THE EVOLUTION OF A SAND-RICH BASIN-FILL SEQUENCE IN THE PENDLEIAN (NAMURIAN, E_{1c}) OF NORTH-WEST ENGLAND.

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ABSTRACT

The Bowland and Lancaster Fells Basins of north-west England contain a Pendleian E_{1c} sequence up to 650m thick, most of which is coarse clastic material. Each of the major stratigraphic units in this sequence, namely the Pendle and Grassington Grit Groups, represents a phase of clastic input to the basins. Initially, the basins were deep marine troughs with water depths of several hundreds of metres. Their northern margins were steep submarine slopes or relict fault scarps rising up onto the partially emergent Askrigg and Lake District Blocks. Within the basins there was significant topographic relief caused by differential compaction over buried Dinantian fault blocks. The Pendle Grit Group, comprising the Pendle Grit and Pendle Shale Formations, represents the development of a sand-rich submarine fan/slope system in these confined basins. The sediment for the fans was probably supplied by a fluvial source to the north-west. Intra-basinal relief within the basins strongly influenced the initial development of the fan system but was later swamped by sedimentation. Only the Waddington Fell High remained present throughout deposition: the "low-seeking" turbidite sediments of the Pendle Grit Formation pinch-out over this structure.

A period of uplift or eustatic sea-level fall at the end of Pendle Grit deposition resulted in the development of a minor unconformity on the southern margin of the Askrigg Block and across reactivated intra-basinal highs. This unconformity heralds a major change in the palaeogeography of north-west England and the beginning of Grassington Grit Group deposition. The fluvial clastic input moved eastwards, supplying sediment directly into shallow water on the Askrigg Block. This resulted in rapid progradation of a coarse clastic dominated braid-delta system over the Askrigg Block and, subsequently, out into the Bowland Basin. Water depths still increased rapidly across the fault controlled boundary between these two palaeogeographic features and this led to a change in the depositional processes on the braid-delta system. The resulting differences between Grassington Grit Group facies sequences across this palaeogeographic boundary form the basis for recognition of two Formations, namely the Grassington Grit and Warley Wise Formations.

Detailed sedimentological study of the sand-rich fans in the Pendle Grit Group shows that they are dominated by in-channel deposition: lobes and basin

plain deposits are very rare. A channel hierarchy has been recognised in the fan sediments, based on the presence of erosion surfaces of different magnitude and extent. First-order erosion surfaces bound channel-complexes up to 1000m in width and 100m deep. These features were cut by infrequent, high energy turbidity currents. They were filled by progradation of a coarse-grained turbidite sand-body deposited from smaller, more frequent turbidity currents trapped in the first-order channel. During this process, second-order channels were cut and filled in the prograding sand-body. Individual beds within the fan system provide evidence for lateral migration of turbidity currents during deposition and also for prolonged flow events. The flow mechanics of such flows are qualitatively examined and their evolution with time and space over the fan system is discussed. A new facies model for sand-rich fans is presented, based on the sedimentological features seen in the Pendle Grit Formation.

The E_{1c} basin-fill sequence was buried to several kilometres depth by end early Westphalian times. It then underwent rapid uplift. The paragenetic sequences in sandstones from this sequence are related to the maximum burial depth and the amount of subsequent uplift: deeply buried sandstones developed illite cements and, if affected by meteoric flushing during uplift, also have extensive pore-filling kaolinite. This relationship allows qualitative predictions of the reservoir quality of the E_{1c} sandstones to be made from an estimate of their burial histories.

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"Another damned, thick, square book!"

William Henry, Duke of Gloucester (1743-1805)

CHAPTER ONE

PROJECT BACKGROUND, DATABASE AND METHODOLOGY.

1.1 Project rationale and thesis outline.

Carboniferous strata crop out over almost a quarter of the land area of England and are present at depth under another third of the country. These strata provided the bulk of the lime, coal and iron ore which fuelled the Industrial Revolution of the last century and, today, are still the principal source of onshore mineral wealth in Britain. Consequently, the Carboniferous system has always attracted much attention from British geologists, both academic and industrial. Unfortunately, the Namurian sediments of north-west England have been overlooked by modern "dynamic" stratigraphers and sedimentologists, inspite of the significant palaeogeographic and stratigraphic position which they occupy. This project was designed to fill partially the gap in our understanding of the Namurian sediments of north-west England.

The current study covers the sedimentology and stratigraphic relationships of a basin-fill sequence developed in rocks of Pendleian (Namurian, E_{1c}) age. These Pendleian rocks contain the earliest major influx of coarse-grained, feldspathic material into the Carboniferous basins of Northern England. The principal stratigraphic units in this interval, namely the Pendle and Grassington Grit Groups, are dominated by such coarse-grained clastics and form a shallowing upwards sequence from deep marine, mass-flow deposits through fan-delta sediments to delta-top coals. The total thickness of the basin-fill sequence exceeds 600m, contrasting with a maximum thickness of 90m for time equivalent shelf sediments (Yoredales) on the adjacent Askrigg Block, a Carboniferous palaeogeographic high. An understanding of the sedimentary and tectonic history of the Askrigg Block can be used to constrain some of the variables (e.g. eustatic sea-level variations) which influenced the depositional environments within the basinal setting. The Pendle and Grassington Grit Groups form the only exposed example of a Namurian basin-fill sequence for which there is such external control available. It is for this reason that that the sequence has such palaeogeographic significance. On a different but equally important level, the Pendle Grit Group provides a unique opportunity to study the sedimentology of a coarse-grained submarine fan sequence at outcrop. Such systems have received scant coverage in the sedimentological literature, although sand-rich fans are increasingly important reservoir targets in hydrocarbon provinces worldwide. Attempts to explain the deposition of sand-rich mass-flow deposits in terms of classical submarine fans have had only limited success in predicting sandbody geometries and reservoir properties. This work erects a new conceptual model for sand-rich fans based on a detailed sedimentological study of the Pendle Grit Group. Hopefully, this model will help in understanding the sandbody architecture within sand-rich fan systems which, in turn, should prove of direct economic value in hydrocarbon exploration and production.

As the project attempts to cover the stratigraphic development of the complete E_{1c} basin-fill sequence as well as the detailed sedimentology of the Pendle Grit Group, there are problems in documenting the results in a concise fashion. The conceptual model for sand-rich fans developed herein is critically dependent on an understanding of the palaeogeography and structural setting of the basin in which the Pendle Grit Group was deposited. Similarly, to describe the dynamic stratigraphy of the E_{1c} basin-fill sequence requires a knowledge of the processes of sand-rich fan growth. This thesis has been written from a stratigraphic viewpoint as this seemed to allow for a more logical presentation of the work. Consequently, Chapter 2 begins by reviewing the E_{1c} stratigraphy of north-west England and formally defines the stratigraphic nomenclature to be used in the rest of the work. Chapters 3 and 4 cover the sedimentology of the Pendle and Grassington Grit Groups respectively while Chapter 5 discusses the provenance and diagenesis of the coarse clastic sediments in the basin-fill sequence. The conceptual model for sand-rich fans is developed, somewhat out of context, in Chapter 3. Chapter 6 then considers the tectonic and palaeogeographic controls on sedimentation in the depositional basin and, together with the data and conceptual models from previous chapters, documents the "dynamic" stratigraphic evolution of the E_{1c} basin-fill sequence. Unfortunately, this treatment somewhat under-emphasises the implications of the sand-rich model which is probably the main contribution of this thesis.

1.2 Geographical and Geological Location of the Study Area.

The study area covers a large triangular tract of ground the corners of which are defined by Harrogate in the east, Preston in the west and Lancaster in the north (Figure 1.1). Most of this ground is underlain by Carboniferous strata of Dinantian and Namurian age (Figure 1.2). The position of the E_{1c} sequence with respect to the overall stratigraphic column is shown in Figure 1.3. The two principal stratigraphic units within the E_{1c} sequence, the Pendle and Grassington Grit Groups, crop out in broad north-east to south-west trending belts in the Lancaster Fells, around Lancaster itself and along the Pendle Escarpment. Sandstone bodies within these Groups form strong topographic ridges; reaching a maximum height of 557m on Pendle Hill. This higher ground is of poor quality and is given over to upland pasture, forestry and grouse moor. Further north around Lancaster, and to the east in the Harrogate area, the ground is lower lying and drift covered. In these areas the land is mainly used for dairy pasture.

The north-east to south-west trend of the Pendleian outcrops corresponds to the tectonic grain of the ground (Figure 1.4). The study area lies within a tectonic province characterised by a series of *en échelon* periclines, some with reverse faulted limbs. This province is known as the Ribblesdale Foldbelt (Phillips, 1836) and formed during latest Carboniferous and early Permian times, although individual periclines began to grow much earlier (Hudson, 1936; Hudson and Turner, 1933). The foldbelt is bounded to the north by Lower Palaeozoic basement blocks with a thin, relatively undeformed cover sequence of Carboniferous age. The boundary between these two tectonic provinces lies along major fault zones, particularly the Craven Fault System (Figure 1.4).

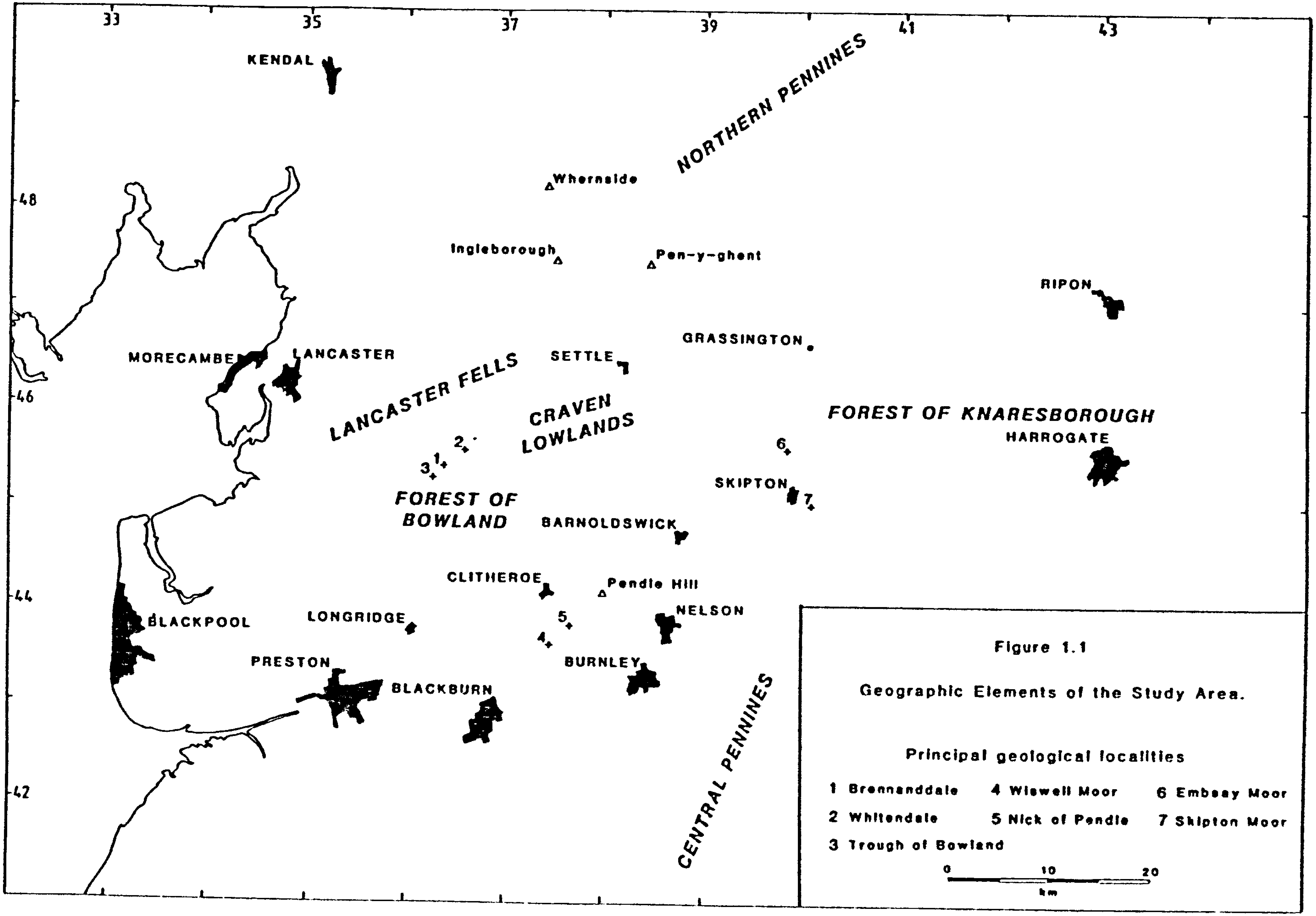


Figure 1.1

Geographic Elements of the Study Area.

Principal geological localities

1 Brennanddale
2 Whitendale
3 Trough of Bowland
4 Wiswell Moor
5 Nick of Pendle
6 Embsay Moor
7 Skipton Moor

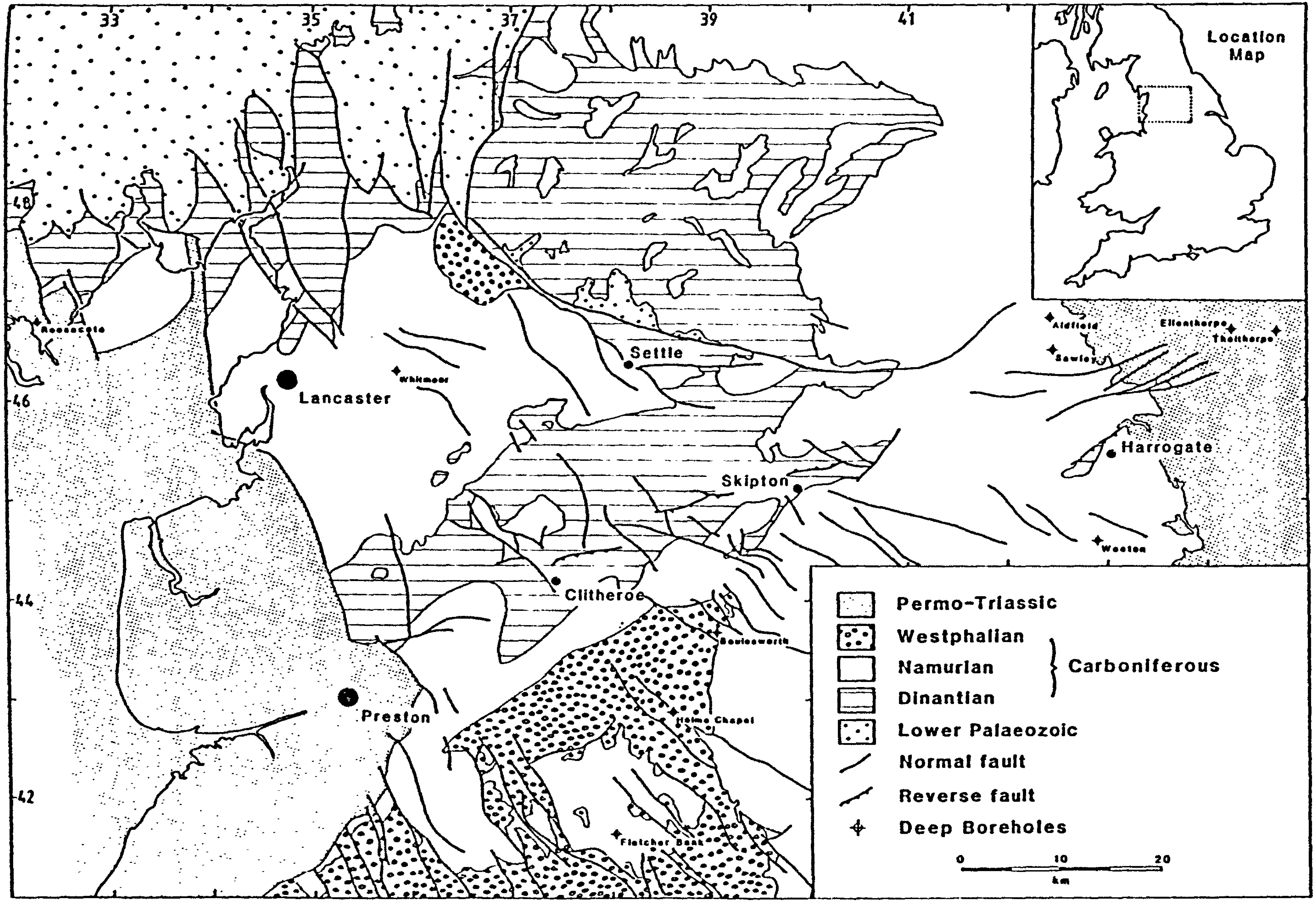


Figure 1.2: General Geological Map of north-western England

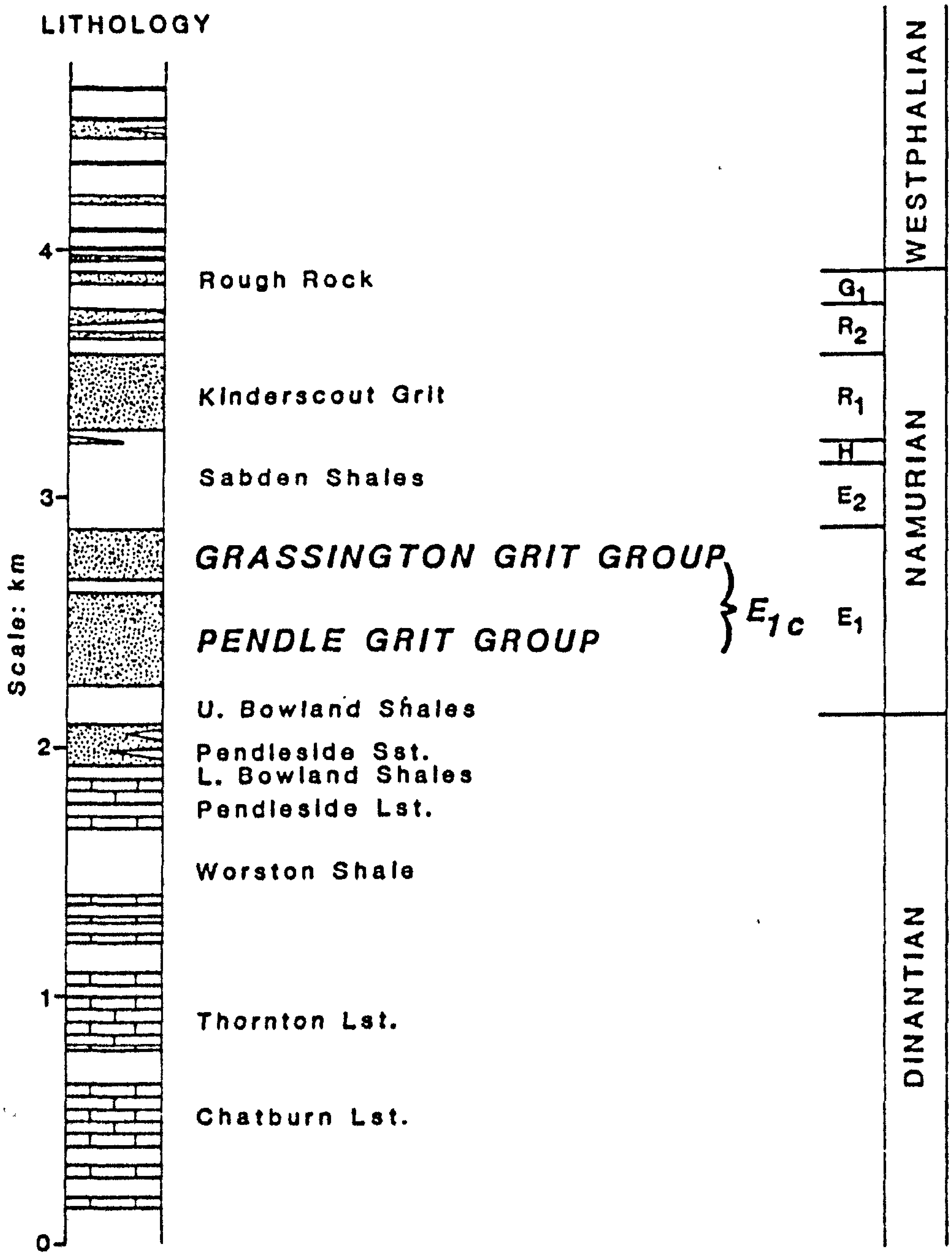


Figure 1.3: Summary section of the Carboniferous strata along the Pendle Monocline (c.f. Figure 1.4) showing the stratigraphic position of the Pendleian (E_{1c}) basin-fill sequence which is the subject of this work. Note that the E_{1c} section lies at the boundary between carbonate dominated Dinantian sediments and later clastic dominated Namurian sediments.
Dinantian section based on Gawthorpe (1985), Namurian section after Earp *et al.* (1961). Goniatite indexes, shown only for the Namurian, are as follows: E_1 Pendleian; E_2 Arnsbergian; H Chokerian and Alportian; R_1 Kinderscoutian; R_2 Marsdenian; G_1 Yeadonian.

Similarly, the southern boundary of the Ribblesdale Foldbelt lies along a major structural lineament, in this case the Pendle Monocline (Figure 1.4).

It has long been recognised that the tectonic provinces defined above coincide with Carboniferous palaeogeographic elements. The Ribblesdale Foldbelt formed by tectonic inversion of the Carboniferous basins within which the E_{1c} sediments covered in this thesis were deposited. These basins are known as the Bowland and Lancaster Fells Basins (Figure 1.5). The undeformed northern tectonic province corresponds to palaeogeographic highs known as the Lake District Massif and the Askrigg Block (Figure 1.5).

1.3 Geological database.

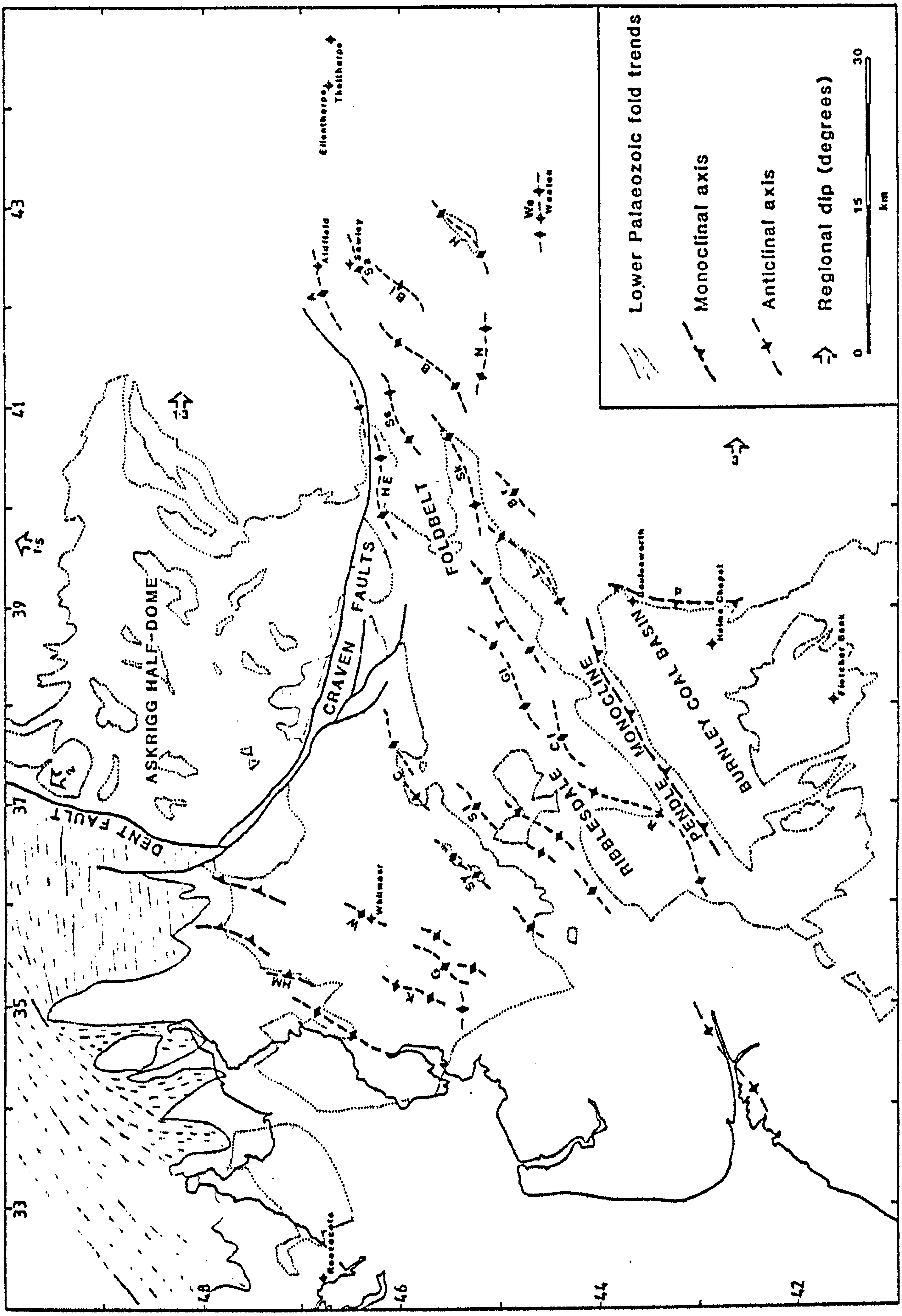
The chief source of data on the distribution and thickness of the various E_{1c} stratigraphic units is from British Geological Survey (B.G.S.) small-scale geological maps (Figure 1.6). Maps older than 1924, which are all too common in the area, were compiled without the benefit of any biostratigraphic control. Consequently, the sandstone correlations are dubious in many cases. Major revision of the maps can be expected following the completion of the resurvey being undertaken by the B.G.S.. More recent maps, such as the Leeds and Bradford sheets, use goniatite bands to constrain the stratigraphic correlations and are much more reliable. They do, however, suffer from miscorrelations brought about by the failure to recognise that some bed cut offs are stratigraphic (e.g. channel margins) rather than structural (i.e. faults). Attempts to match beds across such "faults" has led both to a proliferation of faults and to some unnecessary complications in the stratigraphic nomenclature. Only the Harrogate Sheet has been compiled with the help of both the goniatite stratigraphy and recent palynological and micropalaeontological dating techniques.

In addition to the 1:50,000 and One Inch maps, Six Inch County Series and 1:10,000 scale maps have been consulted. In general, these offer little extra data

Figure 1.4: Major structural elements of north-western England. Fault trends have been omited for clarity but are shown on Figure 1.2. Note that the Ribblesdale Foldbelt is bounded by the Pendle Monocline, the Craven Faults and the Lake District Lower Palaeozoic outcrops. Anticlinal folds within this foldbelt are arranged *en échelon* and are periclinal in form. Interlimb angles range from open to tight with reverse faults developing in some of the tighter folds (e.g. Harrogate, Skipton and Clitheroe anticlines). Broad synclines separate the anticlines. The Pendle and Pennine monoclines have limb dips of 40-60° and bound the major syncline which forms the Burnley Coal Basin.
Data from Moseley (1972), Lawrence *et al.* (1987) and B.G.S. maps.

A	Aldfield Anticline	B	Beamsley Anticline
Bi	Birstwith Anticline	Br	Bradley Anticline
C	Catlow Anticline	Cl	Clitheroe Anticline
G	Grizedale Anticline	Gi	Gisburn Anticline
HM	Hutton Monocline	H	Harrogate Anticline
HE	Hetton-Eshton Anticline	K	Knots Anticline
L	Lothersdale Anticline	N	Norwood Anticline
P	Pennine Monocline	R	Ribble Anticline
Sk	Skipton Anticline	Sl	Slaidburn Anticline
Ss	Simonseat Anticline	Sy	Sykes Anticline
T	Thornton Anticline	W	Whitmoor Anticline
WE	Weeton Anticline		

In this and future figures the limits of the Namurian outcrop are shown as a dotted line. and, hence, are time equivalents.


Lower Palaeozoic fold trends
Monoclinal axis
Anticlinal axis
Regional dip (degrees)
0
15
30
km
ASKRIGG HALF-DOME
DENT FAULT
CRAVEN FAULTS
FOLDBELT
RIBBLESDALE
PENDLE MONOCLINE
BURNLEY COAL BASIN
33
35
37
39
41
43
42
44
46
48

but do allow the calculation of more accurate thicknesses. These maps are held at B.G.S. offices in Keyworth and Newcastle. On a smaller scale the B.G.S. 1:250,000 maps Humber-Trent (53N02W), Tyne-Tees (54N02W), Liverpool Bay (53N03W) and Lake District (54N03W) combine to produce a regional map for the study area which is ideal for the investigation of regional structural trends and stratigraphic relationships.

A substantial amount of stratigraphic data covering the study area has been published over the years. Recent B.G.S. memoirs are available for the Settle, Preston, Bradford, Leeds and Clitheroe sheets (Figure 1.6) while many papers covering the geology of small parts of the area have been published in academic journals. This literature, mostly pre-1960 in age, provides one of the major sources of data for this thesis and is reviewed in Chapter 2 herein.

Recent renewed interest in onshore exploration in this country has led to the award of Exploration Licences (EXL's) covering most of the study area (Figure 1.7). Significant amounts of seismic data have been acquired on these licences although little is currently in the public domain. Two seismic lines have been published from the Lancashire Fells Basin (Lawrence *et al.*, 1987) and five lines over the southern margin of the Bowland Basin have been reproduced as line drawings by Lee (1988a). The current author has also had access to unreleased data from the Harrogate area, subject to its confidential status being maintained. Other geophysical data available include B.G.S. compilations of gravity and aeromagnetic surveys over the whole of the study area. Most of this data set has been interpreted by Lee (1988a, 1988b) and by Gawthorpe (1987). The work of these authors provides much information about the detailed basement structure and early Carboniferous development of the Bowland Basin. Their interpretations have been extended into the Lancaster Fells Basin herein.

Although hydrocarbon and B.G.S. boreholes are not yet common in this area, those that are available provide data which is vital to the arguments presented in this work. Whitmoor-1, Roosecote-1 and Weeton-1 (Figure 1.2) are the most

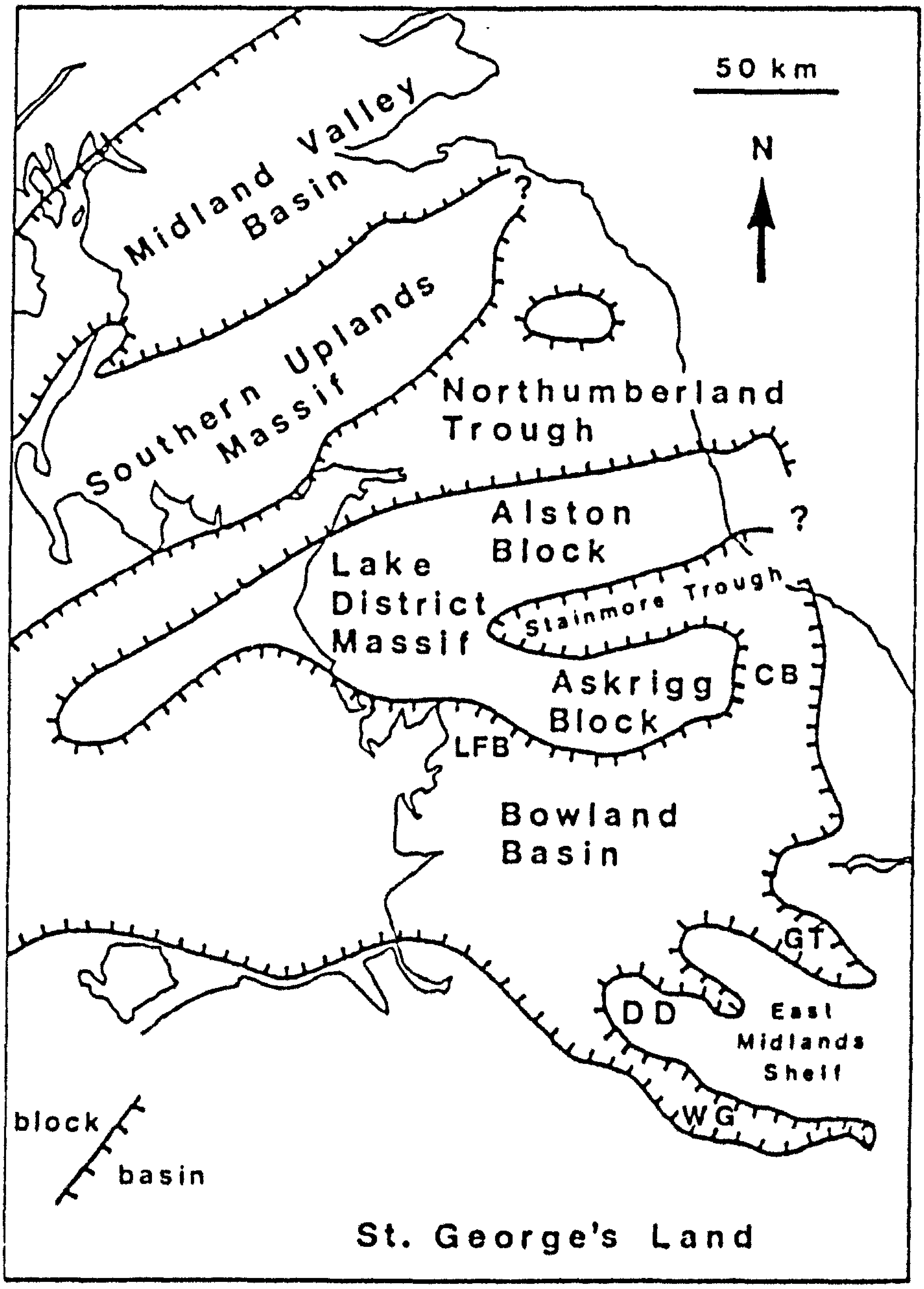
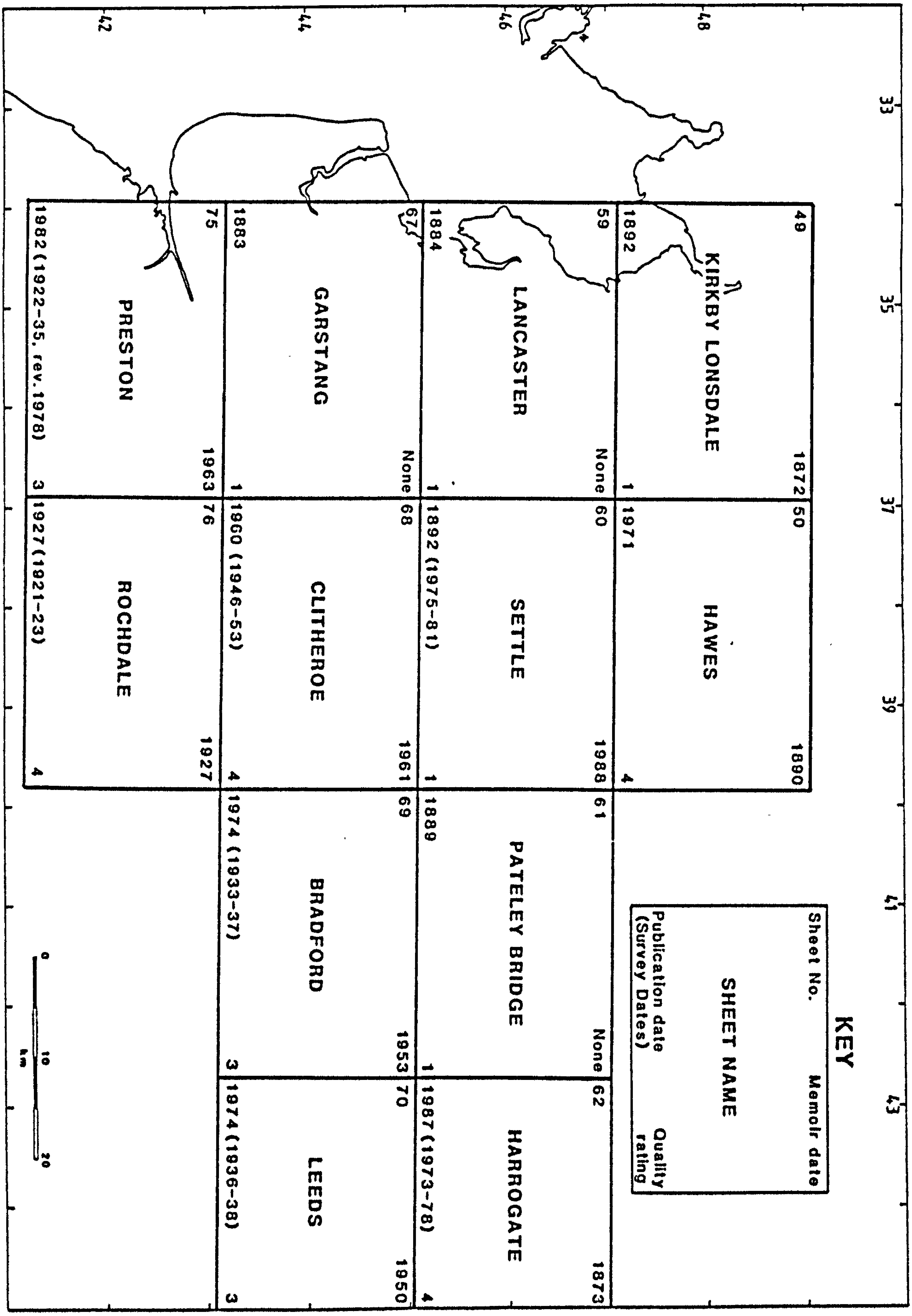


Figure 1.5: Lower Carboniferous Blocks and Basins in Britain. Blocks and Basins first developed in late Devonian or early Carboniferous times and were significant palaeogeographic features throughout the Dinantian. During the Namurian these features became less distinct because of basin infilling by coarse clastics. The distribution of the Blocks shown here is that during Dinantian times. The Alston and Askrigg Blocks together form the structural "Rigid Block" of Marr (1921). Note the correspondence between the palaeogeographic elements and the present day structural provinces shown on Figure 1.4. After Gawthorpe, 1985.

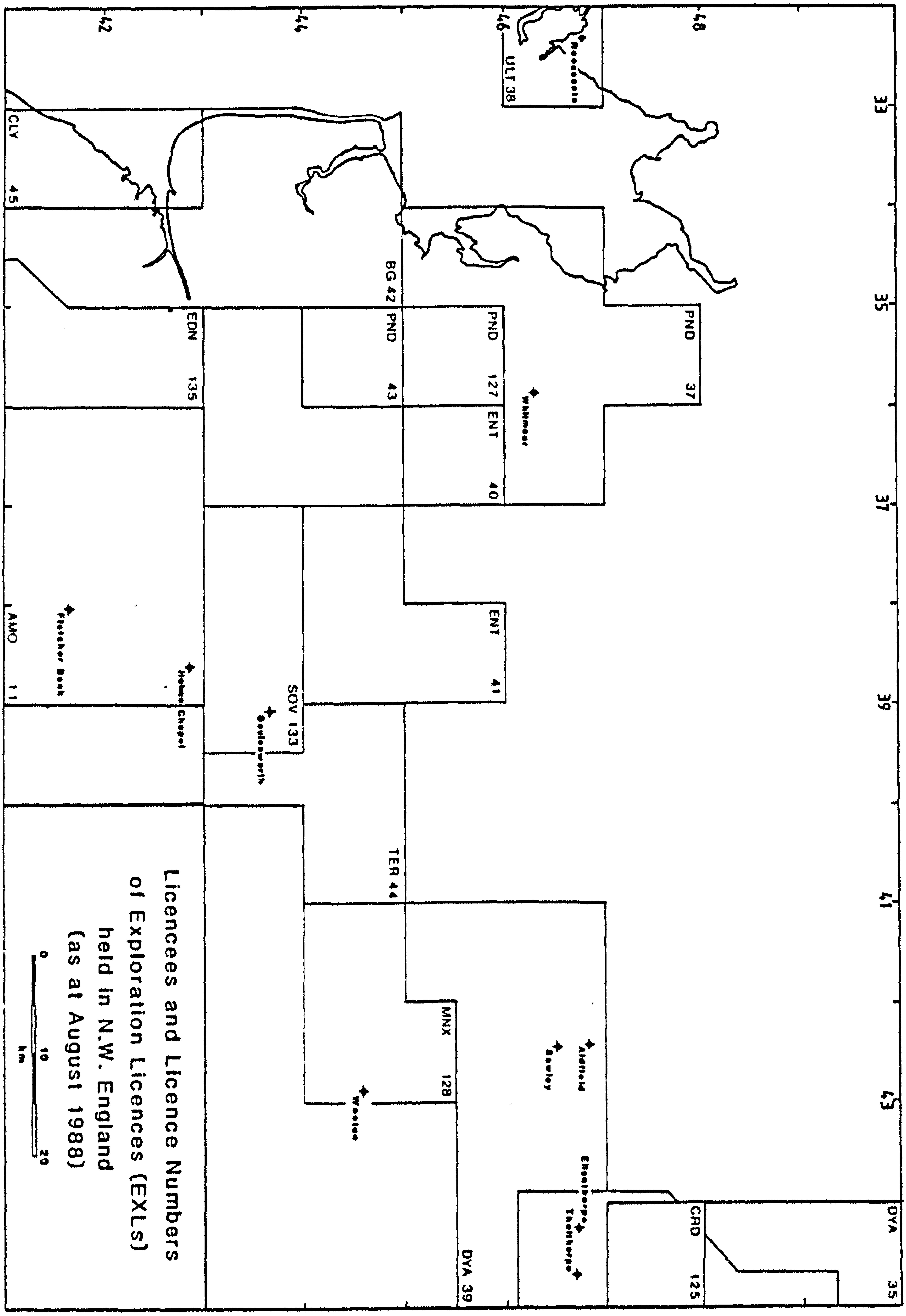
DD Derbyshire Dome
GT Gainsborough Trough
WG Widmerpool Gulf
CB Cleveland Basin
LFB Lancaster Fells Basin


KEY
Sheet No.
Memoir date
SHEET NAME
Publication date (Survey Dates)
Quality rating
49
KIRKBY LONSDALE
1872
1892
1
50
HAWES
1890
1971
4
59
LANCASTER
None
1884
1
60
SETTLE
1988
1892 (1975-81)
1
61
PATELEY BRIDGE
None
1889
1
62
HARROGATE
1873
1987 (1973-78)
4
67
GARSTANG
None
1883
1
68
CLITHEROE
1961
1960 (1946-53)
4
69
BRADFORD
1953
1974 (1933-37)
3
70
LEEDS
1950
1974 (1936-38)
3
75
PRESTON
1963
1982 (1922-35, rev. 1978)
3
76
ROCHDALE
1927
1927 (1921-23)
4
33
35
37
39
41
43
42
44
46
48
0
10
20
km

important boreholes within the area. Data from the latter have been used in the development of palaeogeographic models but details of the well remain confidential and are not reported. Cores, cuttings samples and electric logs from the former wells, however, are in the public domain and have been extensivly used during this research. Other recent boreholes outside the main area of interest include Boulesworth-1, Holme Chapel-1,and Heywood-1, the logs from which have been used to extend data coverage for the Pendle and Grassington Grit Groups south towards to their depositional limits. Many of the older hydrocarbon boreholes from the areas to the east of Harrogate (e.g. Sawley and Aldfield, Figure 1.2) have been summarised by Falcon and Kent (1960) and other authors. These boreholes provide some control along the eastern margin of the study area.

The bulk of the new data reported in this thesis was acquired during two fieldwork seasons, the first spent working along the Pendle Monocline and the second working in the Lancaster Fells and Lancaster district. Exposure in both areas is sparse due to extensive drift cover on the lower ground and to hill peat on the higher ground. Consequently, good sections are confined to streams, slump scars and to old quarry workings. This raises a problem common to many stratigraphic studies, namely that it is possible that exposure, and hence data, is strongly biased towards the sandstone rich parts of the stratigraphic sequence. This problem is discussed in Chapter 3 herein. The lack of exposure becomes acute in the Harrogate area where there is only one section of any real value in an area of some 55km^2.

Figure 1.6: Small scale B.G.S. maps and Memoirs covering the study area. The quality rating is a subjective measure based on the age of the map, the biostratigraphic control available during the mapping and the nature of the outcrop. The rating refers only to the mapping of the Pendleian strata on the maps.


Licencees and Licence Numbers
of Exploration Licences (EXLs)
held in N.W. England
(as at August 1988)
ULT 38
CLY 45
BG 42
EDN 135
PND 43
PND 127
PND 37
ENT 40
ENT 41
AMO 11
SOV 133
TER 44
MNX 128
DYA 39
CRD 125
DYA 35
Fletcher Bank
Holme Chapel
Aldfield
Sawley
Weeton
0 10 20
km

1.4 Methodology

The initial stages of the project involved a synthesis of the extensive stratigraphic data for the study area and the erection of a consistent stratigraphy and nomenclature (Chapter 2). Lithological logging techniques were then used to analyse good sections in the various formations, leading to the recognition of the facies and facies associations present. Where exposure allowed, field mapping was used to relate the facies associations to each other both in a lateral and vertical sense. In areas of poor exposure facies associations could be partially defined by feature mapping and aerial photograph interpretation. The regional and stratigraphic distribution of the facies associations was mapped and related to tectonic models for the basin and to the palaeocurrents measured in the field. Process models were then developed for the various depositional systems leading to an understanding of the "dynamic stratigraphic" evolution of the E_{1c} basin-fill sequence.

The diagenetic and provenance investigations in this work were carried out on samples collected during fieldwork and from the Roosecote-1 and Whitmoor-1 boreholes. Standard petrographic techniques were used to derive

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Figure 1.7: Exploration Licences in north-west England. Most of the study area is licensed and is being actively explored. The map shows the position of deep boreholes from which data were available to the project. Limited seismic data are publically available in EXL 127 (Lawrence, *et al.*, 1987) and in EXL 11 (Lee, 1988a). The abbreviations used for the Licencees are:

AMO	Amoco	BG	British Gas
CLY	Clyde	CRD	Concorde
DYA	Dyas	EDN	Edinburgh Oil and Gas
ENT	Enterprise	MNX	Marinex
PND	Pendle Petroleum	SOV	Sovereign Oil & Gas
TER	Teredo Oils	ULT	Ultramar

paragenetic sequences for the rocks and to relate these to their burial histories. The pore-filling cements and their chemistries were investigated using Energy Dispersive X-ray Analysis (EDAX) on scanning electron microscopes. X-ray diffraction (XRD) and electron microprobe analysis (Wavelength Dispersive X-ray Analysis: WDAX) aided in the investigation of the smaller sized the clay mineral cements. No stable isotope work was possible within the timeframe of the project.

1.5 Conventions used.

Throughout this work grid references are based on the National Grid system. In the text 100km grid square references are given, together with the appropriate grid square prefix. On the illustrations National Grid eastings and northings are labelled with the first two numbers of the full grid coordinates. Hence references prefixed SD in the text lie within the grid square with coordinates $^{3}XXX^{xxx}$ $^{4}XXX^{xxx}$ while those prefixed SE lie within square $^{4}XXX^{xxx}$ $^{4}XXX^{xxx}$.

All depths and thicknesses are given in metres unless otherwise specified. Grainsize grades conform with the Wentworth scale while bed thickness descriptors are as follows (Ingram, 1954):

laminae	less than 1cm.
very thin beds	1-3cm.
thin beds	3-10cm.
medium beds	10-30cm.
thick beds	30-100cm.
very thick beds	over 1m.

CHAPTER 2

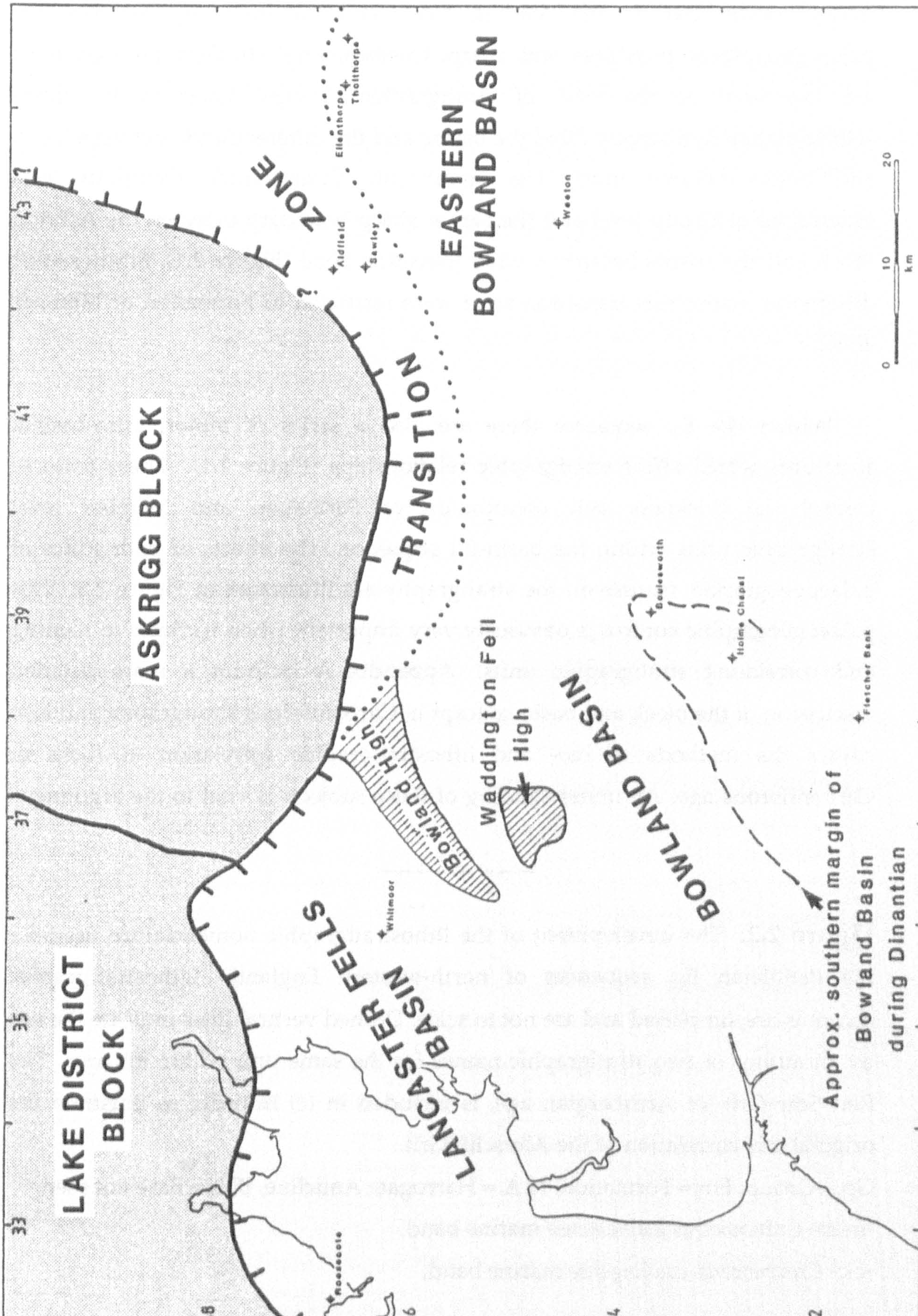
STRATIGRAPHY AND STRATIGRAPHIC RELATIONSHIPS OF THE E_{1c} BASIN-FILL SEQUENCE.

2.1 Introduction.

Carboniferous stratigraphy in the British Isles is strongly influenced by the presence of a series of major palaeogeographic blocks and basins (c.f. Figure 1.5). The basins, probably initiated in Late Devonian times, were infilled by great thicknesses of sediments whilst the blocks subsided more slowly and accumulated thinner sequences. In north-west England the bathymetric relief between the blocks and basins was maintained throughout the Dinantian by differential tectonic subsidence (Gawthorpe, 1987). However, during the Namurian a reduction in tectonic activity and the influx of Millstone Grit clastic sediments resulted in rapid filling of the basins. The Bowland and Lancaster Fells Basins (Figure 2.1) were the first to be affected by this new clastic source, the sediment arriving in the basins during the *Cravenoceras malhamense* zone of the Pendleian (Namurian, E_{1c}). This chapter discusses the stratigraphy and stratigraphic relationships of the resulting E_{1c} basin-fill sequence.

The zone of *Cravenoceras malhamense* (E_{1c}) includes all strata between the *Cravenoceras malhamense* marine band and the succeeding *Cravenoceras cowlingense* band (Figure 2.2). Simplistically, the basinal succession in this zone consists of two thick sandstones overlying a shale sequence (Figure 2.2a). Again simplistically, the much thinner succession on the adjacent Askrigg Block (Figure 2.1) consists of a series of limestone/shale/sandstone cyclothems overlain by a pebbly grit (Figure 2.2b). The major differences between the thicknesses and facies of these successions highlights the influence of the palaeogeography on the E_{1c} stratigraphy in north-western England. At the beginning of E_{1c} times,

Figure 2.1: Principal early Namurian palaeogeographical elements within north-west England. Compare with the tectonic elements map (Figure 1.4). Note that the palaeogeographic basins coincide with the Ribblesdale Foldbelt, some anticlinal axes being related to palaeogeographic intra-basinal highs. The southern margin of the Askrigg Block lies along the present day Craven Fault Zone. During the Carboniferous the latter formed a transition zone across which the stratigraphy changed from typical "block" facies to typical "basin" facies. The nature of the transition zone varied over time (see text). The southern extent of the Bowland Basin during the Namurian is unclear. During the Dinantian the margin lay close to the line of the Pendle Monocline (Gawthorpe, 1987) and, although this line was overlapped by Namurian sediments, it probably had some influence on Namurian (particularly Pendleian) facies distributions.


LAKE DISTRICT BLOCK
ASKRIGG BLOCK
TRANSITION ZONE
EASTERN BOWLAND BASIN
LANCASTER FELLS BASIN
BOWLAND BASIN
Bowland High
Waddington Fell High
Approx. southern margin of Bowland Basin during Dinantian
Roosecote
Whitmoor
Aldfield
Sawley
Ellenthorpe
Tholthorpe
Weeton
Boulsworth
Holme Chapel
Fletcher Bank
33
35
37
39
41
43
48
46
44
42
0
10
20
km

sedimentation in the Bowland and Lancaster Fells Basins was completely isolated from that on the Askrigg Block and the boundary between the palaeogeographic provinces was sharp. Consequently, stratigraphic variations are expressed at the level of stratigraphic Groups. Later in E_{1c} times sedimentation had largely filled the basins and the differentiation between block and basins became much less significant. Stratigraphic continuity was established at Group level and the earlier sharp boundary between the Askrigg Block and the basins became a wide transition zone (Figure 2.1). Stratigraphic differences across this transition zone were restricted to Formation or Member level.

Within the E_{1c} sequence there are also a series of minor, intra-basinal structures which affect stratigraphic relationships (Figure 2.1). These features control the thickness and distribution of Formation and Member level stratigraphic units within the basin-fill sequence. The effects of both scales of palaeogeographic feature on the stratigraphy are illustrated in Figure 2.3. This palaeogeographic control is obviously very important when it comes to naming and correlating stratigraphic units. Appendix A includes a more detailed discussion of the block and basin concept in Carboniferous stratigraphy and also covers the methods of bio- and lithostratigraphic correlation in rocks of Carboniferous age. An understanding of these subjects is vital to the arguments

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Figure 2.2: The development of the lithostratigraphic nomenclature used for the Pendleian E_{1c} sequences of north-western England. Lithostratigraphic columns are simplified and are not to scale. Dashed vertical lines indicate the use by an author of two stratigraphic names for the same unit within his area. The Red Scar Grit, of Arnsbergian age, is included in (c) in order to illustrate the original mis-correlation of the Almscliff Grit.

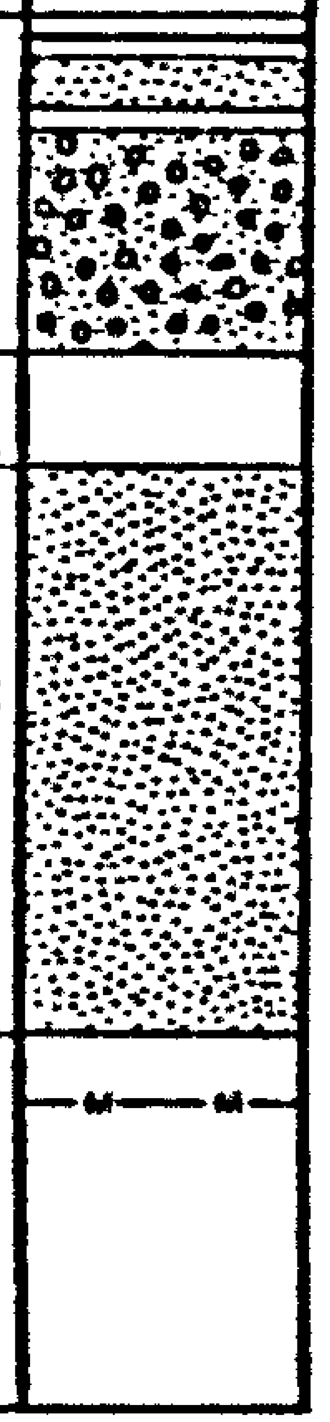
Gp.= Group, Fm.= Formation, H.A.= Harrogate Anticline, BNS= Base not seen.

-m-m- *Cravenoceras malhamense* marine band.

-c-c- *Cravenoceras cowlingense* marine band.

a) Bowland Basin

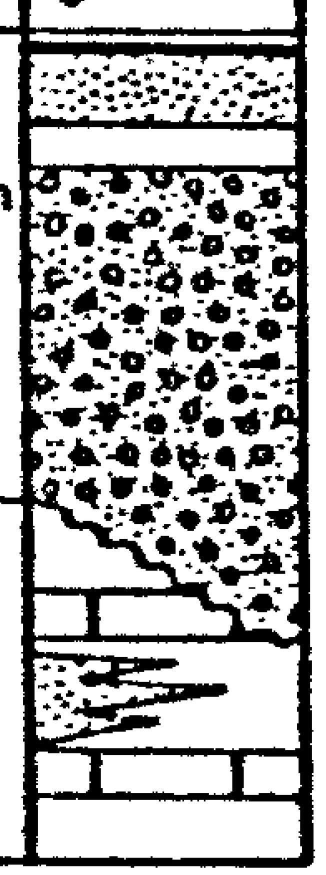
This Work		Lithology
Grass. Grit Gp.	Warley Wise Fm.	
Pendle Grit Gp.	Pendle Shale Fm.	
	Pendle Grit Fm.	
Bowland Sh Gp. (parts)	Upper Bowland Shales	



PHILLIPS 1836	DAKYNS et al. 1879	WRIGHT et al. 1927		BRAY 1927	STEPHENS et al. 1953		EARP et al. 1961	BAINES 1977
Grits of Pendle Hill	Kinderscout Grit	Pendle Grit Group	Upper Wilpshire Grit	Warley Wise Grit	Skipton Moor Grit Group	Lindley Moor Grit Group	Warley Wise Grit	Grassington Grit
				Surgill Shales				Pendle Shale Fm
	Pendle Grit		Lower Wilpshire Grit	Pendle Top Grit			Pendle Grit	Pendle Grit Fm.
	Bowland Shale	Bowland Shales		Bowland Shales	Upper Bowland Shales		Upper Bowland Shales	Upper Bowland Shales

b) Askrigg Block

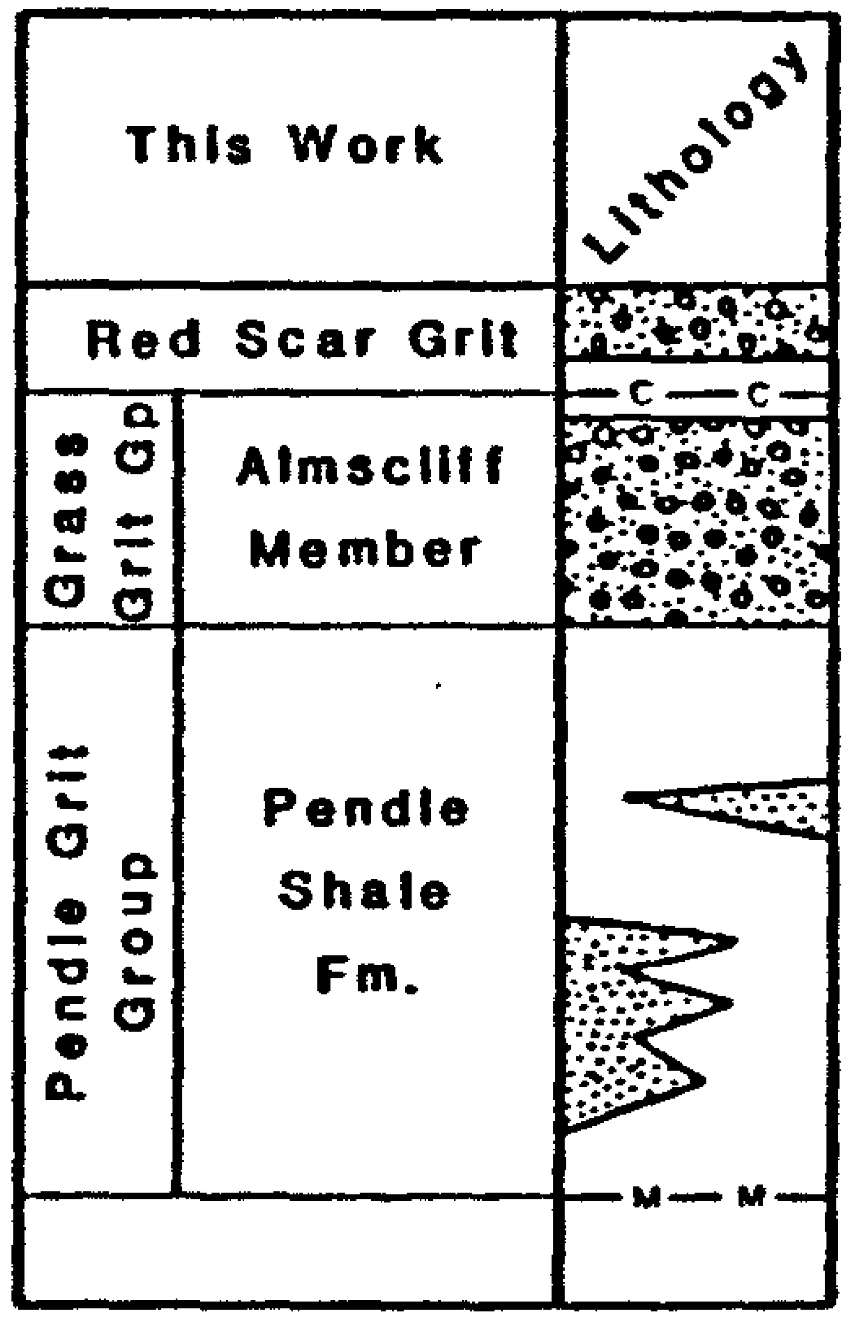
This Work		Lithology
Grassington Grit Gp.	Grassington Grit Fm.	
Wensleydale Group (parts)		



PHILLIPS 1836 West	PHILLIPS 1836 East	DAKYNS et al. 1891	DAKYNS 1892	BLACK 1950	DUNHAM & WILSON 1985
Ingleborough Grit		Howgate Edge Grits	Grassington Grit	Grassington Grit	Grassington Grit Fm.
	Bearing Grit				
Yoredales	Yoredales	Yoredales	Yoredales	Yoredales	

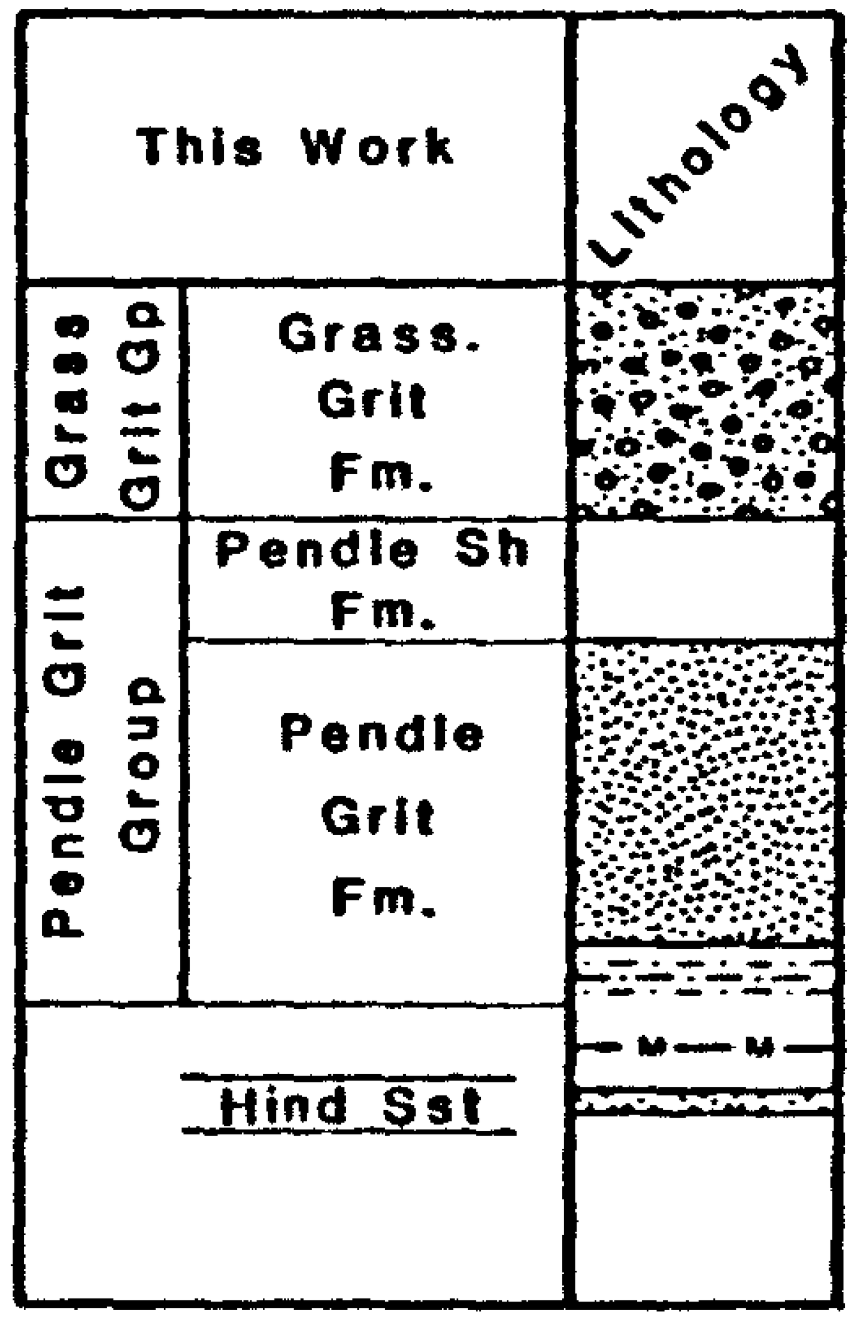
For lithology Key see Encl. 1

c) Eastern Bowland Basin/Transistion Zone



PHILLIPS 1865 North H.A.	FOX-STRANGEWAY 1873	HUDSON 1934		HUDSON 1939	JONES 1943
Almscliff Grit	Kinderscout Grit	Harrogate Grit Gp.	Almscliff Gt		
Harrogate Tunnel Grit			Harrogate Tunnel Grit	Pockstones Grit BNS	Beamsley Grit
		Bowland Shale			Beamsley Shales
					Deerstones Grit
					Upper Bowland Shale

d) Lancaster Fells



PARKINSON 1936	MOSELEY 1954	ARTHURTON et al. 1988
Grassington/ Ingleborough Grit	Brennand Grit	Grassington (Brennand) Grit
		Pendle Grit Formation
Pendle Top Grit	Pendle Top Grit	
Upper Bowland Shale	Upper Newton Fells Grit Bowland Shale	Upper Bowland Shale

For lithology Key see Encl. 1

presented throughout this work.

As a consequence of the large area over which the E_{1c} basin-fill sequence crops out, individual workers have rarely covered sufficient ground to appreciate fully the continuity of basinal stratigraphic units. Workers in different areas usually followed the general practise in Carboniferous stratigraphy of naming only the sandstone and limestone units. This, coupled with the complicated lateral variations in the stratigraphy of the Askrigg Block transition zone, has led to a proliferation in the number of informal stratigraphic names in the literature. Equally, the lack of biostratigraphic control prior to 1924 (See Appendix A) has sometimes resulted in the same name being applied to two stratigraphically distinct units. In this Chapter the stratigraphic nomenclature of the E_{1c} interval is formally redefined based on an understanding of the sedimentological relationships of the various units and of the palaeogeographic controls discussed above.

2.2 Previous stratigraphic research

Phillips (1836) was the first to document detailed stratigraphic sections in the northern Pennines. He recognised that pebbly grits at the top of the E_{1c} succession on the Askrigg Block were different to the underlying Yoredale cycles and consequently referred the unit to the Millstone Grit. He named the unit the Ingleborough Grit and was able to correlate it across the Block-Basin transition from Ingleborough to Bolton Abbey. Later, he traced the unit eastwards into the ore-rich "Bearing Grit" of the Greenhow area and then into the Almscliff Grit north of Harrogate (Phillips, 1865. Figure 2.2b,c). However, Phillips failed to recognise that there was an additional sandstone below the equivalent of the Ingleborough Grit in the Bowland Basin. Consequently, he suggested that the Ingleborough Grit was equivalent to both sandstones seen at Pendle Hill out in the Bowland Basin. He named the latter sequence the "Grits of Pendle Hill and Longridge" (Figure 2.2a).

The systematic survey of the Askrigg Block (Green *et al.*, 1878; Dakyns, 1890, 1893; Dakyns *et al.*, 1879) confirmed Phillips' original correlations and established a nomenclature for the Yoredale Series which has changed little since (Figure 2.4). Continued Survey work led to the Bearing Grit being renamed the Grassington Grit (Dakyns, 1890; 1892) and to the recognition that this unit passed northwards via the Howgate Edge Grit into a ganister (Figure 2.2b). The survey also recognised Phillips' miscorrelation in the Bowland Basin and restricted the name Pendle Grits to the lower sand in the basin (Figure 2.2a). Kinderscout Grit was used for the upper sand in the basin and also for the Bearing Grit and equivalents on the Askrigg Block. This followed the nomenclature of the Derbyshire section where the Kinderscout Grit was the first pebbly unit in the Millstone Grit sequence.

Later research (Hind, 1902) led to the recognition of an unconformity at the base of the Grassington Grit on the Askrigg Block. According to Chubb and Hudson (1925), this unconformity was first suspected by Dakyns. Chubb and Hudson mapped the unconformity and showed that it was extensive across the southern margin of the Askrigg Block. They also proved that southwards thinning of the Yoredales cyclothems was partially the result of the unconformity and not due to lateral transition of the Yoredales into Millstone Grit facies. However, because of the lack of biostratigraphic control, they believed that the unconformity represented an area of non-deposition between a retreating Yoredales depositional system and a transgressing, time-equivalent Millstone Grit system. Consequently, they proposed that the basal grit above the unconformity became younger northwards: the Lower Howgate Edge Grit was

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Figure 2.3: Correlation of Pendleian E_{1c} sections. The section shows the relationships of the various stratigraphic units to the palaeogeographic elements shown in Figure 2.1. Sections 8 and 9 are from Whitmoor and Roosecote boreholes respectively. All other sections are based on outcrop. The stratigraphic nomenclature used on this diagram is formally defined in the text.


MORECAMBE HIGH
LANCASTER FELLS BASIN
WADDINGTON FELL HIGH
BOWLAND BASIN
CRAVEN FAULT ZONE
ASKRIGG BLOCK
9
8
7
6
5
4
3
2
1
GRASS. GRIT FM.
PSF
PENDLE GRIT FM.
W.M.
WARLEY WISE FM.
PSF
PENDLE GRIT FM.
BOWLAND SHALES
G.G.
"CARB. LST."
WENSLEYDALE GP.
100m
MARINE BANDS
Cravenoceras cowlingense
Cravenoceras malhamense
W.M. WHITENDALE MEMBER
P.S.F. PENDLE SHALE FORMATION
G.G. GRASSINGTON GRIT FORMATION
Pebbly sandstone
Massive turbidite sandstone
Thin-bedded turbidite sandstone and mudstone
Carbonaceous siltstone and mudstone
Mudstone
N
LOCATION MAP

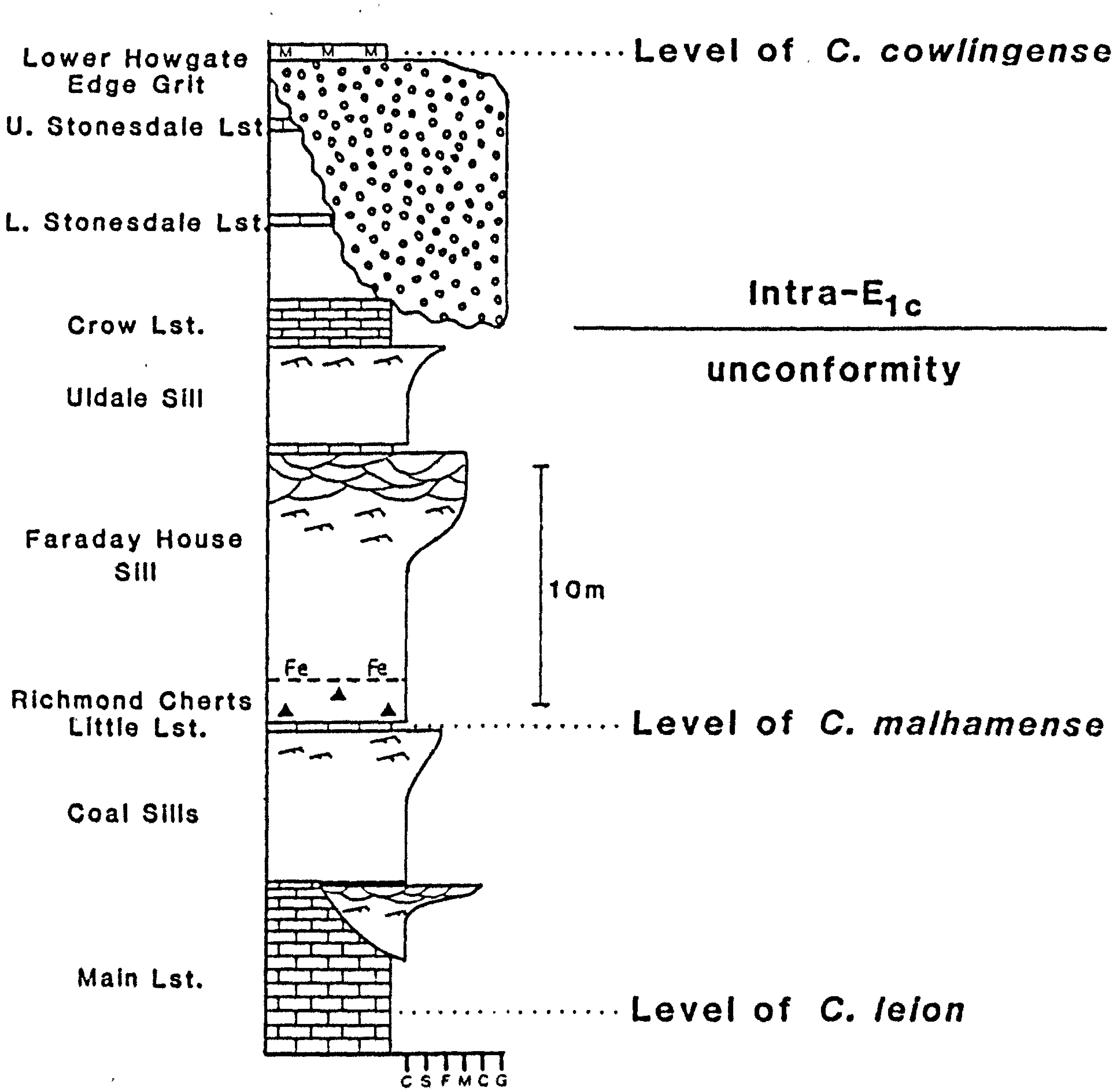

Figure 2.4: Schematic log of the Pendleian parts of the Wensleydale Group (Yoredales) on the Askrigg Block. Note the occurrence of coarsening upwards cycles begining with a marine limestone, passing up through muds into silts and sands. Each cyclothem is recognisable across most of the Askrigg Block and represents progradation of a small delta lobe. The thicknesses and sand contents of the cyclothems vary across the Block and so the thicknesses shown are representative rather than exact. The Little Limestone is thought to represent the base of E_{1c} on the Askrigg Block.

thought to overlap the Grassington Grit (Figure 2.5a) rather than being its lateral equivalent. Through discovery of new goniatite localities, Rowell and Scanlon (1957a, 1957b) were able to re-establish the lateral equivalence of these grits. Their careful mapping showed that the sub-Grassington Grit unconformity was not diachronous as Chubb and Hudson had believed. Rather, they showed that below the unconformity the Yoredales were time-equivalents of the Bowland Shale/Pendle Grit interval in the Bowland Basin. Only above the unconformity was there a lateral transition from Millstone Grit facies into Yoredale facies (Figure 2.5b). Rowell and Scanlon were thus the first authors to fully appreciate the changing nature of the block-basin transition during E_{1c} times.

Further work on the E_{1c} stratigraphy of the Askrigg Block was undertaken by various authors. Tonks (1925) recognised sub-Grassington Grit channels in Nidderdale and Wilson (1957) subsequently showed that relief on the intra E_{1c} unconformity surface in this area was also due to channelling. Dunham and Stubblefield (1945) and Black (1950) refined the mapping of the Grassington Grit along the southern margin of the Askrigg Block. The latter recognised overstep of the Grassington Grit across Bowland Shales which were restricted by palaeo fault scarps cutting the underlying Yoredale rocks.

The development of goniatite biostratigraphy around 1924 led to major changes in the correlations of the two sandstones in the Bowland Basin. Using his new goniatite zones, Bisat (1924) was able to prove that the so-called "Kinderscout Grit" of the Bowland Basin was, in fact, much older than the Kinderscout Grits of the type area in Derbyshire. Wright *et al.* (1927) were the first to use marine bands to constrain the mapping of grit packages. They used the term Pendle Grit Group to cover both E_{1c} sandstones in the Bowland Basin. To distinguish the two packages they introduced the local names Upper and Lower Wilpshire Grit (Figure 2.2a). In mapping the area around Lothersdale Bray (1927) used the names Pendle Top Grit and Warley Wise Grit for the lower and upper sand packets respectively (Figure 2.2a). For the first time he also named the shale intervals in the succession, refering to the barren muds between the two grits as the Surgill Shales.

The resurvey of the Skipton Sheet (Stephens *et al.*, 1942; 1953) introduced two further stratigraphic names inspite of covering ground immediately adjacent to that of Bray. On the south side of the Skipton Anticline the whole of the E_{1c} interval above the Bowland Shales was referred to the Skipton Moor Grit Group. To the north of the anticline the same interval was called the Lindley Moor Grits (Figure 2.2a). Earp *et al.* (1961) reverted to Bray's nomenclature for the resurvey of the Clitheroe district (Figure 2.2a). During this resurvey they recognised that pebbly grits in the Waddington/Newton Fells outlier (for location see Figure 1.1 and 1.2) rested unconformably on Bowland Shales. These grits they correlated with the Warley Wise Grit of the main crop and they suggested that the unconformity was equivalent to the intra-E_{1c} break on the Askrigg Block.

Parkinson (1936) had previously named the grit on Waddington/Newton Fell the Newton Fells Grit, thinking that the unit lay in E_{1b} (i.e. that it was older than the Pendle Grit). He had failed to identify the unconformity below this unit. Before Parkinson's mistake was recognised, Moseley (1954) used the name Newton Fells Grit for a thin E_{1b} sandstone in the Lancaster Fells Basin (Figure 2.2d), believing it to be the lateral equivalent of the grits on the outlier. The B.G.S. have recently amended the stratigraphic nomenclature by renaming Moseley's E_{1b} sandstone the Hind Sandstone (Figure 2.2d). In the Lancaster Fells Basin both

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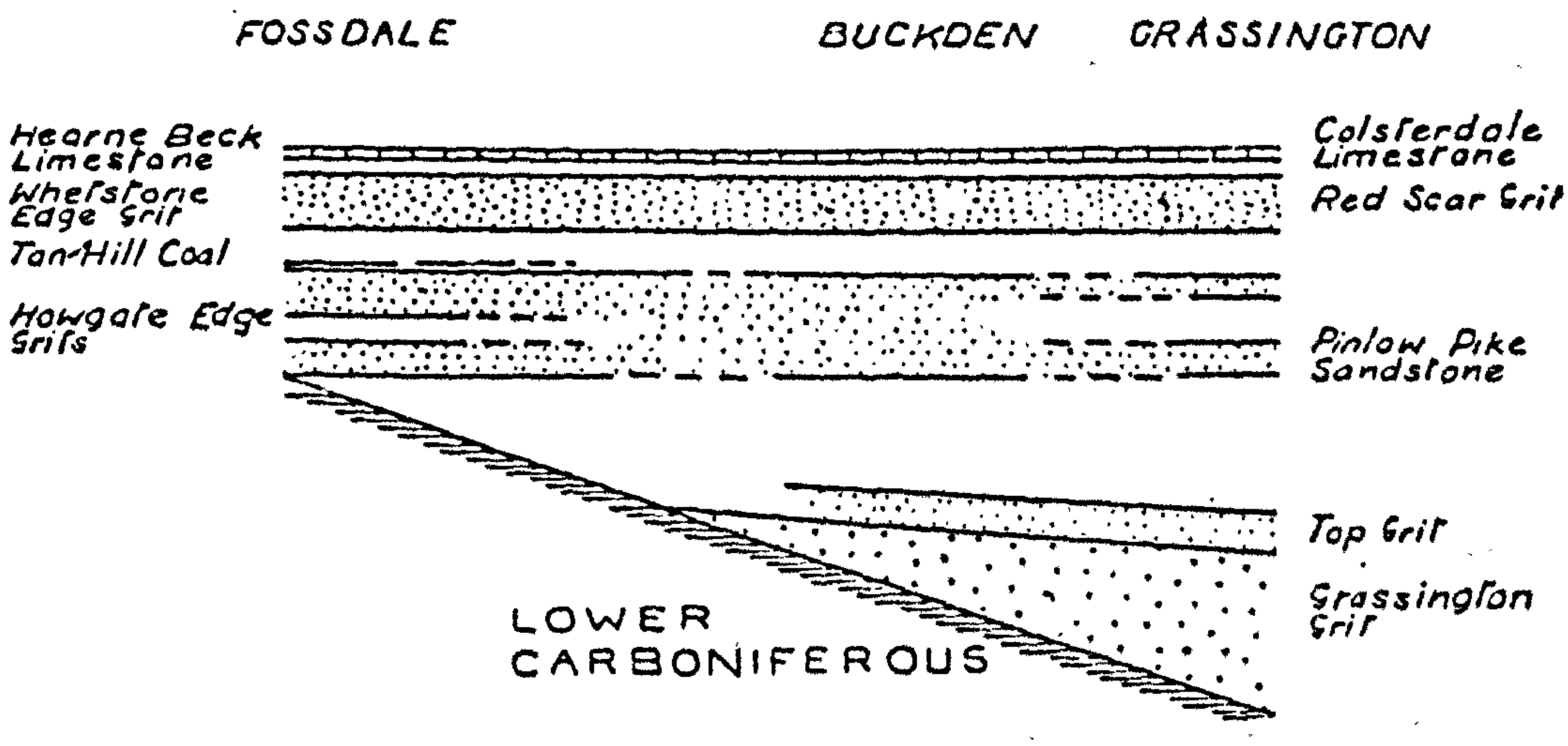
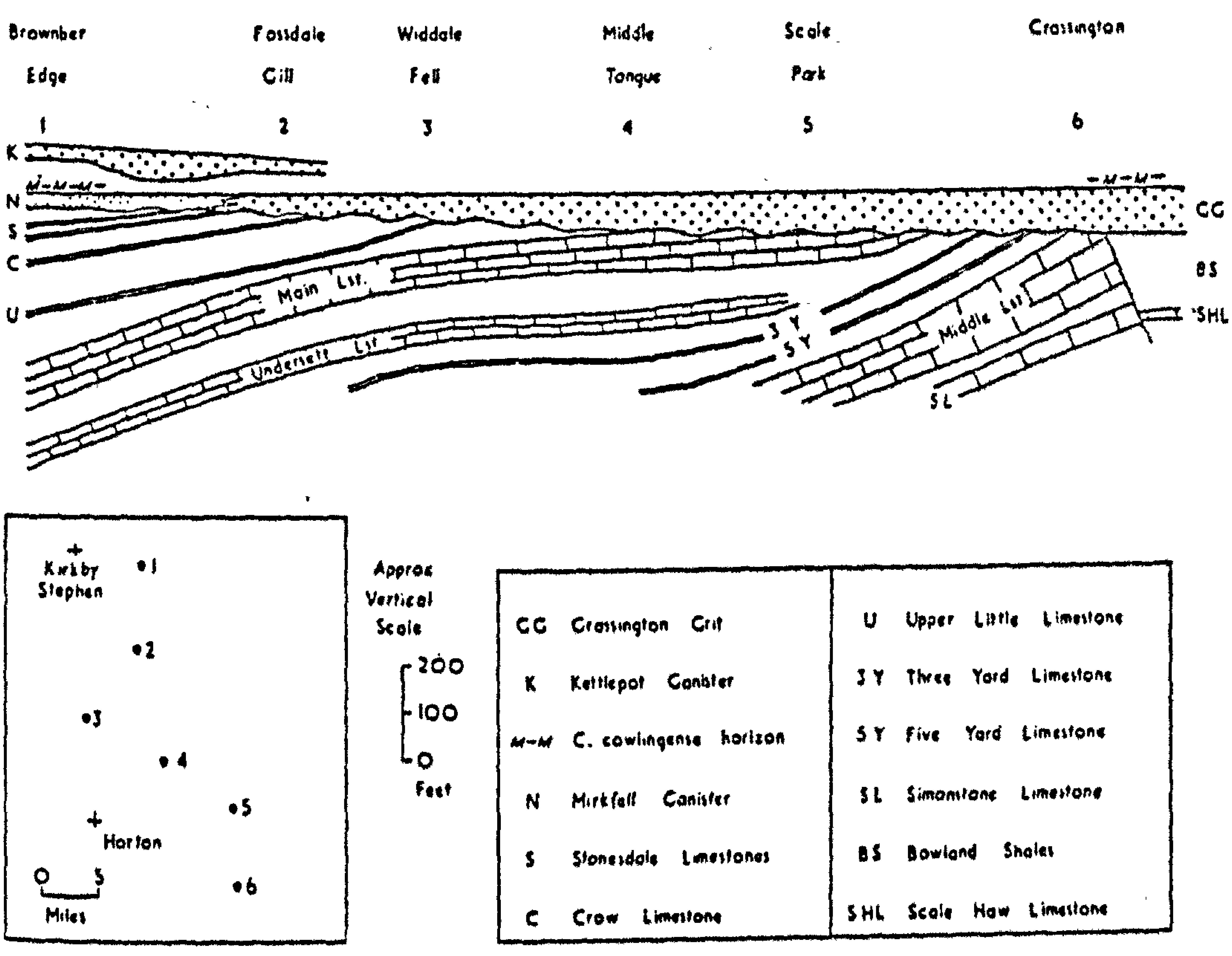
Figure 2.5: Interpretations of the stratigraphic relationships across the Intra-E_{1c} unconformity. Chubb and Hudson's interpretation of overlapping beds above the unconformity was based on a mis-correlation of the Howgate Edge and Grassington Grits. The discovery of new goniatite localities by Rowell and Scanlon allowed them to demonstrate that there was no overlap: the Mirkfell Ganister (laterally equivalent to the Howgate Edge Grit) and the Grassington Grit are time equivalents as both are overlain by beds containing *Cravenoceras cowlingense*.

a) Chubb & Hudson (1925)



b) Rowell & Scanlon (1957a)



Parkinson and Moseley referred to the lower E_{1c} sand package as the Pendle Top Grit. Moseley introduced the name Brennand Grit for the upper sand (Figure 2.2d) which Parkinson had correctly correlated with the Ingleborough and Grassington Grits of the Askrigg Block.

The E_{1c} nomenclature for the Eastern Bowland Basin (Figure 2.1) was developed independently to that in the rest of the Bowland Basin. Although, as mentioned above, Phillips had correctly correlated the Almscliff Grit with the Grassington Grit on the Askrigg Block, he traced the grit incorrectly eastwards towards the Harrogate anticline. The poor exposure obscured the fact that the Almscliff Grit was faulted along strike against a very similar grit in E_{2a}: Phillips believed that the outcrops on either side of the fault were of the same stratigraphic unit. Consequently he referred the true representative of the Almscliff Grit to a new unit, the Harrogate Tunnel Grit, while assigning the name Almscliff Grit to what is, in reality, the stratigraphically younger Red Scar Grit. (Figure 2.2c).

The Geological Survey originally mapped three sandstone units on the south side of the Harrogate anticline, referring them all to the Kinderscout Grit (Fox-Strangeways, 1873. Figure 2.2c). The upper of these units was Phillips' Almscliff Grit; implying that the Survey had made the same mapping error as had Phillips. Hudson accepted the Survey mapping (1930b, 1934, 1938) but revived Phillips' nomenclature as, following the application of goniatite stratigraphy to the area, the use of Kinderscout Grit for the Pendleian section was clearly incorrect (Figure 2.2c). Hudson (1930b) did notice, however, the apparent abnormal thickness of the Bowland Shale which resulted from the miscorrelation of the grits. Noting that the interval also appeared to be more sandy, he suggested that the "Bowland Shale" at Harrogate was at least in part laterally equivalent to the Pendle Grit of the main Bowland Basin. In fact this is very close to the truth, as revealed by the recent B.G.S. resurvey. Hudson, however, seems to have dropped the idea prior to his later contributions (1934, 1938).

Finally, two other stratigraphic schemes have been used for the E_{1c} interval in the transition zone between the Bowland Basin and the Askrigg Block. Hudson (1939) used Pockstones Grit for the upper of the two E_{1c} grits in the Simonseat Anticline while Jones (1943) used Deerstones Grit and Beamsley Grit for the two grits in in the Beamsley area (Figure 2.2c).

2.3 Discussion of the boundaries of the E_{1c} zone.

The previous sections have assumed a knowledge of which units lie within the *Cravenoceras malhamense* zone. There is some debate, however, as to the actual position of the zonal boundaries in some parts of the area. The main problem has been the positioning of the upper boundary in areas in which the marine band defining the top of the zone (*Cravenoceras cowlingense*) is not developed. In many cases the absence of this marine band is due to sedimentological conditions at the end of E_{1c} which were inimicable to goniatites (e.g. high sand supply). Similarly the basal marine band, (*Cravenoceras malhamense*) may also be missing, usually due to erosion beneath the intra-E_{1c} unconformity. Generally, however, the base of E_{1c} is well defined across the area. In the basins the *Cravenoceras malhamense* band is usually present although it may be locally cut out over intrabasinal highs. On the Askrigg Block C. aff. *malhamense* was recovered from the Little Limestone in Allendale (Dunham and Johnson, 1962. See Figure 2.4) and consequently this limestone has been taken as the base of the E_{1c} zone. The Little Limestone cyclothem is also the most sandy unit in the Pendleian succession on the Askrigg Block which correlates well with the enhanced sand supply to the Bowland Basin during E_{1c}.

Fixing the top of the E_{1c} zone poses more serious problems. By the end of this time most of the basin had been filled to base-level and conditions were rarely suitable for deposition of marine bands. On the Askrigg Block the end E_{1c} transgression switched off the sand supply to the Grassington Grit system which led to the deposition of abandonment facies such as the Mirkfell Ironstone and the Cockhill Limestone. Both of these units contain *Cravenoceras cowlingense* and hence their bases are taken as the top of E_{1c}. Similarly the presence of the C. *cowlingense* band above the Almscliff Grit in the Harrogate district (Godwin, 1971) ties down the upper zonal boundary at the eastern end of the Bowland Basin. In the main part of the basin Moseley (1954) recorded C. *cowlingense* and *Eumorphoceras bisulcatum* in marine bands above the Brennand Grit in the Lancaster Fells. Yates (1962) showed that these were equivalent to the Cockhill Limestone on the Askrigg Block and, thus, confirmed the correlation of the Brennand and Grassington Grits. The B.G.S. have recently proved the presence of the C. *cowlingense* band at Grizedale (SD 52 48) in the western Lancaster Fells. Here the band lies immediately above the Pendle Grit but the Brennand Grit has probably failed by this point (Aitkenhead pers. comm.). Consequently, the discovery does not challenge the age relationships established further east in the Fells.

Yates (*op. cit.*) also showed that the Edge and Warley Wise Marine bands (which are exposed above the Warley Wise Grit along the Pendle Monocline) were younger than the Cockhill Limestone: the presence in these bands of *E. bisulcatum* alone is indicative of a high E_{2a} age. The stratigraphic position of the underlying Warley Wise Grit is, therefore, debatable. Ramsbottom (1974) originally suggested that this Grit should be placed in the Arnsbergian and that the level of the C. *cowlingense* marine band lay between the Pendle and Warley Wise Grits. Baines (1977), however, correlated the Bradley Flags (which overlie the Warley Wise Grit) with the C. *cowlingense* band. He recognised a series of wave influenced facies in the Bradley Flags which he interpreted as being deposited during the end E_{1c} transgression. This correlation was subsequently accepted by Ramsbottom *et al.* (1978) for incorporation in the Geological Society of London Silesian Correlation report. The lithological and sedimentological similarity between the Warley Wise, Grassington and Brennand Grits is very strong evidence to support this correlation (Chapter 4) but it still possible that the Warley Wise Grit might extend up into the early Arnsbergian.

2.4 Formal Stratigraphy of the E_{1c} zone.

No formal stratigraphic divisions have been proposed for the E_{1c} sediments of the Bowland Basin. Arthurton *et al.* (1988) did use the term Pendle Grit Formation for sub-Brennand Grit sands and shales in the Settle area but did not formally erect the division (Figure 2.2d). Similarly, Baines (1977) used both Pendle Grit Formation and Pendle Shale Formation without formal definitions being published (Figure 2.2a). To clarify the situation all the major E_{1c} units which have been studied in the field by the current author are formally defined below. Type and reference sections have been erected for these units. Formal names have also been suggested for units which are outside the scope of the current research but which are clearly part of the E_{1c} lithostratigraphic package. The data for these units comes from literature reviews. Relationships between the stratigraphic units defined herein are summarised in Figure 2.6 (also see Figure 2.3). The formal stratigraphy conforms with the American Code on Stratigraphic Nomenclature (American Commission on Stratigraphic Nomenclature, 1961) and is based on the latest biostratigraphic and sedimentological data. An understanding of how the sedimentological development of the area has been influenced by block-basin controls has been critical in determining the boundaries between stratigraphic units.

Bowland Shale Group

The Bowland Shale Group is found only in the Bowland Basin and across the transition zone with the Askrigg Block. It spans the Dinantian-Namurian boundary which is placed at the level of the *Cravenoceras leion* marine band. The sediments above this level are known as the Upper Bowland Shales and only their uppermost part of lies within E_{1c}.

Above the C. *malhamense* marine band the Upper Bowland Shales are lithologically very variable. In the Skipton district they consist of blocky mudstones with a poorly developed millimetre scale colour banding resulting from slight variations in grainsize. The mudstones are barren and contain little carbonaceous material. Thin silt and sand laminae become more abundant upwards, heralding the arrival of the Pendle Grit Formation (q.v.). Along the length of the Pendle Monocline the Upper Bowland Shales vary little from this norm (see Moore, 1933; Gill, 1940) although the thickness of the unit does drop towards the south-west. This is probably due to it becoming condensed over minor structural highs (Clemmey, pers. comm.). In the Lancaster Fells and Settle areas there is a similar thinning to the south-west. Around the Sykes Anticline the unit consists of laminated mudstones with occasional thin calcisiltite bands. Further north the section is thicker and more silty, carbonaceous and micaceous (Arthurton *et al.* 1988). A similar thick Upper Bowland Shale section is present in the Keer Valley on the North side of the Lancaster Fells Basin (Toswill, pers. comm.).

No formal Formations have been recognised within the E_{1c} parts of the Bowland Shale Group: the unit is too uniform to allow mappable units to be distinguished. However, in other parts of the Bowland Shale Group there are units which should have Formation or Member status and which will require formal definition at some stage. (For example, the Pendleside Sandstone, the Embsay Limestone etc.). The only Upper Bowland Shale unit referred to later in

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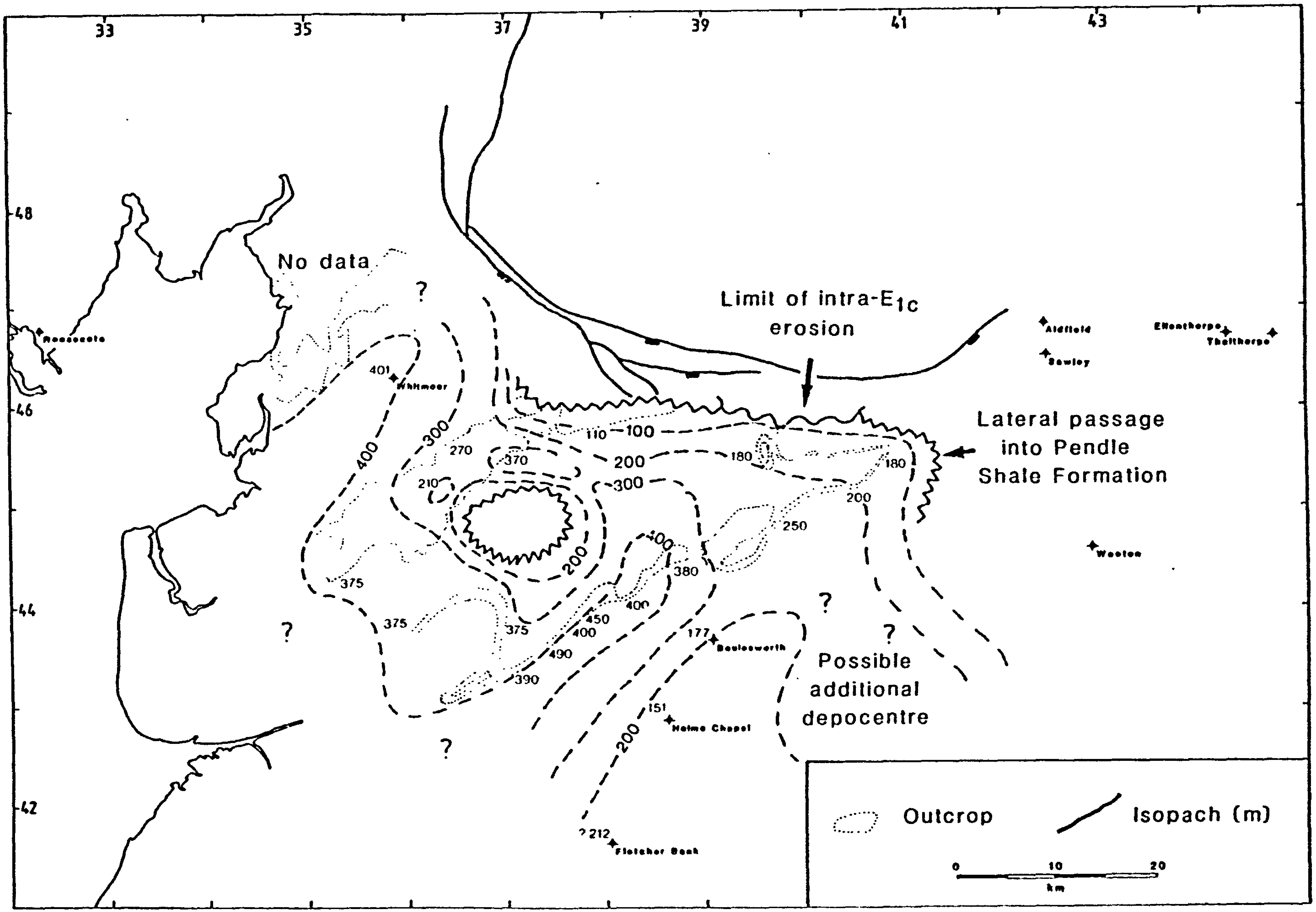
Figure 2.6: E_{1c} stratigraphic relationships in north-western England. Note the profound influence of the Craven Fault Zone on the distribution of the stratigraphic units. During early E_{1c} times the complete palaeogeographic differentiation between the Askrigg Block and the Bowland Basin led to Group level stratigraphic differences. Above the intra-E_{1c} unconformity, however, the block/basin differentiation broke down and stratigraphic continuity at the Group level was established. The relationship of the *Cravenoceras cowlingense* marine band to the Pendle Grit Group at the west end of the main section (D) is partly based on unpublished B.G.S. data (Aitkenhead, pers. comm). Diagram is not to scale.

D
C
B
A
BOWLAND BASIN
CRAVEN FAULT ZONE
ASKRIGG BLOCK
C L
GRASSINGTON GRIT GROUP
GRASSINGTON GRIT FORMATION
M M
WENSLEYDALE GROUP (parts)
WARLEY WISE FORMATION
Intra-E1c unconformity
PENDLE SHALE FORMATION
PENDLE GRIT FORMATION
PENDLE GRIT GROUP
WHITENDALE MEMBER
BOWLAND SHALE GP. (parts)
B
E
EASTERN BOWLAND BASIN
GRASSINGTON GRIT GROUP
GRASSINGTON GRIT FORMATION
ALMSCLIFF MEMBER
PENDLE SHALE FORMATION
PENDLE GRIT GROUP
SMM
PENDLE GRIT FM
Location Map
m Level of C. malhamense
c Level of C. cowlingense



No data
Limit of intra-E1c erosion
Lateral passage into Pendle Shale Formation
Possible additional depocentre
Outcrop
Isopach (m)
0 10 20 km
33 35 37 39 41 43
48 46 44 42
400
300
200
100

this work is the Hind Sandstone which is given provisional member status herein (Figure 2.2d).

Pendle Grit Group

The Pendle Grit Group includes all the stratigraphic units between the top of the Bowland Shale Group and the base of the Grassington Grit Group. The Group is only developed in the basinal areas and consists of a series of formations which are genetically linked (Figure 2.6).

Pendle Grit Group, Pendle Grit Formation.

Name: Named after Pendle Hill, the location of the type section. This formal definition of the Pendle Grit Formation covers the pre-existing local names Deerstones Grit, Pendle Top Grit, and Pendle Grit. "Grit" has been retained in the formal name to maintain consistency with the widespread use of the name Pendle Grit. Pendle Grit has historical precedence over the other local names. The new definition contains only the basal, sandy parts of the the Pendle Grit Formation of Arthurton *et al.*, (1988). The latter's definition of the formation was based on too small an area to be useful on the scale of the basin as a whole.

Type section: No single complete section through the Formation is known. Little Mearley Clough (SD7852 4109 to SD7890 4119) on the north flanks on Pendle Hill provides the type section for the base and middle parts of the Formation. Quarries at Wiswell Moor (SD755 373), some 3.5km south-west of Little Mearley Clough, extend the type section to the top of the Formation (Refer to Figures 3.16 and 3.20)

Figure 2.7: Distribution and thickness of the Pendle Grit Formation. Thicknesses include those of the Skipton Moor and Whitendale Members (q.v.). The contouring of the essentially linear data strips is based on palaeocurrent data and on interpretations of the tectonic structure of the basin during deposition (from Lee 1988a). Data compiled from many sources.

Reference sections: Good stream sections in the Formation are present in Whitendale (SD6537 5590 to SD6560 5680) and the Brennand Valley (SD6400 5456 to SD6402 5550) in the southern Lancaster Fells. A section in the River Hodder (SD7000 5870 to SD7082 5893) provides a reference section for the more mudstone rich parts of the Formation.

Thickness and distribution: The Pendle Grit Formation is present throughout the main parts of the Bowland and Lancaster Fells Basins but is locally absent over intrabasinal highs such as Waddington Fell (Figure 2.7). In the Bowland Basin the Formation reaches a maximum thickness of around 450m in the area between Pendle Hill and the River Ribble (Figure 2.7). South-westwards along the Pendle Monocline the unit thins to 370m at the limit of the outcrop. North-eastwards, towards Skipton, there is a marked thinning as the Formation passes laterally into the Pendle Shale Formation (Figure 2.7). In the Lancaster Fells Basin the Formation thins from 450m in the west to less than 100m over the Bowland High. Again, the Formation passes eastwards into the Pendle Shale Formation. In the centre of this basin 401m of the Formation was proved in the Whitmoor-1 borehole. Further north around Lancaster thicknesses appear to be similar but measurement of accurate thicknesses is difficult due to the structural complexity of the area. North-west of Lancaster, in the Barrow-in-Furness district, the Formation passes laterally into the Roosecote Mudstones (Rose and Dunham, 1977. Figure 2.3, Section 9).

In the subsurface, the Formation extends south of the Pendle Monocline into the Huddersfield Basin: 151m of sands referred to this Formation were present in the Holme Chapel-1 borehole and 177m were encountered in Boulesworth-1 (Figure 2.7). It seems likely that there may be an additional Pendle Grit Formation depocentre in this region. The south-westerly extent of the formation under the Fylde and East Irish Sea is unknown.

Lithology: In the type sections the Formation is characterised by coarse to fine feldspathic sandstones of light grey to buff colour. The sandstones are poorly sorted and often contain large numbers of mudflakes. Bed thickness varies from over 1m to 10cm or less, the thinner beds usually being finer grained. Thin siltstone or mudstone partings between beds are sometimes developed. These become more common upwards in the sequence. Mica and plant debris are common in the finer grained sandstones. Elsewhere the sandstones are similar but there is a general increase in the maximum grainsize northwards across the Lancaster Fells Basin: granule and pebbly sandstones are present in the southern Lancaster Fells while pebbly sandstones and rare orthoconglomerates are found around Lancaster itself. Thick mudstone packets, although not present in the type section, are present in the eastern parts of the Basin where the Formation is transitional with the Pendle Shale Formation. In these areas the sandstones form large lenticular bodies set within a silty mudstone sequence. The sandstone bodies are volumetrically the most significant part of the sequence.

Boundaries: Where the Whitendale Formation (q.v.) is absent, the base of the Formation is taken at the base of the lowest massive sandstone. The base is often sharp due to channelling but, in some cases, there are a few thinner sand beds beneath the major sandbodies. The top of the Formation is a gradational boundary with the overlying Pendle Shale Formation and is rarely exposed. While mapping, the boundary is usually placed using topographic features which correspond to large drops in the sand/shale ratio. Consequently, to be consistent with this practise, the top of the formation is defined as the level at which the sandstone body/background shale ratio drops to less than 0.25.

Pendle Grit Formation, Whitendale Member.

Name: Named after the valley in which the type section is located.

Type section: The type section lies in the Whitendale River of the Lancaster Fells (SD6539 5588 to SD6538 5590). The whole of the Member is well exposed and is highly water polished (See Figure 3.21).

Reference sections: Two well-exposed sections are proposed as the reference sections for this Member. The first is in river cliffs on the south side of Calf Clough (SD6625 5495 to SD6643 5498) and the second at Trough Scar on the west side of the Trough of Bowland (SD624 528).

Thickness and distribution: The Member is confined to the south-western margin of the Lancaster Fells where it varies in thickness from 13 to over 100m (Figure

2.8). The Member is also present in the Whitmoor-1 borehole in the centre of the Lancaster Fells Basin where it is 82m thick. It is not seen on the northern margin of the Lancaster Fells Basin around Lancaster.

Lithology: The Member consists of muddy, feldspathic sandstones (greywackes) interbedded with silty, carbonaceous mudstones. Together these form a series of metre and decimetre scale fining-upwards units. The sand/shale ratio is low at the base of the Member but increases upwards as the graded units become progressively thicker and coarser. Sorting improves upwards through the Member and the sands are generally cleaner: the topmost sands are identical to the finer sands in the Pendle Grit Formation. Ball and pillow, flame and convolute structures are characteristic of this Member.

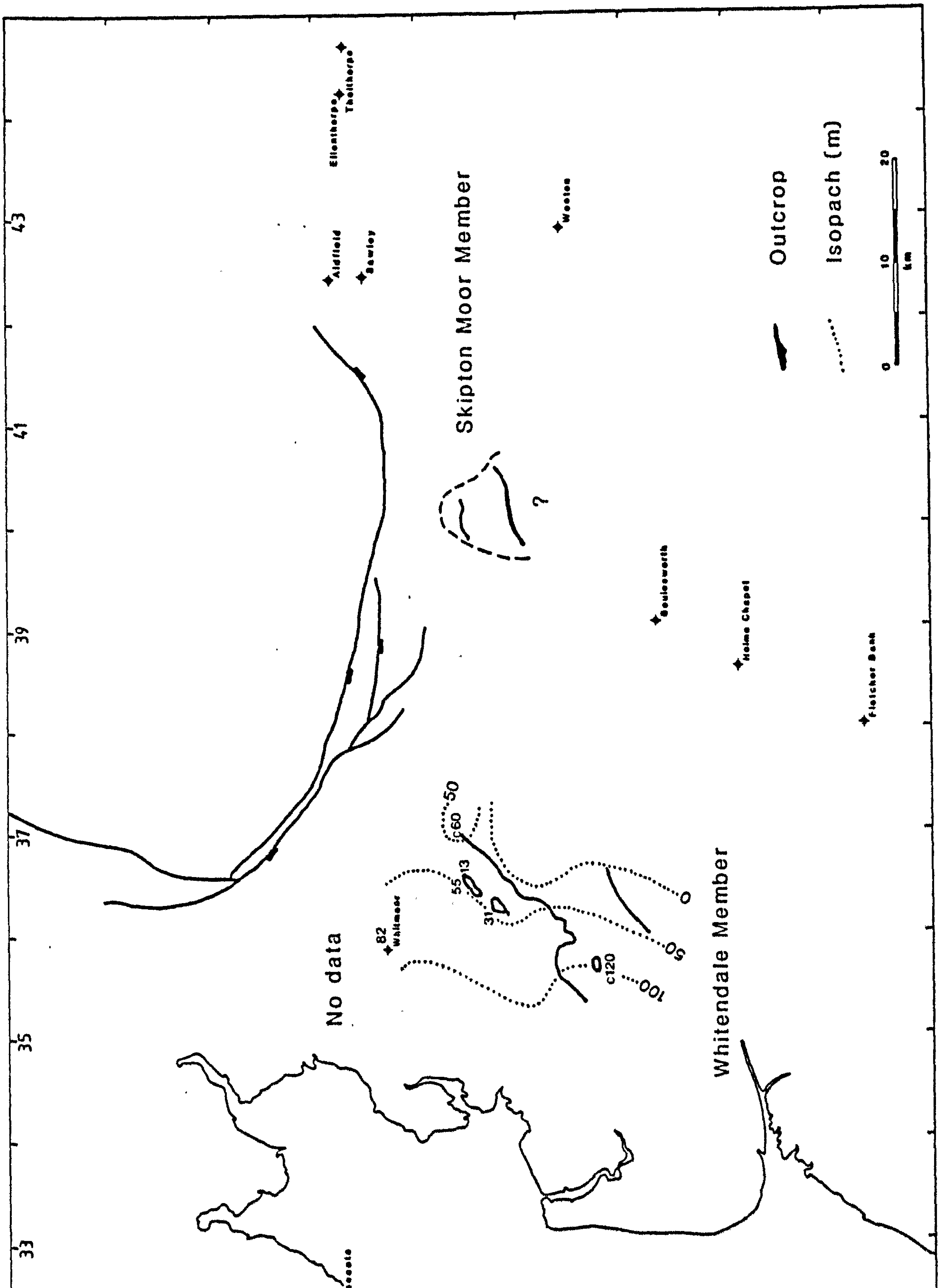
Boundaries: This Member sits between the Bowland Shale Group and the Pendle Grit Formation s.s. (Figure 2.6). The base is gradational with the former and,in the type section, is placed 13m above the *C. malhamense* marine band at the bottom of a 50cm thick, coarsening-up bed of silty sandstone. The base is defined elsewhere at the base of the first sandstone unit greater than 10cm in thickness. The top of the Member is placed at the first appearance of granule grade or pebbly sandstones at the base of the Pendle Grit Formation s.s..

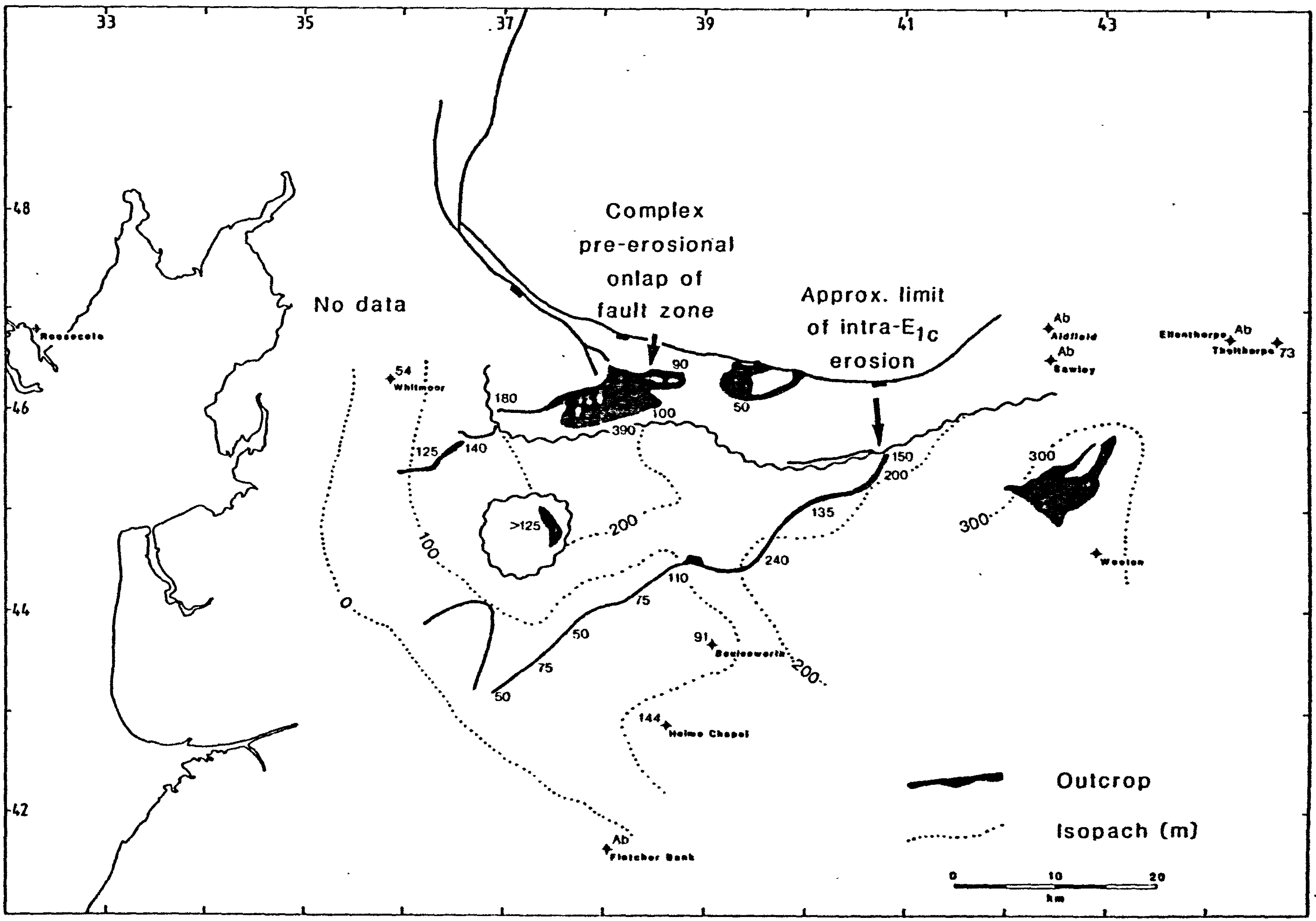
Pendle Grit Formation, Skipton Moor Member.

Name: Derived from the name of the geographical feature on which the member is best exposed.

Type Section: There are no good vertical successions through this member. Outcrops of the sandstone units occur widely on the hillside south-east of Jenny Gill (SE003 509) but the intervening finer facies are very poorly exposed.

Figure 2.8: Distribution of the Skipton Moor and Whitendale Members. Representative thicknesses are shown for the Whitendale Member but the data are sparse and the isopachs unreliable. The data does suggest that the unit thins over the Sykes Anticline system.


Skipton Moor Member
Whitendale Member
No data
Outcrop
Isopach (m)
0
10
20
km
Aldfield
Holme Chapel
Fletcher Bank
43
41
39
37
35
33
48
46
44
42
50
c60
55
13
31
82
c120
100
50
0
?


No data
Complex
pre-erosional
onlap of
fault zone
Approx. limit
of intra-E_{1c}
erosion
Outcrop
Isopach (m)
km

Reference Sections: None.

Thickness and distribution. The member is only present in a restricted area of the Bowland Basin around the Skipton anticline. In this area it reaches a maximum thickness of 170m (Figure 2.8).

Lithology: The member consists of lenticular bodies of coarse to granule grade sandstones sitting in carbonaceous siltstones with thin laminae of fine sand. The coarse sandstones are feldspathic and moderately well sorted. The sandstone bodies, some 25-50m thick, contain a distinctive sedimentary structure named ropy weathering by Baines (1977). This structure (illustrated and discussed in detail in Chapter 4) distinguishes the sandstones in the member from those in the main part of the Pendle Grit Formation.

Boundaries: The base is placed at the base of the first sandbody with well-developed ropy weathering. The top is placed at the top of the youngest similar unit.

Pendle Grit Group, Pendle Shale Formation.

Name: The name is again taken from Pendle Hill and was chosen to highlight the close sedimentological association with the Pendle Grit Formation. The usage roughly equates with that of Baines (1977). Neither of the local names (Surgill Shales and Beamsley Shales) were considered well-enough established to warrant their retention in the formal name.

Figure 2.9: Distribution and thickness of the Pendle Shale Formation. The Formation is thickest in the north and east where it represents the whole of the Pendle Grit Group (Compare with Figure 2.7). It originally thinned to zero across the Craven Fault zone but the nature of the onlap was complicated by topography on minor fault blocks (Arthurton *et al.*, 1988; Hudson, 1930a; Black,1950). Subsequent erosion beneath the Intra-E_{1c} unconformity further complicates the thicknesses and, hence, no attempt has been made to contour this zone. The south-western zero edge is probably the depositional limit of the Formation. Data compiled from published sources.

Type section: The Formation is poorly exposed in the main parts of the Bowland Basin. Consequently a section in a tributary of Oak Beck, Harrogate (SE2781 5420 to SE2754 5436) has been chosen as the type section. The section, in the Eastern Bowland Basin, exposes approximately 250m of the middle and upper parts of the Formation.

Reference Sections: A series of reference sections are defined because of the lateral variability of the Formation. These sections tend to be of limited stratigraphic thickness. Sections in Lancashire Gill (SD897 432 to SD905 433), Harden Beck (SD910 448 to SD913 444) and Low Burn Gill (SD995 549 to SD992 553) are representative of the Formation in the Bowland Basin. In the Craven Lowlands, reasonably well exposed sections are found in Rathmell Beck (SD7943 5995 to SD7870 5967) and in Long Gill Brook (discontinuous section from SD784 584 to SD768 772). Arthurton *et al.* (1988) give details of further sections in this area.

Thickness and distribution: The Formation is present over most of the Bowland Basin and the Askrigg Block transition zone (Figure 2.9). At the eastern end of the Bowland Basin around Harrogate, the Formation replaces the Pendle Grit Formation and reaches 350m in thickness. As the Pendle Grit Formation thickens westwards, so the Pendle Shale Formation thins to less than 50m at Pendle Hill. West of this again the Formation is not proved although it has been mapped by feature on the 1:50,000 Clitheroe and Preston Sheets. Along the Bowland High, the Formation varies from 70 to 300m in thickness but the top is often eroded beneath the intra-E_{1c} unconformity. In the Lancaster Fells Basin the Formation is present in the Whitmoor-1 borehole where it is 54m thick. It has not been proved at outcrop further north in the basin.

Lithology: Interbedded silty mudstones, laminated siltstones and thin, fine-grained sandstones form the bulk of the Formation. When the thin sandstone beds form a significant proportion of the section, these lithologies are frequently referred to as "striped beds" (e.g. Arthurton *et al.* 1988). All of the lithologies contain abundant mica and platy carbonaceous material while trace faunas are well developed at some levels in the Formation (Baines 1977, Eag.r *et al.* 1985). Large-scale slumping within the finer lithologies is characteristic of the Formation.

Across the Bowland High and in the Askrigg Block transition zone, lenticular bodies of coarse to fine grained sandstone occur at all levels in the Formation. The lenticular units are generally less than 5m thick and are generally unmappable due to their small lateral extents. The sandstones in the units are massive or feintly cross-bedded and, in this sense, are distinctly different from the lenticular sandstones in the Pendle Grit Formation in this area.

Boundaries: The base of the Formation is gradational with the Pendle Grit Formation or, where the latter is not developed, with the Bowland Shale Group (Figure 2.6). The boundary with the Pendle Grit Formation is arbitrarily placed at the level at which the overall lenticular (i.e. channel) sand to encasing shale ratio drops to below O.25. In the outcrops around the western Lancaster Fells and along the Pendle Monocline, this boundary is distinct. It is much more difficult to place in the area between the Catlow anticline and the Craven Faults where the Pendle Grit Formation is more shale rich. Beyond the pinchout of the Pendle Grit Formation, the boundary of the Pendle Shale Formation with the Bowland Shale Group is placed at the base of the first sandstone body or at the level at which siltstones account for more than 50% of the section (whichever is the lower). The top of the Formation is placed at the base of the pebbly sandstones of the overlying Grassington Grit Group. Often, this is an unconformable contact (See Figure 2.6).

Grassington Grit Group.

The term Grassington Grit Group was first used by Hudson (1939) to refer to the succession between the intra E_{1c} unconformity and the Cockhill Limestone on the Askrigg Block. In this work the Group is redefined to include the lateral equivalents of this succession in the Bowland and Lancaster Fells Basins. As with the Pendle Grit Group, the Grassington Grit Group contains a number of formations which are genetically and sedimentologically linked.

Grassington Grit Group, Grassington Grit Formation.

This Formation has already been defined by Dunham and Wilson (1985) and the comments below expand only on the distribution of the unit in the area of the Bowland High.

Name: The name comes from the village around which the Formation was first recognised.

Type and reference sections: See Dunham and Wilson (*op. cit.*) and Figure 2.10. An additional reference section is defined in the Brennand Valley (SD6315 5523 to SD6278 5595) which illustrates the nature of the Formation across the Bowland High.

Thickness and distribution: The Formation is present over much of the Askrigg Block and the Bowland High. In these areas the thickness averages 50m and 80m respectively but local thickness variations are considerable (Figure 2.11). On the Bowland High, for example, the Formation locally attains a thickness of 220m (Arthuton *et al.*, 1988) due to tectonic stacking of channel sediments (See Chapter 6). Dramatic thickness variations on the northern margin of the Askrigg Block are also attributable to channelling (Figure 2.11). To the south of the Craven Fault transition zone the Formation passes laterally into the Warley Wise Formation (Figure 2.11). In the main parts of the Lancaster Fells Basin, the Formation is present in the Whitmoor-1 borehole and has been recorded in the Lune Valley by Moseley (1954). However, it does not appear to extend as far as the northern margin of the Basin: the Formation is absent in the Lancaster area and on the northern margin of the Ingleton coalfield.

Lithology: The lower parts of the Formation consist of pebbly, coarse or granule grade sandstones with abundant feldspar. Kaolinite cements are very well developed. The sandstones are usually moderately well to well-sorted and have

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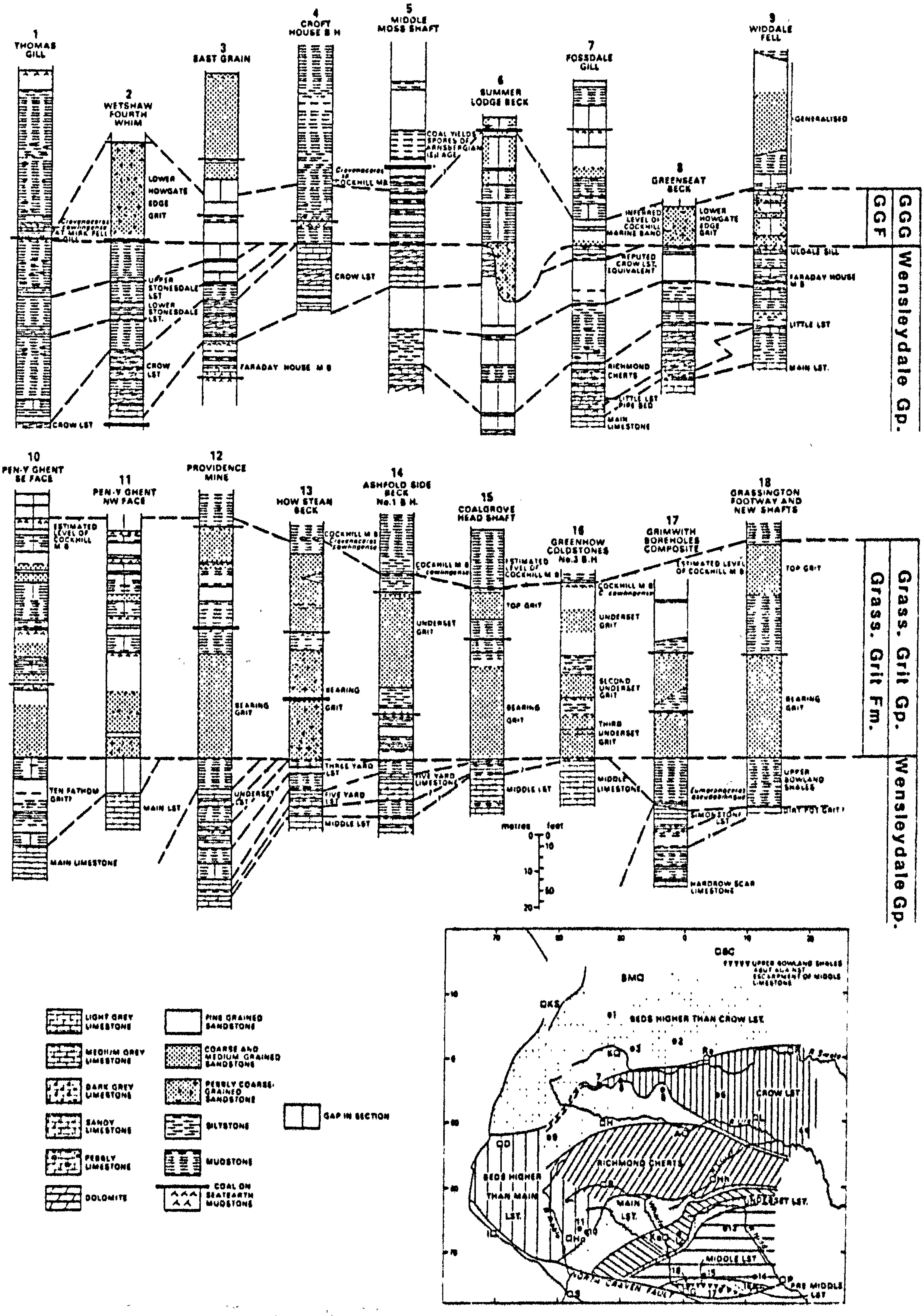
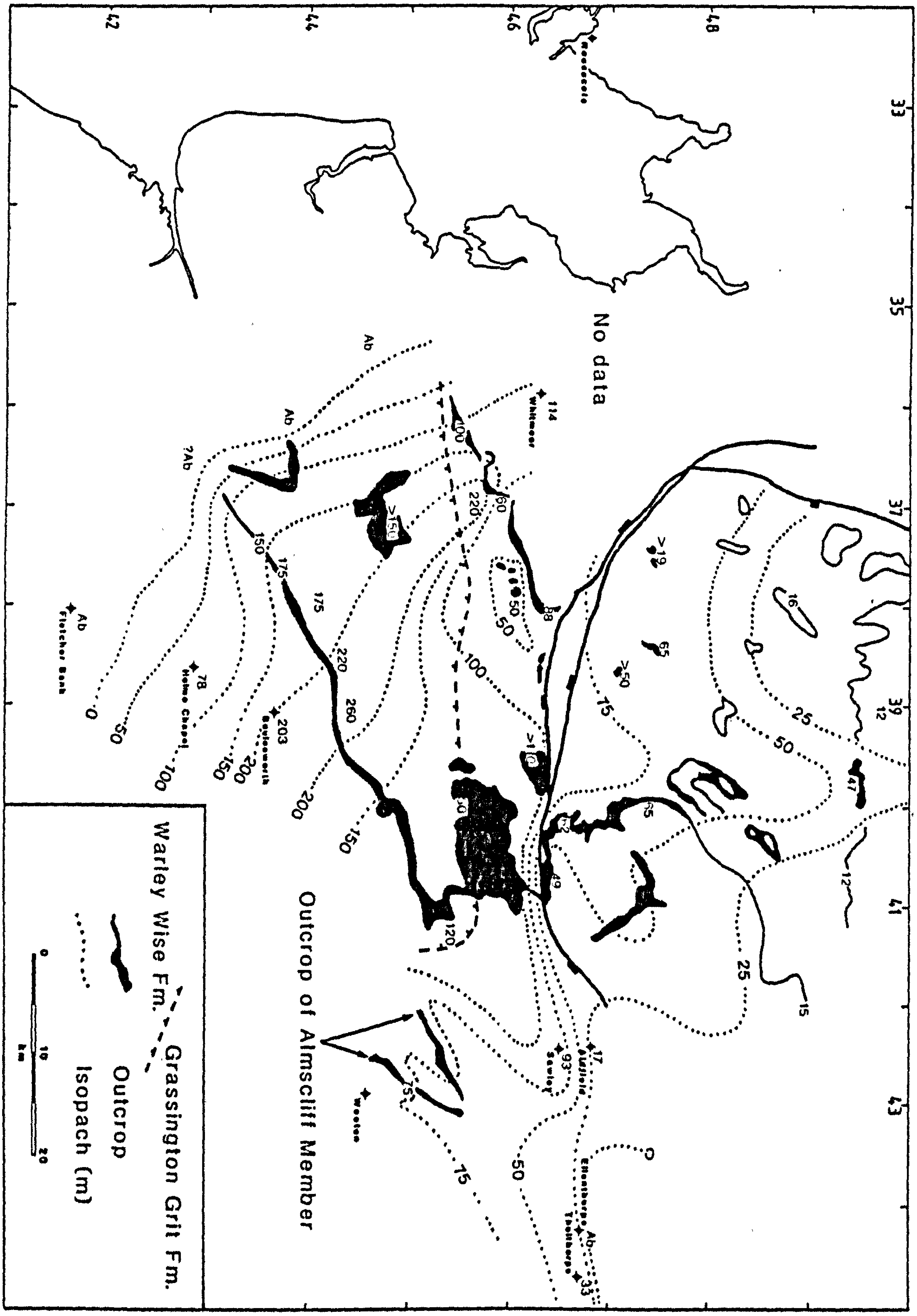
Figure 2.10: Stratigraphic sections in the Grassington Grit Formation, from Dunham and Wilson (1985). The sections demonstrate the lateral variability of the of the Formation and the local names given to some of the lithological units. The map shows the location of the sections and the subcrop beneath the Intra-E_{1c} unconformity.

GGG Grassington Grit Group GGF Grassington Grit Formation



1 THOMAS GILL
2 WETSHAW FOURTH WHIM
3 EAST GRAIN
4 CROFT HOUSE B H
5 MIDDLE MOSS SHAFT
6 SUMMER LODGE BECK
7 FOSSDALE GILL
8 GREENSEAT BECK
9 WIDDALE FELL
LOWER HOWGATE EDGE GRIT
UPPER STONESDALE LST
LOWER STONESDALE LST.
CROW LST
FARADAY HOUSE M B
COCKHILL M B
COAL YIELDS SPORES OF ARNSBERGIAN (E) AGE
INFERRED LEVEL OF COCKHILL MARINE BAND
REPUTED CROW LST. EQUIVALENT
RICHMOND CHERTS
LITTLE LST
PIPE BED
MAIN LIMESTONE
GENERALISED
ULDALE SILL
FARADAY HOUSE M B
LITTLE LST
MAIN LST.
GGG
GGF
Wensleydale Gp.
10 PEN-Y GHENT SE FACE
11 PEN-Y GHENT NW FACE
12 PROVIDENCE MINE
13 HOW STEAN BECK
14 ASHFOLD SIDE BECK No.1 B H.
15 COALGROVE HEAD SHAFT
16 GREENHOW COLDSTONES No.3 B.H
17 GRIMWITH BOREHOLES COMPOSITE
18 GRASSINGTON FOOTWAY AND NEW SHAFTS
ESTIMATED LEVEL OF COCKHILL M B
TEN FATHOM GRIT?
MAIN LST
MAIN LIMESTONE
BEARING GRIT
UNDERSET LST
THREE YARD LST
FIVE YARD LST
MIDDLE LST
UNDERSET GRIT
FIVE YARD LIMESTONE
TOP GRIT
SECOND UNDERSET GRIT
THIRD UNDERSET GRIT
MIDDLE LIMESTONE
SIMONSTONE LST
HARDROW SCAR LIMESTONE
UPPER BOWLAND SHALES
DIRT POT GRIT?
Grass. Grit Gp.
Grass. Grit Fm.
Wensleydale Gp.
metres
feet
LIGHT GREY LIMESTONE
MEDIUM GREY LIMESTONE
DARK GREY LIMESTONE
SANDY LIMESTONE
PEBBLY LIMESTONE
DOLOMITE
FINE GRAINED SANDSTONE
COARSE AND MEDIUM GRAINED SANDSTONE
PEBBLY COARSE-GRAINED SANDSTONE
SILTSTONE
MUDSTONE
COAL ON SEATEARTH MUDSTONE
GAP IN SECTION
UPPER BOWLAND SHALES ABUT AGAINST ESCARPMENT OF MIDDLE LIMESTONE
BEDS HIGHER THAN CROW LST.
CROW LST
RICHMOND CHERTS
BEDS HIGHER THAN MAIN LST.
MAIN LST.
UNDERSET LST.
MIDDLE LST
PRE MIDDLE LST
NORTH CRAVEN FAULT


No data
Outcrop of Almscliff Member
Grassington Grit Fm.
Warley Wise Fm.
Outcrop
Isopach (m)

a cleaner appearance than those of the Pendle Grit Group. Large and small scale soft-sedimentary deformation structures, particularly destratification due to water escape, are characteristic of the lower parts of the Formation. Higher in the Formation, and also northwards, finer lenticular sandstones set in micaceous mudstones and siltstones are predominant. Thin coals and seat earths are present at various levels in the Formation.

Boundaries: Across all of its outcrop the base of the Formation is placed at the intra-E_{1c} unconformity. In some places the unconformity is subtle or inferred by dip discrepancies with the underlying Pendle Grit Group sediments (Arthurton *et al.*, 1988). The top of the Formation is taken at the base of the Cockhill Limestone or the *C. cowlingense* marine band, where present. In the Settle area, where the top of the Formation is not exposed, the boundary is placed at the top of the last feature forming sandstone (Moseley, 1956; Arthurton *et al.*, 1988). At the south-western end of the Bowland High the top of the Formation is placed at the base of the Tarnbrook-Wyre marine beds (i.e. the *C. cowlingense* marine band).

Grassington Grit Formation, Almscliff Member.

Name: The name is taken from a prominent landmark, Almscliff Crag, which forms the most significant exposure in the Member.

Type section: The most complete vertical section through the Member lies immediately above the type section of the Pendle Shale Formation at Oak Beck, Harrogate (SE2754 5436 to SE2755 5465).

Reference section: Almscliff Crag (SE490 268) provides an excellent three-dimensional section in the main part of the Member.

Figure 2.11: Grassington Grit Group isopachs. The north-south trending "thick" on the Askrigg Block is caused by infill of a large channel at the base of the Group. The sandstone filling this channel was given the local name Lower Howgate Edge Grit. The north-west to south-east trending "thick" in the Warley Wise Formation is thought to be fault controlled. Data from measured sections, Dunham and Wilson (1985), Wilson (1960), Arthurton *et al.* (1988) and B.G.S. maps.

Thickness and Distribution: The Almscliff Member is only present in the Eastern Bowland Basin where it forms shoe-string sand bodies with a north-east to south-west orientation (Figure 2.11). Within these sand bodies the thickness is locally extremely variable due to the development of syndepositional growth faults (See Chisholm ,1981 and Chapter 4). Maximum thickness of the Member is around 70m, rising to 100m in growth-faulted sections (Chisholm, *op cit.*).
Lithology: Pebbly, coarse sandstone with well-developed cross-bedding is the dominant lithology. The sandstones are well sorted, feldspathic and have strong kaolinite cements. Rafted logs are abundant in the Member.
Boundaries: The base is placed at the sharp, erosive contact at the bottom of the sand body. It is not clear whether this erosion surface represents the intra-E_{1c} unconformity or whether it is simply due to erosion at the base of a fluvial channel. The top of the member is placed at the C. *cowlingense* marine band, although this is not currently exposed. The top of the member is mapped by a feature which corresponds to the top of the sand body.

Grassington Grit Formation, Mirkfell Member.

The Mirkfell Ganister, as defined by Rowell and Scanlon (1957), is here renamed the Mirkfell Member and is provisionally referred to the Grassington Grit Group. Being a thin, ganisteroid sandstone, the Member has more in common with the Yoredale facies rocks of the northern Askrigg Block. However, as the unit is the feather edge of a Grassington Grit Formation sandbody it seems sensible to include the Member within this Formation (See Figure 2.10, Column 1).

Grassington Grit Group, Warley Wise Formation.

Name: From Warley Wise Farm north of Lothersdale where the Formation was first described. The name follows Bray's (1927) original usage of Warley Wise Grit. This name has historical precedence over other informal names and is the most widely known name amongst the geological community.
Type section: The are no complete sections in the thickest parts of the Formation along the Pendle Monocline. The original sections described by Bray (1927) are now poorly exposed. Consequently, no type section is defined for this Formation.
Reference sections: Several reference sections are necessary because of the lack of a type section and due to the lateral variation within the Formation. Along the

Pendle Monocline good vertical sections are exposed in the River Ribble (SD6762 3587 to SD6755 3583) and in an adjacent deep gully to the north (SD6822 3725 to SD6820 3750). Further north-east, representative sections in the lower and middle parts of the Formation are exposed at Churn Clough Reservoir (SD7865 3840) and at Noyna Hill (SD902 426) respectively. Good sections within the lower parts of the Formation are also seen in Waddington Fell Quarries (SD716 475)

Thickness and distribution: The Formation extends southwards from the Askrigg Block transition zone out into the Bowland Basin (Figure 2.11). It reaches its greatest development in the area of Noyna Hill where the thickness exceeds 260m. Further south-west along the Pendle Monocline the Formation is thinner and on Longridge Fell it fails completely: topographic features formed by the sands in the Formation die out to the south-west on the Fell (Bridges, pers. comm.). As previously mentioned, a similar south-westerly failing of the Formation can be seen around Grizedale in the Lancaster Fells.

Lithology: Pebbly sandstones forming massive, thick-bedded units predominate in the lower parts of the formation. In the middle parts of the formation these pebbly grits form thick, laterally discontinuous beds with a high sedimentary dip. Cross-bedded units become common upwards as the grainsize decreases. Carbonaceous, micaceous shales and siltstones are present in the upper half of the Formation, some with very extensive trace faunas (Baines, 1977; Eager *et al.*, 1985).

Boundaries: The base is taken at the level of the first pebbly sandstones above the Pendle Shale Formation: this is usually a sharp boundary. The top is sharp in the more distal areas but is gradational close to the sediment source (See Chapter 4). In the Lancaster Fells the top is placed at the base of the Tarnbrook-Wyre marine beds (the C. *cowlingense* marine band), if they are developed. When these beds are not present, the top of the Formation is placed below a coarsening-upwards sequence found at the base of the overlying Roeburndale Formation (Arthurton *et al.*, 1988; Croxford, pers comm.). At the western end of the Pendle Monocline the top of the Formation is sharp and is placed at the top of the last sandstone rib exposed in the Ribble (See reference sections above). Around Lothersdale the Formation passes gradationally up into the Bradley Formation (q.v.). Following Baines (1977), the boundary between the formations is placed at the base of the first sandstone unit with well-developed wave formed sedimentary structures.

Comments: The Warley Wise Formation is essentially a thicker version of the Grassington Grit Formation but with a package of submarine mass-flow deposits at its base. The decision to define two separate formations is arbitrary but is justified in that it highlights the continuing affect of Block-Basin controls on the stratigraphy and sedimentology of the E_{1c} interval.

Grassington Grit Group, Bradley Formation.

This unit is not formally defined herein as it probably lies outside E_{1c} (see 2.3) and, therefore, has not been studied by the author. However, lithostratigraphically the Formation clearly belongs in the Grassington Grit Group: it represents continued sedimentation from the Grassington Grit Group sediment source during the end E_{1c} transgression. Essentially, the unit consists of micaceous sandstones and siltstones with abundant wave induced sedimentary structures. These lithologies represent marine reworking of a retreating fluvial system. North-eastwards, towards the sediment source, the grainsize increases and the unit becomes very similar to the underlying Warley Wise Formation (Stephens *et al.*, 1953). A full description of the Bradley Formation is given by Baines (1977).

2.5 Summary.

The formalised stratigraphy described above represents a compilation of many author's contributions undertaken with the benefit of a sedimentological understanding of the various E_{1c} stratigraphic units. The new nomenclature is an advance over previous schemes in that all the units, including the mudstone sequences, are formally named. Also, as the nomenclature has been derived from study of E_{1c} stratigraphic units over a wide geographical area, it reflects sedimentological links between units which may only be recognizable at such a large scale. This formalised stratigraphy, based as it is on sedimentological considerations, forms a sensible framework within which the "dynamic" stratigraphy of the E_{1c} basin-fill sequence can be discussed.

CHAPTER THREE

SEDIMENTOLOGY OF THE PENDLE GRIT GROUP.

3.1 Introduction.

The formations and members of the Pendle Grit Group are distinct stratigraphic units recognised on the basis of the facies associations which they contain. Consequently, the stratigraphic units also represent high-level sedimentological units which reflect the temporal and physical extent of depositional environments within a depositional system. This chapter presents a facies analysis of the sediments within the various stratigraphic units of the Pendle Grit Group and shows how the distribution of these units can be understood in terms of an evolving submarine fan system.

The term facies, or lithofacies, is commonly defined along the lines of "*a body of rock with certain specified attributes that distinguish it from other rock units*" (Leeder, 1982). All definitions of the term stress that a facies should be defined on purely descriptive grounds. This is a straight-forward task when stratigraphic sequences contain lithological and sedimentological units with distinct attributes. However, in the Pendle Grit Group, where the sedimentological units form part of a continuously variable spectrum of deposits, subjective decisions have had to be made about where to place facies boundaries. Ultimately, these decisions were based on a knowledge of the facies associations developed within the Group and on interpretations of the depositional processes operating in the system. The facies described below, then, represent convenient "pigeon-holes" for groups of rocks with similar sedimentological features and which may have been depositied by similar processes. As discussed, the facies boundaries are arbitrary: the descriptions attempt to bring out the gross variability within individual facies and the relationships between facies. Section 3.3 shows how the facies can be related to each other in terms of process and explains why mass-flow facies must form a continuous spectrum of deposits.

3.2 Facies descriptions and interpretations.

At first sight the Pendle Grit Group sandstones appear to show little in the way of facies variation: everywhere the sandstones are thick or medium-bedded, usually featureless and generally monotonous. However, in detail there are a series of sedimentological features which allow subdivision into descriptive facies. Chief among these are:

1) The thickness of depositional units.
2) The grainsize and the nature of size-grading within beds (if present).
3) The presence or absence of traction structures within beds.
4) The sand/shale ratio of individual depositional units.
5) The lateral extent of lithological units.

These features are exactly the same as those used by Mutti and Ricci Lucchi (1972) to erect a facies scheme for the excellently exposed turbidite sequences of the Italian Apennines. This scheme, summarised in Table 1, is essentially simple and is widely applicable to mass-flow systems. The facies scheme is well known and has been adapted by many authors to suit the specific requirements of individual systems. Alternative facies schemes have been applied to mass-flow systems but many are too process based to be useful (e.g. Carter, 1975; Postma, 1986). More recent schemes have expanded the range of facies distinguished. The scheme of Pickering *et al.* (1986), for example, has a three-level hierarchy: facies classes, facies groups and individual facies. The facies classes are based on grain-size and are subdivided into facies groups based on whether the sediments are "organised" or" disorganised" (i.e. whether or not they contain grading or traction structures). The individual facies are based on detailed variations in the type of structures developed and on the sand/shale ratio of depositional units (Figure 3.1). This scheme is claimed by the authors to be more universal in its application than are former schemes. However, the scheme involves a proliferation of facies (38 as opposed to 14 in the Mutti and Ricci Lucchi scheme) which, in the current authors opinion, serves only to obscure the basic similarity and continuity between the facies. The great advantage of the Mutti and Ricci Lucchi scheme is that it encompasses the wide variation within turbidite systems without an overabundance of arbitrary facies boundaries.

	FACIES A		Very thick and very coarse-grained, absence of tractive structure
	Subfacies	A_1:	"Organized" Cgl, pbly ss, ss
		A_2:	"Disorganized" Cgl, pbly mdst, "slurried beds"
	FACIES B		Thick, medium- to coarse-grained, inclined plane laminae for base upward
	Subfacies	B_1:	Relatively continuous beds, 30-200 cm thick
		B_2:	Lenticular, wedging, top molded by dunes and ripples, 20-50 cm thick
	FACIES C		Complete graded Bouma sequence beds
	Subfacies	C_1:	Coarse- to fine-grained sand, Cs-tail grading, T_{ace} and T_{ac}
CLASSIC PROXIMAL		C_2:	Medium- to fine-grained sand, distribution grading, T_{abcde}, T_{abce}', T_{abde}
	FACIES D		Base-missing graded Bouma sequence beds
	Subfacies	D_1:	T_{b-e}, T_{c-e}, T_{de}, ss:sh > 1, 3-40 cm thick
CLASSIC DISTAL		D_2:	T_{b-e}, T_{c-e}, T_{dc}, ss:sh < 1, 30-150 cm thick
		D_3:	T_e mudstone only, 3-200 cm thick
	FACIES E		Thin bedded, medium- to coarse-grained ss with discontinuous sh partings, high ss:sh ratio, irregular geometry, ungraded, high-angle cross-bedding, sharp tops, 3-20 cm thick
TRACTION LAG DEPOSITS			
SLIDES & SLUMPS	FACIES F		CHAOTIC deposits except debris flows (A_2) from gravity sliding and slumping
	FACIES G		Pelites from hemipelagic deposition and "rainfall"

Table 1: Summary of the Mutti and Ricci Lucchi (1972) turbidite facies scheme prepared by T.H. Nilsen. (From Nelson and Nilsen, 1984).

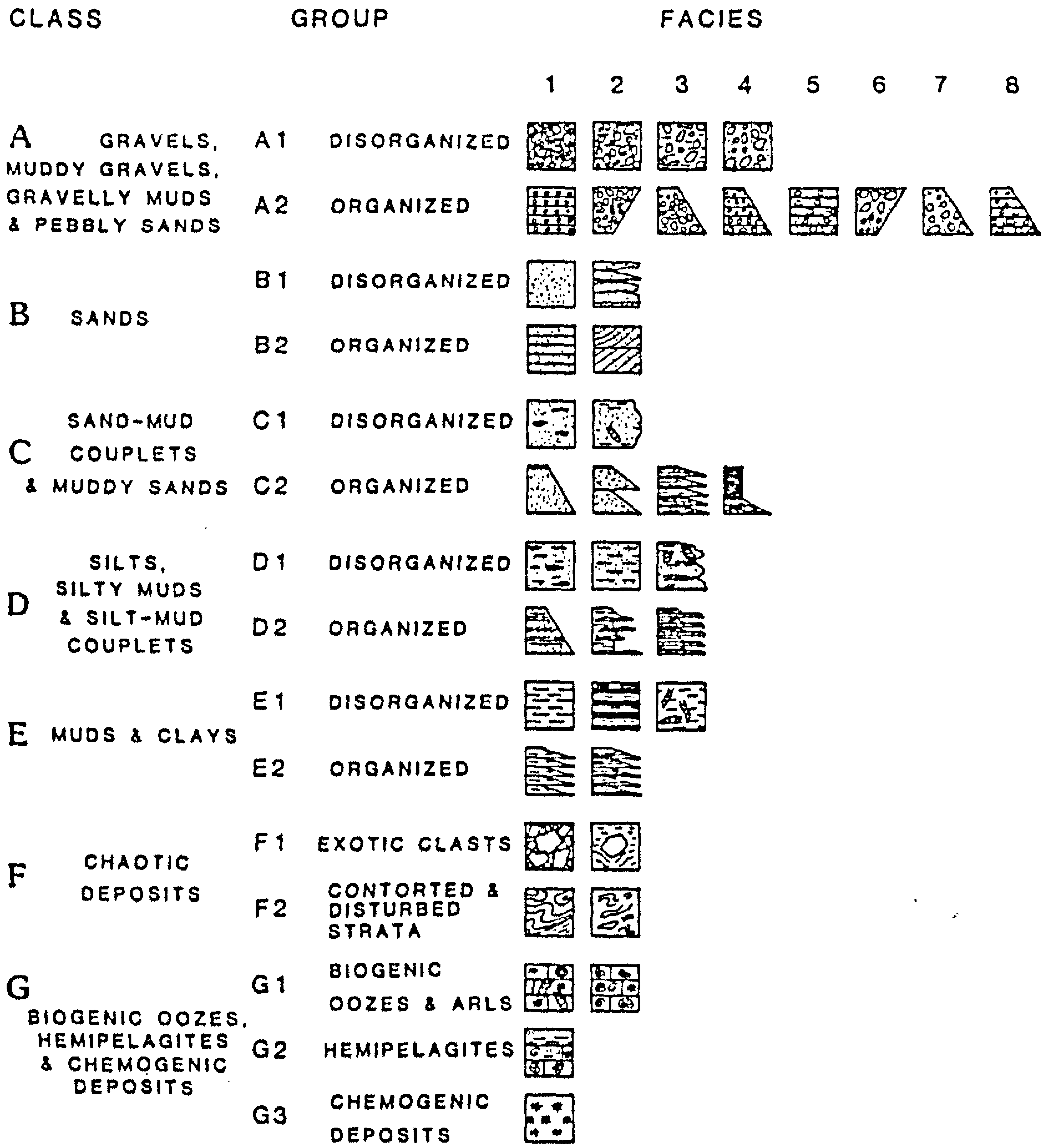


Figure 3.1: Summary of the facies scheme of Pickering *et al.* (1986). Notice the basic similarity at Facies Class level with the Mutti and Ricci Lucchi scheme (Table 1). The large number of individual facies into which the classes are split tends to obscure the continuity of the facies spectrum in turbidite systems. For example the differences between facies A2.4 and A2.8 and between facies C2.1 and C2.2 are minimal.

In the light of the above discussion, the facies descriptions in this work are based around the facies scheme of Mutti and Ricci Lucchi. The gross facies divisions, namely A to G, remain exactly as in the original scheme but the facies-types are redefined because of the recognition of features which have not previously been described in mass-flow systems. In order to avoid confusion with the Mutti and Ricci Lucchi scheme letters rather than numbers have been used in the subscripts for the facies-types.

3.2.1 Facies A.

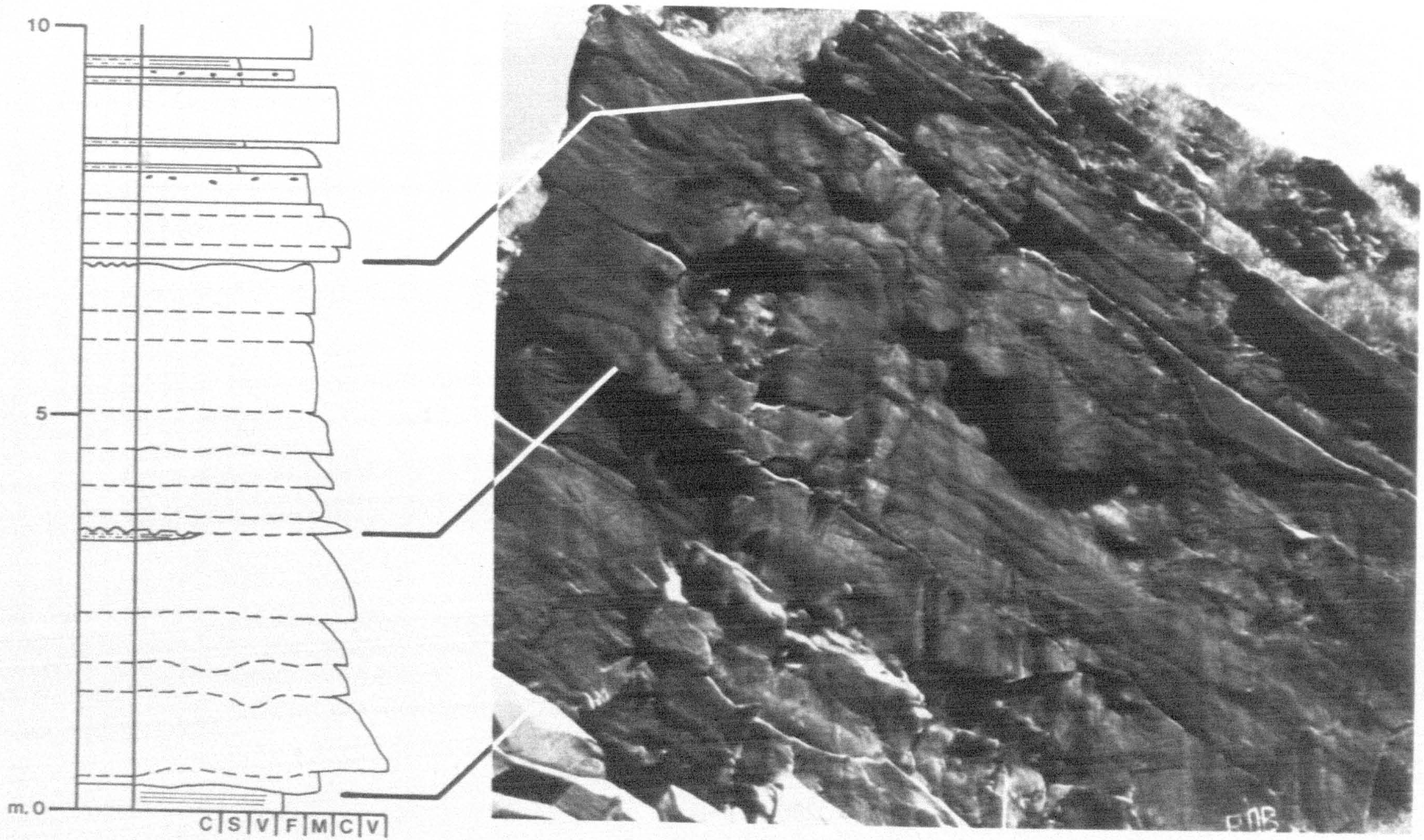
Description
Facies A forms approximately 25% by volume of the Pendle Grit Formation. It is characterised by medium to very thick-bedded sandstones which show abundant evidence of amalgamation. Grainsize varies from pebbly sandstone to fine grade, the coarser grades being predominant. Thicker beds are usually coarser but exceptions to this rule are common. Lateral continuity of the facies is generally low. Three facies-types are recognised on the basis of the scale of lateral continuity and the types of internal structures present.

Facies-type A_a: is the most common facies-type in Facies A. It consists of thick to very thick-bedded units of granule to medium grained, poorly sorted sandstones. The base of the facies-type is either sharp or erosional and is frequently modified by post-depositional loading. Maximum bed thickness is around three metres and beds contain a series of amalgamation surfaces which are sub-parallel to the bed boundaries. These surfaces are defined by abrupt changes in the grainsize which may also be picked out by jointing (Plate 1). Often the surfaces are non-planar with an irregular, wavy topography. This can occasionally be proved to result from erosional scour when the surfaces cut weakly developed lamination within the underlying units. Individual amalgamated units within the beds are massive or show coarse-tail grading (Plate 1) and there may be an overall decrease in maximum grainsize between each unit. Mudflakes up to 80cm in diameter and 5cm thick are scattered throughout the beds, frequently concentrating along the amalgamation surfaces. Some large mudflakes have been observed sitting vertically at the top of massive amalgamated units. These types of amalgamation surface compare with those described in similar facies in the Shale Grit of Derbyshire (Walker, 1966. See Figure 3.2). Large *Calamites* impressions are often present.

Plate 1: Thick-bedded, massive, amalgamated sandstones of Facies-type A_a. The sedimentary log is drawn at the same scale as the photograph. Note the almost perfect amalgamation of the beds between the three marked horizons. This style of amalgamation is interpreted as being due to intra-event erosion due to migrating or pulsing flows. The three marked horizons represent longer time breaks between events. The middle horizon passes along its length from a silt-draped wavy top to an amalgamation surface similar to Type 1 of Walker (1966): refer to Figure 3.2. Note the small winged scour fill at the level of the top marked horizon. This scour is illustrated again on Figure 3.16.

Wiswell Moor Quarries, SD753 371. For Key see Enclosure 1.

Plate 1



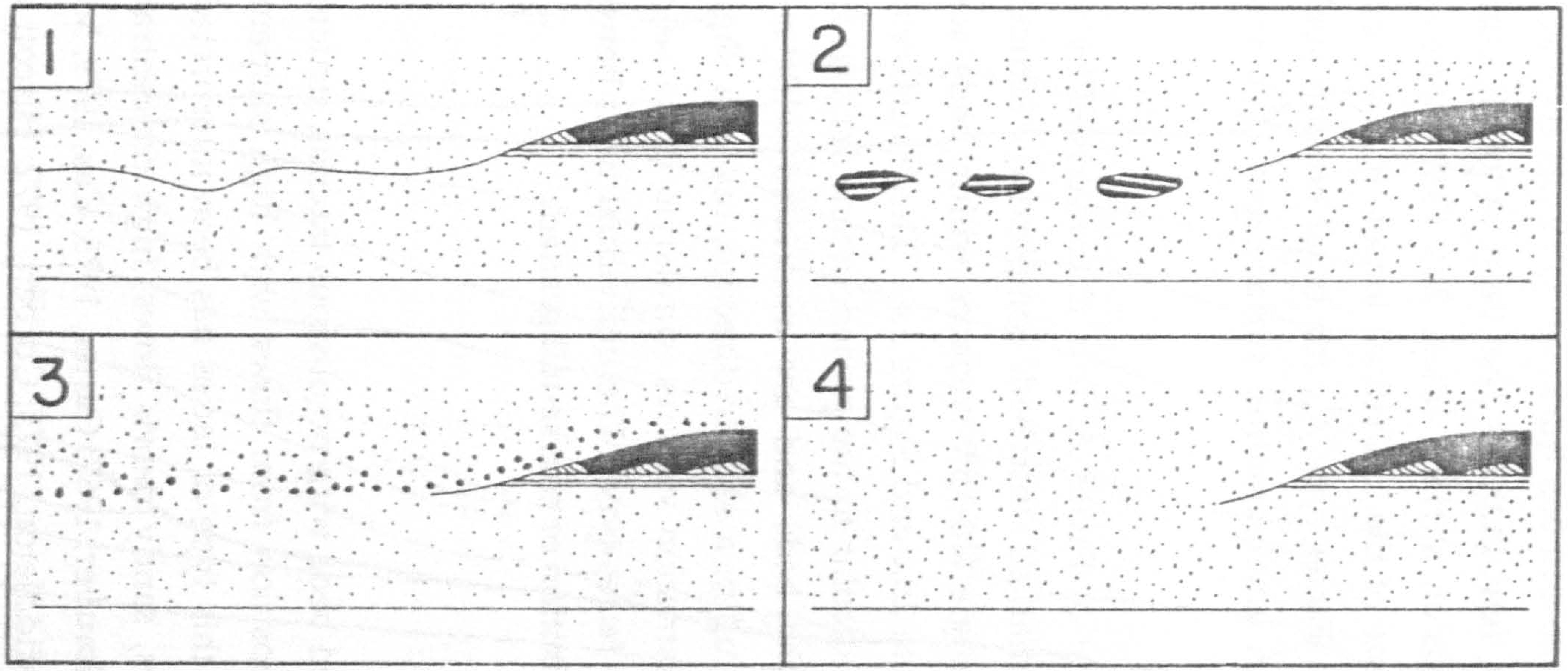


FIG. 8.—Amalgamation of the graded divisions of two successive currents. 1—the line of amalgamation is shown up by a horizontal joint, sometimes with small load casts. 2—the line of amalgamation is shown up by a row of angular or rounded mud flakes. 3—the line of amalgamation is shown up by a change of grain size. 4—the line of amalgamation is invisible due to perfect annealing between the two beds.

Figure 3.2: Types of amalgamation surface recognised by Walker in the Shale Grit (Namurian, R_1) of Derbyshire. All of these types of inter-bed amalgamation have been seen in Facies A deposits in the Pendle Grit Formation. However, a further type of amalgamation surface in which there is no sign of lateral passage into a bedding plane is the most common example in the Pendle Grit Formation. (see Plate 1). These latter surfaces are interpreted as resulting from intra-event erosion, rather than inter-bed erosion. Figure and caption from Walker (1966).

Facies A_a beds are usually laterally continuous on the scale of small outcrops but can be seen to be lenticular in large quarry exposures. In such exposures this facies-type can be seen to grade laterally into Facies B or Facies-type A_b. Additional examples of this facies-type are shown on Figures 3.15, 3.16, 3.19 and Plates 2b, 2c and 7b.

Facies-type A_b: is similar in grainsize and bed thickness to Facies-type A_a but beds are lenticular at a scale of a few metres. The base is often convex downwards and sits on a minor erosion surface (Plate 2a). Centimetre-scale lamination is present at the top of some of the individual lenses. In some cases, particularly when in association with Facies-type A_a, the basal erosion surface can be seen to pass laterally into an amalgamation surface and then into a massive bed (Figure 3.3).

Facies-type A_c: is a rare component of Facies A. It is characterised by very thick-bedded sandstones showing marked intrabed complexity. Structures within this facies-type include low angle (10-15°) surfaces within beds across which grainsize changes abruptly (Plate 2b,c and Figure 3.4). These surfaces dip in a direction which is almost perpendicular to the palaeocurrent derived from erosion structures on the base of the beds and are associated with crude centimetre scale ?cross-bedding dipping in a similar direction but at a slightly higher angle (Figure 3.4). Flat amalgamation surfaces, similar to those seen in facies-type A_a, are also present in this facies-type. The individual units between the various internal surfaces are distribution or coarse-tail graded.

Interpretation

The presence of massive or graded beds with no obvious traction structures is indicative of extremely rapid deposition from flows with high suspended sediment concentrations. Classically, this type of facies has been interpreted in terms of deposition from grain flows, sandy debris flows, high concentration turbidity currents or liquified flows (Stauffer, 1967; Carter, 1975; Cas, 1979; Mutti and Ricci Lucchi, 1972; Skipper and Middleton, 1975; Corbett, 1972; Hiscott and Middleton, 1979). The grain support mechanisms in these gravity driven flows are dispersive pressure, matrix strength, turbulence and upwards flow of pore fluid respectively Lowe, 1982; Middleton and Hampton, 1976. Figure 3.5). Facies A sediments in the Pendle Grit Group are thought to have been deposited from a

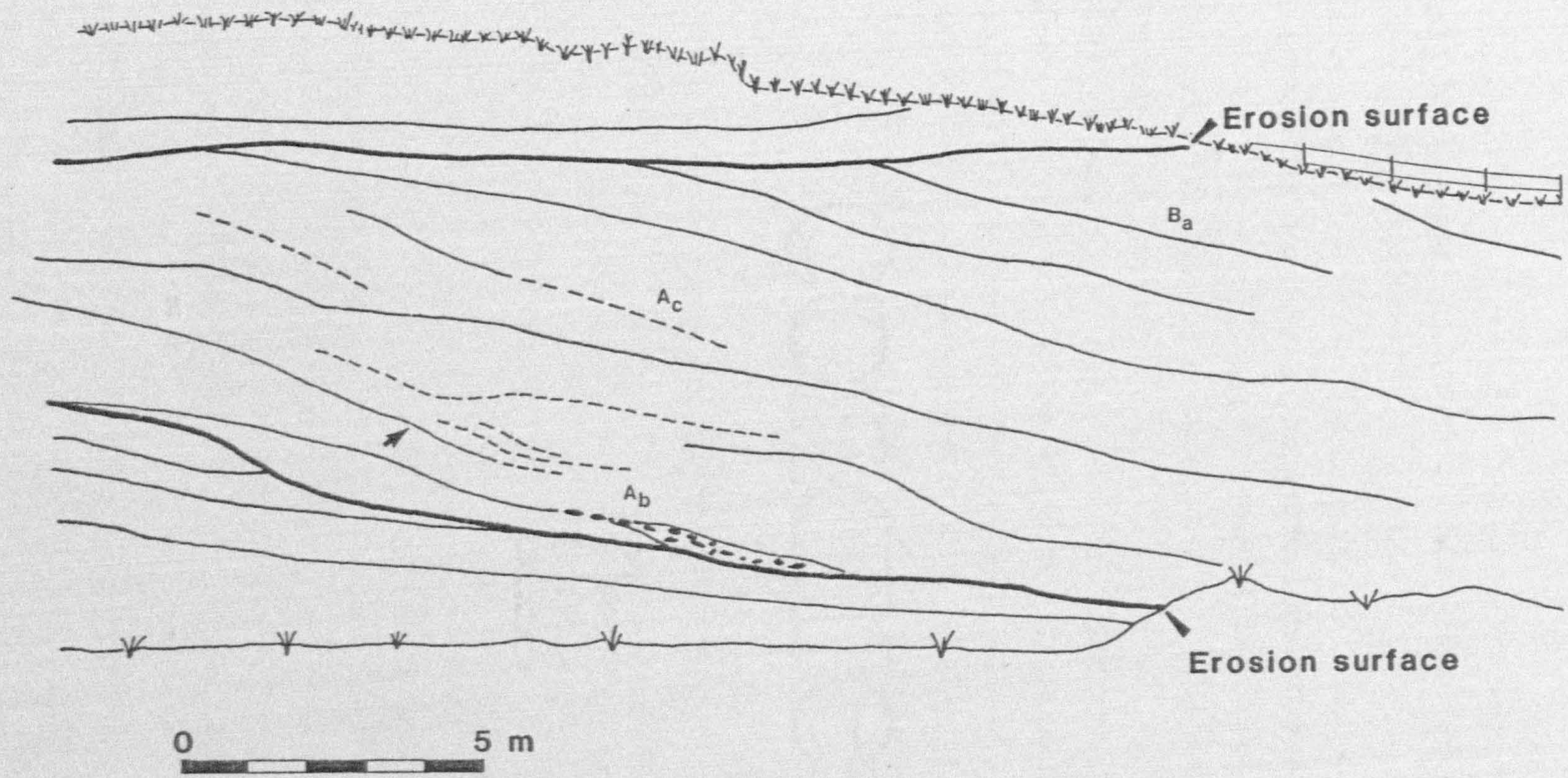


Figure 3.3: Line drawing of Facies-type A_b amalgamation surfaces and scours. The Facies-type A_b bed immediately overlies an erosion surface (Second-order surface: see Section 3.4.1) cutting down into the underlying beds. A small mudflake conglomerate unit is present at the base of the bed. The minor scour surface within the bed (arrowed) passes laterally into an amalgamation surface which then disappears into a massive sandstone. A series of amalgamation surfaces (dashed) are similarly discontinuous. Obviously, there was considerable spatial variation of erosion and deposition rates during deposition of this bed. Note the laterally dipping surfaces in the overlying Facies-type A_c bed. Palaeocurrent is into the page. Salterforth Quarry, Barnoldswick (SD877 452).
Lithology and bedding symbols on this and future line drawings are equivalent to those shown on the Graphic Log Key (Enclosure 1).

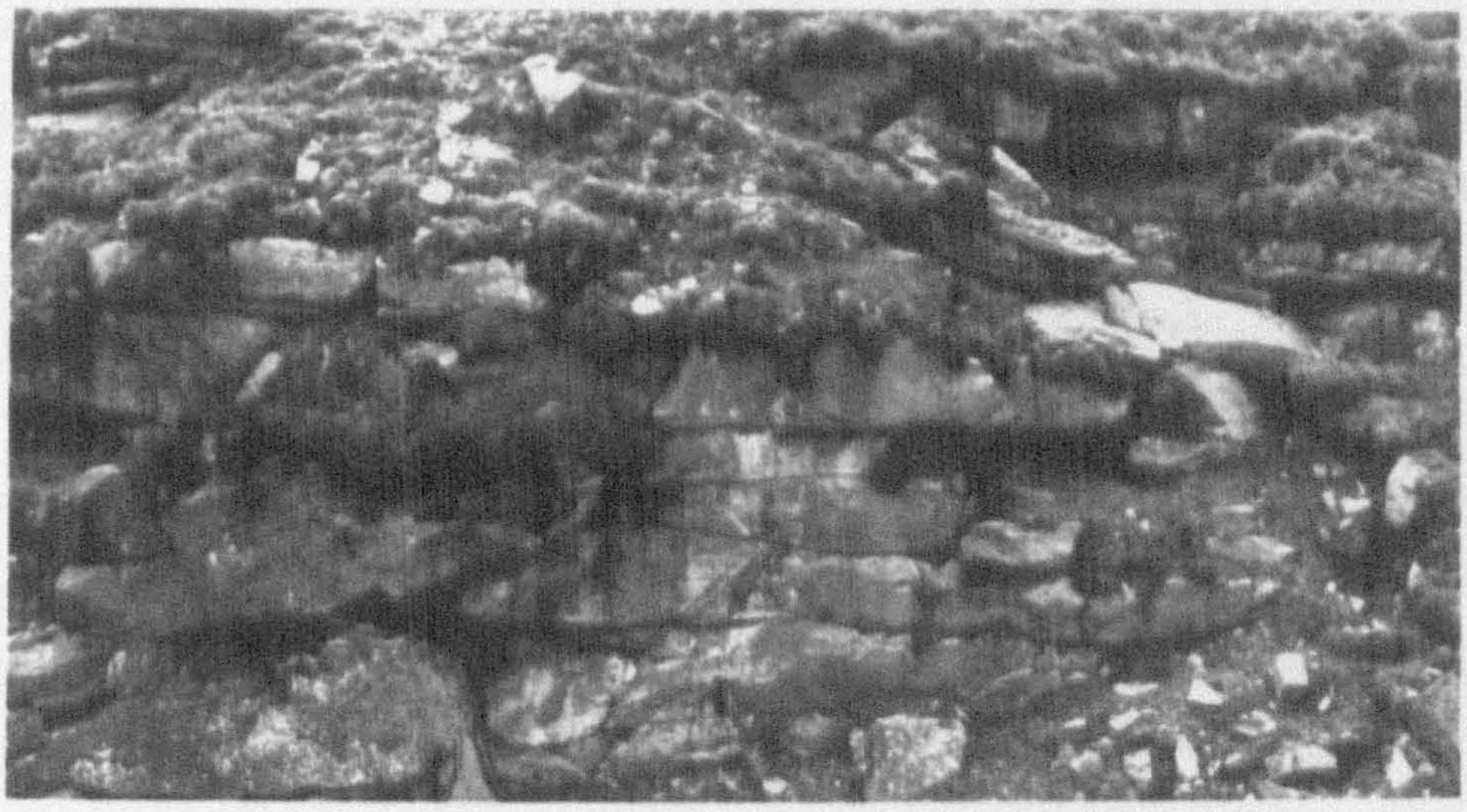
Plate 2: Illustration of the facies-types in Facies A deposits.

a) Facies-type A_b deposits. Note the lenticular nature of the individual beds. Bed above rucksack is 90cm thick. Ogden Clough, Pendle Hill (SD791 406).

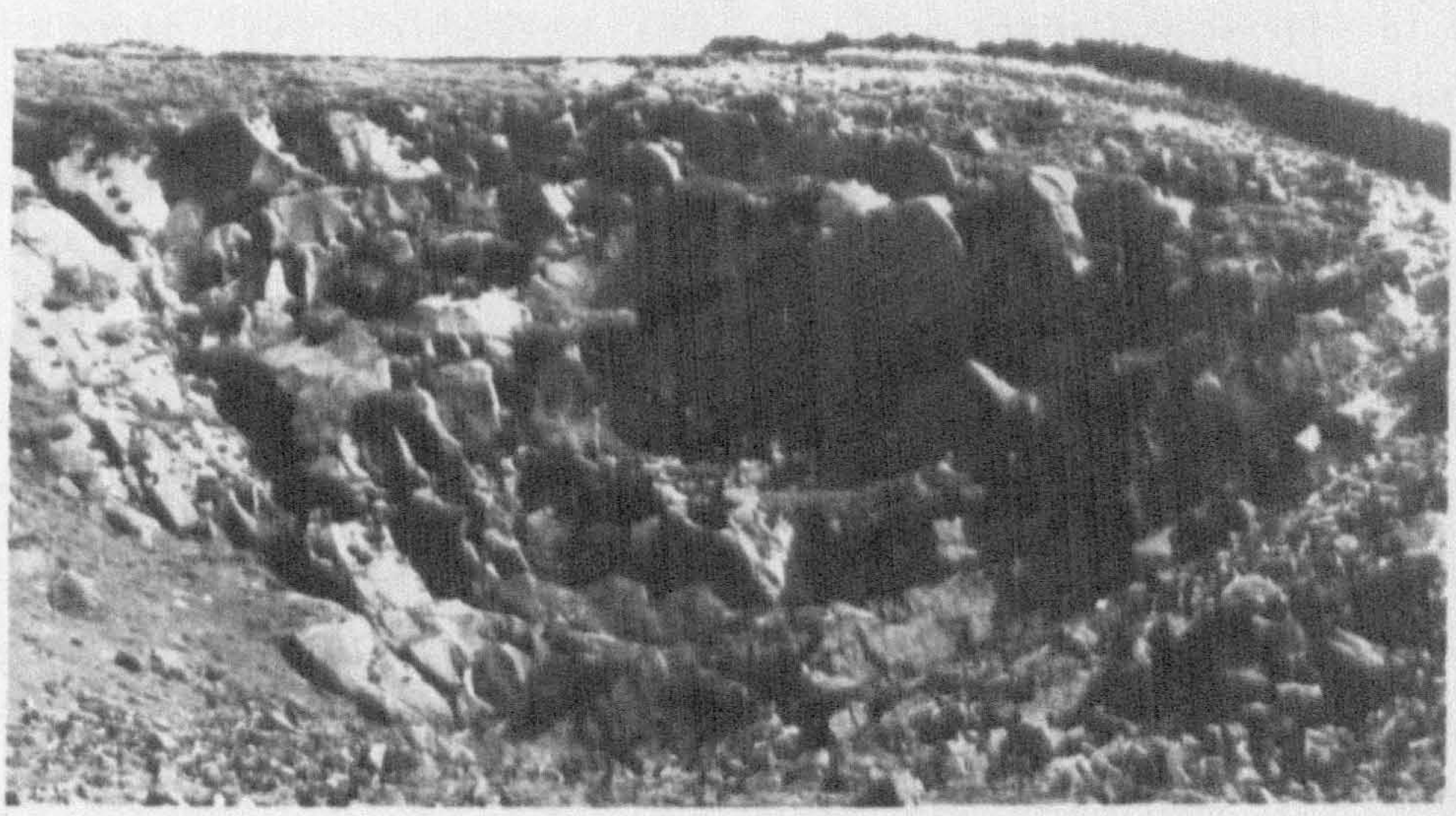
b) Section through interbedded Facies-types A_a and A_c deposits. Quarry face is 25m high. Whitshaw Bank Quarry, SE002 548.

c) Line drawing of (b). Heavy lines show second-order erosion surfaces within the channel complex (See Section 3.4.1 for terminology). Thin lines indicate the bedding surfaces within the second-order channels. Note the dipping amalgamation surfaces (dashed lines) shown in the Facies-type A_c bed and the debris flow deposit (Facies-type F_a at the base of one of the second-order channels. Figure 3.4 is a detailed drawing of the Facies-type A_c bed labelled on this plate.

A



B



C

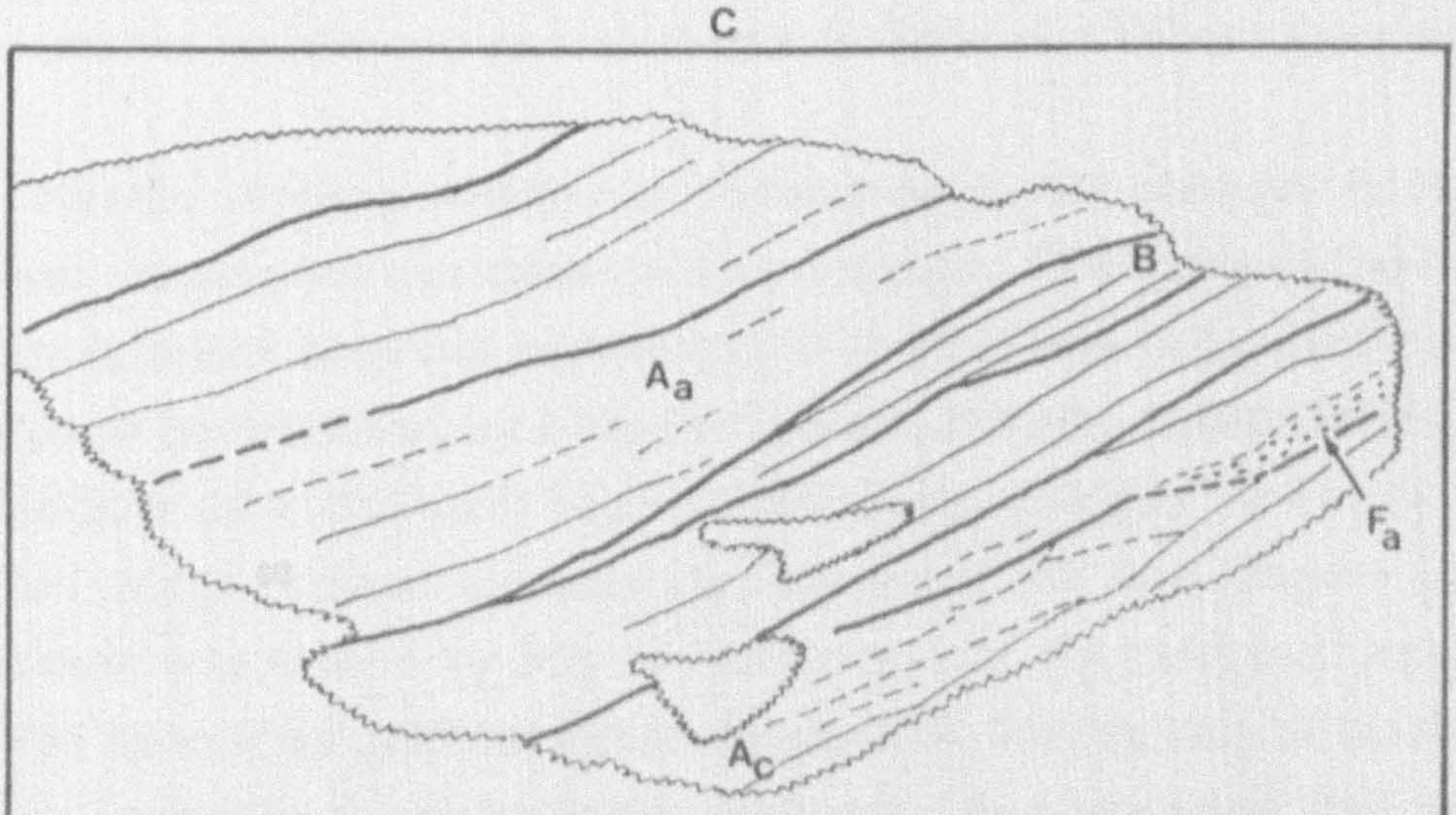


Plate 2

high concentration turbidity current which was sufficiently powerful to support a basal "boundary layer" in the flow (See section 3.3). Grain support in this boundary layer was by one of these other mechanisms. The lateral facies variations and the facies associations of Facies A deposits indicate that the whole high-concentration turbidity current, or at least the thalweg of the flow, was laterally confined.

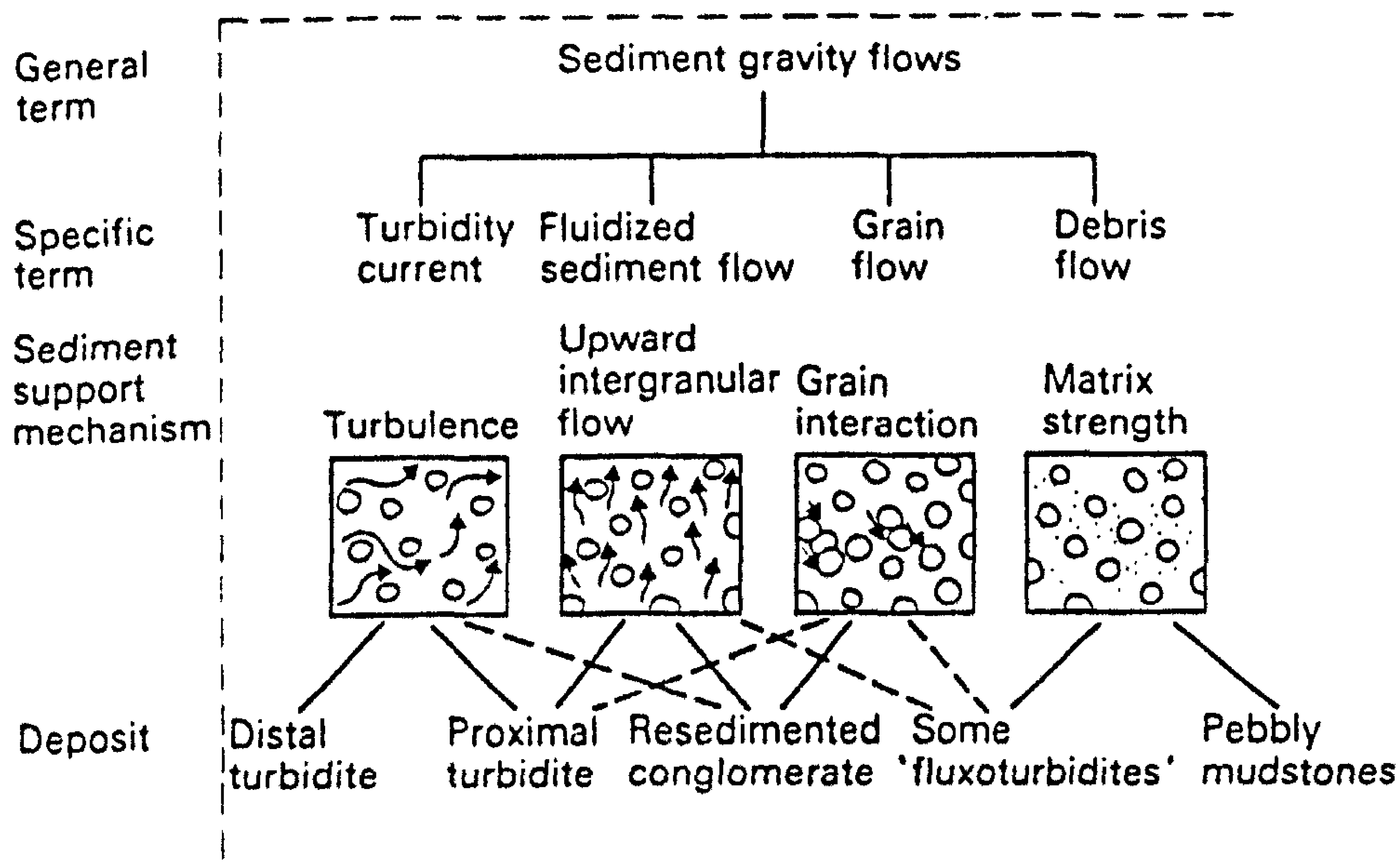
Facies A_a beds without amalgamation surfaces appear to have been deposited very rapidly as a single unit. The occurrence of vertical mudflakes at the top of some of these beds shows that, in some cases, the basal zone of the flow had cohesive strength during the final phases of deposition. The coarse-tail grading in other Facies-type A_a beds suggests rapid deposition from high-concentration, turbulent suspensions which had a liquified (quick bed) "boundary layer" flow (Middleton, 1967). In both cases, deposition of the bed appears to have been a continuous process. In Facies-types A_b and A_c, however, the balance between erosion and deposition seems to have been very close. Lenticular erosion surfaces in Facies-type A_b and the rapid grainsize variations and amalgamation surfaces in Facies-type A_c, suggest spatial and temporal variations in the power of the flow. These variations must have been very rapid and local in order to produce the scour and fill structures seen in Facies-type A_b. By analogy with turbulent systems in river flow (c.f. Leeder, 1982), these structures might result from "sweeps" of high velocity fluid from the main turbulent flow into the basal boundary layer of the flow. Normally, supply of suspended sediment to the boundary layer was such that nett deposition occurred. However, during sweep events the energy balance was tipped in favour of erosion and the scours at the bases of the lenticular beds were cut.

The laterally dipping surfaces in Facies-type A_c are thought to represent longer term, larger scale variations in flow strength. The surfaces have similar geometries to lateral accretion surfaces on in-channel bars in braided rivers and are thought to have formed by a similar process. Initially, sedimentation on the lateral margins of a confined, high-concentration turbidity flow produced a graded bed which thinned towards the thalweg of the flow (Figure 3.6). This sedimentation was caused by loss of competence in the marginal parts of the flow as they moved up-gradient out of the channel. The thalweg of the flow was sufficiently energetic to maintain all grainsizes in suspension and to cause erosion of unconsolidated, cohesionless sand from the channel base. As

Figure 3.4: Detail of intra-bed complexity within Facies-type A_c. This diagram is a detail of the beds labelled A_c and F_a on Plate 2c. The abrupt grainsize changes in the illustration mark perfect amalgamation surfaces within the bed. These surfaces dip at a low angle to the palaeocurrent (which is towards the observer, out of the page). Crude cross-bedding with a similar orientation is present (X) within the bed. At (Y), the small mudflake conglomerate grades laterally into both the cross-stratified unit and the unit above the amalgamation surface. This suggests complex spatial variations in the deposition and erosion rates across the bed surface. Where the erosion surface (Z) cuts through the lower bed the two are perfectly amalgamated: the surface is picked out only by rip-up clasts. Note how the erosion surface has been prevented from cutting down further by a cohesive silty mudstone at the top of the underlying Facies C deposits (V). Even the thinner silty mudstones (e.g. at W) appear to resist erosion. Once these layers are breached, however, the underlying sandy part of the Facies C bed is rapidly eroded. The undercut silty mudstone layer then breaks up into a series of mudflakes.

Whitshaw Bank Quarry, SE002 548.


W
E
Y
X
X
Z
W
V
0
1
2
3m
MUDFLAKES
GRANULES
PEBBLES
RIP-UP CLASTS
AMALGAMATION SURFACE
ABRUPT GRAINSIZE CHANGE



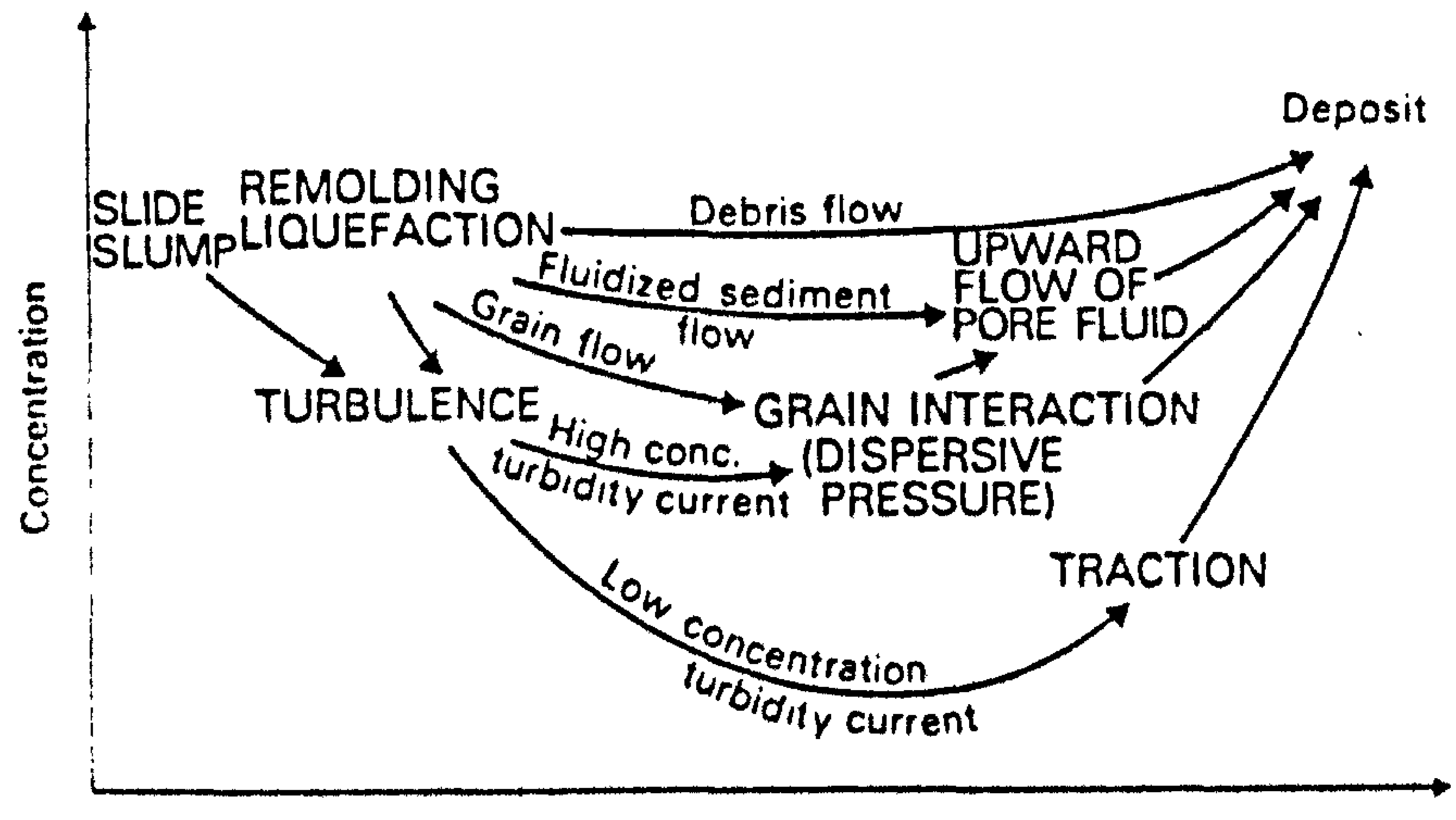


Figure 3.5: Grain support mechanisms in sediment gravity flows, from Middleton and Hampton (1976). Middleton and Hampton described separate classes of gravity driven flows with these different grain support mechanisms. The lower diagram shows how they believed such independent flows might relate in time and space. Mathematical and experimental work have shown that grain and liquified flows require high slopes or high rates of pore-fluid expulsion respectively to maintain sediment in suspension. Hence, they can only exist as independent flows in geologically rare conditions and cannot be responsible for the majority of the coarse facies seen in turbidite systems. The sedimentary structures in Facies A deposits of the Pendle Grit Formation, however, indicate that grain support mechanisms just prior to sedimentation did involve dispersive pressure and/or liquifaction. Consequently, it is proposed that these mechanisms were active *in the basal zone* of the flows responsible for Facies A deposits. (See discussion in Section 3.3).

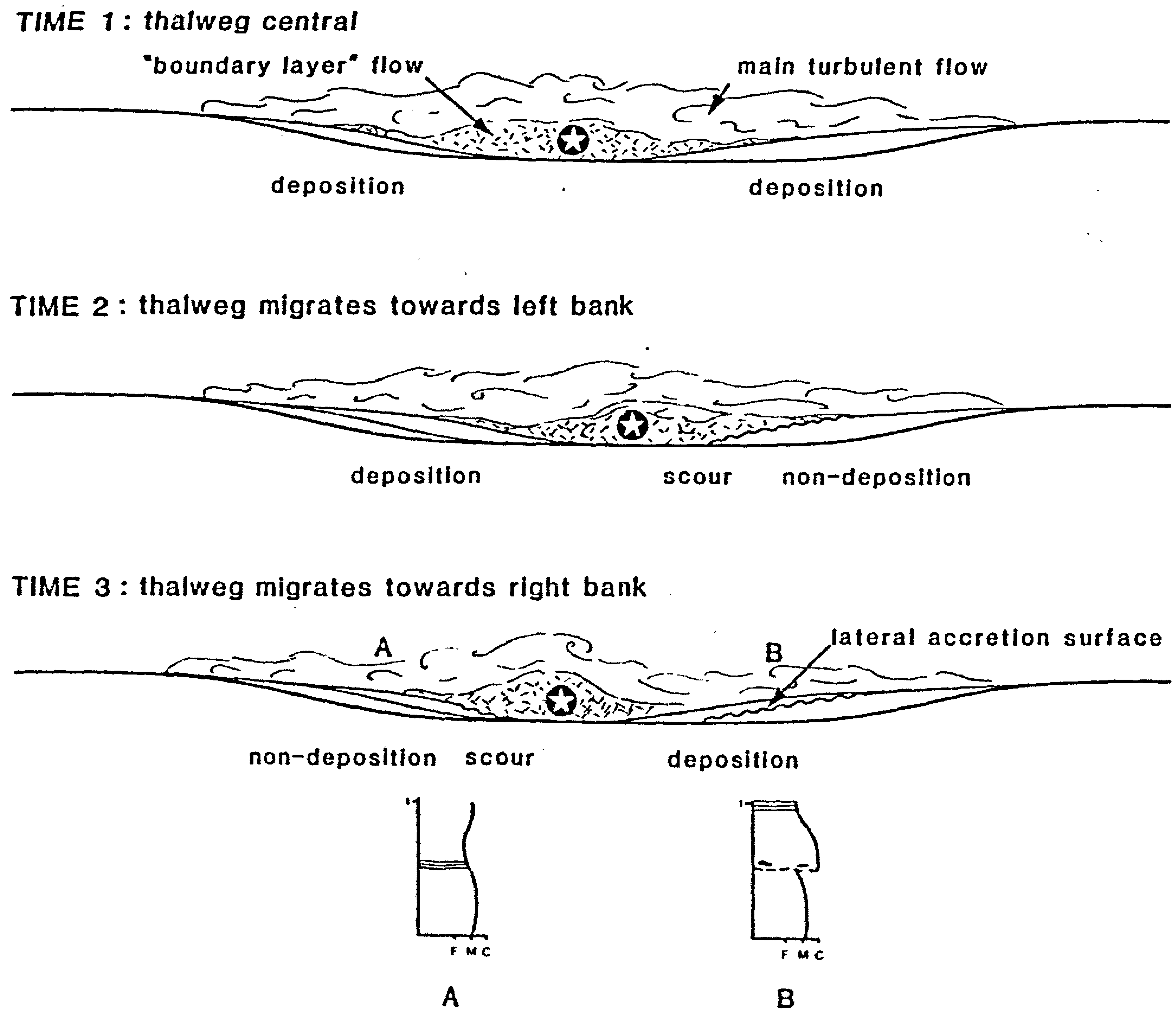


Figure 3.6: Model to explain the origin of lateral accretion surfaces in Facies-type A_c deposits. Flow is towards the observer. For explanation refer to the text. Logs A and B indicate the type of intrabed variation that might result from a laterally migrating flow.

migration of the thalweg took place, so lateral accretion on one bank continued while partial erosion occurred on the other (Figure 3.6). Reversal of the thalweg migration direction would reverse the situation. Repeated cycles of migration could produce the intrabed complexities seen in Facies-type A_c (Figure 3.6). Alternatively, the dipping surfaces might have formed as a result of a "pulsing" flow: deposition of graded beds took place during the lower energy flow periods while the dipping surfaces were cut by entrenchment of the thalweg into its own bed during the high energy flow periods. The former explanation is preferred: the presence of coarse ?cross-bedding (traction carpet shearing?) sub-parallel to the surfaces and clear evidence of lateral migration of larger structures (see Section 3.4) is support for the presence of laterally migrating flows.

The laterally continuous, sub-horizontal amalgamation surfaces seen in Facies-type A_a are, however, interpreted as resulting from long period spatial and temporal variations in flow power during a single depositional event. Some of these amalgamation surfaces clearly represent inter-event erosion and the "welding" together of separate sandstones (Walker 1966. See Figure 3.2 and Plate 1). However, many of the amalgamation surfaces in Facies A deposits probably represent intra-event erosion (Plate 1). The evidence for this is:

1) The thick beds of Facies-type A_a always lie at the base of the Channel Facies Association (See 3.4.1). Consequently, the facies-type was probably deposited by the most powerful and longest flow events. Lateral migration or pulsing of such flows seems intuitively likely.
2) The amalgamation surfaces in Facies-type A_a are very different in nature to the boundaries of overlying Facies B beds of similar grainsize (Compare Plates 1 and 3). These surfaces must, therefore, have a different origin. Walker (*op. cit.*) suggested that this difference was due to variations in the time interval between depositional events: amalgamation surfaces represented short breaks while bedding planes represented longer intervals. Here, it is suggested that the shorter breaks in sedimentation actually occurred during a single flow event rather than between flows.
3) There is evidence from other sedimentological features in Facies A for flow power variations large enough to cause inter-event erosion.

A scaled-up version of the model used to explain the lateral accretion in Facies-type A_c could produce such large-scale amalgamation surfaces. Migration of the thalweg over a large lateral distance, coupled with a higher nett sedimentation rate, would produce almost flat amalgamation surfaces. Alternatively, large variations in flow power during single events could cause repeated phases of deposition and erosion.

In summary, Facies A deposits represent deposition from confined high density turbidity currents. These currents probably had boundary layers in which grain support was by a mechanism other than fluid turbulence. The energy relationships between the boundary layer and the main flow can explain some of the sedimentary features of the facies. Longer term lateral migration and pulsing flow, perhaps overs days rather than hours, can explain the larger features of the facies.

3.2.2 Facies B.

Description

Facies B consists of thick to very thick-bedded sandstones with sub-planar bases and tops. Grainsize varies from granule to fine sand grade and is, on average, less than that of associated Facies A deposits. Crude internal lamination is characteristic of this facies and the deposits tend to have good lateral continuity at the scale of the largest exposures (tens to hundreds of metres). Around 30% of the sands within the Pendle Grit Formation are developed in this facies.

Facies-type B_a: is very similar to a single amalgamated unit in Facies-type A_a but is bounded by good bedding planes or by thin pelitic or carbonaceous fine sandstone laminae (Plate 3). There is a complete gradation between this facies-type and both Facies-type A_a and Facies-type C_a. In Facies-type B_a individual beds are massive or coarse-tail graded. Frequently grading is only present in the top few centimetres of the units. Internally the beds may show crude centimetre scale laminations, usually only visible on very well weathered surfaces. In rare instances isolated sets of low-angle planar cross-stratification are present and more distinct, finer lamination is developed (Plate 3c,d). Dish structures have been recognised in association with all other sedimentary structures.

Plate 3: Facies B deposits.

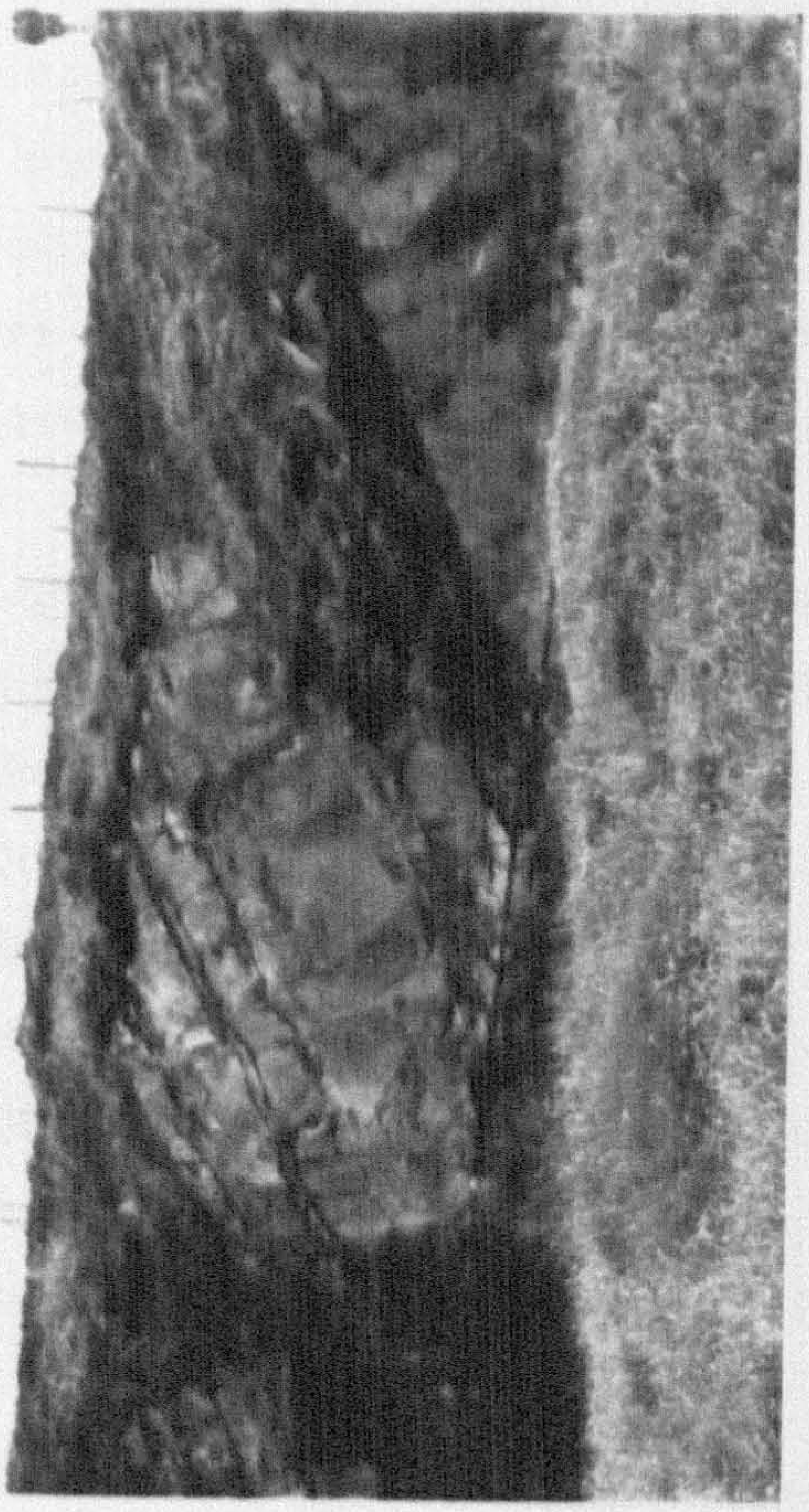
a) Parallel bedded, massive, structureless sandstones of Facies-type B_a. Note that the basal contacts of the beds are slightly loaded. There are no pelitic drapes on the weathered back bedding planes. Maximum bed thickness here is 1.05m. Kemple End, SD690 405.

b) Facies-type B_b beds showing greater variation in thickness between beds. The thickest beds (2.2m) are gradational with Facies-type A_a beds. Hill Top, Longridge (SD612 382).

c) Facies-type B_a bed with well-developed internal structures. The base of the bed has crude thick laminations in granule grade sand. A single set of tabular cross-bedding overlies this, passing to the left into massive sand. The plane-laminated top of the bed is in medium sand: the lamination is caused by mm thick laminae of coarse sand. The bed is 95cm thick in total. Park Farm Quarry, SD629 303. d) More Facies-type B_a beds with well-developed internal structures. To the left of the notebook a set of ?trough cross-stratification is seen at the top of a bed full of pillow structures. Plane-lamination is present within the underlying bed. The clay drape underneath the overhang at the top of the photograph drapes a First-order erosion surface (see Section 3.4.1). Notebook is 20cm long. Longridge Caravan Park, SD615 382.

A

B



C

D

PLATE 3

The bases of Facies-type B_a beds are usually planar or strongly deformed by post-depositional loading. This may also result in deformation of the internal laminae within beds (Plate 3c). These loads are often elongate parallel to the palaeocurrent which suggests that they may have originated as erosional bottom structures (e.g.flutes). Loads develop only when the top of the underlying bed is very fine or when there is a thin clay or silt parting between sand beds.

The tops of Facies-type B_a beds often have a low amplitude, long wavelength wavy appearance in vertical sections. On bedding plane exposures (e.g. Wiswell Moor Quarries, SD753 371) this is seen to result from the moulding of the bed into regularly spaced domes and hollows with amplitudes of 10-15cm. and wavelengths of 1-2m. Mudflake lags in the overlying beds are sometimes concentrated in the hollows. At Nick of Pendle quarries (SD772 385) similar structures are seen but with a higher amplitude/wavelength ratio (Plate 4a,b). At this locality, laminae within the top few centimetres of the bed can be seen to pass over the domes and thicken into the hollows.

Additional examples of this facies-type are shown on Figures 3.15, 3.16 3.18-20 and Plates 7 and 8.

Facies-type B_b: is an unusual facies which does not seem to have been described in any other mass-flow systems. In the Pendle Grit Group sandstones in the Skipton Moor Member are developed entirely in this facies-type. It is rare in other stratigraphic units within the Group. Baines (1977) gave a brief description of the facies-type, referring to it as "ropy weathering sandstone". His mapping in the Skipton area showed that it was found only in channel fills. The facies-type is characterised by thick to very thick-beds of coarse to granule grade sandstone with a weathering pattern that resembles a ropy lava flow (Plate 4c,d). Weathering picks out irregular shaped lenses within the rock which tend to be elongate parallel to palaeocurrents measured in adjacent facies. Discontinuous weathering surfaces within individual lenses have a geometry similar to that of the foresets within a single set of low-angle trough cross-stratification. Individual lenses have a feint grain fabric parallel to the boundaries and may be reverse graded. In general, the overall size of the lenses decreases up through a bed. Vertical or sub-vertical surfaces often join the boundaries of individual lenses forming "staircases" which extend up through much of the bed. Metre scale scours are seen within and at the base of Facies-type B_b beds (Plate 4d).

Plate 4:

a) Dome and hollow structures on the top of a Facies-type B_a bed. Notebook is 20cm long. Nick of Pendle quarries (north quarry), SD772 386.

b) Detail of one dome at the above location. Note the lamination passing over the structure and the thickening of the individual laminae into the hollows. Notebook is 20cm long.

c) Ropy weathering in Facies-type B_b granule grade sandstones of the Skipton Moor Member. This weathering pattern is picking out a weak grain fabric in the rock. Note how some of the weathering surfaces define narrow lenses (best seen in bottom left of photograph) while others appear to subdivide lenses. See text for discussion of these structures. Hammer is 40cm long, palaeocurrent is perpendicular to bottom left face of outcrop. Skipton Moor, SE003 508.

d) Larger scour in Facies-type B_b sandstones. The scour cuts into and is filled by ropy weathering sandstones of identical grainsize (granule grade). The relationship of the "ropes" to the scour shows the former are reflecting an existing sedimentary structure and are not just a weathering feature. Lens cap is 5cm in diameter. Skipton Moor, SD007 510.

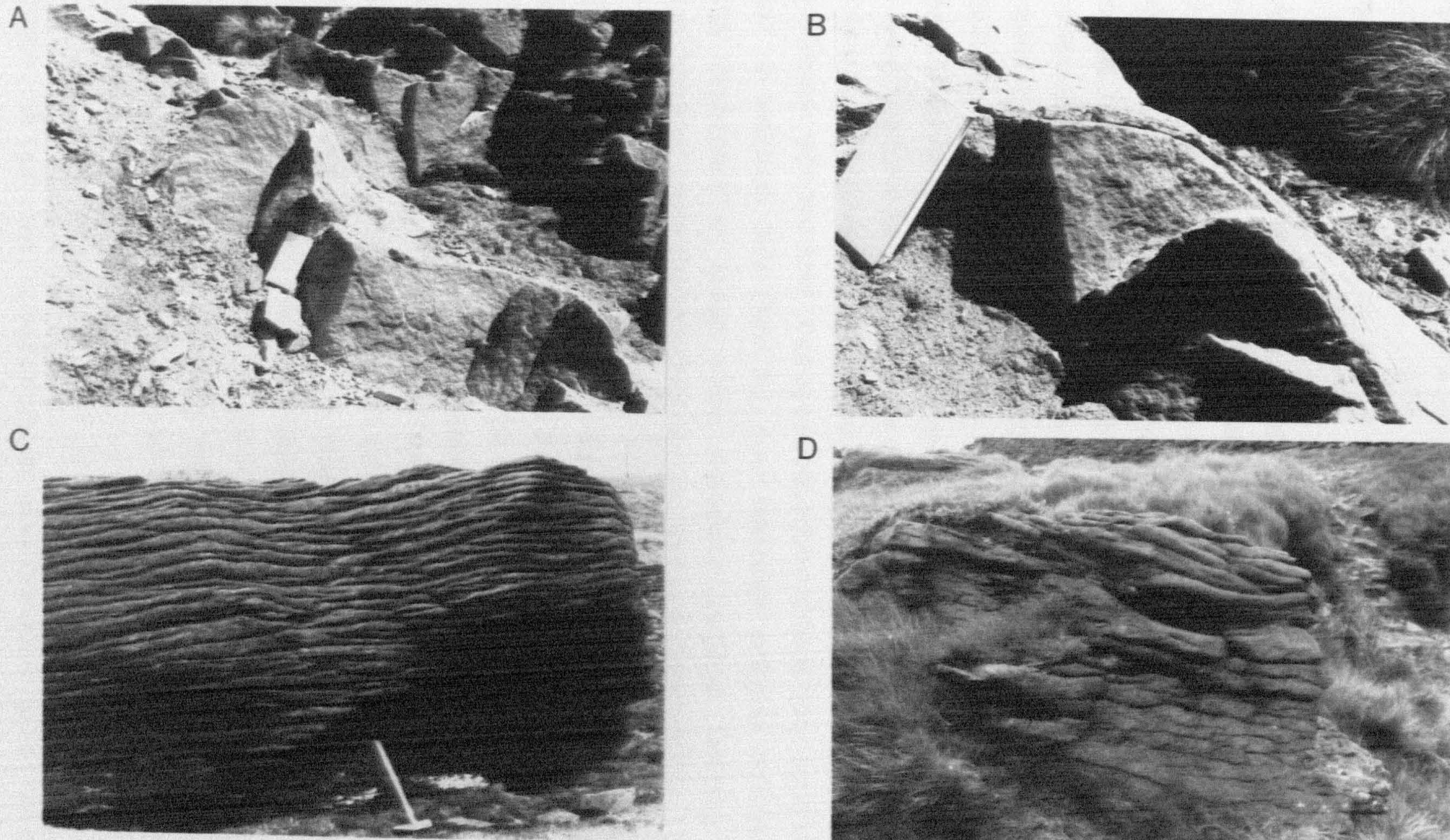


Plate 4

Large areas of destratification are found within some beds, usually towards the top of sequences dominated by this facies-type.

Interpretation

Although grossly different in appearance, the two facies-types in Facies B are believed to result from the operation of similar near-bed processes during deposition from high-concentration turbidity currents. Facies-type B_a is interpreted as the product of rapid deposition from a current with a well-developed "boundary layer" supported by dispersive pressure or upwards movement of porewater during the final stages of deposition. The sedimentary structures in Facies-type B_a reflect the processes operating in this boundary layer. Crude lamination within the beds may be due to shearing in the boundary layer (Postma, 1986) or to slight power variations during deposition. Sedimentation rates appear to have been too high for the widespread development of traction structures but the isolated sets of cross-strata indicate that bedforms were locally in equilibrium with the flow. Perhaps in situations of local flow expansion over "banks" of sediment (See Plate 3c), flow velocities dropped sufficiently for traction deposition to occur while elsewhere frictional freezing of the traction/liquified carpet was the dominant sedimentation mechanism.

There are probably several origins for the dome and hollow topography on the upper surfaces of Facies-type B_a beds. Wavy bedding has been widely described in sequences of Facies B type deposits (e.g. Mutti and Ricci Lucchi, 1972) but has rarely been given any sedimentological significance. Similar dome and basin topography described by Chipping (1972) was interpreted as being due to later tectonic deformation. Such an origin is unlikely for the Pendle Grit Formation structures as there is no other evidence for significant tectonic deformation. However, low amplitude-long wavelength, load-induced deformation of the overlying bed might have caused the topography in some instances. In other cases, particularly when there is a mudflake lag in the hollows, irregular scour at the base of the overlying bed may explain the development of the topography.

Clearly, though, the topography can also represent preservation of a bedform: this is proved by the behavior of internal laminae in the underlying bed (Plate 4b). It is possible that these bedforms represent antidunes caused by supercritical flow in the transporting fluid. Walker (1967) interpreted a ridge-hollow topography on turbidite bedding planes as preserved antidunes. This topography was of similar wavelength/amplitude ratio to that of the dome-hollow topography of Facies-type B_a. However, Hand (1974) has shown from experimental modelling that antidunes in turbidity currents should be two orders of magnitude larger than the structures reported herein and by Walker. Smaller scale antidunes seem unlikely to exist unless they scale with the proposed boundary layer flow rather than turbidity current as a whole. The morphology of the domes and hollows also invites comparison with the hummocky cross-stratification of clastic shelves (Dott and Bourgeois, 1982). However, the latter requires the interaction of oscillatory waves with the sediment during deposition and it is difficult to envisage such waves being generated within mass-flows. Both these processes, then, are unlikely to have caused the dome and basin topography in Facies B.

A more likely explanation is that the dome-hollow bedforms represent the preservation of a shear-stress induced wave instability at the boundary between two zones within the flow. Modelling by Middleton (1967) shows that such instabilities develop between a "boundary layer" supported by dispersive pressure/porefluid escape and the main parts of a high concentration turbidity flow. Although these waves disappeared before final consolidation of the bed this might have been due to the rapid flow deceleration during the experiment. In the real situation the bedform might be preserved if deceleration was such that frictional freezing of the liquified bed occurred prior to the shear-stress from the overlying bed disappearing. The laminae at the top of the bed might represent suspension deposition with slight traction reworking during the waning of the remainder of the flow. The dish structures present in this facies-type lend support to an origin related to deposition from a fluidised/liquified "boundary layer" as they suggest major dewatering during the final stages of deposition.

Facies-type B_b is similarly interpreted as the result of deposition at high rates from high-concentration turbidity currents. Again the currents are thought to have been highly non-uniform but, in this case, to have had insufficient energy to maintain a continuous "boundary layer" flow. However, random variations in energy impinging on the depositional surface (again similar to bursts and sweeps in fluvial flow) caused repeated resuspension of small areas of the accreting bed surface (Figure 3.7). This sediment then moved downcurrent as small grainflows

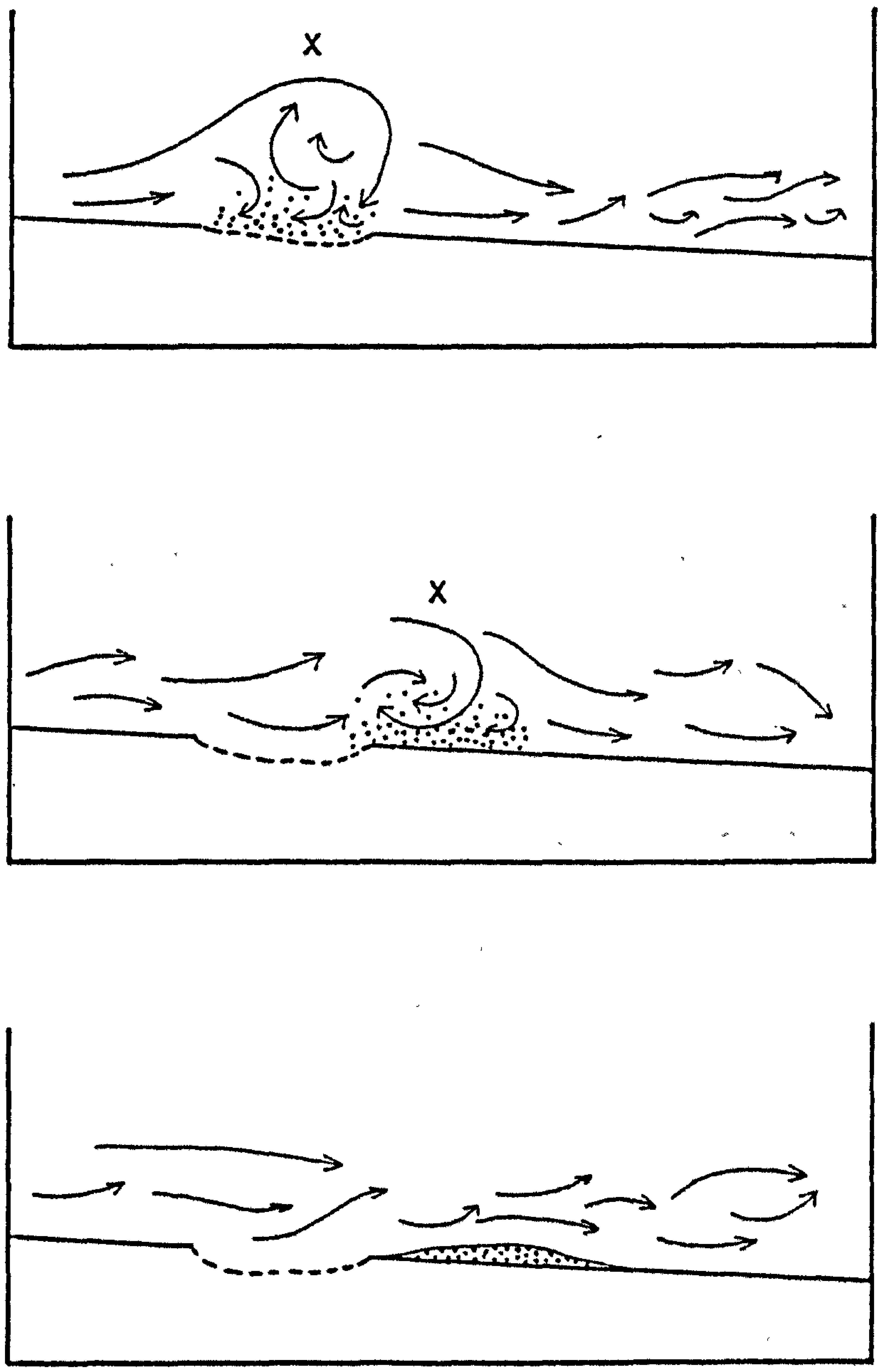


Figure 3.7: Possible origin of the sedimentary lenses in Facies-type B_b. In (a) a high energy sweep event in the turbidity current impinges on the bed surface (X). This causes resuspension of some coarse material which is then carried downcurrent by the sweep event as a mini-grainflow (b). As the event decays (c), frictional freezing of the grainflow deposits a tongue of reverse graded sediment.

for the duration of the sweep event. As the sweep decayed so frictional freezing of the grain flow took place and small tongues of sediment were deposited. Nett deposition from the flow allowed these to be preserved as the "ropy" lenses seen today. The scale of the lenses, and the presence of reverse grading within them, is good evidence for the operation of such a mechanism (Hein, 1982; Hiscott and Middleton, 1979). Higher energy sweep events might be sufficient to cause the larger erosion scours seen in this facies-type.

Excess pore water seems to have been expelled along the boundaries of the lenses and also to have jumped upwards between lenses. The latter mechanism explains the vertical "staircases" of the weathering pattern. Whether or not this dewatering is related to deposition is unclear. The larger scale destratification in this facies-type appears to be post-depositional but cannot be directly linked to the water escape structures in the rest of the beds. Baines (1977) thought that the ropy weathering was entirely due to differential erosion along water escape paths. Other authors have described structures which look something like the ropy weathering in this facies but have interpreted them as dish structures (Hein, 1982; Corbett, 1972). However, the structures in Facies-type B_b are believed mainly to reflect sedimentary processes because of the grain fabrics within the lenses and because of erosional cut-offs of the "ropes" (Plate 4d).

3.2.3 Facies C.

Description.

Facies C consists of medium to thick beds of graded sandstone alternating with pelitic beds. Units within this facies can be described with reference to the classical Bouma sequence subdivisions, T_a to T_e (Bouma, 1962. See Figure 3.8). The presence of a well developed T_a division and a high sand/mud ratio is characteristic of this facies. About 30% of the Pendle Grit Formation is developed in this facies.

Facies-type C_a: consists of a thick T_a division overlain by a thin T_b or T_c division. The T_a division usually has a sharp base with well developed flutes, tool marks or linear loads similar to those seen in Facies B. Grainsize in this division varies from coarse to fine sand grade, coarser units being thicker. The division may show distribution or coarse-tail grading with the bulk of the grainsize reduction occurring over a very thin interval just beneath the fine-scale lamination or cross-lamination of the T_b or T_c divisions. The latter divisions are usually less than 5cm thick: often only a single set of cross-lamination is present. Linguoid ripples (wavelength 10-15cm, amplitude 1-5cm) are often preserved on bedding planes in this facies-type. In some cases ripples have been washed out to leave rib and furrow structure on the bed surface. Where the T_c division is absent the top surfaces of beds often display primary current lineation.

There is a distinct grainsize gap above the T_b or T_c division which produces a sharp top to the sand bed. The overlying pelitic interval is thin (usually less than 10cm) and consists of carbonaceous siltstone and silty mud. Thin laminae of rippled very fine sandstone are sometimes present within this interval.

This facies-type is equivalent to Facies-type C_1 of Mutti and Ricci Lucchi. Examples of the facies-type are shown on Figures 3.15, 3.18-20 and Plate 5.

Facies-type C_b: is similar to Facies-type C_a but there is no grainsize gap between the sandy and muddy parts of the bed. The T_{b-c} division is expanded relative to T_a and the pelitic intervals are thicker and continuous with the sand bed. This facies-type corresponds to Facies-type C_2 of Mutti and Ricci Lucchi and is rarely developed in the Pendle Grit Formation. One example of this facies-type is shown on Figure 3.16, Log I.

Facies-type C_c: is superficially similar to C_b but the sediment involved is much more mud-rich throughout. The basal T_a division is effectively a greywacke: abundant mudflakes and very poorly sorted sand grains float in matrix of silty mud. The coarser grainsize fraction rapidly disappears upwards through division T_a. Division T_{b-c} consists of alternating laminae of mud-rich and mud-poor micaceous sandstone which are strongly deformed by ball and pillow formation, often producing slurries of apparent "mudflakes" in muddy sand (Plate 6). The sandstone laminae become thinner and finer upwards through Division T_d while internal deformation declines in intensity. Division T_e shows a grading from silty, carbonaceous mudstone to blocky mudstone. Small sand dykes with ptygmatic folds extend from the lower sand divisions into the pelitic interval.

This facies-type forms the bulk of the lower parts of the Whitendale Member (See Figure 3.20).

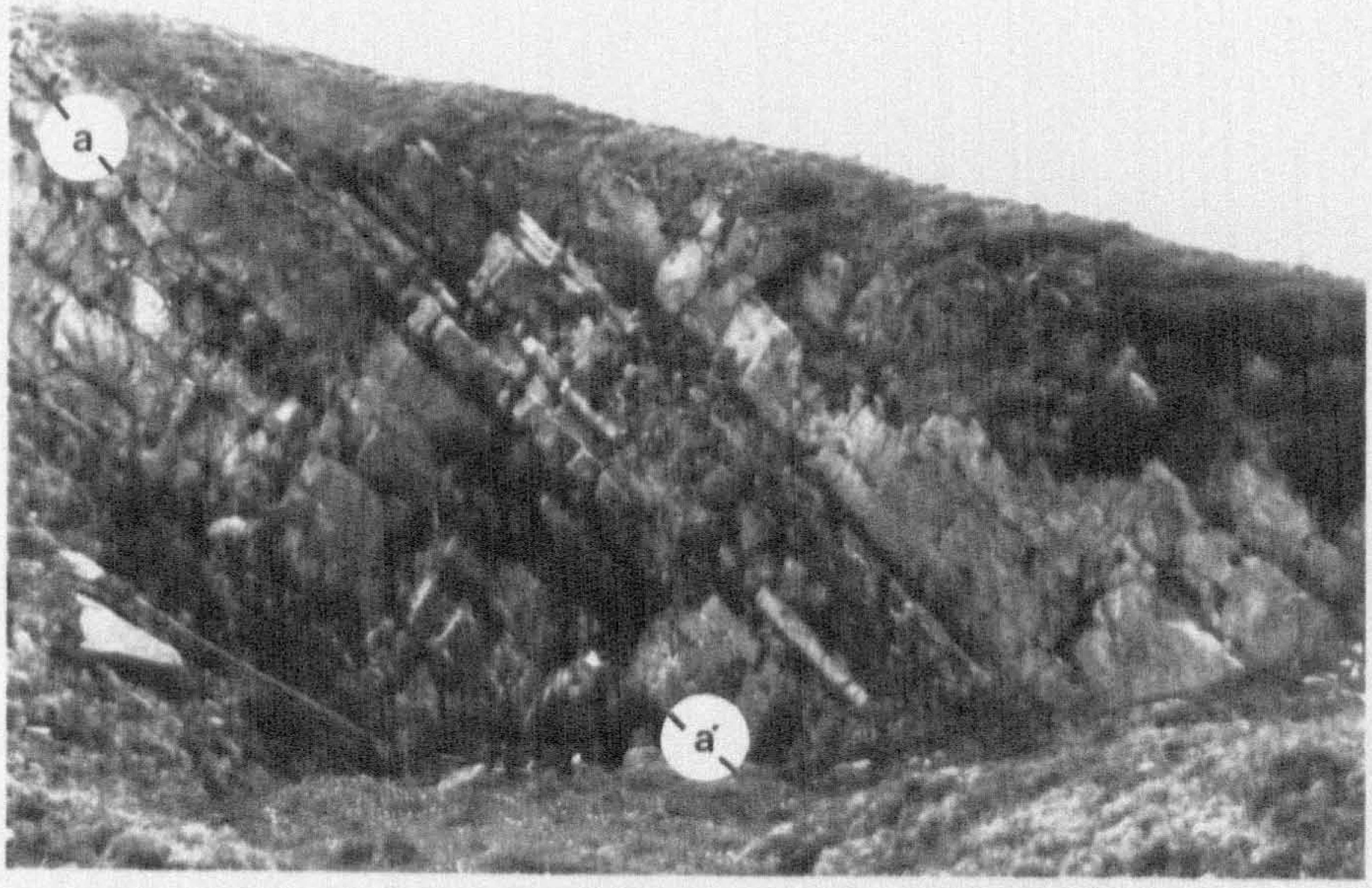
Plate 5: Facies C and E deposits.

a) A series of second-order channel fills (see sub-section 3.4.1) including a packet of Facies C, E and G deposits. Note the marked angular discordance across the second-order erosion surface a-a' and the thickening down-dip of the overlying Facies A/B beds into the channel axis. The top of this pelitic package is another, less marked second-order erosion surface. Wiswell Moor Quarries, SD753 371. Quarry face is 10m high.

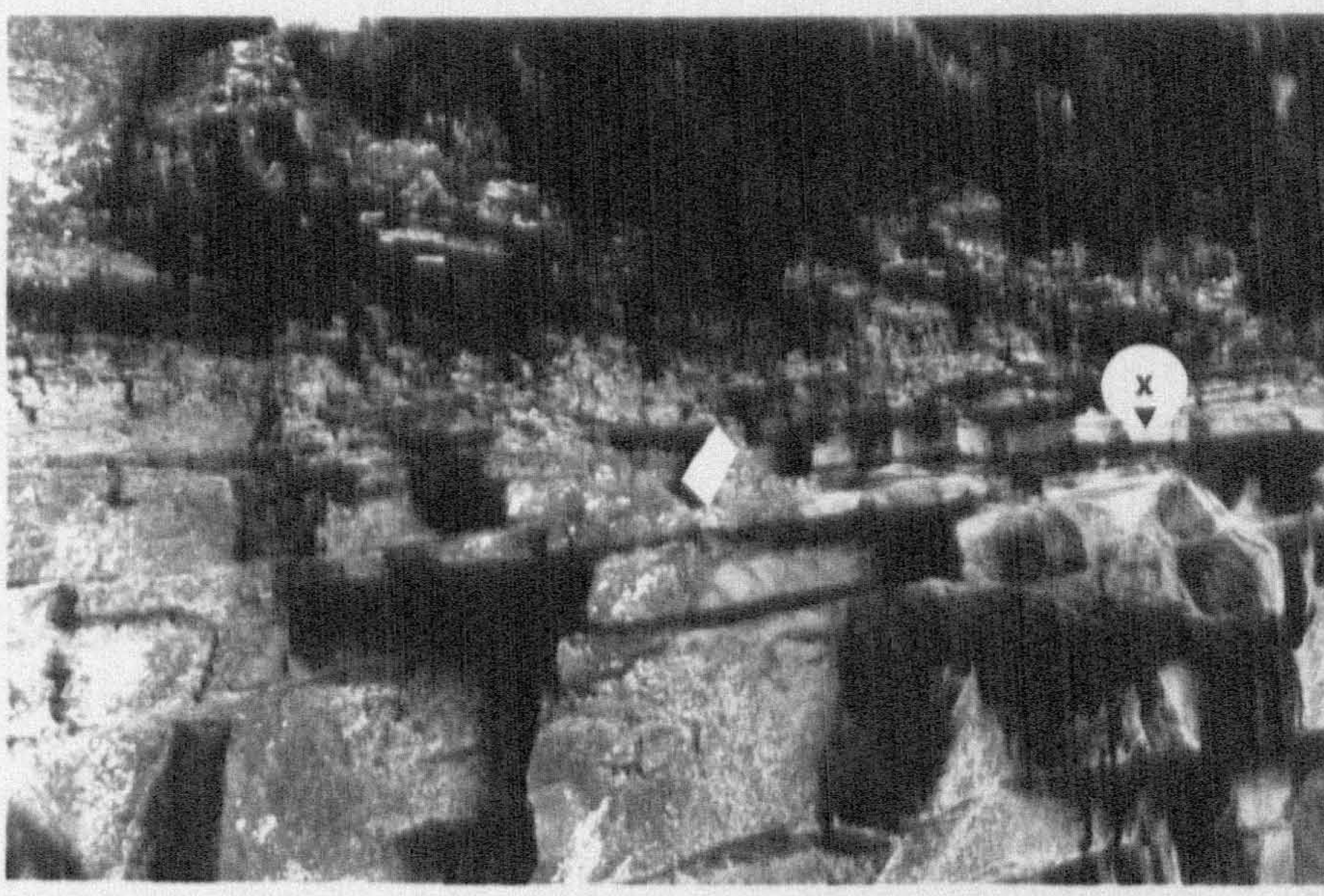
b) Complex lateral relationships between Facies E and D beds at the top of the pelitic unit in (a) above. Note that the Facies E bed immediately under the notebook pinches out abruptly to the right. Overlying facies D beds (x) thicken in the opposite direction. This may be due to the slight topographic relief caused by the pinchout of the underlying bed. Notebook is 20cm long.

c) Facies E deposit from the pelitic unit in (a) above. The grainsize of this unit is slightly greater than in the associated Facies C deposits. Note the starved ripple train on top of the bed has a very sharp base, a feature which has not been observed elsewhere in Facies C, D or E deposits of the Pendle Grit Formation. It is thought likely that this ripple train marks deposition of sand by a flow event separate to that from which the bulk of the bed was formed. The grainsize of these ripples is similar to that of the ripples in the overlying Facies G unit. Consequently, they may belong more properly to the latter rather than the Facies E bed. Scale increment on notebook is 5cm.

A



B



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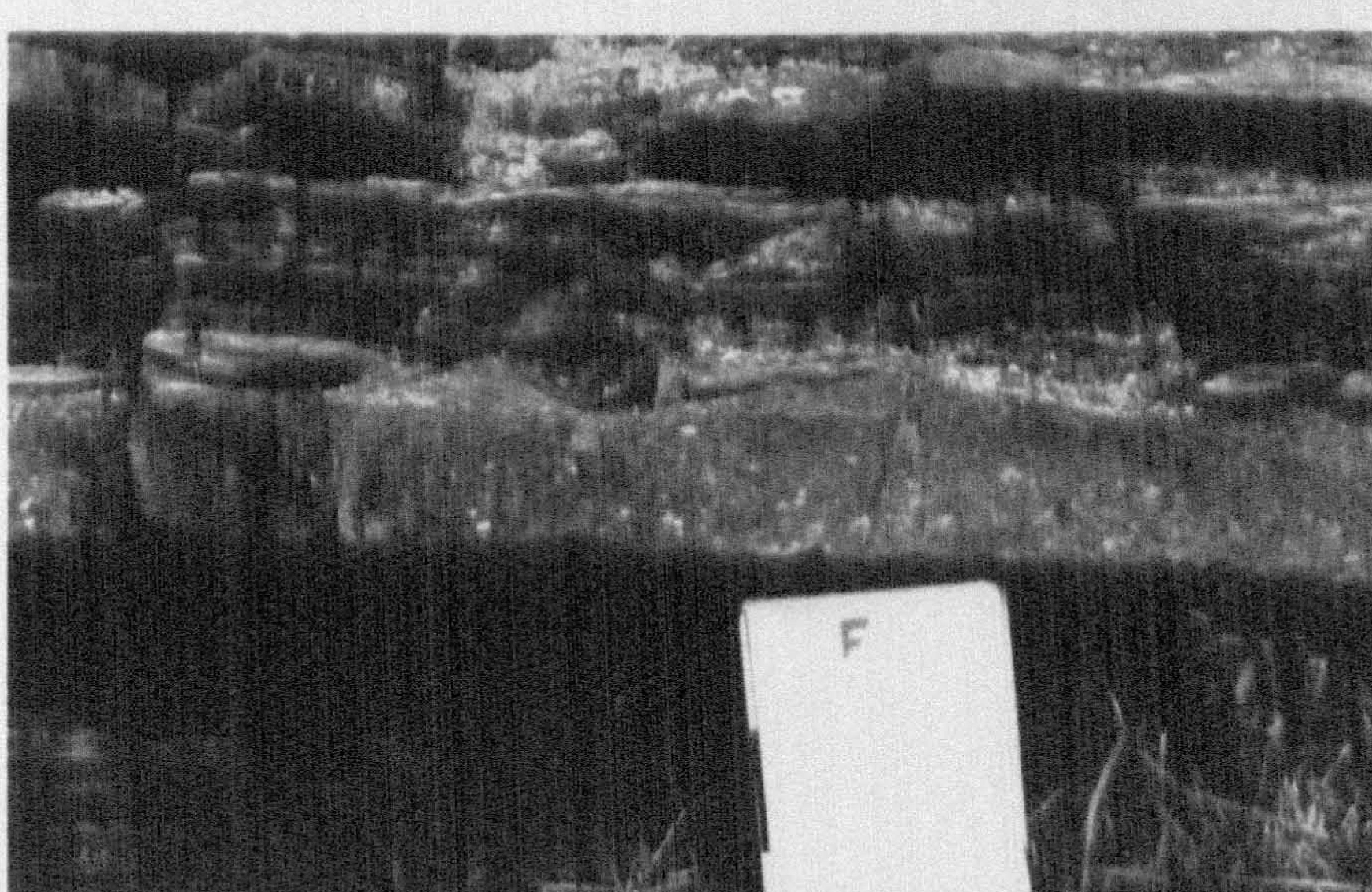


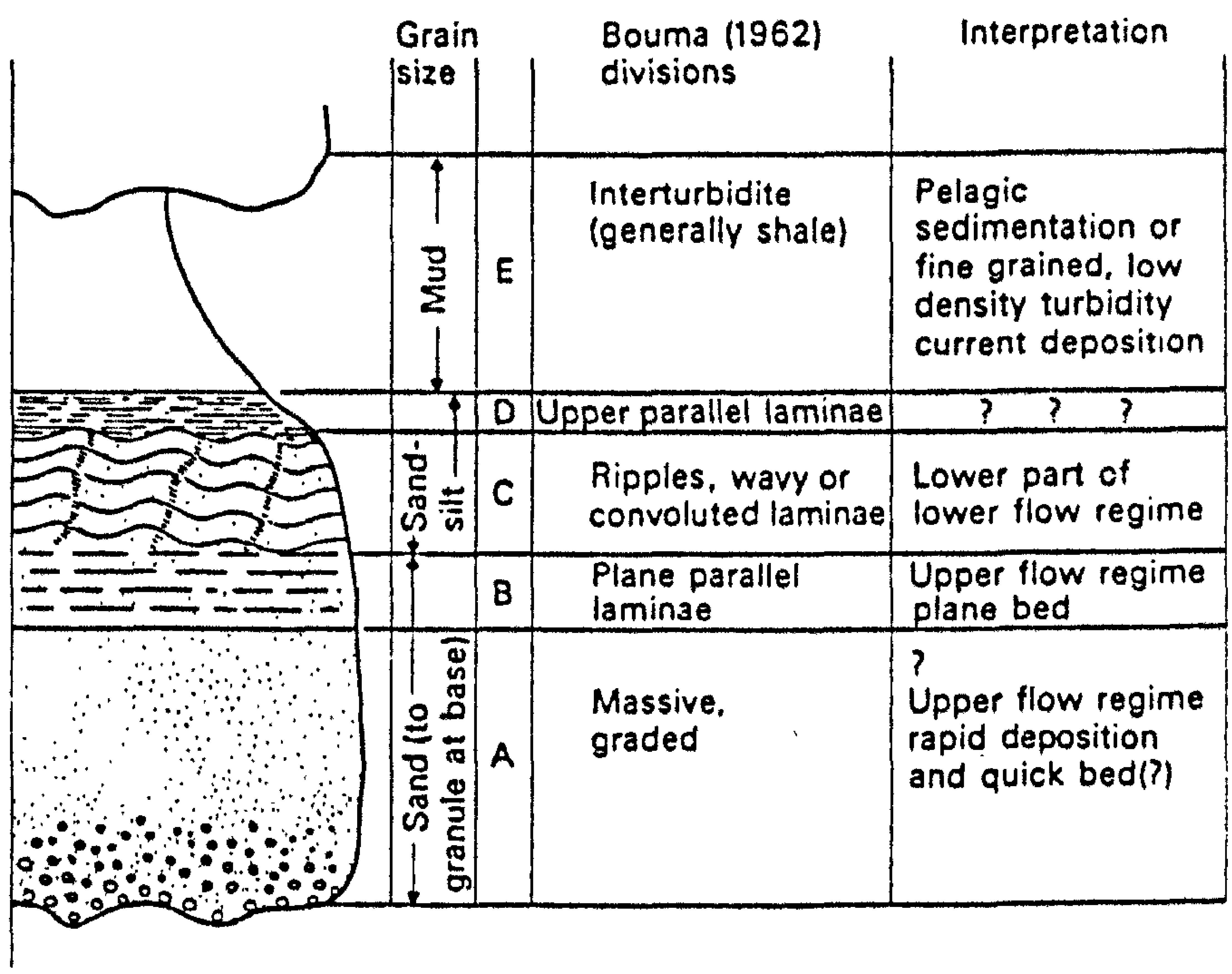
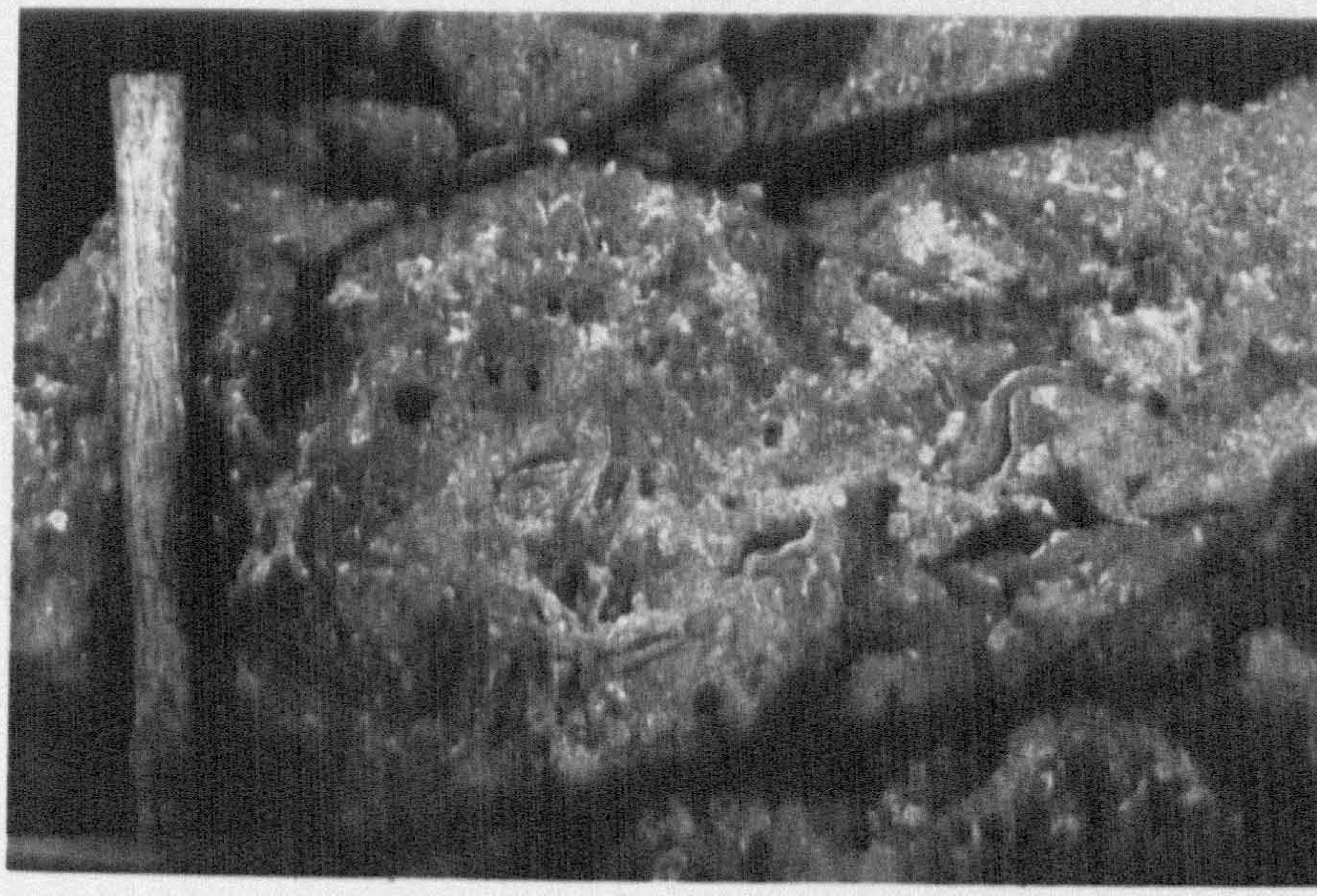


Figure 3.8: The idealised Bouma sequence in turbidites illustrating the sedimentological divisions within a single bed and their hydrodynamic interpretation. From Middleton and Hampton (1976).

Plate 6: Log and photograph of a Facies-type C_c bed. The log shows the boundaries of the Bouma Divisions. Note the muddy nature of the basal divisions and the complex T_c division made of a mixture of "blebs" of silty sandstone and claystone. These "blebs" are thought to form during dewatering of the bed. See text and Figure 3.9 for further discussion. Photograph shows a T_c division from a better exposed bed higher in the section. Hammer is 40cm long. Whitendale River, SD653 558.



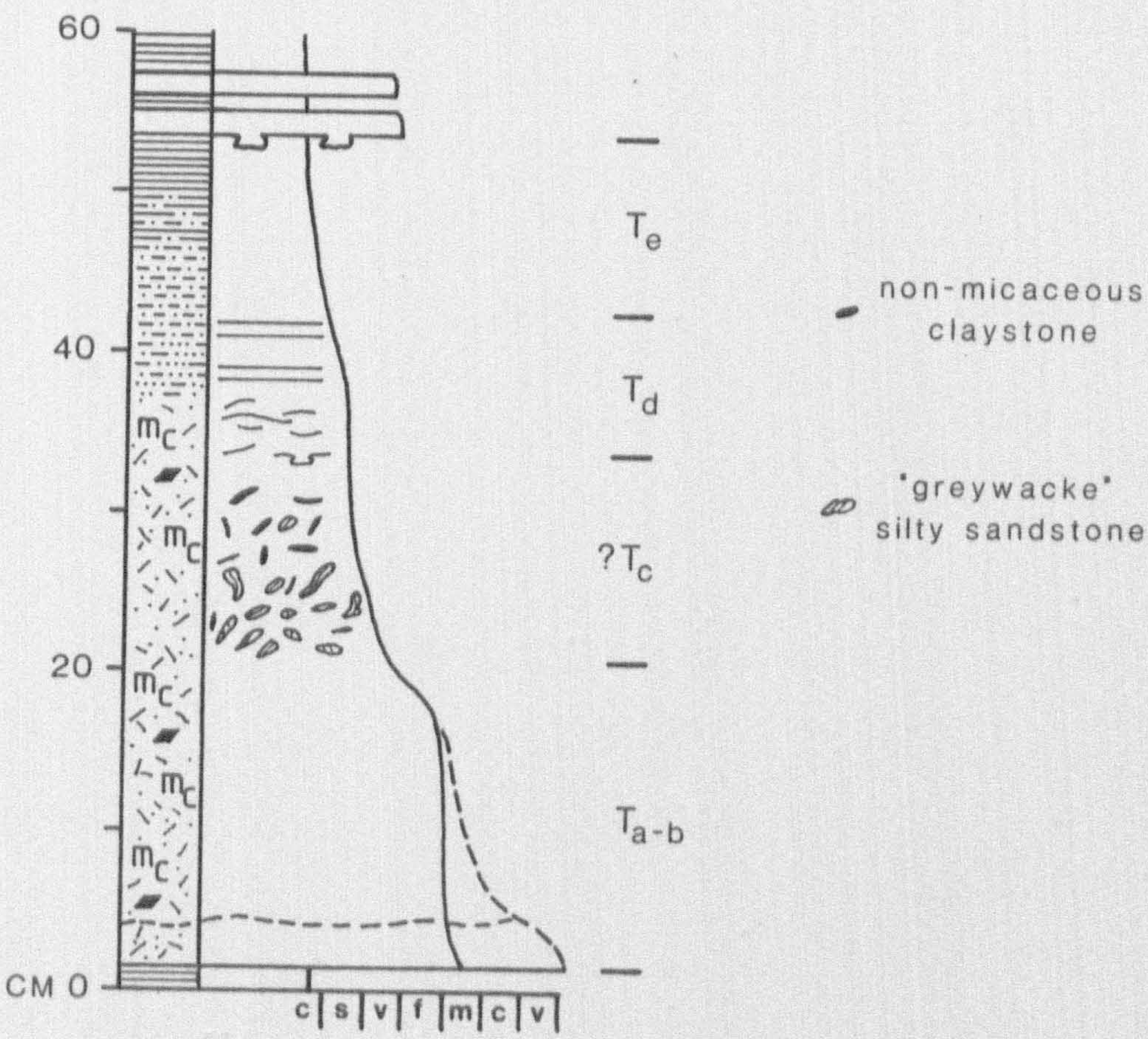

60
40
20
CM 0
T_e
T_d
$?T_c$
T_{a-b}
m_c
c s v f m c v
non-micaceous claystone
'greywacke' silty sandstone

Plate 6

Interpretation.

Facies C represents deposition from turbidity flows with lower flow power and lower sediment concentrations than those responsible for deposition of Facies A and B. The basal T_a divisions of Facies-type C_a beds indicate rapid sedimentation due to fallout of sediment from suspension. Sedimentation rates were sufficient to dampen turbulence in the basal zone of the turbidity current but a boundary layer flow was not developed because of the low flow energy. As sedimentation rates declined there was some movement of sediment as bedload. This produced the T_b and T_c divisions of the beds. The overlying pelitic interval, division T_e, may represent either suspension fall-out from the "tail" of the flow or deposition from low-density turbidity currents between major flow events.

Facies-type C_b represents the complete, classical Bouma sequence in which the divisions are interpreted as representing deposition from a waning turbidity current (Figure 3.8). Experimental work confirms such an interpretation: crude Bouma sequences have been produced from decelerating model turbidity currents in which sediment concentrations were less than 30% (Middleton, 1967). The thick T_e divisions in this facies, which are continuous with the sand bed, are clearly suspension fall-out deposits from the same flow. This suggests that the flows responsible for deposition of Facies-type C_b had a higher mud/sand ratio than the flows discussed so far.

An even higher mud/sand ratio seems to characterise the flows from which Facies-type C_c was deposited. The muddy nature of the T_a division suggests that the deceleration of the flow was extremely rapid and that much of the sediment load was simply dumped. The rapid deceleration might be due to flow expansion at a channel mouth causing a hydraulic jump (Komar, 1971) or to the flow running into an obstruction or reverse gradient. If the latter is the case, the laminated $T_{b\text{-}c}$ interval may represent the interaction of the tail of the turbidity current with the reverse flow generated by reflection (See Pantin and Leeder, 1987 and Figure 3.9 herein). Deposition rates were obviously high as the mud-rich sediment was unable to dewater. The extensive soft sedimentary deformation is due to this inability to dewater. It was caused either by the internal gravity collapse of the $T_{b\text{-}c}$ divisions as T_a dewatered by dyke intrusion or by deformation as the entire water-saturated unit moved on downslope by gravity sliding.

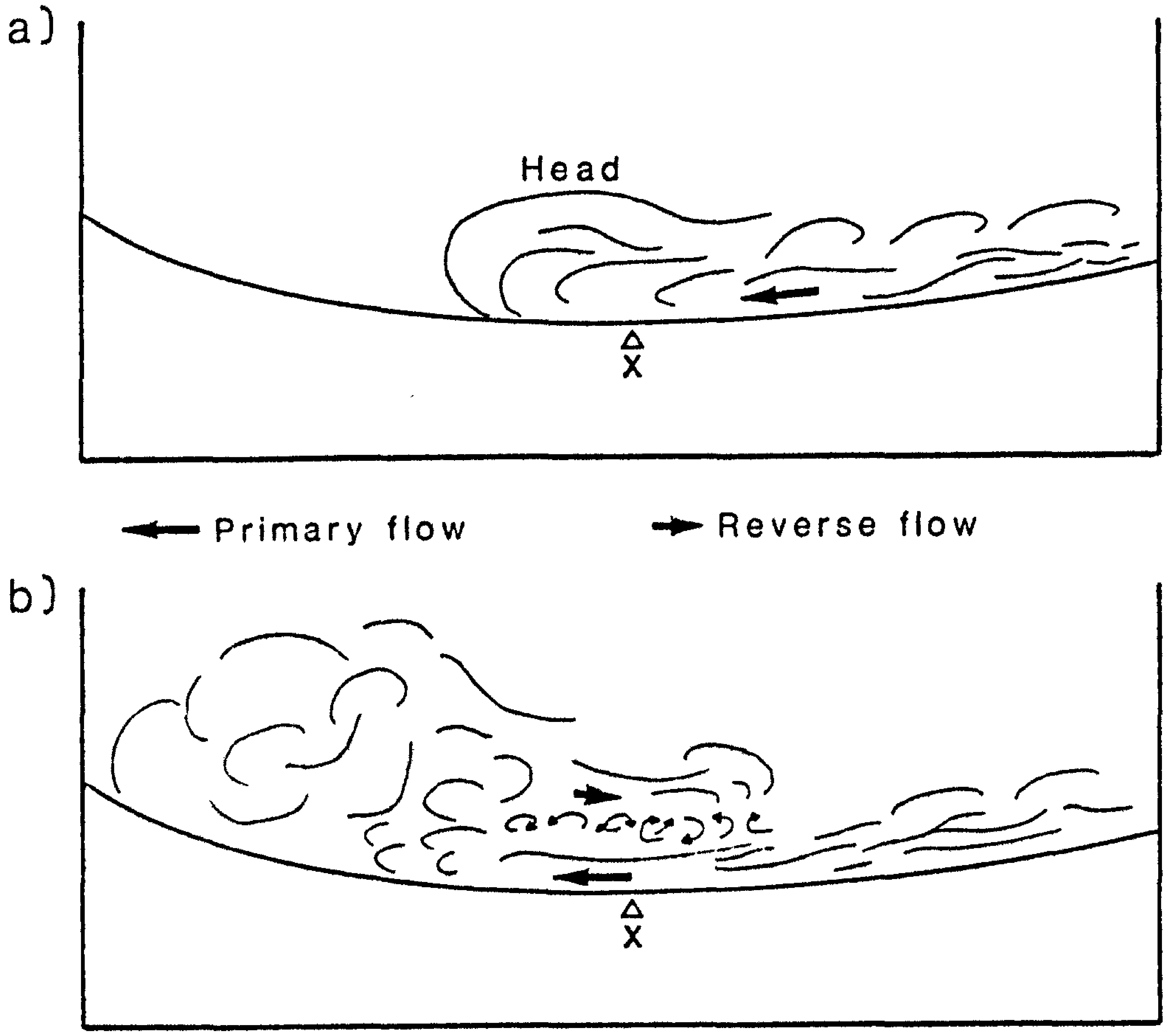


Figure 3.9: One explanation of the origin of Facies-type C_c deposits. In (a) a low density, mud-rich turbidity current approaches a reverse gradient. At point X, fallout of sediment from suspension (caused by deceleration) produces Bouma Division T_a (See Plate 6). As the body of the current continues past X, laminated sands are deposited. Meanwhile, the head of the current climbs up the reverse gradient until its hydraulic head is exhausted. Continued supply of sediment from the body of the current builds up the head which then collapses back over the primary flow (b). The reversed flow contains finer sediment than the primary flow because of deposition of its coarser grade material during flow reversal. This finer sediment falls back into the primary flow across the zone of extreme turbulence at the boundary between the two flows. In turn, this dampens the turbulence of the primary flow and the whole flow system becomes unstable and collapses. At point X the resulting rapid sedimentation produces a muddy (greywacke) sand lamina followed by a silty mud lamina while the primary flow re-establishes itself. This cycle repeats itself until the primary flow is exhausted. The resulting sand/mud laminae are then broken up by dewatering to produce the "slurried" T_{b-c} division of the bed. Division T_d may represent reworking by the tail of either the primary or the reverse flow. Division T_e results from post-flow settling of the suspension cloud built up above the flows.

3.2.4 Facies D

Description.

This facies includes all beds which can be described in terms of base absent Bouma sequences. Grainsize of the sandy interval ($T_{b\text{-}d}$) varies from fine sandstone to siltstone while the muddy interval (T_e) is usually thicker than in Facies C deposits. The facies is poorly represented in the Pendle Grit Formation, unlike in the classic Apennine turbidite sequences where it is the most abundant mass-flow type.

Facies-type D_a: comprises thin- or occasionally medium-bedded fine sandstone overlain by a thick pelitic unit. Bouma divisions T_b and T_c are well developed and overlie a sharp but non-erosive base. The pelitic interval grades upwards from carbonaceous silty mudstone (with laminae or flasers of rippled fine sandstone) to blocky mudstones. A well developed trace fauna is present in this facies-type: *Rhyzocorallium, Bergaueria* and *Planolites* are the most common forms observed by the author. Baines (1977) and Eager *et al.* (1985) give details of a wider range of trace fauna from the Pendle Grit Group, most of which appear to come from this facies-type and from Facies E.

Facies-type D_a corresponds to Facies-types D_1 and D_2 of Mutti and Ricci Lucchi. Examples are illustrated on Figures 3.16, 3.18 and 3.20.

Facies-type D_b: comprises very thin to thin-bedded units of graded silty mudstone. Only Bouma divisions T_d and T_e are developed. Division T_d is represented by micro-laminated or rippled carbonaceous siltstones which grade up into a T_e division of blocky mudstone (Figure 3.20). Often groups of Facies-type D_b beds form cycles in which the T_d/T_e thickness ratio decreases upwards from bed to bed (refer to Figure 3.20).

Interpretation.

Facies D represents a further decrease in flow power and increase in mud/sand ratio in the turbidity flows. Again, waning flow led to deposition of the coarse fraction as the traction dominated $T_{b\text{-}d}$ divisions. Suspension fall-out of the mud fraction then produced division T_e. In the case of Facies-type D_b the flows must have been extremely weak and dilute.

3.2.5 Facies E

Description

As in the Mutti and Ricci Lucchi scheme, this facies includes thin or medium-bedded sandstones with sharp tops and bases. The grainsize varies from coarse to fine sand grade and is higher than in associated Facies D deposits. The sharp bases are fluted and have a range of tool marks. Mudflake and granule lags at the base of the bed are common while the rest of the bed may be graded, massive or cross-bedded throughout (Figure 3.10a). Linguoid ripples are often preserved on the sharp top of the bed. A trace fauna similar to that in Facies D is associated with this facies. Facies E Beds are usually discontinuous over metres or tens of metres. The facies accounts for 5% or so of the Pendle Grit Formation and is also seen locally in the Pendle Shale Formation. Additional examples of this Facies are illustrated in Figures 3.16, 3.18 and 3.19.

Interpretation.

The thin, coarse-grained, discontinuous beds in Facies E indicate small, high energy flows of limited lateral extent. The abundance of flutes shows that the flows were capable of causing erosion while the abundant tool marks suggests saltation transport of some of the sediment load. The origin of these flows is enigmatic. Mutti and Ricci Lucchi interpret Facies E deposits as "traction lags" due to reworking of sediment by non-depositional turbidity currents. However, this interpretation seems unlikely for the Facies E deposits of the Pendle Grit Formation because of the lack of reworking in other facies with which they are associated. Rather, the Facies E beds are thought to be related minor gradient changes in the depositional system. Turbidity flows meeting an increased gradient would accelerate and develop enhanced turbulence. In turn, this would cause erosion and resuspension of bottom sediment (Parker *et al.*, 1986) which would be rapidly deposited as a thin "lag" deposit should the gradient decrease again. Figure 3.10b illustrates one scenario in which such a process could operate: that of a spill current (Komar, 1972) entering an abandoned channel. This setting is compatible with Facies E being most common in channel-fill facies associations (See 3.4.1).

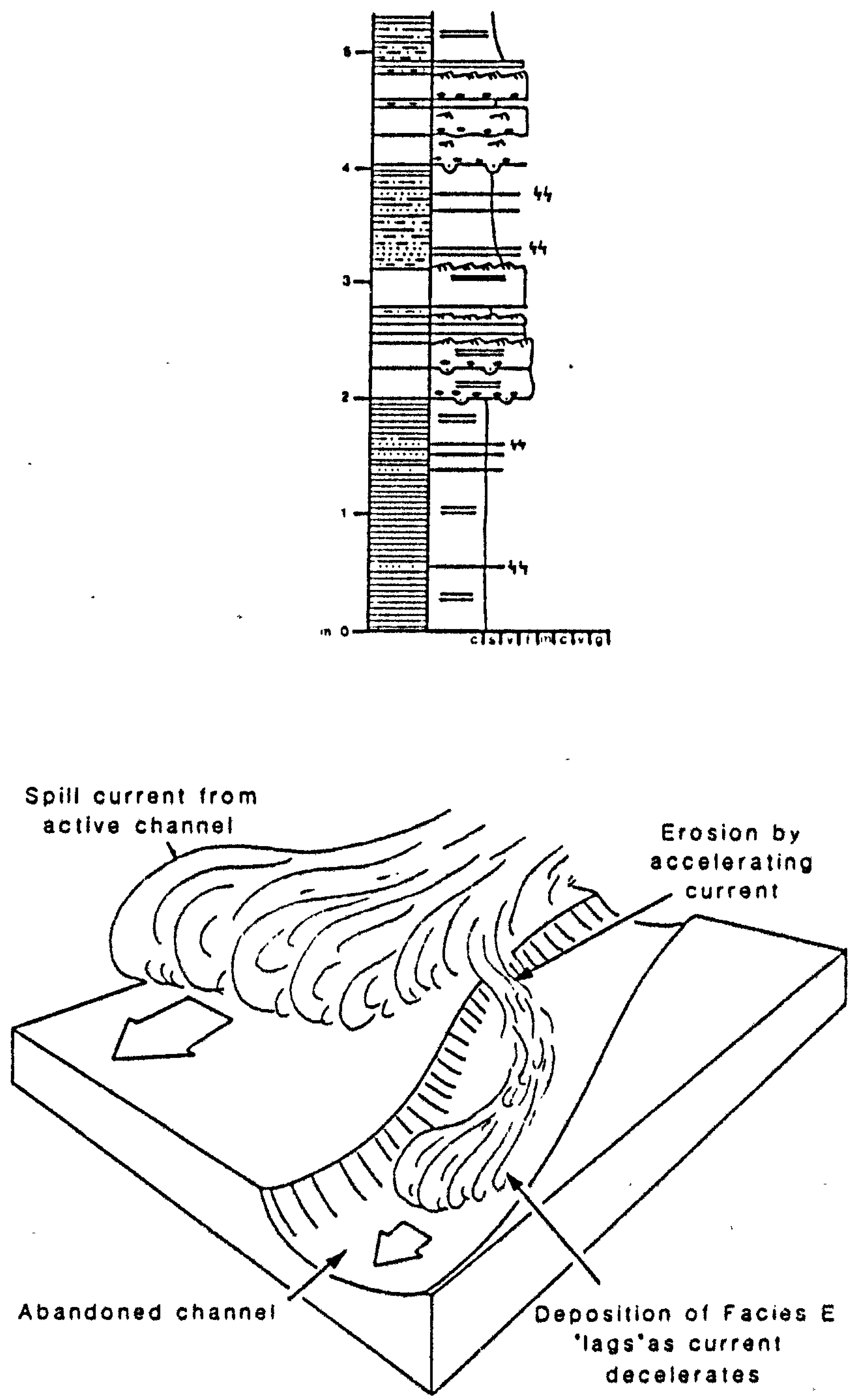


Figure 3.10: Facies E deposits and their origin. The log shows a typical section of Facies E deposits showing their thin lag-like nature (Section at Croasdale Bridge SD685 561). The bottom diagram shows how such deposits could form by the acceleration and deceleration of a non-depositing (i.e equilibrium) current upon meeting changing gradients. The example illustrated is discussed in the text.
For Key see Enclosure 1.

3.2.6 Facies F

Description

Chaotic beds and beds with very high mudflake concentrations are included in this facies. It is equivalent to a combination of Facies-type A_2 and Facies F of the Mutti and Ricci Lucchi scheme. The facies is rare in the Pendle Grit Group.

Facies-type F_a: consists of thick to medium beds of mudflake conglomerate. Mudflakes up to 1m in size are set in a matrix of sheared, slurried sand. Mudflakes account for approximately 50% of the bed and are usually orientated sub-parallel to bedding (Plate 7a, Figure 3.18).

Facies-type F_b: is similar to Facies-type F_a but has higher mudflake concentrations. The mudflakes may be either randomly oriented or crudely aligned with bedding. Beds are upto 5m thick and usually have steep depositional margins against which other facies are banked up (Plate 7b).

Facies-type F_c: comprises blocks or slabs of sediment which have been incorporated into other facies by gravity sliding. Usually the blocks show only slight internal deformation. (Plate 7c).

Interpretation.

These facies result from mass-movement down slope in flows with significant cohesive strength. Facies-types F_a and F_b represent sandy and muddy debris flows respectively (Hampton, 1972; Enos, 1977) while Facies-type F_c represents slump blocks emplaced by gravity sliding over short distances. These often result from bank collapse in open channels.

3.2.7 Facies G

Description.

This facies is minor in the Pendle Grit Formation but forms sequences hundreds of metres thick in the Pendle Shale Formation. It includes all the "background" pelitic beds which do not form part of the other facies. Lithologically the facies consists of laminated, carbonaceous, micaceous siltstones and muddy siltstones interbedded with blocky mudstones. Laminae and very thin beds of rippled very-fine sandstone are common. A varied trace fossil assemblage is developed (See Baines, 1977; Eager *et al.*, 1985) while specimens of the bivalve *Sanguinolites* have occasionally been recovered. Figures 3.11, 3.16 and 3.18 show thin units of this facies.

Interpretation.

The muds and the carbonaceous siltstone laminae are thought to represent suspension fall-out either from the spill currents of channelised turbidity flows or from river generated turbid plumes. The presence of thin, rippled sand beds indicates some slight current activity, possibly as the result of fluvial underflows or higher energy spill currents. Sedimentation rates were relatively high: there is no significant bioturbation in the muddy sediments. However, the intervals between depositional events were sufficient for the development of a soft bodied trace fauna and the introduction of *Sanguinolities*. This latter form is characteristic of Carboniferous delta-front and slope sequences (N. Riley, pers. comm.) and appears to be tolerant of slightly reduced salinities. Reduced basin salinity during deposition has been used to explain the lack of a more extensive fauna in this facies (Eager *et al.*, 1985).

3.2.8 Facies Z

Description.

This facies consists of lenticular bodies of sandstone some 2-10m thick sitting on an erosion surface cut into shales of Facies G. The sandbodies are laterally discontinuous over metres or tens of metres and may have very steep margins. Grainsize within the sandbodies varies from pebbles to coarse sand grade. Large mud clasts and logs are sometimes found. Internally the beds are massive or have crude large-scale trough cross-stratification (sets 2-3m wide and 50cm deep; individual foresets 5-10cm thick). Bases of the beds do not show typical mass-flow bottom structures and the tops are sharp (Figure 3.11). This facies is only found within the Pendle Shale Formation.

Interpretation.

Facies Z is enigmatic. The sedimentary structures in the facies vary from being very similar to those in Facies A beds to being reminiscent of those in a

Plate 7: Facies F deposits.

a) Facies-type F_a bed lying above a first-order erosion surface (arrowed). See section 3.4.1 for terminology. Note the load balls and soft-sedimentary deformation in the Facies B beds underlying the erosion surface. Scale bar is 50cm long. Longridge Caravan Park, SD615 382.

b) Facies-type F_b mud plug lying on a first-order erosion surface (arrowed). The right hand margin of this plug is steep and is onlapped by Facies A and B beds deposited in the channel. The mud-rich debris flow obviously had significant cohesive strength and had marked relief on its top surface after it finally stopped moving. Scale bar is 2.5m long. Salterforth Quarry, Barnoldswick (SD877 453). This part of the exposure has now been covered.

c) Facies-type F_c slumped block. This block has moved about 5m downslope into a minor channel: it represents the results of gravity instability in interbedded muds and sands at a small scale. Scale bar is 50cm long. Lower Ogden Clough Reservoir, SD398 817.

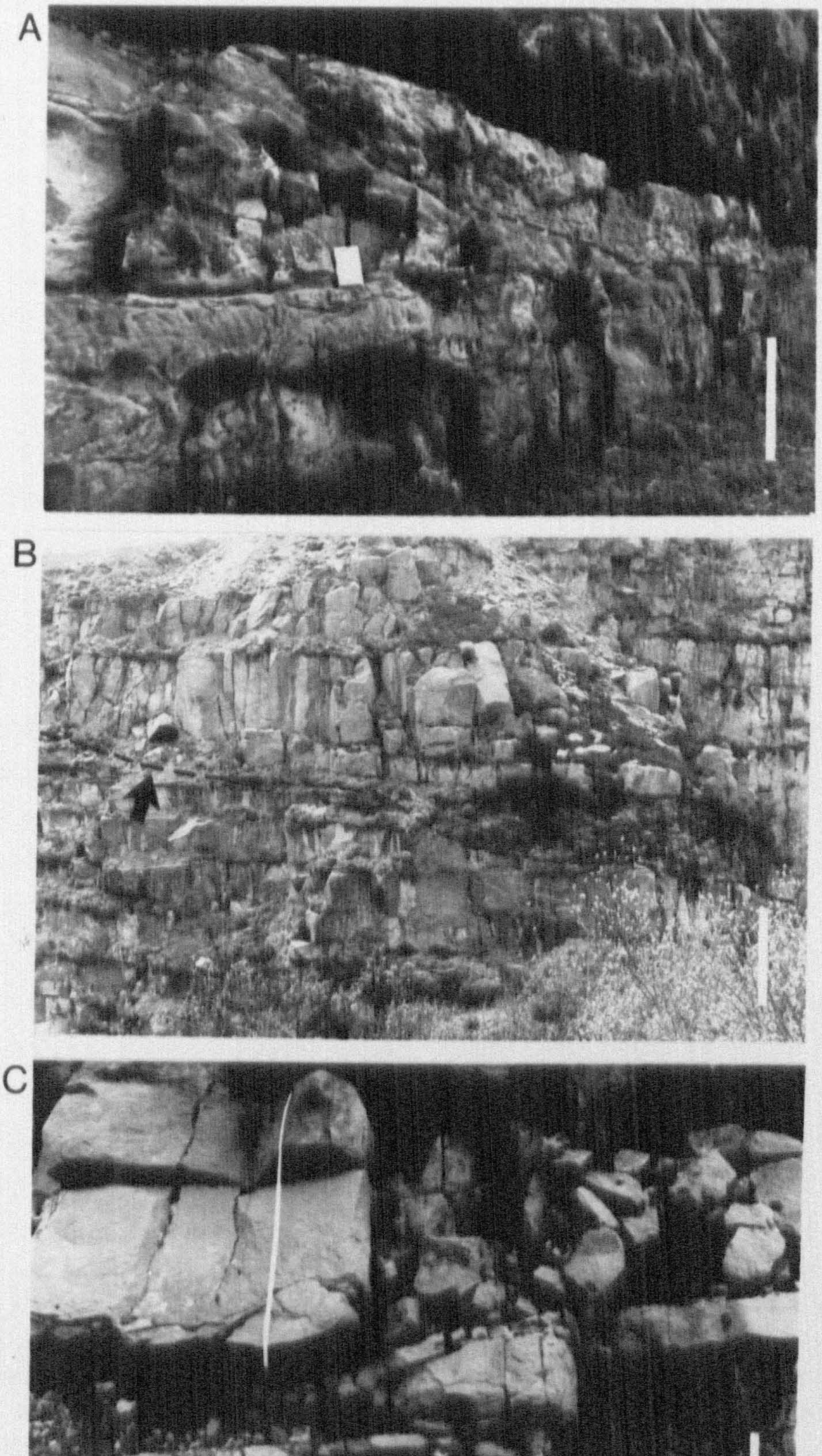


Plate 7

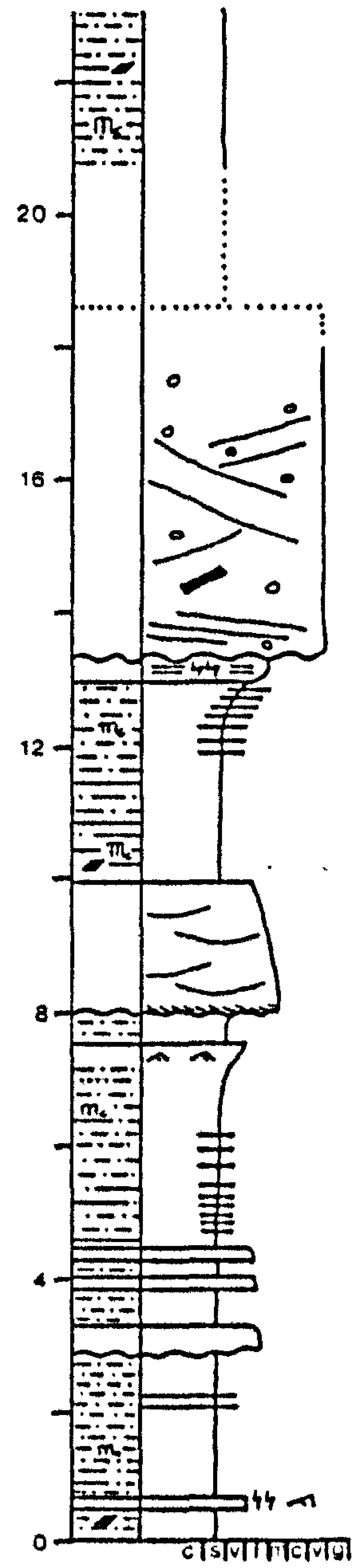


Figure 3.11: Log through a section of Facies Z deposits set in Facies G "background" silts. The sandstone bodies pinch out rapidly away from the line of the log. The beds at the 4m level are only metres in width while the other sand-bodies extend across most of the outcrop but appear from feature mapping to be restricted to steep sided channels. Just below the level at which this log starts there is a ten metre thick sequence with disturbed and slumped bedding. This is thought to have been caused by gravity collapse of the Facies G slope sediments. The sandstone channels may have been trapped in the relief caused by the slumping (see text). Section in Hesley Beck, SD794 599.
For Key see Enclosure 1.

fluvial channel fill. Unfortunately, there are too few exposures of this facies-type to come to a firm conclusion as to its origin. However, the available data suggest that the facies is equivalent to that of sandstones in the Hareelv Formation of eastern Greenland (Surlyk, 1987). Consequently a similar origin is envisaged: namely from cutting and filling of shelf gullies either by very high-energy turbidity currents or by strong (river generated?) traction underflows.

3.3 Summary and discussion of flow types and flow evolution.

In the Pendle Grit Formation the bulk of the sediments are developed in Facies A to C, indicating that the whole fan system was dominated by coarse clastic sedimentation. In such sand-rich systems much of the sediment must have been transported by turbidity flows with high suspended sediment concentrations. Unfortunately, the mechanics of such flows are poorly understood as process data is scarce: only a few high suspended sediment concentration flows have been directly observed in modern submarine environments (see Prior *et al.*, 1987). Consequently, in order to reconstruct the flow types responsible for deposition of the sand-rich facies, it is necessary to extrapolate data from density and low-concentration turbidity flows and to use results from experimental and theoretical work. Experimental modelling of high concentration flows, especially channelised flows, is extremely difficult to scale accurately. (See Middleton 1966a, 1966b, 1967; Kersey and Hsu, 1976; Kelts and Hsu, 1980; Luthi, 1981a; Siegenthaler *et al.*, 1984). Mathematical models of turbidity flows can provide some constraints on how flows behave but, as yet, are too simplistic to deal with the complex fluid mechanics of these flows (See Bagnold, 1962; Komar, 1970, 1971; Hand, 1974; Pantin, 1979; Luthi, 1981b; Parker, 1982; Fukushima, *et al.*, 1985; Parker *et al.*, 1986). In particular, mathematical models tend to be based on energy conservation approaches which consider the flows as single units with single sediment transport mechanisms. More and more sedimentological evidence, including that presented above for Facies A and B, suggests that high concentration turbidity flows have at least two layers with differing grain support mechanisms. Lowe (1976a, 1976b, 1979, 1982) and Middleton and Hampton (1976) have reviewed the possible grain support mechanisms within such flows (dispersive pressure, fluidisation, cohesive strength, turbulence. See Figure 3.5) and the sedimentary structures which would result during deposition from one of these flow types. As the authors recognise, however, one grain support mechanism is unlikely to be dominant throughout a flow: the near bed flow mechanics are almost certainly different to those in the main body of the flow. Intuitively this is to be expected: it is exactly analogous to river flows in which sediment may be carried in suspension or, if the flow is sufficiently powerful, as bedload. At a first approximation turbidity flows can be regarded in much the same way as river flows: if the turbidity flow has more than sufficient energy to produce autosuspension of a certain grainsize population (Bagnold, 1962; Pantin, 1979) then this excess energy could be used to support a population of larger grainsize moving as an equivalent to bedload. The resulting basal layer in the flow would have extremely high sediment concentrations and grain support in the zone might involve some of the other mechanisms mentioned above. Which of the grain support mechanisms was dominant in the basal zone would depend on the gradient down which the flow was moving, the mechanism of energy transfer from the main flow and the sedimentation rate from the basal zone (refer to Lowe, 1982): liquified beds might develop if sedimentation rates were very high while traction carpets (supported by dispersive pressure) would be more likely at lower sedimentation rates.

It is not suggested that the basal zone of the flow is a grain or fluidised flow in its own right. The basal zone can only exist because of excess fluid shear stress in the main flow. This excess shear stress can generate and maintain a dispersive stress in the highly concentrated basal zone of the flow such that the resulting deposits have the characteristics of grain flows. Alternatively, when the sedimentation rate is so high that the mechanics of the base of the flow are dominated by pore-fluid expulsion, the shear stress may physically move the whole "quick bed" downslope. The resulting deposits would show structures characteristic of fluidised flows. These mechanisms explain why sediments with sedimentary structures reminiscent of fluidised or grain flows can exist on slopes much less than those theoretically necessary to generate such flows: the sedimentary structures represent the transport process in a basal zone of a high-concentration turbidity current rather than in an homogenous flow. In the facies interpretations above, such basal zones have been referred to as boundary layer flows. The use of this term is justified because of the completely different flow mechanics within the basal zone compared to the main flow. The presence or absence of a boundary layer flow, and the possible processes within it, have been used to explain the differences between the various facies. It is the importance placed on the variation of this boundary layer, coupled with the suggestion of lengthy flow events, which makes the facies descriptions herein different to those of previous workers.

To summarise, the facies-types have been interpreted as resulting from deposition from flows with varying boundary layer thickness and grain support mechanisms. The sedimentary structures in the coarser facies (A and B) have been related to rapid spatial and temporal variations in a boundary layer associated with a high concentration turbidity current. During prolonged flow events the thalwegs of these flows were able to migrate laterally resulting in the production of amalgamation and lateral accretion surfaces. The finer-grained facies (C and D) represent deposition from flows with increased mud/sand ratio and with insufficient energy to support a boundary layer. However, just before deposition some material was carried as a more "orthodox" bedload as shown by the development of traction structures within these facies.

In general then, Facies A to D represent deposition from flows of decreasing power. These flows, however, probably represent stages in a single turbidity current evolving downslope via deposition of its sediment load and via energy loss. Consider a fully autosuspended turbidity flow entering a grossly simplified "fan" system with upcurrent channelised flow and downcurrent unconfined flow (Figure 3.12). At all points upcurrent of the base of slope in the depositional system there are three distinct phases in the history of the flow (Figure 3.12):

1) Rapid flow velocity increase as the head of the turbidity current passes. The absolute maximum velocities at any point will decline downslope due to the factors discussed above.
2) A period of relative velocity stability as the body of the current passes. The duration of this phase will depend on the ability of the turbidity current generation mechanism to sustain the flow.
3) Velocity decline as the tail of the current passes.

At points beyond the base of slope there will be a much more rapid velocity decline: when the tail of the current reaches the base of slope the hydraulic head driving the flow is exhausted and the flow collapses (Figure 3.12,. See Middleton, 1967). The effect of these downslope changes in flow velocity on the facies deposited from the flow can be understood using the sediment continuity equation. This states that the rate of sedimentation is directly proportional to the rate of change of flow velocity with distance. In mass-flow systems flow velocity is mainly dependent on downslope gradient and the excess density of the flow with respect to the basin water. On average, both of these parameters decrease along the length of the flow: all modern fan systems are graded (in a similar fashion to rivers) while gradual entrainment of basin water into the flow reduces the density contrast. Flow expansion, for example at the mouths of channels, causes more rapid entrainment due to the much increased frontal and surface area of the flow (Luthi, 1981a). Consequently, by the continuity equation, the rate of nett sedimentation will be greatest where velocity changes and water entrainment are at a maximum. Figure 3.12b shows how this fact can be used to "map" facies fields in time and space for a single flow in the simple fan system discussed above. Up-current, in the confined parts of the system, the passing of the head causes erosion which is followed by deposition of Facies A or B sediments during the main flow period. In this part of the system the flows are sufficiently energetic to maintain the "boundary layer" flows discussed previously. Rates of deposition are high (because of rapid gradient change) and thick beds result. Further downslope the flows have reduced velocity and erosion by the head produces bottom structures only. The remaining sediment load, now considerably depleted of the coarser grainsize fraction, is deposited as Facies C and then D deposits. In reality, the boundaries between the "stability fields" shown on Figure 3.12b are broad zones which explains why there is a complete gradation between facies-types in mass-flow systems. Note that the bulk of the sedimentation, particularly of the coarser grained facies, is thought to occur during the period of relative flow stability (phase 2): sedimentation occurs because the flow's energy balance is evolving with distance down-slope rather than with time. Note also that the bulk of the sediment volume is probably deposited on the fan slope (Facies A and B) where the rates of gradient change are highest. The final phase of waning flow (i.e. as the tail of the current passes) represents the only stage at which the flow mechanics at a point evolve with time. The length of this third flow phase will depend on the updip flow generation mechanism. For example, if the initial flows are generated by slump events in the source area then the third flow phase may be short. Turbidity currents generated by flood events in rivers may decline more gradually. Whatever the length of the third flow phase it is unlikely to have any great effect in the up-current areas. The tail of the turbidity current probably carries very little coarse-grained material and is very dilute with respect to the rest of the flow. Consequently, it will not cause significant deposition in the upcurrent areas. Some thin silts and reworked, rippled sandstones at the top of Facies A, B and C_a beds might represent deposition during this interval. Downslope, particularly beyond the break of slope, suspension settling during and after the passage of the tail probably accounts for much more of the rock in Facies C_b and D, as per the original interpretation by Bouma.

Figure 3.12: Qualitative facies stability fields for a flow event in a hypothetical fan system. The initial flow entering the system is considered to be fully autosuspended. (a) shows how the flow velocities vary down the fan as a result of gradient and density changes (See Text). (b) indicates how the sediment continuity equation can be used to translate these velocity variations into facies stability fields in time and space across the fan system. It should be possible to produce numerical simulations of this model which would accurately predict the slopes of the curves in (b) and hence allow better prediction of the facies boundaries.

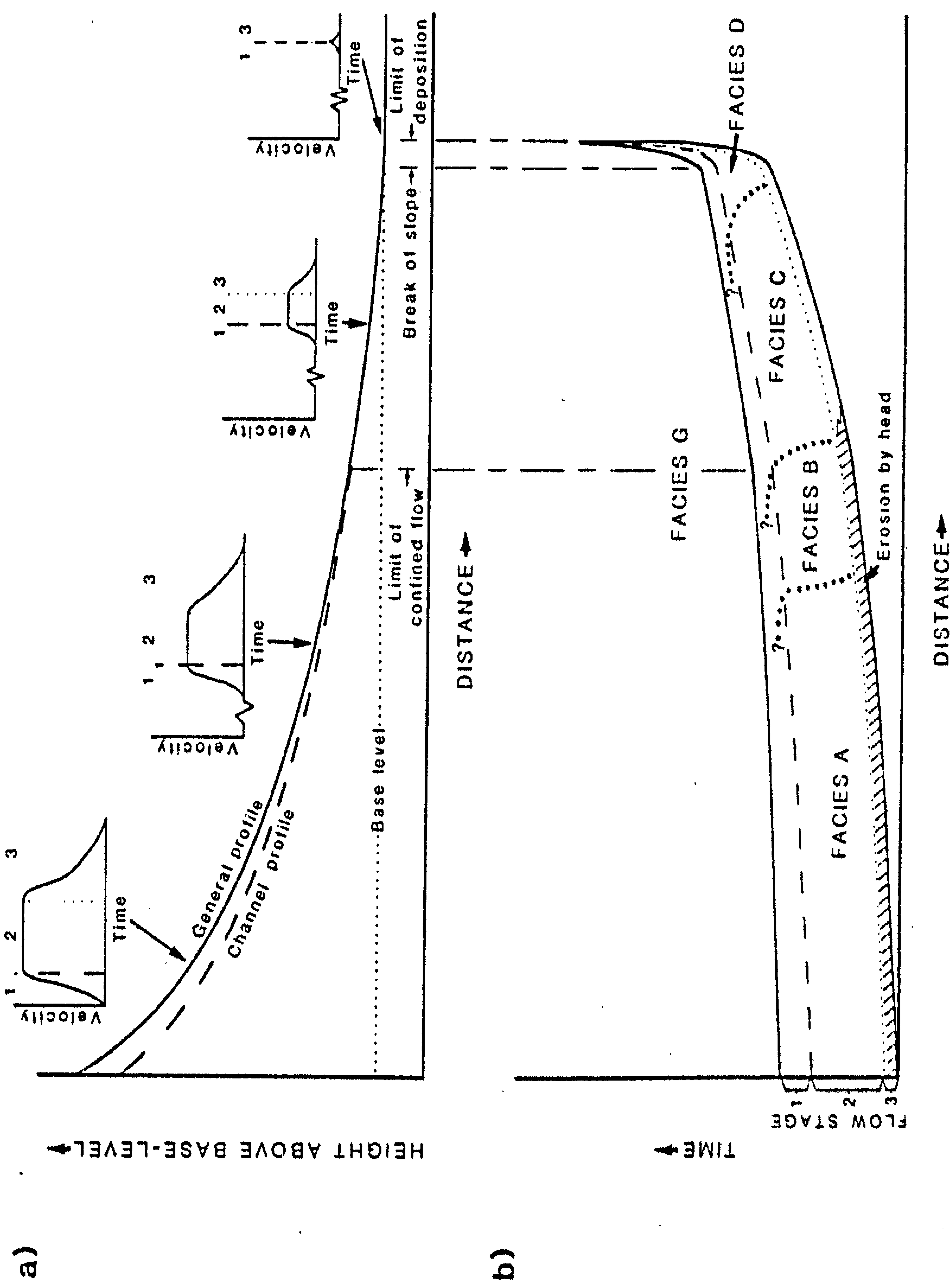

a)
HEIGHT ABOVE BASE-LEVEL
Velocity
Time
1
2
3
General profile
Channel profile
Base level
Limit of confined flow
Break of slope
Limit of deposition
DISTANCE
b)
TIME
FLOW STAGE
FACIES G
FACIES D
FACIES C
FACIES B
FACIES A
Erosion by head
DISTANCE

Figure 3.12b is obviously only one of an infinite suite of such diagrams: there will be a different diagram for each separate flow event in each system. The time and distance scales and the resulting facies stability fields for any event will depend on the initial flow power, the flow period, the grainsize population in the flow and the route the flow takes through the depositional system. During pulsed flows the positions of the stability fields will rapidly change in space during a single event. This is one explanation for the presence of amalgamation surfaces in Facies A (See 3.2.1).

Figure 3.12b shows only proximal to distal variations in a system. It is clear from the facies associations (Section 3.4.1), however, that there is a similar pattern of flow evolution away from confined flows as they expand laterally up and out of their channels (see interpretation of Facies-type A_c and Facies E deposits). The principal impact of the figure, though, is not compromised: the facies are clearly linked spatially due to flow evolution. This point is critical as it justifies the use of Walther's Law in the interpretation of turbidite facies associations.

3.4 Facies Associations and Sandbody relationships.

3.4.1 Facies Association 1: Channel Complexes.

This is the most widely developed association in the Pendle Grit Formation, accounting for about 70% of the rock volume. The facies architecture of this association is best understood in terms of a series of stacked channel-fills related to channels of various scales. A hierarchy of channel sizes can be recognised: smaller channels are developed within the fills of larger scale features. In this sense, the facies architecture is analogous to that of sand-bed braided rivers. The existence of a channel hierarchy is based upon the recognition in the field of a series of convex downwards erosion surfaces of different magnitudes. The largest erosion surfaces, here called first-order erosion surfaces, bound lenticular sand bodies up to a kilometre wide and 100m deep (See Plates 2, 7, 8 and Figure 3.16 for illustrations of these erosion surfaces). These sand bodies are referred to as channel complexes to distinguish them from channel fills which relate to second order erosion surfaces (see below). In the Skipton area, where the first-order surfaces are cut into shales, the cross-sectional geometry of the channel complexes is well defined: topographic features allow the complexes to be mapped (Baines, 1977). Elsewhere the channel complexes are more difficult to define as the bounding first-order surfaces cut into sands of older channel complexes. However, some of these surfaces are exposed (e.g. Nick of Pendle, SD772 385; Longridge Caravan Park, SD618 384; Salterforth Quarry, SD877 453) and can again be mapped by feature. Where this is possible the lateral extent of the channel complexes is comparable with that in the Skipton area, although the margins appear less steep. The along-palaeocurrent dimension of the channel complex sandbodies is not directly measurable in the field. Feature mapping between Wiswell Moor and Nick of Pendle (Grid squares SD75 37 and SD76 38) suggests that the bodies are at least two or three kilometres in length. If the analogy with braided river systems holds then the channel complexes might be expected to form linear sand bodies with a very high length/width ratio.

The internal architecture of the channel complexes is very variable but two general styles have been recognised, corresponding to the architectural styles recognised in turbidite channels in the Apennines (Ricci Lucchi, 1984):

1) Channels whose fill was mainly vertically accreted (Figure 3.13a).
2) Channels showing evidence of significant lateral migration (Figure 3.13b).

As shown in Figure 3.13b the final phase of fill of the Type (2) channels must also be by vertical accretion. Consequently, the two architectures are not mutually exclusive.

In the Pendle Grit Formation the architectures of the channel complexes are significantly more complicated than the simple models above. In vertically accreted channel complexes the base of the fill, immediately overlying the first-order erosion surface, is dominated by thick packets of Facies A sandstones (Figure 3.14a). These pass laterally and upwards into Facies B deposits. Debris flow units (Facies F) are usually found at this level, commonly sitting on the first-order surface (Plate 7a,b). Higher in the channel complex, Facies B and C deposits are dominant while the uppermost part of the sequence is usually a pelitic interval consisting of Facies G silty mudstones and Facies D and E sands. This gross fining-upwards trend is complicated by the presence of numerous channels which represent the second level in the channel hierarchy. These channels are bounded by second-order erosion surfaces with a vertical extent of several tens of metres and which downcut by upto 10m (See Plates 2, 7 and 8). The erosion surfaces generally dip at a very shallow angle and pass laterally into bedding planes beyond the limit of the channels which they define (Plate 8). In vertical

Plate 8: Relationships of facies and erosion surfaces in sand-rich channel complexes.

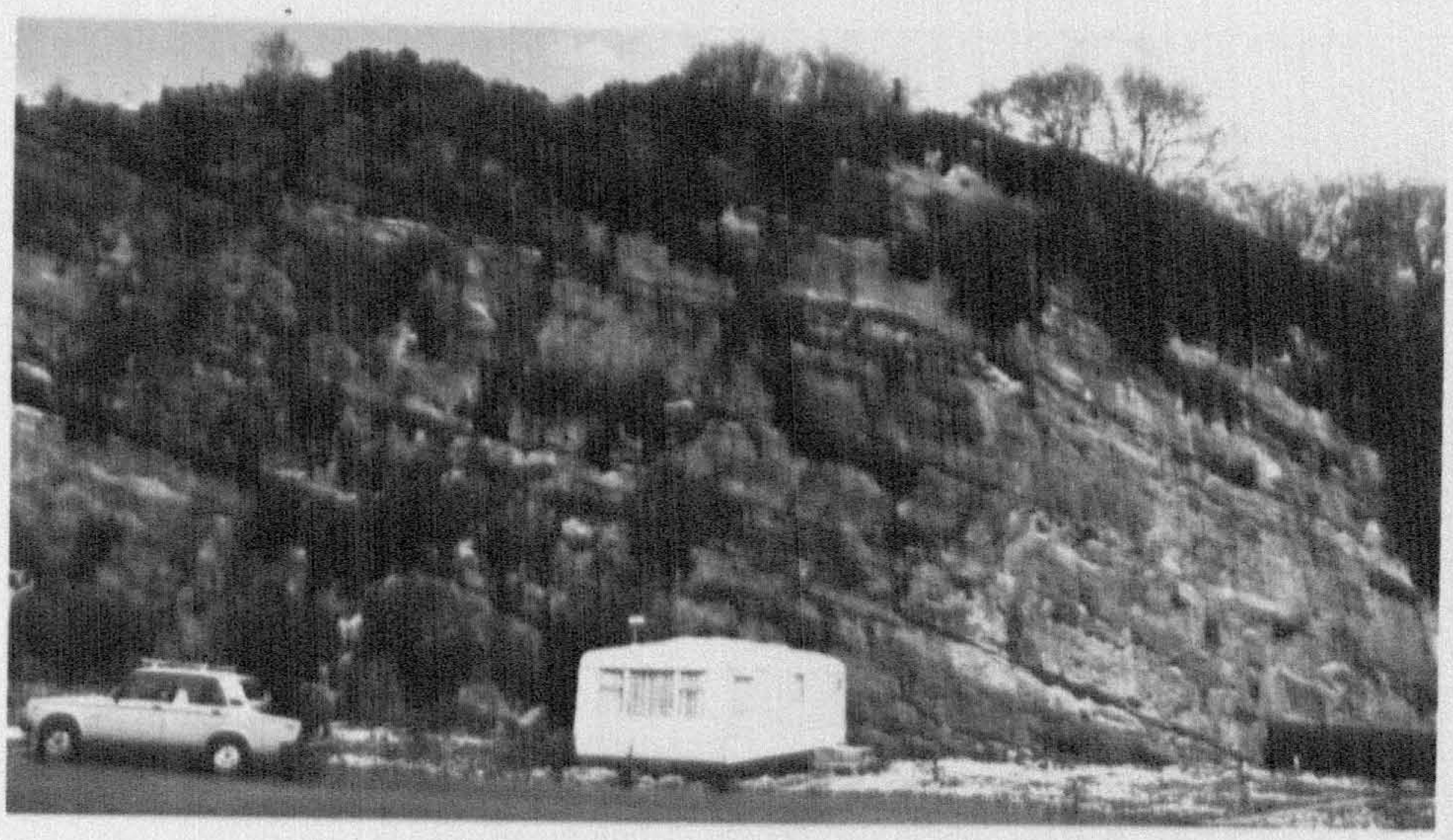
a) The basal units of a channel complex at Nick of Pendle Quarries, SD772 385. The massive, amalgamated Facies A beds on the left of the photograph lie above a first-order erosion surface which is exposed in the quarry and can be mapped by feature on the surrounding hillside. Facies A and B deposits higher in the quarry appear to be parallel bedded but close examination reveals at lest two second -order erosion surfaces (marked). These low-angle surfaces are defined by slight dip changes, erosive cutoffs and onlap relationships. It would be very difficult to recognise such surfaces in vertical section or in boreholes. Quarry face is 17m high.

b) Another sand dominated channel complex showing the internal features of the fill. Palaeocurrent is towards the observer. Quarry face is 13m high. Longridge Caravan Park, SD615 382.

c) Line drawing of (b) to highlight the architecture of the sandbody. Individual facies are labelled and the various erosion surfaces are shown. Second-order surfaces show erosional truncation of underlying beds and are onlapped by overlying beds (see arrows). The latter shows that the erosion surfaces were bounding channels which were active for more than a single flow event rather than being simple scours at the base of a single flow. Note the first-order erosion surface is overlain by a thin unit of claystone. This unit drapes the erosion surface for over 300m up palaeocurrent and seems to represent a phase of non-turbidity current deposition in the first-order channel after its initiation. For further discussion of this point see text. Scale as above.

A

B



C

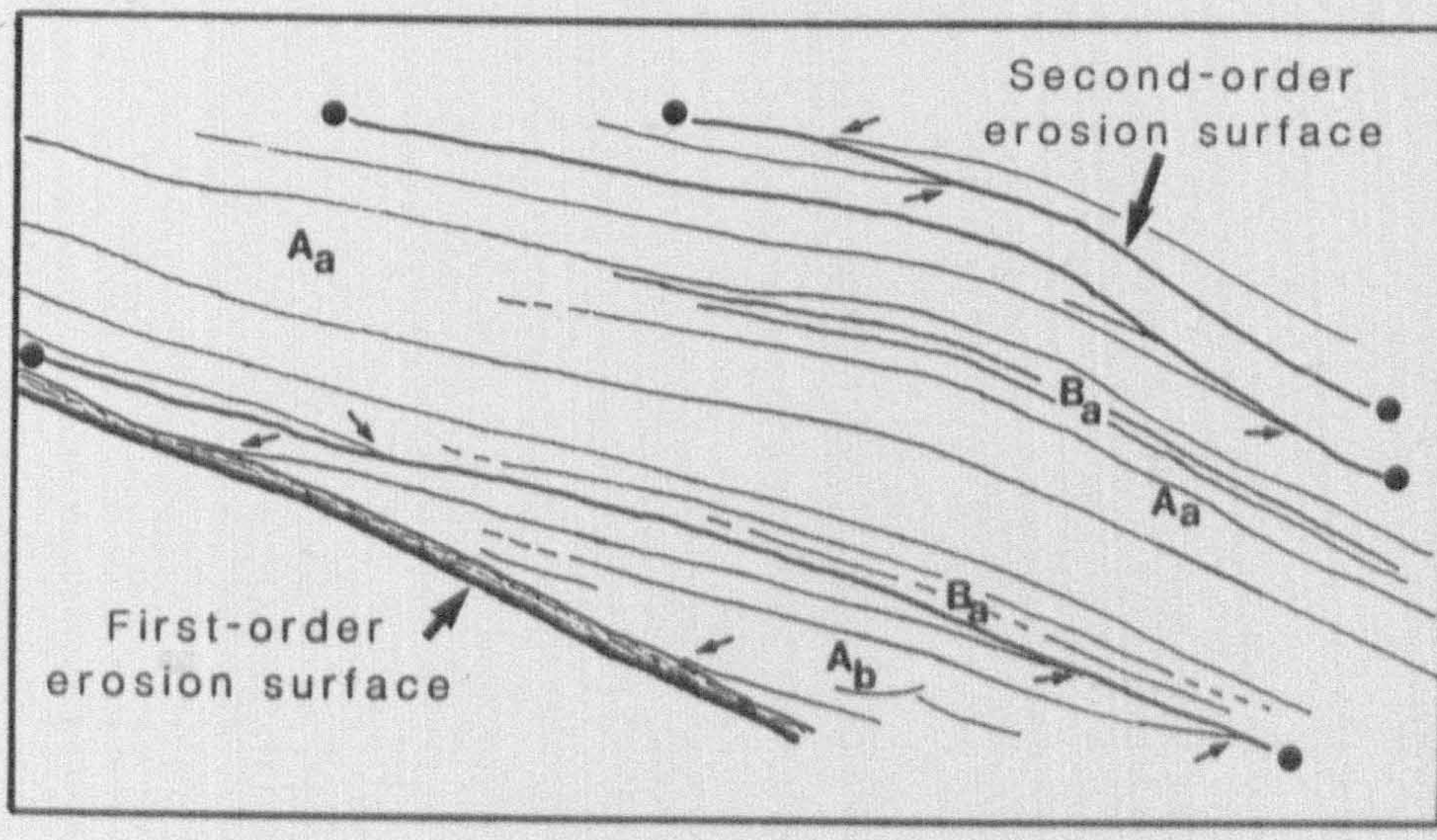

Second-order
erosion surface
A_a
B_a
A_a
B_a
A_b
First-order
erosion surface

Plate 8

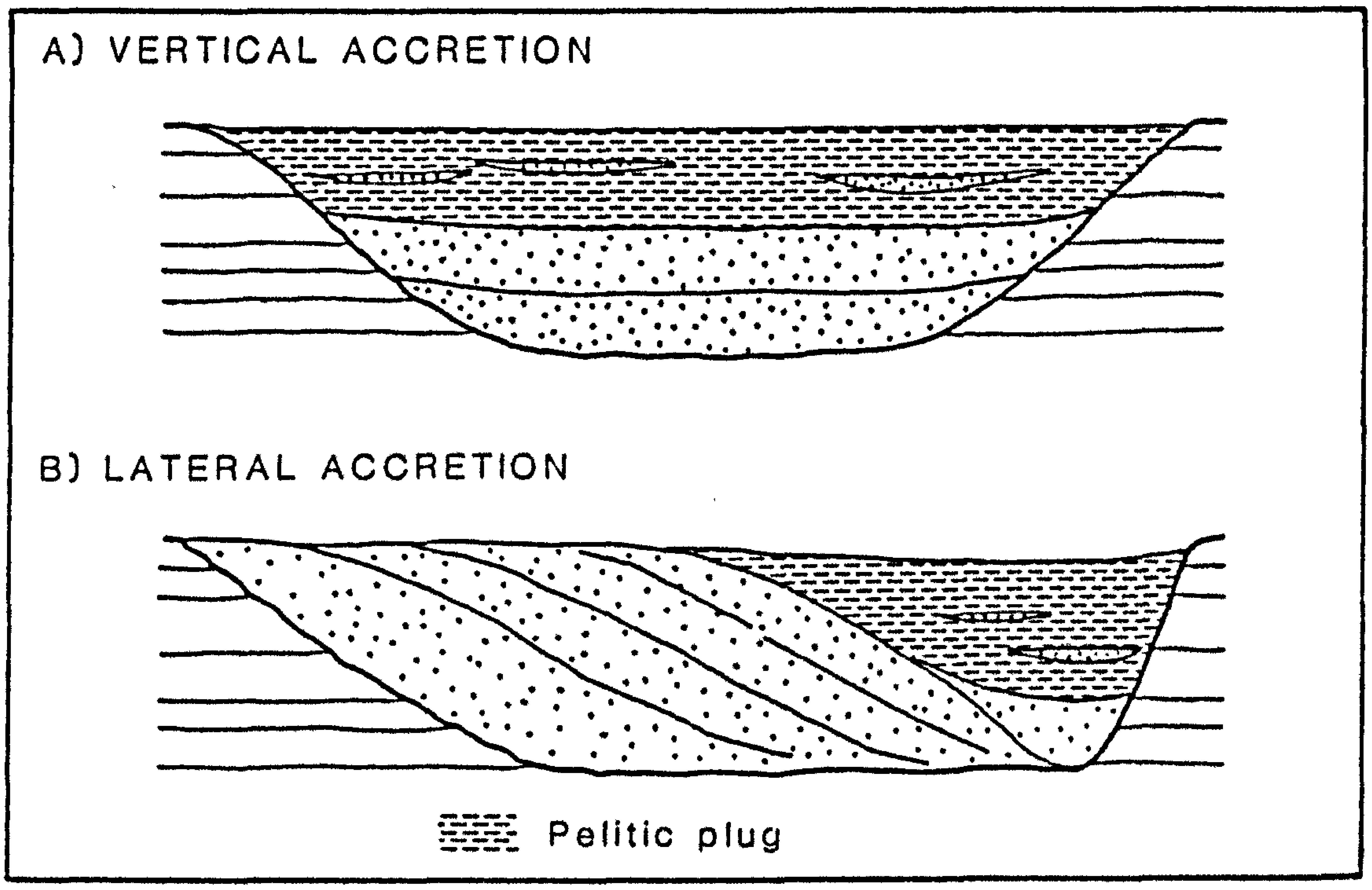


Figure 3.13: Simple models for the architecture of turbidite channel fills. Note that the final stage in the filling of a laterally accreting channel is by vertical accretion. Based on Ricci Lucchi (1984, his Figure 3).

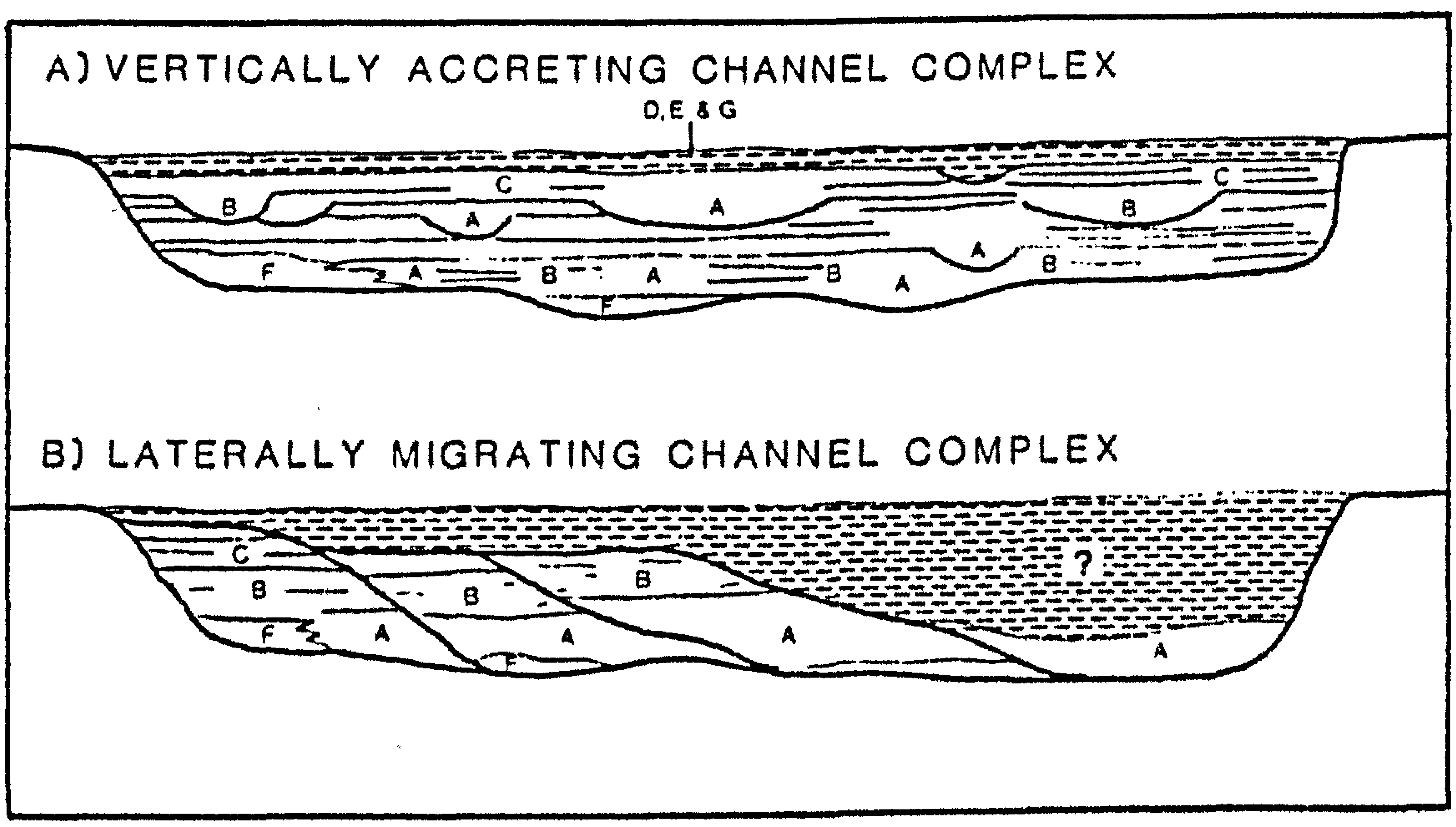


Figure 3.14: Simplified diagrams representing the architectures of channel complexes in the Pendle Grit Formation. Compare with Figure 3.13 and see Plates 2b,c and 8b,c. The width of a channel complex is usually about 1km and the depth some 30-100m. These sketches are not to scale.

Figure 3.15: Log through part of a sand dominated channel complex. Examples of individual facies and facies-types are labelled. The basal first-order erosion surface is defined by feature mapping along strike from the outcrop. Second-order surfaces are defined by bed cut-offs and slight dip discordances between units. In less well-exposed sections these surfaces would be very hard to define. The sketch shows a small Facies E filled scour cut into the top of a Facies-type A_a bed. Such scours could be regarded as third-order erosion surfaces. This scour and the basal 8m of the section are shown on Plate 1. Wiswell Moor Quarries, SD753 371, part of the type section for the Pendle Grit Formation.

For Key see Enclosure 1.

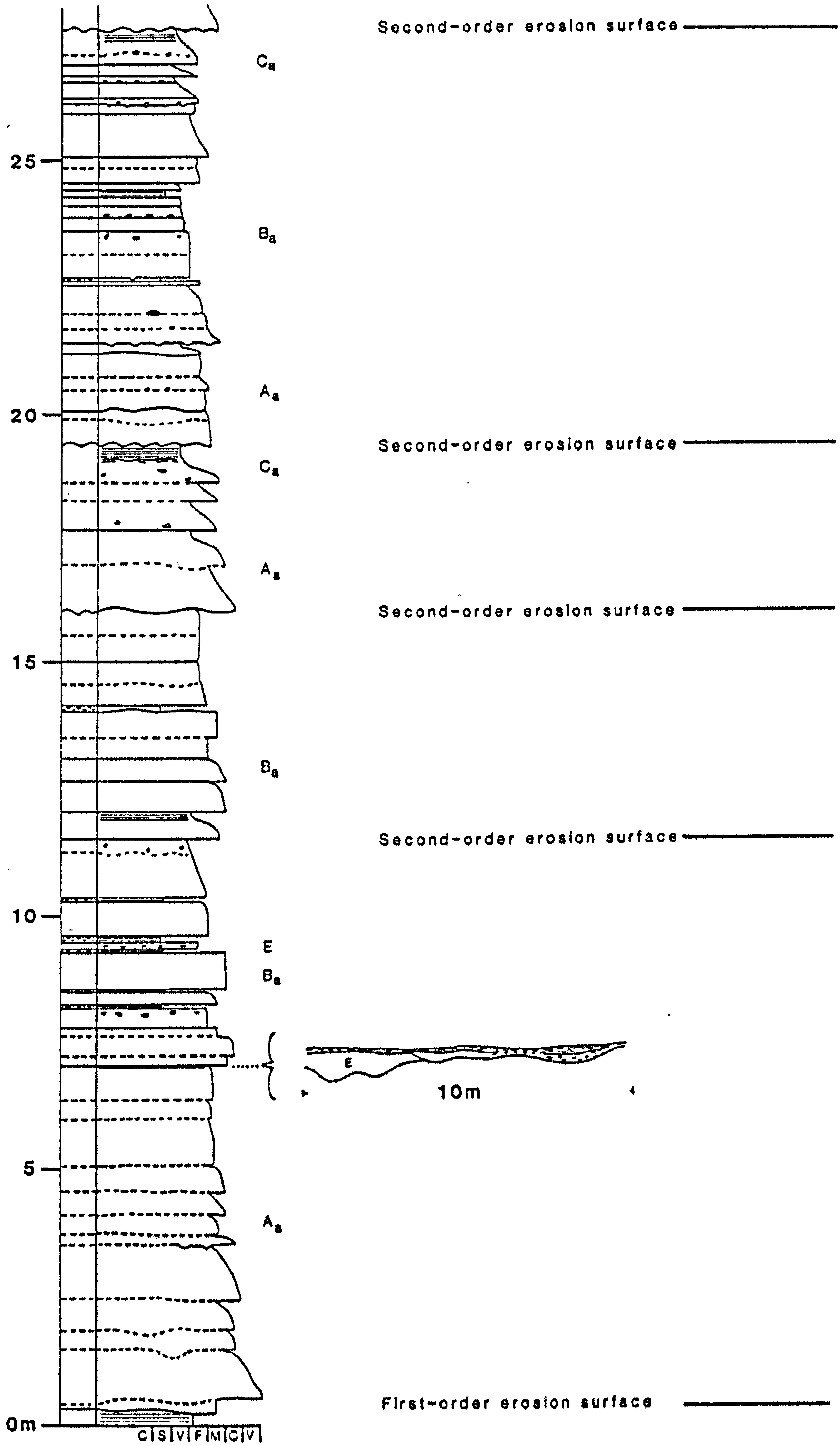

Second-order erosion surface
Ca
25
Ba
Aa
20
Second-order erosion surface
Ca
Aa
Second-order erosion surface
15
Ba
Second-order erosion surface
10
E
Ba
E
10m
5
Aa
First-order erosion surface
0m
C S V F M C V

Figure 3.16: A pelitic plugged channel-complex at Jenny Gill Quarry, Skipton (SE003 509). The Quarry contains three separate channels (1-3) bounded by first-order erosion surfaces (heavy, ornamented lines on diagram). The lowest of these, Complex 1, has a fairly simple fining upwards fill (Log I) ranging from Facies A deposits at the base to Facies D at the top. The channel margins are not exposed but the mapped relationship with the Bowland Shales suggests that they are steep. The dip of the sandstone beds at the base of the section on the north-east wall rapidly reverses. Projection of these reversed dips shows that there must be a step in the level of the base of the channel. Palaeocurrent measurements show this step is almost perpendicular to the flow in the channel. Consequently, the step is believed to be due to a growth fault into which the basal beds of the complex thicken, hence producing the dip roll-over. The second channel complex (2) is much coarser in grade and cuts deeply into Complex 1. There are steep cut banks on the margins of this channel where it cuts through the pelitic units at the top of Complex 1. In turn, Complex 2 is cut by Complex 3. This latter complex contains a basal sandstone lag but the remainder of the fill appears to be entirely pelitic (Log III).
For Key see Enclosure 1 but note that the lithologies on the logs are keyed to the 3-D diagram. Letters on the logs indicate the Facies developed.

3-D INTERPRETATION OF QUARRY

GEOLOGICAL MAP
(showing palaeocurrents numbered by sandbody)

sequences these second-order erosion surfaces are very difficult to distinguish from normal bed boundaries because of this low dip. Towards the base of channel complexes there is often very little variation in the facies across these second-order erosion surfaces: in some cases channels cut in Facies B deposits are filled by Facies A amalgamated sandstones while others may be filled with similar Facies B deposits (Plates 7 and 8). Higher in the channel complexes the second-order channels appear to become more common and to be mutually erosive. Each channel is individually filled by small fining-up packets comprising Facies A, B and C beds (Figure 3.15).

Individual channel complexes vary greatly from the schematic version of Figure 3.14a. Some are dominated by thick Facies A and B packages while others consist mainly of stacked second-order channel fills. The main sand supply to a channel complex may switch off at any time during its infilling. Many complexes are totally sand filled (e.g. Figure 3.15, Plate 8) while there is a complete gradation through to complexes which are almost wholy filled with Facies G and E deposits (Figure 3.16). In all cases where a Facies G/E plug is found the base of the plug is sharp: cessation of sand supply was obviously abrupt.

Laterally migrating channel complexes (Figure 3.14b) have only been recognised at one locality: a partially backfilled quarry at Longridge (SD618 383). The exposure consists of a vertical face 4-10m high which extends along tectonic strike for 300m (Enclosure 2). It seems likely that this scale of exposure is necessary if the geometries described below are to be recognised. Consequently, the lack of other examples of lateral migration is thought to reflect the lack of suitable exposure rather than the architecture being rare.

Enclosure 2 is a simplified line drawing of the exposure. The principal features are two series of low-angle, first-order erosion surfaces[1]. Each series represents the repeated reexcavation of a major first-order channel with each new first-order channel offset with respect to its predecessor. Between the erosion surfaces the facies are similar to those in the vertically accreted complexes: Facies A

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[1] The interpretation of these erosion surfaces as first- rather than second-order surfaces is based on the known smaller scale of second-order surfaces in other quarries along strike from this exposure.

or B deposits onlap the lower levels of the surfaces while Facies C deposits onlap further up. This shows that sedimentation between the events which caused lateral migration of the first-order channel was via vertical accretion processes. In other words, the erosion surfaces do not bound lateral accretion sets. It is interesting to speculate whether the process responsible for the formation of these surfaces might also explain the origin of the so called point bars seen on the modern Mississippi fan (Pickering *et al.*, 1986).

The channel hierarchy discussed above is probably the most important sedimentological feature of the Pendle Grit Formation. The hierarchy implies a variation in the size and power of flows within the depositional system: the first-order erosion surfaces must scale with flows of very large magnitude while the second-order surfaces scale with much smaller flows. Thinking of the braided river analogy again, the first-order channels may scale with the largest flows in the system while the second-order channels scale with smaller, more regular flows trapped by the first-order channel. If this is the case it implies either that the initial mass-flow generation mechanism was able to produce a wide range of flow sizes or, that there were two generation mechanisms. Alternatively, all flows may have been of similar initial size but split up into smaller flows each of which occupied a different first-order channel. These smaller flows then cut the second-order erosion surfaces. In both cases the origin of the mass-flows is likely to have been a point source, or a series of point sources, as it seems unlikely that a linear source (e.g. shelf margin collapse) could produce strongly confined flows in the depositional system down-current.

It is difficult to test which of the two explanations for the flow power variation indicated by the erosion surface hierarchy is most appropriate. Indeed, both may have operated in tandem. The facies variations and the overall fining-up shows that average flow power declined during the filling of a channel complex. This would be compatible with the process of gradual channel abandonment (as in river cutoffs). However, bearing in mind the model for flow evolution discussed in Section 3.3, it is also possible to explain the facies architecture by a multi-energy flow system. In such a system the initial first-order channel, cut by a high energy flow, would trap all subsequent minor flows. These minor flows, being less powerful than the original flow, would deposit sediment within the upper reaches of the first-order channel (Figure 3.17a). Sedimentation in the upper reaches of the channel would reduce the gradient such that the next flow would decelerate even more quickly. Hence the facies stability fields for each flow would migrate up-channel (Figure 3.17b). As this process continues so a fining-up sequence will be deposited at any point in the system up-current of the initial deposits. However, at a certain stage the backfilled first-order channel would reach a geomorphic threshold (Schumm, 1977) whereby it became unstable. Any minor flow with sufficient velocity to reach the enhanced slope down the front of the new depositional body would be accelerated sufficiently to cause erosion (Figure 3.17c). This, and subsequent flows, would regrade the system by headwards erosion. Second-order erosion surfaces, then, might represent flow entrenchment during this regrading process. If the power spectrum of the minor flows is wide then the back-filling/regrading process would be extremely complex and would be capable of producing the complicated architectures seen in the channel complexes.

Other evidence is compatible with this second scenario:

1) Some first-order erosion surfaces are lined by clay beds prior to their being filled (Plates 3d, 8b and 8c). Such clay beds may represent sedimentation in the open first-order channel before the arrival of the prograding second-order channel system.
2) The sharp based pelitic plugs at the top of some channel complexes indicate abrupt cessation of sand supply. The most likely explanation of this is that sudden cutting of a new first-order channel led to complete abandonment of the old channel complex.
3) The lateral migration seen at Longridge suggests that major flow events repeatedly reactivated a single, partially filled first-order channel. Such a process seems unlikely if the fill of a channel complex is by partial flow trapping from an adjacent active channel. This adjacent channel should be more able to take the enhanced flow during a high-power event and reactivation of the older channel would be unnecessary.
4) Observation of a modern sand-rich fan in Bute Inlet, Canada, shows that turbidity currents do not always extend the full length of the channel systems (Prior *et al.*, 1987. See Section 3.6).

In summary, then, facies associations within the channel complexes are believed to represent erosion and depositional events of widely varying power. Large, powerful flows cut first-order erosion surfaces and began filling the

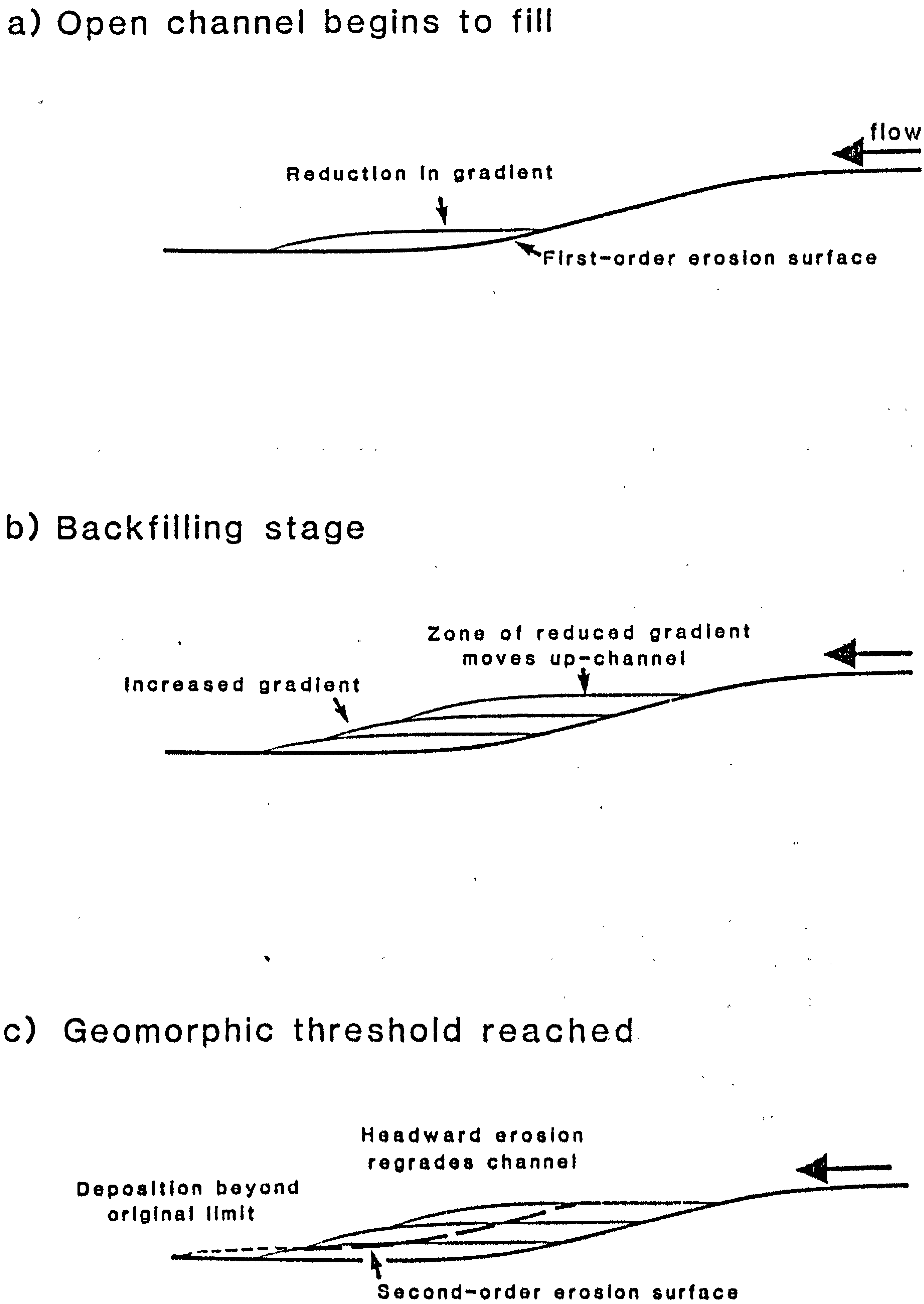


Figure 3.17: Model to explain how first-order channels are filled with coarse sediment. For explanation see text.

resulting major channels. The remainder of the channel-fill was deposited from less energetic flows trapped subsequently by the main first-order channel. These latter flows cut the second-order erosion surfaces. Pelitic plugs represent the total abandonment of the channel complex and its fill by head or body spill currents from other channels. The vertical facies associations in the channel complexes are compatible with the ideas of flow evolution discussed in section 3.3.

3.4.2 Facies Association 2: Parallel Bedded Association.

This association is present throughout the Pendle Grit Formation but is most common in the basal 50m or so. It has a much lower sand/shale ratio than Facies Association 1. The association comprises packets of Facies B and C beds separated by pelitic intervals (Facies G) with interbedded Facies C,D and E beds (Figure 3.18). The sandy intervals are usually two or three metres thick and are parallel bedded. Rarely, they may fill shallow scours upto 1m deep and 10-20m wide. Thin Facies-type F_a debris flows are sometimes found within the pelitic units. When developed at the base of the Pendle Grit Formation, the association contains thick beds of slurried, mud-rich sandstone (Figure 3.19). These beds are gradational between Facies-type F_a debris flows and Facies-type C_c greywackes. There is no overall bed thickness or grainsize trend upwards through Facies Association 2. In vertical sections this association is always overlain by Facies Association 1.

Due to the shale-rich, parallel bedded nature of the sequence, the association is interpreted as representing deposition beyond the mouths of channels. The fact that this association is best developed at the base of the Pendle Grit Formation and that it is always followed by Facies Association 1 suggests that it results from deposition in front of first-order channels. The thick debris-flow/greywacke beds may represent deposition from the major flows which cut the original first-order erosion surfaces. As the flows expanded at the mouths of their channels rapid deposition of the sediment load would occur: the sediment would be especially rich in mud and mud-flakes produced by erosion up-current.

3.4.3 Facies Association 3: Coarsening-upwards Sequence.

The Whitendale Member of the Pendle Grit Formation is built from a single coarsening-upwards sequence. This sequence is recognised as a facies association

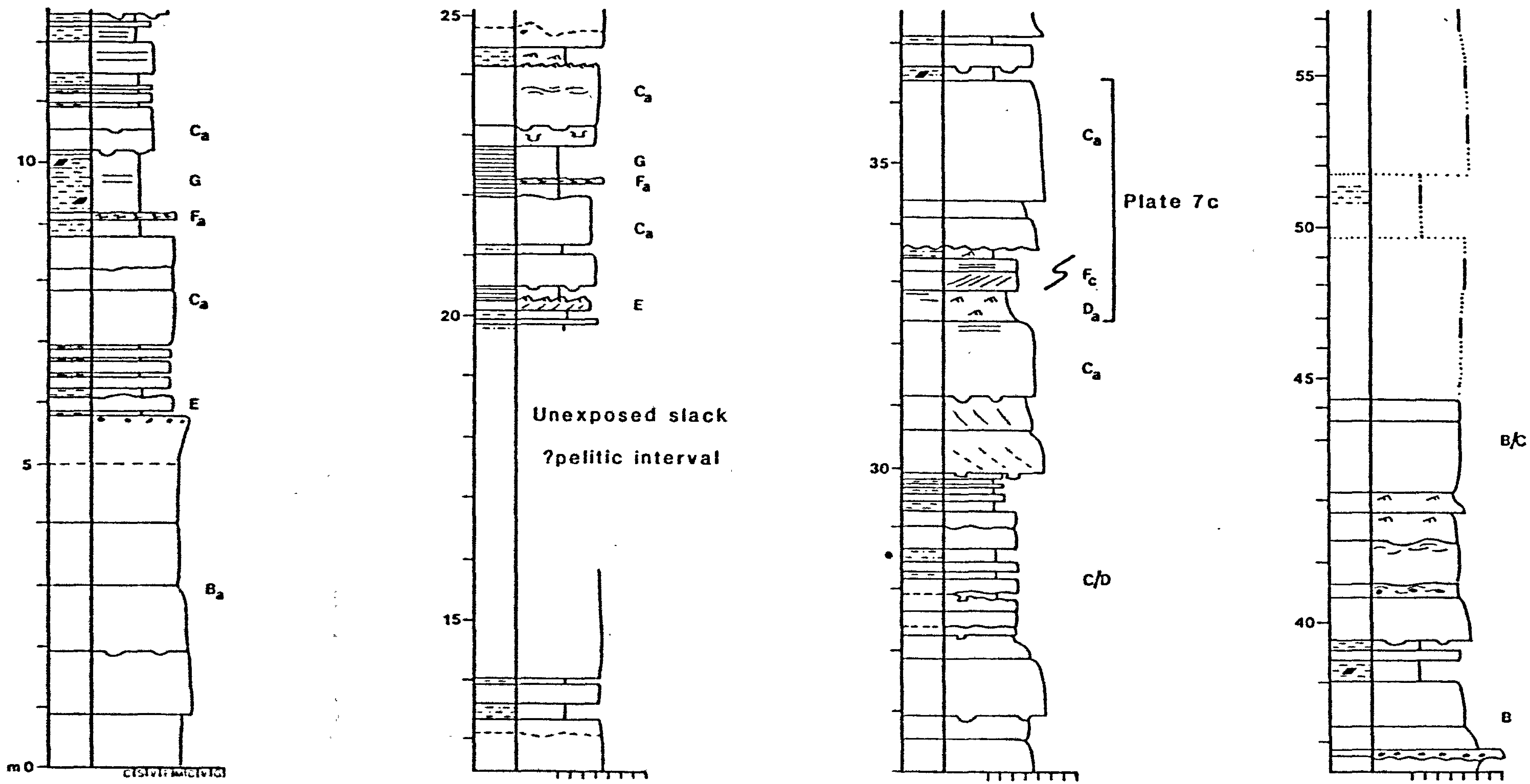


Figure 3.18: Log through Facies Association 2: Parallel bedded Association. Individual facies and facies types are labelled. Note the lower sand/shale ratio (compare with Figure 3.16) of the section and the lack of Facies A deposits. Part of this section is illustrated on Plate 7c. Lower Ogden Clough reservoir SD398 817.
For Key see Enclosure 1.

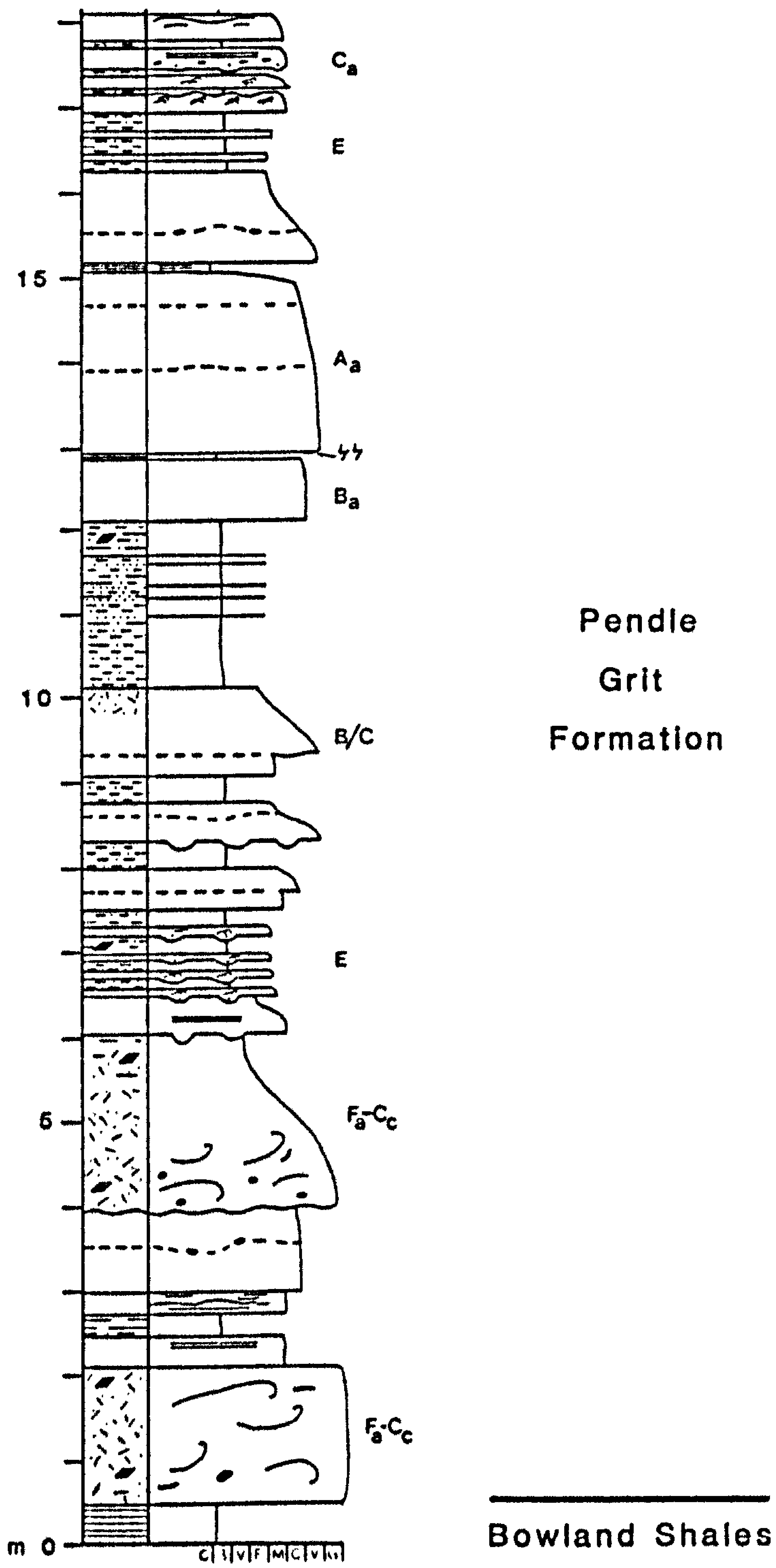


Figure 3.19: Facies Association 2 as developed at the base of the Pendle Grit Formation. This section contains a series of thick "slurried beds" which are characteristic of the basal few metres of the Pendle Grit Formation along the Pendle Monocline. Little Mearley Clough (SD785 411), part of the type section for the Pendle Grit Formation.
For Key see Enclosure 1.

distinct from any others in the Pendle Grit Group. In this association parallel bedding is the norm: the individual beds are laterally continuous on the scale of all the exposures visited. The base of the association is characterised by Facies-type D_b micro fining-upwards units interbedded with thin or medium beds of Facies-type C_c greywacke sandstones (Figure 3.20). Upwards, Facies-type D_b is replaced by coarser Facies-type D_a and cleaner sand beds (Facies-type C_a) become more frequent. The sand/shale ratio continues to increase upwards by the gradual replacement of the pelitic facies by sandy facies. Facies A and B deposits come in at the top of the sequence. Again, the association is succeeded by Facies Association 1 deposits.

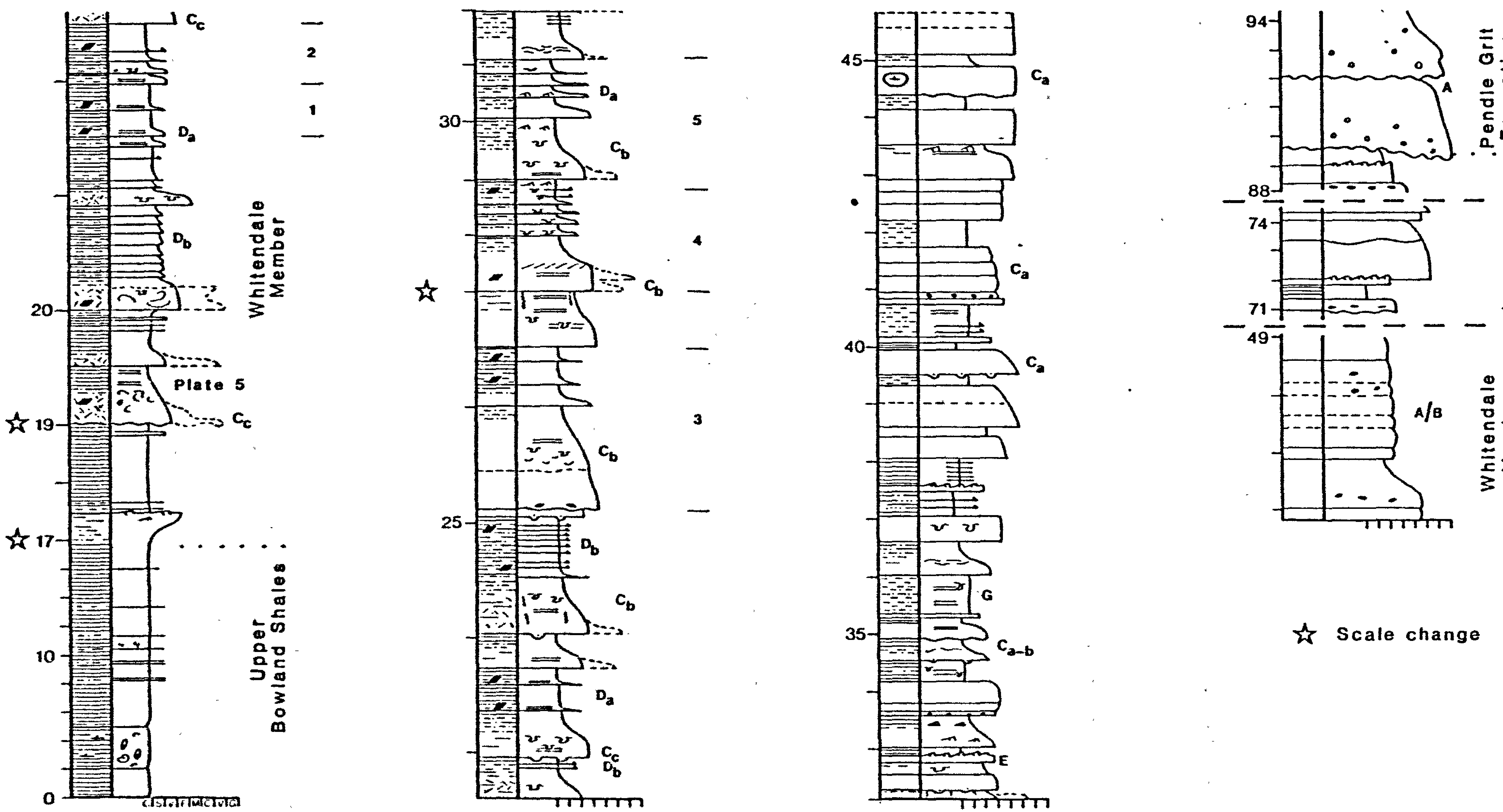
As with Facies Association 2, this association is interpreted as resulting from deposition beyond a channel mouth. In this case, however, sediment was prograding from the channel mouth, hence producing the coarsening-upwards sequence. The limited distribution of this Association (see Figure 2.8) suggests that development of this environment may have been restricted by some palaeogeographic factor.

3.4.4 Facies Association 4: Slope association.

The Pendle Shale Formation is built entirely of this association. The association always lies above the coarser facies associations of the Pendle Grit Formation. It consists of thick sequences of Facies G siltstones and mudstones with rare lenticular sand-bodies throughout. At the base of the sequence these lenticular channel-fills contain mainly Facies A,B and C deposits. They look similar to the second-order channel fills of Facies Association 1. Higher up, the lenticular bodies are built from Facies Z sandstones. Large scale slumping within the Facies G pelitic sediments is common and the Facies Z sandbodies appears to be associated with these slump horizons (Figure 3.11).

This association is interpreted as a pelitic slope deposit. The sandbodies at the base of the sequence are thought to be feeders for the main Pendle Grit Formation fan. The Facies Z sandbodies probably represent gully sandstones (See Section 3.2.8). The association of these sandbodies with slump horizons parallels that seen today in the Mississippi delta/fan system (H. Neilson, pers. comm.). In the Mississippi system, slumping creates topographic depressions on the delta slope which subsequently act as clastic traps. Some may control the

Figure 3.20: Log through Facies Association 3 (Coarsening upwards sequence). Note how the average grainsize of both the coarsest and finest beds in the section increases upwards suggesting progradation of the whole system. This is best shown by looking at the packets of beds labelled 1 to 5. These packets form small "cycles" in which each fining upwards unit contains less coarse material than the underlying bed. Each of the "cycles" is, on average, coarser than earlier "cycles": compare numbers 2 and 5 for example. The origin of these cycles is unknown but they bear some resemblance to the compensation cycles of Mutti and Sonnino (1981). In such cycles the depocentre of each flow event is situated in the hollows between the deposits of earlier events. Thinner beds are deposited over the latter. Hence, at any single location thinner beds will follow each other until the relief due to the first thick bed is "compensated" (at which point the whole process may be repeated).
This is the Type Section for the Whitendale Member of the Pendle Grit Formation. Whitendale River (beginning at SD653 558). The bed at 19m is also illustrated on Plate 6.
For Key see Enclosure 1. Letters refer to individual facies-types.


Upper Bowland Shales
Whitendale Member
Plate 5
Pendle Grit Formation
Scale change
0
10
17
19
20
25
30
35
40
45
49
71
74
88
94
1
2
3
4
5
C_c
D_a
D_b
C_b
C_a
C_{a-b}
G
E
A
A/B
C S vf f m c vc G

positions at which distributaries feed on to the slope. A similar mechanism may have produced the initial relief necessary to trap the gully sandstones (Facies Z) of the Pendle Shale Formation.

3.4.5 Facies Association 5: Ropy Channels.

The Skipton Moor Member of the Pendle Grit Formation is entirely built of this association. The association comprises channels of ropy weathering sandstone (Facies-type B_b) set in Facies G silty muds. The relationship of these channels to those in the Pendle Grit Formation *sensu stricto* is not clear. The erosion surfaces bounding the channels are similar in scale to the first-order erosion surfaces of Facies Association 1 and, stratigraphically, the association lies above Facies Association 1 deposits and below the slope deposits of Facies Association 4. Consequently, by Walther's Law, the ropy sandstone association probably represents slope feeder channels to the first-order channels of the Pendle Grit Fans. The limited distribution of the ropy channels (See Figure 2.8) suggests there may be a palaeogeographic factor controlling the development of this facies association.

3.5 Development of a facies model for sand-rich fans based on the sedimentological features of the Pendle Grit Group.

In Chapter 2 it was said that all the stratigraphic units in the Pendle Grit Group were linked sedimentologically. The nature of this link is now apparent: the Pendle Grit Formation represents a submarine fan system and the Pendle Shale Formation the related slope system bypassed by the bulk of the coarse sediment. The areal extent of individual environments within this system can be determined by mapping the facies associations. As each of the stratigraphic units within the Group (with the exception of the Pendle Grit Formation) is also a single facies association, the isopach maps presented in Chapter 2 (Figures 2.7, 2.8, 2.9 and 2.11) show the distribution of the various sedimentary environments within the depositional system. Similarly, Figures 2.3 and 2.6 illustrate the vertical distribution of the facies associations. Although the spatial relationships of the various depositional environments can be mapped in this way, the temporal relationships remain uncertain due to the absence of time lines within the Pendle Grit Group. This leaves much scope for modelling how the depositional system actually filled the basin. Without time control, it is difficult to determine whether the system was prograding or simply aggrading. Further discussion regarding this question will be left to Chapter 6 where data from the Grassington Grit Group depositional system and the petrology will be used to help address this question. In this section the facies associations and facies distributions seen in the Pendle Grit Group will be used to develop a new conceptual facies model for sand-rich submarine fan systems.

From the isopach maps and the interpretations of the various facies and facies associations, the following important points about the Pendle Grit Formation fan system can be made:

1) The fan system is very sand-rich with a gross sand/shale ratio exceeding 5:1.
2) There is only minor vertical and downcurrent variation within the fan. Facies sequences are dominated by Facies Association 1 channel complexes, particularly along the Pendle Monocline. Most of the channels are cut into pre-existing channel deposits of the same system.
3) There is a hierarchy of channel forms which is believed to result from dramatic variations in the sizes of flows within the system. The coarse-grained facies within the channel fills have features which suggest prolonged flow events.
4) In volume terms, the proportion of Facies Association 2 and 3 pro-channel deposits ("lobes") is very low. Except where the Whitendale Member is developed (i.e. Facies Association 3), there is no evidence of progradation of the whole system into the basin: the channels arrive very early in the history of sand deposition, often right at the base of the Formation.
5) There are no obvious patterns to the facies distributions or isopachs which would suggest radial growth of the sandbodies. Palaeocurrents parallel the north-east to south-west structural grain in most areas (refer to Chapter 6).
6) The Pendle Shale Formation slope deposits always overlie the Pendle Grit Formation fan, when both are developed. The Pendle Shale Formation, then, must represent slope progradation either after or during deposition of the fan system. The latter seems more likely because of the presence of the turbidite filled channels within the slope association (Facies Association 4).

These features cannot be reconciled with classical models of fan sedimentation (Mutti and Ricci Lucchi, 1972; Walker, 1978. See review in Pickering *et al.*, 1986). In these classical models (Figures 3.21 and 3.22) fans are seen as approximately radial sediment bodies with a distinct proximal to distal decrease in grain-size and sand/shale ratio. Channels are confined to the up-current parts of the fans (Upper and Middle fans). Progradation of the fan system should produce a coarsening-upwards sequence with sand-rich channel deposits only present at the top of the section. These predictions from the classical models are obviously incompatible with the features of the Pendle Grit Formation sand-rich fan listed above. Consequently, a new conceptual facies model has been developed for sand-rich fans.

The new facies model is based on the architectural elements discussed in Section 3.4 and on the distribution of these elements as shown by the isopach maps presented in Chapter 2. In the model sand-rich fans are seen as developing by a process of stacking channel complex sandbodies (Figure 3.23). First-order channels, cut by the high power flows, extend right across the depositional system. They are filled by the deposits of smaller, more frequent flows which prograde down the first-order channels, cutting and filling second-order channels in the process (Figure 3.23).

At the mouths of the first-order channels small lobes are deposited (Facies Association 2). These lobes contain sediment deposited from both the initial major flow (the greywacke/debris-flow beds) and from occasional minor flows which extend beyond the first-order channel mouth. More classical, coarsening-upward lobe deposits are produced (Facies Association 3) when progradation of the channel-filling sediments extends beyond the mouth of the first-order channels (Figure 3.23). Periodically, new first-order channels are initiated by major flow events, leading to the rapid abandonment of the older first-order channel. The latter are then filled by pelitic material deposited by spill currents from the new channel (Figure 3.23).

Not all new first-order channels extend to the margins of the fan: some reach the upper parts of the fan only. Small lobes developed in front of these channels. Such lobes provide a mechanism for the deposition of Facies Association 2 at any level in the sand-rich fan system.

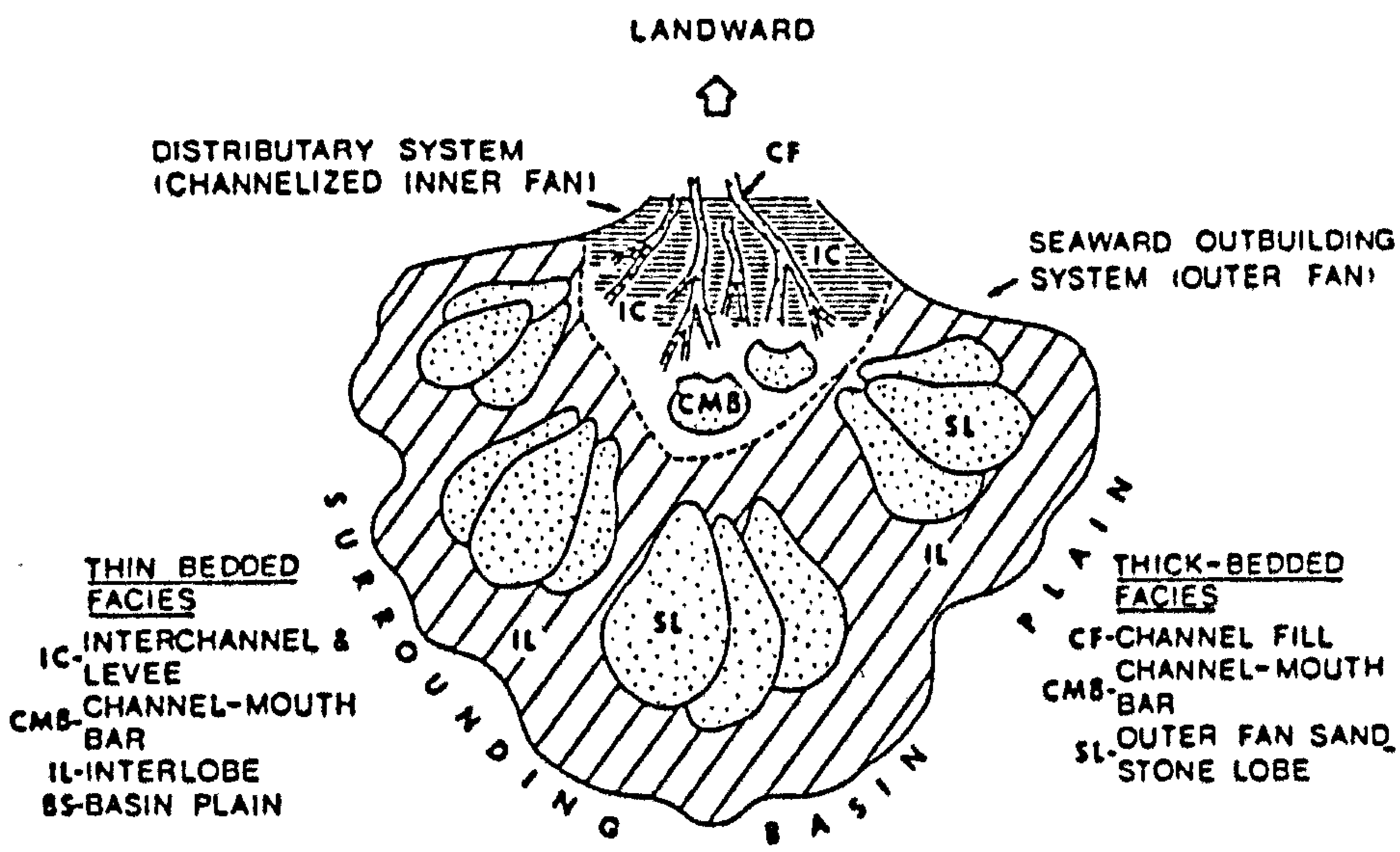


Figure 3.21: Mutti and Ricci Lucchi's 1975 facies model for submarine fan systems. Note the model fan is dominated by lobe deposits with feeder channels confined to a small part of the fan. The bulk of the facies sequence produced by progradation of such a fan would comprise thin-bedded turbidites representing the basin plain and the outer fan lobes.

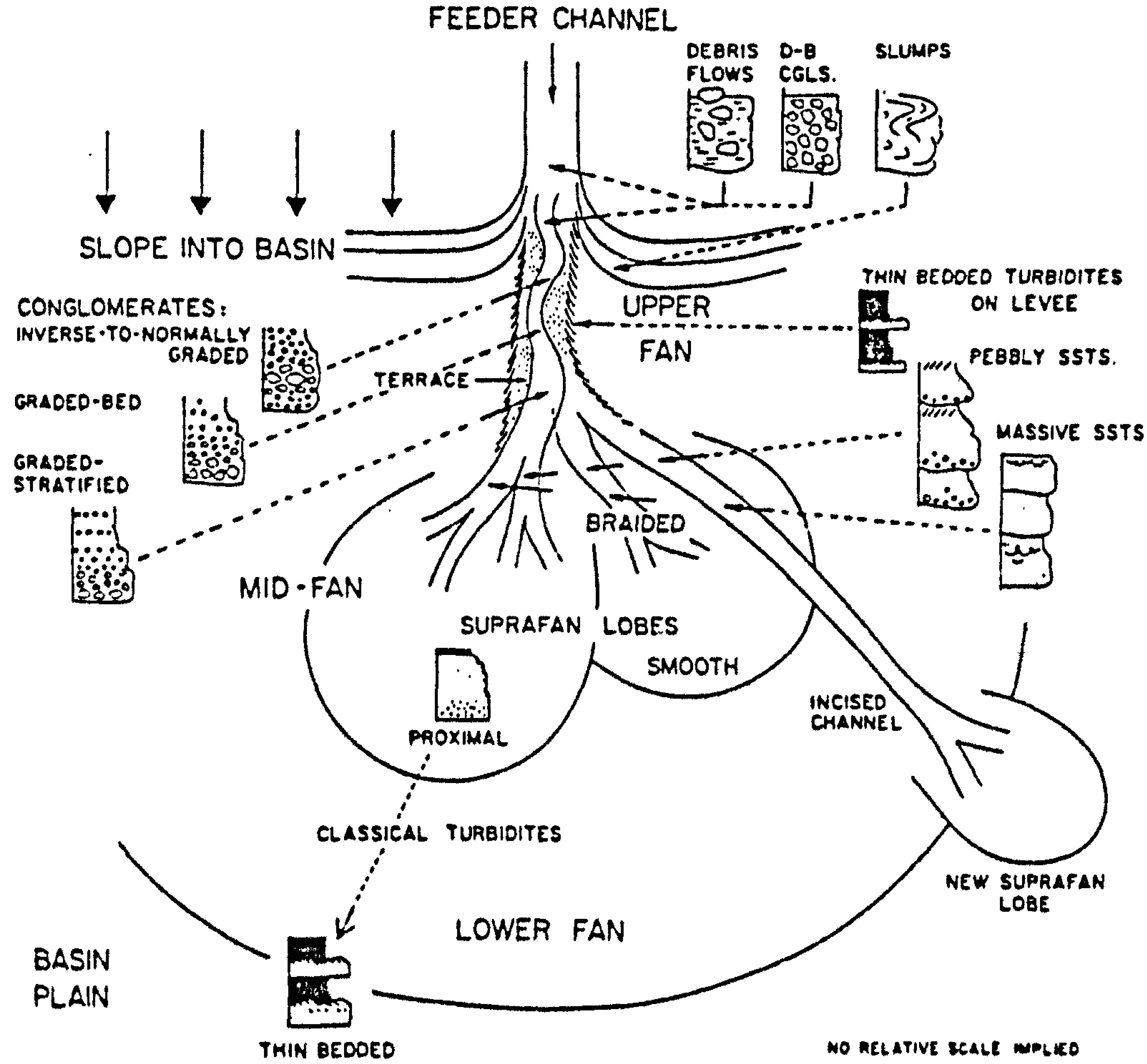


Figure 3.22: Walker's classic 1978 fan model, still widely and successfully used today. Channel sediments are more important compared to the Mutti and Ricci Lucchi model (Figure 3.21) but classical turbidites of the outer fan are still very prominent in the model. The suprafan lobe of the model was an attempt to link the morphology of modern fans with facies associations in ancient examples. The success of such an approach has been discussed by Mutti and Normark (1987).

1) Abandoned channel complex filling with Facies G pelitic deposits and Facies E sandstones sourced from spill currents. Note the "point bar" formed by lateral migration of the channel complex during its active stage.

2) Active channel complex filling by progradation of Facies A, B and C deposits down the first-order channel. Second-order channels cut into the top of the prograding sediment package. Some Facies F deposits are produced by bank collapse.

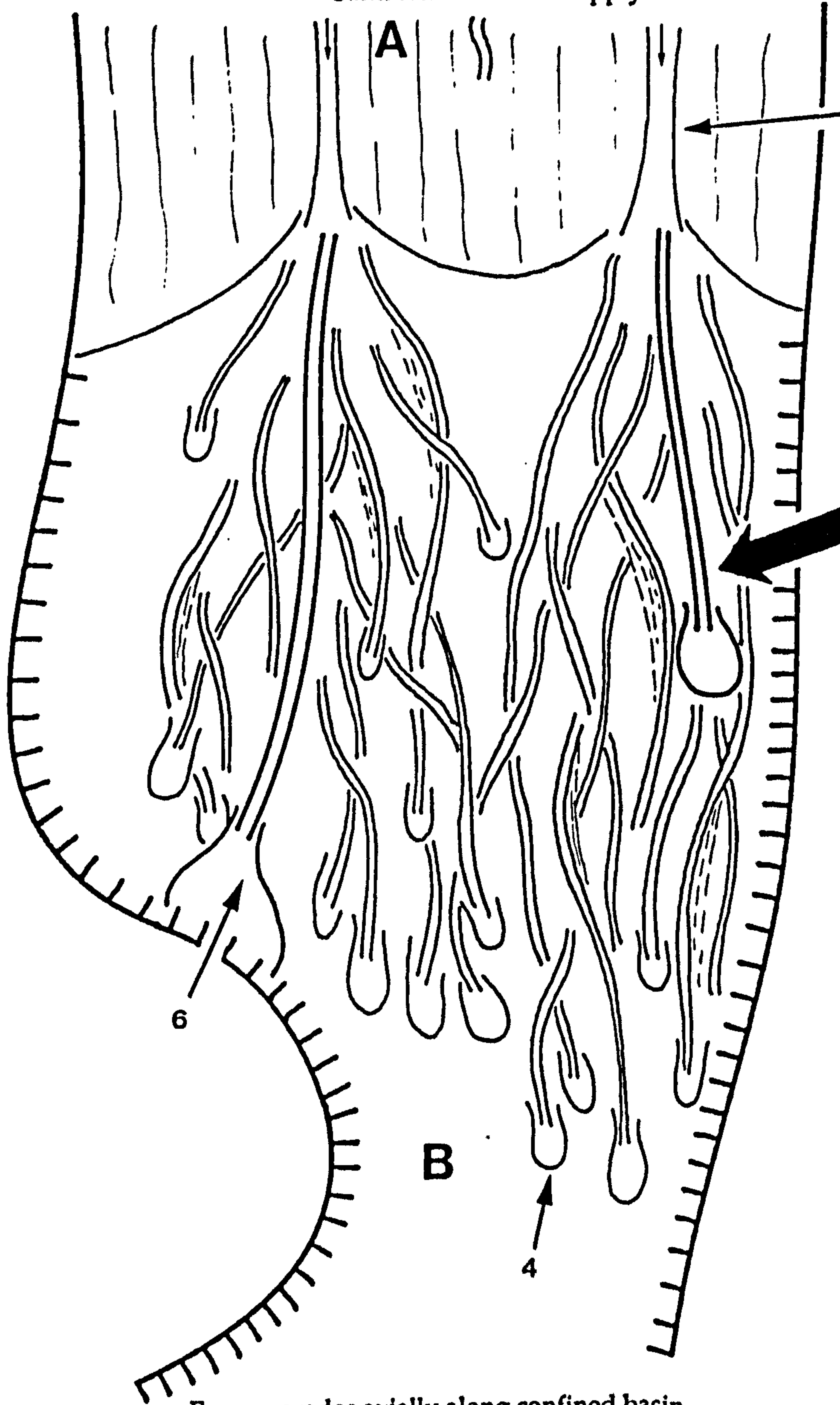
3) Major cause of basin confinement is likely to be active faulting or differential subsidence of the fill of older fault bounded basins. This enhances the stacking of the channel complex sandbodies. Palaeocurrents parallel local topographic (i.e. tectonic) trends.

4) Terminal lobes to first-order channels (Facies Association 2). Note small size of lobes with respect to the channel length.

5) Feeder channels (Facies Association 5) cutting across pelitic slope with minor gullies (Facies Association 4).

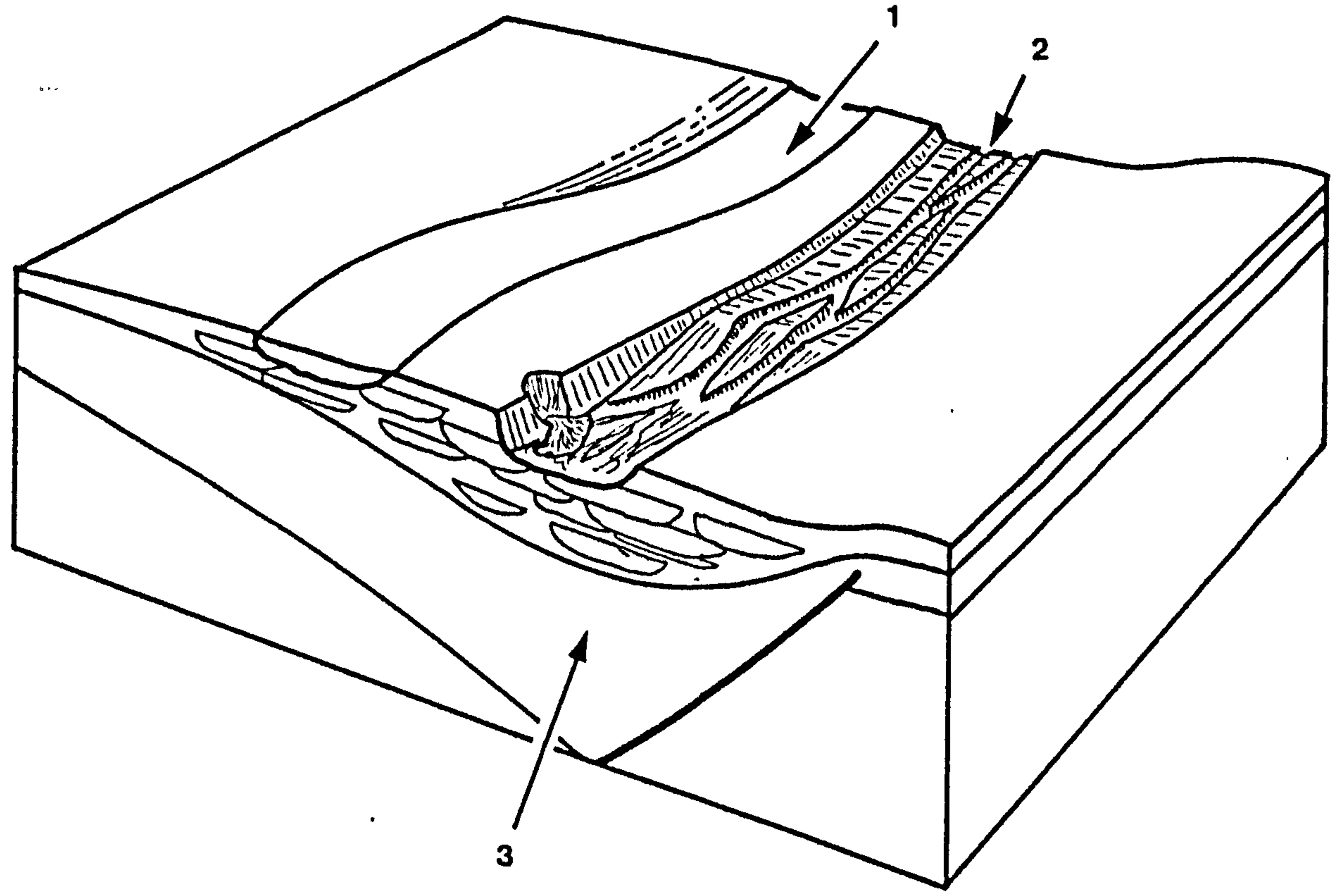
6) Sediment progrades beyond end of first-order channel. Lobe enlarges and starts climbing reverse gradient at basin margin. Coarsening upwards sequence deposited (Facies Association 3).

CHANNEL COMPLEX FAN MODEL

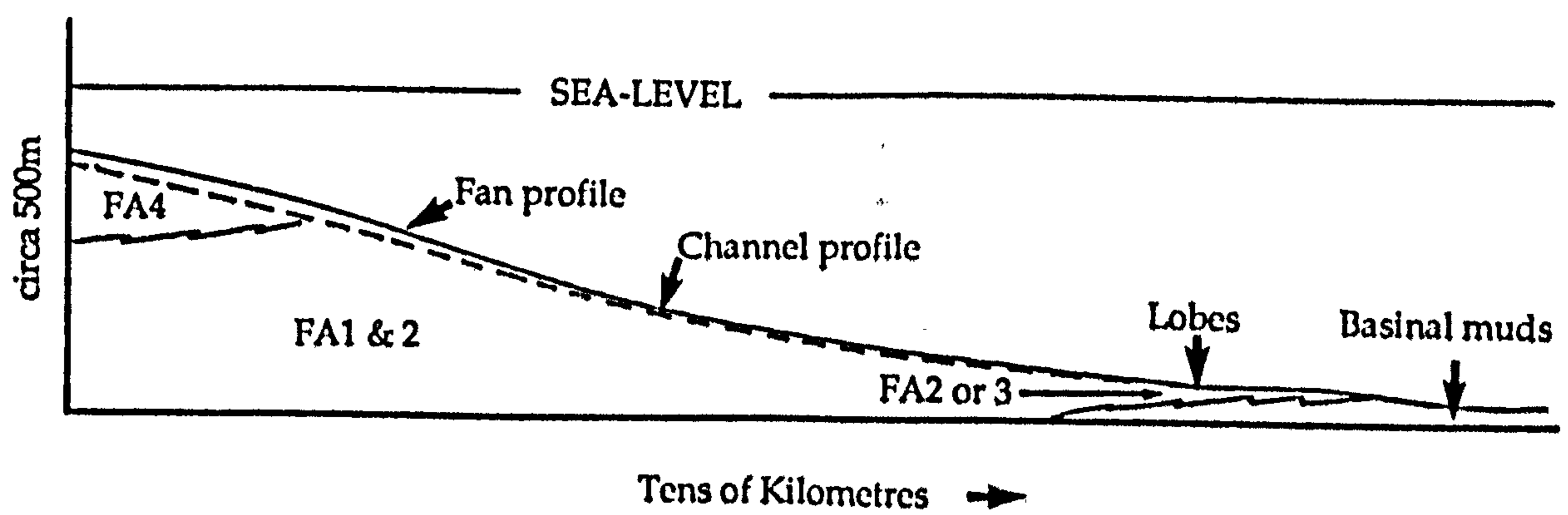


Fan progrades axially along confined basin.

DETAIL SHOWING PROCESSES IN FIRST-ORDER CHANNELS



SECTION A-B SHOWING RESULTS OF FAN PROGRADATION



FA1-3 equivalent to the Pendle Grit Formation.
FA4 equivalent to the Pendle Shale Formation.
Basinal muds equivalent to the Bowland Shale Formation.

Note: scales are those appropriate for the Pendle Grit Formation.

Figure 3.23: Conceptual facies model for sand-rich, channel dominated submarine fans. Based on the facies associations and sequences seen in the Pendle Grit Group. Facies Associations are shown on the diagram by numbers. For discussion see text.

The orientation of the channel complex sandbodies is controlled by the palaeotopography of the basin which may be dependent on active faulting or differential subsidence over buried faults. Confinement of the receiving basin by such mechanisms prevents radial distribution of sediment and enhances the stacking process. Obviously, a sand-rich sediment source to the basin is a prerequisite. Channels dominate the fan because of this sand-rich nature: the sandy facies from which the fan is built would be easily erodeable compared with the mud dominated sediments of more classical fans.

The slope deposits overlying the fan are deposited by the turbid plume thrown up by the current passing through slope channels (Facies Association 4 and 5). Progradation of the whole system advances the slope deposits over the fan deposits.

Ultimately, the source of the turbidity currents is irrelevant to the model: the source can be a delta local to the fan or a shelf many tens of kilometres up-gradient from the fan. So long the turbidity currents can reach the top of the slope in equilibrium, and are carrying the right grainsizes in suspension, a sand-rich fan can develop. The present author's experience of Jurassic sand-rich fan systems in the North Sea suggests that turbidity currents may move sediment over distances in excess of 150km before deposition begins. The nature of the turbidity current generation mechanism and the transport distances for the specific case of the Pendle Grit Formation fan are dealt with in Chapter 6.

3.6 Discussion of the channel complex model and comparison with other fan models.

The channel complex fan model differs from models described to date in that the whole fan is dominated by channels within which sand deposition occurs. There is very little lobe development in front of the channel mouths. The basis of this model is the observed field relationships between the various facies and the hierarchy of erosion surfaces. The model is critically dependant on the observed relationships being representative of the system as a whole. Two questions arise:

1) Does the present outcrop represent the whole of the depositional system or just the channelised, sand-rich part of a large classical turbidite system whose distal facies are present under the East Irish Sea Permo-Triassic deposits?

ADDENDA

BAINES, J.G., 1977. The stratigraphy and sedimentology of the Skipton Moor Grits (Namurian E_{1c}) and their lateral equivalents. Ph.D. Thesis, Keele.

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Page 68 line 18: for inter-event read intra-event.

Page 111 line 31: for Neilson read Nilson.

Pages 165 line 15 and 192 line: for oligioclase read oligoclase.

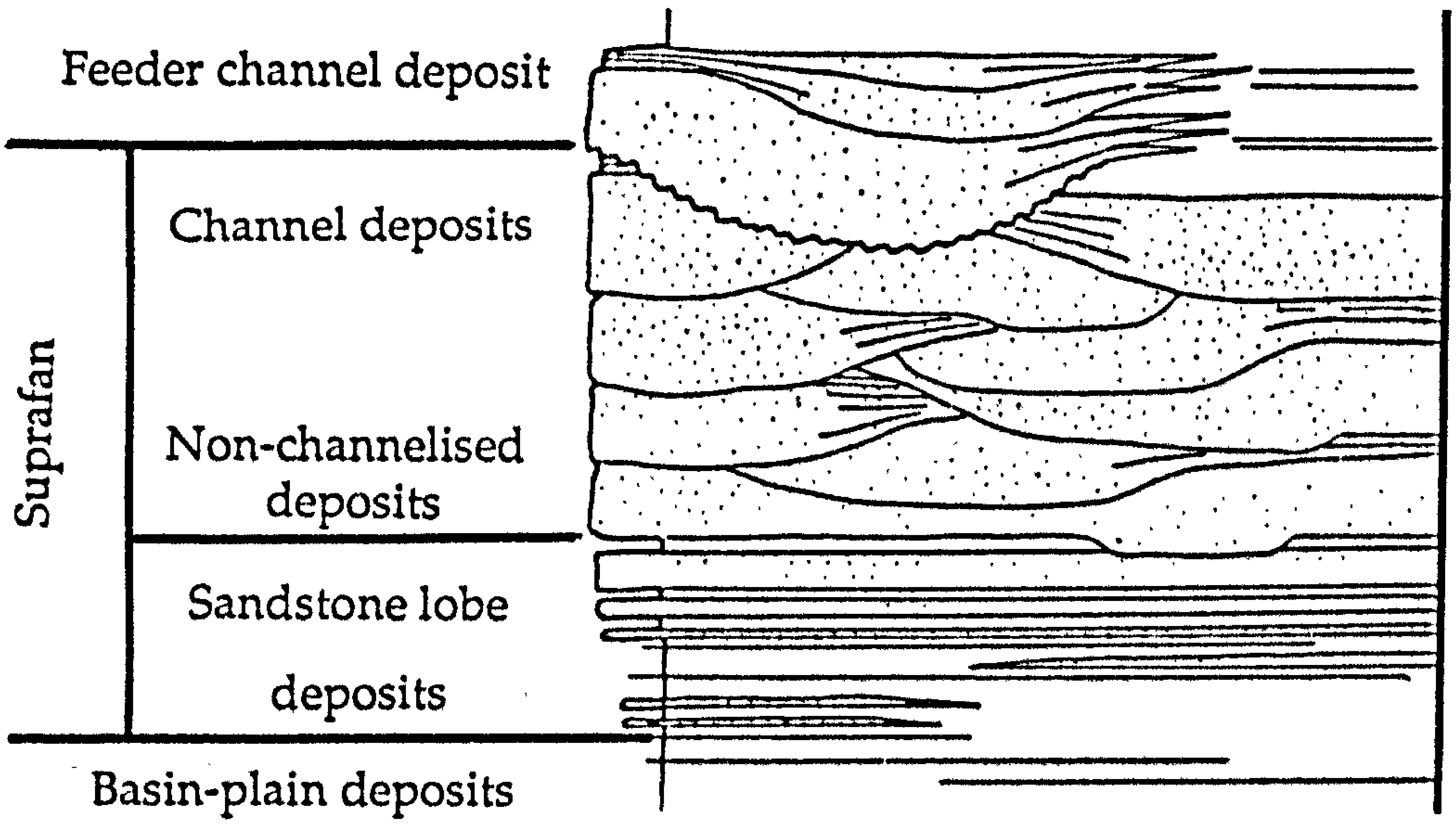


Figure 3.24: Model for small, sand-rich fans of the Apennines (Mutti, 1978). Note the implied presence of a thickening and coarsening-upwards sequence beneath the main channelled units. This sequence would be equivalent to Facies Association 2 or 3 of the Pendle Grit Formation. There is no channel hierarchy described in the channel deposits of these fans.

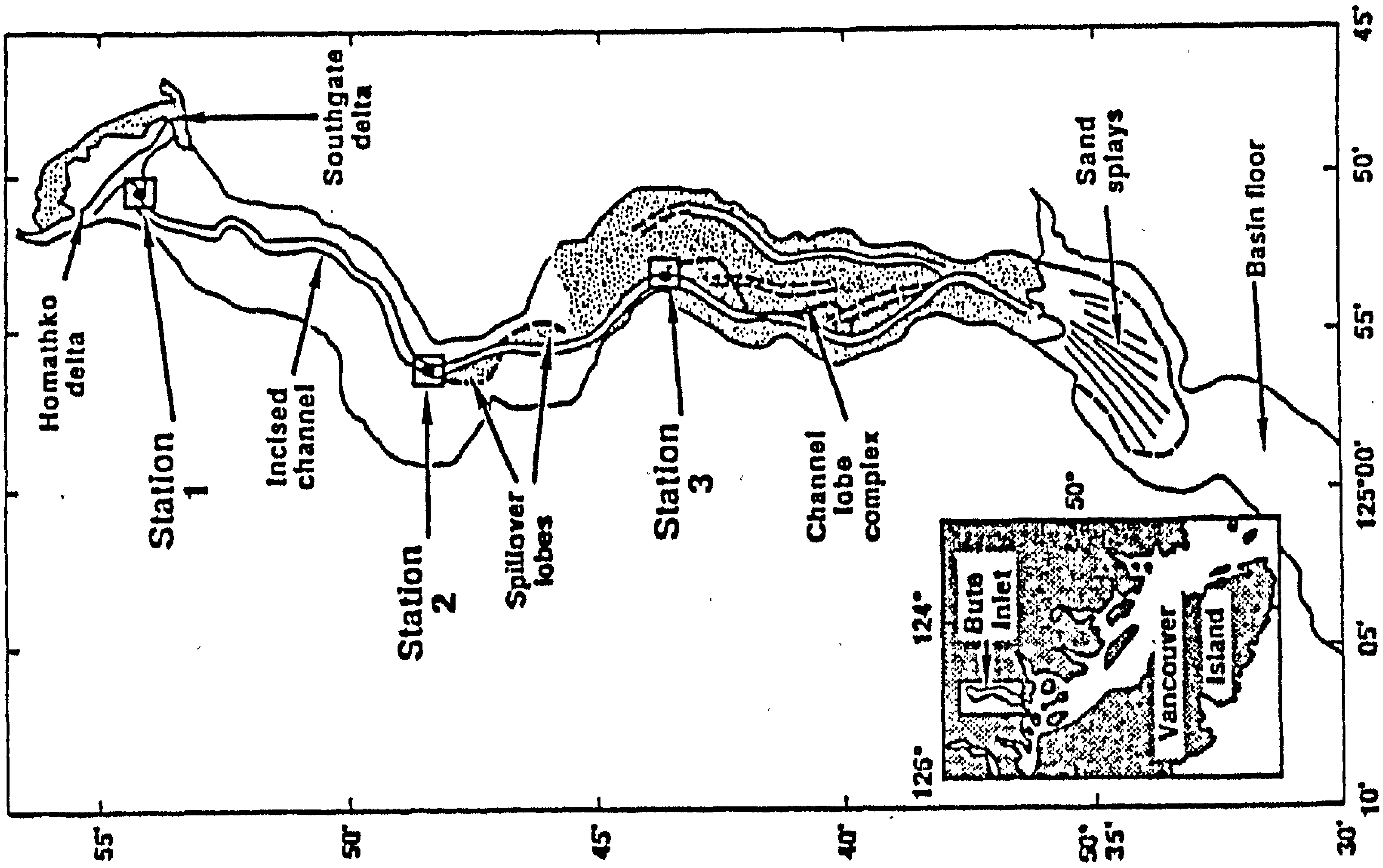
development of the sand-rich Apennine fans. Both factors are important elements of the channel complex fan model and it seems likely that they are prerequisites for the development of a sand-rich fan.

The problems of comparing modern and ancient mass-flow systems have been admirably summarised by Mutti and Normark (1987). They show how the differences in scale, observation techniques and data types between modern and ancient systems prevents any one-to-one mapping of the physical morphology of modern fans with the facies associations in the rock record. Consequently, no attempt will be made to do this for the Pendle Grit Formation fan. In fact, even at the level of the whole fan system it is impossible to find examples of modern fans which are similar in setting, style and facies to the Pendle Grit Formation fan. However, there is one fjord fan system which provides a modern analogue for the channel complex model. Bute Inlet, Canada, is floored by a channel dominated fan system fed by a braid-delta system at the head of the fjord (Prior *et al.*, 1987. See Figure 3.25). Water depths and gradients in this system are probably comparable with those which existed in the Bowland Basin during deposition of the Pendle Grit Formation (See Chapter 6). Similarly, the limited width of the fjord mimics the confined nature of the Bowland Basin. Sediment supply to the system is very sand rich (15% gravel, 65% sand, 15% silt and 5% clay). This sand is transported from the braid-delta to the fan system by turbidity currents. Significantly, current meters show two frequencies and magnitudes of events. Quoting from Prior *et al., op. cit.*:

"*(There are) frequent low-velocity, limited distance flows and relatively infrequent, large-scale events that travel long distances at high velocities*"

This modern power and distance spectrum for the flows on the fan is exactly the same as that proposed herein to explain the channel hierarchy in the Pendle Grit Formation. Note this model was developed from outcrop data alone before the Bute inlet data was available to the current author. Not only are the processes comparable but the facies appear very similar. Sampling from the fan surface shows the modern deposits are directly comparable with the facies in the Pendle Grit Group. Figure 3.25 shows how the facies associations of the Pendle Grit Formation might relate to the environments of the Bute Inlet Fan.

Clearly, this modern analogue provides powerful support for the channel complex model developed from the Pendle Grit Formation. Whether or not the new model is useful will depend on its predictive powers when applied to other



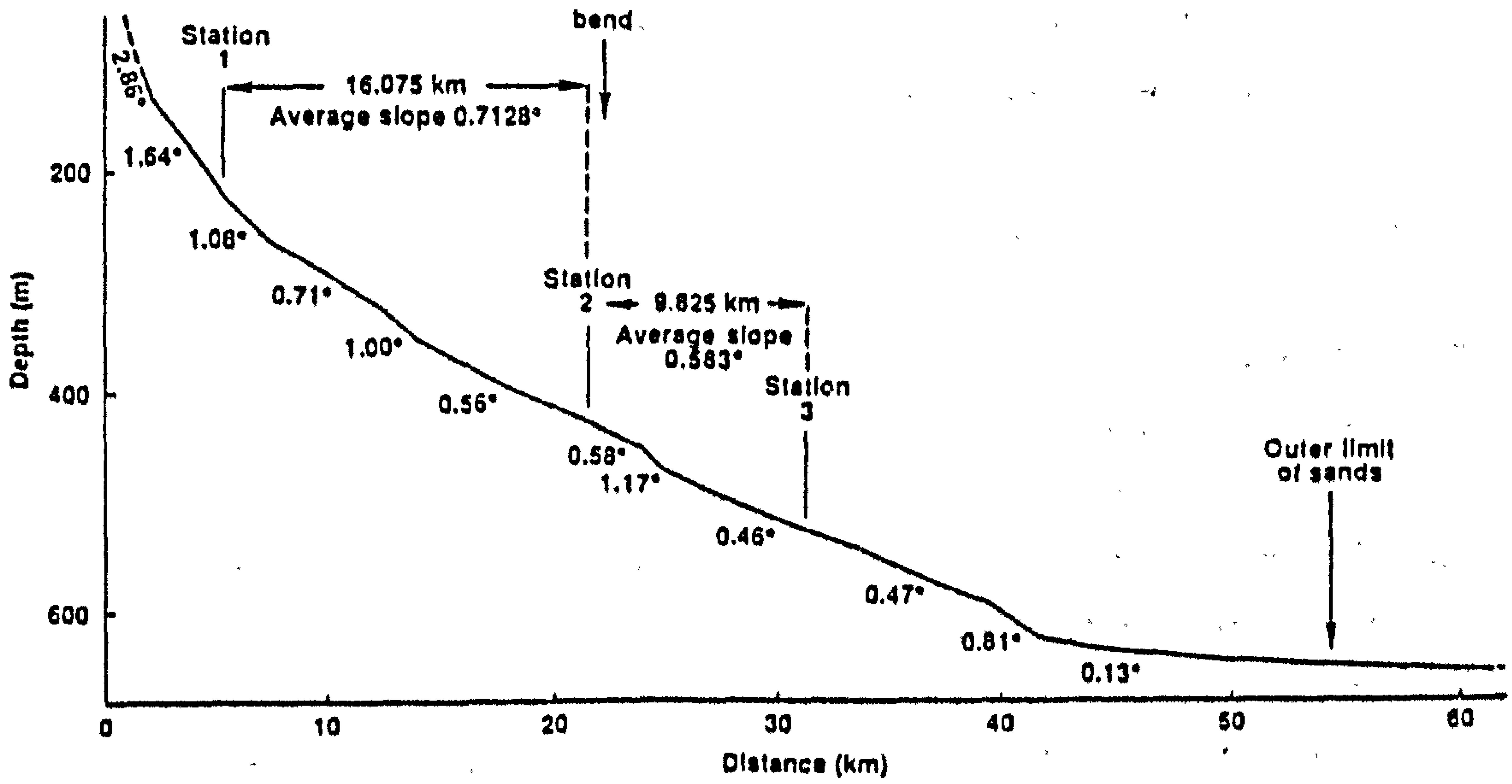


Fig. 3. Longitudinal profile along the sinuous channel axis from the delta front to the basin, showing the locations of monitoring stations.

Figure 3.25: Map and section of the Bute Inlet fan system (Prior *et al.*, 1987). This system is a good modern analogue for the channel-complex model. See text for details. Stations are the positions on the fan at which current measurements and current sediment samples were taken.

sand-rich fans. Many such fans have been described from the North Sea Basin where they form extremely productive reservoirs (e.g. the Forties, Andrew and Maureen Formations of Tertiary age and the Jurassic Magnus Formation. See Deegan and Scull, 1977). These fan systems developed in an intra-cratonic extensional setting similar to that of the Pendle Grit Formation. Facies Associations in these fans are also comparable. It would be a useful exercise to apply the channel complex model to these North Sea systems to see if it could predict sandbody geometries and reservoir characteristics better than existing models. The channel complex model predicts the following:

i) At the level of a single channel complex

a) Linear, palaeocurrent-parallel sandbodies with a width/thickness ratio of 10-30 bounded by major erosion surfaces and built by numerous fining-up sequences.
b) Within channel complexes the best poroperm characteristics should be at the base of the sequence where the sediment is coarsest, the sand/shale ratio highest and the beds thickest. However, lateral variation in poroperm characteristics may be very rapid due to the intrabed complexities within facies A beds and due to the presence of second erosion surfaces. Horizontal permeability barriers (shale breaks) should be very rare and, if present, laterally impersistent.
c) The finer sediments higher in the complex should be, on average, of less good reservoir quality although the lateral continuity of the units may be better. Horizontal permeability barriers (i.e. shale breaks between Facies C and D beds and Facies G plugs) should be more abundant, thicker and more laterally extensive.

ii) At the level of the overall depositional system.

a) Sandbody connectivity should be excellent: each channel complex generally cuts into several pre-existing complexes.
b) Overall, the total sand-packet geometry should be very dependent on the topographic relief of the receiving basin

(often fault controlled). This should lead to better connectivity in the depositional lows and to thinner, poorer quality sandbody development on the intra-basinal highs.

c) Correlation of minor reservoir units in these systems will be extremely difficult over distances greater then about a hundred metres (maximum). Recognition of first order erosion surfaces in vertical sections may provide the best means of correlation. Massive, amalgamated facies at the bottom of channel complexes and shale breaks at the top are likely to be the only reasonably continuous units.

The three-dimensional data available from the producing fields in these North Sea sand-rich fans should be sufficient to test these predictions.

3.7 Conclusions.

1) The Pendle Grit Group represents a sand-rich submarine fan (the Pendle Grit Formation) overlain by shale dominated submarine slope deposits (the Pendle Shale Formation).
2) The Pendle Grit Formation fan is very sand-rich and dominated by channel deposition.
3) A channel hierarchy exists within the fan system. First-order channels were cut by rare, large turbidity flows. Filling of these channels produced sandbodies referred to as channel complexes.
4) Channel complexes were filled by vertical accretion of sediment deposited from minor turbidity flows trapped within the first-order channels. Filling was by repeated steps of backfilling and regrading of the first-order channel during which second-order channels were cut and filled. The whole of the first-order channel may be re-excavated by subsequent major flows, causing lateral migration of the whole complex.
5) Within the fan system the individual sandstone facies (Facies A to D) can be understood in terms of the evolution of an initial high density turbidity current with distance from source and also through time. A qualitative prediction of the facies stability fields for any flow at any place in the fan system can be made using the sediment continuity equation.
6) Facies A and B represent deposition from high-concentration turbidity currents which were initially capable of maintaining a basal boundary layer.

In this boundary layer grain support mechanisms included dispersive pressure and pore-fluid expulsion. Sedimentary structures in these facies reflect the mechanics of the boundary layer rather than of the flow as a whole. Through deposition of these facies the flow became less powerful and more mud-rich. Facies C and D represent deposition from this evolved flow.

7) The coarser-grained facies within the channel complexes (Facies A) have larger scale internal features which indicate considerable variation in flow power during individual flow events. Flow events are believed to have been hours or days long.

7) A conceptual facies model (the channel complex model) has been constructed in order to explain the above features of the Pendle Grit Group sediments. It differs from other models in having channels developed across the whole length of the fan. The bulk of the sedimentation was in these channels.

8) Comparisons with other sand-rich fans described in the literature suggest that tectonic or palaeogeographic confinement of the fan system, a sand-dominated sediment supply and a variation in flow sizes in the system are necessary conditions for channel dominated fans to develop.

9) Predictions of the sandbody geometries and reservoir characteristics of sand-rich fans have been made using the model. Testing of these predictions in other sand-rich fans will indicate whether or not the model is valid.

CHAPTER FOUR

SEDIMENTOLOGY OF THE GRASSINGTON GRIT GROUP.

4.1 Introduction.

The Grassington Grit Group is the oldest Namurian stratigraphic unit which is present both on the Askrigg Block and in the Bowland Basin (Chapter 2, Figure 2.11). Although the formations in these two palaeogeographic provinces comprise distinct facies sequences, they have many facies and facies associations in common. On the Askrigg Block and across the Craven Fault transition zone (Figure 2.1), sedimentological data were first acquired from the Grassington Grit Formation during stratigraphic studies by Thompson (1957) and Wilson (1957). More recently, Baines (1977) has presented a rigorous facies analysis of the Formation. This contrasts with the Warley Wise Formation in the Bowland Basin for which there is no published sedimentological data. The lack of work on the Warley Wise Formation is probably a reflection of the very poor exposure in the main crop along the Pendle Monocline.

In this Chapter the results of a reconnaissance sedimentological study of the Warley Wise Formation are reported and the facies sequences in the Formation are compared with those in the Grassington Grit Formation. The present author has measured sections in the latter formation only in the Craven Fault transition zone and in the Eastern Bowland Basin. Only a few of the thinner sections on the Askrigg Block were studied. This look at the Grassington Grit Formation provided an overlap with the previous work by Baines (*op. cit.*) and allowed for integration of the two studies. Baines' study is comprehensive in extent and only small elements of his facies association interpretations are challenged in this work.

The figures illustrating this chapter are each referred to several times. Consequently, for ease of access, they have been gathered together at the end of the text.

4.2 Facies Analysis of the Grassington Grit Group.

A fairly simple facies subdivision has been made for the sediments in this Group because of the reconnaissance nature of the study. As the majority of the facies are less controversial than those in the Pendle Grit Formation their distinguishing features and interpretations are tabulated (Table 2) rather than discussed in the text. The facies are interpreted to fall into three groups:

1) Mass-flow deposits (Facies 1-3 and 9).
2) Fluvial deposits (Facies 4-8).
3) Flood plain/delta-top and channel abandonment deposits (Facies 10-12).

Most of the mass-flow deposits are comparable with the thick-bedded facies in the Pendle Grit Formation (Facies A and B) and resulted from similar sedimentary processes (refer to Chapter 3). Facies 9 (Giant cross-beds), however, represent true grain-flow deposits caused by avalanching down the face of a prograding Gilbert-type delta (Gilbert, 1885). True grain-flows are able to exist in such environments because of steep submarine gradients and the cohesionless nature of the coarse sediment being supplied to the delta-front (See discussion of mass-flow types on Figure 3.5). The interpretation of this facies as a series of braid-delta foresets is based on the facies associations and sequences discussed below. Foresets on a similar scale, and in sediments of similar grainsize, have been recorded from Plio-Pliestocene fan-deltas in the Corinth area of Greece (Collier, pers. comm.). This facies is difficult to distinguish from the other mass-flow facies because of the vertical scale of the foresets. It is particularly difficult to identify in small exposures, particularly along the Pendle Monocline where the structural dip is high. Several of the dip measurements recorded on geological maps of this structure are exaggerated because of a failure to recognise the sedimentary component of dip in this facies (c.f. Figure 4.6).

The fluvial and overbank facies are very similar to those reported from other Namurian sedimentary cycles such as the Roaches and Kinderscout Grits (Jones, 1980; Collinson, 1968)). The fluvial facies are sand-rich and contain many features characteristic of sand-bed braided or low-sinuosity rivers (Bristow, 1987). The interpretations of these facies are discussed further below.

4.3 Facies Associations.

4.3.1 Facies Association 1: Mass-flow Association.

This association consists of units of Facies 1 (Thick-bedded sandstones) and Facies 3 (Chaotic deposits). The association is found only at the base of the Warley Wise Formation and is best exposed at Waddington Fell Quarries (SD716 475), Turnhill (SD738 355) and at Churn Clough reservoir (SD786 384). Figures 4.1 and 4.2 show representative sedimentary logs through this association. In streams north of Waddington Fell Quarries, sediments at the base of the association are seen to contain packets of fine-grained sediments (Facies 11) similar in lithology to Facies G deposits of the Pendle Grit Formation. These finer sediments are probably developed elsewhere but are not well-exposed. As a whole, the Association appears to be laterally continuous although individual beds are lenticular on the scale of tens to hundreds of metres. Around Foulridge it is possible that some of the thick-bedded sandstones of this Association lie within channels cut into the Pendle Shale Formation: feature-forming pebbly sandstones crop-out as a series of lenticular bodies (Figure 4.7). These have been mapped as fault-bounded bodies by B.G.S.. The sharp lateral boundaries of the features might, however, be channel margins. Elsewhere, there is no evidence of channelling at the base of the Association.

The top of the Association has not been seen in sections visited by the current author. However, from composite stratigraphic sections (e.g. Figure 4.7) the association appears to be between 40 and 60m thick. Over Waddington Fell it expands to around 100m in thickness.

On the basis of the facies sequences discussed in Section 4.4 below, the thick-bedded sandstones of this Association are believed to have been deposited in a mass-flow apron at the base of a braid-delta sub-aqueous slope (See Figure 4.8). The mass-flows were probably generated by collapse of distributary channel mouth bars and by sub-aqueous slope failure. Mouth bar collapse during flood events has been reported from modern deltas (Heezen, 1955) and seismic or gravity induced collapse/slumping on delta-fronts is also a major process (Prior and Bornhold, 1988; Coleman *et al.*, 1974). Turbidity currents generated by such processes (Heezen *op. cit.*), would transport coarse sediment down to the basin where it would form a mass-flow apron around the base of the delta-slope.

TABLE 2

FACIES NAME	GRAINSIZE	UNIT THICKNESS	BOUNDARIES	SEDIMENTARY STRUCTURES	INTERPRETATION	COMMENTS
1 Thick-bedded sandstones.	Coarse and pebbly. Poor sorting.	Up to 1.5m.	Sharp/scoured, loaded & fluted.	Normal grading, rare diffuse lamination or cross-bedding.	High concentration turbidity current deposits.	Similar to Facies A and B of the PGF. See Figs 4.1, 4.2.
2 Ropy sandstones.	Coarse/pebbly moderate sorting.	Up to 4m.	Sharp or erosive base, gradational top.	Ropy bedding, destratification common. Internal scours.	Coarse grained channel mouth bars.	Similar to Facies-type B_b of PGF. See Plate 9a.
3 Chaotic deposits.	a) Silt and sand dominated.	50cm to 3m.	Sharp base and top.	Abundant mud-flakes, slurried appearance.	Debris flows.	See Figs 4.1, 4.2. Similar to Facies-type $F_{a/b}$ of PGF.
	b) Mud rich.	up to 3m.	Sharp base, irregular top.	Retains original bedding and structures.	Slumped blocks of delta-top deposits.	Similar to Facies-type F_c of PGF.
4 Lenticular sandstones.	Granule or pebble grade.	Up to 5m.	Very erosive base. Top sharp or gradational.	Ropy bedding at base. Diffuse bedding in main sandbody.	Chute channel fill Channel filled by rapid, vertical accretion.	Width/depth ratio of channels varies from 5-10. See Fig 4.3 and Plate 9.
5 Plane-laminated sandstones.	Medium to granule, occ. pebbly.	Circa 1m.	Sharp top and base.	Cm scale laminations due to variations in grainsize.	Sheet-flood deposits between fluvial and delta front channels.	See Plate 9c and Fig 4.3.
6 Tabular cross-bedded sandstones.	Medium to Pebbly. Well-sorted.	Cosets up to 1.5m.	Sharp top and base.	Thick (5cm) foresets. Foresets asymptotic and normally graded. Pebble lags at toe and on foresets.	Straight-crested fluvial bars: scroll bars of Bristow (1987).	See Figs 4.4 and 4.9.

TABLE 2 (CONTINUED)

7	Large-scale trough cross-bedded sandstones.	Coarse to pebbly. Moderately well-sorted.	Cosets 1-2m. Beds up to 8m thick.	Planar erosion surfaces at top and base of cosets.	Asymptotic or angular foresets up to 2cm thick. Foresets coarsen down-dip. Down-current dipping reactivation surfaces common.	Fluvial braid-bars deposited during high flow-stage. Reactivation surfaces represent falling stage erosion of bar?.	Individual cosets can be traced up to 80m along palaeo-current. See Figs 4.5 and 4.9.
8	Small-scale trough cross-bedded sandstones.	Medium to granule, occasionally pebbly.	Cosets 10 to 50cm. Beds up to 10m.	Gradational base with Facies 4, otherwise erosional base and sharp top.	Asymptotic or angular foresets, mm thick. Troughs irregular in section.	a) deposits of in-channel dunes. Size of dune scales with channel depth. b) Parasitic dunes on larger bar-forms.	Individual cosets can be traced up to 10m along palaeo-current. Cosets thin upwards through beds. See Figs 4.5, 4.9 and Plate 9c.
9	Giant cross-bedded sandstones.	Pebbly.	Foresets up to 50cm.	Eroded tops, base not seen.	Massive or graded foresets which thin or thicken down sedimentary dip.	Foresets of Gilbert-type braid-delta. Formed by grainflow avalanching down delta-front.	Intrasets recorded in this facies by Baines (1977). See Figs 4.6, 4.7 and 4.8.
10	Coarsening-up sandstones.	Fine to medium.	2 to 3m.	Gradational or sharp base. Top sharp.	Planar lamination, ripples, climbing ripples. Bioturbated top. Sometimes rootlets present.	a) Crevasse splay deposits. b) Mouth bars of delta-top bay-fill deltas.	Description partly based on Dunham and Wilson (1985).
11	Laminated mudstones and siltstones.	Clay and silt with laminae of very fine sandstone.	Sequences up to 20m thick, usually thinner.	Usually sharp based. Top sharp but gradational with Facies 10.	Laminated and/or rippled. Isolated ripples of sand common. Often bio-turbated. Rare rootlets.	Delta-top or flood plain lake deposits. May also represent fine-grained channel plugs in some cases.	Facies rarely well exposed. Details based on Baines (1977) and Dunham and Wilson (1985).
12	Coals and carbonaceous smuts.		2-10cm.	Sharp top, gradational base.	Rootlets, thin silt and sand laminae.	Coal formation in abandoned channels, on stabilised fluvial bars and in flood plain lakes.	Coals have been worked on Embsay, Boss and Skipton Moors (Embsay/Bradley coal).

PGF= Pendle Grit Formation.

4.3.2 Facies Association 2; Deep water delta-front association.

Giant cross-bedded sandstones (Facies 9) dominate this Association (Figures 4.6, 4.7). Set sizes appear to be very large although the bases of the sets are never seen. Facies sequences suggest that set sizes might reach up to 100m in thickness (refer to Figure 4.7) although the maximum set height exposed is 15m (seen at quarries on Noyna Hill, SD902 426). Above the giant cross-sets the facies association may contain trough cross-bedded sandstones (Facies 7 or 8), massive lenticular sandstones (Facies 4) or finer deposits containing muds, silts and sands of Facies 10 and 11. (Figure 4.6)

The Association is believed to represent the progradation of a sand-dominated braid-delta into deep water (Figure 4.8). Braid-delta is used in the sense of McPherson *et al.* (1988): that is, a sand-rich delta formed by progradation of a braided river system into a body of standing water. The giant foresets (Facies 9) represent the progradation of the delta front by avalanching and their height, therefore, gives an indication of the water depth of the receiving basin. It can be estimated from Figure 4.7 that maximum water depths were around 100m. The lenticular and cross-bedded sandstones within the association represent filling of short-lived distributary channels on the braid-delta. The finer facies may represent either mud plugging of channels after abandonment or deposition in inter-distributary areas.

4.3.3 Facies Association 3: Shallow water Braid-delta Association.

This Association is very similar to, and gradational with, the deep water braid-delta association (Facies Association 2). It consists of sequences dominated by ropy sandstones (Facies 2) passing up into lenticular sandstones (Facies 4) interbedded with parallel-laminated sandstones (Facies 5). The relationships of the facies in this association are shown on Figures 4.3 and on Plate 9. Exposures of this Association are widespread over the Bowland High and across the Craven Fault transition zone (e.g. Giggleswick, SD809 641; Flasby Fell and Sharp Haw, Grid squares SD95 55 and SD96 55).

The association is interpreted as representing progradation of a fluvial braid-plain into shallow water (Figure 4.9). Unlike in Facies Association 2, water depths were insufficient for the development of large foresets and a base of slope mass-flow apron. Instead, the submarine parts of the braid-delta are represented by ropy sandstones (Facies 2) containing channels filled by massive sandstones (Facies 4). These channels may be very similar to the chute channels reported from the sub-aqueous delta-slopes of fjord fan-deltas (Prior and Bornhold, 1988). A vital exposure at The Strid (Grid square SE06 56) shows how the delta front deposits relate to distributary channel deposits. The river cliff exposure runs parallel to the palaeocurrent for approximately 500m and, at the down-palaeocurrent (south) end of the section, is dominated by ropy-bedded pebbly sandstones (Facies 2). These can be traced up-palaeocurrent at the same horizon into sandstones with crudely developed trough cross-bedding. Further up-current the trough cross-sets become better defined forming intervals of true Facies 7 deposits. The crude troughs in the transition zone between the two facies are some 50cm to 1.5m in width and up to 20cm deep. Individual foresets are around 4cm thick and the tops of the sets are sub-horizontal because of erosion in front of subsequent dunes. The troughs in the Facies 7 deposits are less deeply eroded with thicker sets for any given trough width. Individual foresets are also somewhat thinner.

The facies transition described above shows that Facies 2 and Facies 7 deposits represent down-current changes in depositional processes. The up-current troughs of Facies 7 obviously represent lower flow regime dunes, probably migrating down a distributary channel on the braid-delta. The partially washed out troughs suggest that bed shear stress was increasing down-current. This is thought to have been caused by shallowing of the channel towards its mouth: deposition in the channel mouth resulted in shallowing and the consequent increase in flow-velocity raised the bed shear stress. The ropy-bedding might have resulted from continued shallowing causing the flow to reach the boundary between the upper and lower flow regimes. At this stage turbulent burst and sweep events (Leeder, 1977) may have remobilised sediment down the seawards face of the coarse grained mouth bar by a process similar to that described for Facies-type B_b ropy-bedded deposits in the Pendle Grit Formation. Alternatively, flow expansion seawards of the mouth bar might have led to deposition rates so high that bed-form lag effects prevented the formation of true dunes by the expanding fluvial flow. Because of the relationships in this exposure, the ropy bedded facies (Facies 2) is interpreted as representing deposition at channel mouths: either as low-stage distributary mouth bars or as high-stage chute channel mouth bars (See below and Figure 4.9).

High flow-stage activity at the braid-delta front was dominated by deposition of massive lenticular sandstones (Facies 4) in short-lived submarine chute channels (Prior and Bornhold, 1988). High flow velocities during flood events enabled the distributary channels to cut across their mouth bars and to produce sediment laden underflows down the braid-delta front. These underflows cut the chute channels while flow expansion at the chute channel mouths caused deposition of Facies 2 deposits (Figure 4.9). The whole braid-delta system prograded rapidly because of the vast sediment influx during the flood event. Progradation took place via a process of infill of the chute channels and forwards migration of the fluvial distributaries. Between the chute channels, and in the interdistributary areas of the delta top, sheet flow caused by overbank flooding gave rise to the plane-laminated sandstones of Facies 5 (Figure 4.9). The lateral continuity and thick internal laminations of Facies 5 beds is reminiscent of beach deposits. However, there is no other indication of wave activity in this facies associations and, hence, the former interpretation is favoured.

4.3.4 Facies Association 4: Fluvial Association.

The fluvial association consists of channelised cross-bedded sandstones (Facies 6, 7 & 8). The sandbodies built by these facies reach 60m in thickness and 3km in width. Baines (1977) recorded channel widths of up to 700m and suggested that the channels became smaller and thinner upwards through the Grassington Grit Group. Stacking of channels produces local multistorey sandbodies. For example, local thickening of the Grassington Grit Formation over the Bowland High (Figure 2.11) is due to stacking of three separate channel units. Trough cross-bedded facies (Facies 7 and 8) form the bulk of the channel sandbodies. These facies are regularly interbedded with the small trough cross-beds (Facies 8) sometimes forming down-current dipping cosets accreted to the downstream end of a set of larger scale trough cross-bedding (Facies 7). This architecture (Figure 4.5) is interpreted to represent dune migration over medial bars in a braided fluvial system (c.f. review of fluvial bedforms in Bristow, 1987). The small trough cross-sets also form thick packets independent of the larger sets (Figure 4.5). These packets result from in-channel deposition by migrating dunes. The size of the individual sets often decreases upwards through these packets. This is compatible with a reduction in dune size as the channel shallowed and was partially abandoned with time.

Rarely, the fluvial association contains fine grained sandstones of Facies 10 and mudstones of Facies 11 (Figure 4.4). Both lithologies may contain rootlet beds (e.g. Far Costy Clough SD687 592. See also Figure 2.10 columns 10, 11, 14 and 16). Coals and carbonaceous smutts (Facies 12) are also present (See Figure 2.10, columns 5, 6, 13). These facies represent the filling of abandoned channels by overbank floods and the eventual colonisation of the abandoned channels by plants.

4.3.5 Facies Association 5: Floodplain/Delta platform Association.

This association is very poorly exposed in the areas covered by the present author. Data from Baines (1977) and Dunham & Wilson (1985) suggests it is better developed over the Askrigg Block and in the Craven Fault transition zone. The principal constituents of the association are siltstones and mudstones with abundant bioturbation (Facies 11). Coals are present in some areas, as are thin channelised units of trough cross-bedded sandstones (Facies 7). Thin coarsening upwards sandstones (Facies 10) are also present. Baines (*op. cit.*) suggested that a northwards increase in the abundance of this association relative to the Fluvial Association was a result of a change from floodplain to delta platform deposition across the Askrigg Block.

4.4 Facies Sequences.

Facies sequences show the stratigraphic sequence of facies associations in any particular part of a depositional area: a borehole section will, therefore, record the complete facies sequence at the location of the borehole. Such sequences show how depositional environments change with time at any location. The regional distribution of facies sequences can be used to define stratigraphic units which are sedimentologically sensible and internally consistent. The recognition of two areally distinct facies sequences in the Grassington Grit Group led to its subdivision into the two Formations formally defined in Chapter 2. The facies sequences on which this subdivision was based are outlined below. Obviously, the areal extent of the facies sequences coincides with the Formation boundaries as mapped on Figure 2.10 and as shown on Figure 2.6.

4.4.1 Facies Sequence 1: Warley Wise Formation.

Throughout the Bowland Basin there appears to be a consistent facies sequence consisting of Facies Association 1 (Mass-flow Association) overlain by Facies Association 2 (Deep water braid-delta Association) and capped by Facies Association 4 (Fluvial Association). Facies Association 5 (Delta-platform Association) is also developed locally (Figure 4.7). The full facies sequence is not exposed at any point: the stratigraphic positions of various outcrops have been used in its construction (Figure 4.7). Generally, the complete Facies Sequence is over 150m thick and the mass-flow and giant cross-bedded facies at the base of the sequence account for over half of this total thickness. This sequence is interpreted to represent the progradation of the various sub-environments of the braid-delta system (as defined by the facies associations) into deep water in the the Bowland Basin (Figure 4.8). The implication of such a model is that the Warley Wise Formation is strongly diachronous across its outcrop.

In the current author's experience the facies sequence, although constructed from limited outcrop data, can be used to predict accurately the facies association at any given level in the Formation. The predictive model has only failed with respect to one section on the River Ribble (SD675 358). At this locality the fluvial association (Facies Association 4) lies directly over the Pendle Shale Formation. The difference in this section may be due its position close to the depositional limit of the Warley Wise Formation (Figure 2.11) or, more likely, to uplift of the Longridge Fell area during the tectonic phase which produced the intra-E_{1c} unconformity on the Askrigg Block.

4.4.2 Facies Sequence 2: Grassington Grit Formation.

In contrast to Facies Sequence 1, this sequence is dominated by shallow water braid-delta and fluvial deposits (Facies Associations 3, 4 and 5). In general it is much thinner than Facies Sequence 1 (Figure 2.11). Usually, the sequence begins with shallow water braid-delta deposits (Facies Association 3) and passes up into fluvial deposits (Facies Association 4 and 5). From south to north across its depositional area Facies Association 3 (Shallow water braid-delta association) becomes progressively less well represented before disappearing completely in the north-west corner of the Askrigg Block. However, southwards towards the Bowland Basin there is a rapid expansion of this part of the Facies Sequence which can be traced laterally into the giant cross-beds and mass-flow deposits at the base of Facies Sequence 1. The boundary between the two Facies Sequences is, therefore, gradational (Figure 4.8).

Facies Sequence 2 was produced by the progradation of a braided fluvial system into a shallow body of standing water. The resulting braid-delta at the mouth of the fluvial system (Figure 4.9) quickly filled the receiving basin and allowed the fluvial system to extend progressively further out into the basin. This progradational model for the system implies that the Grassington Grit Formation is strongly diachronous across its outcrop.

One feature of this Facies Sequence is the presence of localised thickening of the sand-rich parts of the sequence due to growth or syn-sedimentary faulting. Chisholm (in Mundy and Arthurton, 1980) suggested that such faults might explain a thickening of the Grassington Grit Formation across faults on Sharp Haw, Flasby (Grid square SD96 55). Major soft sedimentary slumps and an apparent increase in the abundance of channel sands (Facies 4) in the hanging wall of these faults supports this interpretation. Unfortunately, the case cannot be proven because present day erosion of the footwall block prevents detailed correlation across the faults. However, the presence of true growth faults in the Almscliff Member of the Grassington Grit Formation can be demonstrated without doubt. Chisholm (1981) first recognised these structures. He suggested that dip and thickness changes in fluvial sediments at Almscliff Crag (SE268 490) could only be explained by inferring the presence of a listric normal fault with a pronounced rollover anticline in the hanging wall block. He interpreted this fault as a growth fault because it appeared to affect only the base of the Member (Figure 4.10). Further investigation of the outcrop has revealed the presence of a series of minor synthetic and antithetic faults to the main fault (Plate 10). These faults were formed prior to any cementation in the rock: there is no grain breaking or diminution along the fault planes. Displacement occurred by independent particulate flow along either discrete fault-planes (Plate 10b) or over zones upto 5cm thick (kink bands). The minor synthetic and antithetic faults probably developed sequentially away from the major fault and as new faults formed so the earlier faults became inactive and were rotated passively by continuing movement on the major fault (Plate 10c). This rotation proves the listric geometry of the main fault and suggests that the basal detachment on the fault is very close to the base of the outcrop (Figure 4.10). The minor faults are not growth faults, they simply accommodate the extension in the hanging wall caused by the rollover on to the main listric fault. However, their relationship with the sediments at the top of the crag proves that the main fault was moving during sedimentation: the youngest faults are cut off by erosion at the base of a set of trough cross-strata (Plate 10c).

Similar minor structures have been recognised in the Almscliff Member in other areas where there is a local thickening of the unit (e.g. Birk Crag, Grid square SE27 54). By analogy with the Almscliff Crag exposure, these localised thickenings are also attributed to growth faulting. The occurrence of growth faulting seems to be limited to areas in which coarse grained facies immediately overlie thick slope deposits of the Pendle Shale Formation. Presumably, the slope sequence was undercompacted and failed when loaded by rapid progradation of the braid-delta and fluvial systems.

4.5 Discussion and Conclusions

The differences between the two facies sequences (Formations) reflect the control of receiving basin water depth on the braid-delta system. The initial sediment influx was into very shallow water on the Askrigg Block and the resulting braid-delta prograded very rapidly. Water depths were insufficient for development of giant foresets or a mass-flow apron. In fact, progradation was so rapid that the braid-delta probably reduced its seawards gradient sufficiently to cause enhanced braiding at the seawards margin of the delta and choking of the delta front by sediment (as on the Yellow River delta today). When deeper water in the Bowland Basin was reached, the progradation rate slowed as more sediment was required per metre of forwards progradation. The morphology and sedimentary processes on the braid-delta evolved so as to equilibriate with these new conditions. Sediment transport on the sub-aqueous delta front became less influenced by traction current processes and more dependent on gravity driven mass-movement. Because sedimentation rates were high and there was little reworking of the delta front by marine processes, excess sediment built up at the distributary mouths. Avalanching of this sediment down the delta front, together with flood or slump generated turbidity currents, produced the giant foresets and mass-flow apron which characterise the deep water Facies Sequence.

The Grassington and Warley Wise Formations, therefore, provide an excellent case study of how sedimentary processes on a braid-delta system evolved as it prograded across a major downwards step in the floor of the receiving basin. The distribution of the Formations (Facies Sequences) shown on Figure 2.11 indicates that this step in the basin floor lay across the Craven Fault transition zone. Obviously, there was still a strong block/basin control on sedimentation at this time. The details of this control and the way the Grassington Grit Group fits into the "dynamic stratigraphy" of the E_{1c} basin-fill sequence are discussed further in Chapter 6.

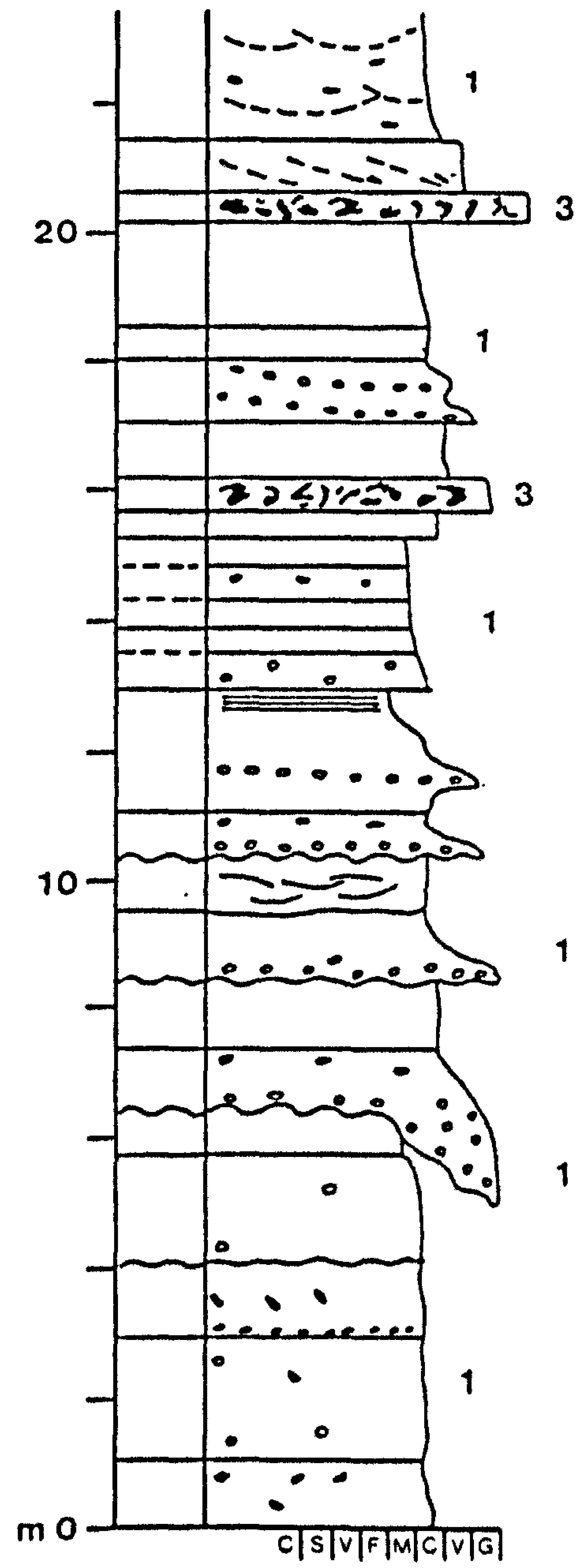


Figure 4.1: Representative sedimentary log through Facies Association 1 (mass-flow association). The association consists of thick-bedded sandstones (Facies 1) interbedded with debris-flow deposits (Facies 3). Many of the Facies 1 beds lie on erosion surfaces which define shallow scours some 2-3m deep (c.f. bed at 7m level). The debris-flow deposits are comparable with Facies-types F_a and F_b of the Pendle Grit Formation. For Key see Enclosure 1. Churn Clough Reservoir, SD786 384.

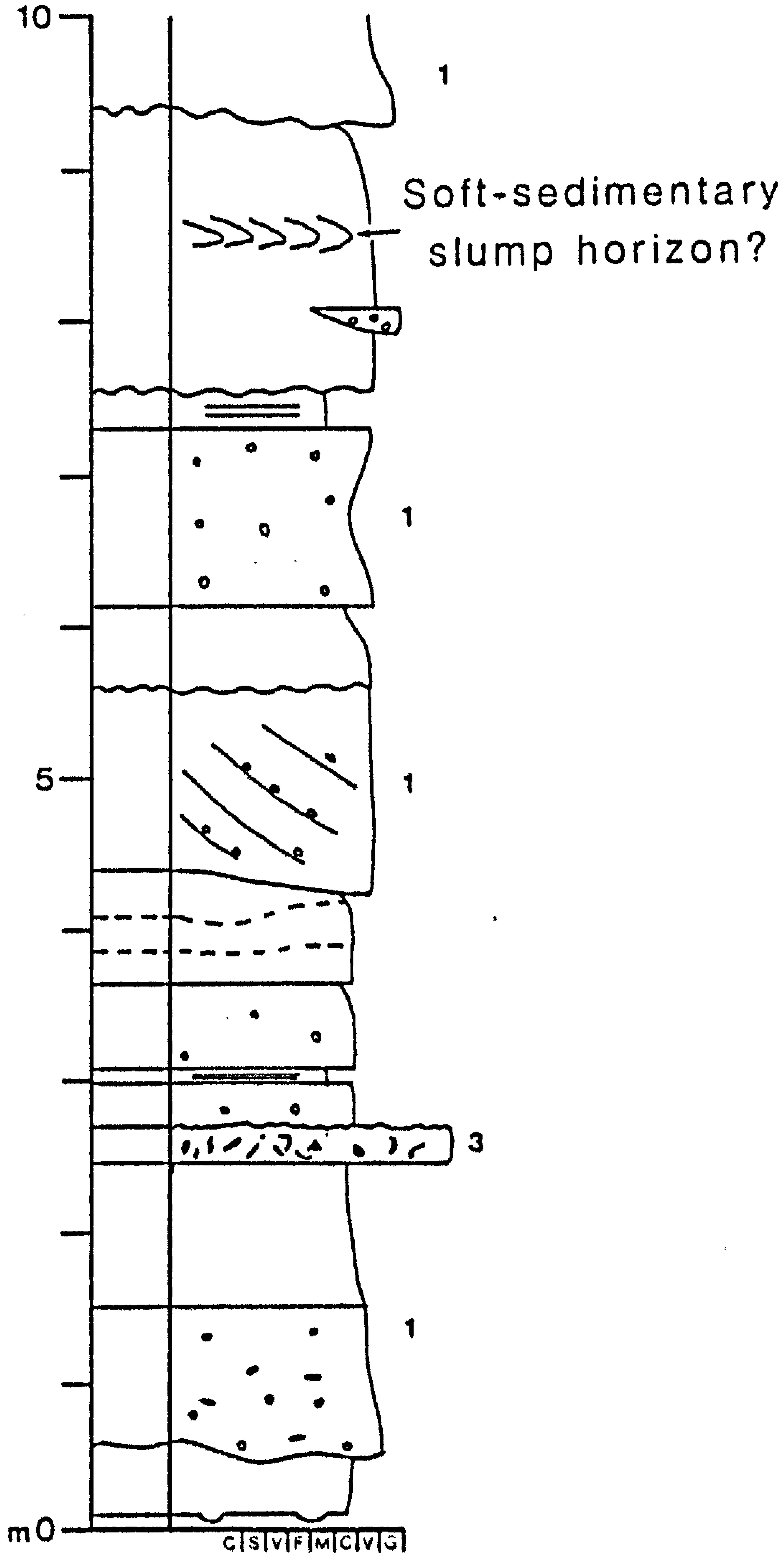
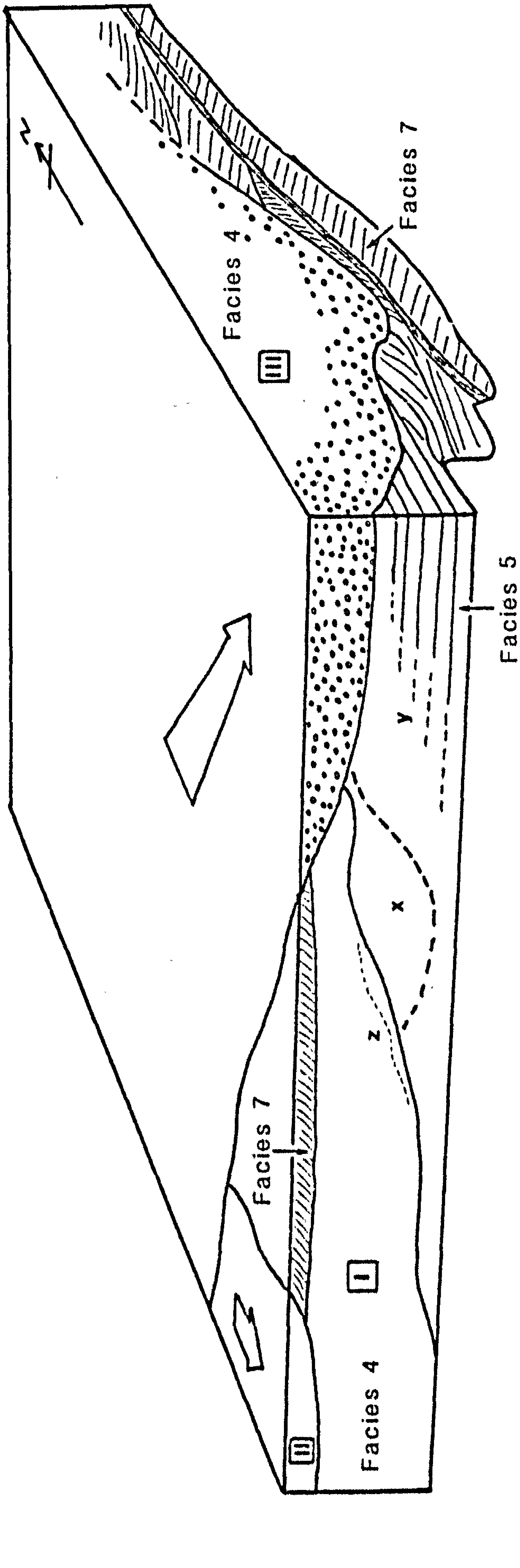


Figure 4.2: Sedimentary log through Facies Association 1 (mass-flow association) from the base of the Warley Wise Formation. For Key see Enclosure 1. Turn Hill, SD738 355.

Figure 4.3: Block diagram showing the relationships of facies within Facies Association 3 (Shallow water braid-delta Association). Three Facies 4 lenticular sand-bodies (I,II and III) are shown. The large arrows show the palaeocurrents of two of these bodies. Note how the bedding and lamination in the underlying Facies 5 deposits (y) dies out westwards into a zone of complete destratification (x). This zone was probably created by dewatering during the deposition of sandbody I. Around z the sediment immediately above the channel base has been sheared in a plastic fashion. This too may have occurred during deposition of sandbody I: bed shear stress causing further motion of sediment partially liquified by the porewater expelled from zone x. Note the asymmetry of the fill of sandbody III: pebbles are restricted to the true right bank of the channel where they form a vertically accreted plug. This suggests that the channel had a radius of curvature large enough to develop a helical flow cell. This swept the coarser sediment into the right bank. The steep right bank and the shallow left bank support such an argument.

Based on quarry exposures at Giggleswick, SD809 641. South wall is 30m long, east wall 12m long (drawn on larger scale to show details). Vertical scale is equal to the horizontal scale on each wall.


N
Facies 4
III
Facies 7
Facies 5
y
x
z
Facies 7
I
Facies 4
III

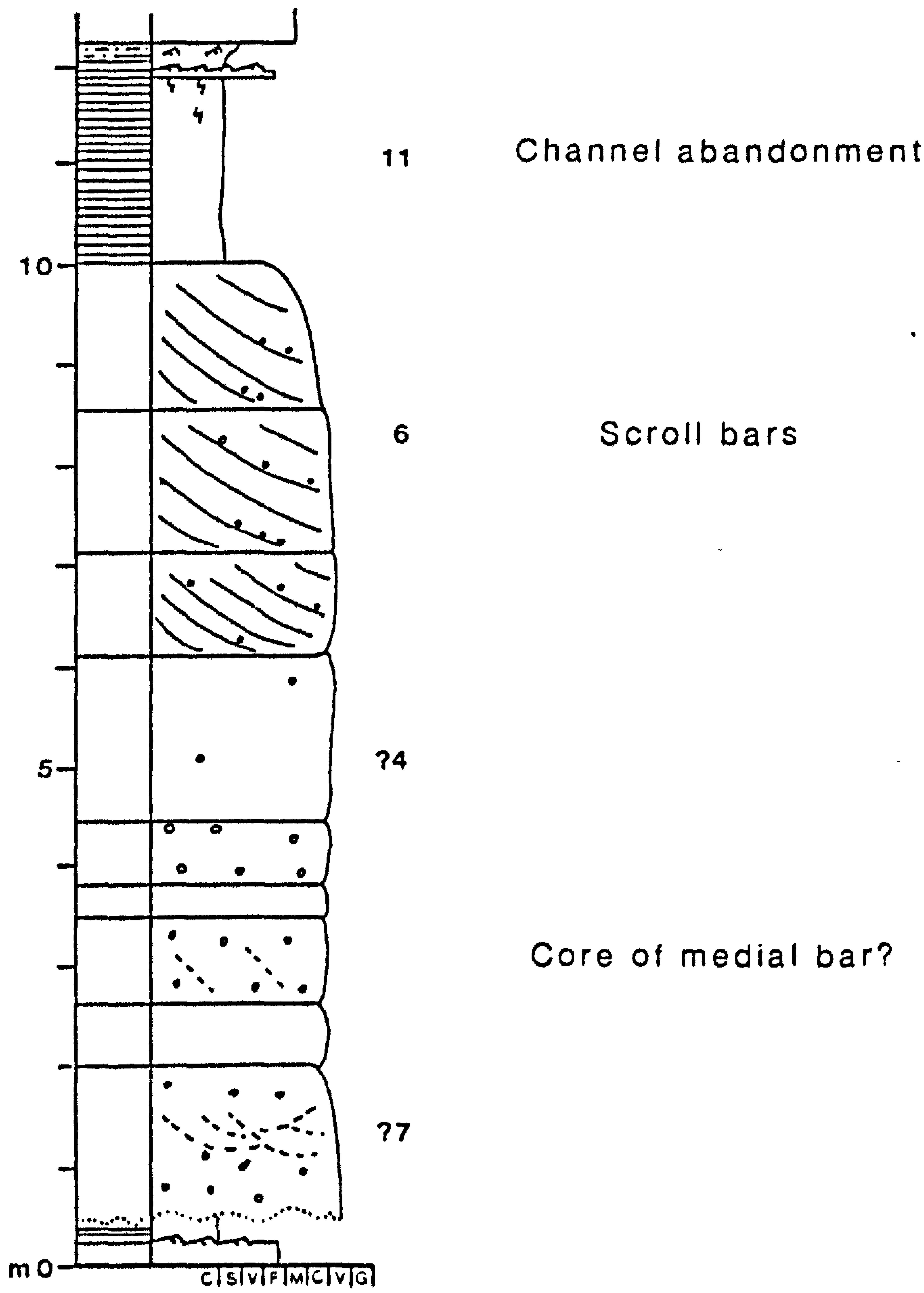


Figure 4.4: Sedimentary log of section at the top of the Warley Wise Formation. This section is representative of Facies Association 4 (Fluvial association) in this Formation. Individual facies are numbered. The presence of sediments apparently developed in Facies 1 is enigmatic: these are usually found only at the base of the Formation. Here they are crudely cross-bedded and may have a sedimentary dip of about 10°. Consequently they are thought to represent the core of a medial bar deposited during flood-stage of a fluvial channel. The ripples in the finer facies are symmetrical wave ripples. They are thought to represent reworking by wind driven waves of fine sediment in an abandoned channel.
For Key see Enclosure 1. White Haugh quarry, SD832 402.

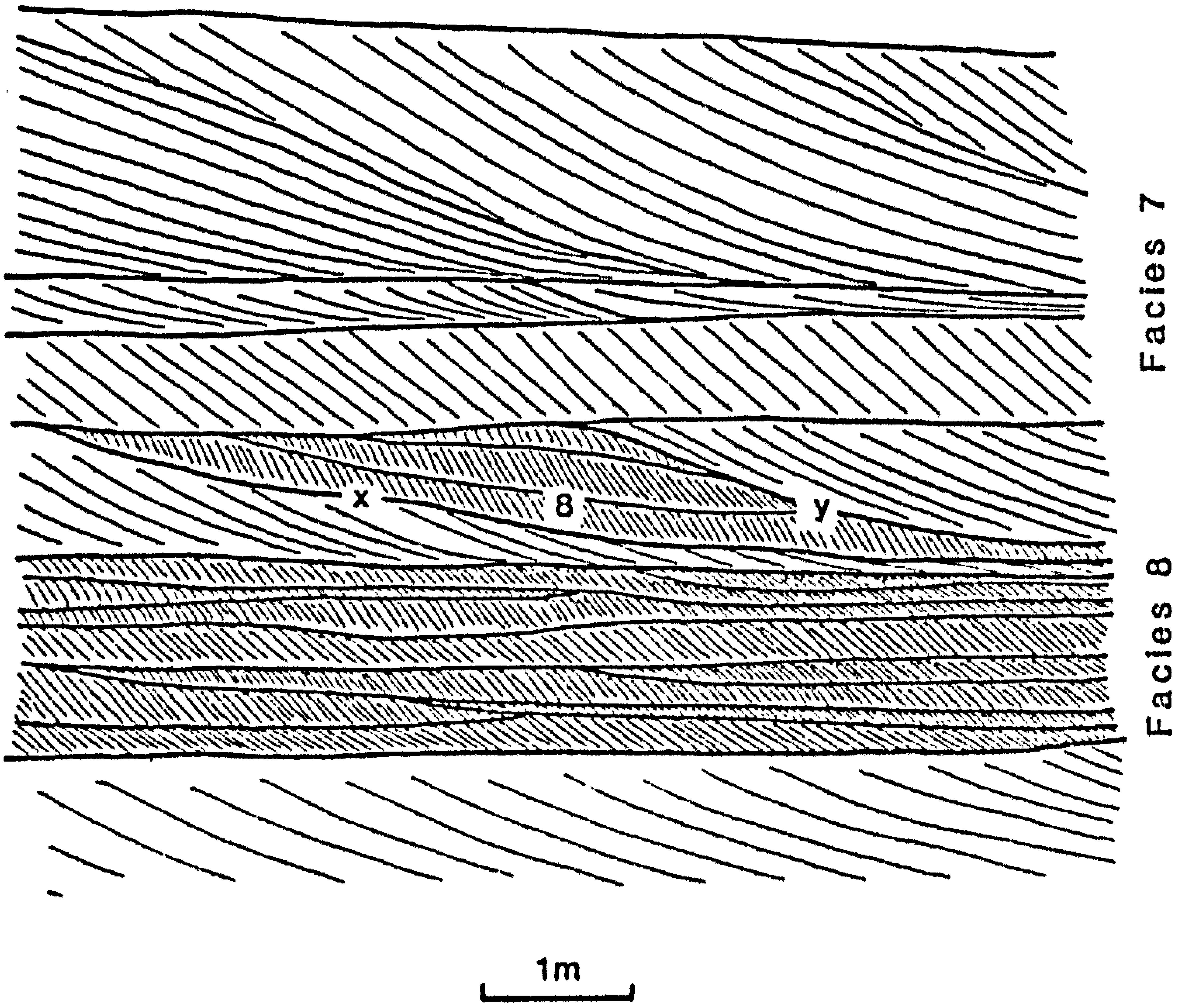


Figure 4.5: Line drawing of photograph showing the architecture of Facies 7 and 8 trough cross-bedded sandstones in the Fluvial Association (Facies Association 2). Facies 7 represents the deposits of major bar forms migrating during high flow stage in the braided river system. Note the reactivation surfaces in the upper set. Reactivation surface (x) is overlain by down-current dipping sets of Facies 8 sandstones. These represent deposition by dunes migrating over the bar, possibly during falling and low flow stages. The main bar began high flow stage. The 2m thick unit of Facies 8 sandstones represents sedimentation by dunes in the channels between and through the larger bars. Crookwise Crag SD989 555.

Figure 4.6: 3-D relationships of facies within Faughs Quarry, SD818 392. This quarry is representative of Facies Association 2 (Deep water braid-delta association). The structural dip of this quarry is defined by the base of the unit of Facies 7 sandstones. Two sets of Facies 9 sandstones are present in the quarry, the dip of one of these (set 1) was taken as the structural dip in this exposure by B.G.S..

Relationships on the back wall of the quarry have been projected into the vertical. Vertical height of the east wall is 8.5m.

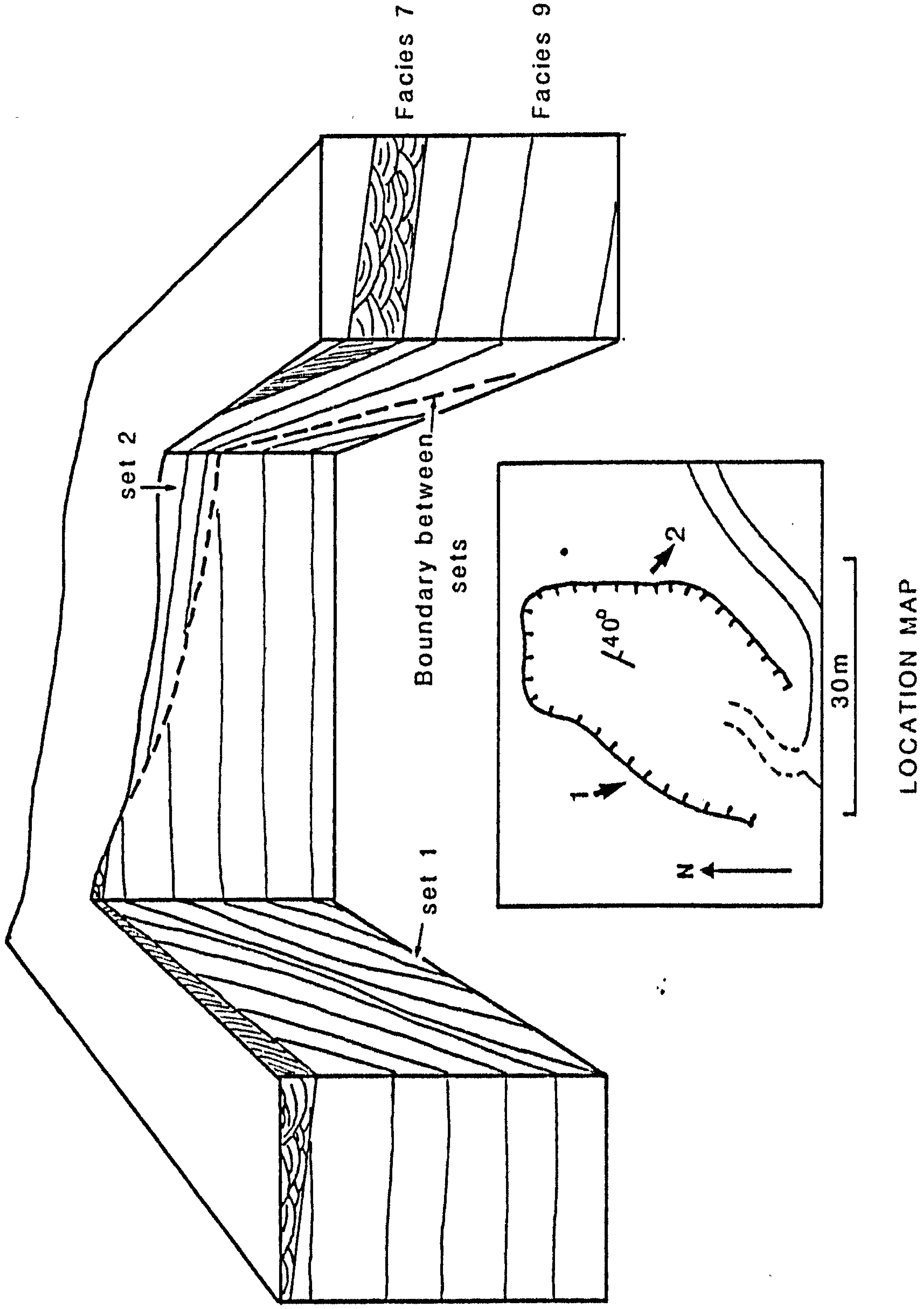

Facies 7
Facies 9
set 2
Boundary between sets
set 1
40^o
1
2
N
30m
LOCATION MAP

Figure 4.7: True scale stratigraphic section along the Pendle Monocline showing the relationships of Facies Associations within Facies Sequence 1 (Warley Wise Formation). The lines at a, b, c and d represent the extent of the stratigraphic sequences in this area on which the Facies Sequence interpretation is based. Locations b, c and d have been illustrated in Figures 4.4, 4.6 and 4.1 respectively. Note the possible channelling at the base of the mass-flow association around Noyna. This, and the presence in the same area of discrete fluvial sandbodies, is based on a reinterpretation of the topographic features in this area (see text).

WWF Warley Wise Formation.

PGF Pendle Grit Formation.


SW
NE
Pendle Hill
Barley
Foulridge
Noyna
200
100
0 m
a
b
c
d
5km
Location Map
5km
PGF
WWF
a
b
c
d
38
39
44
Key to Facies Associations
Overbank/delta-top (FA 5)
Fluvial (FA 4)
Deep water braid-delta (FA 2)
Mass-flow (FA 1)
Unnamed equivalent of Pendle Shale Fm.

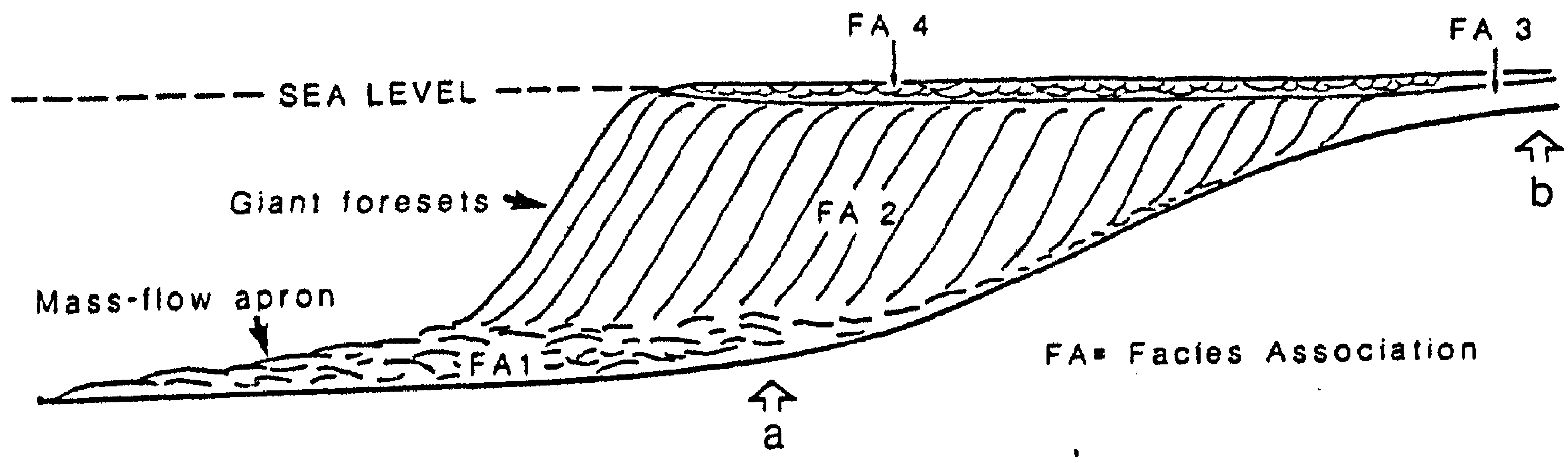


Figure 4.8: Interpretation of relationships between Facies Associations and Facies Sequences in the Grassington Grit Group. The diagram shows a braid delta progrding from shallow to deep water and changing its style to equilibriate with the new conditions. Progradation of the braid-delta in the deeper water (e.g across point a) produces Facies Sequence 1, the Warley Wise Formation. Previous progradation through the shallower areas (e.g across point b) produced Facies sequence 2, the Grassington Grit Group.

Figure 4.9 (opposite): Interpretation of the facies relationships within Facies Sequence 2 (Grassington Grit Formation). At low-water stage deposition is mostly up system in the fluvial channels (out of picture). Some coarse sediment moves through the low stage channels and forms fluvial mouth bars. As flow rates increase during a flood event the mouth bars are eroded and chute channels cut down the mouth of the braid-delta. The flood remobilises the coarse sediment trapped in the fluvial system and this is transported to the delta front where it is deposited in the chute channels and as chute channel mouth bars. Meanwhile the major bar forms are reactivated in the fluvial channels and there is overbank sheet flooding between major distributaries. The bulk of the progradation of the braid-delta occurs during flood events.
The locations in which the various facies are deposited are shown by numbers (refer to Table 2 for Key to facies numbers).

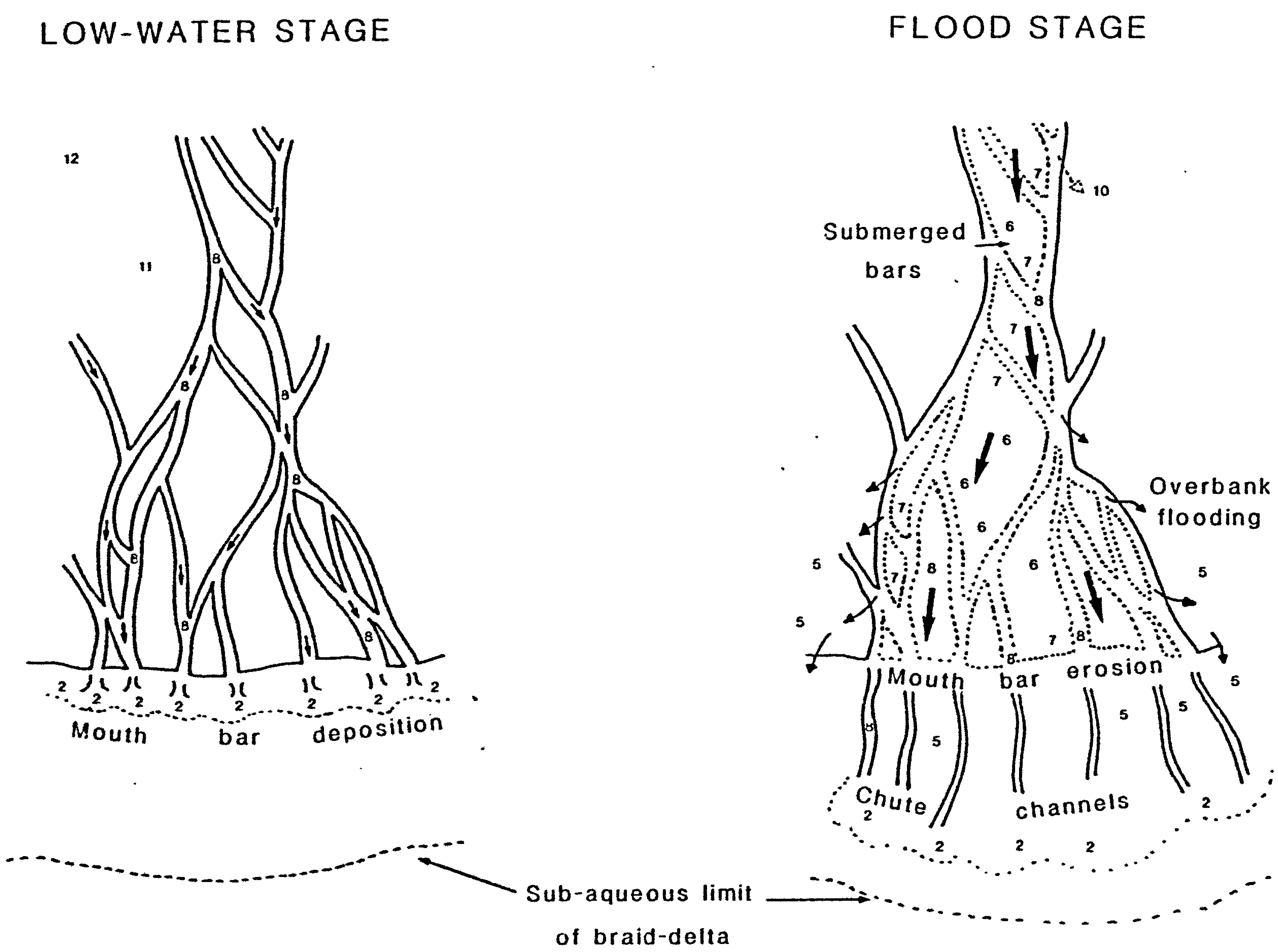
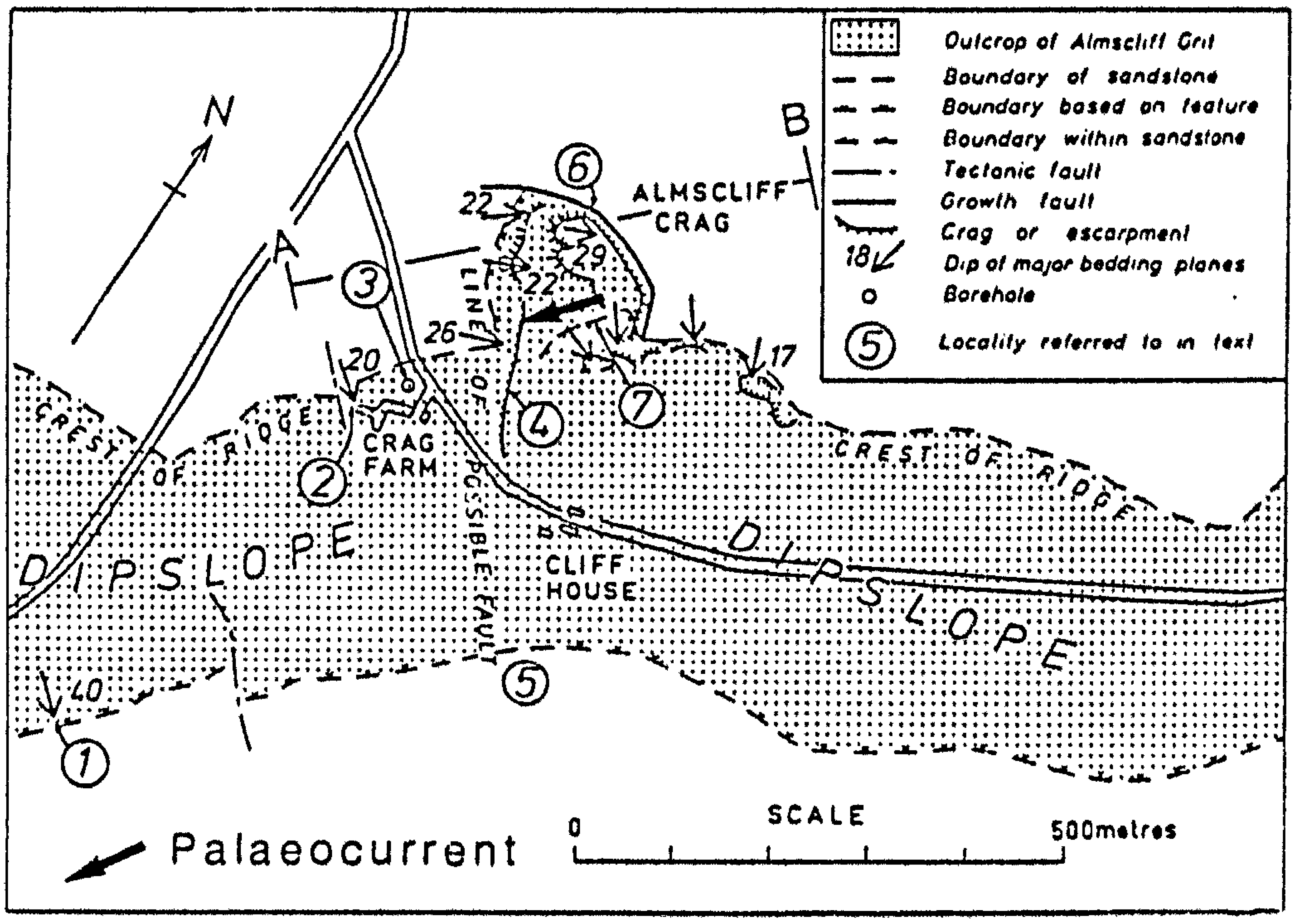

LOW-WATER STAGE
FLOOD STAGE
12
11
Submerged
bars
10
Overbank
flooding
Mouth bar deposition
Mouth bar erosion
Chute channels
Sub-aqueous limit
of braid-delta

Figure 4.10: Geological map and section of Almscliff Crag. The fault trajectory shown on the section has been drawn by eye so that it is compatible with the variation of dips in the outcrop and the rotation of minor faults discussed in the text. Section is true scale, the boundaries of the Almscliff Member are projected updip. Map taken from Chisholm (1981).


Outcrop of Almscliff Grit
Boundary of sandstone
Boundary based on feature
Boundary within sandstone
Tectonic fault
Growth fault
Crag or escarpment
Dip of major bedding planes
Borehole
Locality referred to in text
ALMSCLIFF CRAG
CRAG FARM
CLIFF HOUSE
LINE OF POSSIBLE FAULT
CREST OF RIDGE
DIPSLOPE
Palaeocurrent
SCALE
500metres

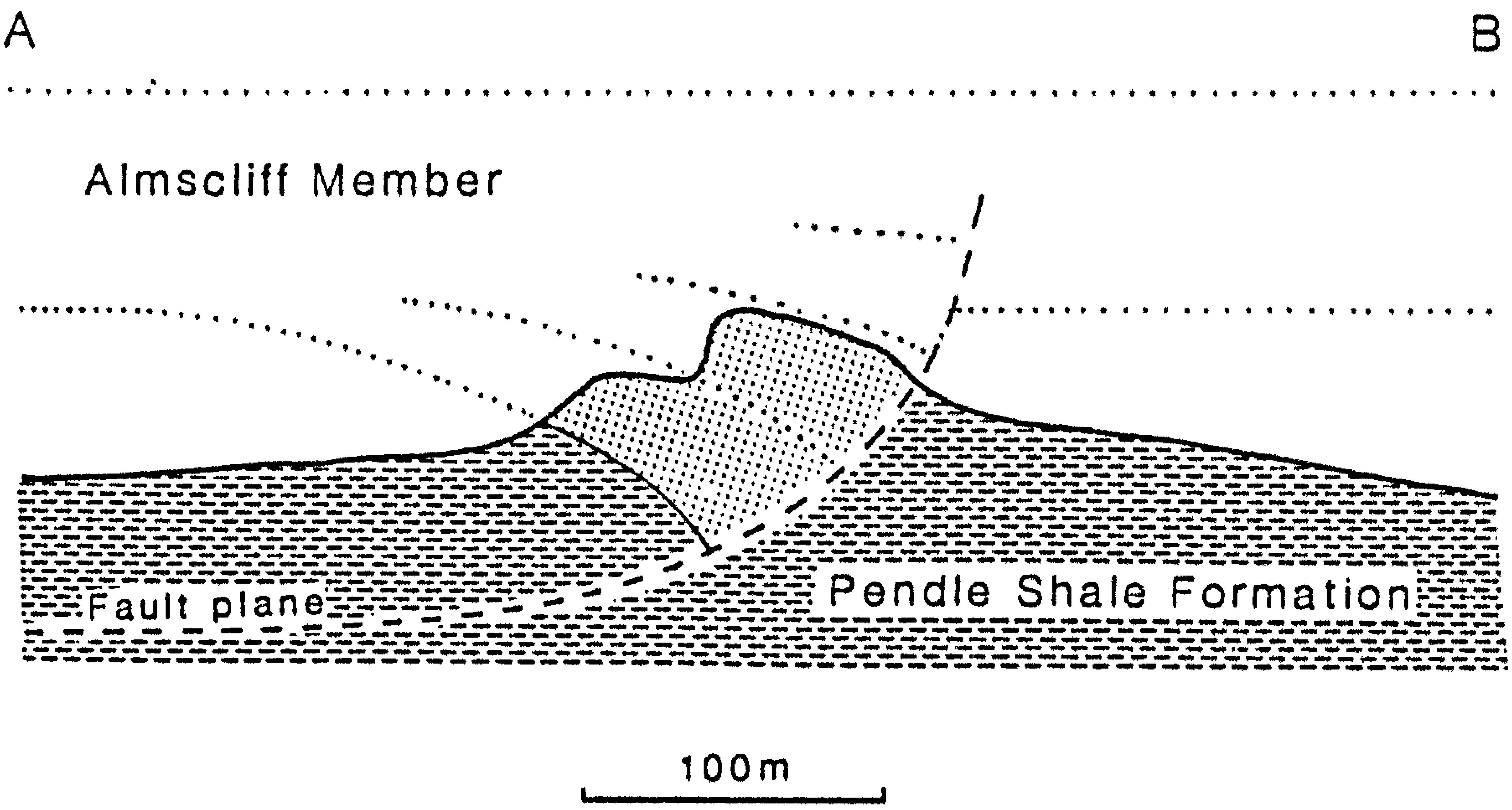

A
B
Almscliff Member
Fault plane
Pendle Shale Formation
100m

Plate 9: Facies 2 and Facies 4 deposits of the Grassington Grit Group.

a) Facies 2 deposits showing ropy bedding. Compare with Plate 4c and d. The Ropy bedded unit is overlain by an erosive based unit of Facies 4 (Lenticular sandstone). Notebook is 20cm long. Embsay Crag. SE005 551.

b) Facies 4 (Lenticular sandstone) channel from Rough Haw, Flasby Fell (SD963 558). Note the ropy bedding at the base and the pseudo-hexagonal weathering pattern of the massive sandstone which probably picks out dewatering paths in the channel sand. Map is 25cm in length.

c) Line drawing and log of the lenticular sandstone body in (b) above. Individual facies are numbered. Note the crude trough cross-bedding at the top of the Facies 4 channel and the overlying unit of Facies 8 trough cross-bedded sandstones which are not confined to the channel. Diagram is drawn to the scale of the log.

a)

b)

c)

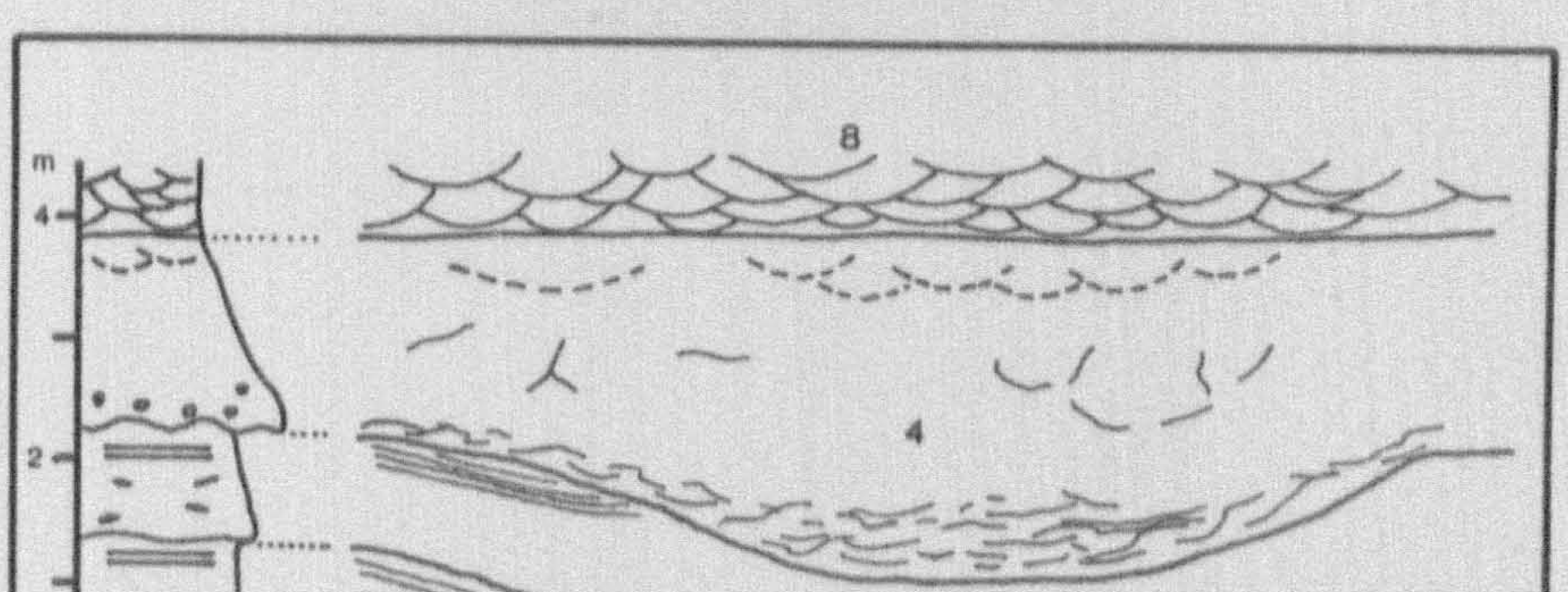


PLATE 9

Plates continue overleaf.

Plate 10: Growth faulting at Almscliff Crag, SE268 490.

a) View of the north-west face of Almscliff Crag (See Figure 4.10 for location). The top surface of the outcrop approximates a bedding plane within the Facies 7 (trough cross-bedded sandstones) which make up the crag. The major listric fault lies at the left end of the photograph. The extent of the rollover in the hanging wall is obvious. Crag is 15m high. See (c) for locations of the minor faults.

b) Two of the minor synthetic faults on the north-west wall of the crag (location arrowed on (a) above). The faults offset the trough cross-sets by approximately 20cm, the width of the sets. The structural dip in the rollover at this point is such that the downlapping foresets now appear as "onlapping sedimentary fills" of small half-grabens produced by the faulting (e.g. at X). This is an unfortunate coincidence as the small faults in fact show no evidence of growth: each set restores by removal of the same throw. The streaky appearance of the fault plane suggests that it formed a route for water escape either during or post movement. The short edge of the photograph is 35cm in length.

c) Simplified line drawing of (a) above to show the position of the minor synthetic and antithetic faults at the crag. Note the progressive rotation of the faults into the major fault plane which is just off the drawing to the left. Bedding plane A marks a level above which none of the faults extend. They appear either to have broken surface at this point and then to have been covered by renewed sedimentation or to have been eroded down to this level during later sedimentation. See text for further discussion. The destratified zones probably represent area of which were partially liquified during the movement on the main fault.

a)

b)

c)

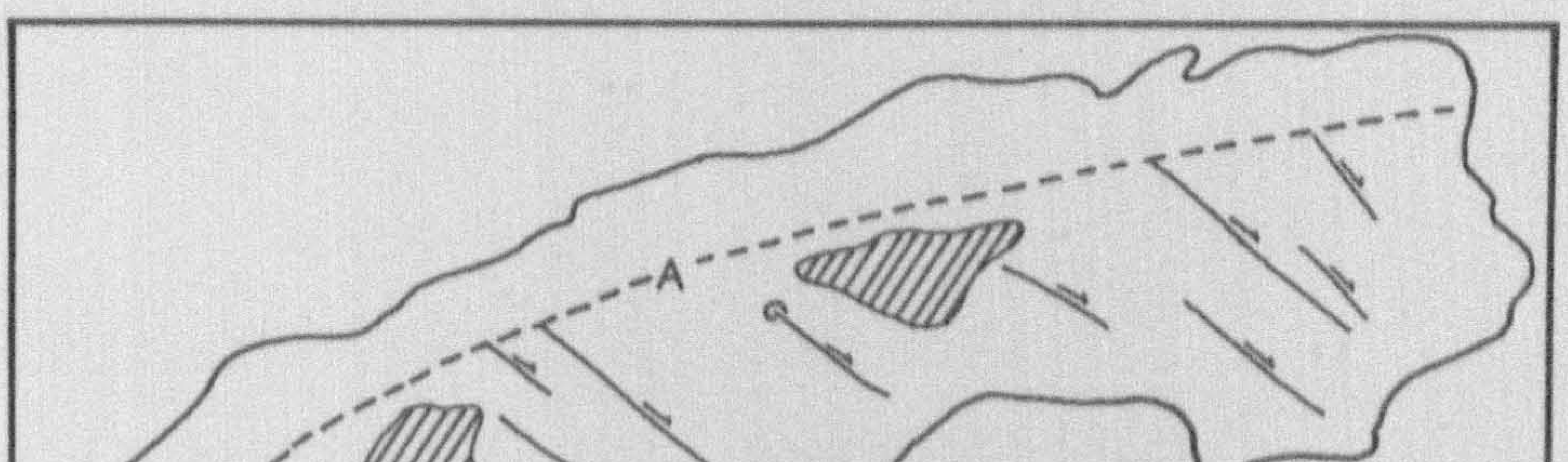


PLATE 10

CHAPTER FIVE

PETROLOGY AND DIAGENESIS OF PENDLE AND GRASSINGTON GRIT GROUP SANDSTONES.

5.1 Introduction and Aims.

This chapter presents the results of a reconnaissance study of the petrography of sandstones developed within the E_{1c} basin-fill sequence. The study began as an investigation of the provenance of the sediments and as an attempt to determine whether the Pendle and Grassington Grit Groups could be distinguished by petrographic criteria. Also, it was hoped that there might be proximal to distal petrographic variations within each of the two Groups. These, it was thought, would help in producing a more constrained palaeogeographic reconstruction for the area during the Pendleian.

Initially 35 samples were collected from exposures of the Pendle and Grassington Grit Groups across a wide geographical area (Figure 5.1). Samples were taken from sandstones of both coarse and fine facies so that facies/grainsize controls on petrology could be determined (Table 3). This initial study showed a remarkable uniformity in the bulk composition of the sediments throughout the two stratigraphic Groups (See below) but revealed complex paragenetic sequences which seemed to help constrain the burial and uplift history of the sandstones. Consequently, the scope of the study was broadened to include the diagenetic aspects of the sediments. At this stage, additional samples were acquired from boreholes in areas with different burial histories to that of the maincrop. Most significantly, samples from the B.G.S. Roosecote and Place Oil Whitmoor boreholes were made available courtesy of the B.G.S. (Figure 5.1). The E_{1c} sandstones in these boreholes have dramatically different burial histories to the sandstones at outcrop and provide excellent calibration points unaffected by surface weathering.

5.2 Sandstone Texture.

Thin-sections and hand specimens were visually assessed for grainsize, rounding and sorting. In general, sorting is closely linked to facies: the traction

Table 3: Sample data by stratigraphic units.

PENDLE GRIT FORMATION

Figure Ref.*	Sample Number	Facies	Grainsize (Went.)	Range (mm)	Sorting	Location	Depth (m)
A	48038	A	Medium	0.125-2.0	P	SD772385	
B	48039	B	Medium	0.25-0.75	P-M	SD772385	
C	48040	C	Fine	0.0625-0.25	P-M	SD654312	
D	48043	C	Fine	0.0625-0.75	VP	SD815397	
E	48256	D	Silt	0.0625-0.125	W	SE005508	
F	48257	B	Medium	0.0625-0.5	P	SE689406	
G	48258	C	Medium	0.125-0.5	M/W	SE077544	
H	49218	A	Coarse	0.5-2.0	PM	SE002548	
J	49219	B	Medium	0.125-1.0	P	SE004510	
K	49220	A	Coarse	0.125-2.0	VP	SE004510	
L	49224	C	Fine'	0.125-1.0	PM	SD652560	
M	49225	C	V.Fine'	0.0625-1.0	PB	SD652560	
N	49227	B	Medium	0.125-2.5	VP	SD876452	
O	49228	C	Fine'	0.125-1.75	VPB	SD625528	
P	49229	B	Fine	0.0625-1.0	PM	SE084538	
Q	49230	C	Medium	0.0625-0.75	PM	SD662550	
S	49232	A	Coarse	0.125-3.0	P	SD545455	
T	49234	B	Medium	0.125-0.5	MW	SD523677	
U	49235	F/C	Coarse'	0.0625-2.0	VPB	SD785411	
W	49236	B	Medium	0.125-1.5	VP	SD787412	
X	49419	B	Fine	0.0625-0.25	MW	SD948449	
R1	49418	C	Fine	0.125-0.25	M	SD230687	487
R2	49420	C	Fine	0.125-0.25	W	SD230687	324
R3	48421	C	Fine	0.125-0.25	M	SD230687	491
R4	49422	?C	Medium	0.125-0.5	W	SD230687	199
R5	49423	?C	Fine	0.125-0.25	M	SD230687	477

PENDLE SHALE FORMATION

Figure Ref.*	Sample Number	Facies	Grainsize (Went.)	Range (mm)	Sorting	Location
I	48047	E	V.Fine	0.0625-0.125	W	SD782404
II	48255	G	Silt	0.03125	W	SD782404
III	48259	A	Medium	0.125-1.0	P	SE291516
IV	48260	B	Medium	0.125-1	M	SE278542
V	49231	B	Fine	0.125-0.25	W	SD685580
VI	49233	B	Medium	0.125-0.75	MW	SD683563

GRASSINGTON GRIT GROUP

Figure Ref.*	Sample Number	Facies	Grainsize (Went.)	Range (mm)	Sorting	Location
1	48036	1	Medium	0.125-1.5	P	SD716476
2	48041	9	Medium	0.25-0.75	M	SE034552
3	48042	7	Coarse	0.25-0.75	W	SE277547
4	48044	?8	Medium	0.125-0.5	MW	SD833402
5	48045	1	Coarse	0.25-2.0	M	SD786384
6	48046	1	Fine	0.125-1.0	M	SD786384
7	49226	1	Coarse	0.25-0.75	W	SD676359

P= Poor, M= Moderate, W= Well, V= Very, B= Bimodal.

* See Figure 5.1.

' Greywacke facies.

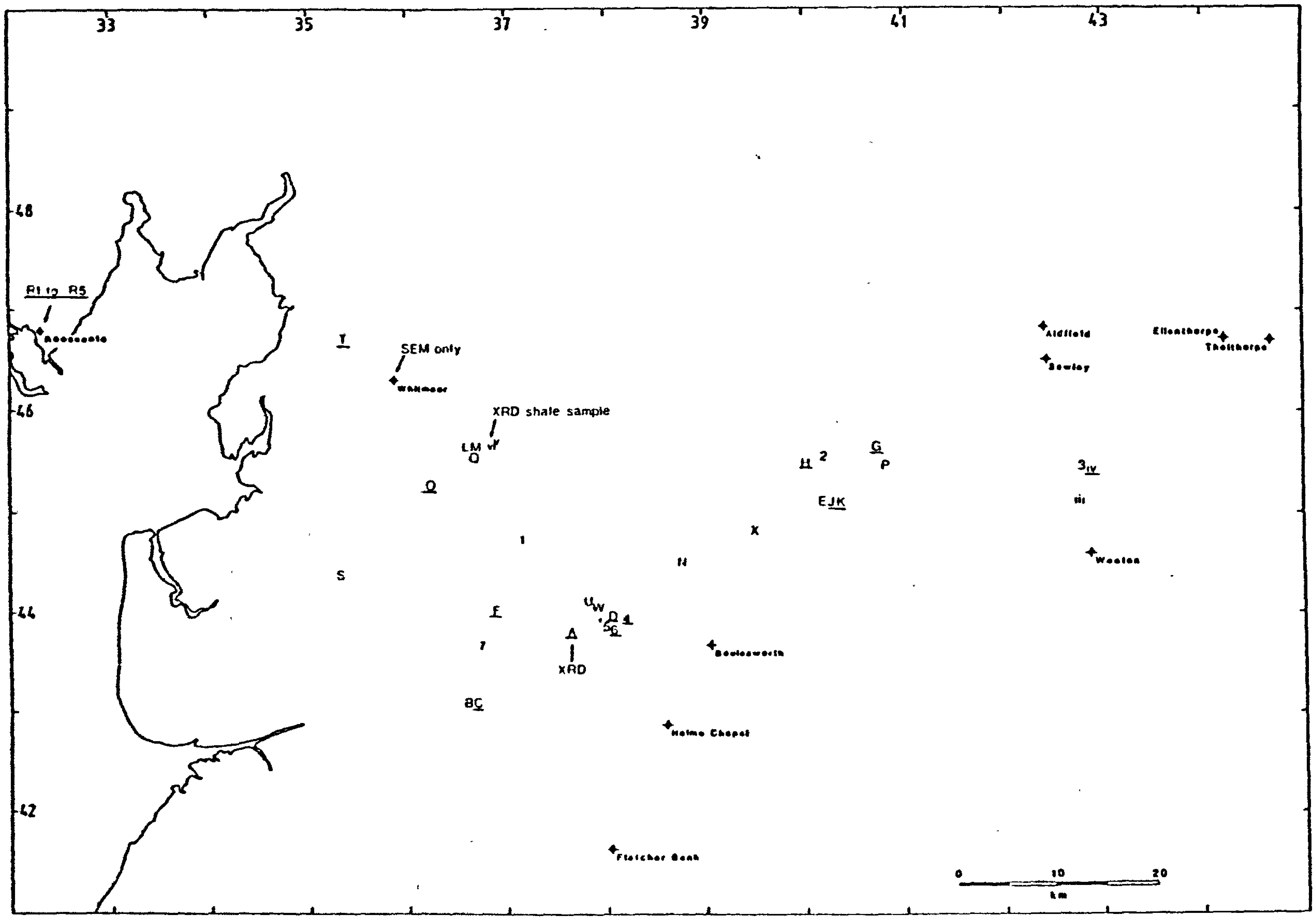


Figure 5.1: Distribution of sample locations. Refer to Table 3 for key to localities.

dominated facies in both the Pendle and Grassington Grit Groups are moderately-well to well sorted regardless of grainsize (Table 3) while the mass-flow sediments show very variable sorting in the range poor to moderate. Mass-flow sandstones in the Pendle Grit Group are always more poorly sorted than equivalent facies in the Grassington Grit Group (refer to Figure 5.5). This probably reflects enhanced sorting of the latter sediments during storage in the fluvial system and associated braid-delta from which they were sourced (See Chapter 4). The "greywacke" sediments of the Pendle Grit Group (Facies-type C_c) tend to have a bimodal grain population with one mode in the range medium to coarse sand and the other in the clay to very fine sand range. The latter grainsizes form the greywacke matrix within which the coarser grains float.

The roundness of the individual grains varies widely within and between samples: the principal framework grains, quartz and feldspar, can vary from angular to well-rounded within a single thin-section. This variation does not appear to be grainsize or facies dependent. Cathodoluminescence studies of quartz fabrics confirm that the variation in rounding remains when the effects of later quartz authigenesis are removed. Consequently, it is thought likely that this wide divergence in rounding reflects an original difference in provenance: the more rounded grains may have travelled very much further or, possibly, be a second-cycle component (See sub-section 5.3.2).

A further feature of the fabric of some of the Pendle Grit Group sandstones is an alignment of the long axes of the quartz grains (Prior and Sims, 1986). Most quartz grains show slight elongation parallel to the crystallographic c-axis and this elongation is sufficient for sedimentary processes to cause alignment of grains during deposition. The alignment is invisible in most hand specimens but, in thin section, it results in grains going into extinction at approximately the same angle of stage rotation. Comparison of the quartz fabrics with mica orientations and palaeocurrents shows that the grains are aligned with their long axes pointing along current, parallel to bedding. Hiscott and Middleton (1980) measured similar fabrics in coarse grained turbidites from the Tourelle Formation, Quebec, and suggested that these were due to shear during the final stages of deposition. Such an interpretation is consistent with the depositional mechanics suggested for the coarser facies in the Pendle Grit Group (See Chapter 3). Further study of quartz fabrics in such facies might confirm that they can be used as palaeocurrent indicators. Such indicators would be very valuable for sub-surface studies of turbidite systems and also in surface studies of coarse-grained systems in which conventional palaeocurrent indicators are rare.

5.3 Original composition and classification of the E_{1c} sandstones.

Thin-sections of all samples (excepting those from Whitmoor-1 which were unsuitable for thin-section preparation) were point counted using an automatic stage capable of recording 12 separate channels. A maximum of 11 channels were needed to record the phases present in these sandstones. 400 counts were made on most thin-sections with selected samples being counted several times to ensure repeatability of the measurements. Results of the point counting are included as Tables 4, 5 and 6. In this section the individual framework constituents are described and then the provenance is discussed in Section 5.4. Diagenetic phases are covered in Section 5.5.

5.3.1 Quartz.

The majority of the framework grains are composed of monocrystalline, polycrystalline or strained quartz. Most of these grains appear to be derived from acid igneous or acid gneissic rocks. The monocrystalline grains are frequently full of inclusion trails and fluid inclusions which give the grain a dusty appearance. Although the inclusions are usually sub-microscopic some larger examples are identifiable and include rutile, zircon, apatite and muscovite. Strained quartz grains are monocrystalline grains which show undulose extinction. Such straining can be caused during growth of the original quartz crystal or by subsequent tectonic stress. Consequently, strained quartz grains are not diagnostic of a specific type of source terrain as was once argued (See Pettijohn *et al.*, 1972).

Within the polycrystalline quartz grains there are two fabric types. The first consists of uniformly sized subgrains with straight contacts and the second comprises highly elongate subgrains with sutured contacts. The former can be difficult to distinguish from a collection of monocrystalline grains "welded" together by diagenetic quartz overgrowths. Both fabric types are characteristic of acid igneous rocks and acid pegmatites (Pettijohn *et al.*, *op. cit.*).

Table 4: Point count results for Pendle Grit Formation Sandstones.

Sample	Quartz	Ortho-clase	Micro-cline	Olig-clase	Musc-ovite	Bio-tite	Matrix *	Kaol-inite	Others	Porosity +	Carb-onate
A	80.5	5.5	4.0	Tr.	Tr.	1.5	5.5	3.0	Tr.	Tr.	
B	77.7	5.2	0.7	0.7	1.2	1.2	6.2	Tr.	7.0	Tr.	
C	66.5	4.7	2.2	0.2	1.5	9.5	14.0	Tr.	1.0		
D	75.0	5.0	0.8	0.4	0.4	5.6	10.6	Tr.	2.0		
E	59.0	5.5	Tr.		3.0	4.5	28.0	Tr.	Tr.		
F	79.5	6.7	1.0		0.5	0.5	Tr.	3.0	Tr.	9.0	
G	70.2	7.7	1.0	Tr.	0.5	14.2	3.7	1.5	0.5		
H	81.2	6.0	4.0		0.6	Tr.	3.6	3.3	1.0	Tr.	
J	54.5	4.6	1.0	Tr.	2.3	Tr.	Tr.		Tr.		37.3
K	79.9	8.6	2.0		Tr.	3.6	5.6		Tr.		
L	66.0	4.6	Tr.		1.0	Tr.	27.3	1.0	Tr.		
M	73.6	4.6	1.6	Tr.	0.3	Tr.	14.0	Tr.	Tr.		5.0
N	76.0	6.6	2.0		0.6		13.3		1.3		
O	64.3	7.9[1]			2.0	12.0	12.0	1.0	0.6		
P	65.3	5.0	0.6	Tr.	0.6	Tr.	16.6	Tr.	Tr.	11.6	
Q	75.3	7.3	1.6		0.6		8.3	Tr.		6.6	
S	72.0	4.6	1.6		2.6	Tr.	12.6	1.0		5.3	
T	84.0	5.0	2.6		Tr.		3.3	2.6	0.3	2.0	
U	47.6	5.3	0.6		1.3	2.6	39.3	2.0	1.0		Tr.
W	71.9	1.0	1.3		0.6		10.3	0.3	0.4	14.0	
X	66.4	4.6	1.6		1.6		25.6				Tr.
R1	78.6	7.0	0.3	Tr.	1.3		Tr.	6.3	1.0		4.6
R2	38.3	4.0	1.6		1.6				0.6		53.6
R3	75.3	5.3	1.0		2.0		2.0	4.3	1.3		6.3
R4	64.3	7.6	1.6		0.6			25.3	0.3		
R5	81.0	4.3	0.3		0.6			6.3	0.3		5.3

* See text for description.
+ Apparent macro-porosity, mostly caused by damage during thin-section preparation.
[1] Includes all feldspars: extensive alteration obscures individual mineralogies.

Table 5: Point count results for Pendle Shale Formation Sandstones.

Sample	Quartz	Ortho-clase	Micro-cline	Olig-clase	Musc-ovite	Bio-tite	Matrix *	Kaol-inite	Others	Porosity +	Carb-onate
I	69.2	7.5	0.7		1.7	27.2	0.7	Tr.	0.2		
II	69.5	7.0	1.5		2.0	3.5	Tr.	1.0	Tr.		2.5
III	78.2	6.5	0.5	0.2	2.2	Tr.	Tr.	1.7		10.5	
IV	73.0	14.5	1.5	0.2	0.5	3.0	Tr.	6.2	Tr.		
V	82.6	6.6	0.6		Tr.	3.3	3.0	3.0		0.6	
VI	83.0	3.3	1.0	Tr.	1.0	1.0	4.0	1.3	Tr.	5.3	

Table 5 (cont.): Point count results for Grassington Grit Group Sandstones.

Sample	Quartz	Ortho-clase	Micro-cline	Olig-clase	Musc-ovite	Bio-tite	Matrix *	Kaol-inite	Others	Porosity +	Carb-onate
1	82.3	4.3	1.6	0.6	1.6	1.3	Tr.	6.3	Tr.		
2	78.5	2.3	Tr.	Tr.	1.3	1.0	Tr.	3.0	3.0	10.6	
3	75.0	9.0	2.0	0.6	0.4	2.0	1.0	7.8	0.4	2.8	
4	80.1	8.2	0.7		Tr.	0.5	7.7	Tr.	1.2	1.0	
5	72.0	7.6	3.2	0.4			2.0	Tr.	1.6	13.2	
6	77.9	8.5	2.0		0.7	0.2	2.2	Tr.	0.5	8.0	
7	85.0	4.3	2.3	0.6	2.3		0.3	4.3	0.6	Tr.	

* See text for description.
+ Apparent macro-porosity, mostly caused by damage during thin-section preparation.

Table 6: Quartz grain types recalculated to 100%

Pendle Grit Formation

Sample	Mono-crystalline	Poly-crystalline	Strained	Grain-size
A	65	17	18	M
B	77	15	8	F
C	93	5	2	F
D	84	13	3	M
F	85	13	2	M
G	86	12	2	M
H	90	7	3	C

Grassington Grit Group

Sample	Mono-crystalline	Poly-crystalline	Strained	Grain-size
1	81	6	12	M
2	79	16	5	M
3	74	15	11	C
4	86	5	9	M
5	68	25	7	C
6	78	16	6	F

F=Fine, M=Medium, C=Coarse

Point counting of representative thin-sections from the Pendle and Grassington Grit Groups shows that the relative proportions of the three quartz types are quite variable (Table 6). Monocrystalline quartz is always dominant and shows a slight increase in relative abundance in the finer sandstones, as would be expected. However, this increase is not universal as shown by samples such as H and 4 (Table 6). Perhaps more interesting is the slight decrease in the proportion of monocrystalline grains between the Pendle Grit Group samples and those from the Grassington Grit Group. With the data available it not possible to say whether this is a random result or whether it arises from a sedimentological or source control. If the apparent trend is real it might be explained either by the grain-size factor mentioned above or, more significantly, it might represent minor changes in the provenance. This aspect of the petrography might be worth expanding upon in future work.

5.3.2 Feldspar.

Orthoclase, microcline, orthoclase microperthite and oligioclase are present throughout the suite of samples. Orthoclase is always the most abundant feldspar, with a maximum of 14.5% being recorded in sample (iv) from the Pendle Shale Formation. When thin-section damage and diagenesis is allowed for (see 5.3.7), maximum original orthoclase percentages may have been closer to 16%. The ratio of orthoclase to microcline is variable but averages 5:1. Microperthite and oligioclase are present in trace amounts in most samples. The latter is usually very fresh (Plate 11a).

The most noticeable feature of the feldspar framework grains is the wide variation in weathering/diagenesis which they show. Often fresh and heavily altered feldspars of the same mineralogy are present within a single thin-section (Plates 11b, 12c), sometimes even in contact with each other. Similarly, fresh feldspars and kaolinite pseudomorphs after feldspar are commonly seen in close association (Plates 12b, 13a, 13b). The presence of both fresh and altered examples of the same mineral species indicates that the alteration is not controlled entirely by feldspar chemistry. Instead, it is suggested that the original framework feldspars consist of two separate suites. One suite represents first cycle potassium feldspars supplied relatively unaltered to the depositional site (note that the fresh oligioclase probably also belongs to this suite). The second suite consists of potassium feldspars which were already strongly altered prior to incorporation into the sandstones of the Pendle and Grassington Grit Groups. This alteration might have been due to a long transport path from the source area, intense chemical weathering in the source area or to the feldspars being second-cycle erosion products. These are not mutually exclusive possibilities as discussed further in Section 5.4.

5.3.3 Rock fragments.

Rock fragments are poorly represented among the framework grains. Rare chert, schist and granitic fragments have been observed but never at an abundance sufficient to warrant separate classification during point counting. Small mudrock fragments are slightly more common but are usually highly deformed during compaction to produce a "pseudomatrix" (c.f. sub-section 5.3.5). In this study polycrystalline quartz grains have been considered as quartz rather than as rock fragments.

5.3.4 Micas.

White mica is present in most of the sandstones. It occurs as large flakes (Plate 11c) in most facies and as micro-flakes in the matrix of the mud-rich facies. The larger flakes are most abundant in finer sandstones where they may be concentrated on ripple foresets or in thin mica and plant rich laminae. Chemical analyses on the Electron Microprobe (WDAX) and on the SEM (EDAX) show that the mica is always of muscovitic composition.

Biotite was an important component of the finer grained sediments, particularly in the Pendle Grit Group. In all cases it has been heavily altered to a green-brown pleochroic hydromica and hydrous iron oxides (Plate 13b).

5.3.5 Matrix.

In the Pendle Grit Group and the mass-flow facies of the Grassington Grit Group there is a significant proportion of depositional matrix. Today, this is composed of a mixture of poorly crystalline kaolinite, micromica, illite, quartz silt and pseudomatrix formed by compaction of mudrock fragments. This matrix has undergone some recrystallisation during diagenesis (Sub-section 5.5.5). In order to try and determine the original mineralogy of the matrix, it was decided to look at the mineralogy of the shales in the Pendle Shale Formation. The clays in these shales are believed to be representative of the allogenic clays incorporated into the sandstones because of the close sedimentological link between the sandstones and the shales (See Chapter 3). X-ray diffraction (XRD) slides were prepared from a shale sample taken from just above the highest sandstone in the Pendle Grit Formation (Figure 5.1). Whole rock, sub-2μm and sub-0.5μm grainsize fractions were separated from the shale using the technique of Ord (1988). The results of the XRD work (Figures 5.2, 5.3 and 5.4) show that kaolinite, quartz and a trace of illite are present. SEM studies of this shale sample indicate that the illite is authigenic (See sub-section 5.5.5 and Plate 19b) and hence it is concluded that the main allogenic clay mineral is kaolinite. The kaolinite in the matrix of the sandstones then is also likely to be allogenic.

5.3.6 Heavy Minerals and other trace components.

A varied heavy mineral suite has been recognised in these rocks. Zircon (rounded and euhedral), monazite, apatite and tourmaline (in that order) are the most common elements of this suite. Rutile, ?magnetite and sphene have been detected on the microprobe but have not been recognised in thin-sections (Plate 19c). This heavy mineral suite is similar to that recorded in other Namurian sandstones by Gilligan (1920) and by Drewery (1987).

Wood and plant fragments are the only other allogenic components of the sandstones. They are present in most of the sandstones from the Pendle Grit Group and in some of the Grassington Grit Group samples. Small, sub-millimetre platy fragments tend to follow mica in abundance, being relatively more common at the top of finer sandstones and, particularly, in rippled units. Larger, more blocky fragments are found dispersed throughout the rock volume of other facies. All of the carbonaceous material preserved is vitrinite or inertinite.

5.3.7 Classification and related issues.

Sandstone nomenclature has always proved troublesome in rocks which have undergone substantial diagenetic change or which include significant allogenic matrix (Pettijohn *et al.*, 1972). The sandstones of the Pendle and Grassington Grit Groups suffer from both these classification problems. At the

Figures 5.2, 5.3 and 5.4.

X-ray diffractograms for a mudrock sample from the Pendle Shale Formation (Whitendale River, SD656 568). Figures 5.2, 5.3 and 5.4 are for Whole Rock, sub-2μm and sub-0.5μm grainsize fractions respectively. Intensity peaks are as follows:

I	Illite (001).
I_1	Illite (002).
K	Kaolinite (001), destroyed on heating.
K_1	Kaolinite (002), destroyed on heating.
Q	Quartz (101).
Q_1	Quartz (100)

Note the illite peaks become stronger with decreasing grainsize and the kaolinite peaks *vise versa*. This shows that the illite is dominant within the smaller sized grain fractions, as suggested by S.E.M. studies. The presence of quartz peaks in all the grain fractions shows that clay and silt grade quartz is a component of the detrital mudrock. There is no evidence of swelling clay minerals on these diffractograms

(compare with Figures 5.9 to 5.11). The reason for the enhanced reflections following glycolation of the samples is unknown.

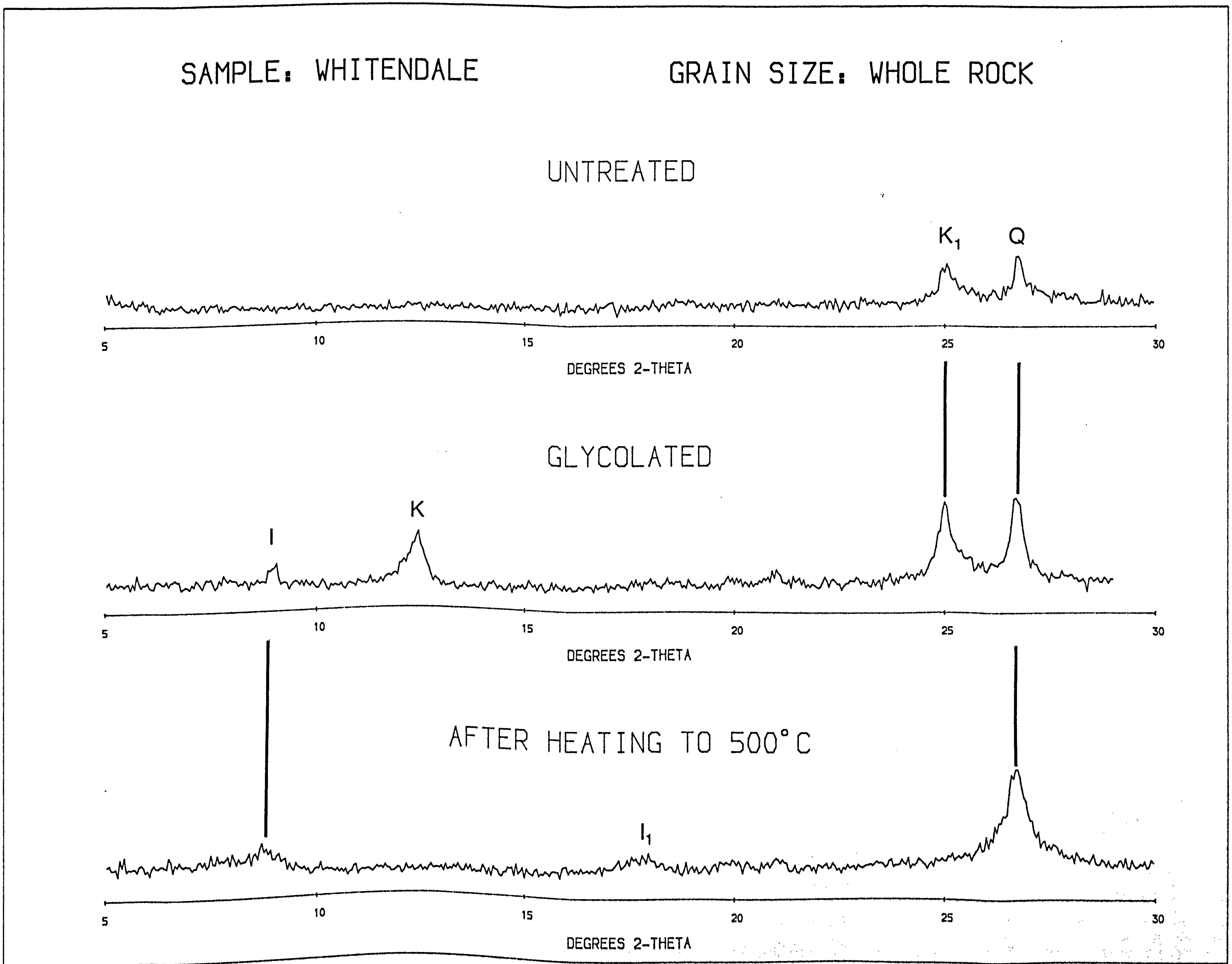

SAMPLE: WHITENDALE
GRAIN SIZE: WHOLE ROCK
UNTREATED
K1
Q
5
10
15
20
25
30
DEGREES 2-THETA
GLYCOLATED
I
K
5
10
15
20
25
30
DEGREES 2-THETA
AFTER HEATING TO 500°C
I1
5
10
15
20
25
30
DEGREES 2-THETA

Figure 5.2

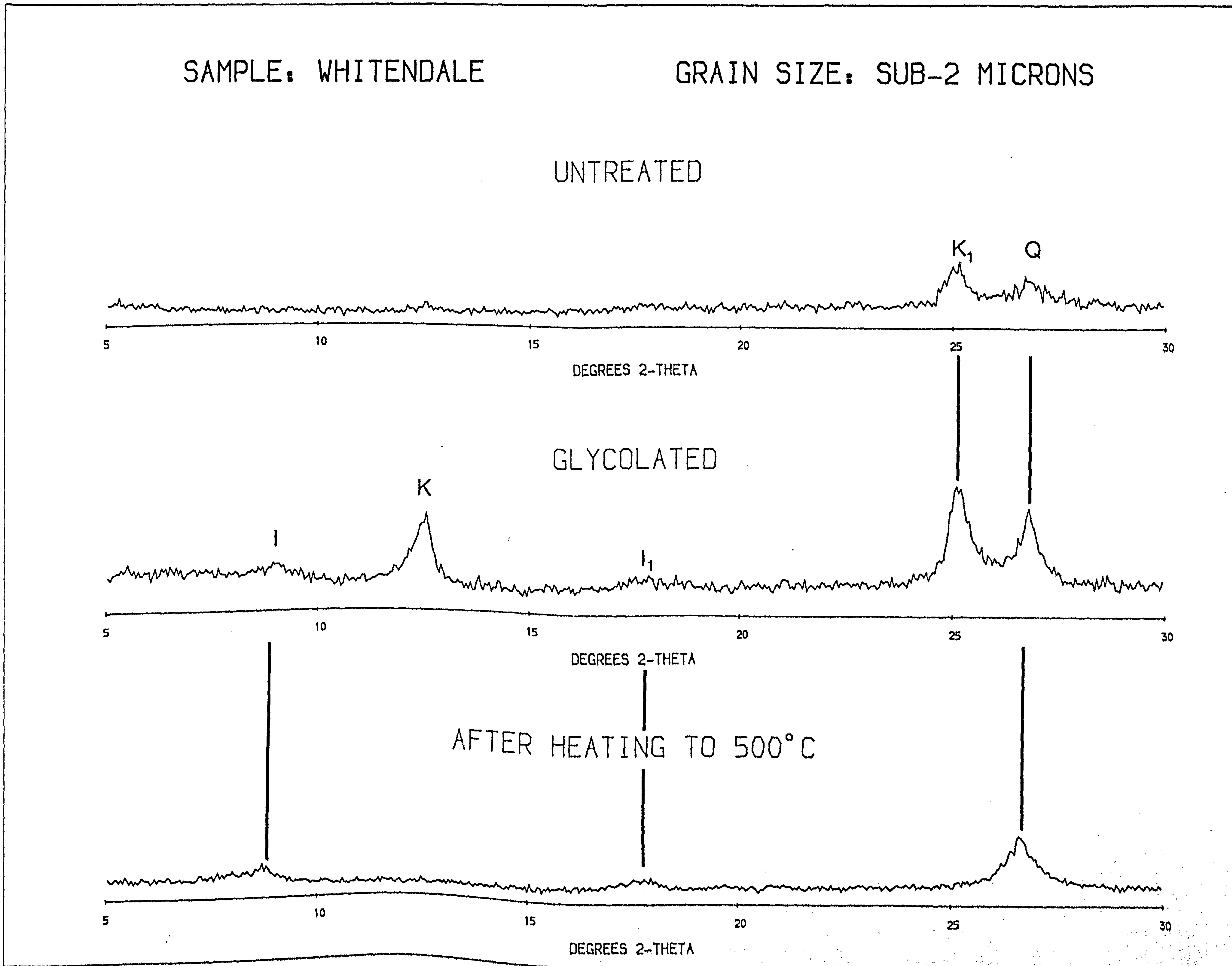
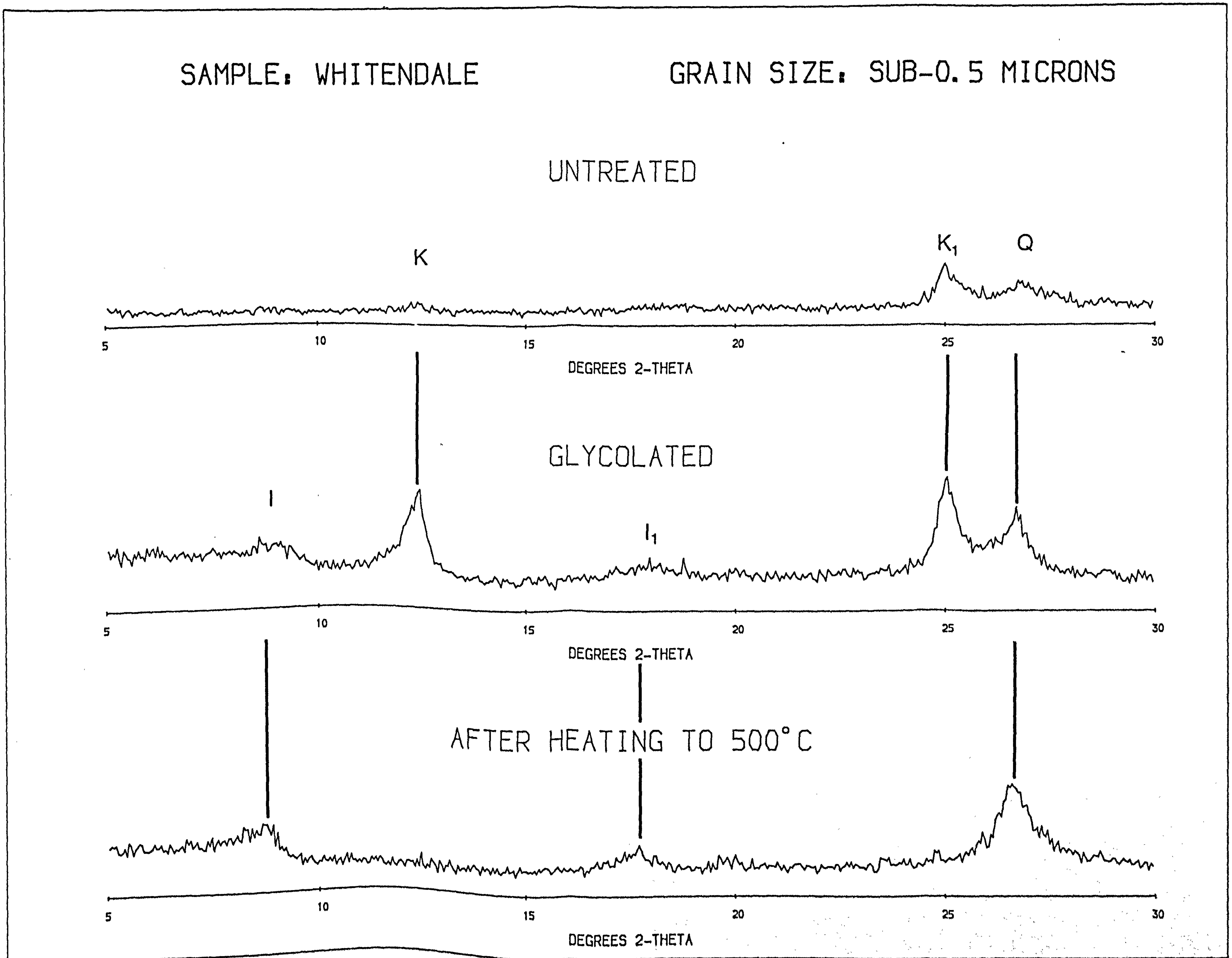

SAMPLE: WHITENDALE
GRAIN SIZE: SUB-2 MICRONS
UNTREATED
K1
Q
5
10
15
20
25
30
DEGREES 2-THETA
GLYCOLATED
K
I
I1
5
10
15
20
25
30
DEGREES 2-THETA
AFTER HEATING TO 500°C
5
10
15
20
25
30
DEGREES 2-THETA

Figure 5.3



SAMPLE: WHITENDALE
GRAIN SIZE: SUB-0.5 MICRONS
UNTREATED
K
K_1
Q
5
10
15
20
25
30
DEGREES 2-THETA
GLYCOLATED
I
I_1
5
10
15
20
25
30
DEGREES 2-THETA
AFTER HEATING TO 500°C
5
10
15
20
25
30
DEGREES 2-THETA

Figure 5.4

present time all the sandstones with less than 15% matrix fall into the subarkose field of the Pettijohn *et al.* (*op. cit.*) nomenclature, regardless of whether polycrystalline quartz is counted as quartz or as rock fragments for the purposes of classification. The matrix-rich sands classify as arkosic wackes. However, in order to understand the provenance it is necessary to try and unravel the diagenetic alterations so as to arrive at a knowledge of the original sandstone composition. In this study the effects of diagenesis have been crudely removed by including the kaolinite and apparent porosity point count classes within the feldspar totals. This is justified in that much of the kaolinite is found as pseudomorphs after feldspar while all the apparent porosity in thin-sections is caused by plucking of rotted feldspars and kaolinite during thin-section preparation (the latter can be proved by comparing SEM and thin-section samples of the same sandstone). In classification terms this recalculation of the original feldspar content shifts the more feldspathic sandstones into the Arkose field of Pettijohn *et al.*. However, because of the limited number of rock fragments, the classic quartz/feldspar/rock fragment triangular plots of these authors are ineffective at showing any compositional trends within the sandstones. Instead, Figure 5.5 shows triangular plots of quartz/feldspar/matrix for the sandstones of the Pendle Grit Formation and the Grassington Grit Group. These plots do show certain trends. In the Pendle Grit Formation, the plot shows that the quartz/feldspar ratio varies little regardless of grainsize but that the finer sediments are usually more matrix rich. Comparison of the two plots suggests that there is no difference in quartz/feldspar ratio between the Groups, although the Grassington Grit Group sediments are texturally more mature at any given grainsize. As these variations can all be attributed to facies and grainsize factors (See comments in 5.2) there appear to be no compositional trends which might suggest differences in provenance either within or between the sandstones of the two Groups.

5.4 Discussion of Provenance.

The lack of compositional changes noted above indicates that the ultimate source area for the sandstones of the Pendle and Grassington Grit Groups did not change substantially during E_{1c} times. The nature of the framework grains suggests that the parts of the source area supplying coarse clastic material were coarse grained acid igneous and metamorphic rocks or, possibly, sediments derived from such rocks. Fresh feldspar suites in the E_{1c} sandstones confirm that plutonic acid igneous rocks were a significant component of the source area. The weathered feldspar suite, however, may represent the erosion product of older arkosic sandstones or of deeply weathered acid igneous or metamorphic rocks. In order to help pin down the source terrains for Carboniferous clastic sediments, Drewery *et al.* (1987) and Drewery (1987) investigated the ages of the zircon suites in a wide range of Carboniferous sandstones. The results of these authors showed that most of the zircons were ultimately of Archean origin. Based on the colours and textures of the various zircon populations Drewery (*op. cit.*) was able to show that the following zircon types were present in all of the sandstones studied*:

1) Zircons derived from direct erosion of Archean terrains.
2) Zircons produced by erosion of sediments derived from Archean terrains.
3) Zircons eroded from Caledonian granites which had scavenged Archean zircons during petrogenesis and emplacement.

These results are entirely compatible with the bulk petrography above, particularly the feldspar data: the weathered feldspars may be sourced from the weathering of Archean terrains or to recycling of sediments derived from these terrains (e.g. Torridonian Sandstone, Devonian sandstones). The fresh feldspars would correspond to erosion of Caledonian granites.

Drewery's work has highlighted the need to look beyond the immediate environs of northern Britain for the source terrains of the Carboniferous sandstones, including those discussed herein. Her generalised palaeogeographic map showing the position of terrains capable of supplying the various zircon populations is included as Figure 5.6. Areas from which the feldspar suites recognised herein might have been sourced have been added to the diagram. As Drewery *et al.* (1987) note, it must have required a truly continental scale drainage system to move material over the distances shown.

The nature of the feldspars, together with the allogenic clay mineralogy of the sandstones and shales of the Pendle and Grassington Grit Groups, may give

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* Note: this is a rather simplistic summary of a detailed and complicated argument based on isotopic evidence. Refer to Drewery (1987) for a full discussion of the interpretation.

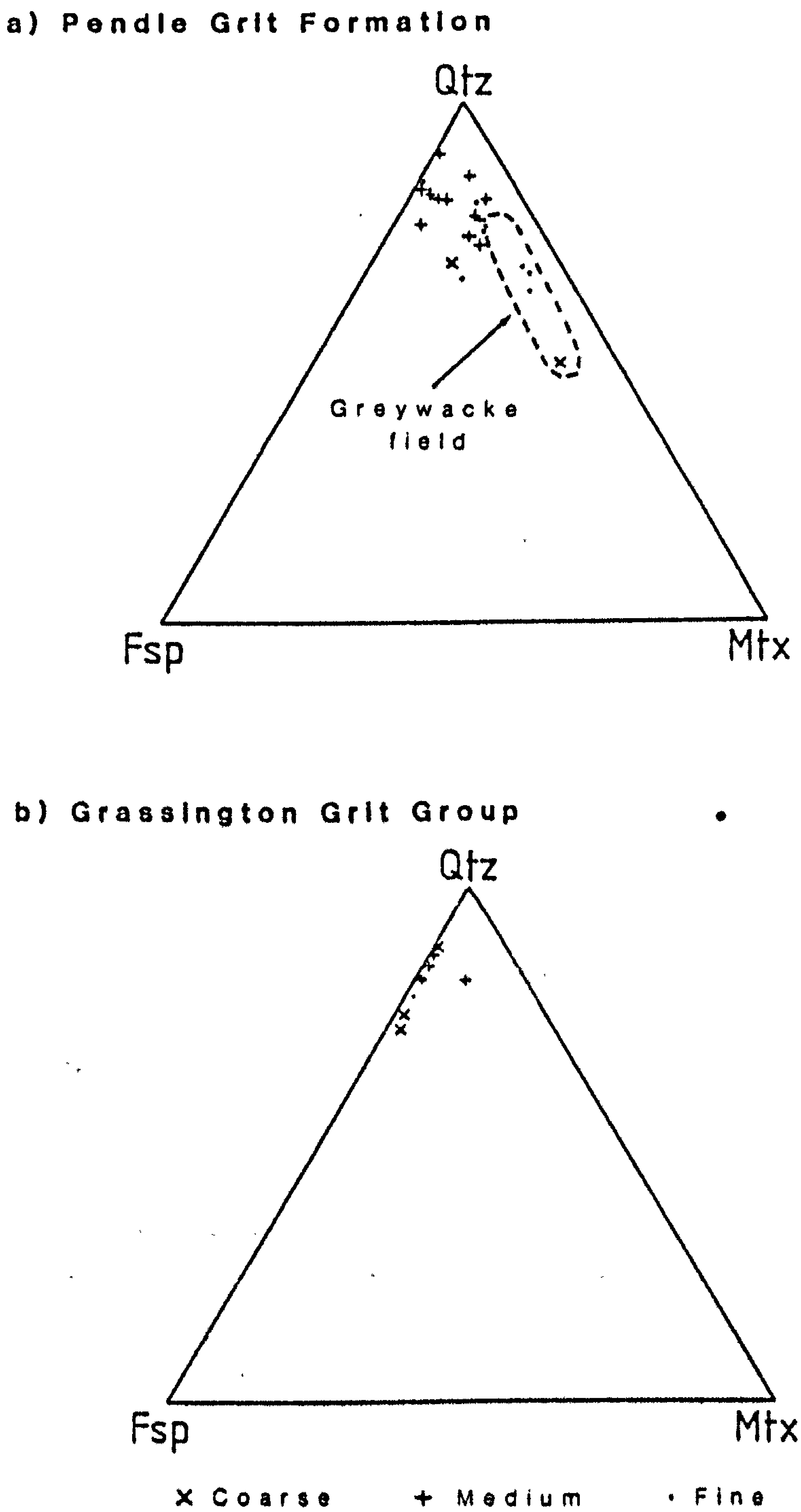


Figure 5.5: Quartz/feldspar/matrix plots for sandstones of the Pendle and Grassington Grit Groups. For discussion see text.

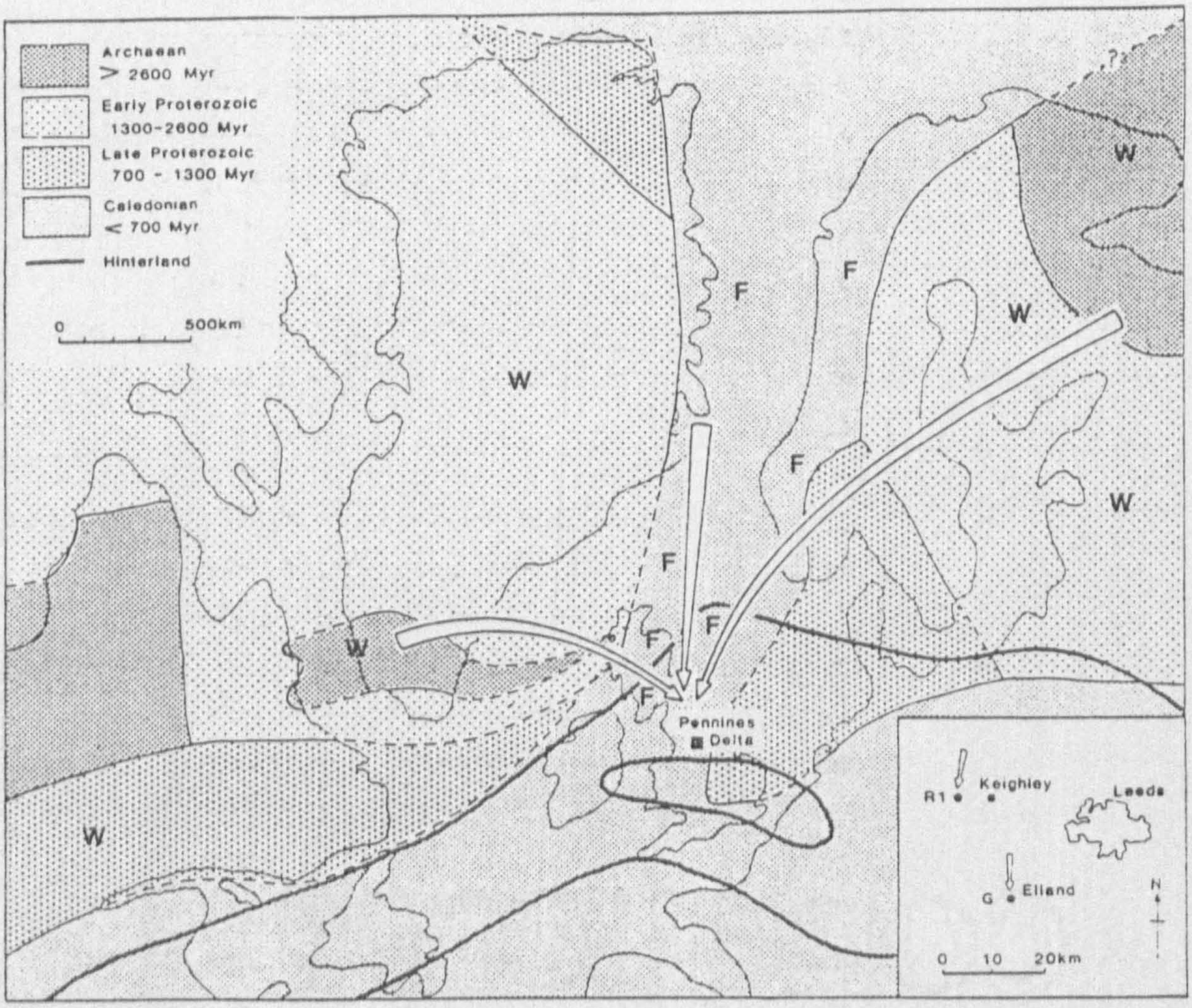


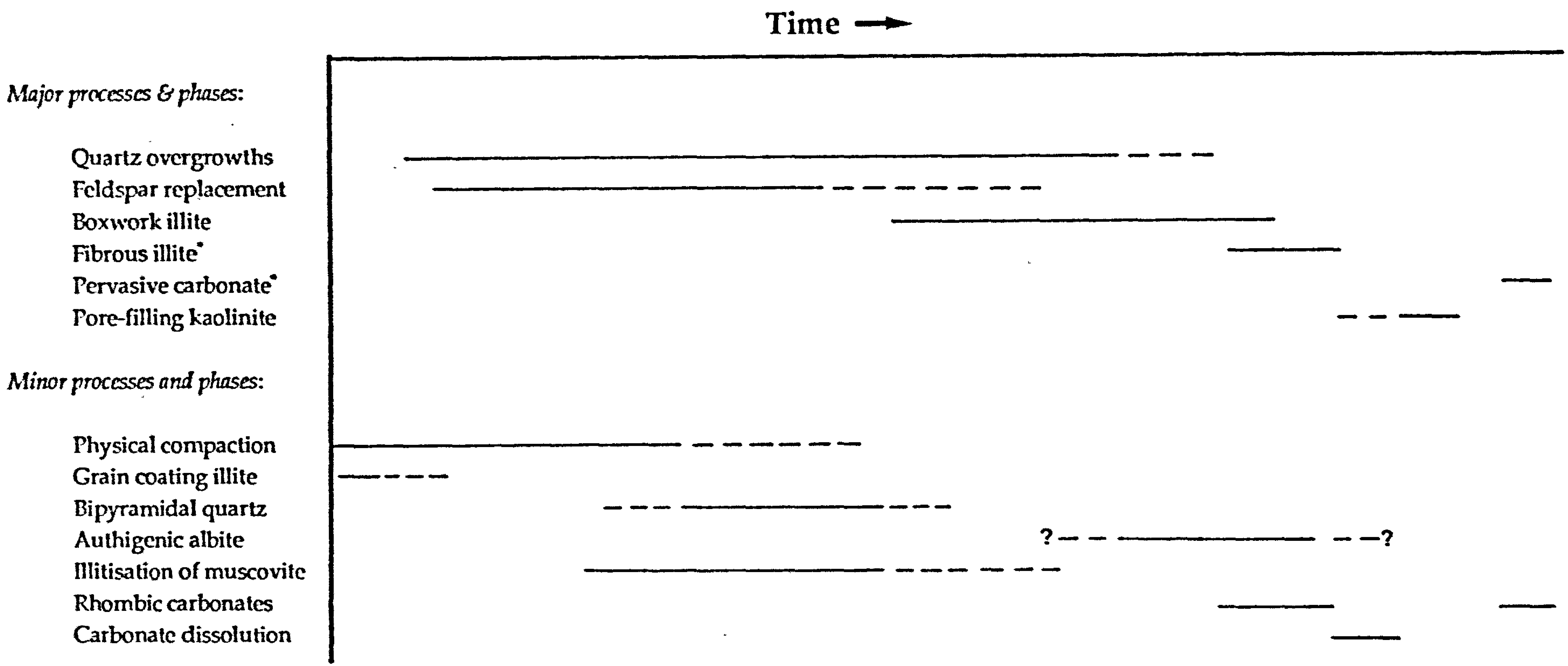
Figure 5.6: Generalized palaeogeographic map for Carboniferous Europe showing the likely source terrains for the sandstones of the Pendle and Grassington Grit Groups. The areas from which the feldspar suites discussed in the text were probably derived are highlighted: F= fresh feldspars, W= weathered feldspars. Modified after Drewery *et al.* (1987).

some insight into erosion processes operative in the sources areas. The fresh feldspars are indicative of high rates of mechanical erosion, probably related to significant relief in the source terrains. Kaolinite, however, suggests very intense chemical weathering under acid conditions. At the present time, kaolinite is the most common clay mineral in tropical soils subject to extreme leaching by high rainfalls. As a consequence of this, most of the clay minerals supplied by rivers to low-latitude deltas are composed of kaolinite (H. Jenkyns, pers. comm.). The abundance of allogenic kaolinite in the sediments of the Pendle and Grassington Grit Groups are, therefore, interpreted as being due to tropical, high rainfall conditions in at least part of the source terrain.

Comparison of the petrography of the Pendle and Grassington Grit Group sandstones with time-equivalent sandstones on the Askrigg Block (Wensleydale Group/Yoredales), shows that the latter are much less feldspathic and texturally more mature (data from Dunham and Wilson, 1985). This apparent difference in petrography is problematic if all the sediment supplied to the Carboniferous basins of northern England came from a single, massive fluvial system. One explanation for the petrographic differences is that there were two fluvial systems supplying the same basin. While the terrains undergoing erosion in the drainage basins of these systems may have been of similar ages, there was a slight difference in the clastic sediments produced in each system. This scenario is compatible with the data of Drewery (1987) and is hinted at by the three large arrows shown on her palaeogeographic map (Figure 5.6).

5.5 Diagenetic phases and paragenetic sequences.

Lithification of the Pendle and Grassington Grit Group sandstones took place via a combination of compaction and cementation. This resulted in the complete destruction of all primary macroporosity. In general, the paragenetic sequences seen in the sandstones are similar, although the importance of individual diagenetic processes is dependent on the initial amount of allogenic matrix. Hence, the muddy sandstones and greywackes of the Pendle Grit Formation have behaved differently during diagenesis from the cleaner sandstones in the two Groups. The paragenetic sequences of these two facies types are illustrated in Figures 5.7 and 5.8. Individual diagenetic modifications are discussed in approximate time sequence below.



*Some sandstones only: see text.

Figure 5.7: Generalised paragenetic sequence for matrix-poor sandstones of the Pendle and Grassington Grit Groups

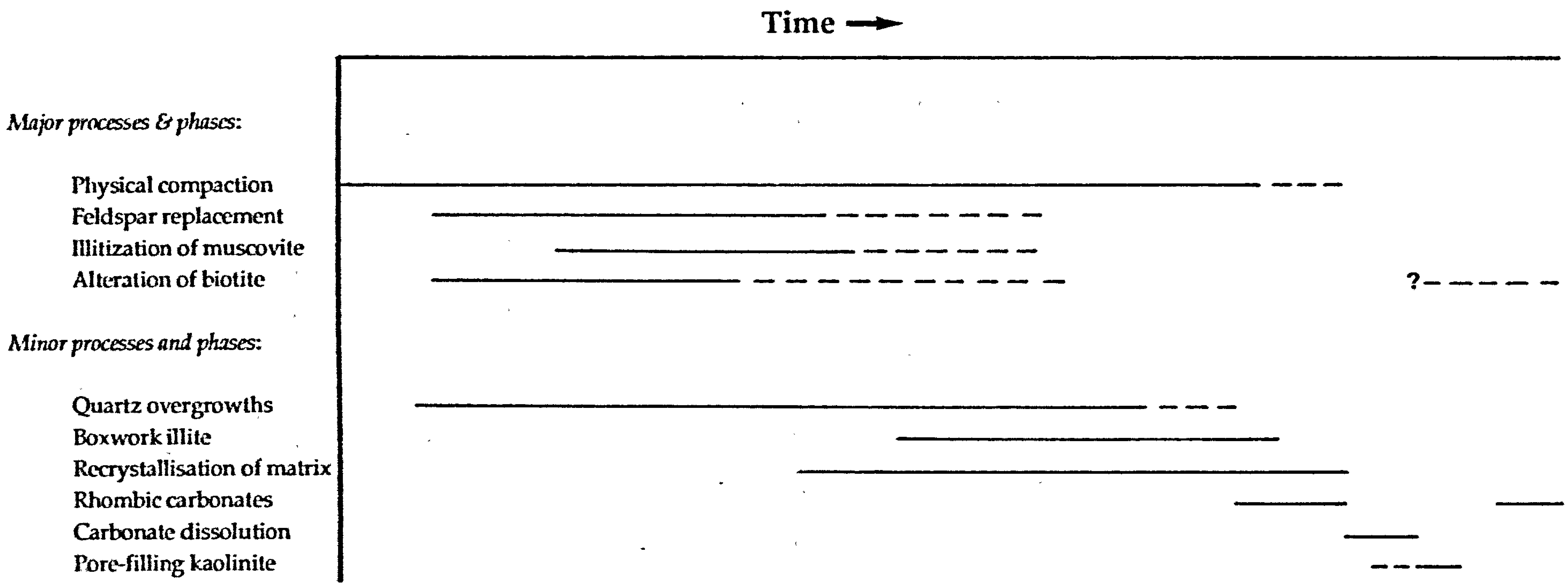


Figure 5.8: Generalised paragenetic sequence for matrix-rich sandstones of the Pendle and Grassington Grit Groups

5.5.1 Compaction.

This is the most significant porosity reducing mechanism in the muddy facies. Severe compaction is evidenced by dislocation or bending of mica fragments (Plates 11c and 14d) and by the production of pseudomatrix from mudrock fragments (Plate 12c). Clay fabrics in the allogenic matrix suggest that there has been expulsion of clays from between larger grains during compaction. In the coarser sandstones compaction was a less important porosity reducting process. Minor grain bending and brittle fracturing of some feldspar grains are the only signs of mechanical compaction in these cleaner facies.

Analysis of the original grain/grain contacts in these sediments is made difficult by quartz authigenesis (see below). However, cathodoluminescence studies produced no evidence for significant pressure solution (chemical compaction) between grains: most of the original grain boundaries are point or tangential contacts.

5.5.2 Quartz diagenesis.

Authigenic quartz is present in all the samples but is much more abundant in the cleaner facies where it is the major porosity reducing phase. It forms either as syntaxial overgrowths on clastic quartz grains or, more rarely, as small pore-filling aggregates of bipyramidal crystals. The latter are usually less than 30µm long and 5µm in width and are optically unresolvable. The syntaxial overgrowths, however, result in obvious compromise boundaries in thin-section (e.g. Plates 11a, 12a 12b). "Dust rims" outlining the original grains are sometimes present (e.g. Plate 12d) but, rather than being grain coats, are usually lines of fluid inclusions. These inclusions form due to multiple nucleation of quartz on the original grain and the growth of many euhedral crystallites away from the grain surface (Plate 14a). Continued growth of these syntaxial crystallites causes them to coalesce into a single large overgrowth but, in the process, small cavities are left close to the original grain surface (Plate 14b). This process is enhanced where there is an early patchy grain-coating cement (Plate 14b).

Cathodoluminescence work suggests that authigenic quartz may account for up to 10-15% of the whole rock in the cleaner facies. The source of the silica required for this cement is unclear. No mass balance/volume calculations have been attempted, but it seems unlikely that the feldspar alteration reactions discussed below can release sufficient excess silica to account for the authigenic quartz. As mentioned in 5.5.1, there is no evidence that pressure solution of fabric grains has contributed significantly to the silica supply. This implies an external source of silica and that the diagenetic system was open. More work is required on this aspect of the diagenesis of the E_{1c} sandstones.

Further examples of the syntaxial quartz overgrowths are shown on Plates 14c, 16, 17 and 18.

5.5.3 Feldspar diagenesis.

In the Pendle and Grassington Grit Group sandstones feldspar is present in the following states:

1) Relatively fresh potassium feldspar and minor fresh oligioclase (see 5.3.2).
2) Skeletal grains of potassium feldspar, some with albite overgrowths (Plate 13c).
3) Euhedral authigenic albite in porespaces and intergrown with authigenic quartz (Plates 13a, 14d, 15a and 15c).
4) Potassium feldspar partially replaced by illitic and/or kaolinitic clay minerals (Plates 13a, 13b and 12c).

In addition, large patches of kaolinite and kaolinite+illite+muscovite appear to pseudomorph feldspar (Plate 12b). The latter patches are always more illitic at the margins and have kaolinite dominated centres.

Timing of the various stages of alteration is difficult to ascertain because of confusing textural relationships. However, the following sequence of diagenetic processes seems to fit the observations most closely:

1) Initial replacement of some of the second-cycle feldspars produces an illite rich reaction rim.
2) Continuing dissolution/replacement reactions under conditions of lower K^+ ion activity result in the rest of the feldspar being converted into a mixture of kaolinite and minor illite (c.f. Warren, 1987).
3) Conditions change such that potassium feldspar dissolution occurs without the formation of new clay minerals (i.e. pore fluids undersaturated with respect to both feldspars and clay minerals).

Boxwork feldspars and minor secondary porosity were produced at this time.

4) Dissolution halts. Albite becomes the stable feldspar phase and grows on the preexisting framework grains and as new euhedral crystals within porespace.

These observations show how complex the feldspar diagensis has been in these rocks. Equally, they suggest that the diagenetic processes have been extremely selective: the fresh feldspars have survived intact through all the above processes. The feldspars which have been affected are those that have been referred to as second cycle feldspars in previous Sections. Possibly, weathering on the surfaces and along the cleavage planes of these feldspars grains may have made them more vulnerable to diagenetic alteration during subsequent burial. The geochemistry of this situation is particularly interesting and requires detailed investigation.

5.5.4 Mica alteration.

The alteration of biotite to hydrobiotite and iron oxy-hydroxides appears to have begun very early as partially altered hydromicas with splayed ends are preserved within quartz overgrowths (Plate 12c). However, further oxidation probably continues at the surface today due to weathering.

Muscovite transformation to illite is seen in many samples from both the muddy and clean sandstones of the Pendle and Grassington Grit Groups. In thin-section muscovite with splayed, illitised ends is common and marginal birefringence decrease in muscovite grains is ubiquitous. SEM samples reveal almost total illitisation of some muscovite (Plate 15d and 19c). All the muscovite alteration appears synchronous with the early phases of feldspar replacement.

5.5.5 Clay mineral diagenesis (excepting pore filling Kaolinite).

Clay mineral authigenesis directly related to feldspar dissolution/replacement has been discussed in sub-section 5.5.3. Here the development of other independent clay minerals is reviewed, with the exception of late pore-filling kaolinite which is considered in sub-section 5.5.7.

Clay mineral authigenesis is an important aspect of the diagenesis of both the muddy and clean facies types. However, while authigenic clays are common in the cleaner facies they are more sparse in the matrix-rich rocks. The authigenic clays seen are as follows:

1) Early, amorphous grain-coating clays (Plate 14b and 14c)
2) Pore-lining clays with a regular or irregular, crenulated boxwork morphology (Plates 16 and 17).
3) Pore-bridging clays with a fibrous or lath-like morphology, usually overgrowing or growing from boxwork clays (Plate 17).
4) Pore-filling boxwork clays (Plate 15)

These clays are all highly birefringent and qualitative chemical analysis on the SEM (EDAX) shows that they all have an illitic composition. The clay morphologies are all consistent with a true illite composition rather than a mixed-layer smectitic composition (Macchi, 1987). The identity of these clays has been confirmed by XRD analysis of the sub-2μm and sub-0.5μm grainsize fractions of a sample from the Pendle Grit Group (See Figure 5.1 for location and Figures 5.9, 5.10 and 5.11 for results). These size fractions were used as all of the morphological types described above are represented while larger sized pore-filling kaolinite does not significantly dilute the sample.

In the muddier facies all the above clay morphologies are present but are confined to the limited primary porosity preserved after compaction. Some recrystallization of the allogenic kaolinite has occurred in these facies: marginal illitisation of the kaolinite is seen, mirroring the diagenetic changes found in the shale sample collected for XRD analysis. Also, the crystallinity of the allogenic kaolinite in the sandstones seems higher than that in the shale (Compare Plates 15a and 19b).

5.5.6 Carbonate diagenesis.

There are two phases of carbonate growth in the sandstones of the Pendle and Grassington Grit Groups, separated by a phase of dissolution and secondary porosity production. The early phase of carbonate growth is difficult to quantify as much of the carbonate prepicitated was removed during the subsequent dissolution phase. The only evidence of its existence in most rocks is the presence of oversized angular pores (Plate 12c and 12d) and occasional large rhombic pores (Plate 12c). Both pore types are always filled by kaolinite (See below).

Figures 5.9, 5.10 and 5.11.

X-ray diffractograms of clays separated from Pendle Grit Formation sample A (See Table 3 for details). Figures 5.10 and 5.11 are for sub-2μm and sub-0.5μm grainsize fractions respectively. Figure 5.12 is a slow, high-accuracy scan of the low angle reflections from the sub-0.5μm fraction. Peaks are as follows:

I	Illite (001).
I_1	Illite (002).
K	Kaolinite (001), destroyed on heating.
K_1	Kaolinite (002), destroyed on heating.
Q	Quartz (101).
Q_1	Quartz (100)
V	?Hydrobiotite: Vermiculite (001)/Biotite (001).

Note that there is a well-defined illite peak, particularly in the sub-0.5μm grainsize fraction. Also, kaolinite is much less abundant in this fraction. This is compatible with the size differences between kaolinite and illite seen in S.E.M. samples.

A swelling component in the clays in this sample is shown by the movement of peak V towards lower angles (larger lattice spacings) on glycolation and its collapse on heating show. This component is interpreted to be due to hydrobiotite in the sample rather than due to smectitic clays (for which reflections would be expected at around 5.9° and 5.2° in the untreated and glycolated states respectively). Hydrobiotite consists of interlayered vermiculite and biotite lattices and the former cause the swelling and collapse of the lattice during glycolation and heating respectively. This interpretation is consistent with the bulk petrography of the rock (hydrobiotite content=1.5%) and the complete lack of clays of smectitic composition revealed by electron microprobe work.

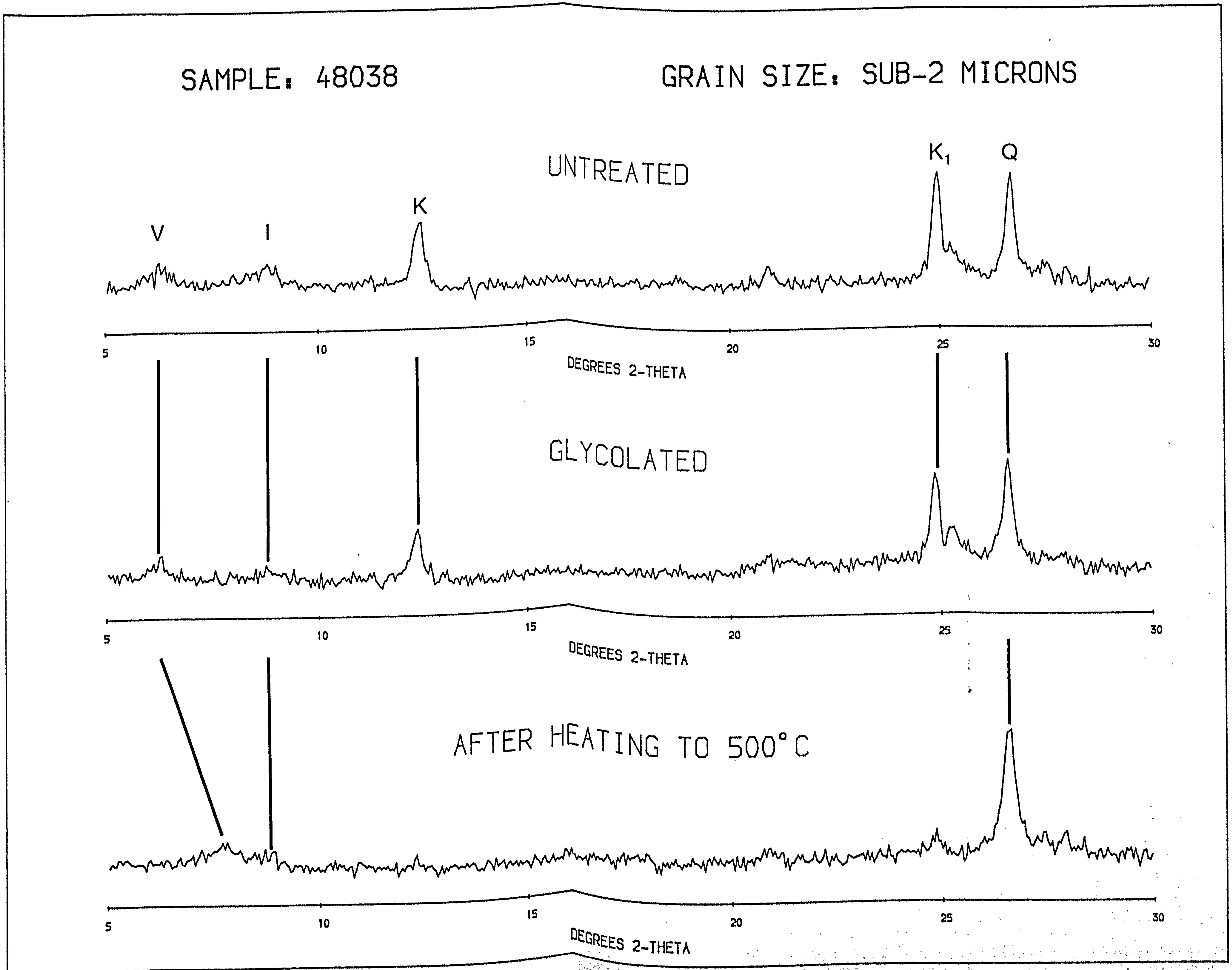

SAMPLE: 48038
GRAIN SIZE: SUB-2 MICRONS
UNTREATED
V
I
K
K_1
Q
5
10
15
20
25
30
DEGREES 2-THETA
GLYCOLATED
5
10
15
20
25
30
DEGREES 2-THETA
AFTER HEATING TO 500°C
5
10
15
20
25
30
DEGREES 2-THETA

Figure 5.9

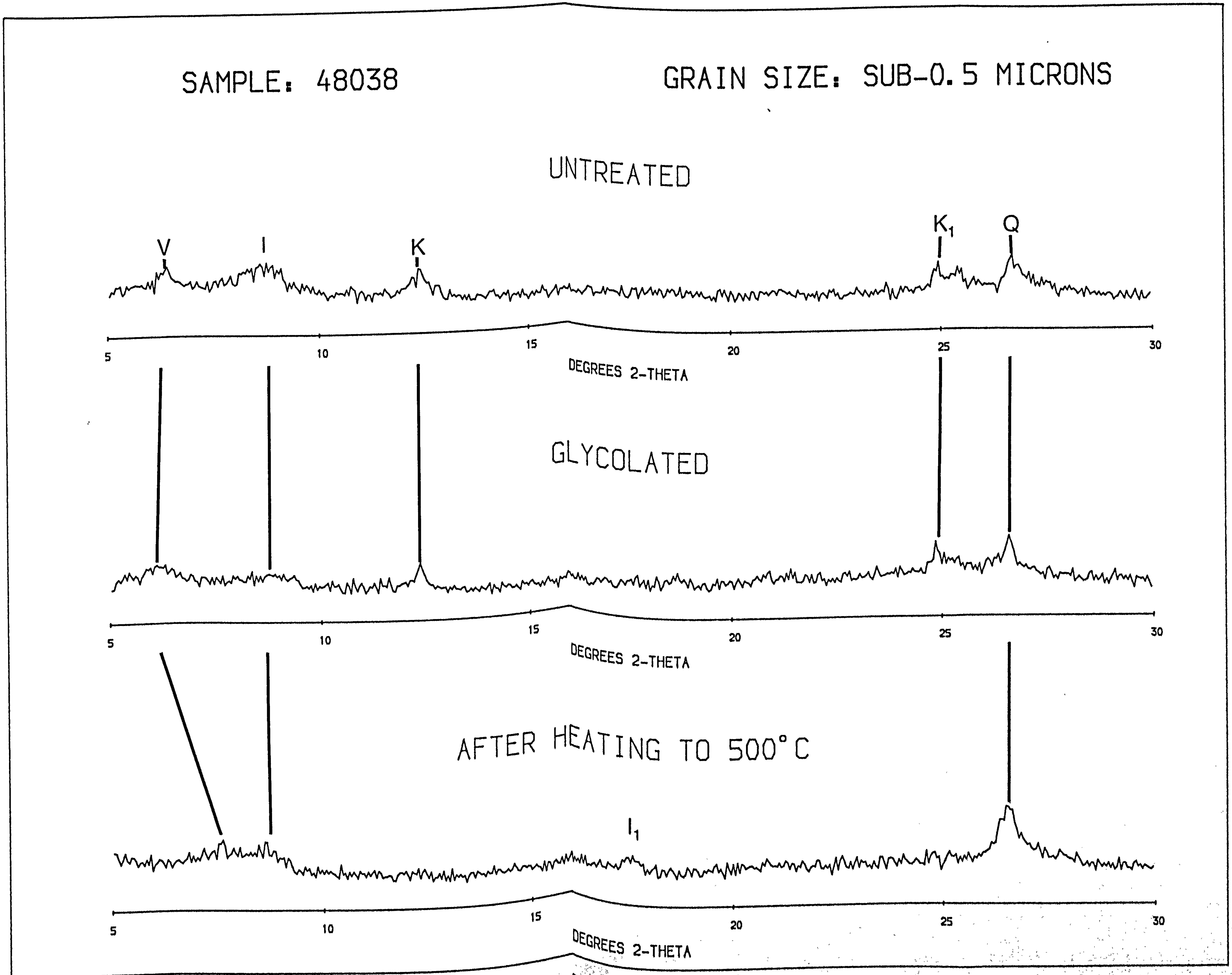

SAMPLE: 48038
GRAIN SIZE: SUB-0.5 MICRONS
UNTREATED
V
I
K
K_1
Q
5
10
15
20
25
30
DEGREES 2-THETA
GLYCOLATED
5
10
15
20
25
30
DEGREES 2-THETA
AFTER HEATING TO 500° C
I_1
5
10
15
20
25
30
DEGREES 2-THETA

Figure 5.10

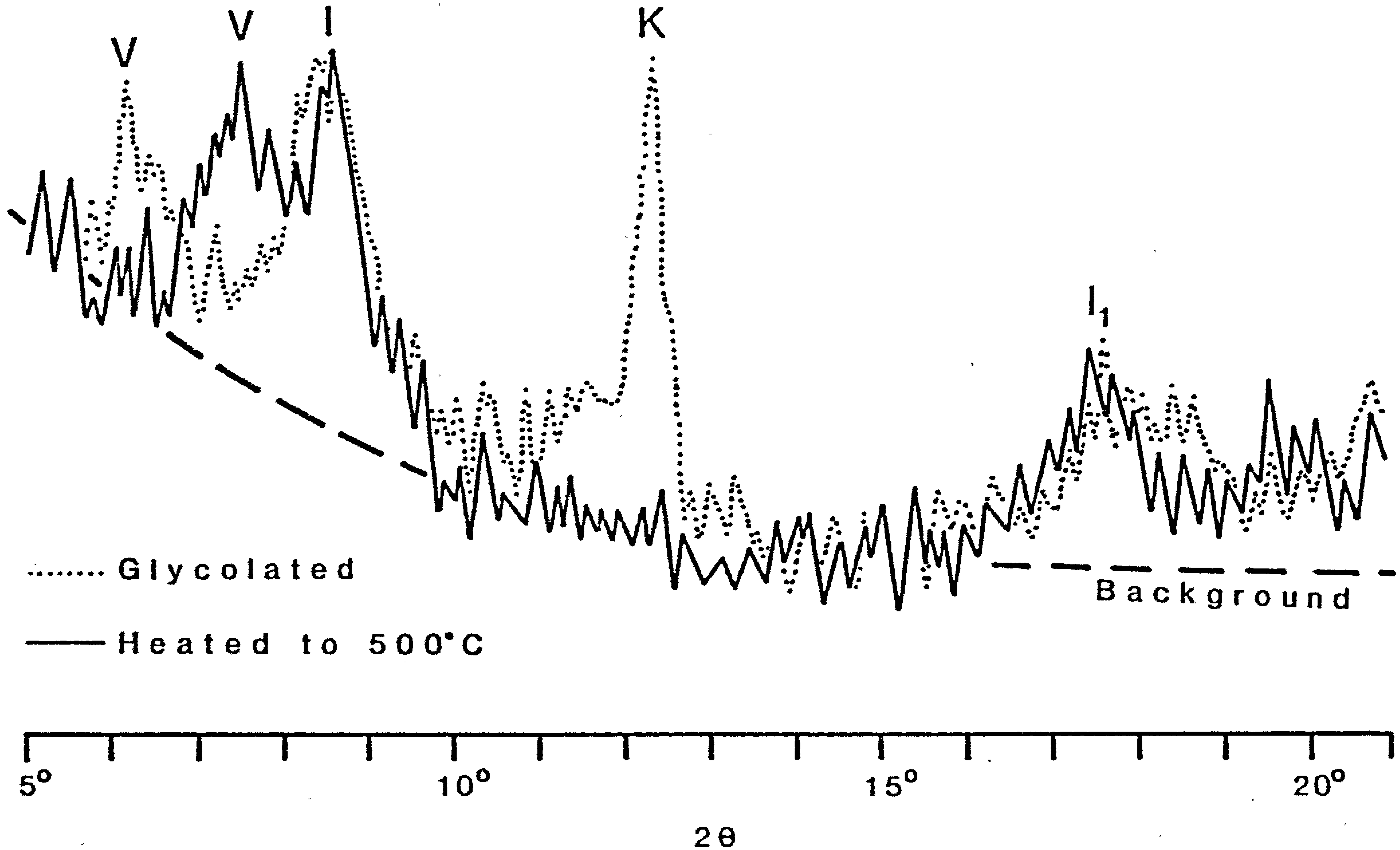


Figure 5.11

These textural relationships show that the carbonate growth post-dates the lithification of the rock (as the secondary pores have not collapsed) but was prior to major kaolinite authigenesis. The carbonate may have been supplied by expulsion of carbonate rich fluids from mudrocks in the Bowland Shales during deep burial. The change to more acidic porewaters indicated by the dissolution of the early carbonate phase probably relates to end-Westphalian flushing by meteoric water, as discussed in section 5.6.

Diagenetic carbonate within the E_{1c} sandstones is present in three forms:

1) Euhedral rhombs of ferroan calcite or ferroan dolomite growing replacively outwards from nucleation sites in the microporosity around clay minerals (Plates 12c, 12d, 14c, 18a and 18c). Individual rhombs rarely exceed 100μm in size but may coalesce to form larger patches of cement. Some rhombs are zoned, with ferroan calcite cores and ferroan dolomite rims (Plate 18a). It is possible that the calcite cores represent remnant early carbonate subsequently overgrown by dolomite during the second carbonate growth phase. This is the most common type of diagenetic carbonate and is ubiquitous in subsurface samples. It is rarely seen in samples from outcrop, probably due to recent weathering.
2) Pervasive carbonate cements, mainly of ferroan dolomite (Plates 11c, 13c and 18b). The fluids from which these cements were derived were highly corrosive and etch all the silicate phases in the rock. Clearly, they were severely undersaturated with respect to silica. Pervasive cements are rare and seem to be confined to shale-encased sandstone bodies near to major faults (e.g. those exposed at Jenny Gill Quarry, Figure 3.16, and those in Roosecote sample R2). This relationship may indicate that the cements are related to physical trapping of porewaters migrating from depth along faults.
3) Nodular carbonate concretions (Plate 1), mainly of ferroan dolomite and calcite. The diagenetic history of these nodules has been fully described by Batupe (1987).

All these carbonate types appear to be post-kaolinite in age although there is little textural evidence to confirm this in the case of the carbonate concretions.

5.5.7 Pore-filling and replacive kaolinite.

Large booklets of kaolinite upto 100µm long and 20µm in diameter are found growing in all the secondary porosity generated during the carbonate dissolution phase discussed above (Plates 12a, 12c, 12d, 16a, 16c, 17 and 18c). Any remaining primary porosity was also plugged at this time. The kaolinite crystals within the booklets are very well formed. This morphology, and their larger size, serve to distinguish them from the allogenic kaolinite and that formed by the alteration of feldspar. XRD data (Figures 5.9 to 5.11) confirm this phase is true kaolinite and not dickite.

In one sample from Roosecote borehole (R4), kaolinite forms 25.3% of the total rock. In this case the kaolinite appears to be replacive of all other phases (Plates 11d, 18d and 19a). This may possibly result from the replacement of a pervasive early carbonate cement but there is no textural evidence for this. Instead, it appears that the kaolinite precipitating fluid was the corrosive agent. This is discussed more fully below.

5.6 Timing of diagenetic processes.

The paragenetic sequences of Figures 5.7 and 5.8 represent generalised sequences developed from all the samples studied. It has been possible, however, to constrain the absolute timing of some of the diagenetic stages using burial history reconstructions for Roosecote and Whitmoor boreholes. Figures 5.12 to 5.14 show the reconstructed burial histories of these wells and of the outcrops along the Pendle Monocline. All the curves show an initial phase of very rapid burial followed by equally rapid uplift. However, the magnitude of the burial and uplift varies between the three localities. As a result of this the paragenetic sequences in the three areas are subtly different. The pertinent differences are as follows:

Roosecote

Sample R4, from only 40m beneath the Permian unconformity in the well, is the only sample with replacive kaolinite. During the time of maximum uplift at this location, the sandstone involved was almost certainly connected updip with the Permian surface. Sandstones at some 330m beneath the Permian unconformity (Samples R1, R3 and R5) have only minor amounts of pore-filling kaolinite. Illite

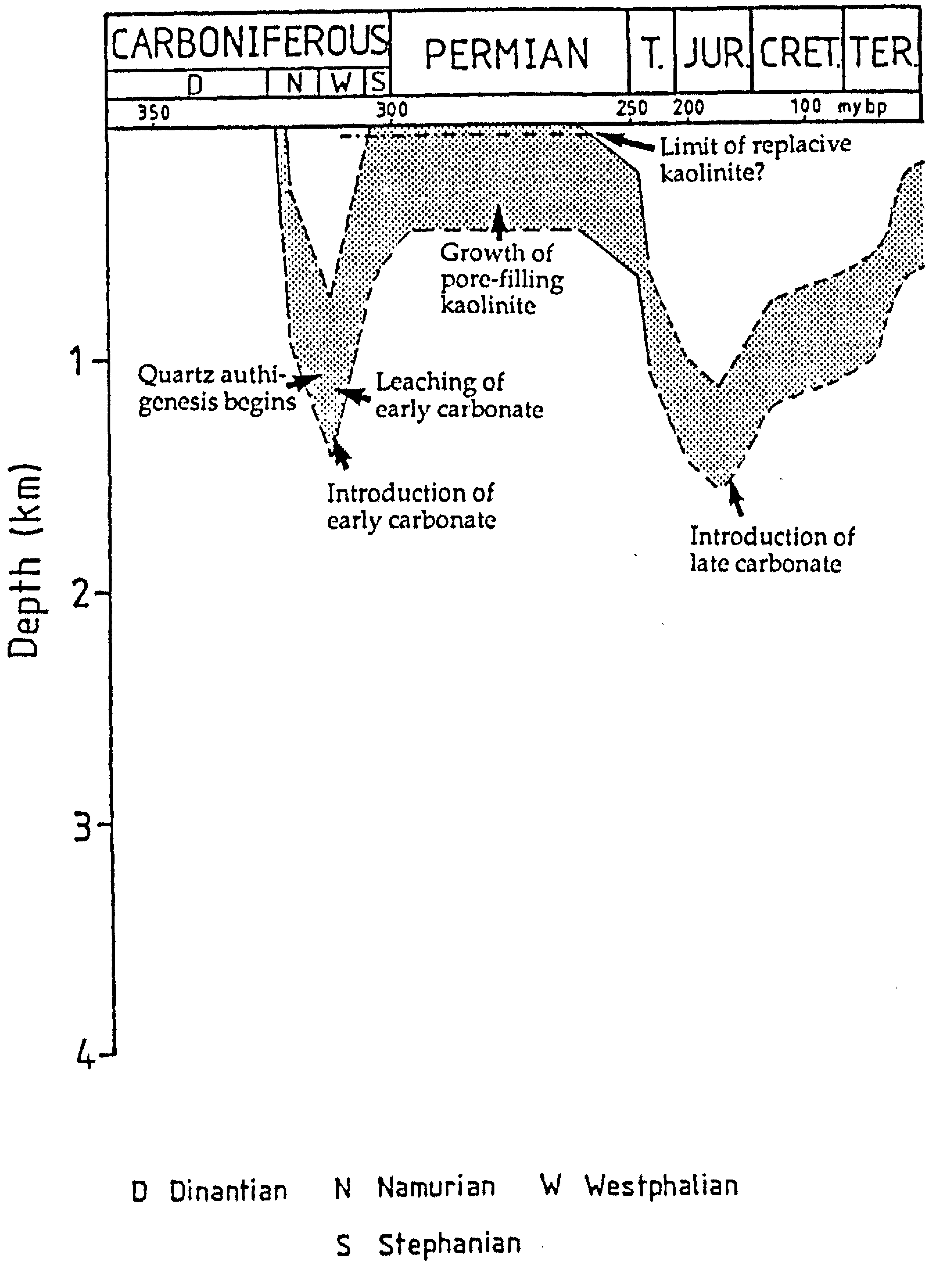


Figure 5.12: Burial history curve for E_{1c} sediments in Roosecote borehole (See Figure 5.1 for location). Solid lines indicate data derived from the borehole, dashed lines indicate where the burial history has been modelled using geological data from adjacent areas. Sections have not been decompacted.
Note particularly:

1) Initial burial was limited.
2) Uplift exposed the E_{1c} section at the surface during the Permian.

The interpreted timing of the diagenetic phases seen in this borehole are shown.

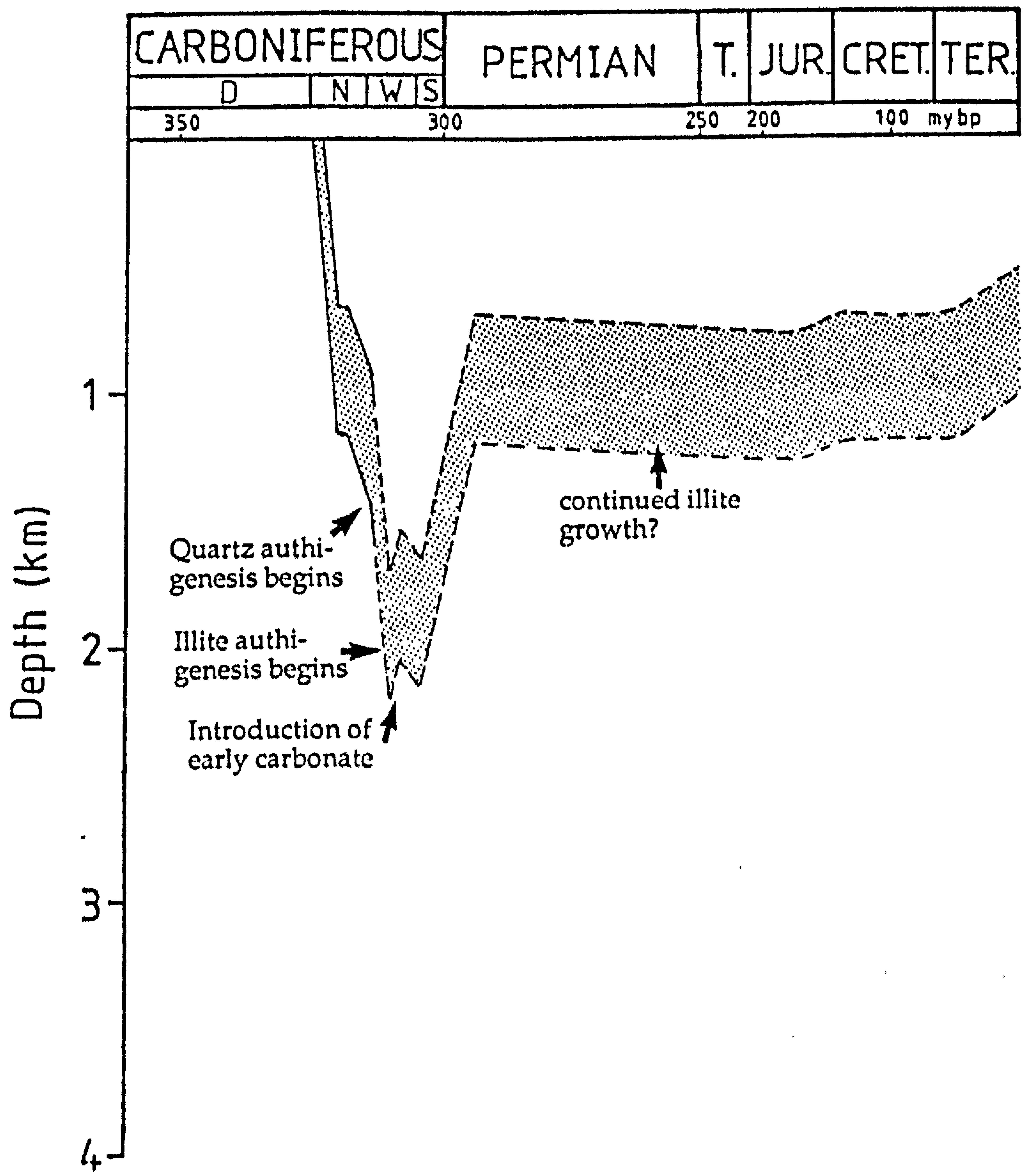


Figure 5.13: Burial history curve for E_{1c} sediments in Whitmoor-1 borehole (See Figure 5.1 for location). Solid lines indicate data derived from the borehole, dashed lines indicate where the burial history has been modelled using geological data the Ingleton Coalfield (Ford, 1954) and from adjacent areas. Sections have not been decompacted. The Mesozoic to Recent burial history is very uncertain. However, it is clear from comparing maturity model data with vitrinite reflectance data (reported by Lawrence *et al.*, 1987) that the E_{1c} section has never been buried to depths greater than those attained during the Westphalian.

Note particularly:

1) Moderate initial burial.
2) E_{1c} section has never returned to surface.

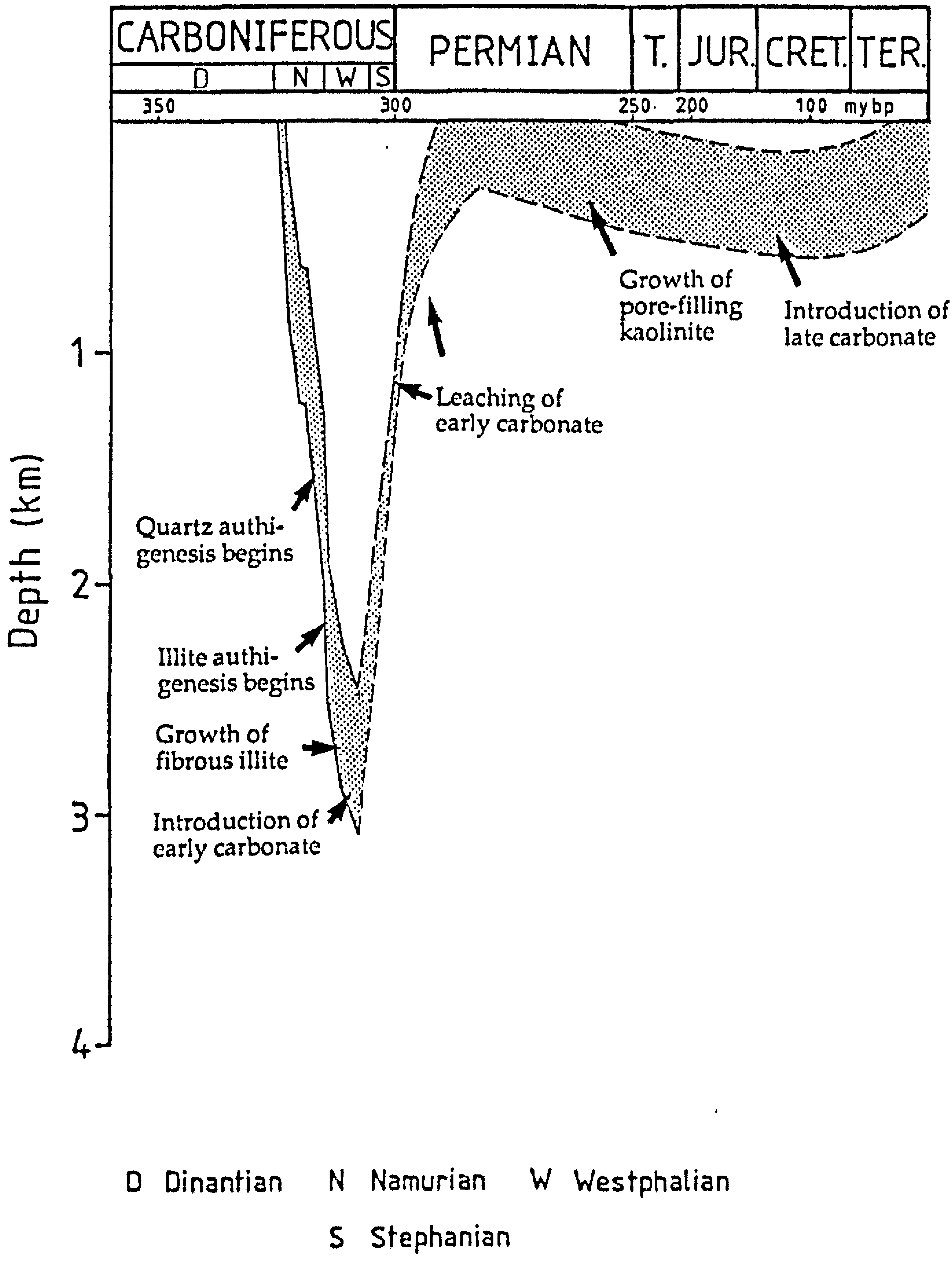


Figure 5.14: Burial history curve for E_{1c} sediments along the Pendle Monocline. Again, solid lines indicate data derived from true stratigraphic thicknesses measured in the immediate area (data from Earp *et al.*, 1961). Dashed lines indicate where the burial history has been modelled. Sections have not been decompacted.

Note particularly:

1) Deep burial during the Namurian and Westphalian.
2) Uplift to surface during the late Carboniferous.

is only a minor authigenic phase and fibrous illite is never seen. Note that sandstones in this well have been buried to a shallower depth than those discussed below.

Whitmoor

SEM samples from fine-grained Pendle Grit Formation sandstones at 629m, 726m and 765m all show well developed pore-filling illite. No kaolinite was recorded although this might reflect the limited number and size of the cuttings samples studied. Significantly, the sandstones in this well were never closer than 629m to the Permian surface and have remained at depth longer than the sandstones studied elsewhere (compare Figures 5.12, 5.13 and 5.14).

Pendle Monocline

These samples contain illite and pore-filling kaolinite in significant proportions. Fibrous illite is well-developed (e.g. Plate 17). Note that the sandstones in this area have been buried deeper than in other areas but returned to the surface during the Permian.

The differences in the paragenetic sequences outlined above clearly point to the following conclusions:

1) Compaction, quartz diagenesis and feldspar dissolution/replacement began during early burial and continued throughout the initial subsidence phase. Illite authigenesis, particularly of the fibrous variety, probably occurred during deeper burial: these clays are best developed in the more deeply buried samples (i.e. the Monocline samples) but are rare in the less deeply buried sandstones (i.e. Roosecote samples). The early phase of carbonate growth probably took place late during this subsidence phase.
2) During uplift the pore water chemistry changed dramatically due to influx of meteoric water from the Permian surface. The change in pore water (probably to more acidic, low K^+ ion activity solutions: Warren, 1987) resulted in a phase of secondary porosity production through dissolution of early carbonate and, probably, feldspar. This porosity was immediately filled by kaolinite. Sandstones closest to the surface were most affected by this process and ?soil-forming processes resulted in extreme kaolinitisation of the rock (e.g. Roosecote, sample R4). Sandstones which remained deeply buried and which were not

affected by the meteoric water influx did not develop secondary porosity or kaolinite (e.g. Whitmoor samples).

3) During later reburial illite continued to grow in the sandstones which had remained deeply buried (e.g. Whitmoor) while clay authigenesis seems to have stopped at shallower levels. Carbonate rich fluids migrating up-dip through the sandstones or along fault-planes may have been responsible for the late carbonate diagenesis seen in some samples.

The interpreted relationships between the burial histories and the authigenic phases at the locations considered are shown on Figures 5.12 to 5.14.

5.7 Discussion.

If the interpretation of the absolute timing of diagenetic events discussed above is correct, then a predictive model for the reservoir quality of the E_{1c} sandstones is suggested. The paragenetic sequences developed in any sample seem to depend on the depth of burial prior to the Hercynian inversion and whether or not this inversion was sufficient to allow penetration of Permian groundwaters into the sediments. Only those samples initially buried to significant depths develop illite and, particularly, fibrous illite. Similarly, only those subjected to meteoric flushing during the Permian have significant pore-filling kaolinite cements. In theory, if the burial history of a sub-surface representative of the Pendle and Grassington Grit Groups can be estimated then this simple model will predict the diagenetic assemblage present in that sandstone.

Unfortunately, the data collected so far suggest that macroporosity in these sandstones will always be poor regardless of the burial history. Similarly, permeability reduction caused by pore-bridging clays will always be a problem. However, there are certain scenarios in which reservoir quality might be enhanced:

1) Leaching of late carbonate cement at burial depths greater than those studied.
2) Early hydrocarbon migration preventing significant clay mineral authigenesis (c.f. Hawkins, 1978, for an example from Carboniferous sandstones similar to those discussed herein).

3) Production of secondary porosity during the uplift phase but no significant kaolinite authigenesis. This would require that the leaching front and the limit of kaolinite authigenesis extended to different depths within sandstones affected by meteoric water influx. There is no evidence from the samples studied that this is a realistic possibility.

There is scope to extend the diagenetic study of the E_{1c} sandstones into the subsurface south of the Pendle Monocline, using data from existing boreholes (c.f. Figure 5.1). This would allow Scenario (1) above to tested in an area which has remained more deeply buried than that around Whitmoor. Testing the other scenarios will require further drilling. It is the current author's opinion that all of these scenarios are extremely risky models upon which to base future drilling for Pendle and Grassington Grit Group targets in the north of England.

5.8 Conclusions.

1) The sandstones within the Pendle and Grassington Grit Groups were originally arkoses, sub-arkoses and arkosic wackes.
2) The only obvious petrographic differences between the sandstones of the two Groups are that those of the Pendle Grit Group are more poorly sorted and have higher allogenic mud components. When this factor is removed. there is no evidence for any variation in provenance between the two Groups.
3) The principal framework constituents are quartz and potassium feldspar grains which appear to have been sourced (ultimately) from plutonic acid-igneous rocks and granite gneisses.
4) There are probably two suites of feldspars amongst the framework grains. Fresh, unaltered feldspars are interpreted as first-cycle erosion products derived from Caledonide granites. A heavily altered feldspar suite of similar composition to the fresh suite may be a second-cycle erosion product.
5) The petrographic data is consistent with the zircon age data of Drewery (1987).
6) All the petrographic data points to a generally northern provenance. The major source areas were probably the cratonic areas of the Laurentian and Baltic Shields together with an uplifted Caledonian terrain.
7) The dominance of allogenic kaolinite in the muddy sandstones and in associated shales may imply intense tropical weathering in the source area.

8) Diagenesis in the sandstones of both stratigraphic Groups is similar but is influenced by the original facies. Porosity reduction in more muddy sediments was dominated by compaction while quartz and clay authigenesis was predominant in cleaner sandstones. Feldspar alteration and replacement is common to both sediment types.
9) The paragenetic sequence developed in any sandstone is critically dependent on the burial history. Initial rapid burial during the Namurian and Westphalian caused compaction , quartz diagenesis and feldspar replacement. If sandstones were buried sufficiently deeply, illite authigenesis became significant. Uplift and erosion during end-Carboniferous basin inversion event led to secondary porosity production by leaching of carbonate cements. This porosity, together with all remaining porosity, was then filled by kaolinite. Both these stages of diagenesis are thought to relate to flushing by acidic meteoric water during the uplift phase. Beneath the level reached by Permian meteoric waters, the uplift event had little diagenetic effect. However, at these levels authigenic illite growth may have continued during or after uplift.
9) The links between the paragenetic sequences and burial histories discussed above lead to a predictive model of reservoir potential in the Pendle and Grassington Grit Groups. Unfortunately, this model suggests that the reservoir quality of these Groups will usually be very low.

Plates accompanying Chapter 5.

The Plates for Chapter 5 are gathered together in the following pages. Individual plates follow a specific theme and, hence, are not referenced in order through the text. Each sub-plate is separately described and sample numbers and figure references are given thus: Sample 48038 (A). Locations and other data pertaining to these samples can be obtained from Figure 5.1 and Table 3. Abbreviations used in the following descriptions are as follows:

XP	Crossed polars.
XPST	Crossed polars plus sensitive tint plate.
PM	Photomicrograph
SEM	Scaning Electron Microscope.
EM	Electron micrograph.
BSEM	Back-scattered electron micrograph.

Note that all the mineral identifications shown on the BSEMs and the EMs are backed up by X-ray dispersive analysis.

Plate 11:

Principal Sandstone types.

a) Clean, moderately sorted, sub-arkosic sandstone from the Pendle Grit Formation. This rock is strongly quartz cemented and compromise boundaries between quartz grains are well developed . Note the fresh oligioclase (centre) and compare this with the altered orthoclase grain (top centre).
Sample 49230 (Q). XP. Scale bar= 1mm.

b) Very poorly sorted, sub-arkosic wacke ("greywacke") from the Pendle Grit Formation. Note the virtual absence of quartz overgrowths and the significantly higher "matrix" content compared with (a). Again, both fresh (F) and weathered (W) feldspars are present.
Sample 49235 (U). XPST. Scale bar= 1mm.

c) Pervasive carbonate cement in shale encased sandstone from the Pendle Grit Formation. All the other phases within the rock are partially replaced and many have strongly corroded margins. The darker patches of cement (A) are probably the remains of illitised feldspars subsequently replaced carbonate. Note the deformed mica which shows that there was significant compaction prior to carbonate cementation. See also Plate 18b.
Sample 49420 (R2): Roosecote Borehole, 324m. XP. Scale bar= 1mm.

d) Pervasive kaolinite cement in sandstone from the Pendle Grit Formation. As in (c), all the non-cement phases are highly corroded. Some quartz overgrowths are preserved around grain A. See also Plates 18d and 19a.
Sample 49422 (R4): Roosecote Borehole, 199m. XP. Scale bar= 1mm.

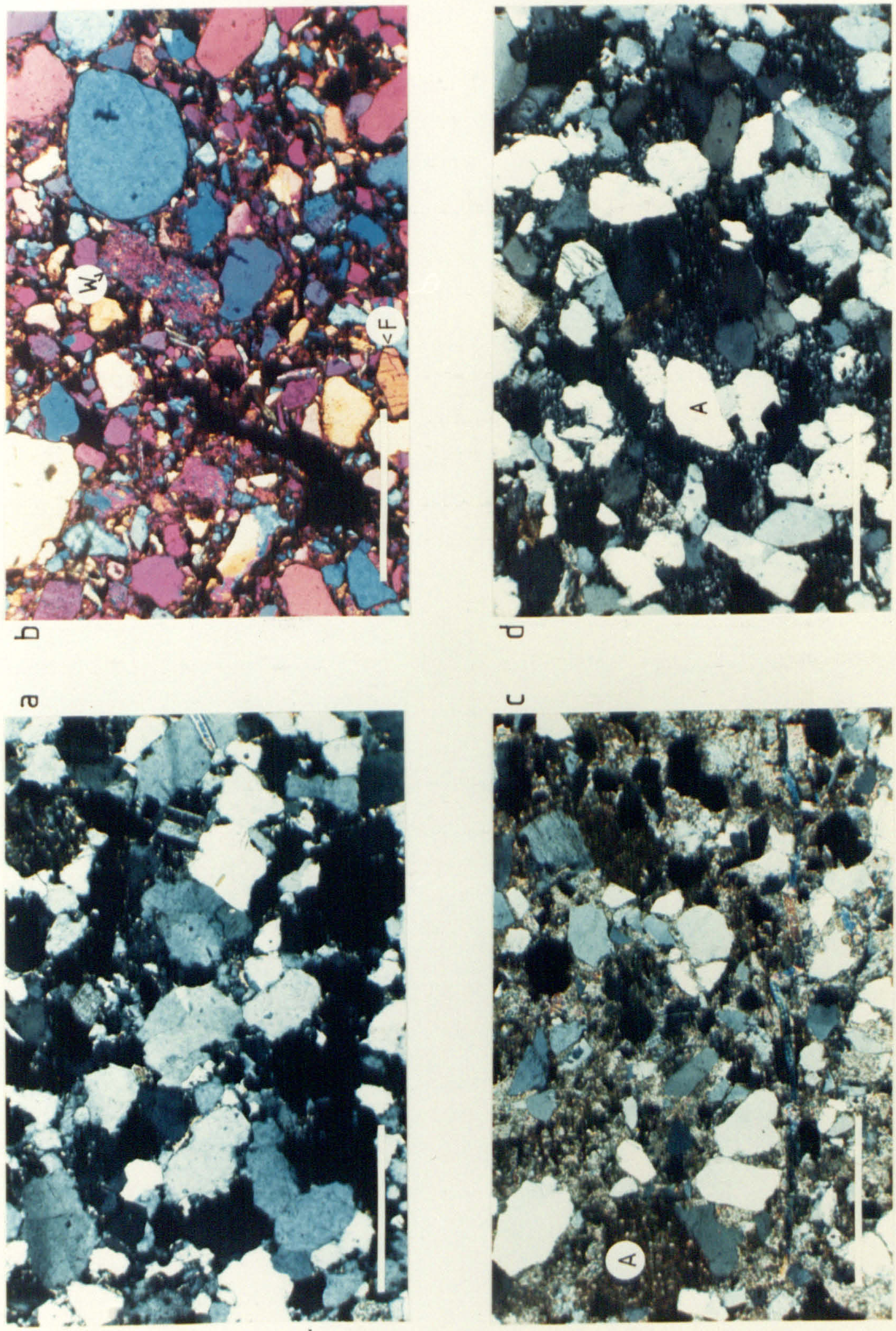


Plate 11

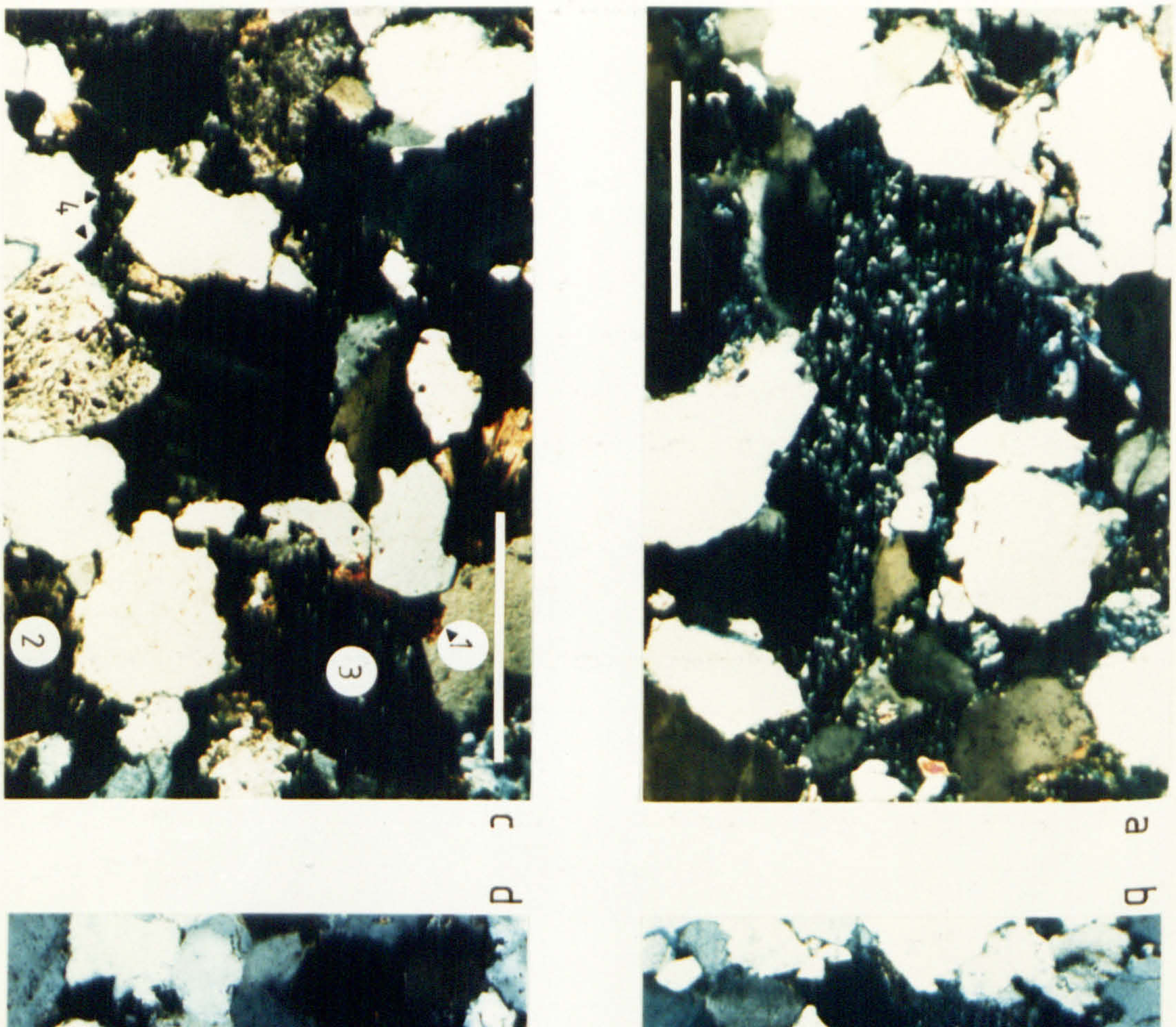


Plate 12

Plate 12

Feldspar and carbonate replacement textures.

a) Oversized angular pore filled with kaolinite. This pore must have formed by mineral dissolution: note how the muscovite grain (top left) has been deformed by compaction around the original mineral. The angularity of the pore margin suggests that the pore was created by dissolution of a rhombic carbonate.
Sample 49042 (3): Grassington Grit Group. XP. Scale bar= 1mm.

b) Photomicrograph showing probable replacement products of feldspar. The kaolinite patch (upper left) appears to be pseudomorphing feldspar. The patch of kaolinite and illite (upper right) is the end product of *in situ* feldspar alteration. Note how the margin of the patch is very illite rich (high birefringence). Note also the relatively fresh microcline (M) and orthoclase (O).
Sample 49218 (H): Pendle Grit Formation. XP. Scale bar= 1mm.

c) Further evidence for two feldspar suites within the E_{1c} sandstones. A fresh microcline (centre) has weathered orthoclase all round. Note also the splayed hydrobiotite overgrown by quartz (1) and the pseudomatrix at (2). The large rhombic pore (3) and the smaller angular pores (4) filled with kaolinite are further evidence for dissolution of a former carbonate cement.
Sample 48260 (iv): Pendle Shale Formation?. XP. Scale bar= 1mm.

d) Kaolinite filing an angular pore, probably after carbonate, and being displaced by later ferroan dolomite. The dolomite has nucleated in the microporosity of the kaolinite. A dust rim between the original quartz grain and its overgrowth is seen at (A).
Sample 49418 (R1): Pendle Grit Formation. XP. Scale bar= 0.5mm.

Plate 13

Back-scattered electron micrographs and photomicrographs of equivalent areas showing details of cement phases.

a) Detail of pore-filling phases. The history of this pore space is complex. Illite fills patches around the margin of the pore (1, 2, 3) and a large angular area towards its centre (4). The shape of this latter area suggests that the remainder of the pore was occupied by a solid phase during illite growth. The illite patches, therefore, probably represent filling of the primary porosity in the rock. The remainder of the pore must be secondary and formed by leaching of the original solid phase (carbonate or feldspar?). This secondary porosity is now filled with kaolinite (K). The bright material on the BSEM is authigenic iron oxide intergrown with authigenic albite (A).
Sample 48038 (A): Pendle Grit Formation. XPST and BSEM. Scale bar= 100μm.

b) Kaolinite (K) filling secondary porosity created by feldspar dissolution. orthoclase has been illitised at its margins (1). Pore-filling illite is present at (2) and expanded hydrobiotite (=biotite/vermiculite interlayers) at (H).
Sample 48038 (A): Pendle Grit Formation. XPST and BSEM. Scale bar= 100μm.

c) Pervasive calcite cement. The dark patch on the PM (K) is kaolinite. Note that the borders of this patch are angular and that they are formed by growth of the carbonate. The carbonate, then, is later than the kaolinite. Grain (O) is an orthoclase grain with marginal albite overgrowths.
Sample 49219 (J): Pendle Grit Formation. XPST and BSEM. Scale bar= 100μm.

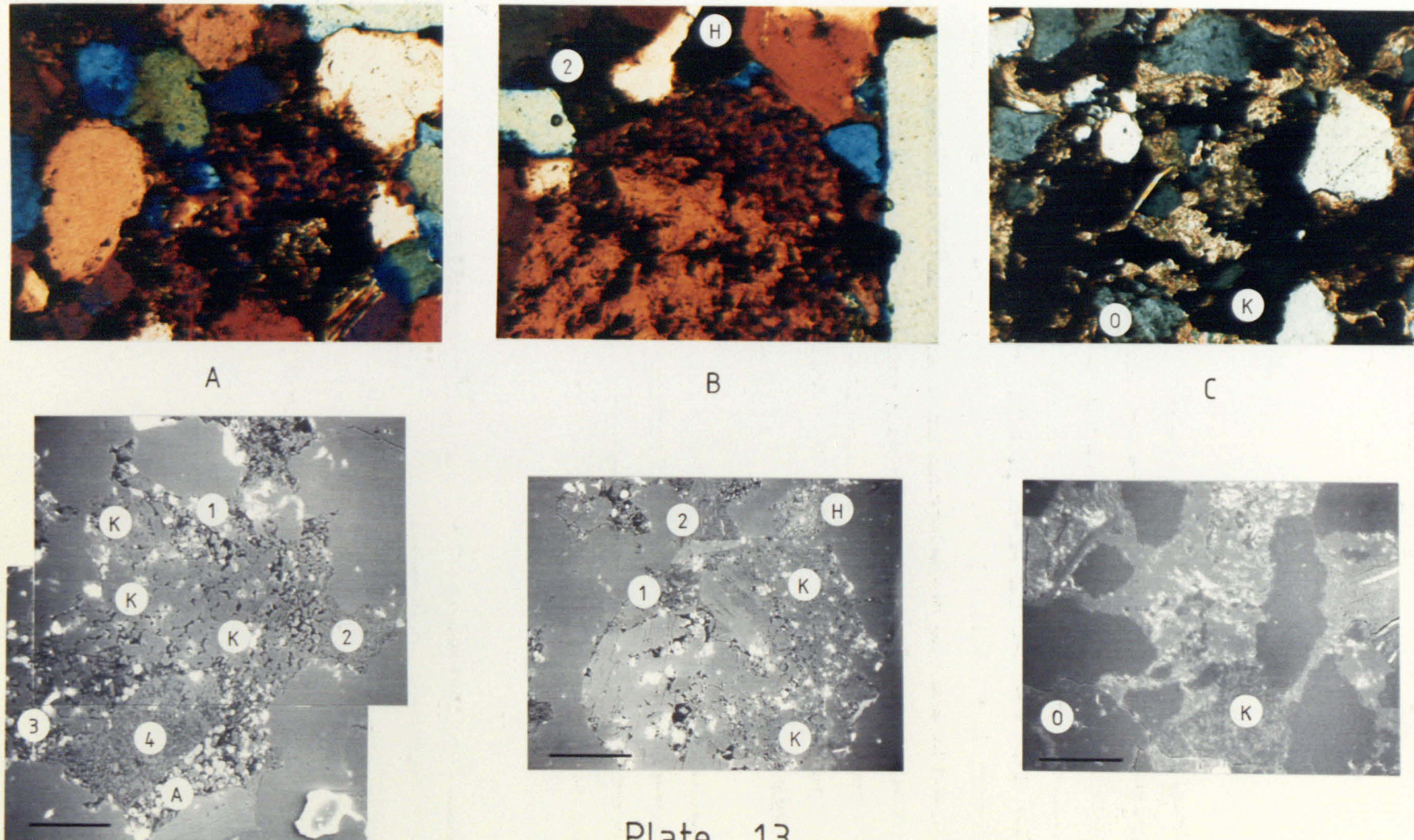


Plate 13

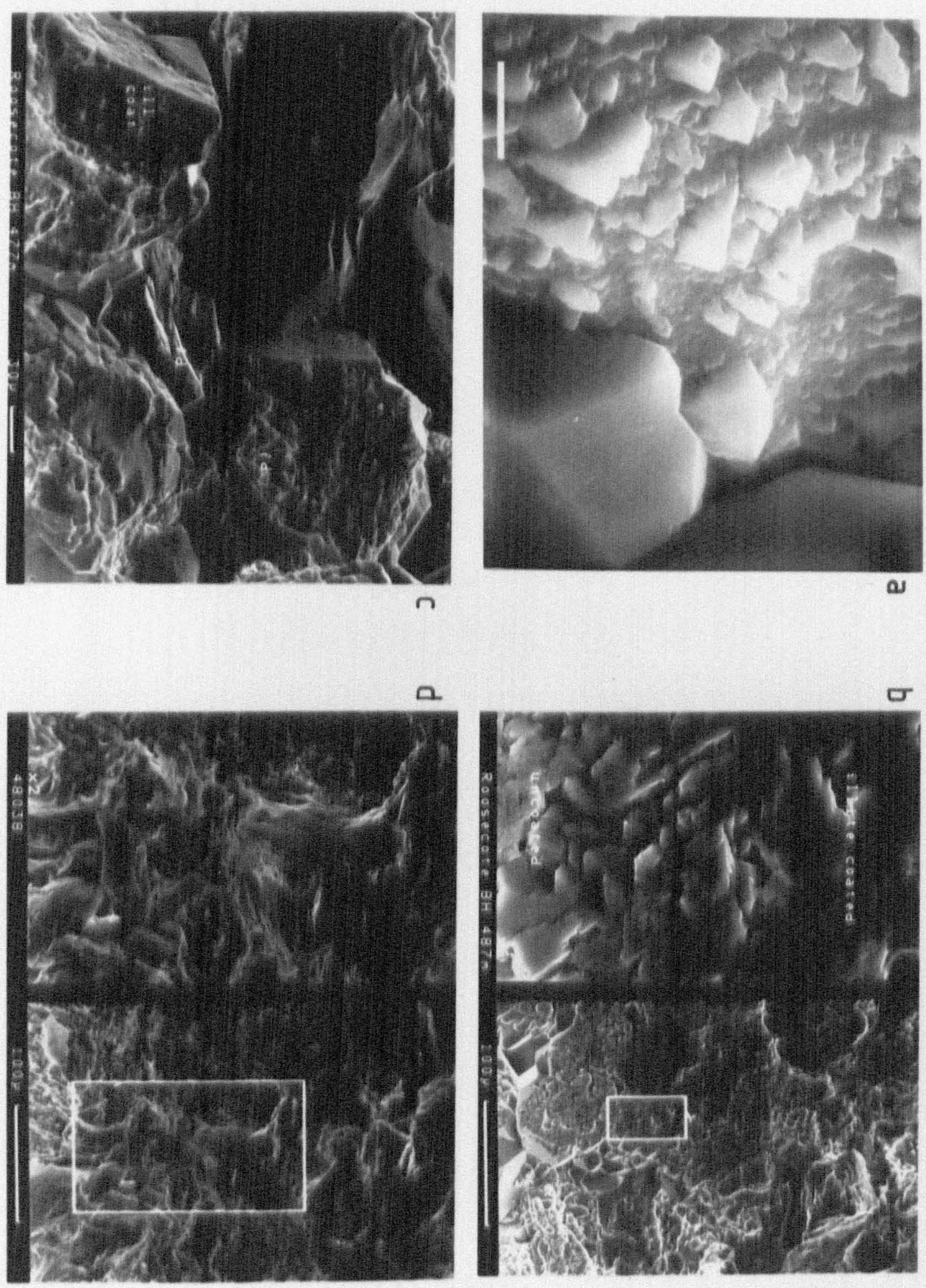


Plate 14

Plate 14

Electron micrographs of authigenic quartz and framework feldspars.

a) Microprismatic authigenic quartz. This is the first stage in the growth of large, syntaxial overgrowths such as those in (c) below. In the process of growth the individual crystallites merge but may leave small fluid inclusions trapped close to the grain boundary. This may result in a "dust rim" outlining the original grain in thin-section.
Sample 48258 (F): Pendle Grit Formation. EM. Scale bar= 5μm.

b) This sample has broken along a grain contact during preparation. The texture seen results from quartz overgrowths similar to those in (a) which have grown against the removed grain. Bypyramidal quartz, seen in the background has also grown against this boundary. An amorphous grain-coating illite from the removed grain appears to have been left on the quartz overgrowths.
Sample 49419 (R1): Pendle Grit Formation. EM. Enlargement is x5.

c) Well-developed syntaxial quartz overgrowths filling the bulk of the primary porosity in this sandstone. Ferroan dolomite (fd) has nucleated in the remaining porosity and grown at the expense of the overgrowths. Note the etching of the overgrowth at top right.
Sample 49423 (R5): Pendle Grit Formation. EM.

d) Matrix of micro-muscovite deformed around a framework feldspar. The muscovite is partially illitised.
Sample 48038 (A): Pendle Grit Formation. EM.

Plate 15

Scanning electron micrographs from Whitmoor borehole.

a) Authigenic albite intergrown with bipyramidal quartz. The remainder of the pore is filed with poorly crystalline kaolinite (probably allogenic) and with boxwork illite.
Whitmoor Borehole, 765m. Pendle Grit Formation. EM.

b) General view (right) of sandstone showing quartz grains with well-developed authigenic overgrowths surrounded by pore-filling illite. The enlargement (x5) shows the delicate boxwork form of this illite.
Whitmoor Borehole, 765m. Pendle Grit Formation. EM.

c) Intimately intergrown authigenic quartz and authigenic albite. It appears from this image that these phases were growing simultaneously and competing for porespace.
Whitmoor Borehole, 629m. Pendle Grit Formation. EM.

d) Fibrous illite between two quartz grains. This morphology contrasts markedly with the boxwork, pore-filling illite in the bottom left corner of the enlargement. The fibrous illite, in this case, is thought to be due to the complete transformation of a muscovite grain.
Whitmoor Borehole, 727m. Pendle Grit Formation. EM.

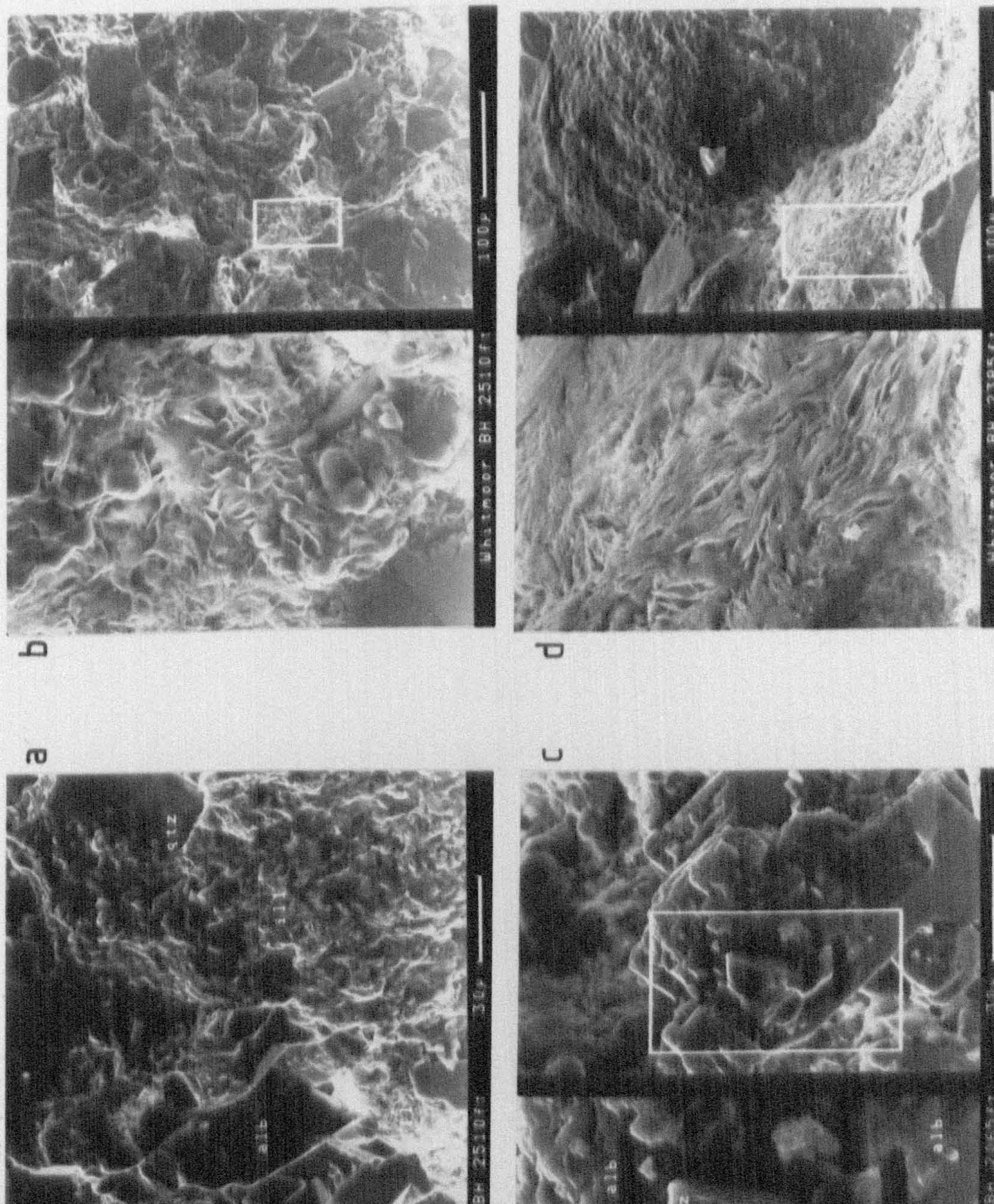

a
b
c
d
qtz
ill
alb
Whitmoor BH 2510ft
30μ
100μ
Whitmoor BH 2385ft

Plate 15

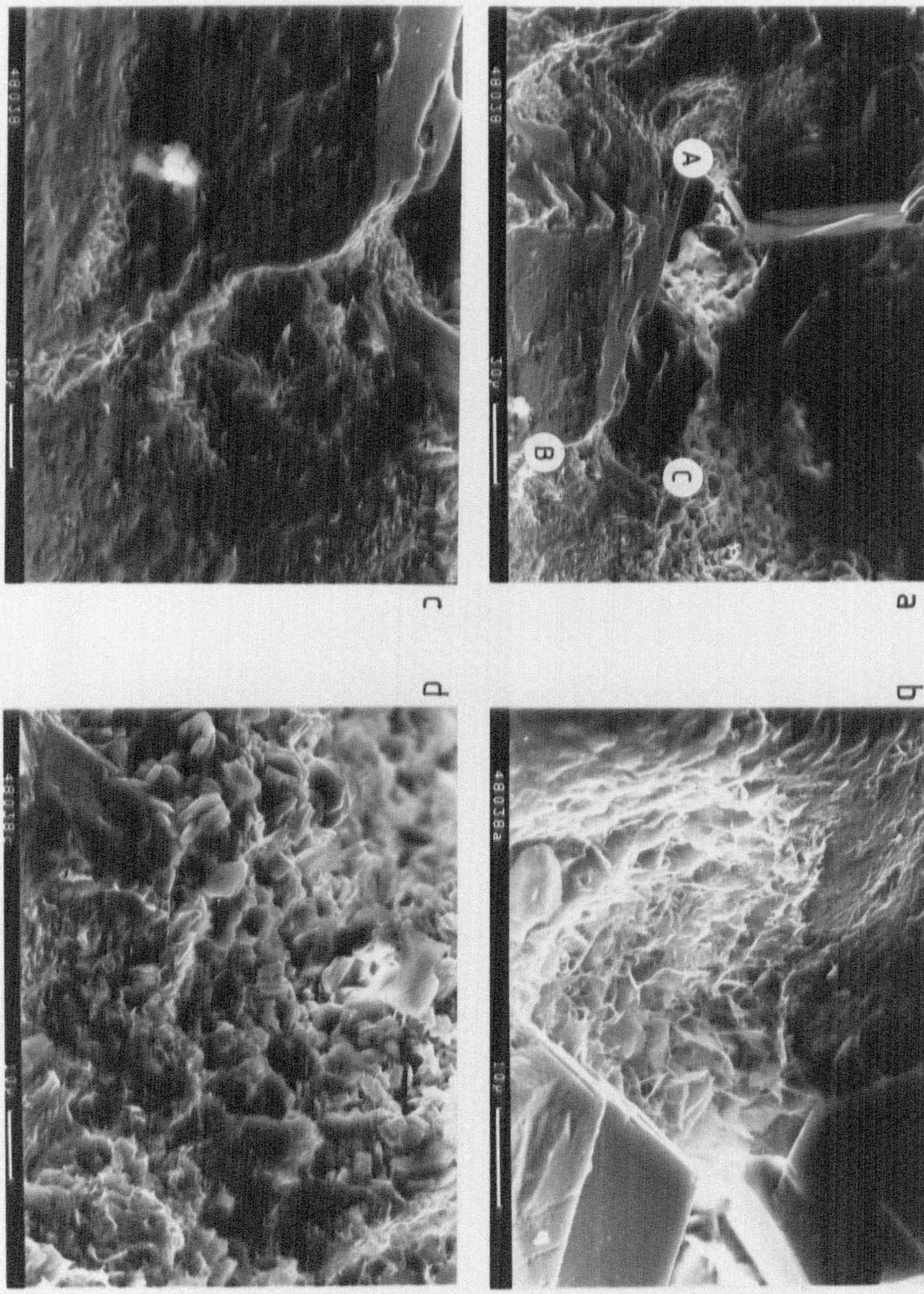


Plate 16

Plate 16

Authigenic clay morphologies (1).

a) General view of a large pore with pore-lining illite and pore-filling kaolinite (mostly washed out during preparation). Note also the well-developed quartz overgrowths. The remaining EMs on this Plate show details of this view.
Sample 48038 (A): Pendle Grit Formation. EMs.

b) Detail of area A showing the delicate morphology of the boxwork illite. Small "hairs" are beginning to grow on some of the pseudo-platelets of the illite. Note the grain contact area immediately above the illite (exposed during sample preparation). This area shows no evidence of major pressure solution.

c) Detail of area B. Boxwork illite grows in all the microporosity between the quartz overgrowths. The rough texture on the adjacent quartz surface is another grain contact area.

d) Detail of area C. Kaolinite booklets are seen growing over the boxwork illite and authigenic quartz. Note the euhedral and highly crystalline nature of the kaolinite.

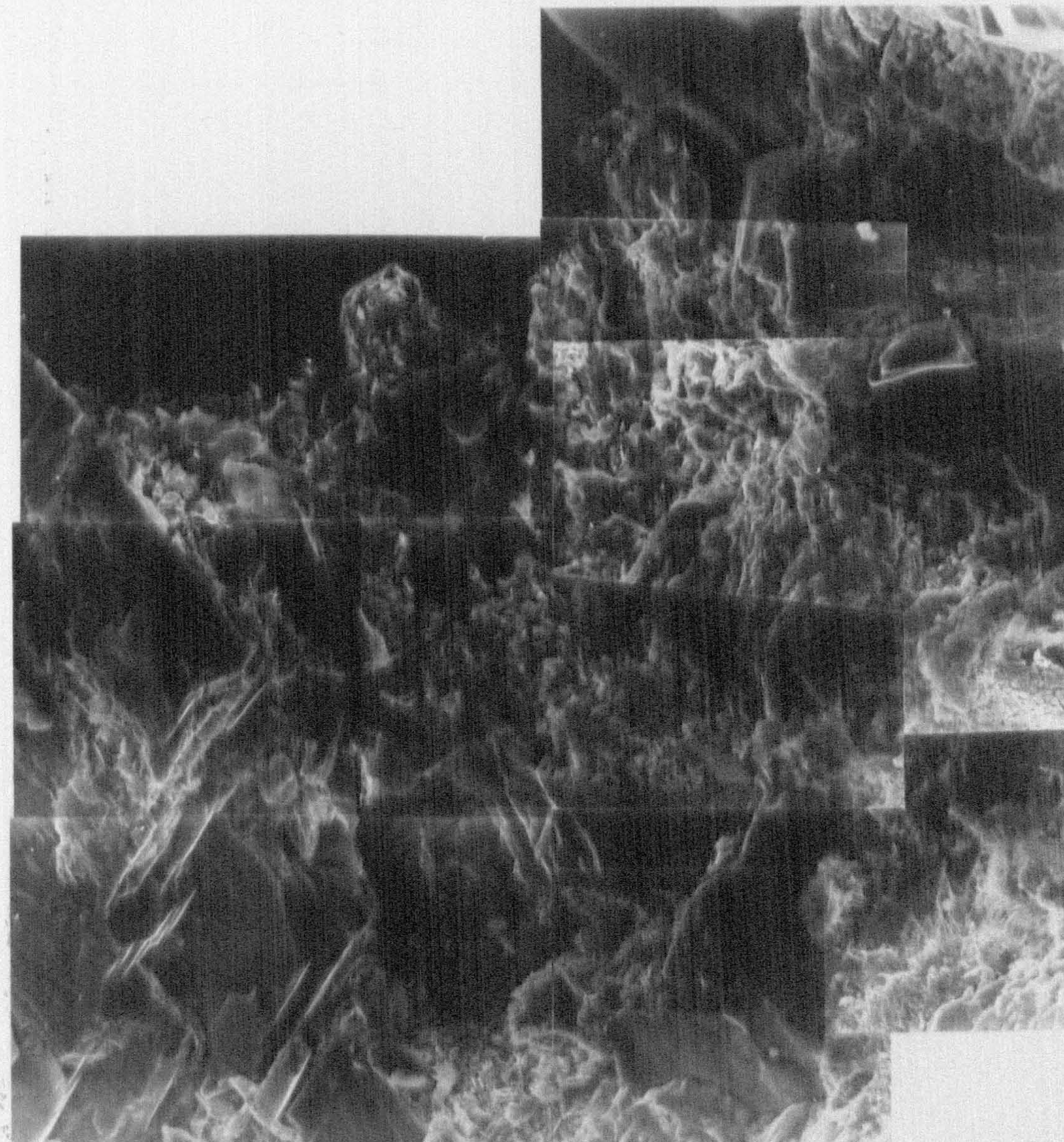
Plate 17

Authigenic clay morphologies (2).

Montage of EMs showing clay morphologies within a large pore. The left of the montage is dominated by a large, fractured orthoclase grain. This fracturing occurred during sample preparation. Partial alteration of the orthoclase can be seen along the cleavage planes. At top right another heavily corroded orthoclase is seen.

There is a thin development of pore-lining boxwork illite overgrowing a mixture of poorly crystalline kaolinite, thought to be recrystallised allogenic material. Well-developed fibrous illite is seen at bottom right.

Sample 48038 (A): Pendle Grit Formation. EM.


1,48038
30μ

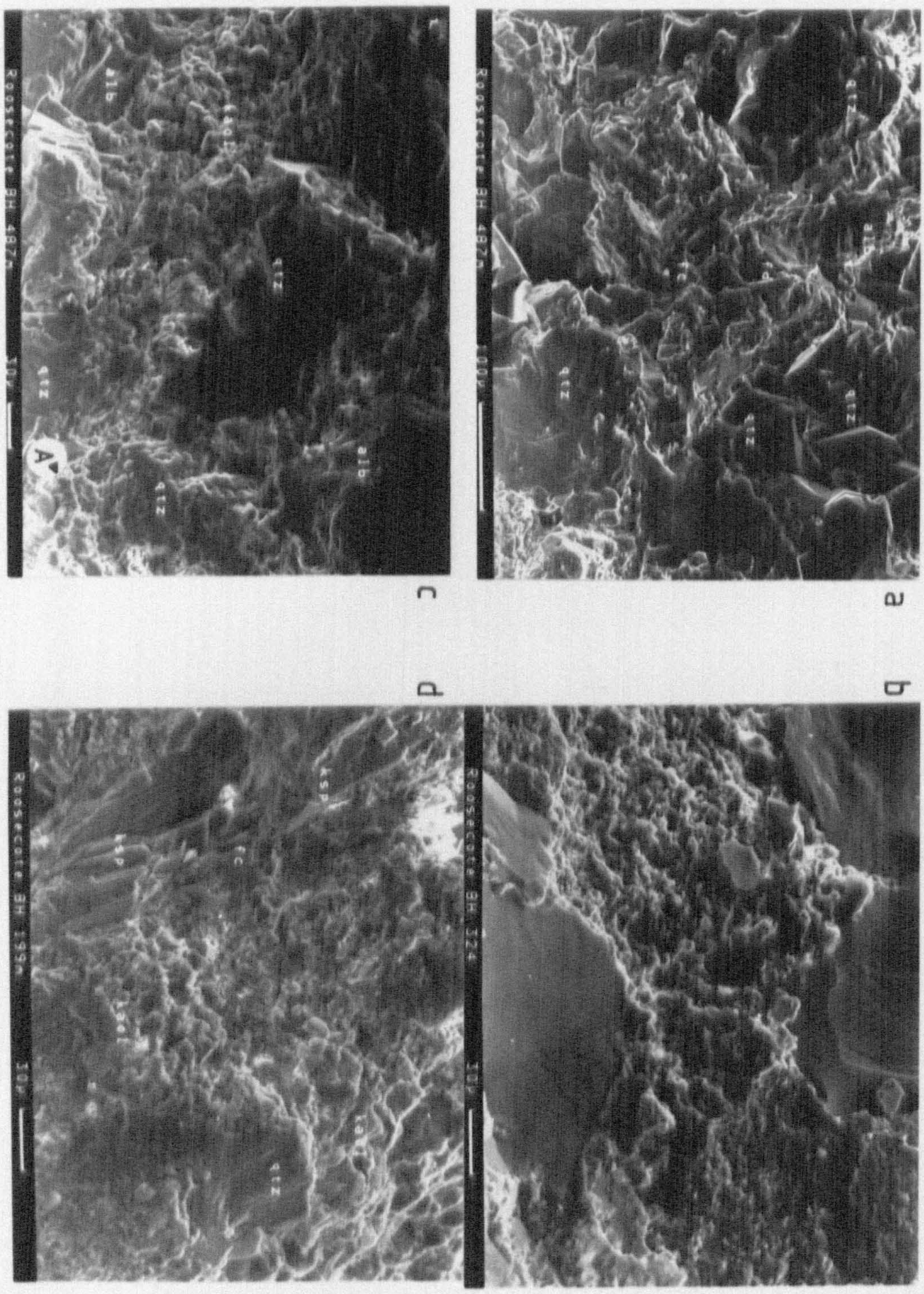


Plate 18

Plate 18

Electron micrographs from Roosecote Borehole.

a) Sandstone cemented by authigenic silica and later carbonate and albite. Note that the carbonate has a ferroan calcite (fc) core and a ferroan dolomite (fd) rim.
Sample 49418 (R1): Pendle Grit Formation. EM.

b) Pervasive carbonate cement (see also Plate 11c). The cement is microcrystalline ferroan dolomite and is replacive of all other phases. Note the strongly etched quartz surface at top right.
Sample 49420 (R2): Pendle Grit Formation. EM.

c) Kaolinite completely plugging any porosity remaining after quartz authigenesis. Note the euhedral morphology and the vermicular habit seen at (A). Also, note the absence of a pore-lining illitic clay (compare with Plate 16).
Sample 49418 (R1): Pendle Grit Formation. EM.

d) Kaolinite cemented sandstone (see Plate 11d). The kaolinite has a similar habit to the pore-filling variety seen in (c) except near grain margins where it is almost amorphous. The cement is replacive of all other phases: note the etched margin on the left of the labelled quartz grain.
Sample 49422 (R4): Pendle Grit Formation. EM.

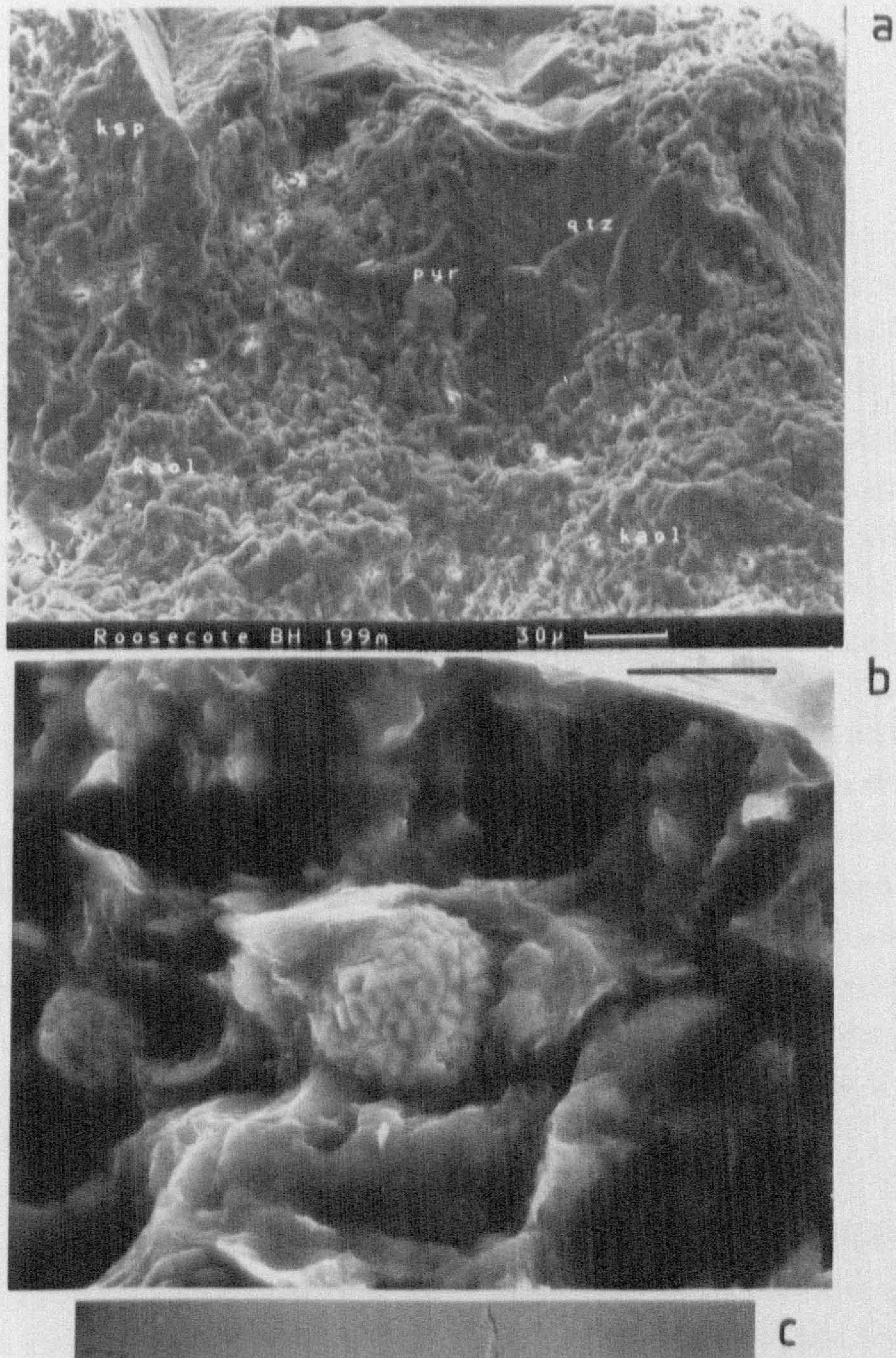
Plate 19

Miscellaneous electron micrographs.

a) Late authigenic pyrite growing within kaolinite cement(see Plate 18d). Note the etch pits on the quartz and feldspar grains.
Sample 49422 (R4): Pendle Grit Formation. EM.

b) Electron micrograph showing the morphology of the allogenic kaolinite in the Pendle Shale Formation. Note the poor crystallinity of the kaolinite compared to that of the pore-filling varieties in the sandstones. Micro-framboidal pyrite is seen growing within the kaolinite.
Sample from the Pendle Shale Formation, Whitendale (SD656 568). EM.
Scale bar= 5μm.

c) Back-scattered electron micrograph showing illitisation of muscovite. Where the muscovite grain extends into the porespace (M) it has been split along its basal cleavage planes and splayed apart by hydration of the muscovite lattice to form illite. Further growth of illite, possibly nucleating on that formed by muscovite alteration, has partially filled the porespace. Micro-muscovite (m) from the allogenic matrix has been similarly illitised and is intergrown with authigenic albite (A). Again, kaolinite (K) fills the remaining pores. The bright mineral at (R) is allogenic rutile.
Sample 48038 (A): Pendle Grit Formation. EM. Scale bar- 100μm.


a
ksp
qtz
pyr
kaol
kaol
Roosecote BH 199m
30μ
b
c
K
A
I
M
m
R

CHAPTER SIX
INTEGRATED BASIN ANALYSIS.

6.1 Introduction.

The previous chapters in this volume have reviewed various elements of the sedimentary geology of the Pendleian E_{1c} sequence within the Bowland and Lancaster Fells Basins. Chapter 2 described the distribution and thicknesses of the stratigraphic units within the E_{1c} basin-fill sequence and outlined the broad relationships of these units to sediments on the Askrigg Block. It was shown that the basins were stratigraphically distinct from the Askrigg Block during early E_{1c} times but that stratigraphic continuity between the two palaeogeographic units was established later during this zone. Also, it was noted that the boundary between the two main stratigraphic units, the Pendle and Grassington Grit Groups, is unconformable in some areas (the intra-E_{1c} unconformity). Chapters 3 and 4 then looked in detail at the sedimentology of the Pendle and Grassington Grit Groups. The Pendle Grit Group sediments were interpreted as the deposits of a sand-dominated submarine fan while those of the Grassington Grit Group were deposited by a braid-delta system. These two sand-rich packages were shown to represent separate pulses of sedimentation within the Bowland and Lancaster Fells Basins. Finally, Chapter 5 looked at the petrography and provenance of the E_{1c} sandstones. The data presented showed that the source area for both the Pendle and Grassington Grit Group sediments was a major continental landmass lying generally to the north of the British Isles.

The aspects of the geology of the Pendleian E_{1c} sequence covered in the individual chapters to date have been considered in relative isolation. In this chapter they are drawn together to show how the basins and their included sediments evolved with time.

6.2 Origin and Geometry of the receiving basins.

The gross geometry of the Bowland and Lancaster Fells basins during the early Namurian has been illustrated in Figure 2.1. Clearly, the basins shown on this figure were significant topographic depressions prior to the onset of Pendle Grit deposition: the submarine fans of the latter Formation compare with shallow-water facies deposited on the Askrigg Block during the same interval.

The development of the Pendle Grit Formation fans must have been strongly dependent on the initial geometry and relief in the basins because of the "low-seeking" nature of gravity driven turbidity currents. Fan growth and, hence, basin-filling would have been controlled by the palaeo-seafloor relief (cf. contributions to Bouma *et al.*, 1985). Consequently, the following sections discuss how the basin evolved prior to E_{1c} times and the probable distribution of topographic features within the basins at the beginning of Pendle Grit deposition.

6.2.1 Basin Initiation and Development.

The early history of the Bowland Basin has been considered by Gawthorpe (1985, 1987) and by Lee (1988a). Gawthorpe showed from thickness and facies variations in Dinantian sediments that the basin originated as a simple half-graben structure in the late Devonian or early Carboniferous (Figure 6.1). This half-graben trended north-east to south-west and resulted from fault-controlled extension along the south-eastern margin of the basin, coincident today with the Pendle Monocline (c.f. Figures 1.4 and 2.1). During this early stage of basin subsidence a carbonate ramp developed across the basin (Figure 6.1).

Later in the Dinantian the facies distributions indicate increasing segmentation of the basin into small fault-controlled sub-basins. This began with a phase of renewed tectonic extension during the Chadian to early Arundian, at which time the Craven Fault system appears to have developed as a major topographic feature (Gawthorpe *ops. cit.*. See Figure 6.1). Gawthorpe suggested that the segmentation was cause by the development of a series of syn- and antithetic faults to the main basin bounding structure, together with several transfer-fault zones (Figure 6.1). The facies developed during this interval show that there was substantial topographic relief both across the basin margins and the intra-basinal structures: carbonate gravity flows sourced from the basin margins were funnelled into sub-basins on the hanging-walls of the smaller fault-blocks while condensed sections with disconformities are present on the foot-wall blocks. Similarly, large sedimentary slides are found around the intra-basinal highs (Gawthorpe and Clemmey, 1985).

A later phase of tectonic extension during the late-Asbian to early Brigantian appears to have reactivated faults in the Craven Fault Zone and to have re-emphasised the topographic expression of these features. Further emplacement of thick carbonate debris-flows took place in the sub-basins close to the fault zone. These debris-flows were produced by erosion of the southern margin of the Askrigg Block: several minor unconformities developed in the Craven Fault Zone at this time (Hudson, 1930a; Hudson and Versey, 1935; Dunham and Stubblefield, 1945). The unconformities may have resulted from footwall uplift or unloading during the tectonic phase.

Following the tectonic pulse, sedimentation in the Bowland Basin was dominated by shale deposition: carbonate production virtually ceased. Gawthorpe (1985) suggests that this may have been due either to swamping of the carbonate dominated basin margins by clastics or to sea-level rise. An alternative possibility is that the tectonic phase, coupled with regional subsidence, resulted in basin deepening and drowning of the carbonate platform along the Bowland High. Whatever, by late Brigantian times the whole of the Bowland Basin was blanketed by marine mudstones of the Lower Bowland Shales and there is no evidence for shallow water conditions anywhere in the basin. The Askrigg Block however was still close to sea-level, as shown by facies in the Wensleydale Group sediments (Arthurton *et al*, 1988). Obviously, there was now substantial topographic relief across the fault zone. Hudson (1930a, 1944) showed that some of this relief was due to relict fault scarps along the component faults of the Craven Fault System. However, much of the relief may simply have been due to differential compaction across the fault-zone (Figure 6.2).

That topographic relief was still present on the intra-basinal structures is shown by the development of small turbidite sandbodies (Pendleside Sandstone) in the structural lows (Lee, 1988a). Further evidence indicates that this remained important through to the beginning of Pendle Grit deposition: condensed shale sequences with lipid-rich kerogen are developed in the Lower and Upper Bowland Shales across many intrabasinal highs. These compare with thicker shale sequences with a larger component of inert kerogen carried into adjacent "lows", presumably by muddy density currents (H. Clemmey pers. comm.).

The history of the south-eastern margin of the Bowland Basin during the late Dinantian is less certain. Gawthorpe (1985) used data from Boulesworth and Holme Chapel boreholes (see Figure 2.1 for locations) to show that a carbonate

Figure 6.1: Summary evolution of the Bowland Basin during the Dinantian.

a) Tournasian to early Chadian.

b) Mid-late Chadian to early Namurian.

Note the development of intra-basinal fault blocks during the latter interval, probably due to segmentation of the original hanging-wall block of the Pendle Monocline fault. The Askrigg Block also became fault-bounded at this time resulting in the development of steep topographic slopes along the northern basin margins. By Pendleian times this relief was substantial, at least along the South Craven Fault. This relief, together with that caused by compaction over intrabasinal fault blocks, controlled the development of the Pendle Grit Formation fan systems.

From Gawthorpe (1985).

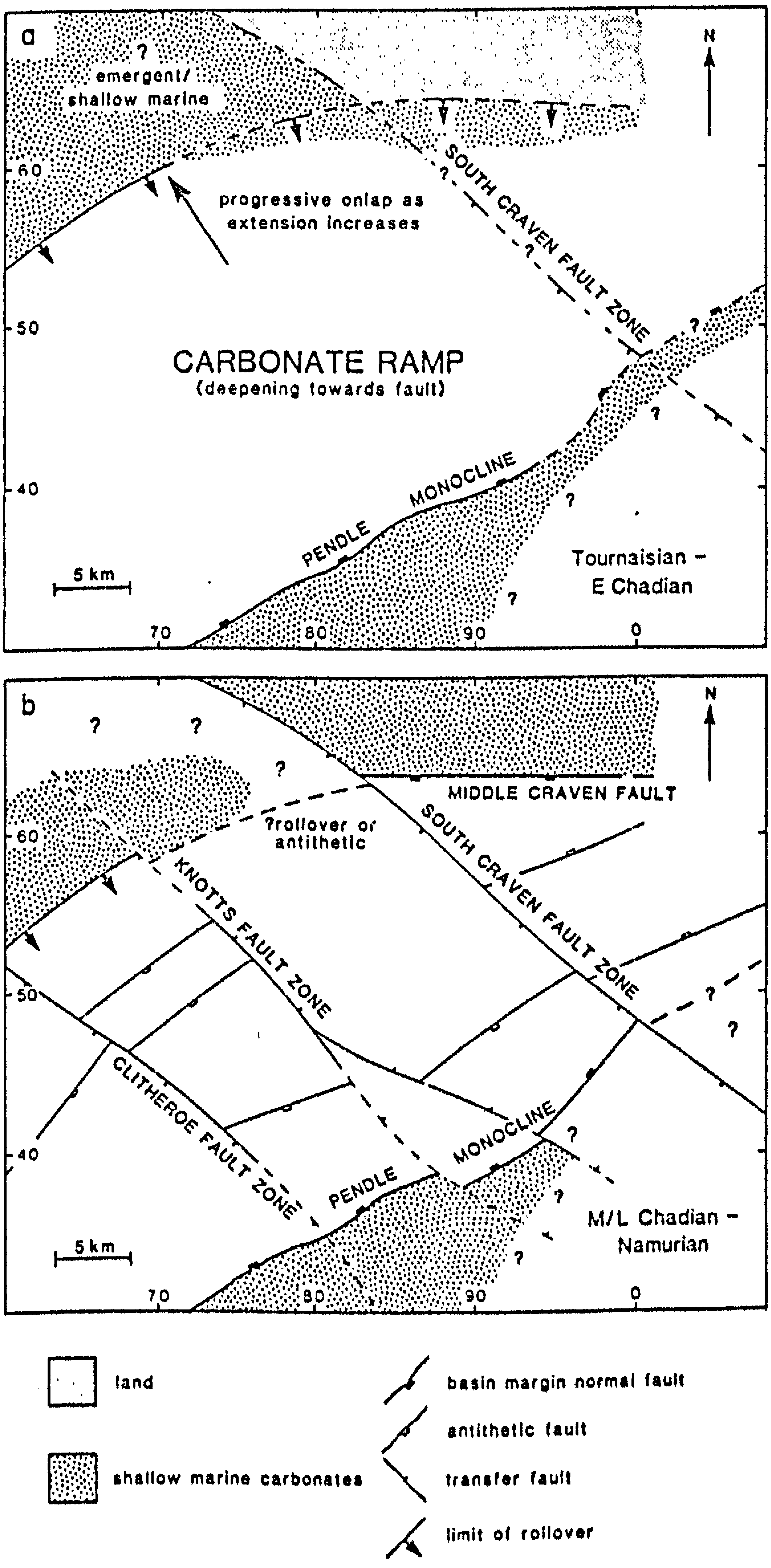
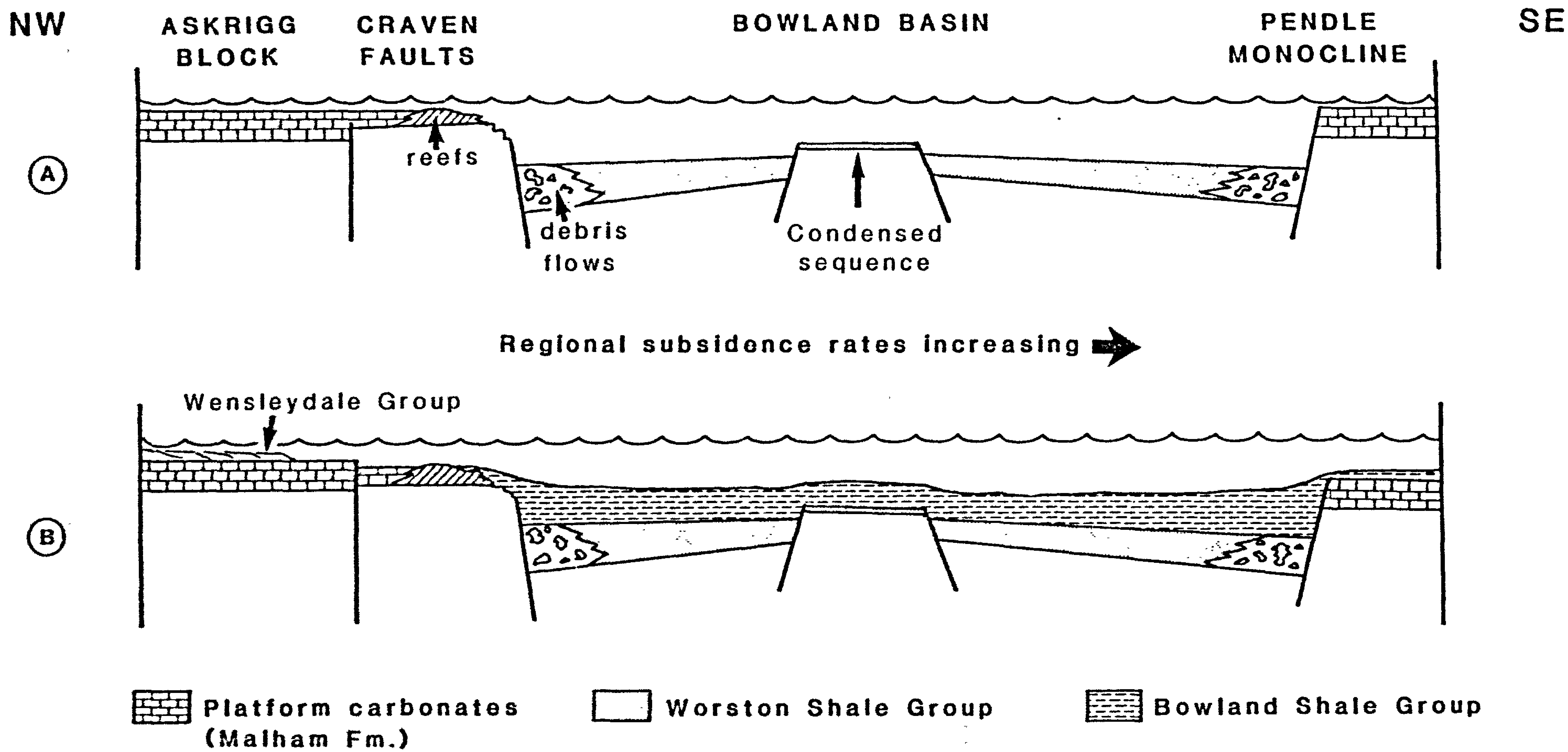

a
?
emergent/
shallow marine
60
progressive onlap as
extension increases
SOUTH CRAVEN FAULT ZONE
50
CARBONATE RAMP
(deepening towards fault)
40
PENDLE
MONOCLINE
Tournaisian -
E Chadian
5 km
70
80
90
0
N
b
MIDDLE CRAVEN FAULT
?rollover or
antithetic
SOUTH CRAVEN FAULT ZONE
KNOTTS FAULT ZONE
CLITHEROE FAULT ZONE
PENDLE
MONOCLINE
M/L Chadian -
Namurian
land
shallow marine carbonates
basin margin normal fault
antithetic fault
transfer fault
limit of rollover

Figure 6.2: Schematic models showing the effects of regional subsidence on the Bowland Basin.

a) Mid-Dinantian: sedimentation is controlled by fault activity. Debris flows sourced from the basin margins are emplaced into hanging-wall sub-basins with condensed sequences developing on the highs.

b) Situation prior to Pendle Grit Group deposition (mid Pendleian). Fault relief was partially smoothed out by deposition of the Bowland Shales but water depths were maintained because regional subsidence rates were higher than sedimentation rates. Higher regional subsidence rates to the south (i.e. effectively a regional tilting) caused the southern margin of the basin to be drowned and overstepped by the Bowland Shales. Consequently, from this time onwards, the Bowland Basin connected directly with the other basins of the Central Pennines Province. The regional subsidence was probably due to thermal cooling of the lithosphere (see sub-section 6.5.2). Note the remnant intrabasinal high caused by differential compaction over the buried Dinantian fault block.

Data from Gawthorpe (1985), Arthurton *et al.* (1988) and Lee (1988a). No scale implied by diagram.


NW
ASKRIGG BLOCK
CRAVEN FAULTS
BOWLAND BASIN
PENDLE MONOCLINE
SE
A
reefs
debris flows
Condensed sequence
Regional subsidence rates increasing
B
Wensleydale Group
Platform carbonates (Malham Fm.)
Worston Shale Group
Bowland Shale Group

platform was developed on the footwall block of the Pendle Monocline fault through most of the Dinantian (Figure 6.1). During this time the platform acted as the southern margin of the Bowland Basin. However, the presence of Bowland Shales and thick Pendle Grit Formation sediments in these wells shows that the carbonate platform had foundered (or been transgressed) during the late Brigantian or early Pendleian and that the Bowland Basin had expanded southwards (Figure 6.2). The timing of this foundering ties in with the basin-deepening event proposed above and fits well with a model involving asymmetric regional subsidence (Figure 6.2).

At the beginning of Pendle Grit Formation deposition, then, the Bowland Basin was open to the south but bounded to the north by a submarine slope, possibly with relict fault scarps along the Craven Faults. Within the basin minor topography was still present, this being the result of differential compaction across the fault-blocks formed during the Dinantian extensional phases (Figure 6.2). Although there is substantially less published data on the Dinantian evolution of the Lancaster Fells Basin, the same stages of basin evolution can be defined (A. Horbury, pers. comm.). Consequently, it is assumed that the early-Pendleian topography was similar to that in the Bowland Basin. However, the margin with the Askrigg Block is thought to have been a major submarine fault scarp. The gravity field in this area (Figures 6.3 and 6.4) shows a major "low" south of the South Craven Fault, centred beneath the Ingleton Coalfield. Calculations by Lee (1988a) show that this low probably represents a sedimentary sequence at least 2.5km thick. As the recorded thickness of the Westphalian and late Namurian strata in this area is only 1100m (Ford, 1954) there must be a substantial thickness of early Namurian and Dinantian sediments at depth (assuming the basin developed at a similar time to the Bowland Basin). It seems highly likely, given the history of the Craven Fault zone further south (see above), that this depocentre was formed by Dinantian movement on the South Craven Fault and that a substantial fault-scarp had developed by the early Namurian: the relief between the Askrigg Block and the basin floor had to be taken up across this one fault while further south it was spread across the many discrete faults in the Craven Fault Zone.

The possible locations of intra-basinal structures in the Lancaster Fells Basin have been predicted from the gravity data, using the method of Lee (1988a). These are shown, together with basement features which Lee recognised in the

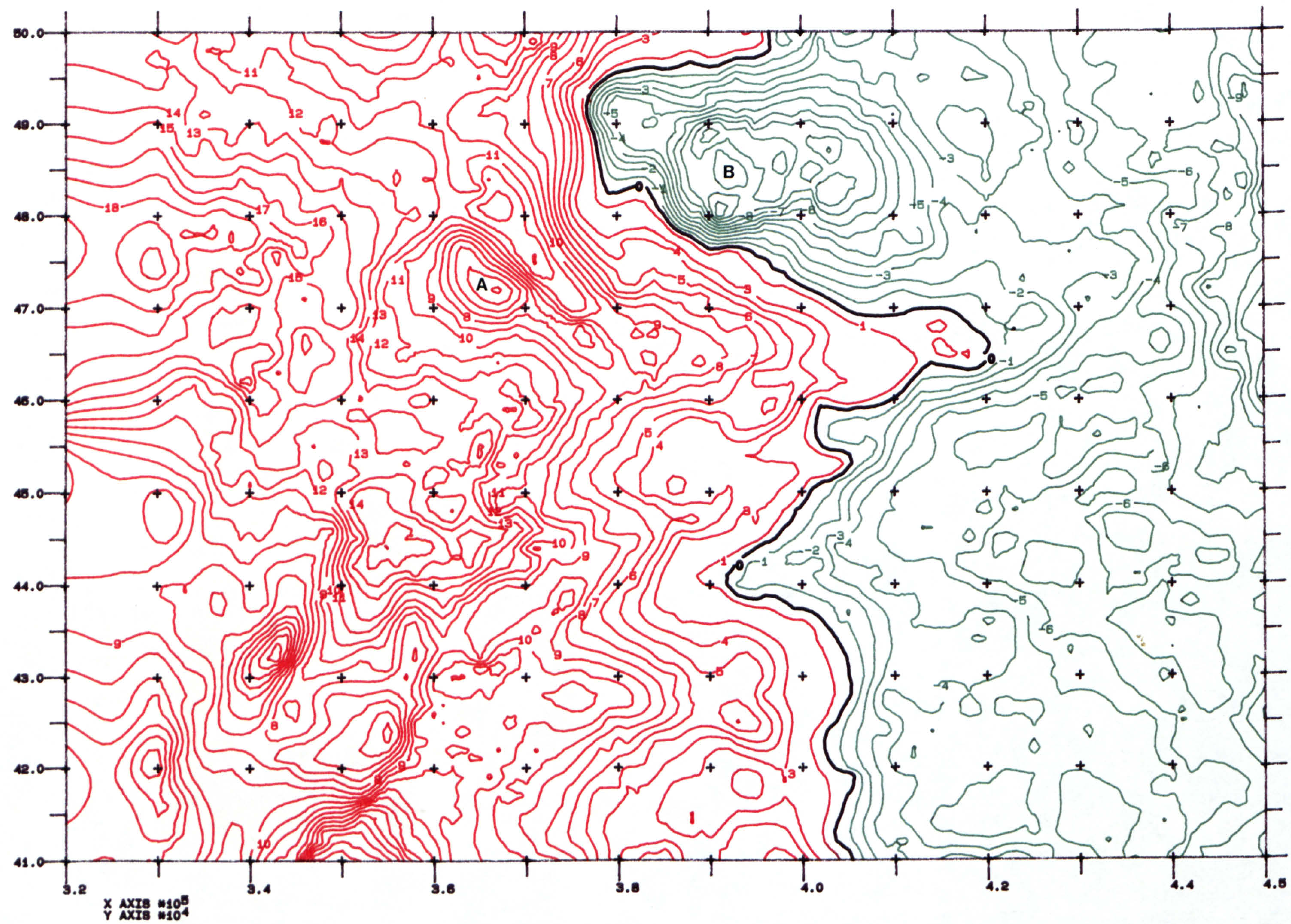

50.0
49.0
48.0
47.0
46.0
45.0
44.0
43.0
42.0
41.0
3.2
3.4
3.6
3.8
4.0
4.2
4.4
4.5
X AXIS *10^5
Y AXIS *10^4
A
B

Figure 6.3: Bouguer Gravity Field of north-western England. The gravity field is dominated by a long-wavelength regional trend which results in a decrease in the Bouguer values from west to east. However, several shorter period anomalies can be seen. Lee (1988) has shown that in this area most of these short period anomalies are generated by the density contrast between more dense Lower Palaeozoic basement/Dinantian carbonates and less dense Dinantian/Namurian clastic sediments. Consequently, gravity "lows" represent areas of thicker clastic sedimentation. The most obvious example is the Ingleton Coalfield (A). Low (B) is an exception in that it is generated by the high-level Wensleydale Granite which has a lower density than the country rocks. This map is difficult to interpret further because of the regional anomaly.

Map produced by A. Lee from B.G.S. data. Contour interval is 1 milligal: red contours are positive values of Bouguer Anomaly, green contours are negative values. There are no data points west of Easting 330 so the map west of this line is spurious.

Bowland Basin, on Enclosure 3. If any of these structures retained topographic relief into the Pendleian they may have influenced the progradation of the Pendle Grit Formation fan systems.

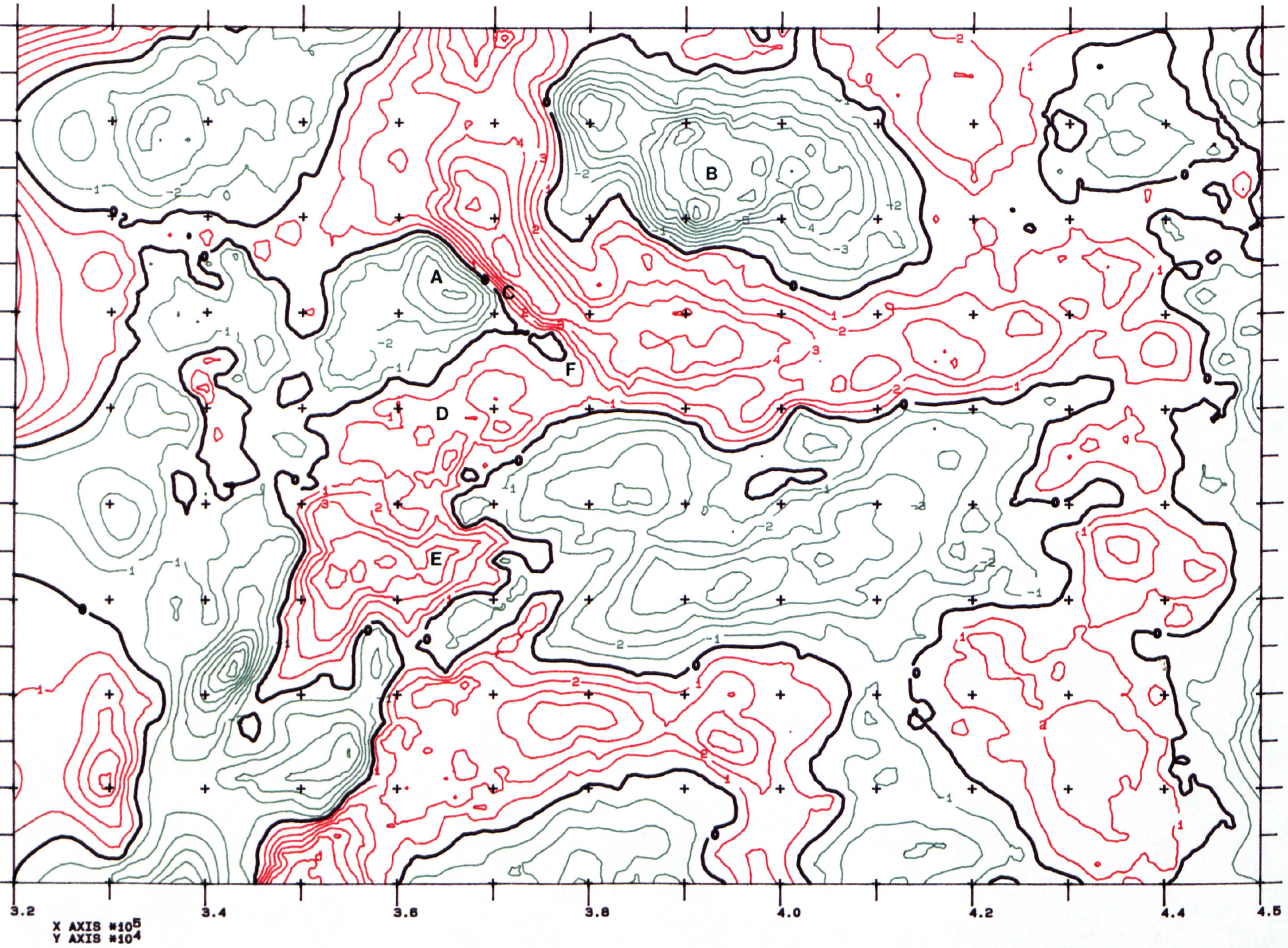
6.2.2. Estimation of initial topographic relief and water depths.

Estimates of the topographic relief across the basin margins can be arrived at by comparing the stratigraphic thicknesses of the E_{1c} sections along the margins of the Askrigg Block and in the immediately adjacent parts of the basin. This method works because the final level of the sediments both on the Askrigg Block and in the Bowland Basin was close to sea-level at the end of E_{1c} times. Consequently, the difference in the stratigraphic thicknesses of the block and basin sediments represents the amount of topographic relief infilled (not allowing for load-induced, compaction-induced or tectonic subsidence). Bearing in mind these assumptions, the isopach maps from Chapter 2 (Figures 2.7, 2.9 and 2.11) indicate that the relief across the Craven Fault Transition zone north of Skipton was around 450m. In the Ingleton Coalfield area, the relief on the South Craven Fault scarp may have reached 500m (assuming that the fault was the palaeo-basin margin and that Whitmoor borehole gives an estimate of the E_{1c} thickness close to the fault). Obviously these values are highly inaccurate but they do give a "feel" for the amount of submarine topography which existed prior to deposition of the Pendle Grit Formation.

By an extension of the above argument, as the Askrigg Block was close to sea-level at the beginning as well as the end of E_{1c} times, the above estimates of basin margin relief also correspond roughly to the water depths in the basins at the start of Pendle Grit Deposition. It is interesting to note that the values for the basin margin relief and the water depth indicate a geomorphological setting very similar to that of Bute Inlet, Canada. It is in this Inlet that the only modern analogue for the Pendle Grit Formation sand-rich fans is found (See Chapter 3).

6.3 Origin and rates of "Millstone Grit" sediment supply.

The flood of feldspathic clastics ("Millstone Grit") which marks the beginning of E_{1c} times in the Lancaster Fells and Bowland Basins is frequently thought to have resulted from tectonic uplift in the northerly hinterlands



A
B
C
D
E
F
50.0
49.0
48.0
47.0
46.0
45.0
44.0
43.0
42.0
41.0
3.2
3.4
3.6
3.8
4.0
4.2
4.4
4.5
X AXIS *10⁵
Y AXIS *10⁴

Figure 6.4: Residual gravity field of north-west England produced by subtracting a forth-order polynomial approximation of the regional anomaly from the Bouguer field (for details see Lee, 1988a). The residual field highlights the shorter period anomalies and shows their correct orientations. Consequently, the geological structures which give rise to these anomalies can be defined more accurately. The Ingleton Coalfield (A) and the Wensleydale Granite (B) are again very prominent. However, a lot more detail is now apparent within the basins. Steep gradients on residual anomaly maps of the Bowland Basin have been shown to relate to basement faults (Lee, *op. cit.*). The gradients relating to the current day Craven Faults are clear (C). The position of possible intra-basinal basement faults in the Lancaster Fells Basin has been determined from this map and from the 1:250,000 B.G.S. geological maps (See Enclosure 3).

Note how the Bowland High (D) and the Waddington Fells High (E) show up as residual gravity highs. The residual gravity trough at the eastern end of the Bowland High (F) may indicate the presence of a down-faulted trough which might have linked the Lancaster Fells and Bowland Basins (See text).

Map produced by A. Lee from B.G.S. data. Contour interval is 1 milligal: red contours are positive values of residual anomaly, green contours are negative values. There are no data points west of Easting 330 so the map west of this line is spurious.

(e.g. Gilligan, 1920). However, when looked at on the regional scale, the appearance of feldspathic material is much less sudden than is sometimes realised. Coarse feldspathic sandstones were deposited in the north-western province of Ireland from late Chadian to Asbian times (Sevastopulo, 1981); all of them were sourced from the north and north-west. In the Bowland Basin itself, feldspathic sandstones are developed in the Brigantian and early Pendleian (Pendleside Sandstone, Hind Sandstone). The arrival of the Millstone Grit in the English Central Pennines Province, then, is simply a repeat of what had occurred previously in related Carboniferous basins and there is no need to infer a major tectonic uplift in order to explain the production of this feldspathic material.

A change to more pluvial climatic conditions, caused by northwards migration of the Pangean supercontinent through the tropical rain-belt, has also been called upon as an explanation for the apparently enhanced sediment supply (Drewery *et al.*, 1987; Leeder, 1988). Although there is strong evidence for such a climatic change, it unlikely that this can explain the initiation of Pendle Grit Formation deposition. Again, the Irish data suggest that major fluvial systems were active throughout the early Carboniferous. Similarly, the flood of quartzitic clastics into the Northumberland Trough during the Dinantian shows that major fluvial systems were also present in some of the English basins.

An alternative explanation for the sudden arrival of the feldspathic clastics in the English basins is that there were major changes in the drainage systems of the northerly hinterlands. These changes resulted in the output from the drainage basins moving north-west from Ireland into northern England. Such a migration could have been caused by tectonic or geomorphic events. Right lateral strike-slip motions on Caledonide suture zones could provide a mechanism for sweeping the fluvial outlet north-westwards but there is little evidence for such motions from palaeomagnetic studies (Smethurst, 1987). Alternatively, block faulting in the hinterlands might have produced different pathways through the Caledonian highlands at different times. There is, however, no need to infer any process other than geomorphological change: river avulsion or river capture are capable of causing major changes to sediment distribution systems without any significant external influences. Such processes are likely to be enhanced by rapid changes in river base-level which allow geomorphic thresholds to be breached (Schumm, 1977): perhaps the eustatic rise which led to development of the

Cravenoceras malhamense marine band triggered the breaching of such a threshold and caused diversion of the lower reaches of the fluvial system into the English basins. A general change in palaeoslope in the hinterlands brought about by west-to-east tilting might also explain the eastwards migration of the fluvial systems.

Unfortunately, it is impossible to distinguish between these possibilities with the data available. However, the present author prefers an explanation which does not call upon major major tectonic activity in the hinterlands. Any such activity would probably produce contemporaneous tectonic effects in northern England. As there is no evidence for such activity at the beginning of E_{1c} times, the more simple explanations for the initiation of feldspathic sedimentation involving migration of river systems are preferred.

An "order of magnitude" estimate of the average sediment supply rate from the fluvial system to the Lancaster Fells and Bowland Basins has been made, based on the total volume of sediment introduced during E_{1c} times. Using an average of 500m for the stratigraphic thickness of the E_{1c} sediments in the basins, the volume of sediment supplied was $1.6 \times 10^3 km^3$. Holdsworth and Collinson (1988) estimate that the average length of a goniatite zone in the Namurian was 180,000 years which, if correct, indicates average sediment inputs of $8.9 \times 10^6 m^3 yr^{-1}$. As most of this was quartz clastic material, an average of 2×10^7 tonnes yr^{-1} of sediment must have been carried by the fluvial system which supplied the basins (assuming an initial porosity of 20% for the sediments). This rate is about an order of magnitude less than the total (suspended+bedload) sediment supply rate of the Brahmaputra river at the present time (Bristow, 1988). The E_{1c} river systems were probably dominated by bedload sediment so these figures are not directly comparable. However, the large discrepancy between the total sediment transport rates is worrying given that the petrological data discussed in Chapter 5 indicate Carboniferous drainage basins comparable in size to modern continental systems such as the Brahmaputra. Further research work on the size of Carboniferous fluvial systems and the rates of sediment supply to the Carboniferous basins are required to define better which modern rivers provide good models for the Carboniferous fluvial systems.

6.4 Basin infill.

The complete lack of time control within the E_{1c} sequence of the Lancaster Fells and Bowland Basins means that the detailed evolution of the basin fill can only be guessed at. Particularly, it is impossible to prove whether or not the Pendle and Grassington Grit Groups are, in part, time equivalents. Baines (1977), who looked only at the Grassington and Skipton areas of the basins, interpreted the vertical sequence from the Pendle Grit Formation through to the Grassington Grit Formation as resulting from the progradation of a submarine-fan fronted delta system southwards off the Askrigg Block. His model mirrors that proposed for other Namurian sequences in the Central Pennines Province (e.g. Walker, 1966; Collinson, 1970; McCabe, 1977; Jones, 1980). Such a model clearly implies that the Pendle Grit Formation is the down-palaeocurrent equivalent of the Grassington Grit Group and that deposition of the two stratigraphic units occurred synchronously but offset in space. While this argument is mostly consistent with Baines' limited data set, it breaks down when data from the whole of the Lancaster Fells and Bowland Basins are looked at together. The following specific problems arise:

1) Much of the Pendle Grit Formation was deposited synchronously with Wensleydale Group (Yoredale) sediments on the Askrigg Block. Consequently, the sediment supply for the Pendle Grit Formation cannot have come across the Askrigg Block as Baines had proposed (but see below).

2) An unconformity (the Intra-E_{1c} event) is often seen between the Pendle and Grassington Grit Groups over structural highs and in the Craven Fault Transition Zone. This implies some eustatic/tectonic event which caused palaeogeographic change during E_{1c} times. It is only after this event that feldspathic clastics entered the basins via the Askrigg Block.

3) The sedimentology of the Grassington Grit Group braid-deltas, particularly those of the Warley Wise Formation, suggests that the sediment supplied by the fluvial system was all deposited in the braid-delta and its associated mass-flow apron. This local mass-flow apron, rather than a linked submarine fan, caught all the sediment supplied down-slope by gravity processes.

4) Palaeocurrent directions in the Pendle Grit Group are significantly different from those in the Grassington Grit Group (Figures 6.5 and 6.6).

Bearing in mind the above points, it is tempting to develop a model in which the deposition of the Pendle Grit Group, the production of the intra-E_{1c} unconformity and the deposition of the Grassington Grit Group occurred in discrete time intervals. However, given the short timespan of E_{1c} it seems inevitable that there was some overlap of these events. The model presented below is the current authors preferred explanation for the relationships seen. It is accepted that this is not a unique solution.

6.4.1 Deposition of Pendle Grit Group sediments.

One of the principal problems of early E_{1c} palaeogeography is where the feldspathic sediments of the Pendle Grit Formation entered the Lancaster Fells and Bowland Basins. Palaeocurrent directions in the Pendle Grit Group (Figure 6.5) allow two alternatives:

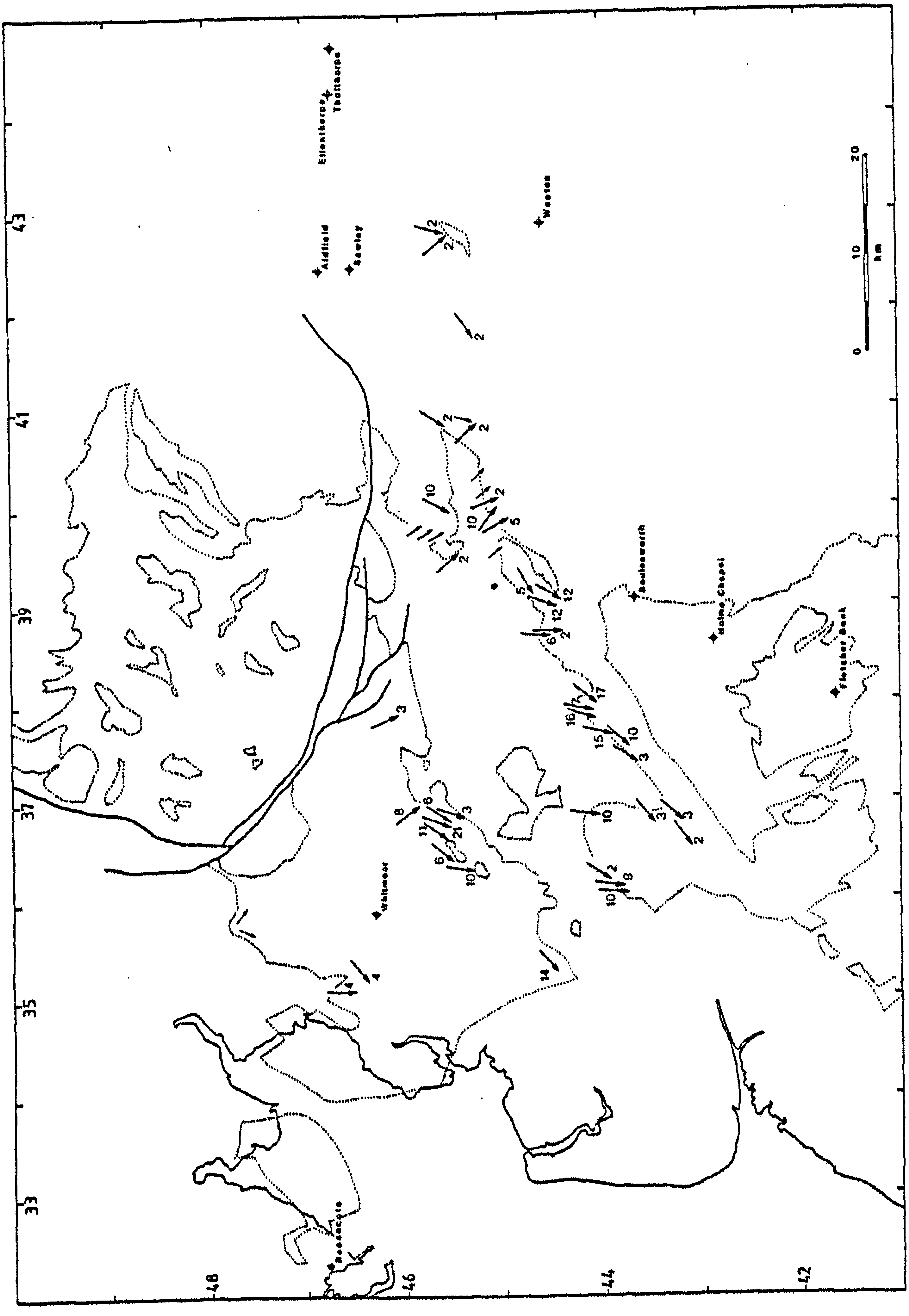
1) Sediment was supplied from the north-west and entered the Lancaster Fells Basin first. Once in the basin, turbidity currents flowed axially (i.e. towards the south-west). Sediment was supplied to the Bowland Basin around and across the Bowland High once the latter ceased to form a topographic obstruction to flow.

2) Sediment was supplied from the north-east and entered the Bowland Basin via the Eastern Bowland Basin. Overspill into the Lancaster Fells Basin occurred after the topographic relief on the Bowland High had been swamped by sediment.

Previously it has been assumed that the "Millstone Grit" sediments all came from the north-east. This assumption was based on the petrographic work of Gilligan (1920) and on sedimentological studies of other Namurian clastic sequences (e.g. Sorby, 1859; Walker, 1966; Collinson, 1968). However, in the case of the Bowland Basin, this would require that coarse sediment bypassed the Eastern Bowland Basin throughout deposition of the Pendle Grit Formation. The

Figure 6.5: Palaeocurrent directions in the Pendle Grit Group. Most palaeocurrents are from flutes, linear loads or rippled bed tops. Large arrows show the vector means of the given number of palaeocurrent measurements at the locality. Angular dispersions at any locality are usually small except in the case of some Facies C and D beds (see Chapter 3) confined to channels. In this setting the rippled tops of beds may give a palaeocurrent direction up to 90° different from the axial direction given by the bottom structures. This is thought to be caused by collapse of the suspension cloud created by the initial turbidity current after the end of the main flow period. The collapse would be controlled by the local palaeoslope: that is the gradient down the channel margin into the channel axis. The ripples may be produced by such flows.

The low number of measurements at most localities reflect the lack of palaeocurrent indicators within this sand dominated system. Small arrows show palaeocurrent directions reported by Baines (1977) and by Toswill (pers. comm.).


33
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0
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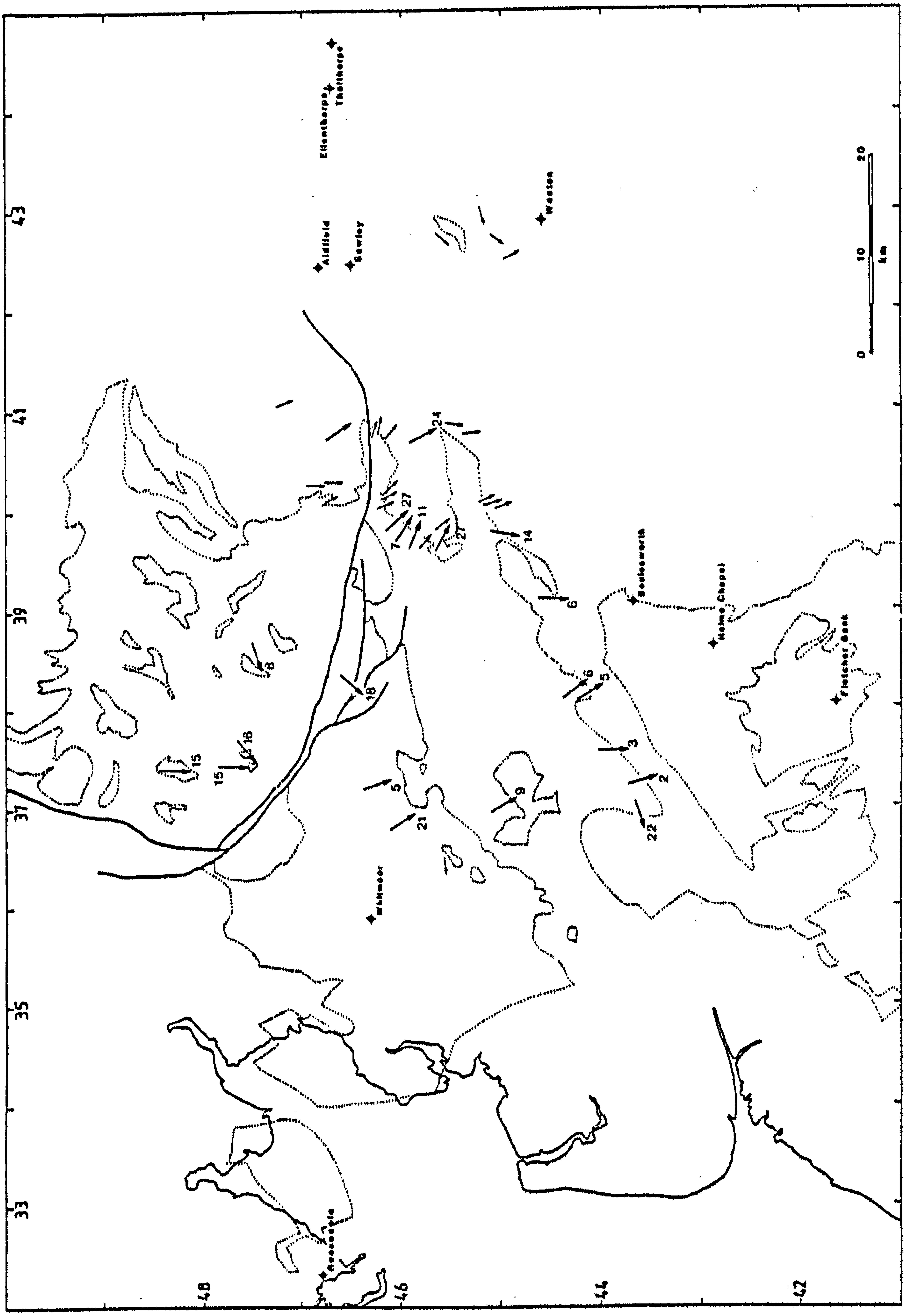
km

Figure 6.6: Palaeocurrent directions in the Grassington Grit Group. Palaeocurrents have been measured from trough axes, ripples on bed tops and from large and giant foresets. Individual foresets of trough cross-sets have not been included to cut the angular dispersion of the measurements. Large arrows show vector mean directions of the given number of measurements. The larger number of measurements at most sites reflects the greater proportion of traction dominated sediments in the Grassington Grit Group. Small arrows show palaeocurrent directions recorded by Baines (1977), Wilson (1957) and Chisholm (1981).


Aldfield
Sawley
Weston
Holme Chapel
Fletcher Bank
km

sediment would also have had to bypass the zone of south-easterly directed palaeocurrents which lies along the South Craven Fault (use Enclosure 3 as an overlay to Figure 6.5). A further problem with the north-easterly source is that some of the boreholes to the east of the study area have thick sequences of Pendleian quartzitic sandstones: there is no evidence for a feldspathic influx into these "up-current" basins during Pendleian times (Bentham, pers. comm.). Consequently, sediment input from the north-east is thought unlikely: a supply of sediment into the Lancaster Fells Basin from the north-west is preferred. In the current author's opinion, a north-westerly input is more consistent with the palaeocurrent data set. It would also account for the increased grade of the Pendle Grit Formation sediments in the Lancaster Fells Basin and the abundance of silty, carbonaceous material in the Bowland Shales of the Keer Valley (see Chapter 2).

There is some evidence that the Lake District Block was not an exposed land area during E_{1c} times: isopachs for Brigantian and early Pendleian cyclothems in the Wensleydale Group show a thickening of the sediments on the Askrigg Block towards the Lake District Block (Dunham and Wilson, 1985; their Figures 10 to 12). There is also a substantial westerly thickening of earlier Dinantian sediments across the line of the Dent Fault (George, 1958) which suggests that a this line may have acted as a down-to-the-west structure at this time (See Figure 6.7a). The feldspathic sediments of the Pendle Grit Group could, therefore, have been supplied across the Lake District Block. Support for this hypothesis comes from the Hensingham Grit of West Cumbria. This sandstone unit is probably time-equivalent to the Pendle Grit Formation and, from a brief description in Taylor *et al.* (1971), is a coarse-grained, feldspathic unit developed in fluvial facies. It may represent the fluvial system which supplied the Pendle Grit Formation fans. Unfortunately, the current author has been unable to find any published sedimentological data relating to this unit which could establish a firm link between the systems. Further investigation of the Hensingham Grit is clearly necessary.

The sand-rich fans of the Pendle Grit Formation, then, were first developed in the Lancaster Fells Basin (Figure 6.7a). As discussed in Chapter 3, the detailed sedimentological development of these fans resulted from the interaction of the coarse-grained sediment supply with the confined basin. The lack of lateral and vertical facies changes in the main body of the Pendle Grit Formation suggests that it is the deposit of a single fan that rapidly expanded to occupy the whole of the Lancaster Fells Basin (Figure 6.7b). The only evidence of the initial progradation across the basin is the presence of the thin lobe sequences of the Whitendale Member (See Chapter 3). Once established, the fan system began to fill the basin by vertical aggradation.

This model implies that the Pendle Grit Formation fans extended over at least 50km in the down-palaeocurrent direction. Such a scale is entirely consistent with the sizes of other small, topographically confined fans in modern settings (See review by Nelson and Nielson, 1984). Once again, the Bute Inlet fan system (See Figure 3.25) provides a good analogue: the along palaeocurrent dimension of the system is identical to that proposed for the Pendle Grit Formation.

The aggradational style of basin-filling does not imply that the fan system developed synchronously across the whole basin complex: sedimentation in the Bowland Basin probably began at a later time due to the "ponding" effect of the intra-basinal topographic relief discussed in Section 6.2. "Low-seeking" turbidity currents carrying sediments into the basin were initially confined to the Lancaster Fells Basin by the relief across the Bowland High: they entered the Bowland Basin only after this relief had been eliminated by sedimentation. Spillover may have first occurred along a topographic low located along the line of the South Craven Fault (Figure 6.7b). There is evidence from residual gravity data (Figure 6.4) for a trough separating the Bowland High and Askrigg Block in this area. Diversion of turbidity currents along this feature would also explain the south-easterly directed palaeocurrents in this part of the basin (Figure 6.5). Later, the fan system transgressed the whole of the Bowland High and supplied sediment direct into the Bowland Basin (Figure 6.7c).

Once in the Bowland Basin, the fan system again expanded rapidly to occupy the whole basin. By this stage the sand-rich fan was active across most of the basin complex: only the Waddington Fell High remained above the level of fan sedimentation throughout (Figure 6.7c) This feature obviously maintained sufficient relief to divert turbidity currents throughout this period and, as a result, the fan facies pinch out onto this feature (Figure 2.3).

Figure 6.7: Palaeogeographic evolution of north-west England during E_{1c} times.

a) Initiation of Pendle Grit Formation deposition. Sand-rich fans develop in the Lancaster Fells Basin at the foot of the steep basin margin slope. This slope, and the position of the feeder channels on it, was probably fault controlled. Limestone/deltaic cyclothems are deposited on the Askrigg Block. These sequences thin southwards towards the Block margin. The Dent Fault (dotted) may have been active as a down-to-the-west structure, causing water depths to increase westwards off the Askrigg Block. BH= Bowland High, WH= Waddington Fells High, PM=Pendle Monocline High.

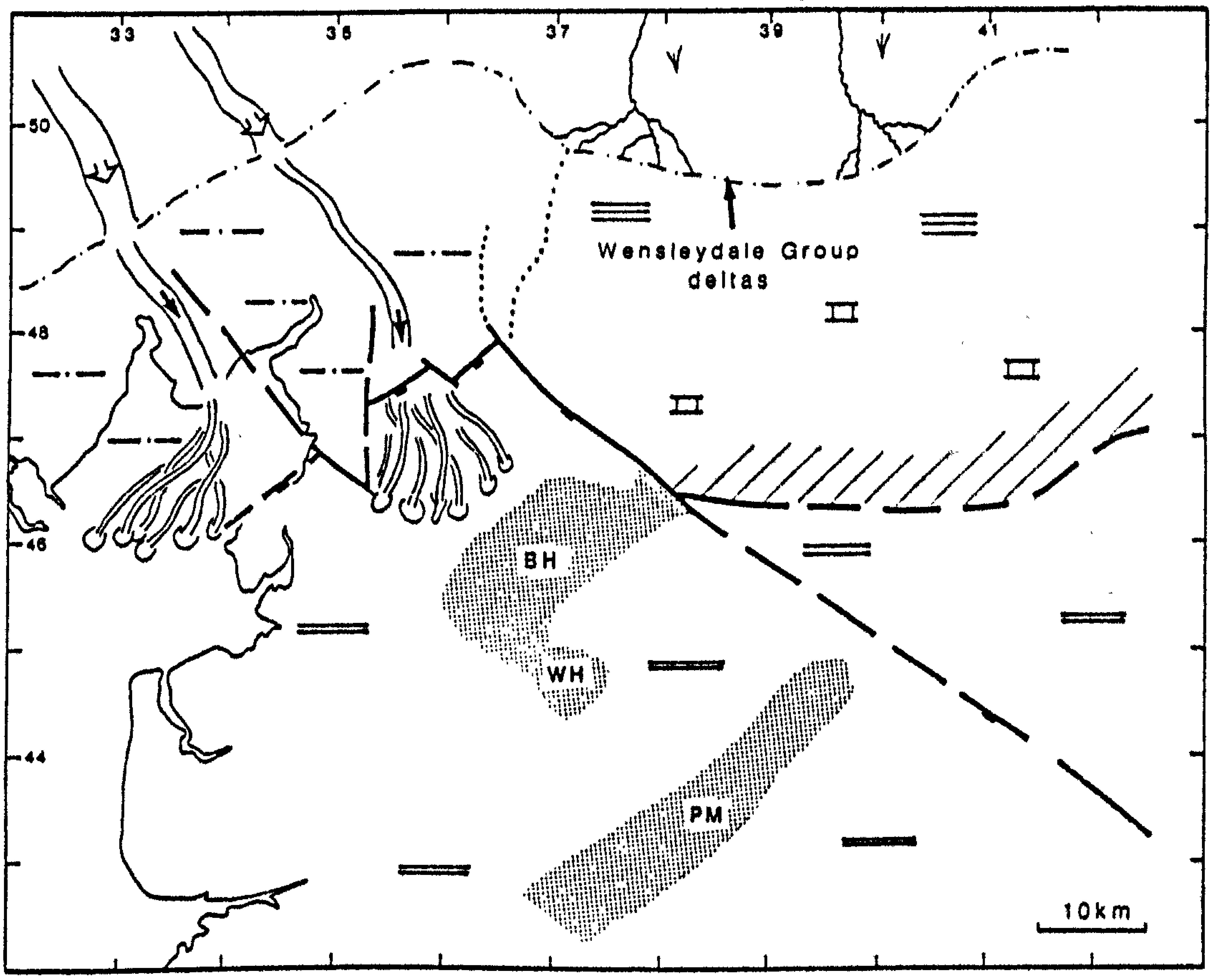
b) *Overleaf.* Fans prograde rapidly down the axis of the Lancaster Fells Basin. As topographic relief is infilled, the fans overspill into the Bowland Basin probably via a trough running along the South Craven Fault. Mud-rich hypopycnal flows issuing from the Wensleydale Group deltas may have supplied mud across the basin margin high and into the basins.

c) *Overleaf.* Fan sedimentation expands to fill the whole of the basin complex. Only the Waddington Fells High remains unaffected (See Figure 2.3). The fans prograde south across the northern end of the buried Pendle Monocline High into the Huddersfield Basin (name after Lee, 1988a). Further south the effects of differential compaction across this feature are sufficient to control the direction of fan progradation. Tilting of the Askrigg Block leads to expansion of the marginal high and, possibly, to minor erosion along the Block edge.

d) *Overleaf.* Initiation of Grassington Grit Group sedimentation. The feldspathic fluvial system has avulsed eastwards and now debouches onto the Askrigg Block. A shallow water braid-delta develops and progrades rapidly towards the break of slope across the Craven Fault zone. These faults had been partially reactivated during the intra-E_{1c} tectonic event.

e) *Overleaf.* Once into deeper water, progradation slows and the braid-delta develops "Gilbertian" foresets and a mass-flow apron. Delta top and flood plain environments develop on the Askrigg Block.

A)



Intra-basinal high

Block margin high

Active fault or relict fault plane

Down-to-basin slope controlled by faults

Basinal shales

Delta-front shales

Slope deposits

Limestones

Slope feeder channels

Fluvial channels

Fan channels

Fan lobes

Braid-delta distributary channels

Shallow-water braid-delta front deposits

Giant avalanche sets

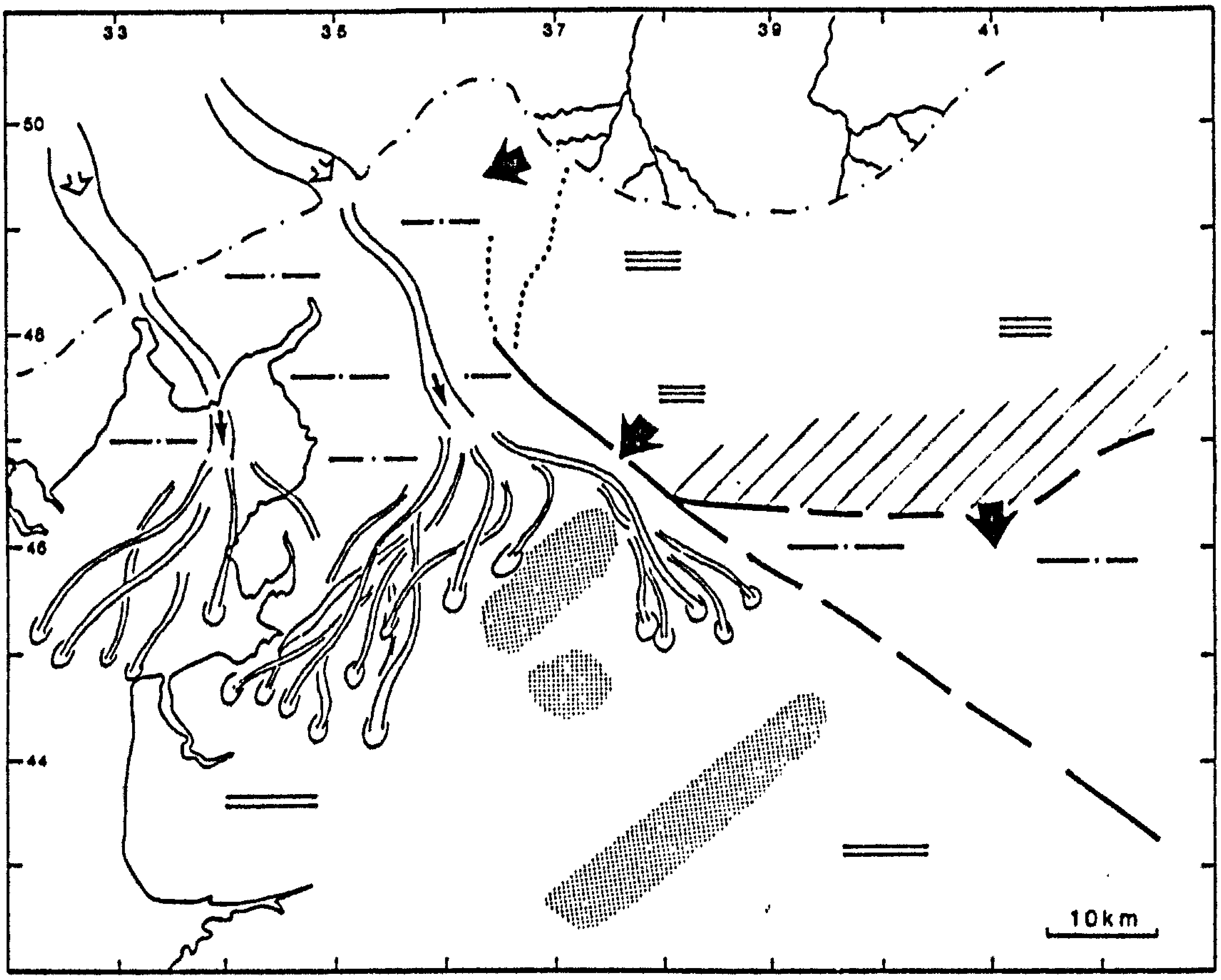
Mass-flow apron

shoreline

Mud plumes carrying sediment from block to basin

Delta-top/flood plain environments

B)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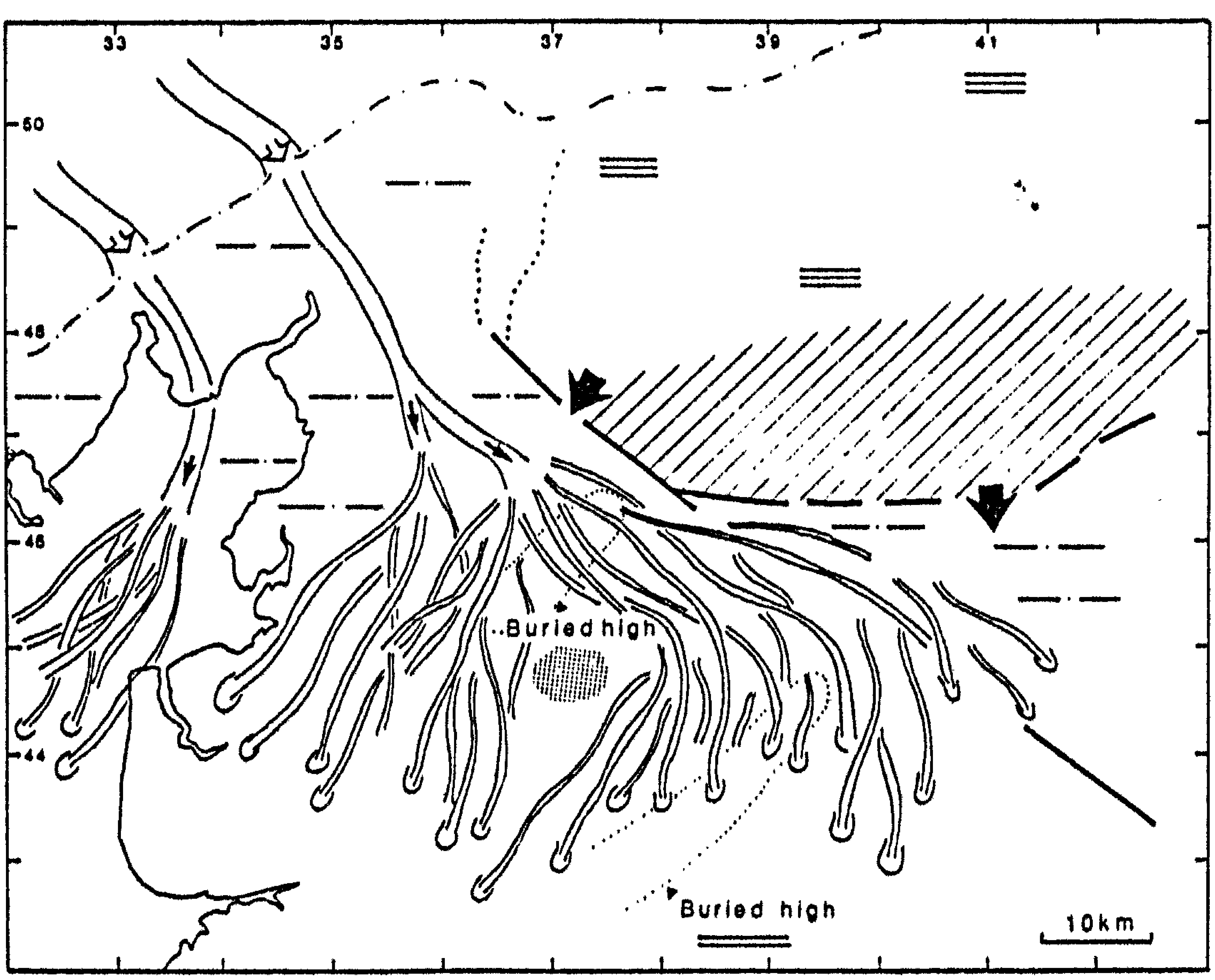
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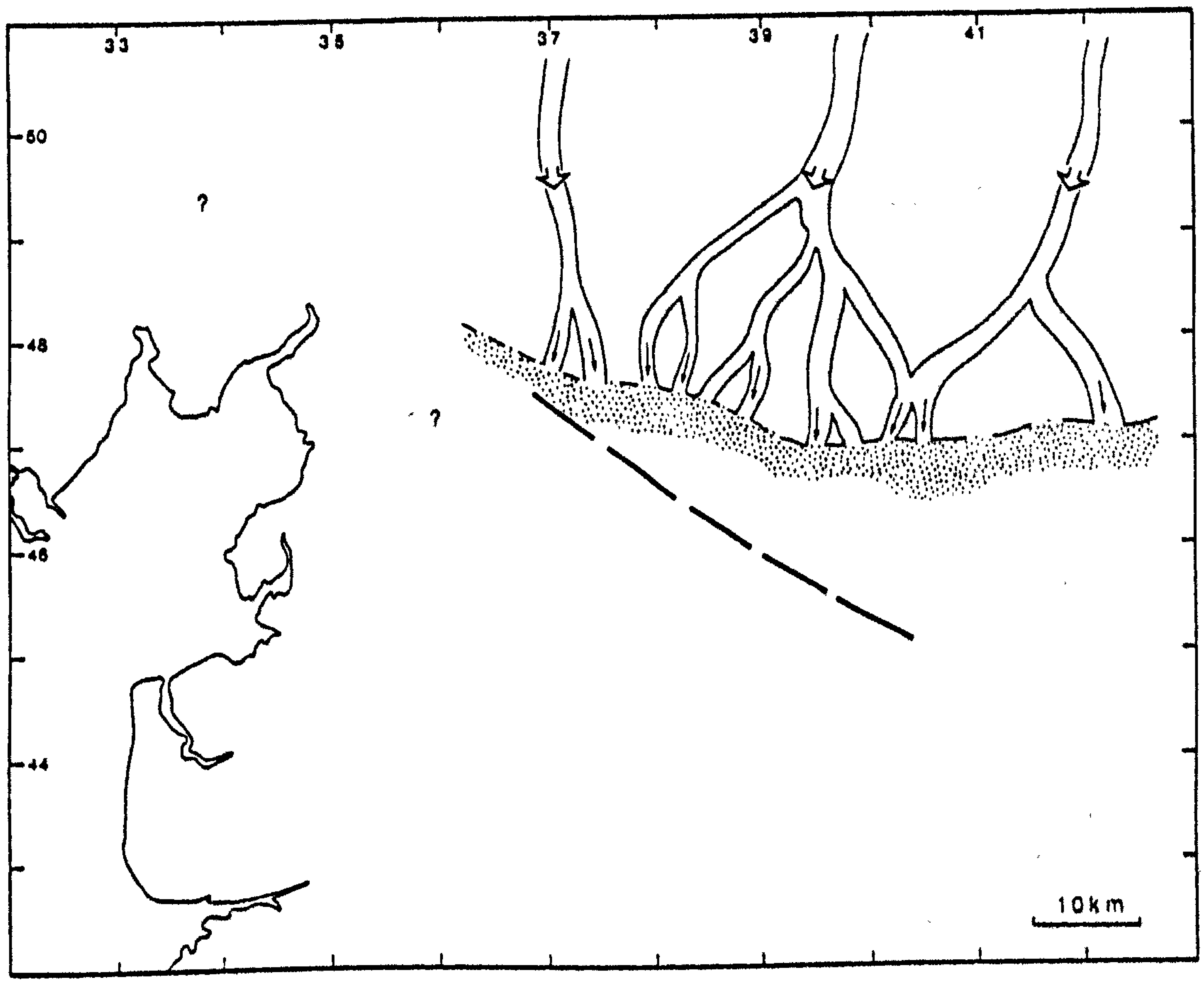
D)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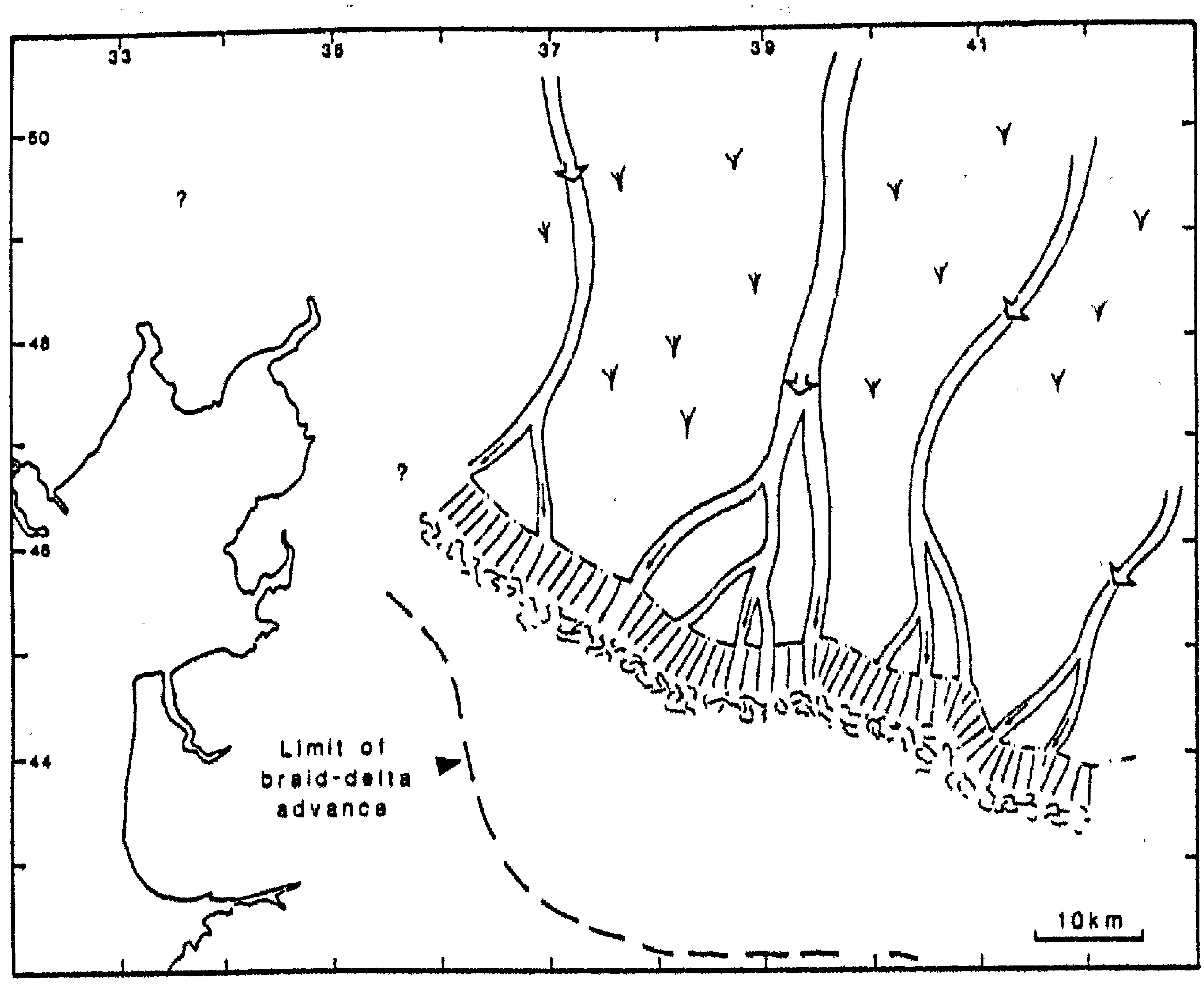
E)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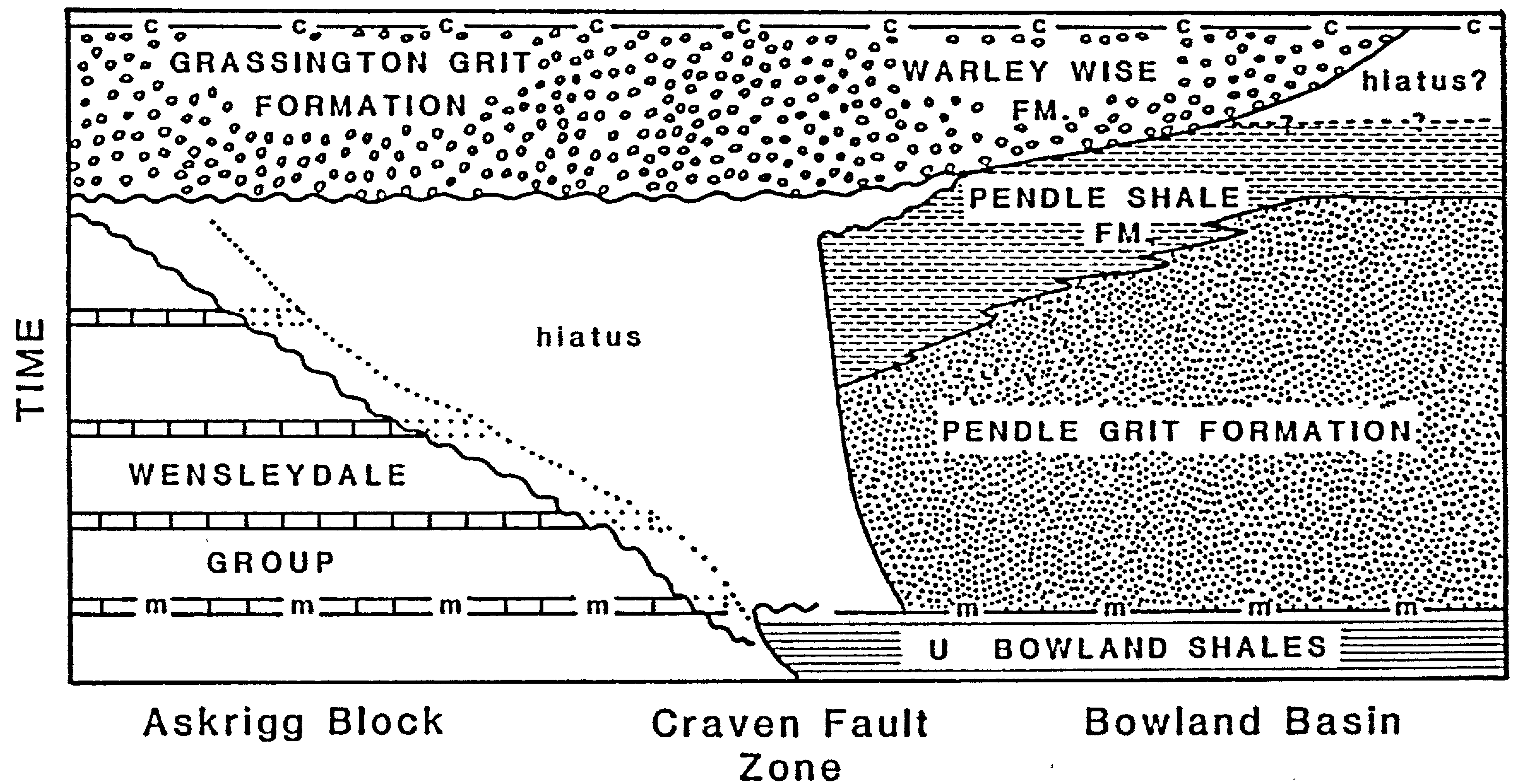

Figure 6.8: Inferred time stratigraphic relationships of the E_{1c} sediments on the Askrigg Block and in the Bowland Basin. Dotted lines show the proposed pre-unconformity offlap of the Wensleydale Group cyclothems. Section runs approximately NNW-SSE and is not to scale.
-c-c- *Cravenoceras malhamense* marine band. -m-m- *Cravenoceras cowlingense* marine band.

Initiation of the turbidity currents which supplied the Pendle Grit Formation fans was probably entirely due to river flood events. As discussed in Chapter 3, there was significant variation in the magnitude of the turbidity currents on the fans. This is believed to have resulted from a variation in the intensity of flood events in the fluvial system. Major flood events produced large magnitude, long turbidity currents as sediment stored in the fluvial system was flushed out on to the steep topographic slopes forming the basin margins. These large magnitude flows cut first-order channels on the fan system and deposited thick amalgamated facies (Chapter 3). Less severe floods produced smaller turbidity currents which in-filled the channel complexes by the processes discussed in Chapter 3. Leeder (1988) summarises the evidence for a pluvial climate during the Namurian and shows that monsoonal conditions are highly probable. Such a climatic regime would clearly lead to rivers with strongly seasonal discharges: exactly what is required for the above mechanism to work effectively.

The palaeogeographic reconstructions above (Figures 6.7a-c) all show a minor sediment supply to the Bowland Basin from the Askrigg Block. The Wensleydale Group cyclothems on the Askrigg Block almost certainly reached close to the Block edge during the earlier phases of Pendle Grit Formation deposition. At this time it is thought likely that mud and silt was carried into the basins by hyperpycnal flows sourced from the Wensleydale Group deltas. This may explain why the Pendle Shale Formation is particularly well developed close to the Block boundaries.

6.4.2 Intra-E_{1c} palaeogeographic changes and unconformities.

The intra-E_{1c} unconformity marks a renewed phase of tectonism which influenced both the Askrigg Block and the Bowland Basin. On the Askrigg Block the unconformity is a sub-planar, sub-horizontal surface across which there is little angular discordance (Dunham and Wilson, 1985; Rowell and Scanlon, 1957). Maximum erosion took place across the south-east corner of the Block (Figure 2.10 inset) suggesting gentle tilting about a north-west to south-east trending axis. A similar sub-planar erosion surface is seen cutting across the intra-basinal highs in the Bowland Basin. Here, however, the unconformity truncates fold structures within the Pendle Grit Group (Arthurton *et al.*, 1988). These folds generally overlie the buried Dinantian fault blocks and probably formed by minor normal faulting on the bounding faults of these structures: the cover sediments taking up the extension by forming drape folds. Renewed down-to the-south faulting took place on the Craven Faults at this time (Arthurton *et al, op. cit.*).

E_{1c} sedimentation on the Askrigg Block suggests that this phase of regional tectonic activity lasted for much of the Pendleian: isopachs show dramatic thinning of E_{1c} cyclothems in the Wensleydale Group onto the south-east corner of the Askrigg Block (Dunham and Wilson, 1885; their Figure 12). Similarly, the younger cyclothems appear to be progressively more confined to the north-west corner of the Askrigg Block, suggesting that tectonic tilting may have been continuing throughout the Pendleian[1]. Production of the intra-E_{1c} unconformity represents the culmination of this process. It seems possible that a minor eustatic regression may have coincided with this final tectonic pulse and that the two processes combined to cause significant erosion on fault block crests. While there is no direct evidence for such a regressive event it does explain why apparently minor tectonic activity caused such widespread erosion.

The current author believes that the end result of these combined tectonic/eustatic changes was a sudden palaeogeographic rearrangement in north-west England. The fluvial system supplying the Pendle Grit Formation fans avulsed (or was captured) to flow further eastwards. It now debouched directly onto the Askrigg Block (Figure 6.7d). Rapid progradation of the fluvial system across the Block initiated deposition of the Grassington Grit Formation. Obviously, the avulsion of the river system also cut off supply to the Pendle Grit Formation fans which became inactive. The implied time relationships of the events discussed above are summarised on Figure 6.8.

The origin of the planar erosion surface which forms the intra-E_{1c} unconformity is enigmatic. Planar erosion surfaces are often associated with marine erosion but there is little evidence that marine processes in the Namurian basins were able to cause reworking of loose sediment, let alone erosion of consolidated limestones within the Wensleydale Group. An alternative is that

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[1] Note that Leeder and Strudwick (1987) have argued from sedimentological data that the cyclicity in the Wensleydale Group is best explained with periodic tectonic subsidence followed by delta progradation causing regression.

the unconformity represents the base-level of fluvial channels in the Grassington Grit Formation: erosion was caused by in-channel processes during lateral migration of these channels. This explanation may be feasible for the Askrigg Block sequences but can not explain the presence of the unconformity across the Waddington Fells High where mass-flow deposits of the Warley Wise Formation braid-delta overlie the unconformity. Perhaps the unconformity was cut by different processes across different structures.

6.4.3 Final basin-filling: deposition of the Grassington Grit Group.

As a consequence of the avulsion of the fluvial system eastwards onto the Askrigg Block, a series of major channels were cut into the Wenlseydale Group sediments. These channels at the base of the Grassington Grit Formation are well documented (Wilson, 1957; Dunham and Wilson, 1985) and show up on the isopachs of the Formation as linear "thicks" (Figure 2.11). They probably remained open for some considerable time while sediment was transported through them into the braid-deltas along the southern margin of the Askrigg Block (Figure 6.7d). Progradation rates were very high in the relatively shallow water across this margin: the shoreline pushed rapidly southwards. However, water depths increased substantially across the Craven Fault Transition Zone, showing that Pendle Grit Group deposition had only partially filled the topographic relief between block and basin. Judging by the increase in the thickness of the Grassington Grit Group across this fault zone (Figure 2.11) water depths in the basin may have been around 100-150m.

Once into this deeper water, progradation rates slowed and the braid-delta changed style (see Chapter 4). Mass-flow aprons built up at the foot of the braid-delta slope which, itself became an avalanche face. Progradation of the resulting "Gilbert" deltas deposited the Warley Wise Formation and pushed the shoreline progressively southwards into the Bowland Basin (Figure 6.7e). Meanwhile, compaction and regional subsidence allowed vertical accretion of sediment on the Askrigg Block. The fluvial channels aggraded and "delta top"/flood plain environments were developed.

Evidence for continued tectonic influence on sedimentation is slight. However, destratification due to violent water escape shortly after deposition is common in Grassington Grit Group sandstones along the line of the Craven

Fault System. This may indicate some seismic activity on the fault systems. Similarly, a pronounced thickening of the Grassington Grit Group on both sides of the Bowland Basin seems to coincide with the Knotts Fault Zone (use Enclosure 3 as an overlay for Figure 2.11). This thickening may have resulted from further movement or differential compaction across this fault zone. Palaeocurrents from the Grassington Grit (Figure 6.6) show little other variation which requires a tectonic explanation. They demonstrate a generally south-easterly palaeoslope in the Bowland Basin which is consistent with continued regional subsidence centred further south in the Central Pennines Province (See Figure 6.2). Obviously, the palaeocurrents also pick out the local palaeoslope across the Craven Fault Zone.

By the end of E_{1c} times progradation of the Warley Wise Formation braid-deltas had infilled most of the Bowland Basin: the shoreline had advanced well to the south (Figure 6.7e). Sedimentation was finally halted when the end E_{1c} transgression re-established marine conditions across both the Askrigg Block and the Bowland Basin. With the deposition of the *Cravenoceras cowlingense* marine band the depositional evolution of the E_{1c} basin-fill sequence was complete.

6.5 Discussion

6.5.1 Eustacy.

The dynamic stratigraphy of the E_{1c} basin-fill sequence in the Lancaster Fells and Bowland Basins reflects the interaction of a sand-rich sediment supply and a deep topographic basin. Most of the evolutionary changes in this sequence have been interpreted as resulting from palaeogeographic changes brought about by tectonic activity: eustatic controls on sedimentation are not obvious. It could be argued that this is because the tectonic influences and high sedimentation rates overwhelmed any eustatic signature. Alternatively, eustatic variations might have had a time period greater than that of the E_{1c} goniatite zone. However, the Askrigg Block would have been very sensitive to eustatic variations because of the depositional surface being so close to sea-level. Equally, the evidence for early Carboniferous glaciation in Gondwanaland is growing (see review in Leeder, 1988): this should have resulted in cyclic glacio-eustatic sea-level fluctuations well within the period represented by the basin-fill sequence (Leeder, *op. cit.*). The interpretation of the Wensleydale Group cyclothems as tectonic in origin (Leeder and Strudwick, 1987) might now require revision in the light of this new evidence for eustatic change, as recognised by the authors (Leeder, pers. comm.). However, even if these cycles are eustatic in origin, there is little evidence of direct eustatic controls on sedimentation in the adjacent basins.

The current author is strongly of the opinion that such glacio-eustatic cycles played a significant rôle in controlling the evolution of the basin-fill sequence. The possibility of regression contributing to the erosion of the intra-E_{1c} unconformity has already been mentioned. Other glacio-eustatic influences were probably indirect. For example, the avulsion of the sediment supplying fluvial systems may have been promoted by base-level changes associated with transgression and regression. Similarly, Holdsworth and Collinson (1988) have argued that a major indirect effect of eustatic cycles on the Central Pennine Province was desalination of the basins during lowstands of sea-level. At such times, fluvial discharge may have been sufficient to displace marine waters from the mainly land-locked Carboniferous basins (See Figure 5.6 for a generalised palaeogeographic map of the U.K during the Carboniferous). During highstands the enlarged connection to the main Carboniferous oceans would allow re-establishment of marine conditions and the influx of goniatites. Marine bands, then, would correspond to deposition during highstands. If this argument is correct, it has very significant sedimentological implications: turbidity currents would be significantly more powerful during lowstands because of the greater density contrast with the low-density, fresh water in the basins. Initial turbidity current generation by fluvial processes might also be enhanced as hypo- rather than hyperpycnal flows would dominate at river mouths. Most of the Wensleydale Group cyclothems contain well developed shelly faunas which indicates marine conditions were common on the Askrigg Block during Pendle Grit Formation deposition. However, most cyclothems also contain barren sediments which might indicate some limited periods of desalination. It is interesting to speculate whether the cutting of first-order channels on the Pendle Grit Formation fans might have been enhanced by generation of powerful turbidity currents during low-salinity periods.

6.5.2 Tectonics.

The origin of the subsidence and subsequent uplift in the Carboniferous basins of the British Isles is currently hotly debated. Two schools of thought hold centre stage in this discussion. One school favours a phase of McKenzie type extension (McKenzie, 1978) caused by back-arc stretching north of the Hercynian collision zone (e.g. Leeder, 1988; Leeder and McMahon, 1987). The other school explains Carboniferous extension in terms of rifting related to an early phase of sea-floor spreading in a proto-North Atlantic ocean (e.g. Russell, 1976; Hazeldine, 1984). Strike-slip controlled basin formation has also been proposed (Arthurton; 1983, 1984). The data presented in this work have some bearing on the mechanisms of basin formation and uplift. The following points are significant:

1) Repeated phases of mainly normal faulting occurred over a period of at least 25m.y. from the time of basin initiation to the middle of E_{1c} times. If this faulting is the result of crustal extension then simple McKenzie type models (which have instantaneous stretching followed by thermal relaxation of the lithosphere) are inappropriate. Models which include finite extension rates must be considered (e.g. Jarvis and McKenzie, 1980).

2) β-values calculated from the sediment accumulations in the Carboniferous basins of northern England imply a doubling of the crustal length between the limits of the stretched zone (Leeder and McMahon, 1987). There is little evidence for this sort of extension on faults known to have been active during the Carboniferous: all Carboniferous faults in northern England have relatively small heaves at outcrop. Unpublished seismic data examined by the current author do not suggest that extra extension is present at depth due to the faults having listric profiles. Consequently, if the sediments in the Carboniferous basins do result from a lithospheric stretching event they must mainly represent the effects of the thermal sagging: the bulk of the upper crustal stretching must be accommodated elsewhere. Gibbs (1987) shows how this can be done by transferring extension from the lower lithosphere into the upper crust on low-angle shear zones. The resulting upper crustal extension can be

significantly displaced relative to the lower crustal thinning by such a linked model (Figure 6.9). However, areas of crust with large amounts of proven Carboniferous extension are not seen in the British Isles. Either this extension was occurring in areas now underwater (e.g. the West Shetland Platform, Rockall Trough or the North Sea) or the subsidence seen onshore is not related to a McKenzie-type stretching event.

3) The end-Carboniferous uplift in the Bowland Basin cannot be accounted for simply by reverse movements on the earlier normal faults across which subsidence took place. The inversion event is regional and involves uplift of upto 3km (c.f. Figures 5.12-5.14).

In the light of point (2) above, there is clearly a need to consider whether stretching models are appropriate for explaining the origin of the Carboniferous basins of northern England. Other models for basin formation which can produce subsidence without major extension need to be examined and tested. For example, Carboniferous volcanic activity has always been explained as a side-effect of extensional basin formation. However, such volcanism could also relate directly to the cause of subsidence: changes in the heat flux into the lithosphere caused by igneous activity are capable of causing regional changes in surface elevation, as McKenzie (1978) noted. There is extensive volcanic activity associated with Carboniferous sedimentation in both the Midland Valley of Scotland and the Northumberland Trough. Similarly, the Devonian was a period of intense volcanism. Such activity represents a high heat flux into the lithosphere. If this activity was mirrored by intrusion at greater crustal depths then the lithosphere would have been substantially thinned. Thermal decay following the end of these phases of intrusive activity could produce regional subsidence on the scale seen in the Carboniferous Basins. Obviously, the thermal decay would look identical to the "sag" phase of a McKenzie-type stretching event. While this mechanism is not proposed as *the* answer to Carboniferous development it has been raised as one possible alternative to a McKenzie type stretching event: until there is a proven causal link between the later Carboniferous regional subsidence and the earlier block faulting the idea of a stretching event must be regarded with caution.

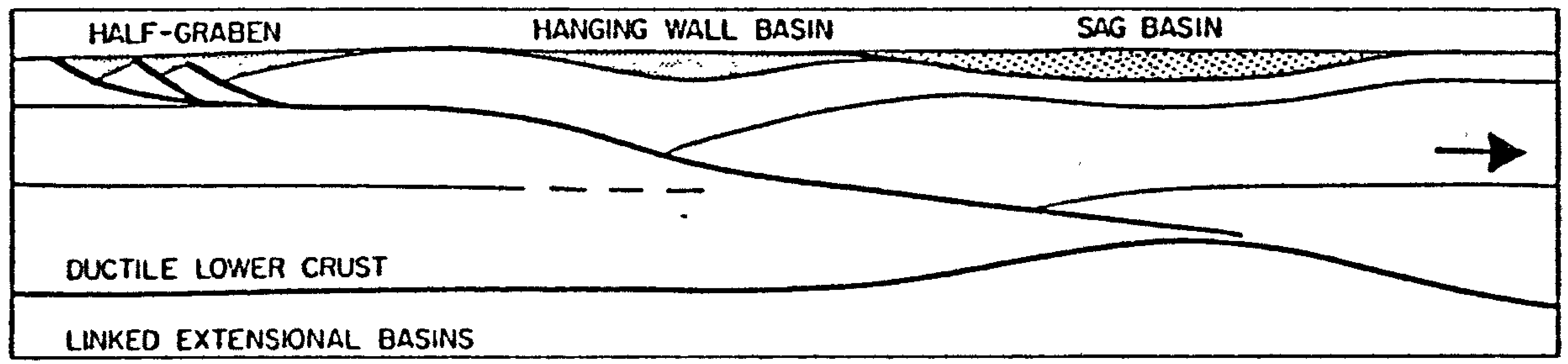

Figure 6.9: Model of linked basins formed by a crustal stretching event. The model shows that upper and lower crustal extension need not occur at the same location. In this case the extensional half-grabens in the upper crust are offset from the thermal "sag" basin which relates to the extension in the lower crust. Note that there are no major extensional faults associated with the "sag" basin.
From Gibbs (1987). Horizontal scale 200-300km.

The regional uplift of upto 3km at the end of the Carboniferous appears to affect the blocks and basins of northern England equally. This is confirmed by vitrinite reflectance data which show little variation across the major block-bounding structures (H. Clemmey, pers. comm.). Consequently, the basin inversion cannot be explained by reverse movement on these structures alone. Previously, the inversion has been related directly to Hercynian collision tectonics to the south of Britain (Hudson and Turner, 1933; Leeder and McMahon, 1987; Leeder, 1988). However, it is possible that the regional element of the inversion was caused by a mechanism entirely separate to that which caused the folding in the foreland to the Hercynian front (which lay through South Wales). The current author accepts that the effects of the northerly migrating Hercynian thrust front are seen in northern England: shortening in the foreland was accommodated mainly by folding in the Carboniferous basins with strike-slip movement along the block/basin boundary faults transferring this shortening northwards. Shortening was particularly intense in those basins where block bounding faults changed direction or died out. The Bowland Basin, in which this situation is most marked (due to the sudden north-westerly disappearance of the South Craven Fault) was one of the most strongly shortened basins, resulting in formation of the Ribblesdale foldbelt. While the shortening may have caused minor uplift, the bulk of the regional basin inversion is believed to relate to a thermal event which caused thinning of the lithosphere north of the Hercynian front. This thermal event may be related to a phase of attempted seafloor spreading to the west of Britain (Russell and Smythe, 1978). The additional heat flux into the crust caused by igneous activity during this event may have caused crustal thinning and regional doming, as in the East African Rift System today. The north-south rifting which occurred on the west side of Britain during the Permian may be the culmination of such a failed spreading episode. The current author, then, does not agree with Leeder's statement that

"the massive (Carboniferous) inversion event is clearly not compatible with the North Atlantic rifting theory (of Russell and Smythe, 1978 and Hazeldine, 1984)"

(Leeder, 1988 page 83): there is still room within the data to explain the inversion by crustal heating resulting from a failed spreading event in the North Atlantic.

A high heat flux into the Hercynian foreland, caused by any mechanism, might also help explain how Hercynian deformation was transferred northwards along the block/basin marginal structures. Heating of the crust would have decreased its strength which, in turn, may have enabled reactivation of basement lineaments to have occurred. Northwards transfer of Hercynian compressive stresses could thus have been greatly enhanced by crustal heating.

In summary, the above discussion has attempted to broaden the argument regarding the formation and inversion of the northern England Carboniferous basins. A few problems with simple extensional models have been pointed out although it is accepted that thermal subsidence played a major rôle in basin formation. With regards to basin inversion, it is necessary to try and separate cause and effect: the regional uplift may have influenced reversed reactivation of basin-bounding faults but cannot be explained by this process. Consequently, a more regional uplift mechanism must also have been operative. We know that Hercynian collision was occurring to the south during the inversion of the Carboniferous basins and that immediately post-inversion a series of major north-south rift systems on the west side of Great Britain. Ultimately, trying to link the inversion of the Carboniferous basins to either Hercynian compression or to the subsequent rifting alone is too simplistic: all the events must be related and form a continuum of processes. This is a clear consequence of a whole world model such as the Plate Tectonic paradigm. Consequently, the inversion must be looked at as a multicause event, perhaps along the lines of the model suggested above. It is vital to start considering such multi-cause models and to test their predictions against the large amount of data available. This work must clearly involve geologists of many different disciplines but its results might lead to major revisions of our ideas about end-Carboniferous palaeogeographies.

6.6 Conclusions.

1) At the beginning of E_{1c} times the Lancaster Fells and Bowland Basins formed a marine trough up to 600m deep. Steep, fault-controlled margins to the basins rose up onto the shallow-marine Askrigg Block.

2) Deposition of the sand-rich fans of the Pendle Grit Formation began in the Lancaster Fells Basin following the initiation of a coarse clastic input from the north-west of the basin.

3) This sediment supply was probably from a major fluvial system which avulsed onto the area at the beginning of E_{1c} times. This system probably flowed across the Lake District Block.

4) The Pendle Grit Formation fan expanded rapidly to fill the Lancaster Fells Basin and, as sedimentation proceeded, spilled over into the Bowland Basin. Deposition then occurred mainly by vertical aggradation, leading to partial infilling and shallowing of the basin.

5) Further avulsion of the fluvial system switched off sediment supply to the Pendle Grit Formation fans and began deposition of the Grassington Grit Group on the Askrigg Block. This major palaeogeographic change may be related to renewed tectonic activity which caused tilting of the Askrigg Block and rejuvenation of the block bounding faults.

6) The intra-E_{1c} unconformity is a sub-planar erosion surface cut during the culmination of this tectonic activity.

7) Basin-filling was completed by progradation of the Grassington Grit Group braid-deltas off the Askrigg Block and into the Bowland Basin.

8) Tectonic controls on E_{1c} sedimentation in the Lancaster Fells and Bowland Basins appear to be more important than direct eustatic controls. However, indirect effects of glacio-eustacy may have been significant.

9) The wider implications of this study with respect to mechanisms of Carboniferous basin formation and inversion have been discussed. Simple McKenzie-type extension models are not entirely compatible with the geological evolution of the basins. Similarly, the regional end-Carboniferous inversion may not be causally related to the Hercynian compression which generated the Ribblesdale foldbelt.

CHAPTER SEVEN

CONCLUDING REMARKS

Rather than reiterating the conclusions drawn at the end of each chapter, these concluding remarks seek to highlight the significant geological advances made during work on the project reported on herein. Some of the areas which the project has shown to require further investigation are also pointed out.

The geology of the E_{1c} basin-fill sequence discussed in this work was previously poorly documented: the stratigraphic nomenclature was confused and incomplete, sedimentological data was sparse and the relationships of sedimentation in the Lancaster Fells and Bowland Basins to that on the Askrigg Block was uncertain. Within the constraints set by the lack of internal time markers, this work has begun to remedy this situation. A consistent stratigraphic nomenclature for the E_{1c} sediments has been proposed which is applicable to the whole of the basin complex. As the individual stratigraphic units have been defined on the basis of a sedimentological understanding of the sediments, the nomenclature should be capable of projection into the sub-surface south of the Pendle Monocline (where additional E_{1c} are predicted to exist). The formal definition of the stratigraphic units made herein has been a worthwhile step forward: it provides a solid framework upon which other studies can be based. The isopach maps for the E_{1c} stratigraphic units presented herein, while very crude, are the only attempts made so far to illustrate the regional thickness changes in these sediments.

Most of the new data in this thesis are sedimentological and petrographic details of the Pendle and Grassington Grit Group sediments. The interpretation of the data from the Pendle Grit Group in terms of a new style of sand-rich submarine fan is the most significant result of this project. Features within this fan system, particularly the channel hierarchy and the intra-bed complexity, have not previously been described in the literature on ancient fan systems. The channel complex facies model proposed in Chapter 3 is, therefore, a complete departure from earlier facies models for submarine fans. Notable features of the new model are its lack of lobe deposits, the low morphological variation predicted on the fan and the dominance of in-channel deposition. The model was developed on a non-actualistic basis with fan morphology being inferred from outcrop observation rather than being based on comparison with modern fans. Similarly, the suggested variation in the power spectrum of turbidity currents in the fan system was based on the hierarchy of erosion surfaces seen in the field rather than on current data measured in modern submarine fan environments. The recent description of an active sand-rich fan in Bute Inlet, Canada (Prior *et al.*, 1987) provides confirmation that the processes and morphologies predicted by the channel complex model do exist in the real world. It is now necessary to see if it can be used as a predictive tool to explain the sedimentary geology of other sand-rich fan systems. It has been suggested (Chapter 3) that some of the Jurassic and Tertiary fans of the North Sea basins may be suitable systems upon which to test the model. At outcrop, systems like the Gres d'Annot of the French Alps might also be suitable. Even if the channel complex model as a whole breaks down, the concept of a channel hierarchy may prove of great help in understanding sandbody development in these systems.

The discussion of the flow mechanics of high-concentration turbidity currents in Chapter 3 has attempted to highlight the problems of interpreting transport processes from the facies deposited from these flows. Previous authors have pointed out that transport mechanisms in turbidity currents evolve with time and space and have suggested the existence of flows in which more than one grain-support mechanism is operative at any one time. The concept of turbidity currents with a definite "boundary layer" flow supported by processes other than turbulence does not appear to have received detailed treatment in the literature. Hopefully, the qualitative discussion herein will lead to detailed physical and mathematical modelling to see whether or not such flows can exist in reality. The modelling of such turbidity currents should be possible using large scale finite element computer simulations. While developing such simulations will require massive investment in time and computer power, the current author believes that this approach stands the greatest chance of producing real progress in our understanding of deposition from high-concentration turbidity currents.

The reconnaissance study of the Grassington Grit Group sediments has significantly expanded the data available on this stratigraphic unit. Of particular interest is the change in style of the braid-delta systems seen in this Group as they prograde from shallow to deep water. This case study is important because the change in style can be attributed entirely to changing water depths: there is no evidence for changes in sediment supply, tectonic activity or basin energy. Further study of this transition may provide valuable insights into how the change in processes occurred. However, poor exposure in the critical areas may limit the amount of data which can be collected.

A possible palaeogeographic evolution of the Lancaster Fells and Bowland Basins during E_{1c} has been discussed in this work. Perhaps the most notable feature of the proposed palaeogeography is the north-westerly source of feldspathic sediment and the eastwards migration of this sediment source with time. This does, in fact, fit well with regional temporal changes in sediment type. During the Dinantian the northern British basins were dominated by a north-easterly source of quartzitic material which gradually spread southwards from the Northumberland Trough onto the Alston and Askrigg Blocks. Meanwhile, basins in north-west Ireland were dominated by feldspathic sediments supplied from the north-west. The Namurian marks a time of rapid expansion of feldspathic sedimentation across the British basin while quartzitic sedimentation was supressed. Feldspathic sedimentation is seen first in the Lancaster Fells and Bowland Basins during early E_{1c} times but, subsequently, it spread eastwards to include the Askrigg Block and then the rest of the Central Pennines Province. This sweep of feldspathic sediment east and southwards is accompanied by changes in the palaeocurrents: later sediments were supplied from a progressively more north-easterly direction. The gross expansion of the feldspathic source eastwards, and the concommitent cessation of quartzitic sedimentation, must reflect long term changes brought about by changes in general palaeoslope of the hinterlands, possibly a gradual tilting from west to east causing eastwards migration of fluvial systems. The controls on these hinterland changes and the interaction of the two types of sediment remain the least understood aspects of Carboniferous geology. There is considerable scope for further study of this problem in the E_{1c} sediments of the Bowland Basin and the Askrigg Block: both sediment sources were active at the same time and within a few kilometres of each other.

The lack of precise time-control within the E_{1c} sediments of northern England has been a major limit on the accuracy of the palaeogeographic modelling and, to a lesser extent, on the sedimentological examination of these sediments: the relationships of the Pendle and Grassington Grit Group to each other and to the intra-E_{1c} unconformity are still unclear. This is a real problem because an understanding of the events which led to the production of this unconformity and to the break down of the "block/basin" palaeogeographic provinces in north-west England is critical. The current author is of the opinion that solving this particular set of problems will provide the key to answering the wider questions of how extra-basinal events controlled the evolution of the Carboniferous basins. While there is scope for more detailed fieldwork to address these problems, the lack of time control will always be a stumbling block. There is a definite need for detailed palynological and micropalaeontological studies of the E_{1c} sediments to provide a good local biostratigraphic framework. It is strongly recommended that such work be given a high priority.

The possibility that there were salinity changes in the E_{1c} basins also requires further research. Salinity changes might be expected to produce a geochemical signature in shales both on the Askrigg Block and in the basins. If so, then this might provide another means of time-correlation within these sediments, assuming a eustatic origin for the salinity changes (Holdsworth and Collinson, 1988).

To conclude, this thesis has presented much new data on one of the least studied "Millstone Grit" basin-fill sequences in the Central Pennines Province. The sequence is significant in that it marks the onset of major clastic deposition in this Province and also because the relationships of the basin-fill sequence to time-equivalent sedimentation on an adjacent "block" can be studied. The basin-fill sequence also includes a type of sand-rich submarine fan which has not previously been documented. Hopefully, the ideas and data presented herein will provide a firm base for further detailed study into the Namurian palaeogeographic evolution of north-west England and the nature and processes on sand-rich submarine fans.

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APPENDIX A

A DISCUSSION OF BLOCKS AND BASINS IN CARBONIFEROUS STRATIGRAPHY

A.1 Introduction.

As a consequence of the differentiation of Carboniferous Britain into "blocks and basins", there are marked regional variations in the thicknesses and facies of Carboniferous sequences. Indeed, it was the recognition of these gross variations which led to the original concepts of Carboniferous differential subsidence and, later, to models of extensional tilt-block control on facies development. Unfortunately, well-exposed transitions from block to basin are uncommon: it is rare to find time-equivalent sections exposed both on the blocks and in their surrounding basins. This is because of preservation of younger Carboniferous strata in the basins or erosion of the block sequences. However, across the boundary between the Bowland Basin and the Askrigg Block time equivalent sediments are exposed in rocks of both Dinantian and Namurian age. By understanding the stratigraphic development across this tectonic hinge zone, constraints can be placed on the timing of local tectonic activity which might influence sedimentation both on the Askrigg Block and in the Bowland Basin. Similarly, knowledge of Carboniferous eustatic sea-level variations derived from the stratigraphy of the Askrigg Block can help in the development of stratigraphic models for the basin-fill sequence.

An understanding of the block and basin concept in Carboniferous stratigraphy, then, is vital when considering the stratigraphic and sedimentological evolution of the E_{1c} basin-fill sequence. Consequently, this Appendix discusses the development of the concept in some detail. It also covers the basis upon which blocks have been identified and the methods of stratigraphic correlation from block to basin.

A.2 "Blocks and Basins" and their effect on Carboniferous stratigraphy.

Much of the early data on thickness and facies changes within the Carboniferous came from work on the Dinantian Limestone districts of Derbyshire and the Northern Pennines. This was due to the combination of good exposure and the presence of economically important mineralization in the limestones. It was from the study of the Northern Pennines that the concept of blocks and basins was to evolve. Phillips (1836) was the first to record the lateral variation in the Carboniferous rocks of this area. He noticed the upwards transition of the Mountain Limestone into Yoredales cyclothems in the Yorkshire Dales and recognised the apparent southwards passage of the Yoredales into thick shale sequences in the Craven Lowlands. This, he suggested, was due to a rapid gradation between the facies types in the vicinity of the Craven Fault system. He considered that the facies change represented the influence of two separate source areas, one supplying argillaceous material from the west and another supplying course clastics from the north. Following the first systematic mapping in the Pennines, Tidderman (1889, 1890) concluded that there was no evidence for a gradational facies change. Instead, he proposed that the two stratigraphic units were deposited in different "provinces" which were subsiding at different rates. The provinces were separated by the Craven Faults which he believed to be active during deposition. He was thus the first person to recognise the importance of tectonic controls on Carboniferous sedimentation in the area. Although Hind (1902) strongly disagreed with this interpretation of the relationships along the Craven Faults, the work of later authors produced more evidence for the existance of slowly subsiding blocks and more rapidly subsiding basins. Kendal (1911), for instance, described the Craven, Dent and Pennine fault systems and presented evidence for pre-Carboniferous activity on these zones. He also noticed the lack of deformation in the ground bounded by the faults. Marr (1921) further developed the concept of structurally simple blocks surrounded by fold-belts. He defined a Northern Pennines "rigid block" which was fault-bounded and unaffected by Hercynian folding (refer to Figure 1.5).

Trotter and Hollingworth (1928, 1932) first linked the ideas of Marr with the stratigraphic data. From mapping in the area extending from the northern half of Marr's "rigid block" into Northumberland they were able to show that sedimentation began later on the stable "Alston Block" than in the adjoining

"Northumberland Trough". They also recognised the change from limestone-dominated sedimentation on the Block to clastic-dominated deposition in the Trough. Meanwhile, Hudson had developed similar models for the southern part of the "rigid block". He found evidence of syn-sedimentary activity on the Craven Faults (Hudson, 1930a) and concluded that the variations in Carboniferous stratigraphy were related to the structure of the basement (Hudson, 1933): thin Yoredale cyclothems were confined to the Askrigg Block (named by Hudson in 1938) and thick shale and sandstone deposits to a geosynclinal trough in the Craven Lowlands. The Craven Fault zone was implicitly considered as the boundary between the block and the trough.

By the time of Rayner's (1953) review of the Dinantian rocks of northern England, the concept of blocks and basins was well established. Rayner linked the concept to Dinantian palaeogeography on a regional basis: troughs were either marine gulfs or clastic-sediment traps surrounding islands formed by the rigid blocks. In time, regional subsidence led to drowning of the blocks and the establishment of limestone deposition. Subsequent southwards progradation of deltas coming out of the Northumberland Trough led to deposition of the Yoredale cycles on the Blocks while the Bowland (Craven) Basin remained deep-marine.

At around this time geophysical investigations, particularly gravity and magnetic surveys, confirmed that the northern blocks were basement highs and advanced the idea that the stability of the blocks through time was due to the presence of granitic batholiths in the basement (see summaries in Bott *et al.*, 1957 and Bott, 1967). The subsequent drilling of the Raydale and Rookhope boreholes confirmed the presence of granites under the blocks.

Exploration drilling provided much more data on the Carboniferous thickness variations beneath the Mesozoic cover in eastern England and around the margins of the main coalfields (Lees and Taitt, 1946; Falcon and Kent, 1960). Kent (1966) used the borehole data to trace the pattern of blocks and basins throughout north-east England and suggested criteria by which blocks and basins could be defined.

His main criteria were:-

1) Lower Visean sediments are absent on blocks but present in the troughs.
2) Upper Visean sediments are shale dominated in the troughs but limestone dominated on the blocks.
3) Namurian and Lower Coal Measures sequences are expanded in the troughs, particularly those of Pendleian age.
4) Blocks appear as structurally undeformed zones and as structural highs on geophysical data.

Kent was careful to point out that "block" did not always imply faulted margins but rather a slowly subsiding area. Bott (1967) and Johnson (1967), however, implied that all the boundaries of the Northern Pennines blocks were faults (or faulted hinge zones) which had been active during Carboniferous sedimentation.

Kent's suggestion that the development of thick Namurian sediments implied the presence of a Carboniferous trough extended the block and basin concept beyond the foundations upon which it had been built: Namurian basins were now to imply the presence of underlying Dinantian basins. Although this is actually the case for all of Kent's troughs, it is not a general truth. For example, the Holme Chapel borehole in the middle of the Central Pennines Province (Figure 1.2) drilled through a thick, basinal Namurian sequence into Dinantian carbonates of block facies. The original concept of blocks and basins had developed from the recognition of large-scale palaeogeographical features which acted in the same fashion throughout the Carboniferous. As more data became available at the local level, so it became obvious that there were other tectonically controlled features which were not so stable with time. Consequently, more sophisticated models were necessary to explain the smaller scale and local variations in Carboniferous stratigraphy.

Kent (*op. cit.*) had already noted that the basins appeared to be asymmetric, with one sharp margin and the other "an even transition from deep-basin to shelf". Miller and Grayson (1982) also noticed this asymmetry and proposed that Dinantian facies changes could be modelled more effectivly using a tilt-block model in which basins were simply the the rapidly subsiding hanging-wall blocks of major normal faults. Collinson produced predictive models for Carboniferous facies variations around such blocks in confidential work for

Britoil plc, subsequently published in review form (Collinson, 1988). Further support for the new model came from Smith *et al.* (1985) who used gravity data to model the deep structure of the Derbyshire Dome as a series of major tilt-blocks which were active during the Carboniferous.

The power of the tilt-block model came from its predictive capacity when coupled with modern theories on the origin of extensional basins (McKenzie, 1978) and on the geometries of extensional faults (Gibbs, 1984). In particular, the model predicted that there should be a series of sub-basins and intra-basinal highs within any major Trough. These minor features were necessary to solve the space problems associated with the extension of an inhomogenous crust, particularly when extension took place on faults with a ramp and flat or listric geometry. The model also predicted that some original "blocks" might later become "basins" due to continued extension on a ramp and flat fault system or due to the onset of a thermal subsidence phase (Mckenzie, *op. cit.*). The sequence in the Holme Chapel borehole was, therefore, explicable by the new models.

The Bowland Basin was the first area in which the tilt-block model was applied on a detailed basis. Gawthorpe (1985, 1987) showed that the thickness and facies patterns in the Dinantian sediments of the basin were best explained by topographic relief over small-scale intrabasinal fault blocks, some of which were reactivated during deposition. Using the facies patterns, together with an elementary gravity interpretation, he was able to map minor tilt-blocks within the basin (Figure 2.1). These tilt-blocks have a profound influence on the subsequent Namurian basin-fill. Lee (1988a, 1988b) refined the gravity interpretation for the Bowland Basin and extended the gravity analysis to the remainder of the Central Pennines Province. He recognised a series of gravity residual highs and lows from which he determined the distribution of minor Carboniferous blocks in the subsurface over the whole of central and northern England.

There is, then, a range of scales of "blocks and basins" affecting Carboniferous stratigraphy all of which affect the evolution of the particular E_{1c} basin-fill sequence discussed in this thesis. On the larger scale, deep-seated crustal features form regional highs which subsided slowly during the Carboniferous. These, the classic blocks, control stratigraphic variations at the Group level. For example, the Wensleydale Group (Yoredales, see Arthurton *et* *al.*, 1988) is confined to the Askrigg Block while the time-equivalent Pendle Grit Group is developed only in the Bowland Basin. The major blocks may also affect the basin stratigraphy at the Formation and Member level if they act as sediment sources to the basin margin. The smaller scale tilt-blocks within the major basins, and also on some of the blocks, cause stratigraphic variation mainly at the Formation and Member level and control the thickness of individual stratigraphic units.

A further difference between the scales of "blocks" is the timespan over which they remained significant. The classic blocks were major palaeogeographic features throughout the Carboniferous while the smaller tilt-blocks and intra-basinal highs tended to have progressively less stratigraphic effect with time.

A.3 Stratigraphic correlation between blocks and basins.

In the early stages of development of the block and basin concept, stratigraphic correlation relied entirely on lithostratigraphic methods. The two main horizon used across the country were the top of the Carboniferous Limestone (Mountain Limestone of Phillips, 1836) and the incoming of pebbly grits refered to the Millstone Grit. Between these horizons lithostratigraphic correlation was possible within palaeogeographic provinces but rarely across the boundary between provinces (See discussion in A.2). Hence, the original correlation of lithological units within the Wensleydale Group (Yoredales) of the Askrigg Block has changed little since the original surveys (Dakyns *et al.*, 1890 and 1891). Similarly, correlation within the Worston Shale-Bowland Shale interval of the Bowland Basin has remained fairly similar to that originally proposed (Pendleside Series of Hind, 1905). However, the correlation of a lithological unit in the Bowland Basin with a lithological unit on the Askrigg Block was a matter for much debate because of the lack of any time control.

Biostratigraphic correlation first became possible thanks to the recognition of a series of biozones within the Dinantian. These were based on changes in coral-brachiopod faunas (Vaughan, 1905). Use of this zonal scheme was only partially successful in solving correlation problems because of the facies dependence of the faunas: they are only well- developed in shelf limestones and, therefore, are mostly absent in the basinal successions of the Dinantian and the clastic sequences of the Namurian.

The major breakthrough in Carboniferous correlation was the development by Hind (1918) and Bisat (1924) of a zonal scheme based on goniatites. Goniatites, being nektonic, are found in all marine facies and can be used to tie together shelf limestones with basinal muds. In general, however, goniatites within the basinal mudstone sequences are restricted to thin bands surrounded by barren sediments. These "marine bands" are thought to represent the only times at which the basin waters were fully marine and, because of their European extent, are believed to represent times of eustatic transgression (See discussion in Holdsworth and Collinson, 1988). If this interpretation is correct then the marine bands approximate to time planes. Holdsworth and Collinson (*op. cit.*) have calculated that the time interval between Namurian marine bands is approximately 185,000 years. The time control given by marine bands allowed much more precise lithostratigraphic correlations to be made. The marine band stratigraphy for the Carboniferous (See Ramsbottom *et al.*, 1978) is still the main constraint on correlation between blocks and basins.

Marine bands tend to occur at the base of sedimentary cycles. Upwards through these cycles there is an increase in sand percentage, possibly representing regression prior to the next transgressive marine band. Ramsbottom (1977) recognised that the cycles were not all equally extensive over the palaeogeographic blocks. Using an approach closely analogous to that used by seismic stratigraphers, he mapped the relative extent onto the Askrigg and Alston Blocks of the marine bands associated with the cycles. In seismic stratigraphic terms this equates with mapping the relative coastal onlap of seismic events which, like marine bands, are thought to represent time lines (Vail *et al.*, 1977). His analysis showed that there is a series of larger cycles (mesothems) composed of groups of minor cycles. At the base of each mesothem minor cycles were confined to the basins while subsequent minor cycles extended progressively further onto the Blocks (Figure A.1). Genus level changes in the goniatite faunas appeared to occur only at the base of the mesothems, leading Ramsbottom to conclude that mesothems were related to long-period eustatic cycles. He thought the introduction of new faunas represented renewed transgression after a pronounced regression: the coastal onlap of the minor cycles representing the gradual return to high average sea-levels.

Holdsworth and Collinson (*op. cit.*) have argued that the mesothem hypothesis should be rejected on the grounds that similar patterns of coastal

Diagrammatic chronostratigraphical section of the Namurian successions of Figs. 4 and 5 redrawn with time as the vertical axis rather than thickness. The increasing hiatuses which develop northwards on the block area and the limited distribution of some cycles are now apparent.

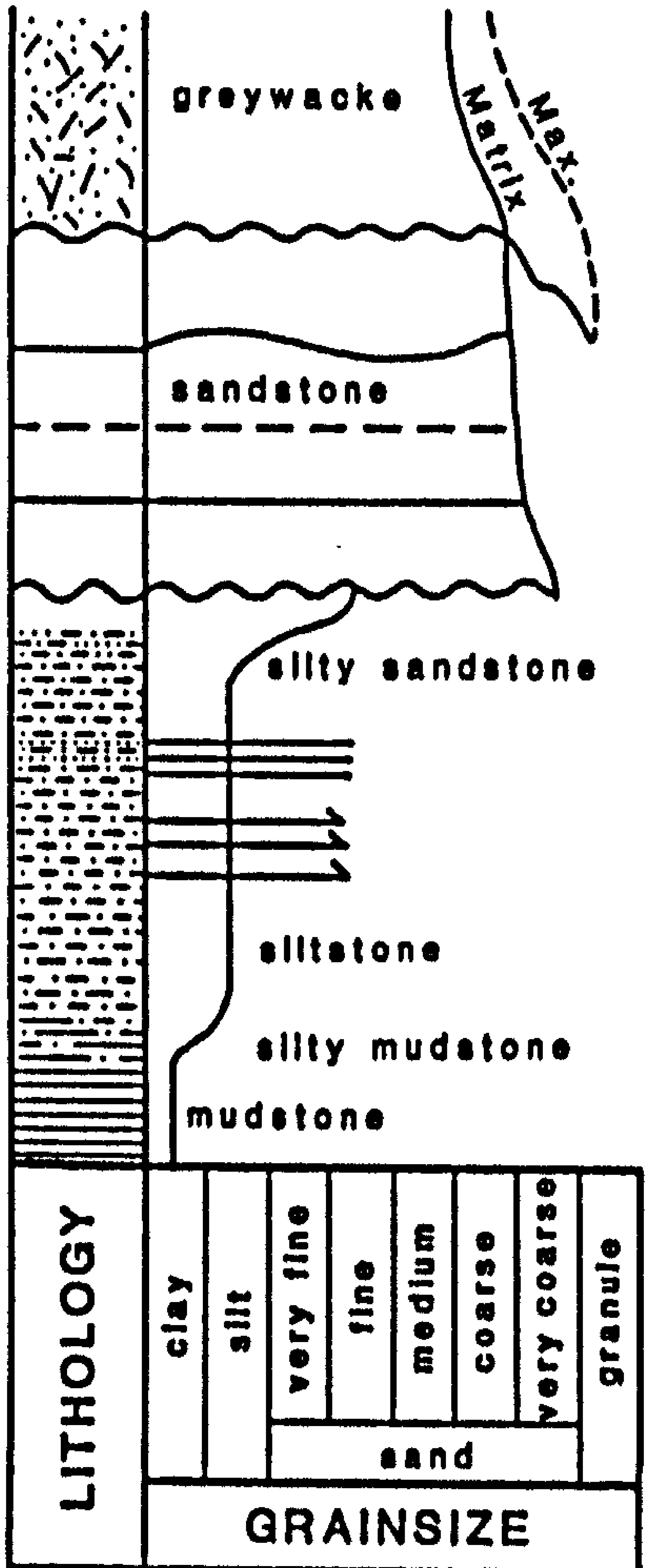
Figure A.1: Mesothems in the Namurian. Ramsbottom's original 1977 diagram (right) shows the progressive onlap of the Blocks during a mesothemic cycle. The relative sea-level curve (left) was drawn using the method of Vail *et al.*(1977) and shows at least three periodicities of sea-level variation. The longest period change (shown as a dashed line) is almost certainly eustatic in origin: the maximum lowstand corresponds in time with unconformities in Carboniferous sequences throughout the world. The shorter period variations, responsible for the details of the curve, are interpreted by Ramsbottom as also representing eustatic variations. However, it seems likely that they actually represent a combination of tectonic, sedimentological and eustatic events.

onlap are not seen on every Namurian shelf and that basinal sands (representing regressive events) are not equally developed at the top of each mesothem. Although Holdsworth and Collinson do point out some significant problems with the model, their arguments for rejecting the overall hypothesis fall open to the same criticisms as does Ramsbottom's original work: namely the failure to take sufficient account of local tectonic and sediment supply factors which could influence the detailed pattern of coastal onlap. Ramsbottom's approach was novel and suggested a method of combined litho- and biostratigraphic correlation which could prove of great importance in understanding the dynamic aspects of Carboniferous stratigraphy. The importance of Ramsbottom's "seismic stratigraphic" approach will be evident should seismic data become available in the Carboniferous basins.

Other zonal schemes based on palynology have been used for correlation purposes in the Namurian sequences. The miospore zones (Owens *et al.*, 1977) are not directly comparable with the goniatite zones and have much less resolution. They are, however, very valuable in areas where goniatites are absent and in marginal marine to non-marine depositional environments. Work by Campion (pers. comm.) suggests that a much more precise zonation may be possible. Such a zonation would be invaluable as it would allow for time control within the minor cycles between goniatite bands. Conodont zonations (Higgin, 1975) suffer from a similar lack of resolution but may, again, have potential in the future.

ENCLOSURE 1

GRAPHIC LOG



BED CONTACTS

Erosive, downcutting

Wavy

Amalgamated

Sharp

Erosive

Transistional

Interbedded

Micro fining-up units

LITHOLOGY MODIFIERS

Calcareous

Carbonaceous

Mc Micaceous

Pyritic

Fe Sideritic

SEDIMENTARYSTRUCTURES

Flutes, linear loads

Tool marks

Tabular cross-bedding

Trough cross-bedding

Ripples

Climbing ripples

Plane-lamination

Slurried

Load balls/pillows

Slumped horizon

Dish structures

Clastic dykes

Wavy bedding

Mudflakes

Pebbles

Dashes imply structure is crudely or indistinctly developed

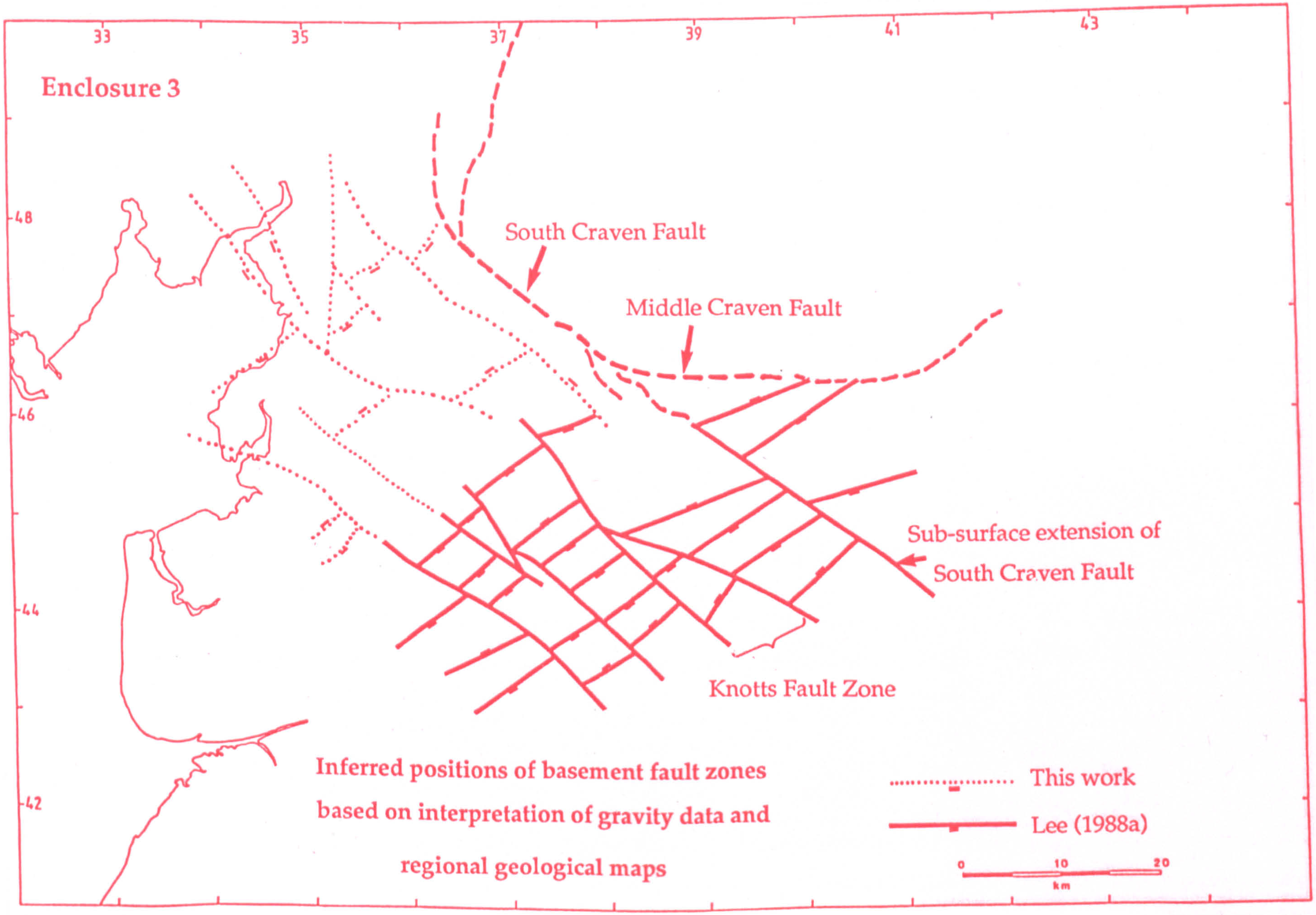
MISC.

Bioturbation

Goniatites

Bivalves

<u>Lingula</u>


Enclosure 3
33
35
37
39
41
43
48
46
44
42
South Craven Fault
Middle Craven Fault
Sub-surface extension of
South Craven Fault
Knotts Fault Zone
Inferred positions of basement fault zones
based on interpretation of gravity data and
regional geological maps
This work
Lee (1988a)
0
10
20
km

ENCLOSURE 3

ADDENDA

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Page 68 line 18: for inter-event read intra-event.
Page 111 line 31: for Neilson read Nilson.
Pages 165 line 15 and 192 line: for oligioclase read oligoclase.