coarse sediment supply before a further phase of progradation.

ACKNOWLEDGEMENTS

This research was carried out whilst I held a University Demonstratorship in the Geology Department, Keele University.

Dr. John Collinson has supervised the work and I wish to thank him for advice and encouragement over 3½ years and for allowing me a free hand in the choice and presentation of material.

Many people have discussed various aspects of the work and in particular I wish to thank Dr. Neil Aitkenhead, Mr. John Baines, Mr. Ian Chisholm, Dr. Alan Heward, Dr. Peter McCabe and Mr. Aidrian Middleton, together with members of the Geology Department at Keele whose research interests were less closely connected to mine.

Mr. W.B. Evans of the Institute of Geological Sciences kindly made available unpublished maps and photographs. Mr. Julian Orford at Keele Geography Department wrote the computer programmes used in the analysis of the Pennant Sandstone sequences.

I thank Mr. M.J. Stead and his staff for providing technical facilities particularly Mr. David Kelsall for taking numerous photographs and Mrs. Linda Bradshaw for diagrams.

CONTENTS

Page

Chapter 1.	Introduction	
1.1.	Aims of research	1
1.2.	Layout of thesis	3
1.3.	Facies Analysis - a definition	3
1.4.	Terminology	4
1.5.	Methods of study	4
SECTION I. BEHAVIOUR OF MODERN BRAIDED AND LOW-SINUOSITY RIVERS		
Chapter 2.		
2.1.	Introduction	7
2.2.	Meandering rivers	8
2.3.	Behaviour of non-meandering rivers	11
2.3.1.	Introduction	11
2.3.2.	Bedform types	12
2.3.3.	Effects of varying discharge	13
2.3.4.	Straight or low sinuosity channels	18
2.3.5.	Straight channels braided at low or falling stage	22
2.3.6.	Multiple channel braided rivers	22
2.3.6.1.	Formation of islands	23
2.3.6.2.	Channel behaviour	25
2.3.6.3.	Lateral movement of the river	27
2.3.6.4.	Scour and fill	27
2.3.6.5.	Conclusions	28
2.3.7.	Palaeocurrents	29
2.3.8.	Formation and preservation of vertical accretion sediments: comparisons between meandering and non-meandering rivers	31
2.4.	Conclusions	33

SECTION II. THE SEDIMENTOLOGY OF THE ROACHES GRIT GROUP IN
NORTH STAFFORDSHIRE AND ADJACENT AREAS

Chapter 3.		Page
3.1.	The area	36
3.2.	Structure	36
3.3.	Stratigraphic position	37
3.4.	Previous research on the Upper Namurian of north Staffordshire	38
3.5.	Previous research on the sedimentology of the Namurian of northern England	39
Chapter 4.	Association A. Deep Water Sediments	
4.1.	Lithofacies description and interpretation	43
4.1.1.	Lithofacies 1; Mudstone	43
4.1.2.	Lithofacies 2; Silty Mudstone	43
4.1.3.	Lithofacies 3; Goniatite Faunal Bed	44
4.1.4.	Lithofacies 4; Thin-bedded turbidites	45
4.1.5.	Lithofacies 5; Thick Sandstones	49
4.1.6.	Lithofacies 6; Parallel-sided Sandstones	53
4.2.	Facies relationships within the association	54

Chapter 5.	Association B. Delta Slope	Page
5.1.	Limits of the Association	57
5.2.	Lithofacies description and interpretation	57
5.2.1.	Lithofacies 7; Ripple-laminated Sandstone	57
5.2.2.	Lithofacies 8; Micaceous, Carbonaceous Sandstone	58
5.2.3.	Lithofacies 9; Parallel-laminated Sandstone	59
5.2.4.	Lithofacies 10; Trough Cross-bedded Sandstone	59
5.2.5.	Lithofacies 5; Thick Sandstones	59
5.2.6.	Lithofacies 2; Silty Mudstone	60
5.2.7.	Lithofacies 6; Parallel-side Sandstones	60
5.3.	Discussion	61
5.4.	Origin of Lithofacies 6	64
Chapter 6.	Association C. Upper Slope and Delta Top	
6.1.	Introduction	66
6.2.	Assemblage A; Sub and interchannel deposits	66
6.2.1.	Lithofacies description and interpretation	66
6.2.1.1.	Lithofacies 12; Lenticularly-bedded Sandstone	66
6.2.1.2.	Lithofacies 13; Channel fill structureless sandstone	67
6.2.1.3.	Lithofacies 2; Silty Mudstone	68
6.2.1.4.	Lithofacies 7; Ripple-laminated Sandstone	68
6.2.1.5.	Lithofacies 8; Micaceous, Carbonaceous Sandstone	68
6.2.1.6.	Lithofacies 9; Parallel-laminated Sandstone	68
6.2.1.7.	Lithofacies 10; Trough Cross-bedding	69

	Page	
6.2.2.	Inclined bedding	69
6.2.3.	Discussion	70
6.3.	Assemblage B; Fluvial Channel Sediments	75
6.3.1.	Lithofacies description and interpretation	75
6.3.1.1.	Lithofacies 14; Medium Scale Cross-bedding	75
6.3.1.2.	Lithofacies 15; Large Scale Cross-bedding	87
6.3.1.3.	Lithofacies 16; Faintly-laminated Coarse Sandstone	91
6.3.1.4.	Lithofacies 16a; Undulatory-bedded Sandstone	93
6.3.1.5.	Lithofacies 7; Ripple-laminated Sandstone	94
6.3.1.6.	Lithofacies 10; Trough Cross-bedding	94
6.3.2.	Lithofacies relationships with channels	95
6.3.3.	Interpretation of the channel fill sequences	97
6.3.4.	Discharge regime	103
6.4.	Relationships within the Association	105
6.5.	Morphology of the delta	107
Chapter 7.	Association D. Delta Margin Sediments	
7.1.	Introduction	112
7.2.	Lithofacies description and interpretation	112
7.2.1.	Lithofacies 17; Seat-earth and Coal	112
7.2.2.	Lithofacies 18; Wave-laminated Sandstone	112
7.2.3.	Lithofacies 19; Homogeneous Siltstone	114
7.2.4.	Lithofacies 1 and 2; Mudstone and Silty Mudstone	114
7.2.5.	Lithofacies 3; Goniatite Faunal Bed	114
7.3.	Vertical sequences and discussion	114

Chapter 8.	Trace Fossils	Page
8.1.	Introduction	117
8.2.	Description and Interpretation	117
8.2.1.	Pelecypodichnus	117
8.2.2.	Inclined bivalve resting traces	120
8.2.3.	Bergaueria	120
8.2.4.	Cigar shaped Ridge	121
8.2.5.	Cochlichnus	121
8.2.6.	Discontinuous Sole Trails	122
8.2.7.	Large Pipe	123
8.2.8.	Planolites	123
8.2.9.	Rhizocorallidae	124
8.2.10.	Rounded Sole Trail	124
8.2.11.	Small Simple Pipes	124
8.3.	Vertical distribution of Trace Fossils	125
Chapter 9.	Aerial Variatons, Basin Analysis and Regional Significance	
9.1.	Earlier Namurian Sedimentation	128
9.2.	Late R2b sequence in the southern Pennines	130
9.2.1.	Thicknesses	130
9.2.2.	Local Variations	130
9.2.3.	Source of the sediment	134
9.2.4.	Controls on sedimentation	135
9.3.	Regional significance	136

SECTION III THE SEDIMENTOLOGY OF THE LOWER PENNANT MEASURES IN
THE RHONDDA VALLEYS

Chapter 10.	Page
10.1. Introduction	138
10.1.2. Previous Research	139
10.2. Lithofacies description and interpretation	142
10.2.1. Association A; Fluvial Channel Sediments	143
10.2.1.1. Lithofacies 1; Tabular Cross-bedding	143
10.2.1.2. Lithofacies 2; Trough Cross-bedding	143
10.2.1.3. Lithofacies 3; Cross-bedding with undulating basal erosion surfaces	144
10.2.1.4. Lithofacies 4; Parallel-laminated Sandstone	146
10.2.1.5. Lithofacies 5; Ripple-laminated Sandstone	146
10.2.1.6. Lithofacies 6; Structureless Sandstone	147
10.2.1.7. Lithofacies 7; Conglomerate	148
10.2.1.8. Lithofacies 8; Laminated Siltstone	150
10.2.1.9. Lithofacies 9; Seat-earth and Coal	150
10.2.2. Association B; Delta or Crevasse Delta Sediments	151
10.2.2.1. Lithofacies 10; Ripple laminated Siltstone and fine Sandstone	151
10.2.2.2. Lithofacies 11; Laminated Silty Sandstone	152
10.2.2.3. Lithofacies 12; Unlaminated Siltstone	152
10.2.2.4. Lithofacies 13; Mudstone	153
10.3. Description and interpretation of vertical sequences in Association B.	153
10.4. Description and interpretation of sequences in Association A,	156
10.4.1. Introduction	156
10.4.2. Problems of vertical sequence analysis	157

	Page	
10.4.3.	Statistical Analysis	159
10.4.4.	Key Sequences	162
10.4.5.	Channel Size and Morphology	163
10.4.6.	Palaeocurrents	165
10.4.6.1.	Previous Research	165
10.4.6.2.	Present Study	166
10.4.7.	Interpretation of the Composite Vertical Sequence	167
10.4.7.1.	Meandering River Models	167
10.4.7.2.	Low sinuosity and braided river models	168
10.5.	Evolution of the Pennant Cycles	171
10.6.	Discussion: ancient braided river sequences, comparisons and problems of interpretation	174

Chapter I. Introduction

1.1. Aims of Research

A major feature of the rapid development of sedimentology during the 1960's was the erection of facies models for different environments. One of the earliest, and most successful, was the meandering river model, developed largely by Allen (1963a, 1964a, 1965a, b) and Bernard and Major (1963).

Since that time thinking on fluvial sequences has been dominated by the concept of the 'fining upwards cyclothem'. Many have been described, and compared, sometimes uncritically, to those described by Allen, op.cit., from the British Old Red Sandstone. More recently, a few workers have attempted to produce models for braided, and low sinuosity sandy river sequences. Most suggest that these will contain few flood plain deposits, although Moody-Stuart (1966), has described shallow low sinuosity channels within thick overbank fine sediments. Commonly, channel fills of braided rivers are assumed not to fine upwards, but Thompson (1970) regards regularly fining upward channel fills as braided in origin because of the low palaeocurrent range and absence of overbank fines. Selley (1969) suggests that braided rivers deposit symmetrical fine→coarse→fine cycles in contrast to the fining upwards meandering cycles.

Most of these models seem to have little in common, and at present there is no real consensus of opinion as to what vertical sequences should look like, nor how the deposits of different types of braided rivers may be distinguished.

This thesis describes two low sinuosity fluvial sequences from the British Carboniferous. Both form parts of larger deltaic sequences, which have also been investigated but the greatest parts of the exposure are of the fluvial channel fills.

The first of these, which forms the major part of the research is

the Roaches Grit Group (Namurian R2b) from north Staffordshire. [fig.1] During the 1960's the Namurian deltaic sequences of the Central Pennines were the subject of several sedimentological studies. Allen (1960), Walker (1966 a and b) and Collinson (1966, 1967, 1968, 1969, 1970a) investigated the Namurian R1C succession of north Derbyshire and established a broadly coarsening upwards sequence of turbidites, delta slope, and delta top sediments. Within the delta top, Collinson, (op.cit), described large "Gilbert Type" deltas. In the early 1970's, McCabe (1975) extended this research to the north, and Baines (p.comm) investigated a similar succession of E₁c age in the north of the basin. McCabe disagreed with Collinson's interpretation of the delta top Kinderscout Grit, and demonstrated that the supposed Gilbert type deltas occurred within channels and were therefore really large bedforms.

During the same period, Holdsworth (1963), Trewin (1969), Ashton (1974) and Bolton (1976) investigated the lower and middle Namurian sequence in the North Staffordshire Basin. They showed that this area had a different sedimentological history being dominated by thin, proto-quartzite turbidite sequences with shallow water conditions restricted to the south western basin margin. In the R2b zone a profound change occurs in the north Staffordshire sequence with the first widespread appearance of coarse, shallow water sandstones similar in petrography to the delta top sediments of the Central Pennine Basin.

The research in north Staffordshire reported here is a full basin analysis type study with two major aims. Firstly to compare the sequence with those in the Central Pennines, particularly, the delta top channel facies which when the work began, (August 1973), were the subject of considerable disagreement. Secondly to extend the work of Holdsworth and co-workers on the sedimentation of the north Staffordshire basin.

The second part of the thesis describes the Lower Pennant Measures (Westphalian C) in the Rhondda Valleys' area. [fig.1] Part of this

succession has already been the subject of basin wide studies by Bluck and Kelling (1963) and Kelling (1964, 1968, 1969, 1974), who showed it to be a cyclical, fluvial sequence dominated by rivers flowing from the south east. Kelling considered the rivers to have been mainly braided but also partly meandering. Reading (1971) also suggested that the fluvial sediments were mainly deposited by meandering rivers.

Here shortage of time has not permitted a full basin wide study. The aim has been to look in detail at a small area and to make detailed comparisons with modern fluvial environments information on which was not available at the time of Bluck, Kelling and Reading's work.

1.2. Layout of thesis

The text of the thesis is divided into three sections. Section I reviews the literature on modern, sandy, low sinuosity rivers, erects a classification and discusses the type of sequences which result and how different forms may be distinguished. Section II describes the Roaches Grit Group in north Staffordshire and adjacent areas looking particularly at the delta top fluvial sediments. Section III describes the Lower Pennant Measures in the Rhondda Valley's area of South Wales and compares the two sequences.

1.3. Facies Analysis - a definition

The rock types observed in this thesis are mainly clastic. They have been divided into lithofacies to facilitate description and interpretation. The term lithofacies is defined as a group of rocks with a characteristic association of grain size and sedimentary structures. Lithofacies were erected on the basis of detailed descriptions of rock units in the field. Separate lithofacies schemes have been used for the two sequences studied, for although similar lithofacies types occur, for example medium scale tabular cross bedding, they differ in detail.

Many of the lithofacies from the Roaches succession are similar to those described from the R1 Central Pennine succession by Walker, (1966

a and b), Collinson (1966, 1967, 1968, 1969, 1970) and McCabe, (1975, 1976). No attempt has been made to apply their facies classification because of the dangers of being subjective but where similarities exist they have been noted. [Table I].

The lithofacies from the Roaches succession have been subdivided into four facies associations, no particular lithofacies is necessarily restricted to one association.

The facies scheme used for the Pennant succession differs considerably from that used by Kelling (op.cit.), but again there is an overlap and similarities have been noted.

1.4. Terminology

Grain sizes are given throughout on the Wentworth (1922) scale or on the phi transformation (Krumbein 1934). Most sedimentary structures, rock types, process and environmental terms are used as by Blatt, Middleton and Murray (1972) unless stated otherwise. The term mudstone is preferred to mudrock. The word ripple is used for all bedforms less than 4cm in height and lamination caused by ripples is termed ripple lamination. The words used for all larger bedforms are explained in the text, and in Fig.2. Lamination produced by these large bedforms is termed cross bedding or cross lamination. Trace fossils names follow the nomenclature of Häntzschel, ed., Treatise on Invertebrate Palaeontology, Part W1 1962, unless otherwise stated.

1.5. Methods of study

In north Staffordshire exposure consists of either stream sections or large gritstone escarpments with rare quarries. Stream sections have been measured in detail with tape and abney level and drawn up on varying scales depending on the complexity of the sequence. Many of the gritstone edges have been photographed on a large scale and the photographs worked on in the field. All exposures of any size in the field area have been visited.

In South Wales the exposure consists mainly of road cuttings and quarries with some natural escarpments but few stream sections. The larger outcrops have been measured in vertical sections and drawn up. Photography has facilitated the recording of lateral variations in some of the largest outcrops.

Whenever possible sections have been correlated by feature mapping. Generally the 6 inch to the mile maps of the Institute of Geological Sciences have been used as a basis for interpreting the structure and little re-mapping has been carried out. The stratigraphic correlation of the sequences also follows that in the I.G.S. maps and memoirs.

Numerous directional sedimentary structures have been recorded for palaeocurrent determination. Where the beds were dipping they have been re-orientated to horizontal by the method of Potter and Pettijohn (1963). Vector means and tests of randomness have been calculated following Curray (1965) using a computer programme written by P.J. McCabe and modified by A.D. Middleton (see appendix).

Table I

Comparison of Lithofacies defined in the Roaches Grit Group
with those described from the Kinderscout Grit Group

Lithofacies	Walker (1966a)	Collinson (1969)	McCabe (1975)
1. Mudstone	D+F	1*	1*
2. Silty mudstone	G*	1*	1*
3. Goniatite Faunal Bed			2
4. Thin-bedded Turbidities	A	14	3
5. Thick Sandstones	B+C		4*
6. Parallel-sided Sandstones	K	3*	8
7. Ripple-laminated Sandstone		5	10
8. Micaceous, Carbonaceous Sandstone		?4*	?9
9. Parallel-laminated Sandstone		6	18
10. Trough Cross-bedding		9*	13*
12. Lenticularly-bedded Sandstone		11	12
13. Channel-fill Coarse Sandstone		7*	4*
14. Medium-scale Cross Bedding		9*	13
15. Large-scale Cross Bedding		10	20
16. Faintly-laminated Coarse Sandstone		7*	4*
16a. Undulatory-bedded Sandstone			21
17. Seat-earth and Coal		12	22
18. Wave-laminated Sandstone			
19. Homogenous siltstone			

* Part of

SECTION I. BEHAVIOUR OF MODERN BRAIDED AND LOW-SINUOSITY RIVERS

Chapter 2

2.1. Introduction

Leopold and Wolman (1957), recognise three types of river channel; meandering, braided and straight. It is often difficult to make sharp distinctions between these patterns; channels are seldom straight for more than ten channel widths, (Leopold, Wolman and Miller 1964), and the last category includes those which may be irregular or of low sinuosity. Where the sinuosity is greater than 1.5, channels are regarded as meandering (Leopold et.al. 1964).

Meandering rivers have a single, sinuous channel, often approaching the shape of a sine curve (Leopold and Langbein, 1966). In braided rivers the flow is split into a series of separate channels at some stage in the flood cycle, and these constantly split and rejoin. In straight, or low sinuosity rivers, the banks are essentially straight, but at high stage the talweg meanders from bank to bank. If, during low stage, the flow is split into separate channels it is more appropriate to term these braided.

The behaviour of meandering rivers is fairly well understood thanks to the work of Mackin (1937), Fisk (1944, 1947), Sundborg (1956) and Wolman and Leopold (1957). A model for point bar sedimentation in meandering sequences has been proposed by Allen (1963, 1965, 1970a) and Bernard and Major (1963). Recently, variability within modern meandering rivers has been recognised by McGowan and Garner (1970) and Jackson (1975 1976b). Straight and braided rivers, are by comparison even more variable in behaviour and until the last decade there have been few descriptions of modern examples which are of much use to geologists attempting to understand ancient sequences.

There is a threefold sequence of problems associated with ancient braided river deposits. Firstly, to be able to reliably distinguish between

ancient meandering and non-meandering fluvial sequences. Secondly, to distinguish between straight or low sinuosity and braided river channels. Thirdly, to differentiate between different types of braided river channels and in particular to assess the sedimentary response to different discharge curves.

In order to approach these problems the behaviour of modern meandering rivers, and some of the models proposed for the generation of sequences by them, will be briefly discussed.

2.2. Meandering Rivers

In meandering rivers the channel is asymmetrical in section across the meander bend. Erosion cuts a steep bank on the outside of the meander and sediment is transported to and deposited on the gently inclined inner bank or point bar, (Matthes 1941, Friedkin 1945). Because of the stress distribution across the channel, coarser sediment is usually concentrated at the bottom of the point bar and sediment becomes finer upwards, (Sundborg 1956, Bersier 1953).

In the 'classical' meandering river models of Allen (1963a, 1964, 1965a and b, 1970a) and Bernard and Major (1963), the lateral migration of the meander by cut bank recession and point bar accretion produces a sheet sandstone with an erosive base in which sediment becomes finer upwards. Cross-bedding sets should become smaller upwards, because smaller bed forms occur in the shallower water towards the top of the point bar. In small channels, where point bars are steeper, successive increments of lateral accretion may be recorded as Epsilon cross-bedding (Allen 1963c). These have been recognised in ancient sequences by Allen (1965c), Moody-Stuart (1966), Beutner et.al. (1967), Cotter (1971), Puidefabrigas (1973) and Elliott (1976).

Schumm (1960, a, b, 1963, 1968) has shown that many meandering rivers have a high proportion of suspended load to bed load. In addition, frequent meander cut offs should lead to a stabilisation of the meander

belt in a particular part of the flood plain, preventing the river from migrating laterally. Thus ancient sequences should contain a thick development of fine sediment deposited in flood plains during overbank flooding as in the models of Allen (1965a, b, 1974) and Schumm (1968).

Because of their high sinuosity, ancient meandering river sequences should show a palaeocurrent pattern with a high variance (Allen 1965a). There has yet to be any systematic study of this in modern rivers however.

The principal components of this 'classical' meandering river model are listed in Table II.

Some of the basic assumptions of this model are now open to question. In particular, the sinuosity of meandering rivers varies greatly, from 1.5 to about 4 (Leopold, Wolman and Miller 1964). It now seems doubtful whether one unifying model will cover the complete range of sinuosities possible.

For example, with high sinuosity meandering rivers meander cut offs should be particularly frequent. This should lead to a sandstone composed of lots of mutually erosive, curved channel fills, as shown in aerial photographs of ancient meandering sequences by Puidefabrigas (1973) and Nami (1976). The sheet sandstones described by Allen, (1964, 1965b, 1970a) from the British Old Red Sandstone are unusual in their scarcity of abandoned channels and of lateral accretion surfaces.

It is in the high sinuosity meandering rivers that preservation of flood plain deposits is most likely because of the stabilisation of the meander belt. With more moderate sinuosities, meander cut off will be less frequent allowing easier lateral shifting of the river channel. In addition, preservation of overbank fines depends greatly on external factors such as tectonic and climatic changes as demonstrated by Allen (1974). Where the rate of subsidence is low, a meandering river may be able to erode away most of its overbank sediment.

Not all meandering rivers produce fining upwards channel fill

sequences. McGowan and Garner (1970) have described bed load meandering streams of comparatively low sinuosity, and with a fairly irregular discharge. At high stage, the thread of maximum velocity cuts across the point bar as the course tends to straighten, cutting 'chute channels' and depositing 'chute bars' at the ends. Trough cross-bedding formed in the deeper part of the channel is overlain by larger sets of tabular cross bedding representing the chute bars. The overall vertical sequence does not fine upwards. Supposed ancient examples occur in Texas Tertiary deposits.

Recently Jackson (1975a, 1976b) has shown that in tightly curved meandering reaches of the Wabash River the channel fill also does not fine upwards. Thus fining upwards fills appear to be restricted to meandering rivers of 'intermediate sinuosity'. In this type of river Jackson (1975a, 1976b) has demonstrated the development of a 'transitional' zone developed in channel sections between successive meanders, where the talweg crosses from bank to bank. In this zone, a vertical section through the sediment deposited showed no fining upwards, nor an upwards decrease in the size of sedimentary structures. Preservation of different zones in the river depends on the way in which the meander migrates laterally. With some types of meander migration observed in the Wabash River largely transitional zone sequences are preserved.

Finally, the palaeocurrent variance in a river with a sinuosity of 4 would be expected to be much higher than in a river with a sinuosity of only 1.5.

This review suggests that the classical meandering river model of Allen and Bernard and Major cannot be applied to many meandering rivers. Other types of meandering river will deposit channel fill sequences which do not fine upwards and some may show a poor preservation of fine grained overbank sediments. At present, it seems doubtful whether their deposits could be distinguished from those of braided and low sinuosity rivers in ancient sequences.

2.3. Behaviour of non-meandering rivers

2.3.1. Introduction

Channel fill sequences in non-meandering rivers depend on channel morphology and the discharge regime. Some rivers have only one channel at high stage, whereas others may have several. Two types of channel occur; [Fig.3a, b].

(i) Straight channels, where the thalweg meanders from bank to bank between side bars or alternate bars. At low stage these bedforms become emergent at the sides of the channel and are eroded by the meandering thalweg which now occupies only the centre of the channel.

(ii) Straight channels braided at low stage, where at high stage the thalweg meanders between side bars as in type (i). These are covered by smaller bedforms usually lingoid bars, and at low stage the bars are emergent and split the flow into many smaller components. This process of braiding was first described from the River Platte by Ore (1964).

In many cases rivers have only one main channel at high stage so that channel type and river type are synonymous. In some large rivers such as the Niger, Brahmaputra and Yellow River, the flow is split by islands into more than one channel even at high stage. [Fig.3c]. Within these multiple-channel braided rivers individual channels may be of either of the two types described above, but the channels do not behave completely independently. This type of braiding is quite different from the process described by Ore.

In this section, after a description of the types of bedforms present in non-meandering rivers and the effects of different discharge regimes, the behaviour of the two channel types described above will be discussed. This will be followed by a discussion of the more complicated multiple channel rivers.

2.3.2. Bedform types

A wide variety of bedform types occurs in non-meandering rivers. Some forms have only recently been described, and it seems likely that others still await recognition. It is not the intention here to discuss bedforms in detail, but these will be referred to in the text and a brief description is necessary.

Bedforms in modern rivers exhibit a wide variety of sizes. Jackson (1975c), has recently proposed a hierarchical classification and his scheme will be followed here.

- (i) Microforms - These are the smallest category present and are represented by current ripples, forming at relatively low flow powers in the lower part of the lower flow regime of Simons and Richardson (1961).
- (ii) Mesoforms - Most of the bedload in rivers is transported by the movement of intermediate sized bedforms, large scale ripples, forming in the upper part of the lower flow regime. Two varieties of mesoform are common but others are also present. Dunes are sinuous crested bedforms with a triangular profile and a height to chord length ratio of the order of 1:10. Deep leeside scour troughs are characteristic. The resulting deposit is a coset of trough cross bedding, (Harms et.al. 1975). Sandwaves also called lingoid bars (Allen 1968, Collinson 1970b) and transverse bars (Ore 1964, Smith 1970, 1971) may be straight crested or horseshoe shaped in plan, and have steep avalanche faces. The height to chord length ratio is much lower than in dunes. Leeside scour troughs are usually absent and in these cases the resulting deposit is a coset, or isolated set of tabular cross bedding, possibly with interbedded cross laminated units deposited from superimposed ripples. Sandwaves form at lower flow powers than dunes, (Harms et.al. 1975). Longitudinal bars. These normally occur in gravel bed braided streams and are composed of very coarse grade material, (Ore 1964, Smith 1970). Sandy longitudinal bars have been described from fluvial regimes by Williams (1971), Boothroyd and Ashley (1975) and

Jackson (1975a). Little is known about the formation of these bedforms at present but their lozenge shape is similar to the larger braid bars.

Probably they develop in response to excess bed load in the channel..

(iii) Macroforms - These large bedforms occur at a variety of flow powers in response to the large scale flow circulation in the channel. The three common types are side bars, alternate bars and islands or large braid bars [Fig.2]. These will be discussed later. They usually carry superimposed meso- and macroforms.

(iv) Upper Flow Regime Bedforms - Because of the depths frequently attained in large, sandy, low-sinuosity rivers, and the consequent lowering of the Froude Number, upper flow regime bedforms occur rarely. In the Brahmaputra, (Coleman 1969), however, current velocities during flood are so great that an upper phase plane bed is developed in depths up to at least 15m. Turbulent cells arranged in rows parallel to the current are often associated with upper flow regime conditions in straight sections of rivers, forming longitudinal ridges 0.75m-1.5m in height and 8-30m apart.

2.3.3. Effects of varying discharge

The formation of mesoforms and microforms in a river is closely dependent on flow strength. This may be expressed in a variety of ways; simply as mean flow velocity, (\bar{u}), bed shear stress, (T_0), or stream power, ($w = T_0\bar{u}$). Where there are large variations in water depth, these lead to changes in the Froude Number

$$Fr = \frac{\bar{u}}{\sqrt{gd}}$$

where d = depth, g = acceleration due to gravity

so that upper flow regime bedforms may occur in shallow water at quite moderate flow strengths.

Flume experiments show that a particular type of bedform is only in equilibrium within particular ranges of flow parameters (Simons and Richardson 1961, Southard 1971). In a river of variable discharge, there are large variations in flow characteristics with the result that bedforms

are often not in equilibrium with the flow during the flood cycle. (Allen 1973b, Allen and Collinson 1974).

In assessing the effects of variable discharge two parameters are important. Firstly, the ratio of maximum to minimum discharge

$$\frac{Q_{\max}}{Q_{\min}}$$

secondly, the characteristic rate of rise and fall between peak discharge and low water. Following Allen and Collinson (1974) this may be expressed as

$$\left(\frac{\Delta Q}{\Delta T}\right) - Q_{\max}^{-1}$$

where ΔQ is the rate of change of discharge in time increment ΔT .

River hydrographs are closely dependent on climate (Beckinsale 1969). In Megathermal Regimes developed under tropical climates three types are common. In some equatorial areas the heaviest rains occur in spring and autumn following the equinoxes, and there is no dry season. These areas exhibit a double maximum, for example the River Congo. In areas watered by summer monsoon rains there is just one flood peak, the length of which depends on the length and intensity of the rainy season. The Brahmaputra River in Bangladesh (Coleman 1969) has a five month flood peak [Fig.4]. With increasing aridity the flood peak becomes shorter until eventually permanent watercourses cannot be maintained and favoured watercourses have an episodic flood once or twice a year. Williams (1971) describes an example from the Lake Eyre Basin, Australia.

Mesothermal Regimes are developed under sub-tropical and temperate climates (Beckinsale 1969). Many of these have permanent flow with only slight seasonal variations and often the rivers are meandering, such as the Mississippi. The River Platte (Smith 1974), is a braided river with a fairly regular hydrograph [Fig.4].

Microthermal Regimes develop in colder areas which experience winter snowfall. These usually have a very short, sharp flood peak in May or June following the snow melt. The hydrograph rises and falls very rapidly and

during the winter the rivers are often iced over. The Tana River in northern Norway (Collinson 1970b), [Fig.4] is the best documented example, where the flood lasts for just one month.

The varying responses possible with different discharge curves have an important effect on channel sequences. Indeed it seems likely that many sequences in rivers with variable discharges will be flood dependent. A variety of bedform responses is possible.

Bedform response

When there is a change in discharge, bedforms have a minimum relaxation time in which they may respond to the change in flow conditions (Chorley and Kenneday 1971, Allen and Friend 1976a). Most natural populations of bedforms are not stable, even under steady stage conditions, however, (Allen 1973a). The rate at which new forms are constantly being created and destroyed under steady state conditions has been termed the creation time (Allen 1976). Generally speaking, larger bedforms have high relaxation and creation times.

The types of sequences of sedimentary structures which may be produced, for example, during a fall in discharge, depend therefore on a relationship between the rate of fall of discharge, the natural lifespan of the bedform at constant discharge (or the creation time), and finally on the relaxation time for a given discharge change. For convenience this may be expressed as a ratio between bedform size and rate of change of discharge. Where the bedform is dunes, four different responses may be possible [Fig.5].

1. Where the period of change in flow strength is very low in relation to the creation time the dune population will not be immediately affected. However as new forms are created they will be in equilibrium with the new, reduced flow strength and will be smaller. The older large dunes will eventually be replaced. Under conditions of net sedimentation a thinning upwards coset of cross bedding will form [Fig.5].

2. Where the rate of fall of discharge is fast in relation to the creation time, older, larger forms will lag behind the change in discharge, (Allen 1976). A mixed population, composed of older larger dunes and smaller younger dunes will be created. In the Brahmaputra, (Coleman 1969) there is rapid sedimentation during falling stage so that dune shapes are preserved, but the frontal scour pools disappear. This produces large scale ripple drift with broad shallow troughs in sections normal to the dip.

3. Where the period of change in flow strength is shorter but still longer than the reaction time, the dune field will disappear and new bedforms will take their place. Commonly, where a reduction in flow strength is accompanied by a shallowing of the flow, the new bedforms will be ripples. Occasionally, extreme shallowing may cause an increase in Froude Number so that a plane bed develops. As a result, trough cross bedding deposited when the regime allowed dunes to form will be succeeded erosively by either ripple lamination or upper phase parallel lamination. This will probably only occur where the dunes are small (small relaxation time), as during shallowing the reduced discharge and flow strength limit sediment transport. Allen and Friend (1976b) describe small dunes in a tidal regime which have been substantially modified by ripples.

4. If the period of change of flow strength is shorter than the relaxation time, the dune shape will be substantially unaltered. This is common in tidal channels with their rapidly varying currents, (Allen and Friend 1976a, b), but also occurs in some rivers (Coleman 1969, Williams 1971, Singh and Kumar 1974). If there is a large amount of suspended sediment in the river at low stage, then the dune may be buried by the finer sediment and its profile potentially preserved in the geological record.

These observations also apply to sand waves, but here the range of sizes is greater, and sand wave height often responds to changes in flow depth (Harms et.al. 1975). In large braided rivers such as the Brahmaputra, Q_{max}/Q_{min} is great (Coleman 1969) and water depths almost double during floods.

The discharge increases fairly slowly and high discharges persist for over five months. At flood peak, very large sandwaves up to 17m in height appear in some parts of the river, though a wide variety of other sizes of mesoform are also present. As yet, the detailed pattern of growth of these bedforms has not been documented, and little is known about the internal sedimentary structures which result. Possibly they may resemble, on a much larger scale, the climbing ripple drift described by Stanley (1974 Fig.12-B) where a coset of small crossbedding passes downcurrent into a single larger set, with the erosion surfaces at the base of the smaller set left hanging.

During falling stage the variety of responses possible is the same as with dunes but the actual sedimentary response is slightly different [Fig.6]. A very slow fall in discharge may produce a coset of thinning upwards cross bedding as with dunes. This is probably the case in the Weser and Parana Rivers (Stükrath 1969, Wasner 1974) where superimposition of bedforms does not occur. With a slightly more rapid fall a mixed population is developed and the smaller bedforms adjusted to the reduced discharge appear on the backs of the larger forms no longer in equilibrium with the flow. This is well seen in a sequence of profiles from the Fraser River (Pretious and Blench 1951), [Fig.7]. Migration of the smaller superimposed forms down the front of the larger bedforms produces multiple convex upwards erosion surfaces (McCabe and Jones in press), [Fig.8].

If the flood falls very quickly, faster than the relaxation time, the sandwaves will be left standing and often partially emergent, as in the Tana River (Collinson 1970b). Flow around the front of the bar erodes away part of the slip face, sometimes depositing a coset of small scale trough cross bedding orientated normally to the main foresets. During the next high stage the bedform is reactivated, burying the falling stage features.

With the different types of hydrograph discussed earlier it seems clear that these varieties of responses may occur. This may be illustrated by comparing the river Tana (Collinson 1970b) with the River Platte (Smith 1970,

1971, 1972, 1974). Both are relatively straight rivers at high stage, when the channels are covered by lingoid sand waves. As seen in Fig.4, however, the hydrographs vary considerably. In the Tana, most of the sediment movement takes place during a very short flood peak. The discharge decreases rapidly and at low stage it is very small, the river being also iced up for 6 - 7 months. Most of the sedimentary structures are formed at high stage with limited falling stage modifications. Cosets of tabular cross bedding produced by the migration of the sandwaves are the dominant structure.

In the Platte, however, the discharge is much less irregular throughout the year [Fig.4]. Few of the bedforms visible at high stage survive falling and low stage modification. The result is a much greater variety of structures (Smith 1974). Dunes coalesce to form sandwaves and the characteristic tabular cross stratification is only locally developed, forming at the front of cosets of trough cross bedding produced during the earlier dune phase. During low stage channels often cut completely through the sandwaves which are substantially modified. This process was the basis for Ore's original criteria for braiding in sandy rivers in 1964.

Williams (1971) describes another interesting example from a sandbed ephemeral stream in Australia. Trough cross bedding produced by migrating dunes active at high stage is overlain by tabular cross bedding produced by sandwaves moving during falling stage.

2.3.4. Straight or low sinuosity channels

These are the least understood of all three channel types. They are not common, although straight segments occur within longer braided or meandering reaches as in parts of the Benue, (Nedeco 1959). In addition, delta distributaries are commonly straight. Much of our knowledge of these forms comes from man-made channels with rigid sides, (Harms and Fahnstock 1965, Maddock 1969, Culbertson and Scott 1970); and from flume studies

(Friedkin 1945, Vanoni 1946, -Wolman and Brush 1961, Stebbing 1963, Shepard and Komura 1968, Ackers and Charlton 1971, Schumm and Khan 1972).

Ackers and Charlton (1971) and Schumm and Khan (1972), claim from flume experiment studies that straight channels have lower stages than either meandering or braided types. Leopold et.al. (1964), however, showed that naturally straight channels have a wide range of slopes, many as steep or steeper than those of braided channels. Because of the meandering thalweg, straight channels often have a tendency to become meandering with time, (Vanoni 1946, Schumm and Khan 1972). This may be checked where the bank material is made of cohesive clay as with the delta distributary channels of the birds foot Mississippi delta (Shepard 1960).

Two types of flow patterns are known from straight channels. Several flume workers have observed rows of circulation cells parallel to the channel axis. Vanoni (1946), noted that these formed longitudinal ribbons of sediment along the flume bed. It is still unclear, however, why these circulation cells form, nor how generally important they are, (Shen 1971). The parallel ridges described by Coleman (1969) from the Brahmaputra, may have formed because of this type of circulation.

Much better documented is the tendency for the flow to meander between the straight banks. This is more pronounced in channels of large width depth ratio, and so may be difficult to observe in flumes, but is common in natural channels. The meandering leads to alternating erosion and deposition on the channel sides, forming either side bars¹ or alternate bars². These then move downstream and are only preserved after channel abandonment.

Side bars are semi-elliptical in plan with gentle upstream and downstream faces. In straight reaches of the Benue (Nedeco 1959) and Rio Grande (Harms and Fahnstock 1965), they occupy over three-quarters of the channel width. They invariably have smaller bedforms superimposed on their backs, in the Rio Grande these are largely dunes, (Harms and Fahnstock 1965).

1. These are often termed alternate bars.

2. This term is preferred to transverse bar as used by Allen 1968.

Sections cut through side bars in this river show predominantly cosets of trough cross bedding. The tops of the bars have a thin veneer of ripple lamination deposited during the low stage period preceding the excavation, (Harms and Fahnstock 1965).

Alternate bars are triangular in shape with a gentle cross-side but with a steep avalanche face downstream. In the Rio Grande, the side bars occasionally develop avalanche faces at their downstream ends where they are building into slack water (Harms et.al. 1965). In aerial photographs of alternate bars in artificial channels (Maddock 1969) the crest lines have a pronounced horseshoe shape similar to those in longitudinal sand waves.

Up to now none of the alternate bars described from active channels have been much more than 1m high. In straight channels within some braided reaches of the Niger, however, bedforms of a similar type reach up to 3m (Nedeco 1959), although they tend to occur as isolated forms rather than in an alternating series. Baker (1973), has described much larger structures formed at very high discharges during glacial flooding on the Columbia Plateau of eastern Washington. These, which he calls Pendant Bars, are streamlined mounds of gravel which occur just downstream from bedrock projections in the former channel floor. Although formed in a different way from alternate bars of straight river channels, they are similar in shape and reach up to 30m in height and 3km in length. Smaller straight-crested bars occur on their backs. Sections through these bedforms show very large solitary sets of cross bedding. Straight channels with large alternate bars should be dominated by large single sets of cross-bedding.

Both side and alternate bars are constructed during high stage flow and move very little during low stage, (Harms and Fahnstock 1965, Maddock 1969). Because there is much less water in the channel at low stage, the meandering thalweg may have a small wavelength and the bars will be eroded [Fig.9], (Harms and Fahnstock 1965, Popov 1965, Maddock 1969). The erosion should be particularly noticeable where there are large alternate bars as

large sections of these may be removed. Reactivation during the next high stage then buries the erosion surface [Fig.10].

The depth of the thalweg in straight rivers is least at the point of cross-over and greatest at the point of scour on alternate sides of the channel. During low stage in straight reaches of the Niger, the water ponds in the deeper part, whilst erosion occurs in the shallow cross-overs (Nedeco 1959). As a result a 'sand trumpet' or channel delta is constructed out into the deep ponds [Fig.10]. It seems unlikely that these structures will have a particularly high preservation potential as during the succeeding high stage there will again be erosion in the pond areas, (Nedeco 1959).

Straight channels with side bars will be difficult to distinguish from some types of braided river channels. The nature of the channel fill will depend largely on the type of smaller bedforms (mesoforms) present and on the degree of low stage reworking. In the Rio Grande, where dunes are superimposed on the side bars, sections through these show cosets of trough cross bedding mantled by ripple lamination. The edges of the bars are cut by shallow channels filled with tabular cross bedded and ripple laminated sands (Harms and Fahnstock 1965). Abandonment of the channel may preserve this sequence. If the bars are covered by larger lingoid sandwaves cosets of tabular cross bedding will result.

Straight channels are unlikely to migrate laterally. Flume experiments show that bankside erosion leads to the development of a meandering pattern, (Schumm and Khan 1972). Straight channels in the birdsfoot Mississippi delta have not shifted laterally for at least the last 100 years (Scruton 1960). Instead the entire discharge tends to shift to other areas, and develop new channels (Coleman 1969). Preservation of the channel fill will therefore only take place after abandonment.

2.3.5. Straight channels braided at low and falling stage

In this type of channel islands are rare, and at high stage the thalweg meanders from bank to bank forming side bars. Examples are the Tana River (Collinson 1970) Platte River (Ore 1964, Smith 1970, 1971, 1972, 1974), the Loup River (Brice 1964) and Red River (Neill 1969).

Usually the channel areas are covered at high stage by horseshoe shaped sandwaves. During falling and low stage these bedforms emerge, splitting the flow into many smaller components. At these times the river may be said to be braided. It is difficult to make a clear distinction between this type of river channel and the straight channels with side bars previously described. However, with the latter type, the Rio Grande (Harms and Fahnstock 1965) there appears to be just one sinuous thalweg occupying the centre part of the river with the side bars emergent at the edges [Fig.3].

The channel fill sequence will probably depend largely on the nature of the hydrograph as previously described. In the River Tana, where most sediment movement takes place at high stage during a short flood peak, extensive sets of tabular cross bedding produced by migrating sandwaves will predominate. In the River Platte, however, extensive falling and low stage reworking produces a much more variable sequence, and many of the sandwaves are cut through by channels (Ore 1964, Smith 1974). In the Red River (Neill 1969) where Q_{max}/Q_{min} is large, scour and fill during floods is important.

This type of river may possibly be distinguished from the multiple channel braided type by the absence of channel abandonment sequences. Indeed it is unlikely whether the channel fill will fine upwards very much as there appears to be no mechanism which could cause a pronounced grain size segregation.

2.3.6. Multiple channel braided rivers

In this type of river the flow is split up into smaller channels by islands and submerged bars. These are semi-permanent and may often be recognised through several flood cycles even though their shape and size

may change considerably. All of the channels will be active at high stage but not necessarily at low stage. Modern examples are the Niger (Nedeco 1959) Yellow River (Chien 1961) and Brahmaputra (Coleman 1969). This type of river usually has a highly variable discharge.

2.3.6.1. Formation of Islands

Islands and bars are typically lozenge shaped with their longest part parallel to the flow. They occur not only in sandy rivers but also in gravel bed braided streams where they have sometimes been called longitudinal bars (Ore 1964) or spool bars (Krigstrom 1962). An acceptable general term is braid bar (Allen 1968). The larger islands are macro forms following the terminology of Jackson (1975c) and form not at a particular flow regime, but as a result of the inability of the river to transport all its sediment through a given reach. This will particularly tend to happen with a sandy river of variable discharge carrying a large bed load.

In the Niger islands often join together, becoming very elongate, some reaching 5 kilometers in length (Nedeco 1959). During floods scour along their sides may break them up. Some islands have retained essentially the same position for at least the last 100 years. In the Brahmaputra islands, called Chars, are very numerous and very variable in size; larger ones exceeding 10km. Smaller islands may be formed or destroyed during one flood cycle (Coleman 1969).

The development of a braided channel has been investigated in flume experiments by Leopold and Wolman (1957). Initial deposition of coarser material splits the channel into two and causes erosion of the adjacent banks adding to the sediment load and enlarging the cross section. The water surface is lowered and eventually the initial bar emerges as an island. Further bars form adjacent to the first which continues to enlarge. Eventually the flow is split into a number of separate channels.

Where the sediment load is well sorted as in the Niger, (Nedeco 1959), and Brahmaputra (Coleman 1969), island formation takes place in a different

way. When the river is in flood it will tend to widen its cross section. This usually results in a deeper channel along each edge. During falling stage the river will decrease its now too wide cross section by deposition in the channel. This accretion, combined with a fall in water level may cause the mid river sand bank to become exposed. The intensified turbulence along the original river banks will tend to maintain the two channels. During the next high stage the river will enlarge its cross section again, depositing some of the sediment eroded from the channel edge on to the newly born island.

In the Niger (Nedeco 1959) islands sometimes form as a result of differences in the erodibility of the river banks. If resistant bank material projects into the river the channel will scour on its upstream side, and deposit on the downstream side forming a sand bank. This bank will eventually grow out into and down the channel. As the meandering thalweg moves down the channel it will eventually be able to attack the sand bank immediately downstream of the resistant projection eventually severing it from the bank and leaving an island in the middle of the channel.

Once formed the islands in the Brahmaputra (Coleman 1969) undergo scour on their upstream sides and are added to on their downstream sides, so moving down the river. Usually islands have smaller bedforms, small and large scale ripples on their backs. In the Niger emergent islands are quickly colonised by vegetation, (Nedeco 1959), whilst in the Brahmaputra aeolian dunes form. Internally therefore islands will consist of cosets of various types of cross stratification. In the Volga (Shantzer 1951), islands grow by a process of lateral accretion so that they resemble double point bars. Conceivably lateral accretion surfaces and fining upwards sequences may be preserved.

As yet, there have been no detailed descriptions of the internal sedimentary structures of islands and they have not been recognised as discrete bedforms in ancient sequences.

2.3.6.2. Channel Behaviour

Islands split the flow into separate channels. Usually the discharge is not divided equally between the channels, and often a channel carries a large proportion of the flow. Within individual channels the thalweg meanders from bank to bank between side bars or alternate bars as in rivers where the flow is not split. Within multiple channel rivers, however, the channels do not behave independently.

In the Niger, one channel is commonly partly blocked off by the growth of a side bar which plugs the entrance. The water level immediately downstream of the side bar is low because of the reduced discharge, but in the case of a short island, is raised by the backwater effect from the downstream ^{stream} end of the island. Thus the discharge through the partly blocked channel is low. With a long island, however, the backwater does not affect the water level so far upstream, resulting in a drop in water level downstream from the bar and little reduction in discharge. During low-water, however, the channel is effectively cut off by the side bar and may be abandoned completely.

Downstream sediment transport causes side bars to slowly move down the channel with an associated movement of the thalweg. Channel configurations alter constantly as channels are blocked off and eventually abandoned. In the Yellow River (Chien 1961), the main channel may be completely filled and abandoned during one flood cycle. In the Niger major changes take place more slowly, and none were observed during the study period, but Nedeco (1959), suggests how channel abandonment may take place over a period of 40 to 70 years.

On the basis of the Nedeco (1959) observations on the Niger the following model for channel development is proposed [Fig.12]. The flow is split up into two deeper components [Fig.12-A]. As a result the channel widens and an island forms in the middle [Fig.12-B]. Side bar X moves down the river partially blocking the larger channel A [Fig.12-C].

The thalweg is diverted into Channel B which now becomes the new main channel. At high stage channel A is still active as the fall in water level immediately downstream of the side bar maintains the discharge. During low stage, the side bar is emergent and flow is restricted to the deeper main channel. Water ponds in channel A slowly draining. With further growth of the side bar channel A is active for a progressively shorter period during each flood. Eventually it silts up and is abandoned, tying the island into the flood plain [Fig.12 - D]. It is quite likely that the downriver movement of the next side bar, Y, will again deflect the thalweg so that channel A will be re-opened and channel B will now start to be abandoned. Only with a permanent sideways shift in the position of the main river will abandoned channels be preserved.

This process, which also takes place in the Yellow River (Chien 1961), produces a fining upwards channel abandonment sequence, which should often show alternating phases of bedform movement and stagnation. Of course, if the channel is rapidly cut off during one flood cycle the inactive channel will slowly silt up with fine grained sediment as happens with ox-bow cut offs in meandering regimes. In the Yellow River, (Chien 1961), rapid cut off may occur because the main channel is filled with sediment. This causes the river stage to rise and induces the river to spill more water into another channel which has a lower elevation and more favourable alignment with respect to the upper river course. After the passage of the main flood, the flow completely shifts to the new course.

Prior to the abandonment, the channel will behave in a similar way to those in braided reaches which are not permanently divided. The channel fill will depend on the type of mesoforms or macroforms which have developed, and on the nature of the hydrograph.

2.3.6.3. Lateral Movement of the river

Because of the constant changes in the configuration of channels within the river, bank erosion is variable. In the Brahmaputra (Coleman 1969), both sides of the river may widen or narrow simultaneously. In other parts of the river one bank may widen and the other contract for a few years and then the situation may reverse. It seems quite possible that this process will produce lateral accretion surfaces. Annual rates of lateral erosion and deposition range from 50m to 500m per year. Where the river banks cross old clay plugged channels erosion is much reduced.

In addition to this switching of erosion and deposition, the river may progressively shift in one direction over a longer time period. The Brahmaputra has steadily been shifting its course westwards over the last 150 years in response to faulting and tilting of the adjacent Pleistocene Sediments. As a result, a sheet of sand averaging about 20m in thickness has been constructed, although locally, in node areas, thicknesses are nearly twice as great.

The Kosi River (Gole and Chitale 1966) is also shifting laterally in one direction, and has moved about 115km in the last 200 years. The fluvial sediments are forming a very large alluvial fan and the river is progressively shifting from one side of the fan to the other.

2.3.6.4. Scour and fill

Within multiple channel braided rivers islands are not developed everywhere. Usually there are narrow parts of the river, termed nodes (Chien 1961, Coleman 1969), where the flow is undivided and the channel deep. These are often located where the banks are composed of clay. During floods these areas accommodate the increased discharge by scouring more deeply. Sediment eroded from narrow parts of the river channel tends to be deposited in the next wide area downstream, (Lane and Borland 1954, Colby 1964, Coleman 1969). There is little information on how much scour occurs. In the Colorado River at Cross Ferry, during the 1956 flood where the channel

was about 1.5m deep at low stage and 7m deep at high stage the bed was eroded by about 3m (Leopold et.al. 1964), [Fig.13-A]. In the Yellow River at Chiang Kou, in 1919, (Freeman 1922), the channel was about 4m deep at maximum during low stage and 12m deep at high stage. The river eroded down by 4m as far as bedrock, and also widened its channel, [Fig.13-B]. The Colorado River at Black Canyon has been known to scour by over 38m during flood, (Leopold et.al. 1964), but these high values seem exceptional.

Wider areas of the river channel do not scour so deeply, but here erosion may still occur because of the wandering of the thalweg. In the Brahmaputra, (Coleman 1969) rapid deposition of sediment during falling stage and a reduction of the amount of water in the channel causes the thalweg to move by over 1000m in only a few days. Many new channels are formed, function for a short time and are then abandoned. Similar observations have been made on the Yellow River (Chien 1961). During low stage in the Brahmaputra, little thalweg movement takes place, but during the next rising stage, the river seeks a path with less curvature and again the thalweg shifts.

From these observations, it is apparent that the location and timing of scour and fill in a large river is very variable. Some parts of the channel may fill whilst simultaneously other parts are being eroded. Some parts of the river scour during falling stage and fill during rising stage and in other places the sequence is reversed. The importance for the geologist studying ancient sequences is that few channels are likely to be filled completely and then abandoned without going through a phase of erosion, and erosion surfaces should be common in braided sequences of this type.

2.3.6.5. Conclusions

The preceding review illustrates the variety and complexity of processes operating in multiple channel braided rivers. It is difficult to generalise about the resulting sequences. Probably different multiple

channel rivers will produce different types of sequences according to the dominance of the various processes described. Also, different types of sequences may be produced by one river in different parts of the channels. For example, two mechanisms, progressive channel abandonment and lateral accretion on islands may produce fining upwards sequence. Some channels may be straight with alternate bars whereas adjacent channels become braided at low stage. Much more work is needed on modern rivers of this type before ancient examples can be recognised and interpreted.

2.3.7. Palaeocurrents

Measurement of palaeocurrents has traditionally played an important part in the study of ancient fluvial sediments. The subject is complicated, not least because of the different types and scales of structure which may be measured (Allen 1966, 1967, Miall 1974), and which give different palaeocurrent patterns. Cross bedding is usually the most abundantly preserved structure in fluvial sequences and most palaeocurrent studies have used this. Most cross bedding is produced by the migration of mesoforms (Jackson 1975c), sand waves and dunes (Potter and Pettijohn 1963).

Two main sources of variation occur; variations of cross bedding within individual channels and variable orientation of different channels. Thus because of their wide range of channel orientations meandering rivers should show more variability of palaeocurrent vectors than braided and straight river channels, where supposedly most channels are orientated more nearly parallel to each other, parallel to the regional palaeoslope (Allen 1968). Of course, deltas are a separate case because of the tendency for channels to radiate.

Unfortunately, in spite of the very large number of studies of palaeocurrent directions in ancient fluvial sediments, there has yet to be a systematic comparison of current vectors in modern channels of different types. Some doubts may be cast on the assumptions made above, however. Smith (1972) measured foreset dip directions of lobate sandwaves in the

River Platte and found there was a considerable dispersion (vector magnitude 42.4% variance 5,073). Because the lobate shaped bedforms have sideways components of movement, he argued that ancient sequences with tabular cross bedding formed by similar bedforms should also have a wide dispersion because of this in-channel variability. Where sandwaves are straight crested, however, a much lower dispersion would be expected.

Smith did not actually measure cross bedding directions in the River Platte, so as yet his theory is unproved. Banks and Collinson (1974) argued that most sandwave movement takes place at high stage where lateral movement is limited. With only a forwards movement of the sandwave cross bedding formed at the sides will be volumetrically much lower than cross bedding formed at the front. Thus the variance of the preserved cross bedding should be much less than predicted by Smith.

Coleman (1969) undertook a large study of cross-bedding orientation within three different reaches of the Brahmaputra River. These showed a much lower variance (547-1109) than that predicted by Smith (1972), but only small scale cross bedding (sets < 1 - 3m) was measured and the bedforms were probably not the same as those in the Platte. Coleman did not compare the variance between the different channel reaches, but in all these the vector mean was similar (153° - 165°) and so the total variance would not be very great.

The nature of the river hydrograph may have an important effect on palaeocurrent range. In the River Platte substantial sand movement occurs during low and falling stage and lateral movement of sandwaves is considerable (Smith 1974). Small scale structures produced by bedforms generated during falling stage also have a higher variance (Collinson 1971, Miall 1974). Thus a river with a variable discharge and undergoing considerable low and falling stage reworking may exhibit a higher variance than one exhibiting mainly high stage structures. There a system of weighting of different scales of structure as proposed by Miall (1976), may be necessary in order to make meaningful comparisons.

In the meandering Red River, Harms et.al.(1963), measured the orientation of trough cross bedding exposed in the point bar. They found a variance of 1220 - 4740, lower than the predicted by Smith (1972) but higher than any value recorded by Coleman. Presumably, if several point bars of different orientation had been measured, the variance would have been considerably higher. It is important to remember that the sinuosity within meandering rivers varies considerably, from 1.5 to about 4 (Leopold et.al. 1964). Potter and Pettijohn (1963) measured cross bedding in point bar sands in the Vermillion River, Indiana. This is a meandering river of fairly moderate sinuosity. The variance calculated from their measurements is only 2,445 about half that predicted by Smith for the braided river Platte. Probably it would be greater if allowance was made for movement of the meander belt over time.

In conclusion, present methods of palaeocurrent analysis, usually consisting of cross-bedding orientation measurement and calculations of variance and vector magnitude, do not provide a very reliable means of distinguishing between meandering and non-meandering rivers in ancient sequences. In particular, straight and braided rivers with lobate sand waves such as are common in many modern rivers of this type, may show a high variance where there is substantial lateral movement during low and falling stages. This appears to be the case in the sequences described by High and Picard (1974) and Cant and Walker (1976). Meandering rivers of moderate sinuosity may have a similar or even lower variance. Other techniques, such as concentrating on trough cross bedding, or considering only higher ranked structures such as channel sides may prove to be more reliable.

2.3.8. Formation and preservation of vertical accretion sediments, comparisons between meandering and non-meandering rivers.

Schumm (1960, 1963) found a relationship between the percentage of total load carried as bed load and channel shape in Great Plains rivers. Highly sinuous or meandering channels carry a high suspended load of clay

or silt (bedload < 3% of total load), and have large amounts of clay and silt in their banks. Low sinuosity rivers on the other hand carry less suspended silt and clay (bed load > 11%), and have more sandy banks.

This has three important implications for the construction and preservation of vertical accretion flood plain sediments.. Firstly, for rivers of a given size, meandering channels carry a greater volume of fine grained sediment. Secondly, clayey banks are more resistant to erosion, therefore restricting lateral migration of the channel and favouring a higher preservation of overbank sediments. Thirdly, highly sinuous rivers do not shift their courses laterally beyond a certain point, because of the likelihood of meander cut offs. These cause the meander belt to become stabilised in one position favouring preservation of overbank sediments.

As a result, ancient meandering sequences should preserve relatively large amounts of vertically deposited flood plain sediment as in the models of Allen (1964, 1965, 1970) and Schumm (1968). In low sinuosity rivers, however, thick overbank fines will tend to be absent because of frequent lateral shifts of the channel.

Whether the above arguments are always true is open to question. Not all meandering rivers carry high suspended loads, McGowan and Garner (1970) have described meandering rivers of moderate sinuosity with little suspended sediment. The preservation of thick fine grained flood plain sediments depends also on external factors such as the rate of subsidence. If this is low, even meandering rivers may erode away their overbank sediment as envisaged by Yeakel (1962), particularly in rivers of moderate sinuosity where cut offs will be less common.

Not all braided and low sinuosity rivers have small suspended loads. In Schumm's (1960) study silt clay percentages in low sinuosity rivers varied from 12% to nearly 90% of the total load. Generally speaking the concentration of suspended sediment in rivers increases away from mountains towards the centre of the drainage basin (Colby 1963). All large rivers,

braided or not, will carry large suspended loads. Few published figures of suspended loads in large braided rivers are available. The Brahmaputra transports 7×10^6 tons of suspended sediment a day during floods, (Coleman 1969) and has an annual average suspended load of over 7×10^8 tons (Holeman 1968). The Colorado River transported up to 1 million tons of suspended sediment a day during floods in the year 1967-1968 (U.S. Geol. Survey Water Supply Paper 2098) and has an annual average of 135 million tons (Holeman 1968). The Yellow River has an annual average of over 1.8×10^8 tons, 50 times as great as the Mississippi (Holeman 1968).

A considerable amount of deposition of fine material seems possible in drainage basins of braided regimes. Whether this is preserved depends on the nature of the movement of the main channel, and this will be determined to a considerable extent by external, particularly tectonic factors. The Brahmaputra River (Coleman 1969), for example, is steadily shifting its course westwards in response to tectonic movements. During floods vast areas of the flood basin are inundated, and up to 1cm of fine grained sediment a year accumulates. Because of the climate, rapid growth of vegetation is common and peat deposits 1 - 4m thick accumulate (Khan 1957). These vertical accretion deposits overly the channel sand. With continual subsidence a considerable thickness will accumulate and these may be preserved from erosion during the next change in course of the river.

2.4. Conclusions

In this chapter some of the major factors influencing sedimentation in non-meandering river regimes have been discussed and compared with the present models of sedimentation in meandering rivers. Distinguishing between the two types of regime is not as simple as has sometimes been implied.

A number of different types of channel sequences may be produced in meandering rivers and the presence or absence of a fining upwards channel fill is not diagnostic. McGowan and Garner (1970) and Jackson (1975) have both shown how meandering rivers may deposit channel fills which do not fine

upwards, whilst the process of progressive channel abandonment which takes place in multiple channel braided rivers, such as the Niger, probably also produces channel fills which fine upwards. The presence of lateral accretion surfaces is probably not diagnostic. Coleman (1969) has shown how progressive sideways migration of the banks takes place in the braided Brahmaputra River. Shantzer (1951) has shown that braid bars may also grow by a lateral accretion.

Palaeocurrent variance is not a particularly reliable criterion at present. The behaviour of lenticular sandwaves in braided rivers is discussed by Smith (1972, 1974), and possible high variance of the result cross bedding suggests that not all braided regimes have a low variance. This is the case in ancient presumed braided sequences described by High and Picard (1974) and Cant and Walker (1976).

The proportion of overbank to channel sediment in the sequence may not always be diagnostic; its preservation depends to some extent on the tectonic setting of the basin of deposition.

At present, perhaps the only undisputed ancient meandering sequences described are those where aerial photographs show the preserved meandering channels with their scroll bar topography (Puidefabrigas 1973, Nami 1976).

The detailed characteristics of the channel fill sequence will probably prove to be the most important distinguishing criterion. As discussed above, most braided rivers have a variable discharge, and this has an important influence on sequences. Scour and fill is more common in braided rivers, where there are multiple channels, abandoned channels should be commonly preserved. Straight channels with alternate bars should also be distinguishable.

In Sections II and III two different examples of probable ancient non-meandering channels will be described and compared with others, described in the literature. All are quite different from any of the meandering river channel models proposed so far and are thought to represent different types of braided and straight river regimes.

Table II

Essential elements of the 'classical' meandering river models

Channel Fill.	Fines upwards, cross bedding becomes smaller upwards. May show lateral accretion surfaces.
Overbank Sediments.	Extensive preservation between channel fills.
Palaeocurrents.	High variance.

SECTION II. THE SEDIMENTOLOGY OF THE ROACHES GRIT GROUP IN NORTH
STAFFORDSHIRE AND ADJACENT AREAS

Chapter 3

3.1. The area

The Group outcrops in the extreme south-west of the Pennine district of northern England between the towns of Cheadle in the south east, Macclesfield in the west, and Chapel-en-le-Frith in the north, most lying within the Peak District National Park. The greater part of the area is high moorland, the remaining parts being mainly devoted to dairy farming. Many roads criss-cross the area and no outcrops are more than a mile from a road.

Most of the outcrops in the upper parts of the sequence are large natural gritstone escarpments. With the exception of a large working quarry at Walker Barn (SS 952 763) stream sections provide most of the outcrop in the lower and middle parts of the sequence. Most of the streams are small, and there are only small areas of outcrop with many gaps in the usual vertical sequences.

3.2. Structure

The area lies on the western side of the main Pennine Anticline [Fig.14]. The rocks have been folded into a number of subsidiary anticlines and synclines with north south axes [Fig.15]. The most important of these is the Goyt syncline which extends from Upper Hulme, just north of Leek, to the northern edge of the area. It is cut by a number of strike slip faults. The other major structures are the Todd Brook anticline in the north west of the area, the Rushton Hall anticline in the mid-west and the flanks of the Potteries syncline in the south. Small exposures occur in the much thinner sequence on the northern flanks of the Cheadle Coalfield syncline. Exposure is terminated abruptly in the west by the north-south trending Red Rock Fault, which forms the boundary of the Permo-Triassic Cheshire Basin and has a throw of several thousands of metres.

3.3. Stratigraphic Position

The sequences lies in the upper part of the Marsdenian between the Reticuloceras bilingue late form goniatite band and the R. superbilingue band. It conformably follows the lower and middle Namurian sequences described by Holdsworth and co-workers (op.cit.) [Fig.16].

Recent work in Lancashire (Ramsbottom 1969a), and elsewhere, has shown that there are three horizons with different mutations of R. bilingue. These are, in ascending order, R. bilingue late form, (Bisat), R. eometabilingue, (Bisat and Ramsbottom) and R. metabilingue (Wright).

All these forms have now been recognised in north Staffordshire (N. Aitkenhead P.comm.) but not yet within the same section.

The highest form is only known for certain from three localities at present, Brownsett Farm (SK 9926 6371) (N. Aitkenhead, P. Comm), and two in Macclesfield Forest, (SK 9657 7130), (Evans et.al. 1968) and SK 9721 7293 (N. Aitkenhead, P. Comm). The first and third localities are below the lowest mappable sandstones in the sequence, but the second locality is mapped as within some of the lower sandstones. It seems likely that the two bands are not time equivalents, and the Macclesfield Forest occurrence may represent a local horizon. Morris (1969) describes R. metabilingue from Cotton (SK 0617 4640), but Ashton (1975) suggests that this is a mis-identification of R. eometabilingue.

R. bilingue late form and R. eometabilingue have been recognised from a number of widespread sites within the study area. The R. superbilingue band which defines the top of the sequence studied is a widespread and easily recognised horizon throughout the Central Pennines.

Sandstones within the sequence are known by a variety of lithostratigraphic names (Cope 1946, Evans et.al. 1968, Stevenson and Gaunt 1971). These have been retained. No new palaeontological work has been attempted, but the sequence between R. metabilingue and R. superbilingue is here termed the Roaches Grit Group [Table II].

3.4. Previous Research on the Upper Namurian of north Staffordshire

The first important research on the area was at the beginning of the Nineteenth Century. White Watson (1811) in his 'Strata of Derbyshire' published a section across the Derbyshire Dome showing the main anticlinal structure. Between Buxton and Coombs Moss, he recorded above the limestone, a thick 'Aluminous Shale sequence' followed by a thin 'Shale Grit' and above this a thick sandstone which he called 'Millstone Sandstone' and is probably the Roaches Grit. He realised that the sequence was widespread and could also be found on the eastern side of the anticline at Ashover and Matlock.

In the same year, Farey, (1811) published the first geological map of the area showing major structures such as the Potteries and Goyt synclines and Derbyshire anticline.

Hull and Green (1864, 1866) substantially improved Farey's map and delineated most of the important structures. Following the work of Phillips in Yorkshire they recognised in the upper Carboniferous three subdivisions.

Coal Measures

Millstone Grit

Yoredale Quartzites

Within the Millstone Grit they identified, from the top downwards, the '1st Grit' or Rough Rock, the '3rd Grit', the thickest sandstone in the area, and now called the Roaches Grit and the '4th' and '5th' Grits which they correlated with the Kinderscout Grit in Derbyshire. They noticed the 'coarse, often conglomerific nature' of the 3rd Grit and the 'changeable nature' of the 4th and 5th Grits which were often interbedded with shales.

Challinor (1921, 1929, 1930), mapped around the Roaches area. Following Pocock (1906) he called the lowest coarse grained grit the Roaches Grit, and fine sandstones beneath the Five Clouds Sandstone. His discovery of R. bilingue in the upper Churnet valley, showed that here the base of the coarse grained 'Millstone Grit' is stratigraphically higher

than in the northern Pennines. Subsequently Cope (1946) showed that the Roaches Grit Group lay between the R. *bilingue* late form and R. *superbilingue* band and is therefore of upper Marsdenian age. Mapping by Evans et.al. (1968) and Stevenson and Gaunt (1971) has shown that a similar sequence extends into the Macclesfield and Chapel-en-le-Frith areas. Morris (1969) proved a similar sequence around the northern flanks of the Cheadle Coalfield, but here the equivalent of the Roaches Grit Group is much thinner.

The most important work on the Namurian of north Staffordshire since the war is by Holdsworth (1963). Although he did not study the Roaches Grit Group in detail he noted the sandstones within it were rich in K felspar and quite different petrographically from the protoquartzites which dominate the lower and middle Namurian and which had, he considered, a southerly derivation. He showed that during the lower and middle Namurian the north Staffs Basin [Fig.53] was dominated by relatively deep water turbidite sedimentation with shallow water conditions restricted to the basin margin. The western margin of the basin appeared to lie along the line of the present Red Rock Fault and to the west of this lay a platform of shallow water or emergent condition extending into North Wales. The eastern edge of the basin was marked by the Derbyshire Block.

The area of exposure of the present sequence [within the rectangle Fig.53] lies mostly within the North Staffs Basin. There is no exposure west of the Red Rock Fault, but a little exposure at the edge of the Derbyshire Block.

Holdsworth suggested that the Roaches Grit represented the first widespread occurrence of shallow water sedimentation within the north Staffs Basin.

3.5. Previous research on the sedimentology of the Namurian of northern England

Phillips (1836), who worked mainly in Yorkshire, was the first to carry out a facies analysis of the Namurian. He recognised seven rock types,

shale, gray beds (alternatives of shale and sandstone), flagstone, galliard (gannister), gritstone, ironstone and coal. He recorded trace fossils in the flagstones and suggested they indicated a littoral environment.

Sorby (1859) compared the "drift bedding" (cross bedding) of the Namurian of Yorkshire and Derbyshire with that which he produced experimentally. Recognising its palaeocurrent significant, he found that "on average the current was from the north east", and after also looking at the pebble petrography he concluded that the Namurian of that area was derived "from the waste of a south westwards prolongation of an ancient Scandinavia, the site of which is now occupied by the northern North Sea".

Gilligan (1920) carried out a detailed petrographic study of the Namurian and agreed with Sorby as to the sediment source. He was the first to suggest that it was deposited by a river of very large size and produced the well known diagram of a river Nile type delta superimposed on a map of northern England.

The Geological Survey Memoir on the Rossendale area, (Wright et.al. 1927) was the first work which recognised coarsening upwards sequences and interpreted them as delta sequences infilling a basin of water with repeated transgressions producing rhythmic variation in sedimentation.

Trotter (1951) recognised seven sedimentation facies in the Namurian of north-west England. The two most important are, the "Fluvial grit", facies comprising the coarse grained cross bedded sandstones, which he suggested was laid down in river valleys. Flanking this, a "grit-shale" facies which he interpreted as estuarine to open sea deposits with variations due to wave and current action.

The modern era of research on the sedimentology of the north of England Namurian begins with Allen (1960). He showed that the Mam Tor Beds in north Derbyshire, part of Trotters "grit-shale" facies were turbidites. These he considered were derived off the front of deltas to the north.

Holdsworth (1963), described the lower and middle Namurian in north Staffordshire. He showed that this area had a different sedimentological history being dominated by thin turbidite sequences derived from the south.

Reading (1964) suggested that there were two main stages of infill in the Central Pennine Basin. The first in Eic, the Skipton Moor Grit, the second in Ric, the Kinderscout Grit, which extended deltaic sedimentation into north Derbyshire. He suggested that the Shale Grit and Grindslow Shales of the Derbyshire sequences were the deposits of the upper slope of an advancing delta.

Mayhew (1967) investigated the sedimentology of the Ashover and Chatsworth Grits in south east Derbyshire. He recognised two major facies: a lower trough cross bedded non-pebbly sandstone which he suggested was a distributary mouth bar deposit, and an upper coarse grained pebbly facies dominated by planar cross bedding which was deposited within fluvial channels.

Between 1961 and 1975 there has been detailed sedimentological research on the Kinderscout sequence by Walker, Collinson and McCabe. Because of its close similarities with the Roaches Grit Group, this will be described in detail later in the text.

Table III

Stratigraphic position of the sequence studied

Sequence Studied	}	<u>R. superbilingus</u>	}	Roaches
		ROACHES GRIT (locally SHIRLEY HOLLOW SANDSTONE)		Grit
		FIVE CLOUDS SANDSTONE (locally WALKER BARN GRIT, RUSHTON HALL GRIT, CORBAR GRIT)		Group
		[Lower part includes a probable local occurrence of <u>R. metabilingue</u>]		
		<u>R. metabilingue</u>		
		<u>R. eometabilingue</u>		
		<u>R. bilingue</u> late mut.		
		<u>R. bilingue</u> ss.		
Sequence studied by Holdsworth 1963	}			

Chapter 4

Association A. Deep Water Sediments

4.1. Lithofacies description and interpretation4.1.1. Lithofacies 1. Mudstone

This contains clay, scattered fine silt and sometimes scattered large mica plates. Bivalves and fish fragments together with coprolites may also occur. No trace fossils have been found but the lithology makes their identification difficult. Small, flat lying plant fragments are common.

Its appearance varies considerably depending on the state of weathering. When fresh it is usually light to medium grey in colour, homogenous, breaking with a conchoidal fracture. Occasionally it shows a faint light dark colour banding a few millimetres in thickness. When weathered it commonly has a very thin, fissile lamination with powdery Jarosite along parting planes. Alternatively it may be partly stained red by iron oxide and break irregularly.

Spheroidal carbonate concretions; up to 0.15m in diameter are common and thin, 1-2mm carbonate bands parallel to the bedding may also occur. Sections through concretions show disseminated pyrite along bedding planes. The fissile lamination invariably bends around the concretions, it has almost certainly a strong secondary component to its development resulting from compaction and is not entirely related to an original sedimentary fabric.

Interpretation

The features of the lithofacies indicate deposition from suspension in quiet water. It was deposited in a variety of associations but is most common in the Deep Water Association.

4.1.2. Lithofacies 2. Silty Mudstone

Light grey in colour with abundant scattered silt-sized quartz grains and mica plates. Plant fragments and carbonaceous debris are common. No

fauna has been found within this lithofacies.

When weathered, it appears completely homogenous except for a poor fissility similar to that found in LF 1. When traced laterally into fresh exposure its appearance sometimes changes to a well laminated rock consisting of alternating light and dark layers. Light layers are 0.25 - 1cm thick with gradational boundaries and are mainly quartz silt.

No trace fossils have been found within this lithofacies but the normal weathered condition of the rocks which tends to obliterate faint lamination would also destroy any trace fossil evidence.

Interpretation

The lack of traction formed sedimentary structures implies deposition from suspension in fairly still water. Again this lithofacies was deposited in a variety of water depths and is common in the Deep Water Association.

4.1.3. Lithofacies 3. Goniatite Faunal Bed

This facies is very characteristic and easily recognised from a distance in the field as a sooty black mudstone much darker than in LF 1. Beds are thin, 0.3 - 1.3m and vary in thickness laterally. When weathered it exhibits a very poor lamination with Jarosite and Limonite coating bedding planes. Mica plates are absent but plant material fairly common.

The fauna consists of goniatites, pectinid bivalves, and Posidonia sp. Most of the fauna is adult but mollusc spat may also occur. Preservation is mainly as impressions in mudstone but uncrushed preservation of goniatites occurs in bullions. Most goniatites are flat lying.

No systematic study of the faunal composition has been carried out but there appear to be no marked lateral variations in fauna even though the bands in the south-eastern part of the area were probably deposited in a much shallower water than their equivalents from the north.

Interpretation

It is generally accepted that goniatite faunal bands represent periods of high salinity, marine or near marine conditions in an otherwise brackish

or fresh water basin. For many years goniatite bands were actually called marine bands in Geological Survey maps and memoirs, although the non-marine origin of the associated non-goniatite bearing sediments has still to be conclusively proved.

Many bands are widespread, the R. metabilingue and R. eometabilingue bands at the base of the sequence have also been recognised in Lancashire (Ramsbottom 1969a). The R. superbilingue band at the top of the sequence is a widely recognised horizon in the Central Pennines. Exact time equivalence is difficult to prove however. The probable occurrence of two R. metabilingue bands in the area, the one a local development suggests that goniatites could survive in local areas if conditions were suitable. Had there been no coarser sandstones in this part of the sequence to serve as an independent marker horizon the two bands would probably have been correlated as time equivalents.

The goniatites and pectinid bivalves were nektonic and presumably fell to the basin floor after death. The presence of the benthonic Lingula in the R. superbilingue band shows that here at least, conditions were not euxinic. The R. superbilingue band which belongs to Association D was probably deposited in considerably less depth of water than the four bands in Association A. Except for the presence of Lingula casual observations show no differences.

4.1.4. Lithofacies 4. Thin-bedded Turbidites

Sandstone and coarse siltstones from 0.1cm to 50cm in thickness. Generally they have parallel bases and tops and may be traced laterally for up to 20m, the limits of exposure, without showing any thickness variation [Plates 1 and 2]. A few are lenticular, and disappear and reappear at the same horizon. The beds alternate with LF 1 and LF 2.

Bases are always sharp and usually flat, but a variety of bottom structures may occur.

- (i) Endichnial trace fossils. These are described in Chapter 8.
- (ii) Load casts. Bulbous depressions usually a few centimeters in diameter, but some elliptical depressions are up to 50cm in length and 3cm deep. Occasionally they are found in rows [Plate 3].
- (iii) Flute casts. These are small, the largest recorded being 25cm in length.
- (iv) Furrows. Parallel sided rows 0.75cm - 1.5cm, symmetrical in cross section. They extend lengthways for at least 40cm sometimes pinching out between adjacent structures. [Plate 4].
- (v) Groove casts. Long straight ridges often asymmetrical in cross section. [Plate 5].
- (vi) Prod casts. Up to 10cm in length but generally less than 3cm and with blunt ends.
- (vii) Plant impressions. Large depressions resembling prod casts but where the form of a plant is clearly defined [Plate 6].

Of 32 bottom structures from which current directions have been obtained the distribution is: Prods 16 (50%), grooves 12 (38%), flutes 2 (6%) and furrows 2 (6%).

Within the beds up to four types of sedimentary structure may occur; a massive division, two types of plane lamination and a current ripple division. The vertical order of the structures always follows the Bouma (1962) sequence. The massive division is typically a fairly clean and fairly well sorted structureless sand with little clay. Towards the top

there is an increase in the proportion of carbonaceous debris and mica, together with a decrease in the median grain size of the quartz. Scattered angular mudflakes and large plants sometimes occur near the base.

The lower interval of plane lamination usually contains a much larger proportion of clay, carbonaceous debris and mica. The lamination, about 1mm thick, develops from an alternation of quartz rich layers, with either carbonaceous material plus clay or mica rich layers the latter showing primary current lineation.

The current ripples are linguoid with a maximum height of $2\frac{1}{2}$ cm. In thin beds, containing this division only, the ripples exist as solitary form sets and heights may be as little as 1mm. Cross-laminations are marked by mica rich layers and show an enrichment in mud towards their bases. Normally single sets occur and ripples are slightly erosive into the underlying plane laminated division. Where there are two or more sets they occur as mutually erosive troughs. Ripple drift cross lamination has not been seen.

The form of the topmost ripple is preserved and buried by the overlying mudstone. Sometimes in unweathered exposures it is possible to see finely laminated micaceous silt draping over the ripple and forming an upper division of plane lamination. Graded laminated beds (Piper 1972) do not occur.

Not every bed has the full sequence of structures. The massive division in particular tends to be absent in the thinner beds. No attempt has been made to assess types of vertical sequences quantitatively because of the difficulty of recognising the two plane laminated divisions, particularly the upper one, in weathered exposures. Common sequences using Bouma (1962) notation are: ABCE, BCE, CE, but AE and BE sequences also occur. In the latter, the plane laminations become progressively finer grained towards the top.

Interpretation

The sharp based sandstones interbedded with fine grained mudstones suggests a sudden influx of coarse sediment into an otherwise quiet water area receiving sediment from suspension. The vertical sequence of sedimentary structures suggests a waning flow (Harms and Fahnestock, 1965, Walker 1965). The assemblage is similar to that seen in turbidites. Evidence discussed later suggests deposition at depths in excess of 100m.

Three features which require explanation are the predominance of single sets of ripples, the erosive ripple bases, and presence of AE and BE sequences where ripples are absent.

Ripples in turbidites may form in the tail of the current (Walker 1965), by water swept aside by the turbidity current sweeping back after the current has passed (Kelling 1964a) or by bottom currents (Hsü 1964). The similarity in current mean direction between the ripples and turbidite bottom structures [Fig.17] is consistent with the first explanation but does not necessarily invalidate the other two. Other turbidites of Namurian age in the Central Pennines, in the Todmorden Grit (McCabe 1975) and in the Roosecote Mudstones (Eic) (personal observations) also have dominantly solitary sets of ripples. In all three cases ripple palaeocurrent means and bottom structure palaeocurrent means are the same, to the south-west in the latter two cases and to the north-west in the Roaches sequence.

It is unlikely that bottom currents were responsible as the coincidence between ripple and bottom structure orientations seem too great. The second mechanism fails to account for beds where a rippled top is absent. The first mechanism, formation in the tail of the turbidite seems the most likely. Although Trewin and Holdsworth (1973) have suggested that erosively based ripples result from basinal reworking of the turbidite tops, McCabe (1975) has shown that ripples with erosive bases may form during net sedimentation providing the sedimentation rate is slow enough. The predominance of single sets and absence of ripple drift cross lamination does point to a slow rate of sedimentation.

Where turbidites do not have rippled tops the rate of deposition may have been too fast to allow time for ripples to form. Alternatively, large amounts of suspended sediment may have inhibited their development.

Whether the interbedded mudstones are of turbidite or 'hemipelagic' origin is difficult to assess. Walker (1966a) suggested that within the Shale Grit the mudstones belonged to the E division of the turbidite. Sometimes finely laminated silts above the turbidite C division pass upwards gradationally into the overlying mudstone and this probably represents the fine tail of the turbidite.

4.1.5. Lithofacies 5. Thick Sandstones

Sandstones medium to coarse grained in beds of 50cm to 13.5m, the majority being less than 5m. The greater part of the bed is a structureless fairly well sorted clean sand. The lower parts frequently contain bands of angular mudstone flakes and large plant fragments scattered within a sandstone matrix. Towards the top there is often an increase in the proportion of small carbonaceous fragments and mica together with a decrease in the average quartz grain size. A plane lamination, and occasionally above this ripple lamination may develop towards the top in a similar way to LF 4.

The bases of beds are usually flat. The beds are parallel-sided and are laterally extensive over tens of meters [Plates 7 and 8]. Often they amalgamate together to form thick sandstones in the manner described by Walker (1966a). Lines of amalgamation are commonly marked by rows of mudstone flakes [Plate 9] or horizontal joints. The percentage of mudstone between beds is much lower than in LF 4.

Flute and tool marks are absent from the bases, but load casts are fairly common. These disturb the underlying lamination, and at Walker Barn Quarry (SS 952 736) one underlying bed has vertical to slightly inclined pipes 5mm in diameter and up to 40cm in length filled with clean sand. Some bases show small scale channeling [Plate 10] and two larger channels occur. At Danes Mill (SK 0123 6119) part of the side of a channel is poorly exposed and is at least 3.2m deep. A much larger channel occurs within Association B. This will be described in Chapter 5.

In a series of quarries near Walker Barn (SS 952 763) large spherical concretions up to 3m in diameter occur [Plates 7 and 8]. Hardy (1970), analysed several of these and found them to be chiefly ankerite or calcite with some siderite. Concretions of this size have not been found elsewhere.

Interpretation

These beds are similar to LF 4 Turbidites with which they occur in close association. The main differences are; the channeled bases, amalgamation is the rule rather than the exception, the beds are predominantly structureless, and much thicker than in LF 4.

Walker (1966a) described similar beds from the Shale Grit (Namurian Ric) in north Derbyshire (his Facies C), interpreting them as turbidites. At that time few types of density underflow currents were recognised, but now three other types of current must be considered.

Grain Flows. These are concentrated dispersions of cohesionless sediment which move downslope in response to the pull of gravity, the dispersion being maintained by intergranular collision. McCabe (1975), suggested that grain flows may have played a part in the transport of a similar facies, (his F4) in the Todmorden Grit and lithostratigraphic equivalents. They are necessary he claimed to explain the presence of mudstone clasts floating near the tops of the beds and also to transport the very large clasts near the base.

Grain flows may be observed on the leeward faces of bedforms and deltas. These invariably involve relatively small volumes of sediment and

form beds 5cm thick or less. Large examples are unknown. Middleton and Hampton (1973), suggested that grain flow deposits are massive upgraded, except for a possible reverse grading near the base and with flat tops. More recent work, Lowe (1976a) suggests, however, that grain flows of just sand only occur on slopes near the angle of repose and cannot form very thick units of sediment. In other words, they always resemble the small flows observed on the slip faces of bedforms.

Thick grain flows of gravel sized debris and sandy flows with interstitial mud can move on much lower slopes and form thick sedimentation units (Middleton 1970, Lowe 1976a). The sediments under discussion do not contain gravel sized debris, the clay content is less than 3% and most of this appears to be a breakdown product of K feldspar. The mechanism cannot therefore have operated as a primary process.

In the absence of an interstitial matrix, sandy grain flows might possibly be driven by overriding turbidity currents which shear them from above. It seems unlikely, however, that very thick beds could be transported in this way (Lowe 1976a).

Debris Flows. These are downslope movements of poorly sorted mixtures of granular solids including abundant clays and water in response to the pull of gravity. Subaerial debris-flow deposits have a poorly sorted fabric with large clasts 'floating' in a finer matrix. They resemble tills. Lithofacies 5 is quite different. Hampton, (1972), however, has suggested that only 10% of clay may be necessary to support the sand sized material. Such flows may be regarded as intermediate between debris flows and grain flows. Again, the low clay content suggests this mechanism is unlikely.

Liquified Flow. Coarse sediment liquified after slumping or by compaction of underlying underconsolidated clay layers can flow down slopes of only a few degrees (Lowe 1976b). The grains settle through the liquid displacing it upwards. The resulting deposits show de-watering structures such as

fluid escape pipes with sand volcanos and dish structure (Stauffer 1967). Because of the concentration of the sediment it is unlikely that fluid traction structures will form. (Middleton and Hampton 1973).

Coarse grained liquified flows will resediment after moving only a few metres (Lowe 1976b) and it is unlikely that liquified flow transported the sediment to the basin floor. Many turbidity currents however, go through a short-lived liquified phase immediately prior to deposition (Lowe 1976b). No dish structures have been observed in LF 5. 'Fake' dish structure in the form of ropey weathering occurs but this has also been seen in a variety of other lithofacies and is probably solely a weathering phenomena. Fluid escape pipes occur in one bed, but here dewatering appears to have been caused by the shear of the overlying bed. Plane laminations, and rare ripple lamination at the tops of some beds show that grain traction occurred.

The absence of dewatering structures suggests that a liquified phase was not normally developed.

Turbidity Currents. These seem the most likely agents of transport for the sands. The absence of flutes is probably a function of the relatively coarse grain size. Allen (1969) has calculated that velocities of 1.5m sec^{-1} for medium sand, and 2m sec^{-1} for coarse sand are necessary for flute growth. Walker (1966a) recorded flutes on the base of a similar facies (his Facies C) and flutes also occur on a similar facies in the Alum Crag Grit (R2) of Lancashire (Collinson, Jones and Wilson, in press).

The deposition of a thick bed of structureless sand is still difficult to explain, although a large number of mechanism have now been proposed. Walker (1965) suggested that structureless beds may result from the sudden freezing of a traction carpet with a high concentration of dispersed grains, the shear becoming insufficient to keep the mass of grains in transport. Walton (1967) has speculated that a rapid deceleration of the current may prevent an equilibrium bed from developing. Deposition in this

"metastable" field would consequently be unlaminated. Middleton (1967) suggested that syndepositional shearing may produce a structureless bed, as was observed in experimental highly concentrated turbidites. Rapid deposition from suspension on a surface without traction may also result in a structureless bed.

If, during deposition a turbidity current changes into one of the types of flow discussed above a structureless bed may form. Lowe (1976b) suggests that many high concentration turbidity currents go through a liquified phase immediately prior to deposition. The absence of water escape structures suggests this is unlikely in the present case. Sanders (1965) suggested that a grain flow may develop beneath turbidity currents, but Lowe (1976a) now thinks this unlikely.

4.1.6. Lithofacies 6. Parallel-sided Sandstones

Sandstones in beds 2 - 20cm in thickness. Bases and tops are parallel and may be sharp or slightly gradational. Internally the beds are a poorly sorted fine to medium sand with scattered mica, plant fragments and clay, giving a characteristic 'greywacke' appearance. Usually they are homogenous, but they may show a poorly developed flat lamination. [Plates 11 and 12.] Unspecific bioturbation is common and one bed contains Pelecypodichnus.

Grain size varies irregularly through the bed and grading has never been observed. Current formed bottom structures are absent, as also is ripple lamination. Amalgamation between beds never occurs.

The sandstones may occur as isolated beds within sequences of LF 4 and LF 5 turbidites. Usually, they form thick sequences and are interbedded with LF 2 and 3 mudstones. Generally speaking bed thicknesses and bed spacing are more regular than in the turbidites.

Interpretation

This lithofacies is quite different from LF 4 and LF 5 turbidites from which it may be distinguished by the absence of grading and Bouma

sequences, by being generally more poorly sorted and by sometimes having gradational bases.

The absence of traction formed sedimentary structures suggests that deposition was mainly from suspension. The nature of the currents depositing the beds is uncertain. A similar lithofacies occurs in Association B and further discussion is given in Chapter 5.

4.2. Facies relationships within the association

Three LF 3 Goniatite faunal bands occur close together at the base of the association. The interbedded sediment is mainly LF 1. LF 4 turbidites first appear above the highest Goniatite band, R. metabilingue. A second band of R. metabilingue is mapped as occurring within a turbidite sequence of Macclesfield Forest (SK 9657 7130) (Evans et.al. 1968) but exposure is very poor in this area.

Turbidites of LF 4 and LF 5 are the most important members of the deep water association except in the northern and southern edges of the area where the sequence begins to thin rapidly (Chapter 9). Walker (1966a) suggested that the turbidites of the Shale Grit formed submarine fans, which were preceded by thinner 'basin plain' type turbidites. Following more recent work on active deep water fans in the present day ocean basins (Normark 1970, Haner 1971, Nelson 1971, Nelson and Kulm 1973), a fan model for ancient deep water turbidites has been developed, (Ricci Lucchi 1969, 1975, Mutti and Ricci Lucchi 1972, Walker and Mutti 1973).

It is now important to compare the delta front turbidites of the Central Pennine Basin with these more common, largely shelf derived turbidites. In the model of Mutti and Ricci Lucchi (1972) progradation of the fan results in an overall coarsening upwards sequence developed between the basin plain and base of slope environments. Within the fan channel fills show thinning upwards sequences, believed to result from the progressive abandonment of the channel. Coarse sediments of a variety of

facies predominate. Interchannel areas have a greater proportion of fine sediment with thin turbidite CDE sequences. In the outer and middle fan channels are rare and in the middle fan turbidites are arranged in thickening upwards sequences which are believed to form beyond distributory mouths as prograding suprafan lobes.

Unfortunately, exposure in the Deep Water Association is poor and there are few long continuously exposed sections. Large scale sections, particularly 4, 5 and 6 in Fig.18 show a rough overall coarsening upwards with an increase in the abundance of LF 5 at the expense of LF 4. Three detailed sections are shown in Fig.18, where A shows a well developed thinning upwards sequence of 11m. The base is not exposed but mapping into a nearby stream section shows that the sequence is virtually complete. Two other sections, one with predominantly LF 4, the other with predominantly LF 5 show no vertical organisation, although the thickness of the individual beds varies considerably. Another thinning upwards sequence occurs at Corbar Caravan Park (SK 051 740).

The thick multiple sandstone beds of LF 5, cannot be traced laterally in exposed sections for more than 200m and except for amalgamation show little lateral variation. Feature mapping near Walker Barn and at the Five Clouds (SK 005 625) shows that these thick sandstones split into thinner beds over distances of a few kilometers, and that the proportion of interbedded fine sediment increases. [Fig.15]. It would be particularly interesting to know whether these thick sandstones are parts of channel fills, as Ricci Lucchi (1975) records channels up to several kilometers in width. Channels of this size would be virtually impossible to detect here where most of the exposure is within narrow stream sections. Hamilton (1967) has described 'depositional' channels from abyssal plain turbidites off Alaska. These channels are not erosive, the channel infill sediment passing laterally into leveés. This type of channel may account for the lateral splitting observed at the Five Clouds and at Walker Barn.

Mapping on a larger scale [Fig.15] shows that the thick turbidite sandstones are developed in two areas. One of these, beneath the Roaches at the Five Clouds thins both to the north and south and mappable sandstones extend for about 12km normal to the palaeocurrent mean. A second thick turbidite sandstone sequence occurs around Buxton and to the north-west at Walker Barn near Macclesfield. This also thins to the north and south.

These two areas with a thick development of turbidite sandstones are thought to be two small fans which developed locally around distributary feeder channels. The overall coarsening upwards sequences in the turbidites result from fan progradation. The detailed vertical sequences with rapid changes in turbidite thickness are similar to those described from other ancient fan sequences by Ricci Lucchi (op.cit.) and Kruit et.al. (1975). The rarity of erosive channels probably reflects the limited exposure, unless the channels are of the "depositional" type.

In the north-east of the area, between Buxton and Chapel-en-le-Frith the turbidites thin out and disappear. Here LF 6 parallel-sided sandstones become increasingly important and at Ridge Clough (SK 060 788) near Chapel-en-le-Frith [Section 9 in Fig.18] constitute the greater part of the Association.

In the south-east of the area, on the flanks of the Cheadle Coalfield, where the entire sequence has thinned to only 70m, [Sections 1 and 2 in Fig.18] the Deep Water Association is not recognised.

Chapter 5. Association B Delta Slope.

5.1. Limits of the Association

The base of the association is usually taken at the first appearance in the sequence of LF 7 Ripple laminated Sandstone. This usually appears immediately above the highest bedded turbidites. At Ridge Clough (SK 060 788) where turbidites are largely absent from the Deep Water Association the base of the delta slope cannot be accurately fixed. At Shirley Hollow (SK 0428 4870 - SK 0407 4835) and Cotton (SK 0604 4580 - SK 0601 4556) where the Roaches Grit Group is much thinner and the Deep Water Association is absent, the base of the slope is also difficult to delineate, but it probably comes not far above the highest goniatite faunal band.

The top of the Association is also difficult to fix. At Upper Hulme (SK 0123 6121 - SK 0129 6247) [Section 4 in Fig.18] it is marked by the base of a large alluvial channel. At Cisterns Clough (SK 0330 6983 - SK 0370 6980) [Section 6 in Fig.18], Ridge Clough, Section 9 in Fig.18 and Hogshaw Brook (SK 0552 7483 - SK 0560 7471) [Section 8 in Fig.18] beds of LF 12 lenticularly bedded sandstone appear below the lowest fluvial channel in the sequence. The base of the lowest defines the top of the Slope Association.

As defined the Association is 70m thick at Upper Hulme, 60m at Burbage Reservoir (SK 9654 6517 - SK 9663 6570) [Section 5 in Fig.18] and at least 100m in Cisterns Clough. The differences in thickness partly reflect the difficulty of fixing the limits of the association accurately, though in some sections the upper parts have probably been cut out by erosion in deep delta top distributory channels.

5.2. Lithofacies - description and interpretation

5.2.1. Lithofacies 7. Ripple laminated sandstone

Medium grained sandstones with cosets of trough shaped cross lamination 0.5 - 2cm thick [Plate 13]. The sediment is very micaceous and carbonaceous partings pick out the lamination. Cosets occur in beds up to

10m thick but usually they are 4 - 20cm thick and interbedded with LF 8 [Fig.19]. They have parallel bases and tops and superficially resemble turbidites in field exposures [Plate 14], the coarser sandstones weathering out. Within the beds, the ripple lamination may pass up through crudely flat laminated sandstones into thicker laminations of darker unlaminated fine sandstone. [Plate 15.] Bases of beds are sharp, or sharply gradational. Only rarely is the ripple form preserved at the top of the bed, and these may be sinuous crested or lingoid in plan. Ripple orientations show a mean towards the north-west [Fig.20].

Bioturbation is very common and often all traces of the original lamination are destroyed. Pelcypodichnus (Seilacher 1953a) is ubiquitous and single beds continuously burrowed have been traced laterally for over 100m. Often this trace fossil appears at the base of the bed and disappears somewhere in the middle, new forms appearing higher up above a carbonaceous lamination [Plate 15]. Most have a prominent preferred orientation parallel to the orientation of the ripple lamination. Cochlichnus (Hitchcock 1853), 'simple vertical pipes' (c.f. Chisolm 1968) and small cylinders (Chapter 8) also occur, but these are all rare.

Interpretation

Small scale cross lamination is produced by the migration of ripples. These occur in the lower flow regime (Simons et.al. 1961), and cosets form when there is net sedimentation. Further discussion of this lithofacies will be given in section 5.3.

5.2.2. Lithofacies 8. Micaceous Carbonaceous Sandstone.

Dark grey, poorly sorted fine sandstone with abundant carbonaceous material and large mica flakes. The sediment is usually unlaminated but breaks along sub-horizontal partings caused by the alignment of plant material and mica. Occasionally thin quartz rich streaks and isolated ripples occur.

Interpretation

In spite of the relatively coarse grain size, the general lack of lamination suggests that deposition was largely from suspension. Periods of traction movement produced isolated patches of ripple lamination.

5.2.3. Lithofacies 9. Parallel-laminated sandstone.

Fine to medium grained sandstones in erosively based beds up to 0.35m in thickness. These contain thin, flat, well sorted laminations marked by mica rich layers. This lithofacies is rare in Association B, it is always succeeded by ripple laminated or trough cross laminated sandstone.

Interpretation

The presence of discreet well sorted laminations, suggests that deposition took place in the lower part of the upper flow regime, (Simons and Richardson 1961).

5.2.4. Lithofacies 10. Trough Cross-bedded Sandstone

Medium grained sandstone with trough sets 4 - 20cm in thickness forming cosets up to 35cm. The orientation of the troughs is similar to the orientation of the ripple lamination. [Fig.20] This lithofacies is rare in Association B, it is always seen in small exposures and whether the sets are filling small channels or forming wider sheets is unknown.

Interpretation

Trough cross bedding is the deposit of sinuous crested bedforms called dunes. (Allen 1963b, Harms et.al. 1975). These form in the upper part of the largescale ripple field of Simons and Richardson, (Harms et.al. op.cit.). Turbulent eddies in the lee of the dunes cut scoop shaped hollows which are then filled by the avalanche foresets of the advancing bedform, so that sections through the YZ plane show intersecting trough shaped erosion surfaces (Allen 1963b).

5.2.5. Lithofacies 5. Thick Sandstones

In Association B this lithofacies is restricted to a large channel at

the northern end of the Roaches. (SJ 997 636). The edges of the channel are not exposed but the coarse grained channel fill forms a strong feature which weathers out against the finer grained background sediments. It is markedly assymmetrical in cross section [Plate 49].

The channel fill is approximately 20m thick and 500m wide at outcrop although the true width is probably less than this as the outcrop is not normal to the channel axis. The fill consists entirely of well sorted, structureless medium sand with rare scattered shale flakes. It closely resembles the LF 5 sandstones in Association A, but lines of amalgamation are not present.

Interpretation

This is regarded as a turbidite feeder channel cut into the delta slope. Similar channels have been described by Walker (1966a, b) from the Grindslow Shales, the delta slope of the 'Kinderscout delta'. He argued that the channels were both cut and filled by turbidites and this explanation is accepted here.

Only one channel has been recognised, but at many places the outcrop of Association B in the Roaches Grit Group is covered by soliflucted material derived from coarse grained sediment in Association C. This mantle of recent sediment has probably obscured other channels.

5.2.6. Lithofacies 2. Silty mudstone

This lithofacies is rare in Association B. It is usually very micaceous and often shows thin quartz rich laminations. It was deposited from suspension in quiet water.

5.2.7. Lithofacies 6. Parallel-sided sandstones

This is similar to that described from Association A except that traces of ripple lamination are occasionally present. It occurs in the lower parts of the slope association. Its origin is discussed in section 5.4.

5.3. Discussion

The thickness of the Slope Association suggests that the base of the slope was very approximately 100m below the water level in the basin. LF 7, Ripple laminated Sandstone, usually in thin parallel sided beds, occurs throughout the slope sequence. Trough cross bedding and upper phase plane beds occur rarely. A typical detailed log is shown in Fig.19a. Intensive bioturbation often destroys most of the original structures in the sandstones. Palaeocurrents show a pronounced mean towards the northwest [Fig.20], parallel to the palaeocurrent mean in the turbidites and normal to the presumed orientation of the delta slope.

Wave activity is unlikely to be felt at water depths of 100m, and the lithofacies do not have the characteristics of tidal sediments. The orientation of directional structures suggests that gravitationally propelled density currents were responsible. McCabe (1975) has suggested that density currents operated during the formation of the delta slope in the Kinderscout Grit Group, though the resulting deposits (his Facies 12) have prod and flute marked bases and are often structureless.

Density currents occur where the sediment laden river water has a greater density than the receiving basin and so flows down the basin slope (Bates 1953). This usually occurs where a river is flowing into a fresh-water lake or reservoir, and present day examples have been described from Lake Mead (Gould 1960), Lake Geneva (Forel 1887, Shepard and Dill 1966, Houbolt and Jonker 1968) and the Walensee (Lambert, Kells and Marshall 1976). The presence of the trace fossil Pelecypodichnus the resting trace of 'non-marine' bivalves (Hardy 1970, Eager 1971, 1974) [see Chapter 8] provides independent evidence that the basin water was fairly fresh.

In the Walensee (Lambert et.al. 1976) and in Lake Mead (Gould 1960) currents of up to 30cm/sec have been recorded from the upper part of the slope. In both cases current velocities diminish downslope particularly on the flat floor of the basin. These velocities are capable of forming

ripples in medium sand (data from Harms et.al. 1975) but are too low for dunes, and for upper phase plane beds, unless the flow was exceptionally shallow. Shepard and Dill (1966), however, have recorded bedforms up to 50cm in height moving down the upper part of the slope of the Rhone Delta in Lake Geneva. The predominance of ripple lamination in the Slope Association suggests that current velocities were similar to those in present day density underflows.

In all the modern examples the density currents have cut channels on the delta slope. These are 200m in width in the Rhone delta, and range from 100m - 300m wide by 2m - 6m deep in the Walensee. Channels have not been recognised in the Slope Association except for the one filled with turbidites. Although this may be a result of the limited exposure, the laterally extensive parallel sided sandstones suggest that the currents spread out widely. Houbolt and Jonker (1968) have described in detail the sediments of the Rhone delta slope and submarine fan. Thin graded beds, sometimes with Bouma (1962) sequences are common, but cross lamination is rare. The sediments are therefore quite different from those described here. In the Brushy Canyon Formation, another alleged ancient density current deposit, (Harms 1974), typical graded turbidites are also rare.

Table IV summarises the differences between classical turbidites, such as LF 4 from the Deep Water Association, and the density current deposits of the delta slope. Turbidites have erosive bases often with current formed bottom structures, and a vertical sequence of structures the Bouma (1962) sequence indicative of a progressively waning flow regime, (Harms and Fahnestock 1965, Walker 1965). Although the deposition of a natural turbidite has never been observed it is probably a relatively short lived event brought about when bottom friction slows the current. This most commonly occurs at the base of a slope when the gradient of the basin floor is reduced. River generated density currents are closely related to periods of increased discharge and suspended load during floods (Gould

1960, Lambert et.al. 1976). In Lake Mead, density currents take several days to cross the lake (Gould 1960). In the Walensee, Lambert et. al. (1976) continuously monitored one flow for 192 hours and fluctuations in strength and direction are common.

Although river generated density currents have a large suspended load they are also powerful enough to form ripples, and sometimes dunes, as Shepard and Dill (1966) have observed on the delta slope of Lake Geneva. With net deposition during prolonged periods of flow, superimposed sets of ripple and cross lamination are the likely deposits. Within a flood event there is a period of build up and decline and discharge may fluctuate, so that the deposits of a single flood may show variable sequences.

Most of the features seen in LF 7, 8 and 9 in the Slope Association seem consistent with a density current origin. Parallel sided beds of rippled sand probably formed during individual flood events which transported sand down the delta slope. Fluctuations in current strength formed silty laminations. During low discharges the delta slope received sediment largely from suspension, and was colonised by bivalves. During floods these moved up through the rippled sand to form Pelecypodichnus escape structures.

Hardy (1970) described Pelecypodichnus from what is probably the delta slope of the Bullion Mine Rock Sandstone (Westphalian) of Lancashire. This too contains parallel sided beds of sandstone with escape structures but he did not describe the sediments in any detail. From a careful measurement of the sizes of Pelecypodichnus at different horizons he was able to show that the bivalves represented different populations related to separate periods of spat fall. These he believed occurred in different years during periods of low stage.

5.4. Origin of Lithofacies 6.

In Chapter 4 it was suggested that LF 6 parallel sided sandstones were deposited from suspension, but the reason for these periodic concentrations of coarser sediment was unknown. In the Slope Association ripple laminated sandstones, believed to be deposited by density currents extend right to the base of the slope, but are never found within the Deep Water Association.

In the Walensee density currents decline in strength on the basin floor where they range from 3 - 12cm/sec. These velocities are too slow for ripple formation in fine to medium sand (data of Harms et.al 1975). In Lake Mead currents on the basin floor often become too slow to measure.

Density currents slowing down and spreading out at the base of the slope are believed to be responsible for depositing the beds of LF 6. Current velocities were here too low for traction movement of the sediment, most of which was deposited from suspension.

Table IV

Comparison of the major features of 'classical' turbidites and density current deposits

TURBIDITES	DENSITY CURRENT DEPOSITS
(1) Beds are laterally extensive with parallel bases and tops, interbedded with finer sediment of mud or silt grade.	Similar
(2) Always sharp based.	May be sharp or gradationally based.
(3) Amalgamation may occur.	Amalgamation does not occur.
(4) Flute and tool marks common.	Flute and tool marks are absent.
(5) Beds show Bouma (1962) sequences.	Bouma sequences are usually absent.
(6) Upwards decrease in grain size.	Grain size is variable through the bed.
(7) Ripple form normally preserved at the top of bed.	Ripple form occasionally preserved.
(8) Superimposed sets of ripple cross lamination rare.	Superimposed sets of ripple cross lamination common.
(9) Bioturbation through the bed rare.	Bioturbation common.

Chapter 6.

Association C. Upper Slope and Delta Top.

6.1. Introduction

This association contains two distinct sediment assemblages. Large fluvial channels dominate the delta top sequence, and these constitute the greater part of the exposure. Below and occasionally between these occurs a distinctive assemblage containing sharp based lenticularly bedded sandstones and sometimes units of inclined bedding.

6.2. Assemblage A. Sub and inter channel deposits

6.2.1. Lithofacies description and interpretation

This assemblage contains 7 lithofacies. Four detailed sections are shown in Fig.21.

6.2.1.1. Lithofacies 12. Lenticularly bedded sandstone

Fine to coarse grained sandstones occur in beds of 1cm to 1m in thickness. The bases are sharp and usually flat, but may contain rare grooves, flutes and shallow scours. The beds are often markedly lenticular, usually because the top of the bed is convex upwards with a well developed crest near the middle [Fig.22]. Many of even the thickest beds thin out and disappear over distances of 20 - 30m, though most exposures are too small to trace the beds very far [Plates 16 and 17].

Two lithotypes occur. Firstly a massive or structureless division, moderately sorted with scattered pebbles up to $\frac{1}{2}$ cm., and sometimes small flat lying tree stems. Secondly a parallel laminated division [Plate 14] where the sediment is fairly well sorted into discreet laminations 1mm - 1cm in thickness, marked by concentrations of very fine plant material. The lamination is parallel to the top of the bed, and so may be sometimes markedly convex upwards where the bed top is similarly shaped. Some beds contain only the massive division and here there is a poorly developed distribution grading in the top few centimeters of the bed. Others are

wholly plane laminated and some contain both divisions, the massive division always being the lower one. Where plane laminations are developed there is usually a fairly good distribution grading through the bed.

Individual beds are usually distinct, even when they rest one above the other. Sometimes, amalgamation occurs. This takes a form similar to that in LF 4, the join being marked by a shale flake conglomerate.

Interpretation

This lithofacies resembles LF 4 turbidites with the erosive bases, Bouma (1962) sequences, grading and amalgamation. The principal difference is the common lenticularity and lateral disappearance. The structures and lithotypes suggest fairly rapid deposition from some kind of erosive turbulent, but laterally restricted current. Further discussion will be given in Section 6.2.3.

6.2.1.2. Lithofacies 13. Channel fill structureless sandstone.

Medium to coarse grained sandstones in units 1m to 10m thick. The base is always an erosion surface with a relief varying from 3m to 1m. [Plate 18]. The channel sides are commonly stepped with parallel orientated flutes [Plate 19]. The channel fill consists almost entirely of a fairly well sorted structureless sandstone with scattered pebbles up to $\frac{1}{2}$ cm. Occasionally thin beds of plane laminated sandstone and thin cosets of trough cross bedding occur but these are always cut out from above by further beds of structureless sandstone. Rows of shale flakes sometimes occur within the beds.

In the Dane Valley (SS 9683 6609 - 9730 6623) a bed is at least 520m wide with neither edge exposed and is probably a wide channel. The fill is 10m thick. Elsewhere exposure is too poor to estimate channel widths, but the thickness of the channel fills are considerably less and it seems likely that these are not so wide.

Interpretation

This lithofacies occurs in close association with LF 12 lenticularly bedded sandstone. It closely resembles LF 5, turbidites with the predominantly structureless sediment, evidence of amalgamation and tendency to fill channels. For LF 5 it was argued that amalgamation of a series of separate turbidity current flows formed thick structureless sandstones. Sometimes these cut and filled channels. A similar process has probably operated here. Currents, probably of a similar origin to those forming LF 12, have cut channels and filled them with thick structureless sandstones. Further discussion of this lithofacies will be given in section 6.2.3.

6.2.1.3. Lithofacies 2. Silty Mudstone

This constitutes the finest grained background sediment in the assemblage. It commonly contains thin laminations of silt and sometimes even fine sand. It was deposited from suspension in quiet water.

6.2.1.4. Lithofacies 7. Ripple-laminated Sandstone

This is the same as that described from Association B. It forms beds up to 10m, often with interbedded finer sediments. Pelecypodichnus is very common [Plate 20] and inclined bivalve resting traces and knob shaped burrows may also occur. Often it is interbedded with LF 12 lenticularly-bedded sandstone, on which it rests with slight erosion. At Hogshaw Brook (SK 0546 7491), and Folly Mill (SS 9709 6641), where the tops of two beds of LF 12 are strongly convex in shape the overlying rippled sandstone forms small lateral accretion units about 0.5 thick and fills in the irregular topography. [Fig.22, Plates 18 and 21].

6.2.1.5. Lithofacies 8. Micaceous, Carbonaceous Sandstone

This lithofacies is interbedded with the coarser sediments. At Chapel Station Quarry, (SK 0558 7932) and Dane Valley, (SS 9706 6621), it fills two small channels, 0.5m and 0.7m deep, cut through beds of LF 12.

6.2.1.6. Lithofacies 9. Parallel-laminated sandstone

This lithofacies is rare. At Hogshaw Brook, (SK 0546 7491), beds

of Parallel-laminated Sandstone alternate with beds of LF 7, Ripple-laminated Sandstone which cut into it erosively. The tops of the ripples are truncated by the succeeding bed of plane lamination.

6.2.1.7. Lithofacies 10. Trough Cross-bedded Sandstone

Sets of trough cross-bedding, 0.08m - 0.35m in thickness usually isolated, or forming thin cosets, interbedded with LF 12 and LF 13. At Folly Mill, some sets fill in concave depressions formed by the irregular tops of LF 12 [Fig.22].

6.2.2. Inclined bedding

Most of the bedded units in the assemblage lie horizontally as far as is possible to tell in small stream exposures. In the Dane Valley, however, two units of inclined bedding occur.

The first of these, (SS 9663 6548) is 8m thick and the base is unexposed. The dip is about 30° to the south east, in a roughly reverse direction to the inferred regional palaeoslope. The sediment is a coarse grained, laminated sand with traces of ripple lamination and Pelcypodichnus. The upper part is cut into by a channel with an irregular base, [Plate 22] filled with structureless coarse sandstone, possibly LF 16, and then by two sets of LF 14 Medium Scale Cross Bedding. The erosion surfaces at the base of the cross bedding dip parallel to the local tectonic dip and the channel fill appears to be undisturbed. The relationship suggests that the sediment in the inclined units was tilted prior to being cut into by the channel.

Regional mapping suggests that a much larger unit of inclined bedding occurs in the Dane Valley between Bartomley Bottoms (SJ 9684 6605) and Folly Mill (SS 9709 6641). This section is shown in Fig.21a and is the best exposed in the assemblage. The top of the section is marked by a channel of Assemblage B filled with a thick coset of coarse grained LF 14 Medium Scale Cross-bedding. Mapping into a tributary valley to the east shows that the base of this channel gradually cuts out the entire section and at SJ 9744 6622 the base of the channel may be seen resting on fine grained

sediments which occur below the section [Fig.23]. Altogether about 70m of sediment has been removed.

It is unlikely that this is a straight forward erosion surface beneath a large channel as nowhere are channel fills of this thickness seen. Dips in the section cut out are steeper than the regional tectonic dip by $5 - 10^\circ$, the inclined unit dipping northwards. Again there appears to have been a large scale movement on the upper part of the slope prior to the incoming of a channel. The moved block is at least 600m in width.

6.2.3. Discussion

The upper part of the delta slope is probably the least understood environment in the major delta fills of the Central Pennines Namurian. In particular the generation of the turbidity currents which cut channels through the lower slope has still to be satisfactorily explained. Often this part of the sequence is cut out by deep distributary channels in the delta top.

In large modern deltas, such as the Mississippi, rapid deposition of sand at the distributary mouth causes oversteepening of the front of the mouth bar. This is relieved by periods translational and rotational, shear slumping and faulting, displaced blocks reaching 6km in width (Coleman, Suhayda, Whelan and Wright 1974). Cores through faulted blocks often show lamination dipping by up to 30° in a landward direction (Coleman, Ferm and Saxena 1972).

The units of inclined bedding in the assemblage are believed to have a similar origin. In neither case is the actual fault exposed, but the angle and direction of dip in the first unit described is consistent with a shear fault origin. The second unit is unusual in that the dip appears to be down the delta slope. In the Niger delta, however, Merki (1972) has described complex tectonics in growth faults associated with clay plugs. Here bedding units dip both towards and away from the major delta slope and are several kilometers across.

Inclined units should be common at the top of the slope, but often they will be eroded by the prograding distributary channels. J.I. Chisolm (p.comm) has described large 'growth faults' from the delta slope sequence below the Ashover Grit. McCabe (1975, 1976), has described inclined units up to 12m thick from the Slope Association of the Kinderscout Grit Group. Here, the faults are exposed in section. He originally interpreted them as slope gullies, but now believes them to be tensional grabens produced by lateral flowage of sediment, of the type described by Coleman et.al. (1974) from the Mississippi delta slope.

Origin of LF 12 and LF 13

Four detailed sections in the assemblage are shown in Fig.21. LF 12 Lenticularly-bedded Sandstone and LF 13 Channel Fill Structureless Sandstone are the most common sediments. In section A, there is a pronounced coarsening upwards sequence, from LF 12 up to LF 13, between the 40m and 64m levels. Apart from this the sequences appear random.

Collinson (1967), described a facies similar to LF 12, (his facies 13 sharp based clean sandstone) from the Delta Top Association of the Kinderscout Grit Group. He interpreted it as a crevasse splay sediment occurring within an interdistributary bay. Stanley (1968) has also interpreted turbidite-like sandstones occurring within a fluvial flood basin sequence as crevasse splays. Leeder, (1973), offers a similar explanation for thin graded beds also a flood basin sequence.

Critical in the interpretation of LF 12 and 13 is the environment of deposition of the assemblage. Although a crevasse splay origin cannot be totally excluded, the following lines of evidence suggest deposition on the upper part of the delta slope, rather than in an interdistributary bay.

- (i) Three of the sections (B, C and D in Fig.21), occur immediately above the slope sequence of Association B and below the first large fluvial channel in the succession.

- (ii) The thickest sequence (A), occurs above a fluvial channel, (see section 5 in Fig.18), but here the section has probably been disturbed by syn-sedimentary faulting. It is 75m thick and appears to be too thick for an interdistributary bay sequence, as modern bays are only a few meters deep at most (Shepard 1960, Saxena 1976).
- (iii) Palaeocurrents from the assemblage are shown in Fig.24. Bottom structures show little preferred orientation, but associated ripple and trough cross-stratification shows a pronounced mean towards the north-west down the regional palaeoslope.
- (iv) Root beds are completely absent, but modern delta interdistributary bays are areas of active vegetation growth once they have been filled by crevasse splay sediments, (Saxena 1976).
- (v) The sequence in section A has probably been displaced prior to the incoming of the overlying fluvial channel. This is most likely to occur at the top of the slope.

With this explanation, the total slope sequence in section 5 of Fig.18 is over 200m thick, approximately twice as great as in other sections. This increased thickness can be accommodated where there is syn-sedimentary faulting. Merki, (1972), has shown that sequences on the down throw sides of growth faults may increase in thickness by up to $2\frac{1}{2}$ times.

Because of the rapid dumping of sediment at distributary mouths and oversteeping of the bar front, the upper part of the slope is an area of considerable sediment instability. Sometimes, large blocks of sediment will be displaced, but within these, the original characteristics of the sediment will be preserved, as in the two examples described. Where the sediment is a fairly well sorted coarse silt shear failure may be closely followed by liquefaction (Lowe 1976b). The sediment will then move down-slope as a liquified flow.

This process requires a slope of between 22° - 27° for normally consolidated sand. Estimating the angle of the delta slope is difficult. The modern Mississippi delta has a slope of only 1° (Coleman et.al. 1974) and it is likely that the slope in the present sequence was also comparably low. Towards the mouth bar crest, however, slope angles steepen, and it is possible that slumping will occur there.

A much more likely method of liquefaction is the rapid dumping of coarser sediment onto clay. With rapid consolidation of the clay, water is expelled into the overlying sand which liquifies and fails, (Moore 1961, Shepard and Dill 1966, Lowe 1975, 1976b). This process seems quite likely to have happened at the mouth bar crest where sand is deposited rapidly on the clays of the delta slope. It requires no slope, but if the liquified mass is to propagate down-slope, this is unlikely to happen if the angle is less than $3 - 4^{\circ}$ (Lowe 1976b). If the liquified flow becomes sufficiently dilute, it may turn into a turbulent density current, (Middleton and Hampton 1973).

This process, outlined in Fig.25 is suggested as a possible origin of LF 12 and LF 13. Turbidity currents generated from liquified flows formed at the mouth bar crest move down the upper slope sometimes cutting channels. Some of these channels on the upper slope probably connected with the large turbidite filled channels in the lower slope Fig.25 as LF 13 and LF 5 are virtually indistinguishable. Through these channels, the sediment was transported to the basin floor to form the turbidites of the Deep Water Association. Where the flows did not become channelised, they deposited sediments on the upper part of the slope to form LF 12 Lenticularly-bedded Sandstone. The ripple-laminated and trough cross-bedded sandstones in the assemblage were presumably deposited by density currents flowing continuously out of the river mouths, in a similar way to the rippled sands in Association B.

Two principal arguments may be laid against this theory. Firstly there is no evidence of sediment build up at the distributary mouth, and the deposits of these postulated liquified flows have never been observed. Secondly, there is only a limited amount of time and space for the liquified flows to dilute and become turbidity currents. The upper part of the slope, however, is unlikely to be preserved because of erosion by the prograding distributary. Only after abandonment is there any likelihood of seeing this environment. The second objection is more difficult to explain at present.

This theory of turbidity current genesis differs from that proposed by Collinson, (1967, 197) for the Kinderscout Grit Group. He suggested that fluvial distributary channels passed directly into turbidite filled channels on the delta slope, although this transition has never been actually observed. The principal argument against Collinson's model is the difficulty of having river generated turbidity currents form at the same time as the river generated density currents postulated to have formed the rippled sands on the delta slope. Possibly turbidity currents were only generated for a short period during peak discharges when sediment transport through the fluvial channels would be greatest.

6.3. Assemblage B. Fluvial Channel Sediments

6.3.1. Lithofacies description and interpretation

6.3.1.1. Lithofacies 14. Medium Scale Cross-bedding (MSXB)

General Description

Tabular cross-bedded coarse pebbly sandstones in sets up to 3m thick, [Plates 23 and 24] forming cosets up to 35m, but mostly 10-20m. Units of this lithofacies may be traced continuously exposed for up to 1km parallel to the mean foreset dip direction and up to 300m normal to it. It has been mapped for distances of up to 3km and occurs therefore as sheets of wide lateral extent.

Foresets may be angular, [Fig.27, A], but are commonly associated with small sets of trough cross bedding 5-15cm in thickness. These may occur as intrasets interfingering with tangential foresets [Fig.27, Band C, Plate 25], or as bottomsets lying in front of angular foresets, [Fig.27,D, Plates 26 and 27]. In the latter cases, the bottomsets extend for up to one metre in front of the main foresets forming cosets which climb towards the foresets. There is a sharp angular contact with the base of the foresets.

The foreset laminations, 1½-7cm thick are generally well sorted and weather out very distinctly, [Plate 28]. Most are normally graded, but reverse grading also occurs. They become coarser towards the bottom of the set. Most dip at between 25 - 30°.

The erosion surfaces at the bases of major sets are normally flat, but shallow troughs occasionally occur. Commonly they are overlain by a layer of small pebbles or a thin bed of coarse structureless sand [Plate 23]. The bases are usually very near depositional horizontal in attitude, and individual sets extend for tens, and at times hundreds of meters before wedging out. At one locality (The Roaches SK 0052 6246) the erosion surfaces gradually increase in dip down the XY plane over a distance of 200m becoming inclined by about 10° with respect to the original horizontal.

Cross set orientation

This lithofacies has proved to be the most useful palaeocurrent indicator. For each reading the dip and strike of the avalanche foresets has been measured, together with the dip of the basal erosion surface, the latter being used to determine the original horizontal. Because many exposures are two dimensional, only about a half of the sets have been measured. No special sampling procedure has been used. Foreset dip directions have been re-orientated, and vector means and variance calculated, following the methods outlined in Chapter 1. The orientation of the small toeset cross bedding with respect to the main foreset was also measured.

Palaeocurrents for different areas are shown in Fig.28 and the total in Fig.29. In most cases the current roses include readings from more than one coset, and therefore from more than one channel. It has proved impossible to take measurements from individual channels. Thick cosets probably represent the fills of two or more separate channels stacked one upon the other, but because the erosion surfaces marking channel bases are no different from those at the bases of individual sets, separate channel fills cannot be recognised.

In spite of this, most of the local current roses show a unimodal distribution with a fairly low variance (Table V). Within an individual coset the variance is less. Typically a group of 3 or 4 consecutive sets have a virtually identical orientation, successive groups differing by 20 - 40 degrees (Table VI).

The bottomsets have a wide range of orientation with respect to the main foreset. Most dip in a roughly reverse direction but some dip roughly the same way. The orientation of bottomset strike with respect to foreset strike is shown in Fig.30C. Some bottomsets dip almost normal to the main foresets.

Internal Erosion Surfaces

Inclined erosion surfaces separate adjacent groups of foreset lamination, truncating the lower set. They occur within individual sets, and (using Allen's 1963c definition of set), are therefore termed internal. They resemble the 'Diastems' of Boersma (1968) and 'Reactivation Surfaces' of Collinson (1970). In some cases however, the detailed geometry, and; probably also the mode of origin is different from the interpretations of the above authors, and for this reason genetic terms are rejected.

Internal erosion surfaces are common in this lithofacies and also in LF 15. A descriptive terminology is shown in Fig 31. Most are distinctly concave or convex upwards in shape although some may be virtually straight.

(a) Convex Upwards Forms

When fully preserved, these gradually increase in inclination down the XY plane from a near horizontal top (Plate 29). Some have a height equal to the full set thickness ending at the basal erosion surface of the set, whereas others disappear within the set when their inclination reaches the angle of foreset dip. The maximum height recorded is 1.5m and lengths are typically 1 - 6m. The shorter surfaces have inclinations of 25 - 30°, whereas the longer surfaces, which usually extend throughout the full thickness of the set, have inclinations as low as 8°. The strikes of the foresets and of the associated erosion surfaces are usually identical within the limits of measurement. No surfaces have been traced along strike for more than a few meters and all are planar over these distances.

Compared to the main sets the overlying foreset laminations tend to be thinner and always have an angular base. Usually the foresets are straight in the XZ plane, [Plate 30] as in the main sets although one curved, vaguely trough shaped set is known. The orientation of the foresets is not significantly different from those in the associated main sets.

The erosion surfaces always occur in groups and are closely spaced so that they resemble those in the compound cross stratification of Harms, Southard, Spearing and Walker (1975). The maximum spacing at the bases is rarely greater than 30cm decreasing towards the top, where the surfaces often join. No topset occurs, unless there is a down dip passage into LF 15. Frequently the join is not seen because of erosion at the base of the overlying set. Groups of inclined erosion surfaces may be traced for distances of up to 30m in the XY plane before being cut out by adjacent sets or no longer exposed. In most groups the surfaces all have a remarkably similar geometry and spacing so that they are all essentially parallel to each other. Successive foresets all have a similar orientation.

Three exceptions to this normal regular pattern are known. At the northern end of the Roaches (SJ 9999 6410) [Fig.32, Plate 31] the inclination of the erosion surfaces varies from about 8 to 20°. Groups of more steeply dipping surfaces are truncated by surfaces of low inclination, which in turn form the bases of another series of steep surfaces. All seem to have a similar strike although exposure along strike is poor.

At the Roaches (SK 0003 6380) [Fig.33] avalanche foresets can be traced for 11m and then pass abruptly down the XY plane into foresets with inclined erosion surfaces. The first to appear have lengths of about 1m and inclinations of 25°. They merge into the foresets in the middle part of the set. Further down the XY plane they gradually increase in length and decrease in inclination until eventually lengths of about 5m and inclinations of only 8° are attained. These extend for the full set thickness. After 17.6m they disappear.

A similar series of structures occurs at Hen Cloud. Altogether 16 groups of multiple convex upwards erosion surfaces have been seen in LF 14. What proportion of the sets originally contained these structures is unknown, as all of them have had their upper parts eroded away prior to the deposition of the next set.

Concave upwards forms

These are rarer than the convex type. They occur in groups of 2 or 3 up to 15, and are generally spaced about 1m to 1.5m apart. Three different types occur. [Fig.34]

Type a

Here there is an abrupt passage from tangential foresets to angular foresets. The latter hardly truncate the lower set. They usually become more tangential down the XY plane. [Plate 32].

Type b

This truncates the lower set, and the overlying foresets are angular. In rare cases the erosion surface is overlain by a coset of small scale trough cross bedding orientated obliquely to the foreset strike. [Fig.34-B, Plate 33]. The example is part of an intermittently exposed series of at least 10 inclined erosion surfaces which extend for a distance of 100m in the XY plane. Most do not have the smaller trough sets.

Type c

This truncates the lower set and decreases in inclination in the X direction until it reaches horizontal, after which it extends parallel to the main set bounding erosion surfaces.

It has not been possible to trace these surfaces for more than a few meters along strike, but they appear to run parallel to the strike of the foresets.

Disturbed bedding

Occasionally the foreset laminae are disturbed into a series of tight antiformal folds with corrugated limbs [Plates 34, 35]. The folds are circular in plan and asymmetrical in section, pointing down the foreset dip. In places the axes are nearly horizontal and sometimes slightly faulted. In some cases nearly all traces of the original lamination have disappeared.

At the Roaches (SK 006 622) the disturbances extend through two sets for 1.7m appearing abruptly 1m above the base of the lower set, and reaching

a maximum intensity towards the top. The second set shows progressively less deformation upwards, and the disturbed bedding is truncated by the erosion surface at the base of the overlying set. This horizon can be traced for 400m to the limits of exposure.

Interpretation

Cosets of cross bedding with flat based erosion surfaces have been interpreted as the deposits formed by the migration of bedforms with large length to height ratios, usually called sand waves (Harms et.al. 1975). In flume experiments sand waves form in medium to coarse sand in the lower part of the large scale ripple field of Simons et.al. (1961), at lower flow powers than the smaller, more sinuous crested dunes, (Pratt and Smith 1972, Pratt 1973, Costello 1974). In modern rivers these bedforms may be straight crested (Shantzer 1951, fig 28, Jordan 1962, Williams 1971), but more commonly they are lobate or horseshoe shaped. (Rich 1942, Brice 1964, Collinson 1970b, Smith 1970, 1971, 1972, Williams 1971).

With straight crested forms the avalanche face will usually be aligned normally to the flow direction, and there is a simple separation eddy directed in a reverse direction to the main flow. With lobate sandwaves, the crest line is largely oblique to the flow direction. In this case, the flow lines of the separation eddy are orientated obliquely, and a spiral eddy develops, (Nedeco 1959, Allen 1968, Collinson 1970).

The separation eddy may be powerful enough to produce small, regressive bedforms, the direction of movement of which, with respect to the main avalanche face, will depend on the type of eddy. With reverse eddy direction the strike of the intraset cross bedding will be similar to that of the main avalanche foresets, [Fig.30-A]. Most intrasets should fall in the $0 - 15^{\circ}$ class of Fig.A. With a spiral separation eddy, however, where the reversely directed current is orientated at approximately 45° to the dip direction of the main avalanche face, the strike of the intraset cross bedding will be like that in Fig. 30-B with most falling in the

30 - 45° and 45 - 60° classes.

Intrasets are quite common within the sets of MSXB. The orientation of the strikes of these relative to the strike of the associated major foreset shown in Fig.30-C, is more similar to the distribution shown in Fig.30-B, expected with a skewed crestline. Thus the sets of MSXB are tentatively interpreted as the product of lobate rather than straight crested beforms.

The shape of the foresets probably depends on the position of the reattachment point relative to the avalanche face. Where the reattachment point is just in front of the base of the slip face the intrasets and foresets will interfinger as in Fig.27-C and Plate 25 the intrasets being directed in a roughly reverse direction. Where the reattachment point is actually on the lower part of the avalanche foresets, there will again be an interfingering of foreset and intraset, the latter being directed in roughly the same direction, Fig.27-B. Where the reattachment point is some distance in front of the main foresets a pattern as in Fig. 27-D and Plate 27 will be produced.

The cross bedding orientation variability in LF 14 is quite low (vector magnitude 73.3% variance 2025), even though these readings were taken from several different channel fills. Miall (1974, 1976), has found a similar low variance in the orientation of tabular cross bedding of fluvial origin. On the other hand Williams, (1966) High and Picard (1974) and Cant and Walker (1976) all found a relatively high variance within individual channel fills, much greater than that of associated trough cross bedding.

The reason for these marked differences in variance is not clear. Smith (1972) measured foreset dip directions in lobate sandwaves in the River Platte and found a considerable dispersion, (vector magnitude 42.4% variance 5,073). Because the sandwaves have a considerable sideways component of movement, particularly during falling stage, he argued that

ancient sequences with tabular cross bedding formed by lobate sandwaves should similarly have a wide dispersion. With straight crested forms the variance should be much lower.

This argument appears to conflict with the present data, where there is a relatively low variance, even though the bedforms responsible are believed to have had skewed crestlines. Banks and Collinson (1974) have argued that in regimes where most sandwave movement takes place during high stage, there is little lateral movement. With forwards movement, cross bedding formed at the sides of the bedform will be volumetrically much smaller than that formed at the front. Thus the variance should be much lower than that predicted by Smith.

The present results seem to support Banks' and Collinson's argument. Here at least, probable lobate shaped sandwaves produced cross bedding with a low variance. The much higher variances noted by other workers suggests that sideways movement was considerable and that a difference of regime might be implied.

Brice (1964), Smith (1970) and Culbertson and Scott (1970), all suggest that sandwaves develop from a coalescing field of dunes. If this is the case, then tabular cross-bedding should pass upcurrent into smaller trough sets, as observed by Smith (1970). This relationship is never found in the Roaches Grit. Cosets of tabular cross bedding are surprisingly regular and monotonous, with hardly any interspersed trough cross bedding.

Internal erosion surfaces

Multiple convex upwards internal erosion surfaces have been described from aeolian dunes (Reiche 1938, McKee 1966, Bigarella 1965, Thompson 1968); deltas, (Howard 1966); tidal sand waves (Johnson 1975) and recent and ancient fluvial sandwaves (Williams 1971, McGowan and Groat 1971, Jackson 1976a). Mayhew (1967), Collinson (1967) and McCabe (1975) described similar structures from tabular cross bedded sandstones very similar to LF 14 MSXB, in the Namurian, Ashover and Kinderscout Grits of

the Pennines. Banks (1973) described downcurrent dipping crosssets from fluvial sediments in the Pre-Cambrian of Finnmark, Norway. Although these have a maximum inclination of only 8° , and are essentially straight in profile, the presence of similar erosion surfaces in the Roaches Grit in close association with the more common steeper convex forms, suggests that the two structures may be genetically related.

McGowan and Groat (1971) suggested that each separate inclined erosion surface formed as a result of a sudden increase in water depth. A new sandwave forms upcurrent from the position of the avalanche face of the sandwave prior to the increase in water depth. After a slight water rise this moves downstream eventually overtaking and burying the old sand wave avalanche face. A series of several inclined erosion surfaces therefore result from a series of small but sudden increases in water depth.

Banks (1972) suggested that low angle downcurrent dipping cross-sets are formed where smaller bedforms are superimposed on top of a larger one, under conditions of net sedimentation. Each moves up the back and then down the front of the large bedform. With this mechanism, each separate set forms at constant stage and does not require a change of depth.

The formation of multiple convex erosion surfaces has recently been investigated in flume experiments (at Keele University), (McCabe and Jones, in press). An artificial delta 10cm. high was constructed in sand and the flume run for several hours so that an equilibrium assemblage of current ripples formed on the delta top. The rate of erosion of the ripple stoss side increased as the ripple washed out over the delta crest so that the crest was rounded off. A convex upwards erosion surface is formed which is buried by the avalanche sets of the next ripple. [Fig.8]. The migration of a series of ripples produced a train of convex upwards surfaces.

These experiments show, contrary to the ideas of McGowan and Groat that multiple convex erosion surfaces can be produced at constant stage where smaller bedforms are superimposed on the top of a larger one.

The origin of the low angle surfaces is still uncertain. The close association of these with the steeper convex forms suggests that a similar process operated. It is emphasised that the process observed in the flume experiments is quite different from that envisaged by Banks (1972). In Banks' model a series of smaller bedforms occur, one after the other, on the inclined lee side of a larger bedform which does not possess an avalanche face of its own. In the flume experiments the large bedform retains its avalanche face and is still asymmetrical in profile. The mechanism proposed by Banks might operate on a very large scale, for instance where sand waves are moving down the front of larger macroforms such as side bars or large braid bars. It seems unlikely to have operated on the small scale observed here.

Attempts to produce low angle inclined surfaces by reducing the height of the artificial delta and keeping the ripple height constant were not successful. Possibly they are formed near the edges of the major bedform where the superimposed forms move across the obliquely orientated bar edge. Detailed sectioning of modern sand waves is necessary to solve this problem.

As outlined in Chapter 2, there is evidence to suggest that the size of sand waves in rivers is somehow related to the scale of the river discharge. Some cases of superimposition of different sizes of bedform are probably lag effects (e.g. Pretious and Blench, 1951). Most of the evidence in favour of size lag so far produced consists of two dimensional echo profiles of the river floor. These need to be interpreted with care. If, as now seems likely, there are hydrodynamically distinct classes of large scale ripple, dunes and sandwaves (see review in Harms et.al. 1975), then some observed cases of superimposition may not be lag, but the result of dunes moving in equilibrium, on the top of larger sandwaves. Jackson (1975a, 1976a) claims to recognise this situation in the Wabash River and dunes are commonly

superimposed on top of sandwaves in tidal regimes, (Boothroyd and Hubbard 1972, Harms et.al. 1975).

As yet, the effects of different forms of bedform superimposition on sedimentation have not been investigated. Where the superimposed bedforms are dunes, however, the resulting deposit will probably be a coset of trough cross bedding overlying a solitary tabular set. This type of sequence is common in ancient braided fluvial sequences. (Williams 1966, Smith 1970, McGowan and Groat 1971, Cant and Walker 1976).

The multiple convex upwards erosion surfaces in LF 14 are flat and the associated cross sets are essentially a series of descending tabular cross sets. Superimposed sets of trough cross bedding are extremely rare. These multiple convex upwards erosion surfaces are interpreted as the result of the superimposition of smaller sandwaves on the tops of larger ones, when the larger bedforms are no longer in equilibrium with the flow. It seems very unlikely that superimposed sinuous crested dunes would produce this type of structure.

The two unusual sequences shown in Figs.32 and 33 probably represent changes in sandwave size and shape in response to continuously changing flow conditions. In the Fig.32 sequence it seems possible that two scales of bedforms were superimposed the larger one cutting the low angle surfaces, this in turn carrying smaller bedforms which cut the steeper surfaces. Clearly much more detailed observations of sand wave response to changing flow conditions are needed before these complex structures are adequately understood.

Concave Erosion Surfaces

The origin of the concave erosion surfaces is uncertain. The one example where a coset of trough cross bedding covers the erosion surface [Fig.34-A and Plate 33] resembles the Reactivation Surfaces of Collinson, (1970). These are produced by falling stage erosion of and deposition on the bar slip face. [Fig.35-A]. If they do represent bar slip face

modification of the type described by Collinson, it is difficult to explain why they always occur in groups whereas elsewhere foresets lacking erosion surfaces extend for tens and sometimes hundreds of meters. In the Tana, Collinson found that some Reactivation Surfaces formed as a result of fluctuation of discharge within one flood cycle. If it could be proved that the concave erosion surfaces formed in the same way, their distribution may provide evidence of the nature of the river hydrograph.

Unfortunately, it is likely that a number of other processes can produce similar structures. Allen, (1973), investigated the behaviour of ripples in a flume run at constant discharge. He found that although the ripple field remained constant in a statistical sense, there was a continuous creation and destruction of individual ripples. The formation of a new ripple usually associated with a temporarily invigorated eddy resulted in a concave erosion surface which flattened out down the XY plane [Fig.35-B].

Whether this process operates on a much larger scale, and with a different class of bedform is as yet unknown. There have been no detailed observations on the movement of sand waves at high discharges. The types of surface shown in Fig.32-C may have formed in this way similar to the small scale example figured by Allen.

The shape of the foresets vary with fluctuations in the strength of the leeside eddy. The passage from tangential foresets to angular foresets shown in Fig.34-A probably records a decrease in the strength of the eddy [Fig.35-C]. (Jopling 1965). An increase in the strength of the leeside eddy might result in erosion of the previously deposited foresets. Allen (1973), observed this in his flume experiments with ripples with change in discharge. This process could form concave erosion surfaces of type A where the overlying coset of trough cross bedding is absent.

6.3.1.2. Lithofacies 15. Large Scale Cross-bedding (LSXB)

Cross bedded coarse pebbly sandstones in single sets 3 - 20m thick. They reach a maximum observed width of at least 380m at Ramshaw Rocks (SK 02 62), but the southern edge is unexposed, and the northern edge has been eroded. At the Roaches (SK 00 62) a set may be traced for 460m down the XY plane but the downcurrent edge has been eroded. At Hen Cloud (SK 0080 6170) the sets are also of fairly large dimensions but elsewhere exposure is too limited to give any indication of their true extent.

Three types of foresets occur [Fig.38].

(i) Angular foresets, where the bottomsets are less than 1m thick.

Individual laminae are 2 - 10cm thick, well sorted, with very rare scattered pebbles. They fine up very slightly in the top 1cm. They rarely exceed 1m in length, and the foresets consist of a series of impersistent, interlocking laminations, [Plate 36]. Dips are 25 - 30°.

(ii) Tangential foresets - concave upwards cross beds, horizontal near the base, and increasing up dip to a maximum of 30°, [Plate 37]. Individual foresets reach 80cm in thickness but most are 5 - 10cm. Many contain intrasets though they are not always easily seen because of weathering. These are small cross sets, usually 5 - 10cm in thickness, reaching a maximum of 20cm. In sections parallel to the XY plane they occur in thick cosets, with flat parallel bases with pebble lags. In the YZ plane they are usually trough shaped, [Plate 38]. The orientation of the intraset current directions is variable. At Ramshaw Rocks, the mean is almost parallel to the foreset strike, whereas at Hangingstone, (SJ 9739 6540), the mean is approximately down the foreset dip. At Hen Cloud, (SK 0030 6170), the mean is approximately up the foreset dip. Intrasets extend for up to 10m from the base of the main sets and disappear when the dip approaches 30°.

At Hen Cloud, the large foresets become trough shaped near the base, [Fig.37]. These are truncated on the downcurrent side of the trough, by a low angle erosion surface which extends up into the main set. The erosion

surface is succeeded by normally dipping foresets which are parallel to it. Here the foresets do not contain intrasets.

(iii) Concave upwards foresets without intrasets. Here individual foresets are around 50cm thick, poorly sorted and the bedding is indistinct. Downdip they pass gradationally into LF 16 Faintly-laminated Coarse Sandstone.

Inclined erosion surfaces which extend completely, or most of the way through the set and separate two groups of adjacent foresets are very common in this lithofacies. The terminology used for LF 14 will also be adopted here. The main types are shown in Fig.38.

Convex upwards erosion surfaces

These are similar to those described from LF 14 but occur on a much larger scale. Where fully developed they occur where LSXB^s overlain by cosets of LF 14 Medium Scale Cross-bedding. At the Roaches (SK 004 627), over 40 of these erosion surfaces occur within a horizontal distance of 460m is the XY plane [Fig.39]. The first 12 of these extend from top to bottom of the large sets which here reach 5.8m in thickness. Foresets are truncated by the horizontal erosion surface at the base of a coset of MSXB. This surface gradually increases in inclination in the X direction, descending the large foresets until it eventually disappears over a distance of 35m. Further down the XY plane each succeeding erosion surface at the base of the medium scale sets descends the large foresets.

The main difference from much smaller convex surfaces in LF 14, apart from size, is that the cosets of LF 14 form a topset to the large foresets of LF 15. In the section described above, the foresets of the large cross-bedding increase from 5.8m to 10m down the XY plane; but the overlying coset of MSXB does not change appreciably in thickness, new sets appearing successively to take the place of those which merged into the large foresets.

Multiple sets of convex erosion surfaces also occur at Back Forest, (SJ 981 655).

Concave upwards erosion surfaces

These increase in inclination up the XY plane and eventually disappear into the main foresets when they reach the angle of foreset dip, [Fig.38-2a]. Sometimes they may pass up into a convex erosion surface, Fig.38-2c. Down the XY plane they gradually flatten out and pass into flat, horizontal erosion surfaces. Sometimes these surfaces are irregular, and overlain by LF 16 Faintly-laminated Coarse Sandstone, [Fig.38-2b]. This passes up into the large foresets.

The main face of Hen Cloud (SK 0080 6170), Fig.37, shows most of the features described above. One large concave surface, disappearing up into the main foresets, occurs in the middle of the Face [A-A]. Below is an irregular surface overlain by LF 16, [B-B]. Sometimes only the toesets of the cross-bedding are preserved [2 and 3]. The dip of the foresets overlying A-A changes from 300° to 250° down the XY plane. The foresets are truncated by an erosion surface, [C-C] which differs in strike by 50° . The succeeding foresets dip parallel to the erosion surface.

At the main face of the Roaches [Figs.39, 40] (SK 004 627), most of the erosion surfaces are of the convex upwards type. Seven concave upwards surfaces occur towards the northern end. In three cases the overlying foresets pass down into LF 16, FLCS, which rests erosively on the underlying cross-bedding, [Fig.40]. These three erosion surfaces extend for distances of up to 200m, each cutting at a successively higher level into the main set. Two of these pass up dip into convex upwards surfaces and eventually form the base of a typical medium scale set of LF 14.

In most cases the strike directions of the internal erosion surfaces and of the preceding and succeeding foresets do not differ by more than 10° . The second erosion surface from the right in Fig.40, however, cuts obliquely into the preceding foresets and differs in strike by 65° , the succeeding foresets dipping almost parallel to it.

In conclusion, multiple convex upwards erosion surfaces tend to be associated with angular foresets [Fig.39], whereas concave erosion surfaces, tangential foresets, and foresets passing downdip into LF 16 FLCS tend to occur together, [Figs. 37 and 40]. At the Roaches, where a set of LSXB extends out from a channel edge, the sets nearer the channel edge are angular [Fig.39], but farther out they pass down dip into FLCS with associated concave erosion surfaces, [Fig.40].

Palaeocurrents

Cross-set orientations have been measured for each set of LSXB, and for each subset between major concave erosion surfaces. The sharp differences in orientation between some successive subsets have already been noted. The vector mean, however, is very similar to that in LF 14 [Fig.41].

Interpretation

Collinson, (1967, 1969) described large scale cross bedding in sets up to 40m thick from the Kinderscout Grit in Derbyshire. He considered them to be the deposits of 'Gilbert Type' deltas built into standing water following a sharp rise in water level.

McCabe (1975), who worked on the northern Kinderscout Grit, where there is a similar lithofacies, has shown beyond doubt that the large sets lie within channels. He considers a variety of origins are possible; channel confluence deltas, channel infill deltas or large sandwaves. He suggests that the majority, however, were large transverse bars. (This follows the definition of transverse bar in Allen 1968. The term "alternate bar" is now preferred, see Chapter 2).

Baines (p.com.) has described large scale cross-bedding from the Skipton Moor Grit of Yorkshire, also suggesting that they lie within channels.

In other sequences cross bedding of this scale has been interpreted as aeolian, (McKee 1966, Thompson 1968, Gradzinski and Jerzy Kiewicz 1974), as the deposits of large marine sandwaves (Jerzy Kiewicz 1968) and as 'Gilbert Type' deltas, (Gradstein and Van Gelder 1971, Cotter 1975).

In the present sequence the lithofacies is clearly the product of slip face accretion because of the high dip. The coarse grain size discounts an aeolian origin. Further discussion of this lithofacies will be given after its relationship to the other channel infill sediments has been outlined.

6.3.1.3. Lithofacies 16. Faintly Laminated Coarse Sandstone (FLCS)

Coarse grained sandstones, usually with scattered pebbles up to 3cm. Beds reach 10m in thickness and have been traced laterally for up to 675m.

The bases are erosive, usually with considerable relief. [Plate 39]. At Ramshaw Rocks, the section consists of 4 sets of LF 15 Large-scale Cross Bedding in strike section alternating with beds of LF 16 the bases of which cut erosively into the cross bedding. One erosion surface can be traced laterally continuously for 150m. At the southern end of the outcrop it consists of a series of undulations up to 5m in height and 10 - 25m in width with gently dipping sides. Towards the north it becomes extremely irregular, the sides having dips of up to 70° , and a relief of up to 10m, [Fig.42]. Most of the steep sides have approximately the same orientation and are roughly parallel to the dip of the associated cross bedding, but some may be orientated normally [Plate 40].

The lowest parts of the erosion surface are roughly concave in shape and 10 - 40m in width, [Plate 41], some cutting into each other Plate 42. As a result of this extreme irregularity the overlying beds of LF 16 pinch out and locally disappear laterally between succeeding sets of cross bedding. The surfaces are very sharp and make a clean cut into the underlying sand which is usually undisturbed, although at the Roaches (SK 0053 6247) the lamination in the adjacent sand is bent.

The overlying beds often appear completely structureless, particularly in relatively young exposures or where there is intense weathering or jointing which fractures the rock. In favourably weathered

outcrops, however, faint lamination is visible. This shows up because the coarser pebbles have a tendency to be slightly concentrated in thin bands, the pebbles showing a poor alignment; or because of a crude bedding 10 - 50cm in thickness marked by differential weathering.

The lamination is undulatory. The smallest undulations are as little as 5 - 10cm in height and 2m in width; [Fig.43a, Plate 43], many are 50cm - 1m in height and up to 7m in width. [Fig.43b, Plate 44]. At the Roaches near Rockhall (SK 0061 6218) even larger scale structures occur but these are not sufficiently well weathered to determine their true size. Dips on the limbs may reach 30° but many are only $5 - 10^{\circ}$. Usually the lamination is parallel to the erosion surface at the base of the bed. At Ramshaw Rocks (SK 0190 6206), for example, a concave upwards surface 15m in width is overlain by LF 16 with concave upwards lamination parallel to the sides, towards the top it gradually flattens. [Plate 41].

It is difficult to trace the structures for large distances laterally as they frequently disappear in unfavourably weathered outcrops. It is clear, however, that in some cases the undulations change in size laterally and often flatten out, [Fig.43-C]. They usually also flatten out towards the tops of beds, this being accompanied by a crude coarse tail grading. Very little is known about the three dimensional geometry of the structures but the limited evidence suggests they are ridges.

The beds commonly contain internal erosion surfaces. These are similar to those at the base, some are flat or gently undulating [Fig.43d,e,f, Plate 45], others have steeply dipping sides. Again, the overlying lamination is parallel to the erosion surfaces. [Fig.43g, h, Plates 45, 46].

Usually the beds are overlain erosively by LF 14 or LF 15. At several localities along the Roaches however LF 16 in beds 10cm - 1m alternates with LF 10 trough cross bedding in cosets up to 0.7m [Plate 47]. Sometimes the cross sets are overturned.

Interpretation

Collinson (1966, 1967, 1970a) described structureless sandstone, his Facies 8, Massive Beds, from the lower part of the Kinderscout Grit in Derbyshire. He suggested that they were the product of very rapid deposition in the upper flow regime where sorting processes had insufficient time to operate.

McCabe (1975) described a similar facies from the northern Kinderscout Grit. (His F 4 unlaminated sandstone.) He showed that some beds occur in fluvial channels in association with large scale cross bedding, whereas others, slightly lower down in the sequence occur in channels where large scale cross bedding is absent. The latter he suggests are coarse grained turbidites. McCabe did not describe any lamination but X radiographs of the beds lying within fluvial channels showed a concentration of certain grain sizes along planes.

In this sequence LF 16 is regarded as the equivalent of the parts of Collinson's Facies 8 and McCabes Facies 4 which lie within fluvial channels. Favourable weathering shows that the beds are not structureless but contain faint lamination.

McCabe (op.cit) suggested that the unlaminated sandstones within fluvial channels were formed as a result of intense scour and fill at the reattachment point of eddies within the lee of large bedforms represented by the large scale cross bedding. Further discussion of the origin of this lithofacies will be given after the relationship to other channel infill sediments has been outlined in more detail.

6.3.1.4. Lithofacies 16a. Undulatory-bedded Sandstone.

This lithofacies is rare being found only at the base of the main escarpment at Hen Cloud (SK 008 616) [Fig.37]. It is a coarse pebbly sandstone well bedded in units of variable thickness from 5cm to 90cm. The bedding is undulatory, about 6m in wavelength and about 1m in amplitude.

Small intrasets of cross bedding are directed away from the crest, otherwise the beds are completely structureless.

Interpretation

McCabe (1975) described a much better exposed example of a similar facies from channel fill sequences in the Kinderscout Grit Group. He suggested the bedding represented sections through ridges formed at the foot of, and aligned normal to the foresets of large scale cross bedding. These are related to the development of parallel cork-screw eddies in front of avalanche faces orientated oblique to the mean flow direction. (Allen 1968).

The lithofacies described here may have formed in a similar way. Although McCabe did not observe a lateral passage from undulatory bedding into large scale cross bedding, the beginnings of such a relationship are seen in the main large scale set at Hen Cloud [Fig.37] described in section 6.3.1.2. Here the tangential foresets pass into undulatory bedding at the base.

Many examples of much fainter undulatory lamination occur in LF 14, and this may have a similar origin. Further discussion of this will be given in section 6.3.5. after the lithofacies relationship have been described in detail.

6.3.1.5. Lithofacies 7. Ripple laminated Sandstone

This is similar to that described earlier. Ripple drift never occurs and most sets are less than 2cm thick. The sediment is usually very micaceous. Pelcypodichnus is ubiquitous.

6.3.1.6. Lithofacies 10. Trough Cross-bedded Sandstone

Medium and coarse grained sandstones. Within the assemblage this lithofacies occurs in association with LF 14 MSXB and LF 16 FLCS.

In association with LF 14 it forms cosets up to 2.7m. Individual sets are 10 - 20m thick and 1.5 - 4.0m wide. Exposures are too limited to determine its lateral extent. In association with LF 16 it occurs as

isolated sets or thin cosets alternating with beds of structureless sandstone [Plate 47], beneath which the cross bedding may be overturned. Sometimes a coset of trough cross bedding occurs at the top of a bed of LF 16 [Plate 48], and this may be cut out laterally by an erosion surface at the base of another bed of LF 16 [Fig.43d]. At Ramshaw Rocks (SK 0196 6228), small trough sets overlie a concave shaped erosion surface 4m deep, cut into a bed of LF 16. Fig.44 shows a field sketch where most of the trough sets are seen in the XY plane. Near the bottom the bases of the sets lie at an angle slightly less steep than the erosion surface. Towards the top they decrease in inclination and eventually become horizontal. The orientation of the individual sets is difficult to measure accurately but appears to be fairly variable.

Interpretation

Trough cross bedding is the deposit of sinuous crested and linguoid dunes (Allen 1963b). (see section 5.2.4.)

6.3.2. Lithofacies relationships within channels.

The six lithofacies described are all parts of channel fills. The maximum dimensions of the channels are unknown, but they are certainly great. The great extent, transverse to the current of beds of FLCS at Ramshaw Rocks (section 6.3.1.3.), suggests that channel widths of up to 1km seem likely. Cosets of MSXB have been mapped for larger distances, (section 6.3.1.1.), but whether these large sheets represent just one channel is uncertain. Where MSXB is the channel fill lithofacies, steep channel sides are invariably absent, and channel bases usually flat without substantial lags. Thus many of the wide sheets may represent more than one channel.

The vertical thickness of channel fills is also difficult to estimate because of the tendency for channels to be stacked one upon the other, and the difficulty of sometimes recognising channel bases. At the Roaches and Hen Cloud, channel fills reach 23m and 20m respectively, sets of LSXB accounting for the greater part of this.

Where all three major lithofacies are present in a channel, FLCS usually lies in the deepest part, overlain by a set of LSXB and this by a coset of MSXB. At the northern end of the Roaches (SK 0008 6366 to SJ 9999 6386) a coset of MSXB 10m in thickness representing an almost complete channel fill appears to pass downdip into a single set of LSXB at least 9m thick over a distance of 300m. Unfortunately, the transition is not well exposed because of the severe frost shattering of the escarpment.

At two localities sets of LSXB are orientated obliquely to steep erosion surfaces at channel edges. At the Roaches (SK 0053 6246) a channel side dips at 45° and is mantled by a thin bed of FLCS [Fig.45]. Above this foresets extend out, their strike diverging by 18° from the orientation of the channel side. These reach 4.5m in thickness before being cut out by another channel. A little further to the north (SK 0053 6256), [Figs. 39, 45], a large set extends out from the left hand side of a channel. Here the orientation of the channel side can only be estimated, the foresets diverging from the channel margin at an angle of 70° .

In this large set, which extends down dip for 460m until being cut out by another channel, the foresets nearest the channel edge are angular. Here the LSXB is overlain by a coset of MSXB and each of the smaller superimposed sets merges down dip with the larger foresets as previously described, [Fig.39]. Further down dip [Figs. 39, 40] concave erosion surfaces are present within the large set and the foresets become tangential in shape. Eventually they pass into beds of FLCS as explained in section 6.3.1.2.

The other three lithofacies found in the fluvial channels are much less common. LF 16a Undulatory-bedded Sandstone occurs only at Hen Cloud where it is overlain by toesets of LSXB. LF 10 Trough Cross-bedded Sandstone usually occurs in association with FLCS. Thin cosets of trough cross-bedding occasionally occur within thicker cosets of MSXB. A 2m thick coset of Ripple-laminated Sandstone with Pelecypodichnus overlies a bed of FLCS in a quarry near Coldspring Reservoir (SK 0447 7463). This is overlain erosively by another bed of FLCS. At a small quarry south of Old Dollands Farm

(SJ 9538 6919), towards the top of a sandstone feature, a coset of MSXB is overlain by an alternation of Ripple-laminated Sandstone with Pelecypodichnus and Trough Cross-bedding.

The apparent rarity of these last two lithofacies is probably partly a result of the nature of the exposure. Most of this is within natural gritstone edges which only weather out where the sandstone is coarse. It is clear, however, that the greater part of the channel fills are represented by lithofacies 14, 15 and 16.

The types of relationships observed, summarised in Fig.46 show that some parts of the channel were occupied largely by the smaller transverse sandwaves responsible for the cosets of MSXB, and that sometimes these passed downcurrent into the larger bedforms responsible for the sets of LSXB. The smaller bedforms were also superimposed on the backs of the larger ones. LF 16 FLCs forms in the lee of the large bedforms.

6.3.3. Interpretation of the channel fill sequences

Interpretation of the fluvial channel assemblage is difficult because of the unusual lithofacies present, and the absence of an apparent modern analogue. Crucial in the reconstruction of the channel morphology is to know what type of bedform was responsible for the sets of Large Scale Cross Bedding. It is also important to know whether the two types of bedform responsible for the two scales of cross bedding were in equilibrium with the flow at the same time, or whether they represent different responses to different discharges.

At present a number of alternative models seem possible, though not all are equally likely. McCabe (1975) has discussed four possibilities for a similar facies association in the Kinderscout Grit. Two of these, assuming the large bedforms were either channel confluence deltas or channel infill delta have already been adequately discussed by McCabe (1975) and seem unlikely. Any model must not only account for the facies relationships discussed in the preceding section, but also for the internal erosion surfaces common in the Large Scale Cross Bedding and also for the association of different types of erosion surfaces with different types of foreset. Four possibilities are discussed below.

1. Sandwave Model. Fig.41.

This assumes that the bedforms represented by the LSXB were very large sandwaves, such as those described by Coleman (1969) from the Brahmaputra. They are thus the same class of bedform, mesoforms, as those responsible for the cosets of MSXB, but as in the Brahmaputra, the larger bedforms developed at peak discharges. The common superimposition of cosets of MSXB on top of the single sets of LSXB, which merge down dip into the large foresets may represent a lag effect similar to that described in section 6.3.1.1. to account for the smaller multiple convex upwards erosion surfaces within individual medium scale sets. The concave erosion surfaces may have been produced by dissection during falling stage as the main thalweg shifted across the channel.

Although this theory cannot be entirely discounted at present, it seems unlikely for two reasons. Firstly, there are very few sets of cross bedding intermediate in size between the medium scale sets, which are always 3m or less in thickness and the large scale sets which generally exceed 10m. Secondly, except for the poorly exposed example at the northern end of the Roaches described in section 6.3.1.2. there are no cases of cosets of MS B passing downcurrent into sets of LSXB. This would be expected if the two lithofacies do represent a similar type of bedform developed at different discharges.

2. Alternate Bar Model.

This assumes that the large bedforms were alternate bars. These form in straight channels where the thalweg meanders from bank to bank. [Chapter 2, Figs. 9, 10]. They are macroforms (Jackson 1975c), a higher class of bedform than sandwaves. McCabe (1975) has suggested a similar origin for large scale cross bedding within channels in the Kinderscout Grit.

The main evidence for an alternate bar model in the Roaches Grit is the oblique relationship of the LSXB to steep channel sides. As explained

in Chapter 2, alternate bars tend to be eroded during low and falling stage when the thalweg reduces its wavelength to correspond to the lower discharge [Figs. 9, 10]. The concave erosion surfaces within the LSXB may have formed in this way. Sometimes as at Hen Cloud, only the toesets were preserved. If the thalweg shifts downriver between successive floods, as occurs in straight sections of the Yellow River (Chien 1961), then the bars may begin to build forwards in a different direction in response to the new flow pattern. This appears to happen at Hen Cloud and the Roaches where successive groups of foresets differ significantly in strike.

Three main types of foresets occur in LSXB [Fig.36]. With angular foresets the separation eddy was weak. A much stronger separation eddy was present with tangential foresets. Where the intrasets point down the foreset dip the reattachment point was relatively high up on the lee side of the bedform. With intrasets dipping in a reverse direction to the main foresets it was in front of the bedform. Often the intrasets are orientated obliquely to the main foreset dip. This should occur where the crestline of a bedform is aligned at an angle to the flow direction, so that a spiral separation eddy forms (Allen 1968). This type of flow pattern has been observed in large natural bedforms with skewed crestlines by Nedeco (1959) and Collinson (1970b).

A third type of foreset is represented by the downdip passage into FLCS. This lithofacies contains faint undulatory lamination, often of a large size. It is not dissimilar to the structure in LF 16a Undulatory bedded sandstone, and the two lithofacies are probably of similar origin. The stepsided irregular erosion surfaces at the base of the lithofacies, and the poorly developed lamination suggest rapid erosion and deposition, probably within a powerful turbulent eddy. Sets of trough cross bedding within the more structureless beds were formed during slower periods of deposition when smaller dunes had time to form. These eddies occurred periodically in the lee of the large bedforms. Coleman (1969) has described

intense turbulent cells in the lee of giant sandwaves in the Brahmaputra River.

Where the bedform crestline is highly skewed relative to the flow a series of parallel helicoidal cells develop, (Allen 1968). Spurs in front of bedforms with skewed crestlines such as those described in front of alternate bars by Harms and Fahnstock (1965) were probably cut by this type of eddy pattern. McCabe (1975) suggested that eddies of this type formed undulatory bedding (his Facies 21) in front of large alternate bars in the Kinderscout Grit Channels. The presence of faint undulatory lamination within the FLCS lithofacies suggests that the short-lived turbulent cells envisaged may also have had a helicoidal shape.

The different types of foreset described above may have developed at different discharges or at different positions of the bar crest, or due to a combination of both. With the alternate bar model envisaged, [Fig.48] the area nearest to the channel edge probably contains a relatively weak eddy. Here angular foresets should occur. Further out towards the edge of the bedform a more powerful eddy should occur. Here the crestline is skewed relative to the flow direction favouring the development of spiral or paired helicoidal eddies. The FLCS lithofacies was probably deposited very quickly at high discharges.

Evidence in support of the ideas expressed above comes from the section along the main face of the Roaches [Figs.39, 40, 48]. This may be interpreted as forming from the movement of an alternate bar down a channel. Because of the orientation of the exposed section, relative to the channel edge, the section shows different parts of the bar, beginning with the parts nearest to the channel edge and then moving downdip out towards the centre of the channel. The LSXB forming near the channel edge has angular foresets with convex erosion surfaces. Here the separation eddy was weak. Further down dip, the foresets become more tangential in

shape and eventually pass down into FLCS. Several concave erosion surfaces occur here, formed during periods of thalweg migration during falling stage. The overlying foresets often pass down into FLCS, probably deposited very quickly at high discharges as a result of lee scour and rapid dumping of the sand in front of the large bedforms. As the bars continue to build out into the channel the large foresets overlie the beds of FLCS. These are markedly tangential with well developed regressive intrasets. No angular foresets occur in this part of the section.

Accepting an alternate bar origin for the LSXB it is still necessary to explain the relationship to the MSXB, as in some parts of the channel the alternate bars are not present. Three relationships seem possible [Fig.47].

1. Here the alternate bars are assumed to occur in narrow deep channels in a braided river, whereas the smaller sand waves largely occurred in wider areas. A similar type of relationship appears to occur in the Niger (Nedeco 1959) where side bars and alternate bars are best developed in the narrower channels. Lateral migration of the shallower parts of the channel may lead to the common superimposition of MS+B cosets over solitary sets of LSXB.

This is essentially the model proposed by McCabe (1975). The main evidence in favour is the wider lateral extent of the MSXB cosets. As explained earlier, however, it is not certain whether these large sheets represent just one channel. Another problem is the similarity in orientation of the large scale and medium scale sets. Collinson (1967, 1969) notes a similar close association in the Kinderscout Grit. A third difficulty is the absence of channel abandonment sequences. These should occur in multiple channel braided rivers when channels are blocked off (Chapter 2).

2. Here the channel is assumed to be straight. Alternate bars form on opposite sides. Smaller sandwaves are superimposed on their backs, and also

upstream in areas where the thalweg crosses over from one side of the channel to the other, [Fig.47a]. Thus in some parts of the channel where only sandwaves are present, cosets of MSXB are produced. In other parts of the channel occupied by alternate bars solitary sets of LSXB are overlain by cosets of MSXB.

The main argument against this model is the wide lateral extent of the MSXB cosets. This can be discounted, however, if the wide sheets are assumed to represent the fills of more than one channel. Another problem is the absence of steep channel sides and erosion surfaces within the MSXB cosets. With the falling stage reworking envisaged to account for the large concave erosion surfaces with the sets of LSXB, it is difficult to see why similar structures should not occur within the MSXB cosets. A third difficulty is the mutually erosive relationship of the separate channel fills which often occur as sheet sandstones [Fig.45]. This relationship is more likely to occur with braided channels rather than straight channels, which in most modern deltas occur as isolated forms surrounded by finer sediments.

3. A final possibility worth consideration assumes that the alternate bars were not in equilibrium with the flow at the same time as the smaller sandwaves [Fig.47a 2]. In a straight channel intense scour at the channel sides by the meandering thalweg during peak discharges forms alternate bars on either side of the channel. During falling stage thalweg migration dissects the bar cutting the concave erosion surfaces. At a lower discharge sand waves form in the channel both on top of the bars, and in the area between the bars where the thalweg crosses the channel. The first occurrence accounts for the usual association of LSXB succeeded by a coset of MSXB. Where the sandwaves reached the crests of the now 'inactive' alternate bars these would continue to move forwards even though no longer in equilibrium with the flow. This may be an alternative explanation

for the association of angular large scale foresets and multiple convex upwards erosion surfaces passing up into MSXB seen at the Roaches [Fig.39]. Sandwaves forming in the cross over area form channel sequences with just thick cosets of MSXB.

Not all channels necessarily developed alternate bars. If the discharge was not great enough channel fills would be dominated by cosets of MSXB as appears to be the case in the Rough Rock (Heath 1973) and Crawshaw Sandstone (Guion 1971).

Ripple-laminated sandstones with Pelecypodichnus may have formed in the deepest parts of the channel at low stage or during periods of channel abandonment. The presence of Pelecypodichnus indicates the ability of basin water with bivalve spat to move up into the channels at some stage. In the Mississippi, a salt water wedge extends for 216km up from the mouth during low stage, (Moore 1970). The major channel fills exposed at the Roaches must have been deposited within only 5km of the river mouth, as beyond this distance the Delta Top Association is absent.

6.3.4. Discharge Regime

In Chapter 2 it was suggested that there is a range of possible bedform responses to different changes in flow strength. Falling stage modifications have the greatest preservation potential, and sandwaves are generally a more useful indicator than dunes because of their greater size and slower response.

Lithofacies 14, Medium Scale Cross-bedding occurs in thick, relatively undisturbed cosets. No structures definitely resembling the Reactivation Surfaces of Collinson (1970a) occur, the main evidence of falling stage modification being the multiple convex upwards erosion surfaces. This suggests, that unlike the Tana River, when the sandwaves are left standing and substantially unmodified during the rapid fall in discharge, here the rate of fall of discharge was sufficiently slow to allow bedform superimposition. These smaller bedforms substantially modified the larger

sandwaves in equilibrium with a higher discharge. In addition, there is no evidence of low stage modification in this part of the river channel, unlike the River Platte, for example, where the considerable low stage flow substantially modified the high stage features (Smith 1974).

This evidence suggests that the river which deposited the Roaches Grit had a regime similar to that of the present day Brahmaputra. In this river, Q_{\max}/Q_{\min} is high, [Fig.4], and the low stage flow is restricted to the narrow deepest part of the channel, many parts of the high stage channel being completely abandoned, (Coleman 1969). The falling stage period however is long enough to allow bedform superimposition over wide areas of the channel.

In LF 15, Large Scale Cross-bedding there are also multiple convex up erosion surfaces indicative of bedform superimposition. It is uncertain, however, whether this is a lag effect or whether the relatively smaller sandwaves were superimposed in equilibrium with a different class of bedform such as an alternate bar. In addition LF 15 contains the large concave erosion surfaces indicative of large scale erosion and dissection. This is only to be expected. Because of their larger size, the bedforms responsible for LF 15 would have a much greater relaxation time for a given discharge than the smaller sandwaves responsible for LF 14 MSXB.

In conclusion the available evidence suggests that during falling stage the smaller sandwaves adjusted to the discharge change by being covered by even smaller superimposed bedforms, whereas the giant bedforms represented by LF 15 could not adjust and suffered extensive dissection as they lay largely inactive in the channel.

6.4. Relationships within the Association

The Delta Top Association is dominated by Assemblage B, fills of the large channels. Assemblage A usually occurs below the lowest channel in the sequence at the top of the Slope Association. The relationship of the separate channel fills is best seen at the southern edge of the Goyt Syncline, particularly between Hen Cloud (SK 008 615) and Roach End (SJ 996 644), where there is 3km of almost continuous exposure. The northern part of this is shown in Fig.49.

In some sections there is only one channel fill in the Association, but elsewhere two or three or four. These may be stacked erosively on top of each other as at Ramshaw Rocks, or separated by finer grained sediments as at the northern end of the Roaches, [Plate 49]. Where there is laterally continuous exposure, as in the middle part of the Roaches, channel fills are sometimes seen to cut into each other, lying side by side to form laterally extensive sandstones [Figs. 45, 49]. Some of these may be mapped for several kilometers.

The Roaches Grit Group between the R. metabilingue goniatite band and the base of the Delta Top Association is about 250m thick at the Roaches. This was deposited immediately prior to the deposition of the Delta Top Association and would still have been relatively uncompactd. About half of this sediment is fine grained mudstone and siltstone, and mudstone will compact eventually by about 50%, (Westoll, 1962). This compaction will result in a sinking of the delta top so that the level of successive channels will be adjusted to a higher relative sea level. This compaction alone will account for most of the thickness of the Delta Top Association.

Between Hen Cloud and Roach End, where there are at least 11 separate channel fills, these do not only lie at successively higher levels in the sequence, but are also systematically offset so that all channel fills become progressively younger towards the north, [Fig.49]. This is particularly well seen at the northern end of the Roaches [Plate 49]. The coarse grained channel fills form features, whereas the relatively finer grained interchannel sediments form slacks, but are not exposed. Here three separate channels occur at successively higher levels separated by 15m - 20m of finer sediment. The channels are offset towards the north.

Brown (1969), has described offsetting of series of 2 or 3 channels from a Pennsylvanian deltaic sequence in Texas. He suggests this is a result of differential compaction. After channel abandonment the sand fill will compact relatively little, whereas the adjacent finer sediments into which the channel has cut will compact more, [Fig.50]. This will result in a slightly irregular relief in the delta top with the areas underlain by abandoned channels being slightly higher. When a second channel occupies the area, its course will run along the zone of lower relief, so that the two channel fills will be offset.

At the Roaches, many more channels are offset, and it seems unlikely that differential compaction alone will account for the relationship. Several modern rivers are systematically shifting their courses in one direction. The Kosi River has moved westwards by about 115km in the last 200 years, (Leopold et.al. 1964, Holmes 1965, Gole and Chitale 1966), each time shifting in a series of short steps. The Indus has likewise shifted laterally a comparable distance, though at a much smaller rate of approximately 0.05km/yr, (Wilhelmy 1966, Holmes 1968). In Bangladesh, the Brahmaputra River is also systematically shifting westwards, whereas the Ganges is moving north-eastwards, (Coleman 1969). Although the causes of these movements are not known for certain, the Bengal Basin is an area of considerable seismic activity and it is likely that structural tilting is affecting the positions of the river courses. (Coleman 1969). Failing has shown that the courses of the Trinity and adjoining rivers have changed as a result of tectonic movements in the Galveston area of the U.S.A. Here the development of an anticline and syncline, probably only in the last 8,000 years, has shifted the river courses to the west.

These examples show that small tectonic movements can have marked effects on the courses of rivers. The northerly offsetting of the channels at the Roaches is most likely the result of slight tectonic tilting to the north. The area is only about 20km north of the southern edge of the basin. Beyond this was a land area still supplying sediment to the South Wales Basin (Kelling 1974). The Roaches Grit Group thins rapidly southwards and is absent to the west and southwest of Stoke, suggesting that this area was still land although no longer supplying sediment to the north. Slight uplift of this area would be sufficient to tilt the basin.

6.5. Morphology of the delta

The Roaches Grit delta was built out into a deep water basin where no evidence of tidal action has been recorded and wave activity was limited. It is therefore of the high constructive fluvially dominated type (Fisher 1969,

Wright and Coleman 1973). The best known modern example is the birdsfoot delta of the Mississippi River (Fisk 1961). McCabe (1975) has suggested that the Kinderscout Grit delta is of this type and that the coarse grained channel fills were elongate distributories similar to those in the Mississippi.

There are, however, important differences between the pattern of sedimentation in the Mississippi delta and that which can be deduced for the deep water deltas in the Central Pennines Namurian. Firstly, most of the deposition of sand in the Mississippi takes place in the mouth bar region. As the distributories build out they construct what Fisk (1961), has called, 'bar finger sands'. Little deposition takes place within the channels. A vertical section through the Mississippi, (Coleman and Wright 1975) shows a thick mouth bar sequence but very little channel fill sediment. The Roaches delta sequence is quite different, here, most of the deposition took place within deep channels and mouth bars have not been recognised. In most cases the mouth bar sediments if they ever existed have been eroded away by the channels.

Secondly, the channels in the Mississippi delta are extending outwards but not shifting laterally. The present birdsfoot delta is about 250 years old (Frazier 1967) and the major channels have not shifted position for at least the last 100 years (Scruton 1960). Eventual abandonment of the delta will leave a series of narrow radiating sand bodies surrounded by finer grained sediment, (Fisher 1969). In the Roaches sequence, however, the delta top contains sheet sandstones composed of many channel fills with erosive contacts.

A birdsfoot model therefore seems inapplicable. The other type of high constructive delta, the lobate form, produces much more widespread sands, but here the flow splits into many small channels. The size of the channels in the Roaches sequence suggests that the flow was confined to just one or at most a very small number of major channels.

The Roaches Grit delta was unusual in that it was building out into a relatively fresh water basin and that the discharge was probably highly variable. No modern example of a large variable discharge river building out into a fresh water basin is known to the author, and as yet there seems no reasonable modern analogue.

The relationship of the channel fills suggests that the channels behaved like those in modern sandy braided rivers as discussed in Chapter 1. Frequent lateral shifts in the course of the river resulted in the construction of sheet sandstones. Little is known of the interchannel areas because of the limited exposure. In the Kinderscout sequence McCabe (op.cit.) and Collinson (op.cit.) describe small scale coarsening upwards sequences which were probably formed as crevasses.

Table V

Directional and other properties in Lithofacies 14 from different areas

<u>Area</u>	<u>N</u>	<u>Vector Direction</u>	<u>Vector Magnitude</u>	<u>Variance</u>
Wildboarclough	24	346	73	2025
Rudyard	12	269	69	2500
Cisterns Clough	11	290	86	1156
Allgreave Wood	36	312	81	1600
Helmesley Wood	38	300	76	1936
Roaches - south	34	323	90	729
Roaches - north	31	347	74	2025
TOTAL*	220	319	73	2025

* Includes a few readings from other areas.

Table VI

Two vertical sections through cosets of LF 14, from base upwards, showing variations in cross-set orientation

A.		B.	
<u>Cross set number</u>	<u>Orientation</u>	<u>Cross set number</u>	<u>Orientation</u>
1	348	1	026
2	305	2	NM
3	306	3	Several small sets
4	308	4	330 NM
5	Not measureable	5 - 6	NM
6	320	7	332
7 - 8	NM	8	338
9	346	9	NM
10	321	10	278
11	NM	11	340
12	334	12	NM
13	NM	13	332
14	336	14	346
15	328	15	292
16	334	16	302
17 - 20	NM	17	NM
21	063	18	330
22	NM	Total thickness 17m	
23	288	Location of sections	
24	NM	A	SK 006 622
25	282	B	SK 005 624
26	NM		
27	316		
28 - 30	NM		
31	324		
32	NM		
33	334		
34	332		
Total thickness 16m			

Chapter 7

Association D. Delta Margin Sediments

7.1. Introduction

This association is generally quite thin, and only exposed in a small number of stream sections. It extends from the top of the highest fluvial channel in the succession up to the R. superbilingue goniatite band. Six lithofacies occur in the association.

7.2. Lithofacies description and interpretation7.2.1. Lithofacies 17. Seat-earth and Coal

This lithofacies is now only exposed in section C [Fig.51] in Shirley Hollow. This coal is just a very thin accumulation of coalised plant debris overlying a coarse sandstone. Roots penetrate through the sandstone and into an underlying mudstone for a total of 0.6 m. A few roots also overlie the coal.

Elsewhere, thin coals have been recorded near Thickwithens Farm (SJ 9561 7044) (Pocock 1906) and in the stream section opposite Westwood Hall (SJ 9592 5602) (IGS 6" map SJ 95 NE). Hull and Green (1864) recorded thin coals in a number of areas. It seems clear, however, that a laterally persistent coal is not developed.

Interpretation

This lithofacies provides evidence of emergent or near emergent conditions. Coals can develop in a variety of sub-environments (Jansa 1972). Further discussion will be given in section 7.3.

7.2.2. Lithofacies 18. Wave-laminated Sandstone

Fine grained sandstones and coarse siltstones in beds 0.5 - 2.5 m thick, with fairly sharp bases. The sediment is parallel laminated with very thin laminae, usually less than 1 mm thick, picked out by concentrations of carbonaceous debris or mica. They may be plane, usually inclined by a few degrees in various directions, or undulatory. Flat or curved erosion

surfaces with a relief of up to about 10 cm occur throughout the beds. The succeeding lamination is usually parallel to an erosion surface or may "back fill" an eroded hollow. [Plates 50, 51]. Some of the more undulatory lamination resembles small scale trough cross-bedding but the regular intersecting trough shaped erosion surfaces do not occur. Instead, when traced laterally, the erosion surfaces here invariably merge with and disappear into the lamination.

Rarely small 'channels' with a relief of up to 0.7 m occur. Thin 'lags' of plant stems are present in the 'channel' bases. Bioturbation is very common in this lithofacies and at many horizons the lamination has been completely destroyed [Plate 52]. Specific ichno species are difficult to identify but some have retrusive spreite similar to those found in the family Rhizocorallidae (Seilacher 1967).

Interpretation

This lithofacies is unusual both in structure and lithology. Nothing resembling it occurs in the other three associations. The parallel lamination and erosion surfaces indicate powerful current activity. Close modern analogues are difficult to find; but the structures bear some similarity to those formed by wave activity. The horizons of intense bioturbation show that there were periods of little net deposition during which the sediment was reworked. Howard (1971), describes a similar situation in shoreface sands from the Georgia Coast, U.S.A. The area lies within the zone of possible wave reworking, but except for storm conditions, it is below strong wave disturbance. The most abundantly preserved structures are parallel-laminated sands, produced by wave activity during storms, and towards the top of these intensely burrowed horizons formed when the animal community re-establishes itself on and near the sediment surface. The interbedded siltstone units in the Roaches sequence were presumably deposited during quieter conditions.

7.2.3. Lithofacies 19. Homogeneous Siltstone

Dark grey siltstone with much dark carbonaceous material and scattered mica. Small plant fragments are common. Sand sized sediment is absent, and the rock is completely unlaminated. Essentially it is a finer grained version of Lithofacies 8.

Interpretation

The characteristics of the lithofacies suggest deposition from suspension in fairly quiet water.

7.2.4. Lithofacies 1 and 2. Mudstone and Silty mudstone.

These resemble the sediments described from Associations A and B. They were deposited from suspension in quiet water.

7.2.5. Lithofacies 3. Goniatite Faunal Bed.

This is represented by the R. superbilingue goniatite band which forms a widespread and apparently persistent horizon at the top of the sequence studied. Like the goniatite bands in the Deep Water Association the sediment is a sooty black colour. The main difference is the presence of the benthonic Lingula.

Interpretation

The R. superbilingue band records the return to quiet water offshore conditions following the final abandonment and modification of the delta. The depth of water was almost certainly much less than when the bands in the Deep Water Association were formed, but except for the presence of Lingula, there is no other obvious difference.

7.3. Vertical sequences and discussion

Three sections in Association D are shown in Fig.51. Only section C from Shirley Hollow represents a complete sequence. Sections A and B are only 3.1 km apart, and the tops and bases are believed to lie roughly at the same horizon. In this part of the basin, where the Roaches Grit Group is thickest, the association reaches 30 m in thickness.

In section B, which starts above the highest fluvial channel there is a well developed coarsening upwards from homogeneous silts through poorly sorted fine sands into a fairly well sorted development of LF 18. [Plate 53]. Section A begins with LF 19 Homogeneous Siltstone into which is cut a small channel. Above this, fine sands and coarse silts of LF 18, often extensively bioturbated, alternate with thinner beds of LF 19. Towards the top the sequences fines upwards through siltstone and mudstone being capped by the R. superbilingue band.

In Section C a thin coal occurs above the highest fluvial channel. Above this occurs a straightforward fining upward sequence up to the R. superbilingue band.

These three sections provide almost the only evidence of the delta top succession marginal to the major fluvial channels. For this reason they have been described in some detail in spite of the limited exposure. Within modern deltas, marginal environments are found in active interdistributary areas and abandoned parts of the delta top. Although all of the sections occur above the highest channels in the vertical sequence it is quite possible that they are partly laterally equivalent to higher channels elsewhere and as such represent overbank as well as abandonment conditions.

If the origin of LF 18 outlined in Section 7.2.2. is accepted then the sequences in sections A and B show that the delta was subject to a fair amount of wave activity.

Wave reworking in modern deltas becomes most important following abandonment, when the development of barriers on top of the former active lobe, and their subsequent shorewards migration produces a coarsening upwards sequence, (Otvos 1970, Elliott 1974a, 1975). This provides a possible explanation for the sequences in sections A and B, but the thickness of the coarsening upwards component (4.5 m) and of the wave formed sandstone unit (14 m) makes this explanation unlikely. More probably sections A and

B represent some form of interdistributary environment where finer sediment swept out of the fluvial channels in suspension was reworked. With this interpretation Sections A and B represent the approach and then cut off of a fluvial channel a few kilometers away.

All of the coals in the Roaches sequence appear to be local developments only. Widespread coal seams of the type which cover abandoned delta lobes (Frazier and Ozanik 1969, Elliott 1974a) and which appear to occur in many of the succeeding cyclothem in the Namurian and lower Westphalian (Stevenson and Gaunt 1971) are absent. The coals in this sequence may have formed in a number of local environments in the delta top such as interdistributary flood basins, abandoned channels and marshes between beach ridges (Jansa 1972). The absence of a widespread abandonment phase coal, may possibly be a result of the high rate of subsidence in the thick deltaic pile.

Chapter 8

Trace Fossils

8.1. Introduction

A considerable variety of trace fossils exist in all four associations, though most occur in Association B, the main delta slope. Some forms occur in more than one association whereas others are restricted. A crude depth zonation exists, but with only limited collecting the complete depth range of the ichnospecies cannot be established.

8.2. Description and interpretation

Trace fossil names in this section follow Häntzschel (1962) and terminology follows Martinsson (1970).

8.2.1. Pelecypodichnus sp Seilacher 1953.

Material A large number of specimens from many localities.

Morphology Oval epichnial depressions and hypichnial casts, also as a V shaped endichnial disturbance usually in groups. [Plate 54]. Endichnial disturbances result from a down turning of the adjacent sedimentary lamination into a V. [Plates 15, 20]. Sometimes the lamination is completely destroyed and replaced by a plug of structureless cleaner sand. [Plate 55]. The disturbances may be as little as 0.5 cm in length, but commonly exceed 30 cm. They are usually vertical, but may be inclined by a few degrees.

Pelecypodichnus varies in length from $1\frac{1}{2}$ cm to $3\frac{1}{2}$ cm. Generally speaking forms at a particular horizon vary little in length. Hardy (1973) found a similar narrow size grouping in Pelecypodichnus from the lower Westphalian in Lancashire. Groups usually occur closely spaced with a strong preferred orientation. [Plate 54].

Occurrence. Pelecypodichnus is found in Lithofacies 6 and 7. It is very common in Association B and single horizons have been traced laterally for over 100 m. Sediments repeatedly disturbed by Pelecypodichnus reach 12.8 m. One occurrence is known from the Deep Water Association, and two

from Association C where it is found in ripple-laminated sandstones within fluvial channels.

Ecology

Seilacher (1953) interpreted Pelecypodichnus as a resting trace formed by bivalves lying edgewise in the sediment. Pelecypodichnus occurs in the Upper Carboniferous Central Pennine Basin associated with the bivalves Carbonicola cf. extenuata, (Hardy 1973, Eager 1971, 1974). Hardy has found the bivalves in life position. He suggested that Pelecypodichnus is a resting trace with the bivalve lying vertical buried up to the umbo with the inhalent siphon pointing downcurrent.

Eager (1971, 1974), on the other hand regards Pelecypodichnus as a burrow, and shows a life reconstruction with the bivalve almost completely buried with the posterior just protruding above the sediment surface. He suggested that the cavity was infilled with later deposited sediment. Osgood (1970) has described Lockeia (James 1879), probably a junior synonym of Pelecypodichnus from the Ordovician of Cincinnati, U.S.A. He also suggested that the bivalves burrowed, subsequent erosion removing most of the cavity.

Burrows result in a break in the lamination which is plugged with sand, whereas resting traces ideally form an unplugged vertical disturbance, (Chisholm, 1970) [Fig.52]. Both forms occur and the V shaped structures are undoubtedly resting traces, similar structures being produced by the recent bivalve Mya arenaria (Reineck 1967). Massive plugs are most commonly found in Lithofacies 7 Ripple-laminated Sandstone, but they do not necessarily indicate a burrow. Flume experiments with the shells of similarly shaped recent bivalves pushed into a bed of sand on which ripples are active bedforms show that the shells offer considerable resistance to the flow. With a parallel orientation, scour on either side of the shell caused the shell to fall over in less than a minute when the ripples were moving quickly.

The living bivalves anchored themselves to the sediment by means of a foot. Active movement and repositioning of the foot was probably necessary to keep them secure. Such movements could easily result in the formation of a massive plug. There is therefore no unequivocal evidence to suggest the bivalves burrowed.

Long endichnical disturbances are produced when the animal moves upwards during periods of rapid sedimentation. At Bosley (SJ 91 64) the Slope Association consists of alternations of Ripple-laminated Sandstone and Micaceous, Carbonaceous Sandstone deposited mainly from suspension. Pelecypodichnus usually occurs at the bases of the rippled sandstones. This suggests that the bivalves colonised the surface during periods of slower suspension sedimentation and then moved upwards through the rippled sandstones during periods of more rapid sedimentation brought about by density currents operative on the slope during floods. The wide lateral extent of some horizons of Pelecypodichnus shows that large colonies of bivalves colonised the delta slope. Presumably later solution removed the shells, as is common in acid rich waters.

Pelecypodichnus is uncommon in the Deep Water Association, but it does not follow that bivalves were also rare in this environment. Most of the sediments are not of a type which would easily preserve resting traces, the mudstones are too fine grained, and the turbidity currents probably swept away the shells. Only in Lithofacies 6 is Pelecypodichnus found and this lithofacies is comparatively rare in Association A.

The presence of Pelecypodichnus in sediments within fluvial channels shows that these had passive connection with the basin at some period. This may have occurred during low stage when as in modern delta distributaries basin water moves up the channel (Moore, 1970) or after abandonment.

8.2.2. Inclined bivalve resting traces

Material A number of specimens from Chapel Station Quarry (SK 055 793).

Morphology

Hypichnial ridges of structureless fine sand, about 0.5 cm in width and 2 cm in height, vertical, but more commonly inclined at between 10° and 30° to vertical. In plan they are oval in shape 1 - 2 cm in width. They are found in groups on the soles of thin, fine sandstone beds, interbedded with finer grained silty mudstone. Usually they have a consistent orientation.

Ecology

They occur within the upper part of the delta slope in fine grained sediments interbedded with coarser LF 7 Ripple-laminated Sandstone and LF 12 Lenticularly-bedded Sandstone. (Section D in Fig.21). Their size and oval shape suggests they are probably bivalve resting traces. No bivalves have been found but associated ripple-laminated sandstones contain Pelecypodichnus.

Wherever bivalves have been found in life position in association with Pelecypodichnus in the Central Pennines they have always been vertical (Hardy 1973). However, Wagner (in Eager 1974) has found Anthraconaia in 'burrowing' position inclined to the bedding by about 40° in the Stephanian of Spain. The shape of the trace fossil from the Roaches Grit Group suggests that here also, the bivalves sometimes rested inclined within the sediment.

8.2.3. Bergaueria sp. Prantl 1946

Material Specimens from Hogshaw Brook (SK 0547 7492) and Dane Bridge (SJ 9661 6521).

Morphology

Sub-rounded hypichnical cast up to 1.5 cm in diameter and 0.75 cm deep. The specimen shown in Plate 56 is asymmetrical in cross section, one half with a near vertical side, the other with the side inclined at about 30° .

Occurrence

Occurs at the base of LF 7 Ripple-laminated Sandstone and LF 4 Turbidites in Association B. Bergaueria has been interpreted by various authors as the cast of burrows made by sedentary organisms such as coelenterates, arthozoans and actinians (Alpert 1973). Baines (p.com.) on the other hand suggests that Bergaueria found on the bases of turbidites within the Pendle Grit (Namurian Eic) in the northern Pennines may be the dwelling place of a worm living at the sandmud interface. No trails have been found leading in to the structure in either of the present examples, or those figured by Baines, and his suggestion seems unlikely. It is just as likely that Bergaueria is a dwelling cavity in the muddy sediment surface, casted by sand brought in later.

8.2.4. Cigar Shaped Ridge

Material Specimen from near Burbage Reservoir (SK 0346 7216)

Morphology

Cigar shaped hypichnical ridge 3.8 cm x 1 cm and about 0.5 cm deep. Semi-circular in cross section with fairly steep sides [Plate 57].

Occurrence

Base of a bed of LF 4 Turbidites within the Deep Water Association

8.2.5. Cochlichnus sp. E. Hitchcock 1858,

This form has also been referred to Sinusites (Demagnet and Van Straelen 1938), which is still widely used.

Material Loose and in situ specimens from a number of localities.

Morphology

Regularly sinuous endichnical burrows, hypichnial depressions and epichnical ridges, resembling sine curves. They are circular in cross section, with a diameter of about 1 mm. Wavelengths vary from 1 - 2 cm and are constant within an individual structure. Length is highly variable, the maximum recorded being 17 cm, though some are as short as 1½ cm. Several trails may criss-cross the same area [Plates 56, 58, 60].

Ecology

Cochlichnus occurs within Lithofacies 4, 7 and 8 in the Deep Water, Delta Slope and Upper Slope and Delta Top Associations. The meandering trail was probably made by a worm moving on or just within the sediment surface.

8.2.6. Discontinuous Sole Trails

Material Loose and in situ from a number of localities.

Morphology

Sinuuous to meandering hypichnial ridges [Plate 59]: 1 - 5 mm diameter and up to 13 cm in length. The structures vary in thickness and often thin out altogether for distances of up to 1 cm forming a discontinuous trail.

They resemble forms described by Holdsworth (1963) referred to Granularia but the occasional branching which he observed is not present. McCabe (1975) and Baines (p.com) describe similar structures and refer them to Planolites, although these are less variable in thickness. The 'interrupted meanders' figured by Ksiazkiewicz (1970) are also very similar.

Ecology

These occur on the soles of Lithofacies 4 Turbidites within the Deep Water Association. It is not clear whether they pre date or post date the casting sediment as critical relationships with turbidite bottom structures have not been seen. In the specimen shown in Plate 59 rare fine grooves occur in un-bioturbated areas, but most of the bioturbation seems to have obliterated the current formed structures.

The trails are probably formed by worms, burrowing along the sand mud interface, periodically passing material through the alimentary canal. Both Seilacher (1962) and Holdsworth (1963) consider the similar Granularia to be post-depositional.

8.2.7. Large Pipe

Material One loose block in stream section above Upper Hulme (SJ 0125 6129).

The specimen is incomplete.

Morphology

Hypichnial cast, circular in cross section, 1 - 2 cm in width, and 4 - 5 cm in length. Slightly sinuous in plan and inclined sub-horizontally. The pipe is entirely sand filled with traces of a concentric lamination parallel to the sides [Plate 61].

Occurrence

Occurs within LF 7 Ripple-laminated Sandstone within Association B.

8.2.8. Planolites sp. Nicholson 1873.

Material In situ material from near Burbage Reservoir (SK 0346 7216)

Morphology

Straight, cylindrical, unbranched endichnical burrows 1 mm in width and up to 5 cm in length. They occur within the Bouma B division in Lithofacies 4 Turbidites, which consists of interlaminated arenaceous and argillaceous layers. They are now preserved along laminations as ridges where the underlying sedimentary lamination is argillaceous or as grooves where the overlying sedimentary lamination is argillaceous.

Occurrence

Planolites occurs within LF4 Turbidites in the Deep Water Association and is probably formed by worms burrowing within the sediment. Where the animal moved along the junction between two laminations of different composition, the cavity was filled by collapse of the upper lamination.

8.2.9. Rhizocorallidae

Material In situ specimens from Cumberland Brook (SK 994 698)

Morphology

Endichnial disturbances up to 1 cm in diameter and 2 cm in length, inclined by about 5° from vertical. They contain a series of thin, gutter shaped laminations or spreite. [Plate 52]. The associated sediment is extensively bioturbated with little structure.

This type of structure occurs with Rhizocorallium, Diplocraterion and Teichichnus. None of the specimens are well enough preserved to assign a genus and are referred to the Family Rhizocorallidae (Seilacher 1967).

Occurrence

Found only in LF 18 Wave-laminated Sandstone, in Association D.

8.2.10. Rounded Sole Trail

Material In situ specimen from stream section Burbage Reservoir (SK 0346 7216)

Morphology

Round hypichnial trail 3.5 cm in diameter and 0.5 cm deep. Slightly asymmetrical in cross section [Plate 62].

Occurrence

Base of a bed of LF 4 Turbidites in the Deep Water Association.

8.2.11. Small Simple Pipes

Material Loose block from Bosley Stream Section (SJ 91 64) and in situ material near Burbage Reservoir (SK 0346 7216).

Morphology

Small simple pipes, cylindrical in shape 0.5 - 1 mm in diameter. They give rise to round hypichnical casts regularly spaced at about 5 per square centimeter. [Plate 57]. In section they occur as vertical to slightly inclined pipes up to 3 cm in length. They are usually only visible in laminated and thinly bedded sediments, where endichnial forms are filled with darker, more argillaceous sediment brought down from the overlying bed, and exichnial forms are filled with lighter coarser grained

sediment. Quite often there is a slight bending down of the adjacent sedimentary lamination [Plate 63].

These trace fossils are very similar to the 'small simple pipes' figured by Chisholm (1968), but neither bent, nor horizontal forms have been seen.

Ecology

Small simple pipes occur within lithofacies 7 and 8 in the Slope Association and in Lithofacies 4 in the Deep Water Association. Chisholm (1968), notes that among present day forms, similar burrows are produced both by small worms which eat sediment below the surface and those which find their food on the surface or above it.

8.3. Vertical Distribution of Trace Fossils

The vertical distribution of the forms described within the main Facies Associations are shown in Table VII. There has been insufficient collecting to establish the complete range of the ichnospecies. The forms most common in the Delta Slope Association, Pelecypodichnus, discontinuous sole trails and small simple pipes also occur in the Deep Water Association, but Pelecypodichnus, very common on the slope, is rare. Within Association D, in spite of the intense bioturbation of the sediment only one form may be actually recognised. Pelecypodichnus does not occur here, perhaps because of the approach to more marine conditions following the cut off of the fluvial supply.

The vertical sequence in the Roaches Grit Group records the filling in of a fairly deep water basin and the establishment of shallow water conditions. It is difficult to estimate the water depth during the deposition of the Deep Water Association because of the unknown effects of compaction and subsidence, but it is likely to have been in excess of 100 m.

Seilacher (1967) has claimed that trace fossil communities are mainly bathymetry-controlled. He has erected four major communities ranging

from the littoral Skolithos Facies with its vertical burrows and other dominichnia, down to the deep water Nereites Facies characterised by more regular grazing trails. This type of depth zonation does not occur in the Roaches Grit Group. The Deep Water Association contains vertical burrows but regular grazing trails are absent. This is almost certainly because the turbidite sequence here was deposited at much shallower depths than the thick Flysch sequences from where Seilacher has recognised deep water communities. It does show, however, that vertical burrows and other probable dominichnia such as Bergaueria can occupy a comparatively wide depth range and are not always restricted to a very shallow water environment.

Table VII Distribution of trace fossils within the AssociationsASSOCIATION D

Rhizocorallidae

Much unspecific bioturbation

ASSOCIATION CFluvial Channel Assemblage

Pelecypodichnus

Upper Slope Assemblage

Bergaueria

Cochlichnus

Pelecypodichnus, very common

Inclined bivalve resting trace

ASSOCIATION B

Bergaueria

Cochlichnus

Pelecypodichnus, very common

Discontinuous sole trails

Large pipe

Small simple pipes

ASSOCIATION A

Pelecypodichnus, Rare

Planolites

Cigar shaped Ridge

Discontinuous sole trails

Rounded sole trail

Small simple pipes

Chapter 9

Areal Variations, Basin Analysis and Regional Significance9.1. Earlier Namurian Sedimentation

By the beginning of the Namurian a fairly deep basin had been established in northern England. The southern edge of the basin lay against the Midland Landmass through mid-Staffordshire and south Derbyshire [Fig.53]. The northern edge is marked by the Craven Fault forming the boundary of the Askrigg Block. The western and eastern limits of the basin are unknown.

During the early part of the Namurian, a number of structural blocks had a considerable effect on sedimentation facies and thicknesses. In the central area, the East Midland Shelf and its south west extension the Derbyshire Block were areas of shallower water and condensed deposition, the E1 sequence being locally unconformable on the edge of the Derbyshire Block around Buxton, (Kent, 1966, Trewin and Holdsworth 1973). In north Staffordshire, the north-south trending Red Rock fault seems to correspond with the western margin of a wide Western Shelf covering most of Cheshire and North Wales. The E1 sequence thins rapidly across the fault (Trewin and Holdsworth 1973). The Askrigg and Alston Blocks were also positive areas of condensed deposition and provided a link to other basins further north.

Between these blocks were three areas of deeper water, [Fig.53]. The large Central Pennine Basin; to the south the small North Staffordshire Basin lying between the Red Rock Fault and Derbyshire Block, and to the east of this, the narrow east-west trending Widmerpool Gulf (Falcon and Kent 1960).

Prior to the beginning of the Namurian paralic conditions were restricted to north of the Craven Fault. The earliest phases of deposition during the E Zone in all three basins were turbidites.

During the Namurian the basin was progressively filled so that by Westphalian times parelic environments were widespread across the entire area. In the Namurian the middle and northern parts of the area experienced three major phases of fill, [Fig.58]. The first, during the Elc Zone, the Pendle and Skipton Moor Grits (Reading 1964, Baines p.com.), the second during the R1 Zone, the Kinderscout Grit Group (Reading 1964, Collinson 1967, 1969, McCabe 1975), and the third during the lower R2 Zone, the Alum Crag Grit in Lancashire (Collinson, Jones and Wilson in press). The Kinderscout Grit Group is the most thoroughly investigated, although the other two incursions are similar, where there is a thick sequence with a threefold subdivision into turbidites, delta slope and delta top environments. Each phase of fill represents the first establishment of shallow water conditions in the area. All of these sediments are rich in feldspar and are considered to be of northerly derivation, (Sorby 1859, Gilligan 1920).

During the same period small volumes of sediment were derived from the southern Midland Landmass into the north Staffordshire Basin and Widmerpool Gulf. These are protoquartzites with little feldspar or mica and are petrologically quite different from the northerly derived felspathic sandstone. Here too, shallow water deltaic sediments are flanked by deeper water turbidites (Trewin and Holdsworth 1973). By the beginning of R2b, progradation from the south had established shallow water conditions around the southern margin of the basin extending just to the south of Leek, (Ashton 1974). Further to the north west, in the central and northern parts of the North Staffordshire Basin, deep water, mainly turbidite sedimentation persisted, (Ashton 1974). In the R1c - lower R2b north Staffs sequence palaeocurrents in protoquartzitic turbidites are towards the north east, whereas palaeocurrents in northerly derived felspathic turbidites are towards the south east, (Ashton 1974, and

personal observations). The latter orientation suggests derivation possibly over a sill around the western margin of the Derbyshire Block which was still a positive feature, and where a relatively condensed mudstone sequence was deposited between the E and lower R zones.

(Ramsbottom et.al. 1962).

9.2. Late R2b sequence in the southern Pennines

9.2.1. Thicknesses

Isopachytes for the R. bilingue late form to R. superbilingue interval in the southern Pennines are shown in Fig.54. In marginal areas, where the R. bilingue late form band has not been recognised, the R. bilingue ss. band has been used as the lower stratigraphic marker. This makes little difference to the thicknesses as the two horizons are normally close together.

The greatest amounts of sediment were deposited in an elongate north-west, south-east trending zone about 80 km long and 30 km wide. The sequence thins rapidly to the north, particularly between Buxton and Chapel-le-Frith. It also thins rapidly south westwards, although there is little detailed information outside north Staffordshire. Unfortunately there is no exposure adjacent to the thesis area to the west of the Red Rock Fault.

9.2.2. Local Variations

Regional variations in the vertical sequence in the thesis area are shown in Fig.18. The full vertical sequence with both deep water, and delta top associations, [Fig.55] is restricted to area A. Almost complete, but intermittently exposed sections through the whole sequence occur at Upper Hulme, [Section 4], the Dane Valley [Section 5], Cisterns Clough [Section 6] and further north in two stream gulleys near Buxton [Sections 7 and 8]. All of these localities are too wide apart to permit any detailed correlation.

At Upper Hulme turbidites in the Deep Water Association are exposed in the stream section below the village [Fig.15]. They pass abruptly into rippled sandstones of the Slope Association at the top of the waterfall below Dane's Mill (SK 0123 6122). The top of the Slope Association is cut into by a large fluvial channel. This and other channels are exposed in the nearby escarpments at Hen Cloud and the Roaches. There is no continuous exposure in the sequence between the top of the fluvial channels and the R. superbilingue band in this area.

In the Dane Valley, the Deep Water Association is exposed in the stream bed upstream from Hammends Hole (SJ 9544 6407) to the weir (SJ 9570 6421). Above this the sequence is repeated by complex folding and faulting. The top of the Deep Water Association occurs below the bridge at Danebridge (SJ 9654 6417). A thick discontinuously exposed section in the Main Slope Association occurs upstream. As explained previously, this part of the sequence has probably been disturbed by syn-sedimentary growth faulting but the actual faults are not exposed.

The top of the slope is cut out by a large fluvial channel at Gibbons Cliff probably at a higher level than the channel above Upper Hulme. This channel consists entirely of a coset of Medium Scale Cross-bedding exposed in the gorge at Allgreave Wood. A coarsening upwards sequence in the Delta Margin Association above the channel occurs at the entrance to the gorge (SJ 9722 6681) and is repeated by faulting upstream from Allgreave Bridge (SJ 9699 6739) [Fig.51B].

None of the stream sections north of this area exposes the Deep Water or Slope Associations. Sections in the Association D, occur in Cumberland Brook (SJ 993 699) [Fig.51.A] and further east in the River Dane below Culthorne Hill (SK 00 67).

A long section through the complete sequence occurs in Cisterns Clough [Fig.18]. The Deep Water Association is poorly exposed, but mapping suggests that the turbidites are much thinner here than further south at

Upper Hulme. Most of the lower part of the Slope Association is now concealed by the new roadworks. The upper part of the slope sequence consists mainly of ripple laminated sandstones, above this a thick coset of medium scale cross bedding represents one or more channels in the Delta Top Association.

In the two sections at Buxton the lower part of the Deep Water Association is dominated by LF 6 parallel sided sandstones. Turbidites appear higher up the sequence forming the lower part of the escarpment at Corbar Hill. They thin out rapidly towards the north. A well developed slope sequence is again present though it appears to be considerably thinner in the River Wye section above Burbage Reservoir (SK 036 721), [Fig.18]. Possibly hidden faults or folds occur in the thick unexposed section.

The Delta Top Association extends westwards as far as Rudyard and northwestwards into Macclesfield Forest. Beyond this it disappears but the actual zone of disappearance is virtually unexposed.

In area B there is only one long section through the sequence [Fig.18-3] and this is very intermittently exposed. No turbidites occur in the Deep Water Association here but there are many gaps. The Delta Slope Association extends to near the top of the sequence close to a mudstone with fish fragments which is probably very near to the horizon with the R. superbilingue band. Good sections in the Slope Association occur at Bosley Works (SK 9138 6477) and in Ravens Clough Wood (SK 915 631) which is mostly mudstone. The R. superbilingue band occurs above this sequence at (SK 9106 6321).

In the southern part of area B poor sections in the Deep Water Association occur in the stream running through Lee Wood (SK 92 63) where turbidites come in above the R. eometabilingue band, and the stream below Woodside Farm (SK 9053 6460) where turbidites pass up into ripple laminated sandstones of Association B.

Further to the north around Macclesfield a very thick turbidite sequence occurs in the Deep Water Association. These are the Walker Barn Grits of Evans et.al. (1968), and are well exposed in quarries around Walker Barn and in the stream section above Lamaload Reservoir (SJ 96 75). Above this ripple laminated sandstones of the Slope Association occur in the River Dean (SJ 9600 7560). The upper part of the Association is not exposed.

Turbidites in the Deep Water Association extend northwards into the Todd Brook area gradually thinning out.

In area D, in which there is just one long section, Ridge Clough (SK 06 79) [Fig.18] the sequence thins to about 200 m in the thesis area. Turbidites are absent from the Deep Water Association which consists mainly of LF 6 Parallel Sided Sandstones. These probably extend up into the Slope Association. Above this the slope contains alternations of lenticularly bedded and ripple laminated sandstone. A fluvial channel with a coset of Medium Scale Cross-bedding caps the sequence. The upper part of the succession is not exposed.

In area C, there are two long sections in Shirley Hollow [Fig.18:2] and near Cotton [Fig.18:1]. Here the sequence has thinned substantially to about 60 m. The Deep Water Association is absent and a Slope Association of about 40 m passes up into a thin fluvial channel sequence (The Shirley Hollow Sandstone of Morris, 1969) with Medium Scale Cross-bedding. The Abandonment Association exposed in Shirley Hollow contains a seat earth and very thin coal at the base followed by a mudstone sequence up to the R. superbilingue band.

In east Derbyshire, the late R2b sequence, dominated by the Ashover Grit shows a similar vertical subdivision with turbidites passing up through slope sediments into delta top fluvials (J.I. Chisholm p.com.). Palaeocurrents show a mean towards the north west.

9.2.3. Source of the sediment

The sandstones of the Roaches Grit Group, and of the Ashover Grits are rich in felspar quite different from the southerly derived protoquartzites [Fig.56], but similar to the northerly derived felspathic sandstones. The palaeocurrents, however, are from the south east, whereas in the three previous major phases of fill palaeocurrents are from the north east [Fig.56].

This apparently anomolous direction has led to a difference of opinion over the last 15 years as to the source of the sediment in the Roaches and Ashover Grits. Mayhew (1967) suggested that the Ashover Grit was derived from the Midland Landmass which was uplifted at this time. This uplift also temporarily established a south to north dipping palaeoslope. Stevenson and Gaunt (1971) also suggested a southerly derivation. Harrison in Evans et.al. (1968) and Holdsworth (1963) on the other hand suggest that the Roaches Grit Group is of northerly derivation on the basis of petrological similarities to the earlier Kinderscout and Skipton Moor Grits. They do not explain the apparently anomolous palaeocurrent directions.

The present investigation of the Roaches Grit Group has shown that the delta top sequence, particularly the large fluvial channels, is similar to the Kinderscout Grit Group. The channel fill sediments are very unusual, and at present seem to be unique to the Central Pennines Namurian. The channels are very large, probably 1 km or more in width. The Midland Land Mass is believed to have been quite small during the Namurian, with a southern margin running through South Wales across to Kent, (Ramsbottom 1971). In modern rivers there is a fairly close relationship between channel size and the size of the catchment area, (Leopold et.al. 1964). It is very unlikely that the Midland Land Mass was large enough to serve as a catchment area for the river which deposited the Roaches Grit.

Three separate lines of evidence, petrology, nature of the channel fill, and size of the channels, all argue against a southerly derivation, and suggest that the sediments in the Roaches Grit Group, together with

the Ashover Grits were, like the older Kinderscout and Skipton Moor Grits, derived from the north.

9.2.4. Controls on sedimentation

The pattern of sedimentation was largely controlled by the earlier sedimentary fill. The sequence thins rapidly to the north because here the basin had been filled during the earlier R_{1c} Kinderscoutian fill phase. Similarly there is a rapid southerly thinning against the shallow water lower and middle Namurian protoquartzite sequences. In between these two areas lay a narrow deep water trough which was filled in by the "Roaches Grit and Ashover Grit delta". By the close of R_{2b}, shallow water conditions had been established everywhere, except possibly in the extreme north west of the basin around Macclesfield and Todd Brook.

Palaeocurrents are parallel to the axis of the trough as the delta prograded from the south east, roughly normal to those in the older proto-quartzites and the reverse of those in K felspathic turbidites of R_{1c} and early R_{2b} age. This unusual palaeocurrent direction is a function of the shape and orientation of the deep water trough into which the delta was building plus the effects of slight tectonic movements along the southern margin of the basin. Although the Midland Land Mass was no longer a supplier of sediment in the north, it was not completely submerged and still supplied sediment to the South Wales Basin (Kelling 1974). The R_{2b} sequence is absent south-east of a line between Stoke and Stafford. The offsetting of channel fills at the Roaches, described earlier in Chapter 6, suggests there was a slight tilting away from the Midland Land Mass towards the north. As the river entered the basin from the north-east, its course was diverted somewhere beyond the area of present outcrop and the channels became aligned nw/se, the delta beginning to prograde from the south-east along the axis of the trough, [Fig.57].

The deeper part of the trough was rapidly filled with sediment and this would compact relatively quickly. To the north and south, a much

thinner sequence was deposited over the older sedimentary fill, which by this time, would already have been relatively well compacted. As a result, subsidence would have been greatest in the deeper parts of the trough. The river channels will have tended to follow the line of greatest subsidence and will become aligned parallel to the trough axis. This last phase of progradation is now exposed as the Roaches Grit Group and Ashover Grits.

The outline of the Derbyshire Block, has no effect on the isopachyte pattern [Fig.54] nor the palaeocurrents. Vertical sections on the edge of the Block near Buxton (sections 7 and 8, Fig.18) are similar to those further south in the central part of the basin. By this time the Block had ceased to be a positive structural feature.

9.3. Regional significance

The Roaches Grit Group represents the fourth, and last stage of fill in the Namurian Central Province Basin. The vertical sequence is similar to those of the earlier phases with a thick turbidite and delta slope succession preceding the establishment of shallow water delta top conditions. This type of sequence generally only occurred where deltas built out into a deep water part of the basin, and for this reason the location of these coupled turbidite-delta systems is closely controlled by the earlier sedimentary fill [Fig.58].

The sediments of each phase are felspathic and are all probably of northerly origin. Following an early influx in the Eic zone, most of the fill took place considerably later during the R zone. It took the greater part of the Namurian period which lasted for 12m years (Ramsbottom 1969) to finally fill the basin.

Although the structural morphology of the basin had an important influence on sedimentation facies and thicknesses during the early part of the Namurian, (Ramsbottom 1971, Trewin and Holdsworth 1973, Baines p.com.),

this became less important as the basin was filled. The Derbyshire Block had no effect on sedimentation of the Roaches Grit Group and Ashover Grits, and the Craven Fault appears to have had little effect from E2 (section in Ramsbottom 1971). It appears that during the Namurian there was little differential subsidence between block and basin and sedimentation gradually blanketed the irregular morphology.

In none of the sequences described is there any evidence of tidal activity and it may now be safely concluded that the basin was largely tideless. Wave reworking of the sediments was also limited, although the upper parts of the 'Roaches Grit delta' show evidence of limited wave activity.

McCabe (1975) found no evidence for sea level changes in the Kinderscout Grit Group, and none has been found in the present investigation. The transgression, and re-establishment of fully marine conditions at the top of the Roaches Grit Group probably resulted from a major abandonment of the delta system. The reason for these four major influxes of sediment is unknown but it seems unnecessary to invoke sea level changes. Climatic or tectonic changes in the source area, or major shifts in the courses of the rivers could lead to periodic influxes of large amounts of coarse sediment into the Central Pennine area.

Section III

THE SEDIMENTOLOGY OF THE LOWER PENNANT MEASURES IN THE RHONDDA VALLEYS

Chapter 10

10.1. Introduction

The area studies lies between Pontypridd in the south east and Glyn Neath in the north west, along the valleys of the Rhondda Fawr, Rhondda Fach Taff and Gawr. The long road section at Earlwood Roundabout near Briton Ferry has also been examined [Fig.59]. Most of the exposure occurs in quarries and road cuttings along the valley sides, supplemented by rarer stream sections.

The main area lies in the central part of the South Wales Syncline and the rocks are horizontal with little subsidiary folding. Faulting splits the sequence into a large number of structural blocks typically hundreds of metres in size, and in some of the larger exposures the structure allows lateral correlation of the vertical sections. The Earlwood Roundabout section lies on the southern limb of the South Wales Syncline and the sequence dips to the north. This provides a continuously exposed section of 130 m. In the main area, most sections are typically 10 - 30 m in length.

The Lower Pennant Measures, as defined by Woodland, Evans and Stephens (1957), occur between the Upper Cwmgorse Marine Band at the base, and Hughes Coal at the top. They correspond to the zone of Anthraconauta phillipsi in the upper part of Westphalian C. [Fig.60.] Within the succession, there are no biostratigraphical marker horizons and correlation is based on the principal coal seams [Fig.60]. These have now been correlated over wide areas of the South Wales Coalfield and within the comparatively small area of study these correlations appear to be reliable.

Three subdivisions of the Lower Pennant Measures are recognised, the Llynfi, Rhondda and Brithdir Beds [Fig.60]. The succession averages

550 m in thickness in the area of study, thinning to the east and thickening to the west.

10.1.2. Previous Research

A summary of previous research on the stratigraphy and structure of the area is given by Woodland and Evans (1964). Work on the sedimentology and palaeogeography of the Westphalian of South Wales has shown that this was deposited in a partially restricted east-west trending basin bounded in the north by the Midland Land Mass and to the south by a large 'Hercynian Continent', (Wills 1951, Calver 1969). This basin probably connected with open sea in the west and extended to the east through Kent and northern France into Belgium and Holland.

The Westphalian A and B succession is broadly similar to other British 'Coal Measure' sequences with numerous small coarsening-upwards cycles with marine or non-marine faunas at the base and coals at the top, (Woodland and Evans 1964). These have been interpreted as largely deltaic in origin, with the occasional development of alluvial flood plain environments, (Williams 1966, Reading 1971, Kelling 1974). Sediment supply was predominantly from the north during Westphalian A, but during Westphalian B, southerly supply became increasingly important. Isopachytes show that the thickest part of the sequence lay in Swansea Bay during Westphalian A and between Swansea and Ammanford during Westphalian B. The basin of deposition in the area of the main South Wales Coalfield was aligned north-west, south-east, slightly oblique to the orientation of the present structural basin, (Owen 1964, Thomas 1974).

At the beginning of Westphalian C, a profound change in the nature of sedimentation occurs. The "cycles" in the Pennant Measures are much thicker than earlier in the Westphalian, consisting mainly of thick sandstones with conglomerate bands, fining upwards into mudstones with coals near the top, (Woodland and Evans 1964). As will be explained in more detail, later, this type of sequence is not established simultaneously

everywhere, and coarsening upwards 'Coal Measure Type' cycles occur locally, particularly in the Llynfi Beds.

The sedimentology at the Pennant Measures has been investigated by Bluck and Kelling (1962), Kelling (1964, 1968, 1969, 1974), Reading (1971) and Stead (1974). These workers have mainly concentrated on the better exposed Rhondda Beds, but Stead (1974) has also worked on the higher Brithdir Beds, and stratigraphic equivalents in the Forest of Dean and Bristol Coalfields. They have shown that the Pennant Measures may be subdivided into two quite distinct lithofacies, a subgreywacke facies, known locally as the 'Blue' Pennant and an 'orthoquartzite' facies restricted to the northern and eastern parts of the present structural basin. The orthoquartzite lithofacies was earlier considered to be fluvial in origin, (Bluck and Kelling 1962, Kelling 1964, 1968) but more recently Kelling (1974) and Stead (1974) have suggested that it is a marine deposit. This re-interpretation may be questioned on several grounds and will be discussed later.

All the workers agree that the subgreywacke lithofacies is mainly fluvial in origin and derived from a landmass, sometimes called the 'Cornubian Landmass', to the south of the basin of deposition. This conclusion is accepted in the present study which has concentrated solely on the subgreywacke lithofacies.

Isopachytes in the Rhondda Beds and entire Lower Pennant Measures (Woodland and Evans 1964, Thomas 1974), show a similar thickness pattern to the preceding Westphalian A and B succession with the locus of deposition in the Swansea Bay area. Palaeocurrents in the Rhondda and Brithdir Beds, show that sediment transport was predominantly from the south-east towards the north-west, roughly parallel to the trend of the isopachytes (Kelling 1974, Stead 1974). Stead, in addition suggests that during the deposition of the Rhondda beds a shoreline existed in the eastern part of the Coalfield,

aligned roughly normal to the palaeocurrent trend and running roughly through the centre of the present main coalfield. Open marine conditions persisted to the north-west, passing upcurrent into fluvial environments to the south-east. This interpretation is not supported by the present study.

Within the fluvial sediments in the Rhondda Beds, Reading (1971) commented on the regular fining upwards pattern seen in some channel fill sequences and suggested that meandering rivers were mainly responsible. Kelling (1964b, 1968, 1969, 1974) made a more extensive facies analysis of the Rhondda Beds and recognised a fully developed fining upwards cycle following the technique of Allen (1964a, 1965). This cycle has five principal components. A basal log conglomerate member with ironstone pebbles, shale flakes and plant fragments is overlain by the main sandstone member which has an upward gradual or abrupt decrease in median grain size, generally near the top. The lowest part of this sandstone member may be massive or irregularly bedded commonly containing numerous logs. This is followed by cross bedded sandstones with sets decreasing in size upwards. Above this are units of flat-bedded and then ripple laminated sandstones. The third member is termed the siltstone mudstone complex and often contains a series of ripple-laminated, or cross-bedded fining upwards units. Finally there are capping members of seat-earth and then coal.

Not all cycles contain the full sequence. In particular the basal conglomerate member and massive and melange units of the sandstone may be absent. Kelling suggested that these formed in the deepest or axial parts of non-meandering channels and laterally banked up against steep channel sides. Thus sections starting higher up in the channel fill sequence at the sides of the channel, would not contain them. The overlying unit of cross-bedded, flat-bedded and ripple-laminated sandstone is interpreted as the point bar complex of a meandering river. The siltstone-mudstone unit is regarded as either the fill of an abandoned channel or sometimes as a

true flood-plain deposit the smaller fining upwards cycles originating as crevasse-splays. Local deviations in cross bedding orientation are regarded as the result of the variable alignment of meandering channels. The thickness of these cycles varies from 2.6 to 35 m. Kelling regards the thicker cycles as resulting from several stacked channel fills resting one above the other. He does not suggest a typical thickness for single channel fills. In the southern part of the basin the proportion of finer grained sediment is much less. Here Kelling suggests the rivers were braided or of low sinuosity.

In an earlier study, Bluck and Kelling (1962) demonstrated the abundance of preserved channel morphologies within the Pennant sequence. These generally have a south-east to north-west alignment parallel to the cross bedding vector mean. They claimed to recognise four types of fill; side fill, vertical fill, downfill and upfill.

The present investigation disagrees with both Kelling's description of the nature of the channel fill sequences and with his interpretation of the type of river responsible. His earlier general ideas on the regional palaeogeography are accepted as a framework for the detailed local study, although the recent reinterpretation of the ortho-quartzite lithofacies (Kelling, 1974; Stead 1974) seems questionable.

10.2. Lithofacies description and interpretation

A separate lithofacies classification has been devised for the Lower Pennant Measures. A comparison with the lithofacies types defined by Kelling (1968) is shown in Table VIII. Two facies associations occur: Association A, largely restricted to the Rhondda and Brithdir Beds consists mainly of sandstones in thick, broadly fining upwards units with subordinate finer grained sediments. These are interpreted as fluvial channel and channel overbank sediments.

Association B, largely restricted to the Llynfi Beds consists of small scale coarsening upwards cycles, sometimes with very small fining upwards

channel fill sequences at the top. These are interpreted as the sediments of small deltas or crevasse deltas. The detailed relationships of the two associations are discussed later.

With four exceptions, the lithofacies are not particularly unusual and the aim is to give them only a brief description. The chief purpose is to look in detail at the facies relationships.

10.2.1. Association A Fluvial Channel Sediments

10.2.1.1. Lithofacies 1, Tabular Cross-bedding [Fig. 61A]

Medium grained generally well sorted sandstones in sets up to 3.5 m thick [Plate 64]. These form cosets up to 5 m thick, usually containing only a few individual sets. The basal erosion surfaces to the sets are flat [Fig. 61] sometimes with lags of plant material. Foresets may be angular or tangential and may contain multiple convex upwards internal erosion surfaces [Chapter 2]. Regressive intrasets do not occur. Individual sets may be traced for tens of metres in the XY plane without substantially altering in thickness, and then are usually cut out by a differing overlying facies.

This lithofacies is quite rare.

Interpretation

Tabular cross bedding is deposited by large scale ripples termed sandwaves. As explained in Chapter 2, these may be either straight crested or linguoid shaped. The presence of multiple convex erosion surfaces shows that sometimes smaller bedforms were superimposed on their backs [Chapter 2].

10.2.1.2. Lithofacies 2, Trough cross-bedding [Fig. 61c]

Fine to medium grained sandstones, generally well sorted. They form erosively based cosets 1 - 10 m in thickness which may be often traced many tens of metres and sometimes hundreds of metres in the XZ and XY planes. Set thicknesses vary considerably from 5 cm to 2½ m but most fall in the 20 - 75 cm size range. Larger sets have troughs up to 20 m in width [Plate 65]. Within cosets there is sometimes a progressive upwards increase in average set thickness by three or four times. Upwards

decreases in set thickness of a similar magnitude also occur very occasionally. Within cosets, set thicknesses are generally fairly similar, but occasionally set sizes appear to vary by a factor of up to three. Unfortunately, the structures are rarely sufficiently well weathered to determine set sizes accurately.

Sometimes the trough form is preserved and filled in with siltstone, as at the roadcutting near Blaenrhondda (SN 9310 0050) [Plate 85] or by an alternation of siltstone and ripple laminated sandstone as in Nant y Gwair (SS 9101 9839), [Section in Fig.64]. The profile of a dune may also be preserved by a siltstone drape as north of Craig yr Hesg, (SS 9229 0153), [Plate 84].

This lithofacies is very common in Association A. When exposed only in the XY plane, however, large sets of trough cross bedding are difficult to differentiate from tabular sets [Plates 65, 66].

Interpretation

Trough cross-bedding is usually formed by sinuous crested bedforms, dunes. Harms and other (1975), suggest that most dunes are no larger than 1 m in height. Clearly the bedform responsible for the larger set sizes in this lithofacies were much bigger. Trough shaped erosion surfaces are known to occur in front of much bigger sandwaves (see detailed discussion in Section 10.2.1.3.) and it is possible that the large sets of trough cross bedding were not formed by dunes.

10.2.1.3. Lithofacies 3, Cross-bedding with undulating basal erosion surfaces [Fig 64c]

This lithofacies is an intermediate form between the types described in sections 10.2.1. and 10.2.2. It may nevertheless be clearly differentiated in sections through the YZ plane.

It consists of fine to medium grained sandstone in cosets of cross-bedding with individual sets usually 10 cm to 1 m thick. In the XY plane the basal erosion surfaces are flat [Figs. 62, 63, Plate 68] and the

lithofacies is indistinguishable from Lithofacies 1, individual sets extending for tens of metres. In the YZ plane the erosion surfaces occur in a series of shallow troughs usually no more than 10 cm in depth [Fig. 61, Plate 69]. These have low angle sides and unlike LF 2 are not intersecting. The overlying cross-lamination is parallel to the erosion surfaces in the XZ plane, and slightly tangential in the XY plane.

It is difficult to assess how often this lithofacies occurs as it is not easily distinguishable except in good three dimensional exposures, but it appears to be more common than LF 1 but less common than LF 2.

Interpretation

Recent flume and field studies of large scale ripples have demonstrated that many of these are of two main types, dunes and sandwaves (Harms and others 1975). The differences have been outlined in previous Chapters. In many ancient fluvial sequences there appears to be a clear distinction between tabular and trough cross-bedding, and it is probable that in many cases these structures are the deposits of these two types of bedforms. In not all modern environments is there such a clear distinction. Coleman (1969), for example, records three types of large scale ripple from the Brahmaputra, 'megaripples', 'dunes' and 'sandwaves' of which the 'dunes' and 'sandwaves' ranging from 1.5 m to 17 m in height, compare in size to the sandwaves of Harms and others (1975). However the 'dunes' have very irregular crestlines and well developed scours in the lee. Even the larger 'sandwaves' develop leeside spoon shaped scours at high discharges. Boothroyd (1969) also notes the development of scour pits in front of tidal sandwaves and Harms and Fahnstock (1965) note a similar development in the Rio Grande.

Thus it seems doubtful that sandwaves (Harms and others 1975) will always give tabular cross bedding with flat erosion surfaces. Lithofacies 3 is interpreted as the deposit formed by sandwaves which developed shallow scourpools in the lee. It is also possible that some of the larger sets of LF 2 Trough Cross-bedding have a similar origin.

10.2.1.4. Lithofacies 4, Parallel-laminated Sandstone

Fine to medium grained sandstones in parallel laminations generally 0.25 - 1.0cm thick. The Laminations are picked out by alternations of quartz rich and carbonaceous layers. Large logs are completely absent. Sometimes the lamination may be markedly undulating and often trough shaped hollows up to a few tens of centimetres deep are cut into the parallel laminations. These occur at many levels through the bed. [Plate 70]. Filling laminae parallel the sides of the trough and are truncated by succeeding parallel lamination.

This lithofacies occurs in beds up to 5.75 m thick and individual beds have been traced laterally for 175 m. Usually, however, beds are generally less than 1 m in thickness and are cut out laterally by overlying lithofacies. It also occurs in beds parallel to channel sides where it is inclined to horizontal at angles of 5 - 10° [Fig.6&].

Interpretation

Parallel lamination is formed at moderately high flow powers in the lower part of the upper flow regime (Allen 1964b). Presumably at these velocities all the larger plant material is swept away giving rise to a much cleaner, better sorted sand than in much of the cross-bedding. The presence of trough shaped erosion surfaces shows that the flow fluctuated and occasionally lower flow regime conditions were established. Simons and Richardson (1965), have noted in flume experiments how the flow may alternate between upper and lower flow regime conditions, large scale ripples temporarily forming and then being washed out. It is also possible that temporary obstacles such as logs may cause local scour pits even though there is no evidence of them to be seen.

10.2.1.5. Lithofacies 5, Ripple-laminated Sandstone

Fine to medium grained sandstones in erosively based beds 10cm to 1.5 m in thickness. Ripples may be straight crested or linguoid in plan and lamination is picked out by fine carbonaceous debris. Ripple cross-

laminated sets are normally 1-2 cm thick and ripple drift occurs rarely. Sometimes ripples may occur within larger troughs of LF 2 which are largely unmodified.

Interpretation

Ripple lamination forms at moderate flow powers in the lower part of the lower flow regime of Simons and Richardson (1961).

10.2.1.6. Lithofacies 6, Structureless Sandstone

Sandstones medium and occasionally coarse grained. At outcrop they appear completely structureless except for very rare flat parting planes. X-radiographs show a poor preferred grain orientation but no lamination [Plate 71]. Usually the sand is very well sorted with little plant material, but occasionally stringers of ironstone pebbles or small logs occur.

The beds are always erosively based, some having a relief of up to 6 m with a distinct channel form, [Plate 72]. The erosion surfaces vary in inclination from the usual 5 - 10°, up to about 60°, and sometimes they are extremely irregular [Plate 73]. Usually they are cut into other sand sized sediments. Log conglomerates are only rarely developed and are generally thin. Bed thicknesses range from 0.5 m to 8 m.

Interpretation

Structureless sandstones may be deposited in several ways and various theories have already been outlined in Chapter 3. Kelling (1968) suggested that the beds may be indicative of antidune deposition, following Collinson's original interpretation of supposed structureless beds in the Namurian of the Central Pennines. However, the faint undulatory lamination described from antidune deposits in flumes, (Middleton 1965) and from probable antidune deposits in ancient fluvial and turbidite sequences (Skipper 1971, Walker 1967 and Hand et.al. 1969) have not been seen.

Antidune sediments are unlikely to be common in fluvial sequences, for except in very shallow parts of the channel, the Froude Number will

be too low for antidunes to occur. Most of the theories of structureless sandstone deposition discussed in Chapter 3 refer to density current deposits and are not applicable here. The ever-present erosive bases, often cut into sandstone at angles greater than the angle of rest of unconsolidated sand, suggest that erosion was immediately followed by rapid deposition. As in LF 16 from the Roaches Grit Group, the irregularity of the erosion surfaces may point to large scale turbulence, but there is no evidence here that this occurred in the lee of large bedforms.

Traces of parallel lamination occur within LF 6 and rapid deposition in the lower part of the upper flow regime as discussed by Collinson (1970) seems the most likely mode of origin. This is supported by the scarcity of plant material, which as in LF 4 was probably swept away by the powerful current.

10.2.1.7. Lithofacies 7, Conglomerate

This lithofacies is common in Association A, and consists of two distinct lithotypes. In the first of these the sediment contains large logs of Calamites and Lepidodendron. Sometimes just the bark remains, but commonly the complete stem is preserved [Plate 74]. Some of these may be very large indeed and at Taff Vale Quarry one log is 10 m long. More usual lengths range from 0.25 - 1 m. The logs may be aligned at any angle to the vertical and the associated sediment is usually very irregularly bedded [Plate 75]. Often the logs are aligned sub-horizontally and Bluck and Kelling, (1962), demonstrated a marked preferred orientation close to the vector mean of associated cross bedding. Found with the logs are subrounded pebbles of vein quartz, ironstone and mudstone typically 2 - 3 cm in diameter the whole occurring within a matrix of medium to coarse sand.

The second lithotype, volumetrically much less important is a conglomerate of vein quartz, ironstone and clay pebbles identical to

those in the first lithotype, and, similar in size and shape. These are distributed in a matrix of medium to coarse sand, and poorly bedded, but plant material is completely absent. [Plate 76]. Typically this lithotype occurs in beds 25 - 50 cm thick alternating with thicker beds of lithotype A. In the section north of Craig yr Hesg (SS 9229 0153), four separate beds of lithotype β alternate with beds of type A in a conglomerate unit 4.85 m thick.

Within the conglomerates it is common to find patches of cleaner sandstone a few tens of centimetres in diameter which contain traces of lamination or cross lamination. Near Craig yr Hesg the clean avalanche foresets of a set of LF 2 pass down dip into foresets covered in small plant fragments. After only 7 m the foreset lamination becomes completely unrecognisable passing into the typical irregular bedding of a plant conglomerate.

Beds of LF 8 Laminated Siltstone sometimes occur within the conglomerates. In one of these, near Craig yr Hesg, the lamination has been bent post-depositionally into an S shape, [Plate 77].

The conglomerates vary in thickness from 6 m down to about 10 cm eventually grading into mere plant stem lags at the bases of sets of cross-bedding. Typically beds are 0.5 m to 2 m thick and may be traced laterally for up to about 100 m. They invariably thin out usually grading into LF 2 Trough Cross-bedding. Typically the plant stems remain as trough lags in the transition area. The 4.85 m thick conglomerate found just above the No.2 Rhondda Coal near Craig yr Hesg passes laterally into a fairly clean cross bedded sandstone over a distance of only 20 m.

Interpretation

As explained later in Section 10.4 many of the conglomerates demonstrably occur at the bases of major channels. Lithotype 6 compares to some extent with the conglomerates described by Allen (1962) from the fluvial channel facies in the Old Red Sandstone although intraformational

clasts are rare. He interpreted them as channel lag deposits. Continued winnowing of the finer sediments in the deepest part of the channel concentrates the coarsest part of the bedload. Similar lags occur at the bottoms of modern river channels (Happe et.al. 1940, Fisk 1944).

Lithotype A is more unusual with its large amounts of plant material often of considerable size. Here the lateral passage into cross bedded sandstone is critical to the interpretation. Many beds were probably originally cross bedded with subsequent plant collapse destroying most of the bedding. The bed of mudstone shown in Plate 77 seems to have clearly been deformed by plant collapse in the underlying sediment.

10.2.1.8. Lithofacies 8. Laminated Siltstone

Grey micaceous silty mudstone, with small plant material and light coloured laminations of quartz silt 1 - 5 mm thick. These are usually gradationally based. Sometimes they thicken laterally into small ripple form sets. This lithofacies occurs in beds up to 5 m in thickness.

Interpretation

The features of this lithofacies suggest deposition from suspension in quiet water, with short lived periods of traction forming isolated sets of ripple cross-lamination.

10.2.1.9. Lithofacies 9. Seat-earth and Coal.

This lithofacies is common in the Lower Pennant Measures. Seat-earth are commonly un laminated, with listric surfaces and ironstone nodules. Stigmarian roots are very common. The thickest coals occur in the Rhondda and Brithdir Beds, and several have been worked, particularly the No.2 Rhondda which is still worked by a small drift mine at Blaenrhondda. Thicknesses range from a thin smut up to 0.5 - 0.75 m. The No.2 Rhondda is thicker reaching 2.8 m. Most of the major coals extend across the research area and over wide areas of the South Wales Coalfield, although over distances of more than a few kilometres they commonly split into

several seams (Woodland and Evans 1964, Downing and Squirrel 1969).

Thinner coals occur locally, [Plate 78] but are generally cut out by overlying channel sandstones.

Interpretation

This lithofacies provides evidence of periods when vegetation was widespread, probably at times when the sediment supply was locally reduced. Further discussion will be given when the origin of the Pennant sequences is discussed [Section 10.5].

10.2.2. Association B, Delta or Crevasse Delta Sediments

Most of the lithotypes present in Association A are also present here, but thick channel fills are rare. The sediment is generally finer grained and cross-bedding sets are smaller. With the exception of the Tormyndd and No.3 Rhondda, the coals are only locally developed. In addition, four further lithofacies are present.

10.2.2.1. Lithofacies 10. Ripple-laminated Siltstone and fine Sandstone

Cross laminated coarse siltstones and fine sandstones in sets up to 10 cm thick. The lamination consists of alternations of silt or sand rich laminae with mud rich laminae. Sets are often gradationally based merging into flat or wavy laminated silty sandstone. The tops of the ripples may be draped by wavy laminated silts a bit like in ripple drift cross-lamination. This form is gradational into LF 11, Laminated Coarse Siltstone [Plate 80]. It bears some resemblance to the 'Silty Streak Facies' of De Raaf, Reading and Walker (1965). Sometimes the sediment resembles small scale trough cross bedding [Plate 79].

Interpretation

This lithofacies is unusual and its origin uncertain. It bears some resemblance to wave ripple lamination, but the range of grain size and high proportion of mudstone makes a wave origin unlikely. It was probably deposited fairly rapidly with a high fall out from suspension of relatively fine grained sediments. The large size of some of the cross sets suggests that at times small 'dunes' developed.

10.2.2.2. Lithofacies 11. Laminated Silty Sandstone. [Plate 80]

Sandstone and coarse siltstone in flat to slightly wavy laminations with gradational boundaries interbedded with silty mudstone. This lithofacies has a characteristic weathering pattern in the field resembling corrugated sheeting [Plate 81]. Beds are gradational and typically 0.5 - 1 m thick. Commonly the lamination bends locally into cross-lamination and this lithofacies interdigitates with LF 10. In coarser developments the bedding may occur in erosively based sets inclined at a few degrees to the horizontal.

Interpretation

The characteristics of this lithofacies suggest deposition from suspension probably at a fairly rapid rate. Traces of ripple cross-lamination point to short-lived periods of bed load transport.

10.2.2.3. Lithofacies 12. Unlaminated Siltstone

Light to dark grey siltstone and silty mudstone. The sediment is poorly sorted and unlaminated, usually breaking with a conchoidal fracture. Siderite nodules are ubiquitous. It may contain roots where it has then been classified as a seat-earth, but most beds lack the distinctive listric surfaces of the seat-earths.

In favourably weathered exposure this lithofacies is quite distinctive from LF 8 Laminated Siltstone and usually more resistant to erosion. Where weathering is less favourable, particularly in wet stream sections the two appear more similar, particularly if LF 8 contains rare ironstone nodules. For this reason the two lithofacies cannot be differentiated in the vertical sections.

Interpretation

LF 12 clearly represents a distinctive environment but its nature is unknown at present. The features of the sediment suggest slow deposition from suspension with less fluctuation in current strength than in LF 8. Possibly it was deposited further away from the sediment supplying

fluvial channels, an argument supported by its position in the vertical sequences [Section 10.3]. The significance of the high iron content is also uncertain at present but may reflect a slower rate of deposition.

10.2.2.4. Lithofacies 13. Mudstone

Dark grey mudstone, sometimes laminated, sometimes unlaminated and breaking with a conchoidal fracture. It is commonly unfossiliferous but may contain bivalves, (Anthraconauta phillipsi) or fish scales.

Interpretation

This lithofacies was deposited from suspension in quiet water. Further discussion will be given later.

10.3. Description and interpretation of vertical sequences in Association B.

Four detailed vertical sections in the upper part of the Llynfi Beds are shown in Figure 63. The No.2 Rhondda Coal at the top of the sequence provides a good basis for correlation. Six major cycles may be recognised in Section C and the top four of these can be correlated into the other three sections, by means of coals and seat-earths. The cycles vary from about 5m to 15m in thickness. Many contain a lower coarsening upwards unit and upper fining upwards unit of approximately the same thickness. Good examples are the sequence between the 39 m and 50 m levels in Section C [Fig.63], and the sequence between the 16 m and 26 m levels in Section A. Other cycles defined by seat-earths or coals are more variable. The sequence between the 5 m and 16 m levels in Section for example cannot be subdivided easily.

The coarsening upward components frequently start above a coal with a unit of LF 13 Mudstone, sometimes with a bivalve fauna. This passes up into LF 12 Unlaminated Siltstone. The higher parts contain ripple lamination, either in sharp based cosets of LF 5, or frequently in gradationally based beds of LF 10, (shown as ripple drift in the sections) interbedded with beds of LF 11, Laminated Silty Sandstone. In none of

the sections is there a regular gradation of grain size change up through the sequence, sharp variations and local decreases in grain size being characteristic.

The transition to the fining upwards part of the cycles is usually marked by sharp based sandstones many of which are probably channel fills. These are very variable in size, the largest channels occurring between the 5 m and 15 m levels in Section A. Here at least two channel fills appear to be stacked one upon the other, the lower channel resting directly on thin Coal. This sequence is much more reminiscent of Association A, the typical Pennant Sandstone facies, but the sediment is finer grained. Clearly it represents an unusual, major channel sequence within the Llynfi Beds in this part of the basin.

Most of the sharp based sandstone units are much smaller with bases which are either flat, or with a relief of less than 0.5 m some having flutes [Plate 81]. The sandstones are typically 2 - 3 m thick often with a unit of LF 5 Ripple-laminated Sandstone at the base. Although all fine upwards in their upper part the coarsest sediment and largest sized sedimentary structures are not always at the base. Exposure is too limited to recognise channel forms and some sharp based beds may have a different origin. The uppermost parts of the cycles fine upwards into seat-earths which are often capped by coals. Palaeocurrents are mostly towards the northwest, as they are in other similar cycles not shown in Fig.63.

Interpretation

Within the sequences there is absolutely no evidence of wave or tidal activity. Faunas are non-marine (Calver 1971) and restricted to a few local horizons; bioturbation is very limited. The nature of the definite channel fills and the strong preferred current direction suggests that these are fluvial. Most of the channels appear to be fairly small

and cannot be correlated between the sections, although the overlying seat-earths and coals are widespread.

The sequences suggest deposition by small, lobate high constructive deltas or crevasse deltas building into brackish water. A modern example is the Colorado delta in Matogora Bay, Texas, (Kanes 1970). This delta grew rapidly into a shallow bay protected from wave and tidal influence by a barrier island, the Matagorda Peninsula. Prior to formation of the delta the water depth in the bay was only 2 m.

The Colorado River splits into numerous small distributary and crevasse channels, indeed it is not possible to meaningfully distinguish between the two types. Sections through the delta show a variety of sequences depending on their position relative to the channels (Kanes 1970, Fig.18), distal sections lacking sand sized sediment, sections through major channels showing thick sands. Similar sequences occur in some of the smaller crevasse deltas of the Mississippi, (Coleman et. al. 1964, Saxena 1976), and in fluviolacustrine parts of the Rhone Delta (Oomkens 1970).

The lack of evidence of wave or tidal action and limited fauna suggests that in the Llynfi Beds the deltas prograded into partially restricted bays. The thickness of the coarsening upwards component suggests deeper water than in Matagorda Bay, of the order of 10 m after allowing for compaction. LF 10 Unlaminated Siltstone probably represents relatively slow deposition within the bays, whereas the beds of LF 10 and LF 11 in the upper parts of the coarsening upwards sequences indicate fairly rapid deposition with a high sediment fall out from suspension in an upper delta slope or mouth bar environment.

Major channels occur between the 5 m and 15 m level in Section A but their widths are unknown. Most channels were only a few metres deep as in the Colorado Delta, (Kanes 1970). The unusual vertical sequences in some of the sharp based sandstone units suggests they may

not be channels, but possibly small crevasse splay lobes. These tend to be dissected by later formed channels, (Elliot 1974a, Saxena 1976). This possibly explains the sequences in the sandstone units at the 22 - 24 m level in Section A and 44 - 46 m level in Section C where cross-bedded sandstone erosively overlies a sharp based bed of ripple-laminated sandstone.

Following abandonment, delta lobes are commonly vegetated, (Frazier and Osanik 1969, Elliott 1974a). This also appears to be the case in Association B as most coals may be correlated through the four sections for a distance of 3 km. Following submergence of the delta top, brackish water bay conditions were re-established sometimes allowing in a restricted bivalve fauna.

10.4. Description and interpretation of sequences in Association A.

10.4.1. Introduction

The chief aim of the research has been to look in detail at vertical and lateral facies changes in the fluvial channel facies of the Pennant Sandstone. Because of the usually near-horizontal attitude of the beds it has been possible to look at lateral changes over considerable distances, particularly in the large roadcutting east of Blaenrhondda where there is over 1.5 km of continuous exposure.

Erosion surfaces are very common in the Pennant Sandstone and attempts to trace lateral facies changes have usually shown that these are erosive, with an unknown amount of sediment being removed. Only rarely has it been possible to trace gradational changes from one facies to another. Thus this aspect of the work has been inconclusive and most of the evidence presented here deals with vertical sequences.

10.4.2. Problems of vertical sequence analysis

The sequences in Association A consist of sandstones, typically 25 - 100 m thick and usually of considerable lateral extent, lying between major widespread coals. [Fig.60]. Within the sandstones are siltstone units rarely more than a few metres in thickness. Mapping shows these to be laterally impersistent, but more widespread siltstones often occur at the top of thick sandstone units immediately underlying coal.

Sections through these thick sandstones show variable sequences and erosion surfaces are common. Unlike other fluvial successions, such as the Old Red Sandstone of the Anglo-Welsh Cuvette (Allen 1963a, 1970a), where there is often a clear differentiation into coarse and fine members, the Pennant sequences can not immediately be split into natural sedimentological units.

For this reason it has been necessary to use a slightly different approach to that adopted by Allen (1970a). Within the succession two major types of sequences have been recognised, termed 'Normal Sequences' and 'Key Sequences' respectively.

'Normal Sequences'

These are sequences in Association A of the Lower Pennant Measures such as would probably be seen in any randomly chosen section. In that sense they compare with the 'modal' sequences of Duff & Walton (1962). The best example is at Earlswood Roadcutting (SS 729 944 - SS 727 945) [Fig.64].

Here the sandstone unit, beginning immediately above a coal [Plate 73] is 78 m thick and contains 22 separate erosion surfaces with reliefs varying from 0.25 m up to 2.5 m. The sandstone units between erosion surfaces vary from less than 1 m up to 13 m in thickness. Clearly this sequence represents the fills of a series of separate channels, stacked one above the other.

The erosion surfaces have two possible origins. Some may represent the bases of major channels so that the sediments above and below belong

to channels of considerably different age. Other erosion surfaces, however, may just represent periods of scour in the channel at some stage in the flood cycle [Chapter 2] so that the deposits above and below belong to the same channel and may differ in age by no more than a few weeks or months (cf. Coleman, 1969).

For a proper understanding of the channel sequences, and in order to be able to split the succession into meaningful sedimentological units, it is important to be able to differentiate between these two possible origins. Unfortunately, this is only occasionally possible. Unambiguous evidence of the base of a major channel only occurs where the erosion surface rests on a non-channel facies such as a coal or thick mudstone. Where the erosion surface rests on older channel fill sediments its precise origin cannot be reliably determined.

Earlier workers on the Pennant Sandstone have failed to fully appreciate this problem. Kelling (1968), in particular appears to have regarded all erosion surfaces with a substantial relief as major channel bases. There is no conclusive evidence that this is so.

The limitations of these 'Normal Sequences' may be more fully appreciated when lateral variations are considered. In the dipping section at Earlswood it is not possible to trace channels very far laterally, but in the main research area, in the Rhondda Valleys, the succession is near horizontal.

Taff Vale Quarry, (ST 078 916) [Plate 82] demonstrates the typical lateral variability. Here there are three erosion surfaces, two of them cross cutting. The lowest surface is undulating showing the original channel morphology. The left hand side dips at about 15° . Banked up against the channel wall is a lateral accretion unit with parallel-laminated sandstone passing up into small scale trough cross-bedding. The bedding gradually flattens off to the right away from the channel wall. A second lateral accretion unit overlies the middle erosion surface. These

are often present against channel sides; Bluck and Kelling (1962), termed them "side filled" channels. They rarely extend very far.

Another example showing the complex erosive fills, from the road-cutting east of Blaenrhondda (SN 9309 0051), is shown in Fig.68. Whether these two examples represent one channel exhibiting several phases of scour and fill, or whether a series of stacked separate channels is unknown. Obviously the vertical section obtained varies considerably depending on where it is measured.

10.4.3. Statistical Analysis

The limited usefulness of randomly chosen sequences can be further demonstrated by a statistical analysis. A Markov Chain Analysis was undertaken with a computer programme written by Mr. J.D. Orford. All measured sections greater than 20 m in length and with 14 or more transitions were used.

It is always difficult to decide how best to subdivide the successions for this type of analysis. The aim here is to make comparisons with fluvial successions elsewhere, particularly the 'meandering' sequences of Devonian age, (Allen 1970a) and the 'braided' Devonian Battery Point Succession, (Cant and Walker 1976). For this reason 7 states have been used, most of which correspond to a particular and distinctive lithofacies. This compares with 8 states of Cant and Walker but is slightly more detailed than Allen's study where only 6 states were used.

The states are :

A. Erosion Surface

These have been distinguished where the relief exceeds 0.75 m. Most of the facies transitions in the sequence are erosive but usually there is no appreciable relief, and in most cases probably little sediment has been removed.

B. Conglomerate

Beds of LF 7, Conglomerate, exceeding 0.5 m in thickness. Thin lags of plant material commonly underlying sets of cross-bedding have not been included.

C. Massive Sandstone (LF 6)

D. Cross-bedding

Three different types of cross-bedding have been recognised, but it is only possible to reliably distinguish these in good three-dimensional exposure. In many sections the cross-bedding type is uncertain, and for the statistical analysis all have been included together.

E. Parallel-laminated Sandstone (LF 4)

F. Ripple-laminated Sandstone (LF 5)

G. Laminated Siltstone (LF 8)

Commonly thin silt laminae thicken laterally into isolated ripple form sets. These are all included in this category and not with Category F.

Seat-earths and coals occasionally occur in the sequences, but they have been excluded here so that the analysis may be kept within a reasonable size. They invariably overlie LF 8, Laminated Siltstone.

Altogether 12 separate logged sections have been used to a total length of 457.4 m. Locations and lengths of individual sections are given in Appendix 2. With the exception of the Earlswood Section most are between 20 m and 55 m in length.

Transitions have only been recorded where one lithofacies changes to another. This is the 'embedded Markov' chain of Krumbein and Dacey (1969), and has zeros in the principal diagonal of the matrix. Altogether 368 transitions occur, Table IX. The Transition Probability Matrix, Table X, shows the probabilities of all transitions from one state up into any of the remaining states. Thus 55.1% of all transitions up from State A, Erosion Surface, pass up into State B, Conglomerate. The matrix in

Table XI shows for each state, the probabilities of transitions down into the remaining states. Thus when State B, Conglomerate occurs, in 74.5% of cases it is preceded by State A, Erosion Surface. Finally, a Difference Matrix is shown in Table XII. Positive entries in this table show transitions which have occurred with greater than random frequency.

Discussion

The analysis shows a definite memory. For instance, over half of the transitions from State A, Erosion Surface, pass up to State B, Conglomerate, and almost two thirds of those from B pass up to State D, Cross-bedding. There are also fairly high values for the transitions from C to D, E to D, and F to G [Fig.66].

Such flow diagrams, however, can easily give the impression of a greater degree of order than is actually the case. The Differences Matrix [Table XII, Fig.67] shows a large number of transitions which may occur with greater than random frequency most of them having a similar statistical significance. Of particular note, is the pivotal position in the flow chart of State D, Cross-bedding, which can be preceded with greater than random frequency by States B, C, E and F and be succeeded with greater than random frequency by States A, F and E.

It is hard to draw any useful conclusions from the analysis. Whilst the sequences are not random, the degree of order is low.

One probable reason for this is the presence of erosion surfaces. This is brought out by the differences between Tables X and XI and Figs. 65 and 66. Where succeeding transitions are considered, clearly the removal of sediment, prior to the burial and preservation of the erosion surface will add an extra random element to the matrix. However, because the erosion surfaces have been included as one of the states in the analysis, their presence will add no extra random influence where preceding transitions are considered, except for preceding transitions down from the erosion surfaces themselves. Thus in nearly 75% of cases in State B, preceded

by State A, and in over 75% of cases in State E preceded by State D [Fig.65], most of the other transition values are also quite high.

Comparisons between these two matrices leave no doubt that erosion of previously deposited sediment has contributed substantially to the random element in the sequences. If it is possible to find sequences which have not been interrupted by erosion, these may provide more accurate information on the actual depositional sequences in the channel fills.

10.4.4. Key Sequences

In some parts of the succession it is possible to find sequences which show several metres exposure without erosion surfaces. Some occur above major channel bases and provide information on the nature of the fill in the lower parts of the channels. Others are found beneath Siltstone Units and show the nature of the fill in the upper parts of the channels.

Such sections, which unfortunately constitute only about 10% of the total exposure in Assemblage A provide much more useful information on the depositional histories of the channels. Theoretically they may be further subdivided into two types:

- (a) Complete Sequences: these show, in a vertical channel sequence all the lithotypes encountered in the Association.
- (b) Incomplete Sequences: in these only some of the total number of lithofacies types are present in one channel sequence, the other being absent because of non-deposition. (Note that in 'Normal Sequences' sediment was usually absent because of subsequent erosion.)

No complete sequences occur in the Lower Pennant Measures. Eight incomplete Key Sequences, all from the upper parts of the channels are shown in Fig.69. Of these, only section H definitely shows a whole channel fill up from major channel base to sediment deposited after channel abandonment. In the others, the erosion surfaces at the bases may not mark the base of a major channel. Most fills are of the order of 10 - 15 m thick. Section A is the nearest to a complete sequence, and a seat-earth/

coal lithofacies occurs about 10 m above the top of the section shown.

By combining evidence from various Key Sequences and simplifying the succession into major components, a rather idealised form of Composite Sequence (Duff and Walton, 1962), may be constructed. This is shown in Fig.70 which also gives a description of the principal components. It cannot be emphasised too strongly that this is not a strictly statistical representation of the succession, like the type produced by Allen (1970a) for example. Nevertheless, the six principal components usually have the same vertical relationship to each other, though these are insufficient sections of this type to provide enough transitions to justify a statistical analysis.

None of the 'Key Sequences' show all of the lithofacies and are therefore 'incomplete' following the previous definition. Four examples are known (none shown in Fig.69), where irregular erosion surfaces, some of which may represent major channel bases are followed immediately by the Siltstone Component. In the stream section in Nant y Gwair (SS 909 988), an erosion surface with a relief of 2.8 m is overlain by LF 8, Laminated Siltstone, to a total thickness of 4.3 m [Fig.64B]. Within this unit the bedding is inclined along large rotational slump scars, where the sediment has slipped away from the sides of the channel. In a number of cases (e.g. Section H in Fig.69), the Main Sandstone Member is succeeded immediately by the Siltstone Member, the intervening Alternating Beds Member being absent.

10.4.5. Channel Size and Morphology

Major channel bases may sometimes be recognised where they rest on fine grained sediments or coals. Some are irregular, sometimes cutting into the top of a coal but rarely cutting through. Mapping by the Geological Survey, (Woodland and Evans 1964), shows that channel sandstones may rest immediately on top of coals for distances of up to several

kilometres. Presumably prior to compaction the vegetation was too thick to enable the river to erode completely through. Nevertheless this has happened occasionally; the No.1 Rhondda Rider Coal has been washed out along Mynydd Blaenrhondda (SN 9180 0018) (Evans 1959) and also over wide areas to the north and east of Blaengawr (Woodland 1960). These washouts appear to be mainly restricted to areas where the coals are unusually thin.

Not all channel bases are irregular. Evidence from underground mine workings shows that some bases are flat and can maintain the same level above shale partings in the underlying coal over distances approaching a kilometre (W.B. Evans p.comm). A flat channel base overlying a mudstone is shown in Plate 83, this can be traced for over 0.5 km and shows little relief, the section is roughly parallel to the channel axis as judged from vector directions in the overlying cross bedding.

Where major channels overlie coals they frequently have log conglomerates at the base. A series of unusually thick conglomerates overlie the No.2 Rhondda Coal, where the channels represent the first widespread development of Association B within the area. They are not present everywhere. Below Swar yr Ogeiriad, (SS 9134 9820), a mudstone roof is preserved above the No.2 Rhondda and the overlying channel sandstone does not have a conglomerate at its base. The channel shown in Plate 83 also lacks a basal conglomerate. Thus the presence or absence of a conglomerate is no criterion for the presence or absence of a major channel.

The widths of channels are not easy to assess because of the difficulty of tracing fills laterally. This is possible, however, in the large road cutting east of Blaenrhondda. Here, palaeocurrents in most of the stacked channel sequence above the No.1 Rhondda Coal are towards the northwest. In one channel fill, however, the vector directions are towards the southwest. By measuring a series of sections along the outcrop it is possible to identify this channel by the abrupt change in the

cross-bedding vectors, even though the channel base may not be particularly distinguishable. The channel fill can be traced laterally for 1.4 km without either edge being seen. The orientation of the channel can nevertheless be determined by calculating the vector mean of the cross-bedding. Coleman (1969) found a very close correlation between bedform orientation and channel orientation in the Brahmaputra river. This shows that the width normal to the channel axis is at least 0.84 km.

It is uncertain whether this represents the channel width at one particular time, or whether the channel constructed a wider sheet sandstone by lateral migration. The channel fill is at least 15 m in thickness, and possibly more. Accepting the 15 m figure this gives a width depth ratio of about 56 : 1. This is a realistic figure if compared to modern, large sandy river channels where width depth ratios of around 65 : 1 are normal (Leopold et. al. 1964).

The rivers which deposited the Pennant Sandstone were therefore quite large by modern standards. In the Niger, for example, among the smaller of the great rivers of the world, the river is generally between 1.5 to 3 km across before it enters the delta. Individual channels are 0.3 to 2 km wide and rarely more than 10 m deep, (Nedeco 1959).

10.4.6. Palaeocurrents

10.4.6.1. Previous Research

Kelling (1969) undertook an extensive palaeocurrent analysis of cross-bedding in the Rhondda Beds. He showed a complicated pattern with local bimodal orientations and a considerable range of variance. This is no doubt partly due to the existence of more than one sediment source, plus the effects of local gradients around the margins of the basin. In the central area of the basin, which includes the present research area, and where all the sediment is of southerly derivation, palaeocurrent variability is less, with a fairly consistent mean towards the northwest.

Kelling has split the area into sectors, corresponding to the O.S. 6 inch County Quarter Sheets, each 15.54 sq km in area. The variance values for different sectors range from 196 to 5,629 and for the whole area the variance is 5,550. Kelling argued, that the pattern supported the existence of predominantly meandering rivers in the research area.

10.4.6.2. Present Study

Numerous directional structures have been measured and the results agree with Kelling's values of the regional variance in the study area. Assessing the variance within individual channel fills is difficult because of the problem of recognising discreet channels discussed earlier. In addition, poor weathering out of the structure often makes it difficult to take accurate measurements. The very limited evidence from the 'Key Sequences' [Fig.69], suggests a fairly low variance in the main body of the fill, but a higher variance in the upper Alternating Beds Component. There is no evidence of bimodal directions for tabular cross-bedding as described by Cant and Walker (1976), and as might occur with the type of sandwave movement described by Smith (1972).

This suggests that the moderately high variance values for the entire sequence are more likely the result of the variable orientation of different channel fills, rather than variation in cross-set orientation within individual fills. An example of this has already been given in Section 10.4.5.

It does not follow that this is the result of meandering river channels. Aerial photographs and diagrams of modern braided rivers, (e.g. Coleman 1969), show that curved reaches are common. In addition, the Pennant Sequences were deposited by rivers which had a number of phases of progradation each associated with several periods of lateral movement across the flood plain (Section 2.3.6.3.). A considerable diversity of channel orientation is likely to result from these events.

10.4.7. Interpretation of the Composite Vertical Sequence

In Chapter 2 a number of different models for various types of meandering and low sinuosity rivers were discussed and these may now be compared with the Pennant Succession.

10.4.7.1. Meandering River Models

'Classical' Model The Pennant Sequences differ from the models of Allen (1963a) and Bernard and Major (1963) in two ways. Firstly the channel fills never show a progressive upwards fining and upwards decrease in the size of sedimentary structures from bottom to top. Secondly, the proportion of overbank fine grained sediment is much lower. Thus it is unlikely that the rivers were of the normal meandering type.

Bedload Meandering Model In this model of McGowan and Garner (1970), straightening of the thalweg during floods forms chute bars on the insides of meander bends. These give rise to large, solitary sets of tabular cross-bedding in the middle part of the channel sequence. Such sets are not present in the Pennant sequences and this model seems inapplicable.

Transitional Meandering Model Jackson (1975a, 1976b), has shown that non-fining-upwards channel sequences may occur in upstream areas of point bars, and in tightly curved meanders. The channel fill sequences in the Pennant Sandstone may be of one or other origin. However, where meanders are tightly curved, meander cut offs should be common, leading to stabilisation of the meander belt, and preservation of thick overbank fines. The scarcity of fine sediment in the Pennant succession suggests that active channel migration occurred and this is unlikely to occur with tightly curved meanders.

Where non-fining upwards sequences occur as a result of deposition in upstream areas of point bars, there will probably be a mixture of these together with the more usual fining upwards sequences produced on the downstream ends of point bars. Such a mixture is not found in the Pennant sequences.

10.4.7.2. Low sinuosity and braided river models

In Chapter 2 low sinuosity rivers were subdivided into straight and braided types. Straight channels in flood-plain regimes are usually fairly rare, being restricted to localised segments of the river. There is no evidence in the Pennant succession of more than one channel type and it is unlikely that the whole Association was produced by straight river channels.

Two types of braided rivers have been distinguished, [Chapter 2], but as yet the vertical sequences produced by these are very poorly known. There is even still disagreement as to just how braided river channels are filled. Moody-Stuart, (1966), who claimed to recognise low sinuosity channels from the Devonian of Spitsbergen, showed field evidence suggestive of vertical accretion of sediment. Allen (1970a), strongly disagreed, maintaining that deposition in low-sinuosity and in meandering channels is similar in kind. More recently, Cant and Walker (1976), have reaffirmed Moody-Stuart's views.

In all braided rivers, there is a sinuous thalweg or thalwegs which meander between topographically higher parts of the channel, usually side bars or islands. Studies on the Niger and Yellow Rivers (Nedeco 1959, Chien 1961) which were discussed by both Allen (1970a) and Moody-Stuart (1966), show beyond doubt that side bars move down the channel by erosion of sediment on their upstream sides and deposition on their downstream sides. As they do so they displace the thalweg area downstream, [Fig.12]. Thus sediment deposited in the thalweg is overlain by sediment deposited in side bars. This process is quite distinct from lateral accretion on the inside of meander bends, and would not give rise to lateral accretion surfaces of the type which are often associated with point bar sedimentation.

In rivers with islands sedimentation is additionally complicated by island growth, some growing by lateral accretion, (Shantzer 1951), some by

a mixture of downstream and vertical accretion, (Leopold and Wolman 1957), [Chapter 2].

A major feature of the channel fills in the Lower Pennant Measures are the thick conglomerates with their large plant stems. Many of these demonstrably lie at the bases of major channels, but they invariably pass laterally into cleaner cross-bedded sandstones. The most reasonable interpretation of the conglomerates is that they represent concentrations of plant material and coarse sediment in the deepest, or thalweg areas of the channels. The laterally equivalent cleaner sandstones are interpreted as having been deposited at roughly the same time in the topographically higher parts of the channels. These were probably side bars and possibly also islands although it is not possible to recognise these large bedforms as such. The upward passage from conglomerate into cleaner cross-bedded sandstones may record the downstream movement of a side bar or island which buries the former thalweg area. The Siltstone beds which sometimes overlie the Conglomerates result from ponding of water in the deepest parts of the channels at low stage.

Within the overlying Main Sandstone Component, rare lateral accretion sets record local accretion along fairly steep channel edges probably shortly after the channel was cut. Vertical sequences in this show very little order and are probably mainly determined by changes in discharge. Many of the erosion surfaces are probably the result of scour during floods. 'High' stage sedimentation is represented by erosively based cosets of cross-bedding whilst the common thin intervening beds of ripple-lamination and parallel lamination record the washing out of larger bedforms during falling or low stage. The preservation of dune profiles in siltstone shows that sometimes the discharge fell too quickly for dune modification, the dunes being preserved by water ponding at low stage. Thus the rivers which deposited the Pennant Measures probably had a fairly variable discharge.

Another effect of the downstream movement of side bars in multiple channel braided rivers is the cutting off and eventual abandonment of channels. This process has already been described in detail in Chapter 2. Progressive channel abandonment is thought to be the cause of the Alternating Beds component. The best example occurs at the Blaenrhondda road cutting (SN 9310 0050) [Section A in Fig.69], where the upper part of the channel fill shows alternations of trough cross-bedded or ripple-laminated sandstone with siltstone [Plates 86, 87], and a more diverse palaeocurrent pattern than lower in the section.

This sequence has already been discussed in detail by Kelling (1968), who suggested that the sandstone beds were crevasse splays. However there seems to be a continuous gradation from the main channel fill into the Alternating Beds Component. Perhaps the first evidence of the beginning of channel abandonment is provided by the siltstone filled trough at the top of the main fill, [Plate 85]. This seems to be the precursor of the thicker siltstone beds higher in the sequence. The sudden cutting off of a channel with a partially blocked entrance, due to a fall in the water level during falling stage would lead to a sudden reduction in discharge in the channel and favour the preservation of dune shapes. Ponding in the inactive channel during low stage allows suspension sedimentation of silt which drapes the dune beds.

The sequence of alternating beds is partly cut out by a siltstone filled channel, [Plate 88]. This is more easily explained in the context of a channel environment, suggesting partial reopening of the channel, rather than in an overbank environment envisaged for a crevasse splay origin. Sections B, D, E, F and G in Fig.69 all show evidence of possible progressive channel abandonment in their upper parts.

Other evidence pointing to frequent channel abandonment, is the common preservation of channel base morphology and the existence of incomplete

fill sequences. The former suggest fairly rapid filling and then abandonment of channels without extensive lateral migration which would tend to favour the production of extensive flat channel bases as in the classical meandering river model. The incomplete fills are believed to occur where a channel is abandoned at an early stage in its life history. Thus siltstone filled channels point to abandonment soon after the channel was cut, whereas channels with a Main Sandstone Component, but no Alternating Beds Component, (e.g. Sections C and H in Fig.69), were probably rapidly abandoned after an active early stage of fill.

Thus a variety of types of fill sequences are possible depending on the channel abandonment history. A similar situation probably prevails with chute cut offs in many meandering systems, but here the limited correspondence to the present meandering river models suggests that the sequences are best interpreted in terms of multiple channel braided rivers.

Comparatively little overbank sediment is preserved in the Association. Most siltstone units cannot be mapped for more than a few hundred metres. As Kelling (1968) has suggested, many probably fill in the tops of abandoned channels rather than represent true flood plain environments. The laterally extensive sheet sandstones are the expected deposit of a braided river which progressively shifts across a flood plain [Chapter 2]. The thickness of the units shows that the major sandstones between extensive coals represent several periods of lateral migration across the flood plain over a considerable period of time.

10.5. Evolution of the Pennant Cycles

The Lower Pennant Measures contain two distinct associations, one interpreted as a result of repeated incursions of small deltas, or crevasse deltas, into shallow bays, the other, as a floodplain environment with laterally migrating braided rivers. Both form part of the same drainage network over most areas of the basin and in the research area palaeocurrents

in both associations are dominantly towards the northwest. The deltaic association is best developed in the Llynfi Beds within the main research area. In the eastern part of this, fluvial channel sediments appear to be completely absent, but in the western half they come in above the Tormyndd Coal as well as just below the No.2 Rhondda Coal. By a further 5 kilometres to the west, in the Maesteg area, the deltaic association has almost completely disappeared. Here the Llynfi Beds consist of the typical Pennant Sandstone of Fluvial Channel Association. A similar east-west facies change takes place as far south as the exposed limits of the sequence.

In the succeeding Rhondda and Brithdir Beds, Association B is poorly developed, but one coarsening upwards sequence occurs between the Daren Rhestyn and No.1 Rhondda Coals [Fig.60].

In between the clastic sediments are major coals which extend for large areas across the basin. These record long periods when the coarse sediment supply was cut off. Within the mudstone roofs to the coals restricted bivalve faunas have commonly been recorded in the maps and memoirs of the Geological Survey. These do not appear to form continuous bands as in the Westphalian of the Central Pennine Basin. They show that at certain times there were widespread areas of standing water in the basin.

Often the coals with their roof faunas are succeeded immediately by thick channel sandstones, some of which have cut right down to the coals themselves. This sharp transition from a brackish fauna to a fluvial floodplain environment is unusual. A possible explanation is that there are two stages in the development of the Pennant "Cycles" [Fig.71]. In the first stage shallow bays were filled in by small deltas or crevasse deltas, developed at the front, and on the sides of the main Pennant river. Following this an alluvial flood plain environment was established as the

main river entered the basin. The river channels were at least 15 m deep and possibly more. This is greater than the thickness of many of the deltaic cycles. Thus as the river channel migrated laterally it would erode away much of the earlier basin fill. Often the channels cut right down to the coals at the base of the cycle, thus removing all evidence of the early phase of progradation.

In the Llynfi Beds, the lowest division of the Pennant Sandstone, the main river channels are restricted to the central parts of the basin where the sequence is thickest. They did not extend as far as the eastern half of the basin and the evidence of the earlier progradational phase is preserved.

At the top of each major cyclothem a switch in the sediment supply allows another phase of coal development. This is followed by a drowning of the coals and further subsidence deepens the basin prior to another phase of progradation.

In none of the progradational cycles is there any evidence of marine activity, suggesting that the basin was cut off from active marine influence. Recently, however, Kelling (1976) and Stead (1974) have suggested that the stratigraphically equivalent orthoquartzite facies, outcropping along the north eastern margin of the basin was deposited as a shallow marine complex of 'beaches, barrier bars and tidal channel deposits'. It is possible of course, that the fluvio-deltaic succession present in the research area was protected from the sea by a barrier complex. This is the case with the present Colorado delta (Kanes 1970), and has been envisaged for similar small coarsening upwards sequences in the Rhône Delta, (Oomkens 1970). However, the geographic position of the orthoquartzite crop, north east of the research area and near to the supposed northern margin of the basin of deposition makes this explanation difficult here. The interpretation of Kelling and Stead, therefore seems

puzzling in the light of the present study, and the earlier explanation envisaging a separate northerly source in the Midland Land Mass to account for the different composition of the sediment, (Kelling 1964b, 1968) may be more reasonable.

10.6. Discussion: ancient braided river sequences, comparisons and problems of interpretation

Harms et.al. (1975) have discussed a methodology for the analysis of fluvial successions. The essentials of this are:

1. Measurement of detailed sections.
2. Subdivision into facies units.
3. Statistical analysis of the succession using Markov Chain Analysis.
4. Construction of a statistically significant vertical sequence, which characterises the succession.
5. Comparison with other statistically derived sequences and formulation of different fluvial models.

This approach, first used by Allen (1970a) for a supposed meandering sequence has since been used by Cant and Walker (1976) for a supposed braided sequence, who emphasise the differences between the two.

It is doubtful whether this approach will work in successions like the Pennant where there are thick sandstones with stacked channel fills. Of the other workers who have described supposed sandy braided river sequence, Selley (1969) did not formulate an ideal sequence, and Thompson (1970) and Guion (1971) erected ideal cycles without a statistical analysis. Graham and Reilly (1972) after carrying out a Markov Chain Analysis of a probable Devonian braided river sequence could detect no statistically significant order. A similar conclusion has been reached in the present study. Although Cant and Walker (1976) claim to be able to erect a statistically significant sequence, the accuracy of this must be doubted as there were only 61 observed transitions for 8 facies states. Considering

the large number of erosion surfaces in the Battery Point Succession and the rarity of overbank fines their rather arbitrary subdivision into major cycles [Fig.2 pl05] seems questionable.

The conclusion of the present study is that within stacked sandstone fluvial sequences, only those parts of the succession which lack major erosion surfaces, and where major channel bases or definite channel tops can also be recognised, will provide meaningful information on the nature of the channel fill. Statistically derived ideal sequences based on the whole succession may be misleading, or impossible to interpret.

Because of the different ways in which ideal sequences have been derived comparisons with other supposed ancient braided river successions must be treated with caution. The Lower Pennant Measures differ substantially from the Bunter and Keuper Sandstones in Cheshire (Thompson 1970), the latter having regularly fining upward channel fills more reminiscent of the classical meandering models, except for the lack of overbank fines. They also differ from the Crawshaw Sandstone (Guion 1971), which above a basal conglomerate contains a fairly uniform non-fining upwards succession of tabular cross-bedding very similar to LF 14 MSXB described in Section II. Guion interpreted this uniformity in terms of a discharge regime which varied little over the year, although in the light of the conclusions reached in Chapter 5, a fairly short flood stage with limited reworking may be envisaged.

The Battery Point succession, (Cant and Walker 1976) is more variable, with an alternation of tabular cross-bedding produced by sand-waves with dune formed trough cross-bedding. Cross-set orientations are bimodal in the tabular sets, but unimodal in the trough sets. The sequences show no overall vertical fining. This succession is clearly different from the Pennant in terms of both palaeocurrent distribution and in lack of evidence for progressive channel abandonment, perhaps

fitting more closely to a Platte type single channel braided river with considerable lateral sand-wave movement.

Finally, the channel fills in the Roaches Grit, are again distinct with the development of very large cross-bedding sets. These may be alternate bars, or possibly, very high discharges in deep channels may have allowed the development of very large sandwaves as in the Brahmaputra. There is no evidence of progressive channel abandonment.

The rivers which deposited the Pennant Sandstone lacked very large bedforms, in spite of channel depths of 15 m or more. Dunes were the most common type of bedform and several phases of scour and fill probably occurred during the filling of a major channel. The complexities of the vertical sequence suggest marked variations in discharge with accompanying changes in the bedform assemblages. Large scale ripples, active at high stage, were sometimes washed out during falling or low stage, and covered with erosively based beds of ripple, or parallel lamination. Sometimes, with more rapid falls in discharge, dune profiles were preserved by siltstones deposited in ponded water.

Table VIII

Comparison of the Lithofacies classification used for the Lower
Pennant Measures with that adopted by Kelling (1968)

	<u>This thesis</u>	<u>Kelling</u>
	1	* B2
	2	* B2
	3	* B2
	4	B3
	5	B4
	6	B1
	7	A
	8	* C
	9	D + E
	10	-
	11	-
	12	-
	13	-

* part of

Table IXTransition Count Matrix

	A	B	C	D	E	F	G	<u>Total</u>
A	-	38	6	18	1	1	5	69
B	3	-	3	31	4	3	8	52
C	5	1	-	7	0	1	0	14
D	28	10	0	-	22	35	8	103
E	5	0	2	19	-	3	0	29
F	9	1	1	22	2	-	25	60
G	15	1	3	9	0	13	-	41
Total	65	51	15	106	29	56	46	368

Table XTransition Probability Matrix

	A	B	C	D	E	F	G	Total
A	-	0.551	0.087	0.261	0.014	0.014	0.072	1.000
B	0.058	-	0.058	0.596	0.077	0.058	0.154	1.000
C	0.357	0.071	-	0.500	0.000	0.071	0.000	1.000
D	0.272	0.097	0.000	-	0.214	0.340	0.078	1.000
E	0.172	0.000	0.069	0.635	-	0.103	0.000	1.000
F	0.150	0.017	0.017	0.367	0.033	-	0.417	1.000
G	0.366	0.024	0.073	0.220	0.000	0.317	-	1.000

Table XIIDifference Matrix

	A	B	C	D	E	F	G
A	-	0.415	0.046	-0.024	-0.065	-0.143	-0.050
B	-0.123	-	0.017	0.311	-0.002	-0.099	0.032
C	0.176	-0.065	-	0.215	-0.079	-0.086	-0.122
D	0.091	-0.039	-0.041	-	0.135	0.183	-0.044
E	-0.009	-0.136	0.028	0.370	-	-0.054	-0.122
F	-0.031	-0.199	-0.024	0.082	-0.046	-	0.295
G	0.185	-0.112	0.032	-0.065	-0.079	0.160	-

References

- Allen, J.R.L. (1960) The Mam Tor Sandstones, a "turbidite" facies of the Namurian deltas of Derbyshire, England. *J. Sedim. Petrol.*, 30, 193-208.
- (1962) Intraformational conglomerates and scoured surfaces in the Lower Old Red Sandstone of the Anglo-Welsh Cuvette. *Lpool. Manchr. Geol. Journ.*, 3, 1-20.
- (1963a) "Henry Clifton Sorby and the sedimentary structures of sands and sandstones in relation to flow conditions." *Geol. en Mijnb.* 42, 223-228.
- (1963b) Asymmetrical ripple marks and the origin of water-laid cosets of cross strata. *Lpool. Manchr. Geol. J.* 3, 187-234.
- (1963c) The classification of cross-stratified units, with notes on their origin. *Sedimentology*, 2, 93-114.
- (1964a) Studies in fluvial sedimentation: six cyclothems from the Lower Old Red Sandstone, Anglo-Welsh Basin. *Sedimentology*, 3, 163-98.
- (1964b) Primary current lineation in the Lower Old Red Sandstone (Devonian), Anglo-Welsh Basin. *Sedimentology* 3, 89-108.
- (1965a) A review of the origin and characteristics of recent alluvial sediments. *Sedimentology*, 5, 89-191.
- (1965b) Fining upwards cycles in alluvial successions. *Lpool. Manchr. Geol. J.* 4, 229-46.
- (1965c) Sedimentation and palaeogeography of the Old Red Sandstone of Anglesey, North Wales. *Proc. Yorks. Geol. Soc.* 35, 139-182.
- (1966) On bedforms and palaeocurrents. *Sedimentology*, 6, 153-90.

- Allen, J.R.L. (1967) Notes on some fundamentals of palaeocurrent analysis, with reference to preservation potential and sources of variance, *Sedimentology* 9, 75-88.
- (1968) "Current Ripples", pp433. North Holland, Amsterdam.
- (1969) Erosional current marks of weakly cohesive mud beds. *J. Sedim. Petrol.* 39, 607-23.
- (1970a) Studies in fluvial sedimentation: a comparison of fining upwards cyclothems, with special reference to coarse member composition and interpretation. *J. Sedim. Petrol.* 40, 298-323.
- (1970b) The sequence of sedimentary structures in turbidites, with special reference to dunes. *Scott. J. Geol.* 6, 147-61.
- (1973a) Features of cross-stratified units due to random and other changes in bed forms. *Sedimentology* 20, 189-202.
- (1973b) Phase differences between bed configuration and flow in natural environments, and their geological relevance, *Sedimentology*, 20, 323-29.
- (1974) Studies in fluvial sedimentation: implications of pedogenic carbonate units, Lower Old Red Sandstone, Anglo-Welsh outcrop. *Lpool. Manchr. Geol. J.* 9, 181-208.
- (1976) Computational models for dune time-lag: population structures and the effects of discharge pattern and coefficient of change. *Sedim. Geol.* 16 99-130.
- Allen, J.R.L. and J.D. Collinson (1974) The superimposition and classification of dunes formed by unidirectional aqueous flows. *Sedim. Geol.* 12, 169-178.
- Allen, J.R.L. and P.F. Friend (1976a) Relaxation times of dunes in decelerating aqueous flows. *J. Geol. Soc. Lond.* 132, 17-26.
- (1976b) Changes in intertidal dunes during two springneap cycles, Lifeboat Station Bank, Wells-next-the Sea, Norfolk (England). *Sedimentology* 23, 329-46.

- Alpert, S.P. (1973) Bergaueria (Cambrian and Ordovician) a probable Actinian trace fossil. *J. of Palaeontology* 47, No.5.
- Ashton, C.A. (1974) Palaeontology, stratigraphy and sedimentology of the Kinderscoutian and Lower Marsdenian (Namurian) of North Staffordshire and adjacent areas. Unpublished PhD. thesis University of Keele.
- Baines, J.G. (1977) The Stratigraphy and Sedimentology of the Skipton Moor Grits (Namurian Eic) and their lateral equivalents. Unpublished Ph.D. thesis, University of Keele.
- Baker, V.R. (1973) Palaeohydrology and sedimentology of Lake Missoula flooding in eastern Washington. *Geol. Soc. Am. Spec. Pap.* 144, pp79.
- Banks, N.L. (1973) The origin and significance of some downcurrent dipping cross-stratified sets. *J. Sedim. Petrol.* 43, 423-27.
- Banks, N.L. and J.D. Collinson (1974) Discussion of "Some sedimentological aspects of planar cross-stratification in a sandy braided river." *J. Sedim. Petrol.* 44, 265-67.
- Bates, C.C. (1953) Rational theory of delta formation. *Bull. Am. Ass. Petrol. Geol.* 37, 2119-2162.
- Beckinsale, R.P. (1969) River Regimes. In: R.J. Chorley (Ed.) *Water, Earth and Man*. Methuen, London, pp455-472.
- Bernard, H.A. and C.J. Major (1963) Recent meander belt of the Brazos River: an alluvial "sand" model. *Bull. Am. Ass. Petrol. Geol.* 47, p.350.
- Bersier A. (1958) Séquences détritiques et divigations fluvials. *Eclog. Geol. Helvetiae* 51, 854-93.
- Beutner, E.C., L.A. Flueckinger and T.M. Gard (1967) Bedding geometry in a Pennsylvanian channel sandstone. *Bull. Geol. Soc. Am.* 78, 911-16.

- Bigarella, J.J. (1965) Sand ridge structures from Parana coastal plain,
Marine Geology 3, 269-79.
- Blatt, H.G.V., Middleton and R. Murray (1972) "Origin of sedimentary rocks",
pp634, Prentice Hall, New Jersey.
- Bluck, B.J. and G. Kelling (1962) Channels from the Upper Carboniferous
Coal Measures of South Wales. Sedimentology 2, 29-53.
- Boersma, J.R. (1969) Internal structure of some tidal mega-ripples on a
shoal in the Westerschelde Estuary, the Netherlands.
Report of a preliminary investigation. Geol. en Mijnb.
48, 409-14.
- Bolton, T. (In prep.) Palaeontology, stratigraphy and sedimentology of the
Arnsbergian, Chokerian and Alportian of North Staffordshire.
Ph.D. thesis, University of Keele.
- Boothroyd, J.C. (1969) Hydraulic conditions controlling the formation of
estuarine bedforms. In: Coastal Environments, Field
Trip of Eastern Section of Society of Economic
Palaeontologists and Mineralogists, Coastal Research
Group, University of Massachusetts, 417-24.
- Boothroyd, J.C. and G.A. Ashley (1975) Processes, bar morphology, and
sedimentary structures on braided outwash fans, north-
eastern Gulf of Alaska. In: Glaciofluvial and glacio-
lacustrine sedimentation. (Ed. by A.C. Jopling and
B.C. McDonald) Spec. Publ. Soc. econ. Palaeont. Minr.,
Tulsa, 23, 193-222.
- Boothroyd, J.C. and D.K. Hubbard (1972) Bedform development and distribution
pattern, Parker and Essex estuaries, Massachusetts:
U.S. Army Corps of Engineers, Coastal Engineering Research
Centre, Misc. Paper 1-74.
- Bouma, A.H. (1962) "Sedimentology of some flysch deposits; a graphic
approach to facies interpretation", pp168 Elsevier,
Amsterdam.

- Brice, J.C. (1964) Channel patterns and terraces of the Loup River in Nebraska. Prof. Pap. U.S. Geol. Surv. 422-D pp41.
- Wells, J.T. (1974) Evolution of meander loops. Bull. Geol. Soc. Am. 85, 581-86
- Brown, L.F. (1969) Geometry and distribution of fluvial and deltaic sandstones (Pennsylvanian and Permian), north-central Texas. Bur. Econ. Geol. Texas University, 69-4, 23-47.
- Calver, M.A. (1971) Westphalian of Britain. C.R. Congr. int. Statig. Geol. Carb. (6th Sheffield) I, 233-54.
- Cant, D.J. and R.G. Walker (1976) Development of a braided-fluvial facies model for the Devonian Battery Point Sandstone, Quebec. Can. J. Earth Sci. 13, 102-19.
- Carey, W.C. and M.D. Keller (1957) Systematic changes in the beds of alluvial rivers. Proc. Am. Soc. Civil Engrs. 83 Paper 1331, Journ. Hydr. Div., pp24.
- Challinor, J. (1921) Notes on the geology of Roches District. Trans. North Staffs. Nat. Field Club, 55 76-87.
- (1929) Notes on the geology of the North Staffordshire moorlands. Trans. N. Staffs. Field Club, 63, 111-12.
- (1930) The stratigraphical position of the Roches "Rough Rock". Trans. N. Staffs. Field Club 64, 1930.
- Chien, N. (1961) The braided stream of the Yellow River Scientia Sinica X No.6 734-54.
- Chisholm, J.I. (1968) Trace fossils from the geological survey boreholes in East Fife, 1963-1964. Bull. Geol. Surv. Gt. Br. 28, 103-119.
- (1970) Teichichnus and related trace fossils in the Lower Carboniferous at St. Monance, Scotland. Bull. Geol. Surv. Gt. Br. 32, 21-51.

- Chorley, R.J. and B.A. Kenneday (1971) Physical Geography. Prentice Hall, London, pp370.
- Colby, B.R. (1963) Fluvial sediments, a summary of source, transportation, deposition and measurement of sediment discharge. Bull. U.S. Geol. Surv. 1181 A.
- (1964) Scour and fill in sand bed streams. Prof. Pap. U.S. Geol. Surv. 462-D pp32.
- Coleman, J.M. (1969) Brahmaputra River: channel processes and sedimentation. Sedim. Geol. 3, 129-239.
- Coleman, J.M. and S. Gagliano (1964) Cyclic sedimentation in the Mississippi River deltaic plain. Trans. Gulf Coast Ass. Geol. Socs. 14, 67-70.
- Coleman, J.M., J.C. Ferm and R.S. Saxena (1972) Deltas: recent and ancient: New Orleans Geological Society, Continuing Education Seminar, Nov. 28-30, pp 169.
- Coleman J.M., J.N. Suhayda, T. Whelan and L.D. Wright, (1974) Mass movement of Mississippi River Delta sediments. Trans. Gulf Coast Ass. Geol. Socs. 24, 49-68.
- Coleman J.M. and L.D. Wright (1975) Modern river deltas: variability of processes and sand bodies. In: Deltas, models for exploration, (Ed. by M.L. Broussard) Houston, Geological Society, pp 555.
- Collinson, J.D. (1966) Antidune bedding in the Namurian of Derbyshire, England. Geol. en Mijnb. 45, 262-64.
- (1967) The sedimentology of the southern part of the Grindslow Shales and the Kinderscout Grit. Unpublished D.Phil. thesis, University of Oxford.
- (1968) Deltaic sedimentation units in the upper Carboniferous of northern England. Sedimentology, 10, 233-54.

- Collinson, J.D. (1969) The sedimentology of the Grindslow Shales and the Kinderscout Grit: a deltaic complex in the Namurian of northern England. *J. Sedim. Petrol.* 39, 194-221.
- (1970a) Deep channels, massive beds and turbidity current genesis in the Central Pennine Basin. *Proc. Yorks. Geol. Soc.*, 37, 495-520.
- (1970b) Bedforms of the Tana River, Norway. *Geogr. Annlr.* 52A, 31-56.
- (1971) Current vector dispersion in a river of fluctuating discharge. *Geol. en Mijnb.*, 50, 671-78.
- Conaster, W.E. (1971) Grand Isle: a barrier island in the Gulf of Mexico, *Bull. Geol. Soc. Am.* 82, 3049-68.
- Cope, F.W. (1946) Intraformational contorted rocks in the Upper Carboniferous of the Southern Pennines. *Ql. Jl. Geol. Soc. Lond.* 101, 139-76.
- Costello, W.R. (1974) Development of bed configurations in coarse sands. Massachusetts Institute of Technology, Dept. of Earth and Planetary Sciences, Report 74-1, pp120.
- Cotter, E. (1975) Deltaic deposits in the Upper Cretaceous Ferron Sandstone, Utah. In: *Deltas, models for exploration*, (Ed. by M.L. Broussard), Houston Geological Society pp555.
- Culbertson, J.K. and C.H. Scott (1970) Sandbar development and movement in an alluvial channel, Rio Grande near Bernardo, New Mexico. *Prof. Pap. U.S. Geol. Surv.* 700B, 237-41.
- Curray, J.R. (1956) Analysis of two dimensional orientation data. *J. Geol.* 64, 117-31.
- Duff, P.MCL. D. and E.K. Walton (1962) Statistical basis for cyclothem: a quantitative study of the sedimentary succession in the East Pennine Coalfield. *Sedimentology* I, 235-55.

- Eager, R.M.C. (1971) A new section in the Lower Coal Measures (Westphalian A) of Up Holland, near Wigan, Lancashire. Proc. Geol. Ass. 82, 71-85.
- (1974) Shape of shell of Carbonicola in relation to burrowing. Lethaia, 7, 219-38.
- Elliot, T. (1974a) Abandonment Facies of high-constructive lobate deltas, with an example from the Yoredale Series. Proc. Geol. Ass. 85, (3), 359-65.
- (1974b) Interdistributary bay sequences and their genesis. Sedimentology 21, 611-22.
- (1975) The sedimentary history of a delta lobe from a Yoredale (Carboniferous) cyclothem. Proc. Yorks. Geol. Soc. 40, 505-36.
- (1976) The morphology, magnitude and regime of a carboniferous fluvial-distributary channel, J. Sedim. Petrol., 46, 70-76.
- Evans, W.B. (1959) Geol. Surv. 6" Map Glamorgan XVII NE.
- Evans, W.B., AA. Wilson, B.J. Taylor and D. Price (1968) The geology of the country Macclesfield, Congleton, Crewe and Middlewich. Mem. Geol Surv. Gt. Brit. pp328.
- Failing, M.S. (19) Comparison of Trinity River terraces and gradients with other Texas Gulf Coast rivers. In Holocene geology of the Galveston Bay area. R.R. Lankford and J.J.W. Rogers ed. Houston Geol. Soc.
- Falcon, N.L. and P.E. Kent (1960) Geological results of petroleum exploration in Britain, 1945-1957. Mem. Geol. Soc. Lond. No.2, pp56.
- Farey, J. (1811) General view of the agriculture and minerals of Derbyshire. Vol. I.
- Fisher, W.L. (1969) Facies characterisation of Gulf Coast Basin delta systems with some Holocene analogues. Trans. Gulf Coast Assn. Geol. Socs. 20, 239-61.

- Fisk, H.N. (1944) Geological investigations of the alluvial valley of the lower Mississippi River. Mississippi River Commission, Vicksburg, Miss. pp78.
- (1947) Fine grained alluvial deposits and their effects on Mississippi River activity. Mississippi River Commission, Vicksburg, Miss. pp82.
- (1955) Sand facies of recent Mississippi Delta deposits. World Petrol. Cong., 4th Rome, Proc., sec. I, 377-98.
- Forel, F.A. (1887) Le ravins sous-lacustres du Rhone dans le Lac Lemain. Bull. Soc. vaud. Sci. nat. 23, 85-107.
- Frazier, D.E. (1967) Recent deltaic deposits of the Mississippi River: their development and chronology. Trans. Gulf Coast Ass. of Geol. Socs. 17, 287-311.
- Frazier D.E. and A. Osanik (1969) Recent peat deposits - Louisiana coastal plain. In: Environments of coal deposition, (Ed. by Dapples, E.C. and M.E. Hopkins), Geol. Soc. Am. Spec. Pap. 114, 63-85.
- Friedkin, J.F. (1945) A laboratory study of the meandering of alluvial rivers: U.S. Waterways Engr. Exper. Sta. pp40.
- Gilligan, A. (1920) The petrography of the Millstone Grit of Yorkshire. Q. Jl. Geol. Soc. Lond. 75, 251-94.
- Gole, C.V. and S.V. Chitale (1966) Inland delta building activity of Kosi River. Proc. Am. Soc. Civil Engr. 92, No. HY2 J. Hydraul. Div. 111-126.
- Gould, H.R. (1960) Turbidity currents. In: Comprehensive Survey of Sedimentation in Lake Mead, 1948-1949. Prof. Pap. U.S. Geol. Surv. 295, 201-207.
- Gradstein, F.M. and A. van Gelder (1971) Prograding clastic fans and transition from a fluvial to a marine environment in Neogene deposits of eastern Crete. Geol. en Kijnb., 50, 383-92.

- Gradzinski, R and T. Jerzykiewicz (1974) Results of the Polish-Mongolian Palaeontological Expeditions - Part V. Sedimentation of the Barun Goyot formation, Palaeontologia Polonica No.30. 111-146.
- Graham, J.R. and T.A. Reilly (1972) The Sherkin Formation (Devonian) of South-west County Cork. Bull. Geol. Surv. Ireland, 1, 281-300.
- Guion, P. (1971) A sedimentological study of the Crawshaw Sandstone (Westphalian A) in the East Midlands Coalfield area. Unpublished M.Sc. thesis, University of Keele.
- Hamilton, E.L. (1967) Marine geology abyssal plains in the Gulf of Alaska. Jour. Geophys. Research. 72, 4189-4213.
- Hand, B.M., J.M. Wessel and M.O. Hayes (1969) Antidunes in the Mount Toby Conglomerate (Triassic), Massachusetts. J. Sedim. Petrol. 39, 1310-16.
- Haner, B.E. (1971) Morphology and sediments of Redondo Submarine Fan, southern California. Bull. Geol. Soc. Am. 82, 2413-32.
- Häntzschel, W. (1962) Trace fossils and problematica. In: Treatise on invertebrate palaeontology, (Ed. by R.C. Moore) Part W University of Kansas Press. 177-245.
- Happe, S.C., Rittenhiuse, G. and G.C. Dobson (1940) Some aspects of accelerated stream and valley sedimentation. U.S. Dept. Agr. Tech. Bull. 695, pp134.
- Hardy, P.G. (1970) Aspects of palaeoecology in the arenaceous sediments of Upper Carboniferous age in the area around Manchester. Unpublished Ph.D. thesis, University of Manchester.
- Harms, J.C. (1974) Brushy Canyon Formation, Texas: a deep water density current deposit. Bull. Geol. Soc. Am. 85, 1763-84.

- Harms, J.C. and R.K. Fahnstock (1965) Stratification, bed forms and flow phenomena (with an example from the Rio Grande).
In: Primary Sedimentary Structures and their hydrodynamic interpretation. (Ed. by G.V. Middleton) Spec. Publ. Soc. econ. Palaeont. Minr., Tulsa, 12, 84-115.
- Harms, J.C., J.B. Southard, D.R. Spearing and R.G. Walker, (1975)
Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Soc. Econ. Pal. and Miner. Short Course No. pp161
Dallas Texas.
- High, L.R. and M.D. Picard (1974) Reliability of cross stratification types as palaeocurrent indicators in fluvial rocks.
J. Sedim. Petrol. 44, 158-68.
- Hitchcock, E. (1858) Ichnology of New England; a report on the sandstone of the Connecticut Valley, especially its footmarks.
White, Boston, pp220.
- Holdsworth, B.K. (1963) The palaeontology, stratigraphy and sedimentology of Namurian rocks in the Longnor-Hollinsclough-Morriddge region, north-east Staffordshire and south-west Derbyshire. Unpublished Ph.D. thesis University of Manchester.
- Holeman, J.N. (1968) The sediment yield of major rivers of the world.
Water Resources Research 4, 737-47.
- Holmes, A. (1965) Principles of Physical Geology pp. 1288, Nelson, London.
- Holmes, D.A. (1958) The recent history of the Indus. Geogr. J. 134, 367.
- Houbolt, J.J.H.C. and J.B.M. Jonker (1968) Recent sediments in the eastern part of the lake of Geneva, Lac Lemman. Geol. Mijnb. 47, No.2, 131-48.
- Howard, J.D. (1966) Sedimentation of the Panther Sandstone Tongue, Utah Geol. and Miner. Survey Bull. 80, 23-33.

- Howard, J.D. (1971) Comparison of the beach-to-offshore sequence in modern and ancient sediments. In: Recent advances in palaeocology and ichnology. (Ed. by J.W. Valentine), A.G.I. Short Course, Washington, 148-83.
- Hsu, K.J. (1964) Cross laminations in graded bed sequences. *J. Sedim. Petrol.* 34, 379-88.
- Hull, E. and A.H. Green (1864) On the Millstone Grit of North Staffordshire and the adjoining parts of Derbyshire, Cheshire and Lancashire. *Q. Jl. Geol. Soc. Lond.* 20, 242-67.
- (1866) The geology of the country around Stockport, Macclesfield, Congleton and Leek. *Mem. Geol. Surv. Gt. Bt.* pp
- Jackson, R.G. (1975a) A depositional model of point bars in the Lower Wabash River. Unpublished Ph.D. thesis, University of Illinois, pp 263.
- (1975b) Velocity-bed form-texture patterns of meander bends in the lower Wabash River of Illinois and Indiana. *Bull. Geol. Soc. Am.* 86, 1511-22.
- (1975c) Hierarchical attributes and a unifying model of bed forms composed of cohesionless material produced by shearing flow. *Bull. Geol. Soc. Am.* 86, 1523-33.
- (1976a) Largescale ripples of the lower Wabash River. *Sedimentology*, 23, 593-624.
- (1976b) Depositional model of point bars in the lower Wabash River. *J. Sedim. Petrol.* 46, No.3, 579-94.
- James, U.P. (1873) *The Paleontologist*, I.
- Jansa, L. (1972) Depositional history of the coalbearing Upper Jurassic-Lower Cretaceous Kootenay Formation, Southern Rocky Mountains, Canada. *Bull. Geol. Soc. Am.*, 83, 3199-3222.

- Jerzykiewicz, T. (1963) Sedymentacja Górnych Piaskowców Ciosowych Niecki Srodsudeckiej (Górna Kreda), *Geologia Sudetica* IV, 1968R. Warszawa, 409-462.
- Jopling, A.V. (1965) Hydraulic factors controlling the shape of laminae in laboratory deltas. *J. Sedim. Petrol.* 35, 777-91.
- Jordon, G.F. (1962) Large submarine sand waves. *Science* 136, 839-48.
- Kanes, W.H. (1970) Facies and development of the Colorado River Delta in Texas. In: "Deltaic Sedimentation, Modern and Ancient" (Ed. by J.P. Morgan). *Spec. Publ. Soc. Econ. Paleont. and Miner.*, Tulsa 15, 78-106.
- Kelling, G. (1964a) The turbidite concept in Britain. In: *Turbidites*, (Ed. by A.H. Bouma and A. Brouwer) Elsevier, Amsterdam, 75-92.
- (1964b) Sediment transport in part of the lower Pennant Measures of South Wales. In: *Deltaic and Shallow Marine Deposits* (Ed. by L.M.J.U. Van Straaten), Elsevier, Amsterdam, 177-84.
- (1968) Patterns of sedimentation in the Rhondda Beds. *Bull. Am. Ass. Petrol. Geol.* 52, 2369-2382.
- (1969) The environmental significance of cross-stratification parameters in an Upper Carboniferous fluvial basin. *J. Sedim. Petrol.* 39, 857-75.
- (1974) Upper Carboniferous sedimentation in South Wales. In: *The Upper Palaeozoic and post-Palaeozoic rocks of Wales*. (Ed. by T.R. Owen) University of Wales Press Cardiff. 185-224.
- Kent, P.E. (1966) The structure of the concealed Carboniferous rocks of north-eastern England. *Proc. Yorks. Geol. Soc.* 35, 323-52.
- Khan, F.H. (1957) Investigations of peat in the Faridpur District, East Pakistan. *Geol. Surv. Pakistan, Min. Industries, Information Release*, 4, pp7.

- Krigstrom, A., (1962) Geomorphological studies of sandur plains and their braided rivers in Iceland, *Geogr. Annlr.* 44, 328-46.
- Kruit, C., Brouwer J., Knox, G., Schollnberger, W., and A. Van Vliet, (1976). Une excursion aux cones d'alluvions en eau profonde d'age Tertiaire poes de San Sebastian. Excursion 23, IX *Congres International de Sedimentologie*, pp75.
- Krumbein, W.C. (1934) Size frequency distribution of sediments. *J. Sedim. Petrol.* 4, 65-77.
- Krumbein, W.C. and M.F. Dacey (1969) Markov chains and embedded chains in geology. *J. Int. Ass. Mathl. Geol.*, 1, 79-96.
- Ksiazkiewicz, M. (1970) Observations on the ichnofauna of the Polish Carpathians. In: *Trace Fossils*, (Ed. by T.P. Crimes and J.C. Harper), Lpool. *Geol. Jour. Spec. Issue*, 3, Steel House Press, Liverpool, 283-322.
- Lambert, A.M., K.R. Kelts and N.F. Marshall (1976) Measurements of density underflows from Walensee, Switzerland. *Sedimentology*, 23, 87-105.
- Lane, E.W. and W.M. Borland (1954) River-bed scour during floods. *Trans. Am. Soc. Civil Engrs.* 119, 1069-80.
- Lane, E.W. and E.W. Eden (1940) Sand waves in the lower Mississippi River. *Jour. of the Western Soc. of Engineers*, 45, 281-91.
- Leeder, M.R. (1973) Sedimentology and palaeogeography of the Upper Old Red Sandstone in the Scottish Border Basin. *Scott. J. Geol.* 9, 117-44.
- Leopold, L.B. and W.B. Langbein (1966) River Meanders, *Scientific American*, 214, No. 6, 60-70.
- Leopold, L.B. and M.G. Wolman (1957) River channel patterns: braided, meandering and straight. *Prof. Pap. U.S. Geol. Surv.* 282-B, pp85.

- Leopold, L.B., M.G. Wolman and J.P. Miller (1964) Fluvial processes in geomorphology. pp522. Freeman, San Francisco and London.
- Lowe, D. (1976a) Grain flow and grain flow deposits. *J. Sedim. Petrol.* 46, 188-99.
- (1976b) Subaqueous liquified and fluidised sediment flows and their deposits. *Sedimentology*, 23, 285-308.
- McCabe, P.J. (1975) The sedimentology and stratigraphy of the Kinderscout Grit Group (Namurian R1) between Wharfedale and Longendale. Unpublished Ph.D. thesis, University of Keele. pp172.
- (in press) The Kinderscoutian Delta (Carboniferous) of Northern England: a slope influenced by density currents. In: *Ancient and Modern Submarine Canyons and Fans*, (Ed. by G. Kelling and D.J. Stanley), Dowden, Hutchinson and Ross (Wiley and Sons).
- McCabe, P.J. and C.M. Jones (in press) The formation of reactivation surfaces within superimposed deltas and bedforms. *J. Sedim. Petrol.*
- McGowen, J.H. and L.E. Garner (1970) Physiographic features and stratification types of coarse grained point bars: modern and ancient examples. *Sedimentology*, 14, 77-111.
- McGowen, J.H. and C.G. Groat (1971) Van Horn Sandstone, West Texas: an alluvial fan model for mineral exploration. Bureau of Economic Geology, University of Texas, Report No. 72, pp57.
- McKee, E.D. (1966) Structures of dunes at White Sands National Monument, New Mexico (and a comparison with structures of dunes from other selected areas), *Sedimentology* 7, 1-69.
- Mackin, J.H. (1937) Erosional history of the Big Horn Basin. *Bull. Geol. Soc. Am.* 48, 813-94.

- Maddock, T. (1969) The behaviour of straight open channels with movable. Prof. Pap. U.S. Geol. Surv. 622A, pp70.
- Martinson, A. (1970) Toponomy of trace fossils. In: Trace Fossils, (Ed. by T.P. Crimes and J.C. Harper) Lpool. Geol. Jour. Spec. Issue 3, Steel House Press, Liverpool, 323-30.
- Matthes, G.H. (1941) Basic aspects of stream meanders. Trans. Amer. Geophys. Union, 22, 632-36.
- Mayhew, C.H. (1967) A sedimentological investigation of the Marsdenian grits and associated measures in north west Derbyshire. Unpublished Ph.D. thesis, University of Sheffield.
- Merki, P. (1972) Structural geology of the Cenozoic Niger Delta, Ibadan University Geol. Conf. Proc. 635-46.
- Miall, A.D. (1974) Palaeocurrent analysis of alluvial sediments; a discussion of directional variance and vector magnitude. J. Sedim. Petrol. 44, 1174-85.
- (1976) Palaeocurrent and palaeohydraulic analysis of some vertical profiles through a Cretaceous braided stream deposit, Banks Island Arctic Canada. Sedimentology 23 459-483.
- Middleton, G.V. (1965) Antidune cross-bedding in a large flume. J. Sedim. Petrol. 35, 922-27.
- (1976) The orientation of concavo-convex particles deposited from experimental turbidity currents. J. Sedim. Petrol. 37, 229-32.
- (1970) Experimental studies related to problems of flysch sedimentation. In: Flysch Sedimentology in North America (Ed. by J. Lajoie), 253-272. Spec. Pap. Geol. Soc. Can. 7, 272.

- Middleton, G.V. and M.A. Hampton (1973) Sediment gravity flows; mechanism of flow and deposition. In: "Turbidites and deep water sedimentation". Soc. econ. Paleont. Miner. Short course, 1-38.
- Moody-Stuart, M. (1966) High and low-sinuosity stream deposits with examples from the Devonian of Spitsbergen, J. Sedim. Petrol. 36, 1102-17.
- Moore, D.G. (1961) Submarine slumps. J. Sedim. Petrol. 31, 343-357.
- Moore, G.T. (1970) Role of salt wedge in Bar-finger sand and delta development. Bull. Am. Ass. Petrol Geol. 54, 326-33.
- Morris, P.G. (1969) Dinantian and Namurian stratigraphy east and south-east of Leek, north Staffordshire. Proc. Geol. Ass. 80, 145-75.
- Mutti, E. and F. Ricci Lucchi, (1972) Le torbiditi dell' Appenino settentrionale: introduzione all'analisi di facies. Mem. Soc. Geol. Italiana, II, 161-99.
- Nami, M. (1976) An exhumed Jurassic meander belt from Yorkshire, England. Geol. Mag. 113(1), 1976, 47-52.
- Nedeco (Netherlands Engineering Consultants) (1959) River Studies and Recommendations for Improvement of Niver and Benue. North-Holland, Amsterdam, pp1000.
- Nelson, C.H. and V. Kulm (1973) Submarine fans and deep channels. In: "Turbidites and deep water sedimentation", Soc. Econ. Paleont. Miner. Short Course, 39-78.
- Nicholson, H.A. (1873) Contributions to the study of Errant Annelides of the older Palaeozoic rocks. Proc. R. Soc., 21, 228-90.
- Oomkens, E. (1970) Depositional sequences and sand distribution in the postglacial Rhone Delta complex. In: Deltaic Sedimentation Modern and Ancient (Ed. by J.P. Morgan). Spec. Publ. Soc. Econ. Paleont. and Miner., Tulsa 15, 198-212.

- Ore, H.T. (1964) Some criteria for recognition of braided stream deposits. Wyoming Univ. Geol. Contrib. 3, 1-14.
- Osgood, R.G. (1970) Trace fossils of the Cincinnati area. Palaeontographica Americana, 6, 308-12.
- Otvos, E.G. (1970) Development and migration of barrier islands, northern Gulf of Mexico. Bull. Geol. Soc. Am. 81, 241-6.
- Owen, T.R. (1964) The tectonic framework of Carboniferous sedimentation in South Wales. In: Deltaic and shallow marine deposits, (Ed. by L.M.J.U. van Straaten), Elsevier, Amsterdam, 301-07.
- Phillips, J. (1836) "Illustrations of the geology of Yorkshire, Part 2. The Mountain Limestone District". London.
- Piper, D.J.W. (1972) Turbidite origin of some laminated mudstones, Geol. Mag. 109, 115-26.
- Popov, I.V. (1965) Hydromorphological principles of the theory of channel processes and their use in hydrotechnical planning. Soviet Hydrology Selected Papers 1964, No. 2. 188-95.
- Potter, P.E. and F.J. Pettijohn, (1963) Palaeocurrents and basin analysis. pp296, Springer-Verlag, Berlin.
- Pratt, C.J. (1973) Bagnold approach and bedform development. Proc. Am. Soc. Civil Eng., J. Hydr. Div. 99, 121-37.
- Pratt, C.J. and K.V.H. Smith (1972) Ripple and Dune phases in a narrowly graded sand, Proc. Am. Soc. Civil Eng., J. Hyd. Div., 98, 859-74.
- Pretious, E.S. and T. Blench (1951) Final report on special observations of bed movement in Lower Fraser River at Badner Reach during 1950 freshet. Nat. Res. Council Canada, Fraser River Model, Vancouver, Canada.
- De Raaf, J.F.M., H.G. Reading and R.G. Walker (1965) Cyclic sedimentation in the Lower Westphalian of North Devon, England. Sedimentology, 4, 1-52.

- Ramsbottom, W.H.C. (1969a) Interim report of the Namurian working group.
C.R. 6th Congr. Stratig. Geol. Carb. Sheffield 1967, I, 71-78.
- (1969b) The Namurian of Britain. C.R. Congr. int.
Stratig. Geol. Carb. (6th Sheffield), 4, 1359-400.
- Ramsbottom, W.H.C., G.H. Rhys and E.G. Smith (1962) Boreholes in the
Carboniferous rocks of the Ashover district, Derbyshire.
Bull. Geol. Surv. Gt. Br. 19, 75-168.
- Reading, H.G. (1964) A review of the factors affecting the sedimentation of
the Millstone Grit (Namurian) in the Central Pennines.
In: "Deltaic and shallow marine deposits" (Ed. by
L.M.J.U. Van Straaten), Elsevier, Amsterdam. 340-6.
- (1971) Sedimentation sequences in the Upper Carboniferous of
northwest Europe. C.R. 6th Congr. Stratig. Geol. Carb.
Sheffield 1967. 4, 1401-12.
- Reiche, P. (1938) An analysis of cross-lamination: The Coconino Sandstone.
Jour. Geol. 46, 905-32.
- Reineck, H.E. (1967) Layered sediments of tidal flats, beaches and shelf
bottoms of the North Sea. LAUFF, 191-206.
- Ricci-Lucchi, F. (1969) Channelised deposits in the Middle Miocene flysch
of Romagna (Italy) Giorn. di Geologia, 36, 208.
- (1975) Depositional cycles in two turbidite formations of
northern Appenines (Italy). J. Sedim. Petrol. 45, 3-43.
- Sanders, J.E. (1965) Primary sedimentary structures formed by turbidites and
related resedimentation mechanisms. In: Primary
Sedimentary Structures and their hydrodynamic interpret-
ation. (Ed. by G.V. Middleton). Spec. Publ. Soc. Econ.
Paleont. and Miner. Tulsa, 12, 192-219.
- Schumm, S.A. (1960) The shape of alluvial channels in relation to sediment
type. Prof. Pap. U.S. Geol. Surv. 352-B, 17-30.

- Schumm, S.A. (1963) Sinuosity of alluvial rivers on the Great Plains. Bull. Geol. Soc. Am. 74, 1089-1100.
- (1968) Speculations concerning paleohydrologic controls of terrestrial sedimentation. Bull. Geol. Soc. Am. 79, 1573-88.
- Schumm, S.A. and H.R. Khan (1972) Experimental study of channel patterns. Bull. Geol. Soc. Am. 83, 1753-70.
- Scruton, P.C. (1960) Delta building and the deltaic sequence. In: Recent sediments, northwest Gulf of Mexico, (Ed. by Tulsa, Oklahoma, Am. Ass. Petrol. Geol., 82-102.
- Seilacher, A. (1953) Studien zur Palichnologie, II; die fossilen Ruhespuren. Neues. Jb. Geol. u. Palaeontol. 98, 87-124.
- (1962) Paleontological studies on turbidite sedimentation and erosion. J. Geol. 70, 227-34.
- (1967) Bathymetry of trace fossils. Marine Geology, 5, 413-28.
- Selley, R.C. (1969) Torridonian alluvium and quicksands. Scott. J. Geol. 4, 328-46.
- Shantzer, E.V. (1951) Alluvium of river plains in a temperate zone and its significance for understanding the laws governing the structure and formation of alluvial sites. Tr. Inst. Geol. Nauk. Akad. Nank SSSR Geol. Ser. 135,I, 1-271.
- Shen, H.W. (1971) River Mechanics. 2 Vols. Shen, Fort Collins Colorado.
- Shen, H.W. and S. Komura (1968) Meandering tendencies in straight alluvial channels. Proc. Am. Soc. Civil Eng. 94 HY4, Paper 6042, Jour. Hyd. Div. 997-1015.
- Shepard, F.P. (1960) Mississippi Delta: marginal environments, sediments and growth. In: Recent sediments in the north west Gulf of Mexico, Ed. by F.P. Shepard, F.B. Phleger, and T.H. van Andel, Am. Ass. Petrol. Geol., Tulsa, 56-81.

- Shepard, F.P. and R.F. Dill (1966) "Submarine canyons and other sea valleys." pp391 Rand McNally, New York.
- Simons, D.B. and E.V. Richardson (1961) Forms of bed roughness in alluvial channels. Proc. Am. Soc. Civil Engr. 87 HY3 Jour. Hydr. Div. 87-105.
- Singh, I.B. and S. Kumar (1974) Mega- and giant ripples in the Ganga, Yamuna and Son Rivers, Uttar Pradesh, India. Sedim. Geol. 12, 53-66.
- Skipper, K. (1971) Antidune cross-stratification in a turbidite sequence, Cloridorme Formation, Gaspé, Quebec. Sedimentology, 17, 51-68.
- Smith, N.D. (1970) The braided stream depositional environment: Comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians. Bull. Geol. Soc. Am. 81, 2993-3014.
- (1971) Transverse bars and braiding in the Lower Platte River, Nebraska. Bull. Geol. Soc. Am. 82, 3407-20.
- (1972) Some sedimentological aspects of planar cross-stratification in a sandy braided river. J. Sedim. Petrol. 42, 624-34.
- (1974) Some sedimentological aspects of planar cross-stratification in a sandy braided river: a reply to N.L. Banks and J.D. Collinson, J. Sedim. Petrol. 44, 267-69.
- Sorby, H.C. (1859) On the structure and origin of the Millstone Grit in south Yorkshire. Proc. Yorks. Geol. Soc. 3, 669-75.
- Squirrel, H.C. and R.A. Downing (1969) Geology of the South Wales Coalfield, Part I. The country around Newport. Mem. Geol. Surv. Gt. Brit. H.M.S.O. London, pp333.

- Stanley, D.J. (1968) Graded bedding-sole marking-greywacke assemblage and related sedimentary structures in some Carboniferous flood deposits, eastern Massachusetts. Spec. Pap. Geol. Soc. Am. 106, 211-39.
- Stanley, K.O. (1974) Morphology and hydraulic significance of climbing ripples with superimposed micro-ripple-drift cross-lamination in lower Quaternary lake silts, Nebraska. J. Sedim. Petrol. 44, 472-83.
- Stauffer, P.H. (1967) Grain-flow deposits and their implications. J. Sedim. Petrol. 37, 487-508.
- Stead, J.T.G. (1974) The sedimentology of the Upper Coal Measures of the Forest of Dean and adjacent areas. Unpublished Ph.D. thesis, University of Swansea, pp217.
- Stevenson, I.P. and G.D. Gaunt (1971) The geology of the country around Chapel-en-le-Frith. Mem. Geol. Surv. Gt. Brit., H.M.S.O. London. pp180.
- Sundborg, A. (1956) The river Klarälven, a study of fluvial processes. Geogr. Annlr. 38, 127-316.
- Thomas, L.P. (1974) The Westphalian (Coal Measures) in South Wales. In: The Upper Palaeozoic and Post-Palaeozoic Rocks of Wales, (Ed. by T.R. Owen) University of Wales Press, Cardiff 133-60.
- Thompson, D.B. (1968) Dome-shaped aeolian dunes in the Frodsham member of the so-called "Keuper" Sandstone Formation (Scythian-Anisian: Triassic) at Frodsham, Cheshire (England), Sedim. Geol. 3, 263-89.
- (1970) Sedimentation of the Triassic (Scythian) Red Pebbly Sandstones in the Cheshire Basin and its margins. Lpool. Manchr. Geol. J. 7, 183-216.

- Trewin, N.H. (1969) The stratigraphy, palaeontology and sedimentology of the upper Pendlian and lower Arnsbergian of the north Staffordshire Basin area. Unpublished Ph.D. thesis, University of Keele, pp426.
- Trewin, N.H. and B.K. Holdsworth (1973) Sedimentation in the lower Namurian rocks of the North Staffordshire Basin. Proc. Yorks. Geol. Soc. 39, 371-408.
- Trotter, F.M. (1951) Sedimentation facies in the Namurian of north-west England and adjoining areas. Lpool. Manchr. Geol. J. I, 77-112.
- U.S. Govt. Printing Office, (1973) Quality of surface waters in the United States, 1968 Wat. Supp. Pap. 2098, Parts 9 and 10, Colorado River Basin and Great River Basin.
- Vanoni, V.I. (1946) Transportation of suspended sediment by Water. Trans. Am. Soc. Civil Engrs. 125, pt. I, 1140-1175.
- Walker, R.G. (1965) The origin and significance of the internal sedimentary structures of turbidites. Proc. Yorks. Geol. Soc. 35, 1-32.
- (1966a) Shale Grit and Grindslow Shales; transition from turbidite to shallow water sediments in the Upper Carboniferous of northern England. J. Sedim. Petrol. 36, 90-114.
- (1966b) Deep channels in turbidite bearing formations. Bull. Am. Ass. Petrol. Geol. 50, 1899-1917.
- (1967) Upper flow regime bedforms in turbidites of the Hatch Formation. Devonian of New York State. J. Sedim. Petrol. 37, 1052-58.
- Walker, R.G. and E. Mutti (1973) Turbidite facies and facies associations. In: "Turbidites and Deep Water Sedimentation", Soc. Econ. Paleont. and Miner. Short Course. 119-158.

- Watson, W. (1811) A deliniation of the strata of Derbyshire. Sheffield.
Reprinted 1973, Moorland Reprints, Hartington.
- Wentworth, C.K. (1922) A scale of grade and class terms for clastic sediments.
J. Geol. 30, 377-92.
- Westoll, T.S. (1962) The standard model cyclothem of the Visean and Namurian
sequence of Northern England. C.r. Congr. Avanc. Etud.
Stratigr. Carb. 4, Heerlen, 1958, 3, 767-73.
- Wilhelm, H. (1966) Der "wanderinde" Ström. Erdkunde 20, 265.
- Williams, P.F. (1968) The sedimentation of Westphalian (Ammanian) measures
in the Little Haven-Amroth Coalfield, Pembrokeshire.
J. Sedim. Petrol. 38, 332-62.
- Williams, G.E. (1971) Flood deposits of the sand bed ephemeral streams of
central Australia. Sedimentology 17 1-40.
- Wills, L.J. (1951) A palaeogeographical atlas of the British Isles and
adjacent parts of Europe, Blackie and Son, London, pp64.
- Woodland, A.W. (1960), Geol. Surv. 6" Map Glamorgan XXVI NE.
- Woodland, A.W. and W.B. Evans (1964) The Geology of the South Wales Coalfield.
Part IV. The country around Pontypridd and Maestod,
Mem. Geol. Surv. Gt. Brit. pp391.
- Woodland, A.W., W.B. Evans and J.V. Stephens (1957) Classification of the
Coal Measures of South Wales with special reference to
the Upper Coal Measures, Bull. Geol. Surv. Gt. Br.
13, 6-13.
- Wolman, M.G. and L.M. Brush (1961) Factors controlling the size and shape of
stream channels in coarse non-cohesive sands. Prof. Pap.
U.S. Geol. Surv. 282-G, 183-210.
- Wright, W.B., R.L. Sherlock, D.A. Wray, W. Lloyd and L.H. Tonks (1927) The
geology of the Rossendale Anticline. Mem. Geol. Surv.
Gt. Brit., H.M.S.O. London, ppl82.
- Yeakel, L.S. (1962) Tuscarora, Juniata and Bald Eagle palaeocurrents and
palaeogeography in the central Appalachians. Bull. Geol.
Soc. Am. 73, 1515-40.

Collinson, J.D., C.M. Jones and A.A. Wilson, (1977).

The Marsdenian (Namurian R2) succession west of Blackburn;
implications for the evolution of Pennine Delta Systems.

Lpod. Manchr. Geol. J. 12, 59-76.

Johnson, H.D. (1975). Tide- and wave-dominated inshore and shoreline
sequences from the late Precambrian, Finnmark, North Norway
Sedimentology, 22, 45-74.

```

0 PRINT "FEED IN DATA IN DEGREES"
1 LET A=0
2 LET Y=0
3 LET N=0
4 LET Z=3.1416/180
5 PRINT " : ";
6 INPUT A
7 IF A<90 THEN 100
8 IF A=90 THEN 150
9 IF A<180 THEN 200
10 IF A=180 THEN 250
11 IF A<270 THEN 300
12 IF A=270 THEN 350
13 IF A<360 THEN 400
14 IF A=360 THEN 450
15 IF A>360 THEN 1000
100 LET B=SIN(A*Z)
105 LET C=COS(A*Z)
110 LET X=X+B
115 LET Y=Y+C
120 GOTO 500
150 LET X=X+1
155 GOTO 500
200 LET D=180-A
205 LET E=SIN(D*Z)
210 LET F=COS(D*Z)
215 LET X=X+E
220 LET Y=Y-F
225 GOTO 500
250 LET Y=Y-1
255 GOTO 500
300 LET H=A-180
305 LET I=SIN(H*Z)
310 LET J=COS(H*Z)
315 LET X=X-I
320 LET Y=Y-J
325 GOTO 500
350 LET X=X-1
355 GOTO 500
400 LET G=360-A
405 LET K=SIN(G*Z)
410 LET L=COS(G*Z)
415 LET X=X-K
420 LET Y=Y+L
425 GOTO 500
450 LET Y=Y+1
455 GOTO 500
500 LET N=N+1
505 GOTO 15
1000 LET M=X/Y
1005 LET M2=AIN(M)
1006 LET M3=M2/Z
1007 IF M3<0 THEN 1012
1009 GOTO 1016
1010 LET M3=M3+180
1011 GOTO 1016
1012 IF Y<0 THEN 1015
1013 LET M3=M3+360
1014 GOTO 1016
1015 LET M3=M3+180
1016 LET T=(X^2)+(Y^2)
1020 LET F=SQR(T)
1025 LET S=F*100/Z
1028 PRINT
1029 PRINT
1035 PRINT "VECTOR DIRN":TOL(C20);FREC(4);M3
1040 PRINT "VECTOR MAGNITUDE":TOL(C20);FREC(4);S
1045 PRINT "NO. OF OBSERVATIONS":TOL(C20);FREC(4);N
2000 END

```

Appendix 1

Computer programme for
calculating Vector Mean and
Magnitude after Curray 1956

```

2000 IF Y<0 THEN 1015

```

Appendix 2.

Sections used in the Markov Chain analysis of the Lower Pennant Measures

1. Stream section in Nant y Gwair (SS 9111 9875 - SS 9089 9879), beginning immediately above the No.2 Rhondda Coal. Rhondda Beds. Length, 39.5m.
2. Outcrop west of Blaenrhondda (SN 9309 0051), between No.2 Rhondda and Daren Rhestyn Coals. Length 16.5m.
3. Roadcutting at Earlswood (SS 729 944 - SS 727 945). Between No.2 Rhondda and No.1 Rhondda Coals, Rhondda Beds. Length 130m.
4. Quarry above Cwmparc (SS 9406 9525). Sandstone above the Tormynydd Coal, Llynfi Beds. Length 26m.
5. Roadcutting east of Blaenrhondda, section between the No.1 Rhondda and No.1 Rhondda Rider Coals. (SN 9288 0134). Length 39m.
6. As (5) (SN 9292 0125). Length 41m.
7. As (5) (SN 9398 0106). Length 27m.
8. As (5) (SN 9301 0091). Length 53m.
9. As (5) (SN 9310 0081). Length 22m.
10. As (5) (SN 9313 0065). Length 54m.
11. As (5) (SN 9332 0022). Length 16.9m.
12. Roadcutting east of Blaenrhondda, section between the No.2 Rhondda and Daren Rhestyn Coals. Rhondda Beds (SN 9308 0049). Length 32m.