THE SEDIMENTOLOGY AND STRATIGRAPHY OF THE KINDERSCOUT GRIT GROUP (NAMURIAN, R_1) BETWEEN WHARFEDALE AND LONGDENDALE

Peter J. McCabe

Thesis for the degree of Doctor of Philosophy

of the University of Keele

1975

VOLUME 1 : Text

ABSTRACT

The thesis is the result of a sedimentological and stratigraphical survey of the Kinderscoutian (R_1), Namurian, in the Central Pennine Basin between Wharfedale and Longdendale. Twenty two facies are described and the relationships between them discussed. Three major "assemblages" are distinguished in the 200 to 570m thick, deltaic sequence.

The lowest assemblage, A, has sandstones interpreted as turbidites and grain flow deposits. Interbedded with the sandstones is mudrock, representing normal deposition from suspension. The sandstones thicken, become less parallel sided and become more abundant higher in the sequence. Channels, up to 30m deep are common in the upper part of the sequence. Assemblage A is interpreted as the deposits of submarine fans.

Assemblage B is essentially a coarsening upward sequence, with silt dominant in the lower part and very coarse, pebbly sandsone dominant in the upper part, but shows great lateral variation. Units, up to 14m thick, with beds parallel to inclined bases, dipping at up to 16°, are interpreted as slope gully deposits. Rippled beds within these units are thought to be due to deposition under density current flow. Channels, over 9m deep, infilled with coarse sandstone are thought to be due to deposition under density current flow. Channels, over 9m deep, infilled with coarse sandstone are thought to have been cut and infilled by immature turbidity currents. Overbank deposits from such channels are also recognised. Assemblage B is interpreted as the delta slope sediments.

At the base of Assemblage C large channels occur infilled with cross-bed sets up to 35m thick. Unlaminated sandstones occur at the

base of the channels. Undulatory bedded sandstone, with crest to crest lengths between 9 and 23m and heights of 1m, occur on some channel sides. The large-scale cross-beds are interpreted as channel infill or bedform features; the majority being thought to be transverse bars. The undulatory beds are thought to be spurs formed by corkscrew vortices in the front of the bedforms with skewed crestlines. The unlaminated sandstones are thought to have formed at the reattachment points of the large bedforms. The remainder of the assemblage consists of; cosets of medium-scale cross-bedded, coarse sandstones interpreted as fluviatile deposits; coarsening and fining-upward sequences of mudrock, wavy bedded sandstone and mudrock and ripple laminated sandstones, interpreted as interdistributary bay deposits; and seatearths and coals representing terrestrial conditions. The whole assemblage is interpreted as a delta top sequence with the large channels at the base being the main delta distributaries.

The facies analysis is used as the basis of a new lithostratigraphic subdivision of the Kinderscout Grit Group. The following formations are defined:- Mam Tor Sandstone Formation, Shale Grit Formation, Todmorden Sandstone Formation, Otley Sandstone Formation (all of Assemblage A), Grindslow Shale Formation, Hebden Bridge Shale and Sandstone Formation, Silsden Shale and Sandstone Formation (all of Assemblage B) and the Kinderscout Grit Formation (Assemblage C). Facies equivalent formations are diachranous between Wharfedale and Longdendale, commencing earlier towards the north/northeast.

All the palaeocurrent indicators suggest a current toward the south-southwest. The thickness of Assemblage A and B is thought to reflect the depth of the basin. Both are thickest in the south of the area, apparently due to northward shallowing against the underlying Skipton Moor Grits. The thickness of Assemblage C is thought to reflect the amount of subsidence and compaction during deposition of the assemblage. The assemblage is thickest in the north due to the earlier establishment of terrestrial conditions in the north. Subsidence, apparently at even rate throughout the basin, and compaction are thought to be the major controlling factors of transgressions in Assemblage C.

The whole sequence is interpreted as the deposits of a river dominated delta entering a partially enclosed basin which was, for most of the time, of low salinity, although occasionally fully marine conditions were established.

ACKNOWLEDGEMENTS

This thesis is the result of work carried out during the tenure of a postgraduate research studentship of the Natural Environment Research Council. Dr. John D. Collinson has been enthusiastic in his supervision and I have benefitted from thought provoking discussions with him, from his critical approach to my "theories" and from his understanding of the problems of research students.

Professor Cope and the academic and technical staff of the Department of Geology, through their friendship, encouragement and helpfulness made my stay at Keele an enjoyable one. The stratigraphy section of the thesis owes a great deal to Dr. Brian K. Holdsworth with whom I have had many stimulating discussions. The ideas expressed in the thesis also owe much to interesting discussions with John G. Baines, Dr. Nigel L. Banks, Colin M. Jones and to other research workers at Keele, whose own interests were less related to my own.

Wakefield and District Water Board, Manchester Corporation Waterworks and the West Pennine Water Board kindly allowed me access to their lands and the Wakefield and West Pennine Water Boards provided geological information from various construction programmes. I am also indebted to the Northwestern Road Construction Unit for information on the M62 Windy Hill cutting and for the opportunity to study the exposure.

My mother has helped considerably in the typing of the thesis and Moira, my sister, helped in a small way with the fieldwork and the diagrams. My greatest debt, however, is to my father (who sadly died as the thesis was being completed) and mother for their encouragement and generous help in many ways.

Thank you.

CONTENTS

Page No.

Page No.

Chapter 6; Stratigraphy

Chapter 7; Regional Details

LIST OF TABLES.

CHAPTER 1 INTRODUCTION

1.1 Aims of Research

During the 1960's Allen (1960), Collinson (1966, 1968, 1969, 1970) and Walker (1966 a and b) carried out a detailed investigation of the upper Kinderscoutian (R_{1c}) in the High Peak area of Derbyshire. They established an upward sequence of basinal mudstones - turbidites - slope deposits - delta top. It was also known (Reading, 1964 and Collinson, 1969) that the rocks of the same age in Wharedale were fluviatile sandstones interbedded with paralic sediments.

The original aim of the current research was to carry out a sedimentological and stratigraphical investigation of the R_{1c} in order to better understand the nature of the transition from the established regressive sequence in the south to the thinner so called "cyclic" sequence in the north. During the course of the study it was found that the regressive sequence did in fact occur throughout the area but was initiated earlier in the north; the research was therefore extended to the entire Kinderscoutian (R_1) succession. The study has also necessitated a reinterpretation of some of the rock types described from the Derbyshire area. This was partly due to the advantage of viewing the succession over a wider area and studying rocks spanning a greater length of geological time. The presence of many quarries within the thesis area, was also an advantage though the area lacks the numerous long stream sections of Derbyshire.

1.2 The Area

The study area is the Central Pennines between Wharfedale in the north and Longdendale in the south (see Fig.1, Vol.2). South of Calderdale the hills are plateau like (Plates 1 and 2) and rise to 460m, south of Todmorden, and to 520m around Saddleworth. To the north, the hills are lower, rising only to 400m, and are in the form of escarpments with the scarp slopes facing northwards.

Steep, narrow valleys dissect the plateau in the south whilst in the north the valleys, such as Airedale and Wharfedale are wide and have much lower slopes. The hills of the south are moorlands covered with heath vegetation and are used for sheep grazing and grouse shooting. Only the highest parts of the northern hills are covered by heath whilst the slopes are grass covered and are used for cattle grazing. The area lies between the major conurbations of South Lancashire and West Yorkshire and is cut by a series of communication systems :- railways, roads, a motorway and canals.

The rocks of the Kinderscout Grit Group are important, at the present day, in underlying the soft water catchment areas of the many reservoirs which supply the nearby conurbations. The rocks are also one of the attractions of the area as a weekend recreational centre. In the past the very coarse sandstones, or "grits" were extensively quarried as a building stone, as a source of millstones and, more recently, as a source of road aggregates.

1.3 Stratigraphic Position

The Kinderscout Grit Group was deposited during the Kinderscoutian (the lower <u>Reticuloceras</u> stage, R_1), within the Namurian, and is one of the major sandstone groups of the Millstone

Grit. The major biostratigraphical division of the Namurian and the lithostratigraphic divisions of the Millstone Grit are given in Table 1.1.

1.4 Structure of the Region

South of Keighley the rocks form part of the Pennine anticline. Most of the rocks occur in the core of the anticline and to the east of the fold axis, which runs north/south. These rocks are near horizontal or dip slightly to the east. A series of north/south, steeply inclined faults and thrusts are associated with the fold axis which runs between Todmorden and Saddleworth. Small outcrops of the Kinderscout Grit Group occur between these faults and dip steeply towards the west. North of Keighley the rocks lie on the southern limb of the Skipton anticline and dip gently southwards. A series of faults, mostly with a north-west/south-east trend, but with a complementary north-east/south-west set, cut the entire area.

1.5 History of Previous Research

1.5.1 Sedimentology

Phillips(1836) was the first to carry out a facies analysis of the Millstone Grit. He identified seven rock types :- "Shale", "Gray Beds" (alternations of shale and sandstone), "Flagstone", "Galliard" (gannister), "Gritstone","Ironstone" (in concretionary form) and "Coal". He noticed trace fossils in the flagstones and interpreted them as indicating a littoral environment. He also recognized, from the petrography and variations in grain size, that the grits, of the Yoredales, at least, were deposited by a current flowing from the north.

Table 1.1

NAMURIAN		MILLSTONE GRIT
Yeadonian	G_1	Haslingden Flags and Rough Rock
Marsdenian	R_2	Middle Grits
Kinderscoutian	R_1	Kinderscout Grit Group
Alportian	H_2	Sabden Shales
Chokerian	H_1	Sabden Shales
Arnsbergian	E_2	Skipton Moor Grits
Pendleian	E_1	Upper Bowland Shales

Sorby (1859), in an outstanding paper, compared the "drift bedding" (cross-bedding) of the Millstone Grit of south Yorkshire and north Derbyshire with that which he produced experimentally. From his extensive observation of the cross-bedding he found "on average the current was from the north-east". Combining an examination of the pebble petrography with the palaeocurrent evidence he concluded that the Millstone Grit was derived "from the waste of a south-westwards prolongation of an ancient Scandanavia, the site of which is now occupied by the North Sea".

Sixty years elapsed before the next sedimentological study of the Millstone Grit was published. Gilligan (1920) in a detailed petrographic description describes the Millstone Grit as an arkose containing pebbles of igneous, metamorphic and sedimentary rocks. He also carried out a survey of the heavy minerals but could not distinguish the assemblages of different sandstones. Gilligan considered the Millstone Grit to be deposited by a river, comparable in size to the Mississippi with its source in mountains, of Himalayan scale, stretching across north Scotland and the northern North Sea.

The work of Wright et al (1927) on the Rossendale area was important in that it was the first work that recognised the coarsening upward sequences as being delta sequences infilling a basin of water, with repeated transgression producing rhythmic variation in the sedimentation.

Trotter (1951) recognized seven sedimentation facies in the Namurian of north-west England but only distinguished three within R_1. The "fluvial-grit facies" of the Kinderscout Grit was, he suggested laid down in river valleys. His second facies, "grit-shale" was placed to the north and west of the main Kinderscout facies and apparently included the Addingham Edge, Caley Crags and Shale Grit. These Trotter interpreted as estuarine to open sea deposits with variations due to wave and current action. In the Fylde to Wirral area he suggested a "marine shale" facies, presumably the equivalent of the Edale Shales.

C.T. Walker (1955) carried out a statistical survey of foreset directions in the R_1 sandstones of Wharfedale and Airedale. He found the foreset dip directions were more variable than described by Sorby and Gilligan and concluded that the sandstone was the deposit of two delta cones with apices towards the north and east.

Allen's (1960) paper on the Mam Tor Sandstones was important in recording the first recognition of turbidites, not only in R_1 but also in the whole of the Namurian of the Central Pennines. He suggested a fairly steep depositional slope because of the presence of crumpled bedding.

Holdsworth (1963) described the southern part of the Central Pennine basin as an autogeosyncline, the final phase of its evolution being the rapid infilling during R_1. The subsidence he suggested was rapid enough to produce turbidites at the base of the sequence.

M.D. Wright (1964 a and b, 1967) studied the whole of the R succession in the Longdendale and Marsden area. He considered the cross-bedded sandstone to be deposited in large fluvial, probably braided, channels. The finer, flaggy and rippled sandstones he considered to be tidal and estuarine in origin whilst shales were deposited in "embayments" and areas of shallow sea near the deltas. He considered there was no evidence to suggest that the Shale Grit was a turbidite formation, as had recently been suggested for the Mam Tor Sandstones.

In his review of the sedimentation of the Millstone Grit, Reading (1964) regarded the R_{1c} sediments as being the deposits of the second of two major periods of infilling of the Central Pennine Basin; the first period depositing the Skipton Moor Grits in E_1. He suggested that the Shale Grit and Grindslow Shale were the deposits of the upper slope of the advancing delta.

R.G.Walker (1966 a and b) and Collinson (1966, 1968, 1969, 1970) carried out detailed facies analysis of the Shale Grit, Grindslow Shale and Kinderscout Grit in north Derbyshire. They suggested that the environments of the formations were submarine fan, slope and interdistributary and delta-top respectively. A detailed comparison of their facies and interpretations with those of the present is given in the following chapter.

1.5.2 Stratigraphy

On his map of England and Wales (1815) William Smith shows the "Mountain Limestone" overlain by the "Coal Measures" despite the description of the "Millstone Grit" and "Shale Grit" by his pupil Farey (1811), in the Derbyshire area. Smith does however refer to a sandstone underlying the Coal Measures, presumably the Millstone Grit. On his map of Yorkshire, six years later, he places the "Moorstone or Millstone Grit" between the "Mountain or Metaliferous Limestone" and the "Coal Measures".

Phillips (1836) places the "Limestone Shale" and "Millstone Grit or Farewell Rock" between the "Mountain Limestone and the "Coal Measures" in Derbyshire and correlates the "Limestone Shale" and the shales of the Todmorden valley with the Yoredales of the Askrigg Block. He also depicted the syntypes for <u>R.reticulatum</u> and <u>H.striolatum</u>.

In 1850 the Geological Survey commenced mapping the area and produced a series of memoirs to complement the maps. Hull and Green (1864) numbered the grits in Derbyshire "in the order in which a well sinker would number them". The lowest, the "4th Grit" was also termed the "Kinder Scout Grit". The beds below were termed "Yoredale Beds" topped by the "Shale Grit". This classification was generally followed throughout the area, although the equivalents of the "Shale Grit" were sometimes termed "Yoredale Grits" (e.g. Hull et al 1875).

Important publications of the second half of the nineteenth century include those by Spencer (1874-1899) on Calderdale and Holroyd and Barnes (1896) on the Saddleworth area. Hind (1897) was the first to recognize the difference between the true Yoredales of the Askrigg Block and the Shales beneath the Kinderscout Grit. Some of the fossiliferous localities of the Todmorden and Hebden Bridge area were described by Brown (1841) and Spencer (1898).

Kendall and Wroot (1924) gave a general account of what was then known about the Namurian. It included the, then supposed, correlation of the Skipton Moor Grits with the Kinderscout Grits. The same year saw the publication of Bisat's (1924) important work on goniatites which permitted biostratigraphical correlation for the first time. Bisat also set up a series of zones, the basis of the present biozones. The _Reticuloceras_ Stage was defined and the lower part, R_1, was subdivided into the _inconstans_ (_circumplicatile_) and _reticulatum_ zones. Bisat (1928) added a middle zone, _R. eoreticulatum_, and named the R_1 "Kinderscoutian". The zonation was considerably refined by Bisat and Hudson (1943) who subdivided the sequence into six zones.

The remapping of the area by the Geological Survey commenced in the twenties. The new palaeontological background allowed a more detailed division of the strata and a much better correlation than had been possible in the old series. Papers resulting from this work include Lloyd et al (1927) on the Todmorden District and Wray (1929) and Stephens _et al_ (1942) on the Rombalds Moor region.

A useful synthesis of the Central Pennine Namurian was produced by Ramsbottom (1966). An interim report of the Namurian Working Group, set up in 1965, was produced in 1968 (Ramsbottom 1968a). The report sets up the stratotypes for the Namurian stages in the Central Pennines. Ramsbottom (1974) gave a resume of R_1 stratigraphy in Yorkshire. He interprets the R_1 sequence as a series of Wright _et al_ (1927) type cycles of sedimentation, caused by a series of marine transgressions resulting in marine bands (Ramsbottom, 1974, fig. 27). The complete coarsening upward sequence ending in a coal is, however, only found in the upper part of the sequence.

1.6 Layout of thesis

The text of the thesis (this volume) is divided into three sections. In the first section (Chapters 2, 3 and 4) facies are defined, described and interpreted. An introduction to the facies analysis is given in the next section; 1.7. The facies analysis is the basis of a new lithostratigraphic subdivision of the Kinderscout Grit Group. This subdivision and a comparison with the biostratigraphy is given in Chapter 6. The final section (Chapters 7 and 8) deal with the regional details and an analysis of the basin evolution.

As most figures and plates are referred to more than once in the thesis, they are bound together for ease of reference. The figures on A4 paper and the plates are bound in Volume 2. Large fold-out maps and diagrams are bound seperately in Volume 3.

1.7 Facies analysis - an introduction

The rock types observed in the thesis area are essentially clastic. A brief description of their petrography is given in Chapter 5. The rocks have been divided into twenty two "facies" to facilitate description and interpretation. Facies (numbered F.1, F.2 etc.) are defined as groups of rocks differing from others in their grain size, range of sedimentary structures and fossil content. As with most rock classifications there are transitions between many of the facies.

The majority of the facies can be classified into ten "associations" of two or more facies which occur intrinsically together. Any one facies, however, is not necessarily confined to one association.

The sedimentary sequence has been divided into three divisions, or "assemblages", each with three or four associations. The assemblages will be described in upward order. Within each assemblage the facies will be described and interpreted in terms of sedimentary processes. They will then be grouped into associations whose interpretation can be taken to a higher level. Relationships within each assemblage will finally be discussed and a depositional environment proposed.

Walker (1966a) and Collinson (1969) made detailed facies analyses of the lateral equivalent of the entire succession, in the Derbyshire area. No attempt was made to apply their facies classification in the field because of the dangers of being subjective. The classification used in this thesis is therefore a completely new one based on field observations. In many ways it is similar to Walker's and Collinson's classification but it differs in ways which help more detailed depositional interpretation. Table 1.2 compares the three facies schemes.

1.8 Terminology

Grain sizes are given throughout on the Wentworth (1922) scale. Most sedimentary structures, processes and environmental terms are used as by Blatt et al (1972) unless otherwise stated. As Allen (1968) points out, there appears to be "an overwhelming case for the reality of two populations of ripples". To avoid lengthier terms the use of the word "ripple" will be confined to bedforms less than 0.6m in chord length and 40mm in height whilst the term "dune" will be used for larger structures. Lamination produced by ripples will be termed "ripple lamination".

Table 1.2

Facies	Walker (1966a)	Collinson (1969)
1) Mudrock	D+F	1+2*
2) Goniatite faunal bed		
3) Laminated sandstones	A+B*+C*	14
4) Unlaminated sandstones	B*+C*	7
5) Intermediate sandstones		
6) Laminated silts		
7) Shell Bed		
8) Gradationally laminated ssts.		2*+3*
9) Micaceous silty sandstones		
10) Ripple laminated sandstones		5*
11) Ripple laminated ssts. and coarse ssts.		
12) Sharp based sandstones		5*+6*+11*+14*
13) Medium-scale cross-bedded ssts.		9
14) Zeta cross-stratification		
15) Horizontally laminated coarse ssts.		8+6*
16) Striped silts and sandstones		2*+3*
17) Wavy bedded sandstone and mudrock		
18) Parallel laminated sandstone		6?
19) Thin sandstone beds		11*
20) Large-scale cross-bedded sst.		10
21) Undulatory bedded sandstone		
22) Seatearth and coal		12

* Part of

Walker's pebbly mudstone (E) was not observed

CHAPTER 2 ASSEMBLAGE A: DEEP WATER SEDIMENTS

The rocks of this assemblage are divided into five facies, which are grouped into three associations.

2.1 Facies

2.1.1 Facies 1; Mudrock

Description

This facies covers unfossiliferous silt and clay grade material and is termed "mudrock", following Blatt et al, 1972. Homogenous and laminated mudrock occurs. The homogenous varieties range in colour from dark grey to black whilst the, more abundant, laminated types are medium to dark grey, with the laminae defined by dark and slightly paler layers. In thin section the carbon and pyrite content appear to be highest in the darker laminae. The laminae are parallel and are normally less than 1mm thick and never exceed 3mm. Most of the facies occurs as shale. Exposures often have a rusty brown coating of limonite or, in the finer grained shales, of yellow powdery jarosite.

Carbonate cemented concretions, up to 0.25m in diameter, have been found in the laminated parts of the facies. They are important because the laminae are thicker than the shales, allowing a more detailed examination. Within the concretion, long continuous parallel laminae of carbon free mudrock, up to 2.5mm thick, form the predominant lamination (Plate 6). The remainder consists of carbonaceous material enclosing elongated pods of carbon free mudrock. This also gives the rock a laminated appearance. Whilst the thicker, continuous laminae can be seen in the shale, it is impossible to seperate thinner continuous laminae from the pseudo-lamination. It is consequently difficult to make accurate comparisons but the concretions appear to have laminae about twice as thick as in surrounding shales.

No trace fossils have been found in the shales but two burrows have been found in concretions (Plate 6) where they are 8mm in diameter and penetrate vertically 18mm.

Interpretation

F.1 was probably deposited very slowly in quiet water. The abundant carbon and pyrite suggest lack of oxygen. Conditions may well have been euxinic, bottom conditions only rarely being favourable for benthos and never lasting long enough for a shelly fauna to become established. The homogenous mudrocks suggest constant supply of mud and carbonaceous material whilst the laminae indicate periods of accelerated mud supply.

The parallel nature of the laminae in the concretions indicate either (a) that the concretions formed rapidly, (b) the rate of sedimentation was very slow, or (c) the concretions grew at depth where the rate of compaction was slow. The thicker laminae within the concretion are probably due to the greater compaction of the surrounding shales rather than expansion during concretionary growth as that would lead to distortion and truncation of laminae. If the concretion grew close to the sediment surface a 50% compaction of the original mud is indicated. If (c) is important, however, the total compaction must be considerably more.

2.1.2 Facies 2; Goniatite Faunal Bed

Description

This facies consists of mudrock, identical to that of F.1, with fossils. The fauna consists mainly of goniatites and pectinoid bivalves. Both occur only in adult form in this assemblage. The fossils are generally aligned with their long axes parallel or at a low angle to bedding.

Several concretions have been observed in the homogenous mudrock. The concretions, dark grey in colour, may be septarian and always have a concentration of pyrite around the edges; a pyritic rind. Goniatites preserved in the concretions are usually uncrushed whilst in the surrounding shale they are flattened.

Interpretation

The importance of this facies is that it indicates that conditions were marine. The restriction of the fauna to this facies will be discussed in Chapter 8. The depositional environment is thought to have been similar to that of F1. The absence of benthonic forms, despite the evidence of a marine environment, is probably further indication that conditions were euxinic.

Septarian concretions are thought to occur by the contraction of the hydroplastic interior with water loss. They appear therefore to be characteristic of highly porous, water laden sediment and indicate early concretionary growth (Raiswell, 1971). At the present day pyrite forms, in sediments rich in organic material, within the top 5m. Both pyrite and carbonate formation can be due to sulphate reducing bacteria metabolising organic matter. Raiswell (1973) has suggested that concretions with pyrite rinds result from bacterial colonies within the sediment; the rate of pyrite formation relative to carbonate being highest as the bacterial colony dies.

No importance is attached to the collection of concretions from only the laminated mudrock of F1 and the homogenous variety in F.2 as this may be due to the low number of concretions seen.

2.1.3 Facies 3; Turbidite Sandstones

Description

The sandstones of this facies have an average thickness of about 0.1m but the sandstones vary in thickness between 0.24m and beds consisting of a single set of ripples with only 1mm height.

The bases of the sandstones are sharp and are usually flat or loaded but may have erosional sole structures. Of fifty three sole structures from which palaeocurrent data were collected 72% were grooves, 11% prods (Plate 7), 9% flutes and 8% furrows (Plate 8). Flute moulds are generally small; less than 200mm in length; and are not well formed. Two sets of groove or prod moulds, differing in orientation by up to 27°, can be seen on some bases. Load moulds (Plate 9) rarely exceed 20mm in depth and average 10mm in depth and width. No detached sandstone balls have been found. Some ripples lying directly on F.1, mudrock, are loaded but show no evidence of loading taking place during deposition. Load moulds are occasionally aligned in rows parallel to the current direction, as indicated by other sedimentary structures; this appears to be due to loading of small erosional structures now destroyed.

Four major lithotypes occur within the sandstone beds; an unlaminated, two planar and a ripple laminated division. The unlaminated division comprises up to 75% of the bed. It varies in grain size between coarse and medium sand grade. It may show a poor grading with either a concentration of a few large grains at the base or a more general upward decline in grain size.

The lower division of parallel lamination occurs in medium to fine grained sandstone. The laminae are generally parallel to the regional dip although they may be slightly undulating. They are defined by layers of mica plates lying parallel to the lamination and a parting lineation (Plate 10) is often present. The maximum thickness of parallel lamination seen is 0.22m.

Ripple lamination, defined by small mica plates, occurs in medium to fine grained sandstone. The ripples occur as either form sets or as climbing ripples. The base of the ripples may be horizontal or may truncate the underlying horizontal lamination (Fig.4, Vol.2). The climbing ripples never show stoss-side preservation and are of Allen's (1973) Type A. The angle of climb is low and ripples never climb for more than one channel length.

As the ripples are usually overlain by shale their surface morphology is often seen. Most ripples form linguoid trains (Plate 10) but straight crested ripples (Plate 11) also occur. Ripple lengths average 90mm and they have vertical form indices of about 6.

The upper division of parallel lamination occurs in fine to very fine grained sandstone and is defined by laminae of varying grain size. Mica and carbonaceous material are also abundant on certain planes but no parting lineation has been observed. Occasionally coarser grained laminae show thin ripple lamination whilst thin muddy laminae also occur. When overlying a rippled surface the basal laminae are thicker over ripple troughs than crests (see Fig.4, Vol.2) resulting in a gradual flattening upwards.

Seven different sequences of these four divisions have been observed and are shown in Fig. 3, Vol.2).

Cross bedding also occurs in the sandstones of F.3 but is rare. Only two examples have been recorded; both from Dovestones Reservoir (SE016035). The sandstone is very coarse grained with occasional pebbles up to 5mm in diameter. The sets are about 0.25m thick. The bases are flat whilst the top surfaces undulate. The cross-laminae are defined by variations in grain size and dip in the regional downcurrent direction. Mudflakes, up to 15mm in diameter, are common on the foreset slope and tend to be more abundant near the bases of the sets.

The top of the sandstones are sharp, unless the upper division of parallel lamination is present. The sequence is usually overlain by F.1, mudrock, but is occasionally overlain by another F.3 sequence, in which case the two sequences are seperated by a sharp surface, often with sole structures.

Rarely individual beds are contorted in a similar fashion to a ruckled tablecloth. Such "crumpled" beds were described by Allen (1960) from the Mam Tor Sandstone Formation, where they are relatively common.

Interpretation

Of the internal divisions, the virtually ubiquitous ripple division is the easiest to interpret. Their size limits them to the small-scale ripple field of Simons <u>et al</u> (1965) (Fig.2, Vol.2) which falls in the lower part of the lower flow regime. Allen (1968, fig.4.61) claims that straight crested ripples are more likely to occur at deeper flow depths and at lower velocities compared to linguoid types.

Some authors (eg. Allen, 1968 and Trewin and Holdsworth, 1973) have suggested that ripples with inclined bases are due to ripple formation under an erosive current. Allen has shown that for a ripple moving downward relative to its base (see Fig.5a, Vol.2):-

$$i_L = A\sigma . U \sin \zeta s$$

where i_L = net loss in unit time (weight)

U = velocity of ripple along a path inclined downward at an angle of ξ

s = specific weight.

As the ripple velocity depends on the stream power, there is a limit beyond which the ripple would be replaced by another bedform. It is therefore obvious that there is a limiting value to ξ. This limit has not, as yet, been calculated but is probably considerably less than the average of 12° seen in F.3 ripples. Ripples with similar inclined bases have however, been observed in a flume with the development of ripples on a plane bed after rising or falling stream power. The ripple crests are built up with material eroded from ripple troughs (see Fig.5b, Vol.2) and no net erosion takes place. Eventually the ripples reach the equilibrium height for the flow and move forward with $\xi = 0^{\circ}$. The ripples with inclined bases were therefore short lived, often not achieving equilibrium. Rather than net erosion, the ripples probably formed under net deposition, as suggested by the climbing ripples.

In their study of flow regimes, Simons et al (1965) identified two fields of plane beds (Fig.2, Vol.2). Both divisions of plane lamination within F.3 are too fine to belong to the lower flow regime plane bed field. As the lower division consists of relatively well sorted sand and has a parting lineation, interpreted by Allen (1964) as being a product of upper flow regime, it is thought that it was deposited in the plane bed phase of the lower part of the upper flow regime.

It is probable that most of the sediment of the upper division of parallel lamination was deposited out of suspension without passing through a traction phase, as is suggested by the infilling of ripple troughs and the thin silty laminae. Deposition from suspension should lead to a perfectly graded division but eddies in the flow could fluctuate shear stress, resulting in a laminated bed. Occasionally shear stress appears to have been strong enough to allow bed load movement of the fine sand, resulting in thin ripple laminae.

With the interpretation of the three major laminated divisions of F.3 it is now possible to discuss the type of depositing current. The upward passage from upper flow regime plane beds, through lower flow regime ripples, to the threshold of sediment movement clearly indicates a waning flow. The flute and groove structures indicate predepositional erosion. The general graded nature of the beds suggests that the current was partially sorted with finer sediment in the tail of the current. The interbedded mudrock of F.1 indicates that deposition between the currents was slow and there is no evidence of current action. It is therefore very probable that turbidity currents were responsible for the deposition of the sandstones. The upward sequence of structures is identical to Bouma's (1962) turbidite sequence (Fig.3(8), Vol.2). Allen (1960) and Walker (1966) have already interpreted the laterally equivalent Mam Tor Sandstones and Shale Grit of Derbyshire as turbidites. Postulating the turbidity current origin of the sandstones several features can be discussed in more detail.

Whilst the rate of erosion of flutes is dependant on the bottom shear stress and the characteristics of the underlying sediment, flutes can only grow if the transported sediment is not deposited in the hollow. The flute's seperation bubble must be sufficiently strong to keep it clear of sediment deposited from suspension. It follows that higher current velocities are necessary for flute growth in coarser sediment. Allen (1969) calculated that velocities over 1.5m/sec for medium sand and 2m/sec for coarse sand. The poorly developed flutes of F.3 presumably represent the general inability of the turbidity currents to sustain these speeds.

The tool marks were probably produced by the shale clasts and wood fragments carried by the current. Some grooves do in fact show a regular pattern closely resembling Calamites.

The initial division deposited in the sequence presumably depends on the flow regime at the initiation of deposition (Walker, 1967a). The origin of the massive or unlaminated division of turbidites has been the subject of much debate (eg. Walker, 1965, and Middleton, 1970). The origin of massive beds in general will be discussed in the next facies (section 2.1.4).

Cross-bedding is rare in turbidite sequences but has been observed in several coarse-grained turbidites, notably by Hubert (1966) and Maschalko (1964). Allen (1970b) hypothesized that a dune phase can only be initiated by turbidity currents carrying relatively coarse grained sediment. The pebbly nature of the cross-beds observed here is compatible with this theory.

The thin nature of the ripple division and the usual absence of the upper division of parallel lamination suggests that deceleration of the turbidity currents was more rapid than was the case for most described turbidite sequences. This may have been due to the small percentage of sediment of less than fine sand grade compared to most turbidity currents.

Walker (1966a) called the interbedded F.1, mudrock, the E division (Bouma, 1962) of the turbidites. In only a few sequences, however, is there complete upward gradation of the sandstones into the mudrock. The two therefore appear to be of completely seperate genesis in contrast to many turbidite sequences.

Allen (1960) suggested that the crumpled beds were formed by downslope sliding of the sand beds as the axes were usually normal to the palaeoslope. Slumping of this type is however unlikely to take place on a slope low enough for the deposition of turbidites. The preservation of recumbent folds necessitates deformation taking place after deposition of the overlying muds. Liquefaction of the upper layers of the sediment is therefore suggested. Similar liquefaction phenomena (eg. Allen and Banks, 1972) have been attributed to earthquake shock. If, however, the sediment has been rapidly deposited and the muds were unconsolidated, liquefaction could be triggered by minor movements associated with a turbidity current. Such a current over the sediment may explain the good orientation of fold axes.

2.1.4 Facies 4; Unlaminated Sandstones

Description

The sandstones of F.4 are, in this assemblage, coarse to very coarse grained with pebbles up to 30mm in diameter. Beds of this facies are usually greater than 1m thick and most are greater than 2m. Beds have been recorded up to 15m thick but measurement of bed thickness is difficult because of amalgamation (Walker 1966a) between adjacent beds. The bases of the sandstones are generally sharp with no sole structures, except loads. Load moulds extend up to 0.15m downwards and occasionally, where the underlying sediment is sandstone, load balls are detached. The top of the sandstone, if not cut into by another sandstone, is also sharp and usually flat but may be irregular.

Mud clasts are common; they average 60mm in length and are generally less than 10mm thick. They may occur scattered throughout the facies but often concentrate at the top of the sandstone bed or may occur along zones which when traced laterally can be seen to be zones of amalgamation. Occasionally large rafts of finer sediment, up to 10m in length, occur. Plant fragments are also common and again are more abundant at the top of the beds. No grading, other than the concentration of the large clasts at the top of the beds, has been observed.

Interpretation

Many theories have been put forward as to the origin of unlaminated or "massive" beds. Below are listed what seem to be the main possible mechanisms. The first six are depositional whilst the last two are post-depositional processes.

1/ Freezing of traction current carpet. Although traction normally results in lamination, it has been suggested that massive beds may form under conditions of rapid deposition. Walker (1965, p11) thought that the massive beds may be due to sudden "freezing" of a traction carpet with a high concentration of dispersed grains, when the shear becomes insufficient to keep the mass of grains in motion; a process described by Bagnold (1955).

2/ Upper flow regime. It has been argued by Walker (1965) and Harms and Fahnestock (1965) that deposition in the upper flow regime can form massive beds. Since then, Middleton (1965) has produced faint antidune lamination in experimental antidunes and supposed antidune lamination has been described by Walker (1967b), Hand et al (1969), Skipper (1971) and Schmincke et al (1973) from various sedimentary environments. It is however possible that certain upper flow regime conditions prevent lamination forming. If very poor lamination did form it may be hard to discern in inland exposures of coarse or highly diagenetically altered sands.

3/ Metastable field. Walton (1967) has speculated that a rapid deceleration of a current may prevent the equilibrium bed form developing. Deposition in this "metastable" field would consequently be unlaminated.

4/ Non-turbulent sediment gravity flow. Fluidized flows, grain flows and debris flow are deposited by mass emplacement and hence are structureless (Middleton and Hampton, 1973). Stauffer (1967) interpreted some thick sandstones in the lower Tertiary of California as grain flow deposits. He suggested that the following features are typical of such deposition; a) large clasts of fine sediment (some of which show partial digestion), b) dish structures, c) unusual sole markings (step-like loads, frondescent and ropy marks) and d) absence of grading or typical flysch sole markings. Middleton and Hampton (1973) show that reverse grading should also be a feature of grain flow deposits, as seen in avalanching sand. Bagnold (1954,1968) showed that the dispersive pressure within a flow would tend to push the larger grains to the area of least shear stress. Middleton (1970), on the other hand, suggested that a grain flow could have a kinetic sieve mechanism in which small grains would be able to move downwards into gaps between larger grains during flow.

5/ Syndepositional shearing. A currents boundary shear stress may be sufficient to produce rotary shearing motion in underlying loosely compacted sediment. Middleton (1967, p.494) observed the formation of such a "quick" bed in experimental, highly concentrated turbidity currents.

6/ Deposition from suspension. Deposition of sand from suspension on a surface without traction would result in unlaminated beds. Grading will result due to the differing fall velocities, although a continuous supply of sand could lead to a homogenous bed except for grading at the top and bottom.

7/ Post-depositional liquefaction and fluidization. Upward moving water, due to the compaction of underlying sediment, especially mud, expands the fabric of sand beds making them unstable. The excess pore pressure may be sufficient to destroy the fabric, in which case fluidization takes place, but a sudden shock may be sufficient to destroy the fabric at lower pore pressures resulting in liquefaction. In both cases the sand would behave like a viscous fluid although liquefaction would be more transient than fluidization.

8/ Extensive bioturbation. Bioturbation destroys the sediment fabric and, if extensive, little or no sedimentary structures would be left.

The last theory can be discounted in interpreting F.4 unlaminated beds because of the coarse grain size and lack of any significant signs of bioturbation in other facies. It is also very unlikely that deposition was from suspension because of the lack of grading and the coarse grain size, including the large clasts.

The facies has some of the characteristics one would expect of a grain flow deposit; large clasts, absence of erosional sole marks and inverse grading in the form of concentrations of clasts at the top of the beds. Unfortunately, as Middleton and Hampton (1973) point out, it is at present impossible to positively identify grain flow deposits. Deposition by turbidity currents with freezing of a traction carpet, upper flow regime or metastable conditions, syndepositional shearing or post-depositional liquefaction and fluidization may all give unlaminated beds similar to F.4.

Walker (1966a and b) favoured deposition by turbidity currents for his facies C sandstones which are roughly equivalent to those of F.4. He found some flute and groove marks, on the bases of thick sandstones, indicating turbulence. The scarcity of erosional sole structures may be a reflection of the coarse grain size; velocities of at least 4m/sec being necessary for flute growth (Allen, 1969). Highly concentrated turbidity currents may be capable of operating the kinetic sieve and clasts would be kept to the top of the flow and so prevent tool marks forming. It is unlikely, however, that the largest clasts could be transported by turbidity currents. Both turbidity currents and grain flow were therefore probably important in the deposition of F.4 in this assemblage.

The larger load structures recorded in this facies, than in F.3 turbidites, can be attributed to the disturbance wavelength increasing with smaller density differences when the bed sits on another sandstone as predicted by Allen (1970a). The thicker nature of the units must also be an important factor. The intensity of the loading where one sandstone overlies another may be due to the poorly cohesive nature of sand in comparison to mud.

2.1.5 Facies 5; Intermediate Sandstones

Description

There are some sandstones which cannot easily be assigned to F.3, turbidites, or F.4, unlaminated sandstones. These have parallel lamination, similar in form and grain size to the lower parallel laminated division of F.3, overlying unlaminated coarse to very coarse sandstone. The latter usually comprises well over 50% of the whole bed. Beds of this facies vary in thickness between 0.5 and 5m. Both bottom and top surfaces are flat with the exception of some beds with loaded bases. Mud clasts are rare but, if present, are concentrated at the base of the bed or, less often, at a level in the lower half of the massive sandstone.

Interpretation

The structureless sandstone with mudclast concentrations is reminiscent of F.4 and was probably deposited from either grain flow or highly concentrated turbidity currents. In some beds the parallel lamination is weak and could be due to shearing between the layers of grains either in the pseudo-laminar flow of grain flow (Stauffer, 1969) or during deposition (Middleton, 1970). Some parallel lamination, however, is well defined with mica and plant fragments, identical to that in F.3, turbidites.

If the F.5 beds are the deposits of turbidity currents the absence of C and D divisions (Bouma, 1962) could be accounted for by either (a) erosion of the top part of the bed in those cases where sandstones directly overlie one another or (b) the failure of the turbidity current to produce ripples because of the rapid deceleration and lack of a fine tail to the current.

The facies appears to be intermediate between F.3 and 4 and suggests that either turbidity currents were capable of producing F.4, unlaminated sandstone, or, if F.4 was a grain flow deposit, that F.5 was a product of a grain flow and turbidity current coexisting.

2.2 Sandstone Dykes

Description

Sandstone dykes have been found at several localities in this assemblage. The sandstone varies between fine and medium grain size and may contain abundant mudflakes. Many mudflakes appear to be partially "digested" and the sandstone has a muddy matrix. Dyke widths are usually between 0.1 and 0.2m. The maximum length observed was only 3m but in only one case was a termination seen. In this case (at Dovestones Reservoir, SE016035) the maximum extent of the dyke is only 0.25m and the sandstone is connected to the top of a F.4, unlaminated sandstone bed with an irregular top surface. A number of dykes (eg. Storris House railway cutting, Otley (SE180447) (Plate 5), Lumbutts Clough (SD953235) and Shaw Clough, Todmorden (SD961245) occur below the lowest medium grade sandstones in the sequence. The dykes cut the bedding at between 20 and 90°. The sides of the dykes are sharp and ptygmatic (Dzulynski and Walton, 1965) in cross-section whilst sides have ridges parallel to the truncated bedding.

Interpretation

The unlaminated nature of the sandstone dykes presumably results from liquefaction associated with their intrusion. The intrusive power appears to have been not only capable of infilling cracks but also of enlarging them. The ptygmatic-ridge nature of the edge of the dykes is almost certainly due to differential compaction of the sandstones and surrounding mudrock.

The sandstone dyke emerging from a F.4 sandstone suggests that, at least in some beds of that facies, the massive nature is due to post-depositional liquefaction, possibly from upward flowing water during compaction of the muds. The apparent extension of some dykes downwards suggests that in some cases gravity and shrinkage of the mud were more important factors than pressure by burial.

2.3 Associations

2.3.1 Association A1; Mudrocks

Description

This association comprises only F.1, mudrock, and F.2, goniatite faunal bed. The former is dominant with F.2 only occuring in beds 1 to 2m thick. Sandstone dykes have been observed in this association within a few metres of overlying sandstones of A2.

Interpretation

The association formed in quiet water with virtually all the sediment settling out of suspension. The surface water was, at least for some of the time, marine whilst bottom conditions were apparently mainly euxinic.

2.3.2 Association A2; Parallel Sided Sandstones

Description

This association comprises F.1, mudrock, and F.3, turbidite sandstones. F.3 beds appear parallel sided but may vary slightly in thickness when traced laterally. Sandstone beds are usually seperated by mudrock although occasionally one sandstone rests on another with little sign of erosion of the underlying bed. The section of part of Hole Bottom Delph, Todmorden (Vol.3, Fig.7) is typical of this association.

Discordances in the association occur in Storris House railway cutting (SE180447) (Plate 4) and at Lumbutts Clough (SD953235) (Fig.6, Vol.2). One has also been observed at Samlesbury Bottoms, Blackburn (SD618289) outside the thesis area. The discordances are relatively smooth and the beds above and below have differing dips and strikes (see Plate 4 and Fig.6 for details).

Interpretation

The association appears to be the product of mature turbidity currents entering an area of quiet water with normal deposition being from suspension. The discordances could be due to erosion by turbidity currents or bottom currents or may be slump scars. The smooth surface of the discordance is more typical of a slump scar than a channel cut into silt (Laird, 1968) and the differing orientations of the sediment seperated by the discordances could be due to the rotary movement of a slump.

2.3.3 Association A3; Wedge-shaped Sandstones

Description

In this association F.1, mudrock; F.3, turbidites; F.4, unlaminated sandstone; and F.5, intermediate sandstones, occur. The sandstones and mudrock may be interbedded but the mudrock percentage is less than 50%. Sandstone beds are less parallel-sided than in A2 and may double in thickness over 1m. The base of the sandstone beds may cut down into the underlying sandstone and in some cases, where the grain size is similar, amalgamation (Walker, 1966a) occurs.

Occasionally F.4 sandstones can be seen to sit in large channels as at Lobb Mill Delph, Todmorden (SD953245) and Dovestones Reservoir cutting (SE016036). In these cases the erosion surface can truncate the underlying bedding at up to 15°. The Lobb Mill channel is probably over 30m deep whilst the Dovestones channel can be seen to downcut by 10m. The absence of mudrock from the channels and the amalgamated nature of the sandstones make it difficult to discern bedding in these channel fills. There are however zones of mudflakes and it is in these channel fills that clasts up to 10m in length have been observed. Two F.5, intermediate sandstone beds, thicken towards and join a channel sandstone in the Dovestones cutting (see Fig.11, Vol.3 and section 7.1.1).

Interpretation

The association is interpreted as being chiefly the deposits of turbidity currents and, possibly, grain flows. The channels were probably cut by gravity flows; there being no evidence of any other currents within the assemblage. Erosion of bedrock by grain flows in the head of submarine canyons has been observed by Dill (1964). Channels cut by turbidity currents are recorded from many deep sea fans (Haner, 1971 and Shepard et al, 1969). Walker (1966b) preferred erosion by turbidity currents, rather than mass movement of sediment, for the channels in the Shale Grit of Derbyshire. He thought it unlikely that grain flow could erode on slopes low enough for turbidite deposition. This argument assumes that the channel is contemporaneous with the sediment it cuts; if the centre of gravity flow is localised, rather than spread out evenly, comparitively thin turbidites could be deposited in proximal parts of a fan, leading to oversteepening.

In one channel Walker (1966b) reported levee deposits of F.3 continuous with the channel sandstone. Here a turbidity current appears to have overflown from the channel producing sandstones thinning away from the channel. Such a turbidity current may have overlain a grain flow or may have filled the channel. A similar pattern of channel/levee deposits occurs in the proximal parts of modern fans (Nelson and Kulm, 1973, Fig.18). No definite turbidite levees have been observed in the thesis area but the F.5 sandstones thinning away from the channel in Dovestone Reservoir appear to be overbank deposits. Some of the associated beds, probably levee deposits, are F.3 turbidites.

2.4 Relationships within Assemblage

The associations are introduced into the assemblage in the order A1, A2 and A3 upwards. Above the lowest beds of A2, parallel sided sandstones, the percentage of A1, mudrock, decreases upward until it is absent in the top part of the assemblage. A2 is progressively replaced by A3, wedge-shaped sandstones, upwards. This is a general trend and there are many variations.

The assemblage is interpreted, following Walker (1966a and b), by a submarine fan model. A1 would be deposited on the basin plain and dormant parts of the fan. A2 sedimentation would occur in the lower fan, only accessible to mature turbidity currents. In the upper fan A3 would be deposited in, and on the flanks of, channels.

When a gravity flow reached the apex of the fan the decrease in gradient would cause it to decelerate. If the current was a grain flow mass emplacement would take place when the gravity stress becomes less than the yield strength of the sediment. This is likely to happen on gradients of 18° or less (Middleton and Hampton, 1973). Such deposition would therefore be restricted to the most proximal parts of the fan. Turbidity current deposits would be spread over a much wider area because a state of autosuspension (Bagnold, 1962) is approached, where gravitational energy input need only equal the friction energy loss. If the gravity flows were grain flows partly converted to turbidity currents, the lower part of the flow would be deposited in the proximal parts of the fan whilst the upper, turbulent part could continue as a turbidity current.

No attempt has been made at working out proximality indices (Walker, 1967a) as there are no long continuous sections for which a realistic index could be calculated. Whilst the mean proximality index rises higher in the assemblage there are many divergences from the general pattern. Distance from the fan channels and channel switching were probably major controls.

No evidence of fining upward sequences, due to the gradual abandonment of major channels (Mutti and Ricci Luchi, 1972) have been observed but Walker and Mutti (1973) report an example from the Shale Grit in Derbyshire.

CHAPTER 3 ASSEMBLAGE B: DELTA SLOPE SEDIMENTS

Thirteen facies are recognized in this assemblage of which three are common with Assemblage A. The facies are generally more distinct than those in Assemblage A, with fewer intermediate varieties. Various sections in the assemblage are given in Fig.8, Vol.3.

3.1 Facies

3.1.1 Facies 1; Mudrock

In this assemblage F.1, mudrock is usually of the homogenous variety. In general it is slightly coarser than that in Assemblage A, going up to coarse silt grade. This may suggest that deposition took place closer to the shoreline.

3.1.2 Facies 2; Goniatite Faunal Bed

Description

This facies is rare in Assemblage B and is generally coarser than in A,with grains up to coarse silt grade. Comminuted carbonaceous material imparts a dark colour to the sediment. As well as an abundance of goniatites and pectenoid bivalves, small gastropods, orthocones and spat occur. This facies usually occurs as shale, in which the fossils are flattened.

Carbonate concretions do however, occur. They are generally irregular in shape and are elongated along bedding planes. In fact, in some outcrops, concretions appear to have replaced entire beds. Internally the concretions may show distortion of the laminae, parallel laminae or cone-in-cone structure. In the centres of some concretions complete shelly fossils and wood fragments, up to 0.15m across and over 10mm in width, have been preserved. The former have usually been infilled by sparry cement or sediment but, if hollow, contain petroleum jelly. The goniatites are orientated randomly, with their long axes parallel, normal or at an angle to bedding. Although no measurements have been taken of the long axis orientation in vertical specimens, no apparent preferential orientation is discernible. No evidence of geopetal infilling has been found; the sparry cement only infilling the enclosed chambers of the goniatites. The cells of the wood have been infilled by carbonate, making it much harder than wood fragments found in sandstone.

In the outer part of the concretions are more closely spaced laminae. Here the goniatites, gastropods and some bivalves are crushed but not flattened and spat generally remain complete (see Plate 21). The percentage of carbonate is less than that at the centre of the concretion. Where lamination with uncrushed fossils passes into lamination with crushed fossils the decrease in thickness causes deformationof the laminae. Where the concretion contains only semicrushed goniatites the laminae are horizontal and parallel. Cone-in-cone structures are associated with parallel laminae and beds up to 0.11m thick occur. Cone apical angles average 80°.

Interpretation

The concretions with uncrushed goniatites obviously represent most closely the original sediment conditions with the carbonate infilling the pore space. There is no evidence that the growth of the carbonate expanded the sediment as this would have caused the goniatites to crack. The high percentage of carbonate may be explained by plant material shrinking during cementation allowing the cement to infill the resulting cavities without collapse of the structure. This theory is supported by areas of sparry cement adjacent to large pieces of plant material.

The shale represents sediment where no early carbonate cementation has taken place and compaction occured before quartz cementation. The plant material has also been compressed and no large fragments remain. The concretions with semi-crushed goniatites or cone-in-cone structures represent carbonate cementation during compaction. Parallel laminae are compatible with formation in partially compacted sediment (Raiswell,1971). The high apical angles of the cone-in-cone suggests that the sediment was of low hydroplasticity (Franks,1969).

The original sediment must have been loosely compacted with a considerable amount of waterlogged plant material; a typical sapropel. The large plant fragments and coarser grain size suggest deposition nearer the shoreline than in Assemblage A. The vertical goniatites suggest a lack of strong currents; there is no evidence of a unidirectional current as reported from the _Gastrioceras cancellatum_ bed by Heptonstall (1964). The random orientation of the goniatites suggests, in fact, that they did not rest on the sediment surface on death but sank down into it. The high percentage of the original sediment made up of plant material probably meant that the sediment / water interface was not clearly defined.

The lack of benthos, the abundant plant material and the lack of any evidence of currents suggests an aerobic environment. The suggested bottom conditions would, however, not be conducive to benthonic life, even if conditions were aerobic. It is a good example of a letal-pantostrat biofacies (Schafer,1972).

The assemblage of fauna in all stages of ontogeny may be explained in two ways. Firstly, it may represent the sudden death of an entire community. Such mass killings could result from upwellings of toxic anaerobic water, as observed in present day fjords (Strom, 1955). Secondly, the presence of spat may be due to proximity to a shoreline if, as suggested by Ashton (1974), goniatites bred near the coast. This theory supposes a high infant mortality rate. The presence of goniatites at all ontogenetic stages is, however, less easily explained.

In conclusion, therefore, it would appear that the facies was deposited in a euxinic, marine, possibly near shore environment.

3.1.3 Facies 4; Unlaminated Sandstone

Description

This facies has a similar grain size as in Assemblage A. Mudclasts are, however, smaller reaching only 0.45m in diameter. These may be more abundant at the base of the massive bed or can be scattered throughout it. To meet Allen's (1971) criticism of Collinson's (1970a) description of this facies as "massive" an X-ray examination of some sandstones was carried out but failed to find laminae (Plates 38 and 39).

The base of the massive beds is usually strongly erosional cutting into the underlying sediment unevenly. On the base large flutes and grooves occur. As in Assemblage A it is difficult to distinguish bedding because of amalgamation.

Carbonate concretions with diameters over 1m occur in this facies. They are almost spherical in shape, hence their local name "mare's balls" (Plate 26).

Interpretation

The origin of unlaminated sandstone has already been discussed in section 2.1.4. Flute casts grow by a "sand blasting" process resulting from flow seperation and attachment. To allow this mechanism to work, the flute must remain free of sediment. Non-turbulent flows may therefore be discounted; as can deposition from suspension without a traction phase. Extensive bioturbation is also unlikely because of the coarse grain size. The remaining possibilities must be examined after a description of the facies' context.

3.1.4 Facies 6; Laminated Silts

Description

The thickness of the laminae in this facies average 1.4mm. The laminae are size graded within the silt grade; the coarser grained laminae grade into finer grained laminae above and below and vice versa. Diastems are therefore absent. The finer laminae have a higher concentration of carbonaceous material which gives a light / dark grey banding.

Occasional small carbonate concretions occur within the facies. These are usually, slightly flattened, spheres in shape. Internally the laminae are more widely spaced than in the non-carbonate silts, averaging 2.5mm.

Interpretation

The absence of diastems indicates that deposition was continuous The lack of traction formed sedimentary structures suggests deposition from suspension. The varying grain size indicates changes in sediment supply. The varve like appearance of the sediments suggests the possibility of seasonal variation but it may be flood dependant.

3.1.5 Facies 7; Shell Bed

Description

This facies is only found at Great Dib, Otley (SE199443). The fauna includes siliceous sponges, crinoids, brachiopods, bivalves, gastropods, goniatites, orthocone nautiloids, trilobites, ostracods and fish. These are found in a coarse siltstone to very fine sandstone, although occasionally the rock is almost entirely organic in origin. The rock is cemented by carbonate which shows no evidence of being concretionary. The fossils are often in a fragmented state and they lie with their long axes parallel to bedding.

Interpretation

Facies 7 is unique in having a preserved benthonic fauna. Although not in growth position the fossils have probably not been transported a great distance. The benthonic fauna, transporting currents and paucity of plant material suggests a fully aerobic environment. On Schafer's (1972) classification the shell bed is best described as a vital lipostrat biofacies although there is no evidence that the fauna is actually <u>in situ</u>. Such a facies would be expected in an open shallow environment.

No evidence of compaction of the shells has been seen which suggests that cementation was early, although not necessarily as early as for the uncrushed goniatites of F.2 because of the thicker nature of the shells.

3.1.6 Facies 8; Gradationally Laminated Sandstone

Description

This facies is similar to F.6, laminated silts, in the nature of the lamination but differs in grain size. The coarsest laminae observed were of fine to medium sand whilst the finest were of coarse silt to very fine sand. Again the laminae have a distinct light / dark banding which easily distinguishes the facies in the field. The thickness of the laminae increases with grain size; the fine to medium sands have average thicknesses of 25mm in contrast to the 10mm for fine to very fine sands. The rate of gradation between finer and coarser grain laminae varies. In most rocks the zone of maximum grain size is very thin and grades gradually down to finer grain sizes. In contrast, the coarser zone may be thick with rapid gradation to the finer zones. The coarser sediments split along the finer laminae into good flags.

A few carbonate concretions occur. These are irregular in shape but are elongated along bedding planes, appearing to have replaced entire beds in some small exposures. The thickness of the laminae is again larger in the concretions; the fine / very fine sandstone laminae increasing to an average of 19mm thick (see Plates 12 and 13).

The finer parts of the facies often contain abundant trace fossils; Scolicia, Planolites and Pelecypodichnus (a full description of trace fossils is given in section 3.2).

Interpretation

Deposition of these sandstones appears to have been similar to that of F.6, laminated silts, that is, from suspension. The grain size indicates relatively strong currents above the bed but there is no evidence of this impinging onto the sediment surface to give tractional reworking. The laminae presumably reflect changes in the strength of the overlying current.

3.1.7 Facies 9; Micaceous Silty Sandstone

Description

This facies is rare in Assemblage B. It has a high mica content with plates up to 3mm in diameter. The grain size of the quartz and feldspar ranges between coarse silt and medium sand. A lamination, produced by differences in grain size, is emphasized by the mica content and often by abundant plant fragments. The latter may be quite large branches. Pelecypodichnus is occasionally found in this facies.

Interpretation

The grain size indicates currents of varying intensity but the abundance of easily transported material suggests deposition in a relatively sheltered area, from suspension.

3.1.8 Facies 10; Ripple Laminated Sandstone

Description

Beds of fine to medium sandstone, entirely cross-laminated on a small scale, make up this facies, Beds up to 2m thick have been observed. The bases of beds are usually sharp, sometimes with prod marks. Individual cross-laminated sets are up to 15mm thick. The ripples have trough shaped bases and curved foresets. The top surface therefore shows a typical rib and furrow pattern. The upper part of the ripple foresets is truncated at varying levels by the overlying cross-laminated set. If the truncation is low in the set the sandstone looks, superficially, parallel laminated.

Pelecypodichnus is common in this facies and its escape traces are up to 0.13m long.

Interpretation

The facies is interpreted as being deposited in the small-scale ripple field of the lower flow regime (Simons et al,1965) (see Fig.2, Vo.2). The ripples obviously migrated under net sedimentation to produce the cosets of cross-lamination. The bases of the sandstone beds indicate slight predepositional erosion.

3.1.9 Facies 11; Ripple Laminated Sandstone with Coarse Sandstone

Description

The majority of this facies consists of small-scale cross-laminated fine to medium sandstone. A combination of mica flakes, carbonaceous material and silt emphasize the lamination of both symmetrical and assymmetrical ripples. The symmetrical ripples have sharp crests, with chords about 80mm and vertical form indices of 7. Internally, lamination dips in both directions. The cross-laminated sets associated with assymetrical ripples occur as cosets. Sets have scoop-shaped bases and the cross-laminae are discordant to the surface. Form sets are rare but, when seen, are about 8mm in height. The orientation of both types of ripple are variable.

Interbedded with the rippled sandstone are beds, mostly between 10 and 40mm thick but up to 0.1m thick, of coarse to very coarse sandstone with no observable internal structure. The base of these beds may be loaded but generally the sandstone rests on the underlying ripple morphology without any evidence of erosion. A few of the thicker beds are distorted and recumbent folds have been seen.

Pelecypodichnus occasionally occurs in the finer part of this facies.

Interpretation

The symmetrical ripples indicate wave action and suggest shallow water conditions. The unsorted nature of the sandstone is, however, atypical of wave worked sediment and wave action was probably limited. The asymmetrical ripples appear to be the result of migration of ripples with irregular crestlines (Allen, 1968). The draping of the massive sandstone over the ripples suggests deposition from suspension despite the coarse grain size. This necessitates occasional strong currents which did not impinge on the sediment surface; a situation which could arise near the mouth of a river. The distorted beds presumably result from liquefaction of the upper layers of sediment, as suggested for the crumpled beds of F.3, turbidite sandstones.

In conclusion, the ripples, unsorted sediment and range of lithotypes indicate a highly variable environment where wave strength and the input of coarse sediment from suspension appear to have been the major factors controlling sedimentation.

3.1.10 Facies 12; Sharp Based Sandstones

Description

The sandstones of this facies range in thickness between 1mm and 1.5m. The base of the sandstones are sharp and may show tool marks, flute or furrow moulds or may be flat. Both prod and groove moulds are often in abundance. Flutes (Plates 20 and 22) are less abundant but are occasionally well developed and broad. One bed, 0.12m thick, has flute marks of maximum width 0.12m with depths of 35mm.

Four lithotypes can be distinguished:- (1) Massive, medium to fine sandstone. (2) Parallel laminated fine sandstone defined by mica plates up to 1.5mm in diameter. The laminae may show a parting lineation. (3) Parallel laminated fine to very fine sandstone defined by large mica plates, plant material and, occasionally, small mudflakes; this lithotype grades from fine sandstone to siltstone upwards. (4) Ripple laminated fine sand. The ripples occur as form sets.

Whilst some beds are composed entirely of (2), others have a massive base or a rippled top surface. These beds have a sharp top. Beds with (3) may have a massive base but the parallel lamination always grade up into the overlying silt. Some thin beds consist entirely of rippled sandstone, (4), whilst the thinnest beds are only a few grains in thickness and no sedimentary structure is discernible.

Beds without (3), parallel lamination, often contain trace fossils; Sinusites, Bergaueria, Pelecypodichnus, bulbous and knob-ended burrows and various unnamed trails.

Interpretation

The origin of unlaminated sandstone overlain by parallel laminated sandstone has already been discussed in reference to turbidites in Assemblage A. Lithotypes (2), parallel lamination, and (4), ripple lamination, are typical of the lower part of the upper and lower flow regimes respectively (Simons et al, 1965). Type (3) plane lamination is more difficult to explain. Whilst it is feasible that plane beds of the upper flow regime could form graded parallel lamination under waning flow conditions (see Fig. 2, Vol. 2) it seems unlikely that it would grade perfectly into the overlying silt. It is, however, too fine to belong to the parallel lamination field of the lower flow regime. A mechanism of deposition from suspension, as suggested for the F.3, turbidite D division, would be unlikely to be directly underlain by massive beds or well-developed flutes. The large amount of plant material may be an important factor as it could have supressed a ripple phase by damping turbulence, thus extending the field of upper flow regime plane beds into lower stream powers.

Although superficially resembling turbidites, in their sole structures and internal lamination, the shallow water conditions, as indicated by the trace fossil assemblage (see sectio 3.3), suggests that other currents may be responsible. Reineck et al (1968) have described rippled and parallel laminated fine sand in "storm layers" from the southern North Sea, which are similar in many ways, including the biogenic structures, to F.12. Such layers are thought to occur when the build up of water near the shore, during a storm, leads to a strong undercurrent transporting the coarse sediment, of the shore area, seawards.

A further discussion of the facies is given after a description of its relationships to other facies.

3.1.11 Facies 13; Medium-scale Cross-bedded Sandstone

Description

Cosets of cross-bedding occur in medium to coarse sandstone. Individual sets range between 0.07 and 0.6m and usually have a trough shaped base. The bases of the cosets generally appear flat but where they overlie silts they may be erosional. Occasionally mud clasts, up to 50mm in diameter, occur on the cross-bed foreset. Mud clasts, especially where they are very abundant, predominate on the lower part of the foresets.

Interpretation

The trough cross-beds are thought to be the deposits of dunes with curved crestlines under net sedimentation (Allen, 1968). This places the flow in the upper part of the lower flow regime (Simons et al, 1965; see Fig. 2, Vol. 2). A further description of this facies is given in Assemblage C where it is more abundant.

3.1.12 Facies 14; Zeta Cross-stratified Sandstone

Description

Zeta cross-stratification (Allen, 1963) with stratification parallel to channel sides, is displayed by some coarse sandstones (Plate 24). Complete channels have not been seen but they appear to be in the order of 8m wide and 1m deep. The channel sides and associated stratification may dip up to 16°. Channels have been seen cutting into siltstone and sandstone.

Interpretation

The absence of a mud drape over the erosion surface indicates that there was no long gap between erosion and infilling of the channel. The even deposition of sand over the entire channel suggests deposition from suspension. The high fall velocity of the sand almost certainly implies a strong current probably extending wellbeyond the confines of the present channel. The cutting and filling of the channels therefore appears to have taken place during one event. Any infrequent current having a large sediment load in suspension, such as a storm rip current or crevasse current, could deposit zeta cross-straification.

3.1.13 Facies 15; Plane Laminated Coarse Sandstone

Description

Grain size in this facies varies between medium and very coarse, slightly pebbly sandstone. The parallel laminae vary in thickness between 5 and 40mm, generally dependant on grain size. In the finer sandstones a parting lineation often occurs. The facies differs from the plane lamination, (2), of F.12, sharp based sandstone, in grain size and in having less mica. The lamination is parallel to the underlying erosion or depositional surface. Where it overlies the massive beds of F.4 the lamination is approximately horizontal. Beds of this facies are up to 2.5m thick. Mud clasts are generally absent but do occur, on some bedding planes in abundance.

Interpretation

The well developed lamination points to deposition by traction currents, presumably in the lower part of the upper flow regime as it seems unlikely that the lower phase of plane beds (see Fig. 2, Vol. 2) could deposit such thick beds. The parting lineation also indicates upper flow regime conditions (Allen, 1964).

. . . .

3.2 Trace Fossils

A great variety of trace fossils occur within this assemblage but they are restricted in their facies distribution.

3.2.1 <u>Bergaueria</u> Prantl 1946

Present in facies: 12, sharp based sandstones.

Vertical or near vertical burrows, approximately 10mm in diameter with convex hyporelief and concave epirelief. The burrows have clear centres surrounded by distorted laminae.

3.2.2 Pelecypodichnus Seilacher 1953

Present in facies: 8, gradationally laminated sandstone
9, micaceous silty sandstone
10, ripple laminated sandstone
11, ripple laminated sandstone with coarse sandstone
12, sharp based sandstone

Pelecypodichnus occurs as convex hyporelief and concave epirelief. The relief is almond shaped (Plate 32) and varies in size and shape with facies:

	Maximum length	Maximum width	Maximum relief (in mm)
Facies 8	8	5	2
Facies 10	40	12	13
Facies 12	28	9	4

The endichnial lamination between the hypo- and epirelief is V-shaped.

Seilacher (1953) suggested that Pelecypodichnus is the resting trace of a bivalve. Hardy (1970) and Eagar (1974) describe Carbonicola and Anthraconaia in burrowing position. Two examples of Anthraconaiad bivalves were found in F.12, sharp based sandstones, (Plate 31), but were not in life position. The presence of Pelecypodichnus therefore suggests fresh or brackish water conditions.

The V-shaped traces are thought to be due to the bivalves "escaping upwards" during sedimentation. Nucula produces similar traces at the present day (Reineck et al, 1968). Eagar (1974) suggested that the bivalves depth of burrowing was related to the substrate and current strength, stronger currents leading to deeper burrows. The difference in size of Pelecypodichnus between F.8 and F.10 bears out this theory. The size of the burrows in the sharp based sandstones may not be directly related to current strength as the current was probably ephemeral.

3.2.3 Planolites Nicholson 1873

Present in facies: 8, gradationally laminated sandstone

Vertical or steeply inclined burrows occur in the coarser laminae and horizontal burrows (Plate 33) occur in the finer laminae. The infill of the burrows is of very clean sand. Burrows are generally 1mm wide. No fixed pattern to the burrow organization can be seen. The clean nature of the infill suggests active infill from an organism foraging in organic rich sediment.

3.2.4 Scolicia de Quatrefegas 1849

Present in facies: 6, laminated silts
8, gradationally laminated sandstone

Scolicia is a trilobed, endichnial, wandering trail, 15mm wide with occasional sharp loops (Plates 34 and 35). The median lobe never exceeds 2mm in width. The larger lateral lobes are ribbed. The organism that produced Scolicia is thought to have burrowed at favourable horizons, usually a sand / shale interface (Seilacher, 1962). In F.6 and F.8 it predominates in the finer laminae.

3.2.5 Sinusites Demonet and van Straelen 1938

Present in facies: 12, sharp based sandstones

This trace fossil has a sinusoidal trail (Plate 29) with an average meander wavelength of 7.5mm and amplitude of 1.2mm. The maximum wavelength is 15mm and amplitudes are up to 3mm. The relief of the trail varies between 1 and 2mm and is dependant on wavelength. Convex hyporelief and concave epirelief suggest that these were surface trails.

3.2.6 U-shaped Burrows

Present in facies: 12, sharp based sandstones

U-shaped burrows range between 1 and 6mm in diameter. The burrows are found in the sandstone bed but the bases may be in the underlying mudrock. They do not, however, penetrate the mudrock by more than the diameter of the burrows. They are therefore endichnial and hypichnial. U-shaped burrows are thought to be dwelling structures (eg. Goldring, 1971).

3.2.7 Bulbous Burrows

Present in facies: 12, sharp based sandstones

These burrows occur as positive hyporelief, up to 30mm in diameter and 14mm in depth (Plate 37). Loading has probably emphasized some of the structures but their widely scattered occurence on an otherwise flat surface suggests that they are more than loading phenomena. Their origin is however uncertain.

3.2.8 Knob-ended Burrows

Present in facies: 12, sharp based sandstones

These burrows are vertical through the sandstone and then become horizontal following the sandstone base. The horizontal part is generally straight, up to 80mm long and up to 6mm wide. It ends in a knob 50% wider than the burrow tube (Plate 36). The burrows are hypichnial to exichnial. The knob may represent the dwelling position occupied by the burrowing organism.

3.2.9 Surface Trails

Present in facies: 12, sharp based sandstone

On the bases and top surfaces of F.12 sandstones there are often long trails about 3mm wide and with 2.5mm relief. Their convex hyporelief and concave epirelief indicates that they were infilled surface trails.

3.2.10 The Trace Fossil Assemblage

Bergaueria and Pelecypodichnus are diagnostic of Seilacher's Cruziana facies whilst Scolicia and U-shaped burrows are also usually found in this facies. The ichnocoenosis therefore suggests a shallow water environment. Pelecypodichnus has been found in association with Sinusites, Planolites and Scolicia which may therefore have been produced by fresh to brackish water organisms.

. . . .

3.3 Associations

It is virtually impossible to split the rocks of Assemblage B into neat associations because of the great variations in facies both laterally and vertically. It is, however, possible to group some of facies into four major associations containing genetically related facies which occur together.

3.3.1 Association B1; Fine Grained Deposits

Description

This association is made up of Facies 1, mudrock, and 6, laminated silts. F.1 is the dominant facies, especially in the lower part of the assemblage. Higher in the sequence F.6 increases in abundance. Although the least common of the four associations, B1 can occur in successions up to 26m thick (eg. Lumb Clough (SE007430), see section in Fig. 8, Vol. 3) but is usually less than 10m thick.

Interpretation

The association represents slow deposition from suspension with virtually no traction.

3.3.2 Association B.2, Inclined Units

Description

F.8, gradationally laminated sandstone, F.10, ripple laminated sandstone, and F.15, plane laminated coarse sandstone, make up this assemblage. F.8 is usually more abundant than F.10 whilst F.15 is uncommon. The facies occur together as beds inclined to the regional dip by up to 16°. The thickness of such inclined units is up to 14m.

Inclined lower surfaces of the unit are often seen cutting into another inclined unit or B3, channeled coarse sandstones. A near horizontal base is only seen in two localities. Plate 19 shows the base of a unit in Hebden Dale (SD972306); there are no sole structures at the base of the sandstone and the underlying mudrock is apparently completely homogenous with no trace of bedding. In Dove Stone Clough (SE028042) the association appears to sit in two flat bottomed channels although only one side of each channel is seen.

The inclined beds are almost parallel to the lower bounding surface of the unit. The dip may gradually decrease or increase upwards. In a 200m section on the Rochdale Canal, Hebden Bridge (SD977266) the dip decreases gradually away from the inclined lower surface (see Fig. 11, Vol. 3). Individual beds are virtually parallel sided and no cases of beds changing their internal structures laterally have been observed. Dip directions of the inclined beds may apparently be of any orientation but are predominantly to the south or east (Fig. 7, Vol. 2). The ripples of F10 always indicate a current towards the south.

Disturbed beds occur at the base of an inclined unit at the western end of the canal section, Hebden Bridge (SD977266, see Fig.11, Vol.3). Faulting occurs at the base of a unit in Hebden Dale (SD972306) see Fig.22, Vol.2 and section 7.2.3).

Interpretation

The smooth nature of the contact, with the underlying sediment, is more characteristic of a slump scar than an erosion surface, where differing resistances of beds to erosion gives rise to a stepped surface (Laird, 1968). Evidence of slumping is also provided by the disturbed beds at the base of the inclined units.

The heterogenous lithology, the concordant relationship of the beds to the underlying, inclined basal surfaces and the orientation of many inclined units normal to the palaeocurrent pattern suggests that the inclined units are the product of lateral accretion. Epsilon cross-bedding (Allen, 1963), interpreted as deposition on a point bar, has heterogenous, wedge-shaped beds, a fining upward sequence and an upward change in sedimentary structures indicating falling stream power (Allen, 1965). The inclined units described here therefore differ in several respects from Allen's model.

Why do the rippled beds in the inclined units not pass downwards into cross-bedded units or thin towards the top of the inclined unit? It is possible that a high concentration of suspended material could have dampened turbulence preventing dunes from forming and thus extending the ripple field. This does not, however, account for the parallel nature of the beds on the inclined units.

Allen (1970a and b) presents a model to show the importance of the main channel hydraulic variables in determining the pattern of sedimentary structures seen in fining upward cycles. This model was examined to see if a pattern of ripples throughout epsilon cross-stratification can occur. The range of depths occupied by ripples can be calculated from the formula (modified after Allen, 1970a (8)):-

$$w = V\,D_f\,g\,S\,y$$

where; w = stream power

V = mean velocity

D_f = fluid density

g = acceleration due to gravity

S = water surface slope

y = local channel depth.

D_f, g and S are constant for any given section in a channel. As V is dependant on depth and if one assumes a constant fall in velocity upwards, a simple relationship exists between w and y. The critical stream power for the development of dunes in fine sand is 750 ergs/cm^2/sec (using data of Guy et al, 1966). Postulating this, the upper limit of the ripple field, as the stream power at the bottom of a channel, the stream power for all depths of the channel can be calculated. Where the stream power falls below 100 ergs/cm^2/sec no sediment movement takes place (Guy et al, 1966 and Fig.2, Vol.2). The top 38% of a channel, with any maximum depth, falls within this field. It therefore follows that for any channel, with ripples at the bottom, ripples will occur only in the lower 62% or less of the channel depth. In B2 it appears that either the flow depth was greater than the thickness of the inclined units or that the units' upper part has been eroded away.

Two models for the deposition of the association are proposed:-

(a) Mouth Bar

This theory assumes that the downcutting basal surfaces are due to erosion. Although no modern mouth bar has been described in sufficient detail it would appear that the bars that occur on the sides and between the bifurcating flow pattern (Mikhailov, 1966) are probably built by lateral accretion. The slopes of submerged bars would provide the regional hydrodynamic situation for the deposition of parallel beds by accretion. The cutting of new channels and abandonment of old ones as the mouth bar develops can explain many of the features seen in this association.

Mikhailov (1966) reports that, in many deltas of the U.S.S.R., silt covers the mouth bar area at normal river stage, whilst river sand is only introduced during high stage. Coleman and Gagliano (1965) report that parallel laminated fine sands, similar to F8, gradationally laminated sandstone, and current ripple lamination are abundant in the Mississippi mouth bars.

(b) Slope Gully

The evidence of slumping rather than erosion suggests that the units may be the result of accretion onto a slump scar. Gullies, thought to be the result of slumping, are common on the oversteepened slope in front of the distributaries of the Mississippi (Shepard, 1955,1956,1960) and the Fraser River delta (Mathews and Shepard, 1962). Gully widths are in the order of 0.8km in the Mississippi and 0.12km in the Fraser delta. They have depths up to 15m and sides have slopes up to 15°. Most gullies occur over a water depth range of 10 to 60m. The gullies evidently infill by accretion on the gully floor and slope, with the same sediment as is being deposited on the main delta slope.

Slope gully deposits must therefore closely resemble Association B2 in geometry. F.8, gradationally laminated sandstone, is a typical deposit from suspension but the ripples of F.10 necessitate occasional down slope currents. Scruton (1956) suggested that at certain times of the year, when the temperature and salinity profiles are near uniform, a slight increase in suspende matter could lead to a turbidity current down the Mississippi slope gullies. Whilst this necessitates exceptional conditions in a marine environment, density currents would be more likely to occur in the less saline environment indicated by Pelecypodichnus. The subaqueous channel of the Rhone River delta in Lake Geneva is well documented (Houbolt and Jonker,1968) and Shepard and Dill (1966) record sand being transported down the channel as ripples with lengths of 100mm and heights of 25mm.

River generated density currents, whether owing their density difference to temperature or the amount of suspended material, are likely to flow for a longer duration than the classical slump generated turbidity current. Such currents could deposit the rippled beds with their sharp bases and prod marks. The down-cutting of ripples into one another may represent the relatively low amount of sediment coming from suspension compared to the high fall out of sediment forming climbing ripples in a typical Bouma (1962) turbidite sequence.

Finally, the occurence of F.15, plane laminated coarse sandstone, in this association is difficult to explain by either mechanism, although it has only been seen in one locality. The problem of upper flow regime in deep channels will be discussed in the next association (B3). If the plane lamination belongs to the lower flow regime (no parting lineation has been found in this association) the lack of intermediate cross-bedding presents no problem.

3.3.3 Association B3; Channeled Coarse Sandstone

Included in this association are F.4, unlaminated sandstone; F.9, micaceous silty sandstone; F.13, medium-scale cross-bedded sandstone, and F.15, horizontally laminated coarse sandstone. The sandstones overlie erosion surfaces with strikes roughly parallel to the palaeocurrent direction, as indicated by flutes on the base and by independant structures above and below the channels. Collinson (1970) described similar sandstones from the Grindslow Shales to the south and showed that the erosion surfaces are actually channel sides. These channels cut into B2, inclined units, or other B3 channels.

Channels are often steep sided with slopes up to 40° to the horizontal. No vertical or overhanging sides to the channels, as described by Collinson.(1970), have been seen. The sides are often stepped, even when one sandstone cuts into another.

The channel dimensions are difficult to ascertain because of the incomplete exposure and because channels frequently cut into older channels. The maximum depth of erosion observed is over 9m in Chew Brook (SE029016). Channel fill sequences have been measured between 4 and 10m thick but many channels were undoubtedly larger and Collinson (1970) records thicknesses of 30m. As no channel shows both sides it is impossible to measure channel width

Exposures like Great Dove Stone Rocks (SE024038) (Plate 1) show thick massive beds continuos over 200m without signs of channel margins. Collinson (1970, p.498) records one well exposed channel as being over 530m wide.

The massive sandstone (F.4) is the dominant facies and occurs in the lower part of the channels. Higher in the channels, the horizontally laminated sandstone, F.15, predominates but the cross-bedded sandstone, F.13, may occur in the top part of the channel, usually above F.15. F.9, micaceous silty sandstone, has been found on the near flat parts of stepped erosion surfaces where a channel cuts into another massive sandstone.

Interpretation

The size of the channels and the flute marks indicate that the channels were cut by strongly erosive turbulent currents. The finer nature of F.9 indicates a near cessation in sedimentation between the erosion of some surfaces and the deposition of the massive sandstone. The restriction of this facies to terraces suggests that infilling was periodic although the general absence of fine grained deposits (F.9 has only been observed in two channels) may indicate that erosion and deposition resulted from the same process. The transition from plane lamination to cross-bedding in the upper part of the channel indicates a waning flow.

As the association often cuts into B2, inclined units, for which two distinct environments have been suggested, it is necessary to discuss the genesis of B3 in two environments:-

Mouth Bar

Three types of turbulent current are known to be capable of eroding channels; tidal, turbidity and fluviatile. It is also possible that hurricanes (Hayes, 1967) and tsunamis (Coleman, 1968) could cause a strong enough surge to cause local erosion. Tidal currents can be discounted because of the lack of any evidence of tidal activity such as a bimodal palaeocurrent distribution or flaser bedding. Turbidity currents gain their energy from gravity. It seems unlikely that the gradient of a shallow water environment could be sufficient to maintain erosive turbidity current flow. Although the importance of catastrophes should not be discounted, it is unlikely that off-surge currents would scour such steep sided, deep channels.

Assuming a shallow water environment the association therefore appears to have been formed by ephemeral, fluviatile currents pushing the river's traction field beyond its normal limit. The sequence of channel erosion followed by a waning flow infilling the channel is probably the result of a flood cycle.

Envisaging such a situation, Collinson (1966) suggested that the massive sandstones were deposited under antidunes. He later (Collinson, 1970) discounted this theory as he thought the depth of the channels was prohibitive to antidune formation. A mean velocity of almost 10m/sec would be necessary to form antidunes to a depth of 10m, assuming a Froude number of 1. Such velocities may seem improbable but velocities of 8.7m/sec are necessary for the plane beds to a depth of 14m, reported in the Brahmaputra (Coleman, 1969), assuming a Froude number as low as 0.75. As bedforms on a Brahmaputran scale occur in the overlying assemblage antidune conditions were possibly present.

Other mechanisms of massive bed formation, discussed in section 2.1.4, that could account for the unlaminated beds of this association, are freezing of a traction carpet, shearing of loosely compacted sediment by a highly concentrated flow or deposition in a metastable field. Collinson (1970) apparently favoured a modification of the latter theory, with the supply of material to the traction carpet being large enough to swamp the sorting process.

In a mouth bar situation the possibility of salt wedges affecting sedimentation should be considered. Scruton (1956), in the Mississippi, and Nelson (1970), in the Po, report areas of "wild turbulence with great eddies" at the hydraulic jump where fresh river water meets saline water. If this turbulence occured at the end of a salt wedge, as in the Po, it could provide a mechanism for the formation of massive beds, either by rapid disorganized deposition or by shearing of loosely compacted sediment. If the flow of the river was in the upper flow regime, the amount of turbulence generated at the hydraulic jump may have been sufficient to erode channels. This mechanism does, however, require rapid erosion and infilling of the channels as the salt wedge was flushed seawards during rising stage.

Slope Channels

Turbidity currents are known to be capable of eroding channels on a slope (eg. Haner, 1971). It is not necessary to explain the entire erosion of the channels by turbidity currents as they may have just modified pre-existing gullies. The thick channel infill, without grading and with cross-bedding is not typical of turbidity current deposition. Such deposits have however been described from other formations and have been termed "fluxoturbidites" (Kuenen, 1958) or "pebbly and conglomeratic facies" (Walker, 1970; Walker and Mutti, 1973). It is suggested that after the peak of the turbidity current flow the current was not competent to carry the coarsest sediment and deposition took place to approach the equilibrium point where frictional energy loss equals input of gravitational energy; a state of autosuspension (Bagnold, 1962).

The flute casts on the channel bases suggest that the current was fully turbulent during all stages of deposition. The formation of massive beds would therefore have to be explained by deposition under upper flow regime, shearing of loosely compacted sediment or deposition in a metastable field. Cross-bedding has been observed in several similar "fluxoturbidite" deposits (eg. Dzulynski et al, 1959) and "slope" turbidites (Thompson and Thomasson, 1969). As Allen (1970) suggested turbidity currents will have a dune field if the grain size is coarse.

The loss of energy by a turbidity current on a steep slope suggests that the current already had considerable energy on reaching the slope. It is therefore suggested that the fluviatile currents of the overlying assemblage were capable, at high discharge, of pushing heavily laden suspension currents over the slope. Once on the slope the current could develop autosuspension. The Congo river and its canyon (Heezen et al, 1964) provide a modern analogue for this process.

Walker (1966a and b) and Collinson (1969 and 1970) report turbidite infilled channels at the base of the assemblage. No definite channels of this type have been seen in the thesis area but the massive sandstones at Lee Wood, Heptonstall (SD991282) are more similar to those of Assemblage A than B. The poor exposure in the lower part of the assemblage may explain the apparent absence of such channels.

Assemblage B3 may therefore represent an upper slope deposit of a turbulent current transitional between fluviatile and turbidity current. The channels below, filled with coarse grained turbidites indicate that these currents fed the underlying submarine fan.

Whilst both interpretations of the origin of the association are plausible their relative merits will be discussed after an examination of the relationships within the assemblage.

3.3.4 Association B4; Parallel Bedded Sandstone and Mudrock

Description

This association includes F.1, mudrock, and F.12, sharp based sandstone; the latter being confined to the association. The two are interbedded and no examples of sandstone beds cutting down into another, or amalgamation, have been seen. The maximum observed thickness of the association is 15m at Colden Clough, Hebden Bridge (SD977281).

In Colden Clough the lower part of the association is contorted destroying the original sedimentary structures. The sandstones are distorted into folds with horizontal axes, large ball structures and boudins (Plate 23). The deformed beds have a sharp top overlain erosionally by a sandstone bed of F.12.

Interpretation

B4 represents rapid influx of coarse sediment with traction into an area of normally quiet deposition from suspension. As the association is often interbedded with the lateral accretion beds of B2, it is unlikely that the sands were deposited by turbidity currents on a submarine fan. It is more probable that flood or turbidity currents responsible for B3, channeled coarse sandstones, also deposited B4 as an overbank deposit.

The contorted structures are typical of slumps with evidence of the sediment having been plastic, under tension and moving before erosion and deposition of overlying beds. Slump folds can be important in indicating the palaeoslope, although the conclusion of Lajoie (1972), after studying slumping snow, suggests they may give erroneous results unless well documented over a large area.

. . . .

3.4 Relationships within Assemblage

Description

Of the three assemblages, B shows the most lateral variation as virtually any configuration of facies may occur. There are, however, some general trends. The assemblage is essentially a coarsening upward sequence with silt dominant in the lower part and coarse pebbly sandstone dominant in the upper part. Association B1, fine grained deposits, tends to pass upwards into B2, inclined units, and B4, parallel bedded sandstone and mudrock, which apparently reach their maximum development at about the same level. B3, channels apparently occur throughout the succession but increase in abundance upwards, being isolated in the lower part and forming multiple units in the upper part of the sequence.

The position of those facies not included in the association can now be discussed. Goniatite faunal beds, F2, occur at several levels in the assemblage and are associated with B1, fine grained deposits. The E7, shell bed occurs at the top of the assemblage.

F.11, ripple laminated sandstone with coarse sandstone, is also rare and occurs at the top of the assemblage overlying massive sandstones of B3 at Eastburn Quarry (SE020440) (see section 7.3.3). Zeta cross-stratification, F14, is uncommon and is usually found with F.12, sharp based sandstone, high in the sequence.

Trace fossils are uncommon in the lowest third of the sequence but above this are abundant.

Interpretation

The whole assemblage is interpreted as a coarsening upward delta front sequence. Two theories are suggested. Firstly, if the B2 inclined units are mouth bar sediments most of the sequence appears to be shallow water in origin. The mouth bar does not seem to have lead to bar fingers, as in the present day Mississippi (Gould, 1970) but covered a large shallow area with the river mouths constantly changing position. With this theory only the bottom third of the sequence is interpreted as the slope down to the underlying turbidite fans.

If the inclined units are, on the other hand, delta front gullies then the majority of the assemblage is slope deposit. It is probable that slumping would be restricted to the steep upper part of a concave slope as in the present day Mississippi and Fraser River delta (Shepard 1955, Mathews and Shepard 1962). At least two types of density current, both river generated, appear to have been important sedimentary processes.

The predominance of accretionary units inclined to the south and east suggests lateral migration of the delta towards the south-east. A higher percentage of units inclined towards the north and west would be removed either by erosion of new river channels or by slumping.

For two reasons the slope gully theory is preferred. The absence of evidence of erosion at the base of the inclined units suggests slumping. Secondly, the thickness of that part of the sequence with inclined units necessitates rapid subsidence if they are all mouth bar deposits, even if these covered a large area.

The facies not included in associations can now be interpreted. Shell beds are common in shelf sediments associated with modern deltas (eg. Coleman and Gagliano, L965, Oomkens, 1970) where they form away from distributary mouths. Whilst both fossiliferous facies were probably deposited distal to the distributaries, F.2, goniatite beds probably formed in a sheltered environment with restricted circulation whilst F.7 was deposited in a more open situation.

F.11, ripple laminated sandstone with coarse sandstone was obviously deposited in shallow water, as evidenced by the wave ripples. The deposition of coarse sand from suspension implies very strong surface currents. It is suggested that this facies is the mouth bar deposit. Both current and wave ripples and abundant wood fragments are present in the Mississippi mouth bars (Coleman and Gagliano, 1965). The mouth bar was probably usually destroyed by the advancing distributary channel, which could explain the scarcity of this facies.

F.14, zeta cross-stratification is probably associated with B3, channeled coarse sandstones. It may have been formed by small offshoots of the major current or may represent the minor currents which never scoured deep channels.

CHAPTER 4 ASSEMBLAGE C : DELTA TOP DEPOSITS

Thirteen facies have been identified in this assemblage; six of which are common with one or both of the previous assemblages. Three associations have been distinguished.

4.1 Facies

4.1.1 Facies 1, mudrock

Description

This facies is generally coarser than in A, the deep water assemblage, but finer than in B, the slope assemblage. It also differs from both in having more abundant plant fragments. These are up to 20mm across and are generally scattered, being far less abundant than in F.9, micaceous silty sandstones.

Interpretation

The facies is again interpreted as the product of quiet deposition from suspension. The abundant plant material suggests a nearshore environment.

4.1.2 Facies 2, Goniatite Faunal Bed

Description

Two faunal beds, of apparent widespread extent, are included in this assemblage; The Butterly and R.gracile beds. The Butterly bed is only poorly exposed within the area studied. The R.gracile bed strictly belongs to R_2 but is included in this assemblage because the basal R_2 sediments seen are identical to those of Assemblage C. The bed is up to 2.7m thick and is composed of clay to silt grade material. The amount of carbonaceous material is substantially less than in this facies in Assemblage B. Both goniatites and pectinoid bivalves occur and, as in B, all stages of ontogeny are preserved. The goniatites may be positioned at any angle to the bedding, although the majority lie with their short axis vertical (Plate 62).

Concretions of ankerite and siderite occur in this facies. Siderite replacement may virtually destroy the fauna and the concretions are often hollow. Ankeritic concretions are septarian and have uncrushed goniatites. These often show geopetal infilling with a few specimens showing inclined internal sediment surfaces.

Interpretation

It would appear that the facies was deposited in a similar way to that suggested in B. The major difference is the smaller amount of carbonaceous material, which probably meant that the sediment-water interface was more clearly defined and shells could not sink down into the sediment as easily; hence the smaller percentage of vertical and inclined shells. The inclined geopetal infillings suggest some post-depositional movement of the goniatites, possibly associated with synerisis of the early concretion.

4.1.3 Facies 4; Unlaminated Sandstone

Description

This facies is similar to that in Assemblage B but may locally be slightly coarser, containing more pebbles. Beds are up to 8m thick.

In some exposures many, but not all, of the pebbles show an apparent alignment though this has not been measured due to the two dimensional nature of the outcrops. When compared to the regional palaeocurrent pattern, the long axes of the pebbles appear to be parallel to the flow and dip downstream at about 10°. In quarry exposures some beds, which at first appear to be massive, may show feint lamination which is emphasized in X-radiographs (Plate 41).

Large plant trunks, up to 0.2m in diameter and 1.25m long, are more common in this assemblage than in B. Both Lepidodendron and Calamites occur. No preferred orientation has been observed and the long axis may dip at a high angle to the horizontal. Pebbles, up to 30mm long are often more abundant in the sandstone associated with the trunks.

Interpretation

Deposition of F.4 appears to have been a slightly more organized process than in B. The attitude of the pebbles with their long axes parallel to and inclined towards the flow direction suggests deposition by avalanching (Johansson 1965, Sengupta 1966); the alignment resulting from shear stress. Johansson (1965) shows that the inclination of pebbles is usually 10 - 20° less than that of the foreset.

The association of pebbles with the plant trunks suggests that floating vegetation may have been involved in transporting the larger grains; a process often suggested to explain the occurrence of boulders in the Coal Measures (Whittle, 1942).

The genesis of this facies will be further discussed after an examination of its relationships to other sediments.

4.1.4 Facies 9, Micaceous Silty Sandstone

Description

This facies is much more abundant in this Assemblage than in B. It occurs in beds, up to 1m thick, draping erosion surfaces in F.20, large-scale cross-bedding. The sandstone may show some thin, asymetrical ripples, where the amount of plant material is small. As well as Pelecypodichnus, sand filled cylindrical burrows occur. These are oval in outline and are inclined at low angles to the lamination. Where several occur together they show an orientation parallel to the original sedimentary dip.

Interpretation

The facies is again thought to have been deposited, mainly from suspension, in sheltered areas. It is suggested that the cylindrical burrows were circular in outline and near vertical in orientation (i.e. at angle to bedding) before compaction of the sediment.

4.1.5 Facies 10, Ripple Laminated Sandstone

Description

The ripple forms in this facies are very similar to those in the previous assemblage but also include rare symmetrical ripples and silt flasers (Plate 66). The symmetrical ripples usually have lengths between 60 and 80mm and heights of 10mm but lengths up to 2m and heights up to 0.1m have been recorded. The smaller symmetrical ripples have pointed crests and foreset laminae dip in both directions with offshoots (Reineck and Singh, 1973). The larger ripples however have more rounded crests and do not show foreset laminae offshoots.

Pelecypodichnus is almost ubiquitous in this facies but has not been observed in association with symmetrical ripples. There are also occasionally, cylindrical burrows up to 5mm across. A few sandstones are so intensly bioturbated that the ripple lamination is almost totally destroyed. Some examples of this lithology show convolute lamination.

Beds of this facies differ from those in Assemblage B in that the beds are not as well defined having gradational upper and lower boundaries. This is partly due to gradation into F.18, wavy bedding.

Interpretation

The asymmetrical ripples are interpreted, as in Assemblage B, as belonging to the lower part of the lower flow regime (see Fig.2, Vol.2). The presence of flasers does however suggest that ripple migration was less continuous than in B (Reineck and Wunderlich, 1968). The shape and internal lamination of the smaller symmetrical ripples indicates that they were produced by wave action (Reineck and Singh, 1973). The larger type may, however, be due to reversals in the current direction.

The lack of sharp bases, with tool marks, to the sandstone beds suggests that there was no pre-depositional erosion. The convolute lamination is presumably due to dewatering.

4.1.6 Facies 13, Medium-scale Cross-bedded Sandstone

Description

This facies is abundant in Assemblage C. It is usually coarser than in Assemblage A, ranging up to very coarse and pebbly sand size. Mudflakes are rare but plant fragments may be large. Set thickness ranges between 0.1 and 3m and sets usually occur in cosets (Plate 58) up to 15m thick. The lower bounding surface of sets are usually erosive and may be either planar or trough-shaped. They are overlain by tabular and trough-shaped sets (Plate 59) respectively. Sets may have angular, tangential or concave based foresets and the base is often marked by a single layer of pebbles. Foresets, which may show a slight mica concentration, usually dip at between 25 and 30°. Tabular sets are generally thicker (mostly between 2 and 0.5m) than trough sets (mostly between 1 and 0.5m).

Convolute bedding occurs near the top of some cosets. This is in the form of irregular upwellings of the strata, between 0.5 and 1m across (Plates 60 and 61). In plan view these give rise to a very irregular, humpy topography with a relief of about 0.3m. The laminae are usually continuous but occasionally the centre of the upwelling has a massive neck which truncates the surrounding laminae.

Interpretation

The origin of trough shaped sets has been discussed in section 3.1.11. Planar cross-beds are probably due to migration of bars or straight crested dunes (Allen, 1963). Further interpretation of the environment is only possible after examining the association with other facies.

The convolute bedding appears to be the result of liquefaction with the laminae being distorted by water escaping from the loosely packed sediment. The massive necks suggest that sometimes fluidization took place and they may be the vertical vents of sand volcanoes. These resemble the second type of sand volcanoe described by Burne (1970) in the Bude Formation, although the depositional origin was completely different; the Bude Formation being a turbidite sequence.

4.1.7 Facies 16, Striped Silts and Sandstones

Description

The facies consists of laminae, which may be normally graded or homogenous, of different grain sizes, varying between fine silt and very fine sand. The laminae are horizontal and are between 1 and 26mm thick. They give the rock a striped appearance (Plate 63). Unlike the superficially similar F.8, gradationally laminated sandstones, the laminae are sharp and there is no apparent relationship between grain size and lamina thickness.

Interpretation

This facies is interpreted as the result of deposition from suspension. The sharp nature of the laminae suggests discontinuous deposition.

4.1.8 Facies 17, Wavy Bedded Sandstone and Mudrock

Description

Wavy bedding is usual here as defined by Reineck and Wunderlich (1968); mudrock and sandstone layers alternating and forming continuous layers (Plates 64 and 65). The sandstone beds range in thickness between 1 and 30mm. The beds vary in thickness laterally and in the thicker beds this can be seen to be due to current ripples. The facies grade into F.1, mudrock, and F.10, ripple laminated sandstone at the finer and coarser ends respectively.

Ripple directions, in cores up to 1.5m long, vary by up to 100°. The ripples frequently occur as form sets. They do not exceed 10mm in height and may have vertical form indices as low as 8. The base of the sandstones are sharp and, in thicker beds, they may show small load features. In the finer parts of the facies the sandstone beds are always separated by mudrock whilst, in the coarser parts, ripples may cut down into one another.

The mudrock is identical to that in F.1. In the finer parts of this facies (F.17) it forms discrete beds up to 5mm thick. In the fine parts the mudrock laminae are thinner and less continuous. Occasionally they form silty drapes over ripple lee sides.

Pelecypodichnus V-shaped escape traces occur in this facies.

Interpretation

Wavy bedding indicates that both sand and mud were available and that periods of traction alternated with periods of slower sedimentation from suspension. The finer parts of the facies presumably represent periods of poor sand supply whilst the coarser parts represent deposition with poor mud preservation potential or more rapid sediment supply.

The continuum between mud and ripple laminated sands differs from that described from the present day tidal environment. There lenticular and flaser bedding separate wavy bedding from mud and ripple laminated sand respectively (Reineck and Wunderlich, 1968). The absence of abundant mudrock flasers suggests that mud was evenly distributed instead of concentrating in ripple troughs. The absence of lenticular bedding suggests that the current and sand supply were related - the stronger currents introducing more sand. This facies therefore differs from present day tidal sediments (Reineck 1960, Reineck et al 1968, Bajard 1966) where bipolar current directions are common, flaser bedding occurs and the sand supply may be insufficient to form continuous ripples.

4.1.9 Facies 18; Parallel Laminated Sandstone

Description

The grain size of this facies varies between very fine and medium sand grade. Laminae are between 0.5 and 3mm thick and are defined by varying grain size. Bedding planes have abundant mica but do not show a lineation. Unfortunately the facies has not been observed in outcrop but is common in the Manshead boreholes (SD9919, see Fig.9, Vol.3). The lamination appears to be near horizontal (Plate 67) but occasional new horizontal erosion surfaces occur truncating the lower laminae whilst those above parallel the erosion surface. Laminae at opposite sides of an erosion surface may differ in inclination by up to 8°.

Pelecypodichnus has been found in this facies but is not common. Occasionally, convolute lamination occurs. This is similar to that observed in F.13, medium-scale cross-bedding, in that the distortion is due to local upwelling of the laminae with occasional truncation of lamination (Plate 68).

Interpretation

The origin of this facies is difficult to ascertain. The sediment is too fine to be formed in the lower flow regime plane bed field (Simons et al, 1965) (Fig.2, Vol.2). Wide variations in grain size between adjacent laminae is not characteristic of upper flow regime plane beds. The nature of the erosion surfaces and the attitude of the lamination suggests that some of the lamination may be low angled concave foresets. The convolute lamination is interpreted in the same way as in F.13. Plate 68 clearly shows post- convolution erosion.

4.1.10 Facies 19; Thin Sandstone Beds.

Description

The sandstones of this facies are medium to coarse grained. Beds are between 0.01 and 1m thick but are mostly between 0.02 and 0.15m thick. The bases of beds are sharp and do not show any erosional structures. The lower part of the sandstone may be virtually structureless although close examination usually shows poor cross-lamination. The majority of most beds consist of cross-lamination which is well defined by grain size variation. Sets have been observed up to 0.15m thick. The top surface is usually irregular if overlain directly by mudrock and may have a relief of up to 50mm. Unfortunately the surface plan is unknown. On the top of some beds, however, small ripples occur. These are preserved as form sets and have wavelengths in the order of 6mm and ripple indices of about 7.5.

Mudfilled endichnial burrows (Plate 69) are found in this facies and exichnial burrows have been found in underlying mudrock. On the base of a few sandstones are irregular traces in hyporelief, averaging 2mm in width and 15mm in length. These traces show no organized pattern and die out gradually at both ends (Plate 71).

Interpretation

The cross-lamination is thought to have been deposited by dunes and the undulating top surface may be a modified dune surface. The change from dunes to ripples upwards indicates a waning flow in the lower flow regime (Fig.2, Vol.2). The sharp bases suggest a sudden influx of the coarse sediment into an area of normally quiet deposition from suspension, as evidenced by the surrounding mudrock. The sharp base, massive lower parts and rippled tops are reminiscent of F.3, turbidites, but the depositing current differed in being less or even non-erosive, by never depositing under upper flow regime plane beds and in always being capable of producing dunes.

It is suggested that the irregular traces are subaqueous shrinkage cracks. These may indicate an environment with salinity variation as suggested by Burst (1965) or may be a result of the rapid flocculation of the underlying muds (White, 1961). Subaqueous shrinkage cracks are found in present day coastal lagoons and lakes.

4.1.11 Facies 20; Large-scale Cross-bedded Sandstone

Description

The facies consists of very coarse pebbly sandstone occurring in cross-bedded sets over 3m thick. The foresets dip in a downcurrent direction as indicated by smaller palaeocurrent indicators. Although there is a continuum between this facies and F.13, medium-scale cross-bedded sandstone, there is a paucity of sets between 2 and 5m thick. The largest set observed in the thesis area is in Derby Delph and the associated bankside (SE019161) and is at least 34m thick, the base is not exposed. Collinson (1968) in the Kinderscout Grit of Derbyshire reports set thicknesses up to 40m. Although the bases of sets are rarely seen it is thought that most sets do not exceed 25m in thickness.

Individual foreset beds are up to 0.5m thick but are mostly less than 0.1m. They may show slight grading or reverse grading. Between the coarse, usually unlaminated beds, the slightly finer sediment often has feint lamination parallel to the cross-beds. The maximum dip observed is 25° and cross beds are either planar or concave up. The planar and near planar cross-beds have sharp bases. Concave beds tend to occur in larger sets and have more gradational bases, merging downwards into horizontally bedded sandstones. A topset is only seen at Chew Hurdles (SE028015, Plate 42) where the top 1m is convex upwards and the highest beds are horizontally bedded.

On both sides of the valley at Colden Water to the west of Heptonstall (SD979278 and 985277) two large sets of cross-bedding can be followed for 300m in the direction of cross bed dip. Collinson (1968) records a figure of 2.4km for a set on the northern edge of the Kinderscout Plateau (SK097898) and it is probable that many sets extend for over 1km. The longest continuous exposure, normal to dip, observed is 200m in Hebden Dale (SD983288). It is difficult to ascertain the true width of the cross-beds because of the lack of exposures. One set, south of the thesis area, at the head of Fairbrook, Kinderscout (SK097891) is about 600m wide (John Collinson, personal communication, and own observation) but even this does not reflect the true original extent as it is eroded into on at least its northern side. The general impression gained in the field is that the sets are often over a kilometre wide.

The cross-beds may show internal cross-stratification; termed "intrasets" by Collinson (1968). Three types of intraset can be distinguished;

(a) Scoop-shaped intrasets. This is the type of intraset described by Collinson and is the most abundant of the three. Sets are up to 0.3m thick and sit in troughs whose axes may be orientated at any angle to the large-scale cross-bedding (Collinson, 1968, Fig.11).

(b) Form intrasets. In these intrasets bedforms showing internal cross-bedding are preserved. The maximum relief seen is 0.8m and bedforms have been observed with internal foresets inclined up the dip or along the strike of the major foresets.

(c) Down-dip intrasets. This type occurs as cosets (Plate 47). The base and top of the sets are generally parallel to the large-scale cross-beds but may be wedge shaped. The intrasets dip downcurrent by up to 10° more than the associated large-scale cross-bedding (see Fig.8). Occasionally internal erosion surfaces are present resembling reactivation surfaces (Collinson, 1970 b).

The large-scale cross-beds have major internal erosion surfaces (Plates 44, 45, 46 and 49). These surfaces have similar strikes to the cross-beds differing by a maximum of 10°. The erosion surface usually truncates the underlying beds more severely higher in the set (Plate 46) but may truncate the lower beds more severely if they are concave upwards (Plates 44 and 45). The overlying cross-beds are usually parallel to the erosion surfaces but may be steeper than it in the upper part of the cross beds or shallower in the lower part.

Spherical carbonate concretions are frequently abundant in this facies (Plate 44), especially in the lower part.

Interpretation

The facies is obviously a product of accretion on a slip face rather than lateral accretion because of the orientation of the foresets and their high dip. Large-scale cross-bedding, produced by a slip face growth, has been observed in several present day environments:-
i) Aeolian dunes. Many of the features, such as set thickness, large internal erosion surfaces and the large lateral extent, are typical of aeolian dunes (McKee, 1966). Gradzinski and Jerzykiewicz (1974) in a description of the Cretaceous of Mongolia interpret large-scale cross-bedding with similar relationships to other facies, including massive sandstones and fine grained deposits on foreset slopes, as aeolian dunes. The coarse and poorly sorted nature of the F.20 sandstones discounts this possibility.
ii) Sand Waves. Water lain dunes with lengths greater than 30m have been described from tidal channels (Reineck, 1963), tidal shelves (Houbolt, 1968) and large fluviatile channels (Coleman, 1969).
Tidal channel sand waves often show a bimodal foreset distribution. Both tidal shelf and fluviatile sand waves produce large-scale cross-bedding. Houbolt (1968) reports heights of 40m for marine sand waves. Jerzykiewicz (1968) described large-scale cross-beds, from the Intrasudetic Cretaceous Basin of the Polish/Czechoslovakia border, similar to the F.20 cross-beds, and interpreted then as nearshore sand waves.

Present day, described fluviatile sand waves are smaller than marine sand waves; reaching 17m in height in the Brahmaputra (Coleman, 1969) and only 6.7m in the Mississippi (Lane and Eden, 1940).
iii) Delta. The forward migration of a classical delta as described

by Gilbert (1883) would produce large-scale cross-bedding on a size determined by water depth. Such delta units can be produced by the flow of a sand laden current; fluviatile or tidal; into a body of relatively still water. Collinson (1968) preferred the last theory and, because of the absence of any tidal deposits, favoured a river delta. Other deposits interpreted as deltaic units include conglomerates from the Neogene of Crete (Gradstein and Gelder, 1971), which have foreset dips of 16^{o}, and the Athabasca Tar Sands (Carrigy 1966 and 1971) where the fine sandstones have foreset dip of 7^{o}.

It is necessary to look at the facies in its context before further discussion of its mode of deposition or of the formation of internal erosion surfaces and intrasets.

4.1.12 Facies 21 - Undulatory Bedded Sandstone

Description

This facies is observed as a series of undulations in sections normal to the regional palaeocurrent direction. The grain size varies between medium and very coarse sand with pebbles up to 20mm in diameter. The undulations have crest to crest lengths between 9 and 23m and heights up to 1m (see Fig.10, Vol.2). Beds are between 0.1 and 0.2m thick and are continuous but often thicken over the undulations (this is well seen on the left of Plate 56) but in some cases the beds thin over the undulations. Beds may show grading or inverse grading but are generally coarsest in the middle and grade into finer material upwards and downwards. The finer parts usually show a poor lamination parallel to bedding but in Derby Delph (SE019161) cross-lamination has also been observed (Plate 57). This cross lamination is rather poor and is only seen on the western side of the undulations where it indicates currents flowing towards the crest (see Fig.11, Vol.2).

The crests and troughs of the undulations are usually horizontal but may dip at a low angle in either an upstream or downstream direction as indicated by independant palaeocurrent indicators. The strike of the flanks differ by up to 90° and may converge or diverge downstream. This is well illustrated on the projected contours on the undulatory beds of Derby Delph (Fig.11,Vol.2).

Beds change height progressively from one undulation to another. In Derby Delph (SE019161) one bed (see top of Fig.11, Vol.2) which can be traced over six undulations, changes height by 7m. Crestlines of the undulations tend to migrate sideways, when traced upwards, in the direction of the higher undulations. The angle of migration may be as high as 50° from the vertical (see Fig.10 and Plates 55 and 56).

Internal erosion surfaces, similar to those of F.20, large-scale cross-bedding, occur. These surfaces generally cut down into the deeper side of the undulating surfaces. In Derby Delph (SE019161) (Fig.10 and Plates 55 and 56) erosion surfaces can be seen to truncate the underlying beds more sharply upwards.

Interpretation

The coarse grain size suggests powerful currents and the thick, generally massive, nature of the beds suggests that they were probably deposited rapidly. Extrapolation of the strike of the undulations suggests that the original bedforms were long, low ridges parallel to the flow. Large-scale current lineations, or "sand ribbons", have been described from the Wadden Sea (van Straaten, 1953), the Bahaman carbonate sands (Imbrie and Buchanan, 1965), the shelf around the southern British Isles (Stride, 1963), ephemeral streams (Karcz, 1967 and Picard and High, 1973) and from the Brahmaputra (Coleman 1969).

The undulations of this facies are on a larger scale than those observed in ephemeral streams and have a larger height to width ratio than the marine ridges. The Brahmaputran ridges, in contrast, are of very similar size in cross-sections ranging from 9 to 30.5m from crest to crest and with heights up to 1.2m. The ridges differ in grain size; the Brahmaputran ridges being composed of fine sand. On excavation Coleman (1969) only found a poor horizontal lamination in trenches cut parallel to the ridges. Whether the laminae are parallel to the bedform is not known.

The origin of sand ribbons is usually attributed to a transverse instability in the flow. Irregularities, such as chance deposition of more sediment in certain areas, leads to large scale downstream "corkscrew" vortices. Such a mechanism was suggested by Houbolt (1968), as the origin of tidal current ridges, and by Allen (1964) on a smaller scale, for the origin of parting lineation. Coleman (1969) does not suggest a mechanism to explain the Brahmaputran ridges. He does however show that under peak flood conditions dunes can be washed out into a plane bed in longitudinal profile whilst in transverse section the profile undulates over 4m. The water surface turbulence pattern shows small turbulent cells parallel to the current, as would be expected with corkscrew vortices. Such strong currents may account for the Brahmaputran ridges, with their poor lamination, and for the ridges of F.22.

In conclusion, it is suggested that the facies represents deposition by a strong current with corkscrew vortices. Further interpretation will be possible after discussion of their relationship with other facies.

4.1.13 Facies 22; Seatearths and Coals

Description

Plant rootlets have been found in sediment varying in size from clay to coarse sand. The fireclays are grey to buff in colour and contain slickensides. The gannisters are white, having a lower feldspar content than the associated sandstones. At Red Scar, Stoodley Pike (SD972242) a concretionary layer of siderite occurs at the top of a gannister, underlying a coal (Fig.12, Vol.2). Within this layer are concretions of sphaerosiderite with individual spheruliths between 1 and 3mm in diameter.

Seatearths are up to 1.8m thick. The grain size appears to be entirely independent of the seatearth profile and the fining upward sequences noted by Collinson (1969) have not been seen. Occasionally remnant structures, such as ripples or cross-bedding can be discerned. The seatearths are superimposed on laterally varying host sediments. The overlying coals are up to 0.2m thick and are usually bituminous but may be silty.

Interpretation

The rootlets indicate that the environment was near to sea or water level (Hemingway, 1968). The fireclays and gannisters appear to have undergone considerable subaerial leaching in acidic conditions. Siderite concretions are common in Carboniferous seatearths (Hemingway, 1968) and are thought to be due to the redeposition of leached iron. The concentration of iron in the upper part of the seatearth, in contrast to the more usual position lower in the profile, suggests upward migration of pore waters, indicating considerable evapotranspiration.

4.2 Associations

Three Associations are distinguished in Assemblage C.

4.2.1 Association C1 - Deep Channel Deposits

Description

This association includes F.4, unlaminated sandstone; F.9, micaceous silty sandstones; F.13, medium-scale cross-bedded sandstone; F.20, large-scale cross-bedded sandstone and F.21, undulatory bedded sandstone. All facies appear to sit within large channels with the exception of medium-scale cross-bedding. Unfortunately the channel bases are rarely exposed and, in fact, Collinson (1968) did not recognize these channels in Derbyshire, although one has now been recognized at the head of Fairbrook, Kinderscout (SK097891).

The size of the channels is difficult to ascertain because of the lack of exposure and the erosion of channels into one another. Their maximum depth is at least equivalent to the thickness of the large-scale cross-bedding, which is the major channel filling facies. Channel sides may locally be vertical but slopes of around 10^{o} are more normal and the largest channels may have average dips less than this.

As discussed earlier, it has been impossible to measure the width of the widest sets of large-scale cross-bedding and therefore the width of the largest channel is unknown. It is thought that many large channels are in the order of 0.5 to 1.5km wide. In Buckton Quarry (SD991016) there is a, 6m deep, channel less than 85m wide which is completely infilled with cross-bedding (Plates 53 and 54). The cross-beds have tangential bases, on the upstream side of the channel, angular bases at the base of the channel, and the sets become more concave upwards up the downstream side of the channel (see Fig.11, Vol.3). This is the largest of several channels, seen to be infilled in this manner, all of which occur near the top of larger sets of cross-bedding. The smallest of these channels is only 1m deep and by definition is infilled with medium-scale cross-beds, though its genesis is obviously the same as the large-scale cross-beds of the deeper channel. The precise relationship between the orientation of the channel and the large cross-beds is unknown as exposures are only in two dimensions.

Unlaminated sandstones occur below large-scale cross-bedding at the base of many of the channels. The channel sides over 20° steep are usually overlain by unlaminated sandstones. The transition from unlaminated beds to cross-bedding takes a variety of forms. If the foresets have a concave upward geometry the near horizontal toesets may pass down into the unlaminated beds with no sharp break (Plate 43). If the foresets are planar there is usually a sharp, near horizontal contact between the two facies (Plate 49 and left side of 48). Some foreset beds may however pass downdip into unlaminated beds or expand into unlaminated beds at the bottom of the set resulting in later foresets being less thick (Plate 45 and right side of 48). Fig.13, Vol.2 shows the only channel base where both facies have been seen to be in contact with the base. In Buckton Quarry (SD991016) a 6m thick, apparently unlaminated bed appears to grade laterally into the large-scale cross-bedding (see Fig.11, Vol.3 and Plate 53).

Mudclasts, although rare in the unlaminated sandstone, may be locally abundant and can be up to 1m in diameter but show no organized orientation. Rarely, a single set of F.13, medium-scale cross-beds occur within the unlaminated beds (Plate 40). Both top and bottom surfaces are sharp and the cross-beds conform with the regional palaeocurrent pattern.

Undulatory bedding blankets a channel erosion surface which is seen to downcut 6m in Buckton Quarry (SD991016). In Derby Delph (see Fig.10, Vol.2 and Plate 55) undulatory bedding is separated from large-scale cross-beds by an inclined erosion surface. It is thought that again the undulatory bedding occurs on the sides of large channels. Assuming that the ridges, inferred from undulatory bedding, were parallel to the channel sides, the large-scale cross-bedding dips approximately at 40° to the channel sides (Fig.11, Vol.2).

F.9, micaceous silty sandstones are found on some of the internal erosion surfaces of large-scale cross-beds. Whilst only seen in the upper part of sets because of the limits of exposure they have never been seen to die out down the foreset. The internal lamination of F.9 is parallel to the foreset bedding.

Cosets of F.13, medium-scale cross-bedding, up to 18m thick, lie on top of the large-scale cross-bedding with a horizontal erosion surface (Plate 44). Occasionally a set of medium cross-beds passes, in a downcurrent direction, into large-scale foresets overlying an internal erosion surface (Plate 46).

Interpretation

The association was obviously laid down in large channels, which in the absence of evidence of tides, were presumably fluviatile. Several theories as to the origin of the large-scale cross-beds can now be proposed:-

(a) Channel Confluence Delta. Where a tributary or anastomised branch, with a high bed load, enters a large relatively slow moving channel it may build out a delta. In the present day Amazon such confluence deltas are large enough to have distributary channels with levees (McIntyre, 1972, p.452). Within a braided river the anastomising channels lead to many such confluences where deltas could build up (Fig.16 (a), Vol.2). Collinson (1970) and Smith (1971) report small deltas built up in this way at the mouths of channels dissecting linguoid bars. The depth of such delta foresets depends on the depth of the host channel. This mechanism may account for some of the large-scale cross-beds but it would be difficult to explain the large areal extent of most sets by this process.

(b) Channel Infill Delta :- An abandoned river channel could be infilled by a delta. The infill of the Buckton Quarry (SD991016) channel and smaller channels shows a very similar geometry to that produced by infilling a trough normal to the flow, as shown by Jopling (1965) in flume experiments. It is thought that such channels have been infilled at an angle to the current flow (Fig.16b, Vol.2); the cutting and filling of the channels possibly taking place during one flood cycle. It is less likely that the largest sets could be channel infill deltas as there is no fine grained sediment beneath that could be interpreted as normal abandoned channel infill.

(c) Fluviatile Sand Waves:- (Fig.16c, Vol.2). No present day fluviatile sand waves have been recorded which are as high as the thickness of some of the large-scale cross-beds. The long distance some foresets can be followed in a downcurrent direction implies that the sand waves had large vertical form indices; a feature of the Brahmaputran sand waves which have lengths over 900m (Coleman, 1969). The theory requires that the channel, containing the sand waves, was abandoned to preserve the bedform as form sets.

(d) Transverse Bars :- The term "transverse bar" has been used for a variety of bedforms but is used here, as by Allen (1968), for large-scale bed forms attached to opposite banks (Fig.16d, Vol.2). The crestlines extend for almost the full width of the channels. The best documented bars are from the Rio Grande (Harms and Fahnestock, 1965) where downstream of the crestline is an avalanche face producing tabular cross-beds. The slight divergence in the large-scale foreset dip direction and the channel axes is typical of what one would expect in transverse bars. The close proximity of at least some large-scale cross-beds to steep channel sides also suggests bar-attached rather than mid-channel bedforms. Preservation of large sets is more likely for transverse bars than sand waves as preservation can take place by channel migration as well as channel abandonment. There is, however, a large difference in scale between F.20 and any transverse bar described in the literature; those described by Harms and Fahnestock (1965) reaching only 0.75m in height. Coleman (Personal Communication), in contrast, reports transverse bars on the sides of the Mississippi delta distributaries. These are spaced about 1.6km apart and the relief between the top of the bar and the thalweg is often in the region of 25 to 30m. Assuming that they migrate downstream they presumably produce large-scale cross-beds of the same order of size as in F.20. Most large-scale cross-beds are therefore thought to be the product of transverse bars.

Allen (1968) suggested that the skewed nature of the transverse bars relative to flow leads to leeward eddies having a spiral motion. To test this theory and examine the consequences of such currents, transverse bars were produced in the Keele, 14m long, 0.8m wide, recirculating flume. In one experiment the sand bed was moulded into a series of bars, approximately 1.5m from crest to crest. In a second run transverse bars were produced by chance from a flat bed on the transition from upper to lower flow regime. In both cases flow depth was shallow, averaging 0.05m, and ripples covered the back of the bars and the thalweg. Three important observations were made in both runs.

Firstly, the separation point was not at the bar crest, as is so with simple deltas or bedforms with crests normal to the flow, but was sited in the upper half of the bar front slope. The siting of the separation point caused the sand to be churned up, by frequent eddies, on the top of the slope. The sand then swept down the slope as a fluidized sediment flow, coming to rest at much lower angles than would be the case in avalanching, and forming concave upward foresets.

The second feature was that the majority of the foreset slope was covered by small ripples and that these migrated towards the bank to which the bar was attached. Those ripples above the separation line had a downward component whilst those below had a small upward component. The ripples were often destroyed either by changes in the position of the separation line or by larger fluidized sediment flows.

The third observation was that spurs parallel to the flow developed in front of the transverse bars. These spurs appear to be similar to those which form in front of swept catenary ripples (Cornish, 1914, Plate 59 and Allen, 1968, Fig.4, 28-31). The flow pattern observed, from small ripples on the spurs, was a series of corkscrew vortices with separation along the spur crests and attachment along the intervening troughs. Allen (1968) suggested that such flow patterns were the result of instability of the separation eddy caused by the skewed nature of the bedforms to which the spurs are attached.

The observation that transverse bars have concave upward foresets is important as it allows an easier explanation of foreset geometry than does any other theory. Concave upward foresets produced by a two dimensional flow necessitate extremely high amounts of very coarse sediment in suspension and a high ratio between depth of water over the crest of the bedform and the depth of water in the thalweg (Jopling, 1965).

The origin of the three types of intraset can now be explained. Scoop-shaped intrasets were probably produced by local shortlived eddies scouring the leeside surface. These then infilled in the direction of the area's dominant leeside current which could be up, down or along the foreset slope. The form intrasets probably represent dunes built by the dominant leeside eddy. The down-dip intrasets resemble the downcurrent dipping cross-stratified sets described by Banks (1973) in the late Precambrian of Finnmark. It was suggested that such cross-stratification could be formed by migration of dunes down the leeside of a larger bedform. The model, as suggested by Allen (1968, Fig.5.16) assumes an even size and migration speed for the dunes in varying depths of water. The down-dip intrasets of the association can be better explained on the transverse bar model by alternately building up the avalanche slope and reducing the angle of slope by leeside erosion. This would therefore be a multiple reactivation (Collinson 1970b) process.

Internal erosion surfaces can be explained by three mechanisms:-
i) Collinson (1970b)suggested that internal erosion surfaces, in cross-stratification of linguoid bars in the Tana River, were due to low stage modification of the bar front. Both wave action and currents, flowing round the abandoned bar, rounded off the top of the avalanche face which was reactivated at the next flood (Fig.17a, Vol.2). The presence of F.9, micaceous, silty sandstone on some reactivation surfaces suggests that at least they were formed by low stage modification of a large bed form. The lee of an abandoned bed form could also be a site of accumulation for drifting plant material.

ii) In experiments on two dimensional deltas in the Keele flume it has been shown that reactivation surfaces can be formed when ripples arrive at the delta crest. Time lapse photography was used to record the process. Drawings of nine frames of one run are shown in Fig.14, Vol.2. The reattachment flow, of a ripple approaching the delta crest, rounds off the crest. When the ripple reaches the crest the slope resumes its original profile. By superimposing earlier frames on the last frame the likely pattern of final internal lamination produced by the run can be drawn (see Fig.15, Vol.2). Smaller bedforms on the back of the bedforms or in channel feeding deltas could therefore form reactivation surfaces in the upper part of the large-scale cross-bed set (Fig.17 b, Vol.2).
iii) A change in flow pattern on the lee of a bedform, possibly associated with changes in the skew of the crest, can change the foreset geometry. The process suggested for the production of down-dip intrasets could take place on a large scale with the development or destruction of a leeside spiral eddy. Internal erosion low in the set is presumably due to changes in the position of the thalweg (Fig. 17c, Vol.2).

The interpretation of the genesis of the massive beds is as problematical as in the previous assemblages. The massive beds in a few, steep-sided channels may be explained by antidunes, freezing of a traction carpet, shearing of loosely compacted sediment or deposition in a metastable field; processes already discussed in section 3.3.3. These processes, however, would probably not occur at the bottom of a river with large bedforms.

The passage of massive beds into large-scale cross-beds in Buckton Quarry (SD991016) suggests that the two are related. Jopling (1964) showed, in flume experiments, that foreset lamination is poor where upper flow regime occurs on the delta top or where the ratio of water depth to delta height is high. In these two situations the sorting processes are at a minimum. In flume experiments at Keele apparent "massive beds" were produced under upper flow regime on the delta but resin peels of this bed showed lamination in relief. It is possible that under high discharges some bedforms could be virtually structureless. This theory does not, however, account for the majority of the massive beds being at the base of channels.

It is suggested that most of the massive beds represent the trough deposits of the large bedforms. The reattachment point, of a flow over a bedform, is an area of erosion with sediment being transported both away and towards the slip face. Given a two dimensional situation this will lead to a pattern of currents and counter current ripples within the troughs of the bedforms (Jopling, 1963). In observed flume experiments, the point of reattachment fluctuates, frequently destroying minor ripples. This presumably occurs in natural rivers. Coleman (1969) observed intense turbulance downstream of sand waves in the Brahmaputra. The possibility of organized lamination forming in this situation must therefore be poor, especially in very coarse pebbly sandstone. Regressive ripples have been noted (Collinson, 1968, Fig.6) and the isolated sets of medium-scale cross-bedding indicate movement away from the foresets. These instances suggest that occasionally conditions were stable enough to allow bedforms to adjust to the flow. The passage of foresets down into massive beds (as in Plate 48) is presumably a readjustment to a change in the water flow pattern.

Slumping may also be a minor factor in the origin of massive beds. Bank collapse could give rise to a massive bed at the base of a channel. This is probably the case in those beds with abundant mud clasts.

F.21, undulatory bedding presumably formed on the river channel sides in a similar situation to the present day Brahmaputra linear ridges (Coleman, 1969). Postulating a transverse bar model the ridges may have occurred as spurs in front of the bars as in the flume experiments. Harms <u>et al</u> (1963) in the Red River and Harms and Fahnestock (1965) in the Rio Grande show scoop shaped hollows downstream of apparent transverse bars. The situation envisaged in the C1 sediments is shown in Fig.18, Vol.2.

The medium-scale cross-bedding is thought to represent smaller bedforms in the shallow parts of the river, dunes, bars and scour infills. The presence of medium scale sets passing downstream into large-scale sets shows the existence of smaller bedforms on the back of the larger bedforms. The wide lateral extent of the cross-bed coset, irrespective of underlying lithology, indicates that shallow water flows were more widespread than the deeply channelised flows. The lack of any fining upward sequence and the low palaeocurrent variance of the facies (see section 8.1) suggests that it was a braided river deposit. The continuous nature of the coset laterally is probably due to the migration of the braided river forming a sheet deposit. Convolute lamination at the top of channel sands is common at the present day (McKee <u>et al</u> 1967 and Coleman 1969).

In conclusion, the association appears to have been the deposit of a very large river. The scarcity of fine sediment suggests that most of it was transported through the system in suspension. The balance of evidence suggests a braided river, of fluctuating discharge, with one or more deep channels in which transverse bars were the major bedform. The shallower parts of the river appear to have migrated laterally whilst the major channel or channels changed position by avulsion - a process which is taking place in the Brahmaputra (Coleman, 1969).

4.2.2 Association C2; Minor Channel Deposits

Description

This association includes F.1, mudrock; F.13, medium-scale cross-beds, and F.19, thin sandstones. F.13 is generally less extensive than in C1 but some coset beds can be followed over several kilometres. The maximum coset thickness is 20m. As in the previous association there are no lag conglomerates at the base of the cosets, no fining upward sequences and no apparent upward change in set size. The bases of coset beds are sharp and erosive. In the Windy Hill Motorway Cutting (SE981148) a bed can be seen to be erosive by over 3m and has groove marks at the base. Cosets have been seen to thin out into F.19 which may die out laterally. F.19, thin sandstone, are rarely seen far from F.13. Both usually overlay and are overlain by F.1, mudrock.

Interpretation

The low palaeocurrent variance of the medium-scale cross-bed cosets and the lack of fining upward sequences suggest a braided river deposit. The undulatory bases and the wedge shaped nature of some cosets also suggest a braided rather than a meandering stream. At least some of these channels are thought to have migrated through time to produce sheet like sandstone bodies. F.19, thin sandstone beds, are interpreted as crevasse splay (Allen, 1965) deposits. Their coarse nature suggests that they were deposited near the river channels, presumably in levee flank depressions (Coleman and Gagliano, 1965, p.146). Although not usually associated with braided channels levees and crevasse splay deposits have been reported from braided stretches of the Brahmaputra (Coleman, 1969). In contrast to the better known Mississippi-type levee, built up from suspended material, bed load deposits form the majority of the Brahmaputra levees.

4.2.3 Association C3 - Interchannel Deposits

Description

F.1, mudrock; F.10, ripple laminated sandstones; F.16, striped silts and sandstone; F.17, wavy bedded sandstone and mudrock; and F.18, parallel laminated sandstone are included in this association. They may be found in any order but often form coarsening upward sequences with F.1, 16, 17 and 10 or 18 in an upward sequence, with complete gradation between facies. The upper boundary of the sequence is generally more sharply defined than the lower boundary but may pass into a seatearth and coal. The maximum observed thickness is 4m but many sequences are less than 1m thick.

Fining upward sequences also occur with the reverse sequence of facies. Although less common than the previous type they may be up to 4.2m thick. The borehole logs from Manshead (SD9919) (see Fig.9, Vol.3) shows many such coarsening and fining upward sequences.

Interpretation

F.16 and F.17 indicate fluctuating flow conditions, presumably over a considerable time span. F.10 and F.18 indicate more persistent currents whilst F.1 suggests quiet water conditions with deposition from suspension. The presence of seatearths and coals, the occurrence of wave ripples and the fluviatile sediments, which downcut into the Association, suggest that the depth of water was not great. It is thought that the association was deposited in shallow bodies of standing water between the fluviatile channels. The coarsening upward sequences may be deltaic successions due to infilling of the bay by overbank deposition. Fining upward sequences can be explained as relatively distal crevasse splay successions; the supply of sediment gradually decreasing as the levee is rebuilt. Fisher (1969) records both coarsening upward minor delta sequences and fining upward overbank deposit sequences in Holocene Gulf Coast delta systems. The convolute bedding of this association is also common in present day overbank deposits (McKee, 1966 and McKee *et al*, 1967.

4.3 Relationships Within Assemblage

Association C1, deep channels, occurs only at the base of the assemblage. The channels cut down into Assemblage B and often it is difficult to draw an exact boundary between the two assemblages if B3, channeled coarse sandstones, occurs at the top of Assemblage B.

There is usually only one set of F.20, large-scale cross-bedding in a vertical section but occasionally two sets occur, as at Buckton Quarry, Mossley (SD991016) and on the west side of Colden Clough, Heptonstall (SD979277). In the Ladcastle and Den quarries, Uppermill, Saddleworth (SD994060), three sets can be distinguished (see section 7.1.3). Overlying the coset of F.13, medium-scale cross-beds, association C2 and C3 occur in approximately equal proportions. Apart from a fining upward sequence of C3 overlying some C2 beds there is apparently little organised relationship between the two associations.

Seatearths and coals, F.22, are associated with C2 and C3. In the Manshead boreholes, numbers 2 and 5, four seatearths occur within a vertical sequence (Fig.23). F.2, goniatite faunal band is associated with C3, interchannel deposits. The R.gracile bed in the Manshead cores overlies a seatearth and in the Windy Hill Motorway cutting (SE981148) it is not far above a coal. The relationships of the Butterly band to facies within the thesis area is unknown but Collinson (1969, Fig.21) places it within C3, not far above a seatearth.

The assemblage, as a whole, is interpreted as a delta top sequence. The areas of shallow water are represented by association C3. Coleman and Gagliano (1965) describe wavy bedded sands and mud, ripple laminated sands and parallel laminated fine sands, similar to F.17, 10 and 16 respectively, as common in interdistributary bays and subaerial levees of the Mississippi. F.18, parallel laminated sandstone, is very similar to "medium to fine grained, horizontally-bedded sand" described by Oomkens (1974) from the Niger. The shallow water areas appear to have had large variations in current velocity and were wave agitated. The coarsening and fining upward sequences are typical of interdistributaries (Fisher, 1969).

The restriction of C1, deep channels, to the bottom of the assemblage suggests that they are distributory channel deposits. High-constructive deltas, such as the Mississippi (Fisher, 1969; Morgan 1970) and the Texas Colorado (Kanes, 1970) have straight distributaries bounded by levees. Transverse bars are a feature of straight channels (Leopold and Wolman, 1957) and evidently occur in such distributaries. In most deltas, such as the Mississippi, new distributaries are created by avulsion or by the development of two channels from the bifurcation at mouth bars. Abandonment of old channels takes place because of the loss of river flow, due to avulsion upstream, or by the growth of a bar across a channel's upstream junction with a dominant channel (Welder 1959). In the Brahmaputra - Ganges delta, about which unfortunately little is known, the distributaries, like the river channel upstream (Coleman, 1969), could well change by avulsion of the major channels and lateral migration of the shallower parts. Although Chowdhury (1966) maintains that the Ganges has not significantly changed its position in historical time, examination of a Recent sediment map (Morgan 1970, Fig.6) strongly suggests that the distributaries have migrated eastwards since the Pleistocene.

The absence of C1 channels high in the sequence, the thinner nature of the C3 cosets and the absence of evidence of large bedforms indicates that the channels upstream of the major distributaries were shallower; two explanations of this phenomenon are suggested. Firstly, as the river entered the basin the restriction of the channel width imposed by the levees may have led to deep scour during high stage. It is suggested that the distributary levees were better developed than the flood plain levees. This situation is the opposite to that found in the present day Mississippi (Coleman and Gagliano, 1965). It is, however, one that might be expected when a river with a high bed load is checked by a large body of water. In the flood plain of such a river the flow would probably form crevasse splays over a wider area. Secondly, the greater compaction of the sediment underlying the delta front may have caused an increase of the slope resulting in deeper scouring within the distributaries. In conclusion, the C3 association is thought to be the deposit of the river within the floodplain, much of which was covered with shallow water.

The several seatearths and coals in a vertical succession indicate that subsidence and transgression took place, forming interdistributary areas.

4.4 Conclusion to Facies Analysis

The whole sedimentological succession is interpreted as an essentially regressive deltaic sequence. In terms of delta terminology (Scruton, 1960); Assemblage A represents the toeset beds, with progressively more proximal turbidites upwards; the coarsening upwards B represents the delta foreset or slope; whilst C represents the topset beds. In Chapter 8 the evolution of the delta and the importance of various factors affecting the deltaic process will be discussed.

CHAPTER 5 PETROGRAPHY

It was not an aim of the research to study the petrography in detail but thin sections have been studied for clues to the depositional history. Walker (1966a) and Collinson (1969) studied essentially the same rocks. Examination of thin sections from the thesis area showed no significant variation from Collinson and Walker's point count analyses of the non-carbonate cemented sandstones. These sandstones have essentially similar petrography in all facies; quartz 50-70%, feldspar (plagioclase and potash feldspar) 5-20%, mica (biotite and muscovite) 0.1-8.2%, carbonaceous material 0.2-6.6%, and matrix 13-35%. Generally the mica, carbonaceous material and matrix content is less in the coarser rock types. The quartz / feldspar ratio is close to 4 to 1 throughout the sandstone.

Classification of the sandstone is difficult as it is often impossible to distinguish between metamorphic quartzite fragments and other quartz grains. Walker (1966a), on an estimate of the feldspar quartzite ratio, tentatively suggested that the sandstones be classified as lithic greywackes (as defined by Pettijohn, 1957). This classification can be criticised in three ways:

a) If the aim of the classification is to discuss the textural maturity, quartzite fragments should be counted as quartz grains as they weather identically (Pettijohn <u>et al</u>, 1973). Other rock fragments are rare in the sandstone.

b) In the finer sands the quartzite fragments are worn down until they are inevitably single quartz grains. The term "lithic greywacke" cannot therefore be applied to the fine sandstones yet they are basically no more mature than the coarser sandstones.

c) The percentage of quartzite fragments never greatly exceeds the feldspar percentage and as part of the matrix is due to post depositional decomposition of the feldspars, the feldspar may have been more abundant at the time of deposition.

It is therefore suggested that, using the classification of Dott (1964) the sandstones are feldspathic greywackes. Whilst not having the necessary 25% feldspar to be called "arkosic wackes" (see Pettijohn <u>et al</u>, 1973) the feldspar/quartz ratio makes them the wacke equivalent of a sub-arkose. The ratio of feldspar to quartz remains relatively constant throughout the grain sizes and the percentage of rock fragments decreases with decreasing grain size.

Most sandstones have obviously been subject to considerable diagenetic alteration. Quartz grains show pressure solution features and occasionally optically continuous overgrowths can be discerned. Feldspars show sericitisation and their boundaries are often indistinct. Biotite often shows limonite or haematite haloes, similar to those described by Walker (1967) from the Pliocene of the Sonoran desert. Sometimes the iron oxide penetrates the mica's cleavage planes. A highly altered large olivine crystal has also been observed, the irregular patches of unaltered crystal being separated by clay minerals. This evidence suggests that much of the matrix in the coarse sandstone is of diagenetic origin. It is, of course, impossible to seperate diagenetic matrix from original mud, especially in the finer rocks.

The carbonate cemented rocks differ in many ways from the majority of the sediments. Although volumetrically small, their different diagenetic history makes them important. The concretions in F.1 (mudrock), F.2 (goniatite faunal bed) and F.8 (gradationally laminated sandstone) of Assemblage A (deep water) and B (slope) have calcite as the basic carbonate. F.7 (Shell bed), in contrast is cemented by dolomite and the shelly material has been dolomitized.

The large spherical concretions in the coarse sandstones at the base of Assemblage C (delta top) rapidly deteriorate on exposure turning brown (hence their polite name; "red horses") on the development of goethite. The carbonate here is probably a variety of ankerite. The concretions in the R. gracile bed (F.2) consist of either ankerite or siderite. In the former case, the goniatite shells are preserved bur altered to ankerite. In the siderite concretions the goniatites are virtually destroyed leaving only "ghosts".

The iron concentration high in the sequence suggests that the rivers were the source of the iron and that it precipitated early, on entering the basin. Hemingway (1968) discussed in detail the relationship between iron carbonates and the delta top environment.

Early carbonate cementation prevents pressure solution of the quartz and appears to have, at least partially, prevented pore fluids reaching the less stable mineral as both biotite and feldspar are relatively fresh. Brenchley (1969) also noted fresh feldspars and biotites in carbonate concretions within Ordovician greywackes.

The less intense diagenetic alteration of the sandstone allows a better insight to the original grain shape. As can be seen in hand specimen of very coarse pebbly sandstones (Fig. 19, Vol. 2) and in thin section in finer sands, both the feldspar and quartz grains are sub-angular in outline. It is possible that the carbonate has, in part, replaced the silicates, as observed in some Carboniferous concretions (Hemingway, 1968). Holdsworth (1964) showed that many Namurian carbonate concretions, in the equivalent to F.2, contain abundant radiolaria, whilst in associated shales evidence is lacking of their presumed original presence. No radiolaria have been observed in the thesis area and the absence may be due to solution. The delicate nature of radiolaria makes them very susceptible to silica solution. There is, however, no evidence of significant etching of grains, decomposition along lines of weakness, or zones of quartz overgrowths surrounding the concretions. The essential grain shape is therefore probably original.

In conclusion, the sediments are relatively immature but the feldspar / quartz ratio is not as high as in arkoses, which generally have a limited geographical range. The average feldspar content of 15.6% falls within the range typical of present-day rivers (Pettijohn <u>et al</u>, 1973, Table 2.1) but is higher than most beach or shelf sediments. This is compatible with the small amount of reworking of delta sediments envisaged in the above facies analysis.

CHAPTER 6 STRATIGRAPHY

6.1 Lithostratigraphy

The establishment of useful rock stratigraphical units in the R_1 succession is an interesting problem, due to the great variety of sediments and their marked lateral variation. Previous classifications have been based on the contrast between the very coarse sandstones, or "grits", and the finer sediment. A more useful classification for mapping and lithostratigraphical purposes was found to be based on the facies, defined in the previous chapters.

The study area has been split into three regions: the Saddleworth and Longdendale region, Area 1; the Todmorden and Hebden Bridge region, Area 2; and the Airedale and Wharfedale region, Area 3 (see Fig. 20, Vol. 2; detailed location and geological maps of all three areas are given in Vol. 3, Figures 1 to 6). The geographical limits of the new formations are based on these three areas and the classification used is given in Table 6.1. The associations and assemblages are listed in Table 6.2 for easy reference.

6.1.1 The Kinderscout Grit Group

The Kinderscout Grit Group is here defined as consisting of the eight formations in Table 6.1. The Group may be extended at a later date to include sediments of the same delta system, particularly those between and including the Parsonage Sandstone and Kinderscout Grit in the Blackburn area. The term "Kinderscout Grit Group" is not used as a biostratigraphical term, as has been done by previous authors (eg. Wright _et al_, 1927 and Earp _et al_, 1961), but as a

Table 6.1

The Kinderscout Grit Group

Area 1 Saddleworth / Longdendale	Area 2 Todmorden / Hebden Bridge	Area 3 Airedale / Wharfedale
Kinderscout Grit Formation		
Grindslow Shale Formation	Hebden Bridge Shale and Sandstone Formation	Silsden Shale and Sandstone Formation
Shale Grit Formation Mam Tor Sandstone Form.	Todmorden Sandstone Form.	Otley Sandstone Formation

Table 6.2

Assemblage A: Deep Water Sediments

A1; Mudrock

A2; Parallel sided sandstones

A3; Wedge shaped sandstones

Assemblage B: Delta Slope Sediments

B1; Fine grained deposits

B2; Inclined units

B3; Channeled coarse sandstone

B4; Parallel bedded sandstone and mudrock

Assemblage C: Delta Top Sediments

C1; Deep channel deposits

C2; Minor channel deposits

C3; Interchannel deposits

term to cover formations which are related by being deposited by the same delta system infilling the Central Pennine Basin during the Kinderscoutian.

6.1.2 Formations of the Kinderscout Grit Group

6.1.2.1 Mam Tor Sandstone Formation

Synonyms: Mam Tor Sandstones (Jackson, 1927)
Mam Tor Beds (Gaunt, 1960)

Thickness: 0 - 140m; see Stevenson and Gaunt (1972) for details.

Type section: Mam Tor (SK127835)

This formation, which only crops out south of the thesis area, consists of thin parallel sided turbidite sandstones interbedded with shale (Allen, 1960). The boundaries are not well delineated (Stevenson and Gaunt, 1972) but are generally marked by the percentage of arenaceous material.

6.1.2.2 Shale Grit Formation

Synonyms: Farey's Grit (Challinor, 1924)
Shale Grit (Jackson, 1927. The Shale Grit of Farey (1811) and Hull and Green (1864) probably covered the Mam Tor, Shale Grit and Grindslow Shale Formations.

Thickness: 0 - 210m; see Stevenson and Gaunt (1972) for details.

Type section: No type section has yet been proposed although Walker (1966a) proposed Alport Castles (SK143914) as the type area for his "Alport Group" which includes the Shale Grit and Grindslow Shale Formations.

The base of this formation is not seen within the thesis area. Associations A1, mudrock; A2, parallel sided sandstones, and A3, wedge shaped sandstones occur. The F.3, turbidite sandstones are generally less parallel sided than those of the Mam Tor Formation. Sandstone forms a higher percentage of the formation than mudrock and channels are common. The lower limit of the formation is defined by the incoming of channels and the thick massive sandstones of A3.

6.1.2.3 Grindslow Shale Formation

Includes: Grindslow Shale (Jackson, 1927)
Part of the Kinderscout Grit (Jackson, 1927)

Thickness: Approximately 100m in the thesis area but reaches a maximum thickness of 120m in north Derbyshire (figure based Collinson, 1968).

Type section: Grindsbrook, Edale (SK115873)

In the Grindslow Formation Assemblage B, delta top sediments, comprise the majority of the formation. The majority of the coarse sandstones are confined to channels. Walker (1966a and b) describes channels, infilled with proximal turbidite sandstone, low in the formation. The top of the formation is drawn at the boundary between Assemblage B and C (see below) and therefore includes massive sandstones previously classed as Kinderscout Grit.

6.1.2.4 Kinderscout Grit Formation

Includes most of the: Kinder Scout Grit (Hull and Green, 1864)
4th Grit (Hull and Green, 1864)
Kinderscout Grit (Jackson, 1927)
Earl Crag Grit (Stephens *et al*, 1942)
Addingham Edge Grit (Stephens *et al*, 1942)
Bramhope Grit (Stephens *et al*, 1942)
Caley Crags Grit (Stephens et al, 1942)

Thickness: 100 to 180m, the maximum thickness being in the Otley area.

Type area: Kinderscout Plateau (SK0889). This area is adequate only as a type section for the lower part of the formation. It is suggested that definitive sections for the formation should be described from the North Derbyshire area.

Assemblage C, delta top sediments, comprise this formation. The base of the Kinderscout Grit, as used by previous authors, was poorly defined. It was generally taken as the incoming of the thick, very coarse, pebbly sandstone which forms the continuos grit edges. The top of Assemblage B, delta slope sediments, and the base of Assemblage C, delta top sediments, is, however, a level of great lateral variation. The base of the Kinderscout Grit, as previously understood, could be drawn well below the large-scale cross-bedding at the base of C1 if B3, channeled coarse sandstones, occur at the top of Assemblage B. Alternatively, where large-scale cross-bedding was absent and the top of Assemblage B was fine grained, the base of the formation was drawn at the bottom of the first coset of F.13, medium-scale cross-bedding. Thus the lower boundary of the Kinderscout Grit varied by over 100m.

It was therefore decided to define the base of the Kinderscout Formation as the base of Assemblage C; the base of the lowest, thick, widespread deposit of F.13, medium-scale cross-bedding, or the base of the channel infilled with the lowest unit of F.20, large-scale cross-bedding. This new definition restricts the lower bouhdary variation to 40m. It also has the advantage of making mapping easier because the boundary usually coincides with a sharp break in slope (Plates 1, 2 and 3) and isolated outcrops can usually be easily assigned to Assemblage B or C.

The upper limit of the unit is drawn at the base of a bed containing Reticuloceras gracile which appears to occur throughout the area.

6.1.2.5 Todmorden Sandstone Formation

Includes: Yoredale Grit (Hull *et al*, 1875)
Todmorden Grit (Wright *et al*, 1927)
Cobden Sandstone (Bisat and Hudson, 1943)

Thickness: Approximately 70m.

Type section: Here defined as Lumbutts, Todmorden. The base is seen in Lumbutts Clough (SD95382349) and a good section is exposed in the bankside leading up to Causeway (SD95172343).

The Todmorden Formation consists of Assemblage A, deep water sediments. The lower boundary is marked by the entrance of F.3, turbidite sandstones, and the upper boundary by the plane beyond which they are absent. Unfortunately the top of the succession is never seen. The sandstones of Spittle and Black Clough (SD957227) are probably near the top but their exact position is unknown due to faulting.

The facies of the Todmorden Formation are the same as those divided into the Mam Tor and Shale Grit Formations in the Saddleworth and Derbyshire area. The same thickening upwards of the sandstones occurs but it is impossible to split the Todmorden Formation into two divisions because of poor exposure and reduced thickness in comparison to the Derbyshire area.

6.1.2.6 Hebden Bridge Shale and Sandstone Formation

Includes: Sub-Kinderscout Grit (Wright _et al_, 1927)
Sabden Shales (Wray _et al_, 1930)
Todmorden Shales (Bisat and Hudson, 1943)

Thickness: 90m

Type section: The type section is here defined as Hebden Dale, Hebden Bridge. Exposures occur on the banks of Hebden Water between Lee Wood (SD991282), which is thought to be near the base of the formation, and the exposures below Hardcastle Crags (SD971302); the upper boundary lying between the bankside exposure and Hardcastle Crags.

The Hebden Bridge Formation is made up of Assemblage B, delta slope sediments. It is lithologically equivalent to the Grindslow Shale Formation and is again defined by the restriction of the majority of the coarse sandstones to channels. The bottom of the formation is not seen but the shales of Heeley Clough and its tributary (SD962230) are probably low in the formation.

6.1.2.7 Otley Sandstone Formation

Thickness: 15 to 40m

Type section: The type section is here defined as Storris House railway cutting (SE180447) near Otley, where the base of the formation is well exposed.

The Otley Formation consists of Assemblage A, deep water sediments and is lithologically similar to the Todmorden Sandstone Formation. The limits of the formation are again defined by the presence of F.3, turbidite sandstones.

6.1.2.8 Silsden Shale and Sandstone Formation

Includes: Addlethorpe Grit of Pool (Stephens et al, 1942)
Sutton Grit (Bisat and Hudson, 1943)
Otley Shell Bed (Stephens et al, 1942)

Thickness: Approximately 80m

Type section: The type section is here defined as Holden Beck (SE061455 - 066456) and the succession is shown in Fig. 8, Vol. 3.

Assemblage B, delta slope sediments, make up this formation and it is the facies equivalent of the Grindslow Shale and Hebden Bridge Formations. The coarse sandstones are restricted to channels.

6.1.3 Formations underlying the Kinderscout Grit Group

At least part of the formations underlying the Kinderscout Grit Group was deposited during the advance of the delta and it is therefore pertinent to discuss these formations. The Edale Shales (Jackson, 1927) only crop out south of the thesis area where they have an average thickness of 245m. The Edale Shales lie between either the Carboniferous Limestone or Alport Siltstones and the Mam Tor Sandstone Formation. The main outcrop of the Sabden Shales (Hull et al, 1875) also lies outside the thesis area. The Sabden Shales are bounded by sandstones of the Skipton Moor Grit Group below and of the Kinderscout Grit Group above. The shales beneath the Todmorden Formation in the Todmorden inlier have previously been attributed to the Sabden Shales (Wright et al, 1927). There is however no proof of an underlying sandstone and the shales may be better included in the Edale Shales; the accepted nomenclature will be followed in this thesis.

In Airedale and Wharfedale the shales below the Kinderscout Grit Group are underlain by the Middleton and Brocka Bank Grits (Stephens et al, 1953) which are higher than the Skipton Moor Grit Group. As they should not therefore be called Sabden Shales, they are here termed the Ilkley Shale Unit. The top of the unit is exposed beneath the Otley Sandstone Formation in Holden Beck, Silsden (SE061455) and at Storris House railway cutting (SE180447) near Otley. As the nature of their contact with the underlying sandstone is as yet undescribed, it is not thought desirable to formally define the unit.

6.2 Biostratigraphy

The Standard Stratigraphical Scale hierarchy for the Kinderscoutian (R_1) stage, as recommended by the Namurian Working Group (Ramsbottom, 1968a) is given in Table 6.3. The report gives the River Darwen at Samlesbury Bottoms (SD617290) and Crimsworth Dean (SD995325), Hebden Bridge as the type biostratigraphical succession.

Ashton (1974) in her study of the North Staffordshire succession has elucidated the biostratigraphy of R_1, especially in revising the *R. nodosum* zone, within which she recognized five beds with distinctive fauna (see Table 6.4).

The fauna of the Central Pennine Basin has not been re-examined but the level of the faunal localities within the lithostratigraphical succession has been recorded (see Fig.11, Vol.3). The faunal content of the formations of the Kinderscout Grit Group is discussed below.

6.2.1 Area 1

The only R_1 fauna that has been recorded from Area 1,

Table 6.3

STAGE	ZONE	SUBZONE
Kinderscoutian (R_1)	R. reticulatum (R_1c)	R. coreticulatum
		R. reticulatum
	R. nodosum (R_1b)	R. nodosum
		R. dubium
	R. circumplicatile (R_1a)	R. todmordenense
		R. circumplicatile

Table 6.4

Divisions of the *R. nodosum* zone according to Ashton (1974)

R_1b_v	*R. prereticulatum* (Ashton) *R. stubblefieldi* (Bisat and Hudson) *R. moorei* (Bisat and Hudson) *Ht.* aff. *divaricatus*
R_1b_{iv}	*R. stubblefieldi* group
R_1b_{iii}	*R. nodosum* (Bisat and Hudson) *R.* aff. *nodosum* *R. eoreticulatum* group *H. striolatum* (Philips) early form *H. spiraloides* (Bisat and Hudson)
R_1b_{ii}	*R. moorei* /nodosum group *Hd. ornatum* (Foord and Crick) *Homoceras henkei* / *striolatum* group
R_1b_i	*R. eoreticulatum* (Bisat)

Saddleworth and Longdendale, is the Butterly Band (A) (letters refer to fossiliferous levels shown in Fig.11, Vol.3) which is exposed in Redbrook Clough (SE026110) and Wessenden Brook (SE055087) near Marsden. The bed occurs within Assemblage C3, inter-channel deposits, of the Kinderscout Formation. It has a benthonic mollusc and brachiopod fauna and, to the south, contains R. butterlyensis (Ashton, 1974).

In the High Peak area the majority of R_1 is represented by the Edale Shales. The base of the Mam Tor Sandstone Formation appears to be irregular and overlies, according to Stevenson and Gaunt (1972) between 3 and 24m of Edale Shales of R_{1c} age (B). Ashton (1974), however, suggests that the first faunal bed beneath the Mam Tor Formation in Swint Clough, near Alport Castles (SK143914), is in fact equivalent to her R_{1bv} bed.

The R. gracile bed, marking the base of the overlying stage, occurs at several localities within the thesis area including Crowden Great Brook (SE062030) and Windy Hill (SD980147).

6.2.2 Area 2

The Sabden Shales of Area 2, Todmorden and Hebden Bridge, have yielded an extensive R. circumplicatile zone fauna. Five faunal levels have been recognized in Lumbutts Clough (SD951238):

(v) R. adpressum

(iv) R. aff. umbilicatum, R. paucicrenulatum

(iii) R. todmordenense

(ii) R. aff circumplicatile, Homoceras henkei, R. aff. pulchellum, Homoceratoides mutabilis, R. aff. coronatum, Ht. prereticulatus, R. todmordenense

(i) R. aff. circumplicatile, H. aff. henkei, R. aff. umbilicatum (for details see Bisat and Hudson, 1943). Bed (i) apparently belongs to the circumplicatile subzone whilst the rest belong to the todmordenense subzone. Bed (ii) probably correlates with the pulchellum, (iii) and (iv) with the paucicrenulatum / todmordenense and (v) with the dubium / adpressum beds of Ashton (1974). Unfortunately the fossiliferous localities of Lumbutts are now covered by a refuse tip and the relationship to the overlying Todmorden Sandstone Formation is unknown, although there is probably no great thickness of intervening shale.

The Sabden Shales are also fossiliferous in Ewood Clough (SD926246), Todmorden (D). R. moorei, H. striolatum (early form), Ht. cf. divaricatus occur in a bed seperated by 3.6m from a higher bed with R. cf. stubblefieldi, R. aff. davisi and R. reticulatum group (Bisat and Hudson, 1943). The higher bed is 3.6m below the base of the Todmorden Formation. The fauna here is typical of the upper part of R_{1b} (see Table 6.4).

The relationships of the Shewbroard Clough (SD925236) shales, with R. regularum and Hd. ornatum (bisat and Hudson, 1943) are unknown because of extensive faulting in this region. If it is the base of the Todmorden Formation which is exposed in Ewood Clough, it is unlikely that the Shewbroard shales overlie the "Cobden Sandstone" as suggested by Bisat and Hudson (1943, p394). The base of the Todmorden and Mam Tor Formations therefore appear to be of the same age but the Todmorden Formation base is possibly earlier than that of the Mam Tor Formation.

Bisat and Hudson (1943) record many fossiliferous localities within the Hebden Bridge Shale and Sandstone Formation but many have not yielded determinable goniatites. Near the base of the formation a bed containing H. striolatum has been recorded from Cross Stone Road, Todmorden (SD944248) (Wright et al, 1927 and Bisat and Hudson,

1943) and Spittle Clough (SD957227) (E).

Crimsworth Dean, north of Heptonstall, contains several fossiliferous beds. The lowest bed recorded, though not found by the writer, is north of Horse Bridge (SD987293) (F) which is approximately 70m below the base of the Kinderscout Formation. It has yielded H. striolatum early form and Reticuloceras sp. (Bisat and Hudson, 1943). Its position low in the formation suggests that it may be the same bed as at Cross Stones and Spittle Clough.

Further upstream at Black Scouts (SD988298) (G), Hd. ornatum, H. aff. striolatum, R.aff. regularum and R.aff. pulchellum have been recorded by Bisat and Hudson (1943). Lloyd and Stephens (1927) add R.reticulatum. The shales are approximately 55m below the base of the Kinderscout Formation, just below the middle of the Hebden Bridge Formation. The faunal assemblage is unusual but is probably best ascribed to the upper R_{1b} zone.

Continuing upstream two faults downfault the Kinderscout Formation which is exposed in the stream for 0.15km north of Wheat Ing Bridge (SD988303). This sandstone has been mistakenly called "Todmorden Grit" by previous authors (eg. Lloyd and Stephens, 1927, Wray et al, 1930, Bisat and Hudson, 1943, and Ramsbottom, 1968a).

Upstream the Hebden Bridge Formation is faulted up and below Outwood farm (SD989307) black shales yield abundant R.reticulatum s.s. and H.striolatum (Bisat and Hudson, 1943) (H). The same bed (now unexposed) in Hebden Vale, is thought to be the source of the type specimen of R. reticulatum figured by Phillips (1836) from High Green Wood (Bisat, 1924). The Crimsworth Dean locality has long been known for its uncrushed goniatites preserved in carbonate concretions (Spencer, 1898) - see Plate 21. The same bed is again exposed 0.2km upstream (SD99053088) due to further upfaulting.

The structurally complex nature of Crimsworth Dean makes it a poor type section for the upper R_1. If, as Ramsbottom (1968a) infers, there are no better sections it is suggested that a future borehole core could make a better type section.

Probably the stratigraphically highest bed of the Hebden Bridge Formation is in Heeley Clough (I), south of Mankinholes (SD9623, precise locality unknown) reported by Bisat and Hudson (1943). They identified Hd. ornatum, H. aff. striolatum, Ht. divaricatus and R. coreticulatum suggesting a high R_{1c} age, although if the identification of R. coreticulatum is incorrect the bed may well be high R_{1b}.

The delta slope, therefore, appears to have been established in the Todmorden / Hebden Bridge area by the beginning of R_{1c}. If the identification of R. reticulatum s.s. near the base of the Mam Formation is correct, as is presumed by Stevenson and Gaunt (1972), then the Todmorden and Hebden Bridge Formations shows considerable diachraneity with the Mam Tor, Shale Grit and Grindslow Shale Formations. If, however, as Ashton (1974) suggests, R. reticulatum s.s. is not found in Derbyshire then the beds may not be so diachranous.

The only bed with a goniatite fauna recognised within the Kinderscout Formation of area 2 is in upper Crimsworth Dean (SD994325) (J) and yields R. coreticulatum (Stephens et al, 1953). p43 The R. gracile bed is exposed at the head of Crimsworth Dean (SD992335) (Stephens et al, 1953) p 57 and was found in the Manshead boreholes (see Fig.23, Vol.2).

6.2.3 Area 3

The lowest R_1 beds of Area 3, Airedale and Wharfedale, are those of the Ilkley Shale Unit at Storris House railway cutting (SE180447), Otley (K) where Stephens et al (1953) recognized four fossiliferous levels;

(iv) R. aff. pulchellum

(iii) R. umbilicatum, R. circumplicatile

(ii) H. henkei, Ht. aff. varicatus, R. coronatum

(i) H. henkei, R. cf. coronatum

This fauna belongs to the R. circumplicatile subzone with the possible exception of (iv) which may belong to the lowest R. todmordenense subzone - Ashton's (1974) R. pulchellum bed. Ashton (1974) records a bentonite bed above the fossiliferous beds and correlates it with a similar bentonite above a bed with R. circumplicatile in north Staffordshire. The base of the Otley Sandstone Formation overlies the fossiliferous beds with well developed turbidite sandstones 2m above the top of bed (iv). In Poolscar Wood (SE220446) near Otley Stephens et al (1953) record H. striolatum and R. coronatum from the Ilkley Shale Unit. The bed is again probably R_{1a} in age.

Two fossiliferous beds occur in Holden Beck, near Silsden (SE061455) (L). The lower bed, thought to be not in situ (Stephens et al, 1953), contains R. adpressum and R. cf. umbilicatum while a higher bed contains R. aff. dubium. This fauna suggests a high R_{1a} age, probably equivalent to the R. dubium / R. adpressum level of Ashton (1974). The upper shales are at the top of the Ilkley Shale Unit in Holden Beck. It therefore appears that the base of the Otley Formation may be slightly diachronous with the initiation of turbidites commencing in mid-R_{1a} at Otley and late R_{1a} at Silsden.

Stephens et al (1953) record H.striolatum (early form) and Hd. ornatum from Lumb Clough. Unfortunately the faulted nature of the area does not allow one to accurately place the bed in a sedimentological sequence.

At Chevin Hall (SE191442) (M) near Otley Stephens et al (1953) record H. cf. striolatum, Ht. divaricatus, Hd. ornatum and R. cf nodosum which suggests a late R_{1b} age. This bed appears to be about 30m below the base of the Kinderscout Formation. Ramsbottom (1974) places the Otley Shell Bed (N), which locally occurs at the top of the Silsden Formation on Otley Chevin (SE199443), as low R_{1b}.

Within the Keighley area (O); R.reticulatum s.s., H. striolatum and Hd. ornatum occur 36m below the Kinderscout Formation at Sladen Bridge (SE018372); H. striolatum and R. cf. retio latum are recorded 50m below from the borehole in Keighley (SE058412); and H. striolatum occurs 60m below the Kinderscout Formation at Lower Laithe (SE015369) (all data from Stephens et al, 1953). This suggests an upper R_{1b} age for most of the Silsden Formation in the Keighley area and is probably an intermediate stage in the diachraneity shown between the Silsden Formation at Otley and the Hebden Bridge Formation.

The Keighley borehole (P) (SE058412) yielded Hd. ornatum and R. cf. reticulatum immediately above the lowest sandstone of the Kinderscout Formation. Also from above the lowest sandstone R. cf. reticulatum and H. striolatum were found in the Yeadon borehole (SE204410) and Carlton Moor borehole (SE224424) (Q). R. reticulatum (late form) was found higher in the sequence in both boreholes (see Stephens et al, 1953 for details of the three borehole logs). All these faunas are thought to be R_{1c} in age.

The R. gracile bed was encountered in the Yeadon, Carlton Moor and Snail Green boreholes.

6.3 Correlation

The formations, because of their definition, are easy to correlate on their facies association. This correlation, with the changes in thickness of the formations is plotted on the fence diagram (Fig.11, Vol.3). The thickness variations are discussed in section 8.2.

The Kinderscout Grit Group formations are diachranous between Wharfedale and Longdendale. This is best exhibited at the base of the Group which probably commences in middle R_{1a} in Otley and near the R_{1b}/R_{1c} boundary in Derbyshire. Deposition of Assemblage B, slope deposits, appears to have been completed in the Otley area prior to the incoming of the turbidites, of Assemblage A, in Derbyshire. The faunal evidence is lacking to provide more detailed comparisons but the faunal evidence fits into the general pattern of facies equivalent formations commencing earlier towards the north/northeast.

6.4 The Pendle Area

The Pendle area has not been studied in detail and has therefore not been dealt with in the above sections. Localities within the area have, however, been examined and a comparison with the thesis area is possible. The general succession is:-

Kinderscout Grit	=	Assemblage C
Shales with sandstones	=	Assemblage B
Parsonage Sandstone	=	Assemblage A
Sabden Shales		

All thickness and faunal data is based on Earp et al (1961), Price et al (1963) and Wright et al (1927).

At Samlesbury Bottoms (SD617290) (R) the Parsonage Sandstone overlies a bed containing Hd. ornatum and R. reticulatum. The turbidites therefore appear to have been initiated at virtually the same time as in the Mam Tor Formation. To the northeast at Sabden (S) the highest faunal bed seen below the Parsonage Sandstone contains R. eoreticulatum and R. moorei. H. striolatum and R. reticulatum occur at the base of Assemblage B, delta slope shales. Sabden may therefore represent a stage in the diachranaity shown between the Otley Formation and the Parsonage Sandstone at Samlesbury. At Samlesbury Assemblage B continues up to the R. gracile bed in the valley of the Darwen (SD636280) and fluviatile conditions are not established until the Revidge Grit (R_2).

CHAPTER 7 REGIONAL DETAILS

7.1 Area 1

The geological map of Area 1, Saddleworth and Longdendale, is Fig.4, Vol.3 and the complementary location map is Fig.1, Vol.3.

7.1.1 Shale Grit Formation

The base of the Kinderscout Grit Group is not exposed within Area 1 where the lowest beds occur as three inliers of the Shale Grit Formation. The most southerly inlier is Longdendale where patchy exposures of A2, parallel sided sandstones, occur in the lower part of Hollins Clough (SK052981). A temporary exposure occurred in 1971 on the construction of an overflow conduit between Torside and Rhodeswood Reservoirs (SK055982). Here 5 m of A3, wedge-shaped sandstones, with some channels, were observed.

The largest exposure of the Shale Grit is the partially submerged cutting on the banks of Dovestones Reservoir (SE016035). Here all three associations of Assemblage A have been seen. A normal fault with a throw of 8 m divides the cutting. Of particular interest is a large channel, with stepped sides, which cuts down into A2, parallel sided sandstones and A3, wedge shaped sandstones (section through these A2 and A3 beds are given in Fig.7, Vol.3), by 10m over 58m (see Fig.10, Vol.3). Two of the wedge-shaped sandstones thin from 3 and 2.4m to 0.4 and 0.5m respectively over 10m. The lower sandstone bed appears to be a continuation of the channel sandstone whilst the upper bed is truncated by the channel. The whole complex is thought to be a channel which has been reactivated on at least two occasions. The lower wedge-shaped sandstone being an overbank deposit of the first channel, this was then cut into by a channel with the second wedge-shaped sandstone as an overbank deposit. Finally a third channel eroded away most of the second channel.

At Diggle Junction (SE003076) A1 and A2 occur in tight folds with north-south axes. This local folding is probably associated with faulting.

7.1.2 Grindslow Shale Formation

Hollins Clough (SK053992) has good exposures in the Grindslow Shale Formation but unfortunately the contact with the underlying Shale Grit Formation is not seen and the succession is faulted. Most of the formation here consists of B2, inclined units, whilst B3, channeled coarse sandstones, is thin.

The best exposures of the Grindslow Formation are in Chew Brook and its tributaries, near Greenfield (SE030016)(see Plate 2). In the upper part of Chew Brook itself and in Great Gruff (SE028016) approximately 25m of B3, channeled coarse sandstones, occur. In Chew Brook the erosive bases of the channels, with flute and groove moulds, can be seen. The lowest channel, which occurs at the confluence of Great Gruff and Chew Brook (Plate 27) can be seen to downcut by over 3m. In Great Gruff (see section, Fig.8, Vol.3) 8m of B2 inclined units occurs within the otherwise coarse succession. By contrast the rocks of Rams Clough (SE018026) consist mainly of B4, parallel bedded sandstone and mudrock (see section, Fig.8, Vol.3). The top of the section in Rams Clough is probably 30m stratigraphically below the beds at the base of the Great Gruff section.

At Dove Stones (SE025038), north of Chew Brook, 75m of B3, coarse sandstones, occur (Plate 1); all of F.4, unlaminated sandstone. In Dove Stone Clough (SE029041) 15m of B2 inclined units underlie the thick B3 sequence. At least four B2 units occur, up to 9m thick, and downcut into one another. Thick B3 sequences, mainly of F.4, are also seen at Alderman's Hill (SE016046), Pots and Pan Stone Quarries (SE010050) and South Clough (SE026076).

Typical B3 channels are seen in the lowest of the Ladcastle Quarries (SD995059). Here F.4, unlaminated sandstone, F.9, micaceous silty sandstone and F.15, horizontally laminated sandstone, are interbedded. F.9 has <u>Pelecypodichnus</u> traces.

7.1.3 Kinderscout Grit Formation

Of the three formations, the Kinderscout Grit Formation has the most extensive outcrop in Area 1. It underlies most of the peat covered, plateau-like hills of the northern and southern parts of the area (see Plate 2).

In Longdendale F.20, large-scale cross-bedded sandstone occurs on the west side of Crowden valley (SE063000) and at Tintwistle Knarr Quarry (SK043992), 2.2km to the west. In the intervening Hollins Clough (SK052996), however, the facies is absent and Assemblage C commences with F.13, medium-scale cross-bedded sandstone. Laddow Rocks (SE057013) and Rakes Rocks (SE057016) to the west of Crowden Great Brook are unusually good exposures of a higher leaf of F.13 coarse sandstone. This upper coset of medium-scale cross-bedding is over 15m thick. The intervening finer sediments are mostly unexposed. Other outcrops of the upper part of the Kinderscout Formation of this region include a thin coal in Black Chew Grain

(SE047020) and the R.gracile bed in Crowden Great Brook (SE062031).

Buckton Quarry (SD991015), which was worked until 1971, shows the base of Assemblage C (see Fig.10, Vol.3). The 9m deep channel (see Plate 53 and 54) in the centre of the main quarry face appears to have been infilled at a high angle to the channel axis. It is thought that the large-scale cross-bedding infilling it was deposited by a delta infilling an abandoned channel (see Section 4.2.1). Underlying the channel to the west is a further set of large-scale cross-bedding of F.20 which again appears to sit in a channel. F.21, undulatory bedded sandstone occurs on this lower channel side. Underlying this channel and at the same level as the channel on the eastern side of the quarry F.4, unlaminated sandstones occur. Adjacent to the channel this facies contains large mudclasts, with no apparent organization, and is probably the product of bank collapse. Further away from the channel the sandstones contain no mudclasts and there is no evidence of post-depositional movement. The sandstones are not however completely massive and have thin layers of pebbles and rare, isolated, cross-bedded sets (Plate 40). X-ray examination of the "unlaminated" sandstones shows the concentration of certain grain sizes along planes (Plate 41). Trunks of Calamites and Lepidodendron up to 1.5m long and 0.25m in diameter occur within the facies. No preferred orientation has been observed but the plant fragments have almost certainly been transported to their present position as there is no evidence of associated rootlets or smaller plant fragments.

Directly overlying the channel complex is a coset of medium-scale cross-bedding, exposed on the quarry's upper level (Plate 53). Most of the remainder of the quarry consists of F.20, large-scale cross-bedded sandstone which has several internal erosion surfaces draped by F.9 micaceous silty sandstone. The quarry is cut by several faults, most of which appear to be low angled reverse faults.

Large-scale cross-bedding of F.20 is common on the "grit edges" to the east of Greenfield and Saddleworth. A gradational junction between horizontal toesets and underlying F.4, unlaminated sandstone, occurs at Wimberry Stones (SE015024, Plate 43). The same relationship is seen at Chew Hurdles (SE028015) where topset beds are also seen (Plate 44). Good examples of internal erosion surfaces occur in Dean Rocks (SE028040, top left in Plate 1). Greenfield Hall Quarries (SE028040) has an erosion surface overlain by F.9, micaceous sandstone with Pelecypodichnus. A 6m deep channel infilled with F.4, unlaminated sandstone, occurs in the top of this quarry and similar channels occur in the Running Hill Pits (SE018073). Other outcrops of F.20, large-scale cross-bedded sandstone, include Raven Stones (SE036047), Ravenstone Rocks (SE020077) and Dish Stone Rocks (SE030018).

A coset of medium-scale cross-bedding of F.13, up to 10m thick overlies all the natural exposures of large-scale cross-bedding (Plate 58). It is also well exposed at Channel Holes (SE027027), Bramley's Cot (SE026034), Ashway Rocks (SE028048) and Slades Rocks (SE016060).

The most complicated succession is the lower part of the Kinderscout Formation of the whole basin occurs in the Ladcastle and Den Quarries of Saddleworth (SD995060) and is sketched, from oblique aerial photographs, in Fig.10, Vol.3. The interpreted sequence of deposition and erosion for the whole channel complex is as follows :-

1. Erosion of first channel.
2. Infilling of channel with F.4, unlaminated sandstone, at the

base and F.20, large-scale cross-bedding. (This is the sandstone with graffiti at the base of Plate 49).

3. Deposition of fine grained C3, inter-channel deposits (overlying the lowest sandstone in Plate 49).
4. Erosion of second channel into C3 deposits (see Plate 49).
5. Infilling of second channel with F.4 and 20 (Plate 49, and lower part of Plates 50 and 51).
6. Deposition of second layer of fine grained, C3 deposits (left-hand side of Plate 50 and centre of Plate 51).
7. Erosion of third channel into second layer of C3 deposits and second set of large-scale cross-bedding, F.20 (upper part of Plate 50).
8. Infilling of third channel with F.4 and 20. (J and H respectively in Plate 48, sandstone at top of Plates 50 and 51).
9. Erosion of Fourth channel into third set of large-scale cross-bedding, F.20.
10. Infilling of fourth channel with F.4, unlaminated sandstone (K in Plate 48).

In the northern part of Area 1 the M.62 motorway cutting at Windy Hill (SD980147) exposes the upper part of the Kinderscout Formation. The lowest beds seen are of F.13, medium-scale cross-bedded sandstone. Upwards shale intercalations become common and some cosets of F.13 can be seen to thin into F.19, thin sandstones. Overlying the F13 beds is 8m of C3, delta top sediments, mostly F.2, mudrock. Two seatearths (F.22) occur in this sequence. A second coarse sandstone unit of F.13 occurs above. This has an erosive base, downcutting by over 3m and has groove marks on the base. The sandstone is about 6m thick and is overlain by a third F.22, seatearth and coal. Mudrock separates the coal from the overlying R.gracile bed. The whole cutting is folded into a broad anticline with a north/south axis.

At Derby Delph (SE017160) F.20, large-scale cross-bedded sandstone, and F.21, undulatory bedded sandstone are well displayed (see Fig.10, Vol.2 for small-scale sketch and Fig.10, Vol.3 for large-scale sketch showing position of plates, the position of section in Fig.11, Vol.2 and the labelling of erosion surfaces). There is no tectonic dip to the quarry as is exhibited in the quarry on the other side of the road where F.13, medium-scale cross-bedding overlies the large-scale cross-beds. The cross-bed set at the far west of Derby Delph can be followed down the bankside to the reservoir below and has a thickness of at least 34m. Three internal erosion surfaces (A, B and C) divide the large-scale cross-bedding of the quarry face. Carbonaceous mudstones occur at the bottom of the quarry below the eastern-most cross-bed set. This may be the channel base but is more probably a fine grained drape on the erosion surface (D) separating F.20 from F.21. Derby Delph provides the best exposure of F.21, undulatory bedded sandstone (Plates 55 and 56). The facies is divided by two internal erosion surfaces (E and F). The undulatory bedding of the quarry is described in detail in section 4.1.11.

The final outcrop of the Kinderscout Formation in Area 1 is Blackstone Edge (SD972163). The large-scale cross-bedding of F.20 is seen just north of the triangulation point and in Blackstone Edge Delph (SD962176). The cross-beds dip towards the east/north-east in contrast to the southerly direction of the large cross-beds throughout the rest of the area. Trough cross-bedding of F.13, indicating a current towards the south, is well exposed around the triangulation point (Plate 59).

7.2 Area 2

The geological map of Area 2, Todmorden and Hebden Bridge, is Fig.5, Vol.3 and the complementary location map is Fig.2, Vol.3.

7.2.1 Sabden Shales

The important fossiliferous exposure of the Sabden Shales in the lower part of Lumbutts Clough (SD951239) described by Bisat and Hudson (1943) is now covered by a refuse tip. The only remaining exposure of the Sabden Shales within the area mapped, is higher up Lumbutts Clough (SD95332352). The mudrock of F.1 is here intruded by sandstone dykes from the overlying Todmorden Formation.

7.2.2 Todmorden Sandstone Formation

The turbidite beds of F.3 tend to thicken upwards within the Todmorden Formation. This is best illustrated in the area between Lumbutts Clough (SD95402348) and Causeway Wood (SD95232353). At the top of the formation the thick amalgamated sandstones form the flat areas, of Longfield (SD940236), Mankinholes Tops (SD955242), Cross Stone (SD946249) and Rodwell End (SD958249), where the less resistant Hebden Bridge Formation has been removed by erosion.

Sections of the best exposures in the Todmorden Formation are given in Fig.7, Vol.3. The section in Hole Bottom Delph (SD940251) is typical of A2, parallel sided sandstones. The turbidite sandstones are separated from one another by mudrock or have eroded only the top of the underlying sandstone. Most of the (SD95352348) section shows a similar pattern but amalgamation occurs between some sandstone beds and thin beds of F.4, unlaminated sandstone occur.

The lower Longfield Wood (SD942239) section shows the same relationships. In the upper section, however, F.3 turbidites are absent whilst the F.4 beds are thicker and F.5, intermediate sandstones appear. At Doroad Scout (SD953239) thick F.5 beds occur and the mudrock percentage is considerably less than in the lower section. Finally, at Eastwood Wood (SD963256), mudrock is absent and F.4 and 5 are very thick. The beds here show marked lateral changes in thickness and are typical of A3, wedge-shaped sandstones.

Two localities occur near the base of the formation. In Lumbutts Clough (SD95402348) an apparent slump is exposed (see Fig.6, Vol.2 and section 2.2.2). In Shew Clough (SD960244) sandstone dykes are abundant.

At Lobb Mill Delph (SD953245) 30m of very coarse pebbly unlaminated sandstone of F.4 occurs. Mudflake concentrqtions are present and there are large rafts of fine sediment; one of which exceeds 10m in length. Unfortunately, due to the faulted nature of the area, the precise relationships of the sandstone are unknown. On the western side of the quarry the F.4 sandstones are seen to cut down over 6m into thinner wedge shaped sandstones of A3. It is suggested that the sandstone of Lobb Mill Delph was the infill of a major submarine fan channel. Similar coarse pebbly sandstones, though not as thick as at Lobb Mill, occur at Lad Stones (SD956247), Bean Hole Delph (SD950249), Longfield Quarry (SD941238) and Causeway Wood (SD951236). The exposures of Cross Stone Road (SD944247), the Mount (SD940248) and Castle Hill (SD949246) are less pebbly but contain far more mudclasts. These beds are unique in the whole area because, although over 10m thick, there are no beds free of mudclasts. The mudclasts are orientated with their long axes parallel to bedding. At Castle Hill there is a sharp discontinuity with beds below dipping at up to 73°, to the south, with strike 250, and the beds above which dip 80°, to the east, with strike 205. This is thought to be a slump feature. The large and abundant clasts and the evidence of slumping suggest, therefore, that the slope at the top of the submarine fan at least in the Todmorden area, was steep.

7.2.3 Hebden Bridge Shale and Sandstone Formation

There are only a few exposures of the Hebden Bridge Formation in the Todmorden area. In Heeley Clough and its tributary (SD961230) and in Spittle Clough (SD958227) up to 35m of F.1 mudrock and F.6 laminated silt occur; a good example of Association B1, fine grained deposits. These are overlain by medium-grained sandstone with mudflakes, carbonaceous material and poorly developed assymetric ripples. Unlaminated very coarse pebbly sandstone, F.4, of Association B3, channeled coarse sandstone occur at the top of the formation south of Lumbutts at Jail Hole (SD949227), Noon Stone (SD957223) and Jackson Rock (SD966230). At Blue Scar (SD972240), near Stoodley Pike, however, 12m of B1, fine grained deposits occur at the top of the formation. In Greenhurst Hey Clough (SD941257) B2 inclined units appear to be dominant but F.12, sharp based sandstones also occur. Unfortunately the tectonic dip of the rocks in this area is unknown.

To the north-east the formation is faulted down, by a series of faults running between Stoodley Pike (SD972242) and Blackshaw Head (SD960273), and crops out in the steep valley sides of the River Calder. In Jumble Hole Clough (SD965265) 4.25m of B4 parallel bedded sandstone and mudrock are overlain by 53m of B3 channelized coarse sandstone forming the top of the formation. The rippled tops of many of the F.12, sharp based sandstones, are well exposed in the stream bed. The wide variety of ripple types seen on some surfaces (see Fig.21a, Vol.2) suggests that the ripples were not in equilibrium with the current which formed them, probably because the current was short lived.

At Dale Clough (SD972266), only 0.61km from Jumble Hole Clough, the B3 sequence at the top of the formation is reduced to 4m of channels infilled with F.4, unlaminated sandstones, cut in F.1, mudrock. Underlying this (see section in Fig.8, Vol.3) is 10m of B4, parallel bedded sandstone and mudrock, in turn underlain by 62m of B2, inclined units (Plate 17). The dip of the inclined units is given on the section; there is virtually no tectonic dip. Two channels, infilled with F.14, zeta cross-stratification, cut into the B2 association (Plate 24). Heterogenous sandstones of varying dip also occur in Parrock Clough (SD969256) but the tectonic dip is unknown and it is posssible that the rocks are faulted.

Excellent exposures of B2, inclined units, with virtually no tectonic dip, occur on the southern bank of the Rochdale Canal (SD975266). In a small exposure opposite the mill several discordances occur truncating the underlying beds at up to 30° from the horizontal. These truncation surfaces are draped by units inclined parallel to the surface (Plate 18). East of the mill an inclined unit can be followed over 200m along the canal bank (see Fig.10, Vol.3). At the western end a series of sandstone dykes occur. They are infilled with sand of the same grain size as a sandstone bed which truncates the top of the dykes. The silts between the dykes dip at increasing angles towards the west and finally become homogenous, losing all trace of bedding. It is thought that the silt has slumped and that the dykes are the result of sand infilling slump crevasses (see section 3.3.2).

With the exception of Dale Clough, most of the area between the Stoodley Pike - Blackshaw Head faults and Hebden Bridge (SD9927) have thick massive beds of B3 at the top of the formation. These crop out at prominent crags such as Callis Nab (SD973262) and Horsehold Scout (SD982269).

A 15m thick exposure of B4, parallel bedded sandstone and mudrock occurs in the bankside of Colden Clough, just downstream from the waterfall (SD977281). The lowest 4m have been contorted into boudins and recumbent folds (Plate 23), thought to be the result of slumping (see section 3.3.4). The disturbed beds are erosionall overlain by undisturbed beds.

A series of exposures in the Hebden Bridge Formation occur on the banksides of Hebden Dale; the more interesting of which are shown in Fig.12, Vol.3. The exposure at Lee Wood (SD991281) (A) is thought to be near the bottom of the formation. The sandstone resembles that of Cross Stones (see section 7.2.2) but is thought to lie within a channel lateral to the more typical slope deposits north of Lee Mill (SD992285). At the latter locality (B), there is a 1.25m thick apparent scour and fill structure infilled with F.8, gradationally laminated sandstone. The 4m deep channel at locality D is filled with F.4, unlaminated sandstone, but at the downstream end concentrations of carbonaceous material and mud clasts occur in bands parallel to the channel sides, which dip at up to 40°.

A complex of B2, inclined units, and B3, channeled coarse sandstone, occurs at locality E (SD971302). Two B2 units are seen (Plate 16), of which the lower one is virtually horizontal. A truncation surface, cutting into the lower unit by over 8m, is inclined at 8° and the beds of the higher unit are parallel to it.

At the downstream end the upper unit is cut into by a channel infilled with F.4, unlaminated sandstones. Although eroding steeply into the bottom of the inclined unit the upper channel sides appear to be more concordant with the underlying beds. The orientation of the channel also appears to be near the strike of the inclined beds. It is suggested that this B3 channel has deepened a pre-existing channel, already partly infilled with B2 units; a procee suggested in section 3.3.3. Such a situation may arise if turbidity currents travel down a delta slope with slump gullies.

Most of the F.12, sharp based sandstones, of locality F (SD972305) have massive bases, a division of horizontal lamination and a rippled top. Trace fossils are abundant and some of the sandstone bases show erosional structures (Plate 22). Some of the sandstones are not as parallel as they appear at first sight; one sandstone bed thins in a downcurrent direction, as indicated by ripples, from 80 to 40mm over 7m. This reduction in thickness is accompanied by the loss of the massive division - a transition expected with decreasing proximality in turbidites (Walker, 1967). Rapid thinning supports the theory that the F.12 sandstones were local phenomena (see section 3.3.4). The medium-scale cross-bedding in this section contains abundant clasts which concentrate at the base of the sets.

Locality G (SD971306) shows a typical B3 channel with F.4, unlaminated, very coarse, pebbly sandstone overlain by F.15, horizontally laminated coarse sandstone (Plate 25). Large spherical concretions are common in the F.4 sandstone (Plate 26).

At H (SD972306) the base of a B2, inclined unit is seen. The beds of the unit dip at approximately 4°. The base is interesting because it shows faulting penecontemporous with deposition (Plate 19 and Fig. 22, Vol.2). The sequence of deposition would appear to be:

1/ Deposition of more than 0.45m of F.10, rippled fine sand.

2/ Faulting down of blocks of the rippled sand into the underlying mud. This may have been a loading phenomena or may be due to late movements of a slump which preceded stage 1.

3/ Flattening of the irregular topography, resulting from the faults, by erosion.

4/ Continuation of deposition of B2 sediments.

Much of the overlying beds are of very fine sandstone which appears to to be horizontally laminated. Closer examination shows ripple cross lamination in very thin sets. The net rate of sedimentation therefore appears to have been slow.

The bases of the F.12, sharp based sandstones, at locality I, show erosional structures (Plate 20). They have massive lower parts and horizontal lamination with abundant carbonaceous material in the upper part.

Unlaminated sandstones, F.4, occur at the top of the formation at Tom Bell Cave (SD979290).

In Crimsworth Dean, a branch of Hebden Dale, there are again several bankside exposures. At Black Scouts (SD988298) 10m of F1, mudrock and F.2, goniatite faunal bed are exposed and are overlain by F.4, unlaminated very coarse sandstone. South of Wheat Ing Bridge (SD988304) B4, parallel sided sandstones are folded into an anticline plunging towards the southwest.

A B1 association, of F.2, goniatite faunal bed with carbonate concretions (Plate 21) and F.1, mudrock, occurs in the stream north of Outwood (SD989307, 989308, 990308). At Gib Scout the top of the formation shows a near vertical contact between 10m of B1, fine grained deposits and near horizontal B2 deposits. Both are overlain by an apparently continuous bed of coarse sandstone. This unusual contact may have been produced by the infilling of a very steep channel or may be due to faulting, penecontemporaneous with deposition.

Further down the valley B 3, coarse channeled deposits, are seen at the top of the formation in Hollin Hall Wood (SD986299) and Smeekin Hill (SD992292).

7.2.4 Kinderscout Grit Formation

F.20, large-scale cross-bedded sandstone is well developed at the base of the formation throughout most of the area. At Utley Edge Quarry (SD963192), the old quarries near Lodge Hall (SD952206) and at Stony Edge (SD956214) the foreset dip is towards the west to north-west suggesting continuation with the Blackstone Edge area of Area 1. Good examples of downdipping intrasets occur at Stony Edge (Fig.8, Vol.2 and Plate 47).

At Gaddings Hole (SD953223) the large cross-beds dip towards the south west. Thus the edge of the tongue of north-west dipping cross-beds must lie between here and Stony Edge, 0.8km to the south. Two internal erosion surfaces occur in the 20m thick large cross-bedding at Gaddings Hole. The lower surface has a drape of F.9 fine grained sediment, approximately 1m thick. The higher erosion surface, separated from the lower one by 4m of cross-beds in the upper part of the quarry actually meets the lower surface and, in the bottom half of the quarry, the two are inseperable. The fine drape on the upper surface thins upwards and is absent at the top of the quarry.

F.20, large-scale cross-bedding, continues to be present at the base of the Kinderscout Formation along Langfield Common Edge as far as Long Stoup Quarries (SD968232) but further to the north-east, at Blue Scar (SD972240) F.13 medium-scale cross-bedding, occurs at the base. No trace of F.20 has been found in the hills north of Todmorden.

At Jumble Hole Clough (SD964267 and 962269) a set of F.20, large-scale cross-beds, at least 22m thick, is repeated by faults downthrowing to the east. At the upstream exposure an internal erosion surface occurs, which truncates more severely those beds lower in the set. Overlying the erosion surface are foresets parallel to it except at the base where a 3m bed of massive sandstone overlies the new horizontal erosion surface.

A similar 4m thick massive bed overlying the base of an internal erosion surface cut into concave upward foresets occurs at Hell Hole, Heptonstall (SD986277, Plates 44 and 45). Such erosion surfaces with massive beds are thought to be due to changes in flow pattern around the bedform forming the large-scale cross-beds (see section 4.2.1).

On the opposite side of Colden Clough from Heptonstall (SD980277) two sets of F.20, large-scale cross-beds occur (Plate 52). The lower set is at least 11m thick whilst the upper set is at least 9m thick. 4m of F.13 medium-scale cross-bedding separate the two sets. Large-scale cross-bedding appears to be present at the base of the formation in the hills south of Hebden Bridge and is exposed at Foster's Stone (SD977263), Horsehold (SD984268) and Wood Top Delph (SD996265).

F.20 is also well exposed in Hebden Dale north of Slack (SD982288), in Greenwood Lee Clough (SD971298) and at Hardcastle Crags (SD972302). In Crimsworth Dean large-scale cross-beds occur north of Wheat Ing Bridge (SD988304) (the "Todmorden Grit" of previous authors) and in Small Shaw Wood (SD991308). At Lumb Hole (SD992314), however, the formation commences with F.13 medium-scale cross-bedded sandstone.

F.21, undulatory bedded sandstone occurs in Slater Ing Wood (SD976283). The large undulation is cut into at the eastern end by a F.4, massive bed, which is erosional by over 4m. F.13 medium-scale cross-bedding of the lowest sandstone of the Kinderscout Formation is well exposed throughout the area but is best seen in Eaves Wood, Heptonstall (SD984279).

Four areas with exposure in the upper part of the formation occur. At Grinding Stone Hole near Oxenhope (SE014338) F.16, striped sandstone with <u>Scolicia</u> occurs between two F.13 medium-scale cross-bedded sandstones. Similar beds with abundant <u>Pelecypodichnus</u> in the F.16 beds occur at the head of Crimsworth Dean (SD991328). At Stoodley Pike (SD972242) the following succession occurs :-

1m	F.13, medium-scale cross-bedded, very coarse pebbly sandstone.
3m	unexposed (silt?)
1.5m	F.1, silt
1m	wave rippled medium sandstone, F.10.
0.05m	F.22, coal
1m	F.22, seatearth
8m	F.10, rippled medium sandstone. Wave ripples in places
3m	F.4, massive very coarse sandstone.

A sketch section of the seatearth and coal is given in Fig.12, Vol.2. At Cellar Hole Delph (SD992184) F.13, medium-scale cross-beds are seen.

Fortunately detailed examination of the upper part of the Kinderscout Formation was possible through cores drilled through Great Manshead Hill (SE0020). Five boreholes (see Fig.2, Vol.3 for location) were put down before the construction of a tunnel to divert water from Turvin Clough to Baitings Reservoir (SE005187). Detailed logs are given in Fig.9, Vol.3. Although there is considerable lateral variation between the boreholes a correlation, based on two faunal beds and four F.22, seatearth and coal, beds, was possible (see Fig.23, Vol.2).

The C3, interchannel deposits are more clearly seen than in outcrop (Plates 63 to 68 show various facies). Coarsening and fining upward sequences can clearly be seen in the logs. The sedimentry structures in the coarse pebbly sandstones were often difficult to discern but most appeared to be F.13 medium-scale cross-bedding. The coarse sandstones are thought to belong to C2, minor channel deposits although the lowest sandstone in boreholes 1,2 and 3 may be the C1 sandstone at the base of the Formation.

7.3 Area 3

The geological map of Area 3, Airedale and Wharfedale, is Fig.6, Vol.3 and the complementary location map is Fig.3, Vol.3.

7.3.1 Ilkley Shale Unit

The Ilkley Shale Unit has been observed in four localities. In Lumb Clough (SE007435) shales of the H stage are exposed and on excavation for a gas pipeline in 1972 produced ellipsoid concretions up to 0.25m wide. An A1 association of F.1, mudrock, and F.2, goniatite faunal bed, of R_1 age occur in Holden Beck (SE061455), Storris House railway cutting (SE180447) and Poolscar Wood (SE220446).

7.3.2 Otley Sandstone Formation

The Otley Sandstone Formation is only exposed in three places. Good sections are seen in Holden Beck (SE062455) where both A2, parallel sided and A3, wedge-shaped sandstones occur (see bottom of Holden Beck section in Fig.8, Vol.3). In Cow Close Gill (SE125469) 13m of A3 sandstones occur. Amalgamated beds with mudclasts are common. In Storris House railway cutting (SE180447) the base of the formation is seen. A sandstone dyke (Plate 5) and a discontinuity (Plate 4) occur. The latter is thought to be a product of slumping (see section 2.2.2).

7.3.3 Silsden Shale and Sandstone Formation

A good section through the Silsden Formation occurs in Lumb Clough (SE007428, see Fig.8, Vol.3). Most of the lower part of the section is composed of F.8, gradationally laminated sandstone with minor F.10 ripple laminated sandstone and F.1 mudrock. These are typical of the B2 association, though a few sharp-based sandstones, F.12, of B4, parallel bedded sandstone and mudrock also occur. Above this lies a B3, channelized sequence of F.4, unlaminated, F.15, horizontally laminated and F.13 medium-scale cross-bedded coarse sandstones with minor F.8, gradationally laminated fine sandstone. The top 26m belongs to the B1 association and is composed entirely of F.1, black mudrock.

In Eastburn Quarry (SE020440), the top of which is approximately 30m below the top of the formation, 10m of F.11, ripple laminated sandstone with coarse sandstone (Plate 14) overlie 13m of F.1, unlaminated very coarse pebbly sandstone. Two contorted beds occur within the F.11 beds.. Abundant carbonaceous material is present, including Calamites and Stigmaria. Some thin, impersistent coals, up to 5mm thick occur and are presumably the product of water-logged vegetation transported to the area. Pelecypodichnus also occurs within F.11, which is thought to be a mouth-bar deposit (see sections 3.1.9 and 3.4). A fault, downthrowing 10m to the west is well exhibited in the quarry face.

At Holden Beck (SE064456) 21m, up to the top of the formation, is seen (see Fig.8, Vol.3). The sediments are mostly those associated with B2, inclined units. Low in the sequence a 2m thick contorted bed is present, probably the product of slumping. Several B3, channeled coarse sandstones also occur. The thin nature of the Silsden Formation in Holden Beck, compared to Lumb Clough, is problematical. It can best be explained by unexposed faults which may thicken the Lumb Clough sequence and/or conceal the base of the formation in Holden Beck.

Two small channels, infilled with F.4, unlaminated medium sandstone, cut into F.10 ripple laminated sandstone in a 6.8m thick sequence near the top of the formation at Brunthwaite Crag (SE060461).

Three metres of F.6, laminated silts and F.8, gradationally laminated very fine sandstone occur in Heber's Ghyll (SE099472). Further upstream 2m of turbidite-like sandstones occur. These are thought to be F.12, sharp based sandstones, but may belong to the Otley Formation if the stream is faulted.

A 10m thick B4, parallel bedded sandstone and mudrock sequence occurs in Cow Close Gill (SE125467, the lowest 7m is shown in Fig.8, Vol.3) at the top of the formation. Lower in the valley F.8, gradationally laminated very fine sandstone with <u>Scolicia</u> occurs in landslipped material.

In Coldstone Beck a 68m thick section is exposed up to the top of the formation. Almost the entire sequence is composed of F.1, mudrock and F.8, gradationally laminated sandstone. The latter occurs in units of varying orientation (see section Fig.8, Vol.3, tectonic dip approximately 10^{o}/280).

In the lower part of Great Dib, Otley (SE198445) are several small exposures in the Silsden Formation. The sediments are those usually associated with B2; F.6, laminated silts, F.8, gradationally laminated sandstone and F.10, ripple laminated sandstone. Several small discontinuities suggest that inclined units are present. There are also erosive beds, up to 0.7m thick, of F.4 unlaminated fine to medium sandstone. At the top of the formation in Great Dib (SE199443) the 5.2m thick Otley Shell Bed (F.7) is exposed. Only 1.2km to the east, at East Chevin Quarry (SE212444) the shell bed is absent and is replaced by a variable sequence of F.1, mudrock, F.4, unlaminated coarse sandstone and F.8 gradationally laminated sandstone with <u>Scolicia</u> and <u>Pelecypodichnus</u> (Plate 35).

A final outcrop of the Silsden Formation occurs in Willows Court (formerly Pool Station, SE246447). At the rear of number 20 a temporary exposure, in 1973, showed 4m of F.1 mudrock and F.2 goniatite (indeterminate) faunal bed. At the rear of number 13, 3m of horizontally bedded medium sandstone, with thin layers of mudclasts, is exposed.

7.3.4 Kinderscout Grit Formation

F.20, large-scale cross-bedded sandstone is well developed in the west of Area 3. It is best seen at Hangingstone Quarry, Earl Crag (SD992431) where 17.5m of one set are seen (Plate 46). Here an internal erosion surface is draped by 1m of F.9, micaceous silty sandstone. The large-scale cross-beds overlying the internal erosion surface continue, in an upcurrent direction, as a 1m thick cross-bed set overlying the lower set of large cross-beds. The stages of development envisaged are :-

1. Formation of large-scale cross-beds by migration of bedform at high stage.
2. Cessation of movement of bedform, modification of the crest and deposition of a fine grained drape at low stage.
3. Second high stage with bedform height adjusting to the new flow conditions before reactivation of the slip face.

At Strikes Quarry (SE012435) 10m of a F.20 set is seen. In the eastern end of the quarry this is cut into by an F.4, unlaminated very coarse sandstone, by over 6m. The base of the formation in Lumb Clough (SE002426) between Earl Crag and Strikes Quarry, shows no sign of the large-scale cross-beds and commences with a coset of F.13 medium-scale cross-beds.

The F.13 beds which form the lowest sandstone of the Kinderscout Formation in this area (the Earl Crag and Addingham Edge Grit of previous authors) is well exposed at Earl Crag (SD984429, Plate 3) and at Raven Stones (SE014437). Walker's (1955) palaeocurrent histogram of Earl Crag cross-beds showed a bimodal current distribution with maximum readings in a south-west and north-eastwards direction. An examination of the crags showed all the trough cross-bedding which formed the majority of the crag, indicates a palaeocurrent towards the south-west. Only two sets of planar cross-beds indicate a north-eastwards direction. The planar cross-beds are, however, given more weight in a palaeocurrent survey as they are exposed in three dimensions whilst most of the trough cross-beds are only seen in two dimensions. The position of the planar cross-beds at the top of the coset suggests that they may be low stage modification either by development of a slip face on the side of a braid bar, as described by Rust (1972), or by development of a delta lobe in front of a channel dissecting a bar, as described by Collinson (1970).

The only place on Rombalds Moor where the lowest sandstone of the Kinderscout Formation (the Addingham Edge Grit of previous authors) shows F.20, large-scale cross-beds is Ilkley Crags (SE124465) where 6m of one set are seen. F.13, medium-scale cross-bedding of the lowest sandstone forms the prominant crags of Brunthwaite Crag (SE038464), White Crag (SE044467), Addingham Edge (SE075472), Panorama Rocks (SE103470) and Hanging Stones (SE130467). F.13 is the lowest facies of the formation in Holden Beck (SE066456) and at Stead Crag (SE139457).

The top of the lowest sandstone of the Kinderscout Formation between Hugh Teal Hall (SE075471) and Cow and Calf Rocks (SE130467) shows disturbed bedding (Plates 60 and 61). The cross-beds have been deformed after deposition, apparently due to water escaping from the sediment. The localised occurrence of these deformed beds, covering an area 5.5km long, suggests that they may be due to an earthquake.

F.20, large-scale cross-bedding has not been seen at the base of the lowest sandstone in the Otley area (the Caley Crags Grit of previous quthors) though in East Chevin Quarry (SE212444) a 2.5m thick set occurs at the base of a 9m coset of F.13, medium-scale cross-bedded sandstone. The cross-beds of the lowest sandstone are well seen at East Chevin Crags (SE209444) and Caley Crags (SE230445). At the latter locality large dewatering structures occur at the top of the coset. Two deformed beds are present, separated by 2m, and it is possible that one is the equivalent of the Rombalds Moor beds.

6m of the top of the medium-scale cross-bedded sandstones are also seen in Pool Bank Quarries (SE233443), where they are overlain by a 6m thick coarsening upward sequence of F.16, striped silts and sandstones and F.10, medium sandstone with symmetrical and asymmetrical ripples. This is in turn overlain by a further coset, at least 2m thick, of F.13 medium-scale cross-beds.

The sandstones of C2, minor channels deposits, in the upper part of the formation, form scarps throughout the area and are particularly well seen at Black Edge (SE014426), south of Steeton (SE038435), Long Ridge (SE090467), Green Crag (SE130469) and The Chevin (SE205442).

On the laying of a gas pipeline in 1972, between Whitley Head (SE031436) and High Hollins (SE043429) south of Steeton, F.17, striped silts and sandstones, F.10, rippled fine sandstones, F.19, thin sandstone beds and F.22, seatearths were excavated between the very coarse sandstones.

CHAPTER 8: BASIN ANALYSIS AND SYNTHESIS

8.1 Palaeocurrents

Rose diagrams of palaeocurrent data from F.3, turbidite sandstones, F.20, large-scale cross-bedded sandstone and F.13, medium-scale cross-bedded sandstone for the whole of the thesis area are shown in Fig.24, Vol.2. All three facies show a vector mean towards 10 and 20° west of south. The vector magnitude (Curray, 1956) is 49.1 for the turbidites (where both sense and direction were known), 54 for the medium-scale cross-beds and 65 for the large-scale cross-beds. These all have high levels of significance with values of variance less than 10^{-4} (Rayleigh test of significance). All other data from F.10, rippled sandstones, parting lineation in F.15, horizontally laminated sandstone and erosional structures on the channel bases fit the general palaeocurrent pattern. There is no evidence from the palaeocurrent to suggest that the delta sediments were reworked by tides or basin currents.

In Fig. 25, Vol.2 the palaeocurrent data for the turbidites and large and medium-scale cross-bedding is shown for the three areas. Vector means have a wider range than in the grouped data but again all indicate a current towards the south-southwest. The only apparent change throughout the area is that the cross-bedding in the north, Area 3, indicates a more south-westerly current than elsewhere.

8.2 Thickness variations

Namurian sedimentation is broadly controlled by a block and basin structural pattern. The Kinderscout Grit Group was laid down in the Central Pennine Basin, lying between the Askrigg Block to the north and the Derbyshire massif to the south. The Namurian of the basin is at its thickest around Manchester and thins away in all directions (Ramsbottom, 1974). The rocks studied lie to the north of the zone of maximum thickness.

Although it is likely that compaction, subsidence and possibly changes in sea level occurred during the deposition of the Kinderscout Grit Group, it is thought that the thickness of sediment required by the delta to fill the basin (ie. from the base of the first turbidites of Assemblage A to the top of Assemblage B) is a guide to the original basin depth for comparitive purposes. Whilst compaction of the sediment decreases the original sediment thickness, subsidence during deposition increases the sediment thickness compared to the original basin depth.

A fence diagram showing the thickness variations throughout the area is given in Fig.11, Vol.3. The most noticeable feature is the marked thickening of the R_1 sequence to the south; the group being approximately twice as thick in the Peak District as at Otley. All this increase in thickness takes place in Assemblage B, delta slope, and more especially in Assemblage A, deep water sediments. Assemblage C, delta top sediments, is, in fact, thicker at Otley than in the Peak District.

The Sabden area is unusual because of the thickness of Assemblage C (335m) which is far greater than anywhere else in the basin. Remarkably, the assemblage decreases in thickness westwards until it dies out north of Blackburn. The thinnest succession is in the Silsden area where the whole R_1 sequence is only about 200m thick.

At the commencement of the R_1 stage the floor of the basin appears to have increased in depth away from the Askrigg Block. This asymmetry, as Ramsbottom's (1966) correlation diagram suggests, is probably partly due to northward shallowing against the underlying Skipton Moor Grits and sandstones of E_2 and H stages. It is perhaps significant that the thin succession in the Silsden area is close to the area of maximum development of the Skipton Moor Grits (J.G. Baines, personal communication).

8.3 Palaeogeographic evolution

The Skipton Moor Grits had established, by E_{2a}, a delta top environment in the northern part of the Central Pennine Basin (J.G. Baines, personal communication). With the exception of the thin Middleton and Brooka Bank Grits, no coarse sediment entered the basin through the remainder of the Arnsbergian (E_2), Chokerian (H_1) or Alportian (H_2) stages. Transgression took place and at the commencement of the Kinderscoutian a sea, probably between one and four hundred metres deep, covered the entire area.

The first sediments of the Kinderscout delta entered the northern part of the area during R_{1a} as turbidity currents. Growth of the early delta within the thesis area appears to have been slow but gradually the delta advanced until fluviatile conditions were established in the Otley area at about the R_{1b}/R_{1c} boundary. At about the same time the earliest delta front turbidites were reaching north Derbyshire and Sabden (see Fig.11, Vol.3).

The major period of delta advance was during the R_{1c} zone, and by the end of the zone, delta top conditions were established in all but the westernmost part of the basin. Coals, seatearths and fluviatile sandstones mark the establishment of terrestrial conditions. Transgression took place several times during R_{1c} creating wide areas of shallow water which were sometimes marine. These bays were gradually infilled, by minor deltaic sequences, reestablishing terrestrial conditions. In some places (eg. Manshead see Fig.9, Vol.3) up to five seatearths were formed all followed by transgression. Finally at the end of the Kinderscoutian a final transgression, marked by a bed with R. gracile, submerged the entire delta top.

8.4 Major factors controlling delta evolution

8.4.1 Compaction

Carbonate concretions within various facies have been described and presented as evidence of post-depositional compaction of the sediment (see sections 2.1.1, 3.1.2 and 3.1.6). The amount of compaction is dependant on grain size. The high angle of much of the cross bedding shows that the very coarse pebbly sandstone has suffered virtually no compaction whilst the finest mudrocks have compacted by at least 50%. The compaction rate is an important factor in determining the effect of compaction on the delta. It seems probable that considerable compaction of the finer sediment of the delta took place during and after establishment of delta top conditions. With a slow rate of sedimentation, as in a coal swamp, this compaction may well have been sufficient to lead to transgression.

8.4.2 Subsidence

If one assumes that the entire thickness of Assemblage C, delta top sediment, is due to compaction of underlying deposits, as envisaged in section 8.4.1, the overall percentage compaction of the sediments (y) can be approximately calculated:-

$$y = \frac{c.100}{a+b+c}$$

where a,b and c are the thicknesses of Assemblage A, B and C respectively. The equation assumes no compaction of mudrock underlying the lowest turbidites and that full compaction of the sequence had taken place when the last of it was deposited. In the Kindersoout area y has been calculated as 25%. This figure, though rather high, does suggest that compaction may account for all the the thickness of the Kinderscout Formation in that area, especially if one considers the compaction of the Edale Shales. In the Todmorden and Otley areas, however, y has been calculated as 41% and 64% respectively; values which are obviously too high for just compaction. Subsidence must therefore be an important factor leading to transgression.

Evidence of considerable subsidence prior to the R_1 is shown in the northern part of the basin where the basin had subsided from being a delta top environment at the end of E_{2a} to being a deep water environment in R_{1a}.

It is suggested that subsidence was continual throughout the basin during the Kinderscoutian. The differing y values are probably more related to the amount of subsidence than to compaction:

$$s = y - x$$

where s is subsidence as a percentage of the thickness of the three assemblages and x is a constant representing compaction as a percentage of the thickness of the three assemblages. The greater value of s northwards is to be expected because of the greater length of time during which the north had delta top conditions.

Although subsidence undoubtedly took place during the deposition of Assemblage A, deep water sediments, it is not thought that differing subsidence rates could lead to the great differences in thickness of the assemblage. These are probably more related to original basin depth. The smaller differences in the thickness of Assemblage B, delta slope deposits, from 85m in the north to 120m in the south, could be due to differing subsidence or compaction rates. If so, then the top of Assemblage A may have been depth related.

8.4.3 Changes in sea level

No evidence has been found in this study of any rise or fall in sea level. Since Wright et al (1927) first described cycles of sedimentation in the Millstone Grit and related them to delta advances many authors (eg. Reading, 1964; Collinson, 1968 and 1969; Ramsbottom, 1974) have assumed that some of the Namurian goniatite beds are the result of a rise in sea level. This supposition is apparently based on Wright et al's positioning of "marine beds" at the base of their coarsening upward sequences although they did not suggest rises in sea level to explain their cycles, most of which are over 30m thick. Bott and Johnson (1967) discuss the possibility of eustatic rises in sea level as a mechanism of formation of Carboniferous cycles of sedimentation.

The only goniatite bed found near a seatearth within the thesis area was the R. gracile bed which occurs up to 4m above a seatearth. One faunal bed, the Otley Shell bed was certainly not deposited after a rise in sea level as it is cut into by fluviatile sandstones. Any rise in sea level would be expected, through time, to have a complementary fall in sea level, assuming that sea level did not have a net rise in the Kinderscoutian. Such falls in sea level could lead to valleys and widespread exposure of the delta top. There is no evidence of any such phase, at least within the area of observation.

It must be assumed therefore, from the evidence within the Central Pennine Basin, that there was little change, if any, in sea level during the Kinderscoutian.

8.4.4 Salinity changes

The Namurian of the Central Pennines appears to have been deposited in at least a partially enclosed, tideless basin and could therefore have been subject to salinity variations. Most authors have assumed that the salinity did vary and argue that the restriction of the marine fauna to thin beds of wide lateral distribution must be due to the basin only rarely being fully marine. Ramsbottom et al (1962) and Ramsbottom (1969 b) argue that faunal phases within "marine bands" are a reflection of salinity variation. Salinity undoubtedly varied within the basin if the size of the river envisaged for the R_1, possibly with a fluctuating discharge, entered an enclosed basin with restricted circulation. Such variations would however be local to the distributary mouths and are not likely to have affected the entire basin.

Did the salinity of the entire basin change? The goniatites undoubtedly indicate that the basin was marine at times but the barren nature of the rest of the sequence does not neccessarily mean that it was not marine. The writer, however, believes that the rippled beds of the B_2, inclined units, were deposited by a continuous density current which would only have formed if the basin salinity was very low. Thus it would appear that during the deposition of most of the sediment, conditions were non-marine.

Salinity variations are probably due to water flow pattern changes at the connection of the enclosed basin with a more open sea. Such variation could take place due to variation in fresh water supply to the basin, to tectonic control of a bar at the edge of the basin, or to changes in sea level. The last possibility is thought unlikely, as discussed in section 8.4.3.

8.4.5 Tectonic control

The much greater thickness of the Namurian in the Central Pennine Basin compared to the Askrigg Block is apparently due to the greater subsidence rate within the basin. This difference in subsidence was probably taken up by movement on the Craven faults. It has been suggested that such tectonic movements may have affected Carboniferous sedimentation (Bott and Johnson, 1967) causing sudden relative rises in sea level. There appears to be little evidence of varying rates of sedimentation due to such movements within the R_1 sequence. The apparent increase in sedimentation rate during R_{1c} was more probably due to uplift in the area of erosion, North Atlantis, as suggested by Reading (1964) or may simply be due to the longer time span covered by the zone. Until more details are known about the thickness of the R_1 sediments in East Yorkshire and the North Sea the possibility of a major delta switch at the beginning of R_{1c} cannot be discounted.

8.5 Morphology of the delta

The Kinderscoutian delta was obviously a fairly large delta by modern standards. Wright and Coleman (1973) recognized a spectrum of delta types from the fluvial-dominated low-wave-energy type as epitomized by the Mississippi Delta to the wave-dominated, low-fluvial-influence type epitomized by the Senegal Delta. The Kinderscoutian delta was obviously closer to the former. Features of the Mississippi type of delta which the Kinderscoutian delta probably had were; (a) a highly indented coastline with extended distributaries, (b) a marsh and bay environment on the delta top and poorly sorted silts and sands lateral to the channels.

The sand bodies of such deltas are ideally "shoe-string" bodies.

The deep distributary channels of the Kinderscoutian delta did not migrate laterally and individual channel bodies probably have a shoe-string pattern. The sand bodies are, however, more laterally continuous than would be expected. This is thought to be because of the braided nature of the river channel in contrast to the meandering form of the rivers feeding most of the major present day river dominated deltas. Of the large present day braided rivers many, such as the Brahmaputra and Congo have no delta due to the proximity of the continental slope or have wave-dominated deltas as in the Sao Francisco of Brazil (Wright and Coleman, 1973). The fluctuating position of the delta top channels is thought to have given rise to the laterally continuous sand bodies and are are clearly distinguishable from well sorted clean beach ridge sands which would also be laterally continuous (eg. Oomkens, 1974).

A final feature of the river dominated delta according to Wright and Coleman (1973) is that the offshore slope has a low gradient and is convex upwards. This configuration appears at first to be at variance with the suggested presence of various types of gravity flow on the delta slope. A closer examination of the river-dominated Mississippi delta shows, however, that the distributaries push forward to the break in the delta slope and the distributary mouth bar is at the top of a slope steep enough for slumping (Shepard, 1955 and 1956). It is suggested that the current flowing out of the distributaries occasionally formed undercurrents but normally formed a plane jet. During the plane jet flow deposition took place from suspension and the coarse sediment must have accumulated at the mouth bar, thus choking the distributary. This coarse sand may have been washed over the edge of the slope at high stage or during periods of intense wave activity. In either case the sediment appears to have picked up enough kinetic energy on the slope to form turbidity currents. The low gradient of the delta between the distributaries could explain the small effect that waves had on the delta. Even in enclosed present day seas many deltas (eg. the Rhone in the Mediterranean (Oomkens, 1970), the Po (Nelson, 1970) and the Danube (Wright and Coleman, 1973) are wave dominated and the river dominated nature of the Kinderscoutian delta is probably therefore due to the river discharge rather than the sheltered nature of the delta.

The delta top sequence is typical of a delta undergoing subsidence. The cyclic sedimentation of the Mississippi River deltaic plain (Coleman and Gagliano, 1964) is very similar to the upper part of Assemblage C. Transgression takes place during periods of low sedimentation rate and the resulting bays are eventually infilled by "sub-deltas".

There is no precise present day analogue to the delta model proposed but, of the described modern deltas, the Kinderscoutian delta appears to have shown strong similarities in certain respects to the Mississippi (Shepard 1956, 1960 and Coleman and Gagliano , 1965), the Rhone Delta in Lake Geneva (Houbolt and Jonker, 1968) and the Brahmaputra (Coleman, 1969). The delta probably had a similar delta top to the Mississippi with much of it under water and with straight distributaries extending far out on to the delta top. Like the Mississippi, the top of the delta slope was probably liable to slumping. The Rhone delta is similar in that it is forming in a basin of low salinity and has a submarine fan, though it is probable that the density currents of the Kinderscoutian were not all river flow undercurrents. The Kinderscoutian delta was also probably similar to the Rhone delta in that the river flow appears to have been the main factor determining deposition within the basin. The

Kinderscoutian river probably had a high bed load and a fluctuating discharge as in the present day Brahmaputra, though the sand was considerably coarser.

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.

___ (1963a) The classification of cross-stratified units, with notes on their origin. Sedimentology, 2, 93-114.

___ (1963b) Asymmetrical ripple marks and the origin of water-laid cosets of cross strata. Lpool Manchr geol. J. 3, 187-234.

___ (1964) Primary current lineation in the Lower Old Red Sandstone, Anglo-Welsh Basin. Sedimentology, 3, 89-108.

___ (1965a) The sedimentation and palaeogeography of the Old Red Sandstone of Anglesey, North Wales. Proc. Yorks. geol. Soc. 35, 139-86.

___ (1965b) A review of the origin and characteristics of recent alluvial sediments. Sedimentology, 5, 89-191.

___ (1968) "Current ripples", pp433. North Holland, Amsterdam.

___ (1969) Erosional current marks of weakly cohesive mud beds. J. sedim. Petrol. 39, 607-23.

___ (1970a) "Physical Processes of Sedimentation", pp 248. Unwin, London.

___ (1970b) The sequence of sedimentary structures in turbidites with special referance to dunes. Scott. J. Geol. 6, 146-61.

___ (1970c) Studies in fluviatile sedimentation: a comparison of fining upwards cyclothems, with special reference to coarse-member composition and interpretation. J. sedim. Petrol. 40, 298-323.

___ (1970d) A quantitative model of grain size and sedimentary structures in lateral deposits. Geol. J. 7, 129-46.

___ (1971) Massive beds in the Central Pennine Basin: a discussion. Proc. Yorks. geol. Soc. 38, 293-4.

Allen, J.R.L. and N.L. Banks (1972) An interpretation and analysis of recumbent-folded deformed cross-bedding. Sedimentology, 19, 257-83.

Ashton, C.A. (1974) Palaeontology, stratigraphy and sedimentology of the Kinderscoutian and Lower Marsdenian (Namurian) of North Staffordshire and adjacent areas. Unpublished Ph.D. thesis, University of Keele.

Bagnold, R.A. (1954) Experiments on a gravity-free dispersion of large solid spheres in a Newtonian fluid under shear. Roy. Soc. Lond. Proc., Ser. A, 225, 235-94.

___ (1955) Some flume experiments on large grains but little denser than the transporting fluid, and their implications. Proc. Inst. Civ. Engrs. Pt. III, Vol. IV, 174-205.

___ (1962) Auto-suspension of transported sediment; turbidity currents. Roy. Soc. Lond. Proc. Ser. A, 265, 315-9.

___ (1968) Deposition in the process of hydraulic transport. Sedimentology, 10, 45-56.

Bajard, J. (1966) Figures et structures dans la zone intertidal de la partie orientale de la baie du Mont-Saint-Michel. Rev. de Geog. Physique et de Geol. Dynamique, 8, 39-111.

Banks, N.L. (1973) The origin and significance of some downcurrent dipping cross-stratified sets. J. sedim. Petrol. 43, 423-27.

Bisat, W.S. (1924) The Carboniferous goniatites of the north of England and their zones. Proc. Yorks. geol. Soc. 20, 40-124.

___ (1928) The Carboniferous goniatite zones of England and their continental equivalents. Congr. de Strat. Carb., Heerlen, p117.

___ and R.G.S. Hudson (1943) The lower Reticuloceras (R_1) goniatite succession in the Namurian of the north of England. Proc. Yorks. geol. Soc. 24, 383-440.

Blatt, H., G.V. Middleton and R. Murray (1972) "Origin of sedimentary rocks", pp634. Prentice Hall, New Jersey.

Bott, M.H.P., and G.A.L. Johnson (1967) The controlling mechanism of Carboniferous cyclic sedimentation. Q. Jl. geol. Soc. Lond. 122, 421-41.

Bouma, A.H. (1962) "Sedimentology of some flysch deposits; a graphic approach to facies interpretation", pp168. Elsevier, Amsterdam.

Brenchley, P.J. (1969) Origin of matrix in Ordovician greywackes, Berwyn Hills, North Wales. J. sedim. Petrol. 39, 1297-301.

Brown, T. (1841) Description of some new species of fossil shells found chiefly in the Vale of Todmorden, Yorkshire. Trans. Manchr geol Soc. 1, 212-29.

Burne, R.V. (1970) The origin and significance of sand volcanoes in the Bude formation (Cornwall). Sedimentology, 15, 211-28.

Burst, J.F. (1965) Subaqueously formed shrinkage cracks in clay. J. sedim. Petrol., 35, 348-53.

Carrigy, M.A. (1966) Lithology of the Athabasca oil sands. Res. Council Alberta Bull., 18, 48pp.

——— (1971) Deltaic sedimentation units in Athabasca Tar Sands. Bull. Am. Ass. Petrol. Geol. 55, 1155-69.

Chillinor, J. (1924) "Farey's Grit". Geol. Mag. 61, 423-7.

Chowdhury, M.I. (1966) On the gradual shifting of the Ganges from west to east in delta-building operations. In: "Scientific Problems of the Humid Tropical zone Deltas and their Implications", U.N.E.S.C.O. 35-40.

Coleman, J.M. (1969) Brahmaputra River: channel processes and sedimentation. Sediment Geol. 3, 129-239.

† ——— and S.M. Gagliano (1965) Sedimentary structures; Mississippi deltaic plain. In: "Primary Sedimentary Structures and their Hydrodynamic Interpretation" (Ed. by G.V. Middleton).

Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa, 12, 34-52.

Coleman, P.J. (1968) Tsunamis as geological agents. J. geol. Soc. Australia, 15, 267-73.

Collinson, J.D. (1966) Antidune bedding in the Namurian of Derbyshire. Geologie Mijnb. 45, 262-4.

—— (1968) Deltaic sedimentation units in the upper Carboniferous of northern England. Sedimentology, 10, 233-54.

—— (1969) The sedimentology of the Grindlow 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. 52, 31-56.

Cornish, V. (1914) "Waves of sand and snow", pp383. Unwin, London

Curray, J.R. (1956) Analysis of two dimensional orientation data. J.Geol. 64, 117-31.

Dill, R.F. (1964) Features in the head of submarine canyons - narrative of underwater film. In:"Deltaic and Shallow Marine Deposits" (Ed. by L.M.J.U. van Straaten). Elsevier, Amsterdam.

Dott, R.H. (1964) Wacke, greywacke and matrix - what approach to immature sandstone classification? J. sedim. Petrol. 34, 625-32.

Dzulynski, S., M. Ksiazkiewicz and P.H. Keunen (1959) Turbidites in flysch of the Polish Carpathian Mountains. Bull. geol. Soc. Am. 70, 1089-118.

Dzulynski, S. and E.K. Walton (1965) "Sedimentary features of flysch and greywackes". pp274. Elsevier, Amsterdam.

Eagar, R.M.C. (1974) Shape of shell of Carbonicola in relation to burrowing. Lethaia, 7, 173-256.

Earp, J.R., D. Magraw, E.G. Poole, D.H. Land and A.J. Whiteman (1961) Geology of the Country around Clitheroe and Nelson. Mem. geol. Surv. Gt. Britain. H.M.S.O., London.

Farey, J. (1811) General view of the agriculture and minerals of Derbyshire. Vol. 1.

Fisher, W.L. (1969) Facies characterization of Gulf Coast Basin delta systems, with some Holocene analogues. Trans. Gulf Cst Ass. geol. Socs. 19, 239-61.

Franks, P.C. (1969) Nature and origin of cone-in-cone structures in the Kiowa Formation (Early Cretaceous), north-central Kansas. J. sedim. Petrol., 39, 1438-54.

Gaunt, G.D. (1960) In: Summ. Prog. geol. Surv. Gt. Britain for 1959. H.M.S.O., London., p32-3.

† Gilligan, A. (1920) The petrography of the Millstone Grit of Yorkshire. Q. Jl. geol. Soc. Lond. 75, 251-94.

Goldring, R. (1971) Shallow-water sedimentation as illustrated in the Upper Devonian Baggy Beds. Mem. geol. Soc. Lond. 5, pp80.

Gould, H.R. (1970) The Mississippi delta complex. In: "Deltaic Sedimentation, Modern and Ancient" (Ed. by J.P. Morgan). Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa, 15, 3-30.

Gradstein, F.M. and A. van Gelder (1971) Prograding clastic fans and transition from a fluviatile to a marine environment in Neogene deposits of eastern Crete. Geologie Mijnb. 50, 383-92.

Gradzinski, R. and T. Jerzykiewicz (1974) Sedimentation of the Barun Goyot Formation. Results of the Polish-Mongolian Palaeontological expeditions. Palaeont. Pol., 30, 111-46.

Guy, H.H., D.B. Simons and E.V. Richardson (1966) Summary of alluvial channel data from flume experiments, 1951-61. Prof. Pap.

U. S. geol. Surv., 462-I, 96pp.

Hand, B.M., J.M. Wessel and M.O. Hayes (1969) Antidunes in the Mount Toby conglomerate (Triassic), Massachusetts. J. sedim. Petrol. 39, 1310-6.

Haner, B.E. (1971) Morphology and sediments of Redondo Submarine Fan, Southern California. Bull. geol. Soc. Am. 82, 2413-32.

Hardy, P.G. (1970) Aspects of palaeoecology in the arenaceous sediments of Upper Carboniferous age in the Manchester. Unpublished Ph.D thesis, University of Manchester.

Harms, J.C. and R.K. Fahnstock (1965) Stratification, bed form 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. Miner., Tulsa, 12, 84-115.

Harms, J.C, D.B. MacKenzie and D.G. McCubbin (1963) Stratification in modern sands of the Red River, Louisiana. J. Geol. 71, 566-80.

Hayes, M.O. (1967) Hurricanes as geological agents: case studies of Hurricanes Carla, 1961, and Cindy, 1963. Texas Bur. Econ. Geol. Rept. Invest. 61, 54pp.

Heezen, B.C. and C. Hollister (1964) Deep-sea current evidence from abyssal sediments. Marine Geol. 1, 141-74.

Hemingway, J.E. (1968) Sedimentology of coal-bearing strata. In: "Coal and coal-bearing strata" (Ed. Murchison, R. and S. Westoll). Oliver and Boyd, Edinburgh. pp43-69.

Heptonstall, W.B. (1964) Taxonomic, ecological and environmental studies of a marine band in the Namurian of northern England and North Wales. Unpublished Ph.D. thesis, University of Manchester.

Hind, W. (1897) Subdivision of the Carboniferous series in Great Britain and the true position of the beds mapped as Yoredale

Series. Geol. Mag. Decade IV, Vol. IV, 91, 159-69 and 205-13.

Holdsworth, B.K. (1963) Pre-fluvial, autogeosynclinal sedimentation in the Namurian of the Central Pennines. Nature, 199, 133-5.

—— Radiolarian nature of the thicker-shelled goniatite faunal phase in some Namurian limestone "bullions". Nature, 201, 697-9.

Holroyd, W.F. and J. Barnes (1896) On the rocks and fossils of the Yoredale Series of the Marsden and Saddleworth Valleys. Trans. Manchr geol Soc. 24, 70-99.

Houbolt, J.J.H.C. (1968) Recent sediments in the southern bight of the North Sea. Geologie Mijnb. 47, 245-73.

Hubert, J.F. (1966) Modification of the model for internal structures in graded beds to include a dune division. Nature, 211, 614-5.

Hull, E. (1864) The geology of the country aroun Oldham. Mem. Geol. Surv.

—— , J.R. Dakyns, R.H. Tiddeman, J.C. Ward, W. Gunn and C.E. de Rance (1875) Geology of the Burnley Coalfield. Mem. Geol. Surv. 225pp.

Imbrie, J. and Buchanan, H. (1965) Sedimentary structures in modern carbonate sands of the Bahamas. In: "Primary Sedimentary Structures and their Hydrodynamic Interpretation" (Ed. by G.V. Middleton) Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa, 12, 149-72.

Jackson, J.W. (1927) The succession below the Kinderscout Grit in north Derbyshire. J. Manchr geol. Assoc. 1, 15-32.

Jerzyiewicz, T. (1968) Sedimentation of the younger sandstones of the Intrasudetic Cretaceous Basin. Geologia Sudetica, IV, 409-62.

Johansson, C.E. (1965) Structural studies in sedimentary deposits. Geologiska Foneningens i Stockholm Forhardlinger, 87, 3-61.

Jopling, A.V. (1963) Hydraulic studies on the origin of bedding. Sedimentology, 2, 115-21.

—— (1965) Hydraulic factors controlling the shape of laminae in laboratory deltas. J. sedim. Petrol., 35, 777-91.

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. Palaeont. Miner., Tulsa, 15, 78-106.

Karcz, I. (1967) Harrow marks, current-aligned sedimentary structures. J. Geol. 75, 113-21.

Kendall, P.F. and H.E. Wroot (1924) "The Geology of Yorkshire" pp995. Leeds.

Kuenen, P.H. (1958) Experiments in Geology. Trans. geol. Soc. Glasgow, 23, 1-28.

Laird, M.G. (1968) Rotational slumps and slump scars in Silurian rocks, western Ireland. Sedimentology, 10, 111-20.

Lajoie, J. (1972) Slump fold axis orientations: an indication of palaeoslope. J. sedim. Petrol. 42, 584-6.

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.

Leopold, L.B. and M.G. Wolman (1957) River channel patterns: braided, meandering and straight. Prof. Pap. U. S. geol. Surv., 282-B, pp85.

Lloyd, W. and J.V. Stephens (1927) The stratigraphical succession below the Kinderscout Grit in the Todmorden district. Proc. Yorks. geol. Soc. 21, 47-58.

McIntyre, L. (1972) Amazon: the river sea. Nat. Geographic. 142, 445-95.

McKee, E.D. (1966) Significance of climbing ripple structure. Prof. Pap. U. S. geol. Surv., 550-D, 94-103.

——, E.J. Crosby and H.L. Berryhill (1967) Flood deposits, Bijou Creek, Colorado, June 1965. J. sedim. Petrol. 37, 829-51.

Marschalko, R. (1964) Sedimentary structures and palaeocurrents in the marginal lithofacies of the Central-Carpathian flysch. In: "Turbidites" (Ed. by A.H. Bouma and A. Brouwer), Elsevier, Amsterdam, 106-126.

Mathews, W.H. and F.P. Shepard (1962) Sedimentation of Fraser River Delta, British Columbia. Bull. Am. Ass. Petrol. Geol. 46, 1416-42.

Middleton, G.V. (1965) Antidune cross-bedding in a large flume. J. sedim. Petrol. 35, 922-7.

___ (1967) 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) Spec. Publ. geol. Ass. Canada, 7, 253-72.

___ and M.A Hampton (1973) Sediment gravity flows: mechanisms of flow and deposition. In: "Turbidites and Deep Water Sedimentation", Soc. econ. Palaeont. Miner. Short course, 1-38.

Mikhailov, V.N. (1966) Hydrology and formation of river-mouth bars. In: "Scientific Problems of the Humid Tropical Zone Deltas and their implications" U.N.E.S.C.O., 59-64.

Morgan, J.P. (1970) Depositional processes and products in the deltaic environment. In: "Deltaic Sedimentation, Modern and Ancient" (Ed. by J.P. Morgan), Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa, 15, 31-47.

Mutti, E. and F. Ricci Lucchi (1972) Le torbiditi dell'Appenino settentrionale: introduzione all' analisi di facies. Mem. Soc. Geol. Italiana, 11, 161-99.

Nelson, B.W. (1970) Hydrography, sediment dispersal, and recent historical development of the Po river delta, Italy. In: "Deltaic Sedimentation, Modern and Ancient" (Ed. by J.P. Morgan) Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa, 15, 152-184.

Nelson, C.H. and V. Kulm (1973) Submarine fans and deep sea channels. In: "Turbidites and Deep Water Sedimentation", Soc. econ. Palaeont. Miner. Short course, 39-78.

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. Palaeont. Miner., Tulsa, 15, 198-212.

____ (1974) Lithofacies relations in the Late Quaternary Niger Delta complex. Sedimentology, 21, 195-222.

Pettijohn, F.J., P.E. Potter and R. Seiver (1972) "Sand and Sandstone", pp618, Spriger Verlag, Heidelberg.

Phillips, J. (1836) "Illustrations of the Geology of Yorkshire, Part 2. The Mountain Limestone District." London.

Picard, M.D. (1973) "Sedimentary Structures of Ephemeral Streams". pp223. Developments in Sedimentology, 17. Elsevier, Amsterdam.

Price, D, W.B. Wright, R.C.B. Jones, L.H. Tonks and T.H. Whitehead (1963) Geology of the country around Preston. Mem. geol. Surv. Gt. Britain. H.M.S.O., London.

Raiswell, R. (1971) The growth of Cambrian and Liassic concretions. Sedimentology, 17, 147-71.

____ (1973) The growth of carbonate concretions and the microbiological diagenesis of Liassic Shales. In: "The Chemistry of Sedimentary and Diagenetic Processes. Synopses of Contributions" Sheffield symposium. Min. Soc. Gt. Britain and Ireland. pp42-45.

Ramsbottom, W.H.C. (1966) A pictorial diagram of the Namurian rocks of the Pennines. Trans. Leeds Geol. Assoc, 7, 181-4.

Ramsbottom, W.H.C. (1969a) Interim report of the Namurian Working Group. C.R. Congr. int. Stratigr. Geol. Carb. (6th Sheffield), 1, 71-7.

___ (1969b) The Namurian of Britain. C. R. Congr. int. Stratigr. Geol. Carb. (6th Sheffield), 4, 1395-400.

___ (1974) Namurian. In: "The Geology and Mineral Resources of Yorkshire", pp405, Yorks. geol. Soc., Leeds.

___, 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.

Reineck, H.-E. (1960) Uber Zietlucken in rezenten Flachsee Sedimenten. Geol. Rundschau, 49, 149-61.

___ (1963) Sedimentgefuge im Bereich der sudlichen Nordsee. Abh. senckenbergische naturforsch. Ges. 505, 138pp.

___, J. Dorjes, S. Gadow and G. Hertweck (1968) Sedimentologie, faunenzonierung un faziesalfolge vor der ostkuste der inneren Deutschen Bucht. Seckenbergiana Lethaea, 49, 261-309.

___ and I.B. Singh (1973) "Depositional Sedimentary Environments" pp439. Springer Verlag, Heidelberg.

___ and F. Wunderlich (1968) Classification and origin of flaser and lenticular bedding. Sedimentology, 11, 99-104.

Rust, B.R. (1972) Structure and process in a braided river. Sedimentology, 18, 221-45.

Schafer, W. (1972) "Ecology and palaeoecology of marine environments". pp538. Oliver and Boyd, Edinburgh.

Schmincke, H.-V., R.V. Fisher and A.C. Waters (1973) Antidune and chute and pool structures in the base surge deposits of the Laaker See area, Germany. Sedimentology, 20, 553-74.

Scruton, P.C. (1956) Oceanography of Mississippi Delta sedimentary environments. Bull. Am. Ass. Petrol. Geol. 40, 2864-952.

___ (1960) Delta building and the deltaic sequence. In: "Recent sediments, northwest Gulf of Mexico."(Ed. by F.P. Shepard, F.B. Phleger and T.H. van Andel). Am. Ass. Petrol. Geolog., Tulsa, 82-102.

Seilacher, A. (1953) Studien zur Palichnologie, II; die fossilen Ruhespuren (Cubichnia). Neves Jb. Geol. u. Palaeontol. 98, 87-124.

___ (1962) Palaeotological studies on turbidite sedimentation and erosion. J. Geol. 70, 227-34.

___ (1964) Biogenic sedimentary structures. In: "Approaches to Palaeoecology" (Ed. by J. Imbrie and N. Newell). Wiley, New York, 296-316.

Sengupta, S. (1966) Studies on orientation and imbrication of pebbles with respect to cross-stratification. J. sedim. Petrol. 36, 362-9.

Shepard, F.P. (1955) Delta-front valleys bordering the Mississippi distributaries. Bull. geol. Soc. Am. 66, 1489-98.

___ (1956) Marginal sediments of the Mississippi Delta. Bull. Am. Ass. Petrol. Geol. 40, 2537-623.

___ (1960) Mississippi Delta: marginal environments, sediments and growth. In: "Recent sediments, northwest Gulf of Mexico" (Ed. by F.P. Shepard, F.B. Phleger and T.H. van Andel). Am. Ass. Petrol. Geol., Tulsa, 56-81.

___ and R.F. Dill (1966) "Submarine canyons and other sea valleys." pp381. Rand McNally, New York.

___, R.F. Dill and U. von Rad (1969) Physiography and sedimentary processes of La Jolla submarine fan and fan-valley, California.

Bull. Am. Ass. Petrol. Geol. 53, 390-420.

Simons, D.B., E.V. Richardson and C.F. Nordin (1965) Sedimentary structures generated by flow in alluvial channels. In: "Primary Sedimentary Structures and their Hydrodynamic Interpretation" (Ed. by G.V. Middleton). Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa, 12, 34-52.

Skipper, K. (1971) Antidune cross-stratification in a turbidite sequence, Cloridorme Formation, Gaspe, Quebec. Sedimentology, 17, 51-68.

Smith, N.D. (1971) Transverse bars and braiding in the Lower Platte River, Nebraska. Bull. geol. Soc. Am. 82, 3407-20.

Sorby, H.C. (1859) On the structure and origin of the Millstone Grit in south Yorkshire. Proc. Yorks. geol. Soc. 3, 669-75.

Spencer, J. (1874) Geology of the parish of Halifax (Part III). Trans. Manchr geol. Soc. 13, 202-25.

——— (1898) Our local fossils and where to find them. Halifax Naturalist, 3, 6.

——— (1899) Yoredale and Millstone Grit rocks of the upper Calder valley. Proc. Yorks geol. Soc. 13, 375-94.

Stauffer, P.H. (1967) Grain-flow deposits and their implications, Santa Ynes Mountains, California. J. sedim. Petrol. 37, 487-508.

Stephens, J.V., W. Edwards, C.J. Stubblefield and G.H. Mitchell (1942) The faunal divisions of the Millstone Grit Series of Rombalds Moor and neighbourhood. Proc. Yorks. geol. Soc. 24, 344-72.

Stephens, J.V., G.H. Mitchell and W. Edwards (1953) The geology of the country between Bradford and Skipton. Mem. geol. Survey Gt. Brit., H.M.S.O., London, 180pp.

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, 180pp.

Straaten, L.M.J. van (1953) Megaripples in the Dutch Wadden Sea and in the Basin of Arcachon (France). Geologie Mijnb. 15, 1-11.

Stride, A.H. (1963) Current-swept sea floors near the southern half of Great Britain. Q. Jl. geol. Soc. Lond. 119, 175-99.

Strom, K.M. (1955) Land-locked waters and the deposition of black muds. In: "Recent Marine Sediments" (Ed. by P.D. Trask). Am. Assoc. Petrol. Geol., Tulsa, 356-72.

Thomson, A.F. and M.R. Thomasson (1969) Shallow to deep water facies development in the Dimple Limestone (Lower Pennsylvanian), Marathon Region, Texas. In: "Depositional environments in Carbonate Rocks" (Ed. by G.M. Friedman). Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa, 14, 57-78.

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. 1, 77-112.

Walker, C.T. (1955) Current bedding directions in sandstones of Lower *Reticuloceras* age in the Millstone Grit of Wharfedale, Yorkshire. Proc. Yorks. geol. Soc. 30, 115-32.

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, 1899-917.

___ (1967a) Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. J. sedim. Petrol. 37, 25-43.

Walker, R.G. (1967b) Upper flow regime bed forms in turbidites of the Hatch Formation, Devonian of New York State. J. sedim. Petrol. 37, 1052-8.

___ (1970) Review of the geometry and facies organization of turbidites and turbidite-bearing basins. In: "Flysch Sedimentology in North America" (Ed. by J. Lajoie) Spec. Publ. geol. Ass. Canada, 7, 219-52.

___ and E. Mutti (1973) Turbidite facies and facies associations. In: "Turbidites and Deep Water Sedimentation", Soc. econ. Palaeont. Miner. Short course, 119-58.

Walton, E.K. (1967) The sequence of internal structures in turbidites. Scott. J. Geol. 3, 306-17.

Welder, F.A. (1959) Processes of deltaic sedimentation in the lower Mississippi River. Tech. Rept. Louisiana State Univ., Coastal Stud. Inst. 12, 90pp.

White, W.A. (1961) Colloid phenomena in sedimentation of argillaceous rocks. J. sedim. Petrol. 31, 560-70.

Whittle, W.L. (1942) The occurence of quartzite boulders in the Stanley Main Coal at Nostell, near Wakefield. Proc. Yorks. geol. Soc. 25, 140.

Wray, D.A. (1929) The Carboniferous succession in the Central Pennine area. Proc. Yorks. geol. Soc. 21, 228-87.

___ , J.V. Stephens, W.N.Edwards and C.E.N. Bromehead (1930) The geology of the country around Huddersfield and Halifax. Mem. geol. Surv. Gt. Brit., H.M.S.O., London.

Wright, L.D. and J.M. Coleman (1973) Variations in morphology of major river deltas as a function of ocean wave and river discharge regimes. Bull. Am. Ass. Petrol. Geol. 57, 370-98.

Wright, M.D. (1964) Cementation and compaction of the Millstone

Grit of the Central Pennines, England. J. Sedim. Petrol. 34, 756-60.

Wright, M.D. (1964b) Cross-bedding in the Millstone Grit of the Central Pennines and its significance. Geol. Mag. 101, 520-30.

—— (1967) Comparison of Namurian sediments of the Central Pennines, England, and recent deltaic deposits. Sediment. Geol. 1, 83-115.

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, 182pp.

† Coleman, J.M., and S.M. Gagliano (1964) Cyclic sedimentation in the Mississippi River deltaic plain. Trans. Gulf Cst Ass. geol. Socs. 14, 67-80.

Gilbert, G.K. (1883) The topographic features of lake shores. U. S. geol. Surv. Report 5, 69-163.