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# The Sedimentology of the Old Red Sandstone Section from Pease Bay to Horse Roads Rock, Berwickshire

by

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## CERTIFICATE

I hereby certify that John Alan Fyfe has been engaged in part-time research and that he has fulfilled the conditions of the Resolution of the University Court, 1974, No. 2, and that he is qualified to submit the accompanying thesis in application for the degree of Master of Science.

E.K. WALTON

I certify that the following thesis is of my own composition, that it is based on research carried out by me, and that it has not been accepted in partial or complete fulfilment of the requirements of any other degree or professional qualification.

J.A. FYFE

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# ABSTRACT

The Upper Old Red Sandstone succession from Pease Bay to Horse Roads Rock comprises a sequence of sandstones with subordinate mudstones. The beds can be divided into six facies types: argillaceous beds, parallel-bedded sandstone, trough cross-bedded sandstone, trough cross-bedded sandstone with mudclasts, small-scale cross-bedded sandstone and unclassified poorly-bedded sandstone. These facies types are described in detail and interpreted with reference to several recent and ancient fluvial deposits described in the literature. Statistical analysis provides no evidence of cyclicity within the sequence but examination of the facies associations indicates the presence of a number of depositional units in the succession. Grain size analysis of the sandstones provides information in the form of cumulative frequency curves and cross-plots of various computed parameters. The facies types can, to some extent, be recognised by the shapes of the frequency curves though the cross-plots provide a clearer distinction, the most sensitive parameters being the measures of extremes of the distribution rather than of central tendency. Petrographic techniques applied to the sandstones include thin section microscopy, cathodoluminescence and scanning electron microscopy. These are used to determine provenance of the sediments and to examine diagenetic changes. Quartz petrography suggests a very varied sediment source. Much of the sand may be derived from the Lower Old Red Sandstone and may thus be a second- or third-cycle deposit. Palaeocurrent analysis reveals a south-easterly current direction, suggesting that the Pease Bay sequence represents a marginal development of the Midland Valley Upper Old Red Sandstone river system. A study of the carbonate and red bed diagenesis indicates lengthy periods of non-deposition, this being related to climatic controls and the marginal nature of the sequence within the basin.

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# 1. INTRODUCTION

## 1.1. LOCATION OF STUDY AREA

The fieldwork area is situated in southeast Scotland on the Berwickshire coast, some 34 km northwest of Berwick-upon-Tweed and 14 km southeast of Dunbar. The described section runs from the northern end of Pease Bay (NT 791711) to Horse Roads Rock (NT 789716). It is best reached by taking the minor road from Cockburnspath on the A1 down to the Pease Bay caravan site. Fig 1 shows the location of the study area on a map of southern Scotland.

## 1.2. OBJECTIVES OF THE STUDY

The primary aim of conducting a facies study on this section was to determine a palaeoenvironmental interpretation with particular reference to:

1. the palaeogeographic framework
2. the effects of early diagenesis on the sediments

## 1.3. OUTLINE OF PRESENTATION

The thesis begins with an overview of the geology of the Upper Old Red Sandstone in the Midland Valley Basin and the Borders Basin of southern Scotland, based on the work of several previous authors. The Pease Bay section is divided into six facies types and these are described in detail together with a transition probability analysis in Chapter 3. In Chapter 4, environments of deposition are proposed for these facies types with reference to other work carried out on both modern and ancient fluvial and terrestrial sequences. The facies associations are then discussed and the whole sequence divided into a number of depositional units based on these associations.

A number of petrological techniques have been employed to examine in further detail the sedimentological characteristics of the various facies types. Chapter 5 deals with the grain size analysis of the sediments. Frequency and cumulative frequency plots are drawn and discussed and various statistical parameters considered. Thin section petrology and cathodoluminescence are then discussed in Chapters 6 and 7 and Scanning Electron Microscopy in Chapter 8. These petrological techniques are used to give additional information in support of the facies models developed in earlier chapters. Chapter 9 deals with palaeocurrent analysis: the facies models are applied to cross stratal dip measurements made and the results analysed in the palaeogeographic context. Diagenetic considerations are discussed in Chapter 10 with emphasis on carbonate diagenesis of the early cornstone development and also red bed diagenesis. This study of early diagenesis leads to a palaeoclimatic model being developed for the sequence. Chapter 11 draws together all aspects of the research work and develops the palaeoclimatic and palaeogeographic models, leading to a discussion of the Upper Old Red Sandstone Province in the whole of southern Scotland - the Midland Valley and the Borders Basins.

## 1.4. APPROACH AND METHODS USED

The geological section was first described and a detailed log constructed. A synopsis of this log is presented in Fig 2. The section comprises cliff and shore outcrops, the beds dipping at around 20° to 25° to the northwest. In places the cliffs are sheer and the beds are only described where they reach the base of the cliff. The shore section consists of a number of seawardtrending reefs which are difficult to examine because of the seaweed which almost invariably covers these outcrops. The main desciption was carried out at the base of the cliff section and it was here that the measurements were made for the detailed log. In several places a close comparison between the cliff section and the shore section revealed that there was a good correlation between the two, the lateral continuity being apparently good over the small distances (several tens of metres) involved.

Grain size analysis was carried out in the Marine Geology sediment laboratory of the British Geological Survey in Murchison House, Edinburgh. Thin sections were examined using a Vickers petrographic microscope at the British Geological Survey. Cathodoluminescence was done in the Geology Department of St Andrews University using the luminoscope. Scanning Electron Microscopes were used in the Botany Department of Edinburgh University and the Gatty Marine Laboratory of St Andrews University. The Library of the British Geological Survey in Murchison House, Edinburgh was used for searching out bibliographic references. Computer programs, primarily for grain size analysis and palaeocurrent work, were run on the PDP 11/70 and GEC 4090 mini computers run by NERC Computing Services in Murchison House, an Apple IIe microcomputer and a personal Dragon 32 microcomputer. The text was prepared on the Apple II and printed using a Brother HR 15 daisywheel printer.

## 2. GEOLOGICAL OVERVIEW

### 2.1. INTRODUCTION

The term 'Old Red Sandstone' was derived from Werner's 'Oelte rother Sandstein' which in fact referred to the Rotliegende. Jameson (1805,1821) mistakenly applied the name to the pre-Carboniferous red beds in Britain and in this sense it has remained in use to the present day. A few years later, an attempt was made to resolve the confusion between Old and New Red Sandstone in southern England by Buckland and Conybeare (1824). Later, Sedgwick and Murchison (1839), while working on the marine strata of southwest England, first used the term 'Devonian' and applied this both to the limestones and the interbedded Old Red Sandstone. Murchison (1859) divided the Old Red Sandstone into the Lower, Middle and Upper groups and Goodchild (1903) first proposed that the Upper Old Red Sandstone might have been laid down in arid conditions. Macgregor and MacGregor (1936) gave the first major synopsis of the Old Red Sandstone in the Midland Valley and George (1960) published a detailed account of the stratigraphic evolution of the Midland Valley. He described the deposition of Lower Old Red Sandstone in early Devonian times followed by Mid Devonian tectonic movements and Upper Old Red Sandstone sedimentation in the late Devonian.

George (1960) records that the earliest Carboniferous rocks are Visean in age and postulates an unconformity between this and the Upper Old Red Sandstone, which he equated purely with the Upper Devonian. Clayton (1971) and Andrews (1978) offer miospore and vertebrate evidence respectively to suggest that the upper part of the Upper Old Red Sandstone is in fact Lower Carboniferous and a Tournasian age is proposed. This agrees with the evidence from the Upper Old

Red Sandstone of Greenland (Westoll 1951) where a Lower Carboniferous fish zone is found in rocks of Old Red Sandstone facies and is born out by the conformable nature of the boundary at the top of the Upper Old Red Sandstone (Smith 1968, Scott et al 1984).

In southern Scotland, Upper Old Red Sandstone sediments occur in two main areas - the Midland Valley and the Borders Basin (Fig 3). The sedimentary sequence at Pease Bay is probably influenced by the Upper Old Red Sandstone drainage systems in both of these basins. A palaeogeographic reconstruction of southern Scotland, based on Bluck (1978) and Leeder (1973), is shown in Fig 4. In order to identify the geological framework, an outline of previous work in southern Scotland is now presented, consideration being given first to the Midland Valley, secondly to the Borders Basin and lastly to previous work on the Berwickshire coast.

## 2.2. THE MIDLAND VALLEY

Upper Old Red Sandstone outcrops in the Midland Valley of Scotland occur in broadly four main areas - Ayrshire and the Clyde Coast, Stirlingshire, Fife and Kinross and the Pentland Hills. These areas may be identified as the major outcrops on the map in Fig 3 and a correlation diagram of the various sections is presented in Fig 5.

### 2.2.1. Ayrshire and the Clyde coast

Outcrops in the west of Scotland are dominated by coarse-grained fluvial sandstones and conglomerates. In Ayrshire, the Upper Old Red Sandstone lies unconformably on Lower Old Red Sandstone and Lower Palaeozoic rocks (Burgess

1960). The sediments are characterised by current-bedded sandstones and poorly-developed conglomerates. Near the base of the sequence the sandstones are interbedded with cornstones up to 4.5m (15ft) thick. Palaeocurrent directions from the cross-bedded sandstones indicate the main sediment source area to be southwest of the present line of the Southern Uplands Fault. Other significant source areas existed e.g. the Lower Old Red Sandstone lavas of the Carrick Hills and the Lower Palaeozoic sediments of the Lesmahagow Inlier.

Further north, on the Clyde coast, Bluck (1967) has recognised four types of conglomerate and conglomeratic sandstone deposit:

A. Alternating beds of laterally persistent conglomerates and thin sandstone: alluvial fan, sheet flood.

B. Single sets of planar cross-bedded conglomerate and thin sandstone in channels up to 12m (40ft) wide: alluvial fan, stream.

C. Large thicknesses of trough cross-bedded conglomerates and sandstones often in downstream-fining bodies: braided stream.

D. Thinly bedded, grain-supported conglomerates alternating with crudely cross-bedded sandstones: river channel, floodplain.

The sediments are interpreted as being laid down at the foot of fault-bounded highs, rapid fault movement having led to vigorous erosion and steep slopes, forming alluvial fans on the downthrown margin with sheetflood deposits grading upwards into stream deposits. The alluvial fan sedimentation is replaced by braided stream and floodplain stream deposition in response to the down-wearing and back-wearing of the source region. Bluck (1978) has suggested that these fault-bounded highs may be related to pull-apart basins forming on the southern margin of the Highland Boundary Fault by sinistral movement in an offset fault zone.

### 2.2.2. Stirlingshire

To the northeast, in Stirlingshire, the Upper Old Red Sandstone may be divided into two main groups (Read and Johnson 1967). The lower group, the Gargunnock Sandstones overlies the Lower Old Red Sandstone and comprises two lithologies:

a. conglomerates and pebbly sandstones containing clasts of Dalradian and Lower Old Red Sandstone affinity, underlying

b. uniform fine to medium grained cross-bedded sandstones with rare thin beds of red siltstone and silty mudstone.

Read and Johnson suggest that these beds may represent channel-bar deposits. The upper group of the Upper Old Red Sandstone, the Cornstone Beds, comprises a series of fining-upward cycles with pebbly and argillaceous horizons. Cornstones are common and many of the sandstones also contain cornstone clasts up to 13cm long. These beds are interpreted as point-bar deposits, suggesting a river channel of relatively high sinuosity. The authors suggest that some of the thicker sandstones within the Cornstone Beds may each represent the deposits of a whole meander belt. The cross-stratal dip measurements of the two groups both indicate dominantly southeasterly palaeocurrent directions. This is compatible with Bluck's overall model and with the Clyde coast palaeocurrent data.

### 2.2.3. Fife and Kinross

Further to the northeast, in Fife and Kinross, the Upper Old Red Sandstone is generally more fine-grained. Chisholm and Dean (1974) have divided the sequence into six formations. These rest unconformably on the Lower Old Red Sandstone sediments and lavas (Armstrong and Paterson 1970). In the basal Burnside Formation, pebbles up to 15cm across have been recorded. These are of volcanic and quartzitic affinity, both probably derived from the Lower Old Red Sandstone. Above this, in the Glenvale Formation, extraformational clasts are absent. Both formations exhibit flat lamination and trough cross-bedding, the latter being interpreted as slip faces of large sinuous-crested ripples or dunes. Cross-stratal dip directions suggest generally eastward flowing palaeocurrents. In the overlying Knox Pulpit Formation, however, the dominant palaeocurrent direction is northwesterly. Chisholm and Dean also identify bipolar dip patterns in some localities and have recorded cosets of opposed northwest-southeastward dipping foresets. Another important feature of the Knox Pulpit Formation is the presence of Skolithos-type burrows. This, together with the bipolar current directions led the authors to suggest a shallow marine environment of deposition. This interpretation is, however, now subject to revision and Chisholm (pers comm) has suggested that the opposing foresets may represent aeolian deposition, the Skolithos-type burrows possibly formed in pools of standing water. The overlying Kinnesswood Formation contains siliceous pebbles with rather poorly developed nodular cornstone horizons suggesting continuing terrestrial conditions.

### 2.2.4. The Pentland Hills

Patchy outcrops of Upper Old Red Sandstone sediments occur in the Pentland Hills. These have been described by Mitchell and Mykura (1962) but no account has been published on their facies relationships or sedimentology. Bluck (1978) shows palaeocurrent data for these outcrops. These show a predominantly southeasterly trend and have been used in this study in the palaeogeographic reconstruction (Fig 4).

### 2.2.5. Summary

Essentially, the Midland Valley Basin appears to have had a more-or-less easterly palaeoslope, the sediments in the west being predominantly conglomerates and coarse sandstones of local provenance. These are interpreted as alluvial fan deposits, the source area being the highland scarps of the Southern Uplands and the Scottish Highlands, and pull-apart basins related to movement on the Highland Boundary Fault. To the east, in Stirlingshire, Kinross and Fife, the sands appear to become progressively finer grained. These are interpreted as being more distal deposits, laid down on a flood plain further from the main source areas. Cornstones and wind-blown sands indicate significant periods of non-deposition related either to long-term migration of the river channels or to the climatic regime, the fluvial deposition occurring only at times of ephemeral stream flooding.

## 2.3. THE BORDERS BASIN

The main Upper Old Red Sandstone outcrop in the Borders Basin occurs in a broad strip up to 15km wide trending southwest from the Berwickshire coast to near Hawick (Fig 3). Minor isolated outcrops have been mapped from the edge of the main outcrop to the Solway Firth (Leeder and Bridges 1978). The sediments comprise a sequence of siltstones, sandstones, cornstones and conglomerates which rest unconformably on intensely folded Silurian strata. Leeder (1973) notes the presence of a buried topography and argues that the relief could never have been high during Upper Old Red Sandstone times and that burial of the topography was complete by the end of the period. Leeder identifies four facies associations:

1. Thick sandstone units with erosive bases overlain by siltstone with sandstone intercalations.

2. Thin intraformational conglomerate units with erosive bases and overlain by sandstone often succeeded by a concretionary cornstone.

3. Fine to coarse sandstones, pebbly sandstones and conglomerates, sometimes bounded by visible channel forms.

4. Thin sandstones and siltstones with occasional conglomerates and pebbly sandstones.

The sediments are interpreted as being wholly fluviatile, deposited in high-sinuosity (Facies Association 1) and low-sinuosity (Facies Associations 2 and 3) stream channels and by stream flooding (Facies Association 4). The flood deposits may be products of overbank and crevasse splay flooding (finer and coarser beds respectively) or of distal and more proximal stream flooding.

Palaeocurrent directions measured in the sandstones indicate the presence of an interior drainage basin in the Borders. In the southwest the northeasterly palaeocurrent direction is consistent with the Galloway pebble provenance while in the north, the cross stratal dip directions indicate a dominantly southwesterly flow (Leeder 1974a).

Leeder (1976a) considers the presence of cornstones at the top of the Upper Old Red Sandstone to represent a period of basin upwarp. This he relates to partial melting in the upper mantle and postulates that the eruption of the Birrenswark lavas led to the termination of the basin as a geomorphological feature.

Unlike that in the Midland Valley Basin, therefore, the sedimentation in the Borders Basin was dominated by an interior drainage system. Both basins appear to have had significant periods of non-deposition with cornstone development, though in the Borders Basin this may have been related more to tectonic controls than to climatic or depositional environment.

## 2.4. PREVIOUS WORK ON THE BERWICKSHIRE COAST

A sedimentological study of the Upper Old Red Sandstone of Burnmouth, 22 km southeast of Pease Bay, was carried out by Smith (1967, 1968). He identifies a number of fining upward sedimentary cycles which he interprets as channel and overbank deposits laid down by streams in a semi arid environment.

Douglas (1972) in a BSc Honours dissertation studied the sedimentology of the Cockburnspath area and identified several sedimentary cycles in the Pease Bay to Horse Roads Rock section. He concluded that the Old Red Sandstone sediments were deposited in a dominantly fluvial environment with some dune formation. A variety of high and low energy stream conditions were indicated by the sedimentary structures. Douglas (1972) found a dominantly southeasterly palaeocurrent trend.

Paterson et al (1976) took several cross-stratal dip measurements from the Pease Bay area, though not in the context of any sedimentological study. They deduced an easterly palaeocurrent direction and suggested that this was confluent with the Midland Valley drainage. Andrews (1978) identified scales of Remigolepis in fallen rocks along the Pease Bay section and took this to indicate that the Upper Old Red Sandstone was of Lower Carboniferous age. Again this was an isolated study, not related to any sedimentological work.

## 3. DESCRIPTION OF THE SECTION

The cliff and shore section from Pease Bay to Horse Roads Rock was measured and described in detail. A synopsis of the log is given in Fig 2. Six distinct facies types have been identified:

A. Argillaceous facies
B. Parallel-bedded sandstone facies
C. Trough cross-bedded sandstone facies
D. Cross-bedded sandstone with mudclast facies
E. Small-scale cross-bedded sandstone facies
F. Unclassified poorly bedded sandstones

These are described in detail below:

### 3.1. FACIES A - ARGILLACEOUS BEDS (PLATES 1 TO 3)

Facies A beds consist of multistorey bodies usually less than one metre thick and comprising repeated alternations of siltstone, silty mudstone and silty very fine sandstone (Plate 1). The individual beds rarely exceed 100mm in thickness and are mostly in the region of 20mm to 50mm. The sediments are predominantly dark red with pale green patches running parallel to the bedding. The paler beds are often up to 20mm broad and, in places, the colouration is seen to cut across primary sedimentary structures (Plate 2). The basal structures of the composite bodies are planar or irregular but do not appear to show erosion features. Similarly, individual beds have planar tops and bases. Within the siltsones, occasional fine to coarse grained sandstone bands, usually only a few centimetres thick, are found, sometimes exhibitng slight scouring at their bases. Some of these sandstone bands also exhibit small fining-upwards units. In addition to these coarser bands and lenses, intraformational mudstone clasts up to 20mm across are also present and, more rarely, individual disarticulated blackened fish scales. In the lower part of the sequence, some bands in the mudstone take on a nodular-like appearance. This does not appear to be related to differences in carbonate content and it is suggested that they represent insipient dessication structures (Plate 3).

The dominant sedimentary structures are parallel laminations. These are disturbed in places by soft sediment deformation, generally in the form of small slump features, usually discrete and of the order of a few centimetres across. In addition, some siltstones and very fine sandstone beds exhibit ripple cross lamination. The ripples are generally of low amplitude, often less than 0.50mm, with wavelengths in the range 20mm to 40mm. Ripple beds do not appear to be laterally extensive, often occuring as isolated lenses and passing into laterally parallel-laminated sediments. Climbing ripples are not seen. Coarser sand also occurs in lenses but often is present as a single continuous sheet, in places becoming no more than a few grains thick.

In the lower parts of the measured section the beds are generally thicker and more silty. In fact, the multistorey sequences appear to be more common in the very lowest beds and are not seen at all in the upper part of the section, where the average bed thickness becomes 100mm to 200mm. In the upper parts, however, there is frequently an associated development of cornstone. This takes the form of carbonate- and iron-rich concretionary growth, apparently parallel to the bedding. In some cases, this occurs as irregularly shaped nodules and, elsewhere, as thin horizons, sometimes within but frequently capping the silty mudstone beds. These horizons have irregular tops and bases and appear to represent a further developmental stage of nodule coalescence.

### 3.2. FACIES B - PARALLEL BEDDED SANDSTONE (PLATES 4 TO 6)

Facies B beds consist of very fine to medium grained sandstones which vary in thickness from 0.25m and 2m (Plate 4). The beds are commonly pale yellowish red and in places are partially drab. Reduction spots, around 150mm in diameter, are not unusual in the lower parts of the section. The basal surfaces are generally planar, often appearing to be erosional, the amount of relief being negligible and scours rare. The sandstones are normally well sorted but a coarser fraction, up to granule grade quartz, is also present in many beds. Intraformational mud clasts and, in the upper part of the section, cornstone clasts are rare and usually do not exceed 50mm in length, often occurring near the base of the unit but also within the sandstone body (Plate 5).

Sedimentary structures are dominated by flat or low-angle sub-parallel laminations, the maximum angle being in the order of 1 or 2 degrees. The laminations are evenly spaced a few millimetres apart. Some beds, generally those over one metre in thickness, develop small scale cross-bedding or ripple laminations at the top (Plate 6). Set sizes are normally 20mm to 30mm but may reach 100mm. These are considered to represent reactivation of the sediment in waning flood conditions. In places the small scale cross-bedding may be disturbed and some soft sediment deformation structures including convolute lamination are present. Primary current lineations have not been seen on the bedding surfaces as, over much of the section, the sandstones are friable and do not appear to fracture preferentially along the bedding planes.

In some places two or more sets of parallel laminated sandstones are seen, the amalgamated junction being discerned by the partial development of small scale cross-bedding or by soft sediment disturbance features at the top of the underlying annealed bed. More commonly, the overlying beds are cross-bedded sandstones with scoured bases, the relief giving rise to variable thickness in the underlying bed and possible erosion of the top structures. Cornstone concretionary growth is rare in these beds though this may be a reflection of their distribution and comparative rarity in the upper part of the section.

### 3.3. FACIES C - TROUGH CROSS-BEDDED SANDSTONE (PLATES 7 AND 8)

Facies C beds consist of fine to medium grained, well sorted sandstones ranging in thickness from 0.5m to 1.5m (Plate 7). Sediments are commonly pale orangish red though darker variants are also found. The basal surfaces are almost invariably erosional, the scoured relief ranging from a few centimetres to 0.3m. In some cases the surface has an undulatory appearance. Intra-formational mudstone and cornstone clasts are rare but are occasionally seen, apparently as a lag deposit, in some troughs in scoured surfaces (Plate 8). Normally the clasts have a tabular shape and do not exceed 40mm in length. Quartz granules or small pebbles up to 20mm in diameter are also infrequently present.

The dominant sedimentary structure is trough cross-bedding, sets being in the range 0.3m to 0.4m. The laminations are closely spaced and are commonly asymptotic to the bases of the sets. In some cases multiple sets pass laterally into a single set. Reactivation surfaces are common in these beds, the later laminations truncating the older at very low angles. In the thickest beds the scale of the trough cross-bedding may be reduced towards the top of the bed and occasionally soft sediment deformation is also seen.

In the upper part of the sequence concretionary cornstone growth may be found in some Facies C beds. Often this takes the form of a layer of discrete nodules up to 100mm across, parallel to the bedding surface though in places they may coalesce into a sheet-like mat. These concretionary sheets need not be horizontal. Cornstone development can occur under a scoured surface suggesting that the erosion was followed by a period of non-deposition during which time carbonate concretionary growth occurred (Plate 20). Elsewhere cornstone growth at the top of the bed may have been partially removed by a subsequent phase of erosion.

## 3.4. FACIES D - CROSS-BEDDED SANDSTONE WITH MUDCLASTS (PLATES 9 TO 11)

Facies D beds comprise fine to coarse grained sandstones which range from 0.1m to 1m in thickness and are usually dark red to reddish brown in colour. The basal surfaces are invariably scoured though the relief is seldom more than a few centimetres, occurring in shallow scoops eroded into the underlying beds. The scoop-like nature of the troughs is most clearly seen from above (Plate 9). The sediments are generally poorly sorted and consistently contain intra-formational mudstone or cornstone clasts up to 50mm across. The clasts are sometimes tabular but are more frequently disc shaped, often occurring as lags in the base of shallow troughs, their short axes perpendicular to the bedding plane. In some cases the beds are apparently normally graded.

The most common sedimentary structure is trough cross-bedding, the sets which make up the troughs being rarely more than 0.2m thick. The laminations are uneven and sometimes diffuse but appear to be low angle and asymptotic, broadly parallel to the margins of the troughs (Plate 10). Individual sets can be up to 5m wide, the dimensions of their long axes being difficult to determine because of the nature of the outcrop but certainly exceeding 10m in some instances. Most beds comprise three or four superimposed sets but isolated troughs are also present.

In the upper part of the section the beds are frequently associated with the development of cornstone horizons. It appears that the intraformational cornstone clasts act as nucleii for concretionary growth. In many cases the concretions form as sheets parallel to the lamination (Plate 11), the carbonate-rich solutions having apparently exploited the bedding planes. Elsewhere the cornstone growth appears to invade the beds below, often in sub-vertical pipe-like concretions (Plate 17). In the lower part of the section, cornstone clasts are present but concretionary development is relatively rare and apparently not particularly associated with facies D beds.

## 3.5. FACIES E - SMALL-SCALE CROSS-BEDDED SANDSTONE (PLATES 12 TO 14)

Facies E beds consist of very fine to fine grained sandstones in beds with a maximum thickness of 0.85m (Plate 12). They are commonly pale yellowish red in colour. The basal surfaces are generally planar but in places appear irregular or even mildly scoured. The sandstones are normally well sorted and intraformational clasts are rare.

The dominant sedimentary structures are small-scale trough and planar cross-stratification. Set sizes normally range between 50mm and 150mm. In some cases these sets are tabular, the base of each set parallel to the basal surface of the bed and truncating the top of the cross-lamination in the set below (Plate 13). In other cases the small-scale cross-bedding takes the form of small troughs with fine asymptotic laminations (Plate 14). Sets of a sufficiently small scale to be classed as ripples are relatively uncommon, though some rippled horizons and units of climbing ripples can be recognised.

Facies E beds make up less than five per cent of the section but are significant enough to be included as a distinct facies type. They are in general restricted to the lower part of the section and in only one bed is nodular cornstone development recorded.

### 3.6. FACIES F - UNCLASSIFIED POORLY BEDDED SANDSTONES (PLATE 15)

An additional type of sandstone bed can be distinguished in the upper part of the section. The beds range from around 0.5m to 1m in thickness and often have an irregular or gradational base. They are invariably very poorly bedded and are usually shot with both vertical and sub-horizontal cornstone concretions. It appears that cornstone growth has completely destroyed the sedimentary structures though in some beds traces of cross-bedding can be seen (Plate 15). They are interpreted as Facies B and Facies C sandstones which have been subjected to early carbonate diagenesis and are therefore not considered as an additional facies type. The nature of the beds, however, precludes them from being classified as one facies or another.

### 3.7. OVERALL TRANSITION PROBABILITY MATRIX

In order to identify any possible cyclicity in the Pease Bay to Horse Roads Rock Old Red Sandstone sequence, a Markov Chain Analysis was carried out. This stratigraphic analysis technique has been used by several authors on a variety of sedimentary sequences. A theoretical study carried out by Krumbein (1967) showed the power of Markov analysis both in investigating cyclicity in sedimentary sequences and in synthesising stratigraphic sections. Vistelius and Faas (1965) compared Palaeozoic flysch beds from the Urals and Tertiary red beds of the Cheleken Peninsula, finding that the sedimentation in both was sequences was simple, homogeneous and non-periodic. Work by Potter and Blakely (1968) compared a marine carbonate sequence, an Ordovician marine clastic sequence and Carboniferous cyclothems, finding evidence of cyclicity in all three cases. The most successful application of the technique is probably in cyclic alluvial sediments, such as the Namurian of the Midland Valley of Scotland (Read 1969) and the Palaeozoic of Polecat Bench, Wyoming (Gingerich 1969). These studies led to the definition of fully-developed cyclothems. Selley (1969) also found that a sequence from the Carboniferous of the Pennines produced a cyclothem 'very similar to the Yoredale facies cycle'. On the other hand, he showed that the Torridonian red bed sequence showed no evidence of cyclicity. This, he concluded, was a result of the deposits having been laid down by a braided, rather than a meandering, river.

Two principal methods of data sampling for Markov chain analysis have been employed by these authors - fixed intervals and lithological transitions. The fixed interval method appears to be primarily used for theoretical studies and for synthesising stratigraphic sections (Krumbein 1967, Schwarzacher 1967). A suitable sampling interval must be chosen and the size of this is critical to the analysis. In analysing startigraphic sequences, most authors use the occurrence of lithological transitions. This reflects the order in which the beds were laid down rather than their relative thicknesses. This is the method proposed by Selley (1969). Selley also discussed the inclusion of multi-strorey lithologies (i.e. the transition from one facies to the same facies). In terms of stratigraphic analysis, the important transitions are probably those between different facies, and therefore different environments of deposition. In addition, multi-storey lithologies are easier to identify in some facies than in others. Read (1969) included multi-storey lithologies when using a fixed sample interval and in the construction of synthetic fluvial cycles. In the sequence analysis using lithological transitions, he rejected the assumption that a lithological state can pass upward into the same lithology.

In the present study, the aim of Markov analysis is to identify possible cyclicity in the sedimentary sequence. The precedent set by Read (1969) has been followed here. Multi-storey lithologies are ignored - it is assumed that any one facies does not pass upwards into the same facies. Accordingly, the matrix is constructed by counting lithological transitions rather than by using a fixed sampling interval.

The clearest exposition of Markov analysis was found in Gingerich (1969). The equations in this paper were translated into a Basic program which was developed for the Dragon 32 microcomputer. The results are shown in Table 1.

Reference to a table of Critical Values of Chi Square (Siegel 1956) shows that for 15 degrees of freedom a value of chi-square of 3.28 is not significant. It is thus improbable that the Pease Bay - Horse Roads Rock section of Old Red Sandstone was laid down by a Markovian mechanism and it must be concluded that there is no statistical evidence for cyclicity.

Notwithstanding this conclusion, reference to the Difference Matrix shows that there are certain transitions which have a considerably higher than random probability of occurring. These are: Facies A overlain by Facies C and Facies B overlain by Facies D. These transitions will be used in the interpretation of the facies in the next chapter.

# 4. INTERPRETATION OF THE FACIES TYPES

## 4.1.1. FACIES A - ARGILLACEOUS BEDS

Argillaceous beds have been recorded in several modern fluvial environments and in particular in braided rivers by Coleman (1969), Williams and Rust (1969), Cant and Walker (1978) and Rust (1978). Argillaceous beds in dominantly sandy sequences appear generally to receive little more than passing reference and no detailed work appears to have been done on this facies. It is generally agreed (Collinson 1978, Cant 1982) that high-sinuosity streams have a greater proportion of argillaceous beds than do low-sinuosity streams. In view of the relative insignificance of argillaceous beds in the Pease Bay Old Red Sandstone sequence (less than 10%), a high-sinuosity (meandering) stream environment of deposition appears unlikely. The later interpretation of the sandstone facies further substantiates this evidence and a high-sinuosity stream model is rejected.

In low-sinuosity streams, the proportion of argillaceous beds to sands is lower. A braided stream model might then be proposed. In the studies of braided streams by Williams and Rust (1969), the argillaceous facies of the Donjek River is represented by faintly colour-banded laminated silty clay. These sediments are found in low-lying areas which are frequently filled with standing water. They are preserved in abandoned channels which are subject to occasional flooding. Rust (1978) described the occurrence of argillaceous beds in the silt-dominated Slims River (Yukon). Here the principal facies are flat laminated and ripple cross laminated sandy silt. In its silty reaches, the Slims River comprises a single channel. The dominance of sandy beds in the Pease Bay sequence suggests that this model is not applicable to the present facies study.

Alternatively the argillaceous facies beds may represent ephemeral deposits laid down in a playa lake or similar environment. The presence of incipient dessication cracks in the lower part of the sequence and cornstone development in the upper part of the sequence suggests periods of drying-out, these becoming more prolonged in the later sediments. On the other hand, the presence of disarticulated fish scales does suggest some degree of continuity between these bodies of water and other parts of the freshwater system. Waterston (1962) has recorded the presence of fish remains in the Upper Old Red Sandstone of Burnmouth and has pointed out that disarticulated fragments point to a moderately high energy or ephemeral environment. Andrews (1978) has identified scales of Remigolepis from loose boulders found on the shore near Pease Bay.

The presence of ripple lamination suggests an intermittent low-energy current regime in these standing bodies of water. It is suggested here that these may represent waning flow of floodplain deposition or overbank flooding. Coarse sand material, present in single continuous sheets and occasional lenses, is interpreted as representing deposition from flood waters where these reached the isolated playa lakes. It is envisaged that the waning flood, having carried its sand load across the flood plain, is dissipated in the low-energy environment, depositing the remaining sediment as a grain flow across the bed of the lake.

### 4.1.2. FACIES B - PARALLEL-BEDDED SANDSTONE

Horizontal stratification has been recorded in several modern fluvial environments including meandering rivers (Bluck 1971), braided rivers (Doeglas 1962, Coleman 1969, Williams and Rust 1969, Smith 1970, Rust 1972, Miall 1977, Cant and Walker 1978) and flood deposits (McKee et al 1967, Smith 1971, Picard and High 1973). Although it is accepted that this type of stratification is a common feature in a wide variety of depositional environments, it is suggested here that the parallel-laminated sandstones of Pease Bay were laid down in high-energy flood conditions.

McKee et al (1967) studied the distribution, composition and structures of flood sediments laid down in major flooding in Bijou Creek, Colorado in 1965. They found horizontal stratification to be the dominant type of bedding (90 to 95 per cent of all deposits). Parallel-laminated sands were laid down both within the main channel and on the flood plain, the precise environment being identified only by the associated bedding types. Within channels, these bedding types included medium to large scale festoon cross-bedding and megaripples, while the floodplain deposits were typified by climbing ripple lamination and convolute structures, representative of waning flow conditions. The Pease Bay sandstones show all of these associations. In contrast, horizontal stratification in less energetic fluvial environments is associated with low-stage reactivation of the bedding in the form of cross lamination at the top of the unit (Cant and Walker, 1978).

Picard and High (1973) identified two types of horizontal stratification associated with ephemeral deposits in Twelve Mile Wash, Utah. The first - horizontal parallel stratification - comprises thin, even strata in sets up to 15cm thick. They suggest that these beds were laid down by low-velocity currents in upper flow regime conditions. This combination of factors they attribute to the relatively fine grain size and shallow water depths found in waning flood conditions. This type of bedding is rare, however. More common is the horizontal discontinuous stratification, where strata thickness is slightly greater and set sizes range up to a metre thick. Individual strata are found to be laterally discontinuous, pinching out over distances of up to a few metres. The authors interpret these beds as being deposited by high velocity currents in upper flow regime conditions. They are found both as channel fill and as bar deposits. Picard and High suggest that this latter bedding type is more common, especially in coarser sandstones.

The nature of the individual laminae and the set sizes in the Pease Bay parallel-bedded sandstones suggest that these indeed comprise horizontal discontinuous stratification. The occasional presence of mud clasts and coarse material also suggest high energy conditions.

The Facies B sandstones of Pease Bay are, in many respects, similar to the parallel-laminated facies which dominates the Trentishoe Formation (Old Red Sandstone, North Devon) described by Tunbridge (1981). Individual beds in the Pease Bay section appear to be generally thinner than those of the Trentishoe Formation (0.4m to 2.5m) but the grain size range and the bedding features are very similar. The relief on the basal scoured surfaces appears to be greater in the Pease Bay beds. In the Trentishoe Formation Tunbridge remarks on the geometry of the parallel-bedded sands, being able to follow cliff-face exposures for over 100m with little thinning evident. The Pease Bay section does not possess such extensive lateral exposure though where the facies is seen, it is apparent that the beds may not be traced far as frequently the tops of the beds are eroded and overlain by later sandstone bodies.

Tunbridge suggests that the parallel laminations in the Trentishoe Formation represent deposition in upper flow regime conditions. Although no primary current lineation has been seen, the presence of an irregular erosive base and the monotony of the beds suggest that upper flow regime conditions are most likely for the Pease Bay Facies B sandstones. Tunbridge also points to the paucity of intraformational clasts in some beds and suggests that this is indicative of a non-channelised surface flow. The absence of cross-lamination at the top of the parallel-bedded sequence may suggest a rapidly waning flow with no low-stage reactivation of the sediment. These observations lead to the interpretation of the sands as high energy flood deposits.

### 4.1.3. FACIES C - TROUGH CROSS-BEDDED SANDSTONE

Trough cross-bedded sandstones have been recorded in a wide variety of modern fluvial environments although most commonly in braided rivers as channel sands (Williams and Rust 1969, Smith 1970, Rust 1972, Miall 1977, Cant and Walker 1978). They are also recorded to a lesser extent in flood deposits (McKee et al 1967, Williams 1968, Picard and High 1973).

Cant and Walker (1978) have compared sedimentary structures found in dunes and cross-channel bars in the sandy braided South Saskatchewan River in Canada. They found that sets of trough cross-bedding were characteristic of in-channel dunes whereas cross-channel bars deposited large-scale planar tabular sets. In some cases, these cross-channel bars developed into extensive sand flats, these being characterised by smaller-scale planar tabular sets and parallel-bedded sands. They identified three types of stratigraphic sequence, the first being the full development of in-channel trough cross-bedding through cross-channel bar tabular sets to smaller-scale tabular cross-bedded and parallel-bedded sands at the top. In their second sequence, renewed channel aggradation after the cross-channel bar development led to further trough cross-bedded sands being deposited while in the third sequence in-channel trough cross-bedding continued to be the dominant structure though at a smaller scale and with some discontinuous tabular cross sets. They found that in such sequences there was little evidence of fining-upward textures which they attributed to the fact that the modal sediment size varied little between sand samples taken from deep and shallow channels. In some deep channels, however, they recorded the presence of a lag of intraformational mudclasts and pebbles.

The trough cross-bedding recorded by Picard and High (1973) is restricted to small-scale structures and festoon cross stratification. They suggest that these bedding types are more common in braided rivers than in ephemeral streams. Nevertheless they find that they are characteristic of channel-fill sequences deposited as the flood decreases and overlain by cuspate ripples and horizontal parallel stratification. They suggest that in ephemeral stream deposits these bedding types have a poor preservation potential.

It is considered that this facies in the Pease Bay section for the most part represents river channel deposits. The basal surface of these beds is almost invariably erosional suggesting that the deposition of these sediments was preceded by a period of scouring. Although mudstone and cornstone clasts in these sandstones are rare, some troughs apparently contain lag deposits of intraformational material (Plate 8). This implies a fluvial origin for these beds, the scours being eroded by high-energy flood deposits, perhaps carrying gravel grade material which is not normally found in the sediment. In a waning flow, some of this material would remain as a lag deposit and be incorporated in the channel sands. Unlike the sequences in the models developed by Cant and Walker (1978) and Picard and High (1973), the Pease Bay Facies C sandstones are rarely associated with planar tabular cross-bedding or with small-scale trough cross-bedding. Occasional tabular cross sets are seen but these are usually incorporated within the trough cross-bedded units. Laterally-continuous sets are only rarely seen and these are thought to represent cross-channel bars. The paucity of these structures is indicative of the short-lived nature of the streams responsible for the deposition of the sediment. The lack of smaller-scale cross stratification may be attributable to the poor preservation potential, possibly aeolian reworking of the upper part of the channel deposits.

At the base of the section, the sandstone underlying one of these scoured surfaces displays distinct colour banding and concretionary growth, both of which are sub-parallel to the scoured surface (Plate 20). This suggests a significant period of non-deposition between erosion and infill of the scours, giving time for early diagenetic changes in the underlying sands. Between the phases of erosion and deposition, the scoured relief may have dried out completely (see Coleman 1969) and the infill may be, in part, aeolian. Indeed the closely spaced lamination, asymptotic cross-bedding and reactivation surfaces may also suggest such a depositional agent. This hypothesis is discussed in more detail later (chapters 5 and 8). Certainly there is no evidence for large-scale dune bedding in the sequence but it is considered possible that some cross-bedded sands may be wind blown.

### 4.1.4. FACIES D - CROSS-BEDDED SANDSTONE WITH MUDCLASTS

Trough cross-bedded sandstones of the type described as facies D are recorded in fluvial sequences by Miall (1977) and Cant (1982) but are relatively rarely featured in facies models.

Miall (1977) found mudclast conglomerates in some channel bases in the Platte River. These he interpreted as being deposited at high stage, the clasts being ripped up from dried-out floodplain deposits. In the Donjek River, Miall also records thin conglomerate bands which fine upwards into sandier trough cross-bedding. Again these are interpreted as basal channel deposits. In both sequences, however, the major bedding type was tabular (channel bar) cross stratification, which is notably poorly represented in the Pease Bay Old Red Sandstone section. Cant (1982) notes the development of large gravelly trough cross-bedded sets formed by the migration of gravelly dunes or flat topped bars and wedge shaped gravel cross-beds deposited on the bar margins. Although Cant refers only to lithic gravels, it is suggested here that this model may be applied to the Pease Bay Old Red Sandstone. With a gravel impoverished sediment provenance, the sole representative of coarse sediment may be mudclast conglomerates.

Facies descriptions of both modern and ancient flood deposited sand sequences (McKee et al 1967; Picard and High 1973; Tunbridge 1981, 1984) do not record the presence of Facies D type bedding. Sen and Sinha (1985) describe thin pebbly sandstones, some with shallow trough cross-bedding in the Triassic Maladara Formation of Bihar, India. They attribute these beds to sheet flooding or stream flooding, the trough cross stratification being present at the edges of the flood sheets.

In the Pease Bay section, this facies is taken to represent high-energy floodplain or crevasse-splay sediments deposited while the streams were in high flood stage. The abundance of mudstone and cornstone clasts suggests that these were ripped up from a dried-out floodplain, probably by a fully competent current. The shallowness of the troughs suggests shallow water conditions which is not consistent with a river channel model. In addition, the common association with cornstone development suggests that these beds were dried out in areas of comparative non-deposition such as the floodplain.

A bar-top model is not considered likely because of the presence of intra-formational clasts. Additionally the overall transition probability matrix (Table 1) demonstrates that facies D beds most frequently overlie facies B sandstones which have been interpreted here as floodplain deposits.

### 4.1.5. FACIES E - SMALL-SCALE CROSS-BEDDED SANDSTONE

Small-scale cross-bedded sandstones have been recorded in fluvial sequences by several authors including Walker (1963), Williams and Rust (1969), Smith (1970), Cant and Walker (1978), Picard and High (1973) and Tunbridge (1981).

Cant and Walker (1978) identified small-scale trough cross-bedding associated with late-stage shallow channel aggradation. The small-scale trough cross bedded sands may overlie larger-scale trough cross-bedded in-channel deposits or planar tabular cross-bedded channel bar sands. In the former case, the small-scale cross-bedding may be seen to develop into even smaller-scale ripples and the deposition of silt and clay as the channel was cut off. In the latter case, the small-scale cross-bedding is taken to represent deposition in minor shallow channels which cut across the channel bars. Neither of these associations is seen in the Pease Bay Old Red Sandstone sequence.

Picard and High (1973) identify a structure in ephemeral stream deposits which they term micro cross stratification. The structure is present in most channel and bar deposits and is frequently found near the top of these sequences. They suggest that this and festoon cross stratification, which is a medium scaled version of micro cross stratification, are formed in moderate velocity currents in the lower flow regime. Williams and Rust (1969) and Smith (1970) record the presence of these structures in braided river sediments and Picard and High (1973) suggest that they might be more common here than in ephemeral streams. On the other hand, they expect the preservation potential of small scale cross-bedding to be limited on account of their occurrence in the upper part of the sedimentation units.

Walker (1963), in a study of ripple drift sedimentation, described the types of ripple drift associated with a fluvial environment and suggested that they were formed in conditions of net deposition from a traction current which is carrying sediment in suspension. Such climbing ripple lamination is relatively uncommon in the Pease Bay Old Red Sandstone.

Authors seem to agree that small-scale cross-bedding of this type is related to waning current flow. Tunbridge (1981) recognises these beds and interprets tham as waning flow on sheetflood sands. It is worthy of note, however, that they do not always overlie facies B sandstones. They may be related to a river channel environment, representing waning flow of flood currents. The presence of rare climbing ripple sets may also be significant (Walker 1963). It is likely, therefore, that this facies is found in several depositional environments and represents waning flood stage sedimentation.

### 4.1.6. FACIES F - UNCLASSIFIED POORLY-BEDDED SANDSTONES

These beds are not considered as a distinct facies type but rather as Facies B or Facies C sandstones which have been subjected to early carbonate diagenesis. It is therefore not logical to include them in this facies interpretation. Cornstone growth is discussed later in chapter 10.

## 4.2. DEPOSITIONAL UNITS

Fig 6 shows the distribution of facies types A to E in the Pease Bay to Horse Roads Rock section. On the basis of these facies types, the section can be divided into eight depositional units, numbered I to VIII and tabulated in Fig 6. On the basis of the facies interpretations and the lithological relationships within the depositional units, a model of the overall environment of deposition can be proposed.

In Unit I the domination of non-channelised sheetflood sands suggests proximal to mid alluvial fan sedimentation. The parallel laminated sands are interbedded with trough cross-bedded sands which exhibit scoured bases and assymptotic infill. Red and drab colouration and carbonate growth in the underlying parallel laminated beds suggest that the infill may be secondary - a period of time has elapsed between erosion and infill.

In Unit II the sheetflood sands again indicate a proximal to mid alluvial fan environment with overbank deposits suggesting seasonal flooding, possibly from short-lived channelised rivers on to the floodplain. Bodies of standing water may have occupied abandoned channels or playa-type lakes, providing loci for silt accumulation.

Unit III comprises beds of all facies, probably representing the transition between the mid fan environment of Units I and II and the more distal outer fan associations of the upper units. Unit III marks the incoming of cornstone development, indicating significant periods of non-deposition consistent with a more distal alluvial fan regime.

In Unit IV the dominance of trough cross-bedded sandstones interpreted as channel bar deposits suggests an outer alluvial fan environment where flow is more channelised. The presence of shallow scours with assymptotic cross-bedding and cornstone clasts suggests that seasonal flooding probably remains the dominant depositional agent. Many beds in this unit exhibit concretionary cornstone growth.

Unit V comprises a series of thin siltstones interbedded with trough cross-bedded sandstones. The siltstone beds are mottled red and pale green with cornstone nodules and sandy laminae. The unit is interpreted as representing sedimentation in an interdistributary area between alluvial fans.

In Unit VI the reappearance of trough cross-bedded sandstones suggests a return to outer alluvial fan sedimentation and a similar depositional environment to that of Unit IV.

In Unit VII parallel laminated sandstones suggest a return to Unit II type mid fan deposition, possibly indicating the advance of a new alluvial fan. This advance would be contrary to the general trend towards a more distal environment of deposition seen in Units I to VI. It is possible that the change is due to climatic and/or tectonic controls. Whereas cornstone clasts are present in the flat-bedded sands, cornstone concretions are rare.

Unit VIII comprises a mixture of facies but the presence of trough cross-bedded sands suggests that the environment of deposition is again dominated by channelised stream flow. The unit is characterised by thick cornstones which develop in the upper parts of many of the beds, indicating significant periods of non-deposition.

## 5. GRAIN SIZE ANALYSIS

### 5.1. METHOD

Samples from 31 beds were disaggregated, sieved and the fractions weighed. The results were used for plotting frequency and cumulative frequency graphs and various computed parameters e.g. mean and standard deviation then used for comparison and the construction of cross-plots.

The sandstones were predominantly fine grained and little or no gravel material was present. It was considered sufficient, therefore, to use a sample size between 30g and 70g (Folk 1968). The average sample size was 63g.

The sandstones were disaggregated using a mortar and pestle. Care was taken to use a crushing rather than a grinding action (Folk 1968). After the first crushing, in all but the most friable sandstones, the fractions coarser than 2.5 phi appeared to be composed largely of grain aggregates. In these cases a second, and in some cases a third, disaggregation of the coarser fractions was carried out until the final sample sieved and weighed contained less than 2% aggregates. The crushed sandstones were sieved on a bank of sieves ranging from 0 phi to 4 phi in half-phi intervals. In the few samples where grains coarser than 0 phi were present, these were passed through a -3 phi to -0.5 phi bank. The contents of each sieve were turned out and weighed to 0.01g on an Oertling balance. The total weight of sample was measured and percentage frequencies for each class interval calculated.

A number of sandstones analysed had a coarse fraction comprising mudstone and cornstone clasts. It was not possible to crush the rock without destroying the clasts and therefore disaagregation yielded impoverished coarser fractions and enriched finer fractions. To correct this, the sample was photographed before being disaggregated. In most cases the clasts took the form of plates whose flat sides were parallel to the bedding planes. These samples were photographed in a plane parallel to the shorter axes, ie in a section normal to the bedding plane. A double-scale print was made and traced on to 1mm square graph film. Each clast with a minimum diameter exceeding 0.5mm was measured and the cross-sectional area calculated by counting the 1mm squares covered. The tabulated results showed percentage area covered by clasts in each half-phi interval. These figures were added to the weighed fractions. Secondly the percentage of "crushed clast" in the finer fractions was estimated and subtracted from the percentage class intervals. The error of measurement in the procedure was considered to be greater than the errors involved in ignoring hydraulic relationships and the problem of hydraulic equivalence of the clasts and quartz grains was therefore not considered.

## 5.2. GRAPHICAL GRAIN SIZE DISTRIBUTION ANALYSES

Four main methods exist for graphical representation of grain size distributions (Folk 1968) - histogram, frequency curve and arithmetic ordinate and probability ordinate cumulative frequency curves. Folk (1968) recommends cumulative curves as being the best pictorial representation of grain size distributions. Visher (1969) used probability ordinate cumulative frequency curves to demonstrate characteristic curve shapes for samples from several modern environments and ancient examples. Visher recommends the probability scale because the curve (and the extreme values in particular) may easily be divided into a number of straight line segments. These segments are taken to represent the traction, saltation and suspension populations of the sediments. The points where the segments intersect are referred to as truncation points and represent the grain size limits for saltation and suspension (approximately 2 phi and 3.5 phi). In some environments, two distinct saltation populations together with a third truncation point may be identified. Visher (1969) maintains that the shape of the 'curve' in each environment of deposition is characteristic to that environment and that the variations between the curves are significant in terms of transport and deposition.

The results from the samples in the Pease Bay section were drawn on probability scale cumulative frequency plots. Curves from each facies type were superimposed to present a 'bulk picture' for comparison (Figs 7 to 10).

Most of the sandstones from Facies B, C and E generally fulfil the criteria set by Visher (1969) for fluvial sediments: a well developed suspension population, the truncation point of suspension between 2.75 and 3.5 phi, the size ranging from 1.75 to 2.5 phi being within the saltation population and the bedload, where present, coarser than 1 phi. Unfortunately, Visher presented no results for terrestrial aeolian deposits. His aeolian dune graphs refer to beach dune sediments, the main difference between them and other beach environments being the presence of a single saltation population. In addition, the truncation of the traction population occurs between 1 and 2 phi. These characteristics are common to those for fluvial sediments and the technique seems to be of little use in differentiating these environments of deposition.

Some Facies D sandstones have a better developed traction population than the examples presented by Visher (1969). His prediction of a traction load truncation point coarser than 1 phi is, however, demonstrated on the graph (Fig 9). He suggests that the presence of coarse material is provenance controlled and is most frequently developed in the deepest part of the channel. The presence of mudclasts has already been discussed (section 4.1.4) as being a product of high energy shallow transport on the floodplain. The presence of this traction load may be the most significant factor which might be used to distinguish Facies D beds from Facies B and C. It is, however, not conclusive and it is suggested here that more sensitive methods must be used to identify microenvironments in these rocks.

## 5.3. COMPUTED PARAMETERS

For each sample, a number of parameters were computed - mean, mode, median, standard deviation, skewness, kurtosis and first percentile. These parameters were computed using the method of moments (Pettijohn 1975) in a computer program devised for and run on the PDP 11/70 computer in Murchison House, Edinburgh. A summary of the results is given in Table 2.

Using these parameters a number of cross-plots were constructed:

(1) mean v standard deviation (Fig 11)
(2) first percentile v standard deviation (Fig 12)
(3) first percentile v median (Fig 14)
(4) skewness v mean-median (Fig 15)
(5) skewness v standard deviation (Fig 16)
(6) fraction finer than 3.5 phi v first percentile (Fig 17)

### 5.3.1. MEAN v STANDARD DEVIATION

This plot was first used by Friedman (1961) to differentiate dune and river sands. Friedman subsequently used the same plot for beach and river sands (1967) and again for dune and river sands (1979). Moiola and Weiser (1968) used the same graph and plotted limits for beach, river and coastal dune sands. The limits constructed by Moila and Weser (1968) and Friedman (1967) were, therefore, principally concerned with the differentiation of beach sands from other depositional environments. Although the limits defined in the two papers differ to some extent, it is perhaps significant that none of the Pease Bay Old Red Sandstone samples lie within the beach field in either case. For this reason, and to prevent confusion, these lines are omitted from the scatter plot in Fig 11.

The remaining limits defining river and dune fields still present a confusing picture, there being a discrepancy between the results of Friedman's 1961 and 1979 data (only the latter fields are drawn on Fig 11). In both cases, however, it is clear that the Pease Bay sediments fit readily into neither classification. The various Facies types are scattered over the whole plot, with the exception that the only data to plot in the river-only field are Facies D sandstones. This, however, is not considered to be significant as Facies D sandstones also fall within both the dune/river field and the dune-only field. It would appear that this plot is not useful for examining the relationship between depositional environment and Facies type of the Pease Bay sandstones. It is therefore clear that the Mean vs Standard Deviation plot is either too insensitive for palaeoenvironmental analysis or that the Old Red sediments are, in fact, of mixed origin.

It is suggested here that the sediments are indeed of mixed origin. The depositional model implies a semi-arid desert environment with ephemeral streams and significant reworking of the floodplain sediments. It would therefore be expected that wind-blown well-sorted sand would be incorporated in the fluvial deposits to a greater or lesser extent. The presence of intraformational clasts alone may explain the higher standard deviation and greater mean size associated with the three Facies D samples which fall in the (1979) river-only field.

5.3.2. FIRST PERCENTILE v STANDARD DEVIATION

Friedman used this plot to distinguish (a) beach and river sands (1967) and (b) inland dune and river sands (1979). The two fields drawn using Friedman's (1967) data are by no means exclusive. Though for first percentile values greater than -0.5 phi the river field consists entirely of river sand samples, at first percentile values less than -0.5 phi, eleven percent of the results in the river field are in fact beach sands. In the beach field 23% of the samples are river sands. Similarly, the dune and river data (1979) is not entirely consistent. Though the river field is almost exclusively restricted to river sand samples, the dune field comprises 28% river samples and this fraction does not significantly decrease with decreasing grain-size ie into the dune field.

The problem lies in the premise that the first percentile is characteristic of the depositional environment (Passega 1957). The sediment source must be of at least equal importance. The bulk of the results from Pease Bay samples cluster well into the dune field (Fig 12). This is due to the lack of a significant coarse fraction, which is in turn considered to be a function of sediment provenance rather than depositional agent. Scattered results occur in the river field and this can be attributed to the presence of quartz granules and small pebbles or mudclasts. In samples containing mudclasts, an intraformational subsidiary source is clearly present.

### 5.3.3. FIRST PERCENTILE v MEDIAN

This plot, termed the C-M plot, was used by Passega (1957) to define a wide variety of depositional environments from 'quiet water' to turbidity currents (Fig 13). The tractive currents (river) field in fact comprises three overlapping fields - I, IV and V.

As with the previous scatter plot, the sensitivity of the C-M plot is dependant upon the premise that the first percentile is characteristic of depositional environment. In this case, however, there appears to be a better correlation between the Pease Bay Facies types and the limits defined on the plot. The Facies D sediments plot almost exclusively outside the field of tractive currents and above it. The Facies B and Facies C sediments plot mostly within zone IV and below zone I (Fig 14). The apparent separation of facies types on the C-M plot compared with the previous first percentile v standard deviation plot suggests that although the first percentile may indeed be characteristic of the depositional environment, the median grain size is more sensitive than the standard deviation to facies changes in the Pease Bay sediments. It is proposed here that, as the standard deviation is essentially a measure of sorting, it is significantly affected by the original source of the sediment. The median, on the other hand, is related to the overall distribution of the phi fractions which may therefore be more sensitive to the actual environment of deposition.

### 5.3.4. SKEWNESS v MEAN-MEDIAN

In a perfect gaussian distribution curve, the values of the mean and median are identical. In a skewed curve, however, there is a difference in the values. The median is represented by the line which divides the area under the distribution curve into equal proportions. The mean is represented by the line about which there is an equal moment. End values therefore have a more significant effect upon the mean than values nearer the centre of the curve. In a positively skewed curve, therefore, the value of the mean-median should be positive while in a negatively skewed curve it will be negative. A direct relationship between these two parameters may therefore be expected and a cross-plot should be sensitive to aberrations in the distribution. Fig 15 shows the relationship between these parameters in the Pease Bay samples to be far from linear except in the most general terms.

It is clear from the graph that Facies C beds are most positively skewed. This reflects the 'tail' of the finer-grained fraction present in these sediments. Facies D beds appear to be most negatively skewed, reflecting the coarser-grained 'tail' of mudclasts present in these sands. The boundary between Facies D and Facies B sands lies parallel to the general linear relationship seen between the two parameters, and this is taken in part to be a function of sorting. Similarly-skewed distributions in Facies B and Facies D sands differ in the value of mean-median, Facies B generally showing a lesser difference, possibly implying that these beds are better sorted.

### 5.3.5. SKEWNESS v STANDARD DEVIATION

It is clear from cross-plotting the various parameters chosen above that skewness is an important measure in differentiating facies types in the Pease Bay Old Red Sandstone. It is also evident that sorting, although it may not always be truly characteristic of the depositional environment, also has some bearing on facies differentiation. For these reasons it was decided to plot skewness against standard deviation (Fig 16).

As with other plots, the Facies D sands clearly plot out in a field of their own. The field boundary appears to bear a relationship to both parameters and in general, the better sorted sediments are more positively skewed. The field boundary between Facies B and Facies C sands is less distinctive - some overlap of facies type does occur. Despite this, a boundary can be drawn and appears to be related more to differences in skewness than in standard deviation. As in the Skewness v Mean-median plot (Fig 15) the Facies C beds appear to be more positively skewed than the Facies B or indeed the Facies D beds. This is taken to indicate that the Facies C beds are richer in fines than the other beds. This may, in turn, reflect the influence of depositional environment.

5.3.6. PERCENTAGE FINER THAN 3.5 PHI v FIRST PERCENTILE

These two parameters were chosen for cross-plotting because both are independant measures of current competancy. The first percentile reflects the ability of the current to move a traction load, while the fraction finer than 3.5 phi reflects the amount of suspended sediment carried by the flow.

The traction load is dependant on the Froud Number for the flow and therefore on the stream velocity and depth. Shallow, faster streams carry a coarser traction load than those which are deeper and slower. The first percentile is clearly dependant upon particle size in the traction load and therefore related to the stream conditions. As suggested earlier, the first percentile must also be related to sediment source, for where there is no coarse material, the river cannot carry it. The presence of intraformational clasts in the Facies D sandstones represents virtually the only material which is coarser than sand grade.

The suspension load is also dependant upon the Froud number as faster streams are able to carry coarser material in suspension. A significant clay fraction also represents a cohesive influence on the sediment and it has been shown by Sundborg (1956) that the lower the current velocity, the less material finer than ca 3.5 phi is eroded. The greater the current velocity, therefore, the more fine material will be taken into suspension and therefore deposited during waning flow.

The Pease Bay sediments appear to fall into three main fields related to their facies type (Fig 17). According to the interpretation shown above, the deposition of Facies D represents the greatest current velocity. This may be related more to the first percentile parameter rather than the fine fraction. The ability to carry coarse material is entirely consistent with the overbank flood deposit model, high Froud numbers being attained in fast and shallow flows. Deposition of Facies B sediments represents the second greatest current velocity and Facies C sediments the weakest current. This is again consistent with the sheet-flood model for Facies B sandstones and the channel bar model for Facies C sandstones. The Facies C sandstones which are interpreted as wind-blown sediment would also be expected to show an impoverishment in both coarse material (clasts) and fine material which might be winnowed.

The Percentage finer than 3.5 phi v First percentile plot is therefore useful in substantiating the various models for the Facies types seen in Pease Bay. It may have a wider application in depositional environment analysis and further work might be done with both modern and ancient sediments.

## 5.4. CONCLUSION

Graphical representations of grain size distributions proved to be too insensitive to identify microenvironments in the Pease Bay Old Red Sandstone. The general shape of the curves agreed with those presented by Visher (1969) as being fluvial. The aeolian influence could not be identified from the broad shape of the curves. The presence of a traction population in fluvial sediments is taken by Visher to be provenance controlled, though in the Pease Bay sediments the presence of mudclasts is a function of the environment. Nevertheless, the traction load truncation point is significantly coarser than 1 phi as predicted by Visher.

The several cross-plots constructed (Figs 11 to 17) are to a variable extent successful in differentiating between the facies types of the Pease Bay sandstones. Many of the traditional plots, including the C-M plot of Passega (1957) are not as useful as might have been expected. The parameters which are found to be most sensitive to the differences in the facies types are those which rely on measures of the extremes of the distribution - first percentile, skewness and fraction finer than 3.5 phi, rather than measures of central tendency - mean, median and standard deviation. A number of reasons are proposed for this:

1. Cross-plots using parameters such as mean, median and standard deviation

have been traditionally used to distinguish between environments such as dune, beach and river. In the Pease Bay sandstones, it is considered that the source of sediment may have been restricted. No conglomerates are seen in the sequence. In fact the sandstones are quite impoverished in extraformational material coarser than 1 phi. In the semi-arid desert environment which is being suggested for the Upper Old Red Sandstone of southern Scotland, much of the sand will have been transported by wind, reworked by streams and later possibly reworked again by wind. Thus the source of the material for aeolian deposition and fluvial deposition is identical and there is little opportunity for environmental selectivity.

2. Cross-plots using the first percentile are sensitive to the distribution of the coarse fraction of the sand. This is traditionally a useful parameter for measuring current competancy where a 'widespread' sediment source is available. Although the sediment source of the Pease Bay sands may be restricted, this measure has been found useful here. This is attributed to the presence of cornstone and mudstone clasts which make up a significant proportion of some of the Facies D beds. This parameter has thus been found useful in highlighting the micro-environment.

3. The Pease Bay sandstones display a strongly positive (fine) skewness, which reflects the dominant fine-grained fraction present in these sediments. The fraction finer than 3.5 phi is a direct measurement of this distribution and has been associated with current competancy and thus the quantity of material which may be carried in suspension. In this respect, it has been demonstrated that the Facies C sands are most positively skewed. It should be noted, however, that a dominant fine fraction can be caused by several other factors such as syn-depositional abrasion of intraformational clasts and

post-depositional mechanical infiltration of clay and diagenetic clay mineral growth. Evidence of these has been demonstrated using scanning electron microscopy (chapter 8) but it is not known how significant the effects of mechanical infiltration is upon the total grain size distribution.

More work should be carried out on computed-parameter cross plots of grain size analyses in modern micro-environments before the technique can be usefully applied to identifying ancient environments.

# 6. THIN SECTION MICROSCOPY

Samples from 12 beds were sectioned and examined using a petrographic microscope. Some samples were further examined using cathodoluminescence and many of the observations made with transmitted light are further expanded in chapter 7.

## 6.1. SANDSTONE PETROGRAPHY

Samples were examined from sandstones of facies B, C and D. Overall, in thin section, the sandstones were similar, comprising dominantly fine to medium grained quartz with subordinate feldspar. The feldspar content was estimated as 10% in most samples, though in some the percentage was significantly greater. Cathodoluminescence provides a more accurate means of measuring feldspar content (section 7.4). Extinction angles were measured on a number of twinned plagioclase crystals. The angles ranged from 12° to 18°, indicating either albite or andesine composition. In the few specimens found, it was not possible to compare refractive indices and the precise composition could not be determined.

Accessory minerals were very uncommon - mica was identified in only three samples and ferromagnesian minerals were absent altogether. Douglas (1972) records the presence of zircon, rutile, garnet, staurolite, apatite and anatase. Of these, only apatite and tourmaline were identified in thin section in the present study. Separation of heavy minerals from bulk samples would be necessary to identify the full suite of accessory minerals present. In view of the varied sediment sources, however, it is unlikely that this would provide very much useful information on the provenance of the sandstones.

Quartz grains appear to indicate several provenances. In thin section, two dominant quartz types are present, identified as inclusion-rich or inclusion-free. The inclusion-rich grains occasionally exhibited strongly undulose extinction but more normally straight or only slightly undulose extinction. The inclusion-free grains invariably exhibited straight extinction. Folk (1968) proposed an empirical classification for sand grains based on the abundance of inclusions and the nature of the extinction. On the basis of this classification, the inclusion-rich grains are interpreted as having a metamorphic provenance and the inclusion-free grains are considered to be igneous in origin. The two quartz types are discussed further in section 7.3.

## 6.2. CARBONATE CONCRETIONS AND CORNSTONES

Carbonate concretionary sandstone was examined for evidence of replacive and displacive textures. In minor carbonate concretions from Facies B, no definitive replacive textures were observed in transmitted light microscopy, though some detrital grains appeared to be partially etched. Replacive textures have been recorded in quartz grains using scanning electron microscopy (see section 8.3). Volumetric changes in the matrix or void space are described in section 7.2.2 where point counting was carried out using photomicrographs. Brecciation of detrital fragments was seen in only one specimen where mica had been 'exploded' along its cleavage planes. Watts (1978) describes this as the most common form of disruption by calcite. In the Pease Bay sandstones, which are quartz- and feldspar-rich and virtually devoid of detrital fragments of other silicate minerals, this 'explosion' of mica is the only expected form of disruption.

The cornstone examples examined were stained with alizarine red, demonstrating the carbonate to be dominantly calcitic, with some ferroan calcite responsible for the red colour of the rocks in hand specimen. Relationships between the various generations of carbonate are better seen using cathodoluminescence and are discussed more fully in section 7.2.2.

The chief displacive calcite texture seen was described by Watts (1978) as 'floating grains'. Many of the grains appear to be more angular in shape than the grains seen in adjacent sandstones and etching is considered to be responsible for this feature. The etching of quartz grains is seen not only in the floating grains but also in the associated sandstone. This is generally buff in colour, contrasting with the deep red of the cornstone. It may be that at least some of the ferroan content of the calcite may be attributable to iron derived from the dissolved cutans on the etched quartz grains. Iron-rich percolating ground water may also contribute to the red pigmentation in the cornstone.

Many of the large calcite crystals appear to grow into irregular cracks. This feature has been described by Watts (1978) from calcretes in the Kalahari Desert and is further discussed in the next chapter.

# 7. CATHODOLUMINESCENCE

A study of cathodoluminescence was carried out on several samples with the following aims in mind:

1. to aid the interpretation of the carbonate cement history
2. to determine quartz grain history
3. to investigate the occurrence and history of feldspars

## 7.1. METHOD

Six samples were sliced and prepared as thin sections for cathodoluminescence examination. The sections were polished and mounted on glass slides using a heat-resistant epoxy resin as the embedding agent. The upper surfaces of the sections were polished to aid transmitted light examination and cover slides were not used.

The sections were first examined with transmitted light and then in the catodoluminescence microscope at the University of St Andrews. Monochrome photographs were taken of the features of interest in both plane polarised light and with cathodoluminescence for later examination. A selection of these is reproduced in Plates 17 to 22.

## 7.2. CARBONATE CEMENT HISTORY

### 7.2.1. Carbonate cathodoluminescence

Smith and Stenstrom (1965) first demonstrated the use of cathodoluminescence as a petrographic tool, describing the luminescence of various minerals including quartz, apatite, feldspar, calcite and dolomite. Carbonates generally exhibit orange luminescence and this is attributed to the presence of divalent manganese in the lattice. The orange colour is apparently quenched by the presence of any of a number of transition metal ions, notably iron. Smith and Stenstrom (1965) demonstrated the use of cathodoluminescence in investigating dolomitisation of calcitic carbonate rocks. The presence of iron was found to control the carbonate development, the dolomite growing most easily from calcite containing substituted iron.

Sippel and Glover (1965) used cathodoluminescence to demonstrate growth structures in infillings of sparry calcite and dolomite. A detailed growth history of the minerals was provided by the dark and bright banding which could be correlated from grain to grain. The presence of the banding was attributed to variation in the amount of iron in the lattice, areas high in iron tending not to luminesce. Sippel and Glover (1965) discovered that crystals tended to grow in regular geometric forms until they approached the outlines of the enclosing vug. They suggested that as the crystals grew larger, the fluid passages would become more constricted, leading to a more variable supply of ions in solution. The straight leading edges of the crystals thus become more irregular and the grains adopt a less geometric appearance.

Meyers (1974) expanded upon Sippel and Glover's previous work and developed the concept of 'cement stratigraphy'. He was able to correlate carbonate cement compositional zoning over several hundreds of feet of vertical section and, furthermore, date these zones by examining the luminescence of carbonates in later formations. The iron and manganese contents in both the ferroan and non-ferroan cements were too high to have been derived by normal marine waters and they are interpreted as being formed in reducing conditions in the phreatic (subaerial) zone. In all, five cement zones were identified on their luminescent character which in turn was related to their iron and manganese concentrations. The youngest cement was found to exhibit a dull luminescence and was interpreted as a late ferroan stage. This is considered as a common feature in ancient rocks.

### 7.2.2. Pease Bay carbonates

Samples were examined from three beds:

1. Facies B sandstone with minor carbonate concretions (Bed 2)
2. Facies D sandstone (Bed 12)
3. Cornstone (Bed 130)

The carbonate concretions from the Facies B sandstones are described fully in section 10.1.2. They are thought to represent the earliest phase of cornstone development. In thin section the concretions are found to be polycrystalline. Cathodoluminescence demonstrates the presence of only one generation of carbonate growth, the matrix being uniformly pale orange in colour. This conclusion is hardly surprising, considering that the concretions are believed to represent a single phase of early growth at a specific depth in the soil profile, this growth ceasing upon further burial of the sediments.

Photographs were taken of the texture within the concretion and in the surrounding part of the specimen. Point counting demonstrates that the carbonate may be displacive in nature. The void ratio in the host rock is

0.19 whereas the 'matrix' ratio within the concretion is 0.35. Although corrosion of the quartz has been observed (see section 8.3), replacement by calcite is ruled out as a mechanism for the increase in matrix ratio. Watts (1978) shows that, in the Kalahari Desert, replacement in calcrete is limited to ferromagnesian minerals and feldspars. In the Pease Bay examples, however, the quartz/feldspar ratio remains constant (3.1) within and around the concretion. The different matrix ratios may, however, be partially accounted for by the effects of burial, the matrix within the concretion being unaffected by compaction due to the presence of crystalline carbonate in the voids. Whether this would account for the observed difference in ratio is doubtful. It should be noted, however, that further evidence for displacive carbonate exists in the brecciation of detrital fragments (section 6.4).

The Facies D sandstone examined provided an interesting carbonate structure. Carbonate growth within much of the matrix was similar to that seen in the Facies B sandstones, though more patchy. In addition, however, small ooid-like bodies are seen (Plates 21 and 22). The ooids appear to have 'seeds' at the centre but on detailed examination these proved to take the form of lath-shaped voids. It is possible that the original seeds may have been completely corroded by the calcite. In plane-polarised light the crystal structure is apparently radial. Under cathodoluminescence, however, gradational concentric zoning was observed from pale orange in the inner part to pale brownish orange at the extremity,this suggesting that the radial texture represents recrystallisation of the ooid.

The bodies show displacive textures on their outer rims. Small grains appear to be 'pushed aside' by the growing ooid. Large grains, however, seem to restrict the carbonate growth and misshapen concretions develop. These ooids are not the normal development of carbonate growth in the Pease Bay sandstones, nor do they represent displacive calcite in the true sense. It is believed that a specific set of conditions were present during their formation. It should be noted that in this bed, kaolinite was observed to be the chief authigenic clay mineral present and quartz overgrowth (an otherwise rare phenomenon) was commonly seen.

The cornstone samples examined provided evidence of several generations of carbonate development. Plates 23 and 24 serve to demonstrate the typical cement stratigraphy in the Pease Bay to Horse Roads Rock cornstones. Several distinct phases of carbonate growth are seen. The first, "massive", phase is represented by the growth of the groundmass (A). Where quartz grains are encountered, these are incorporated into the carbonate crystal growth. Some faint structure may be seen in parts of the ground mass, especially in the luminescence photograph. Crystal growth is most pronounced where a void is encountered. Immediately surrounding the groundmass is an irregular growth of carbonate which appears bright in plane polarised light, having the highest birefringence colours, but has a very dull luminescence (B). This is followed by a further regular crystalline overgrowth (C), this time with bright yellow luminescence, interpreted as being high in manganese content. The fourth phase (D) can be seen as a bright overgrowth in plane polarised light but under cathodoluminescence appears to be continuous with the previous phase. Finally, the development of a ferroan carbonate is seen (E) - dark red in plane polarised light and non-luminescent. This late-stage ferroan carbonate is the same feature as that observed by Meyers (1974) and which is considered by Bathurst (1971) to be a common feature in ancient rocks.

## 7.3. QUARTZ GRAIN HISTORY

### 7.3.1. Quartz luminescence

Three distinct types of luminescence have been observed in detrital quartz grains (Zinkernagel 1978):

1. "violet" luminescing grains - varying proportions of blue and red emission intensities
2. brown luminescing grains - varying pale to dark brown emission intensities
3. non-luminescing quartz

Comparison between detrital quartz and crystalline quartz from a variety of source rocks reveals a relationship between luminescence type and temperature conditions during rock formation (Zinkernagel 1978). Violet luminescing quartz is generally related to high temperature (>573°) conditions of formation, such as occur in volcanic, plutonic or contact metamorphic rocks. Quartz in low-grade regional metamorphic rocks is of the brown luminescing type. High grade regional metamorphic rocks which have been subject to tempering by slow cooling are also of this type. Non-luminescing quartz is restricted to quartz which has not been subject to temperatures greater than 300°C such as authigenic quartz.

### 7.3.2. The Pease Bay Samples

In the Pease Bay sandstones, both violet and brown luminescing quartz were identified. Non-luminescing quartz was seen in diagenetic overgrowth. Plates 25 and 26 show a sample from Facies D (bed 12) with both violet and brown luminescing quartz in both transmitted light and under cathodoluminescence. Patchy carbonate cement and occasional feldspar grains are also seen. The brown luminescing quartz, which makes up the majority of the sample, is found as rounded to subrounded grains with high sphericity which are commonly filled with inclusions. The violet luminescing quartz is commonly found as well rounded, generally ovoid shaped inclusion-free grains. Non-luminescing overgrowth is present on some grains but appears to be best developed on the violet luminescing quartz.

In the samples from the Facies C sandstones, the dominant quartz type is violet luminescing. Again, these grains are well rounded and free of inclusions though no significant overgrowth is observed. The brown luminescing quartz grains tend to be coarser with a higher sphericity, are less well rounded and full of inclusions. It seems unlikely that there could be any relationship between facies and quartz luminescing type.

It appears that two distinct quartz provenances are present - a high formation temperature source and a low formation temperature source. Possible low formation temperature sources include regional metamorphic rocks or secondarily-derived Lower Palaeozoic sediments of the Southern Uplands. High formation temperature sources include Lower Old Red Sandstone andesites of the Midland Valley and Caledonian granites of the Highlands and Southern Uplands. Additionally, the Southern Uplands granites have contact metamorphic aureoles which may contain high temperature quartz. Braithwaite and Jawad Ali (1978) and van Breemen and Bluck (1981) suggest that in the Highlands, the early Caledonian granites were unroofed during Lower Old Red Sandstone times, in which case, high temperature quartz could be secondarily derived from the Lower Old Red Sandstone.

In the sample described above, the number of feldspar grains approaches 10% of the detrital material. The association between feldspar and violet luminescing quartz may be significant although in other samples, brown luminescing quartz is found with 15% feldspar and a relationship between the presence of feldspar and the dominant quartz type is not considered to be present.

A correlation between quartz luminescence type and grain geometry appears to exist throughout the samples analysed. High temperature quartz grains are commonly finer grained and better rounded than the low temperature grains. The size difference may be attributable to the source of the high temperature grains being a fine grained volcanic rock, though if these are Lower Old Red Sandstone lavas, it is difficult to account for the maturity of the grains. It is more likely that they are secondarily derived from Lower Old Red Sandstone sediments. The coarser grained low temperature quartz with inclusions may be derived from Lower Palaeozoic sediments of the Southern Uplands, although a Highland provenance cannot be ruled out for some of the grains. The apparent dominance of brown luminescing grains in Facies D sandstones and violet luminescing quartz in Facies C sandstones is thought here to be related to the grain size distribution in these two sandstones rather than to different provenances.

## 7.4. OCCURRENCE AND HISTORY OF FELDSPAR

In normal transmitted light microscopy, the presence of alkali feldspars can be difficult to observe. Under cathodoluminescence, feldspars stand out as bright blue in contrast to the dark violet and brown quartz. Point counting from the photographs has revealed that feldspars can make up to 20% of the clastic components of the rock. As discussed earlier, there seems to be no reliable correlation between the abundance of feldspars in the sample and the dominant quartz type. It may be noted that in general, the size of the feldspar grains is considerably finer than the quartz grains. Consequently, the abundance of feldspar in a sample may be related more to the mean grain size and sorting than to the provenance of the particular bed.

A point count was carried out in a Facies B sandstone with small carbonate concretions. It is believed that these concretions formed early in the history of the sandstone (see section 10.1.2) and it was thought that the proportion of feldspar in the clastic component might be an indicator of the amount of possible replacement of feldspar taking place during diagenesis. On first examination, the percentage of feldspar to quartz appeared to be constant over the whole sample. Subsequently, a point count was carried out on the photomicrographs. The percentages of mineral species identified from the point counts are shown in Table 2.

The quartz/feldspar ratio is 3.1 in each case which confirms the visual estimates first made on the slide. For comparison, in the Facies C sandstone, the ratio is 4.9. It is therefore suggested that diagenesis has had little or no effect upon the amount of feldspar present. This is confirmed by the rarity of kaolinite as an authigenic clay mineral (see section 8.4.2).

A number of feldspar grains were observed to have been physically and chemically weathered. These occur in a Facies C sandstone (Bed 51), a channel deposit possibly with aeolian influence. Several grains were worn or broken and one exhibited apparent dissolution of the central part. Nevertheless, the majority of the feldspar grains appear to be fresh and it is suggested that the effects of chemical weathering are negligible. The worn or broken nature of some grains may be attributable to the aeolian transport.

# 8. SCANNING ELECTRON MICROSCOPY

## 8.1. INTRODUCTION

The Scanning Electron Microscopes at the Gatty Marine Laboratory of the University of St Andrews and in the Botany Department of the University of Edinburgh were used to examine several samples from the Pease Bay section with the following aims in mind:

1. to evaluate sand grain sedimentary history
2. to evaluate clay mineral history

## 8.2. METHOD

Chips of samples were taken from ten beds in the sequence, where possible noting the way-up of the specimens. Each was mounted on a stainless steel mount with nail varnish and coated with sprayed gold in an evacuated ionised chamber. Each specimen was examined and photographs made for later reference. Black and white photographs of selected samples are shown in Plates 27 to 35. Selected specimens were examined using the electron-dispersive X-ray analyser (EDX) on the SEM in the Gatty Marine Laboratory. Plots of the profiles have been made for both clay minerals and carbonates.

## 8.3. SAND GRAIN HISTORY

Krinsley and Doornkamp (1973) have demonstrated that surface microtextures of quartz sand grains can be closely related to their past sedimentary history. They were able to differentiate between glacial, beach, subaqueous and aeolian textures. The most diagnostic subaqueous textures were found to be mechanically produced V-shaped patterns or depressions whereas aeolian textures were characterised by small 'upturned plates' on the fractured portions of the grains.

The subaqueous mechanical V-shaped patterns are three-sided, narrowing towards the surface of the grain. They result from chipping of the grains on the cleavage plate edges which are the weakest points on the surface. The chipping is caused by mechanical impact of the saltating grains during subaqueous transport and the precise expression appears to depend upon whether the grain surface is smooth or rough. Krinsley and Doornkamp (1973) also identified straight or curved grooves across the surface of the grains as characteristic of high-energy subaqueous environments. It might be expected that these would be more common in beach rather than in river sediments but the flood deposition of the Pease Bay sandstones may have provided a suitable high-energy environment for the generation of such grooves.

Specimens of individual quartz grains selected to identify surface textures using scanning electron microscopy were boiled in concentrated hydrochloric acid to remove diagenetic clay mineral growth. They were then washed in water, dried and mounted on the stainless steel mounts with double-sided tape before being treated as normal S.E.M. samples.

Examples of textures similar to those described by Krinsley and Doorkamp were not conclusively identified on any of the prepared sand grains. Plate 27, however, shows textures which may be related to 'upturned plates'. These were seen in an unprepared sample. The fact that microtextures were seen on so few samples is probably related to the post depositional history and in particular the effects of carbonate and silicate diagenesis. Krinsley and McCoy (1978) have identified 'upturned plates' on Saharan dune sand. They also point to other indicators of aeolian history, the most obvious of these being the well roundedness of the grains, which they take as indicating that the grains have, at some time in their history, been subject to aeolian transport in a large desert.

Krinsley and McCoy (1978) also observed depressions, both equidimensional and elongate, up to 250 m in diameter, which they related to the effects of direct, rather than glancing, impact between saltating grains. Such features are seen on several grains in the Pease Bay Old Red Sandstone (Plate 28) and are identified as such despite the globular silica overgrowth on the grains. The globular overgrowth itself may be indicative of desert conditions, though it has probably been modified by later diagenesis. Marzolf (1978) has identified silica growth on quartz grains which he considers to be the microtextural representation of sand grain frosting. The reaction between silica and carbonate solution may be controlled by small changes in pH, temperature and partial pressure of carbon dioxide. In the present context, the most significant feature is probably temperature. Engel and Sharp (1958) suggest that desert dew charged with carbon dioxide dissolves minute quantities of silica which is precipitated as the water evaporates during the day. The continual process of solution and precipitation of silica on the grain surface leads to globular quartz overgrowth. This is the most evident surface textural feature on the quartz grains examined here.

In addition to silica diagenesis, carbonate diagenesis has affected the surface textures of the quartz grains. Corrosion by carbonate solutions has been identified in some grains, especially those from beds which feature incipient cornstone growth. An example from a Facies B sandstone is shown in Plate 29. Silica replacement generally requires alkaline conditions (Krauskopf 1967) and it is considered that such conditions must have pertained in the semi arid environment proposed for the Pease Bay Old Red Sandstone.

## 8.4. CLAY MINERAL HISTORY

Clay minerals within the Old Red Sandstone of Pease Bay may be divided into two groups - detrital clays and authigenic clays. While it is likely that detrital clay material has been subject to a degree of diagenesis, it was hoped that some evidence of the original sedimentary regime may nevertheless have been preserved. The authigenic clays result from the chemical weathering of feldspars and other silicate minerals which were present in the original sediment. The clay minerals thus formed may give some clue to the diagenetic history of the Old Red Sandstone.

### 8.4.1. Detrital Clays

In the well-sorted ephemeral fluvial sediments of the Pease Bay Old Red Sandstone, the amount of primarily deposited clay material might not be expected to be great (Waugh 1978). Most samples, however, show a significant proportion of detrital clay. Some of this material has undergone a degree of diagenesis and indeed many of the samples showed a significant degree of authigenic clay mineral growth. In several samples, however, evidence of what is considered to be secondary deposition of clays was seen. Walker (1976), Walker et al (1978) and Waugh (1978) have discussed the presence of mechanically infiltrated clay in continental red bed sandstones of Caenozoic age. Fluvial sediments are characteristically mature and essentially free from primary detrital clay. In desert regions, however, during periods of flooding water may be recharged into the deposits. Due to the high porosity-permeability of the sediment, suspended clay is not immediately filtered from the water but settles out within the deposit. Furthermore, in a desert, wind blown loess may also contribute to the source of detrital clay, having been deposited within the upper layer of the sand and washed into the sediment by influent seepage. Walker et al (1978) suggest that in some cases as much as 20% matrix clay is present, almost wholly attributable to secondary infiltration.

Mechanically infiltrated clay may be distinguished from authigenic clay by the detrital nature of the particles which possess no crystalline form and by the characteristic appearance of the grain coatings, the clay particles lying parallel to the grain surfaces. Waugh (1978) has suggested that although infiltrated texture may be seen in sediments older than Caenozoic, this is likely to be masked by authigenic and diagenetic processes. Nevertheless, the Pease Bay sandstones show evidence of mechanical infiltration (Plate 30). A distinct 'bridge' of clay minerals may be observed between quartz grains. In addition, a comparison may be made between the upper faces of the quartz grains and their undersides - it is evident that there is more clay on the upper surfaces. This suggests that subsequent to the sand being deposited, clay was mechanically infiltrated.

The detrital clay minerals have been analysed using the EDX and the profiles shown in Plate 31. The cation ratio, which is purely qualitative, is approximately $Si_6Al_4K_1$ which compares with that of illite - $KAl_4(Si_7AlO_{20})$. Traces of iron (Fe) are also seen. Additional cations (eg Ca) which exist in montmorillonite series clays and smectites are not seen here. Kaolinite series clays do not contain potassium cations and thus this series may also be discounted.

In one instance, analysis showed a significant proportion of titanium, apparently substituting in the Al position in the lattice (Plate 32). Some iron cations are also evident but in very small quantities, as in the 'pure' illite. Titanium has been observed substituting for aluminium in illite by Walker (1950) and Carr et al (1953) who interpreted the clay minerals as weathering products of muscovite. The proportion of titanium in the Pease Bay example is considerably greater than those recorded in the literature, however, and may rather be the result of impurities related to sphene or an alteration product such as anatase or brookite.

Illite is the dominant clay mineral in primary clay deposits (mudstones etc) and is clearly the dominant clay mineral in the interstices between sand grains in the Pease Bay sandstones. It is suggested that much of this clay is secondary mechanically infiltrated clay as suggested above.

### 8.4.2. Authigenic clays

Although much of the clay is clearly primary in origin, there is evidence that diagenesis has had a significant effect and wispy illite can be seen (Plate 34) in nearly all samples. Cation analysis demonstrates this to be illitic in composition. Rarely, kaolinite is present and the hexagonal 'book' structure can be seen in some samples. EDX cation analysis shows comparative similarity in the quantities of Si and Al ions, consistent with the kaolinite formula - $Al_4(Si_4O_{10})$ (Plate 33).

The source of the authigenic clay minerals is most likely to be feldspar. Cathodoluminescence (section 7.4) demonstrates that these can make up up to 20% of the detrital material in some samples. Alteration textures are rarely seen in the feldspars which generally appear to have a fresh appearance. Secondary feldspar growth has not been identified in more than a few samples. More common is a pitted surface texture caused by etching of the grain surface (Plate 35). The etch marks are formed by differential dissolution of the mineral surface at particular sites. Wilson and Jones (1983) suggest that the dissolution sites represent the surface expression of structural dislocations of the lattice. These are attacked preferentially in the weathering process.

The general absence of kaolinite as an authigenic clay mineral and the lack of evidence for significant alteration of the potash feldspars suggests that the environment of deposition was not conducive to the hydrolisation of feldspar to clay minerals. The feldspar-kaolinite reaction can be represented by the equation:

$$2KAlSi_3O_8 + 9H_2O + 2H^+ \rightarrow Al_4Si_4O_{10}(OH)_8 + 4H_4SiO_4 + 2K^+$$

Krauskopf (1967) states that the weathering of feldspar requires acidic conditions and high temperatures. The above equation also implies the need for the presence of water in the reaction. It may be noted that the formation of kaolinite is linked with the production of silicic acid available for quartz overgrowth on sand grains.

The presence of quartz corrosion by carbonate solutions has been discussed above (section 8.3) as requiring a high pH and it is considered that such conditions were the norm during deposition of the Pease Bay Upper Old Red

Sandstone. In addition to the presence of alkaline conditions, the dry semi-arid climate (though with occasional stream flooding) provides an environment which is in general far from ideal for kaolinite formation and the associated development of quartz overgrowth. Such diagenetic features are indeed found only in isolated occurrences.

# 9. PALAEOCURRENT ANALYSIS

## 9.1. HISTORICAL FRAMEWORK

Cross-stratal dip measurements have been made on the Old Red Sandstone of the Berwickshire coast by Smith (1967) and Paterson et al (1976). Paterson et al took measurements at Pease Bay and used their readings as evidence that drainage from the Scottish Border Basin may have been confluent with an eastward-flowing river system in the Midland Valley. Their palaeocurrent readings have vector means between 76° and 113° which would support this hypothesis. Smith (1967) showed Old Red Sandstone cross-stratal dip measurements at Burnmouth which indicate a generally southerly palaeoslope. He interpreted the sequence as channel and overbank sediments deposited in a braided river system and on a semi-arid floodplain. Leeder (1973, 1974b, 1976b) suggests that the Berwickshire dip patterns are consistent with central drainage into a Scottish Border Basin. He does not accept the conjecture of Paterson et al (1976) that the Pease Bay measurements must imply an overall easterly trending drainage system and suggests that the variation in palaeocurrents between east and southwest might arise through a natural variance of channel trends. In the light of this controversy, a palaeocurrent analysis is considered essential to the interpretation of the Old Red Sandstone of the Pease Bay area.

## 9.2. METHOD

Primary sedimentary dip measurements were made where possible in all cross-stratified beds in the Pease Bay sequence. Potter and Pettijohn (1977) suggest that in unidirectional cross-bedding the sampling should be carried out with at least one reading per cross-stratal set. In practice, in an area of dipping strata such as the Berwickshire coast it is not always possible to make reliable readings on one face. Dip measurements usually have to be taken on two intersecting surfaces and a resultant dip calculated. This limits the number of readings which it is possible to take but provides a more reliable set of measurements. For the study, then, readings were taken in all beds and on as many sets as was feasible. A number of other factors limit the number of readings taken. Firstly, beds of facies A and facies B, parallel laminated siltstones and sandstones, by their very nature do not provide cross-stratal dips, though in one case, measurements were made on the reactivation cross-bedding in the upper part of a parallel bedded sandstone. Secondly, in some beds towards the upper part of the sequence, where cornstone growth has destroyed the primary sedimentary structures, no readings were possible. In a number of cases it was possible to take readings of cross-stratal trough axes in Facies D sandstones. Potter and Pettijohn (1977) suggest that these are more reliable estimates of palaeocurrent directions than random foresets.

The palaeocurrent data were corrected for structural dip using a Wolff stereographic net and divided into three groups: readings for all facies types, readings for facies C sandstones and readings for facies D sandstones. The raw data are summarised in Table 3. For each group, poles on the cross-stratification surfaces were plotted on a Schmidt equal-area stereographic projection. These results have been contoured and the raw point data and the contoured data are shown in Figs 18 to 20. Current rose diagrams were constructed and are shown in Fig 21. The rose diagrams take into account not only the cross-stratal dip directions but also the trough axis data. Vector means and magnitudes of the resultant vectors have been calculated for the three groups. These are shown on the current roses. The results data are summarised in Table 4.

The Tukey chi square test (High and Picard 1971) was performed on the readings for the three groups - all beds, facies C and facies D. The values for each of these groups are 60.34, 14.71 and 41.63 respectively. These figures indicate that a level of significance better than 0.01 can be attached to all the results and that they should not be considered as being randomly distributed. In addition, the facies D readings, because of their apparent polymodal distribution, were analysed by a method developed by Tanner (1955). The analysis showed a standard deviation of 4.4 for the palaeocurrent distribution and examination of the data in Table 4 reveals that in none of the intervals is there a number of measurements more than one standard deviation above the expected number of readings. Thus, in no interval is there a significant concentration to indicate that the distribution may be polymodal.

## 9.3. RESULTS AND CONCLUSION

The results of the palaeocurrent analysis demonstrates a strong predominantly southeasterly trend in facies C beds and an apparently more diverse (possibly polymodal) distribution in the facies D beds. The overall trend (all beds) appears to be southeasterly. As discussed earlier, the facies C sandstones are interpreted as low-sinuosity river channel deposits. The prevalent southeasterly palaeocurrent direction would appear to indicate a southeasterly palaeoslope and river drainage in that direction. The facies D sandstones, however, which are interpreted as shallow floodplain deposits initiated by overbank flooding, have a more diverse distribution of directional data. This is consistent with the model of deposition on either side of the main river channel but also controlled by the southeasterly palaeoslope.

The palaeocurrent trends would not appear to support the model of Paterson et al (1976) who envisaged that drainage from the Scottish Borders Basin was confluent with that of the Midland Valley. On the other hand, Leeder's model (1973, 1974b) also does not entirely account for the general southeasterly palaeocurrent direction. Leeder (1974b) assumes a general southerly palaeoslope in the British Isles with the shoreline to the south. In the light of more recent knowledge from the North Sea (Ziegler 1981), a more embracing model may now be presented. It seems likely that the late Devonian - early Carboniferous shoreline lay to the south of Britain and curled round the southern edge of the London Brabant Massif into the southern North Sea area (Fig 22).

The Midland Valley drainage may well have been easterly, controlled by the Highland Boundary and Southern Upland fault scarps to the north and south. At the eastern end of the rejuvenated Southern Uplands block, the drainage would then turn to run southeasterly. A generally southeasterly palaeoslope would then be consistent with this model. The Borders Basin may be considered to be an internal drainage basin, eventually draining to the south to the west of the London Brabant Massif. The cornstones of the Berwickshire and Scottish Borders Old Red Sandstone may indicate that the significant periods of non-deposition were related to poorly drained parts of the land area, the main drainage areas being to the north and south.

# 10. DIAGENETIC CONSIDERATIONS

Diagenesis of the sediments has been discussed briefly in previous chapters. Two aspects deserve further consideration, however. These are cornstone development and red bed genesis.

## 10.1. CORNSTONE DEVELOPMENT

Cornstone classification is first discussed with reference to the literature and this is followed by a consideration of the Pease Bay cornstones.

### 10.1.1. Cornstone classification

There appears to be some confusion over the terminology in the literature. The terms cornstone, calcrete, caliche and duricrust are all used by various authors to describe similar rock-types. I use the word calcrete to refer to the calcitic chemical deposit and cornstone as a more general term which applies also to silcretes and ferricretes. The terms caliche and duricrust are applied specific modes of formation of strongly carbonate-cemented clastic sediments, caliche forming within a soil profile and duricrust at or near the sediment surface.

Allen (1960) has identified two distinct types of cornstone, the first being concretionary, the second deposited as a sediment - conglomeratic cornstone. The term concretionary cornstone is used to describe carbonate-rich or strongly carbonate-cemented clastic sedimentary rocks where the original fabric has been largely destroyed by carbonate growth. In Allen's second type of cornstone, the conglomeratic clasts are largely cornstone which is presumably derived from earlier concretionary cornstones. In the Old Red Sandstone of the present study area, these are quite common in the lower part of the sequence and in some cases the clasts appear to form the loci for later concretionary development.

In the Old Red Sandstone of the Berwickshire coast, it would seem unlikely that any true soil profiles would be present. The palaeoclimate is believed to have been semi-arid but in late Devonian times the amount of land-based vegetation is considered to have been less than at the present day. Although at Rhynie it is known that prolific plant growth was present, this was most likely to have been isolated in aerial extent, possibly an oasis. Elsewhere in the desert, plant growth will have been severely restricted. As a result, although it would be expected that geochemical profiles were present in the sand, these could not be classified as a true pedogenic profiles.

Duricrusts, on the other hand, might be expected to play a significant part in the carbonate history of the Old Red Sandstone environment. These have been described in several modern environments including western South Africa (Knox 1977), Tanzania (Hay and Reeder 1978) and Botswana (Mallick at al 1981). The duricrusts are found at or near the sediment surface, usually overlying desert sand and highly indurated on their upper surfaces. Fractured surfaces penetrating the calcrete often become closed with carbonate cement and primary structures such as bedding and roots are normally eradicated in the development of the caliche. In Botswana, they often appear as cliffs along the major water courses and comprise calcrete overlying silcrete. In some parts the total thickness of the duricrust can be 35m. In major pans, they are commonly overlain by thick sand but are seen very near to the surface on the rims of the pans. Aeolian dunes frequently cover the duricrusts, the sand in the inter-dune areas being very thin or absent.

### 10.1.2. Cornstones of Pease Bay to Horse Roads Rock

The cornstones of the Pease Bay to Horse Road Rock Upper Old Red Sandstone are confined to the upper part of the sequence (Units III - VIII). In the lower of these units, the cornstones are thin and poorly developed while in the upper part, they comprise a significant part of the section, becoming generally thicker in the later beds and culminating in the Main Cornstone which is generally taken as the top of the Old Red Sandstone. Above this, the sequence is quite different, comprising mudstone, marl, cementstone and flaser-bedded very fine sandy siltstone.

In the lower part of the section, where no cornstones are seen, some poorly-developed cornstone clasts are present in Facies D sediments and small carbonate concretions (up to 5mm across) are found in Facies B and other beds (Plate 16). In Unit III the cornstones comprise largely conglomeratic horizons and isolated concretionary growth. These concretions are often vertical or sub-vertical pipe-like bodies (Plate 17) while the conglomeratic horizons appear to form as lags in channels (Plate 8). These appear in many cases to have been subjected to later diagenetic growth as they have formed loci for concretionary cornstone development (Plate 18).

Higher in the section, the cornstones become more significant features, often forming continuous horizons at the top of the sandstone beds (Plate 19). This progression from small concretions to extensive horizons may be seen not only as a stratigraphic trend but also as a representation of the various developmental phases of the cornstones identified by Reeves (1970), Netterberg

(1980) and Netterberg and Caiger (1983). In the lower part of the sequence, it is suggested that the rate of sedimentation was greater than in the upper part. It is anticipated that the various stages of early diagenetic development were 'frozen' when the sediments were buried beneath the vadose zone. In the upper part of the section, extensive periods of non-deposition would result in complete cornstone development.

The chemical reaction involved in calcium carbonate solution and precipitation can be represented by the following equation:

$$CO_2 + H_2O + CaCO_3 \rightleftharpoons Ca^{2+} + 2HCO^-$$

The solubility of calcium carbonate is affected by five independant variables (Klappa 1983) which are temperature, pH, partial pressure of carbon dioxide, addition of salts with common ions and addition of salts lacking common ions. Precipitation of carbonate is enhanced by the removal of carbon dioxide and water from the system. This may be effected both through aeration of the host sediment, reducing the partial pressure of carbon dioxide and by evaporation of water which, if taking place in the daytime may also be accompanied by a temperature increase favourable to carbonate precipitation.

The addition of salts with and without common ions is brought about by interfering or mixing natural waters. Movement of ground water through the sediment is largely due to descending gravitational water and rising capillary water. In semi-arid conditions, no equilibrium state would be reached - rainfall causes an increase in gravitational fresh water to the system and evaporation during dry spells causes a concentration of salts in the capillary water. Lastly carbonate precipitation may be affected by changes in pH. A downward-increasing pH gradient may be controlled by biological agents such as microflora and bacteria (Krumbein 1968).

In the Old Red Sandstone of Berwickshire, the most likely factors in the sediment profile to affect the precipitation of calcium carbonate are evaporation associated with high day-time temperatures, movement of ground water and the presence of a pH gradient. It may be seen that the first two of these factors are primarily climatic and it is accepted that the presence of carbonate concretions and indeed cornstones are clear climatic indicators. The extent of concretion growth is probably related to the length of the periods of non-deposition.

The small concretions which form the earliest phase of development are found in sediments throughout the sequence. They are most common in the Facies B parallel bedded sandstones. The concretions range in diameter from 3mm to 10mm though in any one horizon they appear to be similarly sized. In thin section, they are shown to comprise polycrystalline micrite. Staining with alizarine red shows them to be dominantly calcitic. The concretions occur within the sand body, usually in a zone between 0.3m and 0.4m below the top of the bed. In places, the concretions become more concentrated towards the top of the zone which thus appears to have a 'sharp top'. These features are consistent with the 'early mature' stage of Reeves (1970) or Stage II of Gile et al (1966). The zoning is recognised as part of a geochemical profile which also affects the red-buff colour of the sediments (see section 10.2).

Similar nodules are reported in the Lower Old Red Sandstone of the Anglo-Welsh outcrop (Allen 1974a). Here the nodules or 'glaebules' range in size from a few millimetres to 50mm in diameter. They are found predominantly in mudstones, interpreted as floodplain deposits. This is characteristic of other occurrences of cornstone glaebules in Old Red Sandstone in the Clee Hills (Allen 1974b), in western Norway (Bryhni 1974) and in Antarctica (McPherson 1979). In the Pease Bay to Horse Roads Rock section, the sediments are less argillaceous than in the Anglo-Welsh outcrop and the carbonate growth is consequently found in sandy facies. It is suggested that the dominance of the nodular concretions in Facies B sediments is related to the slow rate of sedimentation on the flood plain.

In places the nodules are found to grow larger and to coalesce into carbonate cemented horizons. This is interpreted as the next stage of cornstone development. A 'honeycombed' texture is apparent, similar to those recognised by Reeves (1970) and Netterberg and Caiger (1983). It corresponds to the 'mature' stage of Reeves (1970) and Stage III of Gile et al (1966). In the higher parts of the section, this type of occurrence is less common than might be expected. Instead, the development of vertical and sub-vertical pipe-like bodies is seen.

Structures of this kind have been described in the literature by Parnell (1983) and he believes them to have been related to plant root growth. On the other hand, they may have developed as mechanical weaknesses in the sand beds where mineral-charged waters have been allowed to percolate, leaving a residue of ferric, magnesium and calcium carbonates.

The final stage of cornstone development is represented by sheet like concretionary growth found along the upper surfaces of the beds. The surfaces tend to be highly indurated and fractured. Associated with some cornstones are cherty concretions. These form in the interstices in the lower part of the profile and represent the development of the 'old' stage of Reeves (1970) and stage IV of Gile et al (1966). Reeves (1970) suggests that the 'old' stage requires 'a considerable time to develop'. Talma and Netterberg (1983) indicate that ages greater than 10000 years may not be uncommon for semi-arid calcrete development.

## 10.2. RED BED GENESIS

The genesis of red beds has been discussed in the literature by several authors and the theoretical considerations will not be discussed in detail here. The reddening of desert sands in particular has been considered by Friend (1966), Walker (1967), Schmalz (1968), Folk (1976), Turner (1980) and Gardner and Pye (1981).

The reddening of the sandstones in the Pease Bay to Horse Roads Rock section may be attributed to two factors - the haematite rims to the quartz grains and the clay fraction. There are two possible modes of red pigmentation - primary, involving oxidation of the sediment penecontemporaneously with deposition and secondary, involving reddening subsequent to burial of the sediment. In addition, there is the possibility of deposition of previously reddened sediment. Much of the Pease Bay section comprises mature quartzose and feldspathic sandstone. It is likely that this represents deposition of eroded Lower Old Red Sandstone sediments.

There is no evidence for the presence of any significant proportion of ferromagnesian detrital minerals in the sandstones, a point also been noted by Douglas (1972). In the small early-stage carbonate concretions, there is no evidence of dissolution or replacement of any detrital components. No ghost forms or skeletal grains are observed. Primary, penecontemporaneous oxidation or in situ mineral breakdown as described by Walker (1967), Folk (1976) and Turner (1980) is therefore not considered to be the main method of reddening of the Pease Bay sandstones.

At the base of the section, at the northern end of Pease Bay, a remarkable example of red - drab colouration can be observed. This takes place in Facies B sandstone which is irregularly eroded - the top surface of the bed is undulatory and infilled by a Facies C cross bedded sandstone. The outcrop is shown in Plate 20. The upper 0.25 to 0.3m comprises red sandstone, the red colour being predominantly attributable to haematite rims on the quartz grains. The junction between the red colouring above and the buff colouring below appears to be partly controlled by the shape of the erosional surface on top of the bed. At a further depth of 0.2 to 0.25m, carbonate concretions are observed. These have a 'sharp top' as previously described (section 10.1.2) and this surface also appears to follow the shape of the upper erosional surface. It is suggested that this represents an early stage geochemical profile forming after the erosion of the top of the bed but before the deposition of the Facies C sandstone.

The presence of this feature above suggests that the red pigmentation is an early feature rather than secondary oxidation subsequent to burial of the sediment. It also confirms that the early cornstone growth can develop over a relatively short period of time i.e. between the erosion of the bed and the subsequent deposition of the cross-bedded sands above.

Although the presence of one example cannot be used as a generalisation for the entire section, it is suggested here that the evidence rules out both primary and secondary forms of red pigmentation. The drab sands at the base of the outcrop must therefore have been reduced from an original red coloured sediment. This red pigmentation in the sand may either be provenance controlled or related to the immediate history of the sediment. As there is a lack of evidence to indicate that ferromagnesian detrital minerals were present in the sediment, it is suggested that the red colour is predominantly provenance controlled, the source being the Lower Old Red Sandstone of the Midland Valley.

## 11. CONCLUSIONS

The study of the Upper Old Red Sandstone of Pease Bay to Horse Roads Rock has provided evidence on the environments of deposition in the area and presents implications for the regional paleogeography of Southern Scotland during the late Devonian and early Carboniferous.

The sequence from Pease Bay to Horse Roads rock has been divided into six distinct facies types. On the basis of various facies associations, eight depositional units have been identified. The lower three units comprise interbedded sandstones which are dominantly fluvial but with evidence of wind-blown sediment. The upper five units comprise similar interbedded sandstone and are characterised by the presence of cornstones. Lithostratigraphic similarities may be noted between these two major divisions and the upper two divisions of the Upper Old Red Sandstone in the Midland Valley of Scotland, the Glen Vale Formation and the Kinnesswood Formation. Evidence from Westoll (1951), Clayton (1971) and M.A.E. Browne (pers comm) suggests that the Glen Vale Formation is Upper Devonian and the Kinnesswood Formation, Lower Carboniferous (Tournasian) in age.

The palaeocurrent analysis using cross-stratal dip measurements indicates a predominantly south-easterly palaeocurrent trend. This suggests that the area of the Berwickshire coast may have been on the margins of the Midland Valley basin which, although having a predominantly easterly palaeoslope, is believed to be confluent with the North Sea Basin which drains to the south (Ziegler 1981). The sediments of the Pease Bay section, being on the margins of a large basinal complex, were possibly more poorly drained than those in the central part of the basin. Extended periods of the non-deposition are indicated by the presence of the cornstones in the upper part of the section. The Borders Basin, to the southwest of Pease Bay, is considered to be a more isolated internal drainage basin.

In the lower part of the sequence, the sediments are shown to be largely fluvial. The sequence is dominated by parallel-bedded and trough cross-bedded sandstones interbedded with occasional silty sandstones. The parallel-bedded sandstones are interpreted as having been laid down by seasonal flooding of an alluvial plain. The trough cross-bedded sandstones are interpreted as channel and high stage floodplain deposits. Some sets of channel sandstones are characterised by closely-spaced asymptotic laminae and this, taken in conjunction with later evidence, suggests wind-blown sedimentation. The sediments interpreted as floodplain deposits generally comprise shallow sets with abundant rip-up mudstone clasts. The argillaceous beds are interpreted as low-energy playa lake sediments, possibly deposited in abandoned channels.

Grain size analysis provides evidence in the form of cumulative frequency curves which approach the shape of those which Visher (1969) constructed from fluvial sand analyses. Cross-plots of a number of grain size parameters were compared with cross-plots drawn by Friedman (1979) and others. The results generally fall into the fluvial fields or span the fluvial and aeolian dune fields. It has been suggested that the grain size distribution of a sediment is related not only to the environment of deposition but also to the past history of the sediment. The Pease Bay sediments, although dominantly fluvial, may well have had a history of aeolian transport.

Scanning electron microscopy demonstrated several surface features on the quartz grains, including modified 'upturned plates', shallow depressions and frosting - all indicative of aeolian transport. The presence of globular diagenetic overgrowth has, however, masked many of the primary surface textures. On the other hand, this globular overgrowth may itself be the microscopic represention of desert varnish caused by repeated solution and precipitation of silica in carbonate-charged dew (Marzolf 1976).

In the upper part of the sequence, the trough cross-bedded channel sandstone becomes the dominant facies type. Floodplain deposits, represented by parallel-bedded sandstones and shallow trough cross-bedded sandstones with mud clasts, are frequently associated with cornstone development. The presence of cornstones within the Pease Bay sequence may be traced through several phases, each representing a stage of calcrete development. This is seen as a stratigraphic trend within the sequence and suggests an increasing tendency towards drying-out of the sediments during long periods of non-deposition.

Leeder (1976) has suggested that the presence of cornstones at the top of the Upper Old Red Sandstone in the Borders Basin signifies a period of basin upwarp related to partial melting of the upper mantle. The significance of cornstones in the upper part of the Pease Bay to Horse Roads Rock sequence suggests that the upwarp of the Borders Basin may have affected the present study area. This, together with climatic controls, gave rise to the progressive cornstone development throughout the upper part of this section.

The lack of rock fragments and ferromagnesian and accessory minerals, noted by Douglas (1972), presents difficulties in determining the provenance of the sandstones. A key to the provenance is provided, however, by a petrographic study of the quartz grains in both transmitted light and cathodoluminescence. These techniques demonstrate that two distinct quartz types are present, one well-rounded, inclusion-free and exhibiting violet luminescence and the other sub-angular to sub-rounded, full of inclusions and luminescing brown. A high-temperature, possibly volcanic, provenance is indicated for the former quartz type and a low-temperature metamorphic provenance for the latter. Source rocks could include Lower Old Red Sandstone andesites and Caledonian granites and regionally metamorphosed rocks of the Highlands and Southern Uplands. The Lower Old Red Sandstone sediments of the Midland Valley may also provide a significant source of sand grade material for the Upper Old Red Sandstone. In view of the wide range of sources, provenance studies provide us with very little useful information.

A high proportion of feldspar (up to 20%) is present in many of the sandstones. This is dominantly potassium feldspar, though some plagioclase is also present. Although the Lower Old Red Sandstone andesites may provide a source for the plagioclase, the potash feldspar is more likely to be derived from more acidic rocks. Although most feldspar grains appear fresh and angular, scanning electron microscopy reveals that the grain surfaces have a pitted texture. This is caused by etching of the surfaces by percolating intrastratal solutions. The overall freshness of the grains, however, suggests that during deposition and subsequent burial of the sandstones, the conditions were not conducive to hydrolisation of feldspar. The evidence from cornstone development suggests that rapid erosion and deposition is an unlikely mechanism and we must look to semi-arid climatic conditions to explain the abundance of feldspar in the sandstones.

The Pease Bay to Horse Roads Rock sequence is, therefore, believed to be a marginal development of the Midland Valley Upper Old Red Sandstone river system. The sediments represent the formation and dissection of a number of alluvial fans, together with the accumulation of wind-blown sands during periods of fluvial abandonment. In the lower part of the section, the argillaceous deposits are taken to indicate the presence of playa lakes, possibly forming in abandoned channels. Towards the top of the sequence, the increasing growth of cornstone suggests significant periods of non-deposition.

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University of St Andrews

# The Sedimentology of the Old Red Sandstone Section from Pease Bay to Horse Roads Rock, Berwickshire

by

J Alan Fyfe

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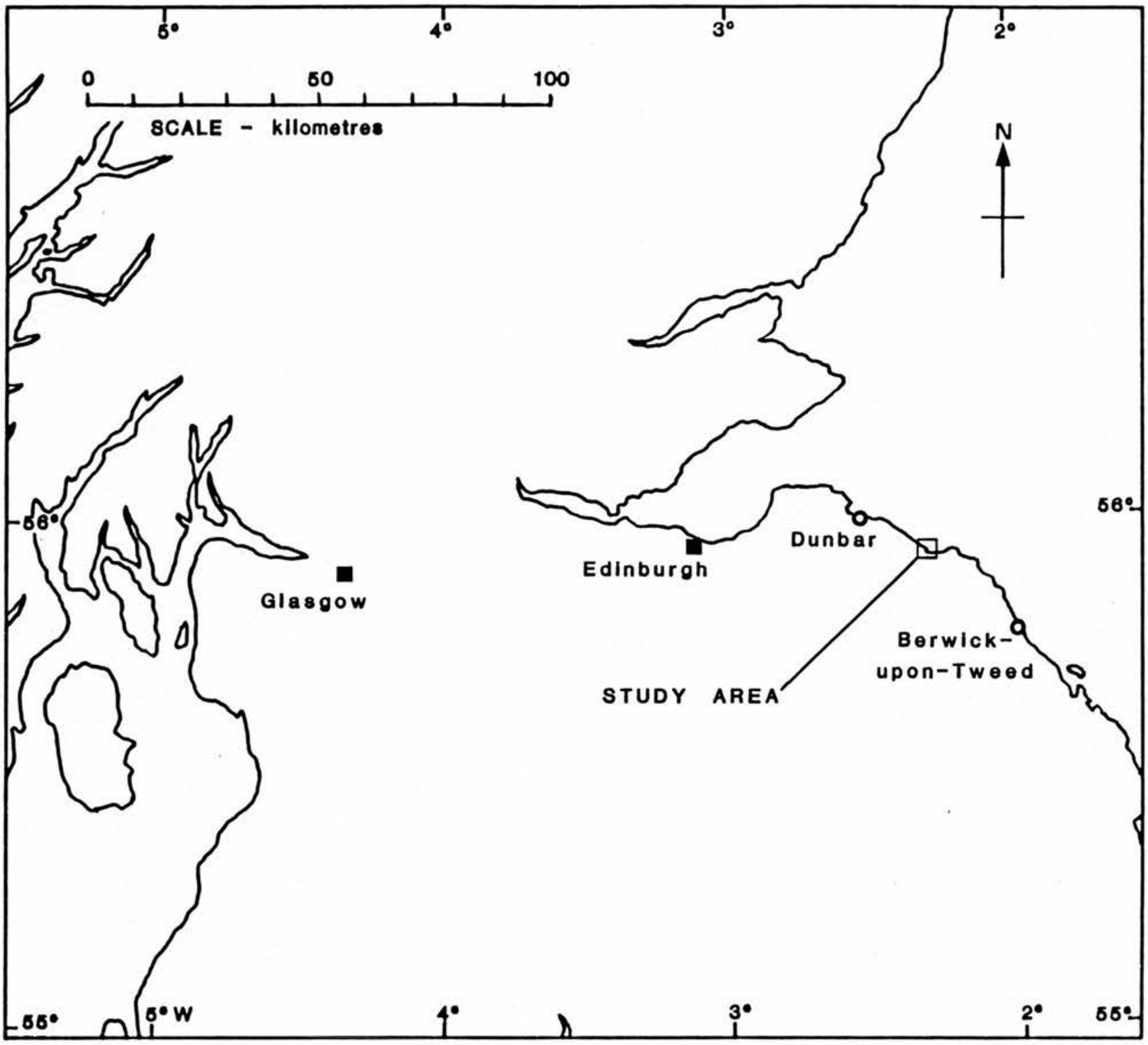



Fig 1. Location map of the study area

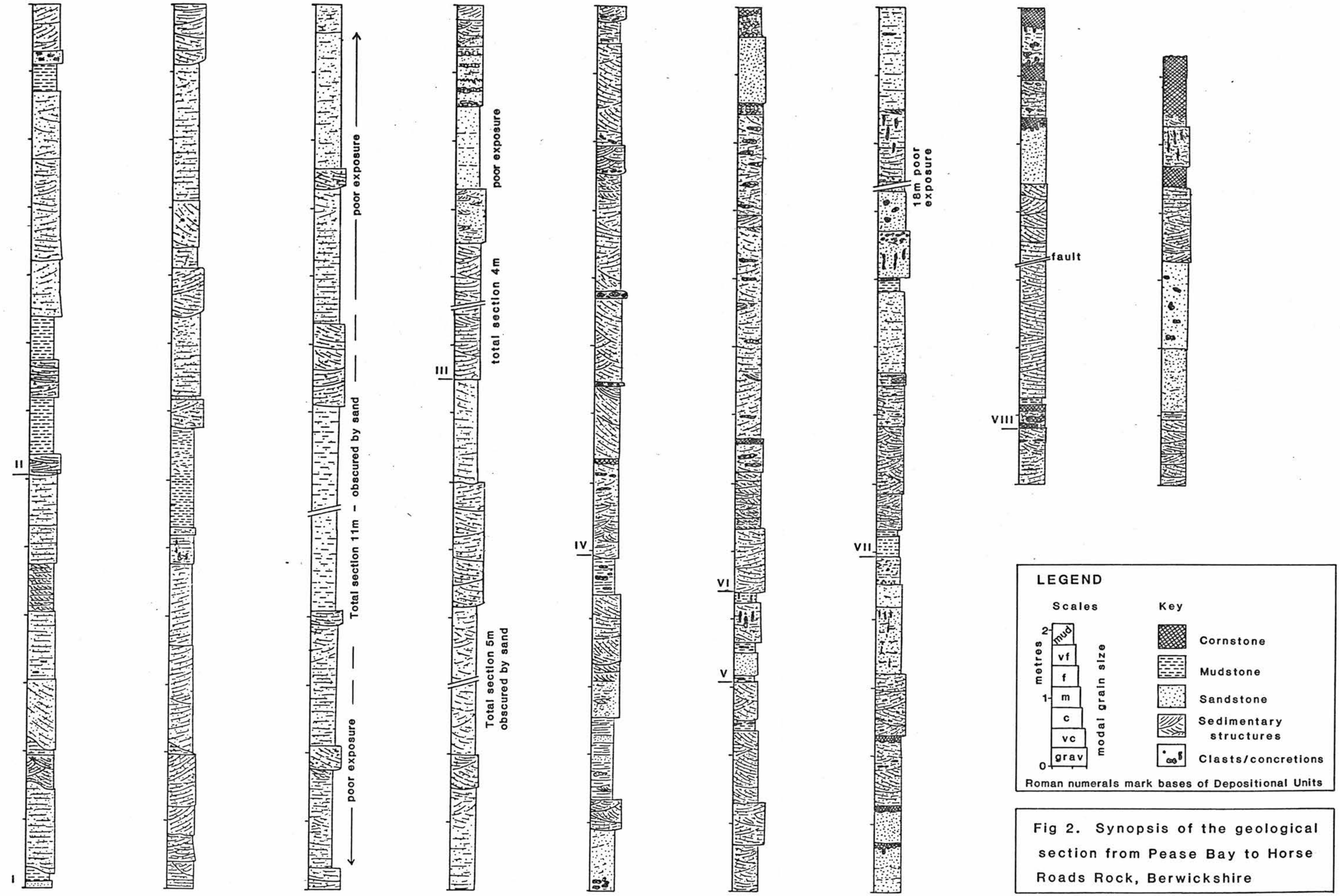



Fig 2. Synopsis of the geological section from Pease Bay to Horse Roads Rock, Berwickshire

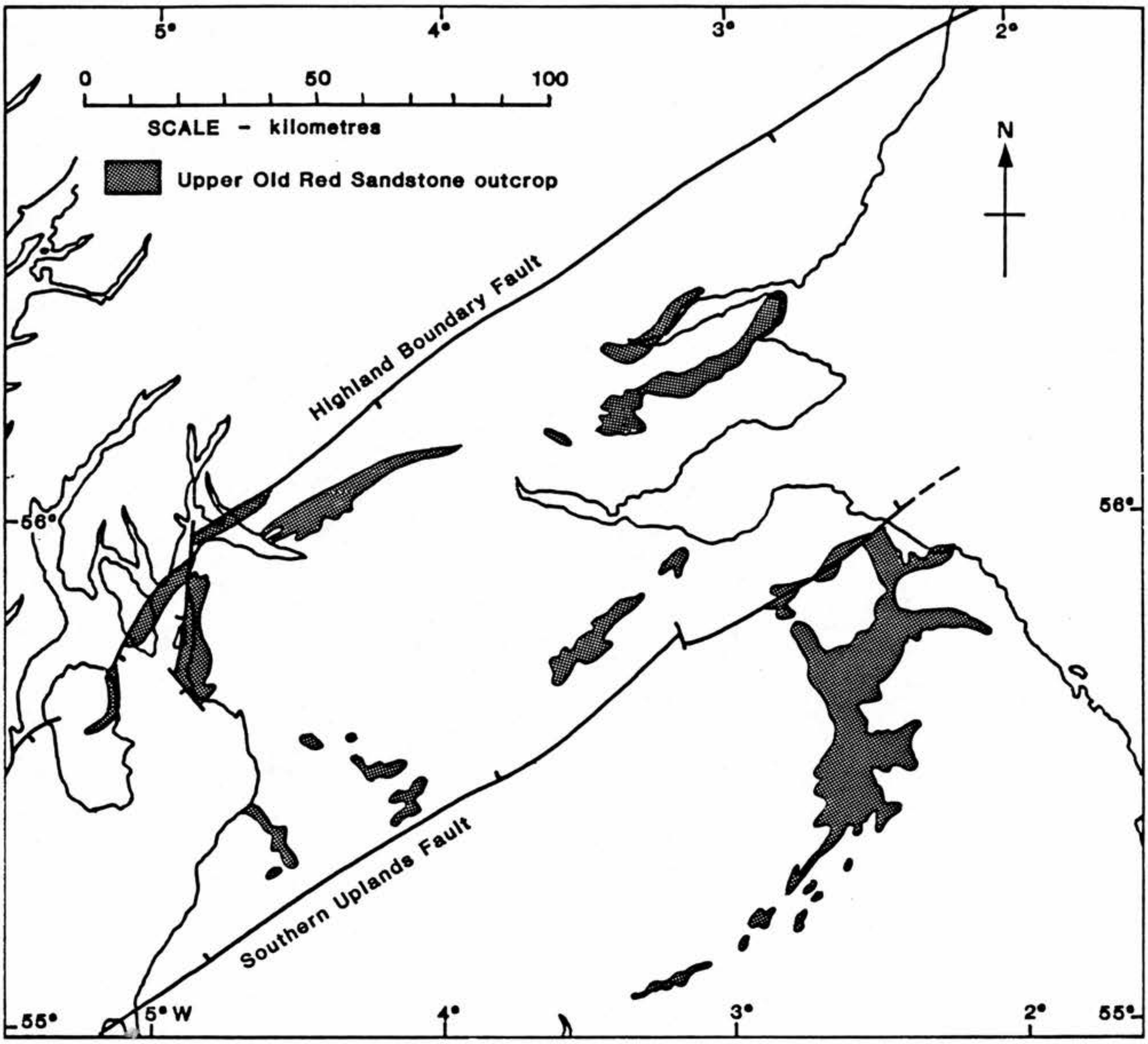



Fig 3. Upper Old Red Sandstone outcrops in southern Scotland

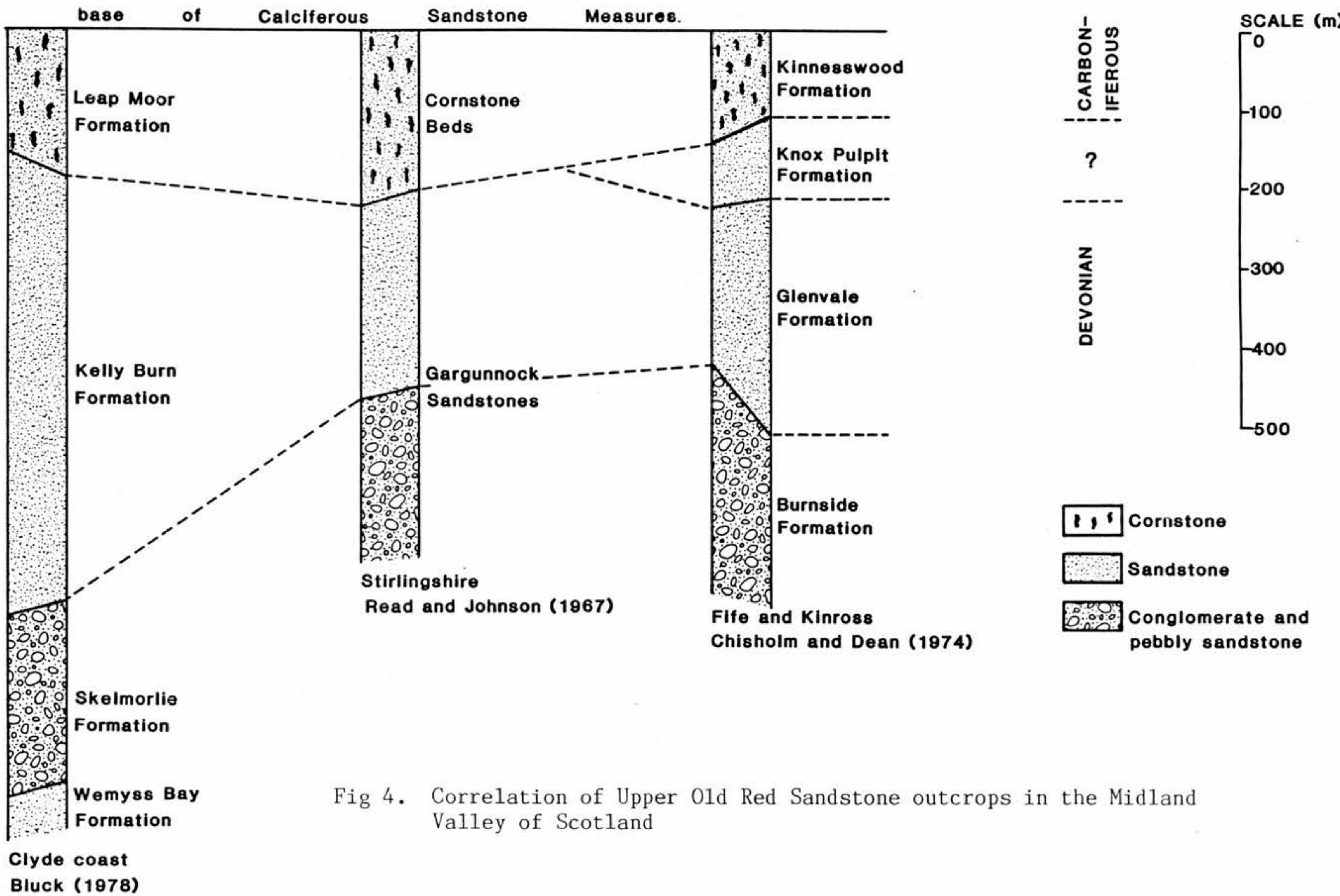



Fig 4. Correlation of Upper Old Red Sandstone outcrops in the Midland Valley of Scotland

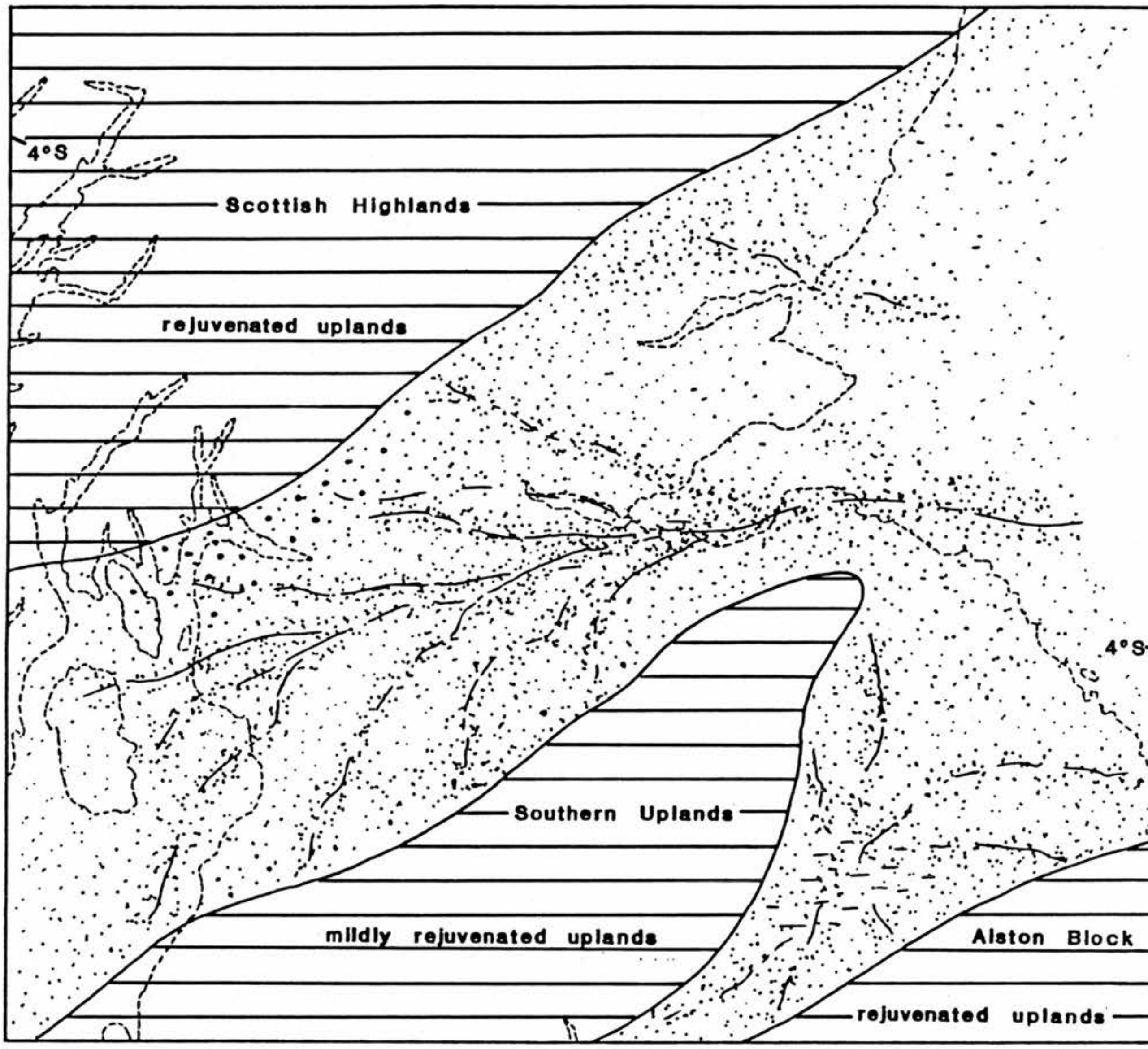
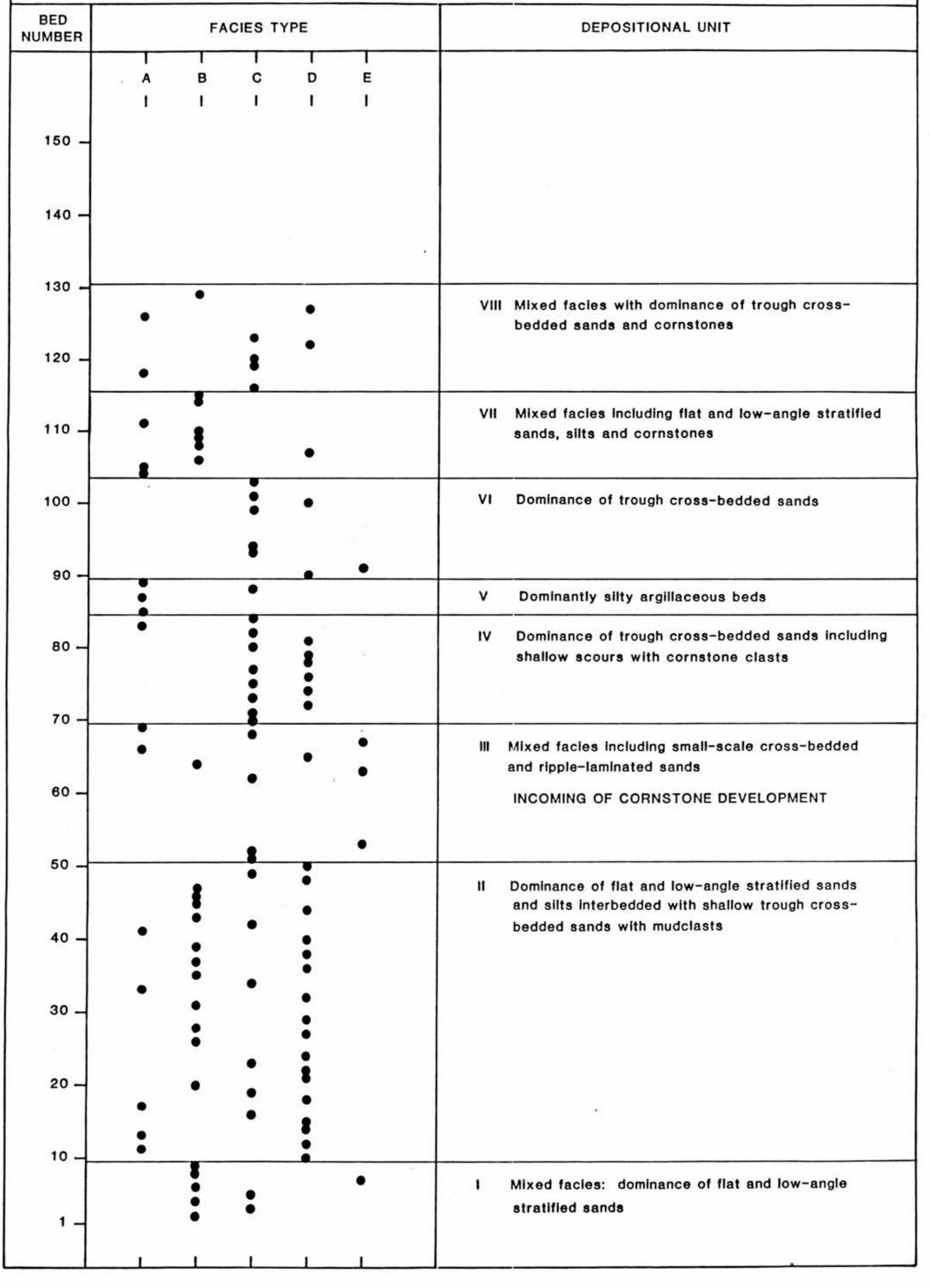



Fig 5. Palaeogeographic reconstruction of the Midland Valley and Borders Basins in Upper Old Red Sandstone times based on Bluck (1978) and Leeder (1973)

Fig 6. Distribution of Facies types and division into depositional units



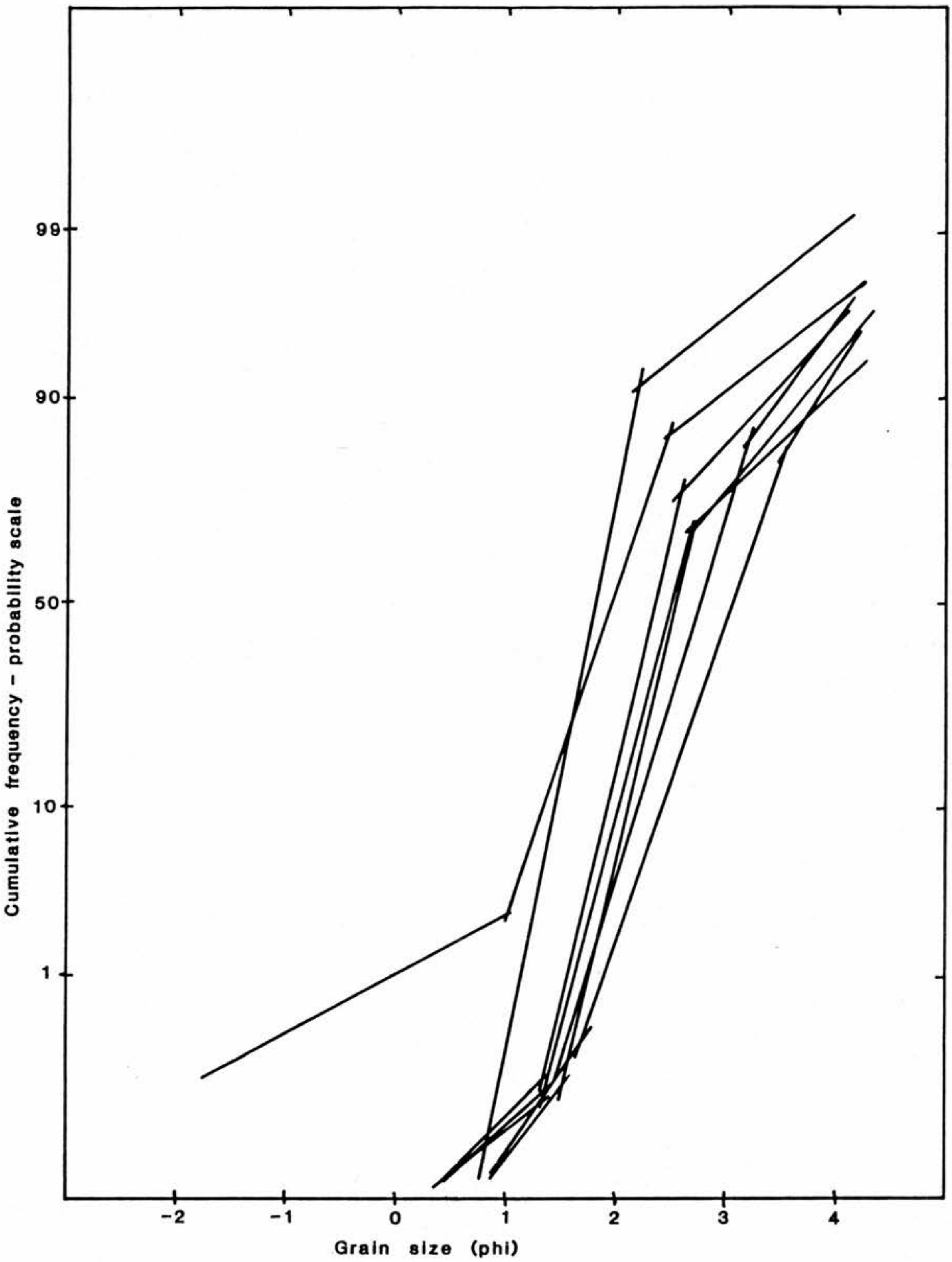



Fig 7. Cumulative frequency curves for Facies B sandstones

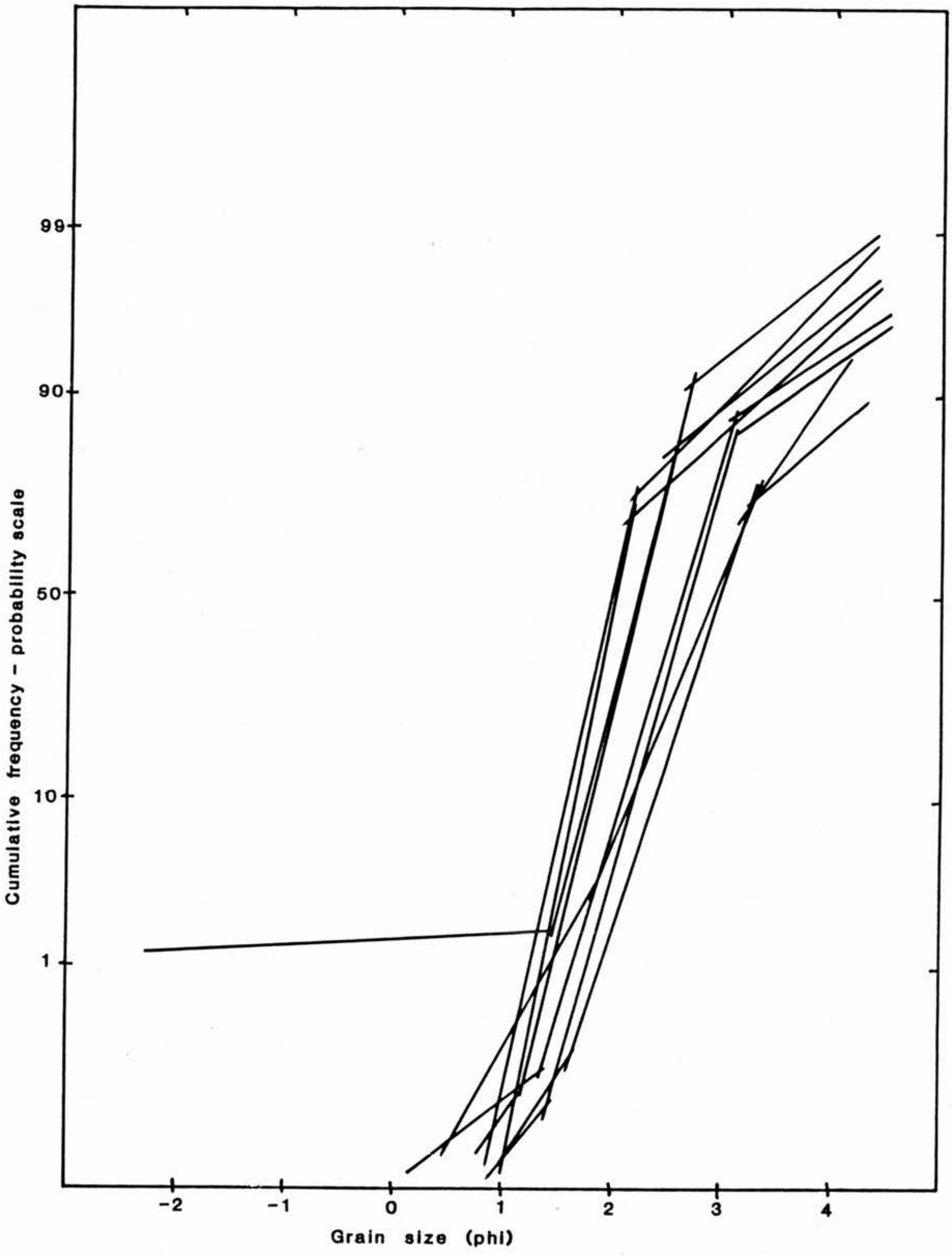



Fig 8. Cumulative frequency curves for Facies C sandstones

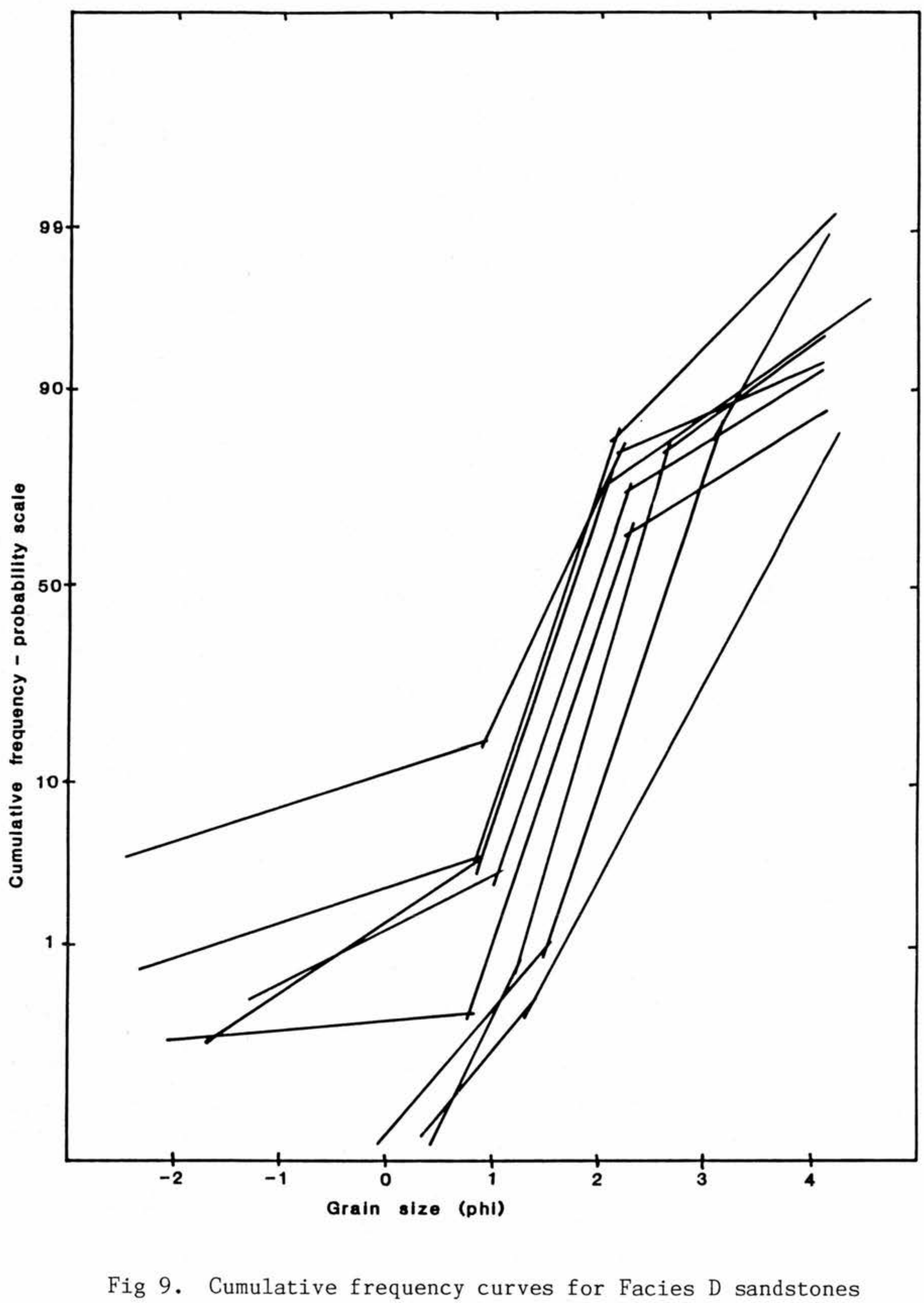



Fig 9. Cumulative frequency curves for Facies D sandstones

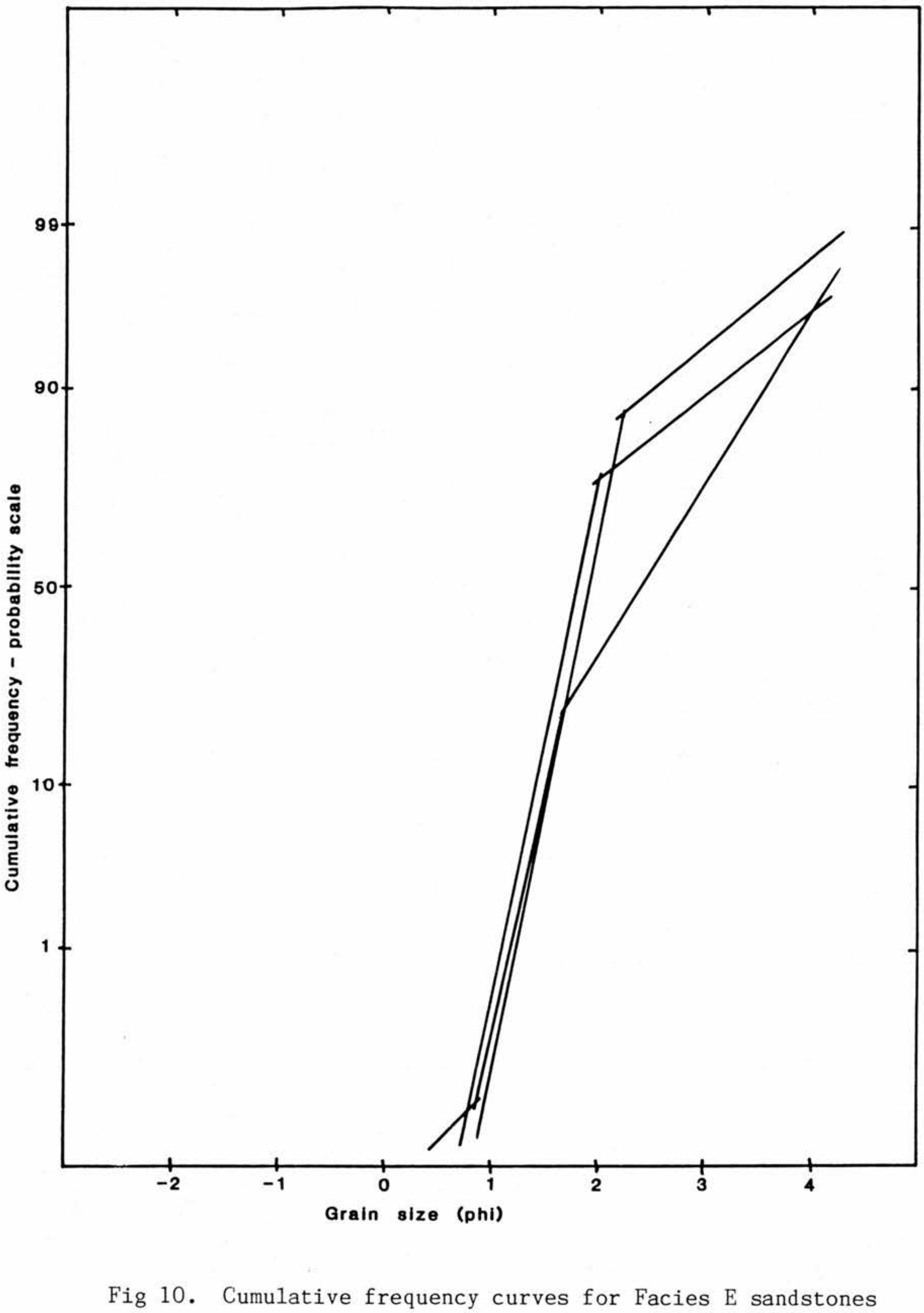



Fig 10. Cumulative frequency curves for Facies E sandstones

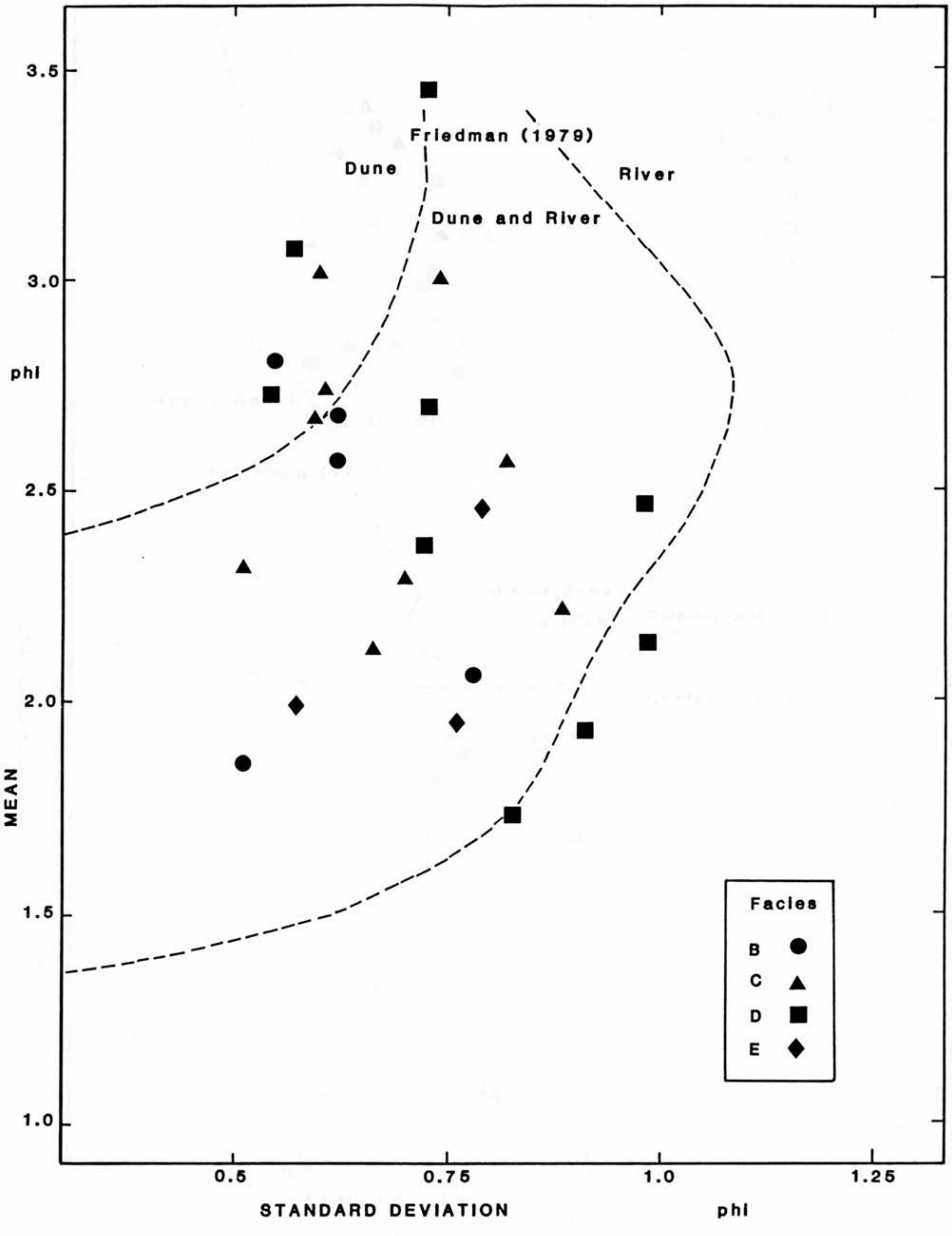



Fig 11. Cross-plot: mean v standard deviation

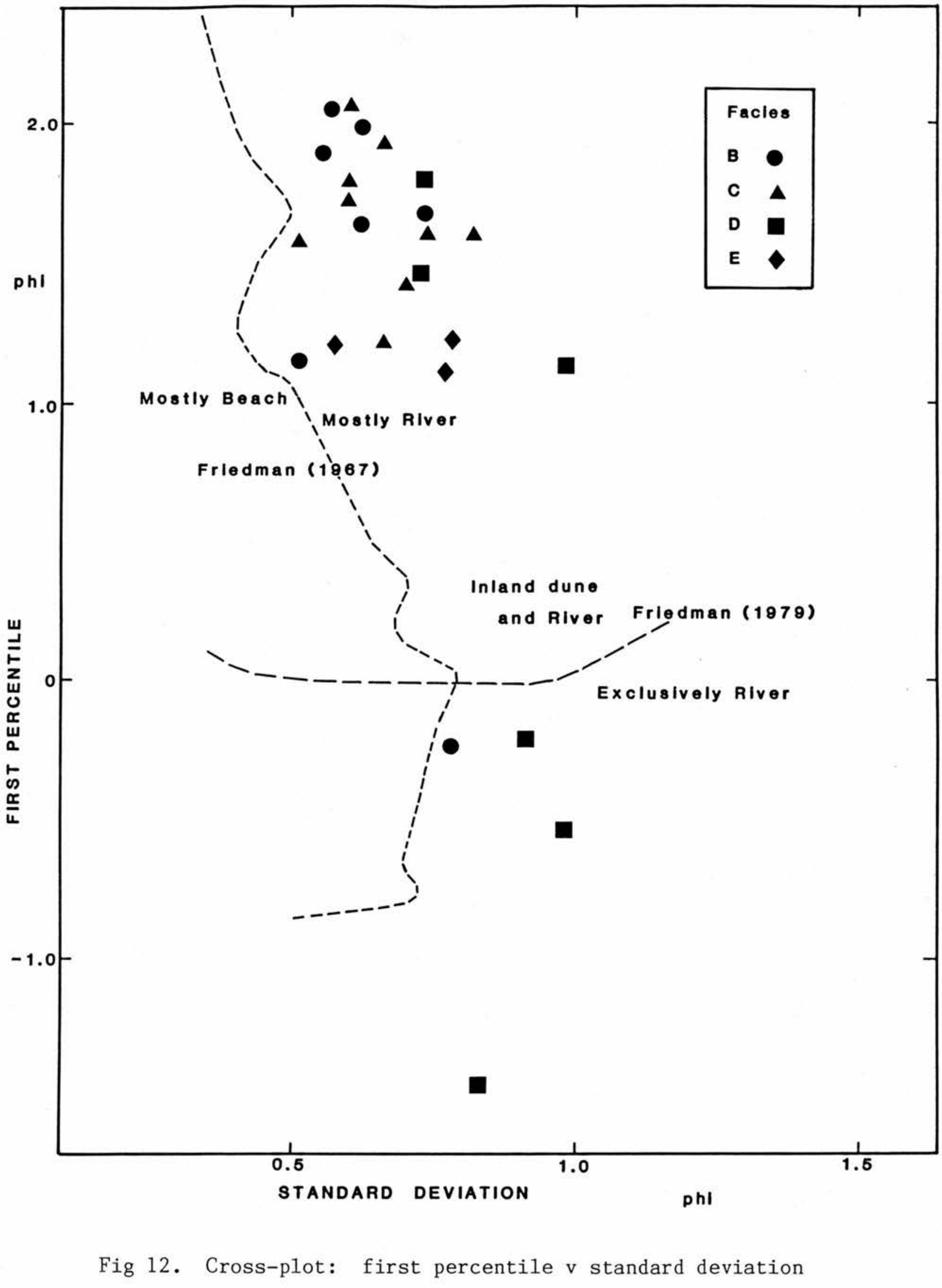



Fig 12. Cross-plot: first percentile v standard deviation

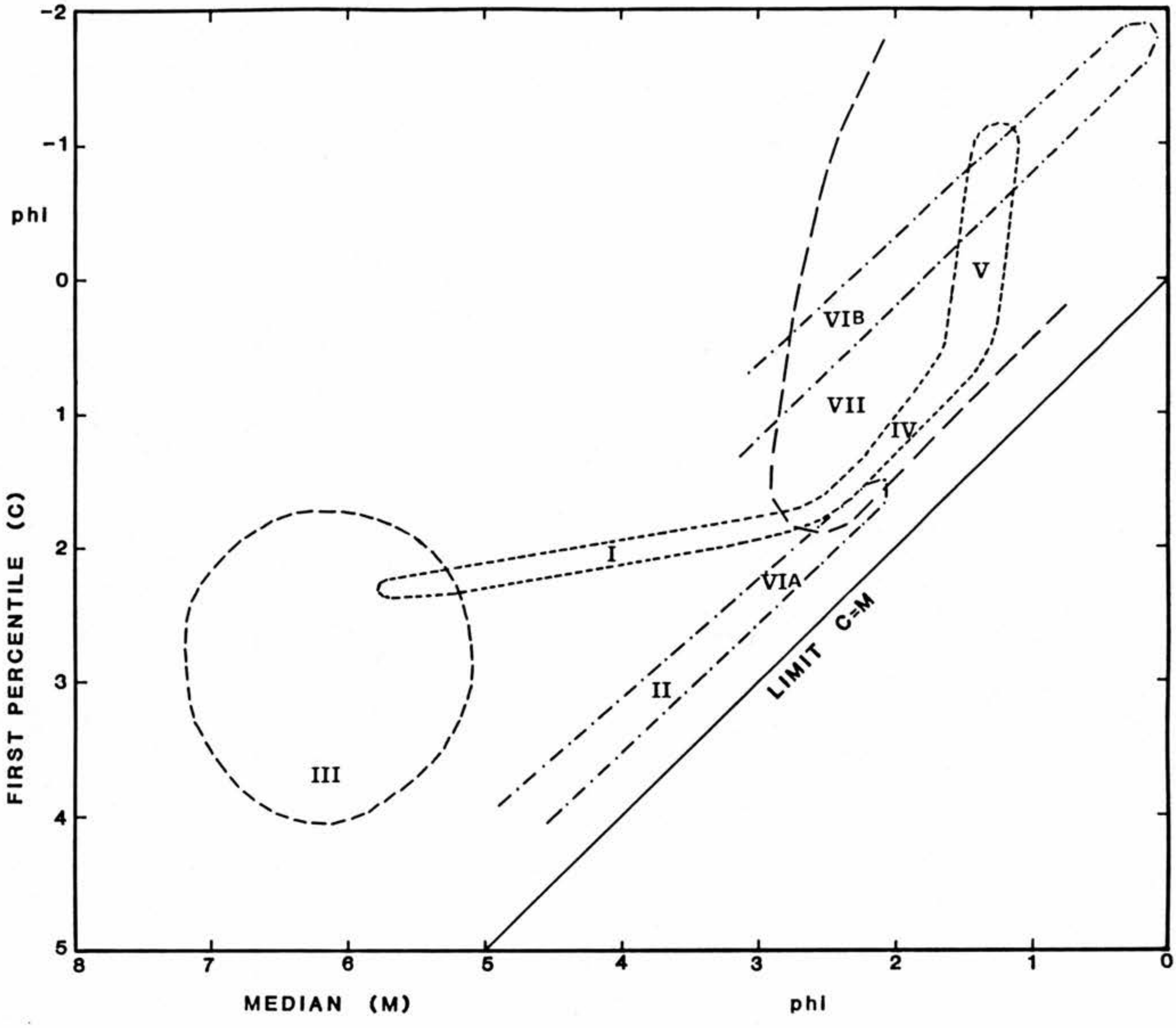



Fig 13. C – M diagram showing fields for various depositional environments (after Passega (1957))

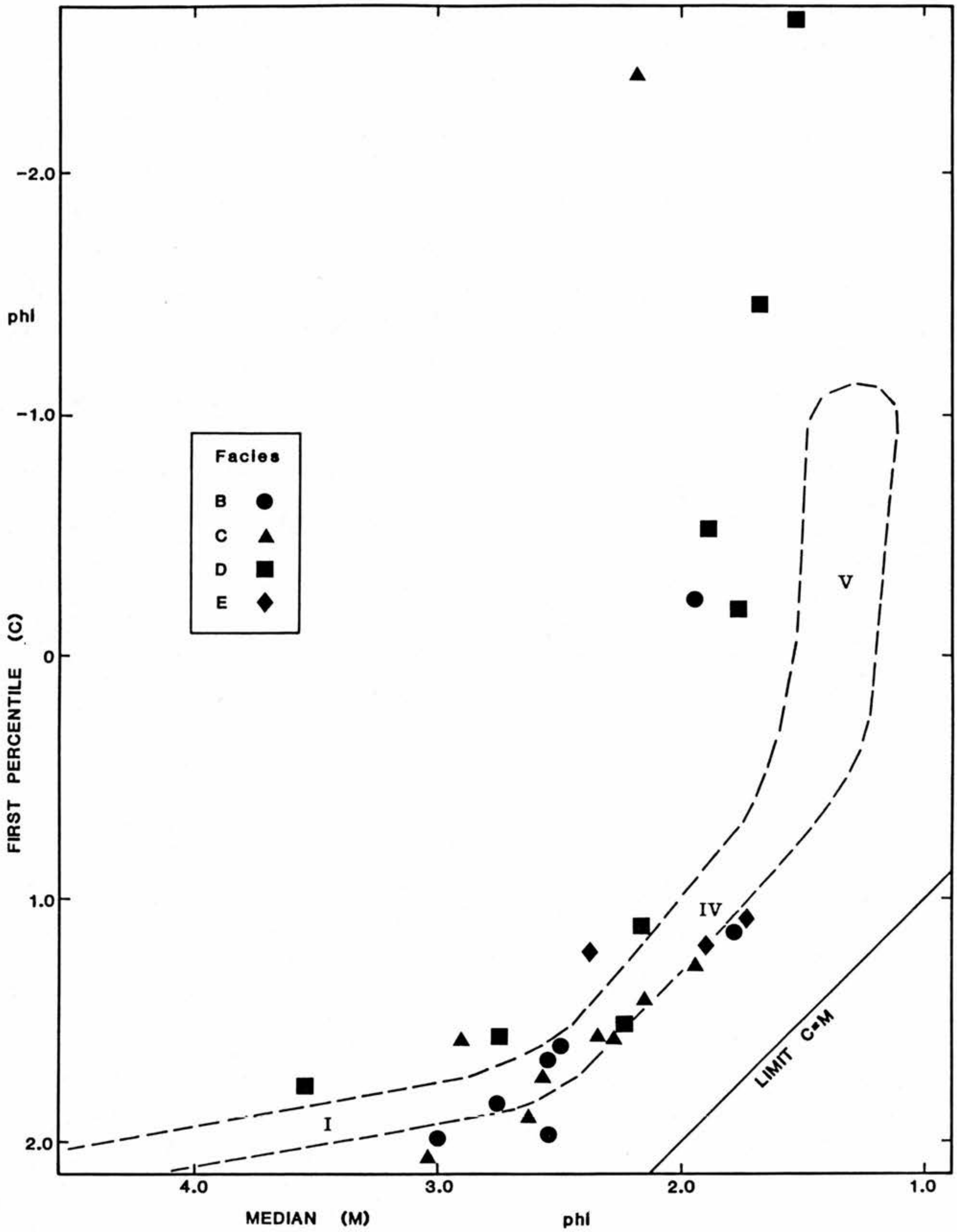



Fig 14. C - M cross-plot: first percentile v median

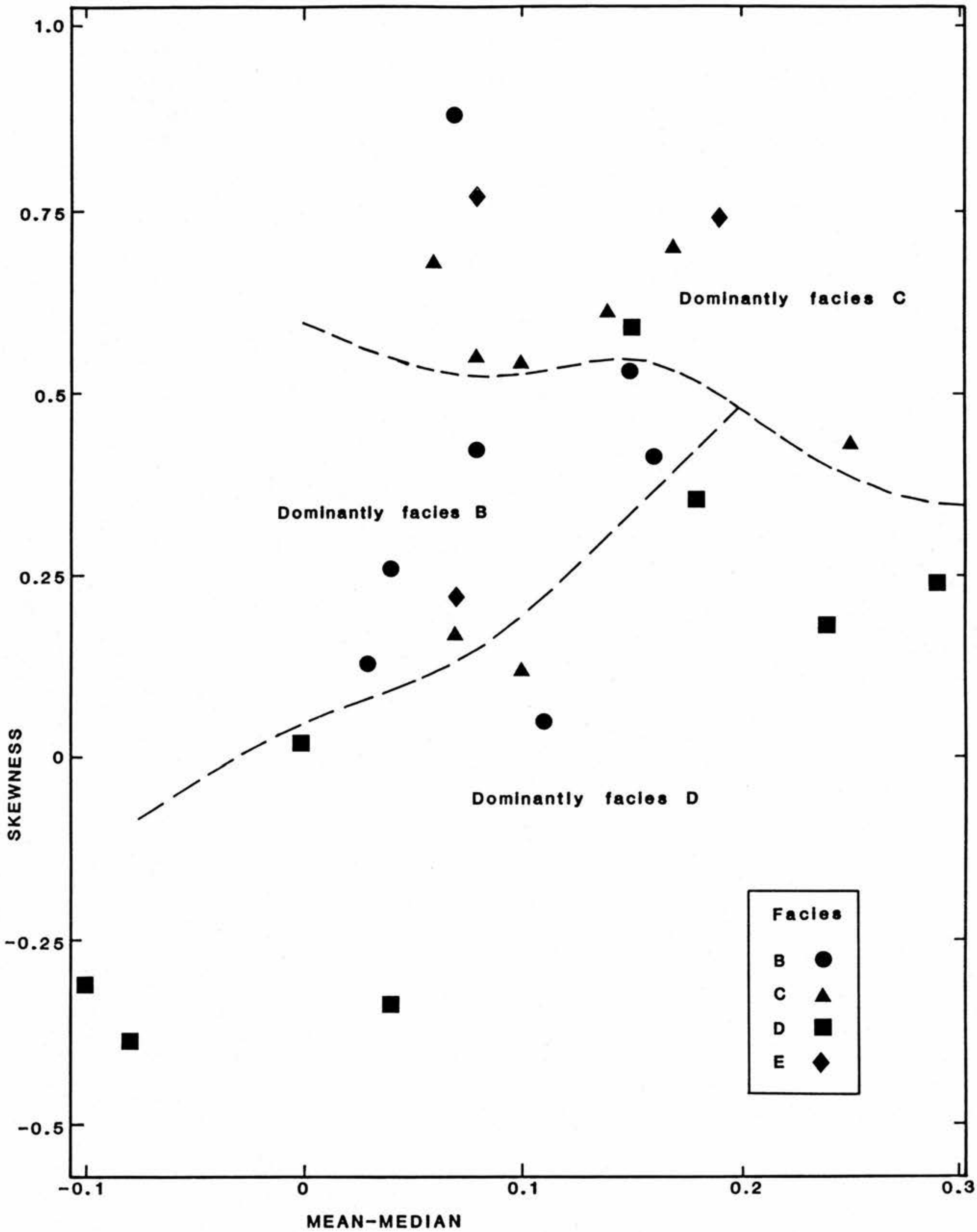



Fig 15. Cross-plot: skewness v mean-median

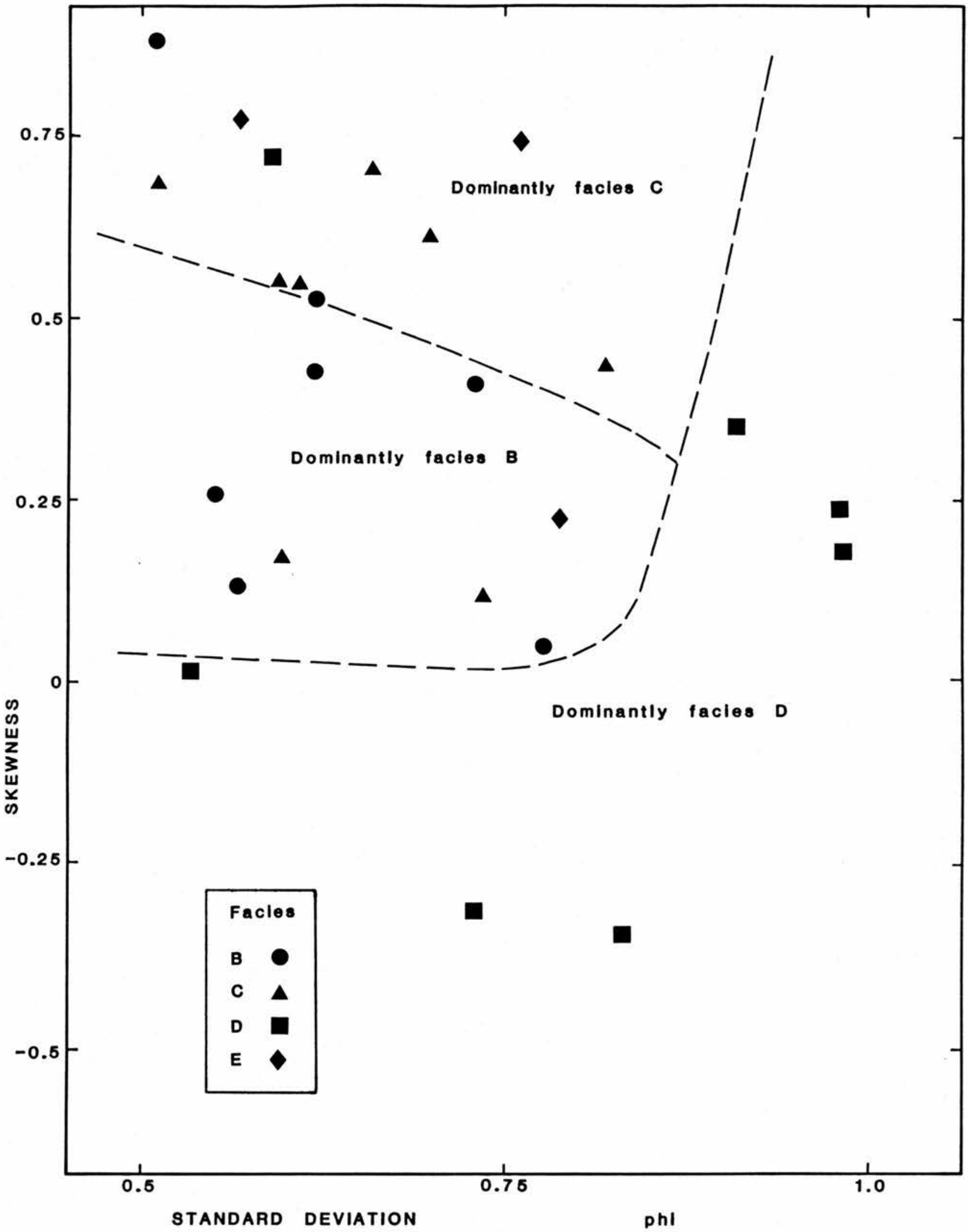



Fig 16. Cross-plot: skewness v standard deviation

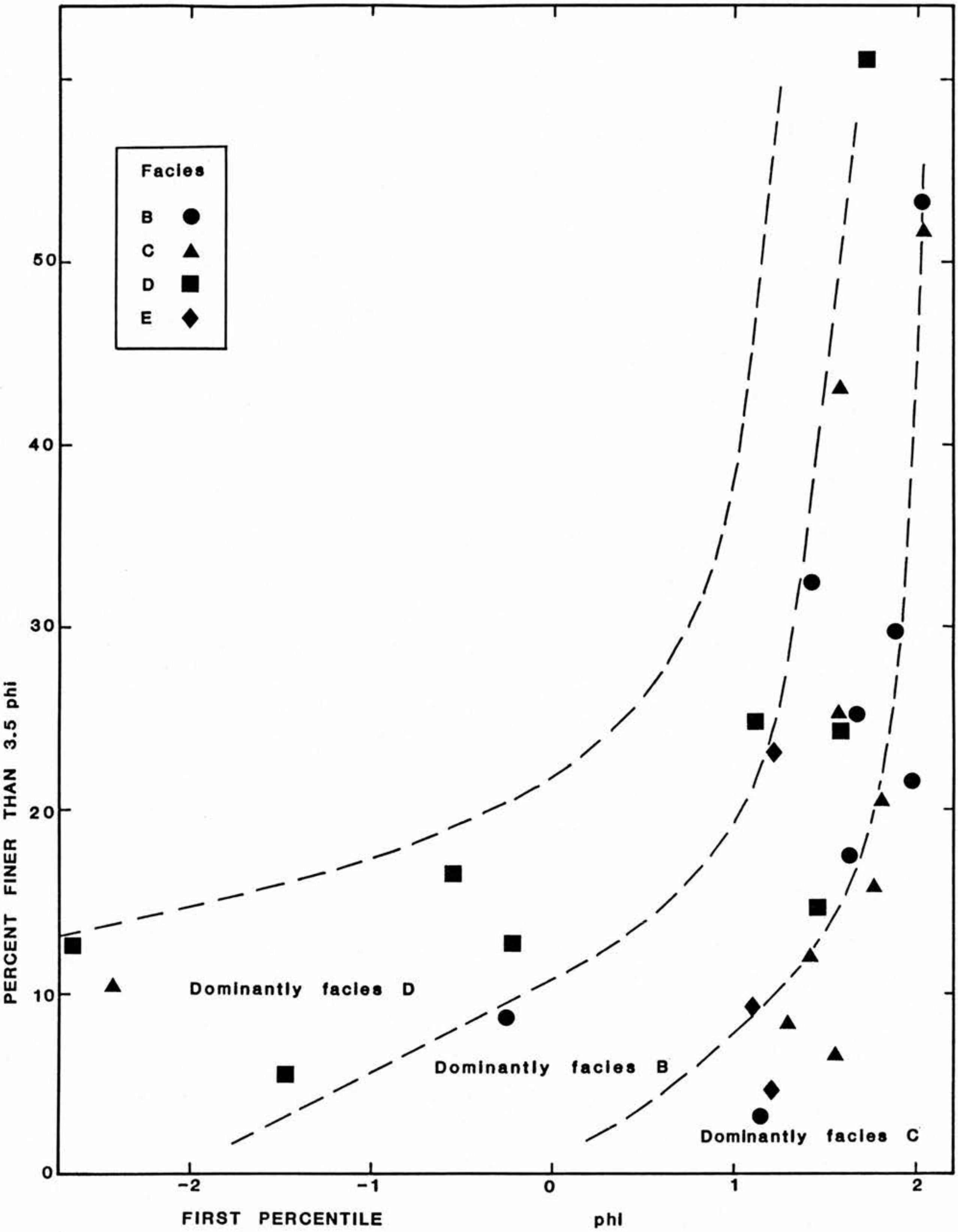



Fig 17. Cross-plot: percent finer than 3.5 phi v first percentile

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Fig 18a. Palaeocurrent directions for all facies - raw data

Fig 18b. Palaeocurrent directions for all facies - contoured

Fig 19a. Palaeocurrent directions facies C - raw data

Fig 19b. Palaeocurrent directions facies C - contoured

Fig 20a. Palaeocurrent directions facies D - raw data

Fig 20b. Palaeocurrent directions facies D - contoured

Key to shading - percentage of readings

>32%

16 - 32%

8 - 16%

4 - 8%

2 - 4%

1 - 2%

<1%

Fig 18a

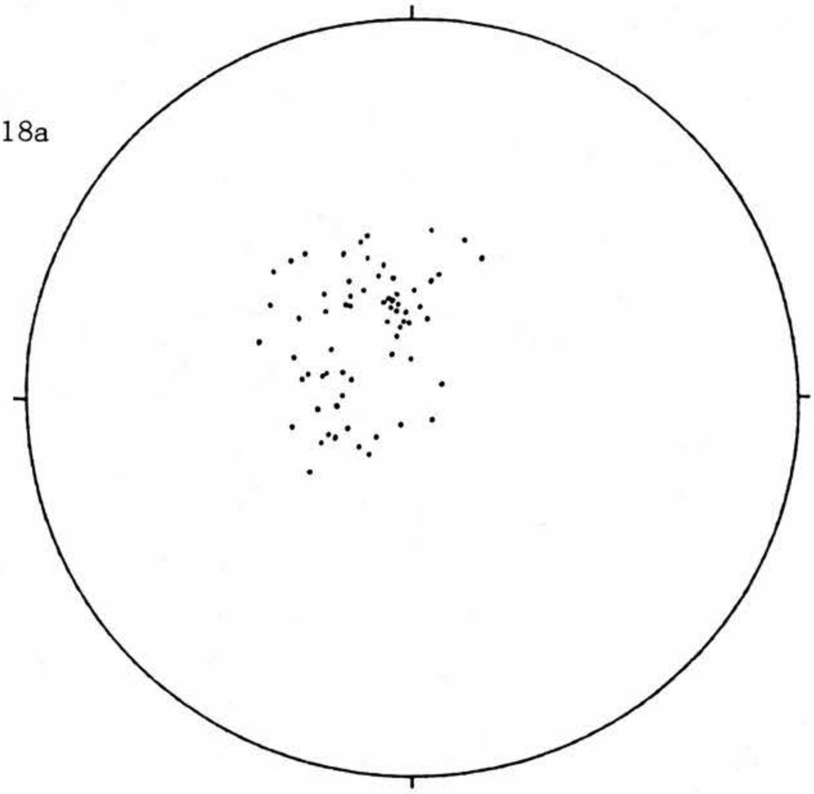


Fig 18b

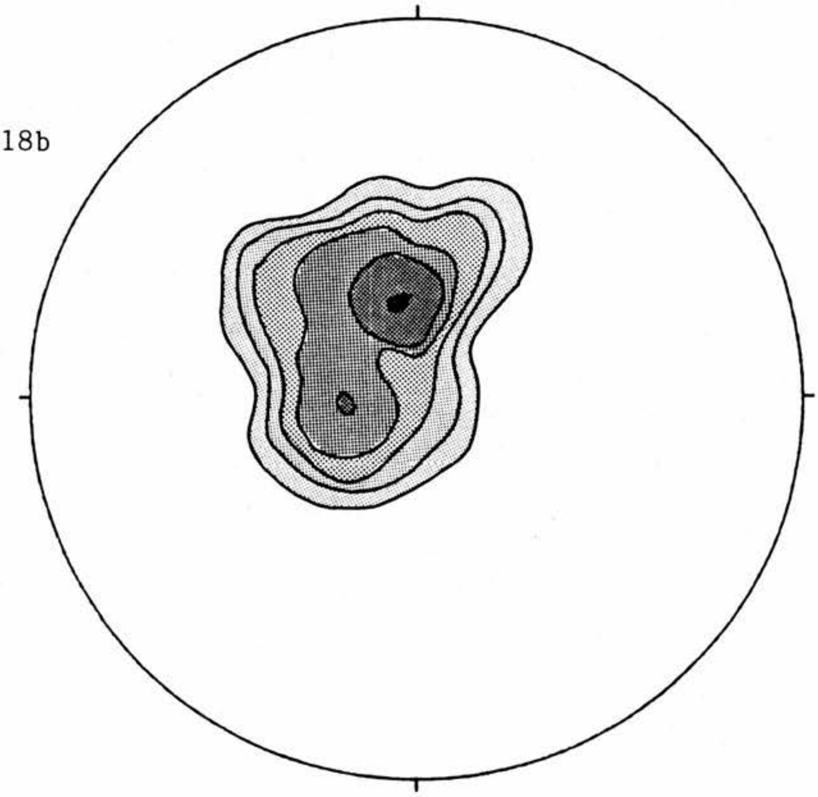


Fig 19a

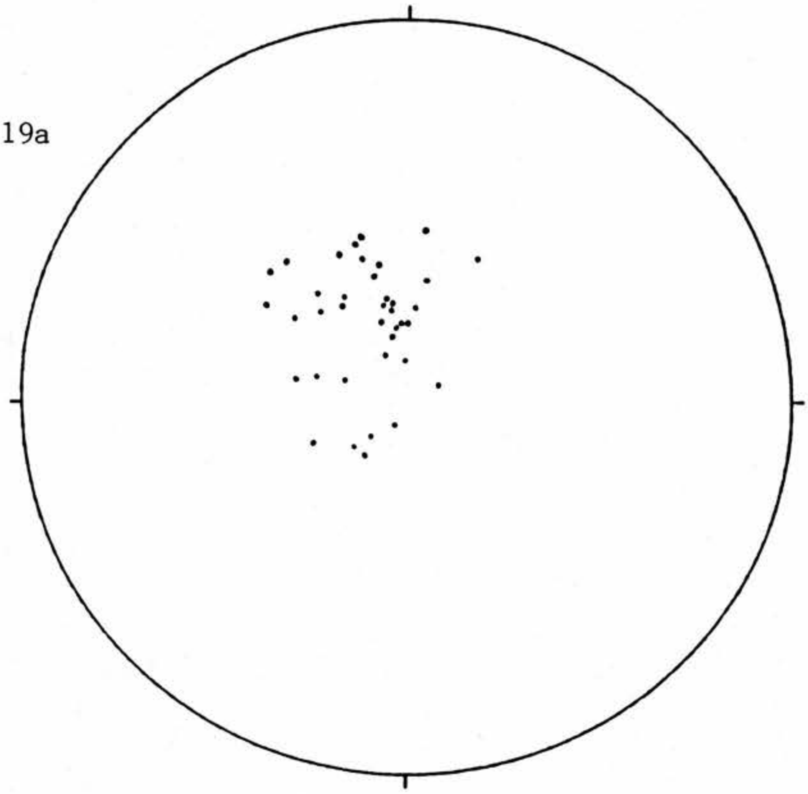


Fig 19b

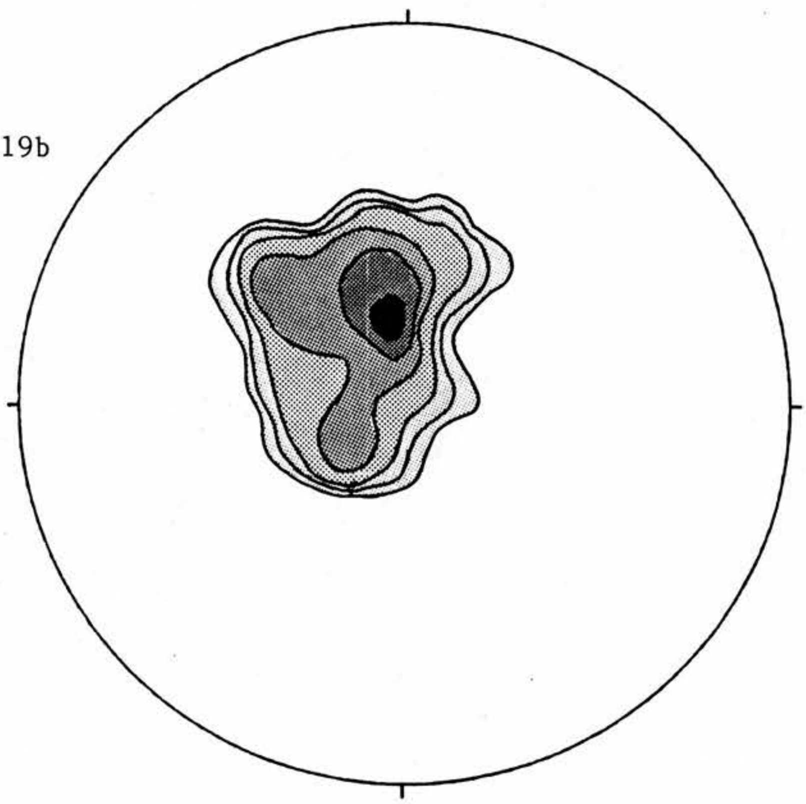


Fig 20a

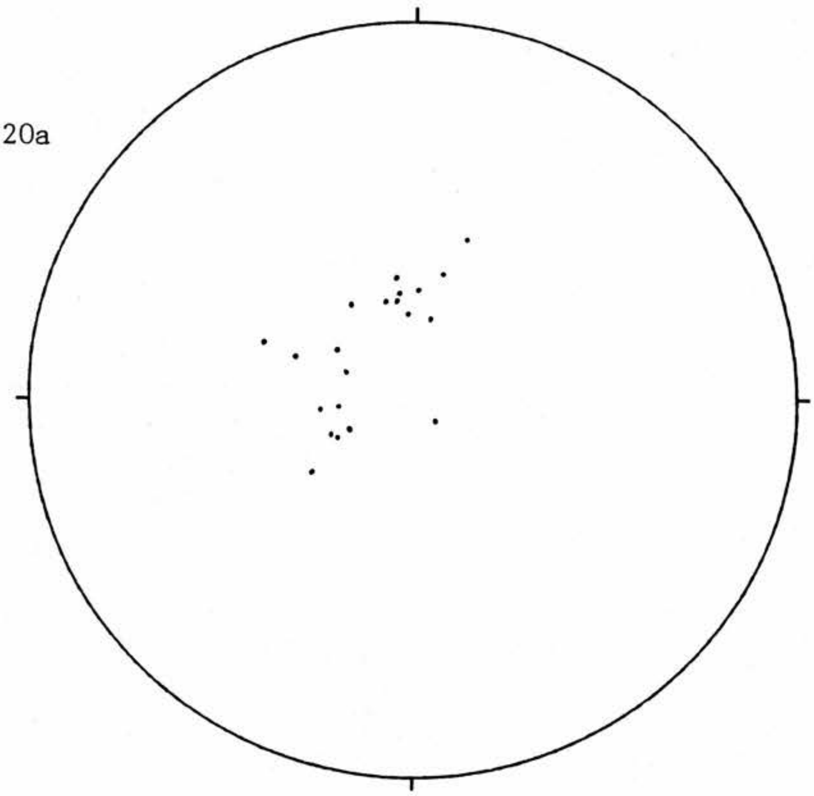
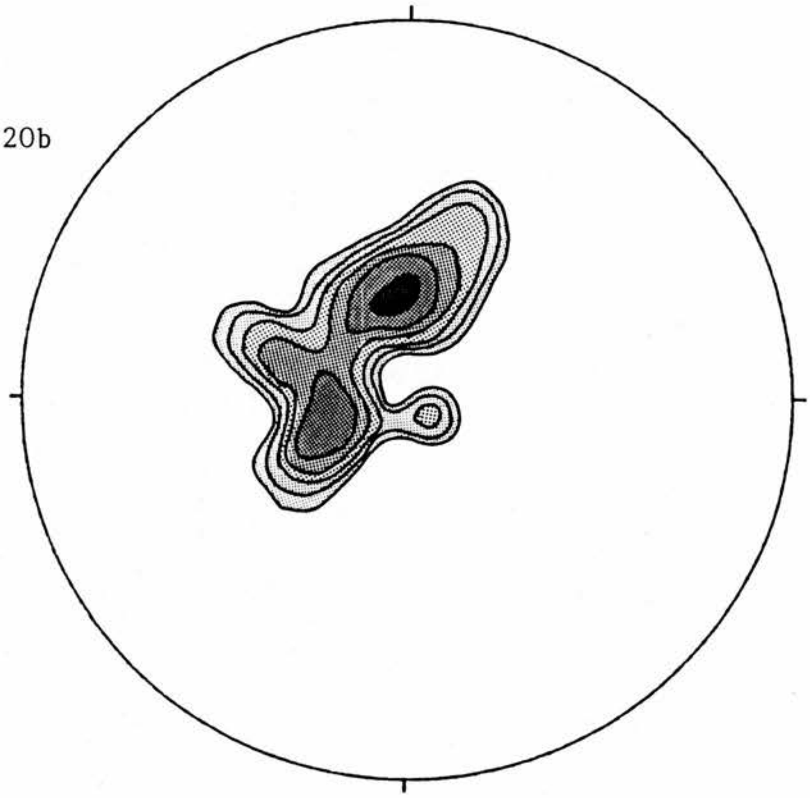
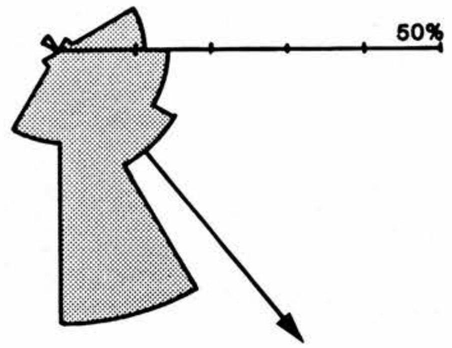


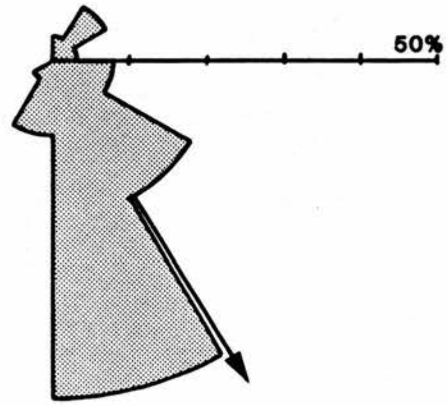
Fig 20b



All facies




Facies C




Facies D

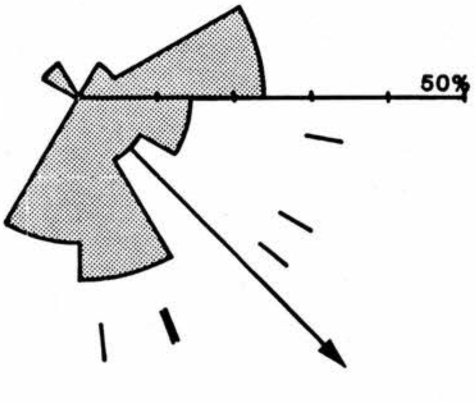



Fig 21. Current rose diagrams for the Pease Bay sandstones

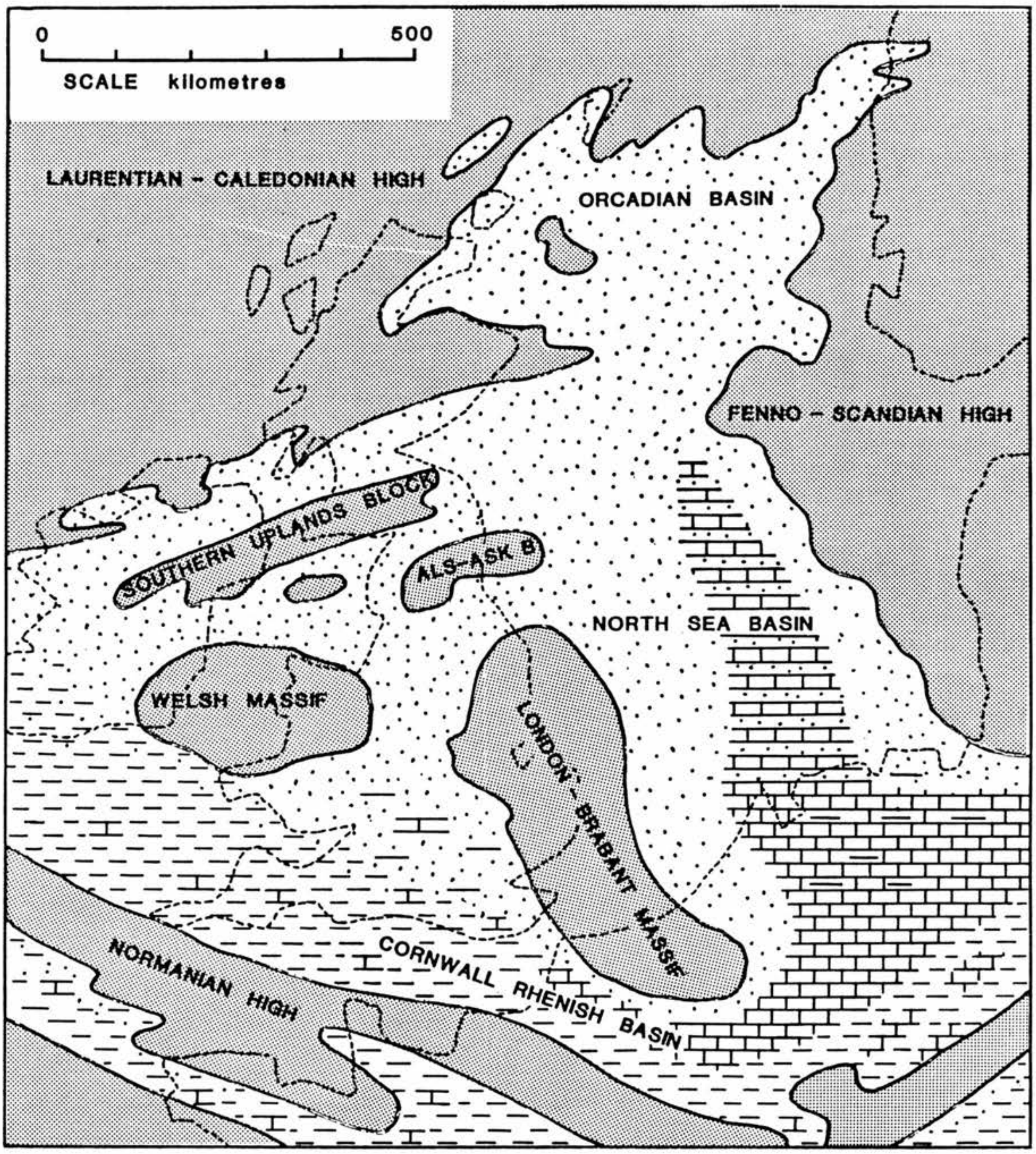



Fig 22. Palaeogeography of western Europe during late Devonian to early Carboniferous (after Ziegler 1981)

# TABLE 1: MARKOV ANALYSIS RESULTS

FREQUENCY OF TRANSITIONS INTO FACIES TYPES

| | |
|---|---|
| A | 15 |
| B | 18 |
| C | 25 |
| D | 27 |
| E | 6 |
| total | 91 |

INDEPENDANT TRIALS MATRIX

| | A | B | C | D | E |
|---|---|---|---|---|---|
| A | 0 | 24 | 33 | 36 | 8 |
| B | 21 | 0 | 34 | 37 | 8 |
| C | 23 | 27 | 0 | 41 | 9 |
| D | 23 | 28 | 39 | 0 | 9 |
| E | 18 | 21 | 29 | 32 | 0 |

TRANSITION COUNT MATRIX

| | A | B | C | D | E |
|---|---|---|---|---|---|
| A | 0 | 1 | 7 | 5 | 2 |
| B | 1 | 0 | 3 | 13 | 1 |
| C | 8 | 6 | 0 | 9 | 2 |
| D | 5 | 8 | 12 | 0 | 1 |
| E | 0 | 3 | 3 | 0 | 0 |

## TRANSITION PROBABILITY MATRIX

| | A | B | C | D | E |
|---|---|---|---|---|---|
| A | 0 | 7 | 47 | 33 | 13 |
| B | 6 | 0 | 17 | 72 | 6 |
| C | 32 | 24 | 0 | 36 | 8 |
| D | 19 | 31 | 46 | 0 | 4 |
| E | 0 | 50 | 50 | 0 | 0 |

## DIFFERENCE MATRIX

| | A | B | C | D | E |
|---|---|---|---|---|---|
| A | 0 | -17 | 14 | -3 | 5 |
| B | -15 | 0 | -17 | 35 | -2 |
| C | 9 | -3 | 0 | -5 | -1 |
| D | -4 | 3 | 7 | 0 | -5 |
| E | -18 | 29 | 21 | -32 | 0 |

DEGREES OF FREEDOM: 15

CHI-SQUARE: 3.28

Table 2: Mineral Species in Facies B and Facies C Sandstones

| Mineral species | Facies B Sandstone - Bed 2 Concretion | Facies B Sandstone - Bed 2 Sandstone | Facies C Sandstone Bed 51 |
|---|---|---|---|
| Quartz | 49 | 61 | 68 |
| Feldspar | 16 | 20 | 14 |
| Carbonate | 35 | - | - |
| Void | - | 19 | 18 |

Table 3: Raw Palaeocurrent Data - Pease Bay to Horse Roads Rock

| Bed no | Facies | Dip (uncorrected) | Azimuth (uncorrected) | Dip (corrected) | Azimuth (corrected) | Regional dip | |
|---|---|---|---|---|---|---|---|
| 3 | C | 20 | 114 | 32 | 140 | | |
| | | 16 | 331 | 7 | 243 | | |
| | | 30 | 105 | 37 | 124 | | |
| 5 | C | 10 | 210 | 26 | 189 | | |
| | | 4 | 230 | 20 | 184 | | |
| 7 | E | 20 | 60 | 19 | 104 | | |
| 10 | D | axial | 175 | axial | 175 | | |
| | | axial | 130 | axial | 129 | | |
| | | axial | 100 | axial | 99 | | |
| 12 | D | axial | 120 | axial | 119 | 17 | 355 |
| 14 | C | 8 | 354 | 9 | 176 | | |
| | | 28 | 123 | 40 | 139 | | |
| | | 21 | 228 | 34 | 206 | | |
| 23 | C | 0 | - | 17 | 175 | | |
| 24 | C | 25 | 15 | 11 | 44 | | |
| | | 16 | 135 | 30 | 155 | | |
| | | 4 | 150 | 21 | 171 | | |
| 25 | C | 10 | 130 | 25 | 158 | | |
| | | 15 | 130 | 29 | 153 | | |
| | | 20 | 120 | 39 | 145 | | |
| 29 | D | 2 | 41 | 22 | 165 | | |
| 34 | C | 19 | 71 | 27 | 136 | | |
| 36 | D | 25 | 60 | 27 | 110 | | |
| | | 30 | 35 | 20 | 84 | | |
| 39 | (B) | 35 | 40 | 26 | 76 | | |
| 40 | D | 20 | 35 | 16 | 112 | 23 | 350 |
| | | 30 | 75 | 35 | 111 | | |
| 42 | C | 30 | 95 | 41 | 133 | | |
| | | 20 | 30 | 14 | 108 | | |
| 48 | D | 30 | 20 | 15 | 67 | | |
| 50 | D | 9 | 305 | 18 | 190 | | |
| 51 | C | 2 | 20 | 21 | 167 | | |
| 62 | C | 5 | 160 | 28 | 165 | | |
| 65 | D | 20 | 225 | 37 | 198 | | |
| | | 26 | 30 | 16 | 86 | | |
| 67 | E | 25 | 30 | 15 | 91 | | |
| | | 25 | 50 | 22 | 102 | | |
| 68 | C | 5 | 150 | 36 | 162 | | |
| | | 20 | 80 | 30 | 126 | | |
| 70 | C | 15 | 210 | 37 | 186 | | |
| | | 27 | 360 | 6 | 27 | | |
| 71 | C | 8 | 360 | 14 | 166 | | |
| 73 | C | 5 | 330 | 17 | 176 | | |
| | | 33 | 15 | 15 | 48 | 22 | 350 |
| 75 | C | 35 | 12 | 15 | 38 | | |
| 76 | D | 17 | 50 | 20 | 122 | | |
| 77 | C | 10 | 145 | 32 | 162 | | |
| 79 | D | 5 | 180 | 27 | 172 | | |
| 80 | C | 15 | 135 | 35 | 155 | | |
| 81 | D | 32 | 25 | 18 | 64 | | |
| 82 | C | 35 | 30 | 22 | 65 | | |
| 88 | C | 6 | 350 | 16 | 170 | | |
| 90 | D | 0 | - | 22 | 170 | | |
| 93 | C | 5 | 20 | 18 | 162 | | |

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| 94 | C | 10 | 96 | 25 | 146 | | |
| 99 | C | 10 | 160 | 30 | 168 | | |
| 100 | D | 25 | 340 | 6 | 313 | | |
| | | 10 | 100 | 25 | 146 | | |
| 101 | C | 25 | 55 | 24 | 100 | | |
| | | 2 | 140 | 22 | 168 | | |
| 107 | D | 2 | 280 | 19 | 176 | | |
| | | axial | 160 | axial | 159 | | |
| | | axial | 160 | axial | 159 | | |
| 112 | C | 0 | - | 20 | 170 | 20 | 350 |
| 116 | C | 20 | 105 | 34 | 136 | | |
| 119 | C | 10 | 105 | 26 | 149 | | |
| 120 | C | 17 | 60 | 37 | 164 | | |
| 122 | D | 5 | 230 | 24 | 182 | | |
| 124 | D | 40 | 30 | 27 | 54 | | |
| | | 30 | 30 | 19 | 67 | | |
| 130 | D | 3 | 180 | 23 | 172 | | |
| | | 12 | 230 | 28 | 192 | | |

Table 4

Palaeocurrent measurements on trough cross bedding

| Azimuth range | Facies C | | Facies D | | All beds | |
|---|---|---|---|---|---|---|
| | n | % | n | % | n | % |
| 1 - 30 | 1 | 3 | 0 | 0 | 1 | 2 |
| 30 - 60 | 3 | 8 | 1 | 5 | 4 | 6 |
| 60 - 90 | 1 | 3 | 5 | 24 | 7 | 11 |
| 90 - 120 | 3 | 8 | 3 | 14 | 9 | 14 |
| 120 - 150 | 8 | 21 | 2 | 9 | 11 | 17 |
| 150 - 180 | 17 | 45 | 5 | 24 | 24 | 36 |
| 180 - 210 | 4 | 10 | 4 | 19 | 8 | 12 |
| 210 - 240 | 1 | 3 | 0 | 0 | 1 | 2 |
| 240 - 270 | 0 | 0 | 0 | 0 | 0 | 0 |
| 270 - 300 | 0 | 0 | 0 | 0 | 0 | 0 |
| 300 - 330 | 0 | 0 | 1 | 5 | 1 | 2 |
| 330 - 360 | 0 | 0 | 0 | 0 | 0 | 0 |
| Totals | 38 | 101 | 21 | 100 | 66 | 102 |
| Vector mean | 149 | | 131 | | 141 | |
| Magnitude % | 74 | | 59 | | 69 | |

Plate 1. Facies A beds - general aspect showing nature of upper and lower surfaces and red and drab colouration (hammer = 30cm)

Plate 2. Facies A beds - red/drab colouration cross-cutting primary sedimentary structures (hammer head = 12cm)

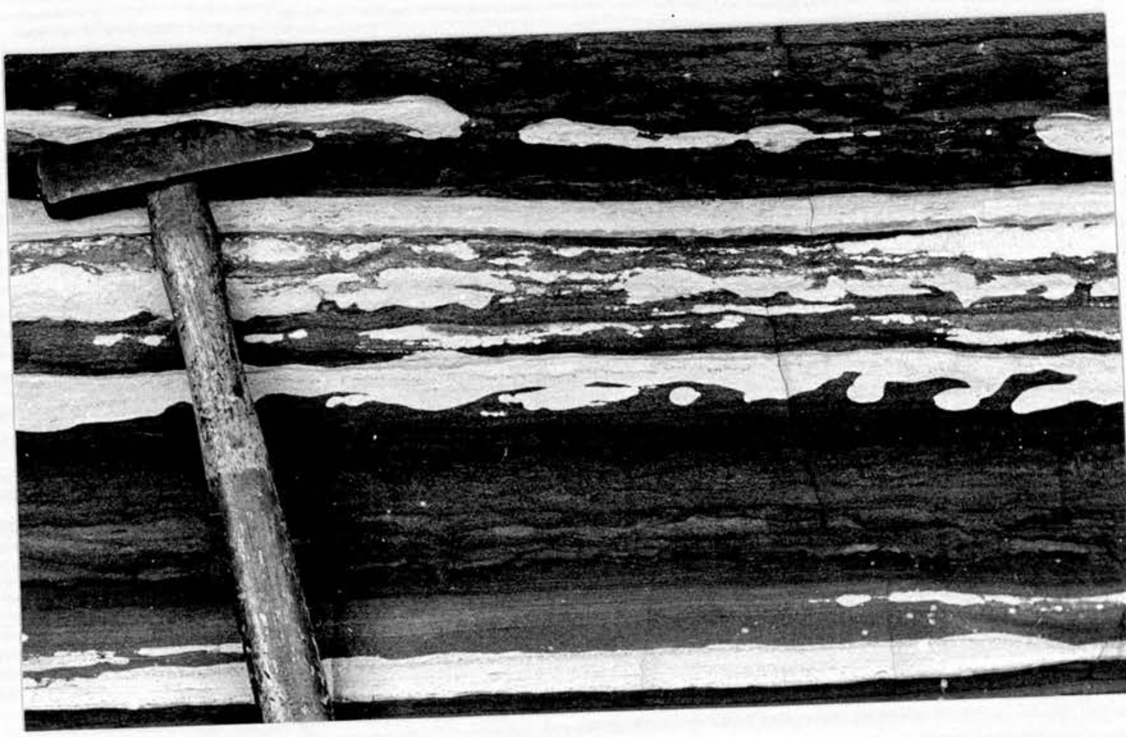


Plate 3. Facies A beds - insipient dessication structures in mudstone (hammer = 30cm)

Plate 4. Facies B beds - general aspect demonstrating parallel lamination (scale = 10cm)

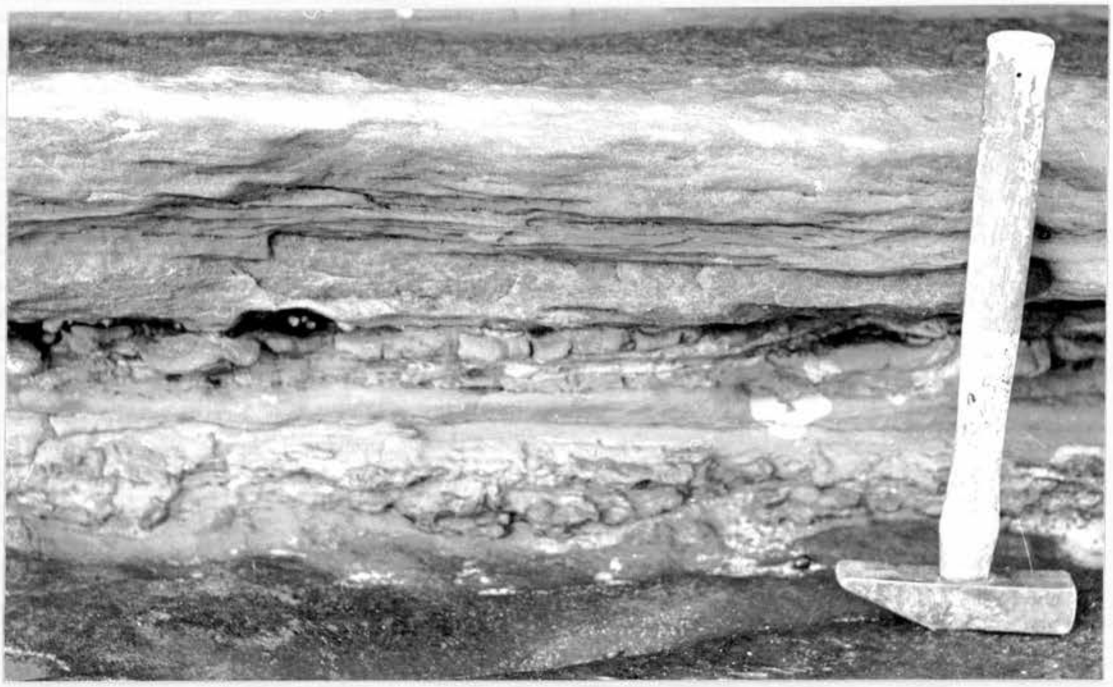
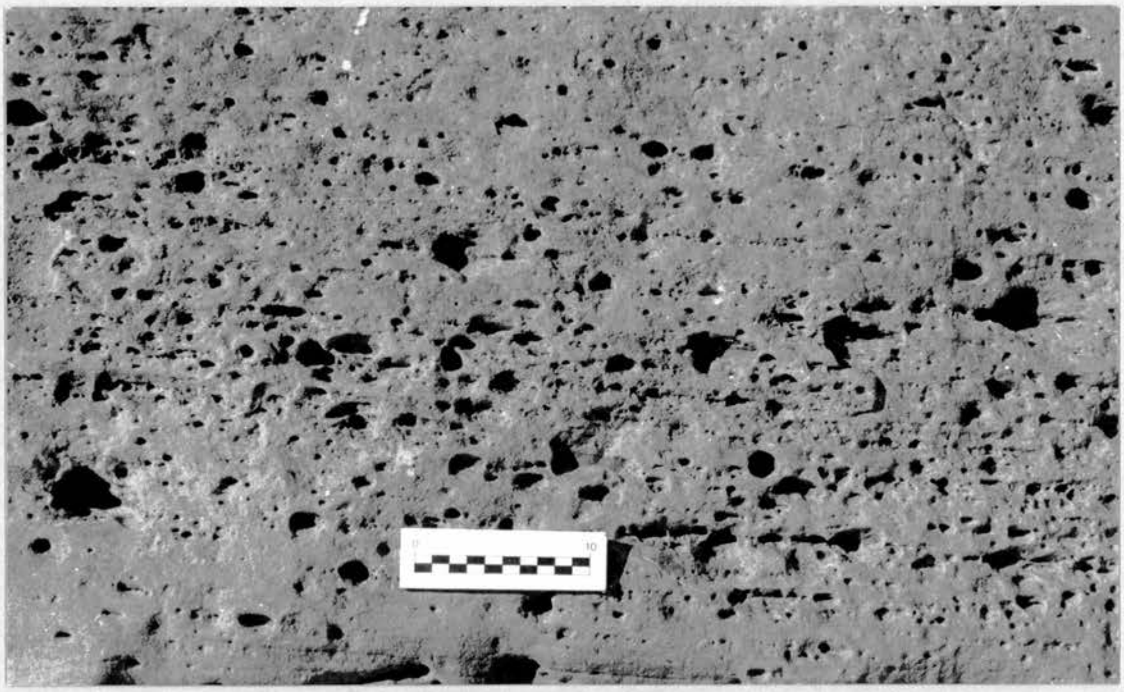


Plate 5. Facies B beds - mudstone clasts near base of bed, weathering out of sandstone (scale = 10cm)

Plate 6. Facies B beds - reactivation of the upper part of the bed (scale = 10cm)



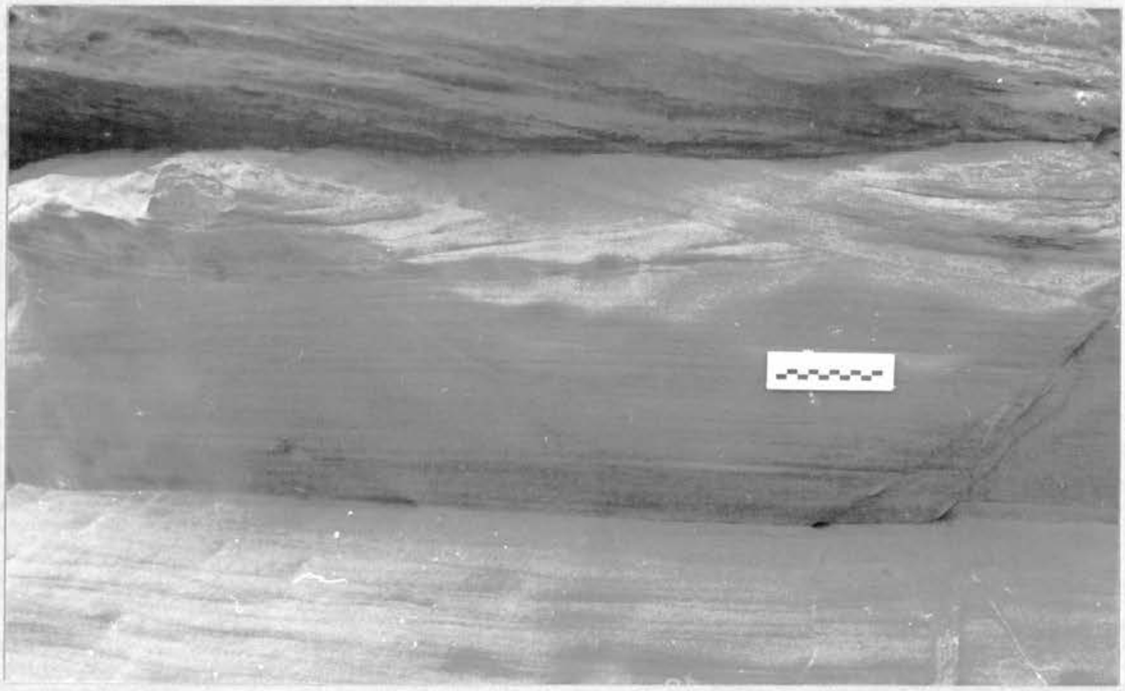


Plate 7. Facies C beds - general aspect showing troughs and asymptotic lamination (scale = 10cm)

Plate 8. Facies C beds - cornstone clasts as lag in scour base (scale = 10cm)

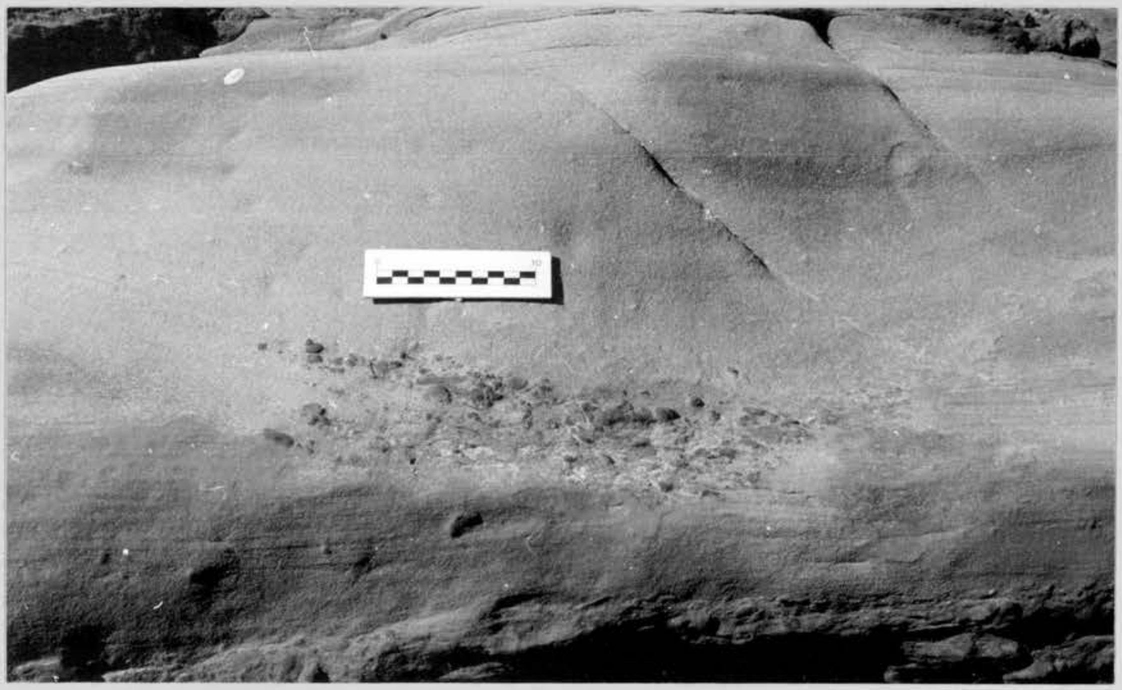
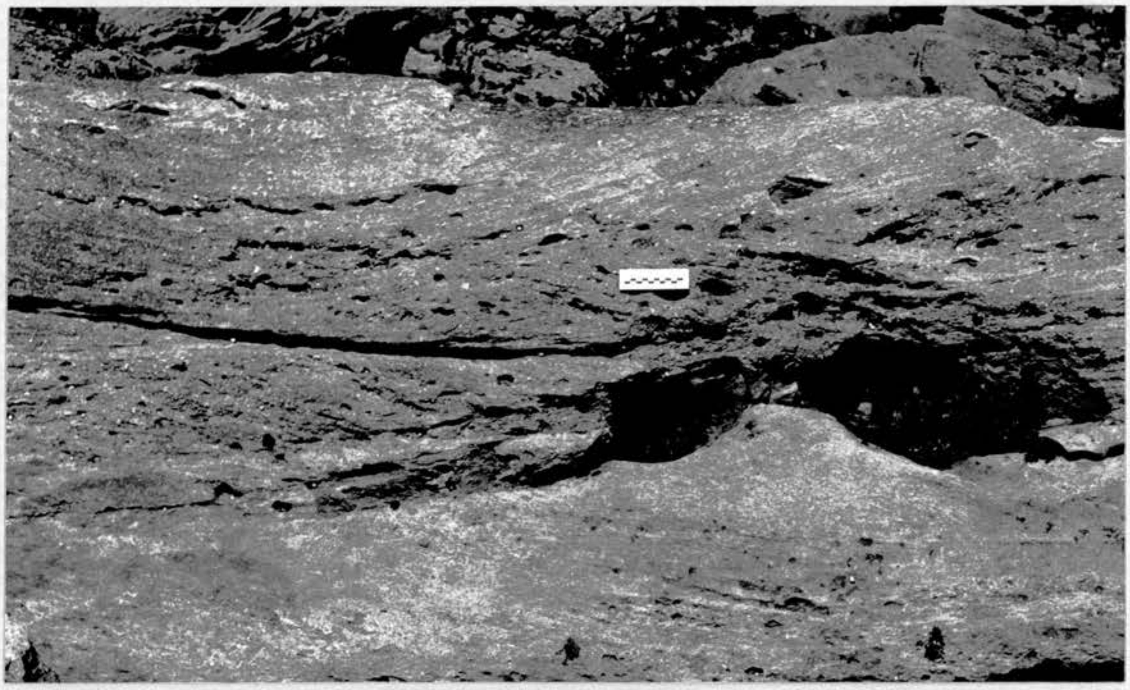


Plate 9. Facies D beds - scoop-like nature of the troughs as seen from above (hammer = 30cm)

Plate 10. Facies D beds - laminations parallel to the trough margins (hammer = 30cm)

Plate 11. Facies D beds - cornstone clasts as concretions lying parallel to the lamination (scale = 10cm)

Plate 12. Facies E beds - general aspect showing planar nature of base of facies B bed above (scale = 10cm)



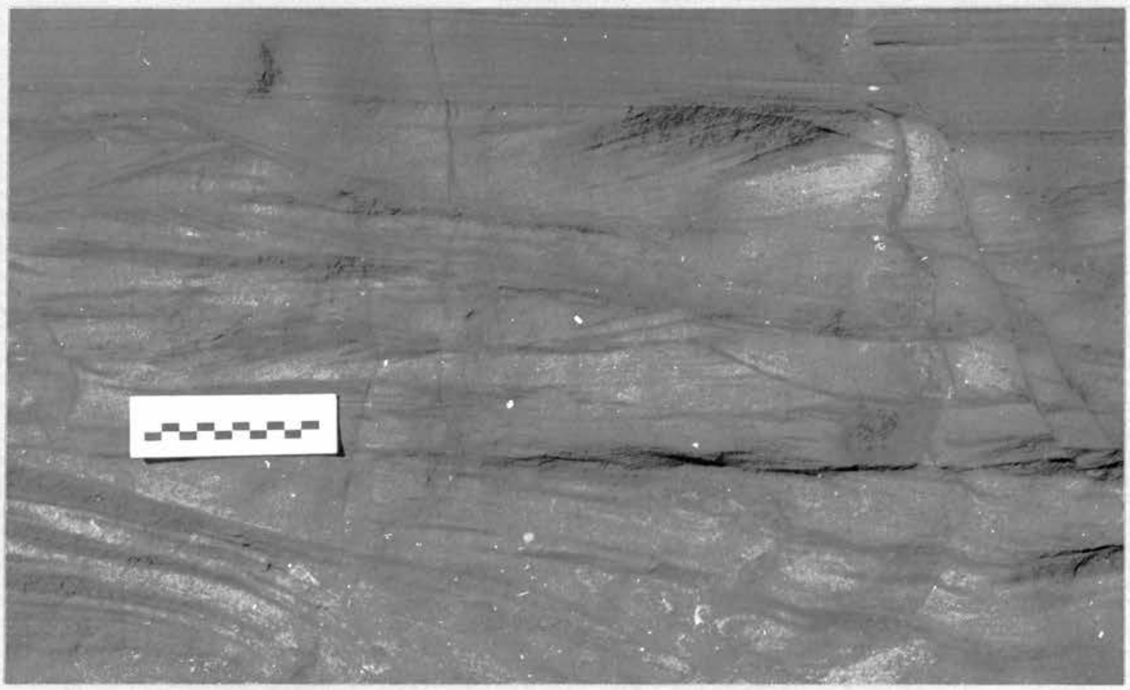


Plate 13. Facies E beds - tabular sets (lens cap = 6cm)

Plate 14. Facies E beds - festoon cross-bedding (pen = 14cm)

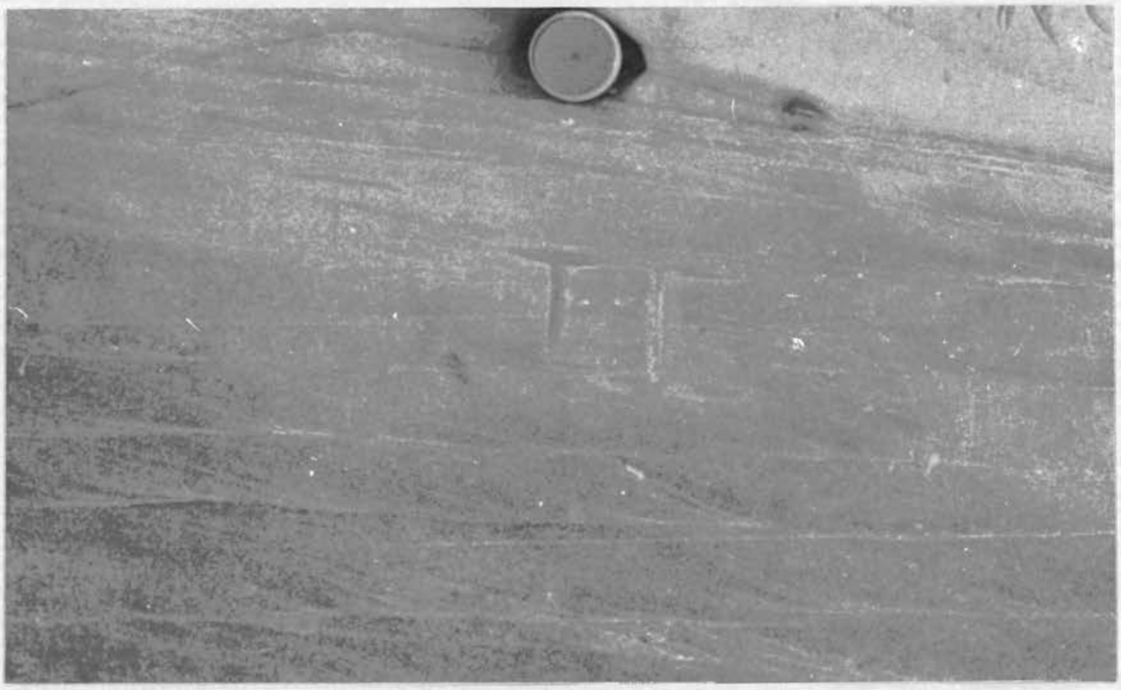
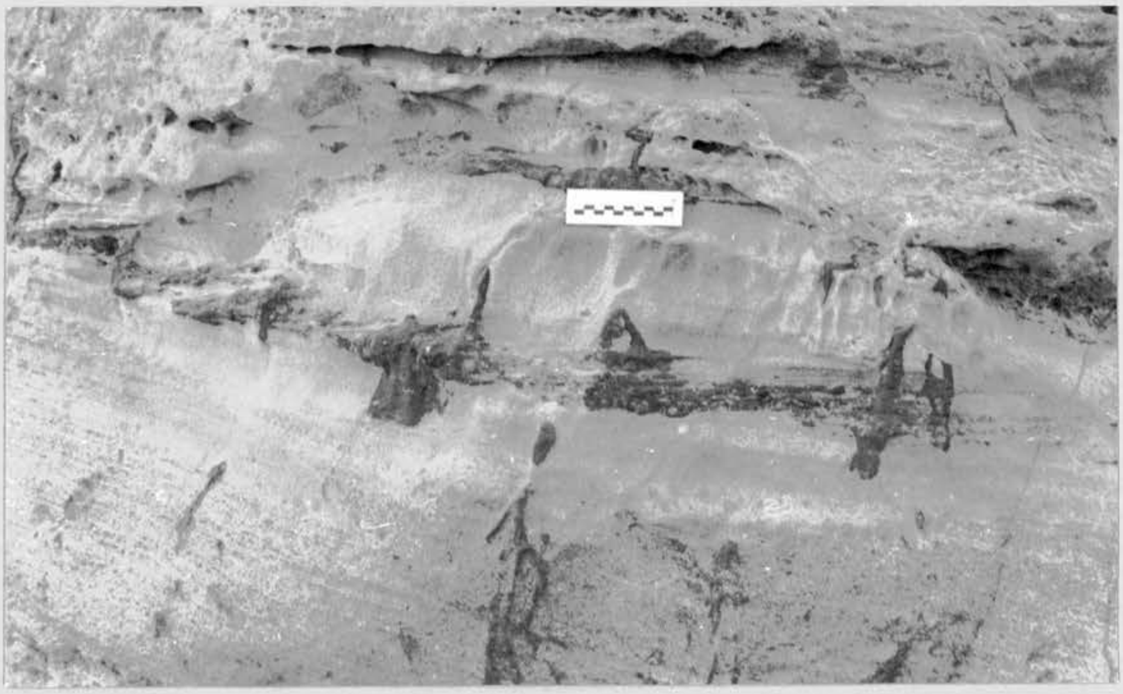


Plate 15. Facies F beds - traces of bedding seen through concretionary growth (scale = 10cm)

Plate 16. Carbonate growth - small concretions or 'glaebules' (scale = 10cm)



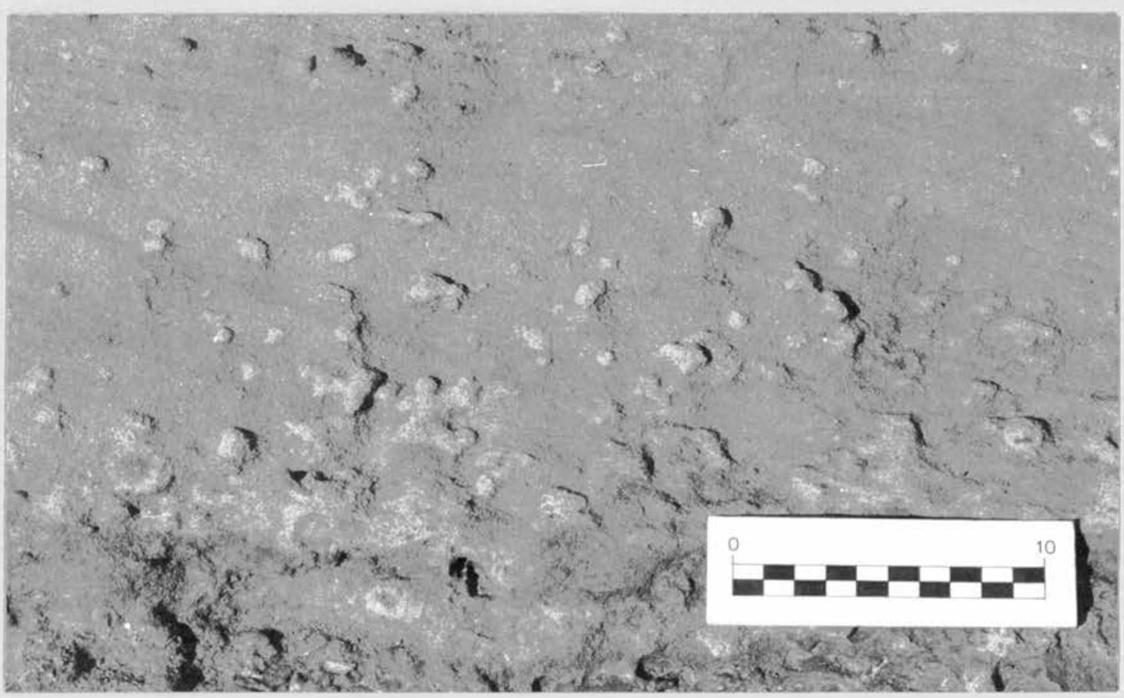
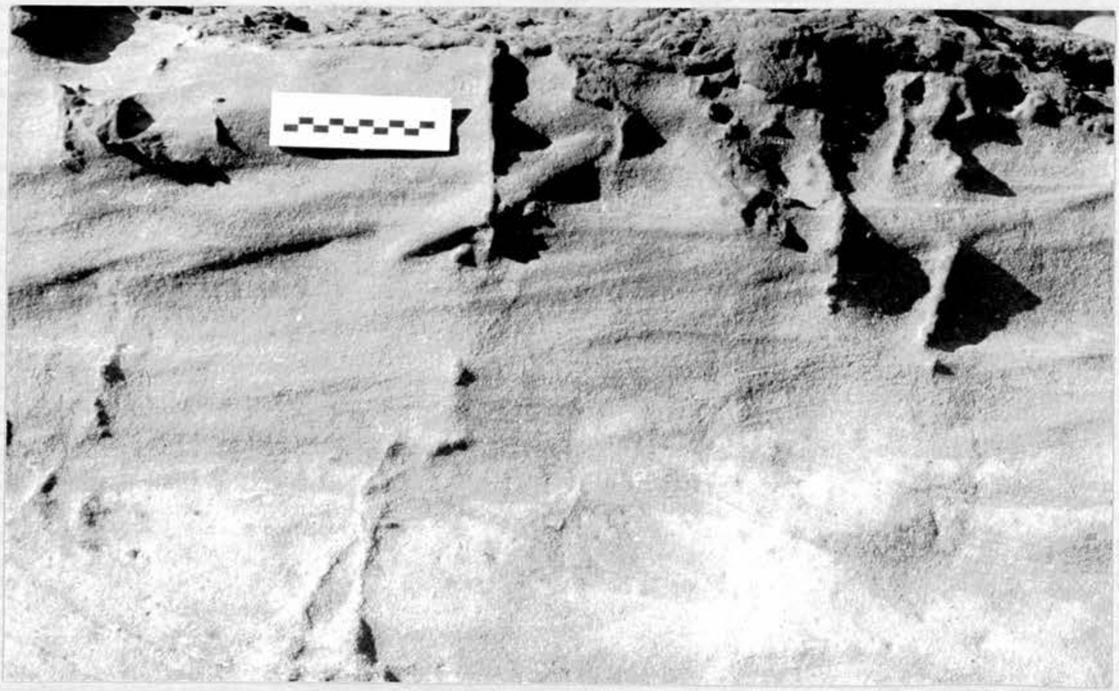

0
10

Plate 17. Carbonate growth - sub-vertical pipe-like concretionary growth, posibly associated with roots or dessication (scale = 10cm)

Plate 18. Carbonate growth - clasts acting as loci for concretionary development (scale = 10cm)



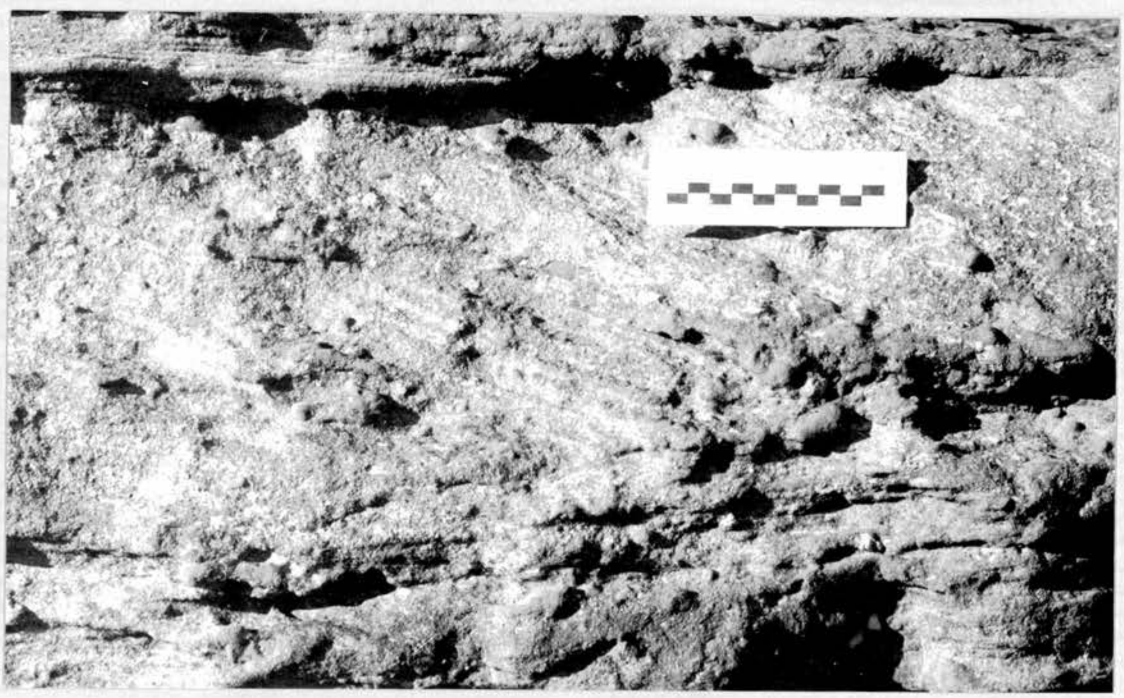


Plate 19. Carbonate growth - continuous cornstone horizon at the top of a sandstone bed (scale = 10cm)

Plate 20. Reddening and carbonate growth - the result of a geochemical profile in a facies B bed overlain by facies C trough cross-bedded sandstone (scale = 10cm)

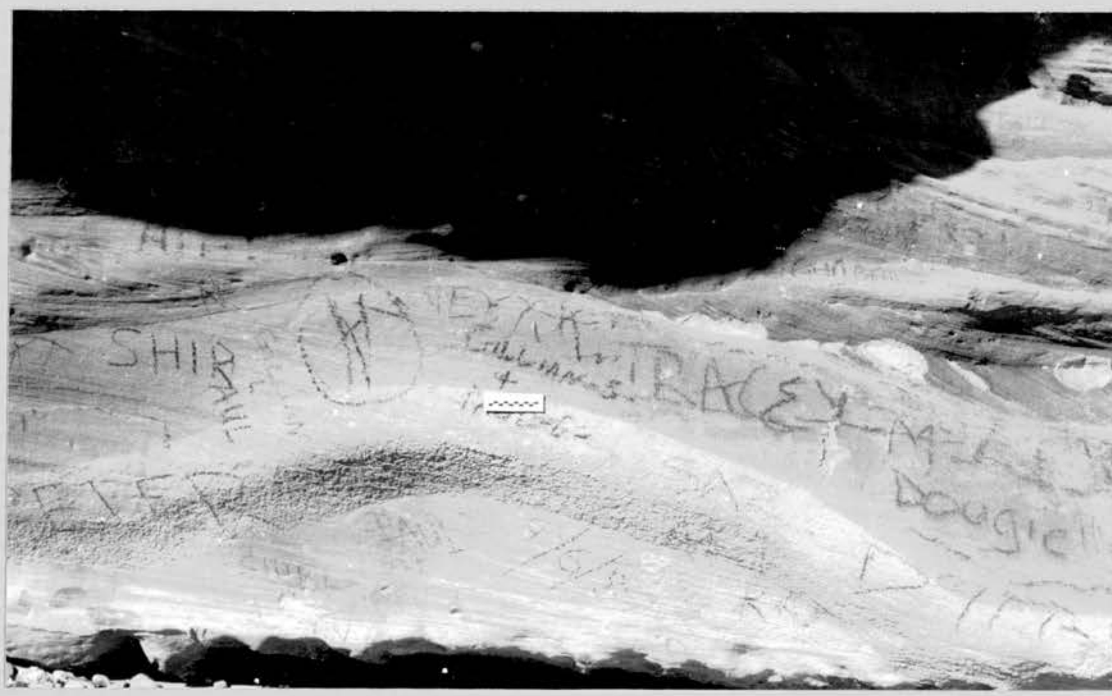
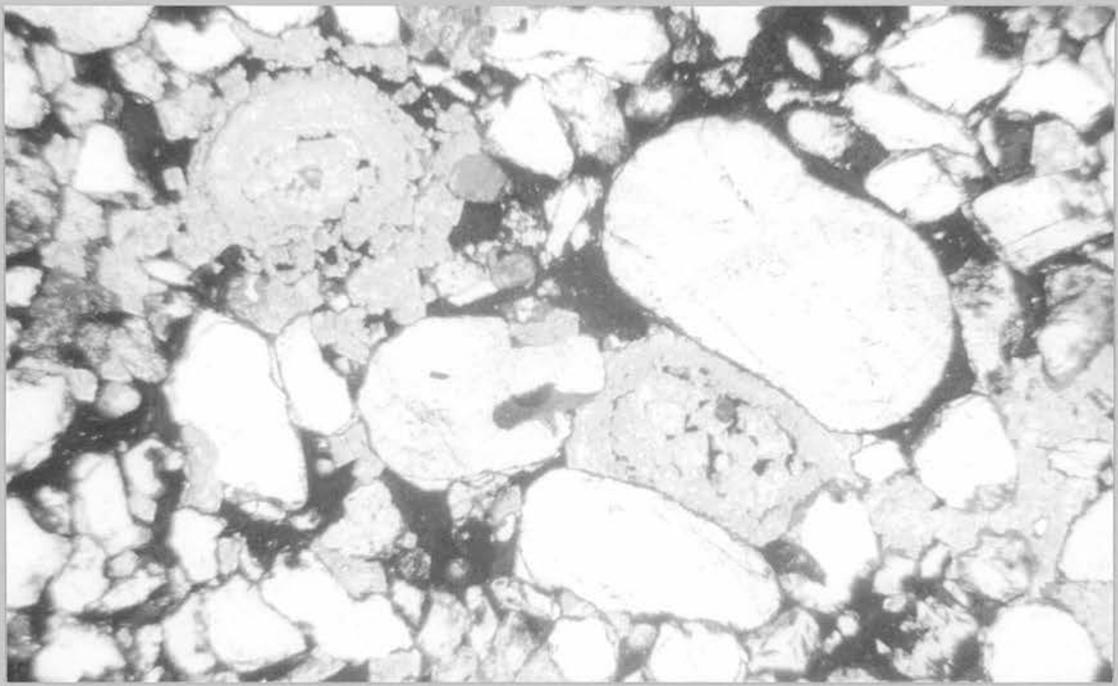

SHIR
TRACEY

Plate 21. Carbonate 'ooids' in Facies D sandstone (plane polarised light)
(scale bar = 1mm)

Plate 22. Carbonate 'ooids' in Facies D sandstone (cathodoluminescence)
(scale bar = 1mm)



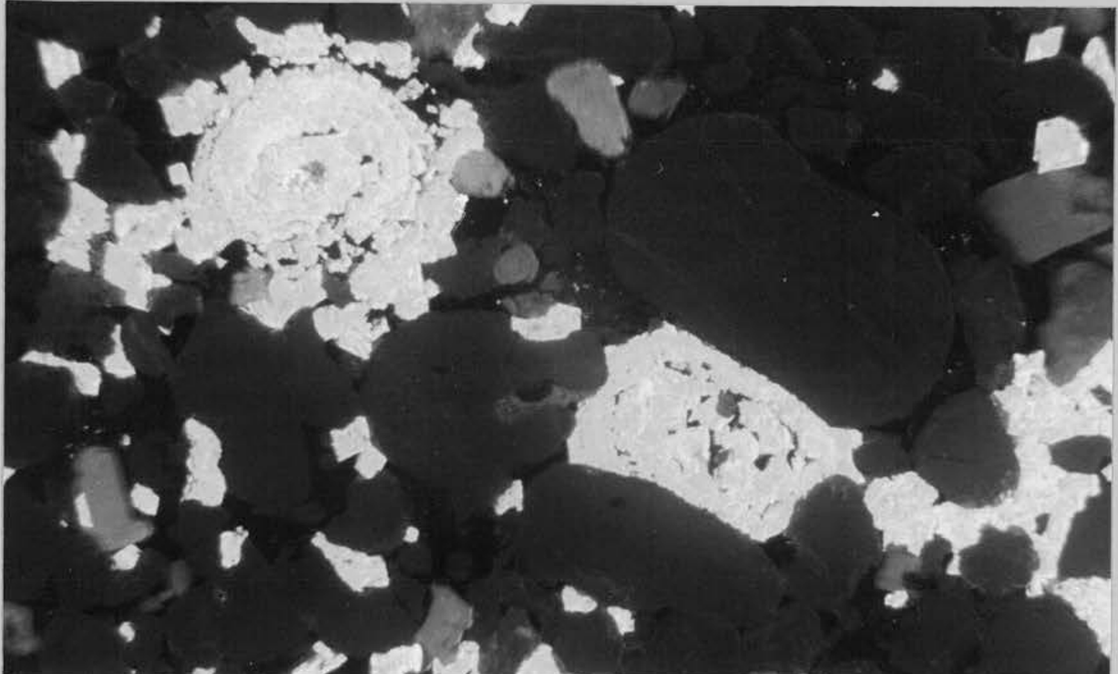
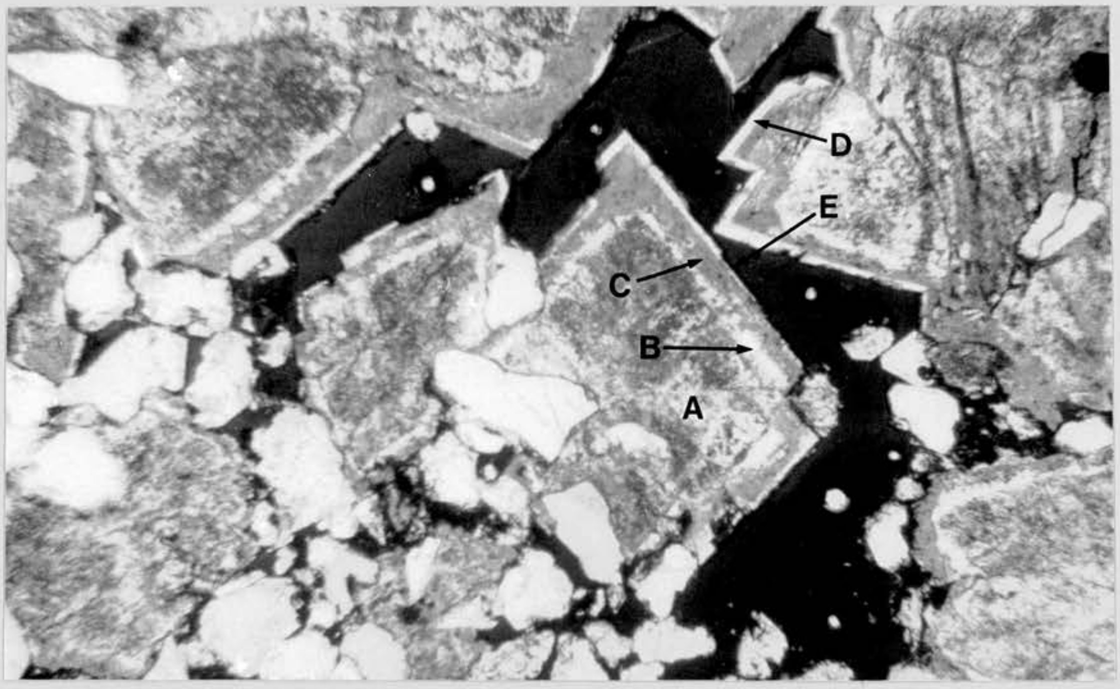


Plate 23. Carbonate cement infill stratigraphy (plane polarised light)
(scale bar = 1mm)

Plate 24. Carbonate cement infill stratigraphy (cathodoluminescence)
(scale bar = 1mm)


D
E
C
B
A

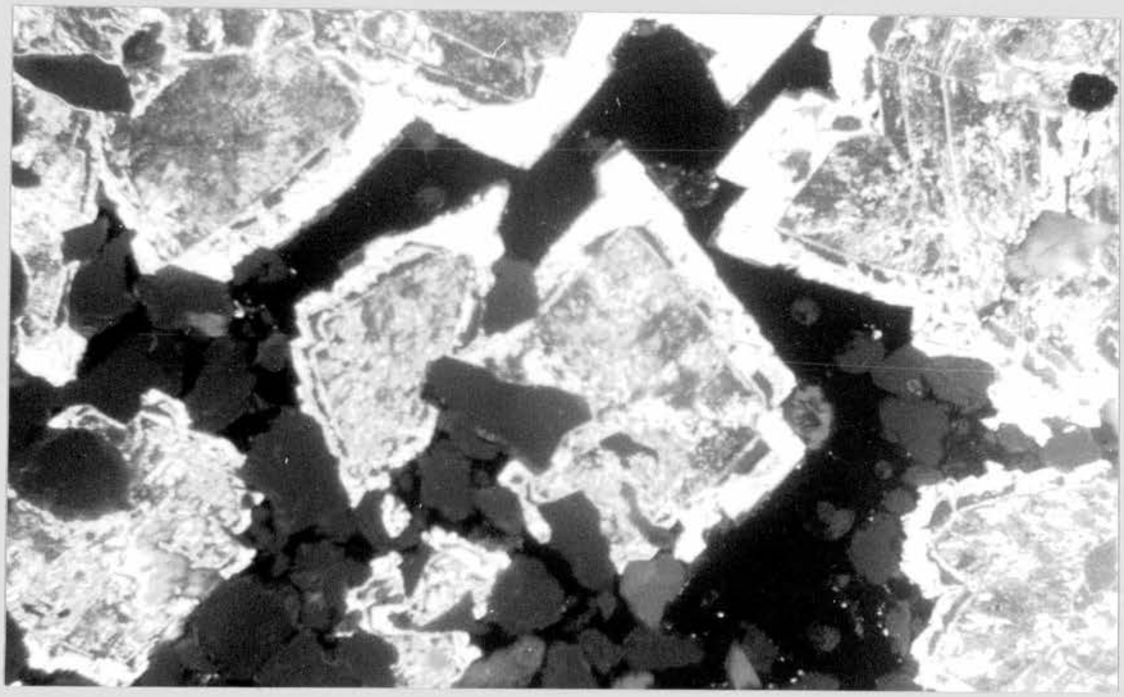
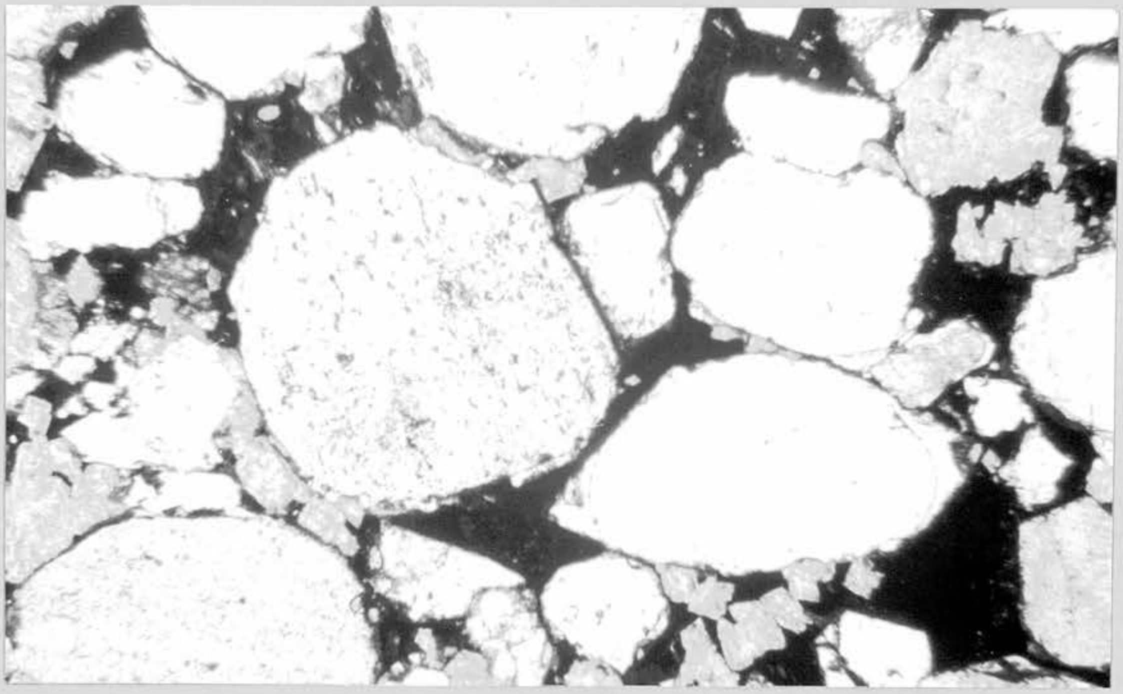


Plate 25. Quartz grains of different provenance in a facies D sandstone (plane polarised light - scale bar = 1mm)

Plate 26. Quartz grains of different provenance in a facies D sandstone (cathodoluminescence - scale bar = 1mm)



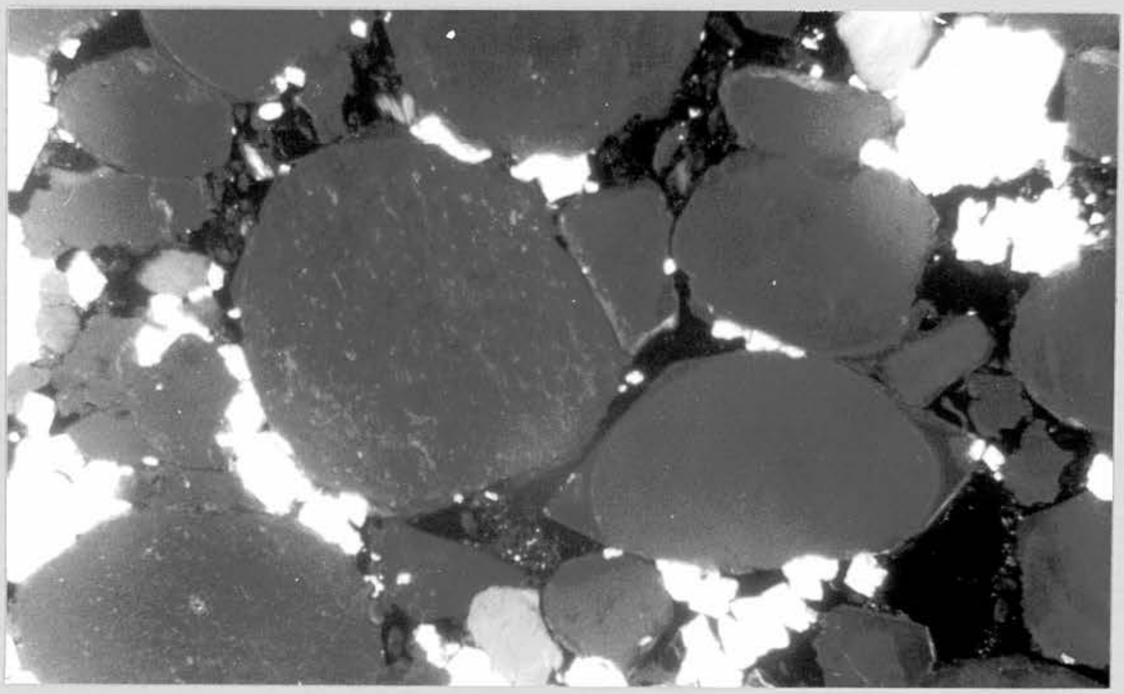
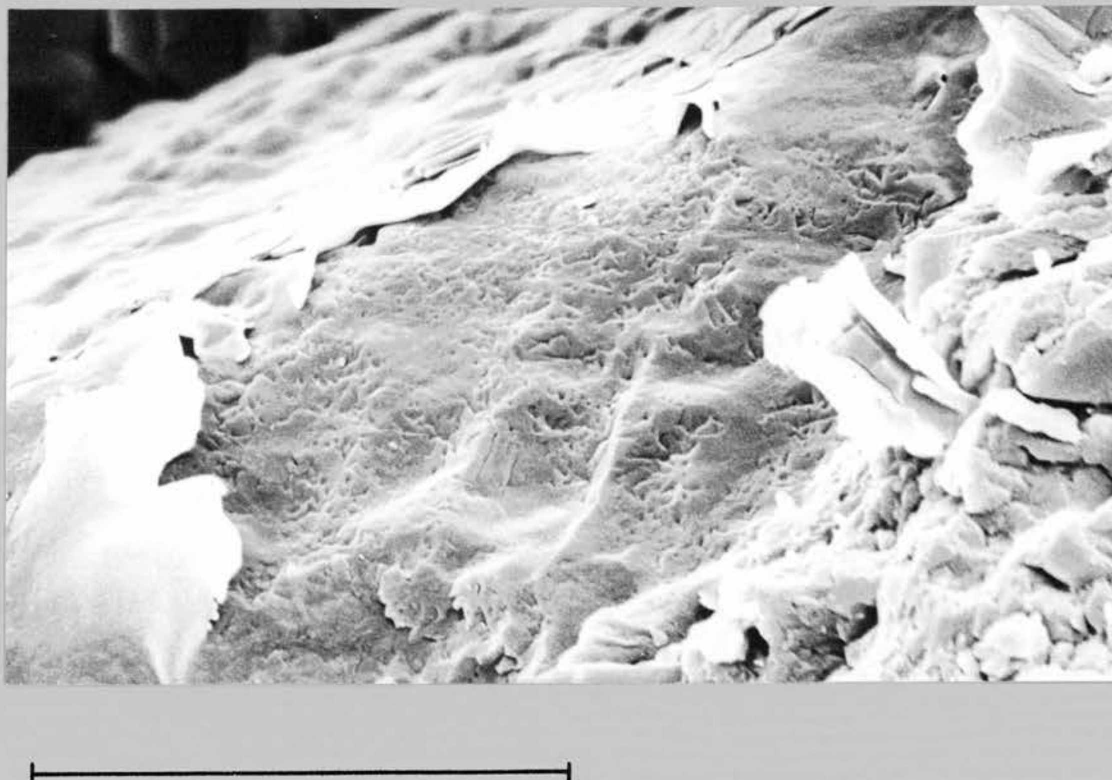


Plate 27. Scanning electron photomicrograph demonstrating 'upturned plates' on quartz grain surface (scale bar = 10 microns)

Plate 28. Scanning electron photomicrograph demonstrating quartz grain surface (scale bar = 40 microns)



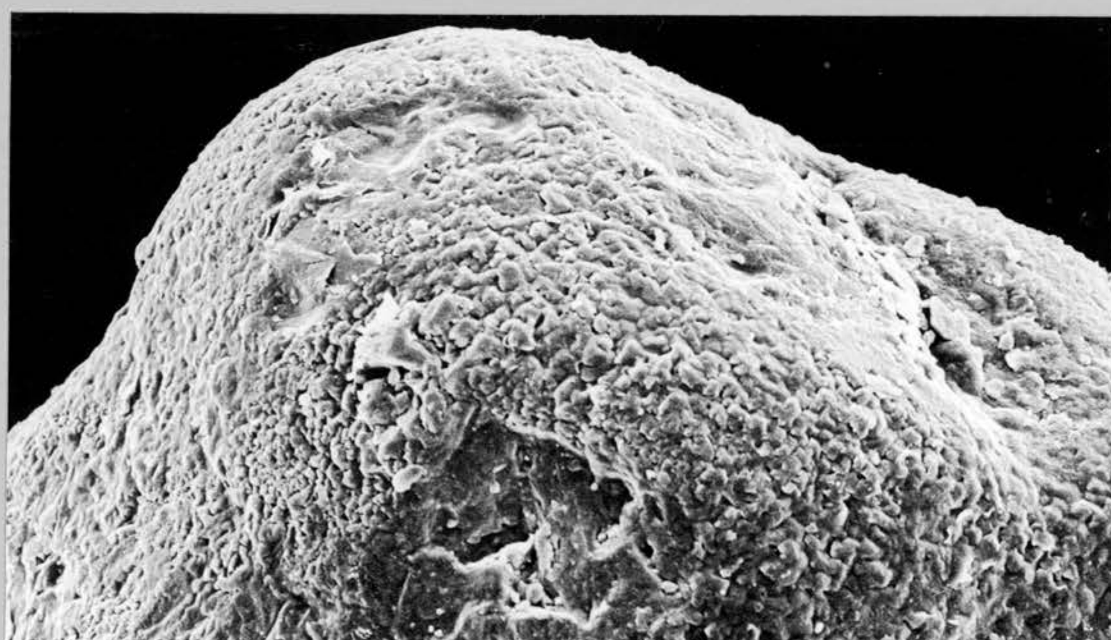
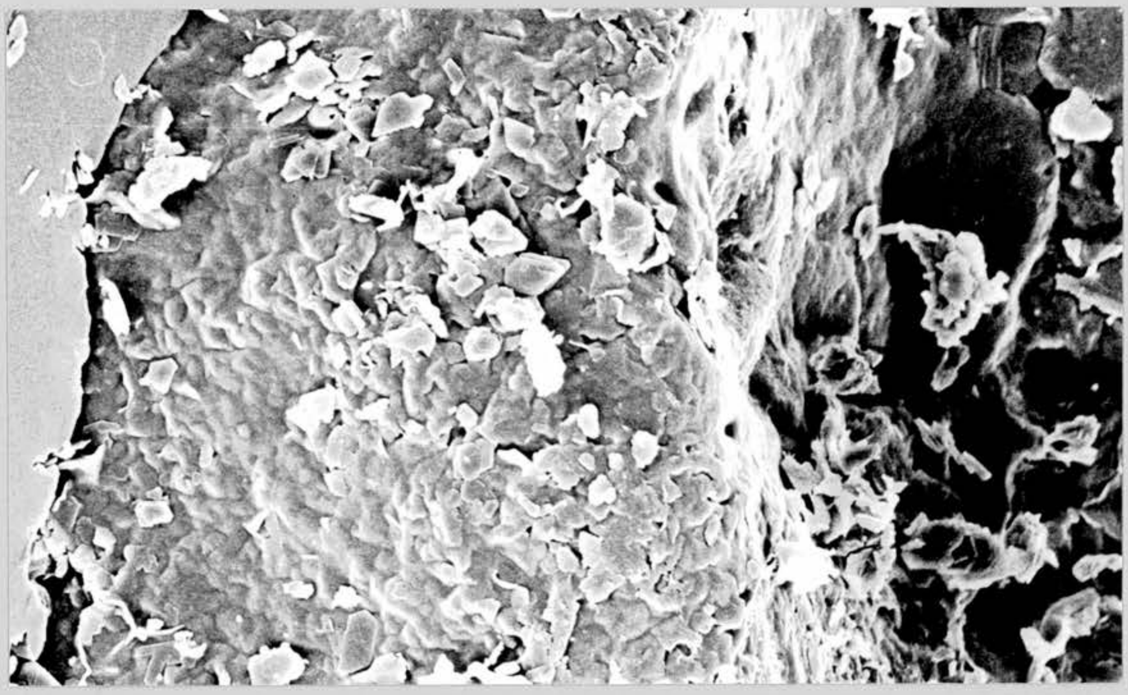


Plate 29. Scanning electron photomicrograph demonstrating carbonate corrosion of a quartz (scale bar = 40 microns)

Plate 30. Scanning electron photomicrograph demonstrating geopetal fabric - mechanical infiltration of clay (scale bar = 100 microns)



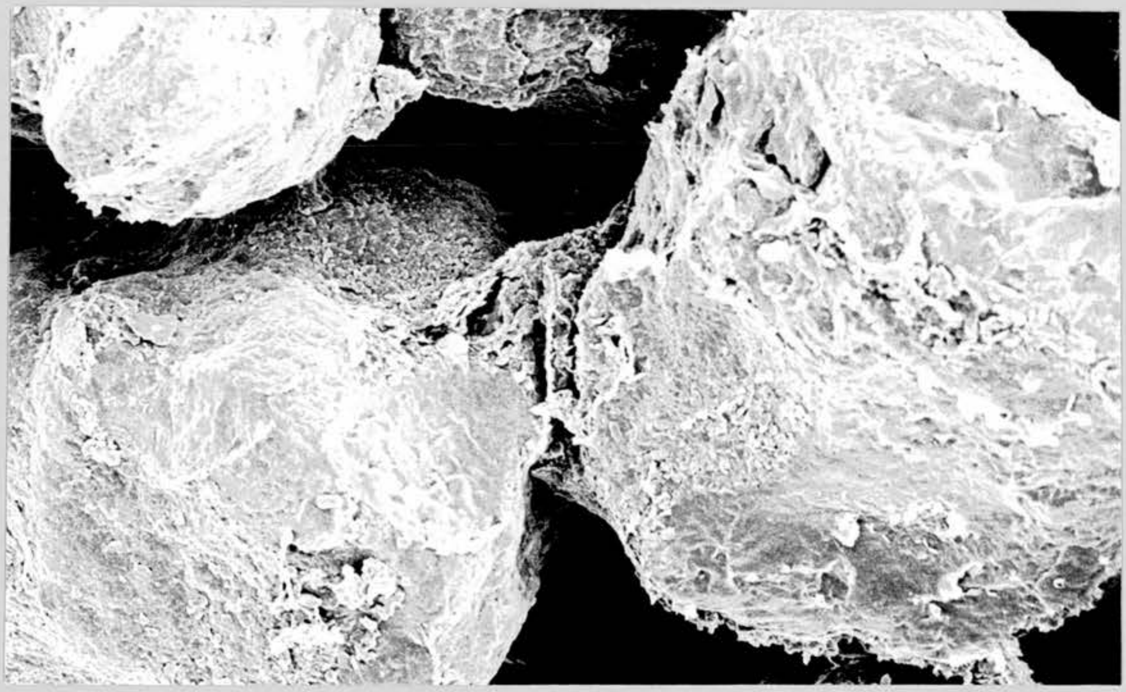
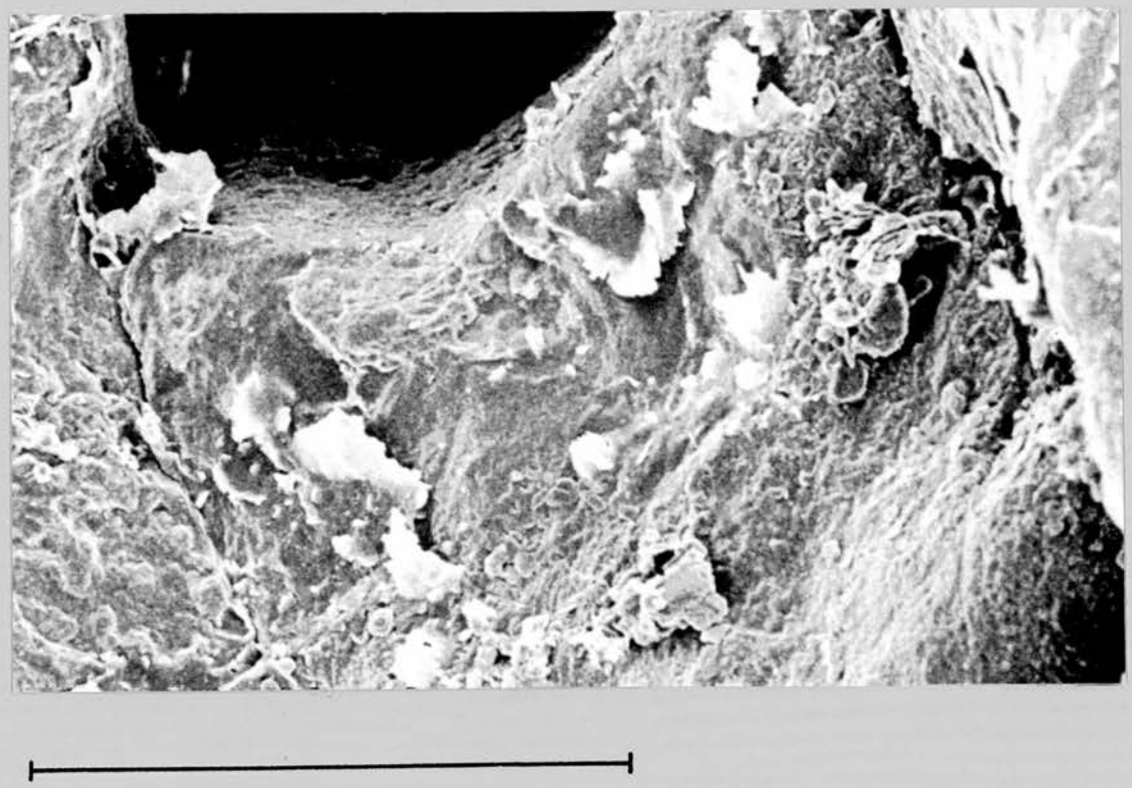


Plate 31. Scanning electron photomicrograph of mechanically infiltrated clay analysed by EDX and EDX graph below (scale bar = 100 microns)



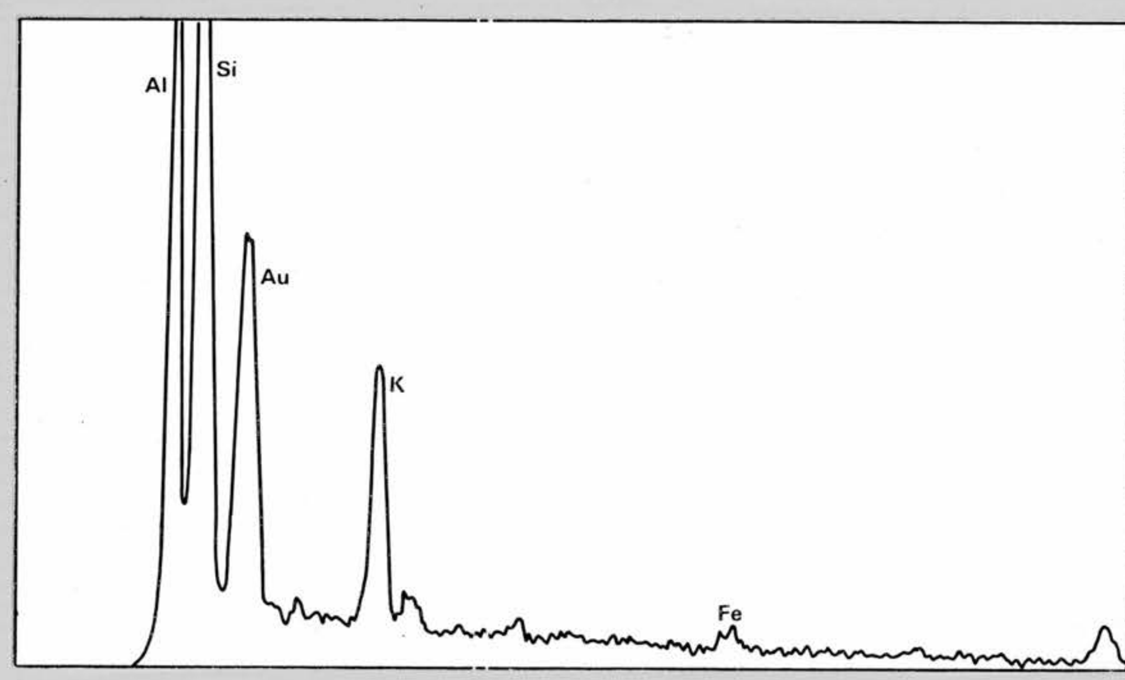
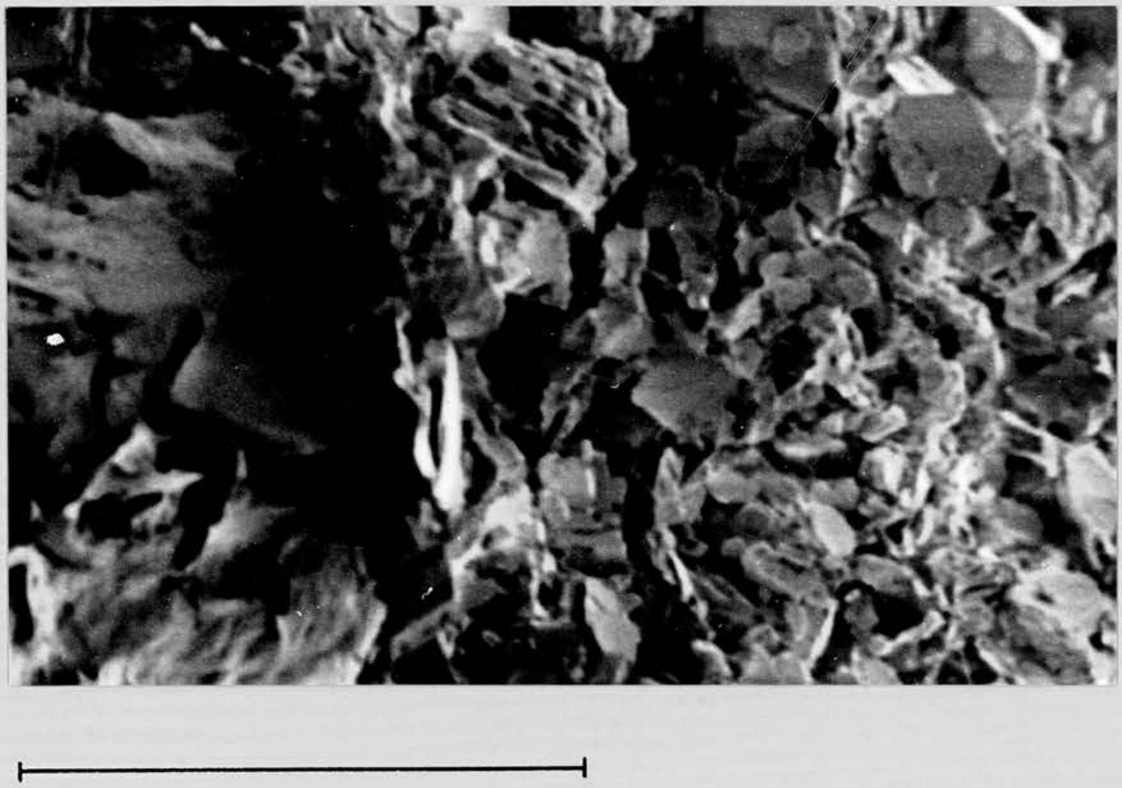
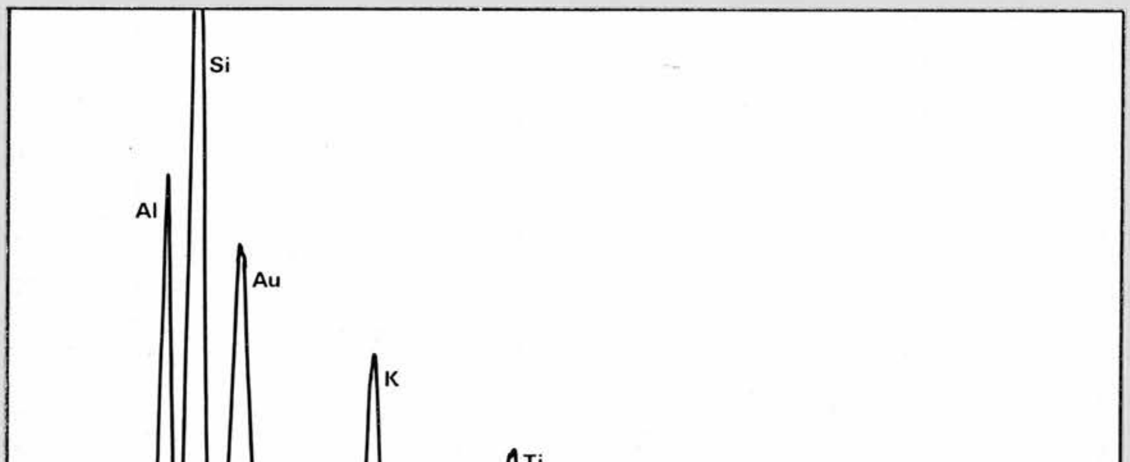

Al
Si
Au
K
Fe

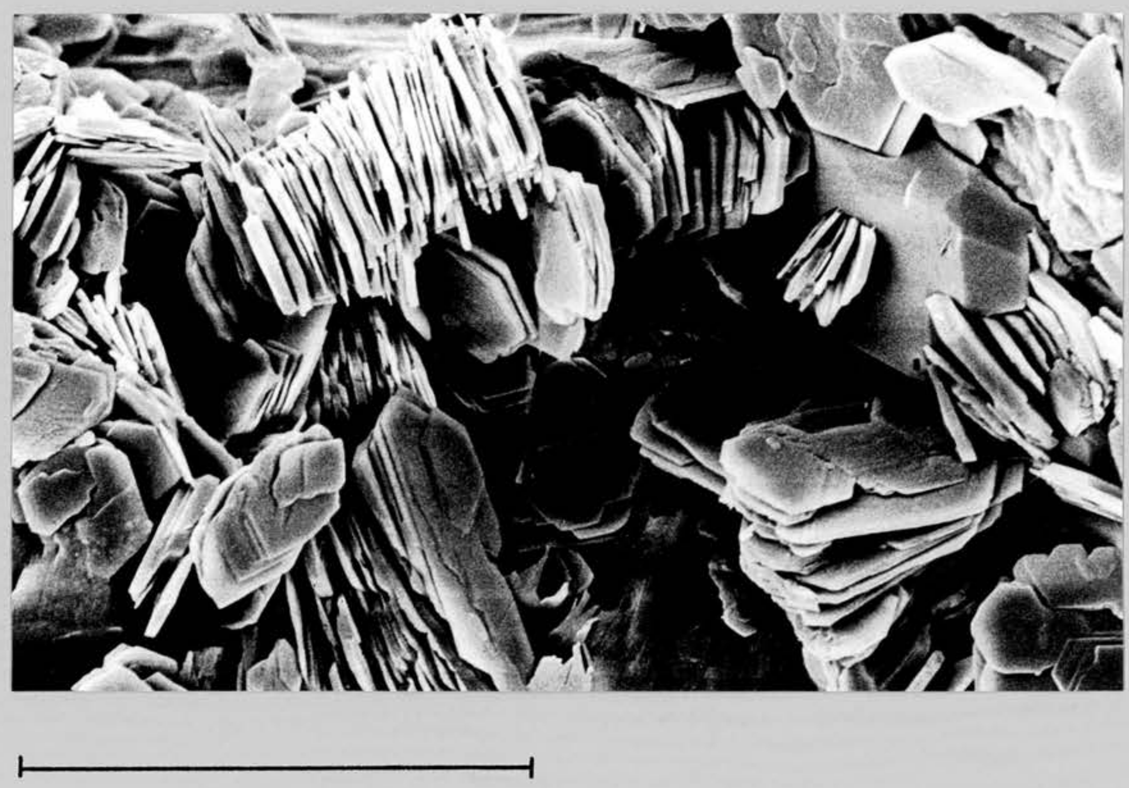
Plate 32. Scanning electron photomicrograph of clay mineral analysed by EDX as being high-titanium illite (scale bar = 10 microns)

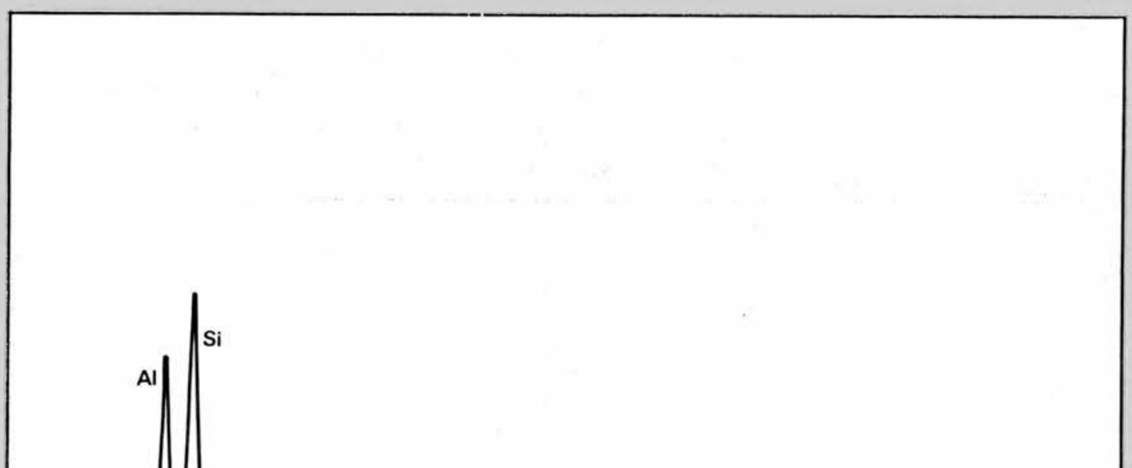




Si
Al
Au
K
Ti
Fe

Plate 33. Scanning electron photomicrograph of kaolinite and EDX graph (scale bar = 20 microns)





Si
Al
Au

Plate 34. Scanning electron photomicrograph demonstrating 'wispy' illite (scale bar = 10 microns)

Plate 35. Scanning electron photomicrograph demonstrating etching of feldspar grain surface (scale bar = 20 microns)