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Sedimentology of the Penrith Sandstone and brockrams (Permo-Triassic) of Cumbria, north-west England

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ABSTRACT

The Penrith Sandstone and brockrams (breccio-conglomerates) are continental red-beds of Lower Permian and Permo-Triassic ages respectively. The brockrams are predominantly the deposits of coalescing alluvial fans which accumulated as the products of subaerial sheetflood, transitional debris flow and braided river processes at the margins of desert basins bordering the upstanding Lake District. The fans were deposited on an irregular topography which is demonstrated to have a probable minimum pre-burial relief of 250 to 300 metres in west Cumbria. The cross-stratified Penrith Sandstone of the Eden Valley accumulated as the foresets of large scale (up to 100 metres width) aeolian dunes orientated in response to a uniform east-south-easterly palaeowind. Dune-bedding within these deposits indicates a crescentic-downwind (barchanoid) configuration for the dune slipfaces. The stratigraphic unit termed the 'Upper Brockram' is believed to represent the deposits of ephemeral streams (arroyos) which flowed through and reworked part of the aeolian sand sea of the Penrith Sandstone.

Nodular carbonate profiles occurring within sandstone/siltstone horizons intercalated with distal alluvial fan deposits in west Cumbria are interpreted as immature caliche soil horizons and represent periods of surface stability (non-deposition) of indeterminate length. Petrographic evidence indicates the most important process of formation to have been that of replacement of the detrital quartz component by microcrystalline non-ferroan calcite.

Extensive post-depositional (diagenetic) modification to the mineralogy and texture of the brockrams of west Cumbria has resulted from the mechanical infiltration of clay, the dissolution of the less stable detrital components (primarily volcanic clasts), replacement by secondary clay and the precipitation of interstitial and void-filling authigenic clay, a

hydrated haematite precursor mineral, ferroan and non-ferroan calcite and gypsum. Most reactions are considered to have occurred as stages within an evolving diagenetic sequence.

CONTENTS

INTRODUCTION

The purpose of research undertaken in the Eden Valley and western areas of Cumbria, north-west England has been to describe various aspects of the sedimentology and petrography of the Permian Penrith Sandstone and Permo-Triassic brockrams in the light of existing descriptions of red-bed sequences and of contemporary arid and semi-arid region sediment successions with the aim of interpreting their precise modes of formation. Petrographic studies of the brockram of west Cumbria have been directed towards an evaluation of the effects of post-depositional textural and mineralogical modification by reference to the many recent advances in the field of diagenesis. It has also been the intent of this aspect of research to attempt a qualification of the chemical evolution of the pore fluids and to relate these changes to the present (observed) mineralogy of the rock. Hitherto unrecorded nodular carbonate profiles occurring in sand-silt horizons intercalated with the breccias in west Cumbria have been selected for a similar study which has been primarily concerned with their genesis and environmental significance. Arising from these objectives, the availability of large amounts of borehole data for the Egremont-Beckermet region of west Cumbria has enabled these studies to be supplemented with computer-based graphical techniques for the construction of maps and perspective block diagrams of brockram isopachytes and palaeotopography, designed to facilitate the detailed appreciation of the form of the breccia accumulations in the region.

To a large extent, these researches have been designed to complement previous studies of the region, notably that of Waugh (1967), and it has not been the intent to reproduce work described in detail elsewhere. In consequence, only the important sites of occurrence of the Penrith Sandstone and brockrams have been selected for detailed investigation and no remapping of the areas was undertaken.

Summary of previous research

The Permian and Triassic rocks of Cumbria were first described by Sedgwick (1832, 1836) and later in more detail by Binney (1855, 1857), Harkness (1862) and Binney (1865). The earliest attempts to correlate and describe the distribution of this series equated the gypsiferous marls occurring below red sandstones in the Eden Valley with the 'lower red marls and gypsum' of Yorkshire (Sedgwick, 1836) whereas the sandstones themselves were considered to be Bunter in age (Binney, 1855). The underlying sandstones and breccias (first noted by Phillips, 1835 and distinguished as 'hard' and 'rotten' brockram by Binney, 1855) were ascribed a Lower Permian age by Harkness (1862) who also described the occurrence of plant beds in Hilton Beck (later termed the Hilton Plant Beds by Goodchild, 1889) and noted the 'hard' brockram of Stenkrith Bridge, south of Kirkby Stephen to be directly overlain by St. Bees Sandstone. Previously, Phillips (1836) had considered the 'hard' brockram to be equivalent to the Magnesian Limestone. The classification was partly revised by Murchison (1864) and Murchison and Harkness (1864) who, similarly to Binney (1857, 1865), proposed an exclusively Permian age on the basis of the continuity of the succession in the area. Subsequently, a Permian age for the St. Bees Sandstone was adopted by Holmes (1881) and Brockbank (1892) although Holmes (op. cit.) recognised a different succession for the Carlisle Basin and referred the two uppermost units of the Stanwix Shales and Kirklinton Sandstone to the Triassic system. Goodchild (1893), however, recognised the difficulties of placing the Permo-Triassic boundary within this succession and proposed that all beds above the Magnesian Limestone Series should be regarded as Triassic and all below as Permian. Goodchild (op. cit.) also omitted the 'Upper Gypsiferous Shale' group of Holmes (1881) from this classification.

This basic scheme was subsequently adopted by the Geological Survey for the Carlisle Basin and northern part of the Eden Valley (e.g. Holmes, 1899;

Dixon et al., 1926; Trotter and Hollingworth, 1932) although the Bunter Marls[1] of Goodchild (1893) were redesignated the St. Bees Shales by Dixon et al. (op. cit.) following the introduction of this term in west Cumbria by Smith (1924) and were only reservedly assigned to the Bunter for the purposes of mapping. The difficulties of assigning an age to this group where the Magnesian Limestone is absent from the succession had been recognised by Dixon et al. (1926) and Trotter and Hollingworth (1932) and its inclusion with the St. Bees Sandstone was based on local stratigraphic considerations rather than correlations with the Upper Permian sequences of north-east England (e.g. Goodchild, 1893; Dixon et al., 1926; Sherlock, 1926; Trotter and Hollingworth, 1932). Similar problems had also been encountered in west Cumbria and the Shales were assigned to a transitional, Permian-Triassic group (Smith, 1924) which was adopted by later workers in the area (e.g. Eastwood et al., 1931; Trotter et al., 1936).

The classification proposed for west Cumbria by Smith (1924; based largely on evidence from surface exposures as at Saltom Bay, NX959160) has since been broadly confirmed by information from borings in the area and adopted by subsequent workers with only minor modification (e.g. Hollingworth, 1942; Taylor, 1961; Meyer, 1965; Arthurton and Hemingway, 1972). The base of the St. Bees Shales was redefined by Arthurton and Hemingway (op. cit.) and the term St. Bees evaporites introduced for the gypsum-anhydrite sequences (the lower part of the St. Bees Shales of Smith, 1924) overlying the Basal Breccia (the Lower Brockram of Smith, op. cit.) of these authors. These authors regarded the St. Bees Shales as being of Permian age and arbitrarily defined the base of the St. Bees Sandstone as the base of the Triassic. Accordingly, the Basal Breccia is considered to be of probable Lower Permian age.

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[1] Corresponding to the Lower Gypsiferous Shales of Holmes (1881).

Following the earlier Survey publications for the Eden Valley and Carlisle Basin, the detailed stratigraphy of the Upper Permian, and particularly the evaporite sequences, was referenced from several boreholes completed in the area (Sherlock and Hollingworth, 1938; Hollingworth, 1942; Meyer, 1965; Arthurton, 1971; Burgess and Holliday, 1974) and has also been summarised by Waugh (1967). The term 'Eden Shales' was introduced by Arthurton (1971) on the basis that this formation does not precisely equate with the St. Bees Shales of west Cumbria and were regarded as being largely of Upper Permian age from palaeontological evidence (Clark, 1965 and Pattison, 1970 _in_ Arthurton, _op. cit._). As with west Cumbria, the base of the Triassic is currently placed at the base of the St. Bees Sandstone whereas the Penrith Sandstone and much of the brockram are regarded to be Lower Permian in age.

The correlation and stratigraphy of the Permo-Triassic sequences have been summarised recently in Arthurton _et al._ (1978) whilst, at the time of writing, memoirs for the Penrith and Brough districts (by Arthurton and Wadge and Burgess and Holliday respectively) are awaiting publication.

Although the earliest references to the breccia sequences of Cumbria are those of Sedgwick (1832), Phillips (1835, 1836) and Sedgwick (1836), it was Binney (1855) who first recognised the characteristics of the 'hard' and 'rotten' brockram occurring in the Eden Valley (previously, Sedgwick, 1836 had noted that the pebbles of the brockram occurring at Burrells were derived exclusively from Carboniferous rocks) and Kendall (1902) who made the first serious attempt to account for the distribution of the Lower and Upper Brockrams. Although Kendall (_op. cit._) envisaged both brockrams as having been derived from the Pennine scarp (with the differing pebble constituents evidencing an intra-Permian movement of the Pennine fault), structural studies by Turner (1927) and Trotter and Hollingworth (1928) suggested the Alston Block to be depressed relative to the Eden Valley in

Permian times and thus unlikely to have provided a source for the detritus. The discovery of Whin Sill pebbles (Holmes and Harwood, 1928; Dunham, 1932), however, revived discussion and again suggested an easterly derivation for part, at least, of the Upper Brockram. Structural evidence (Shotton, 1935; Turner, 1935) continued to contradict such a derivation direction and the matter remained unresolved (e.g. Trotter and Hollingworth, 1932; Versey, 1939; Waugh, 1967) until the model erected by Bott (1974) for the evolution of the Eden Valley reconciled the occurrence of Lower Palaeozoic fragments in the Upper Brockram with a northern Pennine source.

The occurrence of a breccia sequence within the Eden Shales of the southern part of the Eden Valley has been noted by Eccles (1870), Dakyns *et al.* (1891), Burgess (1965) and Waugh (1967). This sequence has been termed the Stenkrith Brockram by Burgess (*op. cit.*) and the St. Bees Brockram by Waugh (1967). Both authors refer to the underlying breccias as the Penrith Brockram (on the basis that its identity as either Upper or Lower Brockram cannot be established); a term which is adopted herein for all breccias which cannot specifically be classed with either the Upper or Stenkrith Brockrams.

Following Sorby's (1880) perceptive comments as to the nature of the 'crystallised sands' of the Penrith Sandstone (a term proposed by Harkness, 1862), and excluding early attempts by Goodchild (1885) to interpret the depositional history of this sequence,[1] the character of the cross-bedding was generally accepted to reflect accumulation within an inland sand sea (e.g. Trotter and Hollingworth, 1932) but was not fully appreciated until Shotton (1956) indicated the possibility of a barchan dune formation. A similar interpretation has been proposed by Waugh (1967) although subsequently Glennie (1970) has compared the structures exhibited in exposures in Hilton

[1] Goodchild (1885) envisaged the sandstone to have been deposited in a large inland lake in which strong easterly currents operated. The brockrams were suggested to have been deposited by melting ice sheets.

Beck with those of a sief dune. Petrographic studies (Waugh, 1965, 1967, 1970) have confirmed a secondary origin for the cement of the silicified Penrith Sandstone, as earlier proposed by Sorby (1880).

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CHAPTER ONE

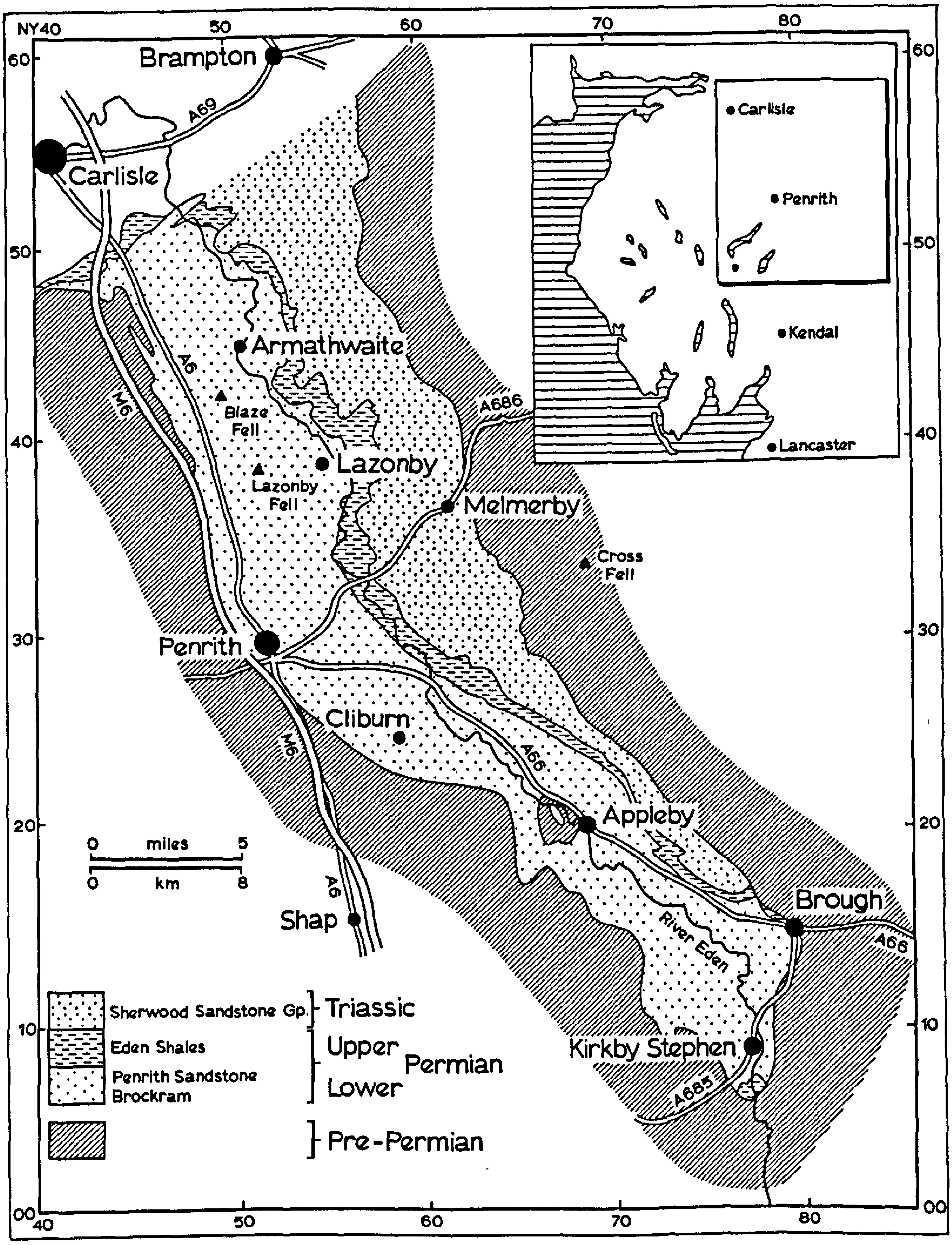
THE SEDIMENTOLOGY OF THE BROCKRAMS

INTRODUCTION

The variable sequences of conglomerates and breccias known as the brockrams form part of a thick succession of continental rocks of Permo-Triassic (New Red Sandstone) age which outcrop in the topographically lower areas flanking the main upstanding block of the Lake District of Cumbria, north-west England. The most extensive exposures occur in the Eden Valley where they form part of a red-bed sequence with an estimated thickness of 1000 metres (Bott, 1978). The breccias are of restricted occurrence within this succession but, where they do occur, generally form the lowermost stratigraphic unit which extends from Kirkby Stephen north-westwards to Crackenthorpe, near Appleby (see Figure 1.1). In other areas of Cumbria, exposures are limited and are found in a few isolated localities in the south, including an exposure at Roughholme Point (SD385740) first described by Sedgwick (1836) and in several river sections occurring mainly to the north and east of Egremont in the west. In the latter area, however, extensive brockram sequences exist underground beneath a cover of St. Bees Shale and Sandstone of Upper Permian and Triassic (Sherwood Sandstone Group) ages respectively. These sequences are known in the Egremont-Beckermet area (see Figure 1.2) from data and specimens supplied from in excess of 500 boreholes drilled by the British Steel Corporation and their antecedents in the search for haematite ore bodies emplaced in the underlying Carboniferous Limestone Series.

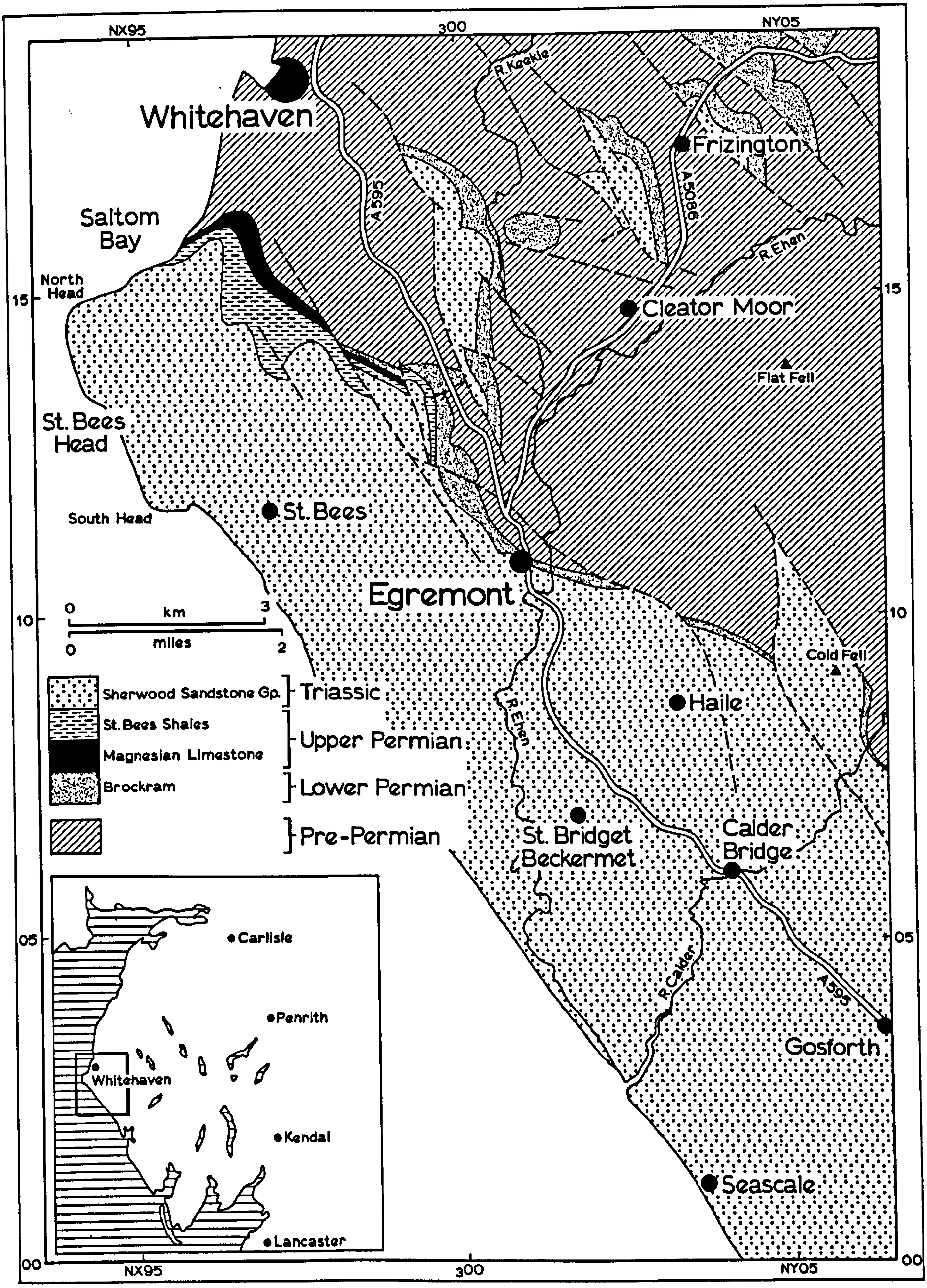
Figure 1.1. Locality map of the Eden Valley showing the simplified distribution of Permo-Triassic strata.

Figure 1.2. Locality map of western Cumbria showing the simplified distribution of Permo-Triassic strata.



NY40
50
60
70
80
Brampton
A69
Carlisle
Armathwaite
A6
M6
Blaze Fell
Lazonby
Lazonby Fell
A686
Melmerby
Cross Fell
Penrith
Cliburn
M6
A66
Appleby
A6
Shap
Brough
A66
River Eden
Kirkby Stephen
A685
0 miles 5
0 km 8
Carlisle
Penrith
Kendal
Lancaster
Sherwood Sandstone Gp.
Triassic
Eden Shales
Upper
Permian
Penrith Sandstone
Brockram
Lower
Pre-Permian

Figure 1.1



NX95
300
NY05
Whitehaven
R. Keekle
Frizington
A595
A5086
Saltom Bay
North Head
R. Ehen
15
Cleator Moor
Flat Fell
St. Bees Head
South Head
St. Bees
Egremont
0
km
3
10
0
miles
2
Cold Fell
Sherwood Sandstone Gp.
Triassic
St. Bees Shales
Magnesian Limestone
Upper Permian
Brockram
Lower Permian
Pre-Permian
Haile
St. Bridget Beckermet
Calder Bridge
Carlisle
05
Penrith
Whitehaven
R. Calder
A595
Gosforth
Kendal
Seascale
Lancaster
00

Figure 1.2

The purpose of this aspect of research has been to describe the geological setting and sedimentology of the brockrams with the aim of determining the dominant mode or modes of formation and with the intent to provide a basis for the proceeding petrographic description (Chapter Two) and studies of associated deposits (Chapters Three to Five). In this respect, no attempt has been made to include detailed descriptions of every known brockram locality and only the main areas of their occurrence have been investigated. Despite the data generously supplied by British Steel, however, the bulk of the environmental interpretation is based on the exposed brockram of Edenside as the nature of the underground sequences usually does not enable individual sedimentary units to be traced laterally across the Beckermet area. The large amount of available data from this area, however, has enabled the overall distribution of the facies to be accurately plotted (see, for example, Figures 1.16 and 1.21).

In view of the diverse terminology applied to these and associated deposits, it is considered necessary to clarify the descriptive nomenclature which will be used in the following chapters. The use of the term 'brockram' is suggested to have originated as a local descriptive name applied to the lithology of the 'broken rock' quarried as a building stone at Burrells and Kirkby Stephen (Waugh, 1967). The term has been used throughout geological literature and has also been used in the stratigraphic nomenclature for the area. To avoid confusion in the use of the term, therefore, it is proposed to adopt the suggestion of Waugh (1967) that 'brockram' should refer to the lithology and 'Brockram' to the stratigraphic division. In western Cumbria, brockram is locally referred to as 'connie' which presumably originated from 'The conglomerate', a term which appears in many borehole records. As Waugh (1967) records, however, this term is indeed a misnomer as rounded and subrounded clasts do not necessarily predominate. For the most part, lithological descriptions based on clast roundness are misleading as fragments ranging from angular to sub-rounded frequently occur within individual beds and the designation of a particular bed as breccia or conglomerate is largely arbitrary. Although Lawson (1913 in Blackwelder, 1928) has ascribed the term 'fanglomerate' to deposits interpreted as the products of alluvial fan sedimentation (see Interpretation, this chapter), it is proposed to refer to all coarse clastic rocks simply as breccia (c.f. Laming, 1966) as this term (although strictly inaccurate) has no specific formational implications.

STRATIGRAPHY OF THE BROCKRAMS AND ASSOCIATED DEPOSITS

The Permo-Triassic deposits of Cumbria were first described by Sedgwick (1832, 1836) and have since been described in more detail by authors such as Binney (1855, 1857), Harkness (1862), Murchison and Harkness (1864), Goodchild (1893), Smith (1924), Versey (1939), Burgess (1965), Meyer (1965), Waugh (1967), Arthurton and Hemingway (1972) and Burgess and Holliday (1974). The stratigraphy and correlation of these sequences is well established, therefore, (see, for example, Arthurton et al., 1978, Table 4, p.200) and need only be described in terms of the occurrence of the brockram facies for the purpose of this research.

The stratigraphic relationship of the Permo-Triassic deposits of the Eden Valley is shown schematically in Figure 1.3. Within this succession, the brockrams occur as a basin margin facies which flanks the uplands to the west and south of the valley and are predominantly of Penrith Sandstone (Lower Permian) age. Whereas previous workers have divided the Edenside brockrams into three stratigraphic units (the Lower, Upper and St. Bees Brockrams of Waugh, 1967), only two divisions are considered in this chapter, these being the lower, Penrith (as termed herein) Brockram and the higher, St. Bees Brockram (now termed the Stenkrith Brockram after Burgess, 1965).

Figure 1.3. Stratigraphic relationship of the Permo-Triassic sequences in the Eden Valley (modified from Waugh, 1967 and Arthurton et al., 1978).

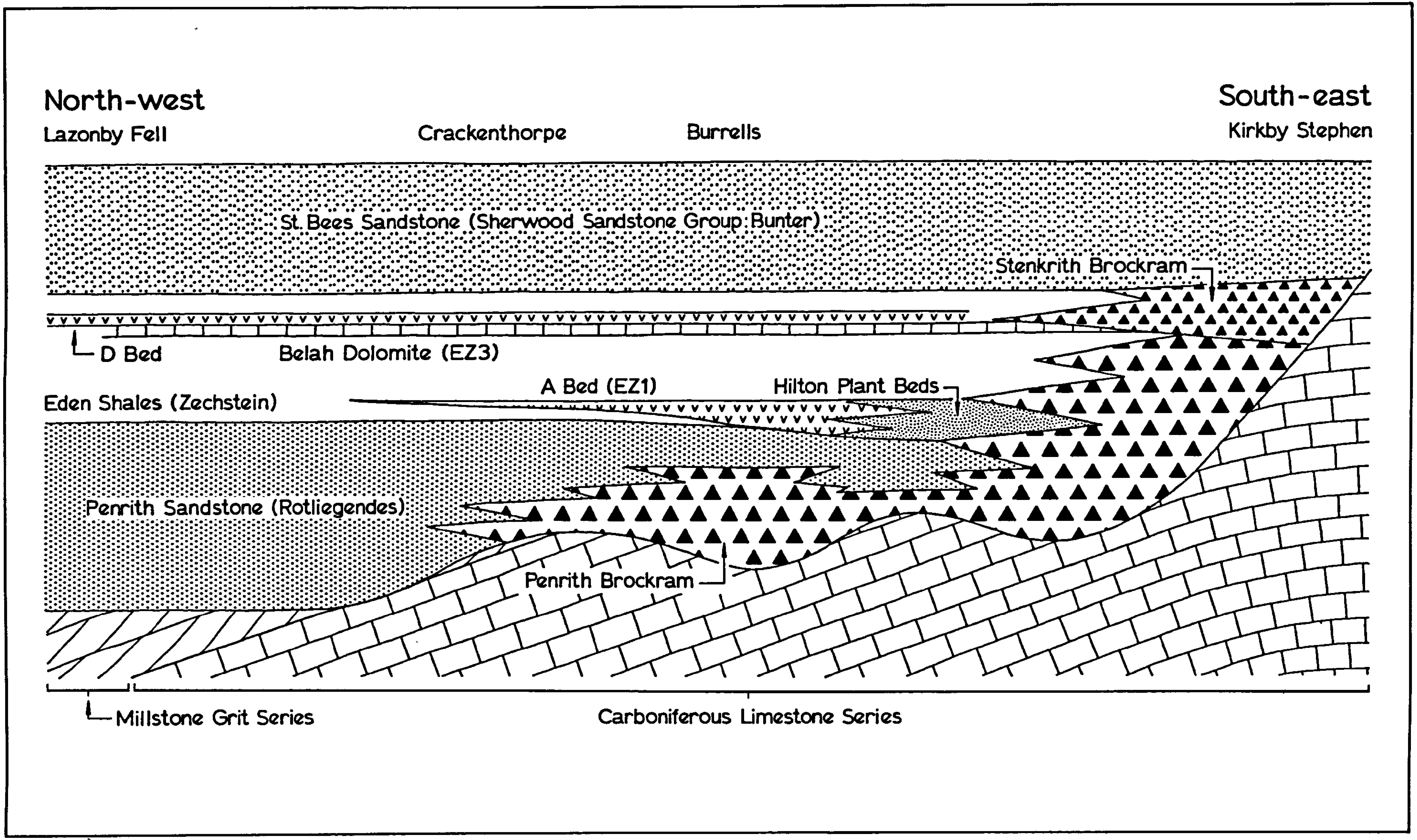

North-west
Lazonby Fell
Crackenthorpe
Burrells
South-east
Kirkby Stephen
St. Bees Sandstone (Sherwood Sandstone Group: Bunter)
Stenkrith Brockram
D Bed
Belah Dolomite (EZ3)
A Bed (EZ1)
Hilton Plant Beds
Eden Shales (Zechstein)
Penrith Sandstone (Rotliegendes)
Penrith Brockram
Millstone Grit Series
Carboniferous Limestone Series

Figure 1.3

The dolomitised breccia and reworked Penrith Sandstone sequences exposed in localities such as Hilton Beck (NY714202) and the River Belah (NY795121) correspond to the Upper Brockram of Waugh (1967) which also includes the dolomitised breccias occurring in the River Eden at Kirkby Stephen (the 'Rotten Brockram' of Binney, 1855) where the Upper and Lower Brockrams were suggested to fuse. Whereas the formation of the former are not considered to have resulted solely from accumulation on alluvial fans and are described separately in Chapter Three, the latter are here incorporated into the Penrith Brockram on the basis that they do not differ significantly from exposures of this stratigraphic unit elsewhere in the Eden Valley. It is suggested that the dolomitisation of the brockrams at Kirkby Stephen and elsewhere is the result of an early post-depositional, localised alteration by saline interstitial fluids in which magnesium ions were concentrated by high evaporation rates and is thus only indirectly related to their mode of formation.

Over most of their occurrence, the brockrams pass laterally into and are the age equivalent of the Penrith Sandstone. In the southern part of the Eden Valley, however, the Penrith Sandstone is successively overlapped by higher beds of Upper Permian (and possibly Triassic) age which ultimately pass directly into and overlie the breccias in the Kirkby Stephen area. The brockrams reach a maximum thickness of 150 metres in the Appleby-Burrells area (Arthurton et al., 1978) but thin rapidly away from this area towards the north-west and pass into Penrith Sandstone in the region of Crackenthorpe. To the south-east, the brockrams thin to less than ten metres thickness in the region of Jeremy Gill (NY694170) and Helm Beck (NY707157) and pass into a sequence of thin breccias with associated, fluvially deposited sandstones. Towards Kirkby Stephen, however, the breccias are exposed in the region of Bermer Scar (NY745146) and increase in thickness until the complete Lower Permian and part of the Upper Permian sequence is formed by the Penrith Brockram which is overlain, in this area, by the Eden Shales.

In the extreme south-east of the Eden Valley, the Eden Shales are succeeded by, and pass laterally into, the Stenkrith Brockram which is thus, in part, the age equivalent to the Eden Shales. Ultimately, the Stenkrith Brockram overlaps both the Eden Shales and the underlying Penrith Brockram and rests unconformably on Carboniferous Limestone in the region of Thringill (NY775061). The exact nature of the upper contact of the Stenkrith Brockram is not entirely certain and it is not clear whether the Eden Shales everywhere overlie the breccia, as in the region of Frank's Bridge (NY776087) and as suggested by Burgess (1965) and Waugh (pers. comm.) or whether the highest parts of the Brockram (as at Stenkrith Park, NY777079) are directly overlain by the St. Bees Sandstone. In the latter case, the upper parts of the Brockram may be lowermost Triassic in age. In the southernmost part of the Eden Valley, therefore, sedimentation was dominated by the brockram facies whose formation spanned the Permian era and may have continued into earliest Triassic times.

Where the base of the brockram sequences are exposed, the breccias can be seen to rest unconformably on several members of the Carboniferous Limestone and Millstone Grit Series which form the low fells flanking the western and southern margins of the Eden Valley against which the brockrams are banked. The nature of the contact is generally represented by an angular unconformity but, as noted by Turner (1936), is particularly irregular in localities such as Crosby Garrett (NY730097) and Waitby (NY752083). Both palaeocurrent data and clast lithology suggest a localised derivation from the western and southern margins and, excluding the interbedded breccia and reworked Penrith Sandstone sequences and the localised brockram sequences recorded from the Roman Fell area (Kendall, 1902; Burgess and Wadge, 1974; Arthurton _et al._, 1978), suggest that the Pennine Fells of the eastern margin did not form a significant source area, even though they were probably upstanding at the time as a result of contemporary subsidence along the line of the Inner Pennine fault zone (Bott, 1974, 1978).

The stratigraphic relationship of the brockrams and associated deposits in western Cumbria (see Figure 1.2) is shown diagrammatically in Figure 1.4. In this area, the brockrams form a thick (133 metres maximum recorded vertical thickness) sequence of breccias banked against the western margin of the Lake District and which form the lowest stratigraphic unit of a sequence of Permo-Triassic rocks deposited on the extreme eastern margin of the East Irish Sea Basin (Bott, 1978). The brockrams thin westwards until, in the region of Saltom Bay (NX959159) only a thin (two metres) basal breccia is present. This breccia corresponds to the Basal Breccia of Arthurton and Hemingway (1972) and is overlain by the St. Bees Evaporites and St. Bees Shales of Zechstein age which thus form the lateral equivalent to the brockrams in the east. The St. Bees Shales are succeeded and overlapped by the Triassic (Bunter) St. Bees Sandstone of the Sherwood Sandstone Group which ultimately overlie the brockrams in the east. The base of this series is, in the west, equivalent to the uppermost parts of the brockram sequence in the east and has been arbitrarily defined as the base of the Triassic system (Arthurton and Hemingway, 1972).

The age of the brockram basal breccia which underlies the St. Bees Evaporites in the west, is regarded as Lower Permian (e.g. Arthurton et al., 1978) although it is not clear whether extensive active deposition occurred in the east of the basin during this period. Lithologically, the lowermost few metres of brockram may be distinguished everywhere from the overlying breccia by the presence of large Carboniferous clasts which probably originated mainly by the mechanical disintegration of the pre-Permian surface. It is likely, therefore, that the base of the sequence is represented everywhere by a basal breccia to which, by analogy with the sequences in the extreme west, can be ascribed a Lower Permian age. It follows, therefore, that the

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Figure 1.4. Stratigraphic relationships of the Permo-Triassic sequences in western Cumbria.


North-west
Saltom Bay
Egremont
South-east
Haile
Cold Fell
St. Bees Sandstone (Sherwood Sandstone Group: Bunter)
St. Bees Shales (Zechstein)
Brockram
Basal Breccia (Rotliegendes)
St. Bees Evaporites (EZ1 and EZ2)
Millstone Grit Series
Carboniferous Limestone Series
Basement Beds
Skiddaw Group and Borrowdale Volcanic Group
Coal Measure Series

Figure 1.4

main phase of sedimentation in this area is probably of Upper Permian (Zechstein) age as distinct from the dominantly Lower Permian (Rotliegendes) sedimentation of the Edenside brockrams.

The Permo-Triassic deposits of western Cumbria rest with marked unconformity on an irregular Carboniferous surface with a westwards dipping palaeoslope (see Figure 1.16, for example). The Carboniferous dips westwards so that in the east of the depositional basin the brockrams overlie the Carboniferous Basement Beds and westwards overlie progressively younger Carboniferous rocks until, at Saltom Bay, the basal breccia is unconformable on the Middle Coal Measures Whitehaven Sandstone. At the extreme eastern margin of the basin, the brockrams are banked against Lower Palaeozoic rocks of the Arenigian Skiddaw Group and Llandeilian and Caradocian Borrowdale Volcanic Group which form the main mass of the Lake District. The nature of the contact between the Lower Palaeozoic strata and the Carboniferous is uncertain and may be either unconformable (e.g. Eastwood et al., 1931, Fig. 22, p.208; Arthurton and Hemingway, 1972, Fig.3, p.569) or faulted (Trotter et al., 1936).

DEPOSITIONAL ENVIRONMENT OF THE BROCKRAMS

The depositional environment in which the Permo-Triassic sequences of Cumbria accumulated has for many years been recognised as that of a predominantly arid to semi-arid continental regime which was initiated subsequent to the earth movements associated with the Hercynian orogeny of the mid-Westphalian to early Permian (e.g. Smith, 1972; Waugh, 1973; Smith, 1976). During this time, desert basins, many of which were fault bounded, and areas of bare rocky pediment flanking regions of high relief were established over much of Britain and Europe (Waugh, op. cit.; Arthurton et al., 1978). Several coalescing basins are considered to have been established around the

Lake District with contemporaneous downwarping leading to the accumulation of thick sedimentary sequences (Arthurton et al., 1978).

Although the generalised palaeogeography is well established for the area, it is considered relevant to include a brief summary of the evidence on which a subaerial mode of formation for the brockrams may be postulated in order to restrict the proceeding interpretive description to a minimum. A more detailed discussion is presented in Chapter Five where such considerations may be regarded of particular relevance in the light of recent studies of rock successions comparable to the Penrith Sandstone whose subaerial mode of formation has been disputed.

The brockrams dominantly comprise coarse, sheet-like deposits of breccia which have a prominent red colouration as a result of oxidation of the iron-bearing component (see Chapter Two) and a clast assemblage which can be closely matched with lithologies outcropping peripherally to the main depositional areas (Waugh, 1967). Considerations of sedimentary structures and fabrics (this chapter) indicate fluvial processes to be the most likely depositional mechanisms of formation and, in conjunction with the above evidence and extensive lateral development of the breccias, has suggested the closest analogue to these deposits to be that of the alluvial fan environment. Although a subaerial (and hence, alluvial fan) formation cannot be conclusively established for the brockrams alone, their lateral associations, which include the Penrith Sandstone (herein interpreted as aeolian dune sands; see Chapter Five) in the Eden Valley and the St. Bees Shales, including the St. Bees Evaporites (interpreted as supratidal sabkha deposits; Arthurton and Hemingway, 1972) in western Cumbria, is strongly indicative of such an interpretation. In addition, finer horizons occurring within the brockrams in western Cumbria include nodular carbonate sequences which are interpreted herein as caliche soil horizons (see Chapter Four) and are thus apparently conclusive of a subaerial formation within an arid or semi-arid environment.

THE EDEN VALLEY BROCKRAM

In any particular exposed sequence of brockram, a number of distinct breccia types can usually be recognised according to non-petrographical[1] variations in the clast assemblage and fabric of the rock and according to any sedimentary structures which may be developed. For the most part, however, these breccia types are arranged in vertical sequence of well-developed and laterally extensive beds which do not normally succeed one another in definite order and which only rarely show non-planar and markedly discordant contacts with one another. In consequence, there is a marked similarity between the majority of brockram exposures in the area which vary only in the relative development of the different breccia types and which otherwise present a rather monotonous aspect. The proceeding section will not incorporate systematic descriptions of individual localities, therefore, (although a generalised account and detailed sedimentary facies logs for the major sites investigated are included) but will be restricted to details of the main sedimentological characteristics developed within the brockram sequences which can be used to distinguish the various breccia types.

Bedding

The most commonly developed and best displayed sedimentary structure of the brockram sequences is bedding. Several distinct but gradational styles can be recognised and range from non-erosive and laterally continuous (sheet-like) to markedly erosive and laterally restricted (channel-like). Intermediary bedding styles are common and typically represented by sheet-like beds with conspicuous internal scour surfaces. Bed thickness is usually in the range of three centimetres to two metres with an average value in the order of 30 centimetres, corresponding to the medium-thickly bedded divisions

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1 Petrographically, the breccias are dominated by derived Carboniferous limestone fragments and show little variation throughout the area.

of Ingram (1954). Statistical analysis (see Figure 1.5) has indicated a significant correlation with maximum clast size, a relationship which has been recognised from comparable deposits of Devonian (Bluck, 1967; Wilson, 1980) and New Red Sandstone age (Steel, 1974a) and which is suggested by these authors to be related to the mode of deposition (see Discussion, this chapter). Data points lying below the regression lines plotted on Figure 1.5 probably represent beds with a reduced thickness as a result of erosion prior to deposition of the overlying unit.

Sheet-like bedding is the most frequently observed style developed within the brockrams although most beds rest with slight discordance on the scoured surfaces of underlying deposits. Successive beds are often marked by conspicuous textural changes which may be related only to the presence of coarser basal horizons.

At the other extreme, the only recorded large scale ($>$0.5 metres thickness) examples of brockram units with markedly erosive bases occur in the Burrells area (NY679177, see Figure 1.8; NY678183, see Figure 1.9). This style of bedding is relatively more common in the interbedded breccias and reworked Penrith Sandstone sequences and examples are exposed at Belah Bridge (NY793120), Hayber Gill (NY751165) and Low House (centred on NY515488). The examples occurring within the Penrith Brockram are generally only intermittently exposed and do not enable accurate lateral correlation although it appears that these units have a restricted lateral development (less than 50 metres; many exposures in the Eden Valley are of too limited extent, therefore to enable accurate determinations of bedding styles) and truncate the bedding of underlying units at a maximum angle of 20 to 30 degrees. Thicknesses increase in proportion to the truncation of successively lower beds to an exposed maximum of two metres and although it has not proved possible to

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Figure 1.5. Locality map of the southern part of the Eden Valley and maximum clast size/bed thickness diagram. The distribution of the brockram facies is based on Versey (1939, p.276) and Waugh (1967, Enc.1) and incorporates the Upper Brockram of these authors.

* Including interbedded sandstones and breccias

Figure 1.5

ascertain whether these units are isolated from surrounding sediment or ultimately pass into beds with a more constant thickness, it is clear that the overall morphology is distinctly channel-like in section.

The breccias overlying these erosion surfaces (and which thus form the channel-fill sequence) occur both as massive infills of poorly sorted coarse breccia (e.g. Figure 1.8, logs 4, 6 and 7) and as a series of distinct sedimentary units separated by marked erosion surfaces with prominent concave-upward sections (e.g. Figure 1.8, logs 9 and 10 and Figure 1.9, log 4 and inset). Well developed sedimentary fabrics are generally absent in the massive type of infill although the clasts may show an approximate horizontal alignment, especially where a crude internal stratification is developed. Upward-fining of the clasts is generally restricted to the presence of a coarser base. The composite type of infill comprises a more ordered sequence in which there is a pronounced segregation of the detritus into smaller, channel-like units separated by curved erosion surfaces and forms the end member of a series of deposits which are transitional in nature to the sheet-like types described above. An example of this type of channel-fill sequence is exposed in a small disused working to the north of the main Burrells exposures at NY678183 (see Figure 1.9). At this locality, a sequence of cross-bedded and horizontally stratified breccia and sandstone forms a series of laterally impersistent and truncated units which occur above the irregular eroded surface of a bed of coarse breccia in an otherwise non-channelled sequence. The sequence thickens from 0.5 to 1.6 metres over a distance of three to four metres and at its thickest development comprises four distinct sedimentary units. With the exception of one coarse grained and structureless unit, trough cross-bedded and horizontally stratified medium to fine breccia and sandstone comprise the remainder of the sequence and, as at other localities where examples of this type of infill are exposed, form the bulk of the sequence.

Cross-bedding

Cross-bedding is a relatively common structure in the brockrams and occurs as low angle festoon-shaped scour infills with a lensoidal section. Planar cross-stratification (McKee and Weir, 1953) has not been recorded. The average dimensions for these infills is in the order of two metres cross-sectional length and 0.5 metres depth. Although examples are exposed in many localities, the availability of data is limited as the structure is usually seen only in two dimensional section and foresets are never well developed. Sections at Burrells Quarry (NY677180), however, are sufficiently well exposed to enable a west-south-west to east-north-east (bearing 041 to 076 degrees) axial palaeocurrent to be inferred in conjunction with measurements of imbricated clast fabrics which concurs with the generalised apparent derivation direction recorded from elsewhere in this area of the Eden Valley. Although the limited amount of available data is insufficient for systematic analysis, field observation indicates a dominantly south to south-westerly derivation for the extreme southern part of the Eden Valley and thus indicates the existence of at least two possible dispersal systems.

SEDIMENTARY TEXTURES

Clast size

Clast size is perhaps the most important single measurable variable of the brockram sequences. The relationship between maximum clast size and bed thickness (see Figure 1.5) indicates that the most useful parameter is the size of the largest particle occurring within a bed which may also provide some measure of the competence of the transporting medium (Blatt *et al.*, 1972). Other textural considerations have also shown the significance of this parameter in that sorting and grading is often restricted to the coarsest grade material only. In the majority of brockram exposures, the difficulty of extracting clasts precludes the possibility of rigorous size determinations and the measurement adopted has been that of the maximum projection plane seen in vertical section. In sections parallel to the local transport direction, this value is likely to be a close approximation to the true long dimension of the clasts as the majority of non-spherical fragments tend to be orientated with a-b planes parallel or at a shallow angle to bedding. Where possible, however, accurate measurements of the a, b and c planes have been recorded for the purposes of detailed clast description.

From field observation of the occurrence of the various clast sizes, a classification into very coarse (> 25 cms), coarse (> 6 cms < 25 cms), medium (> 2 cms < 6 cms) and fine (> 0.2 cms < 2 cms) breccia has been adopted. This scheme loosely corresponds to the gravel grades of the Udden (1898 in Blatt et al., 1972) - Wentworth (1922) scale although the upper limit of the fine breccia grade used herein falls within the pebble grade of these authors and granule grade material is not distinguished. A less subjective approach was not adopted because of the inherent difficulties involved in the measurement of clasts within well cemented sequences and an accuracy of greater than 0.5 cms is not claimed for the larger clasts.

Clast shape

Estimates of clast roundness and sphericity have been obtained for limestone clasts from several localities in the Eden Valley using the methods outlined by Krumbein (1941). Determinations of clast roundness were restricted to a comparison of selected clasts with the visual estimation chart published in Krumbein and Sloss (1963, p.111). Common values lie in the range of 0.3 to 0.9, corresponding to the angular to well-rounded classes of Powers (1953), with average rounding being in the range of 0.5 to 0.7 (sub-angular to rounded). Systematic variations in this parameter have not been observed and clast roundness apparently varies independently of size, excepting a slightly higher value of mean grain roundness for the fine breccia grade (0.53 averaged for 68 samples) as compared to the other clast categories (0.46 averaged for 95 samples distributed in three size classes). Roundness values also appear to vary randomly within individual beds although thinner beds may be expected to have higher average rounding values because of the relationship between particle size and bed thickness.

A limited number of determinations of clast sphericity has been made from comparison of the b/a and c/b dimension ratios measured from 31 separated limestone clasts. The data are presented in Figure 1.6 and samples are grouped according to their size classes. No samples from the very coarse grade have been analysed because of extraction problems. Mean sphericity (Ψ) for all samples is 0.68 but seems to be partly dependent on clast size. Determinations for the fine, medium and coarse grades have given average values of 0.61, 0.70 and 0.75 respectively, indicating sphericity to increase slightly with increasing clast size. It is important to note, however, that these figures are based on a very small sample which, in the case of individual size grades, is less than the minimum sample number of 25 suggested by Krumbein (1941) to afford a reasonable estimate of average sphericity. In addition, determinations for the fine breccia grade are subject to a sample bias inasmuch as low sphericity clasts with high a/c and b/c ratios tend to fracture during extraction and the data are correspondingly skewed towards samples with higher sphericities.

The data plotted on Figure 1.6a allow direct correlation with the particle shape classification scheme of Zingg (1935 in Blatt et al., 1972). The boundaries for the four shape classes have been incorporated into Figure

Figure 1.6. Sphericity data for separated limestone clasts plotted on: a. Krumbein (1941) sphericity chart incorporating Zingg diagram, b. Sneed and Folk (1958) sphericity and form diagram.

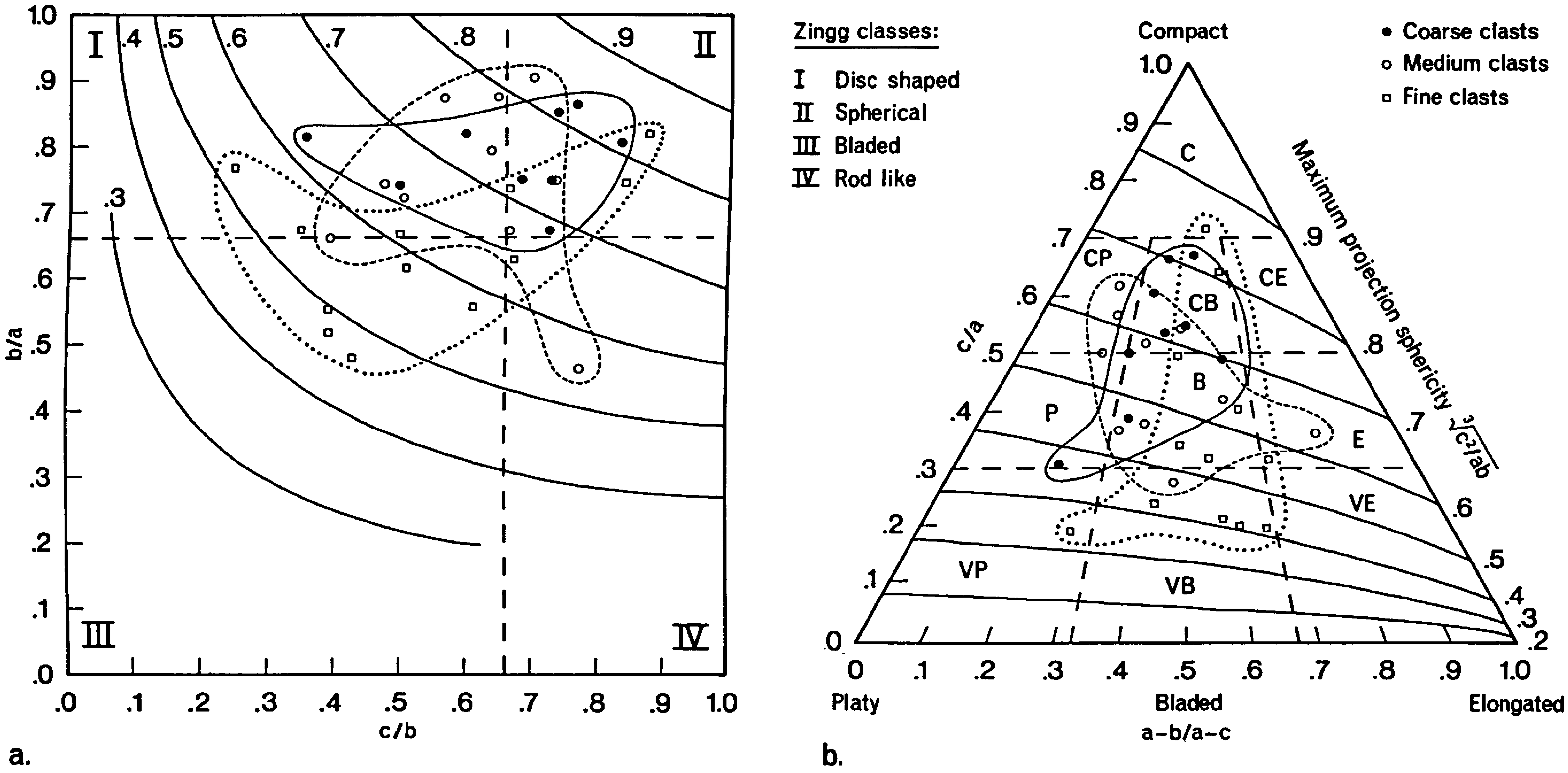
Zingg classes:
I Disc shaped
II Spherical
III Bladed
IV Rod like
b/a
c/b
a.
Compact
Platy
Bladed
Elongated
a−b/a−c
c/a
Maximum projection sphericity $\sqrt[3]{c^2/ab}$
• Coarse clasts
∘ Medium clasts
▫ Fine clasts
C
CP
CB
CE
P
B
E
VP
VB
VE
b.

Figure 1.6

1.6a and indicate the majority of brockram clasts to fall within the disc, spherical and bladed categories (classes I, II and III). According to the available data, rod shaped clasts (class IV) are uncommon. Comparison with the shape classification scheme of Sneed and Folk (1958) indicates very bladed, bladed and compact bladed shapes to predominate with only relatively few clasts falling within the elongated and platy classes of these authors (Figure 1.6b).

Associated with the tendency for mean sphericity values to increase with clast size, the distribution of the clast shapes is correspondingly skewed towards bladed shapes in the fine breccia grade and disc and spherical shapes in the medium and coarse grade. As with roundness, therefore, there is an apparent shape distinction within the fine breccia grade which similarly indicates a degree of dependence on clast size. Definite and systematic variations, however, are not apparent from the limited amount of data available.

Fabric

The most commonly developed depositional fabric recognised from the brockrams is a preferential orientation of non-spherical clasts in which the maximum projection planes are aligned parallel to, or nearly parallel to, bedding. In fewer examples, a distinct imbrication of the clasts is developed in which pebble long axes are inclined upcurrent at an angle of 10 to 40 degrees to bedding, conforming to the contact imbrication style of Laming (1966). Accurate measurement of this fabric is subject to the same constraints as cross-bedding although apparent dip directions can be plotted from the majority of exposures and generally concur with a south-westerly to west-south-westerly derivation direction.

Particle size distribution

Quantitative analysis of particle size distribution within the brockrams is inhibited by the large range of sizes present within individual beds (which precludes measurement by thin section techniques) and by the unsuitability of mechanical methods for lithified rock sequences. Determinations of sorting are thus restricted to visual estimates and a high degree of accuracy is not claimed. In addition, accurate assessment of original depositional textures in coarse grained deposits is also subject to the recognition of diagenetic effects such as the mechanical infiltration of finer grain sizes into the interstitial voids (see Chapter Two and Waugh, 1967, p.251-270 for detailed descriptions) which may result in extensive textural modification (e.g. Walker et al., 1978). Studies of comparable deposits from contemporary environments (e.g. Bull, 1964a, 1972), however, indicate common mean sorting coefficients (Trask) to lie in the range of 1.0 to 4.7 approximating to the moderately to extremely poorly sorted divisions of Folk (1968), although values of 0.48 (well sorted) have been recorded.

As a generalisation, sorting and grading within the brockrams may be regarded as poorly developed and most beds probably correspond to the moderately to extremely poorly sorted divisions of Folk (1968). In extreme examples, disordered clasts ranging in size from 0.2 to in excess of 100 centimetres are supported by a matrix of sand and silt grade material (as at Cuddling Hole, NY673185, at the site of log 37 in Figure 1.7) although there is usually some degree of size consistency of the coarsest clasts and matrix supported textures are uncommon. Where grading occurs within a single bed it is, similarly, often restricted to the coarsest grade material and is most commonly related to the presence of a coarser base. As with clast textures, there is little systematic relationship between particle size and rock fabric although there is an apparent general increase in sedimentary organisation with decreasing clast size.

FACIES ANALYSIS OF THE BROCKRAMS

From the above considerations, five distinct and one transitional breccia facies types can be recognised on the criteria of clast size, sorting and pebble fabric. Considerations of grain parameters such as shape, roundness and sphericity have been excluded on the basis that definite systematic variations have not been observed. A less generalised approach according to lithofacies associations (e.g. the depositional assemblages described by Steel, 1974a) is not considered feasible for the brockrams as a result of the overall uniform nature of the beds.

Facies A. Very coarse (clasts $>$25 cms), extremely poorly sorted breccia. Common occurrence as coarser basal horizons to facies B or C (see below) but also occurs as unstratified channel-fills and as matrix supported beds with a sheet-like form. Pebble fabrics are not extensively developed and are restricted to a bedding parallel alignment of elongate clasts.

Facies B. Coarse (clasts $>$ 6 cms $<$ 25 cms), generally poorly stratified breccia. Bedding usually sheet-like and discordant on underlying units. Clast frameworks predominate but matrix supported textures are occasionally developed. Two sub-facies recognisable according to the development of pebble fabrics and degree of sorting:

B1. Extremely poorly sorted breccia with no recognisable fabric.

B2. Poorly-moderately sorted breccia, often with bedding aligned or imbricated clasts. Coarsest clasts may be normally graded.

Facies C. Medium breccia (clasts $>$2 cms $<$6 cms) with clast supported framework. Bedding usually well developed, often with pronounced internal stratification. Occasional occurrence as cross-bedded scour infills. Two sub-facies recognisable:

C1. Extremely poorly-poorly sorted breccia with no recognisable fabric or ill-defined bedding alignment of clasts. Rare upward-fining of coarsest clasts.

C2. Moderately-moderately well sorted breccia, commonly with bedding aligned and/or imbricated clasts. Grading may be well developed. May occur above type B2 breccia.

Facies C/D Alternating, well-bedded medium-fine (facies E) breccia, usually with aligned clasts sorted into distinct bands. Transitional from facies C to facies D.

Facies D. Fine breccia (clasts > 0.2 cms < 2 cms) with clast supported framework. Bedding, internal stratification and pebble fabrics usually well developed. May occur above type C2 or facies D brockram and rarely above type B2 brockram.

Facies E. Coarse sandstone (grains < 0.2 cms) with occasional fine breccia horizons. Usual occurrence above facies E but may occur as intercalations in facies C, D or E.

For the most part, the different breccia facies types tend to be randomly interbedded and show no definite associations. Occasionally, however, fining-upward cyclothems (in the sense of Allen, 1970a) can be recognised where a coarse member is overlain by a single or two successively less coarse members. The most frequently observed sequence is that of a B-C or C-D association although B-C-D or C-D-E assemblages are also recorded.

DESCRIPTION OF THE EXPOSURES

The majority of Edenside exposures occur in two areas which are centred in the region of Burrells and Appleby towards the north-western limit of the breccias and Kirkby Stephen in the extreme southern part of the valley (see

Figure 1.1). These two areas correspond to the thickest development of the breccia (see Figure 1.3) with the latter including exposures of both the Penrith and Stenkrith Brockrams. Between these two areas, exposures are limited to one or two isolated sections in the River Eden (e.g. Bermer Scar, NY745146) and its tributaries (e.g. Jeremy Gill). It is not proposed to include precise descriptions of individual localities as few of the exposures differ significantly in sedimentological detail and predominantly comprise well-bedded but otherwise haphazard associations of the brockram facies types described previously. However, generalised descriptions of the main areas in which brockram sequences are exposed and representative breccia facies logs of all the principal exposures are included in the following sections.

The Burrells-Appleby region

The exposures of the Burrells-Appleby region occur within the main development of the lower, Penrith Brockram where the breccias have been suggested to reach a maximum thickness of the order of 150 metres (Arthurton et al., 1978). The principal exposure of the area occurs between Burrells and Hoff where the brockrams banked against the underlying Carboniferous rocks are exposed intermittently from Bandley Bridge (NY672189) in the north-west to near Lookingflatt (NY680176) in the south-east; a total distance of 1.4 kilometres. In the north-west, the brockrams overlie reddened sandstones of the Millstone Grit Series which are exposed in Hoff Beck at NY673183, whereas in the south-east, reddened Great Strickland Limestone (Waugh, 1967) is exposed in a disused lime quarry at NY679175. Representative facies logs of the brockram sequences recorded from this section are reproduced in Figures 1.7 and 1.8. Although extensive exposures occur in a series of former river

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Figure 1.7. Representative facies logs of the brockram sequences exposed from Bandley Bridge (NY672189) to Burrells Quarry (NY677180; logs 14, 17 and 20). Hachured lines indicate approximate correlations.

Figure 1.8. Representative facies logs of the brockram sequences exposed from near Haverflatts (NY678179) to near Lookingflatt (NY680176). See Figure 1.7 for legend.


North-west
South-east
46&47
Bandley Bridge
NY672189
42
37
34
Cuddling Hole
30
24
NY674183
20
17
14
Burrells Quarry
NY677180
3
2
1
0
Metres
?
Facies
A
B
C
D/E

Figure 1.7


North-west
South-east
12
Near Haverflatts
NY678179
11
10
9
8
7
Channel-fill
breccia ?
6
5
4
1
Near Lookingflatt
NY680176
0
1
2
3
4
Metres

Figure 1.8

bluffs in the south-eastern part of the section, north of Rowley Wood (centred on NY679177) and laterally continuous crags can be traced for 100 metres or so in the north-western part of the section, centred on Cuddling Hole (NY673185), the only extensive exposure in the central part occurs in a disused quarry, 50 metres north-east of Burrells Gate House, at NY677180. Accurate correlation of the breccias across the unexposed parts of the section is not possible, therefore, and is difficult in the south-eastern area due to the nature of the outcrop. It is clear, however, that all exposures lie very close to the sub-Permian unconformity, the line of which more or less parallels that of the exposures, a short distance to the south-west and thus occur at approximately the same stratigraphic level. Bedding attitudes within the breccias (which generally dip towards the north-east or east at an angle of approximately ten degrees) may thus be expected to be a response to an exposure orientation which is parallel to the basin margin.

With the exception of several coarse breccia units displaying markedly erosive (channel-like) contacts (see Figure 1.8, logs 4, 6 and 7 and logs 8 to 12), the greater proportion of the beds demonstrates a sheet-like form in which successive beds are separated by gently scoured surfaces. When traced laterally, however, specific beds may ultimately be wedged out against overlying units either as a result of truncation by the succeeding bed or as a product of the original depositional form of the bed. Occasionally, as in Burrells Quarry, conspicuous internal, curved erosion surfaces may be developed, conforming to a transitional type of bedding, as described previously. In addition, lensoidal scour structures with crudely cross-bedded medium to fine breccia infills, are also relatively common at this locality. Imbricate and bedding parallel clast fabrics are common throughout these exposures with most of the rock units demonstrating some degree of upward-fining of the coarsest detrital components. Convincing matrix supported textures are rare and are developed only in some of the more westerly exposures towards Cuddling

Hole (Figure 1.7, at the site of log 37, for example). Units of this type are characterised by a sheet-like bedding style, an exceedingly coarse clast component (in excess of 100 centimetres) and a very low degree of sorting but are often overlain by a thin (10 to 20 centimetres) sandstone or fine breccia horizon.

The only other laterally extensive exposures of the area include a small disused working to the north-east of the main exposures at Burrells (NY678183; Figure 1.9) where approximately five metres of breccia is exposed for a distance of 50 metres and Thistley Hill Quarry (NY678205; Figure 1.10) where a series of breccias overlie reddened sandstones in a faulted inlier of Carboniferous to the west of Appleby. With the exception of the cross-bedded and horizontally stratified breccias occurring in the south-eastern part of the Burrells exposure (described previously), the sequences exposed at both localities comprise sheet-like beds in which there is marked absence of prominent erosion surfaces. Small scale bedding scours and internal scour surfaces are of common occurrence, however, and are often associated with poorly developed festoon-shaped cross-beds. Matrix supported textures are not recorded from either locality. Small exposures also occur at Big Clinch (NY664185; Figure 1.11a) and Barwise Hall (NY661177) but are of too limited an extent for bedding styles to be determined accurately. The exposure at Big Clinch is notable, however, in comprising approximately thirteen fining-upward cycles within a vertical thickness of four metres, the coarsest members of which are characterised by a high proportion of disc-shaped clasts with a strong bedding parallel alignment.

Figure 1.9. Representative facies logs of the brockram sequences exposed at NY678183.

Figure 1.10. Representative facies logs of the brockram sequences exposed in Thistley Hill Quarry (NY678205). See Figure 1.9 for legend.

Figure 1.11. Representative facies logs of the brockram sequences exposed at: a. Big Clinch (NY664185), b. Jeremy Gill (NY694170) and c. Bermer Scar (NY745146). See Figure 1.9 for legend.


North-west
South-east
1
2
3
4
3
2
1
0
Metres
Braided channel-fill sequence (see inset)
Facies
A
B
C
D/E
0
Metres
2

Figure 1.9


South
North
1
2
3
4
Metres
0
1
2
3
4
Sandstone-filled vertical
fracture in lower breccia unit

Figure 1.10

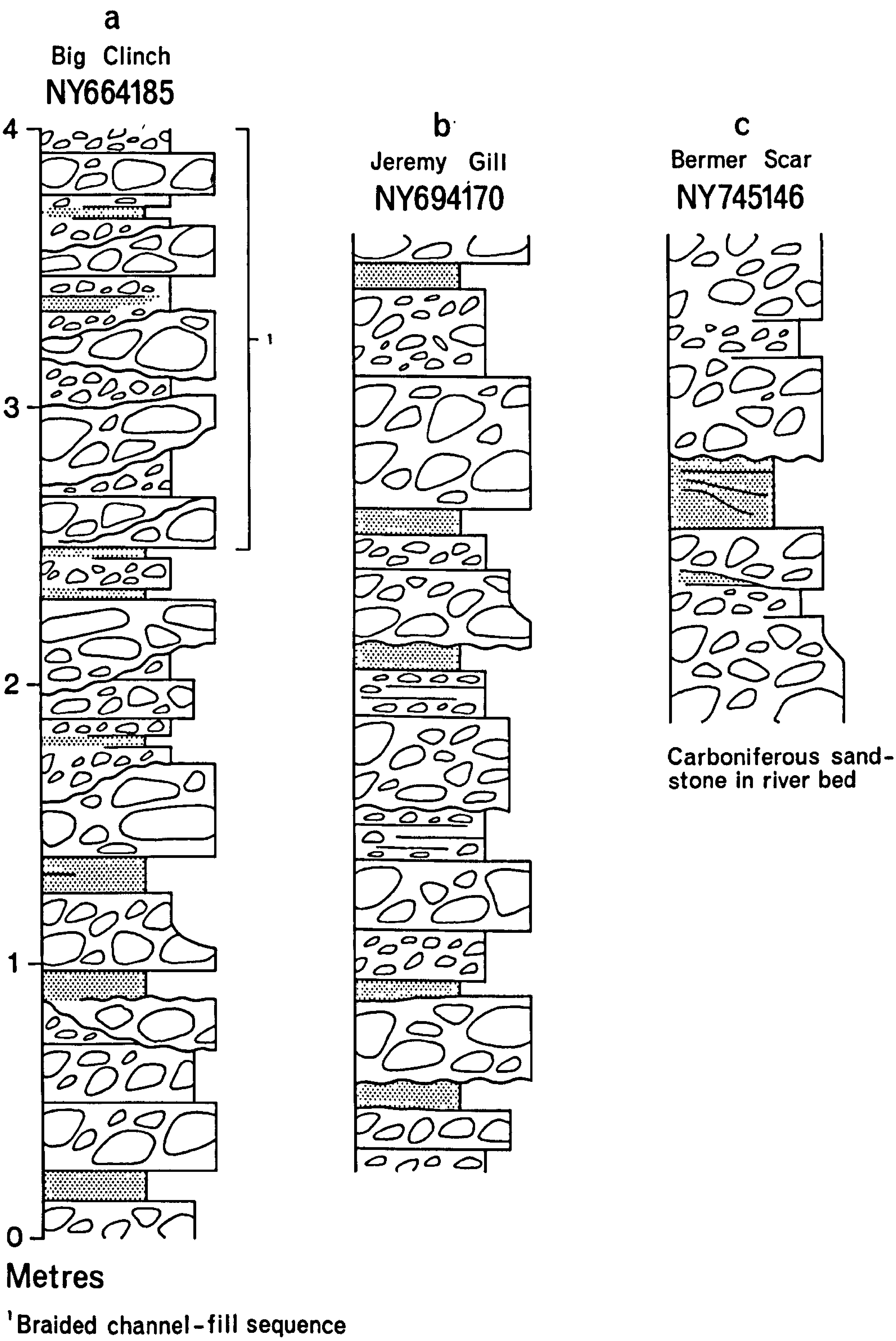
a
Big Clinch
NY664185
b
Jeremy Gill
NY694170
c
Bermer Scar
NY745146
4
3
2
1
0
1
Metres
Carboniferous sand-stone in river bed
[1]Braided channel-fill sequence

Figure 1.11

Exposures in the area between Burrells and Kirkby Stephen are limited to a few isolated and laterally inextensive sections. Facies logs of the sequences recorded in Jeremy Gill and at Bermer Scar are reproduced in Figure 1.11b and c respectively.

The Kirkby Stephen region

Within the southernmost part of the Eden Valley, outcrops of Penrith Brockram are limited to a few poorly exposed sections in the banks and channel of the River Eden although the breccias were seen formerly in the railway cutting at Kirkby Stephen Low Station (Harkness, 1862, p.206). The most extensive section occurs immediately downstream from Frank's Bridge (NY777088) where approximately three metres of dolomitised breccia is exposed in the eastern bank of the river and can be traced for 200 metres or so towards the north from intermittent exposure in the channel of the Eden. Small exposures are also seen on the banks of the Eden at Turn Wheel (NY771065) and at Horn Bank (NY772063) where the breccias are banked against Carboniferous Limestone. The Penrith Brockram is overlapped by the Stenkrith Brockram in the region of Turn Wheel and is absent at Thringill (NY775062) where the Stenkrith Brockram rests directly on Carboniferous Limestone (Burgess, 1965).

The Stenkrith Brockram is well exposed in the area. A complete section is exposed in the River Eden from High Stenkrith (NY771073) to Stenkrith Park at NY775075 (Figure 1.12), where a total thickness of 18.8 metres of breccia has been recorded. The lowermost beds of the Stenkrith Brockram succeed the underlying fine sandstones of the Eden Shales through approximately 1.5 metres of medium to coarse sandstone incorporating isolated clasts and horizons of fine breccia. The medium breccias forming the base of the brockram, however,

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Figure 1.12. Representative facies logs of the Stenkrith Brockram section exposed in the River Eden from High Stenkrith (NY771073) to Stenkrith Park (NY775075). See Figure 1.7 for legend.


South-west
North-east
1
High Stenkrith
NY771073
2
4
5
6
7
9
10
NY775075
Stenkrith Park
11,12 & 13
Metres
0
1
2
3
Eden Shales

Figure 1.12

represent a marked and relatively abrupt increase in clast size. The upper contact of the Stenkrith Brockram is not seen but probably occurs at the top of the highest breccia bed exposed. Burgess (1965) records that three metres of St. Bees Shales (the Eden Shales) are present at this locality and overlie the brockrams. This horizon was not located by the author, however, and the first beds recorded above the brockram are the brick-red sandstones of the Sherwood Sandstone Group which occur approximately 2.5 metres stratigraphically above the highest exposed breccia bed.

The total thickness of the brockram at this locality comprises laterally persistent (sheet-like) beds which vary in thickness from a few centimetres to in excess of one metre. Large scale erosional features are generally absent from the succession although minor scouring and lateral wedging of the beds is common and specific beds may be wedged out over a distance of ten metres or so. The maximum depth of scour is probably only in the order of 30 centimetres and it is considered that the wedging of beds at this locality is partly a primary depositional feature reflecting the aerial extent of particular episodes of sedimentation. Upward-fining of the clasts occurs within single beds and as assemblages of two or more successively less coarse beds. Many of the beds exhibit a strong preferential orientation of the clasts, both parallel to and at an angle (imbricated) to bedding. One of the notable features of this locality is the homogeneity of the rock which, cement and clasts alike, has been cut into a series of pot-holes by the River Eden. A similar style of erosion has been recorded by Glennie (1970, p.34) from petrographically comparable wadi deposits from the Oman Mountains.

A complete sequence of the Stenkrith Brockram is also seen 1.5 kilometres north-east of Stenkrith in the River Eden upstream of Frank's Bridge (NY779087). At this locality, however, the sequence has thinned to less than five metres thickness and is overlain by a medium grained, yellow dolomitic sandstone which passes up into red shaley sandstones interpreted as the Eden Shales (Burgess, 1965).

Exposures of the Stenkrith Brockram also occur at Nateby between the River Eden and the B6259 where 14.7 metres of breccia are recorded from intermittent exposure and in a disused quarry to the west of Stenkrith Hill (NY773077). As with the exposures of this Brockram in the River Eden, the majority of beds possess a pronounced sheet-like form in which small scale scour structures (with a maximum depth in the range of 10 to 20 centimetres) and curved internal erosion surfaces are relatively common in occurrence.

The marked thinning of the brockram from Stenkrith to Frank's Bridge is thus well established in the Kirkby Stephen area and indicates a general derivation direction from the south which accords with the south-westerly directions determined from cross-bedding and clast imbrication.

THE WEST CUMBRIAN BROCKRAM

With the exception of the basal breccia exposed above the Westphalian Whitehaven Sandstone in Saltom Bay (NX959159), the brockrams of western Cumbria are generally less well exposed than those of Edenside and are seen only in a few isolated stream sections to the north and south-east of Egremont. Where seen in the field, e.g. Kirk Beck near Haile (NY038099) and the River Calder at NY062079, the sedimentological characteristics of the sequences are essentially identical to those of the Eden Valley although the lateral extent of the majority of these exposures is insufficient to enable individual units to be traced for more than a few tens of metres. In consequence, the geometry of the beds cannot be assessed in the same detail as for the Eden Valley sequences although it is clear that the dominant bedding styles conform with those described previously.

Petrographically, the western Cumbrian brockrams have a much more varied clast assemblage than those of the Eden Valley and predominantly comprise Borrowdale Volcanic and Skiddaw Slate fragments locally derived from the fells flanking the eastern margin of the depositional basin. Sedimentary structures such as clast imbrication generally concur with this derivation direction and suggest currents flowing towards the west and south-west for exposures in the Egremont-Haile area. Unlike the Eden Valley, Carboniferous limestone fragments form only a minor percentage of the total clast component and are only common at the base of the formation where the breccias unconformably overlie the Carboniferous. Rare clasts of Eskdale Granite have also been recorded by Trotter _et al_. (1936). The effects of diagenesis are extensive within these sequences (see Chapter Two) and have led to the _in situ_ modification of many original depositional textures and the formation of large quantities of interstitial matrix which were not present in the original sediment. In particular, bedding is often partially obscured (except where exposures have been extensively weathered) and the original clast supported texture has often been superseded by a diagenetic matrix supported texture. Where unweathered core samples have been available for study (see below), bedding is difficult to distinguish except where beds can be recognised by obvious differences in clast size. Despite extensive post-depositional alteration, however, sedimentary structures such as preferential clast orientation and textures such as an upward-fining of the coarser clasts within individual beds are still recognisable and indicate the mode of formation for these deposits to be comparable with the Eden Valley brockrams.

As well as a complex diagenetic post-depositional history, the structural setting of the western Cumbrian brockrams also contrasts with that of the Eden Valley in that the Carboniferous and Permo-Triassic sequences are displaced by a series of faults with a north-north-west to south-south-east trend and a less extensively developed series with a north-east to south-west trend. Although these faults are not mappable at the surface over the main development of the Permo-Triassic strata south and west of Egremont, they are known to affect these sequences from the underground haematite workings of the

Beckermet-Egremont area (see Figure 1.17) and can be seen to displace the breccias in exposures in the west bank of the River Calder, six kilometres south-east of Egremont at NY062079. At this locality, the brockrams are successively downthrown to the north-east by a series of north-north-west to south-south-east trending faults.

As many modern desert basins are recorded to be fault bounded (e.g. Collinson, 1978b), it is possible that intra-Permian faults may also be present in this area but not yet recognised. A probable example of contemporary faulting has been recorded, however, from the interbedded breccia and reworked Penrith Sandstone sequences of the Eden Valley (see Chapter Three).

As exposures of the brockram in western Cumbria are not sufficiently informative or extensive to warrant a detailed sedimentological study, the main aim of research in the region has been to examine the petrography and diagenesis of the breccias from core samples supplied by the British Steel Corporation (see Chapter Two). Until 1978, British Steel had a borehole drilling programme in the search for haematite ore bodies in the Carboniferous Limestone underlying the brockrams which was centred on St. John Beckermet, five kilometres south of Egremont. Data from approximately 600 boreholes (see Appendix One) and cored samples from eight of the more recently investigated sites have been made available for detailed study and have formed the basis of a more regional sedimentological study than is possible for the brockrams of the Eden Valley. The large amount of data made available have enabled the construction of computer based isopachyte and palaeotopographical maps and perspective block diagrams for the area around Beckermet using the synagraphic mapping (symap) and symvu programmes produced at the Laboratory for Computer Graphics and Spatial Analysis at Harvard University and modified at the University of Hull for use on ICL 1900 series computers.

Petrographic studies of core samples have also led to the recognition of sequential accumulations of carbonate within finer, sandstone and siltstone horizons which interdigitate with the breccias in more westerly parts of the area and which have been interpreted as caliche soil horizons (see Chapter Four).

The nature of the western Cumbrian brockrams

Over much of the mining district of western Cumbria (see Figure 1.21, for example), the brockrams comprise a continuous succession of poorly sorted breccias with occasional thin sandstone horizons which interdigitate with and are overlain by the St. Bees Sandstone. The contact between the breccia and sandstone is usually fairly abrupt but may be gradational through five to ten metres of interbedded breccias and sandstones. Even where the contact is abrupt, the lowermost St. Bees Sandstone frequently contains thin horizons of gritty Borrowdale Volcanic and Skiddaw Slate clasts. The lowermost two metres or so of the brockrams is distinguished everywhere by being noticeably coarser than the overlying breccias and by containing a significant proportion of large, blocky limestone clasts. The size of these clasts cannot be measured from borehole samples as the diameter of the cores (seven centimetres) is less than that of the larger clasts but long dimensions in excess of 25 centimetres are not uncommon from the exposures of the area. Although, for the most part, clast size (and angularity) varies indiscriminately from bed to bed throughout the remainder of the succession (excepting where a crude fining-upward is evidenced by a relatively finer bed overlying a coarser bed), there is an apparent overall upward decrease in the occurrence of coarse horizons towards the top of the sequence. Several boreholes (e.g. Surface Borehole 296; NY01841035, see Figure 1.13 and SBH 306; NY01970951, see Appendix Two) have penetrated a coarser horizon at the top of the sequence and Trotter _et al._ (1936) have recorded the occurrence of coarse horizons in the upper part of the brockrams elsewhere in the area.

In more westerly parts of the mining area, as at Carlton (NY018094), finer sandstone and siltstone horizons are much more common in occurrence within the brockram sequences and gradually increase in thickness and persistence until the breccias are restricted to a thin bed which is overlain, in these areas, by the St. Bees Shales. As with areas to the east, however, the top of the brockrams may be gradational and thin horizons of gritty clasts may occur within the St. Bees Shales. This facies change is accompanied by a general decrease in the occurrence of coarse beds above the basal breccia and is illustrated in Figure 1.13. In the absence of evidence indicating a formation related to the breccias, the finer horizons are tentatively regarded as belonging to the St. Bees Series which thus interdigitates with the brockrams in the western part of the depositional area.

In the area to the west of Egremont, near Hagget End (NY005105), boreholes have proved a thin dolomitic limestone within the lower parts of the breccia sequence (e.g. SBH 756; NY00751015, SBH 757; NY00551032 and SBH 758; NY00741039) which is regarded by Arthurton and Hemingway (1972) as part of the Saltom Cycle of the St. Bees Evaporites and thus corresponds to the Magnesian Limestone of Smith (1924). This horizon is also recorded from boreholes further to the west (e.g. SBH 287; NX99061081 and SBH 288; NX98861093, near Watson Hill; see Figure 4.2) where it overlies a thin (12 metres) development of the brockrams and is overlain by the succeeding cycles of the St. Bees Evaporites and the St. Bees Shales. This succession corresponds to that exposed on the coast in Saltom Bay. Although, therefore, in the extreme west, the junction between the brockrams and the St. Bees Shales is marked by the evaporites, the Magnesian Limestone occurs within the breccias in the region of Hagget End which are thus, in part, the

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Figure 1.13. East to west facies changes across the northern part of the Beckermet area. SBH 305 (NY00061007) is representative of the thin breccia sequences occurring in westerly areas, SBH 304 (NY01770963) of sandstone - breccia sequences and SBH 296 (NY01841035) of continuous breccia sequences of the eastern areas. See Appendix Two for legend.

¹Undifferentiated

Figure 1.13

stratigraphic equivalent to the upper cycles of the St. Bees Evaporites and lower part of the St. Bees Shales. These breccias are regarded as part of the brockrams and not as coarser horizons within overlying beds.

The breccia sequence exposed in Saltom Bay (NX959160) is illustrated in Figure 1.14. At this locality, the breccias rest with unconformity on, and infill joints and hollows in, an irregular surface of Westphalian Whitehaven Sandstone. Bedding and other sedimentary structures such as preferential clast orientation are well developed in this section and indicate a fluvial mode of formation. The occurrence of limestone clasts (which may be unaltered but are mostly dolomitised) and Borrowdale Volcanic fragments suggests an easterly derivation which is in accordance with the north-east to south-west palaeocurrent direction measured from cross-bedding. However, the high percentage of Carboniferous sandstone clasts at this locality and the widespread occurrence of the basal breccia, which has been recorded in every unfaulted borehole, suggests that a fluviatile origin over the whole of the area is, perhaps, unlikely and that, elsewhere, this horizon may be expected to have originated solely from the mechanical break-up of the underlying Carboniferous strata and thus represent a true lag breccia.

The nature of the sub-Permian surface

The nature of the present sub-Permian surface is shown in Figure 1.15 in the form of a computer constructed (symap) isoline map. Simplified contours representing the height of the surface above mining datum (609.6 metres below ordnance datum) have been reproduced in Figure 1.16 which also shows the

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Figure 1.14. Representative facies logs of the basal breccia exposed in Saltom Bay (NX959160). See Figure 1.9 for legend.

Figure 1.15. Computer constructed isoline map (symap) showing contours on the sub-Permian surface for the Beckermet mining area. Contour values refer to mining datum, 609.6 metres below ordnance datum.

Figure 1.16. Simplified contours on the sub-Permian surface, reproduced from Figure 1.15.


North
South
1
2
3
Magnesian Limestone
2
1
0
Metres
Whitehaven Sandstone

Figure 1.14

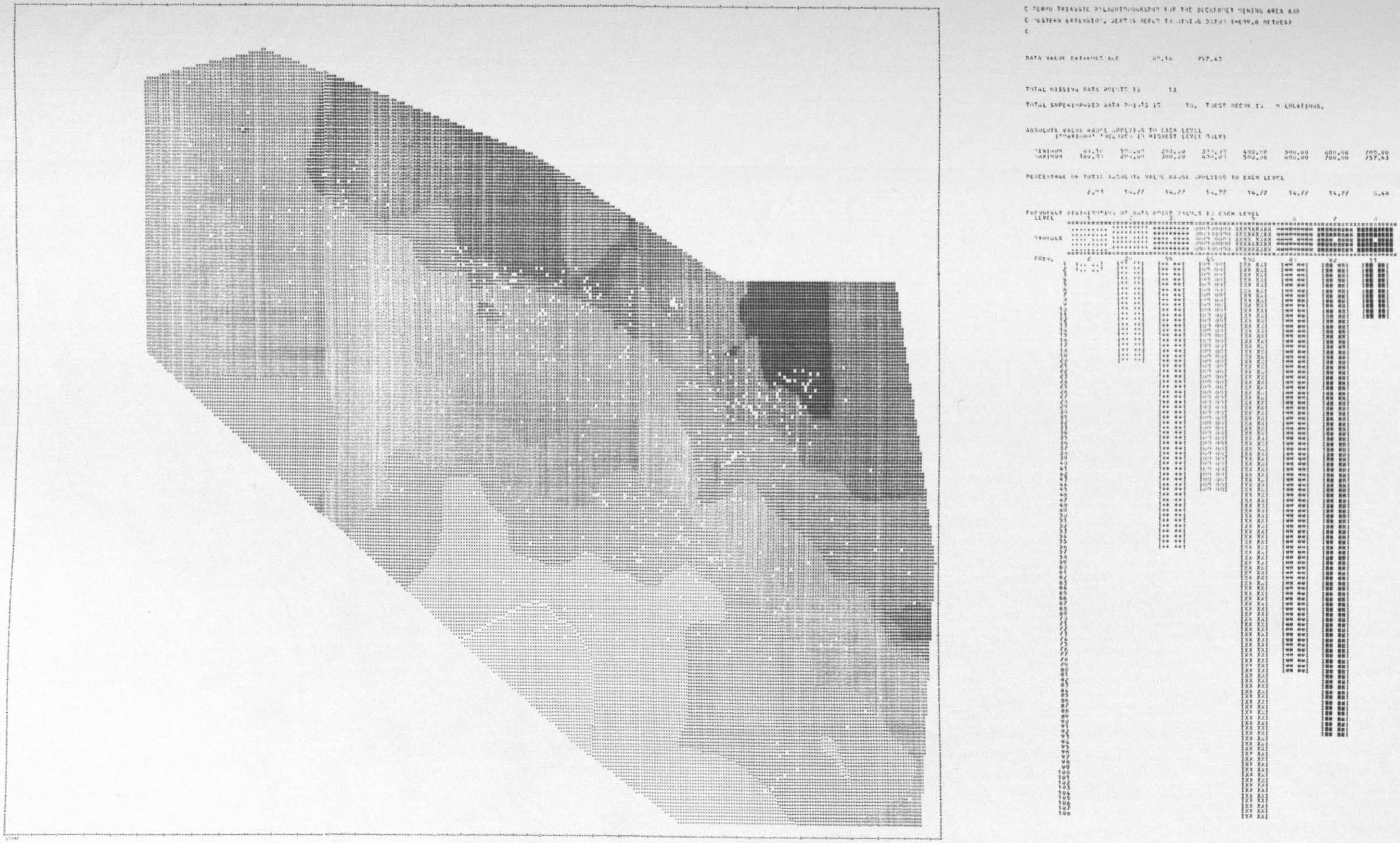
DATA VALUE EXTREMES ARE 757.43
LOCATIONS.
ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
PERCENTAGE OF TOTAL ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
LEVEL
SYMBOLS
FREQ.

Figure 1.15

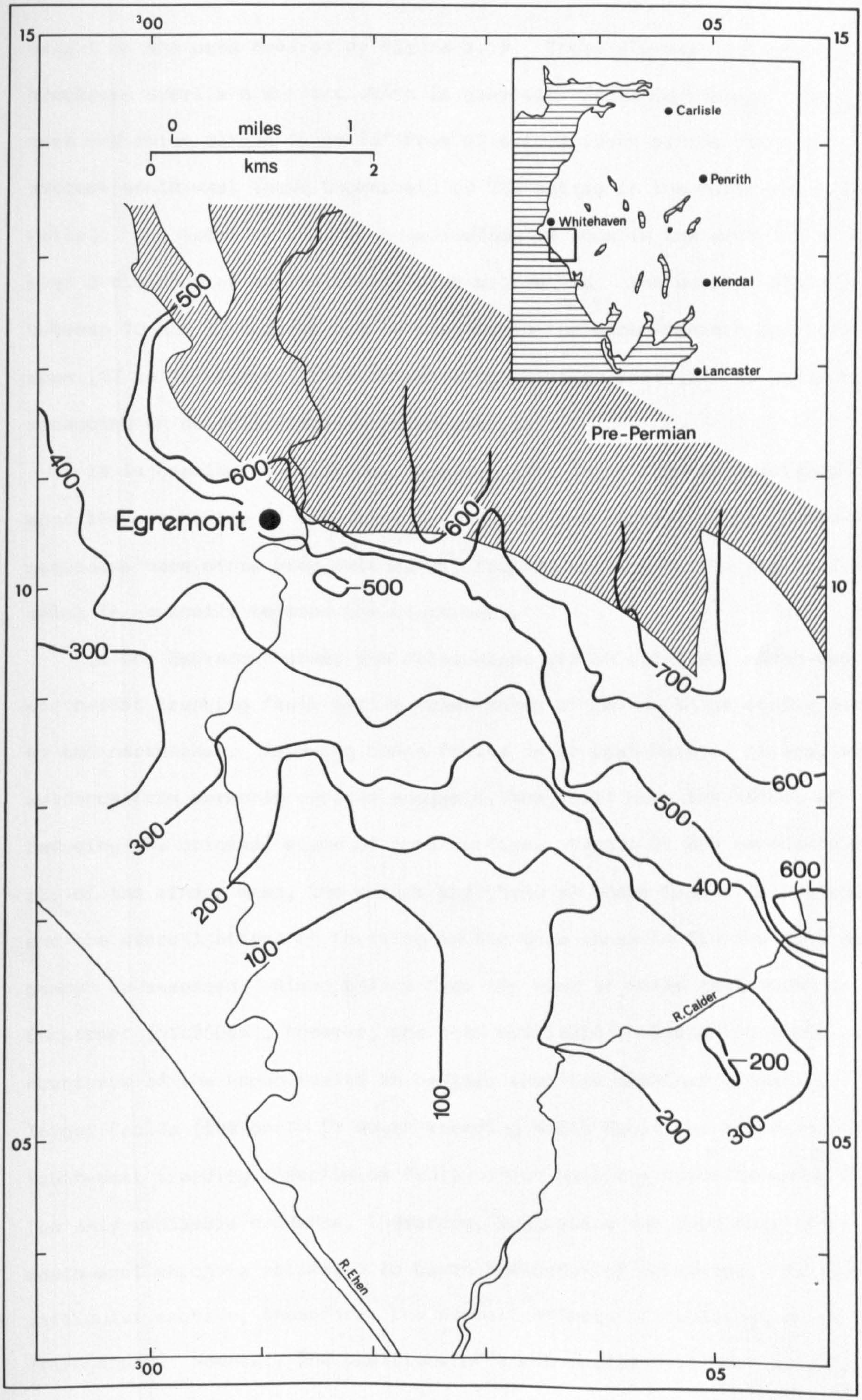
300
05
15
10
05
0 miles 1
0 kms 2
Carlisle
Penrith
Whitehaven
Kendal
Lancaster
Pre-Permian
Egremont
500
600
400
300
200
100
700
R.Calder
R.Ehen

Figure 1.16

extent of the area covered by Figure 1.15. These figures indicate that the brockrams overlie a surface which is generally inclined towards the south-west and which ranges in relief from 80 metres above mining datum in the extreme south-west (near Beckermet) to 700 metres in the north-east (near Haile). The total relief for this surface is thus in the order of 600 metres over a distance of approximately 4.25 kilometres. The average slope is thus between 10 and 12 degrees but is steeper in the north-eastern part of the area (17 to 19 degrees) than in the south-western part (six to seven degrees), producing an overall concave-upward profile.

It is considered unlikely, however, that these figures precisely represent the morphology of the original Permian land surface as the Permo-Triassic sequences have since been extensively faulted and now have a regional tilt which is generally towards the south-west.

In the Beckermet area, the palaeoslope strike parallel, north-west to south-east trending fault series (seen north of Egremont) generally downthrow to the north-east. Assuming these faults to be post-Permian in age, as evidence from borehole records suggests, they will have the effect of reducing the original slope of this surface. Except in the immediate vicinity of the mining area, the extent and throw of these faults is not known and the overall effect of faulting in the area shown in Figures 1.15 and 1.16 cannot be assessed. Along a line from the Head of Haile (NY045095) to Beckermet (NY020065), however, the data available indicate the north-easterly downthrow of the known faults to be less than the combined effects of two larger faults (the north to south trending Haile Fault and the north-west to south-east trending Sheepfields Fault) which have the opposite sense of throw. The only available evidence, therefore, suggests a net downthrow to the south-west which is estimated to be in the order of 50 metres. For this particular section, therefore, the overall effects of faulting may be largely disregarded. However, the positions of known faults have been plotted in

Figure 1.17, superimposed on the contours on the sub-Permian surface and indicate that some of the present topographic features may be related to these faults. The position of the Haile Fault, for example, coincides with a marked break of slope, as does the south-west to north-east trending South Dip Fault. The full degree of fault control on the morphology of this surface, however, has not yet been accurately determined for the Haile-Beckermet area (although it is hoped that a facility for superimposing trend lines on symvu plots will enable a more detailed interpretation of the palaeotopography of this area) and cannot be assessed for the mining area as a whole due to the lack of available data.

The regional dip superimposed on the Permo-Triassic sequences of western Cumbria is generally in the order of five degrees or so towards the south-west, as measured at Saltom Bay although locally, as in the River Calder exposures, dips may be steeper than this figure and attain an angle of 25 to 30 degrees. In the latter example, however, the breccias are successively downfaulted to the north-east by a series of north-west to south-east trending normal faults and may thus be unreliable as an indicator of regional tilt. Whereas bedding within the brockrams may exhibit original depositional dips related to their mode of formation (see Interpretation, this chapter) and the palaeoslope of the depositional basin, bedding in the breccias exposed in Saltom Bay is conformable with that of the overlying Magnesian Limestone and St. Bees Shales and may thus be used as a reliable estimate of regional tilt. If this figure is accepted then corrections for dip indicate the original slope of the sub-Permian surface to be in the order of one to two degrees in the south-western part of the area, steepening to 12 to 14 degrees near Haile. The total relief of the surface would, similarly, be reduced to the range of 250 to 300 metres. The maximum depositional dip of bedding within the

---

Figure 1.17. Position of known faults in the Beckermet mining area and relationship to the configuration of the sub-Permian surface.

³00
05
15
15
0 miles 1
0 kms 2
Carlisle
Penrith
Whitehaven
Kendal
Lancaster
Pre-Permian
500
600
400
Egremont
600
700
500
10
10
300
Carlton Wood F.
Grange Boundary F.
Florence - Winscales F.
John Pickering F.
West drifts F.
Mousegill F.
Haile F.
Haile Moor F.
500
300
Sheepfields F.
400
200
South Dip F.
600
R. Calder
200
100
300
05
05
R. Ehen
³00
05

Figure 1.17

breccias at Saltom Bay would be minimal, therefore, compared with that of breccias deposited further to the east which may be expected, especially in lower parts of the sequence, to adopt an attitude related to the slope of the sub-Permian surface. It is also clear that the westward thinning of the breccias (see Figure 1.20) from a recorded maximum in excess of 120 metres in SBH 258 and SBH 531 (located at NY03710732, near Low Godderthwaite and NY03520981, near Whitehow Head respectively) to two metres at Saltom Bay corresponds to the westerly decreasing palaeoslope.

A further computer technique which has been used to demonstrate the nature of the sub-Permian surface is the symvu programme which enables the construction of perspective block diagrams using data generated by symap. Examples of symvu plots have been reproduced in Figures 1.18 and 1.19. The area covered by these diagrams is outlined in Figure 1.15 and is of approximately 35 square kilometres in extent. The relief of the symvu plots is exaggerated by a factor of 3.5 (approximate) and all of the diagrams have been drawn from an assumed elevation above the horizontal (base) plane (termed altitude) of 50 degrees at a distance of approximately five kilometres (calculated). Mining datum has been taken as the base plane for all symvu plots. Figure 1.18 shows the surface viewed from horizontal angles (azimuths) of 090 and 315 degrees, measured clockwise from azimuth 000 degrees at bearing 180 degrees (south). The azimuths of the symvu plots reproduced in Figure 1.18a and b thus correspond to ~~south~~ west (bearing ~~180~~ 270 degrees) and south-east (bearing 135 degrees) respectively. As the symvu programme has the facility to rotate the data matrix through 360 degrees using the azimuth orientation system, it has proved possible to construct perspective blocks at

---

Figure 1.18. Computer constructed two-point perspective plots (symvu) of the morphology of the sub-Permian surface viewed from azimuths of: a. 090 degrees and b. 315 degrees.

Figure 1.19. Morphology of the sub-Permian surface as shown by symvu stereographic pairs. Azimuths of the plots are 045 and 035 degrees (approximately south-east). Figure 1.19a has been constructed using two-point perspective and 1.19b as an isometric projection.

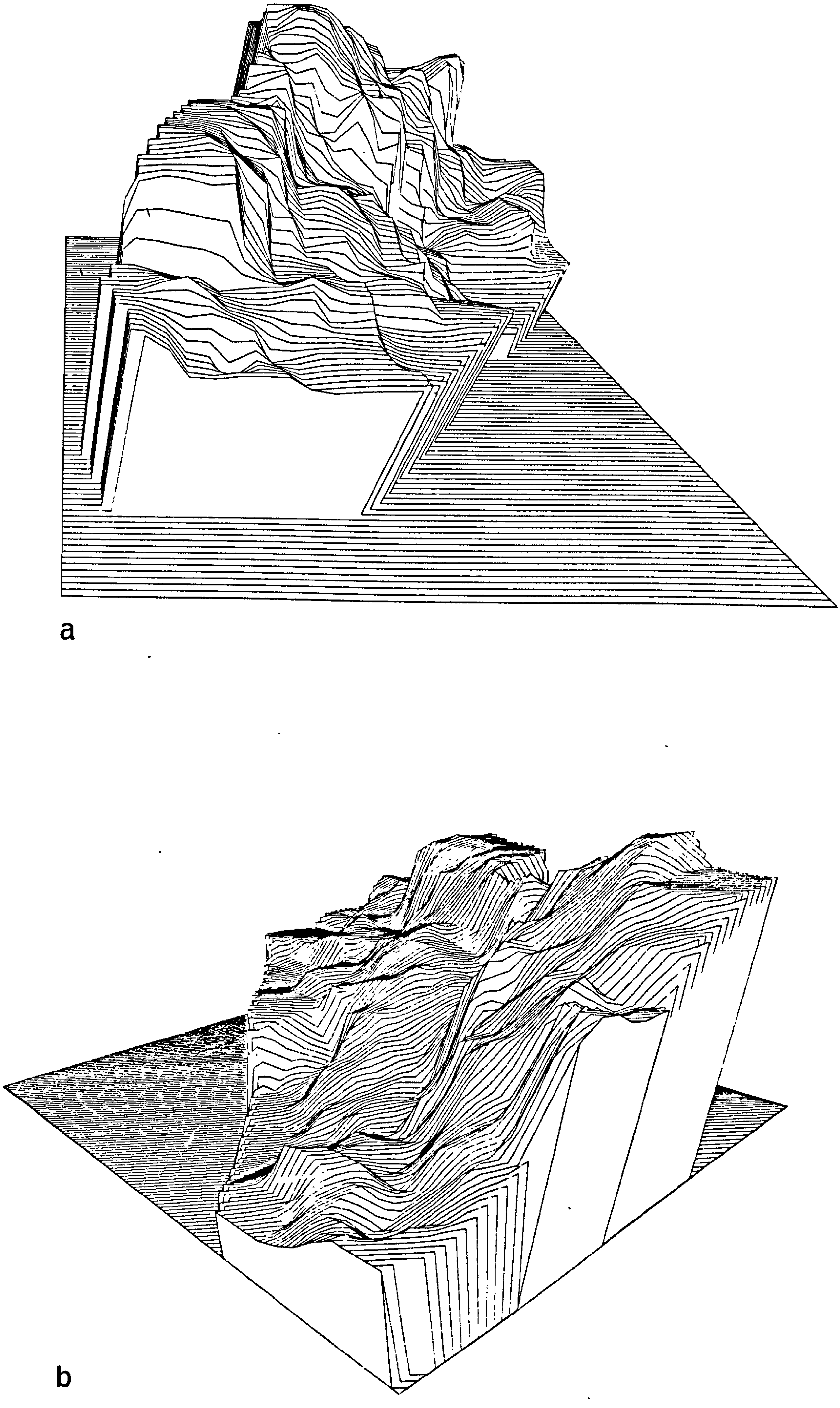

Figure 1.18

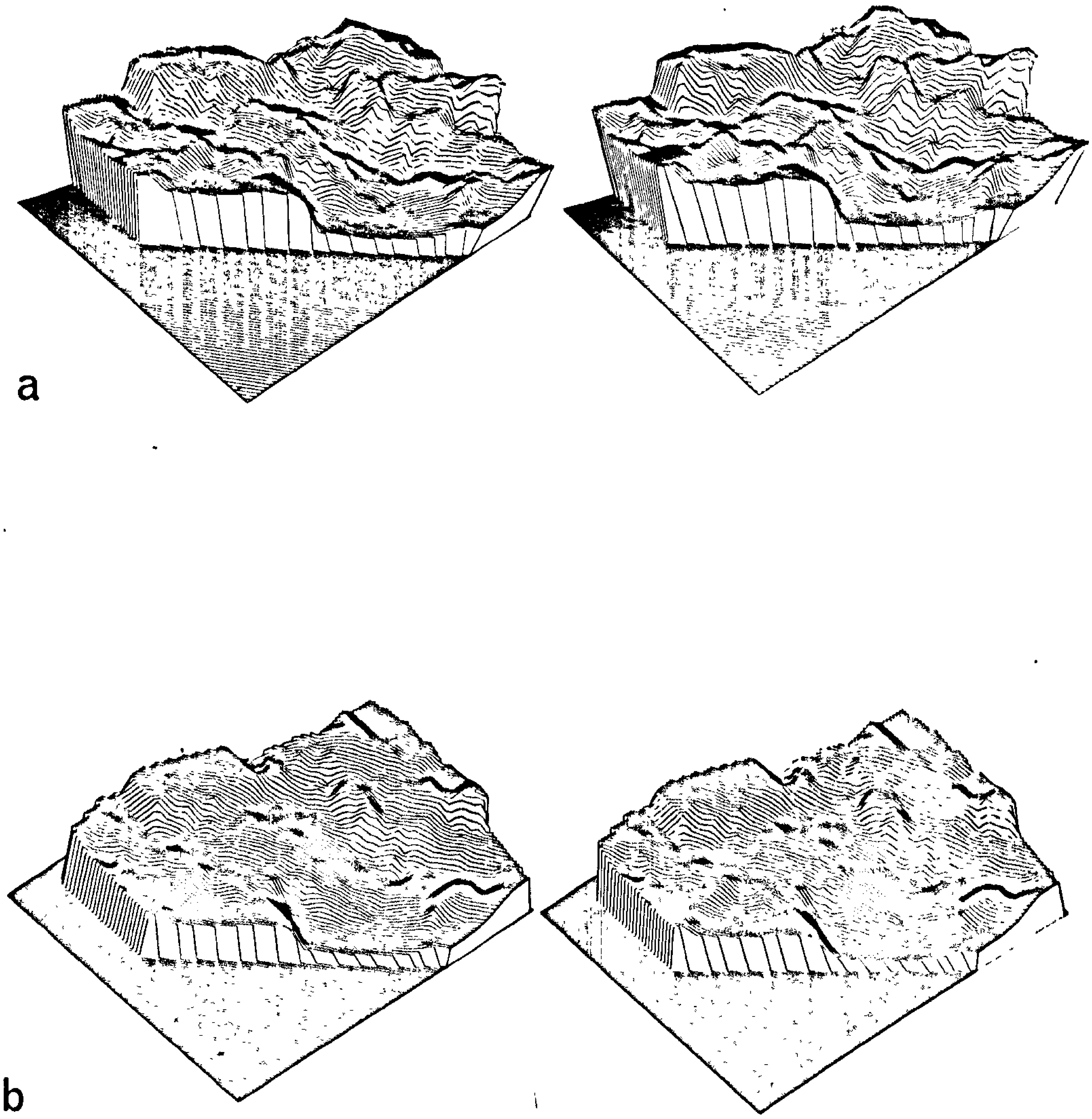

Figure 1.19

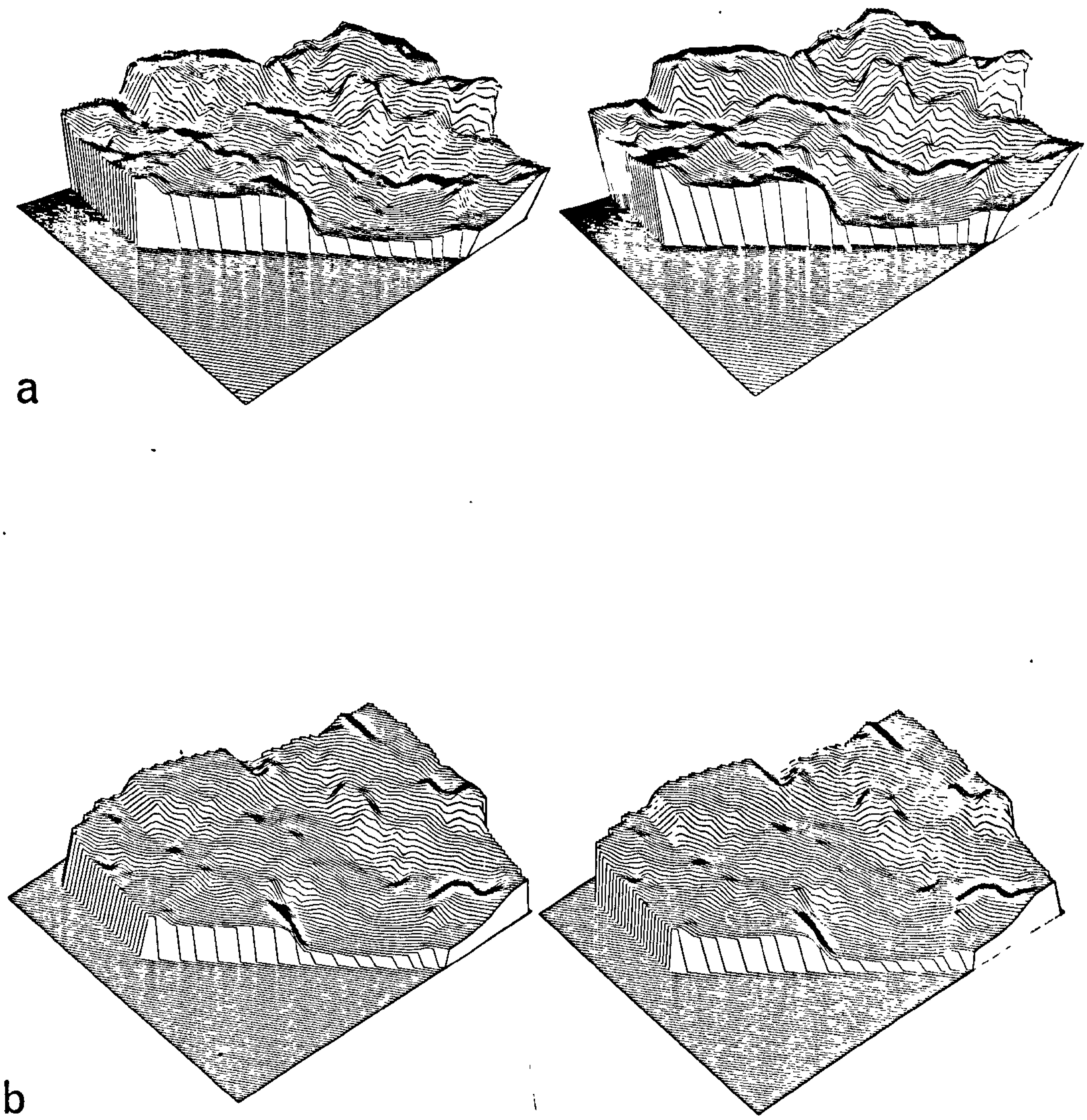

Figure 1.19

ten degree intervals for viewing as stereographic pairs. A stereo pair with azimuths of 045 and 035 degrees (approximately south-west) is reproduced in Figure 1.19a. The blocks shown in Figure 1.19b have the same orientation but, unlike the two-point perspective used in the construction of Figures 1.18 and 1.19a, are drawn as isometric (or orthographic) projections.

From these diagrams, the present sub-Permian surface can be seen to be markedly irregular, with some of the irregularity being related to sharp breaks of slope which predominantly have linear north-north-east to south-south-west and north-west to south-east trends which are undoubtedly related to post-Permian faulting. In particular, the surface in the north-eastern part of the area, in the region of Egremont, appears as an upstanding, steep sided block which is approximately 100 to 150 metres above adjacent areas to the north-west and south. Although features with an obvious structural control can be recognised, most features are less clearly related to post-Permian faulting and may, at least in part, represent the original topography of the basin floor. The configuration of low areas in the south-west, in the region of St. Bridget Beckermet, for example, almost resembles that of a wide (three kilometres) valley system which is enclosed on three sides by higher ground (see also Figure 1.16).

In summary, the eastern margin of the East Irish Sea Basin in which the western Cumbrian brockrams were deposited may be regarded as an irregular north-easterly steepening surface with a generalised concave-upward profile. The precise original topography can no longer be determined as a result of post-Permian faulting although the overall effects, as indicated by the available data, are suggested to be relatively unimportant.

Variations in brockram thickness

The vertical thickness of the brockrams in western Cumbria varies from a recorded maximum of 133.5 metres (SBH 531; NY03520981) near the north-eastern margin of the basin to low values of the order of two metres in the west (viz. Saltom Bay). As with the morphology of the sub-Permian surface, the distribution of the brockrams is exceedingly irregular and does not correspond to a simple westward thinning as suggested by the above figures. Variations in brockram thickness are shown in Figure 1.20 in the form of a computer constructed (symap) isoline map. Simplified isopachytes have been reproduced in Figure 1.21.

Comparison of Figures 1.20 and 1.15 superficially suggests little definite relationship between brockram thickness and palaeotopography. Statistical analysis, however, indicates a highly significant correlation ($p < 0.001$) between thickness and height (to mining datum) of the sub-Permian surface (Figure 1.22). Figure 1.22 indicates that the brockrams are generally thicker where the topography is lower, and, conversely, thinner where the topography is higher, and suggests that the breccias were infilling an irregular relief (which cannot, therefore, be related exclusively to post-Permian faulting). Several data points plotted on Figure 1.22 lie below the regression lines and thus have apparently anomalously low thickness values compared to their topographic setting. It is suggested that these values may be partly explained by the penetration of (unrecorded) normal fault planes within the cored sequences and partly by a distal setting where the influx of detritus was relatively less than in more proximal (easterly) areas. The correlation between thickness and topography, therefore, is only valid for those geographical areas in which active sedimentation occurred throughout the time of brockram formation, with sedimentation being inhibited on topographically high areas until the landscape became sufficiently buried for fluvial activity

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Figure 1.20. Isopachytes for the brockrams in the Beckermet mining area (symap)

Figure 1.21. Simplified isopachytes for the brockrams.

Figure 1.22. Relationship between brockram thickness and elevation above mining datum of the sub-Permian surface.

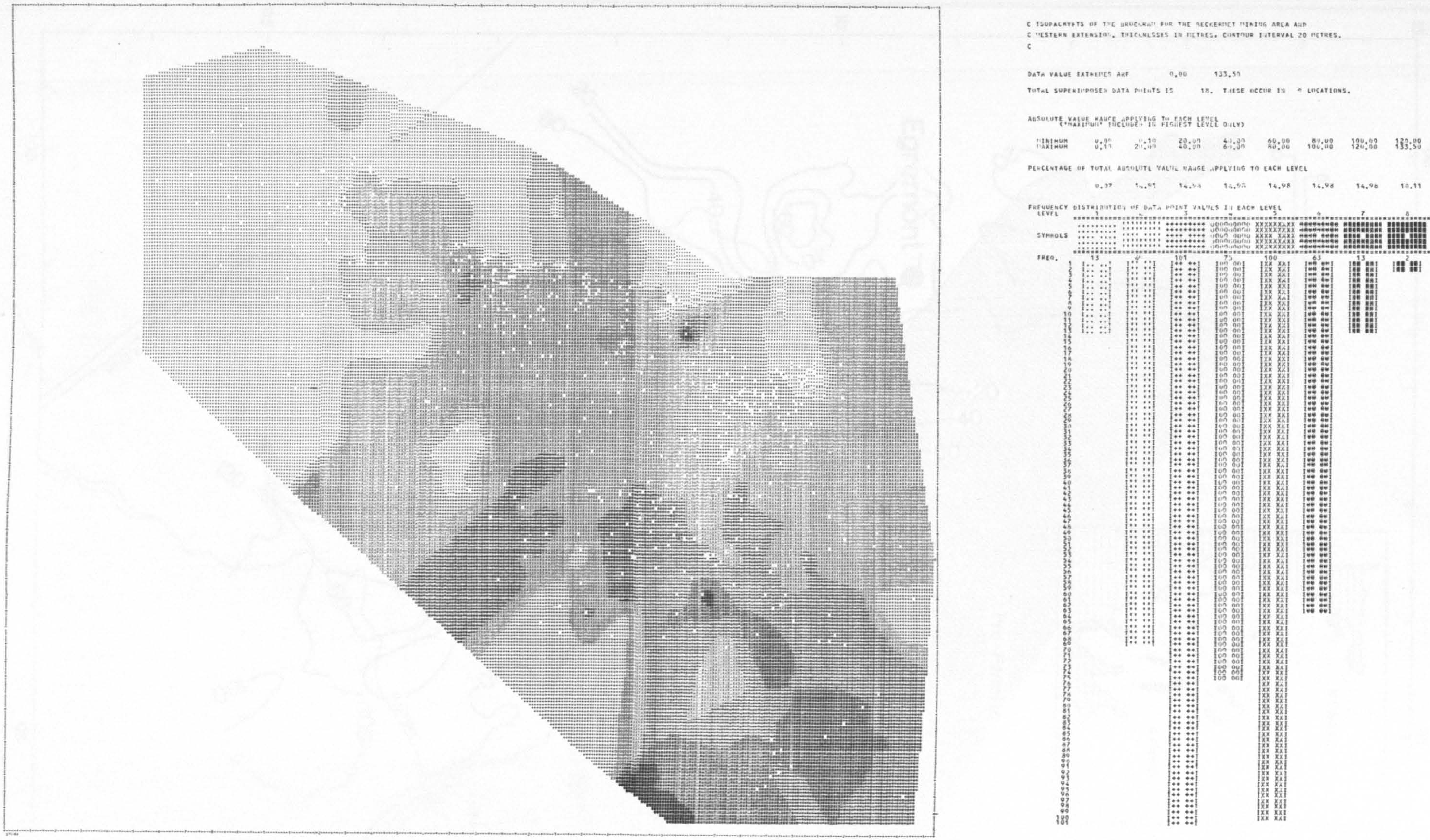
C ISOPACHYTS OF THE BROCKRAM FOR THE BECKERMET MINING AREA AND
C WESTERN EXTENSION, THICKNESSES IN METRES, CONTOUR INTERVAL 20 METRES.
DATA VALUE EXTREMES ARE 0.00 133.50
TOTAL SUPERIMPOSED DATA POINTS IS 18. THESE OCCUR IN 9 LOCATIONS.
ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)
MINIMUM 0.00 0.10 20.00 40.00 60.00 80.00 100.00 120.00
MAXIMUM 0.10 20.00 40.00 60.00 80.00 100.00 120.00 133.50
PERCENTAGE OF TOTAL ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
0.07 14.91 14.98 14.98 14.98 14.98 14.98 10.11
FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL
LEVEL 1 2 3 4 5 6 7 8
SYMBOLS
FREQ. 13 67 101 75 100 63 13 2

Figure 1.20

³00
05
15
10
05
miles
0
1
kms
0
2
Carlisle
Penrith
Whitehaven
Kendal
Lancaster
Pre-Permian
Egremont
R.Calder
R.Ehen
0
20
40
60
80
100
120

Figure 1.21

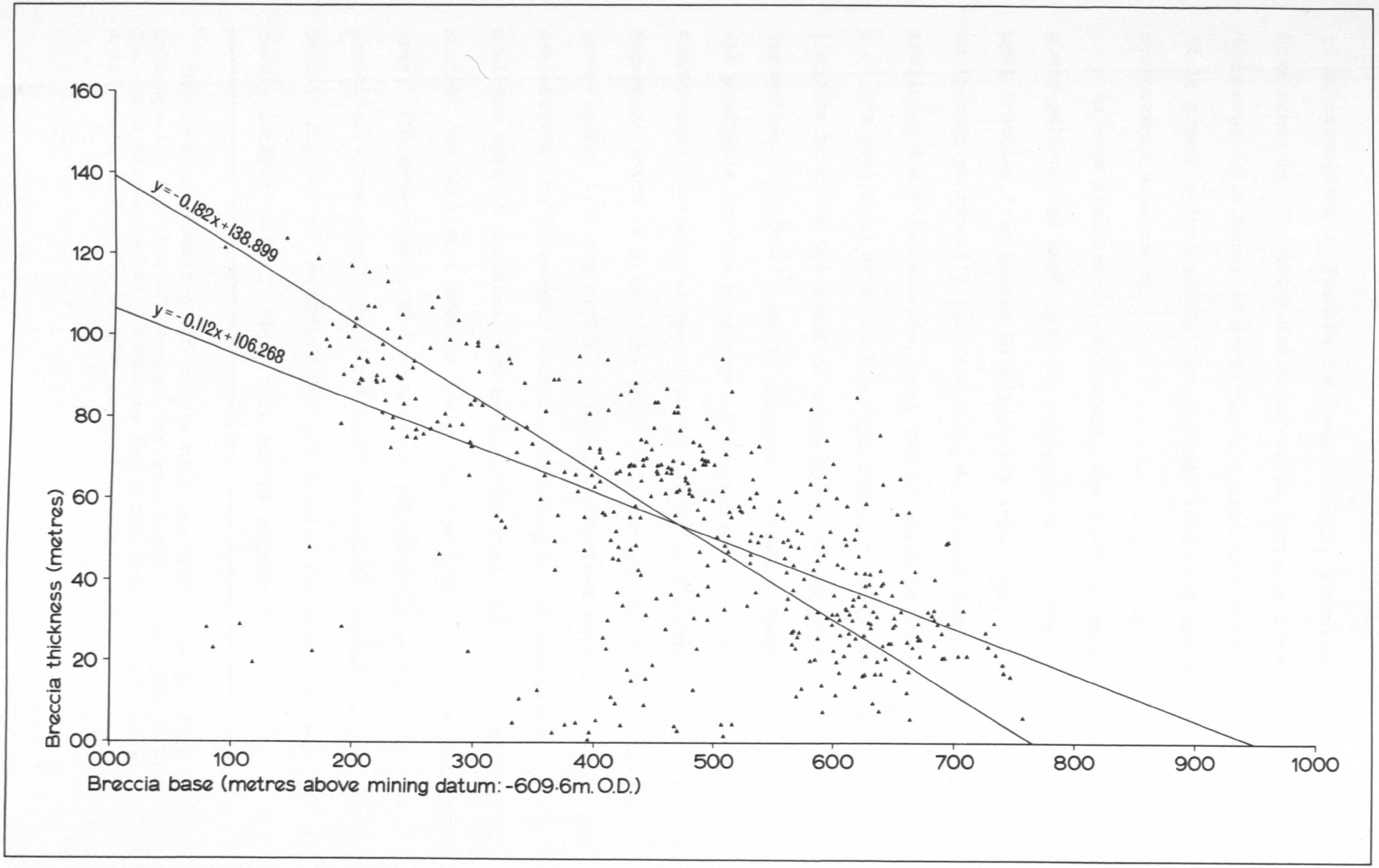
160
140
120
100
80
60
40
20
00
Breccia thickness (metres)
y = -0.182x + 138.899
y = -0.112x + 106.268
000
100
200
300
400
500
600
700
800
900
1000
Breccia base (metres above mining datum: -609·6m. O.D.)

Figure 1.22

to be established. Towards the basin margin, elevated relief appears to have prevented the accumulation of thick (greater than 60 metres) breccia sequences, presumably in response to a balance between deposition and erosion. It is possible that deposition in these locations was also partly determined by tectonic subsidence.

As a generalisation, therefore, the thickest sequences tend to occur in a mid-palaeoslope setting which corresponds to a one to two kilometre wide belt trending from Calder Bridge in the south-east to near Egremont in the north-west whereas thinner sequences occur both towards the steepening basin margin in the north and the lower relief areas to the west.

The available data for the area south and south-west of Calder Bridge indicate brockram thickness to increase to values of 120 metres or so around Yottenfews (NY032051)[1] and to decrease gradually towards the east. Data are not available for the area west of Yottenfews. To the north-west, however, thicknesses decrease to less than 40 metres in the region of St. Bridget Beckermet which is within the area of lowest relief of the sub-Permian surface (see Figures 1.16 and 1.19). Although sufficient data are not available to demonstrate the (presumed) westerly thinning of the brockrams for this southern part of the area, the generalised distribution of the breccias would suggest the Beckermet area to lie within the area of maximum brockram thickness. The development of a relatively thin breccia sequence within a topographically low area is thus difficult to resolve unless such an area was principally one of transportation and erosion (as would be the case in a deeply incised valley) rather than one of deposition.

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1 Thicknesses exceeding 120 metres have not been recorded from this area although 118.6 metres of breccia was penetrated by SBH 846 (NY04250244). The symap programme has a capacity for trend analysis and has extrapolated a 120 metre isopachyte on the basis of adjacent data.

Symvu perspective stereographic pairs of the distribution of the brockrams are presented in Figure 1.23. Azimuths of blocks are 120 and 110 degrees (approximately west-north-west).

Contours on the top of the brockrams have been reproduced in Figure 1.24. As with the sub-Permian palaeotopography, these contours indicate an easterly steepening slope from one to two degrees in distal areas to eight to ten degrees in more proximal areas (corrected for regional tilt). The similarity with the morphology of the sub-Permian surface is suggested to be related partly to the effects of post-Permian faulting whereas the generalised less irregular nature of this surface is suggested to be due to the tendency for the brockrams to infill the underlying topography (compare Figure 1.24 with Figure 1.21, for example). Assuming these figures to reflect the original slope of the upper surface of the brockrams, it is likely that bedding attitudes within the sequences would exhibit a comparable depositional inclination towards the ~~east~~. south west.

INTERPRETATION

From the preceding descriptions of the brockrams, a dominantly fluviatile mode of formation may be postulated on the basis of well-developed sedimentary structures such as bedding, cross-bedding and preferential clast orientation in conjunction with evidence from the brockrams and associated lithologies which strongly implies deposition in a subaerial environment. The development of caliche soil horizons (see Chapter Four) indicates that sedimentation, at least in distal areas, was intermittent.

Figure 1.23. Distribution of the brockrams in the Beckermet mining area as shown by symvu stereographic pairs. Data are represented as a series of bivariate histograms. Azimuths of the plots are 120 and 110 degrees (approximately west-north-west). Figure 1.23a has been constructed using two point perspective and 1.23b as an isometric projection.

Figure 1.24. Contours on the top of the brockram for the Beckermet mining area (symap).

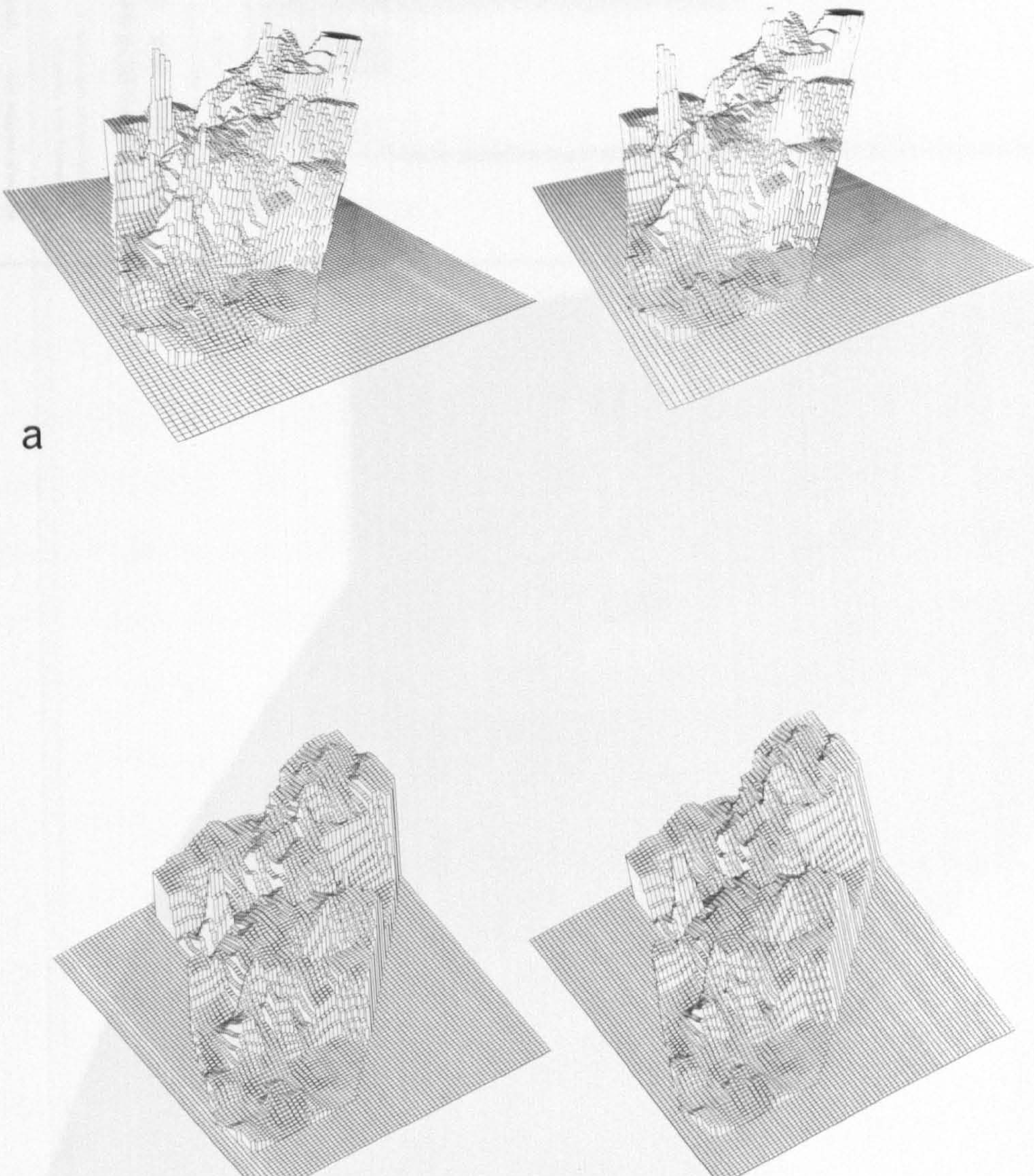

Figure 1.23

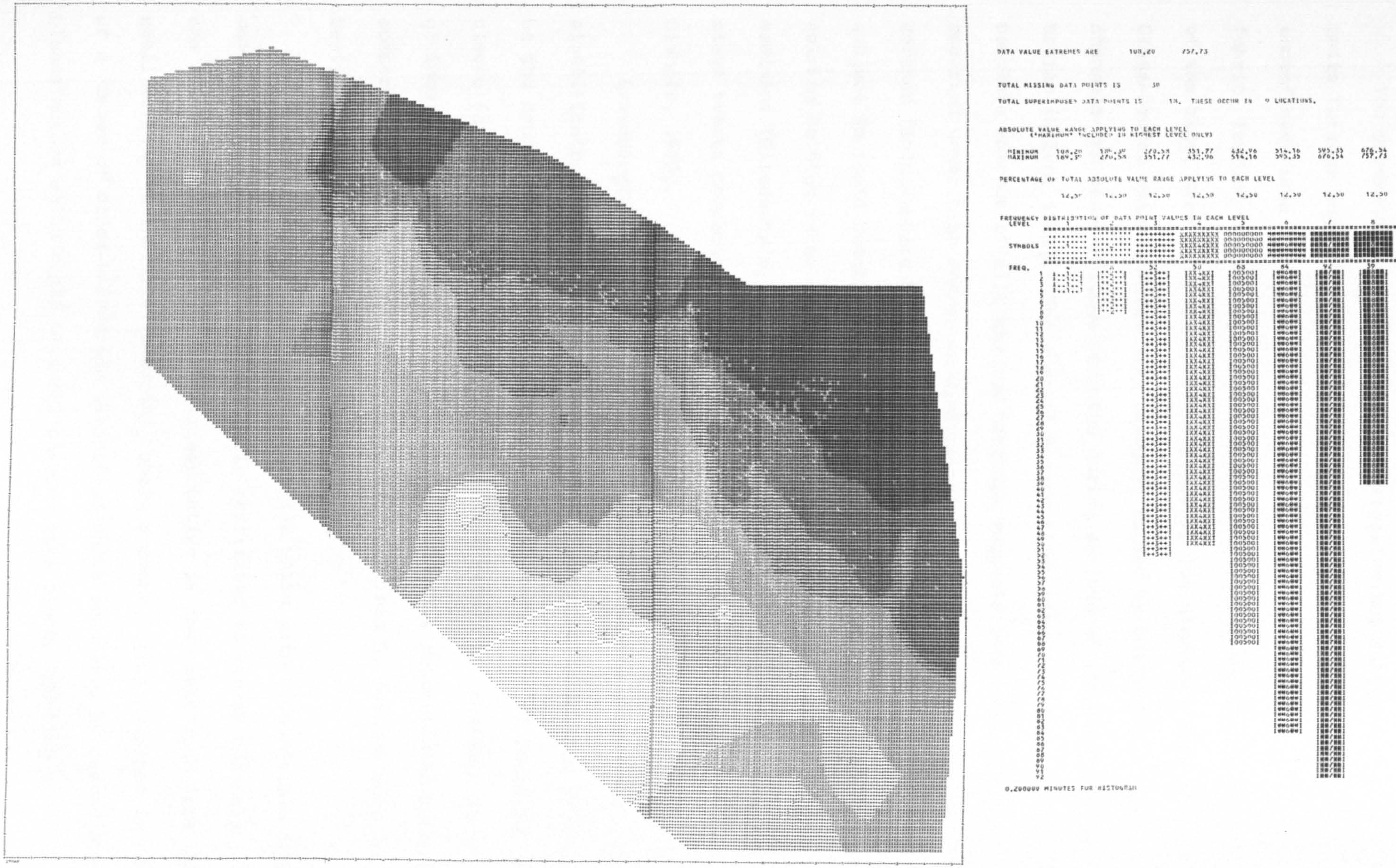
DATA VALUE EXTREMES ARE 108.20 757.73
TOTAL MISSING DATA POINTS IS 30
TOTAL SUPERIMPOSED DATA POINTS IS 18. THESE OCCUR IN 9 LOCATIONS.
ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)
MINIMUM 108.20 189.39 270.58 351.77 432.96 514.16 595.35 676.54
MAXIMUM 189.39 270.58 351.77 432.96 514.16 595.35 676.54 757.73
PERCENTAGE OF TOTAL ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
12.50 12.50 12.50 12.50 12.50 12.50 12.50 12.50
FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL
LEVEL 1 2 3 4 5 6 7 8
SYMBOLS
FREQ. 4 8 52 50 68 84 92 39
0.200000 MINUTES FOR HISTOGRAM

Figure 1.24

The predominant bedding style of irregular but laterally persistent units further suggests that the breccia were mainly deposited by largely unconfined flows which, as shown by the frequent development of a pebble fabric, lack of convincing matrix supported textures and the erosive nature of the beds, were probably mostly turbulent, low viscosity (fluid) flows. Cross-bedded scour infills were probably formed by the burial of small channels and scour hollows produced during periods of increased current action (Allen, 1962). The overall coarse clast size and the more or less complete absence of silt and mud horizons suggests that most deposition occurred under high energy conditions which, as shown by the lack of well defined fining-upward cycles and typical poor degree of sorting, were subject to rapid deceleration of flow. Rapid waning of fluid flows may be related to high rates of infiltration through previously deposited coarse sediment (Blissenbach, 1954) as well as to the violent, but short duration, storm events characteristic of arid and semi-arid environments (Denny, 1967; Glennie, 1970).

Exceedingly coarse and unsorted units with a marked absence of internal stratification (as exposed at Cuddling Hole) are considered less likely to be the products of fluid flow, however, as the occurrence of isolated large clasts (up to one metre in maximum dimension) within an otherwise finer grained matrix suggests a more or less non-selective depositional mechanism comparable to the debris/mudflow mechanism described by authors such as Beaty (1961).

The rare occurrence of markedly erosive units with a channelled form demonstrates that flow was occasionally restricted to stream channels. Of the examples recorded, the most representative is that described from the small working at Burrells (NY678183) where a channel-fill sequence is formed of a number of cross-bedded and stratified units separated by curved erosion planes across which rapid textural changes take place. Sorting and sedimentary organisation is better developed than elsewhere in the Eden Valley although the preservation of coarse, structureless to crudely horizontally stratified gravels probably represents rapid deposition of gravel bars within the main channel. The lack of internal cross-stratification is suggested to be due to a relatively shallow water depth and the failure of the migrating bar to develop recognisable slip faces (Vondra and Burggraf, 1978). Whereas structureless channel-fills suggest exceedingly rapid deposition (and may thus be interpreted as cut and fill structures), the structure of these sequences evidences conditions of relatively more continuous flow where the various fragmental sizes were subject to a more selective deposition which was probably dependent on a highly variable discharge rate. The structure of these sequences is thus consistent with that described from recent braided river sediments (e.g. Doeglas, 1962; Williams and Rust, 1969) and with that of braided channel-fill deposits of New Red Sandstone age described by Steel (1974a) from western Scotland. Sheet-like deposits with prominent internal, curved erosion and scour surfaces are similar to piedmont deposits described by Nagtegaal (1966) from the Southern Pyrenees and may be interpreted as the product of processes transitional to those of an unconfined flow mechanism.

Evidence from localities such as Stenkrith and the Burrells-Appleby area in the Eden Valley and, more particularly, from western Cumbria indicates the gross morphology of the brockram bodies to be strongly wedge-shaped in the direction of sediment transport and to be banked against the upstanding margins of the Lake District. As a generalisation, thickest sequences occur close to the margins of the depositional basins and occupy a proximal position with respect to the source for the detritus and thinner sequences occur in a more distal position (see, for example, Figure 1.21). This thinning is accompanied by a pronounced tendency for the proportion of relatively finer beds to increase, which is related, presumably, to a fining of the clasts within individual beds. Major variations in other grain parameters such as roundness and sphericity have not been observed, excepting a slight increase in mean roundness within the finer grade material.

ANALOGY WITH ALLUVIAL FANS

The closest modern analogue to these deposits are the piedmont zone alluvial fans characteristic of arid and semi-arid areas but which have also been documented from humid (e.g. Winder, 1965; Murata, 1966; McGowen and Groat, 1971) and glacial (e.g. Hoppe and Ekman, 1964; Legget et al., 1966; Wasson, 1977) environments. The structures of alluvial fans and their depositional processes of formation have been widely documented both from modern (e.g. Blackwelder, 1928; Eckis, 1928; Rich, 1935; Davis, 1938; Blissenbach, 1954; Beaty, 1963; Bull, 1963; Bluck, 1964; Bull, 1964a, 1964b; Lustig, 1965; Denny, 1965, 1967; Hooke, 1967; Bull, 1968, 1977; Cehrs, 1979) and ancient examples (e.g. Bluck, 1965, 1967; Miall, 1970; McGowen and Groat, 1971; Steel et al., 1975; Steel and Wilson, 1975; Steel, 1976; Steel et al., 1977; Wasson, 1977; Heward, 1978a, 1978b; Lindholm et al., 1979; Wilson, 1980).

These authors generally agree that semi-arid alluvial fans form in areas of high relief where there is an abundant supply of sediment and where discharge and erosion rates are high due to sporadic but violent storm events (see Blissenbach, 1954, p.177, for example). During these periods, flash-floods, transporting large amounts of rock debris, debouche from the mountains bordering desert basins and rapidly deposit their load as poorly sorted breccias and conglomerates when stream competency is reduced at the foot of the mountains. Deposition may either be initiated by a reduction in gradient (Blissenbach, 1954; Beaty, 1963; Denny, 1967) or, where the slope of the upper fan is continuous with that of the mountain canyon, by a fall in shear stress due to lateral expansion and shallowing of the flow at the fan apex

(Collinson, 1978a; Heward, 1978). Bull (1968) maintains that the second mechanism only is of importance for the majority of alluvial fans. Shallowing may also be effected by infiltration into previously deposited, highly permeable alluvial fan sediments (Blissenbach, 1954; Bull, 1963, 1968). As the rivers emerge onto the desert basin, they spread out from the mountain front and deposit the detritus in the form of a fan which has the section of a segment of a very low cone (Blissenbach, 1954; Denny, 1967). The relief of the fan surface from apex to distal fringe varies with the size of the fan which is controlled largely by the size of the catchment basin (Denny, 1965, 1967; Hooke, 1968) but is also influenced by the erodibility of the source area rock types (Bull, 1962a, 1964a, 1964b; Hooke and Rohrer, 1977). In radial section, most fans have a concave-upward profile (Denny, 1965, 1967; Bull, 1972) in which dips steepen from low values of one to two degrees at the distal margin to angles of five to ten degrees at the apex (Blackwelder, 1931). The angle of dip rarely exceeds ten degrees (Blissenbach, 1954) and is generally greater on small alluvial fans (Eckis, 1928). Alluvial fans are commonly developed along the downthrow sides of active fault zones (e.g. Longwell, 1930; Sharp, 1948; Drewes, 1963; Bluck, 1969; Nilsen, 1969; Wilson, 1970; Steel, 1971; Steel and Wilson, 1975) and may give rise to unusually thick breccia sequences (e.g. Steel _et al._, 1977). Laterally extensive marginal sequences may be formed where adjacent alluvial fans coalesce to form a compound fan or bajada (Trowbridge, 1911; Blackwelder, 1928, 1931).

In comparison, the fans of western Cumbria have a concave-upward longitudinal section in which the slope of the brockram surface increases from one to two degrees at the distal extreme to eight to ten degrees in more proximal areas (see Figure 1.24) and which corresponds to the radial section morphology described from modern fans. As suggested for the western Cumbrian examples, bedding attitudes within fans are recorded to approximate to that of the surfaces (Blackwelder, 1931; Blissenbach, 1954). The overall wedge shaped nature of the deposits also corresponds broadly to that of modern fans although the thickest brockram sequences tend to occupy a mid-fan position and thin both proximally and distally (see Figure 1.21). Although this distribution is partly apparent inasmuch as the pre-Permian outcrop is flanked by a narrow zone of exposed brockram where thicknesses are reduced by erosion, relatively thinner sequences also occur towards the basin margin where there is a cover of St. Bees Sandstone. Blissenbach (1954), however, records that under exceptional circumstances, a sequence may thicken downfan from the apex; a distribution which is presumably related to the morphology of the sub-fan surface. Similar examples have also been recorded by Bull (1972). For western Cumbria, therefore, this thickness distribution is suggested to be due to the infilling of an irregular sub-Permian surface which has a generalised concave-upward form in the direction of sediment transport. The interfingering of the coastal plain sediments of the St. Bees Shales in the west of the mining area indicates that the thickness distribution of the brockrams is also partly related to a decrease in the size of the fan bodies (and hence, competence of flow) through time.

The extensive development of the breccias and the variable thickness relationships perpendicular to the transport direction is suggested to represent accumulation as a series of coalescing fans rather than well defined, isolated fans. The distribution of the Penrith Brockram (see Figure 1.5) gives some indication that two separate areas of development may have existed in the Eden Valley.

Depositional processes on alluvial fans

The surfaces of most alluvial fans are recorded to be characterised by a network of channels which radiate from the fan apex and dissect the fan surface (e.g. Blackwelder, 1931; Denny, 1967). The main channel is generally incised into the upper part of the fan (Eckis, 1928; Bulwalda, 1951; Hooke, 1967; Bull, 1972) but emerges onto the surface at the intersection point (Hooke, 1967; Wasson, 1974). Deposition thus tends to be restricted to channels in the upper part of the fan whereas the lack of a confined channel below the intersection point enables processes such as sheetflooding to predominate over the lower surfaces (Wasson, 1977).

The main depositional products of alluvial fans have been summarised by Bull (1972, 1977) as debris flow[1] deposits, sheetflood deposits, stream channel deposits and sieve deposits, the first being the result of high viscosity, matrix supported flows and the remainder of low viscosity, fluid flows. The characteristics and distribution of these deposits has been extensively documented by authors such as Blackwelder (1928), Chawner (1935), Rich (1935), Ives (1936), Davis (1938), Sharp and Nobles (1953), Blissenbach (1954), Beaty (1961), Bull (1963), Ruhe (1963), Bull (1964a), Hooke (1967), Bull (1972), Wasson (1974), Steel and Wilson (1975) and Brookfield (1980) and need not be summarised here.

As regards the processes of brockram formation, the dominant depositional mechanism has been suggested to be that of unconfined fluid flow which thus corresponds most closely to the shallow braided distributary, sheetflood mechanism of Bull (1972). Debris flow deposits, of the type described by authors such as Krumbein (1942), Sharp and Nobles (1953), Blissenbach (1954), Beaty (1961) and Hooke (1967) have been recognised on the basis that such deposits are recorded generally to form poorly sorted and unstratified beds in which larger clasts are supported by a finer matrix and which usually have non-erosive basal surfaces (Crandell, 1957; Wasson, 1977) as a result of a laminar, rather than turbulent flow mechanism (Fisher, 1971; Blatt <u>et al</u>., 1972). Stream channel deposits, including those of braided channels, form only a minor component of the brockrams. Braided channel deposits, as

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1 This mechanism incorporates mudflow which may be considered as a type of debris flow (Sharp and Nobles, 1953, p.550).

described by Steel and Wilson (1975), have been distinguished from braided distributary, sheetflood deposits on the basis of their laterally more restricted occurrence and abundance of concave-upward erosion surfaces although, clearly, intermediary deposits may be expected to occur. Williams (1966) has made a similar distinction for water-lain alluvial fan sediments of Torridonian age in Scotland. Sieve deposits _sensu stricto_ (Hooke, 1967; Wasson, 1974) have not been recognised. It is believed, however, that rapid infiltration (and hence rapid deceleration of flow) inhibited selective deposition and is thus partly responsible for the typically poor degree of sedimentary organisation which is characteristic of the brockrams as a whole.

Although the significant statistical correlation between maximum clast size and bed thickness ($p < 0.001$; see Figure 1.5) and the relatively high maximum clast size to bed thickness ratio (compare Figure 1.5 with Bluck, 1967, Fig. 3, p.142 and Fig. 4, p.147 and with Steel, 1974a, Fig. 4, p.340 and Fig. 9, p.343) also supports an interpretation of a dominantly unconfined flow origin in which there is little or no reworking of previously deposited sediment (Bluck, 1967), it is important to note that Steel (_op. cit._), Steel _et al._ (1975) and Steel and Wilson (1975) have not recognised sheetflood deposits from comparable New Red Sandstone sequences of Britain and have stressed a debris flow origin for all laterally impersistent beds lacking markedly erosive basal surfaces, including those in which a bedding parallel pebble orientation is developed. Orientated pebble fabrics, including imbrication, have also been described by Lindsay (1966), Heward (1978a) and Kurtz and Anderson (1979) from deposits interpreted as the products of debris flow. Hooke (1967) has also suggested that a selective deposition of the larger particles may be accomplished by this mechanism and both inverse (Fisher, 1971) and normal (Crandell, 1957; Bluck, 1967) grading has been recorded. Although the development of such fabrics is not considered to be inconsistent with a laminar flow mechanism (Allen, 1970b; Blatt _et al._, 1972), Bull (1972) suggests that they are restricted to low viscosity debris flows and do not occur in viscous flows. As low viscosity debris flows are considered to be transitional in character from those of viscous to fluid flow (Bull, 1963), the possibility that significant, although largely unrecognised, debris flow-type deposits occur within the brockrams cannot be discounted (see Discussion, this chapter), especially where such deposits may even have an apparently clast supported framework (Bluck, 1967, p.144; Steel and Wilson, 1975, p.186). Such an interpretation also concurs with observations of Blackwelder (1928, p.474) who records the majority of arid region alluvial fans to be formed of inter-bedded debris flow and sheetflood (referred to as 'washed gravel sheets') deposits.

DISCUSSION

Despite the possibility that deposits of all the main depositional mechanisms recorded from contemporary semi-arid alluvial fans are represented in the brockrams, the apparent preponderance of sheetflood-type deposits initially points towards significant discrepancies with respect to published sedimentary models (e.g. Hooke, 1967; Bull, 1972) but concurs with the author's observations of certain comparable deposits such as the Teignmouth Breccias of Devon which have also been interpreted as the deposits of alluvial fans (Laming, 1966).

According to the models erected by Hooke (op. cit.), Wasson (1977) and others (summarised in Collinson, 1978a), stream channel and debris flow deposits tend to occur near the apex whereas sheetflood deposits occur in more distal areas below the intersection point. Sieve deposits tend to be concentrated in a zone immediately below the intersection point (Wasson, 1974). However, to some extent, both debris flow and sieve deposits are favoured where particular conditions of formation are prevalent in the depositional area. In the case of debris flow, it is suggested that optimum conditions are provided where source rocks can provide a substantial proportion of fine debris (Blackwelder, 1928; Bull, 1964a), usually (but not in every case) including substantial amounts of clay, and by steep slopes which promote rapid erosion and run-off (Blackwelder, 1928; Bull, 1963). Debris flows are also favoured where the volume of sediment is high compared to the volume of water (Wasson, 1977). It is considered that such conditions are more likely to have prevailed in the lithologically varied source area of western Cumbria although Bluck (1964) has recorded the occurrence of debris flow deposits from a limestone dominated area in southern Nevada which is thus comparable with the Eden Valley. For sieve deposits, a prerequisite of formation is that highly permeable previously deposited sediment enables rapid infiltration of subsequent, fine-grade deficient flows and thus results in the deposition of a clast supported gravel lobe. Although, therefore, the apparent absence of such deposits in the brockram sequences may simply be related to formational conditions unsuitable for the production of either debris flow or sieve deposits (for example a critical availability of fine detritus, sufficient to inhibit sieve deposition but insufficient to promote debris flow), the field distinction of these deposits is based on sedimentary criteria such as sorting and pebble fabric which are not uniquely diagnostic of specific processes where the shape of the deposit is uncertain (see Wasson, 1977, p.782). In the case of the unconfined flow deposits of the brockrams, therefore, distinction between debris flow and sheetflood deposits is based solely on the relative degree of sedimentary organisation and the presence or absence of basal erosion surfaces; features which clearly form part of a continuum of rock types and which may thus be interpreted as the products of transitional sheetflood-debris flow mechanisms. Intermediate waterlain-debris flow deposits have been recognised by Bull (1962b, 1963, 1964b, 1972, 1977) but the practical difficulties involved in their precise interpretation has led authors such as Steel (1974a), Steel et al. (1975) and Steel and Wilson (1975) to suggest an exclusively debris flow origin[1] for sheet-like deposits similar to those of the brockrams whereas authors such as Wasson (1977) have adopted a fluvial origin for any deposit showing even the most rudimentary sedimentary structure. Clearly, therefore, considerable confusion exists as far as these deposits are concerned and, for this reason, specific classification of the brockram deposits has not been attempted. Both sheetflood and debris flow type deposits occur as do, undoubtedly, intermediary types. The exact proportion of these various types is, however, impossible to assess in relation to the diverse interpretation of comparable deposits. For the brockrams, therefore, a sheetflood-debris flow formation is postulated on the basis of the shape of the deposits and the variation of structure and fabric from highly organised (fluvial) to disorganised (debris flow).

Whereas the recognition of unconfined flow deposits may present certain difficulties, those of stream channels are more readily identified by virtue of their markedly erosive bases and lateral impersistence and have been recorded from several comparable sequences (e.g. Steel, 1974a; Steel et al., 1975; Steel and Wilson, 1975). Stream channel incision generally occurs in the upper reaches of fans (Bull, 1972) where the gradient exceeds that associated with steady-state streamflooding either, under normal circumstances, as a result of debris flow deposition in the fanhead area (which requires a higher hydraulic gradient; Hooke, 1967; Miall, 1978) or of a shift in the locus of sedimentation (Hooke, 1967). Incision may also result from tectonic

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[1] Similarly, Bluck (1967) and Wilson (1980) refer to sheetflood deposits but do not recognise a distinction from those of debris flow origin. To some extent, however, transitional braided distributary (sheetflood; Bull, 1972) - braided channel deposits may be difficult to resolve in stratigraphic examples and may be represented in the braided stream deposits described by Bluck (op. cit.), Steel (1974a), Steel et al. (1975) and Steel and Wilson (1975).

movement or climatic variation (Bull, 1964b, 1964c, 1977; Wasson, 1979), with the latter possibly including storm events of unusual magnitude (Beaty, 1970). Channels are thus predominantly the sites of erosion and transportation but may be expected to be preserved as infills under aggrading conditions where reworking of the fanhead area during sourceward migration of fan facies is not significant. The apparent absence of stream channel deposits from the brockram sequences (which may be partly due to exposure orientation) thus implies that surface slopes were in equilibrium with fluviatile processes and were only rarely sufficiently steep to promote the formation of channels. Although this assumption is not altogether in agreement with calculations of fan surface slope (which have yielded figures of eight to ten degrees, comparable with the maximum inclinations recorded from contemporary fans; Blissenbach, 1954, p.176), it is in accordance with the high level of statistical correlation between maximum clast size and bed thickness which, according to Bluck (1967) and Steel (1974a), is unlikely to be the case where there is a significant component of braided channel deposits.[1] This interpretation also concurs with the more or less complete absence of convincing debris flow deposits, the formation of which has been suggested to contribute towards channel incision in contemporary fans (e.g. Hooke, 1967). For the brockram sequences, therefore, it is postulated that deposition was mainly effected by sheetflooding, or transitional sheetflood-debris flow mechanisms, initiated by lateral expansion of the flows from confined upland channels at the fan apex (see Bull, 1972, p.66) and that the higher slopes of the fans were approximately continuous with those of the canyons. It is possible that such a situation may have been promoted by a relatively lower relief than is commonly recorded from contemporary analogues and that, in consequence, a

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1 Braided river deposits were recorded to show no significant correlation between maximum clast size and bed thickness.

graded stage (Blissenbach, 1954, p.178), suitable for the production of sheetflood-type flows, was achieved comparatively early in the depositional history of the brockrams.

As a consequence of the monotonous sedimentary style of the brockrams, a division of the sequences into proximal and distal fan facies based on existing models (e.g. McGowen and Groat, 1971; Steel, 1974a; Heward, 1978a) and descriptions of the distribution of fan processes and facies (e.g. Hooke, 1967; Bull, 1972) is not feasible. The only possibilities for distinguishing proximal and distal fan facies within the brockrams are thus restricted to grain parameters such as size and shape. The downfan decrease in particle size is well documented from contemporary fans (e.g. Blissenbach, 1952, 1954; Beaty, 1963; Bluck, 1964, 1965; Bull, 1964a; Ruhe, 1964; Denny, 1965, 1967) and Blissenbach (1954) has demonstrated an associated downfan increase in roundness for a fan at the southern base of the Santa Catalina Mountains, Arizona.[1] As noted previously, however, apparent trends in roundness and shape have not been recognised from the brockrams (c.f. the Van Horn Sandstone; McGowen and Groat, 1971) and deductions of proximity can only be expressed as a relative downfan decrease in clast size which is difficult to quantify.

The overall fining-upward of the western Cumbrian sequences may thus be interpreted as a progressive facies change from proximal to distal style sedimentation through brockram depositional times. Similar trends have been recorded by Bluck (1967), Williams (1969) and Steel and Wilson (1975) and suggested to reflect a gradual migration of fan facies as a result of reduced relief in the source area. This is in accordance with the upward increase in extent of the fan bodies which, because of the relationship between fan size

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1 Sphericity was recorded to be more or less constant over the surface of this example. Similar results have been obtained by Bluck (1964) from an alluvial fan in southern Nevada and by Krumbein (1940, 1942) from flood deposits in California.

and source area (Bull, 1962a, 1964a; Denny, 1965, 1967; Hooke, 1968), implies a corresponding decrease in drainage area.

CONCLUSIONS

The brockrams are regarded as a product of subaerial alluvial fan sedimentation resulting from flash-flood run-off from the Lake District uplands within an arid or semi-arid environment. Deposition occurred at the upland margins of a series of basins peripheral to the Lake District, the breccias of western Cumbria being deposited at the eastern extreme of the East Irish Sea Basin and those of the Eden Valley within a linear basin bounded on its eastern side by the Pennine fault belt (Bott, 1978) and opening northwards into the Carlisle Basin (Smith et al., 1974).

The breccias of western Cumbria were deposited on an easterly steepening and markedly irregular pre-Permian surface with a total relief in the order of 300 metres. The gross morphology of these deposits is wedge shaped with an overall concave-upward profile in the direction of sediment transport. The maximum slope of the brockram surface is in the order of ten degrees in proximal areas, this value being more or less the steepest angle recorded from the surfaces of contemporary fans (Blissenbach, 1954). Bedding attitudes within the sequences may be expected to approximately parallel that of the fan surfaces and, similarly, steepen in a sourceward direction (Blissenbach, op. cit.). The lateral extent and variable thickness distribution of the brockrams along the line of the basin margin suggests sedimentation to be the product of a series of coalescing fans rather than well defined, isolated fans. To a lesser extent, this appears to be true also of the Eden Valley although dispersal patterns and the thinning of the breccias in the region of Helm Beck indicates two possible areas of development.

The dominant depositional mechanism is suggested to correspond to the braided distributary, sheetflood process of Bull (1972) although transitional sheetflood-debris flow processes are considered likely to have formed an important, but indistinguishable, component of the sequences. Deposition is suggested to have been initiated by the loss of confined channels at the basin margins and to have occurred mostly from high energy flows of sufficient competence to transport very coarse debris. Stream channel deposits form only a minor component of the sequences and are mainly the products of braided stream systems. Possible transitional sieve-type deposits exist but have not been distinguished from the products of sheet-flooding.

The recognition of alluvial fan facies based on published models (e.g. McGowen and Groat, 1971) or on descriptions of the distribution of alluvial fan processes (e.g. Collinson, 1978a) has not proved feasible for the brockrams due to the predominance of sheetflood-type deposits, the lack of well-developed structures other than bedding and preferential clast orientation and to the tendency for the different breccia facies types described to be randomly interbedded at any one locality. However, the general westward fining of the western Cumbrian breccias corresponds to the well documented decrease in particle size (e.g. Blissenbach, 1954; Bluck, 1964) and is in accordance with deductions of proximity based on palaeocurrent data and the morphology of the fan bodies and sub-Permian surface. The overall upward-fining of this sequence further suggests a gradual sourceward migration of fan facies which is believed to be related to a lowering of relief in the source area (e.g. Steel and Wilson, 1975). As with sequences of Torridonian age studied by Williams (1969), this implies (but does not confirm: Steel _et al._, 1979) accumulation under relatively stable climatic and tectonic conditions.

CHAPTER TWO

THE PETROGRAPHY AND DIAGENESIS OF THE BROCKRAMS

INTRODUCTION

The petrography of the Permo-Triassic alluvial fan conglomerates and breccias of Cumbria may be considered, in a broad sense, by reference to the clast component and the matrix and/or cement component. Whereas the clasts exclusively comprise derived Palaeozoic fragments, the matrix/cement component is usually formed of a variable proportion of clay, sand, silt and calcite and may be of either a detrital or chemical origin.[1] However, the localised derivation of the clasts has an important effect on the relative abundance and proportion of the various Palaeozoic fragments and clearly differentiates the brockrams of Edenside, including the Stenkrith Brockram, from those of western Cumbria. In particular, the variable clast component has had a considerable effect on the scope and extent of post-depositional alterations which have occurred within these deposits.

With the exception of the interbedded sandstone and breccia sequences, the clast component of the Edenside brockrams is entirely composed of derived Carboniferous rock fragments and comprises limestone, sandstone and chert. In western Cumbria, however, Carboniferous fragments form only a minor percentage of the clasts with the major percentage comprising Ordovician Skiddaw Slate and Borrowdale Volcanic fragments derived from the western margin of the Lake District massif. The scope of the diagenetic processes which can affect these rocks is greater, therefore, for the western Cumbrian deposits as a wider variety of lithologies, and, in consequence, elements, are available for dissolution, mobilisation and precipitation. In particular,

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1 For the purpose of the following descriptions, the term 'matrix' will be applied to the latter component excepting where a specific mode of formation may be inferred for the individual compositional elements of the matrix/cement.

the unstable Borrowdale Volcanic clasts are highly susceptible to such processes.

The petrography of the interbedded breccias and sandstones of exposures such as Belah Crags (NY793120) and Low House (NY515486) is not described in this chapter as these deposits are not considered to have accumulated as alluvial fans. These sequences are discussed separately in the proceeding chapter. Specific mention will not be made of the breccias occurring at Humphrey Head, near Grange over Sands (SD385740) as these deposits are petrographically similar to the brockrams of Edenside.

The petrography of the Edenside brockrams

As noted above, the clast component of the Edenside breccias is composed entirely of Carboniferous fragments. The dominant component is limestone which usually constitutes in excess of 85 per cent. of the clasts. The majority are a light grey to buff coloured unaltered limestone although a small percentage are reddened and, locally, dolomitisation may be extensive. Where clasts are reddened, the reddening may be marginal or, less commonly, occur throughout the clast. The lithology of the limestone clasts is variable and includes shelly bioclastic limestones, foraminiferal limestones, crystalline limestones and micrites. All types, however, occur within the Carboniferous Limestone exposed to the south-west of the Eden Valley (Waugh, 1967) and are thus locally derived.

The other main clast components of the Edenside brockrams are Carboniferous sandstone and chert fragments which together make up less than 15 per cent. of the total volume of the rock. Sandstone usually exceeds chert in quantity in the Edenside brockrams although the converse is true for the southern Cumbrian deposits.

The matrix of the breccias is composed of carbonate and quartz grains which usually have a thin ferric oxide coating and are, for the most part, calcite cemented. Rarely, significant quantities of clay may be present although never in sufficient quantities to form a clay matrix. The proportion of carbonate grains to quartz grains may be of similar magnitude to that of the clasts but is usually much less with some horizons comprising up to 50 per cent. quartz grains. The quartz grains may be either sub-angular to sub-rounded or very well rounded with the latter possessing identical characteristics to those of the Penrith Sandstone. It is considered, therefore, that the former are likely to have been derived directly from the mechanical breakdown of Carboniferous sandstone whereas the latter represent aeolian sand grains (see Chapter Five) which are either derived from the reworking of wind blown sands on the surfaces of the fans or represent the incorporation of the occasional wind blown grain.[1] For the most part, the brockrams of Edenside are cemented by sparry calcite, although horizons with a high percentage of carbonate grains may show localised effects of pressure solution. In such horizons, calcite cement is more or less absent and the grains are tightly packed with long contacts, or, where quartz grains are in contact with carbonate grains, concavo-convex contacts. Sutured contacts are rare.

Locally, as in exposures immediately below Franks Bridge (NY777088) and at the top of the Stenkrith Brockram (NY775075) dolomitisation of the limestone clasts and grains may occur.

Petrography of the western Cumbrian brockrams

The petrography of the western Cumbrian sequences is markedly different to that of the Edenside brockrams in that derived Carboniferous fragments are relatively less significant, mainly occurring immediately above the sub-Permian unconformity, and that Lower Palaeozoic Skiddaw Slate and Borrowdale Volcanic detritus forms the bulk of the fragmental component. The average percentage composition of the clasts comprises Borrowdale Volcanic and Skiddaw Slate fragments 82 per cent., Carboniferous Limestone 3.5 per cent. and

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[1] It should be noted that the aeolian grains may also have an original derivation from the Carboniferous sandstones (Waugh, 1967).

Carboniferous sandstone 13.5 per cent. with the remainder being formed mainly of chert and haematite (see Appendix Three).

Throughout the Beckermet area of western Cumbria, the percentage of Borrowdale Volcanic clasts usually greatly exceeds that of the Skiddaw Slate component and is, therefore, the dominant detrital component of the sequences. It is clear, however, that the relative proportion of the clasts is highly variable and that, elsewhere, Skiddaw Slates may be expected to exceed Borrowdale Volcanics in percentage composition, especially to the north of the Beckermet area where much of the detritus is locally derived from the low fells of Cleator Moor.

The majority of the Borrowdale Volcanic clasts comprise intermediate and acid lavas ranging in composition from andesite to rhyolites. Andesites are more common than rhyolites and usually occur as sub-angular to blocky green-grey clasts with, in thin section, conspicuous plagioclase phenocrysts. Rhyolite clasts are usually of a lighter grey colour and also contain small plagioclase phenocrysts. The composition of the plagioclase is usually in the range of An_{10} to An_{30}, i.e. oligoclase or andesine. Both clast types are marginally reddened and frequently heavily replaced by clay. The green colouration of the clasts is due to the presence of chlorite which is often associated with minor quantities of ferroan calcite and occurs marginally to the feldspar phenocrysts. Where feldspar laths form a framework to the rock, chlorite is restricted to the inter-lath spaces and may thus correspond to a replaced intersertal texture. It is probable, therefore, that the chlorite formed as a replacement of ferromagnesian minerals of the ground mass of the original rock, possibly also deriving aluminium from an associated albitisation of the feldspars (although no evidence of this process is observed in the brockrams). Both the replacement of the ferromagnesian minerals and albitisation of feldspars would provide a calcium source for the formation of calcite. With the exclusion of the marginal reddening of the clasts, however, the mineralogy of the Borrowdale Volcanic clasts, as outlined above, is essentially identical to that of specimens collected from the main outcrops of the Borrowdale Volcanic Group in the Lake District massif (e.g. from Long Sleddale, near Kendal) and, therefore, not related to the diagenetic changes which occurred after the deposition of the brockrams.

Ignimbrite and clasts of other types of pyroclastic deposit are not common in the brockrams although tuff clasts have been recorded from the breccias exposed in Saltom Bay and occur rarely throughout the sequences of the Beckermet area. Where exposures of these rock types are not strongly welded, they are generally highly cleaved and, in consequence, easily erodable. It is considered unlikely, therefore, that pyroclastic detritus would form a significant proportion of the clast component of the breccias.

Skiddaw Slate clasts form the second common fragmental type within the Beckermet area and comprise a variable group of sandstones and siltstones, with siltstones predominating. The grain types of the clasts comprise well sorted, sub-angular quartz grains, lesser amounts of feldspar and muscovite and disseminated opaque ore minerals. The sandstones may be cemented by syntaxial overgrowths of quartz, pressure welded or, more commonly, have an interstitial clay matrix (which may be related to Permo-Triassic diagenesis). Where a clay matrix is present, the clasts may be secondarily reddened; the reddening occurring either marginally or throughout the matrix of the clast. The Skiddaw Slates are also considered to be a likely source for the small percentage of vein quartz clasts present in the breccias.

Carboniferous limestone and sandstone clasts are less common and tend to occur immediately above the sub-Permian unconformity. Whereas limestone clasts occur throughout the Beckermet area, the occurrence of the sandstones tends to be restricted to more westerly areas where the Coal Measure sandstones form the floor of the depositional basin. Locally, as on the coast at Saltom Bay (NX960161), Carboniferous sandstones may form the major clast component. Grey limestone clasts form the largest fragmental type (the dimensions of which cannot be determined precisely as the seven centimetre diameter core sections may penetrate a single clast for 20 centimetres or so) and are generally sub-angular to blocky in shape. Although the majority are unaltered, limestone clasts present in the breccias underlying the Magnesian Limestone in Saltom Bay are dolomitised. Whereas the reddening of the limestone clasts tends to be peripheral, Carboniferous sandstone clasts are frequently completely reddened. As with Edenside brockrams, this variation is probably related to the sub-Permian reddening which is of considerable extent in many areas (Taylor, 1978). The exposures of the Westphalian Whitehaven Sandstone in Saltom Bay, for example, are, in part, coloured a deep purple-red and could provide a source for reddened clasts. It should also be noted, however, that all stages of peripheral reddening are recorded from the brockram sequences and that reddened clasts are usually considered to be the result of <u>in situ</u> diagenetic oxidation of iron bearing minerals in the clasts.

The restricted occurrence of large Carboniferous limestone clasts and sandstone clasts to the base of the sequences reflects the distribution of the sub-Permian basement rocks (see Figure 1.4). The main source for the brockram detritus was formed by the upstanding Lake District massif which thus provided Borrowdale Volcanics and Skiddaw Slates. The floor of the depositional basin, however, was formed of Carboniferous Limestone in the east and Coal Measure sandstone in the west, and would, therefore, provide little detritus, especially once the alluvial fans were initiated. Most of this material, therefore, probably accumulated largely from the mechanical breakup of the bed rock prior to the complete burial of the sub-Permian surface. Much of the detrital component of the thin (less than two metres thick) breccias of Saltom Bay may be regarded to be of this origin, although the occurrence of derived Lower Palaeozoic clasts and the well developed bedding and other sedimentary structures indicates transportation and deposition by stream processes.

The clasts of the sequences are, for the most part, set in a matrix of clay, sand and silt. Ferroan and non-ferroan calcite occur interstitially throughout the sequences and may be present in appreciable quantities, especially in lower parts of the sequence, where the voids may be infilled with drusy-mosaic sparite. Where a calcite cement is present, the clasts and grains nearly always have a well defined clay rim. Excluding the sand component, the average percentage of matrix for the brockram sequences is in the order of 27 per cent., of which 63 per cent. comprises the clay fraction (see Appendix Three). X-Ray Diffraction analysis indicates that illite is the dominant clay phase although kaolinite and possibly chlorite or vermiculite also occur (Figure 2.1). The clay fraction also incorporates appreciable quantities of ferric oxide, as shown by XRD and the predominant red colouration. The average percentage of sand present in the sequences is in the order of 23 per cent. although this value may exceed 80 per cent. in sandstone horizons and may be less than 10 per cent. in coarse breccia horizons. Sand grains are generally poorly sorted, except in sandstone horizons, and sub-angular to sub-rounded in outline. Quartz forms the dominant detrital sand grain although feldspar grains also occur in minor quantities, as do remnant Borrowdale Volcanic and Carboniferous limestone grains. The percentage of matrix is generally greater in lower parts of sequences than in the higher parts with a correspondingly reduced number of clast contacts. In thin section, the clasts and grains of the lower horizons often appear to be matrix supported.

Excepting mudflow deposits, most coarse-grained desert alluvium is recorded to be essentially free of interstitial clay when originally deposited (Walker et al., 1978; Waugh, 1978a), with the clasts forming the framework

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Figure 2.1. XRD traces of -2μ fraction of sample numbers F296-142.04 and F301-123.44 (NY01841035 and NY01551047 respectively) solvated with ethylene glycol, CoKα radiation.

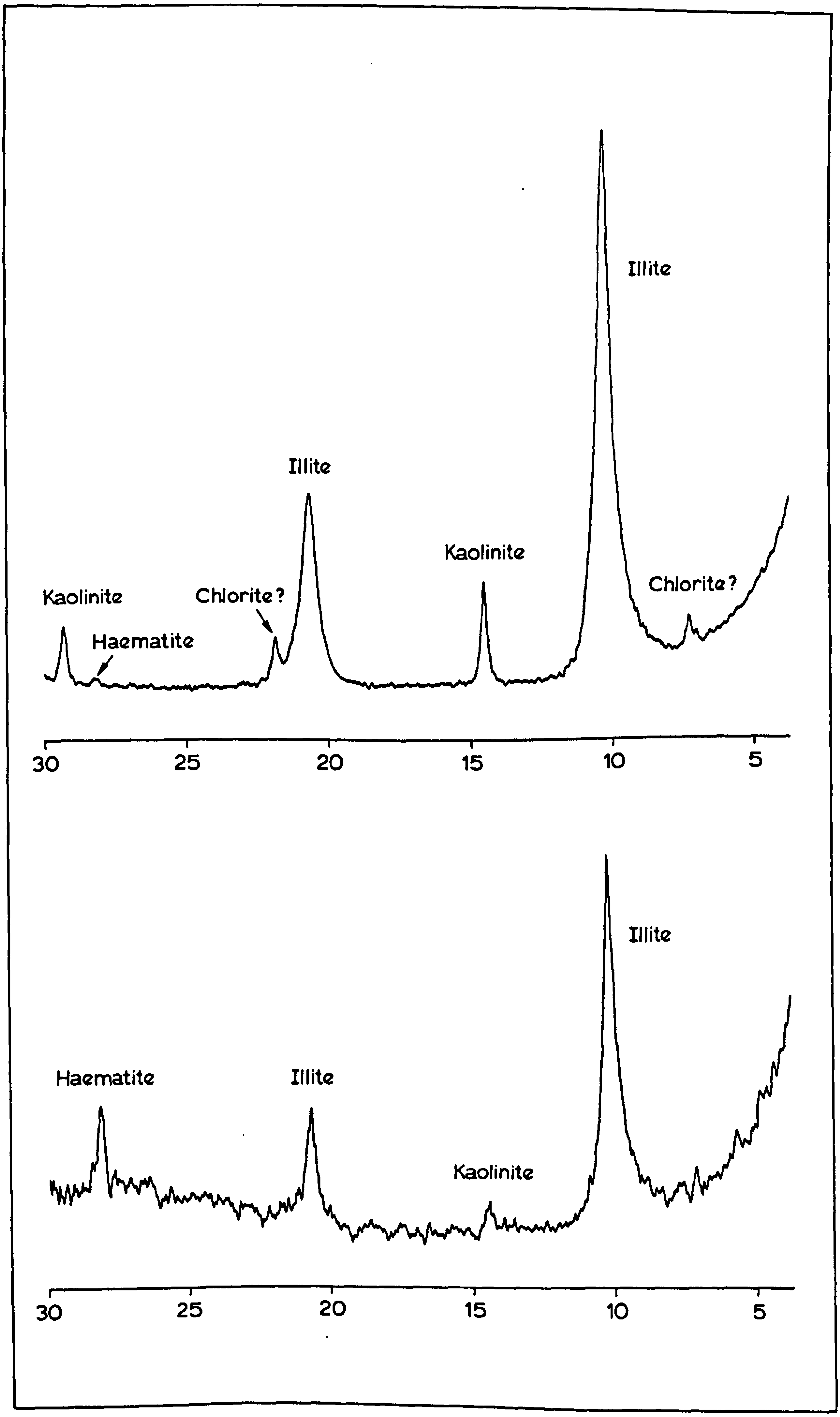
Illite
Illite
Kaolinite
Kaolinite
Chlorite ?
Chlorite ?
Haematite
30
25
20
15
10
5
Illite
Haematite
Illite
Kaolinite
30
25
20
15
10
5

Figure 2.1

of the deposits. Mudflows may occur interbedded with the alluvium, especially where source areas are clay rich, and may be recognised by the high percentage of clay matrix present in the deposit and the characteristically chaotic sorting and size distribution of the fragmental component (Steel, 1974a; Steel and Wilson, 1975). In addition, mudflow deposits typically exhibit a matrix supported texture.

The anomalously high percentage of clay matrix present within the western Cumbrian brockrams may lead to some confusion, therefore, in their field interpretation, especially as particular horizons with greatly in excess of 15 per cent. matrix usually retain some degree of sorting and fabric orientation. Petrographic analysis, however, has indicated that diagenesis has been exceedingly important in modifying the texture and mineralogy of the original deposit to the extent that the present textural features of the clay rich horizons may be almost entirely a result of post-depositional alteration.

DIAGENESIS OF THE BROCKRAMS

The petrography of the western Cumbrian sequences is significantly different from that of Edenside and has resulted in a distinct and more complex diagenetic evolution achieved under physical conditions which were essentially identical. The post depositional changes which have affected the Edenside sequences are mainly related to the formation of the calcite cement, and subsequent partial dolomitisation and dedolomitisation, and have been described by Waugh (1967). It is not proposed, therefore, to include a description of the Edenside sequences in the following section.

The most important diagenetic feature of the western Cumbrian sequences is the formation of the clay matrix which, as noted previously, may be of considerable significance in the interpretation of the depositional processes of formation. The amount of clay present within these deposits is generally such that it completely obscures most of the fine textural details which would otherwise be visible using the Scanning Electron Microscope. Two main mechanisms are proposed to account for the presence of the clay fraction; the early diagenetic mechanical infiltration of clay and the later stage formation of secondary clay from the *in situ* breakdown and replacement of the component clasts. As a general rule, it may be stated that the extent of diagenetic alteration decreases from lower to higher horizons of the sequences although heavily altered horizons may occur at any level.

Mechanical infiltration of clay

Where coarse grained alluvium is deposited by ephemeral fluvial processes, the sediment is recorded to be characteristically clay free (Walker *et al.*, 1978) with detrital fragments forming a framework deposit in which there is a high percentage of interstitial void. The majority of detrital grains in the present sequences, however, possess a pronounced clay rim which, in thin section, usually completely encircles the grain (Plate 2.1A). Clay coatings around detrital grains are recorded from deposits formed in several related depositional environments including alluvial fan and associated deposits (Colter and Ebbern, 1978; Hancock, 1978; Kessler, 1978; Walker *et al.*, 1978 and Waugh, 1978) and deltaic barrier bar sandstones (Hawkins, 1978). Where the interstices of the rock are completely filled with secondary clay, however, clay rims around grains may not be recognised. The use of the SEM, however, has revealed a characteristic texture for the clay of these rims in which the clay plates are usually aligned parallel to the surface of the detrital grain which they coat (Plate 2.1B, C and D). This feature was described by Crone (1974) and has been recognised by authors such as Walker *et al.* (1978) who note the similarity of this structure with that of the soil cutans described by Brewer (1964). The clay plates are typically concavo-convex in shape and are orientated with the concave margin facing the interstitial void. Walker *et al.* (1978) also record the occurrence of infiltrated clay in the form of meniscal bridges between adjacent grains, a feature which is possibly demonstrated in Plate 2.1C. In addition to the characteristic texture of the rim clays, the use of an Energy Dispersive Analysis of X-Rays (EDS) attachment in conjunction with the SEM has, in specific examples such as Plate 2.2A, revealed a slight compositional difference between the rim clays and that of the surrounding matrix. Whereas the general composition of the clays is shown, from XRD, to be illite, EDS has revealed the presence of considerable amounts of Fe in the interstitial illite (Plate 2.2B) which is not present in the rim clays (Plate 2.2C). It is not clear whether the iron is present as free Fe_2O_3 or whether it is bound in the crystal lattice substituting for Al^{3+} in the octahedral sites.

The distinctive texture and composition of the rim clays suggests an origin which is different to that of the Fe-rich clays and which must, because of their location around grains, predate the formation of the interstitial matrix clays. As it is extremely unlikely that the clasts could be deposited with a clay coating, the clay rims must have a secondary origin which precedes the lithification of the deposits (i.e. early post-depositional). Assuming the original sediment to have had open and connecting interstices, then the high porosity, in conjunction with the characteristically low water tables of alluvial fans, would present a favourable pathway for infiltrating surface waters. Following flash-flood events, when high energy, upper flow regime conditions would prevail on the surfaces of the fans and clays would be removed in suspension from the system, water flowing at low velocities would infiltrate into, and flow through, the previously deposited coarse alluvium. The suspended clays are carried into the sediment and are deposited when interstitial conditions are such that the flow of the water is interrupted or when high surface temperatures result in the sub-surface evaporation of the percolating water. The clay settles around detrital grains to produce clay rims which thus form a secondary matrix.

Walker et al. (1978) note that infiltrated clay occurs throughout comparable deposits in the south-western United States and north-western

Mexico although freshly deposited sediments are clay free. The concentration of mechanically infiltrated clay may be increased immediately above relatively less permeable horizons which reduce the flow of the percolating vadose waters but is generally greater at the proximal end of alluvial fans where recharge is most frequent.

The mechanical infiltration of clays into previously clay-free sediment may thus account for appreciable quantities of matrix now present in the brockram sequences and is responsible for both decreasing the textural maturity of the breccias and increasing the mineralogical maturity.

Dissolution of the detrital component

The detrital grains and clasts of the brockram sequences have undergone extensive alteration and dissolution during diagenesis to such an extent that original depositional textures are, in many examples, difficult to recognise. Many of the grains, especially towards lower horizons, have undergone complete dissolution and may only be recognised from the occurrence of clay or calcite pseudomorphs and/or relic rim clays (see Plate 2.8A). Where clay rims completely enclose clasts, however, dissolution cannot occur as the clay forms an impermeable barrier to diagenetic pore fluids. Borrowdale Volcanic fragments form the least stable lithological type recognised from the sequences and are the most extensively altered. Skiddaw Slate and sandstone fragments are more stable although post depositional disaggregation of the component grains is a common feature. Many of the quartz grains are corroded and indicate that silica was unstable during diagenesis. The extent of alteration is dependent on the original mineralogy of the clasts and the relative stability of those minerals with respect to Goldich's (1938) stability series (Walker et al., 1978). Hence, Borrowdale Volcanic clasts, which are more or less composed of framework silicates such as plagioclase feldspar, orthopyroxene and hornblende, are much less stable than the quartzose Skiddaw Slates and sandstones. The apparent relative stability of the limestone fragments is possibly due to an elevated pH in the circulating pore waters; a condition which may also account for the slight etching of the quartz grains.

Most commonly, the initial stages of dissolution occur marginally to the clasts which are progressively dissolved away until, in the final stages of the process, a void is produced (Plate 2.2D). Single crystals possessing marked crystallographic planes (e.g. the twin planes of plagioclase crystals) or grain contacts in rock fragments may form sites of preferential dissolution and result in delicately etched relic grains (Plate 2.3A and B; see also Plate 4.3A and B). Preferential dissolution at grain margins and along crystallographic planes has also been described by Walker et al. (1978), Waugh (1978) and Stanley and Benson (1979). All stages in the dissolution of the clasts may be recognised, from marginally corroded clasts, through partially dissolved clasts to completely dissolved clasts, although this process can usually be recognised only where the voids produced by dissolution have been subsequently infilled with ferroan or non-ferroan calcite or by redistributed replacement clay. Although it is unlikely that all void space produced by dissolution is pseudomorphically preserved by a later infill, the porosity of the brockrams, as seen in thin section, is almost nil and voids are rarely recognised. Small dissolution voids (of 20 microns diameter) are recognised from SEM samples, especially from the finer horizons interdigitating with the distal fan-breccias in western parts of the depositional basin (see, for example, Plate 4.3D). It is likely, therefore, that where infilling did not occur, post-depositional compaction of the sequences has resulted in the collapse of unfilled voids which are thus unlikely to be preserved.

One of the most important textural consequences of the dissolution of clasts is that the different lithological types are selectively reduced in size according to their chemical stability. Thus, the relatively more stable sandstone clasts may retain approximately their original volume, whilst severe corrosion of the Borrowdale Volcanics may lead to a considerable reduction in volume. This process, therefore, results in the diagenetic formation of considerable quantities of sand and silt sized particles which were not present in the original sediment. The production of silt grade material may be further increased where free quartz is present in the Borrowdale Volcanic clasts. Dissolution preferentially removes the alumino-silicate component whilst the quartz crystals remain more or less intact. Where replacement by clay occurs, the quartz grains are left floating in a secondary clay matrix (see Plate 2.4B). Considerable quantities of silt may be produced, in this manner, from the in situ breakdown of a single clast. The net effect of the process of dissolution, therefore, is to produce a marked decrease in the textural maturity of the deposits and, by preferentially removing the unstable grains, a corresponding increase in mineralogical maturity.

Replacement by clay

The in situ replacement of unstable detrital grains and clasts by secondary clay is the most extensive diagenetic process affecting the brockram sequences of western Cumbria and has resulted in the formation of considerable quantities of clay which were not present in the original sediment and which significantly increases the amount of interstitial matrix. In specific examples, the percentage composition of clay may exceed 30 per cent. and form a matrix supported breccia which would probably be interpreted as a mudflow deposit if field examination was not accompanied by detailed laboratory analysis.

X-Ray Diffraction analyses of clays separated from core samples (by mechanical disintegration and sedimentation from suspension) indicates illite to be the dominant clay phase although minor quantities of kaolinite and vermiculite (or chlorite) also occur. The tendency for diagenesis to promote the formation of illite rather than kaolinite or montmorillonite has been noted by Mason (1966). XRD traces also show the presence of significant quantities of Fe in the clay fraction which has been shown, from EDS analyses of SEM samples, to be restricted to the secondary illites. It is likely that the majority of the Fe is now present as free Fe_2O_3 as the clays commonly have a strong red colouration. In most cases, replacement clays are not readily distinguishable from mechanically infiltrated clays as compaction, resulting from overburden pressure, and, to a lesser extent, the displacive growth of calcite cement, has resulted in a redistribution of the clays. In any case, clay rims cannot be identified in thin section where replacement clay forms an interstitial matrix. It is possible, however, that the form of the clay plates, as seen with the SEM and illustrated in Plate 2.3C, may be characteristic of replacement clays and facilitate their identification. In this example, the plates possess a markedly irregular outline and delicately feathered edges which are not evocative of a mechanical origin. Authigenic clay with a comparable form has been illustrated by Colter and Ebbern (1978, Plate 2). Replacement clay also occurs as aggregates of individual clay plates, stacked together with the basal planes of the clays forming the contact between adjacent plates (Plate 2.3D).

As with the process of dissolution, the extent of replacement which affects the various clast types is partially related to their relative chemical stability. In consequence, it is difficult to distinguish between the two processes where dissolution voids have been subsequently infilled with redistributed clay. It is clear, however, that replacement sensu stricto is more likely to take place where the mineralogy of the original grains is such that its chemical breakdown can provide the silica, aluminium and potassium necessary for the formation of the illite. The aluminosilicate rich Borrowdale Volcanic clasts thus form the most extensively replaced lithological type with potassium probably being derived from the breakdown of K-feldspars and muscovite micas which may be present as minor constituents. Potassium may also be derived from the breakdown of muscovite micas present in Skiddaw Slate clasts.

Replacement of the detrital grains is commonly initiated at the periphery of the grains or, as observed by Blanche and Whitaker (1978) and Waugh (1978), may occur preferentially along crystallographic planes and ultimately results in their complete replacement and the formation of clay pseudomorphs. All intermediary stages of this process are recorded from the brockram sequences. As noted by Walker et al. (1978), the contacts between replacement clay and grains are commonly abrupt and transitional zones of partial replacement are not common. Occasionally, replacement may occur selectively within a particular clast (presumably only where pore waters are able to penetrate the clast) and replace the clast internally. In the example shown in Plate 2.4A, for instance, a replaced feldspar phenocryst occurs within a largely unaltered Borrowdale Volcanic clast of andesitic composition. In comparable, unaltered clasts, the feldspar phenocrysts are characteristically not corroded (see petrographic description) and it is considered, therefore, that this feature is a product of the diagenesis of the brockrams. An important feature of this process is the selective reduction in volume of grains and the formation of sand or silt sized particles which were not originally present. The formation of the finer grain sizes is facilitated where differential breakdown of the mineral components of individual clasts can occur, as with Borrowdale Volcanic grains. Both dissolution and, to a lesser extent, replacement, can result in delicately shaped grains which could not, conceivably, have survived transportation and sedimentation. An example is shown in Plate 2.4B, where a quartz vein within a Borrowdale Volcanic fragment is chemically more stable than the grain itself which is extensively replaced by clay. The vein thus forms a delicate projection from the clast surface which clearly cannot be an original feature. Small remnant silt sized quartz grains occur within the replacement clay.

The overall effect of the replacement of the detrital clasts, therefore, is to produce a further reduction in textural maturity whilst increasing mineralogical maturity. Its association with these changes, both replacement and dissolution (see, for example, Plate 4.3A and B) may produce finely etched grains with delicate surface textures which could not have survived transportation.

The formation of authigenic clay

Aluminium and silica, and presumably potassium, liberated by dissolution of the framework silicate minerals is ultimately precipitated as authigenic clay within the brockram sequences. In most instances, the authigenic clay forms an unrecognisable part of the interstitial matrix. Radiating crystals of authigenic illite lining an interstitial void are recorded, however, from one sample in thin section (Plate 2.4C) and can be recognised from a greater proportion of SEM samples. In the latter instance, the illite forms delicate hair-like threads (Plate 2.4D and Plate 2.5A and B), the composition of which has been identified by use of an EDS attachment. Authigenic illite with this texture has been recorded from comparable deposits by Wilson and Pittman (1977) Glennie et al. (1978), Hancock (1978), Sommer (1978) and Odom et al. (1979) and is suggested to have an a-axis elongation (Guven et al., 1980). Where illite of this type occurs, it characteristically contains Fe (Plate 2.5C). Examples are also recorded in which the clay develops the characteristic box-work texture described by Walker et al. (1978) (Plate 2.5D and Plate 2.6A) and which, they state, confirms an authigenic origin. A similar texture has also been recorded by Waugh (1978) from several British Permo-Triassic sandstones and illustrated by Taylor (1978, Plate 1, facing p.88).

Whereas Walker et al. (1978) record that chemically precipitated clay is an important contributor to the interstitial matrix of comparable deposits from south-western United States and north-western Mexico, thin section and

SEM studies of the brockram sequences suggest that clay of this origin may be relatively insignificant, unless clay possessing the texture illustrated in Plate 2.3C is of this origin, as the studies of Colter and Ebbern (1978) suggest. It seems likely, therefore, that considerable quantities of authigenic clay are present but no longer recognisable due to more or less complete intermixing with clays of other origin, as a result of compaction and other, associated effects.

The formation of authigenic calcite

Authigenic calcite is common as a mineral cement throughout the brockram fan-breccias of western Cumbria, but is especially important in lower horizons of the sequences where it may form in excess of 20 per cent. of the total volume of the rock. The calcite is generally coarsely crystalline and often develops a drusy-mosaic fabric in which crystal size increases away from detrital fragments (Plate 2.6B). The nucleation of this fabric may be related to any clast type, although, as in the examples illustrated in Plates 2.6B and 2.6D, the surfaces of detrital limestone grains appear to offer the most favourable sites for early crystallisation. Occasionally, the calcite may form large, poikilotopic plates which enclose rock fragments (Plate 2.6C). It is believed, however, that the formation of large crystal plates is, in part, related to the origin of the haematite ore bodies in the underlying Carboniferous Limestones (discussed in a proceeding section) and thus represents a late recrystallisation phase of an original drusy-mosaic fabric. The calcite may be either ferroan or non-ferroan, as shown by potassium ferricyanide and alizarin red S staining. There is no regular pattern to the development of the calcite types, although, within specific specimens, ferroan calcite often occupies inner zones of interstices and is thus the later phase of precipitation. The interstitial calcite is usually in the form of dog-tooth crystals which successively overgrow earlier crystals and which retain optical continuity with the earlier formed calcite crystals.

Syntaxial overgrowths around detrital grains are recorded only from uni-crystalline calcite fragments such as isolated crinoid ossicles.

The most common occurrence of calcite is as a pore filling cement and as an infill of dissolution voids. Less commonly, calcite can form as disseminated crystal plates within the interstitial, secondary clay matrix. A replacive origin for the calcite, as recorded by Heald (1956) from comparable deposits is not generally accepted for the brockrams as the association of clay-drusy calcite phases within the interstices (see following descriptions) is considered to be indicative of cavity growth rather than replacement. Where calcite is present as an interstitial cement, clay rims are usually preserved around detrital grains and the calcite crystals grow from the surfaces of the clay. Where high proportions of secondary clay are present within the sequences, clay phases may occur between overgrowth phases of calcite (Plates 2.6D and 2.7A). Where clay separates successive overgrowths, the optical continuity of the crystal is usually preserved and indicates that the clay does not form a continuous, unbroken layer within the crystals.

A maximum of four paired overgrowth phases, each separated by a clay phase, have been recorded from one interstice between two detrital grains. In samples of this type, the clay rims enclosing detrital grains may be particularly thick and are considered unlikely to be solely of a mechanical origin. The arrangement of these two stages suggests, therefore, that phases of calcite precipitation alternate with phases in which authigenic clay was precipitated onto the surfaces of the crystals or in which secondary clay was redistributed into the open interstices. However, examples are also recorded where matrix clay has been displaced by the crystallisation of calcite so that calcite-clay-calcite alternations are developed. Displacive crystallisation has also been recorded by Gilbert (1949), Glover (1963), Hancock (1978) and Dapples (1979). SEM studies of samples from the western part of the depositional basin have revealed the presence of small (five to six microns in length) calcite crystals which separate detrital grains from enclosing clay rims and which appear to have grown by dislocating the clay rim from the surface of the grain (Plate 4.5D and 4.8A). A similar example is recorded from the late Carboniferous sandstone reservoirs of the Bothamsall Oilfield (Hawkins, 1978) where rim clays are displaced by the crystallisation of quartz along grain-clay rim interfaces. Within the brockram sequences, many examples of dislocated clay rims may be observed in thin section although, in most cases, the entire rim is not translocated and some clay is left adhering to the surface of the grain. Commonly, only one part of the rim is prised away from the grain surface (Plate 2.7B) but some examples occur in which more or less the entire rims of grains adjoining the same interstice have been dislocated (Plate 2.7C). Occasionally, more complex structures may be observed (Plate 2.7D) in which the rim clays are not only prised away from the surface but have been distorted and/or ruptured during dislocation. However, although it can be demonstrated that some of the carbonate cement forms by displacement of the clay matrix, the continuity of the clay phases described previously and the regularity of their development would suggest a chemical, rather than a mechanical, control to be the more acceptable interpretation (see summary of diagenetic events).

Calcite is also common in the brockram sequences as an infill of voids produced by dissolution of framework silicate minerals; an occurrence which is noted from many comparable deposits (e.g. Kessler, 1978; Walker et al., 1978). Calcite may occur as infills of incompletely dissolved grains or may form pseudomorphs where grains have been completely removed (Plate 2.8A), although, in the latter instance, it is possible that these merely represent recrystallised limestone grains. Where detrital grains have been completely dissolved and there is an interstitial calcite cement, pseudomorphs can often only be recognised from the occurrence of infiltrated or authigenic clay rims. Where the dissolution of the clasts and precipitation of calcite occurs concomitantly, the precipitation of calcite within partial dissolution voids may halt the process of dissolution by reducing the permeability, especially if the clay rims are more or less intact. Contacts between partially dissolved grains and calcite are usually marked by a thin clay coating (see, for example, Plate 2.8D).

Examples with a rather more complex diagenetic history, which may involve multiple phases of dissolution and carbonate precipitation, are illustrated in Plate 2.8B, C and D. In Plate 2.8B, a peripherally altered Borrowdale Volcanic fragment is in part replaced by clay and, possibly, in part by alternations of clay and authigenic calcite with the two phases grading into one another. Although it is probable that a proportion of the clay (with the exception of lesser quantities of indistinguishable detrital clay) is replacive in origin, there is no evidence for assuming a replacive origin for the calcite. As with examples described previously (for example Plates 2.6B and D), the calcite is nucleated on the surface of the grain, has a drusy texture and is optically continuous. It is clear, therefore, that these clay-calcite alternations represent either the interstitial fabric, as described previously, or an infill of a dissolution void with a similar, and genetically related, fabric. The examples shown in Plate 2.8C and D, however, are more obviously the result of dissolution and subsequent infilling by authigenic calcite. In both examples, the calcite is a granular sparite which does not, for the most part, have a drusy texture. In Plate 2.8C, a Borrowdale Volcanic clast has been peripherally dissolved and the resulting dissolution void infilled with authigenic carbonate which incorporates diffuse zones of clay which are aligned sub-parallel to the grain surface and which probably originated from the <u>in situ</u> breakdown of the clast. The zonation thus represents alternating phases of dissolution and replacement with the latter being preceded by precipitation of carbonate within the void. It is possible that the chemical balance of the diagenetic fluids was such that the depletion or liberation of particular elements during one phase initiated the next (see discussion). From Plate

2.8D, at least three distinct phases of carbonate precipitation, the extent of each of which is marked by a thin clay coating, may be observed infilling a void formed as a result of the dissolution of a Skiddaw Slate grain. The original extent of the clast is shown by an infiltrated clay rim which is in contact with surrounding grains. Remnant silt-sized quartz particles occur floating within the calcite. That an initial phase of calcite precipitation can be proceeded by subsequent phases of dissolution of the clast and precipitation of calcite suggests that the diagenetic fluids were able to penetrate the authigenic calcite until a relatively late diagenetic stage, by which time the clast had been reduced to a small remnant core. An alternative explanation could be that permeability was re-established by dissolution of the calcite. It is also interesting to note that the products of dissolution were able to diffuse and, where Borrowdale Volcanic grains are dissolved (as in Plate 2.8C), were precipitated elsewhere, presumably as authigenic clay. Phases of calcite precipitation thus alternate with dissolution and, probably, an associated phase of clay precipitation. Assuming this process to be relatively widespread and the phases of calcite precipitation to occur more or less synchronously throughout particular parts of the brockram sequences, then it is more likely that the clay spacings described from interstitial calcite cements have an authigenic origin with the elements derived from the precipitation of dissolved ions in the pore fluids. One other feature of Plate 2.8D is that the original clast would have been in contact with surrounding clasts to form a clast supported texture. The present texture is, therefore, a product of diagenesis.

Where voids produced by dissolution are not infilled by secondary calcite (see Plate 2.2D), it is likely that compaction, due to overburden, results in collapse of the void and incorporation of the clay rims into the interstitial matrix. Evidence for this process is unlikely, therefore, to be preserved, except in isolated cases. Compaction may be mechanical or the result of hydrostatic pore pressure prior to the formation of the calcite cement. Increased hydrostatic pressure may be brought about simply by overburden pressure but, if the dominant authigenic phase was originally smectite or mixed layer smectite-illite (see discussion), could be caused by the release of large amounts of water from the clay structure during the conversion of smectite to illite (Blatt, 1979). Plate 2.9A shows an almost completely dissolved clast with an intact clay rim which probably owes its present form to the effects of compaction and/or the displacive growth of calcite. It is inconceivable that such a delicately shaped clast could have survived transportation and deposition under high energy, upper flow regime conditions. It is also apparent that if the effects of compaction were removed, then the rim clay would contact adjacent grains and would enclose only a small remnant of the original clast. Where clay rims are not completely intact but pore fluids have only limited access to dissolution voids (or where permeability is effectively sealed by some other mechanism), partial calcite fills may be formed prior to collapse of the rim clays (Plate 2.9B). This feature may also be produced by a post-compactional, second phase dissolution (and subsequent precipitation of calcite) of a remnant grain, although it is considered more likely that the extremely regular form of the example shown in Plate 2.9B is due to the recrystallisation of small calcite crystals which previously lined the rim clay and enclosed a dissolution void.

Where calcite lined dissolution voids are compacted, the clay rim may be fractured and access by the pore fluids re-established. The growth of the calcite crystals may then be reinitiated and cause the partially compacted rim clay to be prised apart (Plate 2.9B) and eventually lead to further fracturing (Plate 2.9B). As a result of the subsequent recrystallisation of the calcite, however, the successive growth phases of the crystals can no longer be established.

In many of the above examples (e.g. Plates 2.9A and B), collapsed or partly collapsed rim clays are in contact with interstitial calcite and, although these features have been interpreted as the result of compaction, it is possible that their collapse is due to the displacive formation of the interstitial cement.

The precipitation of authigenic calcite is, therefore, an extremely important process which is responsible not only for the induration of certain parts of the brockrams but also the preservation and recognition of many features which have been interpreted as the result of dissolution and, in some examples, compaction. The precipitation of calcite has also led to a significant change in the bulk chemical composition of certain horizons within the sequences, either by the addition or removal of Ca^{2+} ions. It is also clear that the growth of the calcite has caused displacement and, possibly, redistribution of the secondary matrix material to the extent that it is difficult to distinguish these effects from those of compaction.

Origin of the calcite cement

Although it is possible that intrastratal solution of plagioclase feldspars and other Ca bearing silicates such as clinopyroxenes may have liberated dissolved Ca^{2+} ions which subsequently combined with carbonated groundwaters to form considerable amounts of calcite, it is considered unlikely that this mechanism could account for the total percentage now present, especially as most of the Ca bearing silicates are contained by the Borrowdale Volcanic clasts which were fairly extensively altered prior to their incorporation into the brockram sequences. In any case, the composition of the plagioclase phenocrysts of the Borrowdale Volcanic clasts are mostly in the range of An_{10} to An_{30} and thus contain proportionally smaller amounts of Ca. The only likely source of Ca derived from intrastratal solution, therefore, is from the relatively minor detrital plagioclase feldspar component of the Skiddaw Slates and Carboniferous sandstones.

It is suggested, therefore, that the majority of the calcite, and particularly the high compositional percentages present in lower parts of the sequences, originated from the dissolution of Carboniferous limestone clasts and grains and the subsequent precipitation of authigenic calcite, following an increase in alkalinity of the interstitial pore fluids. Redistribution of Ca^{2+} ions in solution to higher parts of the sequences may have been effected largely by a mobilisation of the pore fluids in response to the stresses of the compacting sequences, although it is possible that the occurrence of limestone clasts higher in the sequences may have been originally more extensive. Minor quantities of calcite might also have been directly precipitated from downward percolating vadose waters containing Ca^{2+} ions derived from desert loess or dissolved in rainwater (see Chapter Four).

Dissolution of authigenic calcite

The dissolution of calcite represents the final post-depositional alteration which is recorded from the brockram sequences and which is directly attributable to diagenesis sensu stricto. In isolated examples, authigenic calcite has been dissolved due to a change in pore water chemistry and the dissolution voids subsequently infilled with authigenic clay. From Plate 2.9C, several rhomb shaped masses of secondary clay can be observed and are interpreted as clay pseudomorphs after calcite. Occasionally, dissolution may occur preferentially at the contact between successive overgrowth phases which may be, themselves, selectively dissolved (Plate 2.9D), presumably according to minor variations in the structure or chemistry of the crystal.

Occurrence and origin of haematite

Haematite is recorded from both XRD traces of the clay fraction and EDS analyses of SEM samples. The occurrence of discrete crystals, which can be observed with the SEM is, however, uncommon and generally restricted to confined but apparently random horizons within the sequences. More commonly iron oxide imparts a red colouration to the detrital grains and matrix clays. In the latter instance, however, the haematite is either amorphous or occurs in crystals that are too small to be resolved by the SEM (c.f. Walker et al., 1978). Where haematite crystals are observed, the sequences are usually heavily haematised and replaced. It is believed that this haematisation is the result of later, non-diagenetic ferriferous brines penetrating the brockram sequences from the overlying St. Bees Sandstones and ultimately resulting in the emplacement of the ore bodies in the underlying Carboniferous Limestone, following the model erected by Rose and Dunham (1977). The period of mineralisation is dated as either late Triassic or late Cretaceous-Tertiary (Rose and Dunham, 1977). By ore genesis times, therefore, it is likely that the porosity of the brockrams was exceedingly low as a result of the diagenetic formation of interstitial matrix. The most favourable, and possibly the only, pathways available for the descending ore fluids would be restricted, therefore, to any intra-Permian or post-Permian fault planes affecting the sequences which might thus control the apparently random occurrence of the haematised breccia.

Haematite of this origin is illustrated in Plates 2.10A to 2.12B. The haematite occurs commonly as radiating, concentric layers of fibrous crystals with a botryoidal form (Plate 2.10) or as masses of hexagonal tabular crystals (Plate 2.11A, B and C) which may line voids, infill dissolution voids, or, perhaps, replace detrital grains (Plate 2.11D to Plate 2.12B). A remnant (?) infiltrated clay rim is preserved in the example shown in Plate 2.12B.

Although the exact extent of haematisation within the brockram sequences is uncertain, it is clear that the red colouration which is ubiquitous to these deposits cannot wholly be attributed to this process. The origin of the red colouration within deposits comparable to the fan-breccias of western Cumbria is generally attributed to the breakdown of iron-bearing minerals such as biotite, augite and hornblende and, where conditions favour oxidation, their subsequent precipitation (Walker, 1967a; 1967b; Glennie et al., 1978; Hubart and Reed, 1978; Walker et al., 1978; McPherson, 1980). Initially, the iron precipitates as a ferric hydrate such as limonite, which imparts a yellowish-brown stain to the sediments, but in time converts to haematite which produces the bright red colouration characteristic of many ancient continental deposits (e.g. Walker, 1967a; Berner, 1969; Folk, 1976). Turner (1979) has suggested a similar origin for the red colouration of certain marine shales. A primary origin for the red colouration as advocated by Krynine (1950) is no longer accepted (e.g. Walker, 1967b; Walker et al., 1967; Walker and Honea, 1969; Walker, 1976). The model proposed by Rose and Dunham (1977), which explained the red colouration of the St. Bees Sandstone in terms of staining by material derived from the weathering of lateritic soils developed at end-Carboniferous times, is not accepted, therefore, for the Cumbrian deposits (e.g. Ixer et al., 1979).

Direct evidence for a diagenetic reddening, as described by Walker (1967a) from the Sonoran desert of north-eastern Baja, California, is not observed from the brockram deposits as convincing genetic sequences can no longer be recognised and the effects of post-diagenetic haematisation have, in places obscured and/or replaced the pre-existing diagenetic textures. However, it seems likely that Fe^{2+} ions released into solution by intrastratal dissolution of the detrital iron-bearing silicate and non-silicate component of the Borrowdale Volcanic clasts were, as with the other released elements, ultimately precipitated as an oxide within the interstitial matrix. Possible authigenic haematite is recorded from one sample from British Steel Borehole No. 296 (at NY01841035, near Ullcoats) and occurs as tabular crystals (Plate 2.12C to Plate 2.13A) within an otherwise normal diagenetic clay matrix. Further, but inconclusive evidence is provided by the marginal reddening of the clasts and, in particular, the alteration of chlorite to haematitic clay at the periphery of Borrowdale Volcanic clasts. It is considered unlikely that these clasts were reddened or marginally altered to clay prior to deposition, especially as in situ dissolution has, in many examples, reduced these clasts to a small remnant core.

Occurrence and origin of the sulphates

Minor quantities of secondary gypsum and traces of barytes are recorded from the brockram sequences. Occurrences of anhydrite and barytes have been described from Tertiary sandstones of California by Merino (1975) and occurrences of late diagenetic barytes by Hay (1957) and Von Engelhardt (1967) from Middle Eocene sandstones of Wyoming and Middle Keuper sandstones of the Stuttgart area, respectively. In the brockrams, both minerals occur within the interstitial matrix but whereas the former is of limited but widespread occurrence, the latter is always associated with haematisation.

Gypsum may be present as disseminated masses of radiating crystal laths within the interstitial matrix clay or, less commonly, may form as a late stage cement infilling interstices which are incompletely filled by authigenic calcite (Plate 2.13B). Occasionally, gypsum may occur in vein-like fractures associated with heavily haematised parts of the sequence. The compositional percentage of gypsum may attain values in the order of eight *per cent.* within individual samples, as determined by modal analyses, but is generally much less than this value and in the range of nought to two *per cent.* The average for the brockrams as a whole is probably in the order of less than one *per cent.*

Barytes has not been recognised from thin section but its occurrence has been demonstrated by use of the SEM and EDS attachment (Plate 2.13C and D). The occurrence of barytes is always associated with haematisation of the sequences.

The origin of these minerals may, in both cases, be attributed to the precipitation of calcium, barium and sulphate ions from solution in an oxidising environment, with the elements derived from the *in situ* breakdown of Borrowdale Volcanic clasts. Anderson (1974) records the potential sulphur content of andesitic magmas to be in the range of 0.10 to 0.67 *per cent.* and El Sokkary and Abdel Monem (1977) have described an occurrence of barytes for which they suggest the breakdown of granitic rocks to have provided the required barium. Although the latter example is not directly applicable to the brockram sequences, it is possible that barium was derived from the dissolution of orthoclase feldspars present as minor constituents of the Borrowdale Volcanic and Skiddaw Slate clasts. Calcium may be derived from a variety of sources which have been considered previously. It is possible, therefore, that both gypsum and barytes occur as authigenic minerals within these sequences although the association of barytes and haematite, within both the brockrams and the mineralised Carboniferous Limestone, suggests that the occurrence of the latter may be at least partially controlled by mineralisation with Ba^{2+} and SO_4^{2-} ions derived from the ore fluids. Similarly, the occurrence of gypsum may be partly related to the emplacement of the ore but its disseminated occurrences throughout the brockram sequences suggests that a diagenetic origin is the more likely.

If a diagenetic origin for the gypsum is accepted, then its precipitation marks the final phase in the diagenetic history of the brockram sequences of western Cumbria which has been recognised.

SUMMARY OF THE DIAGENETIC SEQUENCE

From the diagenetic associations and the relationships of the various authigenic mineral assemblages, a generalised diagenetic sequence can be established for the brockrams:

1 Mechanical infiltration of clay by percolating vadose waters which were probably slightly acidic from dissolved CO_2. Na^+, K^+, Ca^{2+}, Mg^{2+}, Fe^{2+}, $Al(OH)_4^-$ and H_4SiO_4 ions in weak concentration in solution from leaching of the source area. The production of thick clay rims inhibits further diagenesis by reducing porosity, particularly where unstable grains are completely encased.

2 Intrastratal dissolution of the unstable aluminosilicate and ferro-magnesian minerals by neutral or weakly acidic percolating fluids. Concentration of

dissolved ions increases in the pore fluids.

$$KAlSi_3O_8 + 2H_2O \rightarrow Al(OH)_4^- + K^+ + 3SiO_2$$

Orthoclase

$$3Na(AlSi_3O_8).2Ca(Al_2Si_2O_8) + 14H_2O \rightarrow 7Al(OH)_4^- + 3Na^+ + 2Ca^{2+} + 13SiO_2$$

Andesine

$$CaMgSi_2O_6.CaFeSi_2O_6 + 4H_2O + 6CO_2 \rightarrow Fe(OH)_2 + 2Ca^{2+} + Mg^{2+} + 6HCO_3^- + 4SiO_2$$

Augite

$$MgSiO_3.FeSiO_3 + 2CO_2 + 2H_2O \rightarrow Fe(OH)_2 + Mg^{2+} + 2SiO_2 + 2HCO_3^-$$

Hypersthene

3 Replacement of the unstable framework silicates by Fe-rich clay with minor amounts of early phase carbonate precipitation. This stage is reached when the concentrations of dissolved ions attains equilibrium with respect to illite which might be partly determined by a reduction in solubility of silica and alumina where both are in solution (Hancock, 1978).

$$5KAlSi_3O_8 + 4H_2O + 4CO_2 \rightarrow KAl_4(Si_7AlO_{20})(OH)_4 + 4K^+ + 8SiO_2 + 4HCO_3^-$$

Orthoclase Illite

$$15Na(AlSi_3O_8).10Ca(Al_2Si_2O_8) + 7K^+ + 28H_2O + 28CO_2 \rightarrow$$

Andesine $$7KAl_4(Si_7AlO_{20})(OH)_4 + 15Na^+ + 10Ca^{2+} + 8SiO_2 + 28HCO_3$$

Replacement of the aluminosilicates is accompanied by the alteration of muscovite and chlorite to illite (which provides excess K^+ and Fe^{2+} ions):

$$5KAl_3Si_3O_{10}(OH)_2 + SiO_2 + 2H_2O \rightarrow KAl_4(Si_7AlO_{20})(OH)_4 + Al(OH)_4^- + K^+$$

Muscovite

$$5Mg_4Al_2Fe_6Al_2Si_6O_{20}(OH)_{16} + 4K^+ + 36CO_2 + 16H_2O \rightarrow 4KAl_4(Si_7AlO_{20})(OH)_4 + 2SiO_2 + 20Mg^{2+} + 30Fe(OH)_2 + 36HCO_3^-$$

Chlorite

and by the precipitation of Fe^{2+} as an hydrated ferric oxide precursor from the oxidation of an Fe^{2+} complex in solution. Hydrated oxide later converts to crystalline haematite and imparts the characteristic red colouration to sequence.

$$Fe(OH)_2 + OH^- \longrightarrow FeOOH.H_2O + e^{-1}$$

$$2FeOOH.H_2O \longrightarrow Fe_2O_3 + 3H_2O$$

Most of the above hydration reactions liberate excess silica which would normally form syntaxial overgrowths on detrital quartz grains (Waugh, 1978). As authigenic quartz is not recorded from the brockrams, however, it would appear that excess SiO_2 is diffused from the system or is the limiting factor in the formation of authigenic illite (see below). Diffusion is considered the more likely process, especially as the solubility of silica is recorded to be markedly increased where the pore fluids contain more than 1 molar Na (Beach and King, 1978; Beach, 1979).

4 Precipitation of authigenic Fe-rich illite as clay coats around grains and as linings to voids. Excess $Al(OH)_4^-$ and K ions combine with silica produced by dissolution and replacement:

$$K^+ + 5Al(OH)_4^- + 7SiO_2 + 4CO_2 \longrightarrow KAl_4(Si_7Al\,O_{20})(OH)_4 + 4HCO_3^- + 6H_2O$$

Precipitation of clay may halt the dissolution process by further reducing porosity. Compaction and associated high pore fluid pressures collapse dissolution voids and possibly cause partial redistribution of clay.

The net effect of the precipitation of illite and the alteration of aluminosilicates and ferromagnesians is to liberate HCO_3^- into solution and thus increase the alkalinity of the solution which leads to the precipitation of Fe^{2+} and ultimately of Ca^{2+} ions, by combination with OH^- and HCO_3^- ions. The buffering of the pore fluid towards a neutral pH has been noted by Beach and King (1978) and Beach (1979) for similar, low grade metamorphic reactions during the formation of pressure solution stripes.

5 Precipitation of remaining Fe^{2+} as hydrated ferric oxide precursor of haematite.

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[1] This part equation probably occurs as a component of a more widespread diagenetic reaction.

6 Precipitation of calcite cement. Excess Ca^{2+} ions from early stage dissolution of limestone clasts and aluminosilicates combines with HCO_3^- in carbonated pore waters:

$$Ca^{2+} + 2HCO_3^- \longrightarrow CaCO_3 + H_2CO_3$$

The alternation of clay-calcite cement in the interstices indicates fluctuating pore fluid chemistry and a repeated development of stages 4 and 6. The net effect of carbonate precipitation is to increase the acidity of the solution which may cause the reversion to illite formation until this reaction increases alkalinity and equilibrium with calcite is re-established. It is probable that equilibrium is not maintained by these reactions, however, as the brockrams are an open system which allows the diffusion of excess SiO_2 and the loss of Mg^{2+} and Na^+ ions. It is considered, therefore, that the reversion to illite precipitation is likely to be related to an external control, such as flushing with CO_2 enriched waters.

7 Minor dissolution of calcite and precipitation of clay in dissolution cavities. This stage is not widespread and represents a chemical change comparable with stage 6, although requiring rather more acidic solutions.

8 Precipitation of gypsum and barytes. The formation of these minerals probably reflects a change in the chemistry of the pore fluids, with decreased HCO_3^- concentrations and increased SO_4^{2-} concentrations. Excess Ca^{2+} may originate from the intrastratal hydrolysis of aluminosilicates but it is considered likely that the Ba^{2+} and SO_4^{2-} ions were derived from sulphate rich ferriferous brines. This stage is probably very late, therefore, and associated with the mineralisation of the Carboniferous Limestone underlying the brockrams.

$$Ca^{2+} + SO_4^{2-} \longrightarrow CaSO_4$$
$$Ba^{2+} + SO_4^{2-} \longrightarrow BaSO_4$$

The above reactions (with the possible exception of stage 8) are considered to equate with the redoxomorphic and locomorphic stages of Dapples (1962).

The phyllomorphic stage is not recognised from the brockrams unless the presence of illite is attributed to the recrystallisation of a pre-existing clay phase (see discussion).

DISCUSSION

The diagenetic histories of many sequences comparable to the west Cumbrian brockrams has been well documented in recent years by authors such as Walker (1967a), Walker et al. (1978), Davies et al. (1979), Ixer et al. (1979), Stanley and Benson (1979) and Al-Shaieb et al. (1980) who attribute the alteration of these deposits to reactions between unstable constituents of the rock and percolating water. A prerequisite for diagenesis, therefore, is an initial permeability, the degree of which will be important in determining the amount and extent of diagenesis which can affect a particular rock. As water flows through the sediment, the most unstable grains are dissolved and the fluids become enriched in dissolved ionic species. Eventually, the concentration of ions reaches equilibrium with respect to a particular mineral which begins to precipitate and which may be important in the lithification of the rock. The specific mineral which forms at any point in the diagenetic time continuum is a function of the chemistry of the pore fluids (Davies et al., 1979) which is itself a function of:

1. The original composition of the water. The hydrogen ion potential (pH) and oxidation-reduction potential (Eh) of this fluid are critical in determining the initial diagenetic stage and thus partially control the eventual diagenetic sequence, even though both potentials, particularly pH, are modified during diagenesis. The importance of early oxidation and reduction reactions has been stressed by Dapples (1967).
2. The composition of the rock. The most unstable grains have the earliest and greatest impact on diagenesis (Davies et al., 1979).
3. The length of time the pore fluid has remained in the rock (residence time).

4. The diagenetic stage. Preceding diagenetic stages will result in a relative concentration of specific ions not involved in earlier reactions.

The authigenic mineral assemblages recorded from the brockrams may thus be considered to be the product of these factors with the diagenetic sequence reflecting the chemical evolution of the pore fluids. The relative development of a particular stage, however, may be critical in determining the final diagenetic sequence and may inhibit later diagenetic processes by causing an early reduction in porosity. Where large quantities of mechanically infiltrated clay are present, for example, permeability may be so reduced that the processes of dissolution and replacement can never occur. It is possible that the apparent reduction in the extent of diagenesis towards the top of the sequences is due to an early filling of the pores by infiltrated clay. The infiltration of large quantities of clay may be related to conditions of slightly reduced aridity which are suggested for western Cumbria by the establishment of the fluviatile regime of the St. Bees Sandstone following the deposition of the evaporite coastal plain sediments of the Upper Permian.

As far as the authigenic mineral assemblage of the brockrams is concerned, it is clear that more or less all of the required elements (with the probable exception of Ba^{2+} and SO_4^{2-} ions) may be derived from the hydrolysis of the detrital grains, particularly of the aluminosilicate component. Waugh (1978) records that a greater amount of the contained elements will be released if grains are dissolved than if grains are replaced by clay although significant proportions of ions may be released where replacement is extensive. Although Waugh (1978) and other authors (e.g. Hancock, 1978; Walker et al., 1978) suggest that hydrolysis is accomplished by alkaline fluids, it seems more likely that the processes of dissolution and replacement, in this case, operated under weakly acidic conditions (see summary of diagenetic events) and themselves generated the alkaline conditions necessary for the precipitation of authigenic clay and later calcite cement.

Raam (1968), for example, records the preservation of pyroxenes and amphiboles to be favoured by alkaline solutions and Nagtegaal (1978) suggests that feldspars tend to be stable in alkaline solutions and unstable in neutral to acid solutions.

The formation of authigenic clays has been widely documented and the conditions required for their formation has been discussed by authors such as Carrigy and Mellon (1964), Hancock (1978), Waugh (1978), Hoffman and Hower (1979), Odom et al. (1979) and Tillman and Almon (1979). Authigenic clay phases recorded by these and other authors include smectite (Sheppard and Gude, 1969; Davies et al., 1979; Stanley and Benson, 1979), mixed layer smectite-illite (Walker et al., 1978), kaolinite (Heald, 1950; Blanche and Whitaker, 1978; Hancock, 1978; Almon and Davies, 1979), chlorite (Odom et al., 1979) and illite (Thomas, 1978; Waugh, 1978b; Odom et al., 1979; Tillman and Almon, 1979) of which only the illite phase is present in any appreciable amount in the brockrams. The formation of both illite and smectite is recorded to require alkaline conditions (Odom et al., 1979) although the formation of illite rather than smectite is favoured by high K^+ concentrations in the pore fluids (Mason, 1966; Tillman and Almon, 1979), a higher pH than that required for smectite formation (Odom et al., 1979) and, where the chemistry of the pore fluids does not preferentially favour either phase, by a higher value of free energy of formation (Thomas, 1978). Pore fluids with too high an alkali ion activity, however, are suggested to promote the formation of zeolite and feldspar rather than clay (Sheppard and Gude, 1968; 1969; 1973). Tillman and Almon (1979) also suggest that the relative concentration of Na^+ and K^+ is important in determining the clay phase, as is, presumably, the concentration of Mg^{2+} as both Na^+ and Mg^{2+} are required for the formation of smectite. The presence of illite in the brockram sequences thus suggests saline, K^+ rich and Mg^{2+} and Na^+ depleted pore fluids which were in equilibrium with illite rather than smectite; an interpretation which is not supported by the apparent paucity of K^+ bearing minerals in the brockram detritus compared with the relative abundance of Mg^{2+} and Na^+ bearing silicates in the Borrowdale Volcanic component.

Illite is also recorded to occur as an alteration product of a pre-existing kaolinite (Hancock, 1978; Hancock and Taylor, 1978; Sommer, 1978) or smectite/mixed layer smectite-illite clay phase (e.g. Hoffman and Hower, 1979; Aoyagi and Kazama, 1980). The presence of illite thus need not reflect the original chemistry of the pore fluids. Despite being the stable phase of both smectite and mixed layer smectite-illite, however, it is reported that the smectite phase should persist more or less indefinitely under surface temperature and pressure conditions (Hoffman and Hower, 1979) and that the loss of expandable layers is achieved largely as a result of the elevated temperatures associated with burial diagenesis (Perry and Hower, 1970; Eberl and Hower, 1976; Blanche and Whitaker, 1978; Boles and Franks, 1979; Hoffman and Hower, 1979; Aoyagi and Kazama, 1980; Lahann, 1980). The temperature associated with burial has been measured from several cored shale sequences including one of Mesozoic-Tertiary age in Montana, U. S. A. (Hoffman and Hower, 1979; Table 1, p.58) where it was demonstrated that the complete conversion of smectite to illite occurs in the temperature range 200 - 300°C. These values compare to those obtained experimentally by Eberl and Hower (1976) and approximate to those measured from Cretaceous and Tertiary argillites by Aoyagi and Kazama (1980) who record complete conversion to be achieved at a temperature of 140°C and a pressure of 920 kg cm^{-2}. The presence of well crystallised illite within the brockram sequences may thus be argued to indicate a formational temperature above 200°C although it is unlikely that the temperature exceeded 250°C as Hoffman and Hower (1979) record that illite converts to 2M dioctahedral mica above this temperature. If such a formation could be accepted for the brockrams, then this conversion would be effected at the upper extreme of the arbitrary 0 - 200°C range quoted by

Blatt (1979) as defining the physical conditions included in diagenesis and may be argued to lie within the field of low grade metamorphism. However, if a temperature of $200^{0}C$ is accepted for the crystallisation of illite, then serious problems arise with respect to the mechanism of formation. The simplest method for generating elevated temperatures is deep burial, but a temperature of $200^{0}C$ requires a depth of burial of 10 kilometres (assuming a temperature gradient of $20^{0}C/km$); a figure which is unacceptable for western Cumbria. It is also apparent that no other satisfactory explanation for such temperatures can be invoked for the area as there appear to be no grounds for assuming a higher than normal temperature gradient or an abnormally high heat flow generated by any other mechanism. In addition to the above considerations, Lahann (1980) has suggested that any significant temperature increase associated with illitisation will raise the equilibrium solubility of silica and thus result in the formation of authigenic silica bearing phases (which may derive silica solely from the conversion of smectite to iliite). As determined previously, however, such an association has not been recognised from the brockrams.

Although it seems that the formation of illite is temperature dependent in certain deposits and that diagenetic sequences may similarly be related to increasing temperature and overburden pressure, as proposed by Galloway (1974; 1979) for arc derived sediments of the north-eastern Pacific, evidence for such a control in western Cumbria is lacking and a satisfactory model, based on these considerations, cannot be proposed. The formation of illite within the brockrams can, therefore, only be controlled by the chemistry of the pore fluids although this explanation does not account for the apparent complete loss of Mg^{2+} and Na^{+} ions from the system and the formation of this clay phase rather than smectite.

CONCLUSIONS

The dominant diagenetic processes recognised from the brockram sequences are the mechanical infiltration of clay, dissolution and replacement of unstable grains and the precipitation of dissolved ions as an assemblage of authigenic minerals in the interstices of the rock and in dissolution voids. For the most part, these changes can be considered to have occurred within a closed system[1] where the elements of the authigenic minerals were derived from the intrastratal hydrolysis of pre-existing detrital grains which were not in equilibrium with the prevailing conditions. These reactions represent, therefore, stages in an attempt to attain absolute equilibrium with a new environment which can probably never be achieved (see Curtis, 1978; Walker _et al._, 1978) and which have had a considerable effect on the texture and mineralogy of the original deposit.

These effects can be summarised as an increase in the mineralogical maturity of the deposit, an associated decrease in the textural maturity and a drastic reduction in porosity. The mineralogical maturity is increased by the mechanical infiltration of stable clay minerals and by the breakdown of the unstable detrital grains and the subsequent precipitation of the elements thus released as stable authigenic minerals. Textural maturity is decreased by a combination of all the diagenetic processes, which preferentially remove or reduce in size the least stable grains (and which may lead to the formation of sand or silt sized material) whilst leading to the formation or incorporation of considerable quantities of clay which was not present in the original deposit. Porosity is reduced by the infiltration of clay and by the formation of authigenic minerals although dissolution may secondarily increase porosity

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1 The term is here used to imply an isochemical system rather than a closed system _sensu stricto_. For carbonate, for example, Bjorlykke _et al._ (1979) record that approximately 10^5 pore volumes are required to precipitate 1 cm^3 of cement.

where the voids are not subsequently filled. Primary and secondary porosity has been more or less totally destroyed by authigenesis in the brockram sequence, although, occasionally, intercrystalline voids are present within the calcite cement which may represent the final porosity of the deposit following the complete destruction of permeability. Compaction, due to overburden pressure, also reduces original porosity by plastic deformation of ductile sand grains and, to a lesser extent, by fracture of brittle grains and redistribution of secondary clay.

CHAPTER THREE

THE INTERBEDDED BRECCIA AND REWORKED PENRITH SANDSTONE SEQUENCES OF THE EDEN VALLEY

INTRODUCTION

The interbedded breccia and sandstone sequences occurring within the Penrith Sandstone formation and which form part of the stratigraphic unit elsewhere termed the Upper Brockram (e.g. Waugh, 1967) are considered to be distinct from the Penrith and Stenkrith Brockrams both in their provenance and precise mode of formation and have, accordingly, been distinguished from these deposits for the purposes of this research.

The distribution of this facies has been discussed by Waugh (1967, p.16) and suggested to extend from Kirkby Stephen in the south (where the Upper and Lower Brockrams have been recorded to fuse), through localities such as the River Belah (NY795121), Swindale Beck (NY795145), Hayber Gill (NY752165), George Gill (NY717189) and Hilton Beck (NY714202) to Flakebridge Wood (NY703220). It is, however, by no means certain that the breccias form a continuous sedimentary unit (as indicated in Figure 1.5 which is based on Versey, 1939, p.276 and Waugh, 1967, Enc. 1) as exposure in this part of the Eden Valley is very restricted. It is also clear that sequences exposed near Low House, Armathwaite (centred on NY515486) in the northern part of the Eden Valley are directly comparable in their sedimentological characteristics to the Upper Brockram of the southern part of the valley and have an identical mode of formation, even though they are widely separated, geographically, from this area (see Figure 1.1). As a product of poor exposure, no attempt has been made to modify pre-existing maps of the distribution of this facies and this study concentrates only on the sedimentological aspects of these sequences.

Unlike the breccia dominated alluvial fan deposits described in Chapter One, the generalised sequence exposed is that of a series of poorly sorted breccia[1] units which occur within a sequence of medium to coarse grained, thickly bedded brick-red coloured sandstones with abundant gritty horizons and associated mudstone horizons and mudstone clast breccias. Almost all exposures show evidence of channelling (e.g. Hayber Gill) or wedging (e.g. Hilton Beck) of the breccia units whose formation appears everywhere to be more or less independent of the sandstones in that the breccias rarely grade into the sandstones. Whereas the breccias are typified by an overall lack of well developed sedimentary structure (and are similar in this respect to other brockram exposures), the sandstones display a variety of structures of which the most extensively developed are planar (flat) lamination or bedding and cross-bedding. Cross-bedded sets may be in the form of either festoon-shaped scour infills or tabular and wedge planar (terminology of McKee and Weir, 1953) sets on the scale of dunes (with set thicknesses exceeding one metre in places) but is nowhere comparable to the large scale trough cross-stratification exposed elsewhere in the Eden Valley (see Chapter Five). The association of breccia and sandstone thus indicates marked differences in flow conditions and competence of the transporting medium or mediums (or, conversely, the availability of the fragmental sizes) and is clearly related to a style of sedimentation which could accommodate the periodic adjustment to these contrasting conditions.

Sedimentary structures

The main characteristics of the interbedded sequences, as outlined previously, are that they comprise an association of breccia and sandstone which are, in many respects, similar in nature to the alluvial fan brockrams

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1 Whilst it is clear that these deposits should more correctly be designated as conglomerates, it is proposed to retain the term breccia (as defined in Chapter One) in order to avoid any inconsistency in nomenclature.

and Penrith Sandstone respectively. In addition to these lithologies, well developed mudstone horizons and mudstone clast breccias are a feature of many of the exposures and commonly occur within the sandstones and at the bases of the major breccia units.

Few sedimentary structures are developed within the breccias and, with the exception of one well developed, tabular cross-bedded set at the base of a breccia unit at the northern end of the Low House exposures (see Figure 3.5), are generally restricted to clast fabrics such as a bedding parallel alignment and imbrication. Excluding the often markedly erosive and channelled basal contacts, bedding and internal stratification is very poorly developed and component beds can usually be recognised only where the occurrence of irregularly eroded and impersistent sandstone beds or conspicuous and abrupt changes in clast size indicate deposition in several successive stages. As with the alluvial fan breccias, sorting and grading is poorly developed and is generally restricted to the coarsest grade material only.

Apparent derivation directions measured from imbricated clasts and cross-bedding generally indicate transport from the north-east or north-north-east which concurs with the direction determined by Waugh (1967) and the inferred provenance of these deposits as determined by Kendall (1902), Holmes and Harwood (1928) and Dunham (1932). For the most part, it is difficult to determine the thickening direction of the brockram wedges but where such data can be observed, as in Hayber Gill and Hilton Beck, it is suggested that the wedges thin towards the south-west and may even be completely wedged out in this direction (Waugh, 1967), similarly implying a derivation from the north-east.

Sandstones associated with the breccias display a variety of structures of which the most prominently developed are flat bedding and tabular cross-bedding in which set thicknesses of 50 to 80 centimetres are relatively common. Structureless sandstone units and small scale scour structures are also recorded from these sequences. These characteristics, together with the evidence of the closely associated mudstone horizons and incorporated mudstone and gritty dolomite clasts, suggest the most likely depositional processes to be those of a fluviatile environment although it is possible that certain horizons may represent low angle aeolian sand sheets of the type described by Fryberger (1979). Palaeocurrent directions estimated from two-dimensional exposures of cross-beds concur with a transport direction from the Pennine fells flanking the eastern margin of the Eden Valley.

Petrography and provenance

Despite only a limited superficial resemblance between the interbedded sandstones and the cross-stratified Penrith Sandstone *sensu stricto* in terms of depositional structures, the dominant detrital components of both rock types are similar in comprising well rounded quartz grains with a frosted surface texture (see Chapter Five and Waugh, 1967). Whereas in the Penrith Sandstone this grain type forms more or less the only detrital component, however, the sandstones of the interbedded sequences also contain varying proportions of dolomitised carbonate grains and clasts, derived Lower Palaeozoic fragments and intraformational mudstone clasts which, unlike the quartz grains, are not necessarily restricted to the sand size grade and are thus unlikely to have an aeolian-related origin. Whereas a non-localised source may be proposed for the majority of such fragmental types (see below), the petrographic characteristics of the quartz sand component suggests that a secondary derivation from the Penrith Sandstone by fluvial reworking of the unconsolidated aeolian dune sands is the most likely origin.

Although the breccias are superficially similar to those described from elsewhere in the Eden Valley, the petrography of these deposits is distinct from that of the Penrith and Stenkrith brockram in that the detrital component comprises a more varied assemblage of derived fragments. Waugh (1967, Table 9.1, p.255) records that whereas the clast component recorded from elsewhere in the Eden Valley (the Penrith and Stenkrith Brockrams of this study) is more or less exclusively formed of derived Carboniferous limestone, sandstone and chert fragments (98 per cent., with the remainder comprising shale and haematite fragments), that of the interbedded breccias includes, as well as the above named rock types, clasts of vein quartz, slate, rhyolite (Kendall, 1902), Whin Sill (Holmes and Harwood, 1928; Dunham, 1932) and Borrowdale Volcanics (Versey, 1939) of Lower Palaeozoic age for which, with the exception of Versey (op. cit.), a derivation from the northern Pennines has been inferred. Such clasts are not volumetrically important, however, and constitute only five per cent. of the total component with the remainder being formed mainly of dolomitised limestone fragments. Whin Sill and Borrowdale Volcanic clasts have been recorded only from George Gill (NY717190) and Trough Gill (NY587238) respectively, the latter being reported to occur in a thin horizon at the base of a small waterfall. This horizon has also been recorded by Dakyns et al. (1897) but was not located by the author. Although the provenance of the Whin Sill clasts has been the subject of considerable controversy in the past (see Waugh, 1967, p.241 to 242 and Bott, 1974, p.324 for example), the structural model erected by Bott (1974) for the evolution of the Eden Valley as a contemporaneously subsiding trough adjacent to the Inner Pennine fault or faults, reconciles the occurrence of these fragments with a derivation from the north Pennines, as is also indicated by cross-bedding and imbrication within these deposits. The derivation of the Borrowdale Volcanic clasts recorded from Trough Gill is less certain, however, as this group is exposed on both the western (between Ullswater and Shap Fells) and eastern (the present Cross Fell inlier) flanks of the valley and neither Dakyns et al. (1897) nor Versey (1939) record any directional data from this horizon. The proximity of this exposure to the western margin of the valley, however, probably supports a derivation from the Ullswater-Haweswater area as has been suggested by Versey (op. cit.) on the basis of a greater lithological affinity with this area than with the rock types exposed on Cross Fell. Although a south-westerly source is a possibility for certain localities, therefore, a dominant derivation from the northern Pennines is accepted for the majority of the coarse detritus (c.f. Kendall, 1902; Holmes and Harwood, 1928; Dunham, 1932; Trotter and Hollingworth, 1932; Waugh, 1967) and a westerly or southerly derivation (e.g. Turner, 1927; Trotter and Hollingworth, 1928; Turner, 1935) is not considered likely.

The matrix of the breccias dominantly comprises quartz sand grains which possess identical characteristics to those of the associated sandstones and, therefore, the Penrith Sandstone. Exposures are typically poorly lithified and the breccias are not (apparently) cemented except by a thin iron oxide coating of the grains. This contrasts markedly with the alluvial fan brockrams described in Chapter One which are typically cemented by calcite or dolomite (see Chapter Two and Waugh, 1967). The associated sandstones of the interbedded sequences are also extremely friable and are, at every locality, unsilicified. Although the exposures of Hilton Beck, Hayber Gill and the River Belah occur within the lower, unsilicified Penrith Sandstone (see Chapter Five), those of Low House are within the upper, silicified part of the formation, as can be seen from crags exposed in the River Eden to the north and south of this locality (Hawkcliff Scar, NY515495 and at NY481474 respectively) and thus clearly have a different post-depositional history. In contrast to the poorly lithified sequences exposed in the Hilton Beck section, however, the interbedded breccia and sandstone sequences of the Hilton borehole are recorded to be tightly cemented by gypsum-anhydrite (Burgess and Holliday, 1974), quartz and dolomite (Waugh, pers. comm.) with the gypsum-anhydrite presumably having a formation comparable to the evaporites of the overlying A-Bed or from precipitation from downward percolating fluids derived from these evaprites. The presence of saline fluids is also proposed to account for the post-depositional dolomitisation of the limestone clasts within these sequences. It seems possible, therefore, that the Low House sequences were once also cemented by an early (gypsum-anhydrite?) cement which preceded the silicification of the Penrith Sandstone in this area and which has since been leached from the exposures. As the silicification of the Penrith Sandstone is suggested by Arthurton et al. (1978, p.195) to precede the formation of the B-Bed (sic) and the A-Bed is not recorded to extend as far north as the Armathwaite area (Burgess and Holliday, 1974), the cementation of the Low House exposures is unlikely to be related to downward percolating saline fluids and is considered more likely to represent an early precipitation from saline groundwaters which originated at, or shortly after, the time of deposition. An early cementation and dolomitisation of the limestone component may thus be proposed for all localities, including those of the dolomitised alluvial fan brockrams.

Description of the exposures

The main sedimentological features of the interbedded breccia and sandstone sequences are described in the proceeding section by reference to three well exposed sites which are located in Hayber Gill (NY752165), the River Belah from Belah Bridge (NY793120) to Belah Scar (NY796121) and the River Eden north of Low House, Armathwaite (centred on NY515488). The degree of exposure in localities such as Hilton Beck is not considered of sufficient extent for detailed sedimentological description.

HAYBER GILL, WARCOP

The main exposure at this locality forms a natural stream bluff on the right bank of Hayber Gill, a tributary of Lowgill Beck, at NY752165.

Although of restricted lateral extent (approximately 25 metres), the east to west trending face is unusual in terms of the structures exposed elsewhere in the Eden Valley in providing a well exposed section through an isolated breccia sequence with an apparent channel-like form. This sequence occurs within a series of massive and plane and cross-bedded sandstones and is of a larger scale than comparable structures exposed elsewhere in the area. The main structures of the section and the relationship between the breccias and sandstones is illustrated in Figure 3.1.

The breccias form a sequence of approximately 12 metres width and six metres exposed thickness which appears to be steeply channelled into the sandstones. The western margin of the sequence cuts down at an angle approaching 45 degrees and truncates the structure of a series of plane and cross-bedded sandstones overlying horizontally bedded sandstones with fine breccia horizons forming the western part of the exposure. The eastern margin of the breccia unit is less clearly defined than the western margin and interfingers with flat and irregularly bedded sandstones without any apparent truncation of structure. Thin breccia horizons, overlying irregular erosion surfaces extend laterally into the sandstones of the eastern part of the exposure.

From Figure 3.1, it may be seen that the breccia sequence comprises at least four component units separated by gently upward curved bedding surfaces which progressively lose their identity towards the western contact with the sandstones. It is apparent, however, that the structure of the eastern margin of the sequence is partly determined by the relative lateral extent of these component units. The maximum thickness of any individual unit is of the order of 1.5 metres and, with the exception of the uppermost unit, comprises massive breccias which are more or less totally unsorted and

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Figure 3.1. Channel-fill sequence exposed in Hayber Gill, Warcop (NY752165). Sandstones represented by light stipple, breccias by heavy ornamentation.


West
East
Fluvially reworked Penrith
Sandstone with thin breccia
horizons
Breccias
0
2
4
Metres

Figure 3.1

develop only a rudimentary size grading of the clasts. In consequence, there is little recognisable clast size differentiation between these units. The only sedimentary structure developed in the lower part of the sequence is a crude upward-fining of the very coarsest grade material and a tendency towards a sub-horizontal alignment of the non-spherical clasts. The highest exposed unit is characterised by a crude cross-bedding which progressively steepens westwards until attaining a similar attitude to that of the western margin of the sequence.

The evidence of the abrupt western margin of the breccia sequence and the truncation of the sandstones in this direction thus implies the breccias to be incised into previously deposited sediments which are now represented by the plane and cross-bedded sandstones occurring in the western part of the exposure and which were themselves probably the products of an earlier phase of fluviatile deposition or reworking (as indicated by the incorporation of coarser grade material as distinct horizons and scattered clasts). The interfingering of breccia and sandstone at the eastern margin of the breccia sequence, however, indicates that this margin was not formed by a single (or continuous) phase of incision and suggests the deposition of the sands which now form the sandstones of the eastern part of the exposure to have been closely associated (and concomitant) with the deposition of the breccias. It is considered probable, therefore, that both sands and breccias were deposited within a wide channel system of which only the western (incised) margin is now exposed. The precise relationship between these two facies suggests the breccias to have been deposited at the western extreme of this channel system either as the lateral equivalent to the sands (as might occur in the deeper parts of such channels as bed load) or following a temporary entrenchment of the previously deposited sands during a period of increased current activity.

In summary, therefore, this exposure is considered to represent a transverse section through a channel-fill sequence in which the sediment transport direction was essentially perpendicular to the present face. From previous considerations and measurements of clast fabric and sedimentary structure elsewhere in these sequences, an approximately northerly derivation is inferred. The contrast in detrital grade is suggested to represent deposition from flows of markedly differing competence or capacity either (excluding the sandstones to the west of the breccias) as discrete episodes or in successive phases within a relatively wide channel, as may occur during the falling flood stages of braided channel systems (e.g. Williams and Rust, 1969).

THE RIVER BELAH

At this locality, easily accessible exposures of interbedded breccia sheets and/or channel-fills and fluvially reworked Penrith sandstones occur in a series of west-south-west to east-north-east trending natural crags on the left bank of the River Belah and are exposed over a distance of approximately 250 metres from Belah Bridge (NY793120) to Belah Scar (NY796121). The main face to be described is of approximately 110 metres length and is terminated at its eastern limit by the vertical sandstone bluffs of Belah Scar where access becomes difficult. From the western edge of the main outcrop to Belah Bridge the sequence is exposed only intermittently and is illustrated in Figure 3.2 by reference to two representative sections.

A small outcrop at Belah Bridge (now bypassed by the A685 Brough to Kirkby Stephen road which crosses the river 100 metres to the west of this locality) is orientated approximately perpendicular to the west-south-west to east-north-east strike of the main crags and is dominated by a thick (2.4 metres minimum) breccia unit which forms the upper part of the exposure at

Figure 3.2. Representative sections (1 and 2) of the sequences exposed between Belah Bridge (NY793120) and the main river crags of the Belah (approximately NY795121). Sections 3 and 4 are from the main exposure.


West-south-west
East-north-east
1
Belah Bridge
Channel-fill
2
?
3
Main exposure
N
Imbricated unit
1
4
4
3
2
1
0
Metres
20m.
120m.
9m.
Legend
Very coarse breccia
Coarse/medium breccia
Fine breccia
Sandstone
1 Diameter: 10 measurements

Figure 3.2

the site of section 1 in Figure 3.2. Less than five metres north of this section, however, these breccias are missing from the sequence and in this direction pass directly into sandstones which are largely obscured by vegetation. Where the lower contact of the breccias can be seen (as at the site of section 1), they overlie and truncate the structure of a sequence of flat bedded sandstones containing impersistent fine breccia horizons and scattered clasts of derived dolomitised limestone and Lower Palaeozoic fragments. This contact cuts up section towards the north and, as is apparent from the juxtaposition of sandstone and breccia in the vegetated higher parts of the crags, forms an abrupt lateral margin to the breccias which thus appear to be incised into the sandstones. The breccias predominantly comprise coarse (16 centimetres maximum recorded long dimension), poorly sorted Carboniferous sandstone and limestone fragments in which the only sedimentary fabric developed is a tendency towards a sub-horizontal preferred orientation of the elongate clasts. As elsewhere, the main component of the matrix comprises well rounded, high sphericity quartz grains (presumably secondarily derived from the aeolian dune sands of the Penrith Sandstone) although minor quantities of haematite stained clay are also present.

The lowermost part of the Belah Bridge section is formed of a laterally more persistent breccia unit which can be traced for approximately 30 metres towards the north-east. The top of this unit is gradational through 10 to 15 centimetres of sandstone containing abundant fine clasts.

A short distance to the north-east of Belah Bridge, a cross-bedded sandstone unit is exposed laterally for approximately 20 metres in the upper part of a series of small crags (Figure 3.2, section 2). Only one set of cross-strata, with a minimum thickness of one metre, is exposed and is underlain by flat bedded sandstones and breccias which can be traced to the lower part of the section at Belah Bridge. The cross-bedded sandstone is thus considered to represent the lateral equivalent to the incised breccias described from section 1. Excluding the presence of a thin horizon of fine breccia at the base of the set and the occasional occurrence of intraformational mudstone and other, non-locally derived fragments along the lower surfaces of the foresets, the sandstone is moderately well sorted and is approximately graded into finer and coarser laminae. The cross-beds are inclined at an apparent angle of 27 degrees towards the west and indicate transportation from an easterly source.

Sections 3 and 4 occur at the western extreme of the main exposure. The exact stratigraphic position of these sequences with respect to sections 1 and 2 is uncertain but they are suggested to occur at a slightly higher stratigraphic level on the basis of the field relationships (excluding the indeterminate effects of any unexposed faults).

The main exposure of this locality forms a laterally continuous section of approximately 110 metres length which is illustrated in Figure 3.3. The upper part of the section is generally formed by a thick (three metres maximum exposed thickness) sequence of medium and coarse breccias which irregularly overlies and is interbedded with massive and flat bedded sandstones with subordinate fine breccia and mudstone horizons which predominantly form the lower part of the exposure. These thick sandstones are not seen in the western part of the section where, from NY795121 to 25 metres east of this site, only the breccias (incorporating several, impersistent sandstone horizons) are exposed. The rock sequences of this section are displaced by a number of minor faults and by two larger, normal faults, one of which is exposed 35 metres west of NY796121 (Figure 3.3, upper section) and the other 60 metres west of NY796121 (Figure 3.3, middle section).

Figure 3.3. Sketch section of the sequences exposed from NY795121 to Belah Scar (NY796121) on the left bank of the River Belah. Mudstone horizons (and intraformational clasts in proceeding figures) shown in black.


East-north-east
(NY 796121)
Laminated sandstone
Mudstone
Cross-stratified sandstone
Intraformational fault
Trace of oblique fault
Location of section 3
(Figure 3.2)
Location of
Figure 3.4
0
5
10
Metres
West-south-west
(NY 795121)

Figure 3.3

The breccia sequences exposed at this locality are characterised by a massive appearance and a relatively low degree of internal organisation in which sedimentary structures and clast fabrics are only poorly and impersistently developed. Sedimentary structures are limited to a crude horizontal bedding or internal stratification and fabrics to an approximate bedding parallel or, rarely, imbrication of the elongate clasts. Where internal bedding planes can be recognised, the component beds may exhibit a rudimentary upward-fining or an approximate size differentiation of the clasts which is usually related to the presence of relatively coarser basal horizons. Whereas bedding parallel clast fabrics tend to occur throughout the breccias, imbricated fabrics typically are restricted to beds which have a somewhat higher (moderate) degree of sorting than is usual in these sequences and is most common in the four to six centimetres (long axis) clast range. Long axis inclinations (see Figure 3.2, section 3, for example) are typically towards the north-east which is thus the presumed derivation direction for this detritus.

Although variations in the structure and fabric of the breccia units can be recognised, therefore, and demonstrate the main breccia units illustrated in Figure 3.3 to be formed of a number of component beds (which, in places, interdigitate with the sandstones), such distinctions are rarely sufficiently extensively or well enough developed to enable these beds to be traced laterally for any great distance.

Thick, flat bedded sandstones are exposed in the lower part of the section from NY796121 to 25 metres east of NY795121 and attain a maximum exposed thickness of five metres immediately to the east of the more westerly of the two larger faults (Figure 3.3, middle section). These units typically consist of a number of beds of plane laminated or massive sandstones separated by prominent planar or slightly undulatory surfaces which are frequently draped by thin, discontinuous mudstone laminae. The lowermost few centimetres of succeeding beds are often characterised by the presence of gritty clasts which may form fairly prominent horizons. Larger (up to 50 centimetres), isolated clasts with a bedding parallel alignment may also occur within the sandstones. Lensoidal scour infills of festoon-shaped cross-beds with set thicknesses of the order of 15 centimetres occur sparsely throughout this facies. Only one example of large scale cross-bedding (or stratification) has been recorded from this section and occurs in a sandstone unit which is interbedded with the breccias in an inaccessible, upper part of the rock face in the central area of the exposure (see Figure 3.3, middle section). This unit has an estimated thickness of one metre and comprises a single set of cross-strata in which the foresets are apparently inclined towards the west. This unit is similar, therefore, to the example described from the intermittent exposure of this locality (see Figure 3.2, section 2).

Thinner (less than one metre thick), laterally impersistent beds of sandstone also occur as discrete horizons within the massive breccia sequences (see Figure 3.2, section 3, for example).

Of common occurrence throughout the exposure illustrated in Figure 3.3 are distinct horizons, discontinuous lenses and clasts of red-brown, iron-stained mudstone which typically overlie or occur within the sandstone units. Mudstone horizons have not been recorded from the breccia facies although a few intraformational clasts of this material are locally preserved at the bases of the units. One particularly prominent horizon (indicated in Figure 3.3) is exposed for a distance of 40 metres. Unfortunately, this horizon cannot be examined closely as it occurs approximately three metres from the base of the sequence but, from Figure 3.3 can be seen to define an irregular surface of considerable (1.2 metres) relief. It is also apparent that this horizon is not of a constant thickness throughout its exposed length. Thicker accumulations tend to correspond to positions where the mudstone occupies a lower level in the sequence (i.e. shallow depressions in the underlying sandstone surface; particularly well seen where the westerly development of the mudstone is truncated by the overlying breccia unit) and thinner accumulations to slight elevations in the underlying surface.

The nature of the contacts between the breccias and sandstones and the relationships between these two facies is variable throughout the section illustrated in Figure 3.3. Most commonly, however, this contact is represented by a markedly irregular erosion surface against which (where the breccias overlie the sandstones) the sedimentary structures of the sandstones are truncated. Steeply incised contacts, as described from Hayber Gill and Belah Bridge, are not developed although comparable structures where the breccias can be seen to cut down section and infill shallow, concave-upward depressions occur in several parts of the exposure. One such example can be seen to the west of the more easterly of the two larger faults (Figure 3.3, middle section) where the base of the breccias cuts down towards the west and truncates a prominent mudstone horizon.

In other parts of the section, as at the eastern end of the middle section in Figure 3.3, the breccias directly overlie mudstone horizons without any apparent truncation of structure or, as in the upper part of the exposure in the region of the more easterly of the two larger faults, interdigitate with the sandstones. Apparently conformable contacts are not persistently developed, however, and in the above example the breccias laterally cut up section (in the regions of both larger faults) such that the mudstones are overlain by sandstones which are, in turn, overlain by the breccias.

An unusual relationship between the facies is developed in the western part of the exposure, approximately 30 metres east of NY795121. This part of the exposure is illustrated in Figure 3.4 which is an enlargement of the area

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Figure 3.4. Enlargement of the area shown in Figure 3.3 showing the facies relationships at this site.

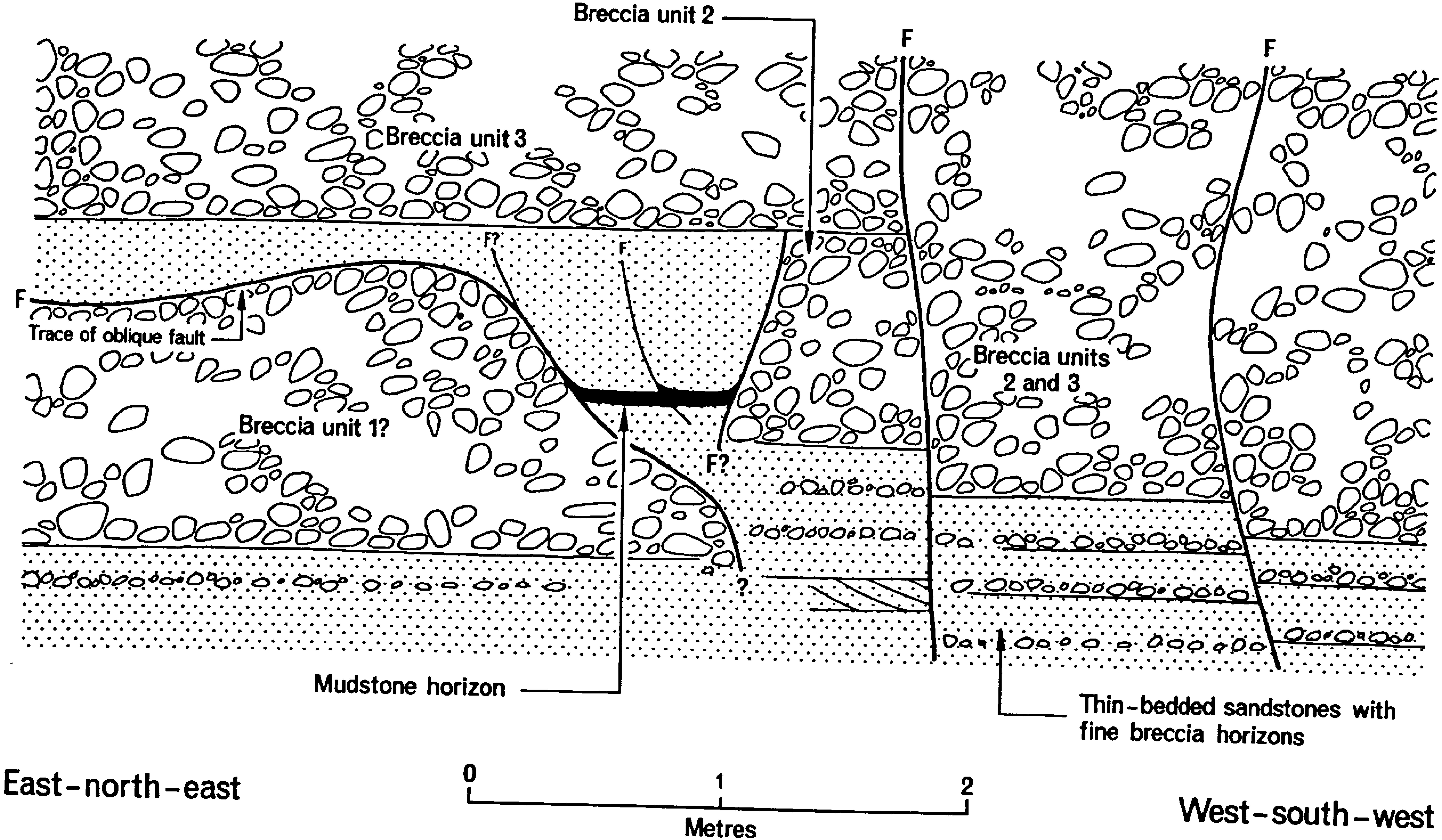
Breccia unit 2
F
F
Breccia unit 3
F?
F
F
Trace of oblique fault
Breccia units
2 and 3
Breccia unit 1?
F?
?
Mudstone horizon
Thin-bedded sandstones with
fine breccia horizons
East-north-east
0
1
2
Metres
West-south-west

Figure 3.4

indicated in Figure 3.3. At this site, a high angle oblique fault (trending east to west into the east-north-east to west-south-west strike of the rock face) displaces breccias (arbitrarily designated as breccia unit 1 in Figure 3.4) against a series of sandstones which otherwise form the lower part of the exposure. These sandstones are overlain by breccia units 2 and 3 which can only be clearly identified as separate beds in the central part of the exposure outlined in Figure 3.4, immediately adjacent to the more easterly of the two minor faults which displace these breccia units. Breccia unit two is terminated abruptly against the sandstones within the exposure and cannot be identified (and presumably is not present) in the eastern part of this section. This contact is notably irregular and does not form a clearly defined plane. The sandstones against which these breccias abut have an apparent wedge-shaped configuration which is partly related to weathering of the obliquely faulted margin of breccia unit 1 (i.e. the present position of the rock face with respect to this fault). The exposed margins of a mudstone horizon occurring within the sandstone are apparently draped against the contacts with breccia units 1 and 3. Two minor fractures may also be observed at this site, one of which displaces the mudstone horizon by two to three centimetres. Traced eastwards from the site of Figure 3.4, the sandstones seen to the south of the oblique fault plane (underlying breccia unit 3 and appearing above breccia unit 1 in the eastern part of the exposure illustrated in Figure 3.4) are concealed by breccia unit 1 which is displaced against breccia unit 3 in the upper part of the exposure.

Although the facies relationships at this site are relatively confused, it is clear that the abrupt lateral termination of breccia unit 2 is dissimilar to the incised (sedimentary) breccia contacts described previously. It is considered more likely, therefore, that this junction represents a displaced (faulted) contact which is not traceable into the overlying breccia unit 3 and thus intraformational in origin. The irregular margin of this unit, which does not form a clearly defined plane, is suggested to result from a movement which occurred prior to lithification of the breccias (c.f. the alluvial fans of the Sheep Range, southern Nevada described by Longwell, 1030). The apparent 'draping' of the mudstone horizon against both this margin and the faulted margin of breccia unit 1 may thus be related to a post-depositional distortion associated with the movement of these faults (although the sense of movement of these faults is uncertain) or to the minor fractures associated with this horizon.

A comparable feature is exposed in the central part of the section illustrated in Figure 3.3 (approximately three metres to the west of a more prominent normal fault with a downthrow of 1.8 metres towards the west or south-west) where a minor fault with a displacement of the order of ten centimetres apparently terminates within the exposure. The displacement associated with this fault appears to exclusively affect the sandstones forming the lower part of the rock face and cannot be traced into the overlying mudstone horizon and breccia unit (which truncates the mudstone one metre to the west of this fault). Due to this inaccessibility of the upper part of the exposure at this point, however, precise relationships cannot be conclusively established.

The Belah section thus comprises a similar sequence to that described from the small exposure in Hayber Gill although the precise morphology of the breccia units at this locality cannot, in general, be so clearly demonstrated. The outcrop at Belah Bridge (see Figure 3.2, section 1), however, comprises breccias which are apparently steeply incised into the sandstones and which may, similarly to the sequence exposed in Hayber Gill, be interpreted as a transverse section through channel-fill of which only the north-western margin is seen. The orientation of this exposure at right angles to the strike of the main face would, therefore, indicate a transport direction approximately parallel to this face which would thus not be expected to transect the cross-sectional morphologies of any channel-fill sequences which may be present. Determinations of palaeocurrent direction (as measured from imbricated clasts) have indicated an approximately north-easterly derivation which concurs with these observations. The configuration of the breccia units at this locality (which, as has been noted, are commonly erosive on underlying beds) is thus considered to represent longitudinal sections through channel-fill sequences of indeterminate width.

Although the bases of the breccia units commonly overlie erosion surfaces and can be seen to be locally incised into the sandstones, it is evident from the preceding descriptions that the lower contact of these units does not everywhere truncate the underlying structures and may, in places, rest with apparent conformity on mudstones. However, it has also been demonstrated that such contacts are not persistently developed and that the bases of the breccias may laterally cut up section to overlie sandstone horizons which, in turn, overlie the mudstones. Although this configuration could be related to a phase of erosion which preceded or accompanied the deposition of the breccias, it is considered unlikely that the erosional level of such a surface could have been partly determined by the mudstones, as would be the case where the breccias directly overlie the mudstones. It is suggested, therefore, that the breccias were probably deposited on an originally (topographically) irregular surface in which the relief was controlled by the occurrence of impersistent sand sheets spread across the underlying mud or mudstone surface. Only a slight reworking or the unlithified sands (which might not be expected to affect a partly dehydrated mud layer) would thus be required to produce the present irregular surface.

It has also been noted that the exposed mudstone horizon illustrated in Figure 3.3 does not retain a constant thickness and that the thicker portions tend to correspond to shallow depressions in the underlying sediment surface and thinner parts to slight rises. Such a relationship is also considered likely to be a primary depositional, rather than an erosional feature whereby the accumulation of thick mud horizons occurred preferentially in topographic depressions in which water could be ponded during the final stages of fluvial activity.

Whilst the sedimentary structure of the majority of the sandstone units is in accordance with a fluvial, rather than aeolian mode of deposition, the large scale cross-bedding described from the upper part of section 2 in Figure 3.2 is evocative of the aeolian dune bedding of the Penrith Sandstone *sensu stricto* (see Chapter Five), especially as these sandstones tend to be graded into finer and coarser laminae. Although an aeolian origin for this structure would be feasible (Glennie, 1970, p.38, for example describes remnant aeolian structures from alluvial fan sediments of Wadi Batha, Eastern Arabia), the presence of coarser detritus along the lower surfaces of the foresets clearly, in this case, indicates a fluviatile mode of formation to be the more likely. By analogy, a fluviatile origin for the large scale cross-bedded set described from the main exposure at this locality seems probable although the possibility of an aeolian formation cannot necessarily be precluded.

In summary, therefore, the sequences exposed in the Belah section comprise a series of breccias which are interbedded with and incised into fluvially deposited sandstones and which may be related to deposition (see Interpretation, this chapter) from flows of markedly differing competence. It has been postulated, however, that certain anomalous facies relationships (as in the abrupt lateral termination of breccia unit 2 in Figure 3.4) cannot simply be explained in terms of these processes and may be more concisely (although by no means fully) explained by reference to contemporaneous displacement.

LOW HOUSE, ARMATHWAITE

The exposures of Low House form a series of north to south trending natural crags on the left bank of the River Eden, two kilometres north of Armathwaite (see Figure 1.1) and were first described by Trotter and Hollingworth (1932, p.127-130). Interbedded breccias and sandstones, which are displaced by three faults of indeterminate throw, are well exposed over a distance of 250 metres from NY515486 to NY515489 although a gap of 80 metres reduces the total length of the exposure to 170 metres. A small gap of one metre towards the northern end of the section effectively divides the exposure into three separate sections which are illustrated in Figure 3.5. From a comparison of this figure with Figure 3.3, it is apparent that the overall distribution of the facies is similar to that exposed in the River Belah section with the breccias tending to form the upper part of the exposure and the sandstones the lower.

The breccias are lithologically similar to those described previously with the main clast components comprising dolomitised limestone and lesser amounts of reddened sandstone. The breccias exhibit a slightly higher degree of internal organisation than is recorded elsewhere from the interbedded sequences and develop conspicuous bedding planes separating beds in which there is usually a moderate degree of size consistency of the coarser clasts. Individual beds are commonly coarser at the base, often developing an exceedingly coarse basal horizon, and fine erratically towards the top. True fining-upward units are relatively uncommon, and as with sorting, tend to be restricted to a size gradation of the coarsest grade material only. Coarse beds, however, are frequently overlain by beds of finer breccia which may, in turn, be overlain by thin and generally impersistent sandstone horizons and thus form fining-upward cycles. Clast fabrics tend to be restricted to

Figure 3.5. Sketch section of the sequences exposed from NY515486 to NY515489 on the left bank of the River Eden at Low House, Armathwaite.


South (NY515486)
F
F
F
78m. not exposed
Location of
Figure 3.7
Sandstone
with isolated clasts
Coarse breccia in
river bed
Location of
Figure 3.6c
Location of
Figure 3.6b
North (NY515489)
1m. not exposed
Location of
Figure 3.6a
0
10
20
Metres
Vertical scale is 1.5 times
that of the horizontal

Figure 3.5

a bedding alignment of the elongate clasts although, as in the River Belah section, imbricated fabrics also occur but are not well developed. Rarely, as in the upper part of the section approximately 25 metres from the northern end of the exposure (Figure 3.5, lower section), the breccias develop a crude internal cross-bedding. Cross-bedded sets are in the range of 50 to 80 centimetres thick and, unlike the lensoidal (trough) sets recorded from the alluvial fan breccias of localities such as Burrells Quarry (see Chapter One), conform to a tabular (or wedge) planar style. The apparent inclination of the foresets is towards the south and, as with clast imbrication at this locality, indicates an approximately northerly derivation for the detritus.

Sandstones are exposed as thick (up to two metres) units, comprised of a number of plane laminated, cross-bedded and massive beds, in the lower part of the section and as thinner horizons interbedded with the breccias. The thick sandstones generally display a wider variety of sedimentary structures than hitherto recorded although weathering of the now poorly lithified exposures frequently obscures the structures and lends these beds a massive, and apparently unstructured, appearance. Successive beds are often separated by thin (and usually discontinuous) mudstone laminae overlain by horizons (which are usually less than one centimetre thick) of scattered fine clasts. An example of a small scale scour structure is illustrated in Figure 3.6a, which is located approximately 35 metres from the northern end of the exposure, one metre from the base of the section (indicated in Figure 3.5, lower section). In this example, a few dolomitised limestone fragments are preserved in a shallow depression which is draped by a mudstone lamina.

Cross-bedding is well developed in the sandstones and occurs either as relatively thick (of the order of 30 centimetres) planar sets or as smaller festoon-shaped scour infills. Rarely, as at the southern end of the lower

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Figure 3.6. Small scale features of the sandstone units: a. cross-section of a scour, b. structureless sandstone/mudstone filled vertical fracture in sandstones and c. desiccated mudstone horizons.


Scour
Mudstone drape
0.5 metres
a.
Sandstone and mudstone filled vertical pipe
0.5 metres
b.
0.5 metres
c.

Figure 3.6

section in Figure 3.5, sets may attain a thickness in excess of 50 centimetres. The development of planar cross-bedding is particularly well seen in central parts of the exposure, approximately 45 metres north of the main gap in the section (Figure 3.5, middle section; see, also, Figure 3.7) where four sets with an average thickness of 35 centimetres are exposed above a prominent mudstone horizon. The lower bounding surfaces of the sets truncate the foresets of underlying sets and are draped by thin, discontinuous mudstone laminae. The apparent inclination of the foresets of these sets (as with all exposed examples of cross-bedding at this locality) is towards the south and indicates an approximately northerly derivation direction.

Clasts of dolomitised limestone and derived sandstone occur throughout the sandstone facies both as distinct horizons interbedded with the sandstones and as sparsely scattered, isolated clasts along bedding surfaces. Occasionally, as in the first exposure north of the main gap in the section (Figure 3.5, middle section), beds of sandstone with a relatively higher proportion of isolated (i.e. non-contacting) clasts may form units of one to two metres thickness which are difficult to designate as either sandstone or breccia.

Sandstone also occurs rarely as infills of small, vertical pipes which cut through beds of breccia or sandstone. Two examples have been recorded from the Low House section, one of which occurs within the breccias forming the upper part of the exposure approximately 20 metres north of the main gap and the second of which truncates sandstones in the lower part of the exposure, approximately 18 metres north of the planar cross-bedded sandstones described previously (Figure 3.5, middle section). This second example is illustrated in Figure 3.6b and can be seen to truncate the structure of cross-bedded and laminated sandstones. The infill is of approximately ten centimetres width and 30 centimetres length and is formed of a structureless mud-rich sandstone which is terminated at its lower limit by a mudstone horizon. Relationships at the upper margin of this structure are less clearly defined, however, and it is uncertain whether the pipe terminates at a mudstone horizon overlying the laminated and cross-bedded sandstones or whether this horizon is, itself, truncated. A similar feature has also been recorded from the lowermost breccia sequence of Thistley Hill Quarry (NY678205; see Chapter One), immediately overlying Carboniferous sandstones.

Discrete horizons, discontinuous laminae and clasts of mudstone are of common occurrence within the thick sandstone units although, in addition, large quantities of mudstone are present in the form of thin (generally less than 0.1 centimetre) drapes within sandstone laminae. The maximum thickness recorded for these horizons is five centimetres although their development tends to be laterally variable and particular horizons may completely thin out over distances of less than two metres. Alternatively, single horizons may split into three of more separate laminae. Most horizons occur above erosion surfaces (which truncate the structures of underlying sandstones; see, for example, Figure 3.7) although they may also be interbedded with plane laminated sandstones or form drapes on the foresets of cross-bedded sets. The most extensive development of the mudstones occurs towards the northern end of the central exposure at this locality (Figure 3.5, middle section) where a prominent mudstone horizon underlying the lowest of the cross-bedded sets described previously is exposed for 15 to 20 metres. When traced northwards, however, this horizon laterally splits into three thin laminae separated by sandstone (illustrated in Figure 3.6c which is an enlargement of the area indicated in Figure 3.5; see, also, Figure 3.7). As these laminae thin northwards, the lower two pass into discontinuous plates with a saucer-like form in which the concave sides face upwards. The scale of the individual plates is in the order of four to five centimetres in horizontal length and less than one centimetre in thickness. From Figure 3.6c and Figure 3.7, it can be seen that several such horizons are exposed at this locality. The plates which comprise individual horizons are usually of a comparable thickness although this may vary from horizon to horizon. Where the thickness exceeds one centimetre, the length of the plate is correspondingly increased and may, in some instances, exceed ten centimetres. Larger examples tend to be upturned only at their edges rather than possessing a regular concavity. Generally, therefore, the scale of these features, and to a lesser extent, the form, is dependent on the thickness of the mudstone.

Angular mudstone clast breccias with a sandstone matrix are also extensively developed towards the northern end of the central exposure (Figure 3.5, middle section) although horizons and isolated clasts of this material occur throughout the sandstones. Rare mudstone clasts are also recorded from the lower horizons of the breccia units at this locality.

For the most part, the distribution of the breccia and sandstone facies in the Low House section is comparable to that described from the River Belah section in that the breccias tend to form the upper part of the exposure and the sandstones the lower. The majority of the main breccia units are laterally extensive (excluding the northernmost exposed breccia unit which thins towards the south and passes into sandstones) and have only slightly irregular bases but can, in places, be seen to overlie marked erosion surfaces against which the structures of the underlying sandstones are truncated. A localised downcutting of the main units (as described from the continuous exposure of the Belah section) occurs at several sites in the section where the breccias infill shallow depressions in the underlying sandstones. One particular example is exposed approximately 40 metres south of NY515489 (Figure 3.5, lower section) where the base of the breccias is locally incised and truncates an underlying mudstone horizon.

Of note at this locality is the presence of several breccia units which occur within, or are interbedded with, the sandstones and which possess a pronounced channel-like morphology. Several examples are exposed in the central part of the exposure (Figure 3.5, middle section) and are generally represented by structures (channel-fills) which have relatively flat upper contacts with the sandstones and asymmetrical, concave-upward bases which truncate the structures of the sandstones. In most examples, the sense of asymmetry is such that the more steeply incised lower contact forms the southern margin of the channel-fill. Part of the central exposure of this locality is illustrated in Figure 3.7 (centred approximately 20 metres from the northern end, as indicated in Figure 3.5) where two breccia filled channels (designated as channels 2 and 3) are exposed in the lower part of the section. A third, comparable structure (channel 1) is exposed at this site where a sandstone supported mudstone clast breccia infills a channel-like scour of 75 centimetres depth.

From Figure 3.7, it can be seen that these channel-like structures are incised into a series of planar cross-bedded, plane laminated and massive sandstones containing a number of conspicuous mudstone horizons (as described previously), one of which can be seen to drape the irregular surface of a prominent cross-bedded unit. Of the two breccia filled channels, the uppermost (channel 3) forms a wide (five metres), shallow (0.5 metre) structure which is apparently isolated from the surrounding breccias. The configuration and structure of the lower of these channels (channel 2), however, is considerably more complex and comprises a breccia-fill with an apparent channel-like section from the northern margin of which the breccias extend laterally to form the lowermost unit of the overlying, main breccia sequence (see, also, Figure 3.5, middle section). Unlike the massive breccia infill of channel 3, channel 2 is infilled with a series of breccias and sandstones and develops a number of marked internal bedding planes which are apparently inclined towards the south. The proportion of sandstone is relatively higher at the southern margin of the channel than towards the northern, in which

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Figure 3.7. Relationships and structures of breccia-filled channels, Low House. The location of this section is marked in Figure 3.5.

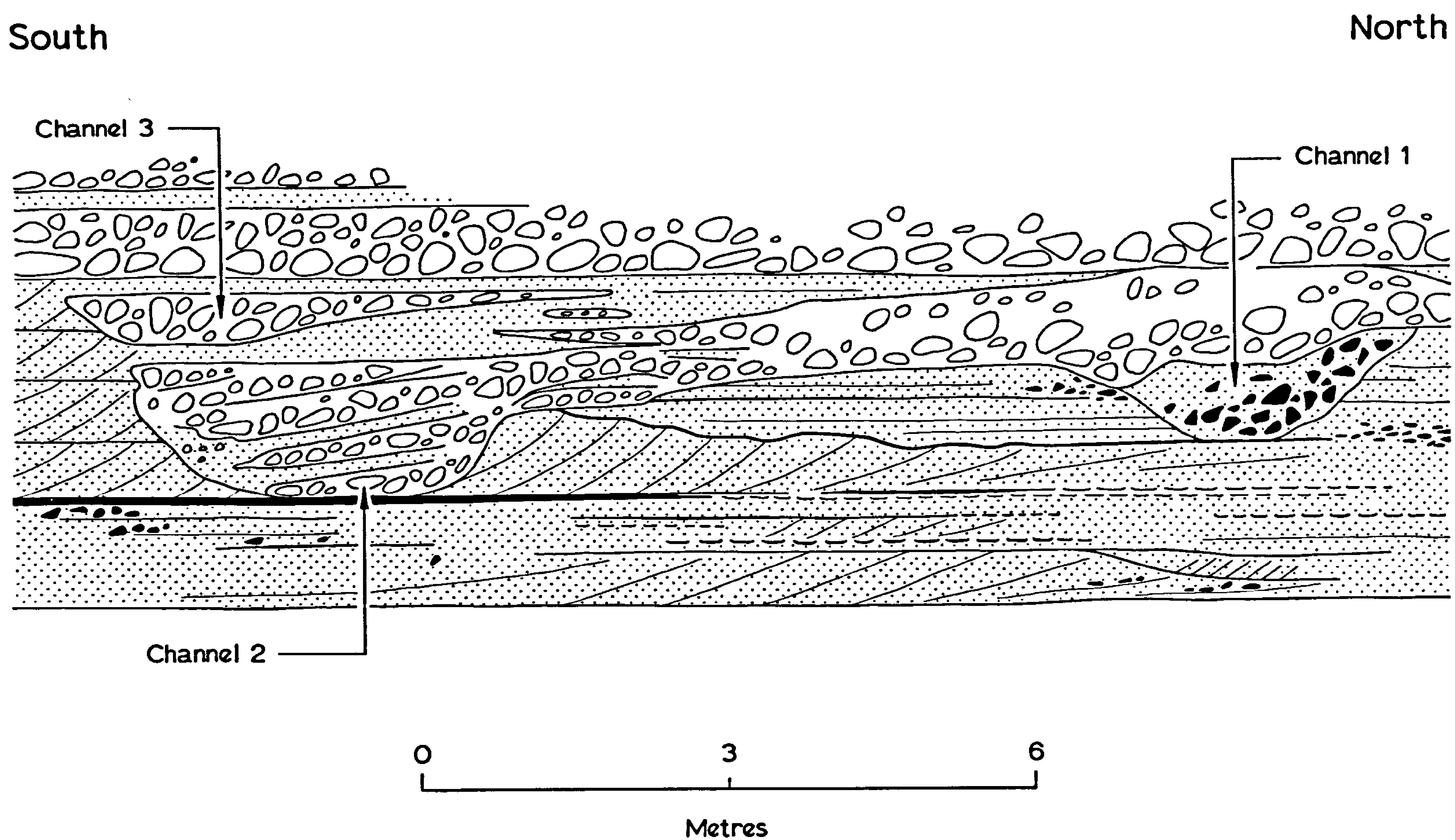

Figure 3.7

direction the sandstone beds appear to be cut out by the breccias. The base of the channel partly overlies a thick (five centimetres) mudstone horizon which elsewhere defines the base of the planar cross-bedded sandstones described previously. The sense of asymmetry of this structure, unlike all other examples at this locality, is such that the northern margin appears to be the steeper. The breccias which extend laterally from the northern margin of this structure cut up section towards the north until they irregularly overlie channel 1 in the northern part of the exposure illustrated in Figure 3.7 and form the lowermost unit of the main breccia sequence. The base of this breccia unit thus defines an irregular surface which truncates the structure of the underlying planar laminated and cross-bedded sandstones. The sandstones overlying channel 2 (into which channel 3 is incised and which are, in part, interbedded with the breccias extending from the northern margin of this structure) are cut out towards the north and pass southwards (apparently) abruptly into planar cross-bedded sandstones.

Several other channel-like structures are exposed in the Low House section, one of which, of comparable form to channel 3 in the preceding descriptions, is located 20 metres from the southern end of the central exposure (Figure 3.5, middle section).

The exposure of the Low House section thus comprises a sequence of deposits which are broadly comparable to those described from the River Belah section and from Hayber Gill. The majority of the breccia units (as in the upper part of the section) are exposed as laterally continuous sequences formed of a number of component beds which are locally incised into the underlying sandstones. Breccia units with a distinct channel-like morphology have also been recorded from this locality and are generally represented by breccia-fill units which have concave-upward, asymmetrical incised lower contacts typical of the cross-sectional configuration of curved channels in which the more steeply incised margin is directed towards the outside of the curve (e.g. Allen, 1970b). For the most part, these structures are represented by simple breccia-fills in which there is little or no internal organisation and which are orientated with their southern margins as the more steeply incised contacts. Only one example has been described in which the northern margin appears to form the more steeply incised contact (channel 2 in Figure 3.7) and occurs in a sequence of sandstones in which two other channel-fills are exposed (channels 1 and 3). Although relationships are uncertain at this site, the sediment size distribution within this structure, with sandstones predominating at the southern margin and breccias at the northern margin, indicates a depositional velocity gradient from relatively low to high which tends to support such an interpretation (assuming the main flow within curved channels to be directed towards the more steeply incised margin).

Whereas the overall coarse fragmental size of the breccias indicates deposition from high energy flows, sedimentary structures and fabrics are better developed at this locality than hitherto recorded and indicate a rather more selective deposition which is related only to the coarser grade material. The lack of well sorted horizons, however, and the restriction of the selective processes of deposition only to the coarser grade material is suggested to reflect deposition from flows which waned rapidly although, perhaps, less rapidly than those of other localities.

The structures of the sandstones are also comparable to those described from other localities although cross-bedded units are generally of commoner occurrence at this locality and comprise both planar cross-bedded sets and trough (lensoidal) sets. The occurrence of mudstone drapes and clasts and of other derived clasts within this facies (and within the cross-bedded units) indicates fluviatile, rather than aeolian, deposition to be the most likely mode of formation. Although the overall smaller detrital grade of this facies (in comparison to the breccia sequences) suggests these deposits to be the products of relatively lower energy flows, the occurrence of coarser horizons and of isolated clasts along the bedding planes indicates deposition to have occurred from flows of variable competence.

Mudstone horizons are extensively developed at this locality and represent deposition of the suspended sediment load under conditions of extremely low or no flow. The occurrence of discontinuous horizons in which the mudstone occurs as upturned plates has also been noted from this locality. Comparable structures have been recorded by several authors (e.g. Laming, 1966; Glennie, 1970; Picard and High, 1973) and interpreted as desiccated horizons. Their occurrence within the interbedded sequences thus suggests fluvial activity to have been intermittent (or ephemeral), with these horizons representing periods of surface exposure during which time the previously deposited muds were desiccated. A relationship between the thickness of the mudstone and the size and form of the desiccation plates has also been recorded from these horizons. A similar size relationship has been noted from desiccated mudstone horizons occurring in the Lower New Red Sandstone of Devon, described by Laming (1966) from Corbyn's Head (SX907632), and it has been suggested that both the spacing (Bull, 1963) and the degree of concavity of the plates (Glennie, 1970) increases proportionally to thickness.

Deposits representing a considerable range of energy environments can be recognised from the Low House exposures, therefore, and evidence a situation in which the depositional processes alternated between conditions of no or exceedingly low flows where clays accumulated and high energy conditions under which streams were able to transport clasts with long dimensions in excess of 35 centimetres. The overall lack of well sorted clast fabrics within the breccias indicates deposition to have occurred from rapidly (although less rapidly than at other localities) decelerating flows, probably as a response to high infiltration rates through previously deposited, highly permeable, sediment. Periods of non-deposition and subaerial exposure can be recognised from the occurrence of desiccated mudstone horizons which are typical of the reworked Penrith Sandstone facies. The processes of deposition are thus regarded to have operated as sporadic and sometimes violent ephemeral floods which were separated by periods of non-deposition.

INTERPRETATION

The association and structures of the sandstone, breccia and mudstone horizons occurring within central and eastern parts of the Eden Valley suggests these deposits to be the products of subaerial (see Chapter One and Five) fluviatile processes operating under conditions which were closely allied to the formation of the Penrith Sandstone and which were similar to those which formed the alluvial fan breccias of the western and southern margins of the valley. The interbedded nature of these deposits apparently indicates highly variable discharge rates which could accommodate the periodic adjustment from high energy conditions, in which coarse detritus could be transported and deposited, to extreme conditions of low flow under which muds were able to accumulate. As opposed to the coarse detritus, for which a derivation from the Pennine fells has been proposed, the petrographic characteristics of the dominant detrital component of the sandstones suggest the most likely source for this material to be from the local unconsolidated aeolian dune sands of the Penrith Sandstone. The lack of any well defined transitional sediment types between the breccias and sandstones further supports differing sources for these two facies although the incorporation of Penrith Sandstone type grains as the dominant clastic component of the matrix of the breccias and of derived dolomite and other grain types within the sandstones indicates the formation of these two facies to be at least partly related. The mudstone horizons are most closely associated with the sandstones and typically separate or occur above sandstone beds. It is not considered that this material is likely to have originated from the reworking of the Penrith Sandstone and, as with the dolomite and other non-aeolian derived grain types, presumably represents weathered detritus from the catchment area.

For these sequences, therefore, the most likely mode of formation is considered to be from flash-flood run-off from the Pennine margin of the Eden Valley which flowed through, reworked, and occasionally flushed large quantities of coarse detritus into the aeolian dune field of the Penrith Sandstone. In many respects, therefore, the breccia-sandstone-mudstone sequences resemble a logical transition from high to low energy conditions in which the influence of coarse detritus derived from the source area lessened away from the basin margin with respect to the quantities of finer, sand-grade material available in the dune field until the stage was ultimately reached where only mud-grade material remained in suspension. Two factors serve to complicate this interpretation and relate to the lack of transitional sediment types and the occasional occurrence of isolated coarse clasts within the sandstones. Whereas the former can be explained in terms of rapid flow deceleration (probably related to high rates of infiltration into previously deposited, highly permeable sediment), the occurrence of isolated large clasts (as in the River Eden section, for example) suggests the transporting flows to be under capacity with respect to their competence. This, in turn, indicates the possible importance of the availability of coarse detritus within this environment although, clearly, such a formation would be dependent on external controls such as a tectonic rejuvenation of the source area. Reference to this point will be made in the following section.

Whatever the precise factors influencing the deposition of these sequences, it is clear that differences exist with respect to the alluvial fan deposits _sensu stricto_ of elsewhere in the Eden Valley inasmuch as a high proportion of the sediment was not derived directly by erosion of the uplands and was not predominantly coarse grained. In addition, markedly erosive contacts are common (e.g. Belah Bridge, Hayber Gill and Low House) and indicate much of the fluvial activity (and at least that transporting coarse detritus) to have been restricted to definite stream channels. The closest modern analogue to these seems to exist in the channel systems of arid and semi-arid lowlands described by authors such as Antevs (1952), Leopold and Miller (1956), Williams (1971), Karcz (1972), Reineck and Singh (1975) and Mabbutt (1977) to which the terms wadi (Arabia), dry-wash and arroyo (south-west United States) have been applied.[1] Ancient examples have not been extensively documented but are known from the Devonian (Wilson, 1980) and Permian (Brookfield, 1980) of the Midland Valley of Scotland.

Arroyos are essentially the sites of ephemeral, short duration fluvial activity in which stream flow occurs only during the periods of sporadic but high intensity rainfall (e.g. Karcz, 1972) which characterises the desert environment (e.g. Reineck and Singh, 1975). These channels are commonly incised into poorly consolidated or unconsolidated sediment consisting of clay, silt, sand or occasionally gravel and typically develop vertical banks and flat floors (Antevs, 1952). They are generally recorded to be wide in relation to their depth, gently winding rather than sinuous and to be locally braided (Mabbutt, 1977). Examples of 30 metres depth and 225 kilometres length have been recorded from south-west United States by Antevs (1952) and Karcz (1972) has described similar valleys of seven kilometres width from southern Israel. Both the high width/depth ratio and the low sinuosity are suggested to be an adjustment to high sediment/water ratios, as is the braided habit of many such rivers (Mabbutt, 1977).

Deposition within the channels is usually initiated by an abrupt deceleration of flow which is partly related to the highly permeable nature of

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1 The use of these terms is restricted herein to the ephemeral channels of desert lowlands although they are equally applicable to the channel systems of the source area uplands and marginal alluvial fans (e.g. Glennie, 1970).

previously deposited, unconsolidated sediment. As the stream channels of arroyos are often braided in nature, the deposits may be similar to those of perennial braided streams (e.g. Doeglas, 1962; Williams and Rust, 1969) with the coarsest material tending to form bars which localise stream flow (and hence sedimentation of the finer material) during the falling flood stages. Other arroyos, however, such as those heading in the Sangre de Cristo mountains of New Mexico, are characterised by remarkably flat channels in which sparse spreads of gravel provide the only elevation above that of the mean channel bed (Leopold et al., 1966). Sands deposited within ephemeral channels are recorded to be typically cross-bedded as a result of the migration of bed forms such as ripples, megaripples and bars (Antevs, 1952; Williams, 1971) and scour infilling (McKee et al., 1967) although, locally, flat bedding may be the predominant structure (as a result of deposition from relatively high velocity, upper flow regime currents) and is a characteristic of the overbank deposits of Bijou Creek, Colorado (McKee et al., op. cit.). Karcz (1972) has also suggested that sheetflood processes may operate within the channel systems. Although flow is intermittent, deep scour pools at river bends may retain water for long periods after flows (Mabbutt, 1977), as can other topographic depressions (Antevs, 1952), and allow the deposition of mud horizons which may be desiccated during the final stages of evaporation (Glennie, 1970; Karcz, 1972; Reineck and Singh, 1975).

Many of the features of the interbedded sequences thus correspond closely to those described from ephemeral river systems. The dominant structure of the sandstones of flat bedding parallels that described from the Bijou Creek, Colorado by McKee et al. (1967) whereas the preservation of tabular cross-bedded units is considered to represent migrating sand bars or dunes within the channels, deposited at lower flow stages. Mud horizons record the final phase of fluvial activity within the channels and represent accumulation from suspension or evaporation. Breccia units usually truncate all underlying structures and represent less typical events in which large quantities of very coarse material[1] were swept over the previously deposited sands and muds, possibly in response to infrequent, high intensity storm events. Although the preservation potential for mudstone horizons may be expected to be exceedingly low (as indicated by the common occurrence of mud clast breccias), examples occur where the breccias directly overlie these horizons which elsewhere form relatively persistent laminae within the sandstones (see Figures 3.3, 3.5 and 3.7). In these instances, it is postulated that the breccias infill an original irregular channel bed without significant reworking of the sediment (as may occur where deposition of the coarse detritus is more or less coincident with the arrival of the flood front or flood surges) and that the sandstones represent the modified remnants of sand sheets or channel bars.

Variations in thickness of the mud horizons are also considered likely to be a primary, rather than erosional feature with thicker portions corresponding to minor topographical depressions in which waters became ponded after flooding and thus allowed the mud to settle during evaporation. Antevs (1952) records that clay films formed in such depressions, which have been termed 'chacos' by Bryan (1952 in Antevs op. cit.), are a common feature of contemporary arroyos. Thinner horizons, which are most commonly preserved as layers of curled mudstone plates in the Low House exposures, were thus probably formed in relatively less deep depressions in the channel floors and are suggested to owe their present form to desiccation following the final drying of the channel (e.g. Picard and High, 1973). This feature is thus strongly indicative of ephemeral flow conditions in which parts, at least, of the channels were subject to periodic subaerial exposure. As with examples described by Laming (1966) from the New Red Sandstone of Devon, the sectional length of the curved plates is related to the thickness of the desiccated

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1 Leopold and Miller (1956) have recorded that cobble grade material may be moved by flows no deeper than half the diameter of the rolling object.

mud horizon. The in situ preservation of the mud plates indicates subsequent deposition of the sands to have occurred under exceedingly low energy conditions. Burial by wind blown sand is a mechanism which has been recorded from a contemporary arroyo (Reineck and Singh, 1975, Fig.290, p.193) and is ideally suited for these purposes. In the Low House exposures, however, supporting evidence for such a mechanism is not available and the presence of wind blown sand sheets has not been established. Where burial does not take place, the mud horizons are unlikely to survive subsequent stream flows and may be eroded and redeposited further downstream as clasts (where transport is minimal), interstitial matrix or a second-cycle horizon. Karcz (1969) has also suggested that mud clast breccias may be derived directly from the collapse of arroyo banks exposing mud horizons. Where such horizons are desiccated, the initial size of the clast may be determined by the spacing of the mudcracks (Karcz, op. cit.).

DISCUSSION

Although the ephemeral streams of desert lowlands provide a suitable model for the formation of the interbedded breccia and sandstone sequences of the Eden Valley, it is pertinent to note that this interpretation is based on the sedimentological characteristics displayed at several, fairly well exposed localities which, perhaps, may not be completely representative. Unfortunately, the degree of exposure in localities such as Hilton Beck is such that the precise morphology of the depositional units cannot be established and the lateral persistence or impersistence of individual beds cannot be ascertained. It is possible, therefore, that these sequences may not everywhere be completely confined to definite channel systems and may also incorporate the products of sheets of water which spread across the dune field (i.e. sheetflood-type processes) as might operate at the distal margins of alluvial fans. Such an interpretation would imply these deposits to be the close counterpart to the alluvial fan deposits described in Chapter One rather than a separate but associated facies. On the basis of the close analogy with contemporary ephemeral stream deposits and the lack of remnant aeolian dune structures (as might be expected if these sequences simply reflect reworking and deposition by sheets of water spreading across the dune field), however, it is proposed to adopt the former interpretation without prejudice to the possible existence of a former, more proximal alluvial fan facies elsewhere along the eastern margin of the Eden Valley. A close modern analogue has been described by Glennie (1970) from an area north-west of Jebel Hafit in Oman.

The causes of arroyo cutting in semi arid regions of south-west United States have been discussed by authors such as Bryan (1925), Bailey (1935), Happ (1948) and Schumm and Hadley (1957) and summarised by Antevs (1952) who concluded that the most important factor is the condition of the plant cover which, under natural conditions, is controlled by the availability of moisture. Cutting occurs during periods of drought when the plant cover is reduced and filling occurs when run-off is slowed by vegetation. A similar mechanism has been proposed by Bull (1964b) for periods of alluvial fan head trenching in California. Whereas the condition of the vegetative cover may be relatively more important in the source area and thus partly control the violence of flood events within desert basins, the Eden Valley sequences are apparently devoid of any kind of preserved organic remains and it is not considered that this mechanism could have significantly contributed towards the formation of filling of the channel systems. The most likely control, in this instance, therefore, is suggested to be related only to the relative reach of the flood-waters which will be a product either of rainfall intensity in the source area or basin (i.e. a climatic control as suggested by Leopold, 1951) or the effectiveness of the source area influence (i.e. a tectonic control).

Incision would thus occur in periods of abnormally intense flow (c.f. Beaty, 1970) related to either of the above mechanisms but may also be partly determined by the gradient of the valley fill (Schumm and Hadley, 1957). Once gulleying has been initiated, however, maximum channel depth is suggested to be attained rapidly (Leopold and Miller, 1956) with subsequent downcutting proceeding only slowly. Channel widening is achieved largely by collapse of the wetted banks following flows (Leopold and Miller op. cit.) although flow diversion around channel bars may also contribute towards bank erosion (Mabbutt, 1977).

Marked variations in flow conditions may also be proposed to account for the contrasting sizes of the detrital components of the breccia and sandstone facies. Two factors, however, suggest that flow competence alone may not be the only control in this respect and relate to the lack of gradational sedimentary facies types and the occurrence of isolated large clasts within the sandstones. Whereas the former may, to some extent, be explained in terms of rapid flow deceleration, the latter suggests that the transporting flows may have been under capacity with respect to the sand grade material and could have carried coarser detritus had it been available (c.f. the flood deposits described from Bijou Creek by McKee et al., 1967). These factors seem to suggest that the influx of large quantities of detritus was sporadic and a response not only to the occurrence of infrequent, violent flood events but to the amounts of source area weathered material available for transport. It is possible that the periodicity of such floods was such that each breccia horizon represents the total weathered detritus from the source area for a particular period and that subsequent, lower power flows were sediment deficient as a result of the previous stripping of the catchment and were thus only able to transport and rework sand grade material from the aeolian dune field. A similar facies recorded from the Old Red Sandstone of the Midland Valley of Scotland (the B2 deposits of Wilson, 1980) have been interpreted as the probable products of local fan reworking by under capacity sheetfloods during periods in which coarse gravel was unavailable or rendered unavailable for transportation.

The recognition of probable contemporaneous faults from the River Belah section is thus of considerable significance in the interpretation of these deposits as such a mechanism may well have had an important affect on the availability of source area detritus, assuming a tectonic control was not restricted only to the known examples. Indeed, a comparison of the rock sequences exposed adjacent to several other faults in the Belah section (see Figure 3.3) indicates the existence of minor discrepancies with respect to the thicknesses and lateral continuity of certain of the sedimentary units and may well point towards a more complex association than is initially obvious. It is suggested, therefore, that some of the larger faults exposed in this section may also be contemporaneous. The possibility that contemporaneous faults occur elsewhere in the Eden Valley (and particularly along the eastern margin) thus cannot be disregarded and concurs with the setting of many modern desert basins in tectonically active areas (e.g. Collinson, 1978a) and the presumed development of the area during Permo-Triassic times (Bott, 1974, 1978). The model erected by Bott (1974) suggests sedimentation to have occurred in a contemporaneously subsiding basin in which the relative depression of the floor was controlled by basement faulting parallel to the Pennine fault line. Such a model provides an ideal explanation for the faulted sequences described above which are thus considered to have an origin directly related to the differential movement of the underlying basement.

A related origin may also be proposed for the formation of the sandstone filled pipes of which one example has been recorded from the Low House exposures and one from the lowermost horizons of the breccia dominated sequences exposed in Thistley Hill Quarry (NY678205). Superficially similar structures have been described by Laming (1966) from the New Red Sandstone of Devon and suggested to result from the mobilisation of saturated sand layers and their upwards injection through beds of sufficient rigidity and low permeability to fracture rather than deform or disaggregate, following the mechanism proposed by Selley et al. (1963). According to Laming (1966) the injection (and hence fracturing) is usually initiated by a sharp movement such as an earthquake which, in the case of the Eden Valley examples, may be related to the effect of contemporaneous fault movement. Bergman (1980) has listed several other mechanisms by which the upwards injection of fluidised sand may be induced but as neither of the described examples have an obvious source horizon or an attenuated sand layer at the base of the pipes, a direct comparison with quicksand injection structures is considered unreasonable. The Thistley Hill examples in particular occur immediately above Carboniferous sandstone and is infilled with a sandstone which can be closely matched with the lithology of the overlying bed. In this instance, therefore, it is postulated that these features represent contemporary infills of irregular fractures produced in semi-lithified beds by earth tremors associated with basement faulting.

CONCLUSIONS

The interbedded breccia and reworked Penrith sandstone sequences of the Eden Valley were deposited by flash floods flowing from the Pennine margin of the basin. It is suggested that these flows were confined to wide, shallow channel systems comparable to those of the lowland ephemeral streams of contemporary arid and semi-arid environments (e.g. Antevs, 1952; McKee et al., 1967) which may or may not have been related to a distal alluvial fan setting (c.f. Glennie, 1970). The coarse detritus was derived from weathered material in the upland catchments whereas the sand grade material (including that forming the matrix of the breccia horizons) was almost exclusively reworked from the aeolian dune sands of the desert basin. Transport of the coarse material is suggested to have been controlled largely by the periodicity of high magnitude flood events in combination with the availability of accumulated weathered detritus in the source area. The recognition of probable synsedimentary faults from the Belah section has suggested the possibility of a more widespread tectonic control which is in accordance with the setting of many modern desert basins (e.g. Waugh, 1973).

CHAPTER FOUR

THE CARBONATE RICH HORIZONS OF WESTERN CUMBRIA

INTRODUCTION

The purpose of this chapter is to describe the occurrence and petrography of carbonate-rich horizons recorded from the brockram sequence and the St. Bees Shale of western Cumbria. Similar horizons have not yet been recorded from the Eden Valley or southern Cumbria due to the predominance of limestone clasts and grains within the deposits which cause the whole sequence to be diagenetically calcareous. The description of these horizons from western Cumbria does not, therefore, preclude the possibility of their existence in other brockram localities, although the recognition of such horizons would present numerous difficulties and be, to some extent, problematical.

Carbonate-rich horizons have been recorded in several boreholes, drilled by the British Steel Corporation, but are not known from the brockram exposures of the area, the limited extent of which has been discussed previously. Detailed examinations of these horizons have, therefore, been restricted to borehole cores, the maximum diameter of which is approximately seven centimetres. Although allowing accurate determinations of the vertical sequence, the spacing of the boreholes and problems of correlation do not enable lateral continuity or lateral variations within particular horizons to be examined.

FIELD OCCURRENCE OF THE CARBONATE-RICH HORIZONS

Carbonate-rich horizons are recorded from a number of boreholes drilled in the brockram depositional basin, such horizons being especially common in the more westerly areas. Although a large number of boreholes have been completed in the area, calcareous horizons have been recorded at only twelve localities. Of these twelve, five[1] have provided core samples for analysis within the past few years and only seven[2] former occurrences are recorded. It is likely, however, that the extent of the carbonate-rich horizons is far more widespread than this small number of boreholes suggests, as the detailed logging of the brockram sequence has only been purposefully undertaken in recent years and information is still comparatively sparse.

Where carbonate-rich horizons occur, they are restricted to the finer sand or silt members of the sequence overlying or separating coarser bands of breccia. The junction with the overlying St. Bees Series is gradational throughout the basin and especially towards the west where coarser horizons frequently occur within the shales and sandstones of the St. Bees Series. Carbonate-rich horizons are not restricted to the brockram facies and commonly extend into the overlying sediment.

The occurrence of these horizons is associated with an important facies variation within the brockram sequence. Figure 4.1 shows the sequence recorded from Surface Borehole (SBH) 304, near the village of Carleton at NY01770963. From a comparison of SBH 304 with previous descriptions of boreholes from elsewhere in the depositional basin, it is immediately apparent that sandstone and siltstone units form a much higher percentage of the brockram sequence. These sandy horizons occur throughout the sequence and vary from thin sandstone partings, less than 0.2 metre thick, to units which exceed six metres in thickness. Six such horizons contain concentrations of carbonate. The total thickness of the brockram sequence, at this locality,

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1 Surface borehole 302 (NY01581024), SBH 303 (NY01860967), SBH 304 (NY01770963), SBH 306 (NY01940950) and SBH 309 (NY00131014).

2 SBH 287 (NX99061081), SBH 288 (NX98861093), SBH 293 (NY01331015), (?) SBH 307 (NY02140935), SBH 651 (NX99731066), SBH 670 (NX99840997) and (?) SBH 804 (NY00881019).

Figure 4.1. Facies log of brockram sequence recorded from SBH 304 (NY01770963) See Appendix Two for legend.


St. Bees Shale 540·36 m.AMD
540
530
520
510
500
490
480
0 50 % Sand
0 50 % Silt
0 50 % Clay
0 50 % $CaCO_3$
0 75 % Clasts
Fifth Limestone 475·95 m.AMD

Figure 4.1

is 64.4 metres. The thickness recorded in boreholes east of Carleton generally indicate a thickening of the sequence to a maximum of over 120 metres in SBH 258 and SBH 531. These boreholes are located at NY03710732, near Low Godderthwaite and at NY03520981, near Whitehow Head respectively. Although detailed records from areas west of Carleton are comparatively sparse, Figure 4.2 shows the successions recorded from SBH 287 and SBH 288. These holes were drilled near Watson Hill at NX99061081 and NX98861093 respectively and both record the brockram sequence to be less than 15 metres thick. At Saltom Bay (NX95901600) the Carboniferous Coal Measures Whitehaven Sandstone is overlain by 1.8 metres of brockram basal breccia which is, in turn, overlain by the Magnesian Limestone. The Magnesian Limestone is also recorded at this horizon in SBH 287 and 288.

The general thinning of the brockram sequence westwards and the associated facies variation from a complete succession of breccias in the east, through alternating breccias and sandstones and siltstones in central parts of the area, to a thin breccia overlain by the St. Bees Evaporites in the west suggests a corresponding environmental succession from a proximal alluvial fan setting to an extreme distal setting where basin topography was so reduced as to allow the formation of the St. Bees Evaporites sabkhas (Arthurton and Hemingway, 1972). Although in these western parts of the basin the Magnesian Limestone and other cycles of the St. Bees Evaporites separate the brockram from the overlying St. Bees Shales, a short distance to the east this junction becomes less distinct as a result of the occurrence of coarser horizons above the Magnesian Limestone. This succession is illustrated in Figure 4.2 which shows the sequence recorded from SBH 651, drilled at NX99731065, near Watson Hill, but to the east of SBH 287 and 288. It is likely that these coarser horizons represent distal alluvial fan sediments

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Figure 4.2. Facies logs of sequences recorded from SBH 287 (NX99061081), 288 (NX98861093) and 651 (NX99731066). See Appendix Two for legend.


SBH 287
St. Bees Sandstone 438·3
440
420
400
St. Bees Shales 376·2
380
St. Bees Evaporites 348·8
360
Basal Breccia
340
Hensingham Group 338·2m.AMD
SBH 288
St. Bees Sst. 426·7
Shales 401·1
Evaporites 376·1
Basal Breccia
Hensingham Group 362·7m.AMD
SBH 651
490
460
Shales 451·3
440
Evaporites 430·0
Shales 428·2m.AMD (Faulted contact)

Figure 4.2

alternating rapidly with sandstones, shales and mudstones which may themselves be part of the distal alluvial fan sequences or which, more probably, represent the coastal plain mudstones of the St. Bees Shales.

The sediments in which the carbonate-rich horizons of the Carleton borehole (SBH 304) occur may, similarly, either represent finer members of the fluvial brockram sequence or part of the basinal facies of the St. Bees Shales with which the basin-margin alluvial fanglomerates interdigitate. It is clear, however, that the occurrence of the carbonate horizons is a feature of distal portions of the alluvial fans and are not known from the more proximal sediments to the east.

Where both the Magnesian Limestone and carbonate-rich horizons are recorded in the same borehole, the carbonate concentrations are generally restricted to the sediment overlying the limestone. The limestone, therefore, forms a useful reference horizon. A similar association is recorded by Waugh (1978, pers. comm.) from the St. Bees Shales exposed in a coastal section at St. Bees North Head (NX949152). At this locality a massive-bedded siltstone is intermittently exposed in a waterfall. The lower three to four metres of exposure is characterised by the presence of carbonate nodules within the sediment. The siltstone also contains a number of gritty horizons of included Borrowdale Volcanic fragments. The nodules are mostly disseminated throughout the siltstone but are occasionally concentrated in more distinct bands. The nodules are arranged with their long-axes parallel to the bedding and are generally less than five centimetres in length. The exposures of North Head are illustrated in Plate 4.1.

Stratigraphically, the exposures of North Head occur above the St. Bees Evaporites and are within the upper parts of the St. Bees Shales. They have, therefore, a similar stratigraphic position to the examples recorded from the Watson Hill boreholes.

CONFIGURATION OF THE CALCAREOUS HORIZONS

From the evidence of borehole samples, the distribution of the calcareous horizons within the sandstone units may be sub-divided into two main forms:

1. Nodular accumulations of carbonate which may occur either as discrete nodules or diffuse patches and which are usually disseminated throughout the sediment. Horizons of this type are the most common mode of occurrence and are usually bounded both above and below by coarse breccia horizons. The form of the 'nodule' rarely differs throughout the thickness of the sandstone unit. Accumulations of this type are illustrated in Plate 4.2A (upper part of the core series).
2. Accumulations which show a vertical sequential increase in carbonate content. These accumulations may be either of the discrete or diffuse types and commonly form distinct bands of carbonate at the top of the sequence. The top of the sequence does not necessarily conform with the top of the sandstone unit in which it occurs. Frequently, two or more such sequences occur in the same unit or occur in association with accumulations of the first type. Accumulations of the second type are illustrated in Plate 4.2A (lower part of the core series) and distinct Carbonate horizons are illustrated in Plate 4.2B.

Disseminated nodular accumulations

Accumulations of this type are common throughout the brockram sequences recorded from the more westerly parts of the area. Although the finer sediments in which they occur, within the brockram sequences, are generally less than five metres thick, similar lithologies of the overlying St. Bees Shales are less restricted by breccia horizons and allow the formation of carbonate concretions over greater thicknesses. For example, over 30 metres of sediment containing carbonate nodules is recorded from the Watson Hill area

(SBH 287). It is clear, therefore, that the comparatively thin carbonate horizons recorded from the brockram facies can be attributed to the correspondingly small thicknesses of the finer sandy and silty horizons.

The nodules are oval or flattened in cross-section and are typically orientated with their long-axes parallel to the bedding of the host siltstone. This relationship is preserved even where the sediment is cross-bedded. The maximum recorded length of the nodules is in the order of five centimetres.

For any one unit containing carbonate concretions, an upward vertical gradation from small, highly disseminated and diffuse carbonate cemented zones through to larger distinct nodules may be observed. The distinct nodules, which are usually more closely spaced, are sometimes characterised by a series of fractures infilled with host sediment or, less commonly, calcite. Similarly with the second form of carbonate accumulation, two or more such sequences may be present in the same sandstone unit. Gritty horizons may also be present.

It is evident that the exposures of St. Bees Shales at North Head fall within this category. The occasional concentration of the nodules into more distinct bands suggests a gradation to the second form of accumulation.

<u>Sequential nodular accumulations</u>

These differ from accumulations of the first type in that a more complete vertical profile is developed. The full sequence may be sub-divided into three main zones:

3. Zone of massive carbonate; usually overlies and grades into zone 2.
2. Zone of disseminated nodular carbonate; usually overlies and grades into zone 1.
1. Zone of disseminated diffuse carbonate.

The zone of massive carbonate forms an obvious carbonate-rich band which may either be diffuse, in which case it occurs as a layer of cemented sediment, or distinct where a hard calcareous band with well-defined upper and lower margins is developed. This layer may also be formed from tightly packed, highly flattened nodules separated by thin clay or silt partings which are deformed around the nodules and give rise to a 'chicken-wire' texture analogous to that described by Bosellini and Hardie (1973) for gypsum. Clay partings and gritty horizons are frequently included within this layer, the clay bands, in particular, often being ruptured by the carbonate. As with the distinct nodules, this layer is characterised by a series of vertical and oblique fractures which are infilled with host sediment or calcite. Whereas the lower margin of the band grades into the underlying zone of nodular accumulation, the upper margin may either form the base of the succeeding carbonate sequence or mark the final phase of carbonate accumulation for that particular siltstone horizon of the brockram. In the latter case, and where the sequence is overlain by a coarser breccia horizon, the upper margin of the carbonate band may be extremely irregular, as the result of solution and erosion. The lowest zones of the breccia horizon frequently contain included angular fragments of carbonate derived from the underlying sequence. The formation of these zones precedes, therefore, the deposition of the overlying sediment and may be regarded as penecontemporaneous. The maximum thickness recorded, for this zone, is in the order of four centimetres.

The lower two zones of the sequential accumulations are equivalent to those of the gradational disseminated accumulations. Within the middle zone of nodular carbonate, however, the density of the nodules noticeably increases towards (as does, therefore, the carbonate content), and grades into, the uppermost zone.

This second form of accumulation, therefore, merely represents a more complete and ordered development of the first form, but is distinguished by being far less common in occurrence.

PETROGRAPHY OF THE CARBONATE-RICH HORIZONS

MINERALOGY AND TEXTURE

Petrological analyses of thin sections of the sand-silt horizons occurring within the brockram sequences of western Cumbria indicate that unicrystalline quartz forms the most common detrital grain. Feldspar and Borrowdale Volcanic fragments form the only other common detrital component, but grains of these types are typically characterised by a high degree of dissolution and replacement by clay (Plate 4.3). In many cases, the occurrence of authigenic clay pseudomorphs denotes what are probably totally replaced feldspar or Borrowdale Volcanic grains (Plate 4.4A and B). Excepting the carbonate accumulations, framework grains are enclosed by a well-defined (mechanically infiltrated?) clay rim and have an interstitial clay matrix.

Zone of disseminated diffuse carbonate

Within the zone of disseminated, diffuse carbonate, the sediment of the 'nodular' masses is distinguished mineralogically in that carbonate cement is contained within the interstitial clay matrix of the framework grains (Plate 4.4C). Unlike the coarsely crystalline sparry calcite of the breccia sequences, the carbonate of these horizons is predominantly micritic or microsparitic in texture and only occasionally forms as sparite. The Scanning Electron Microscope (SEM) indicates that the micrite is formed of disorientated microgranular accumulations of closely packed carbonate crystals (Plate 4.4D). Thin section staining techniques indicate that the carbonate is invariably non-ferroan calcite. The 'nodules' have indistinct lateral and vertical confines which are gradational with surrounding sediment (Plate 4.5A).

Zone of disseminated nodular carbonate

Within the zone of disseminated, nodular carbonate, the accumulations are characterised by their generally well-defined margins and abrupt contacts with surrounding sediment. However, although the nodules commonly have an abrupt upper margin (Plate 4.5B), the lower contacts, especially towards the base of the zone, tend to be gradational and less well defined. Similarly with the zone of diffuse carbonate, the sediment is distinguished mineralogically by containing micritic (Plate 4.5B) or microsparitic (Plate 4.5C) carbonate, but carbonate which forms a cement rather than being contained within the interstitial clay matrix (Plate 4.5D). The cement is commonly microgranular, formed of disorientated calcite crystals, but may also form as interlocking, elongated crystals orientated with their long-axes perpendicular to detrital sand or silt grains. The dimensions of the crystals increase away from the detrital grains (Plate 4.6A and B).

The mineralogy of the nodules differs from that of the underlying zone in that they contain fewer detrital grains, which are enclosed by less obvious clay skins, and larger quantities of carbonate. The clay content of the concretions appears to be less than that of the original sediment. The clay skins are clearly recognisable, however, using the SEM (Plate 4.6C).

The texture of the nodules in this zone may be distinguished from that of the surrounding sediment, especially towards upper parts of the profile where these textural differences become more apparent. Unlike the fresh detrital grains of the surrounding sediment, the quartz grains within the nodules are corroded and have, in thin section, irregular outlines infilled by clay and micritic carbonate (Plate 4.6D and Plate 4.7A and B). Although, from these photomicrographs, it appears that the quartz grains have been exclusively replaced[1] by clay, SEM examination of similar examples shows the clay to contain a high proportion of micritic carbonate (see, for example, Plate 4.8D).

The corrosion of detrital quartz grains is demonstrated particularly well by SEM photomicrographs (Plates 4.7C to 4.9D). In these examples, the

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1 This term is used herein to describe a process of crystallisation and concomitant (consequential) dissolution.

surfaces of the quartz grains are covered by a dense arrangement of replacement pits, many of the grains having been more or less completely destroyed to form 'skeletal' grains. Plate 4.9C demonstrates the marked differences between diagenetically modified quartz and feldspar grains. Both have sustained extensive dissolution, but whereas the quartz grain is only partially replaced by carbonate, the feldspar grain is almost totally replaced by authigenic clay. Although many of the photomicrographs show quartz grain replacement pits to contain small crystals of calcite, the replacement of original detrital quartz by calcite is particularly well seen from Plate 4.10A, B and C. In this example, the interstitial cement is formed partly of a radial growth of elongated carbonate crystals, the growth of which appears to have originated from the surface of a totally replaced (feldspar?) grain. During this growth, an adjacent quartz grain has been corroded and replaced in the immediate 'micro-vicinity' of the carbonate crystals.

The percentage composition of detrital grains within the nodules is very much less than in surrounding sediment with grains 'floating' in the cement rather than forming the framework of the sediment (Plate 4.10D). The nodules are also noticeably deficient in the finer quartz grain sizes. In higher parts of the zone, the nodules are sometimes fractured, these fractures being infilled with either host sediment (Plate 4.11A) or, less commonly, coarsely crystalline calcite (Plate 4.11B). Where fractures are infilled with calcite, the margins of the fracture are generally lined with micrite. There is also a tendency, in these higher zones, for the carbonate directly surrounding detrital grains to be somewhat coarser than the otherwise micritic cement.

The sedimentary laminae of this zone are deformed around the nodules and are in part ruptured by, and domed over the upper surfaces of the nodules (Plate 4.11C). These effects are less well displayed towards the lower margins of the nodules and in lower parts of the profile, where the carbonate is contained within the interstitial clay. This small scale deformation indicates either a displacement of the host sediment during formation of the nodules or, more probably, the effects of a later, differential compaction of the unconsolidated sediment. Any displacement caused by the formation of diffuse carbonate accumulations is accommodated, presumably, by a redistribution of interstitial clay matrix, as indicated by the concentrations of matrix clay above, and to a lesser extent, below the concretions (see Plate 4.5A). The displacement of matrix clay during the formation of the carbonate may account for the apparent replacement of detrital quartz by clay (see Plates 4.6D and 4.7A and B).

Zone of massive carbonate

The mineralogy and texture of the zone of massive carbonate is similar to that recorded from discrete nodules of the underlying zone, mainly differing in the extent to which carbonate is developed. Within massive horizons, carbonate may form up to 60 per cent. of the sediment. As noted in the preceding section, this zone may either be formed of a tightly packed mass of flattened nodules, separated by ruptured and distorted laminae of host sediment or clay, or of distinct bands in which the nodular form is absent, but which may still contain discontinuous and distorted laminae of silt and clay (Plate 4.12A). In Plate 4.12B, a clay lamination is partly contained by a carbonate concretion and partly by original sediment. The deformation of that part of the lamination contained within the nodule clearly demonstrates the disruptive effects of carbonate formation on original sedimentary lamination.

Fracturing is a common feature of this zone and is especially well developed where the upper surface is eroded. Frequently sections also contain sub-circular or oval masses of blocky crystalline calcite, commonly with an outer rim of micrite (Plate 4.12C and D). The dimensions of the crystal plates increase towards the centre of the mass, a texture which suggests the calcite to be infilling cavities, presumably formed during the genesis of the carbonate concretions.

The carbonate of the massive zone may occasionally be formed of recognisable crystal plates which probably represent a post formational recrystallisation of the original micrite.

Secondary silica in the carbonate profiles

Secondary quartz, in the form of overgrowths around detrital grains, occurs rarely throughout the carbonate-rich zones but is generally more common in the lower zones of the profiles (Plate 4.13). Where the overgrowths occur within carbonate accumulations, they may, themselves, be corroded, the surface of the crystals being characterised by replacement pits and hollows (Plates 4.14A to 4.17A). Similarly with the detrital grains, these pits occasionally contain carbonate crystals which are clearly replacing the secondary quartz. Where cross-sections of grains are seen, the overgrowth is commonly corroded preferentially to the original detrital grain (Plate 4.17B, C and D).

As the carbonate profiles are characterised by an overall vertical increase in the carbonate/quartz component ratio and a corresponding increase in quartz dissolution, it is possible that the apparent absence of secondary silica in higher parts of the profile is a product of increased corrosion rather than an 'original' feature. That the overgrowths are corroded suggests that their formation precedes the process of quartz replacement by carbonate.

However, since secondary silica has not been recorded from elsewhere in the brockram sequences, it is considered that the genesis of the quartz overgrowths is related to the formation of the carbonate and that the absence of secondary silica from higher zones of the profiles is not a product of their complete dissolution.

Summary of the petrographic characteristics of the carbonate-rich profiles

In comparison with carbonate-free sand-silt horizons, the carbonate-rich zones possess a number of distinct mineralogical and textural characteristics, which are summarised below:

1. The successive carbonate forms of the complete profile represent a transition from sediment which is essentially carbonate-free to sediment which is carbonate-rich.
2. There is a corresponding transition from sediment with a detrital grain framework and interstitial matrix to sediment in which the grains 'float' in a micritic or microsparitic calcite cement. This transition is also marked by an obvious reduction in the number of grain contacts.
3. Whereas clay rims enclosing framework grains are distinct, those of the carbonate-rich zones are less easily recognised and often dislocated by the growth of calcite crystals (see, for example, Plate 4.5D or Plate 4.8A).
4. Quartz grains of the carbonate cemented zones are extensively corroded and replaced, whereas those of the host sediment are uncorroded.
5. The carbonate-rich zones are deficient in the finer quartz grain sizes.
6. Authigenic silica, in the form of overgrowths around detrital quartz grains, may be common and is especially common in lower zones of the profile. Although occurring in both host sediment and nodules, secondary quartz has not been recorded from comparable, but carbonate-free sand-silt horizons of the brockram sequences.

MINERALOGICAL COMPOSITION

Figure 4.3 demonstrates the compositional variations recorded from sand-silt horizons within the brockram sequence of SBH 304. The data for the graphs included in Figure 4.3 are derived from 35 representative modal analyses, selected from a total of over 200, and indicate that although there

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Figure 4.3. Composition of sand-silt horizons recorded from SBH 304 (NY01770963).

CaCO3%
a.
Detrital quartz sand %
Detrital quartz silt %
Regression analysis for silt/CaCO3 component
b.
Matrix clay %
c.
Combined detrital quartz sand and silt %

Figure 4.3

is little relationship between the sand/carbonate and clay/carbonate components, there is a definite and statistically significant correlation between the silt/carbonate and combined sand and silt/carbonate values ($p < 0.005$ and $p < 0.001$ respectively). These values are simply related in that samples containing a high percentage of carbonate contain less sand and silt than the carbonate-free samples. The relationship between the clay/carbonate ratios indicates the percentage composition of clay to retain approximately the same value for all samples examined.

By reference to the textural features of the sediment and the diagenetic processes described in the preceding section, it is possible to substantiate the data of Figure 4.3. The most important diagenetic process described is the dissolution of detrital quartz sand and silt grains and their replacement by micritic calcite. As a generalisation, it can be seen that the volume of carbonate precipitation more or less equals that of quartz dissolution, so that these horizons may be formed purely by replacement and necessitate no overall volumetric increase for the sediment. However, the distortion of sedimentary laminae around the nodular formations and the fracturing of the nodules does indicate some slight increase in volume and suggests carbonate precipitation to exceed dissolution in volume. Textural considerations have also indicated a reduction in the finer (silt) quartz grain sizes within the carbonate accumulations; a relationship which is implied in Figure 4.3a and which is confirmed statistically, although at a lower level of probability than for the combined quartz value.

As the clay content of the carbonate concretions is apparently less than that of the original deposit, it could be assumed that increases in carbonate content would, similarly, be accompanied by a corresponding, proportional decrease in the clay content of the sediment. However, such a relationship is contradicted by the data of Figure 4.3b, which shows the clay content to retain approximately the same value throughout the range of carbonate content.

This, in association with the deformational features associated with the concretions, suggests that a proportion of the clay fraction may be redistributed during the formation of the concretionary horizons. The clay may either be incorporated within the matrix of surrounding sediment or form laminae within the concretions.

In summary, therefore, the formation of carbonate within these sediments is a process which is largely effected by replacing the detrital quartz component and redistributing or incorporating the clay matrix of the original deposit.

INTERPRETATION

THE CARBONATE PROFILE

From the preceding morphological and textural descriptions, three main carbonate zones may be recognised on the basis of their petrographic characteristics and the gross form of the carbonate. Authigenic silica, in the form of overgrowths around detrital grains, is an important feature of the sequences and has been recorded only in association with the authigenic carbonate of the profiles. The silica does not seem to be restricted to a particular carbonate zone and occurs within both the diffuse and lowermost nodular zone.

It is also apparent that the occurrence of a particular zone is determined according to a definite profile which represents a transition from low carbonate content (diffuse accumulations) to high carbonate content (massive accumulations). Massive carbonate zones commonly occur only above nodular and diffuse accumulations respectively. Where nodular accumulations mark the top of the sequence, underlying carbonate forms are generally restricted to those of the zone of diffuse carbonate. Occasionally, inversions of this sequence may (apparently) exist, but, as will be determined subsequently, are believed to be polygenetic in origin. This ordering of the zones is, therefore, a genetic sequence in which the higher zones represent a more advanced development than that of the underlying zones. Incomplete profiles could have developed, therefore, into complete sequences, had formational conditions persisted. In this manner, nodular accumulations may be recognised as immature stages of the relatively more mature sequential accumulations.

The complete profile is summarised below and illustrated diagrammatically in Figure 4.4a.

Zone 3. Zone of massive carbonate. Forms the uppermost unit of the sequence in which carbonate comprises up to 60 per cent. of the sediment and which forms a distinct horizon. The diagenetic fabric of the zone may exceed 90 per cent. Grades down into the underlying zone 2.

Zone 2. Zone of disseminated nodular carbonate. Carbonate comprises approximately 30 to 10 per cent. of the sediment. Authigenic silica may occur towards the base of the zone. Grades downwards into the underlying zone 1.

Zone 1. Zone of disseminated diffuse carbonate. Carbonate comprises generally less than 15 per cent. of the sediment. Authigenic silica may occur throughout this zone. Grades downwards into authigenic carbonate-free sediment.

The occurrence of eroded fragments of carbonate possessing petrographic characteristics identical to those of the carbonate of the profiles within associated intraformational breccias indicates the formation of the nodules to be penecontemporaneous with that of the host sediment.

Similar sequential arrangements of carbonate have been described from a number of ancient and modern sedimentary environments. In ancient

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Figure 4.4. Idealised carbonate profile (a) compared with caliche profiles recorded from the High Plains, New Mexico and Texas (b and c), reproduced from Aristarain (1970, p.205) and Reeves (1970, p.357) respectively.


a
b
c
2
1
0
Metres
Soil
Conglomerate / breccia
Laminar carbonate
Layered carbonate *
Nodular carbonate
Diagenetic quartz zone
* Terminology of Aristarain (1970)
W.Cumbria
High Plains
(Aristarain, 1970, p. 205)
High Plains
(Reeves, 1970, p. 357;
scale approximate)

Figure 4.4

environments the term 'cornstone' has been applied to these deposits, whereas for modern environments the term 'caliche' has been used. Because of the close analogy between the carbonate-rich profiles of western Cumbria with both caliches and cornstones, it is proposed to include a brief summary and comparison of the main characteristics of these deposits.

ANALOGY WITH CORNSTONES

Carbonate concretions have been described from many continental sequences occurring in the British Isles and elsewhere, e.g. Allen (1960, 1965, 1973), Burgess (1961), Pick (1964), Bruck et al. (1967), Francis et al. (1970), Williams (1973), Steel (1974a) and Higham (1977). The term 'cornstone' has been widely applied to these concretionary masses which are suggested to be the product of a post depositional accumulation of authigenic carbonate (Steel, op. cit.). The similarity between cornstones and caliche has been recognised by many authors who suggest their formation to be the result of an identical pedogenic process.

Cornstones generally occur in the fine sandstone or siltstone members of fluviatile fining upward cyclothems but are occasionally recorded from conglomerates (Bruck et al., 1967; Steel, 1974a). Similarly with brockram examples, there is usually a vertical increase in carbonate content from an underlying carbonate-free zone, through nodular accumulations in which the nodules show a corresponding vertical increase in size, to massive accumulations containing only rare inclusions of original sediment. The concretions characteristically exhibit a wide variety of forms but are commonly irregular or vertically elongate (e.g. Allen, 1965; Steel, 1974a). Only a few examples, such as those described from the Lower and Middle Devonian of northern Scotland by Higham (1977), are recorded to be horizontally elongate; a form which Pick (1964) attributes to the morphological control exerted by sedimentary laminae (viz. the cross-laminated siltstones and concordant concretions recorded from the brockram sequences) and, Burgess (1961), to compression.

The uppermost zone of cornstone development is commonly abrupt but occasionally may be transitional through a poorly developed nodular deposit (Allen, 1973). Profiles are frequently overlain by, and may be eroded by, the coarse-grained basal members of succeeding fining upward units (Burgess, 1961; Bruck et al., 1967; Francis et al., 1970; Steel, 1974a).

The zonal arrangement of these deposits has enabled a number of distinct cornstone profiles to be determined, the most comprehensive of which is probably that described from the New Red Sandstone of western Scotland by Steel (1974a). Steel (op. cit.) arbitrarily categorises cornstones into five main types according to the gross form of the carbonate and the approximate proportion of host rock still distinguishable:

Type 1. Small (one to six centimetres in diameter) irregularly shaped nodules occasionally fractured with veins of host rock. Nodules comprise less than ten per cent. of the rock.

Type 2. Larger (up to ten centimetres in diameter), often vertically elongate nodules. Nodules comprise less than 50 per cent. of the rock in the upper part of the profile. There is usually a downwards gradation into cornstone of type 1.

Type 3. Nodules, vertical pipes or horizontal sheets of carbonate. Carbonate comprises in excess of 50 per cent. of the rock and contains clastic sediment between the carbonate framework. Gradation to type 2 cornstone at the base.

Type 4. Beds of carbonate with rare patches of clastic sediment. Gradation to type 3 cornstone at the base.

Type 4a. Distinct horizons of laminated, brecciated or pisolitic carbonate, commonly capping type 4 carbonate.

Minor discrepancies and differences of terminology allow only approximate correlatives of the zones erected by Steel (1974a) to be recognised from the brockram profiles. As a generalisation, however, the zone of massive carbonate can be equated with types 3 or 4 cornstones on the basis of carbonate content and the similarity between calcite filled voids recorded from this zone and the drusy cavity infilling described by Steel (op. cit.). Both Allen (1965) and Higham (1977) record a comparable development of coarsely crystalline calcite from the higher parts of Old Red Sandstone profiles of Anglesey and northern Scotland, but also note that this zone may be comprised of a compact mass of concretions.

The zones of nodular and diffuse carbonate may be equated, therefore, with Steel's (1974a) types 3 to 1 cornstones. The absence of textures similar to those of the laminated, brecciated and pisolitic type 4a cornstones indicates that the brockram profiles are less completely developed than the New Red Sandstone examples and are, in comparison, immature.

ANALOGY WITH CALICHE

Accumulations of authigenic carbonate in the profiles of desert soils have been described by a number of authors, notably, Breazeale and Smith (1930), Sidwell (1943), Bretz and Horberg (1949), Brown (1956), Gile et al. (1965, 1966), Gile (1967), Ruhe (1967), Aristarain (1970), Reeves (1970), Hay and Wiggins (1980) and Watts (1980). Their formation has been the subject of considerable controversy (for a discussion of which, see Brown, 1956 or Goudie, 1973), but is generally attributed to a pedogenic process in which carbonate is concentrated at specific horizons within the profiles, as a result of sub-surface evaporation of downwards percolating carbonate enriched vadose waters. This mechanism corresponds to the per descensum model of Goudie (1973, p.136). The calcium carbonate is believed to be derived from two main sources, which are summarised below. Caliche crusts formed by the in situ solution and redeposition of carbonate host rock or sediment (Van Siclen, 1957; Blank and Tynes, 1965; Multer and Hoffmeister, 1968; Walls et al., 1975; Kahle, 1977; Chafetz and Butler, 1980) and types such as described by Klappa (1978) from the western Mediterranean are considered to be dissimilar to both the above examples and the brockram profiles and will not be included in the succeeding discussion.

Soil caliche may form from the depletion of upper zone soluble calcium compounds. Downwards percolating waters enriched in carbon dioxide dissolve calcium compounds from the soil and rock but, as a result of surface desiccation and sub-surface evaporation, $CaCO_3$ is precipitated in lower zones of the soil profile at, or near, the maximum depth of water penetration. This process can only operate where the sediment is originally calcaeous. This mechanism has been suggested to account exclusively for the formation of caliche by such authors as Bretz and Horberg (1949). Reeves (1970), however, concludes that only relatively thin profiles are likely to form in this manner, due to the limited availability of A zone solubles.

The majority of caliche profiles are believed to be formed from $CaCO_3$ derived from desert loess (Brown, 1956; Dan and Yaalon, 1966; Gile et al., 1966; Gile, 1967; Ruhe, 1967 and Reeves, 1970), although minor quantities may also originate from dissolved ions in rainwater and from the decay of vegetable matter (Goudie, 1973). The wind blown dust forms a thin surface veneer from which $CaCO_3$ is dissolved during wetting (precipitation or intermittent fluvial activity) and transported into the soil. The $CaCO_3$ is precipitated during dry periods, as the water evaporates. Initially, the carbonate is precipitated interstitially but, with continued alternation of sub-surface saturation and surface desiccation, a zone of intense carbonate impregnation eventually develops. This zone forms at the depth of most frequent wetting and acts as an impervious or 'plugged' horizon which, with time and continued saturation, slowly extends towards the ground surface. In this manner, the existence of equally prominent carbonate horizons in calcareous and non-calcareous sediments can be explained simply (Gile et al., 1966) and without assuming the latter to be the result of the destruction of a pre-existing carbonate component, as proposed by Bretz and Horberg (1949).

Brown (1956) points out that prominent profiles are more likely to develop in aggrading profiles where the thickness of the caliche is not restricted to that of the maximum depth of penetration of the carbonate enriched solutions.

The second mechanism, which may involve considerable expansion of the soil profile, could also provide an explanation for the anticlinal structures recorded by Price (1925) and Jennings and Sweeting (1961). Similar structures have been recorded in association with cornstone sequences by Allen (1973) and Higham (1977).

Morphology of the profiles

Caliche profiles range widely in thickness but are recorded to reach a maximum thickness of about 60 metres (Goudie, 1973). Most caliche profiles, such as those of south-eastern New Mexico (Bretz and Horberg, 1965), form profiles which are in the order of one to five metres in thickness. These profiles are generally comprised of nodular and laminar accumulations of authigenic carbonate and are usually overlain by a variable thickness of soil. Similarly with cornstones, the lower contact of the caliche is characteristically gradational with the underlying sediment. Caliche has been described from alluvial fans and commonly occurs in sediments ranging from gravelly sands to silt loams (Gile, 1961).

The morphologies of many caliche profiles have been described by such authors as Hawker (1927), Bretz and Horberg (1949), Aristarain (1970) and Reeves (1970) and a diverse descriptive terminology has been used for the various zones of carbonate development. However, Gile et al. (1965), studying caliche from southern New Mexico, have proposed the terms K horizon and K fabric for describing prominently developed carbonate accumulations. The K horizon is a master soil profile, formed of a number of sub-horizons, the term having been proposed as an alternative to pre-existing, anomalous soil classifications. The K fabric is the diagenetic fabric of the K horizon and is defined by Gile et al. (1965, p.74) in the following manner:

> 'In K fabric, fine grained authigenic carbonate occurs as an essentially continuous medium. It coats or engulfs and commonly separates and cements, skeletal pebbles. Interstices between skeletal grains are partly or completely filled with carbonate.'

The term 'skeletal' refers to the framework grain texture of the sediment and is not descriptive of individual detrital grains (viz. the skeletal grains of this study). The above definition of K fabric thus excludes those accumulations in which carbonate does not occur as a 'continuous medium'. For carbonate to form a continuous medium, its compositional percentage must attain a certain value (dependent on the textural characteristics of the host sediment), which has been determined as a minimum of between 15 and 40 per cent. However, the concept of the K horizon has been modified subsequently (Gile et al., 1966) in order to incorporate immature, less prominently developed sequences, thought to represent the initial stages of formation of the K horizon. The main features of these profiles are summarised below.

A relatively soft and nodular accumulation of carbonate, containing at least 50 per cent. K fabric, may occur near the top of the soil profile. This is referred to as the K1 horizon and may be represented in western Cumbria by the diffuse accumulations occasionally recorded above massive carbonate zones. Difficulties exist in the interpretation of such horizons as an identical association of carbonate types may be formed polygenetically (Aristarain, 1970, p.207, for example) where a second phase of incomplete carbonate accumulation proceeds further sedimentation. The differentiation of K1 horizons sensu stricto is problematical, therefore, although they have been recognised in ancient examples, e.g. Allen (1973). The K2 horizon is the most prominent horizon of modern caliche profiles and contains at least 90 per cent. K fabric which usually has a blocky or brecciated texture in the lower part and is overlain by a zone of laminated carbonate. The K2 horizon probably equates with the laminated zone (and part of the underlying nodular zone) of Reeves (1970), the layered caliche of Aristarain (1970) and the indurated zone of Williams (1973). In mature K2 (K2m) horizons, laminated carbonate is usually closely associated with amorphous silica and with the development of pisolitic textures (including pelletoid and ooid structures as described by Hay and Wiggins, 1980). The pisolites are often concentrated above laminated horizons but may also occur within the non-laminated K2m and as infills of fractures in the K2m horizon (Hay and Wiggins, 1980). The K2 horizon thus corresponds to the uppermost massive carbonate zone and part of the underlying nodular zone of the brockram sequences. In the K3 horizon, carbonate content and volume of K fabric decrease with increasing depth until the caliche-free horizon is reached. This zone is also referred to as the transitional (Reeves, 1970) or friable caliche zone (Aristarain, 1970; Williams, 1973) and is formed of diffuse carbonate with occasional poorly developed concretions (Aristarain, 1970; Reeves, 1970). This horizon may be equated with the zone of disseminated, diffuse carbonate of the brockram sequences.

Reeves (1970) has recognised a further zone which occurs above the transitional zone (K3 horizon) and which separates it from the overlying nodular and laminar zones (K2 horizon). This zone is characterised by the presence of authigenic silica and is referred to as the diagenetic quartz zone. Similar occurrences have been recorded by Williams (1973) from the New Red Sandstone of the Elgin basin and are represented in the brockram sequences by the occasional development of quartz overgrowths around detrital grains.

Caliche profiles studied by Aristarain (1970, p.205) and Reeves (1970, p.357) are reproduced in Figure 4.4b and 4.4c for comparison with the brockram sequences illustrated in Figure 4.4a.

Gile *et al.* (1966) have recognised two morphological sequences leading to the formation of K horizons. For non-gravelly sediments, the sequence represents an increase of authigenic carbonate content from faint grain coatings (Stage I), through nodular concentrations (Stage II) to nodular accumulations with internodular carbonate (Stage III) and finally to a laminar horizon overlying a plugged horizon (Stage IV). The K horizon is recognisable at stages III and IV. Stages I and II are probably represented by the disseminated accumulations of the brockram sequences, whereas the sequential accumulations correspond to a stage III K horizon.

As opposed to the above morphological sequence, however, the initial stages of carbonate formation within the brockram sequences takes place within the interstitial matrix clay of the framework clastic sediment and does not originate as grain coatings. This indicates that the interstices were at least partially blocked (probably by mechanically infiltrated clay) prior to the formation of the carbonate.

Petrographic considerations

Whereas the sequential morphology of caliche and cornstone profiles has been comprehensively documented, the detailed petrography of these horizons is, in general, less completely described with most authors referring exclusively to higher zones of the profiles, or to the profiles as a whole. A similar statement has been made by Hay and Reeder (1978, p.649). In consequence, few authors discuss the diagenetic processes of caliche formation, the majority assuming that the carbonate is replacive (e.g. Burgess, 1961; Allen, 1965), displacive (e.g. Blake, 1902; Gile *et al.*, 1966; Gile, 1967; Goudie, 1972, 1973) or a combination of both (e.g. Allen, 1973). However, many of the textural and mineralogical features described in the preceding section have been recorded from cornstone (e.g. Burgess, 1961) and caliche (e.g. Nagtegaal, 1969) profiles and indicate a close petrographic (and, therefore, formational) relationship with the brockram sequences. The most important features may be summarised as:

1. Floating quartz grains within the zone of carbonate development (Brown, 1956; Burgess, 1961; Gile et al., 1966; Gile, 1967; Nagtegaal, 1969; Francis et al., 1970; Goudie, 1972, 1973).
2. Dissolution and replacement of detrital quartz grains by carbonate (Bretz and Horberg, 1949; Burgess, 1961; Pick, 1964; Allen, 1965, 1973; Blank and Tynes, 1965; Bruck et al., 1967; Nagtegaal, 1969; Francis et al., 1970; Reeves, 1970; Williams, 1973; Steel, 1974a; Amiel, 1975; Higham, 1977; Watts, 1980). Complete replacement by sparry calcite has been recorded by Francis et al. (1970).
3. Reciprocal silica/carbonate ratios (Aristarain, 1970; Reeves, 1970; Goudie, 1972, 1973) and fixed clay/carbonate ratios (Gile et al., 1966; Gile, 1967).

Although the replacement of quartz by carbonate is widely recognised, several authors suggest that displacement is the most important formational process (e.g. Gile et al., 1966, Gile, 1967; Goudie, 1972, 1973). In particular, the effects of displacement are suggested, by these authors, to account for the separation of detrital grains in higher zones of the profiles. Gile et al. (1966) assume that carbonate is formed exclusively by displacement and suggest that the process of carbonate accretion about framework grains eventually causes separation of the grains and thus creates intergrain regions of nearly pure carbonate. Similarly, Goudie (1972, p.455) states:

> 'Clearly, the growth of calcium carbonate not only fills voids but also disperses sand grains so that they 'float' and do not occur in point-to-point contact as they would do in the parent sediment.'

The occurrence of the occasional detrital grain has led these authors to include, in their hypotheses, a process which is termed 'engulfment' by Gile et al. (1966). Their hypotheses are, therefore, that accreting carbonate displaces most of the original sediment (and thus separates the framework grains) but incorporates a variable percentage of original grains and matrix. In both cases, the relative proportion of matrix in the carbonate-rich horizons is recorded to be increased with respect to detrital grains[1], a feature which must either result from the displacement of grains in preference to matrix or, as suggested alternatively by Goudie (1972), from the downward eluviation of fine material.

Other authors, however, e.g. Pick (1964) and Allen (1965) have recognised the importance of replacement and Francis et al. (1970, p.103) suggest the separation of quartz grains to be the result of both displacement and replacement.

Petrographic studies of the brockram sequences indicate that replacement is the dominant means by which these profiles are formed and that it is possible to explain all the textural and mineralogical features (including the relative increase in the matrix/detrital grain ratio) without necessary reference to displacement, engulfment or eluviation. This hypothesis is supported by the recent studies of calcretes from northern Tanzania by Hay and Reeder (1978). That some degree of displacement does occur has already been ascertained and, although here suggested to be subordinate to replacement, is in accordance with the typically displacive and replacive growth modes of Quaternary calcretes (Allen, 1973).

The extent of quartz replacement in the brockram sequences is indicated by the reciprocal detrital quartz/carbonate ratios (see Figure 4.3c). Although it is not possible for chemical and modal analyses to be compared

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1 Goudie (1972, p.456) records that the silt and clay component of caliche from Libya and Cyprus increases with increasing $CaCO_3$ content. Gile et al. (1966, p.357) record the silicate clay content of upper zones of the K horizon to be approximately the same as that of the overlying calcareous B horizon of soils occurring in southern New Mexico. Gile (1967, p.270) further notes that the concretions in calcareous B horizons have clay contents equal to that of the internodular material.

directly, chemical analyses of caliche profiles (Aristarain, 1970; Goudie, 1973) indicate a similar silica/carbonate relationship. Aristarain (1970) and Reeves (1970), however, record a marked increase in silica values in the uppermost ('layered caliche') zones of well developed profiles, these values being lowest immediately subjacent to this zone. Reeves (1970) suggests that the reduction of the quartz component represents a transition from 'young' to 'mature' caliche and that in 'old' caliche, silica values are again high due to the formation of authigenic silica (terminology of Reeves, 1970). The uppermost 'layered' caliche is considered to be equivalent to the laminated zones of 'old' caliche profiles (Reeves, 1970) and of the laminated K2 horizon (Gile et al., 1965) and, as such, is not represented in the brockram sequences. Unlike the brockram examples, however, Reeves (1970) suggests that alteration affects only a small percentage of the detrital quartz grains and that dissolution is only 'slight'. Reeves (1970) thus implies that the reduction in the silica/carbonate ratio is a result of displacement. As previously ascertained, and in accordance with the work of Gile et al. (1966) and Gile (1967), the clay content retains approximately the value throughout the profiles (see Figure 4.3b).

Reeves (1970) notes that variable silica/calcite ratios are determined mainly by changes in temperature, pressure, partial pressure of carbon dioxide and pH and that the stability and depositional kinetics of silica and calcite are reciprocal. Although it has been suggested (Arkley, in Reeves, 1970) that the complex nature of soil may induce a far more involved relationship between the kinetics of silica and calcite, it is clear that the alkaline nature of the carbonate-enriched solutions would favour the dissolution of quartz. The clay, however, would be relatively more stable and would not be removed, in solution, from the profile. It should be noted that mineralogical changes in the clay fraction of caliche profiles have been recorded by Aristarain (1970) and Hay and Reeder (1978), who also describe clay replacement by micrite. The exact effect of this process, in terms of the overall percentage composition of the clay, is not determined, however.

It is important to note that the reciprocal silica/carbonate ratio could be produced by displacement and is inconclusive, therefore, of a replacive mode of formation. However, displacive formation can be disregarded on the grounds that it would also give rise to a corresponding reciprocal relationship between the clay and carbonate components, which is not the case.

Followed through to its logical conclusion, the process of replacement can also be used to explain the laminar horizons of mature profiles. Gile et al. (1966) explain their formation in terms of displacement and assume that the soil is thickened by an amount equal to that of the laminated horizon. If, however, the hypothesis of replacement is accepted, an increase in soil depth is not necessarily required for the formation of laminated carbonate and it is also possible to explain simply the frequent association of these laminae with amorphous silica. As evaporation of the solutions ponded above the plugged horizon proceeds, $CaCO_3$ is precipitated and SiO_2 is taken into the solution. If precipitation of $CaCO_3$ continues, and the solutions do not escape, the pH of the solutions would be effectively reduced until SiO_2 could no longer be held in solution, at which stage it also would be precipitated. Continued evaporation would, in any case, lead to the precipitation of SiO_2. Complete evaporation would, therefore, result in an underlying lamination of $CaCO_3$ and an overlying lamination of SiO_2 (viz. the occasional alternation of carbonate and 'chert' bands described by Steel, 1974a). Precipitation of both carbonate and silica would require an increase in soil depth unless the volumetric change could be accommodated within the sediment.

In describing occurrences of amorphous silica, Steel (op. cit.) notes that it is generally assumed to be secondary, after carbonate. Although the formation of banded silica can be explained in terms of the model presented above, the nodular and lensiform occurrences described by Sidwell (1943) can, perhaps, best be explained by reference to replacement. A replacive origin has also been proposed by Goudie (1972). It is probable, therefore, that whilst some forms of silica represent replacement of carbonate, other occurrences may be formed directly from a second phase of precipitation from percolating or ponded waters, proceeding an initial phase of carbonate precipitation.

Whilst the formation of authigenic silica may be in some doubt, it is generally agreed that it is derived from silica mobilised during replacement of detrital quartz grains (Steel, 1974a; Watts, 1980). As it has already been ascertained in previous descriptions that replacement (and, consequently, silica mobilisation) is an important feature of the brockram sequences, the absence of any form of silica from the massive carbonate zones requires some explanation. This apparent anomaly is partially solved by Reeves (1970), who states that chert bands are formed only in mature caliche profiles where laminated and pisolitic carbonate overlies (and formed as a result of) an impervious plugged horizon.

Since laminated textures are not recorded from the brockram sequences, it can be postulated that the uppermost zones of the profiles formed only partial barriers to downwards percolating solutions and did not form true 'plugged horizons'. Following precipitation of $CaCO_3$, therefore, the silica-rich solutions were able to migrate to lower zones of the profiles where continued evaporation allowed the precipitation of SiO_2 in the form of overgrowths around detrital quartz grains. In this manner, a silicified zone may be formed below the carbonate-cemented zone which, owing to its semi-permeable nature, could not be expected to favour the formation of overlying laminated carbonate.

Under ideal conditions, development of the diagenetic quartz zone would be restricted to the lower part of the diffuse carbonate zone (where $CaCO_3$

and SiO_2 precipitate together) and to the sediment underlying the lowermost carbonate zone (where all the $CaCO_3$ has been precipitated). The occurrence of authigenic silica in higher parts of the profile (e.g. the nodular zone) and of corroded quartz overgrowths clearly suggests, therefore, variable formational conditions and may be attributed to several factors:

1. Variable rainfall intensity. Assuming potential evaporation (i.e. the rate of evaporation) to remain constant, the effect of a higher rainfall intensity would be to allow percolating waters to attain greater depths in the soil profile. The zone of carbonate precipitation and quartz dissolution may occupy, therefore, a level where silica was precipitated formerly. In this instance the quartz overgrowths would be corroded. The converse would cause silica to precipitate in higher zones. The effects of variable rainfall duration may be similar.
2. Variable potential evaporation. Lower evaporation rates would enable percolating solutions to reach greater depths in the soil profile.
3. Surface degradation or aggradation. A lowering of the soil surface by erosion would effectively lower the zones of carbonate and silica precipitation.

Profiles described by Reeves (1970) and Williams (1973) exhibit a similar association of authigenic silica and carbonate zones in which a diagenetic quartz zone overlies the lowermost 'transitional' carbonate zone. In both examples laminated textures occur at the top of the profile. The formation of the diagenetic quartz presumably precedes the plugging of the profiles.

FACTORS CONTROLLING THE FORMATION AND DISTRIBUTION OF CALICHE

The two most important factors controlling the formation of caliche are climate and time. Goudie (1972, p.449) notes that Tertiary/Quaternary caliches are largely restricted to semi-arid areas. In the case of arid areas, carbonate mobilisation is prevented by the lack of moisture, whereas too much moisture completely leaches the solubles from the soil profile. Dolliver (in Steel, 1974a) states that caliche can form in areas with less than ten centimetres annual rainfall and Blatt et al. (1972, p.258) state that the upper limit of precipitation is generally less than 60 centimetres/ annum. Reeves (1970) suggests that caliche may form in areas with higher rainfalls than this maximum value, providing either temperature is high or infiltration and carbonate supply is high. As a generalisation, however, it can be assumed that caliche profiles imply a semi-arid climate in which surface temperature is sufficiently high to promote sub-surface evaporation.

The formation of caliche profiles also requires surface stability with no sedimentation over long periods of time (Allen, 1965; Gile et al., 1966; Steel, 1974), the amount of accumulated carbonate increasing markedly with the age of the soil (Gile et al., 1965). The time required for the formation of these profiles varies from a suggested timescale of tens of years (Steel, 1974a) to thousands of years (Allen, 1965; Gile et al., 1966), depending on the thickness and maturity of the profiles and the prevailing climatic surroundings.

In addition to these two factors, the availability of carbonate is an important control of the formation of caliche. Goudie (1973) also notes that caliche tends to form more readily on low angle surfaces of less than one degree.

In discussing the distribution of cornstones, Allen (1965, 1973) and Steel (1974a) record that they are commonly associated with floodplain silt-stones of fluviatile fining-upward cyclothems. Steel (op. cit.) notes that this occurrence is far more common than in alluvial fan sequences and concludes that:

> ' ... the critical local relationships in a semi-arid area, between precipitation, relief, run-off and temperature, such as are required for caliche formation, are more likely to be met over floodplains than over areas of piedmont.' (p.365)

In this respect, the occurrence of the carbonate-rich horizons in the finer members of the brockram sequences, rather than in breccia horizons, concurs with that of the New Red Sandstone cornstone sequences of western Scotland described by Steel (1974a) and with those of sequences described by Allen (1965), Bruck et al. (1967) and Francis et al. (1970). The processes which deposited the breccias of the fans are considered unlikely to have facilitated the formation or preservation of caliche, such caliche as may have been developed being eroded by subsequent stream activity.

CONCLUSIONS

From a comparison of the carbonate accumulations of western Cumbria with cornstone sequences and present day soil profiles, it can be concluded that these accumulations are the result of an identical pedogenic process. Several features, however, indicate the brockram profiles to be incompletely developed:

1. The absence of laminated, brecciated or pisolitic carbonate from the top of the profiles.
2. The occurrence of secondary quartz in lower parts of the profiles, indicating that the massive carbonate zones did not form true 'plugged' horizons.
3. The $SiO_2/CaCO_3$ ratios of the brockram profiles show a marked decrease from the lower to higher zones whereas mature profiles have high SiO_2 values in the upper zones due to the formation of authigenic silica (Aristarain, 1970; Reeves, 1970).

The comparative immaturity of the profiles can be related to a variety of factors which include time, $CaCO_3$ availability and climatic conditions.

The fact that a number of carbonate sequences may exist within any one sandstone horizon and that these profiles are commonly overlain by, and may be eroded by, breccias indicates that the development of the profiles was partly controlled by time and sedimentation. The presence of the profiles indicates that sedimentation was ephemeral so that their relative development is a function of sedimentation frequency. Thus, the longer the time interval between sedimentation episodes, the greater the tendency for the complete profile to be developed. The development of a massive carbonate zone at the top of several of the profiles indicates (relatively) prolonged atmospheric exposure.

In this manner, it is possible to interpret unusual thicknesses of immature sequences (non-sequential accumulations) as a result of intermittent sedimentation or a slowly, but continually aggrading soil profile. Similarly, sandstone units containing more than one carbonate sequence can be interpreted as polygenetic (Gile and Hawley, 1966). Where intermittent sedimentation of this kind is assumed, doubts must exist in the interpretation of K1 horizons.

In summary, therefore, the general immaturity of the sequences may be attributed to periods of exposure which are less in duration than that required for a mature profile to be developed and which are controlled by the rate and frequency of sedimentation. However, the formational time for a particular sequence may be extremely variable and is dependent on several factors, including $CaCO_3$ availability and climate. These factors may either retard or promote the development of the profile.

Where $CaCO_3$ is not readily available, the formational time for the profiles may be greatly increased. $CaCO_3$ is generally believed to be derived from loess and from solution of pre-existing carbonate solubles from the soil profiles, including intrastratal solution of iron silicates and feldspars (see Chapter Two). Although a certain amount of the calcium of the brockram sequences is probably derived from the dissolution of feldspars (which are always extensively replaced) and other rock fragments, the high percentage of detrital quartz in these sediments suggests that this source could not account for the amounts of carbonate now present. Loess is suggested, therefore, to be the major source of the carbonate. However, loess accumulates mainly on the downwind sides of major deserts (Reeves, 1970) and probably, therefore, accumulated only relatively slowly in the distal fan environment of the brockram sequences (assuming the prevalent palaeowinds to be easterly, or approximately so; Shotton, 1956; Opdyke, 1958; Waugh, 1967).

The availability of water is a critical factor controlling the rate of caliche development and generally confines these profiles to semi-arid areas. The depth of calichification and the development of caliche diminishes with rainfall to a near-surface rubble at the dry limit (Blatt et al., 1972; Mabbutt, 1977). Where sufficient water is available to mobilise carbonate, but where rainfall events are extremely infrequent (as may occur in arid areas) it is possible to develop mature profiles although the formational time will, again, be extended. Two factors suggest that this may have been the case during the formation of the brockram sequences:

1. The sequences are associated with (and are in part the stratigraphical equivalent of) the St. Bees Evaporites, which formed in a coastal sabkha environment bordering the Bakevellia Sea. Modern analogues, such as the Persian Gulf, are generally arid rather than semi-arid.
2. Many caliche profiles contain vertically elongate carbonaceous and calcite filled tubes interpreted as former root zones (Amiel, 1975; Higham, 1977; see, also, Klappa, 1980). The complete absence of these features from the brockram sequences may imply, therefore, a climate which was too arid to support vegetation. Erosion of the upper zones of the profiles could, however, provide an alternative explanation.

It is possible, therefore, that the formational time may have been appreciably extended, in the western Cumbrian examples, by both the limited availability of $CaCO_3$ and the infrequent availability of water.

The absolute time represented by the soil profiles varies according to their thickness and maturity. Steel (1974a) suggests that although laminated carbonate from the top of mature profiles takes thousands of years to develop, it is possible for caliche to form over much shorter periods of time (tens of years), providing formational conditions are favourable. However, studies by Gile et al. (1966) suggest that even sequences which do not include laminated carbonate may take considerable lengths of time in which to develop. These authors conducted a study of modern carbonate sequences and, in order to obtain some idea of the formational time involved, the morphological surfaces on which they occurred. These studies indicated that although taking longer to form a plugged horizon in non-gravelly soils than in gravelly soils, the time involved was in excess of 5.10^3 years. Laminated surfaces occurred on Late-Mid Pleistocene surfaces. Radiocarbon dates obtained from laminated profiles by Ruhe (1967) and Williams and Polach (1971) give a formational time of between 5.10^3 and 10^4 years. Ruhe (1967), however, notes that the validity of such radiocarbon dates may be limited.

The timescale involved for the formation of the western Cumbrian examples is, thus, debatable. As there is no evidence on which a greater than normal abundance of $CaCO_3$ could be assumed, a comparison with the examples described by Gile et al. (1966) is likely to be as valid as any other. On this basis, it may be postulated that the periods of surface stability, represented by the development of massive carbonate zones, are in the order of 5.10^3 years maximum duration.

One of the concluding statements made by Steel (1974a, p.368) is that 'cornstone is especially frequent and mature in thin stratigraphic sequences, i.e. it is a 'condensed sequence' indicator'. This relationship can be recognised in western Cumbria, where the total thickness of the brockram sequences containing the carbonate profiles is generally less than 30 metres, compared with a maximum thickness of 134 metres in eastern parts of the depositional basin. Such a reduction in thickness reflects a corresponding proximal to distal fan reduction in fluvial activity (and sedimentation) with the conditions prevailing at the distal extremes facilitating the formation of caliche profiles.

CHAPTER FIVE

THE DUNE STRUCTURES OF THE PENRITH SANDSTONE

INTRODUCTION

The Penrith Sandstone is a dull brick-red coloured, medium to coarse grained sandstone which generally exhibits a high degree of sorting. These deposits are typically poorly cemented but may be secondarily silicified, especially in the upper part of the formation between Armathwaite and Cliburn (see Figure 1.1). The predominant sedimentary structure is that of cross-stratification. Cross-bedded units are typically bounded at their upper and lower margins by erosion planes which, although marked, are usually flat or gently sloping. Occasionally, particular sets may be seen to have a 'trough' shaped erosional base, one good example of which may be seen exposed in a railway cutting 800 metres north of Lazonby at NY544406. This exposure will be described in detail in a later section. Due to the erosive nature of these units it is rare that the complete thickness of the sets is represented. However, it is clear that sets of cross-bedding may reach a considerable thickness; thicknesses exceeding five metres have been recorded, although typically only one or two metres of any individual set are preserved.

Where secondary silicification occurs, the rock is hard and coherent and splits readily along planes conforming with the foresets of the cross-stratification. The silicified Penrith Sandstone thus forms an ideal building stone which has been extensively quarried in the past and has provided the material from which many of the surrounding villages such as Great Salkeld, Kirkoswald and Lazonby have been built. At present the sandstones are used mainly as an ornamental stone or for facing so that only a few quarries are now worked. However, because of the vast amounts of stone removed in the past and the nature of the industry which quarried the stone, the area around Penrith has a large number of small disused workings. Occasionally, larger disused quarries are present, e.g. Maidenhill Quarry (NY525332) and Cowraik Quarry (NY542310) but these, as in the case of many of the smaller quarries, have become very overgrown, or, where areas have been reafforested, e.g. Slatequarry Wood (NY550320) and Penrith Beacon (NY525315) completely overgrown In addition, many of the localities have been either partially or completely filled e.g. Cowraik Quarry, Maidenhill Quarry, Blaze Fell quarries (centred on NY498435) and Lazonby Fell quarries (centred on NY390520). Thus, although a cursory examination suggests many quarry exposures in the Penrith Sandstone, the majority of these are either inaccessible or so poorly exposed as to be of little use for sedimentological purposes.

The unsilicified Penrith Sandstone, which forms the lower part of the formation, mainly between Cliburn and Brough, is generally poorly exposed due to its weakly lithified nature and to its lack of economic value. However, natural stream sections are comparatively common and allow for the recognition of structures which are identical to those of the silicified sandstone.

The purpose of this aspect of research has been to examine the sedimentology of these sandstones in the light of detailed descriptions of dunes from modern (McKee and Tibbitts, 1964; McKee, 1966; McKee, 1979a) and ancient (e.g. Thompson, 1969; Brookfield, 1979) aeolian environments with the aim of determining the predominant dune type or types which formed the Penrith Sandstone.

DEPOSITIONAL ENVIRONMENT OF THE PENRITH SANDSTONE

Until recently, the formation of many deposits possessing textural and structural characteristics similar to those of the Penrith Sandstone has been attributed to the aeolian accumulation of sand in an arid or semi-arid environment. Several such rock successions have been advanced as archetypal aeolian sandstones and include:

1. The Pennsylvanian-Jurassic sandstones of the Colorado Plateau of the United States, including the Navajo Sandstone. (Reeside, 1929; Harshbarger et al., 1957; Baars, 1961; Opdyke, 1961; Stokes, 1961; Poole, 1962, 1964).
2. Permo-Triassic sandstones of Great Britain and western Europe (e.g. Shotton, 1937, 1956; Laming, 1966; Smith and Francis, 1967; Thompson, 1969; Glennie, 1970, 1072).
3. The Late Jurassic-Early Cretaceous Botucatu Sandstone of South America (Bigarella and Salamuni, 1961).

However, the aeolian origin of certain deposits has been seriously challenged in recent years, following advances in sedimentary petrology and the detailed description of, and comparison with, previously unconsidered depositional environments. In consequence, several 'aeolian' successions have been reinterpreted. For instance, the aeolian origin of the Colorado Plateau sandstones has been questioned by Jordan (1969), Stanley et al., (1971), Visher (1971) and Freeman and Visher (1975) and the Basal Permian (yellow) sands of north-eastern England by Pryor (1971). In both cases it has been suggested that the deposits are the result of subaqueous dune sedimentation in a shallow shelf sea, following the model erected by Houbolt (1968) from the North Sea.

The criteria of large scale 'aeolian-type' cross-bedding, grain surface frosting, grain roundness and grain sorting, which were once accepted as indicative of an aeolian origin, are no longer thought to be environmentally diagnostic (Bigarella, 1972). Pryor (1971) records the occurrence of large scale cross-bedding from many depositional environments and Pettijohn et al, (1972) suggest that grain roundness cannot be related to a specific mode of formation. It is, however, suggested that grain roundness can be used as a criterion for distinguishing between beach and aeolian dune sands (Shepard and Young, 1961) as can grain size distribution, whereas sorting characteristics reflect the differences between aeolian dune and river sands (Friedman, 1961).

Although grain surface frosting can be produced by a variety of chemical means (Kuenen, 1960), certain surface features such as conchoidal fractures caused by abrasion (percussion marks) can be used as environmental indicators (Krinsley and Doornkamp, 1963; Krinsley and Donahue, 1968; Krinsley and Margolis, 1969; Krinsley *et al.*, 1976).

A detailed analysis of the textural fabric of the Penrith Sandstone was undertaken by Waugh (1967) and has not, therefore, been the purpose of this research project. Waugh (1967, p.81) records that the sandstone 'shows the typical high degree of rounding exhibited by aeolian sands, possessing a pronounced 'millet-seed' character' and that it exhibits a high degree of sorting, although somewhat coarser than comparable sands described by McKee (1934), Bagnold (1935) and Shotton (1937). The average sorting (σ_ϕ) and skewness (Sk_ϕ) values (Folk and Ward, 1957 *in* McBride, 1971) determined for the Penrith Sandstone using the data of Waugh (1967) are 0.50 (corresponding to the well to moderately well sorted classes of Folk, 1968) and 0.14 (symmetrical to fine skewed), respectively. Both values fall within the ranges plotted by Ahlbrandt (1979, p.30) for modern inland dune sands. The grains of the Penrith Sandstone are also recorded (by Waugh, *op. cit.*) to have a frosted surface, in the manner described by Kuenen and Perdok (1962) and to develop larger percussion marks up to eight microns in length which usually bound elongate ridges with rounded crests. This feature has since been described from the Early Triassic Otter Sandstone of Devon by Krinsley *et al.* (1976).

Although the evidence of sandstone fabric, with the possible exception of percussion fractures, is inconclusive in proving an aeolian origin for the Penrith Sandstone, associated lithologies strongly imply a subaerial mode of formation. These include the interdigitation of brockram alluvial fan breccias, the fluvial deposits of the River Belah and Low House sections and the lateral facies associations within the overlying Eden Shales which comprise fan breccias, sheet flood sands, evaporating lake deposits and continental sabkha sands and silts (Burgess and Holliday, 1974, Figure 4, p.9). In addition, wind faceted pebbles with characteristics of dreikanters have been collected by several authors, both from the brockrams and from coarser horizons within the sandstone (Waugh, 1967). The occurrence of rare land vertibrate footprints has also been noted by Nicholson (1868) and Taylor et al. (1971).

In contrast with the Basal Yellow Sands of north-eastern England and the Navajo Sandstone, the only recorded small scale cross-bedding occurs at localities such as Belah Crags and Low House where associated lithologies indicate their formation to be a result of fluvial reworking of the original large scale (aeolian) cross-bedding by ephemeral streams. In particular, there is nothing comparable with the units of current lineated sand inter-bedded with rippled laminae recorded by Freeman and Visher (1975) which, they state, represents a transition from upper to lower flow regime and are, therefore, subaqueous in origin. Other factors discussed by these authors include the angle of repose of the sediment. It is argued that depositional dips of foresets ranging from 20 to 30 degrees, recorded from the Navajo Sandstone, are comparable with values recorded by Hoyt (1967) for subaqueous sand, whereas aeolian dunes, as measured by McKee (1957, 1966) and Bigarella et al. (1969) commonly exhibit a steeper dip ranging from 30 to 34 degrees. The maximum angle of foreset inclination recorded from the Penrith Sandstone is 29 degrees (e.g. Bowscar Quarry and Trough Gill), most values being somewhat less than this and lying in the range of 20 to 28 degrees. Three factors may account for this apparent discrepancy:

1. The formation is tilted at an angle which is generally less than five degrees towards the east-north-east, this direction being opposite to that of the inclination of the cross-stratification (Shotton, 1956; Waugh, 1967). Corrections for regional tilt would indicate, therefore, a

maximum angle of repose of 34 degrees, corresponding to the values proposed by McKee (op. cit.) and others for aeolian deposits

2. For tangentially based cross-strata, the maximum angle of repose is attained in the upper (avalanche) zone of the slipface. The lower, or asymptotic, zone varies considerably in steepness from this maximum to low values in the extreme leeward parts of the dune. In depositional environments, such as that of the Penrith Sandstone, where considerable deflation separates dune-building episodes, the preservation potential is greater for the lower part of the slipface than the upper and, in consequence, exposures of the latter are rare in comparison.

3. The inclined cross-strata may exhibit reduced dips as a result of post-depositional compaction of the sediment.

Although it is clear that both the regional tilt and the preferential preservation of lower zones of the slipface will reduce the dip of the foresets, the effect of compaction is less well documented. Compaction effects for comparable deposits have been discussed by several authors, notably Rittenhouse (1972), Walker and Harms (1972), Freeman and Visher (1975) and Hubert and Mertz (1980). Freeman and Visher (op. cit.) discount these effects (as proposed by Glennie, 1972) on the evidence of studies by Pettijohn et al. (1972) which suggest compaction under load to be small and account for a decrease in porosity of only a few per cent. Walker and Harms (1972), however, assume a seven degree reduction in foreset inclination, based on reduction in porosity of 20 per cent. for the Lyons Sandstone of Colorado.

Porosity values of undisturbed sands are generally in the range of 35 to 50 per cent. with framework grains averaging less than two tangential contacts per grain (Taylor, 1950; Gaither, 1953; Walker and Harms, 1972). Waugh (1967) records the porosity and grain relationships of the Penrith Sandstone to be variable and recognises two main class types. Samples from class 1 are characteristic of lower parts of the formation where the effects of pressure solution cause the number of grain contacts to average 2.87 and porosity to average 16.3 per cent. Class 2 samples represent the silicified upper part of the formation where grain contacts average 1.87 and minus porosity averages 23 per cent. Assuming an initial porosity of 40 per cent., therefore, it is possible that compaction of the Penrith Sandstone has resulted in a reduction of foreset inclination of six to eight degrees, derived from the calculations of Walker and Harms (1972). It is stressed, however, that this figure is based on the assumption that the Penrith Sandstone is directly comparable with the Lyons Sandstone and should merely be considered as an indication of the possible effects of compaction.

In summary, the structure and fabric of the sandstone possess characteristics which, although inconclusive in isolation, are more or less indicative of a subaerial formation which is confirmed by considerations of associated deposits, as prescribed by Bigarella (1972). In addition, dunes with a similar structure and scale as those of the Penrith Sandstone (see Interpretation, this chapter) are, at present, unknown from the subaqueous environment. In conclusion, therefore, an aeolian origin may be postulated for the dunes of the Penrith Sandstone.

DESCRIPTION OF THE EXPOSURES

As a result of the large number of exposures, both man made and natural, and the relatively poor condition of many of these, it has been necessary to select exposures which are typical of those examined. For these purposes, detailed descriptions of eight main exposures are included in the following section, these having been chosen on the following basis:

1. That they are representative of the whole range of exposures examined.
2. That the degree of exposure allowed the maximum amount of detail to be recorded.

3. That the overall extent of the exposure was, where possible, sufficient to show the structural relationship of adjacent dunes.

Description of the structures

The exposures will be described with reference to the main structures exhibited, the most common of which being sets of cross-bedding which are typically bounded both above and below by erosion surfaces. Bounding and internal erosion surfaces have been recognised; the bounding surfaces being markedly more erosive than the internal and analogous to the second order surfaces of Brookfield (1977). Brookfield (op. cit.) also recognises first and third order surfaces; the former having no obvious equivalent within the Penrith Sandstone, whereas the latter are represented by internal erosion surfaces.

Foresets in the lower parts of erosive dune units[1] are commonly parallel to, or exhibit low inclinations to, their lower bounding planes. Towards the upper parts of a dune unit, foresets typically steepen and may attain an inclination of 29 degrees before truncation by the proceeding erosive unit. The foresets have, therefore, a concave-upward form and are asymptotically based. Bounding erosion surfaces may often be recognised by the disparity of cross-strata disposition between the erosive and eroded sets. The erosion planes are usually horizontal or gently inclined, but sets may occasionally be seen to have a trough-shaped erosional base, examples of which will be described in the appropriate section. Bounding surfaces correspond to those of the tabular planar, wedge planar and trough stratified sets described by McKee and Weir (1953).

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1 A dune unit is herein defined as a structural unit which is formed of a set, or a number of sets, of cross-bedding developed above a bounding erosion surface (which truncates all underlying structures) and terminated, at its upper limit, by the lower bounding plane of the subsequent dune unit. This term is not necessarily synonymous with the internal structure developed by a specific dune type as defined by McKee (1979a).

Internal erosion surfaces, being less markedly erosive, rarely cause great disparity of foreset inclination and commonly form internal 'discontinuities' within a particular set of cross-strata. In this instance, it will be suggested that those surfaces mark a break in sedimentation or a change in depositional conditions. Following such a break in sedimentation or a change in conditions (such as a higher or lower wind velocity; see Discussion, this chapter), foresets may be formed with a slightly different inclination and/or orientation. These discontinuities may, therefore, represent reactivation surfaces (analogous to those described by Collinson, 1970 from the subaqueous dunes of the Tana River, Norway), examples of which have been recorded from many localities. Internal erosion surfaces, when traced laterally, may become more or less erosive and, in some circumstances may pass into bounding erosion surfaces.

Small scale structures such as wind ripples have not been recorded from the Penrith Sandstone although a grain lineation analogous to primary current lineation and possible small slipface scour structures occur on foresets exposed in Stoneraise Quarry (NY533352) and will be described in the following section. The occurrence of a lag gravel has been recorded by Waugh (1967) from the base of a cross-stratified unit exposed in George Gill (NY718189).

Accurate drawings of all the main exposures are included. The drawings were compiled from sketches and detailed measurements made in the field,and, with the exception of Trough Gill (NY587240), from photographs (Plates 5.1 to 5.5 and Plate 5.7).

The Bowscar area

The Bowscar area, situated five kilometres north of Penrith between the Great Salkeld Road and the main A6 Trunk Road, comprises a group of quarries which, although generally small, indicate extensive workings in the past. The Penrith Sandstone of the area is well cemented by secondary silica and thus provides an ideal building stone which is still worked in three localities. Of these three, Bowscar Working Quarry[1] (NY520344) and Stoneraise Quarry will be described in detail. The third working quarry exhibits few of the sedimentological characteristics to be described and has, therefore, been omitted. However, Smithy Wood Quarry (NY529348), the first exposure to be described, is situated approximately 150 metres north of this locality.

DESCRIPTION OF REPRESENTATIVE EXPOSURES

SMITHY WOOD QUARRY

At this locality, medium to coarse grained Penrith Sandstone is exposed in a small disused quarry. Quarry fill has greatly reduced access to the exposure and only one face with a north-west to south-east orientation is clearly seen.

From Figure 5.1 it may be seen that this face comprises two main dune units (termed dune 1 and dune 2 in this diagram) of which approximately four metres of the lower unit is exposed and two metres of the upper. These two units are separated by a gently inclined, but marked, erosion surface which has been shown, by detailed examination, to be composed of a number of separate planes. Both units are laterally exposed for 15 metres. The lower cross-bedded units consists of structurally concordant foresets inclined towards the north-west with no apparent internal discontinuities. The foresets are gently dipping towards the base of the exposed dune, at an angle of

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[1] This is now a misnomer as working ceased at this locality in 1978.

Figure 5.1. Smithy Wood Quarry, NY529348. Drawing compiled from photographs (Plate 5.1A) and field sketches. The rock surface only is shown. Bold structural lines indicate bounding erosion surfaces, intermediate lines indicate internal erosion surfaces and fine lines, foreset lamination. The figures refer to the disposition of the cross-strata, the first measurement recording the dip of the foresets in degrees and the second, the direction (bearing) in which they dip.


Northwest
Southeast
Dune 2
10/5
13/353
Dune 1
8/323
12/303
15/312
0
1
2
3
4
5
Metres

Figure 5.1

between eight and ten degrees, and steepen to 15 degrees towards the upper bounding surface.

The upper unit is composed of similar foreset laminae inclined at an angle of five to ten degrees towards the north, but includes at least five internal erosion surfaces. These surfaces may be seen to be horizontal or gently inclined towards the lower bounding surface, whereas towards the top of the exposed dune they steepen and become less obvious. These erosive surfaces are similarly inclined to the foresets, i.e. at five to ten degrees towards the north. It is also apparent that the bounding surface between the upper and lower dune units is formed by the flattening and coalescing of these lower order surfaces.

From the structure of the upper dune unit, it is postulated that the lower bounding surface was formed as a result of several successive phases of sedimentation, which are illustrated diagrammatically in Figure 5.2. The upper dune appears to have advanced up and eroded into the windward slope of the underlying dune (Figure 5.2A). Each phase of sedimentation is separated by a non-depositional phase during which further erosion of the lower dune took place in a downwind direction and, in some instances, erosion of the low angle foresets of the preceding depositional event (Figure 5.2B). Surface deflation of the area immediately downwind of dune slipfaces has been observed by Hoyt (1966) from the Namib Desert of what was formerly South West Africa and is suggested to be the result of reversed lee eddies generated by flow separation at the brink of the slipface. From Figure 5.2C it is

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Figure 5.2. Idealised diagram of the stages in advance of a dune over the partially eroded windward slope of a pre-existing dune. The gradient of the stoss slope of the upper dune is greatly exaggerated. 5.2A: upper dune with erosive base. 5.2B: second stage of advance of upper dune showing erosion of the asymptotic zone of the former slipface and the successive erosion of the lower dune. The initial position of the dune is shown with no increase of height assumed for the second stage. 5.2C: second and third stages showing the internal structure produced by erosion of the low angle foresets of preceding sedimentation stages and successive erosion of the lower unit. Bold structural lines indicate erosion surfaces and fine lines, foresets. Former dune crests are shown.

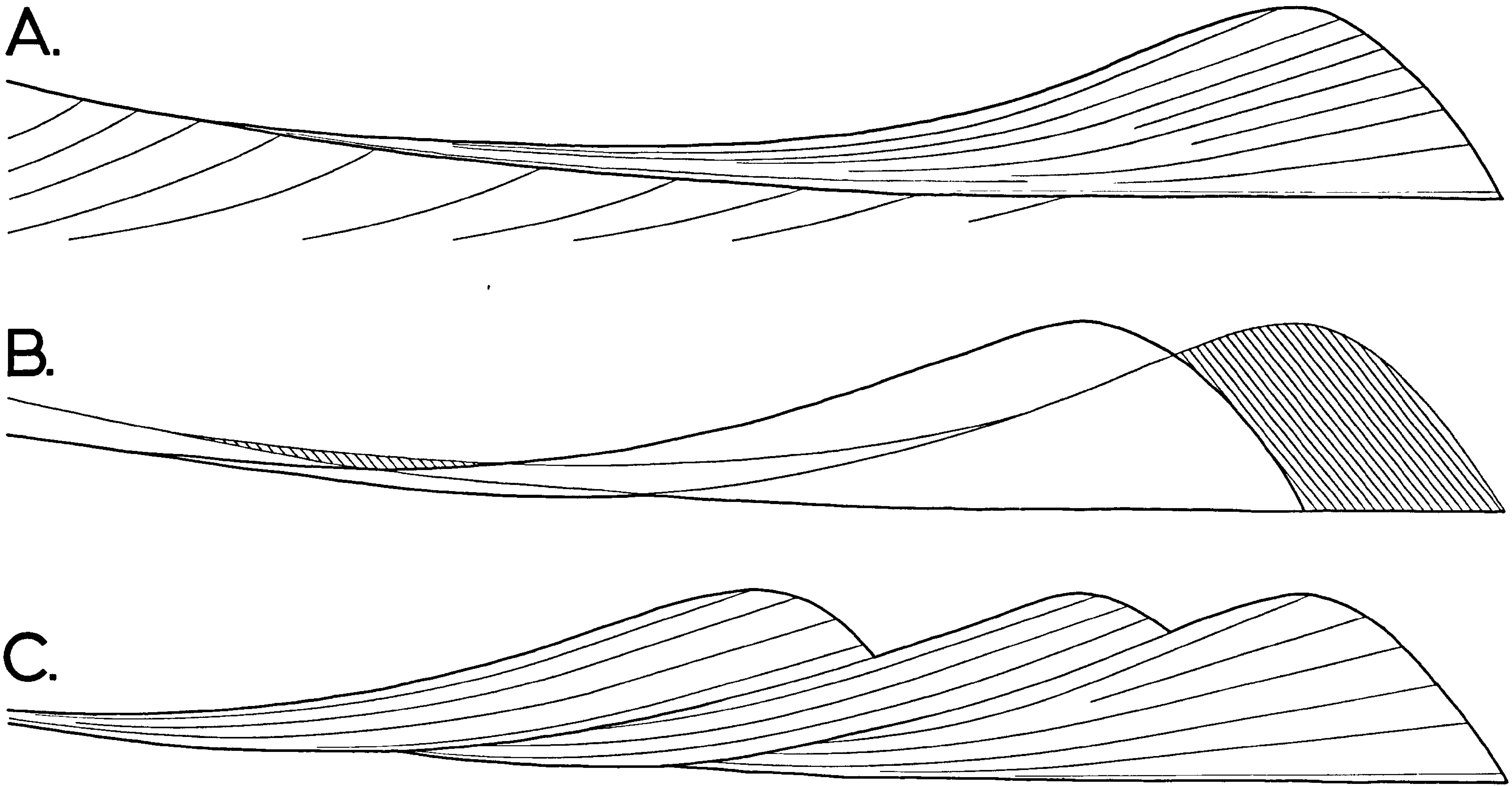

Figure 5.2

apparent that the foresets of the higher part of the upper dune, formed by successive phases of sedimentation, show little or no disparity of inclination. However, towards the lower parts, where the low angle foresets of the preceding event have been truncated, a progressively more pronounced foreset disparity may be observed. This disparity thus forms a readily recognisable erosion plane which may be interpreted as a reactivation surface.

BARONWOOD QUARRY

At this locality medium to coarse grained Penrith Sandstone is exposed in a small disused quarry approximately 3.4 kilometres south-south-east of Armathwaite at NY515428. Long disuse and quarry fill has greatly reduced the size of the exposure and only one face with an east-north-east to west-south-west orientation is clearly seen.

From Figure 5.3 it may be seen that this face comprises four main dune units, each separated by gently inclined bounding erosion surfaces which truncate cross-strata of the underlying unit. The two lower units, dunes 1 and 2 are laterally exposed for 16 metres and the upper units, dunes 3 and 4 for approximately nine metres. Dune 1 has an exposed thickness of 1.8 metres and consists of structurally concordant cross-strata which are gently inclined towards the north-west. The foresets of this unit steepen towards the upper bounding surface, in the extreme north-east of the exposure, to attain a maximum inclination of nine degrees before being truncated by cross-strata of the overlying dune. Dune 2 is bounded both above and below by erosion surfaces and develops a maximum exposed thickness of 2.2 metres. Cross-strata of this unit are concordantly inclined towards the west, the angle of inclination steepening from ten degrees at the base to 13 degrees towards the upper bounding surface. Dune 3 is bounded both above and below by erosion

Figure 5.3. Baronwood Quarry, NY515428. (see Plate 5.1B).


Southwest
Northeast
Dune 3
8/188
8/179
13/284
Dune 4
Dune 2
Dune 1
9/332
6/334
10/282
9/333
0
2
4
6
Metres

Figure 5.3

surfaces, the upper bounding surface truncating the lower in the west-south-west part of the exposure, so that dune 4 overlies dune 2. Foresets of dune 3 are inclined towards the north-west and steepen towards the upper bounding surface. Foresets of dune 4 form parallel laminations and are inclined, as is the lower bounding surface, towards the south.

Although the term 'structurally concordant' has been applied to foresets of other dunes, the parallel laminations of dune 4 are clearly unlike those previously described in that they do not steepen above the lower bounding surface. However, as only 0.6 metre of this dune is exposed, this is not considered to be significant.

Although the lower bounding surface of dune 4 truncates the lower surface of dune 3 and the structure produced superficially resembles that described for the upper dune of Smithy Wood Quarry, three important differences may be observed. First, the foreset inclinations of dunes 3 and 4 indicate opposing palaeowind orientations; those for dune 3 suggesting an approximately south-easterly source (although accurate measurement of this unit is precluded by the nature of the exposure) and those for dune 4 a northerly source. Secondly, the angular disparity of the eroded surfaces is correspondingly greater in this locality and finally, the foresets of the two dune units show an increasing disparity of dip as the thickness of dune 3 increases, the erosion surface separating the two becoming correspondingly more marked. It is clear, therefore, that dunes 3 and 4 may be regarded as separate units and do not represent two advancement stages for the same dune. The surface separating the two is thus a true erosion plane and not a reactivation surface.

The data plotted on Figure 5.4 shows the direction of maximum foreset inclination recorded from all localities examined in the area. The majority

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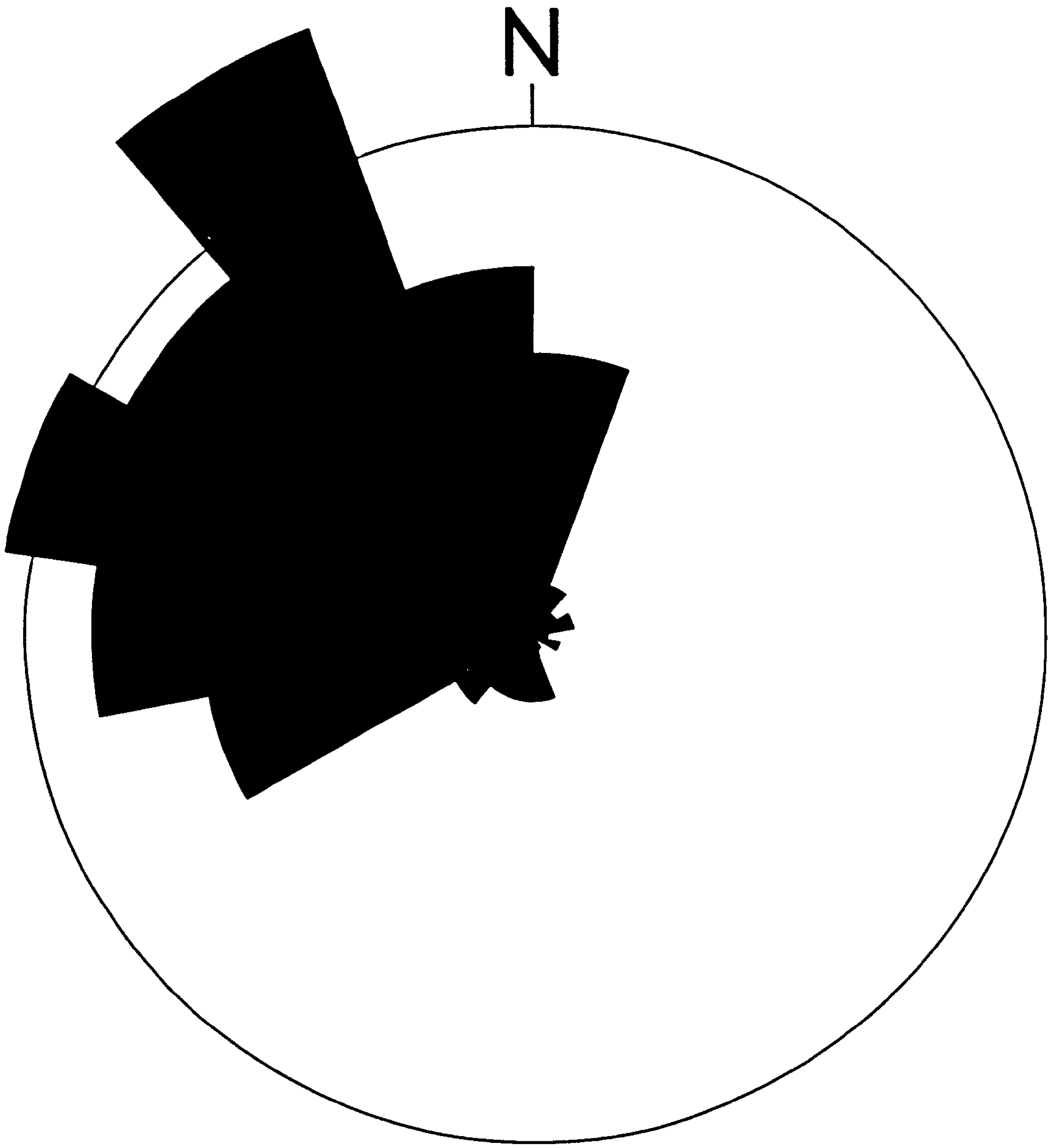
Figure 5.4. Rose diagram showing the direction of inclination of the cross-strata for 277 measurements from all localities examined. The readings are grouped into 18 classes of 20 degrees, the radius of the outer circle of the diagram representing 80 measurements. The values for the mean vector (M_V) and length of the mean vector (R) have been calculated according to the method prescribed by Mardia (1972).



n = 277

Radius of diagram: 80 measurements

Mean Vector (M_v): 281 degrees

Length of Mean Vector (R): 0.5587

Figure 5.4

of the measurements lie in a zone between west 30 degrees south and north 20 degrees east, thus indicating a palaeowind which apparently varied between southerly and easterly. Of the 277 readings taken, only 35 (12.7 per cent.) lie outside of the main zone; their distribution giving no indication of lesser developed, secondary palaeowinds. The mean vector of 281 degrees is close to the value of 284 degrees obtained by Opdyke (1958). The structure of dune 4, with the orientation of the foresets suggesting a northerly palaeowind does not, therefore, correspond with the majority of the data plotted on Figure 5.4, and probably does not represent sedimentation of the main dune building phase. It is suggested that these foresets were formed by the deposition of sediment on the partially eroded windward slope of a pre-existing dune.

WHELPDALE HILL QUARRY

Whelpdale Hill Quarry (NY517338) is a disused quarry situated on the western margin of the Bowscar area, 300 metres south-east of Foresthill. One face, with a north to south orientation shows a number of well defined erosion surfaces separating dune units of highly silicified Penrith Sandstone. The exposure forms an unbroken vertical face of 15 metres length which, although clearly showing a variety of dune structures, does not facilitate detailed measurements of the foresets. The exact relationship of the dunes cannot, therefore, be accurately determined.

From Figure 5.5, it may be seen that this face comprises at least five main dune units, each of which are bounded by erosion surfaces and which contain a number of (apparently) lower order surfaces. It is not clear that these surfaces may all be designated as internal erosion surfaces as any variation in the direction of foreset inclination between adjacent cross-

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Figure 5.5. Whelpdale Hill Quarry, NY517338. (see Plate 5.2).


North
South
Dune 5
6/60
9/16
10/26
Dune 4
Dune 2
Dune 3
26/328
Dune 1
0
1
2
3
4
5
Metres

Figure 5.5

bedded sets cannot be accurately measured. In particular, the discontinuity exposed in the upper part of dune 2 appears to separate cross-strata of markedly different aspect. For the purposes of this description, however, the majority of these surfaces will be assumed to represent internal erosion surfaces.

Dune 1 has an exposed thickness of 0.8 metres and consists of foresets inclined generally towards the north-west. The unit contains two erosion surfaces, the lower of which becomes less marked towards the upper bounding surface where it is replaced by a series of lower order surfaces. It is possible, therefore, that this surface is a reactivation plane, the overlying foresets representing a second stage of sedimentation. The upper erosion plane in dune 1 causes a marked disparity of inclination between the overlying and underlying foresets but whether this represents a second dune or merely a third (exposed) phase of sedimentation for dune 1 cannot be determined. Dune 2 is bounded both above and below by erosion surfaces, the lower surface, especially in the southern part of the exposure, consisting of a complex arrangement of curved 'sub-parallel' and variably developed erosion surfaces which are generally concordant (or nearly so) with adjacent foresets. The lack of any angular disparity between the foresets suggests that little or no erosion has taken place and that these surfaces represent breaks in sedimentation, or minor phases of sedimentation preceding the main dune-building stage. Foresets of dune 2 steepen towards the upper bounding surface both southwards and northwards producing a trough-shaped structure. Although no measurements can be made, this structure indicates the palaeowind direction to be normal or oblique to the quarry face, i.e. approximately easterly or westerly. The data plotted on Figure 5.4 suggest that an easterly palaeocurrent is the more likely. The structure of dune 3 is dominated by four internal erosion surfaces of which the lower two coalesce with, and form part of, the lower bounding surface. The upper two erosion surfaces are both trough-shaped, the uppermost being the more markedly curved. The dune is composed, therefore, of five cross-bedded sets which are generally inclined towards the north-north-east and which attain a maximum exposed thickness of one metre. Foresets of the lower three sets are similar in that towards their lower bounding surfaces they exhibit a low inclination corresponding to that of the bounding surface, whereas towards the upper parts of the set they steepen and may attain an angle similar to that of the adjacent set. The low angle foresets at the base of a particular set are truncated by the lower erosion surface of the succeeding set. In these respects, the structure of dune 3 is identical to that of the upper dune unit at Smithy Wood Quarry and may be interpreted in the same manner. The upper two sets of dune 3 both have trough-shaped erosive bases to which the foresets are inclined at a maximum angle of 18 degrees. The curved nature of the lower erosion surfaces suggests that these sets may owe their formation to infilling of blowout scours.

From the nature of the exposure at Whelpdale Hill, it is not clear whether dune 5 overlies (and truncates) dune 4 or whether the converse is true. For convenience it has been assumed, in Figure 5.5, that dune 5 overlies dune 4. Little of either unit is exposed and in neither case is it possible to obtain detailed measurements. Dune 4 is broadly similar to many dunes described previously but differs in that the lower bounding surface comprises two erosion planes separated by eight centimetres of sandstone in which the lamination is apparently concordant with the erosion planes. Overlying foresets exhibit an apparent disparity of inclination of 15 degrees with the lower bounding surface. The complete exposed thickness (1.3 metres) of dune 5 consists of parallel lamination which is apparently concordant with the lower bounding surface.

It is suggested that the thin sheets of sediment present at the base of dune 4 represent a phase of sedimentation preceding that of the main dune building, the sediment being deposited on the eroded surface of dune 3 in an inter-dune area. The formation of the overlying, more steeply inclined foresets may, therefore, be attributed to the main dune-building phase where the dunes have advanced over the sediment deposited in what was previously the inter-dune area. Careful examination of the foresets associated with the lower bounding surface of dune 2 suggests that a similar process may have been responsible for the structure which they exhibit. Although it is possible for the greater thickness of parallel laminae comprising dune 5 to be formed by this process, it is equally possible that the apparent structure is due to a quarry orientation which is normal to the direction of inclination of the foresets. Assuming a mean vector of 281 degrees (i.e. an easterly palaeowind), the latter explanation seems the more plausible.

BOWSCAR QUARRY

Bowscar Quarry is one of the larger disused workings to be found in the area, the main face having a length of 80 metres and a maximum height of 6.5 metres. It is situated in the centre of the Bowscar area, 500 metres south-east of Bowscar Farm (NY520343). Only one face, with an east-south-east to west-north-west orientation remains for detailed examination, quarry-fill having largely obscured all others, of which only isolated remnants remain.

From Figure 5.6, it may be seen that this face comprises two main dune units, the upper separated from the lower by an erosion surface inclined at 13 degrees to the south. Whereas the structure of the lower unit is determined by a single set of cross-strata, the upper dune shows evidence of several prominent internal surfaces of discontinuity. The lower dune unit has an exposed thickness of 3.5 metres and consists of concordantly aligned foresets inclined at a maximum angle of 29 degrees, this approaching the maximum angle of repose for dry, unlithified sand. At the eastern extreme of

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Figure 5.6. Bowscar Quarry, NY520343. (see Plate 5.3A).


East-south-east
West-north-west
23/263
14/237
Dune 2
Dune 1
27/343
27/345
9/198
15/233
27/320
29/356
13/174
29/3
29/0
0
5
10
15
20
Metres

Figure 5.6

the exposure, foresets of this unit are inclined towards the north-west, whereas in westerly parts they are inclined towards the north. Detailed examination indicates that this variation is a continuous gradation and not one caused by a number of associated sets with slightly different orientations. Further, there is no evidence of internal discontinuity, as required by the latter explanation. The upper dune unit has an exposed thickness of 4.5 metres and is formed of a number of cross-bedded sets, the foresets of which are gently inclined towards the west at the base of the unit and which become more steeply inclined in upper parts of the exposure. The sets are separated by nine apparently non-erosive, but prominent, internal surfaces which have a similar orientation to the foresets. Although there is no evidence of truncation, it can be seen from Figure 5.6 that adjacent foresets (of succeeding sets) are not necessarily concordant with one another and may show considerable (five to ten degrees) disparity of inclination.

Although the structure of the upper dune unit shows important differences to that of the upper dune of Smithy Wood Quarry, it is possible that the process by which it was formed is broadly similar. The main difference in the structure of the two dune units is that the 'reactivation' surfaces are less apparently erosive than those of the Smithy Wood example and may merely represent bedding planes. Careful examination of the more planar lower bounding surface of the upper unit indicates that it was formed by a continuous erosive phase and not, as in the case of Smithy Wood, by several phases which eroded the low angle foresets of the preceding depositional event as well as the underlying dune. Unlike Smithy Wood Quarry, therefore, the more prominent surfaces of the upper dune are not intimately associated with the lower bounding surface and probably represent non-depositional phases separating stages of main dune sedimentation. In this respect, the structures of the upper dune unit of Bowscar Quarry also represent reactivation surfaces.

The foreset disparities noted in the upper dune unit are thought to be a result of varying depositional conditions, e.g. wind velocity, sediment size, wetness etc., and the effect of those variations on the morphology of the dune, particularly the steepness of the avalanche slope.

BOWSCAR WORKING QUARRY

At this locality, well silicified Penrith Sandstone is exposed in a small quarry 150 metres north-east of Bowscar Quarry. The sandstone, which has a medium to very coarse grained texture in which the particles frequently exceed one millimetre in diameter, is worked from one face with a north to south orientation. Seven main dune units may be recognised from the exposure which extends laterally for approximately 30 metres and vertically to a height of six metres.

The bounding surface between dunes 1 and 2, although clearly visible in the field, is obscured by rock debris in Plate 5.38 and is omitted, therefore, from Figure 5.7. Both dunes have an exposed thickness of about two metres which is formed of foresets inclined towards the south-west with no apparent internal discontinuities. Occasional disparities of foreset inclination may be observed, these being similar to those described from the upper dune unit of Bowscar Quarry and, therefore, interpreted in the same manner. Dune 3, which develops a maximum thickness of two metres, is bounded both above and below by erosion surfaces, the upper being markedly erosive along the whole of its length and the lower for only part of its exposed length. The lower bounding surface is replaced northwards by two lower order surfaces, the upper of which forms an internal discontinuity within dune 3, marked by a slight disparity of inclination between overlying and underlying foresets. This

Figure 5.7. Bowscar Working Quarry, NY520344 (see Plate 5.38). Hachured lines indicate the continuation of bounding surfaces exposed behind the main face.


South
North
Dune 7
6/162
?
Dune 6
25/305
12/73
7/173
16/83
Dune 5
Dune 4
11/190
Dune 3
Dune 1
15/257
25/230
Dune 2
7/272
24/229
25/240
11/218
27/340
0
5
10
15
20
Metres

Figure 5.7

disparity progressively decreases towards higher parts of the dune until the internal erosion surface can no longer be recognised. As with the upper dune unit of Smithy Wood Quarry, this structure may be interpreted as a reactivation surface. Foresets of the main part of dune 3 steepen towards the upper bounding surface, both southwards and northwards, to produce a trough-shaped structure. Dune 4 has an exposed thickness of 3.4 metres and is comprised of trough-shaped cross-strata exhibiting the same structure as described for dune 3. Dunes 5, 6 and 7 consist of parallel laminae which have similar orientations to their underlying bounding erosion surfaces, i.e. inclined towards the south (dunes 6 and 7) and east (dune 5).

The parallel laminae of dunes 5, 6 and 7 indicate westerly and northerly palaeowinds, which, in comparison with the prevalent southerly to easterly orientations plotted on Figure 5.4, are clearly atypical. It is probable, in this instance, that the orientation of the strata was controlled by those of the erosion surfaces on which they were developed, these being modified windward slopes of pre-existing dunes. The planar nature of these surfaces may be seen from Plate 5.3B. Although these surfaces are prominently exposed for some distance to the west of the exposure, it is not possible to ascertain whether dune 6 truncates dune 5 (as shown in Figure 5.7) or whether the converse is true. For the purposes of this description, it has been assumed that dune 6 truncates dune 5. Dune 7 can be clearly seen to truncate both dunes 5 and 6 in the ground surface behind the exposure and thus forms the uppermost dune unit. Although foreset inclinations of dunes 3 and 4 are variable, the trough-shaped structure suggests the dominant palaeowind to be normal to or oblique to the quarry face, i.e. either easterly or westerly. The data plotted on Figure 5.4 suggest that an easterly palaeowind is the more likely.

STONERAISE QUARRY

This site, located on the north-eastern margin of the Bowscar area (NY533352), differs from those previously described in two important aspects. First, a more extensive exposure is available for study, the main worked face having a width of approximately 50 metres and a minimum height of seven metres. The sandstone has been quarried back from the main south-south-west to north-north-east trending face so that the structure can also be examined in horizontal plan for a distance of about ten metres. Secondly, therefore, and of greater consequence, the nature of the exposure facilitates detailed examination of the three-dimensional disposition of the strata. The three-dimensional form of the exposure is indicated in Plates 5.4 and 5.5. The minimum height, as quoted above, does not represent the true thickness (approximately 15 metres) of the sandstone due to the benched nature of the quarry and the disposition of the strata which dip generally towards the west-north-west.

A cursory examination of Figure 5.8 indicates that this complete thickness is formed of foresets which are occasionally separated by more marked bedding planes. Foresets throughout the total exposed thickness are ordinarily concordant with one another and are uninterrupted by either bounding or internal erosion surfaces. It is clear, therefore, that the total thickness of the sandstone must be considered part of the same dune unit, this unit being of a much greater magnitude than any formerly described.[1] The morphology of this unit, shown in Figure 5.8, is similar to that of units exposed in

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1 The floor of the quarry has since been lowered and the base of this unit exposed. The lower bounding surface has a slight upward concavity and overlies remnants of approximately five older dunes which are seen in plan view. Foresets of the main dune steepen away from the underlying bounding surface until attaining maximum inclination in the upper part of the dune unit. Quarrying has ceased at the base of the main dune as economic working is inhibited by the more complex structure of the underlying sandstone and the quarry floor is gradually becoming buried by spoil.

Figure 5.8. Stoneraise Quarry, NY533352 (see Plates 5.4 and 5.5).


South-south-west
North-north-east
28/345
27/335
27/340
26/325
25/303
25/275
25/285
26/280
18/335
9/320
23/285
12/285
0
5
10
15
20
Metres

Figure 5.8

other localities such as Whelpdale Hill Quarry and Bowscar Quarry in that the foresets apparently steepen both southwards and northwards, producing a trough-shaped section. However, detailed field measurement indicates that the foresets maintain a relatively constant inclination of approximately 25 degrees, although the direction in which they are inclined is variable. Inclination directions of the cross-strata show a continuous variation from bearing 280 degrees, in northern parts of the exposure, through 300 degrees in central areas, to 345 degrees in southern parts, with all values between these extremes represented. It is the effect of this variation which produces the trough-shaped section shown in Figure 5.8. The asymptotically based foresets of this particular dune unit are, therefore, concave-upward both in transverse (perpendicular to the palaeowind direction) and longitudinal (parallel to the palaeowind direction) sections and produce a 'bowl-shaped' three-dimensional form, which is clearly visible from the exposure.

Occasional small variations of foreset inclination may be observed (Plate 5.6A), similar to those described from Bowscar Quarry. This variation reflects minor changes in the steepness of the slipface and is probably the result of fluctuating depositional conditions.

One other small scale feature of note are the striations seen on upper surfaces of foreset laminae (Plate 5.6B and C). Superficially similar structures, termed 'flute sets' or 'rill sets' have been observed from the lee slopes of modern barchan dunes by Whitney (1978), who ascribes their formation to the erosive effects of 'negative wind flow'. The scale of the flute sets is, however, much greater than that of the striations. Although of widespread occurrence at this locality, striations are commonly better developed on the more prominent bedding planes present in the exposure. This may, however, be merely a product of the tendency for the sandstone to split preferentially along these planes, which are, in consequence, better exposed. In all examples of this structure, the length of the striations is orientated parallel to the direction of maximum foreset inclination. The striations are thus convergent to the axis of the dune in a downwind direction. The flute sets of Whitney (1978), however, tend to be parallel to or slightly convergent to the axis in an upwind direction. Furthermore, petrographic examination indicates that the striations could be the result of an alignment and stacking of individual grains and, therefore, a primary depositional structure as opposed to an erosive lineation, as described by Whitney (op. cit.). The tendency for sand grains to assume preferential orientations in dune foresets has also been described by Schwarzacher (1951), Land (1964) and Waugh (1967) who record the grains to be orientated with their long axes parallel to the wind direction and (Schwarzacher, 1951; Waugh, 1967) with an upslope imbrication. It is suggested, therefore, that the striations are the surface expression of the aligned sandstone fabric and that this structure is analogous to primary current lineation for fluvially deposited sediment. It is proposed that the term avalanche slope lineation be adopted for this structure.

Possible scour structures (c.f. Whitney, op. cit.) occur at this locality as a series of low relief elongate depressions in the surfaces of foresets (Plate 5.6D). Similarly to the striations described above, however, these structures are apparently orientated parallel to the direction of maximum depositional dip (indicated by the hammer handle in Plate 5.6D) and thus do not exactly conform to the erosive features described by Whitney (op. cit.).

The evidence of the foreset inclination directions plotted in Figure 5.9 suggests that the cross-strata have a concave morphology (in plan) which is

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Figure 5.9. Upper diagram: barchan dune orientated with an east-south-east to west-north-west axis (large arrow points to north). Dip trends (bearings) of the cross-strata demonstrate the configuration of the slipface. Lower diagram: approximate wind distribution over a barchan dune (modified from Bagnold, 1941, Fig.73, p.209). Wind flow lines are only slightly deflected by the contours of the dune (see upper diagram). Reversed eddies in the leeward area enclosed by the horns result from flow separation at the brink of the slipface.

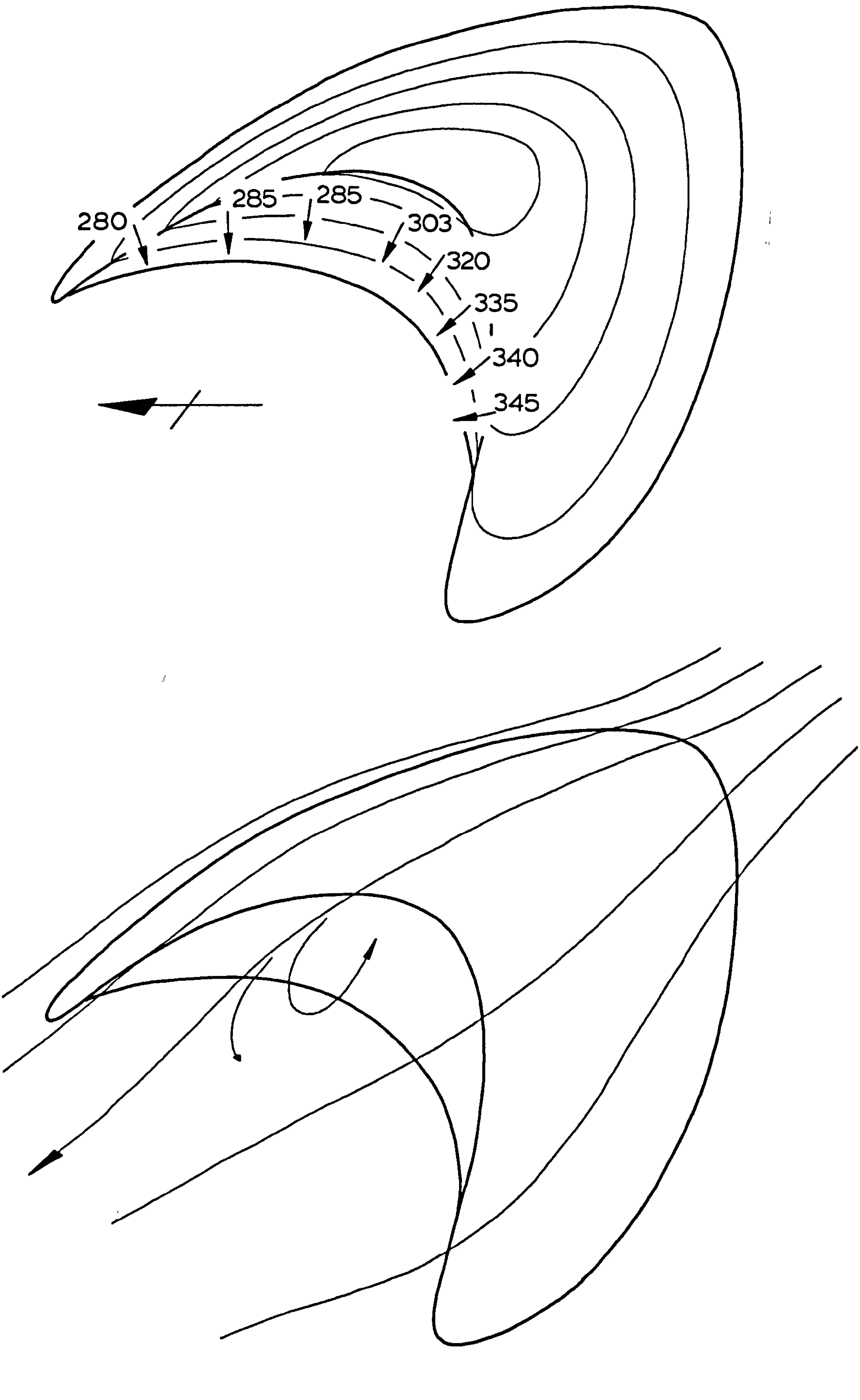

Figure 5.9

clearly that of a crescentic dune slipface in which the horns point in a downwind direction. However, as the relationship of this unit with adjacent dunes is not exposed, it is impossible to determine (from this exposure) whether the morphology is that of an isolated barchan dune or merely a similar section of a transverse barchanoid ridge or sinuous crested dune. For the purposes of Figure 5.9 and following description, this structure will be considered in isolation and the former hypothesis assumed. The lower diagram plotted in Figure 5.9 is a modified version of Bagnold's (1941, p.209) 'approximate wind distribution' over a barchan dune and demonstrates the relationship between the orientation of the dune and the wind flow lines. It is apparent from these diagrams that foreset inclination rarely conforms to the true wind direction and is unreliable, therefore, as an indicator of palaeowind direction. Only where the axes of the structures are exposed, as in Stoneraise Quarry, can true orientations be measured.

Axial measurements of this unit indicate a wind flow towards the north-west (bearing 300 to 320 degrees). The disposition of the horns may be regarded, therefore, as north-east and south-west, the avalanche slope of the former being inclined towards the west and that of the latter towards the north.

LAZONBY RAILWAY CUTTING

A cutting in the Appleby to Carlisle railway line, approximately one kilometre north of Lazonby at NY544406, has exposed Penrith Sandstone along the western side of the track. Although access to this locality is obviously severely limited, the major structures of the exposure may be observed from the opposite embankment and are shown in Figure 5.10. It is not possible, however, to measure the section in detail or to record orientations for the

Figure 5.10. Lazonby Railway Cutting, NY544406 (see Plate 5.7).

Figure 5.10

cross-strata. The cutting is approximately 200 metres in length with sandstone being exposed along most of this distance. Figure 5.10 is constructed from photographs (Plate 5.7) which, due to the limited access, have been angled progressively more obliquely towards northern parts of the exposure and has, therefore, a correspondingly reduced scale in this direction.

A number of dune units, separated by bounding erosion surfaces trending both southwards and northwards are represented. Foresets of these units have an apparent general dip towards the north and are characterised by numerous internal erosion surfaces which, for the most part, are laterally discontinuous. The main feature of interest is the trough-shaped unit exposed mid-way along the cutting. This unit has a markedly erosive, concave-upward lower bounding surface and comprises foresets of similar form which are separated by several minor and discontinuous internal erosion surfaces. With the exception of these minor erosion surfaces, it is clear that the morphology of the cross-strata corresponds to that of Stoneraise Quarry. The exposed width of 35 metres and height of seven metres are also broadly comparable with the dimensions of Stoneraise Quarry. It is also apparent that the structure of this unit is more or less completely dissociated from that of the adjacent units.

Internal erosion surfaces, such as those present immediately above the lower bounding surface of this unit have been noted from many other exposures, such as dune 2 from Whelpdale Hill Quarry. Their formation has been attributed to variable depositional conditions; processes which are in keeping with the initial stages of dune construction. As the main body of sand advances, it is preceded by the horns of the slipface, which may contribute sediment to the area which they enclose. The disposition of this sediment will, therefore, be related to one or other (or both) horns, depending on the extent to which they are developed. Many modern dunes, for example, exhibit an asymmetry which causes one horn to be more prominently developed than the other (see, for example, McKee, 1966; Fig.3, p.12). In this instance it is postulated that the cross-strata of the lower metre or so of sandstone in central parts of the unit are orientated in accordance with the northern horn of the dune slipface with the overlying foresets representing sedimentation of the main part of the slipface. A more detailed interpretation can be made with reference to Figure 5.11.

The upper diagram and section demonstrates the initial stages of sedimentation, where a dune with a prominently developed northern limb overlies the line of the present day exposure. Foresets are aligned obliquely to the exposure and are seen to dip, therefore, at an angle approaching their true maximum value. The lower diagram and section depicts the same dune following sedimentation, the bulk of which having taken place on the southern limb which has become more elongate. The northern limb, receiving relatively less sediment, has retained its original form and, in consequence, the orientation of the slipface has changed to accommodate the modified morphology of the dune. The axis of the dune, however, maintains its trend of 280 degrees.

During the course of the second phase of sedimentation, foresets of both limbs and the main slipface overlie the line of the present day exposure. Cross-strata of the northern limb, unaffected by the remodelling of the slipface, are deposited concordantly on the underlying sediment. However, due to the modified orientation of the slipface, foresets of the main avalanche slope, which strike parallel to (or nearly so) the line of the exposure, now partially overlie sediment deposited by the first phase of sedimentation. Due to the orientation of these foresets, they exhibit an apparent inclination which is much less than that of the underlying sediment. Northwards, however, foresets of the first and second phase of sedimentation have a similar formation on the northern limb of the slipface and are, therefore,

Figure 5.11. Schematic formation of the dune structure exposed in Lazonby Railway Cutting. Dunes are drawn with their axes conforming to a mean palaeowind vector of 280 degrees. The dashed line indicates the orientation of the exposure.

Figure 5.11

structurally concordant. Southwards, sedimentation is of the second phase only. It is apparent from Figure 5.10 that some erosion of the first phase sediment has taken place, probably as a result of the re-orientation of the main slipface. Thus, a small laterally discontinuous internal erosion surface with the same orientation as the overlying foresets may be recognised.

The less marked discontinuity surfaces underlying the surface described above are apparently concordant with cross-strata of the initial phase of sedimentation and mark the southern extent of deposition of this phase. Adjacent foresets to the south were formed by the second phase of sedimentation and mark the infilling of the area previously enclosed between the horns of the dune.

The formation of the concave-upward lower bounding surface is uncertain and will be discussed in more detail in the following section. It should be noted, however, that although the foresets have a corresponding concave-upward form, similar trough-shaped cross-strata can develop above a planar erosion surface, as at Bowscar Working Quarry and Whelpdale Hill Quarry. The morphology of the cross-strata may, therefore, form independently of the lower bounding surface.

TROUGH GILL

The natural sandstone bluffs occurring in Trough Gill (centred on NY588240), 500 metres south of Cliburn, provide the most laterally extensive exposure to be described from the Eden Valley. A series of dune units are exposed more or less continuously for 200 metres on the north-western bank of the gill and for 130 metres on the south-eastern bank. A narrow gorge of 50 metres length reduces the total extent of the exposure to 280 metres. The structure of these units is shown in Figure 5.12, constructed from detailed field sketches and measurement. It has not proved possible to obtain a photographic record of the exposures due to adverse conditions of lighting caused by the narrow character of the heavily wooded valley. The exposure of the south-east bank has been reversed in Figure 5.12 in order that the structural units of the sandstone conform with those recorded from the north-west bank.

From Figure 5.12, it can be seen that the exposure comprises approximately thirteen distinct dune units which are, for the most part, separated by curved or 'trough-shaped' bounding surfaces possessing identical cross-sections to foresets of overlying units. The lower bounding surfaces of the majority of these dune units thus apparently have a concave-upward morphology concordant with that of the foresets. Bounding surfaces are erosive on underlying units and are commonly marked by a considerable disparity of foreset inclination. Assuming the original development of the dunes to have produced structures of more or less comparable dimensions, it is also apparent that, in places, the phases of dune building were separated by periods in which considerable erosion took place. For example, the unit underlying dunes 5a and 6a has dimensions much less than those of surrounding units, presumably as a result of erosion. The present structure bears a remarkable resemblance to that exhibited in Bowscar Working Quarry and similarly demonstrates the formation of concave-upward cross-strata above a planar erosion surface.

As the most important feature of this locality is the lateral extent of the exposure, which is sufficient to enable lateral and vertical relationships

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Figure 5.12. Structures recorded from Trough Gill, NY588240. The exposure of the south-east bank has been laterally reversed to conform with the structures of the north-west bank. Dune 6 has been assumed to represent the highest unit exposed and underlying dune units have been numbered accordingly. Dunes underlying the north-eastern limb of dune 6 with possible lateral equivalents to the south are numbered with the additional suffix 'a'. Dune 4 marks the southern limit of the exposure. Erratum: dune 2 in the central part of the exposure of the south-east bank should refer to the dune unit characterised by 29/280 and is not part of dune 3, as indicated.


South-east bank (reversed)
15/293
21/223
24/007
Dune 5
Dune 6
22/232
26/284
Dune 4
Dune 3
12/250
27/332
25/301
Dune 2
28/005
29/280
Dune 2
23/336
24/311
20/299
Waterfall
22/260
23/309
North-west bank
23/345
22/251
25/295
20/286
20/275
16/262
8/165
Dune 6
Dune 3
Dune 4a
Dune 3a
27/353
21/302
Waterfall
23/309
Dune 1
Dune 2a
21/255
15/287
Dune 2a?
27/335
26/330
33.6m. not exposed
22/349
28/336
30/000
12/302
28/273
28/305
Dune 6a
Dune 5a
24/342
29/338
14/018
22/309
28/012
21/356
30/344
0
10
20
30
40
50
Metres

Figure 5.12

between dune units to be established, it is not proposed to describe individual structures in detail, particularly as a cursory glance adequately demonstrates their morphological similarity to the structure of Stoneraise Quarry. It should be noted, however, that foreset configuration and inclination direction for any individual dune unit conforms exactly with those of the model proposed for the Stoneraise Quarry structure. For example, if the structures of dunes 6 and 3 are considered (these being the most prominently exposed), the concave nature of the foresets is such that while axial trends are towards the west-north-west, cross-strata to the north of the axis indicate a horizontally curving avalanche slope facing the west or west-south-west and foresets to the south of the axis suggest avalanche slopes facing the north. This evidence implies the morphology of the slipfaces to be that of crescentic, barchanoid types with axes trending west-north-west and horns placed north-east and south-west. The maximum distance between the horns of these structures, measured at this locality, is in the order of 100 metres (dune 6), this value being broadly comparable with those of modern barchan dunes from the Salton Sea dune field (e.g. Howard, 1977).

The vertical and lateral relationships exhibited by the dune units of Trough Gill generally show successive structures to overlie earlier formed cross-bedded sets in such a fashion as to result in a maximum disparity of foreset inclination in the order of 90 degrees at the bounding surfaces. This figure approaches that of the maximum spread of the palaeocurrent trends plotted for the Penrith Sandstone (see Figure 5.4). If the structures of dunes 6 and 3 are considered, for example, then it is apparent that depositional dips in the southern parts of the units are approximately towards the north whereas those of the underlying dune units (dunes 3 and 2 respectively) are towards the west. This would suggest (assuming the above model) deposits formed on the south-eastern limbs of the crescentic slipfaces of dunes 6 and 3 to overlie deposits of the north-eastern limbs of dunes 3 and 2.

Similarly, deposits of the north-eastern limb of dune 4 overlie those of the south-western limb of dune 3. Although this latter relationship is less consistently developed throughout the exposure (with the north-eastern limbs of the structures tending to overlie eroded remnants of earlier dune units), the gross cross-sectional configuration of the structures indicates this to be the most likely interpretation.

The vertical arrangement of the dune units thus suggests the successive crescentic slipfaces to occupy or overlie the position between the crescentic slipfaces of the earlier formed dunes. This, in turn, implies the original aerial distribution of the slipfaces to have an 'en echelon' or out of phase arrangement which accords with the dune pattern recorded for many contemporary environments (e.g. Bagnold, 1941, p.212), assuming that the majority of the dune structures recorded from Trough Gill are not separated by periods of complete deflation of earlier formed dunes. That periods of extensive deflation do occur, however, is apparent from the structures of the external southern part of the exposure where the south-western limb of dune 4 overlies the south-western limb of dune 2. In this example, dune 2 is suggested to represent the eroded area between dune 3 to the north and a dune unit of equivalent (unexposed) dune-building phase to the south.

Although evidence from other localities has established a barchanoid morphology for individual dune units, the structures exposed in Trough Gill have enabled the lateral and vertical relationships of these units to be described and have demonstrated their present isolated nature. It is apparent, however, that those units have possible lateral, dune-building phase equivalents which are generally separated by the lower bounding surfaces of succeeding (out of phase) dune structures.

The structures exhibited in the exposure contrast markedly with those recorded by McKee (1966) for comparable dunes from the White Sands dune field in that foresets are remarkably consistent within any specific unit and, with the exception of dune 4a, are uninterrupted by internal erosion surfaces. This implies the sedimentation of the area to be uniform throughout particular phases of dune-building and, therefore, the product of a steady, undeviating palaeowind. The orientation of the dunes, indicating an easterly or south-easterly palaeowind concurs with trade wind patterns characteristic of areas lying within 30 degrees of the equator; the accepted geographical location of the British Isles in Permian times (e.g. Opdyke, 1958; Opdyke and Runcorn, 1960).

Although the dunes of Trough Gill are unquestionably aeolian in origin, several thin beds of breccia have been recorded from the upper reaches of the gorge by Dakyns et al. (1897) and Versey (1974). This horizon was not located by the author but is presumed to represent ephemeral fluvial activity.

INTERPRETATION

Although the majority of Penrith Sandstone exposures in the Eden Valley indicate the complete thickness of the formation to be composed of relatively thin (one to five metres) sets of cross-bedding which truncate underlying and adjacent units and are themselves truncated by overlying units, several less typical exposures have enabled a barchanoid morphology to be established for the preserved dune structures. That the smaller units represent the eroded remnants of similar structures is clearly demonstrable, once this interpretation is accepted. Of the exposures described in the preceding sections, the three localities of Stoneraise Quarry, Lazonby Railway Cutting and Trough Gill are outstanding in importance. These localities have enabled a detailed interpretation of the structures and their three-dimensional reconstruction. In all examples, the concave nature of the cross-strata, shown by the trends of the inclined foresets, indicates a similarly curved slipface which is clearly that of a crescentic dune or segment of a dune. The lateral extent of the Trough Gill exposure, in particular, illustrates the present isolated nature of the units and might be considered to argue against a transverse, sinuous crested dune formation, especially as no convex-upward structures have been recorded, as would be developed on the convex-downwind sectors of such a dune form. However, the preservation of such structures generally is not common (see Allen, 1968, p.119 and Williams, 1971, p.23, for example) and is usually attributed to the simultaneous erosion and deposition related to the downcurrent migration of three-dimensional bedforms. It is suggested for ripples, however, that an estimate of the class of parent bedform may be obtained from the proportion of convex-upward sets, which will be fairly common for sinuous, occasional for catenary and absent for lunate bedforms (Williams, op. cit.). The complete absence of convex-upward structures in the Penrith Sandstone, even where the degree of preservation is as high as in Trough Gill, thus suggests a sinuous crested transverse dune formation to be unlikely, especially as the depth of scour associated with the migrating dunes would have to have been in the order of at least seven metres (assuming the convex-downwind sectors of such dunes to have a comparable height to the concave-downwind sectors). The morphology of the dune units of the Penrith Sandstone, therefore, can more readily be attributed to a dune form which does not develop convex-downwind slipfaces such as the isolated barchan or a barchanoid ridge of the type described by McKee (1979a) and by Inman et al. (1966) from Baja California. As barchanoid ridges comprise parallel rows of coalesced barchans (McKee, 1979a), it is clear that the internal structure of such a ridge would, in transverse section, correspond exactly to a series of adjacent trough-shaped units as described from the Penrith Sandstone. It is also clear that the limbs of any single slipface would overlap the opposite limbs of adjacent slipfaces although the exact nature of this overlap is not yet known. It is considered probable that the internal structure (in transverse section) of this portion of the dune would be akin to that of a sief dune in that its formation would be the product of two opposing avalanche directions. Barchan dunes are recorded to be characteristic of areas where the availability of sand is low (Bagnold, 1941;McKee, 1979a) whereas barchanoid ridges occur where the sand supply is more plentiful, although not so abundant as would favour the formation of transverse dunes (McKee, 1979a). As the dunes of the Penrith Sandstone undoubtedly formed within a sand sea (at least 400 metres of sandstone is recorded from the Appleby-Hilton area; Arthurton et al., 1978), a barchan formation for the structures is the less likely interpretation unless the sands were cemented sufficiently soon after deposition to produce an indurated surface on which subsequent deposition could occur. If a barchanoid ridge interpretation is accepted, then the vertical accretion of the dunes, as described from Trough Gill, would require the crescentic slipfaces of younger dunes to occupy the area between those of the older dunes with the well-defined and solitary bounding surface between adjacent dune units representing an erosion surface of sufficient magnitude to preserve only the trough cross-bedded sets. Although such an interpretation is open to the same constraints as for transverse sinuous crested dunes, it may be possible for a barchanoid ridge to develop a considerable surface relief between the crescentic slipfaces, depending on their lateral spacing. Were this to be the case then the slipfaces of the younger dunes could occupy the topographically lower areas between the older and develop lower bounding surfaces as a modification of a pre-existing topography with relatively little surface deflation. Notwithstanding such an explanation, however, the depth of scour associated with the migration of these bedforms cannot be ascertained and it is proposed to adopt the term 'crescentic dune' for the following discussion, this term referring either to an isolated, barchan dune or to the corresponding segment of a barchanoid ridge.

Whatever the exact nature of the dunes, it is evident that only certain sections enable the true concave-upward nature of the cross-strata to be determined. Exposed axial trends indicate the dunes to be orientated in response to a prevalent east-south-easterly palaeowind direction. Assuming the majority of the crescentic slipfaces to be approximately symmetrical, the mean vector of 281 degrees, as plotted on Figure 5.4, is in accordance with this orientation. Clearly, therefore, it is only in sections perpendicular to the palaeowind direction, i.e. those trending approximately south-south-west to north-north-east, that a complete transect of the preserved dune slipfaces can be obtained and the concave-upward morphology of the cross-strata established. Most other sections cut one part of the slipface only and do not show the true disposition of the sandstone units. The Lazonby Railway Cutting exposure demonstrates that the concave-upward morphology of the cross-strata occasionally may be exhibited by a section other than one trending south-south-west to north-north-east. In this instance, it has been postulated that the morphology of the dune was such that the preserved structure is atypically orientated, as indicated by the structure of adjacent units.

Perpendicular sections, therefore, indicate the minimum width of the crescentic slipfaces of the dunes and longitudinal sections give some indication of the length of the dunes although this value is clearly of little significance in the case of climbing dune forms. The most extensive perpendicular section recorded (Trough Gill) suggests the slipfaces to exceed 100 metres in width and (Stoneraise Quarry) ten metres in height. Continuous longitudinal sections have not been described but occur at several localities such as Cowraik Quarry (NY542310), Maidenhill Quarry (NY525332), Blaze Fell Quarries (NY498435) and Appleby Churchyard (NY688200). In these exposures, cross-bedded dune units, separated by planar or gently curving erosion surfaces, may be traced laterally for several tens of metres. Inclination orientations within any particular set maintain a relatively constant value, although adjacent sets may exhibit opposing orientations, depending on their position of formation with respect to the curving slipface. The most appropriate dimension can, therefore, only be measured from longitudinal sections in which the foresets indicate formation in inner areas of the slipface. The most laterally extensive exposure conforming with these requirements occurs in the upper part of Cowraik Quarry, where a single unit, with foresets inclined towards the north-east (bearing 320 - 340 degrees) can be traced over a distance of approximately 90 metres.

Thus, the minimum dimensions for the dune structures of the Penrith Sandstone are 90 metres across their centre, in the direction of dominant windflow, and 100 metres normal to this, from horn to horn. Although these dimensions are considerably greater than those of barchan dunes examined by McKee (1966), from the White Sands dune field, and Howard (1977), from the Salton Sea dune field, Bagnold (1941) observes the maximum length and width of existing dunes to be approximately 400 metres, with slipfaces of 30 metres height and Norris (1965) records dunes of 165 metres width from Imperial Valley, California. Dune size and form is dependent on a number of factors, including wind velocity and grain size (Wilson, 1972). In particular, variations of these parameters may lead to corresponding changes in height/width and width/length ratios (Howard et al., 1978), the former having an obvious effect on the steepness of the slipface. It is this mechanism which is proposed to account for the non-erosive internal discontinuities noted from many of the exposures examined. The minimum dimensions of the crescentic dunes of the Penrith Sandstone are, therefore, in accordance with those of present day barchan dunes.

A comparison of the main structures within the lesser-developed units exposed elsewhere in the Eden Valley indicates a comparable formation to be plausible over the whole extent of the dune field. For example, the lower unit of Bowscar Quarry (Figure 5.6) is comprised, at its northern extreme, of foresets inclined towards the north, which, interpreted in the manner prescribed above, represents the south-western horn of a crescentic dune.

The erosive upper bounding surface, which is inclined towards the south, can be explained in terms of a modification of the original windward slope of the horn. Following the cessation of sedimentation, incomplete erosion would produce a planar surface partially related to the orientation of the original slope, i.e. one inclined towards the south. Bowscar Working Quarry also shows a similar relationship between the orientation of the erosion surfaces and the original morphology of the dune. The structure of dune 4 in Figure 5.7 represents a transverse (perpendicular to dominant wind) section with both north-eastern and south-western horns exposed. Erosional modification of the windward slopes of the horns has resulted in planar upper surfaces inclined towards the east (bearing 082 degrees) and south (bearing 178 degrees) respectively.

In many localities the degree of exposure is such that accurate reconstructions cannot be made. In these examples the rock surface appears to be composed of numerous sets stacked one on top of the other and in which foresets exhibit a relatively uniform orientation. Frequently, however, adjacent sets have opposing foreset orientation which gives the structure a superficial resemblance to the internal structure of a sief dune. Glennie (1970, p.110) suggests that the exposures of Hilton Beck have such a formation: 'Bedding attitudes in the sands of the Lower Penrith Sandstone at this locality suggest that they were part of a sief dune formed by a wind that blew from the east'. However, foreset inclination orientations at this locality all lie within the zone plotted in Figure 5.4 and do not, therefore, preclude a crescentic dune formation. It is postulated that the surfaces separating the sets are erosional bounding surfaces between subsequent crescentic dunes and not the internal discontinuities of sief dunes. The apparent opposition of the foresets can be related to sedimentation of one limb of a crescentic dune overlying the eroded remnants of the opposite limb of the preceding dune. Clearly, where phases of erosion of this magnitude separate the dune-building phases the planar surfaces produced bear little relationship to the original morphology of the eroded dune.

DISCUSSION

Detailed descriptions of barchanoid ridges have not yet been published and the only work which describes the morphology and internal structure of modern barchan dunes is that of McKee (1966). The structure of a single barchan from the White Sands dune field, New Mexico was examined in two trenches, one being a transverse section and the other a section through one of the horns. Unfortunately, the exact orientations of the trenches, with respect to the dune, are not recorded and some confusion exists in the direct application of McKee's (op. cit.) work. In marked contrast to the dunes described from Stoneraise Quarry and Trough Gill, the example studied by McKee (op. cit.) consisted of a large number of sets of cross-strata bounded by horizontal or downwind dipping surfaces. Neither trench was cut normal to the dominant wind direction so that foresets are generally inclined in response to the wind with only one or two examples exhibiting inclinations to windward. The large number of internal bounding surfaces suggests that the dune is formed of a similar number of sedimentation phases, each separated by periods of erosion in which considerable deflation of the upper surface took place. The internal structure of the dune, as described by McKee (1966), is, therefore, the end product of a number of varied phases of sedimentation and erosion. The dunes described do not appear to be simple barchans in that slipfaces other than the typical solitary crescentic avalanche slope are developed and it is, therefore, possible that some of the lower sets exposed in the trenches are related to these slipfaces. The application of McKee's (op. cit.) work to the dunes described from the Penrith Sandstone is, therefore, limited.

Of note, however, are the trough structures appearing as lenses up to 1.5 metres wide and 0.6 metre deep recorded from the trench cutting the horn of the dune and which McKee (op. cit.) attributes to blowouts resulting from fluctuations in wind regime. The small scale trough structures recorded from dune 3 of Whelpdale Hill Quarry are similar in structure and scale and may, therefore, be formed in an identical manner.

The development of the dunes of the Penrith Sandstone thus represents sedimentation of a more uniform and continuous nature than that of the White Sands dune field, possibly in response to a less varied wind system. The larger dunes described are notable in that the internal structure is characteristically unaffected by erosion surfaces and is remarkably uniform. Such structures can only be formed where winds are consistent not only in direction but also in strength. The smaller structures exposed at many localities, e.g. Hilton Beck, George Gill and Udford Crags (NY570304) may be components of dunes similar to the type described by McKee (1966), but equally may represent the eroded remnants of larger, uniform dunes of the Trough Gill type.

The lower bounding surfaces of many examples exhibit a concave-upward cross-section, the origin of which, although having a possible explanation in Trough Gill, is questionable in Lazonby Railway Cutting. Foresets of adjacent units suggest the exposure to be orientated oblique to the dominant palaeowind direction and are thus approximate longitudinal sections of dune structures. The bounding surface under discussion cannot, therefore, be related to a modification of the morphology of pre-existing dunes and is more likely to be a purely erosive feature. The scale of the structure is comparable with transverse sections of crescentic dunes recorded elsewhere (e.g. Stoneraise Quarry) and is, therefore, thought to be of too great a magnitude to be regarded as a blowout, unless it represents an 'interdune blowout' of the type described by Walker and Harms (1972). However, as Walker and Harms (op. cit.) do not describe the structure of the infilling sediment, the precise origin of this feature is uncertain. It is clear, however, that the concave form of the foresets in transverse section is not dependent on a similarly curved lower bounding surface, as examples have been described in which trough-shaped cross-strata develop above a planar lower bounding surface, e.g. Bowscar Working Quarry. It will be noted that the lower laminae are flat lying, or only gently inclined, and progressively steepen away from the axes of the dunes in higher parts of the set.

Some discussion concerning the orientation of the slipface may be regarded as relevant following the suggestion in previous sections that foreset inclination rarely conforms to the true wind direction. It is, however, considered outside the scope of this research to attempt an evaluation of the origin and development of dunes, those processes being adequately described by Wilson (1972) for a number of aeolian bedforms.

Whilst Bagnold (1941) clearly demonstrates that wind patterns associated with a barchan are vertically and, to some extent, laterally displaced by the contours of the dune, he does not suggest that the crescentic form of the slipface may be directly attributed to this modified windflow. Instead, he states (p.209):

> 'As far as the wind is concerned, therefore, we can regard the dune as a complete circular, or elliptical mound whose surface is continued over the 'bite' by an imaginary 'surface of discontinuity.'

Flow separation at the slipface brink is thus taken to be complete and the wind distribution symmetrical about the crest of the dune. It is clear, therefore, that the form of the slipface is completely dissociated from the almost parallel lateral streamlines. Bagnold (1941, p.210) attributes the form of the slipface to a differential rate of advance:

> 'If we suppose the face to run straight across the dune just in front of the summit, the low extremeties would be exposed to a sand flow very nearly as strong as that at the summit. The extremeties would therefore advance more rapidly than the centre, and the face would become concave.'

In support of this reasoning, Howard et al. (1978) record the horns of barchans to be the site of 'overly-strong' deposition. Thus, these authors maintain that once initiated, the crescentic form of the slipface is preserved by the differential movement of avalanching sand grains and is not dependent on the presence of fixed leeward eddies (e.g. Cornish, 1897, 1900).

Barchans from the Algodones dune field in California (Plate 5.8) show features which superficially support the wind pattern model proposed by Bagnold (1941). In these examples, trains of parallel ripples are superimposed on the windward slopes of the dunes and (assuming ripple crests to be orientated perpendicular to the prevalent winds) indicate windflow over the dune to be uniform and laterally undeflected by the contours. Howard (1977) and Howard et al. (1978) have demonstrated, however, that slope (in this case, the stoss slope of the dune) has an important effect on the transport of sand grains by surface creep, so that ripples tend to assume an orientation corresponding to a direction which is intermediary between downwind and downgradient. Howard (1977, Fig.2, p.854) includes a diagram (reproduced in Figure 5.13) in which the wind directions measured at the surface of a modern barchan dune from the Salton Sea are plotted. These measurements indicate that lateral windflow is considerably modified by the dune, so that flow lines upwind of the crest line are divergent and those over the slipface brink and horns are convergent. Lateral flow patterns may be, therefore, more intimately associated with the morphology of the dune than suggested by the observations of Bagnold (1941), Landsberg (1942) and Sharp (1966), with the crescentic form of the slipface possibly reflecting the configuration of the modified windflow. Foreset inclination directions measured from the horns of barchans may thus correspond to a true (although not the prevalent) wind

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Figure 5.13. Wind directions at the surface (heavy lines) compared with observed (light lines) and predicted (dashed lines) downwind perpendiculars to ripple crests on the stoss slope of a barchan dune near the Salton Sea, California. Reproduced from Howard (1977, Fig.2, p.854).

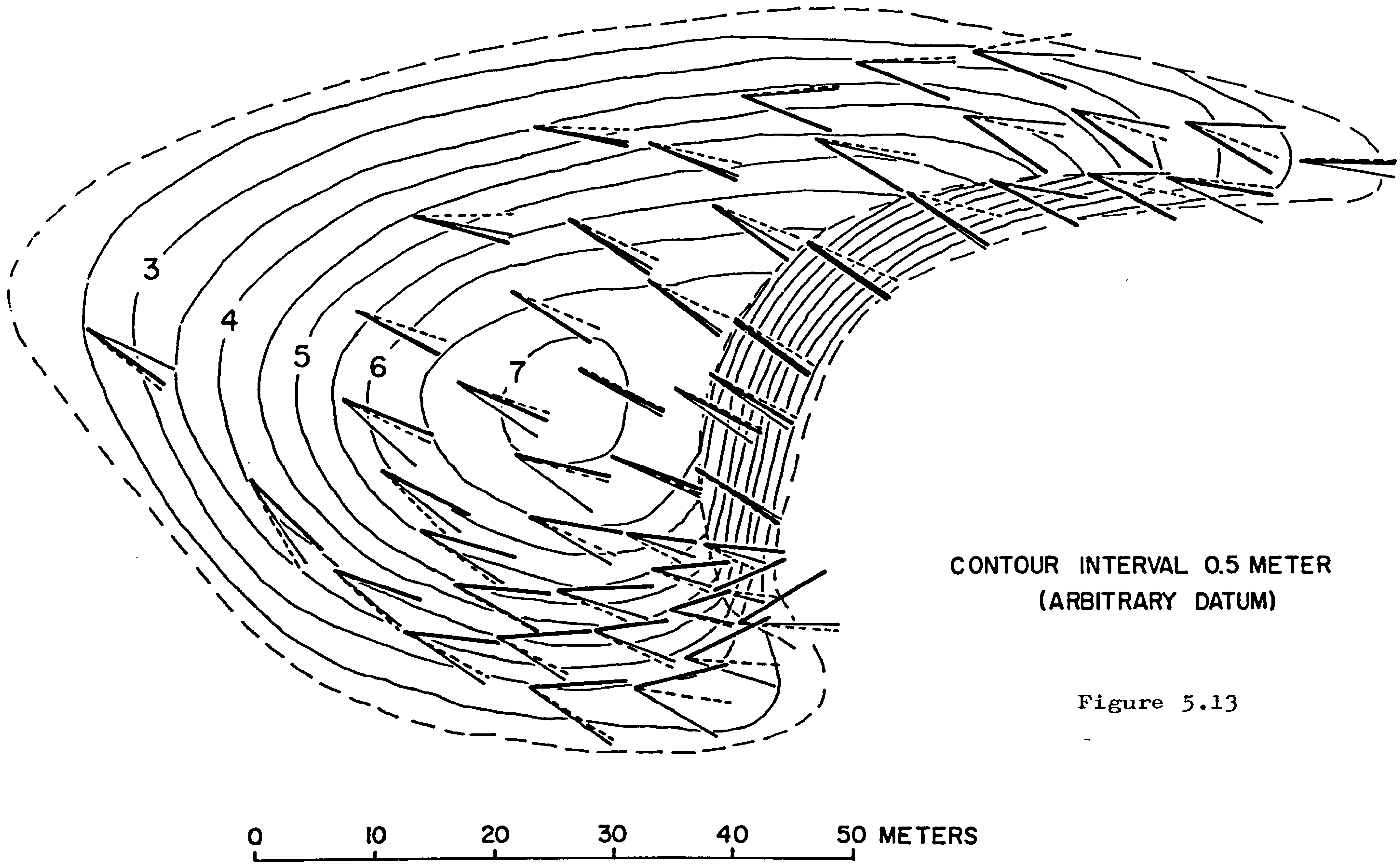
3
4
5
6
7
CONTOUR INTERVAL 0.5 METER
(ARBITRARY DATUM)
0
10
20
30
40
50 METERS

Figure 5.13

direction. In accordance with the work of Bagnold (1941), the slipface indeed appears as a 'bite' which sharply truncates the ripples of the stoss slope (Plate 5.8b).

The barchans of the Algodones field also give some insight into the processes active in the lee of the dunes, in which area wind ripples may be observed. By reference to these leeward ripples, asymptotically and angularly based dunes may be distinguished. In Plate 5.8b and d, the slipface may be seen to be concave in cross-section, i.e. asymptotically based. Furthermore, the slipface appears to be formed of two components; an upper zone and a lower zone which passes downwind into a zone with no sedimentary structures and thence into an area of wind ripples. According to Bagnold (1941), the area enclosed by the horns of the dune is an almost perfect wind shadow in which he envisages the formation of weak reverse eddies as a result of flow separation at the slipface brink. Similarly with the mechanism proposed by Joplin (1965) for subaqueous sediments, these reverse eddies may contribute material to the lower, or asymptotic, zone of the cross-strata, presupposing them to be capable of transporting sand. In this case, only the upper zone forms the true avalanche slope. A comparable mode of formation for subaqueous structures has been proposed by Allen (1970b).

Plate 5.8c shows angular based foresets formed solely by avalanching. In this example, the slipface advances over ripples formed in the lee of the dune without any reworking of the sediment.

The orientation of the ripples formed in the lee of the dunes is also worthy of note and has been briefly described by Bigarella (1972). In inter-dune areas ripple crests are normal to the dominant flow direction, but towards the lee of the dunes this alignment is progressively modified until, between the horns of the barchans, the ripples assume an orientation parallel with the dominant wind (Plate 5.8c and d). This suggests that the area is indeed a wind shadow in which wind cells, formed by flow separation at the slipface brink, not only form reverse eddies but also lateral eddies. The generation of such lateral eddies may be reasonably related to a convergent wind pattern. It is postulated that the orientation of the ripples is due to lateral eddy systems, rather than to temporary wind flow at right angles to the dominant direction, as suggested by Bigarella (1972).

The evidence presented above suggests that the prevalent palaeowind orientation can only be plotted from measurements obtained from exposed axial zones of dunes. Relatively accurate trends can be plotted as a mean vector of a number of measurements, assuming the sample to be great enough to minimise bias and assuming the majority of the dunes to be symmetrical.

CONCLUSIONS

Although an analysis of the dune structures of the Penrith Sandstone has enabled a barchanoid morphology for the individual preserved dune units of several well-exposed localities to be established and considerations of their vertical and lateral relationships in Trough Gill has demonstrated their present isolated nature, it has not proved possible to reconcile these features with any single specific dune type. Of the various possible analogies, however, a barchanoid ridge interpretation seems the most likely on the basis that convex-upward structures have not been recorded (as would favour a sinuous, transverse crested dune model) and on the basis that isolated dune forms are unlikely to have predominated in an area of plentiful sand supply (as would appear to be the case in the Eden Valley). A major consideration in the refinement of any such study must relate to the possible effects of deflation associated with the migration of such large scale bedforms and the exact relationship of the lower bounding surfaces of the dune units. Whereas considerable depths of scour must have been attained in certain localities (viz. Lazonby Railway Cutting), it is by no means clear that this process could have operated throughout the dune field in a sufficiently systemmatic manner as to remove all trace of structure related to the migration of the dunes other than that formed on the crescentic-downwind slipfaces, unless, as has been suggested for barchanoid ridges, this preferential preservation can be related to an external (topographic) control.

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THE UNIVERSITY OF HULL

Sedimentology of the Penrith Sandstone and brockrams (Permo-Triassic) of Cumbria, north-west England

Volume 2

Appendices and Plates

being a thesis submitted for the Degree of

Doctor of Philosophy

in the University of Hull

by

Louis Charles Macchi B.Sc.

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APPENDIX ONE

Borehole data

Summary data from borings in the haematite mining area of west Cumbria, reproduced by permission of the British Steel Corporation. Locations refer to the British Grid Reference System. All levels are given in metres and refer to mining datum, 609.6 metres below ordnance datum.

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
1	NY02520887	722.68	439.83	373.38	66.4
2	NY02830906	702.87	475.18	403.86	71.3
3	NY02630869	694.03	420.01	346.56	73.5
4	NY02450872	717.50	427.94	336.37F	61.6F
5	NY02380855	713.84	438.30	357.23	81.4
6	NY02480864	709.88	419.10	341.68	77.4
7	NY02570899	729.08	467.56	395.02	72.5
8	NY02640915	740.97	489.81	418.80	71.0
9	NY02220917	738.23	517.25	443.18	74.1
10	NY02150904	740.97	496.52	427.33F	69.2F
11	NY02370987	725.42	501.70	433.73	68.0
12	NY02350920	744.63	488.29	421.54	66.8
13	NY02040973	714.76	529.13	461.47	67.7
14	NY02140960	722.68	520.90	452.93	68.0
15	NY02280943	739.44	505.97	430.68	75.3F
16	NY02550941	749.50	511.15	436.78	74.4
17	NY02890924	718.11	484.94	418.19	66.4
18	NY02940966	734.87	486.16	449.58F	36.6F
19	NY02930973	731.52	522.43	486.46	36.0
20	NY02940962	739.14	493.78	438.61	55.2
21	NY02870963	740.66	492.25	434.04	58.2
22	NY02900970	734.26	485.85	459.33	26.5F
23	NY02920968	734.57	487.38	452.32	35.1
24	NY02770952	749.50	475.79	406.30	69.5
25	NY02230960	731.22	523.24	455.07	68.3
26	NY02300973	728.17	509.93	440.44	69.5
27	NY02870880	676.35	446.53		
29	NY02450967	743.71			
30	NY02740819	674.52	327.96	251.16	76.8
31	NY02960838	659.28	356.92	279.20	77.7
32	NY03420930	691.90	553.52	508.71	44.8
33	NY03120927	702.87	462.69	399.90	62.8
34	NY02920825	657.45	348.08	269.14	78.9
35	NY03070816	674.83	327.36	238.05	89.3
36	NY02980799	673.61	328.57	238.96	89.6
37	NY03190877	681.53	449.58	383.74	65.8

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
38	NY03040906	687.63	449.88	386.79F	63.1F
39	NY03770822	720.55	593.75F	556.87	36.9F
40	NY02940937	729.69	468.78	404.77	64.0
41	NY03190886	685.19	450.19	387.10	63.1
42	NY03410813	694.33	306.93	218.54	88.4
43	NY04080806	728.47	534.31	450.49	83.8
44	NY03990879	739.14	606.86	577.60	29.3F
45	NY03550816	764.44	413.00	363.02	50.0F
46	NY03910888	733.96	608.69	577.90	30.8
47	NY03980827	742.19	551.99	469.09	82.9
48	NY04040870	750.42	591.92	533.70	58.2
49	NY03170839	676.35	414.22	354.48	59.7
50	NY03820897	730.91	603.50	588.26F	15.2F
51	NY04080879	749.50	613.26	557.78	55.5
52	NY03840842	736.70	633.07	593.45	39.6
53	NY03840887	728.17	597.41	562.05	35.4
54	NY03850898	735.48	622.40	599.85	22.6
55	NY04100864	764.44	597.71	532.79	64.9
56	NY04170898	756.51	643.43	611.73	31.7
57	NY04790829	732.43	630.63	568.76	61.9
58	NY03720895	723.60	586.13	574.85F	11.3F
59	NY04760897	747.67	724.20	694.94	29.3
60	NY05060922	783.64	744.63	695.25	49.4
61	NY05300851	797.05	677.27	629.41F	47.9F
62	NY03870876	729.39	616.92	583.99	32.9
63	NY04160875	762.00	610.51	554.43	56.1
64	NY04950948	779.37	726.95	715.06F	11.9F
65	NY05330857	804.98	704.09	640.69	63.4
66	NY05030954	787.60	742.80	693.72	49.1
67	NY05030986	780.90	709.27		
68	NY04050890	744.93	617.52	591.62	25.9F
69	NY04940713	722.07	475.49	386.79	88.7
70	NY03740794	705.92	359.05	348.69	10.4F
71	NY03620842	712.32	586.44	537.06	49.4
72	NY03630786	697.99	309.68	232.56	77.1
73	NY04000740	688.85	359.36	262.74	96.6

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
74	NY03490843	701.95	432.82	390.75	42.1F
75	NY04700947	760.48	ABSENT	ABSENT	0.0
76	NY02660831	690.37	369.42	295.66	73.8
77	NY04650934	755.90	ABSENT	734.26	21.6
78	NY03290864	693.72	448.06	386.18	61.9
79	NY04410923	771.14	687.93	664.77	23.2
80	NY04200643	688.85	358.44F	342.60F	15.8F
81	NY04450722	687.02	411.18	334.06	77.1
82	NY05380780	751.94	579.73	523.04	56.7
83	NY03940680	673.30	302.67	230.12	72.5
84	NY04390718	684.28	393.50	302.36	91.1
85	NY01890824	696.77	387.40	304.80	82.6
86	NY05110786	750.72	647.40	588.57	58.8
87	NY01800813	498.30	402.34	322.78	79.6
88	NY01980813	694.03	368.50	290.47	78.0
89	NY04600708	698.91	420.32	326.14	94.2
90	NY02200790	683.97	324.00	240.79	83.2
91	NY01920898	745.54	476.10	403.86	72.2
92	NY02000766	668.73	267.61	189.28	78.3
93	NY04960777	730.30	602.59	530.96	71.6
94	NY01760746	663.85	207.26	113.39	93.9
95	NY03020773	664.77	303.28	209.40	93.9
96	NY04960806	741.88	614.78	555.04	59.7
97	NY03310690	658.06	274.93	176.17	98.8
98	NY01560858	700.74	331.01	220.68	110.3F
99	NY05240798	753.47	626.06	577.29	48.8F
100	NY02720822	679.09	341.07	263.96	77.1
101	NY03210802	680.62	308.15	218.24	89.9
102	NY04250827	765.05	576.68	499.26	77.4
103	NY03450714	676.66	284.07	181.66	102.4
104	NY03840859	735.48	644.65	611.73	32.9
105	NY03860868	732.43	606.86	596.80F	10.1F
106	NY04170747	697.69	368.20	309.37F	58.8F
107	NY03370780	680.31	306.93	217.93	89.0
108	NY03800877	723.29	594.06F	ABSENT	0.0F
109	NY04250908	763.83	655.62	618.44	37.2

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
110	NY02800735	653.49	187.45	151.18	36.3F
111	NY04360754	715.06	480.97	385.88	95.1
112	NY03540755	682.45	281.33	191.11	90.2
113	NY04350927	761.09	701.95	674.22	27.7F
114	NY04380925	764.13	695.55	669.34	26.2
115	NY04230778	728.17	522.73	434.34	88.4
116	NY03070780	669.65	304.19	223.42	80.8
117	NY03030766	663.55	296.57	207.57	89.0
118	NY03900905	735.48	630.02	608.69	21.3
119	NY04530760	711.40	530.66	464.21	66.4
120	NY03620698	675.74	320.95	223.42	97.5
121	NY03650863	711.10	603.81	566.01	37.8
122	NY03450797	690.07	298.09	203.61	94.5
123	NY02000772	676.96	325.53	235.61	89.9
124	NY03610881	709.57	610.82	567.23	43.6
125	NY02540743	647.70	213.36	165.51	47.9
126	NY02540772	655.02	219.46	191.11	28.3
127	NY01680805	674.83	259.38	163.98	95.4
128	NY02850790	662.03	309.68	225.86	83.8
129	NY03500881	702.56	500.48	464.21	36.3
130	NY03240778	671.17	303.58	210.31	93.3
131	NY02500790	677.27	333.45	257.86	75.6
132	NY04070916	744.02	659.59	625.75	33.8
133	NY03430855	693.42	448.06	392.58	55.5
134	NY02810771	650.44	294.74	205.44	89.3
135	NY04660914	768.10	740.05	707.44	32.6
136	NY04570875	767.79	638.25	591.62	46.6
137	NY01470829	678.79	189.59	167.34	22.3
138	NY04710882	748.59	719.33	654.41	64.9
139	NY03230760	668.73	290.78	202.39	88.4
140	NY03980921	738.53	655.93	619.96	36.0
141	NY04320885	778.15	625.45	573.94	51.5
142	NY02330718	644.35	137.16	117.65	19.5
143	NY04370840	778.15	610.51	537.36	73.2
144	NY02320695	640.99	108.20	85.04	23.2
145	NY04320733	694.33	415.14	317.91	97.2

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
147	NY04430903	790.65	673.00	619.96	53.0
148	NY02150704	674.52	108.20	80.16	28.0
149	NY04500890	786.68	651.05	608.08	43.0
150	NY03970910	738.84	620.29	605.64F	14.6F
151	NY02580712	647.09	211.23	125.88	85.3
152	NY04450858	776.33	612.04	586.44	25.6
153	NY04020785	705.00	503.53	409.35	94.2
154	NY02850692	653.48	135.64	106.86	29.0
155	NY03630890	714.45	624.84	594.06	30.8
156	NY03090831	673.91	341.99	248.41	93.6
157	NY03890919	732.13	630.63	598.93	31.7
158	NY04020900	741.27	624.23	598.32F	25.9F
159	NY03490937	695.55	552.60	502.01	50.6
160	NY03910861	741.58	598.63	571.50	27.1
161	NY03940869	737.62	603.81	589.79	14.0
162	NY03580935	678.48	538.28	494.69	43.6
163	NY05810759	744.02	606.86	561.44	45.4
164	NY05620780	753.47	613.56	565.10	48.5
165	NY05410794	759.87	590.70	539.80	50.9
166	NY05140668	713.54			
167	NY04850690	721.77	399.90	301.45	98.5
170	NY03990916	739.75	648.61	614.78	33.8
171	NY03950914	738.53	631.55	605.03	26.5
172	NY03950902	737.62	635.20	604.72	30.5
173	NY04140911	746.76	650.75	625.14	25.6
174	NY04180916	755.29	651.05	615.70	35.4
175	NY04190910	752.86	659.89	630.63	29.3
176	NY04200917	751.94	663.55	639.78	23.8
177	NY04310908	769.32	650.14	611.73F	38.4F
178	NY04130915	747.37	666.29	635.20	31.1
179	NY04200922	751.64	679.09	650.14	29.0
180	NY04290922	760.17	685.50	660.50	25.0
181	NY04070929	740.36	679.09	661.72	17.4
182	NY04210928	749.50	692.20	671.17	21.0
183	NY04140923	744.63	672.69	643.13	29.6F
184	NY03970936	728.47	671.17	646.79	24.4

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
185	NY03860939	714.45	658.06	629.72	28.3
186	NY03900927	729.39	651.66	622.40F	29.3F
187	NY03770911	726.34	601.07	580.95	20.1
188	NY03650900	718.41	585.52	572.12F	13.4F
189	NY03830916	728.47	620.27	593.45F	26.8F
190	NY03780924			608.69F	30.2F
191	NY03720920	713.84	588.87	564.79	24.1
192	NY03730931	705.31	631.85	586.74	45.1
193	NY04150933	740.66	694.94	673.30	21.6F
194	NY04050925	740.05	669.65	645.26	24.4
195	NY04090940	737.01	698.91	669.65	29.3
196	NY03910934			635.20F	28.7F
197	NY04290926	757.12	695.86	670.56	25.3
198	NY04800945	764.74	ABSENT	ABSENT	0.0
199	NY04830930	762.30			
200	NY04540934	754.99	741.88	724.81F	17.1F
201	NY04570919	775.41	715.06	682.45	32.6
202	NY04450935			708.96F	21.3F
203	NY0500927			694.03F	21.6F
204	NY04630899			671.78F	19.5F
205	NY04630950	763.83	ABSENT	757.43	6.4
206	NY04550950	764.44	ABSENT	ABSENT	0.0
207	NY04490952	767.79			
208	NY04400946	759.56	ABSENT	740.36	19.2
209		770.23	674.83	643.13F	31.7F
210	NY04360937	753.77	731.82	710.49	21.3
211	NY04480919			658.06F	33.5F
212	NY04600886	765.66	660.20	618.44	41.8
213	NY04710939	757.12	ABSENT	ABSENT	0.0
214	NY04670936	755.29	ABSENT	ABSENT	0.0
215	NY04490898	790.96	664.16	631.85	32.3
216	NY04620938	756.21	ABSENT	ABSENT	0.0
217	NY04560941	762.61	ABSENT	746.76	15.9
218	NY04510941	763.52	757.73	740.66	17.1
219	NY04440943	761.09	752.55	729.08	23.5
220	NY04480942	762.30	752.55	725.42	27.1

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
221	NY04400932	760.78	714.15	685.19	29.0
222	NY04600908	778.15	706.53	682.75	23.8
223	NY04460929			694.94F	19.2F
224	NY04570931	758.34	ABSENT	ABSENT	0.0
225	NY04600914	777.85	716.58	689.76	26.8
226	NY04750929	758.95	ABSENT	ABSENT	0.0
227	NY04700923	762.91	ABSENT	733.35	29.6
228	NY04520914			674.83F	20.4F
229	NY04430912	784.86	676.35	648.00	28.3
230	NY04550904	786.69	694.64	669.34	25.3
231	NY04370928	761.70	709.88	677.57	32.3
232	NY05000695	734.87	457.50	368.20	89.3
233	NY04350915	769.32	670.86	633.68	37.2
234	NY04960688	730.30	420.62	327.66	93.0
235	NY03740889	722.07	588.57F		
236	NY03690893	719.33	608.99	602.28F	6.7F
237	NY03680886			591.92F	36.0F
238	NY03410790	683.36	293.52	194.16	99.4
239	NY03100820	676.35			
240	NY03090805	675.74	312.42	232.87	79.6
241	NY03510786	689.76	300.53	198.42	102.1
242	NY03780885			572.11F	5.5F
243	NY03820883			575.77F	4.6F
244	NY03830881			573.02F	21.9F
245	NY03760885	722.68	598.32F		
246	NY03010828	675.13	335.28	250.85	84.4
247	NY03400810	691.59	305.41	201.47	103.9
248	NY03750882			605.33F	27.4F
249	NY03810879			584.61F	5.8F
250	NY04040897			607.47F	5.2F
251	NY04000901	739.44			
252		743.71	613.56	601.98	11.6
253	NY03560846	706.53	522.12F	480.36	41.8F
254	NY03620824	707.75	510.54	509.02F	1.5F
255	NY02610829	691.90	374.00	318.52	55.5
256	NY03600820	705.92	474.88F	445.01	29.9F

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
257	NY03530845	704.70	439.22	391.97	47.2
258	NY03710732	678.18	267.00	143.56	123.4
259	NY02630830	691.29	369.11	296.57	72.5
260	NY04100914	744.93	658.06	628.50	29.6
261	NY03600749	681.84	276.15	178.92	97.2
262	NY04150908	750.72	650.44	624.23	26.2
263	NY02640829	690.98	317.91	295.66	22.3
264	NY03980869	742.49	592.23	565.10	27.1
265	NY03970872	739.75	594.06	566.32	27.7
266		718.72	496.21	430.07	66.1
267	NY03940863	739.44	586.74	573.63	13.1
268	NY03920863	738.23	595.27	569.37	25.9
269	NY02950830	658.06	343.20		
270	NY04420760	718.11	519.99	447.75	72.2
271	NY03930861	740.66	580.64	569.37	11.3
272	NY03960828	741.27	549.55	472.44	77.1
273	NY04400760	718.72	503.83	439.52	64.3
274	NY02930830	658.06	317.91	271.58	46.3
275	NY04020826	743.71	548.03	460.86	87.2
276	NY02910822	657.15	343.81		
277	NY04070824	746.15	547.73	466.34	81.4
278	NY04240740			316.69F	50.9F
279	NY03940753	686.41	370.94		
280	NY04210758			374.60F	88.4F
281	NY03890700			216.71F	60.7F
282	NY04030658	685.80	361.80	309.37	52.4F
283	NY04120675	679.70	291.08F	266.09	25.0F
284	NY04350673	680.92	317.30	210.31F	107.0F
285	NY04290658	679.70	245.06	215.49F	29.6F
286	NY05100685	739.75	453.24	363.93	89.3
287	NX99061081	707.60	348.80	338.20	10.6
288	NX98861093			362.71F	64.0F
289	NY03200960	718.11	562.66	511.76	50.9
290	NY03160952	716.89	515.11F	510.54	4.6F
291	NY03080943	716.58	471.83	411.48	60.4
292	NY03380980	722.99	574.24	530.05	44.2

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
293	NY01331015	648.00	551.08	469.70	81.4
294	NY01921038	668.43	629.11		
295	NY01921038	668.73			
296	NY01841035	666.57	579.12	521.94	57.3
297	NY01841040			533.40F	49.4F
298	NY01841034	665.68	579.12	520.90	58.2
299	NY01871035	667.21	578.21	531.88	46.3
300	NY01671042	658.06	569.67	512.98	56.7
301	NY01551047	661.11	590.70	530.35	60.4
302	NY01581024	662.94	553.16	493.78	59.4
303	NY01860967	702.26	556.87	482.80	74.1
304	NY01770963	698.04	540.36	475.95	64.4
305	NY00061007	671.79	451.41	447.65	3.7
306	NY01940951	714.63	542.57	473.84	68.7
307	NY02140935	727.56	540.11	470.31	69.8
308	NY00161003	670.54	417.33	416.34F	1.0F
309	NY00131014	668.34	425.64	423.18	2.5
501	NY04180970	751.03			
502	NY03980959	735.48	713.54	692.51	21.0
503	NY04220936	744.02	710.18	689.46	20.7
504	NY03910966	730.91	726.03	704.70	21.3
505	NY03740961	708.36	695.55	655.93	39.6
506	NY04110948	738.23	712.93	682.45	30.5
507	NY03930944	718.72	687.32	655.02	32.3
508	NY03740970	708.36	ABSENT	652.27	56.1
509	NY03760947	706.83	665.07	626.36	38.7
510	NY03580952	678.79	561.44	527.91	33.5
511	NY03790932	711.17	640.38	623.93	16.5
512	NY03470867	695.86	470.61	418.80	51.8
513	NY03660959	701.95	550.77	548.64F	2.1F
514	NY03371020	732.13	ABSENT	ABSENT	0.0
515	NY03351014	737.62	667.51	ABSENT	0.0
516	NY03341012	738.84	643.13	617.52	25.6
517	NY03331016	740.05	717.88	661.11	56.7
518	NY03361015	737.01	712.01	680.62	31.4
519	NY03351013	737.01	711.40		

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
520	NY03301004	739.14			
522	NY03361014	737.62	708.66	666.90	41.8
523	NY03371012	736.70	713.23	637.34	75.9
524	NY03441012	729.99	ABSENT	ABSENT	0.0
525	NY03401013	732.74	708.36	ABSENT	0.0
526	NY03401010	733.35	723.29	683.06	40.2
527		716.28			
529	NY03331014.	738.84		633.68	
530	NY03520990	716.28	707.75	618.44	89.3
531	NY03520981	704.09	695.55	562.05	133.5
532	NY03480967			572.41F	125.3F
534	NY03331005	737.31	584.00	ABSENT	0.0
535	NY03160995	733.65	579.73	530.05	49.7
536	NY02970995	731.22	582.78	531.27	51.5
537	NY02900989	728.47	559.00	499.26	59.7
538	NY02981013	743.41	612.65	562.66	50.0
539	NY02981020	746.46	619.96	571.80	48.2
540	NY02991027	741.58	623.01	570.59	52.4
541	NY02981005	736.40	599.24	548.34	50.9
542	NY03071030	730.00	662.71	605.64	17.1
543	NY02331046	690.37	663.24	653.80	9.4
544	NY02251033	693.42	662.33	599.85	62.5
545	NY02131046	678.18	651.66	638.86	12.8
546	NY02321055				
547	NY02321051				
554	NY02181041	685.19	ABSENT	626.36	58.8
555	NY02321036	694.03	669.95	611.73	58.2
556	NY02201059				
557	NY02131046	679.09	ABSENT	639.78	39.3
558	NY02081045	676.05	654.71	634.59	20.1
559	NY02221017	716.28	540.72	477.01	63.7
561	NY02451004	722.38	521.21	459.94	61.3
562	NY02581008	727.86	513.59	438.91	74.7
565	NY02581006	727.86	523.34	467.56	55.8
570	NY02571013	729.08	533.40	494.99	38.4
571	NY02501010	723.29	533.10	470.92	62.2

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
575	NY02591019	730.61	591.31	568.15	23.2
577	NY02601021	730.91	668.73	594.06	74.7
578	NY02521035	711.40	671.17	629.72	41.5
579	NY02551037	708.66	677.27	638.25	39.0
581	NY02281014	719.02	533.10	465.43	67.7
582	NY02151018	712.62	540.11	478.84	61.3
583	NY01951039	673.30	615.39F	ABSENT	0.0F
584	NY01861040	665.07	643.13	609.90	33.2
585	NY01851041	664.46	645.57	615.09	30.4
586	NY01931044	667.51	650.75	633.98	16.8
588	NY02541030	719.63	659.28	610.51	48.8
589	NY02561025	726.64	649.83	597.71	52.1
591	NY02591030	724.81	659.89	609.60	50.3
592	NY02671033	729.99	666.90	616.61	50.3
593	NY02691023	738.84	646.79	602.89	43.9
594	NY02661020	734.57	639.47	591.31	48.2
595	NY02001024	696.16	552.30	492.25	60.0
596	NY01991016	698.30	542.24	485.24	57.0
599	NY01761029	664.77	560.22	491.64	68.6
600	NY01561050	662.94	593.45	560.53	32.9
601	NY01581050	662.94	598.32	591.01	7.3
602	NY01601050	663.55	621.79	601.98	19.8
603	NY01591052	663.55	630.94	617.22	13.7
605	NY01761046	660.81	ABSENT	637.03	23.8
606	NY01791030	662.94	524.26	494.08	30.2
607	NY01661014	671.78	546.51	491.34	55.2
608	NY01601007	673.00	580.95	509.93	71.0
609	NY01701040	658.37	569.37	507.49	61.9
610	NY02121007	712.93	526.39	459.94	66.4
611	NY01671038	656.84	569.67	514.20	55.5
612	NY02151002	713.23	519.38	453.24	66.1
613	NY02091003	710.79	520.90	456.29	64.6
614	NY02030997	705.00	513.89	443.80	70.1
615	NY02140993	712.93	508.71	440.44	68.3
616	NY01870999	693.12	518.16	452.32	65.8
617	NY01980988	706.53	527.30	449.88	77.4

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
618	NY01780991	689.15	569.06	500.79	68.3
619	NY01880981	700.13	541.63	476.10	65.5
620	NY01801003	684.89	522.43	486.16	36.3
621	NY01731004	678.48	557.17	508.71	48.5
622	NY01781010	681.23	534.92	465.73	69.2
623	NY01540987	672.38	551.08	483.72	67.4
624	NY01400981	660.81	541.93	479.45	62.5
625	NY01360961	659.28	518.77	452.02	66.8
626	NY01390942	675.74	499.57	430.99	68.6
627	NY01220928	667.21	467.87	400.51	67.4
628	NY01770983	691.59	562.66	493.17	69.5
629	NY01700987	685.80	564.79	495.60	69.2
630	NY01741028	664.46	561.14	492.56	68.6
633	NY01691032	658.67	556.87	505.79	51.1
634	NY01681025	661.42	558.70	488.29	70.4
635	NY02821032	744.02	664.46	611.43	53.0
637	NY02901027	749.50	644.35	588.59	55.8
638	NY01140907	669.65	405.08	333.76	71.3
639	NY01340885	688.24	415.14	403.25	11.9
640	NY01060916	669.65	424.28	358.44	65.8
641	NY01050914	669.34	433.12	361.49	71.6
642	NY00970978	667.51	452.63	439.82	12.8
643	NY00650962	655.32	375.21	359.34	15.9
644	NX99810928	670.56	281.33	279.50	1.8
645	NY00650927	640.08	413.92	366.98	46.9
647	NY02470941				
651	NX99731066	713.23	430.07	428.24	1.8
652	NX98801107	697.69	388.92	384.35	4.6
653	NX99121113	716.58	427.02	417.58	9.4
654	NX99441125	722.68	430.68	427.63	3.1
655	NX99521095	722.07	425.20	414.22	11.0
657	NX99021142	720.85	451.10	441.96	9.1
658	NX98741135	703.48	336.80	332.23	4.6
659	NX98721164	705.92	398.37	395.94	2.4
660	NX98961173	722.07	456.59	448.67	7.9
661	NX99231182	727.25	511.15	506.58	4.6

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
662		729.69	519.68	514.81	4.9
664	NX99251087	712.62	406.60	401.42	5.2
665	NX98941076	703.48	365.15	352.35	12.8
666	NX99051046	692.81	368.20	365.76	2.4
667	NY00261015	667.82	432.51	406.91	25.6
669	NX99961025	679.70	426.72	422.76	4.0
670	NX99840997	671.47	417.88	411.18	6.7
671	NX99691032	697.99	441.55	440.13	1.2
672	NX98411125	706.53	395.63	395.02	0.6
676	NX98721224	714.76	495.30	482.50	12.8
677	NX99591015	693.72	380.39	376.12	4.3
679	NY01101017	644.96	486.77	459.03	27.7
680	NX99760967	672.39	388.92	383.74	5.2
681	NX99710942	669.04	304.50	303.28	1.2
696	NY00760833	642.82	361.19	295.66	65.5
697	NY00530820				
698	NY00280845				
702	NX99111163	727.56	471.22	467.26	4.0
703	NX98921126				
704	NX99291198	727.86	472.14	469.39	2.7
705	NX98851157				
706	NX99181130	721.16	445.92	430.99	14.9
707	NX99041097			350.82	
708	NX99141077			356.01F	
738		743.71	572.11	535.23	36.0
739	NY05760730	694.94	636.42	600.15	36.3
740	NY05800714	694.94	659.59	636.12F	23.5F
756	NY00751016	663.85	492.25	410.57	81.7
757	NY00551022	665.68	463.30	397.46	65.8
758	NY00751039	665.99	468.48	420.62	47.9
765	NY00461135	668.12	664.46	629.11	35.4
766		667.51	ABSENT	626.06	41.5
768	NY00501124	666.60	662.33	604.42	57.9
769	NY00481140	669.04	666.29	662.64	3.7
770	NY00611126	665.68	662.64	650.14	12.5
771	NY00611109	662.64	654.10	602.28	51.8

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
772	NY00251131	679.70	648.92	631.85	17.1
774	NY00201098	682.75	495.00	484.02	11.0
775	NY00281142	680.62	665.38	640.70	14.7
776	NY00301147	680.62	658.37	638.25	20.1
777	NY00141128	689.46	642.21	632.46	9.8
778	NY00101121	701.04	632.76		
779	NY00121118	702.26	523.34	513.28	10.1
780	NX99641152	707.14	473.96	442.87	31.1
781	NY00231111	687.63	627.28	598.32	29.0
782	NY00181120	694.94	642.52	625.45	17.1
783	NX99991107	694.33	480.06	468.48	11.6
784	NY00011135	701.04	627.89	625.45	2.4
785	NY00231105	684.28	541.93	509.02	32.9
786	NY00031086	687.93	458.42	448.36	10.1
787	NY00061012	686.41	426.42	416.05	10.4
788	NY00281047	676.05	433.43	409.96	23.5
789	NX99831080	709.27	420.32	418.80	1.5
790	NX99821102	706.93	439.52	434.95	4.6
791	NX99701101	717.50	420.32	416.66	3.7
792	NY00531046	670.86	445.01	420.01	25.0
793	NY00611047	671.47	453.85	422.76	31.1
794	NY00611052	671.17	490.73	458.72F	32.0F
795	NY00401050	677.57	424.89	407.82	17.1
796	NY00121096	694.03	499.26	488.59	10.7
797	NY00061105	699.21	497.13	488.90	8.2
799	NX99791123	707.14	420.01	413.92	6.1
800	NY00031071	690.37	420.93	409.35	11.6
801	NY00991033	646.79	513.59	483.11	30.5
802	NY00661031	666.90	441.66	411.18	30.5
803	NY01331041	648.61	561.14	487.68	73.5
804	NY00881019	660.50	474.57	421.23	53.3
811	NY02940751	658.06	289.86	197.51	92.4
812	NY03330837	693.12	424.89	369.11	55.8
813	NY03200792	674.52	312.72	224.03	88.7
814	NY04520733	690.98	475.49	394.72	80.8
815	NY03290823	690.37	325.53	250.85	74.7

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
816	NY04720766	698.30	537.06	453.24	83.8
817	NY03240816	685.80	316.99	222.81	94.2
818	NY03850716	679.70	270.66F	ABSENT	0.0F
819		670.86	210.62	89.61	121.1
820	NY03210822	686.10	327.96	226.77	101.2
821	NY03540733	676.05	269.75	177.09F	92.7F
822	NY03650803	705.92	376.73	322.48	54.3
823	NY03460733	676.66	275.84	190.50	85.3
824	NY03930774	693.42	378.87	326.14	52.7
825	NY03770730	678.79	270.36	233.48F	36.9F
826	NY03820762	686.71	377.95	291.69	86.3
827	NY04240693	676.35	327.05	211.53	115.5
828	NY03750755	680.92	332.23	273.71F	58.5F
829	NY03820769	689.46	382.83	298.70	84.1
830	NY03760775	695.86	359.97	293.22	66.8F
831	NY03920758	687.63	370.03	281.03F	89.0F
832	NY03880765	689.15	379.48	296.27	83.2
833	NY03760770	690.07	366.67	292.91	73.8
834	NY03710771	692.20	339.85	247.80	92.0
835	NY03740766	688.85	356.62	275.84	80.8
841	NY04420580	678.48	339.24	226.16	113.1
842	NY04600569	689.15	370.33	263.65	106.7
843	NY04500583	680.31	377.95	268.22	109.7
844	NY04470562	679.09	318.82	243.84	75.0
845	NY05690540	744.32	398.98	301.45	97.5
846	NY04250544	673.61	288.04	169.47	118.6
847	NY04380595	683.97	377.04	278.28	98.8
848	NY05100609	700.74	336.50	236.83	99.7
849	NY04400557	675.13	343.20	238.35	104.9
850	NY04920595	694.03	311.81	194.77	117.0
851	NY04330550	673.30	322.48	215.49	107.0
852	NY05430636	706.53	428.85	340.46	88.4
853	NY05290625	706.83	388.62	284.99	103.6
854	NY04770583	692.20	224.03	192.02F	32.0F
908		658.37	646.79	625.45	21.3
909	NY01161063	658.37	608.08	558.39	49.7

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
912	NY01321065	658.98	616.00	575.46	40.5
914	NY01471046	652.27	597.71	536.14	61.6
915		661.42	605.03	584.61	20.4
916		658.37	637.95	582.17	55.8
917	NY01391067	655.32	ABSENT	638.56	16.8
918	NY01441055	655.32	ABSENT	593.14	62.2
919	NY01331063	658.37	ABSENT	592.53	65.8
920		661.42	644.35	580.64	63.7
922	NY01481076				
923	NY01501078				
924	NY01531071				
925	NY01531075				
926	NY01531077				
931		653.80	638.25	582.47	55.8
932		653.80			
933		653.80	618.44	586.13	32.3
934		653.80	626.36	583.99	42.4
937					28.3
938					19.5
939					56.4
940					50.9
941		673.61	477.62	430.07	47.5
942		670.56	485.24	429.77	55.5
943		678.18	516.94	430.38	86.6
944		679.70	478.54	436.47	42.1
945	NY01321066	658.98	636.42	612.34	24.1
946	NY01281058	658.37	638.56	611.73	26.8
947	NY01291041	649.22	560.22	490.42	69.8
948		652.27			
949	NY01251052	655.32	567.54		
950	NY01351051	650.75	603.20	561.14	42.1
951		650.75	562.66	490.73	71.9
952	NY01311051	652.27	601.98	515.42	86.6
953	NY01451035	652.27	600.15	505.97	94.2
954	NY01561035	653.80	586.44	511.15	75.3
955		664.46	633.98	602.28	31.7

BSC b'h.no.	Location	Surface Level	Base of St. Bees	Base of breccia	Thickness breccia
957		678.18	507.80	447.14	60.7

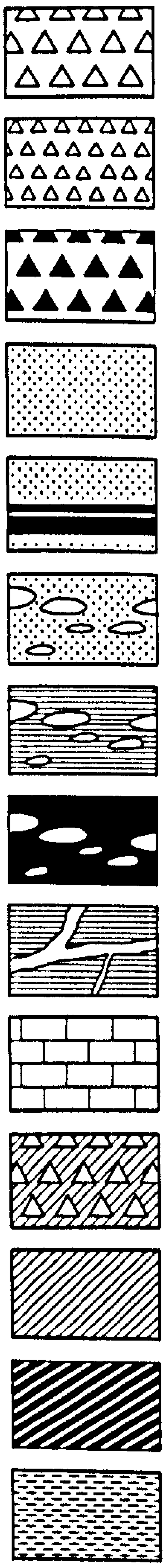
APPENDIX TWO

Borehole logs for west Cumbria

Explanation to the symbols used in the representation of logged Permo-Triassic sequences from west Cumbria. The St. Bees Series is shown in light stipple where undifferentiated. Heights refer to mining datum, 609.6 metres below ordnance datum.

Facies logs for SBH 297 (NY01841040), SBH 298 (NY01841034), SBH 301 (NY01551047), SBH 302 (NY01581024), SBH 306 (NY01940951) and SBH 309 (NY00131014).

Explanation to the text-figure symbols



- Coarse conglomerate / breccia
- Fine conglomerate / breccia
- Breccia with high authigenic clay matrix content
- Sandstone / siltstone
- Sandstone with clay laminations
- Sandstone with carbonate concretions
- Laminated siltstones, marls, sandy silty mudstones etc. with carbonate concretions
- Mudstone with carbonate concretions
- Laminated siltstones with gypsum veins
- Magnesian Limestone
- Haematitic breccia
- Haematitic sandstone
- Haematite
- Collapse breccia

Levels refer to Mining Datum, 609·6 metres below Ordnance Datum


SBH 297
St. Bees Shale 582.9m.
590
580
570
560
550
540
530
Core not recovered
Faulted contact with Basement Beds and Skiddaw Slates 533.4m.
SBH 298
St. Bees Sandstone 579.1m.
580
570
560
550
540
530
520
Seventh Limestone 520.9m.
SBH 301
St. Bees Shale 590.8m.
590
580
570
560
550
540
530
554.5m. to 590.8m. not logged
Seventh Lstn. 530.4m.

SBH 302
St. Bees Shale 553.1m.
560
550
540
530
520
510
500
490
Fifth Limestone 493.7m.
SBH 306
St. Bees Shale 542.6m.
550
540
530
520
510
500
490
480
470
Limestone 473.8m.
SBH 309
St. Bees Sandstone 503.0m.
500
490
480
470
460
450
440
430
420
Shale 434.4m.
Evaporites 425.6m.
Limestone 423.2m.

APPENDIX THREE

Modal analysis data

Modal (percentage composition) analyses of brockram core samples from west Cumbria (standard count: 1000 points/petrographic thin section). The first three characters of the core number refer to the British Steel Corporation Surface Borehole (SBH) reference number and the following characters to the depth (in metres) from surface and sample number respectively. Borehole surface levels (mining datum) are reproduced in Appendix One.

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
296-100.58/4	39.0	0.0	17.3	19.8	1.3	14.6	8.0	0.0
296-100.58/3	52.9	0.0	27.3	3.4	0.3	3.9	12.2	0.0
296-100.58/2	38.8	0.0	20.6	16.6	0.8	16.9	6.3	0.0
296-100.58/1	50.3	1.4	28.7	4.2	0.9	5.2	9.3	0.0
296-114.00/5	20.2	0.9	7.4	32.0	6.6	11.8	21.1	0.0
296-114.00/4	20.0	0.0	14.4	39.6	3.1	9.7	13.2	0.0
296-114.00/3	56.9	0.0	0.8	24.0	3.0	6.0	9.3	0.0
296-114.00/2	47.7	0.5	0.0	16.7	10.8	6.3	18.0	0.0
296-114.00/1	48.0	2.8	0.7	26.1	6.9	7.4	7.9	0.0
296-129.24/5	41.2	0.0	11.8	28.8	9.5	13.0	1.2	0.0
296-129.24/4	31.1	0.0	44.3	10.8	1.6	4.4	7.7	0.0
296-129.24/3	52.4	0.8	8.3	21.0	2.7	9.5	5.3	0.0
296-129.24/2	34.8	0.0	1.0	24.9	4.3	21.1	13.8	0.1
296-129.24/1	49.4	3.0	8.3	23.1	4.4	8.0	3.8	0.0
296-132.89/4	18.3	0.1	44.1	14.3	4.6	12.7	5.7	0.1
296-132.89/3	41.8	0.8	4.5	27.4	7.3	9.1	8.0	1.0
296-132.89/2	46.1	0.0	6.8	24.4	3.1	11.8	7.2	0.5
296-132.89/1	36.8	0.4	32.1	10.5	3.0	13.9	2.9	0.2
296-133.50/2	36.5	6.8	1.4	26.0	3.5	12.6	12.7	0.5
296-133.50/1	23.2	15.8	11.3	20.1	2.1	17.5	10.0	0.0
296-138.68/4	44.5	0.0	8.9	17.6	10.4	16.1	2.2	0.1
296-138.68/3	14.9	0.0	0.7	30.9	12.1	17.6	23.7	0.1
296-138.68/2	19.3	17.0	8.9	19.8	10.3	13.4	11.3	0.2
296-138.68/1	28.9	0.0	3.2	28.0	3.4	17.1	19.4	0.0
296-142.04/10	47.9	0.0	5.3	29.8	3.0	11.6	1.6	0.8
296-142.04/9	43.7	0.0	0.0	6.9	0.4	15.2	31.4	2.3
296-142.04/8	0.7	0.0	0.7	64.5	5.9	22.3	5.6	0.0
296-142.04/7	23.1	0.0	0.0	43.5	4.6	24.1	4.6	0.0
296-142.04/6	4.7	0.0	0.0	54.6	11.7	18.0	11.0	0.0
296-142.04/5	70.5	0.0	2.2	9.5	2.4	6.7	7.5	1.2
296-142.04/4	57.2	0.0	0.0	4.8	4.5	20.5	5.8	7.1
296-142.04/3	55.2	0.0	21.4	8.7	3.9	6.9	3.7	0.1
296-142.04/2	44.9	0.0	13.0	4.0	3.9	15.3	11.6	7.9
296-142.04/1	61.0	0.0	2.3	15.7	2.5	8.1	10.4	0.0
296-142.65/2	29.9	0.0	1.8	28.4	2.0	17.9	19.9	0.0
296-142.65/1	28.5	0.0	44.7	10.6	2.0	9.3	4.9	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
296-144.17/3	30.7	1.1	2.2	29.6	8.8	25.5	1.9	0.0
296-144.17/2	90.3	0.0	0.0	4.5	1.1	4.0	0.0	0.0
296-144.17/1	62.8	0.3	0.0	16.6	5.5	12.8	1.7	0.0
298-89.31/3	42.1	0.0	12.0	29.3	0.7	13.1	2.8	0.0
298-89.31/2	52.9	0.0	0.4	20.3	1.0	22.0	3.4	0.0
298-89.31/1	56.7	0.6	0.6	22.6	1.8	14.2	3.4	0.0
298-98.45/5	21.2	0.2	31.1	19.5	4.1	15.7	7.2	0.0
296-98.45/4	61.1	0.0	0.7	13.6	2.7	15.6	4.4	0.0
298-98.45/3	39.6	0.0	27.3	14.0	2.9	12.4	3.3	0.0
298-98.45/2	40.3	0.0	5.3	21.2	5.2	22.4	3.7	0.0
298-98.45/1	53.4	0.0	7.8	22.3	0.7	9.8	5.8	0.0
298-105.16/3	39.9	0.0	5.5	25.4	1.6	20.7	6.9	0.0
298-105.16/2	20.2	0.0	4.3	12.6	0.4	9.7	2.8	0.0
298-105.16/1	47.3	0.0	13.3	8.4	1.0	16.0	13.3	0.0
298-114.00/5	10.6	8.7	24.8	33.7	3.7	9.0	9.5	0.0
298-114.00/4	52.7	1.9	2.4	17.8	3.2	16.6	5.4	0.0
298-114.00/3	50.0	0.0	9.5	15.7	4.6	10.9	9.3	0.0
298-114.00/2	51.1	2.4	23.1	2.8	0.1	4.3	16.2	0.0
298-114.00/1	22.1	0.9	20.4	22.8	5.1	23.5	5.2	0.0
298-117.04/5	52.8	0.0	8.7	17.1	3.2	13.1	5.1	0.0
298-117.04/4	54.1	0.0	0.9	20.9	2.7	10.7	10.7	0.0
298-117.04/3	37.6	0.8	1.8	31.2	4.0	18.2	6.4	0.0
298-117.04/2	37.5	0.0	44.2	3.9	0.8	9.7	3.9	0.0
298-117.04/1	50.1	0.0	7.1	15.6	1.1	10.7	15.4	0.0
298-117.96/3	66.5	2.5	0.3	6.2	2.0	18.3	4.1	0.0
298-117.96/2	54.0	5.8	0.0	10.8	5.4	18.0	6.0	0.0
298-117.96/1	47.6	2.8	0.0	13.2	10.0	21.2	5.2	0.0
298-122.83/6	1.8	0.0	0.0	40.0	7.6	32.0	18.6	0.0
298-122.83/5	7.1	0.0	14.1	24.5	1.5	35.5	17.1	0.0
298-122.83/4	24.7	0.0	0.5	17.8	1.8	43.2	11.9	0.0
298-122.83/3	12.9	0.0	1.2	25.2	3.5	42.5	14.8	0.0
298-122.83/2	24.9	0.3	0.2	18.1	3.0	40.4	13.0	0.0
298-122.83/1	22.5	0.0	0.0	11.3	0.7	44.8	20.7	0.0
298-123.44/2	8.0	0.0	11.6	26.8	30.2	23.0	0.4	0.0
298-123.44/1	32.4	0.0	0.0	34.2	13.0	19.8	0.6	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
298-124.66/2	32.3	0.0	0.0	39.3	6.7	19.7	2.0	0.0
298-124.66/1	53.9	0.4	0.7	19.7	1.7	14.7	8.9	0.0
298-127.41/2	34.2	2.7	18.9	17.4	3.8	22.2	0.4	0.3
298-127.41/1	56.0	0.0	1.7	11.8	4.3	23.3	2.8	0.0
298-128.02/3	50.0	1.2	0.3	18.4	3.6	24.1	2.2	0.2
298-128.02/2	45.0	3.8	0.0	10.2	5.1	30.0	5.9	0.0
298-128.02/1	39.6	0.0	32.4	7.4	1.6	17.6	1.4	0.0
298-128.63/3	34.4	1.5	1.9	22.9	3.4	22.0	13.8	0.0
298-128.63/2	24.1	0.0	0.0	34.2	8.4	31.6	1.3	0.0
298-128.63/1	47.6	0.0	17.8	11.0	2.6	18.2	2.8	0.0
298-129.84/3	47.8	0.6	0.4	19.2	4.7	22.9	4.4	0.0
298-129.84/2	34.0	0.0	3.6	32.0	1.6	27.2	1.6	0.0
298-129.84/1	30.7	32.2	1.1	11.3	0.1	17.9	6.7	0.0
298-135.64/2	51.2	1.4	23.9	10.4	0.4	9.6	2.8	0.2
298-135.64/1	23.4	0.6	35.5	18.6	1.0	16.6	2.9	1.4
298-138.68/5	50.8	0.7	2.6	20.3	1.7	15.2	8.5	0.3
298-138.68/4	39.6	1.3	2.1	26.1	1.1	17.1	12.7	0.0
298-138.68/3	57.8	1.5	7.0	10.0	0.3	6.7	15.9	0.8
298-138.68/2	23.9	0.9	4.7	40.0	0.7	19.5	10.3	0.0
298-138.68/1	31.5	4.3	12.3	9.8	0.2	11.9	28.4	1.6
298-141.73/4	36.1	10.8	0.0	15.6	1.5	31.4	3.1	1.5
298-141.73/3	60.8	6.3	0.6	10.2	0.4	16.9	3.7	1.1
298-141.73/2	42.6	12.6	0.6	16.4	1.0	22.8	2.6	1.2
298-141.73/1	38.9	0.0	6.3	1.6	0.0	10.9	32.3	10.0
298-142.65/3	50.5	5.2	4.3	15.8	0.7	18.6	2.7	2.1
298-142.65/2	50.1	7.7	0.0	13.8	4.2	21.7	2.0	0.3
298-142.65/1	63.3	1.6	1.8	9.7	0.5	14.1	9.0	0.0
301-106.68/1	26.1	0.0	1.4	31.4	5.6	34.6	0.9	0.0
301-112.47/1	43.7	0.0	2.8	17.7	0.9	28.6	6.1	0.0
301-113.69/2	24.8	0.0	8.2	31.2	6.6	22.6	5.6	0.0
301-113.69/1	17.2	0.0	0.5	24.6	9.8	38.6	4.5	4.7
301-114.30/1	29.1	0.0	2.6	37.2	0.4	19.3	11.3	0.0
301-114.60/3	46.0	5.0	6.8	19.9	0.5	11.7	10.0	0.1
301-114.60/2	24.8	0.0	2.4	48.2	2.4	21.6	0.6	0.0
301-114.60/1	7.4	0.0	19.8	50.2	2.4	13.6	6.8	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
301-115.21/1	41.9	0.0	4.8	27.3	2.3	21.0	2.5	0.2
301-116.74/1	24.0	2.8	0.8	39.2	3.0	26.8	3.4	0.0
301-118.26/2	58.4	1.8	0.0	18.0	1.5	14.1	5.6	0.6
301-118.26/1	30.2	13.0	8.8	5.3	0.2	10.8	30.7	1.0
301-118.57/2	38.7	3.2	1.3	22.4	0.1	11.5	22.8	0.0
301-118.57/1	50.5	1.6	15.9	9.8	0.0	4.7	17.4	0.1
301-118.87/2	34.2	0.0	24.6	16.4	1.4	23.0	0.4	0.0
301-118.87/1	23.8	0.9	1.9	32.9	2.6	31.4	6.5	0.0
301-119.79/2	29.7	0.0	0.0	20.5	4.3	40.9	4.5	0.1
301-119.79/1	41.6	0.0	3.3	24.5	2.7	24.7	3.1	0.0
301-121.92/1	55.3	0.0	3.6	12.7	0.9	20.5	7.0	0.0
301-123.44/2	42.0	0.0	0.0	30.6	0.6	25.4	0.6	0.8
301-123.44/1	40.5	0.0	2.4	17.3	2.5	31.7	4.3	1.3
302-112.78/1	29.1	0.0	2.0	47.0	1.3	12.7	7.9	0.0
302-119.18/3	0.0	0.0	0.0	25.2	18.0	45.6	11.2	0.0
302-119.18/2	0.0	0.0	18.0	26.4	11.6	40.4	2.8	0.0
302-119.18/1	0.0	0.0	0.0	43.2	10.4	39.6	6.8	0.0
302-123.44/3	28.2	5.5	26.6	18.6	0.8	7.1	13.2	0.0
302-123.44/2	21.2	4.8	15.0	31.6	0.6	16.8	10.0	0.0
302-123.44/1	33.3	1.9	29.5	17.2	0.8	11.0	6.2	0.0
302-126.19/1	1.2	0.0	0.0	49.6	9.2	35.2	4.8	0.0
302-127.41/1	1.2	0.0	0.0	18.4	9.2	70.4	0.8	0.0
302-127.71/2	0.0	0.0	0.0	33.0	7.0	60.0	0.0	0.0
302-127.71/1	6.0	0.0	0.0	41.8	5.8	46.2	0.2	0.0
302-128.63/1	3.2	0.0	0.0	30.4	3.0	61.2	2.0	0.0
302-131.37/3	42.3	3.8	0.0	26.3	0.8	11.6	15.1	0.0
302-131.37/2	32.1	3.6	15.3	27.3	0.2	8.3	13.1	0.0
302-131.37/1	34.4	2.1	6.0	27.1	1.5	18.6	10.3	0.0
302-134.42/2	21.8	0.0	2.2	48.6	0.7	16.6	10.2	0.0
302-134.42/1	3.4	0.0	0.0	17.4	19.2	60.0	0.0	0.0
302-135.94/2	44.7	0.0	2.8	27.7	1.8	21.8	1.2	0.0
302-135.94/1	67.0	0.0	0.0	7.9	9.2	15.1	0.1	0.0
302-136.86/1	2.1	0.0	0.0	21.2	11.9	64.9	0.0	0.0
302-138.68/2	36.3	2.1	6.0	21.1	1.0	7.6	25.9	0.0
302-138.68/1	27.7	0.2	0.6	37.6	1.3	18.9	13.6	0.1

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
302-138.99/2	46.8	0.0	7.6	12.6	3.6	22.8	6.6	0.0
302-138.99/1	17.8	3.3	4.4	23.8	2.3	14.8	15.1	18.5
302-140.82/3	3.0	0.0	0.0	40.2	16.0	40.2	0.6	0.0
302-140.82/2	3.6	0.0	0.0	39.4	15.3	39.4	2.2	0.0
302-140.82/1	26.1	29.3	0.0	14.4	0.2	11.8	17.9	0.3
302-142.04/1	0.0	0.0	0.0	36.5	14.5	42.0	7.0	0.0
302-142.65/3	4.6	0.0	5.4	53.2	1.3	28.0	7.5	0.0
302-142.65/2	0.6	0.0	0.6	36.0	8.0	53.4	1.4	0.0
302-142.65/1	21.2	0.0	0.8	42.6	1.6	31.8	2.0	0.0
302-143.56/1	27.6	0.0	2.3	18.7	7.0	40.9	3.5	0.0
302-144.48/1	2.0	0.0	0.0	36.2	8.2	7.8	46.0	0.0
302-146.91/2	57.5	1.7	0.2	23.9	0.6	6.5	9.6	0.0
302-146.91/1	31.5	37.9	0.0	14.8	0.2	2.9	11.1	0.0
302-147.83/3	46.7	0.0	1.0	23.4	0.8	8.1	20.1	0.0
302-147.83/2	40.0	0.6	0.0	40.6	1.0	6.8	10.8	0.0
302-147.83/1	25.8	19.0	0.0	36.6	0.8	12.0	6.0	0.0
302-151.18/2	18.8	0.0	0.4	50.4	1.0	27.6	1.8	0.0
302-151.18/1	54.0	0.0	0.0	27.2	0.7	13.2	4.9	0.0
302-151.49/3	28.0	0.0	1.4	32.4	0.9	24.6	12.5	0.3
302-151.49/2	23.4	0.0	0.0	58.6	1.4	13.4	3.2	0.0
302-151.49/1	19.4	0.0	6.6	55.0	0.5	15.7	2.6	0.1
302-153.31/3	44.7	0.0	0.0	2.4	0.1	39.8	12.9	0.1
302-153.31/2	44.7	1.1	1.6	17.8	0.5	27.7	6.6	0.0
302-153.31/1	31.1	0.0	2.2	43.0	0.8	21.8	1.1	0.0
302-154.53/2	50.0	1.7	1.5	24.7	0.1	14.9	7.0	0.0
302-154.53/1	60.4	0.0	10.4	9.6	0.8	8.2	10.6	0.0
302-156.67/1	45.6	0.3	0.0	18.1	0.8	26.4	8.0	0.8
302-156.97/1	52.0	0.0	23.2	3.6	0.2	6.4	14.6	0.0
302-157.28/2	26.2	9.9	0.0	24.6	1.1	24.4	13.7	0.1
302-157.28/1	25.9	5.3	0.4	35.4	0.2	12.9	19.9	0.0
302-158.80/1	44.3	4.4	0.9	26.1	0.3	12.6	11.4	0.0
302-159.41/1	17.6	0.0	41.2	25.4	0.6	10.0	4.8	0.0
302-160.02/1	32.7	31.0	0.0	8.9	1.3	6.8	19.2	0.0
302-160.93/2	27.2	0.0	0.0	33.0	6.6	31.0	2.2	0.0
302-160.93/1	23.4	4.8	25.9	22.8	2.5	18.6	2.0	0.0
302-162.76/1	31.0	0.0	0.0	33.4	6.6	26.4	2.4	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
304-165.66/2	0.0	0.0	0.0	28.4	47.5	17.9	6.2	0.0
304-165.66/1	2.4	0.0	1.6	22.4	43.9	22.7	7.1	0.0
304-165.76/1	0.0	0.0	0.0	20.1	36.3	21.1	21.5	0.0
304-165.86/1	0.0	0.0	0.0	14.0	54.4	29.6	1.6	0.0
304-165.96/2	0.0	0.0	0.0	30.6	36.4	26.0	7.0	0.0
304-165.96/1	0.0	0.0	0.0	26.0	43.6	22.8	7.6	0.0
304-166.04/1	0.0	0.0	0.0	31.0	36.0	19.4	13.6	0.0
304-166.12/2	0.0	0.0	0.0	27.4	45.2	23.7	3.7	0.0
304-166.12/1	0.0	0.0	0.0	15.6	50.4	25.6	8.4	0.0
304-166.22/2	0.0	0.0	0.0	32.8	34.4	26.8	6.0	0.0
304-166.22/1	0.4	0.0	0.0	29.2	32.4	33.6	4.4	0.0
304-166.32/2	0.0	0.0	0.0	33.6	30.8	32.4	3.2	0.0
304-166.32/1	0.0	0.0	0.0	14.4	41.2	21.6	22.8	0.0
304-166.42/4	0.0	0.0	0.0	20.4	43.2	24.8	11.6	0.0
304-166.42/3	0.0	0.0	0.0	28.0	30.4	32.0	9.2	0.0
304-166.42/2	0.8	0.0	0.0	28.8	38.0	27.2	5.2	0.0
304-166.42/1	0.0	0.0	0.0	21.2	40.4	35.2	3.2	0.0
304-166.73/3	4.0	0.0	1.2	40.4	22.8	21.6	10.0	0.0
304-166.73/2	0.0	0.0	0.0	18.4	50.0	26.4	4.8	0.0
304-166.73/1	0.0	0.0	0.0	19.2	47.6	23.6	9.6	0.0
304-166.83/1	0.0	0.0	0.0	32.0	27.6	28.4	12.0	0.0
304-166.88/2	0.0	0.0	0.0	21.6	38.4	30.8	9.2	0.0
304-166.88/1	0.0	0.0	0.0	20.8	37.2	25.2	16.8	0.0
304-167.03/6	0.0	0.0	0.0	24.0	48.8	18.4	8.8	0.0
304-167.03/5	0.0	0.0	0.0	18.8	46.8	28.0	6.4	0.0
304-167.03/4	0.0	0.0	0.0	19.2	44.8	23.2	12.8	0.0
304-167.03/3	0.0	0.0	0.0	16.8	48.0	24.4	10.8	0.0
304-167.03/2	0.0	0.0	0.0	5.6	59.6	28.4	6.4	0.0
304-167.03/1	0.0	0.0	0.0	16.4	45.2	14.0	23.2	0.0
304-167.44/4	0.0	0.0	0.0	9.6	60.0	18.8	11.6	0.0
304-167.44/3	2.0	0.0	0.0	46.8	20.0	24.0	7.2	0.0
304-167.44/2	7.6	0.0	2.4	33.6	28.4	16.8	11.2	0.0
304-167.44/1	7.6	0.0	0.0	36.0	24.4	20.4	11.6	0.0
304-167.87/7	17.3	0.0	2.6	32.7	39.1	13.9	6.4	0.0
304-167.87/6	11.3	0.4	2.8	59.4	6.4	9.9	11.3	0.0
304-167.87/5	3.6	0.0	1.6	29.6	37.6	15.6	12.0	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
304-167.87/4	22.4	0.0	0.8	48.8	9.2	11.2	7.6	0.0
304-167.87/3	8.0	0.0	0.0	60.0	7.6	14.8	9.6	0.0
304-167.87/2	12.1	0.0	0.0	39.9	23.2	16.0	8.8	0.0
304-167.87/1	51.5	0.0	2.3	29.1	1.0	7.4	8.7	0.0
304-173.28/2	0.7	0.0	0.0	26.9	38.9	30.0	3.5	0.0
304-173.28/1	0.0	0.0	0.0	29.6	41.2	25.2	4.0	0.0
304-173.43/3	0.0	0.0	0.0	26.8	40.8	28.4	4.0	0.0
304-173.43/2	2.0	0.0	0.0	41.6	30.4	23.6	2.4	0.0
304-173.43/1	1.5	0.0	0.0	32.4	37.4	26.7	1.9	0.0
304-173.74/3	2.8	0.0	0.0	46.4	26.0	16.4	8.4	0.0
304-173.74/2	0.0	0.0	0.0	31.0	32.1	27.4	9.5	0.0
304-173.74/1	0.0	0.0	0.0	34.0	47.0	16.6	2.4	0.0
304-174.02/3	0.4	0.0	0.0	33.6	35.6	29.2	1.2	0.0
304-174.02/2	2.0	0.0	0.0	33.6	39.1	17.8	7.5	0.0
304-174.02/1	0.0	0.0	0.0	33.2	35.6	26.8	4.4	0.0
304-174.19/6	0.4	0.0	0.0	33.6	38.0	21.2	6.8	0.0
304-174.19/5	11.2	0.0	0.0	30.8	30.0	26.0	2.0	0.0
304-174.19/4	27.6	0.0	0.0	34.8	15.6	19.2	2.8	0.0
304-174.19/3	35.8	0.0	1.1	34.2	13.1	13.3	2.5	0.0
304-174.19/2	15.2	0.8	0.0	45.6	11.2	21.6	5.6	0.0
304-174.19/1	6.7	0.0	0.0	17.4	40.0	28.5	7.4	0.0
304-174.93/4	10.8	0.0	5.2	46.0	3.2	32.4	2.4	0.0
304-174.93/3	38.8	0.0	3.1	25.6	11.6	15.7	5.1	0.0
304-174.93/2	44.2	0.0	6.4	24.8	3.0	14.0	7.8	0.0
304-174.93/1	0.0	0.0	0.0	15.6	38.8	41.6	4.0	0.0
304-175.26/7	0.0	0.0	0.0	8.0	70.4	10.4	11.2	0.0
304-175.26/6	0.0	0.0	0.0	14.8	60.4	21.2	3.2	0.0
304-175.26/5	2.0	0.0	0.0	15.2	61.2	18.8	2.8	0.0
304-175.26/4	1.2	0.0	0.0	12.4	61.6	22.8	2.0	0.0
304-175.26/3	0.0	0.0	0.0	16.1	67.3	13.0	3.6	0.0
304-175.26/2	0.0	0.0	0.0	18.8	64.0	13.2	4.0	0.0
304-175.26/1	0.0	0.0	0.0	17.6	66.8	13.6	2.0	0.0
304-175.56/2	0.0	0.0	0.0	27.2	50.4	16.8	5.6	0.0
304-175.56/1	0.0	0.0	0.0	37.9	35.9	14.8	11.3	0.0
304-175.87/3	21.6	0.0	0.0	30.0	26.0	16.0	6.4	0.0
304-175.87/2	15.6	0.0	0.0	39.6	20.0	24.0	0.4	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
304-175.87/1	36.0	0.0	0.0	38.8	0.4	8.4	14.8	1.6
304-177.70/2	4.4	0.0	0.0	41.6	33.6	19.6	0.8	0.0
304-177.70/1	10.8	0.0	4.8	56.8	11.2	12.4	4.0	0.0
304-195.22/4	65.2	1.8	2.2	16.2	0.4	5.8	8.0	0.2
304-195.22/3	55.0	4.0	9.2	8.6	7.0	10.2	6.0	0.0
304-195.22/2	53.7	5.0	0.2	24.9	1.0	7.6	7.8	0.0
304-195.22/1	56.6	8.2	1.0	22.8	0.2	6.4	5.0	0.0
304-195.99/4	36.6	9.7	0.5	33.5	1.2	13.0	5.7	0.0
304-195.99/3	0.0	0.0	0.0	76.4	3.6	11.6	8.4	0.0
304-195.99/2	33.2	2.4	5.6	32.0	0.4	5.2	21.2	0.0
304-196.29/2	0.0	0.0	0.0	9.6	19.2	26.0	44.8	0.0
304-196.29/1	0.0	0.0	0.0	18.4	24.4	50.8	6.4	0.0
304-197.41/2	0.0	0.0	0.0	14.8	27.6	57.6	0.0	0.0
304-197.41/1	0.0	0.0	0.0	11.1	24.6	64.3	0.0	0.0
304-200.25/3	1.5	0.0	0.0	35.1	13.7	49.6	0.0	0.0
304-200.25/2	56.4	0.0	0.0	25.6	0.4	9.2	8.4	0.0
304-200.25/1	24.0	0.0	2.4	52.0	1.6	9.2	10.8	0.0
304-201.17/3	23.2	0.0	0.0	40.0	2.4	24.8	9.6	0.0
304-201.17/1	0.0	0.0	0.0	14.1	46.3	39.6	0.0	0.0
304-201.78/2	0.0	0.0	0.0	26.5	42.5	31.0	0.0	0.0
304-201.78/1	43.2	0.0	8.0	27.6	1.8	13.6	5.8	0.0
304-202.01/2	70.0	0.0	0.0	18.5	0.7	8.7	2.0	0.0
304-202.01/1	61.5	0.0	0.0	16.3	0.8	17.8	3.5	0.0
304-203.30/3	42.8	0.0	10.2	13.4	2.0	17.8	12.4	1.4
304-203.30/2	0.0	0.0	0.0	9.9	16.7	73.5	0.0	0.0
304-203.53/1	0.0	0.0	0.0	11.9	63.0	24.7	0.4	0.0
304-203.66/3	0.8	0.0	0.0	22.8	14.8	57.2	4.4	0.0
304-203.66/2	6.0	0.0	1.6	42.8	9.6	39.2	0.8	0.0
304-203.66/1	0.0	0.0	0.0	2.8	15.5	48.8	28.6	4.2
304-204.44/3	0.0	0.0	0.0	45.6	33.9	19.3	1.2	0.0
304-204.44/2	0.4	0.0	0.0	49.2	24.8	24.4	1.2	0.0
304-204.44/1	24.6	0.0	0.0	47.5	3.7	23.8	0.4	0.0
304-205.59/2	29.7	0.0	0.0	47.1	3.6	11.6	3.6	0.0
304-205.59/1	42.2	0.0	0.0	35.2	2.4	14.6	5.6	0.0
304-205.74/3	0.0	0.0	0.0	0.4	37.2	14.3	48.1	0.0
304-205.74/1	0.0	0.0	0.0	1.2	31.6	39.4	28.8	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
304-206.04/4	0.0	0.0	0.0	12.0	28.8	17.6	41.6	0.0
304-206.04/3	0.0	0.0	0.0	1.2	45.2	28.4	25.2	0.0
304-206.04/2	0.0	0.0	0.0	9.2	29.6	18.4	42.8	0.0
304-206.15/2	0.8	0.0	0.0	28.4	42.4	22.8	5.6	0.0
304-206.15/1	0.0	0.0	0.0	14.4	56.0	16.8	12.8	0.0
304-206.45/3	4.4	0.0	0.0	29.2	48.8	12.4	5.2	0.0
304-206.45/2	2.4	0.0	0.0	8.0	30.8	14.4	44.4	0.0
304-206.45/1	0.0	0.0	0.0	24.4	40.8	27.2	7.6	0.0
304-206.55/1	0.0	0.0	0.0	22.4	31.2	23.2	23.2	0.0
304-207.16/2	5.6	2.0	0.8	54.8	4.8	10.8	21.2	0.0
304-207.16/1	2.0	0.0	2.8	29.6	9.2	13.2	43.2	0.0
304-207.32/4	0.0	0.0	0.0	22.0	60.0	13.6	4.4	0.0
304-207.32/3	18.0	0.8	0.0	58.4	3.2	13.6	14.8	0.0
304-207.32/2	8.8	0.0	0.0	49.6	19.2	19.2	3.2	0.0
304-207.32/1	0.0	0.0	0.0	42.0	42.4	11.6	3.6	0.0
304-207.57/5	10.9	0.0	0.0	51.2	7.2	10.0	20.0	0.0
304-207.57/4	0.0	0.0	0.0	41.6	12.0	16.3	30.0	0.0
304-207.57/3	3.6	0.0	0.0	58.8	10.4	14.8	12.4	0.0
304-207.57/2	5.2	0.0	0.0	48.0	13.2	21.2	12.4	0.0
304-207.57/1	0.0	0.0	0.0	34.4	27.6	28.4	9.6	0.0
304-207.95/3	13.6	0.0	0.0	41.6	0.8	5.6	38.4	0.0
304-207.95/2	17.2	0.0	0.0	30.5	0.4	4.8	47.1	0.0
304-207.95/1	3.6	0.0	0.0	49.3	8.7	28.7	9.5	0.0
304-208.31/2	3.2	0.0	0.0	54.2	10.8	23.6	8.2	0.0
304-208.31/1	0.0	0.0	0.0	39.9	14.7	27.2	18.0	0.0
304-208.56/3	0.0	0.0	0.0	14.1	52.3	30.8	2.8	0.0
304-208.56/2	0.4	0.0	0.0	6.0	63.2	28.4	2.0	0.0
304-208.56/1	2.0	0.0	0.0	25.2	40.0	23.6	1.2	0.0
304-208.78/2	0.0	0.0	0.0	36.1	4.4	14.7	45.2	0.0
304-208.78/1	0.0	0.0	0.0	38.8	28.5	24.3	8.4	0.0
304-209.09/3	0.0	0.0	0.0	39.6	25.6	26.8	8.0	0.0
304-209.09/2	0.0	0.0	0.0	33.6	23.2	16.0	27.2	0.0
304-209.09/1	2.8	0.0	0.0	35.2	30.4	27.6	4.0	0.0
304-209.19/2	0.0	0.0	0.0	35.1	27.9	36.9	0.0	0.0
304-209.19/1	0.0	0.0	0.0	32.6	32.0	34.2	1.2	0.0
304-209.40/2	47.0	0.0	0.0	26.0	5.0	17.8	3.2	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
304-209.40/1	39.5	8.9	0.0	23.5	0.2	11.9	15.8	0.1
304-209.70/4	33.2	3.6	0.7	35.5	0.7	17.5	18.8	0.0
304-209.70/3	40.6	0.5	0.3	29.8	0.8	15.3	12.7	0.0
304-209.70/2	28.8	30.8	0.0	17.0	0.0	8.6	14.8	0.0
304-209.70/1	32.5	15.6	0.8	23.9	1.3	17.8	8.1	0.0
304-210.62/3	31.3	2.2	2.0	29.4	0.2	10.3	24.6	0.0
304-210.62/2	32.4	6.0	0.0	31.4	0.4	9.6	20.1	0.1
304-210.62/1	49.7	7.9	0.0	9.6	0.2	11.0	21.1	0.3
304-211.02/4	58.7	3.1	0.0	16.5	0.9	10.8	9.9	0.0
304-211.02/2	70.1	0.0	0.0	14.6	0.6	7.5	7.1	0.0
304-211.02/1	29.8	0.0	0.0	36.6	1.6	24.8	7.2	0.0
304-213.06/3	2.6	0.0	0.0	28.0	9.0	23.4	37.0	0.0
304-213.06/2	3.6	0.0	0.0	41.0	13.2	26.0	16.2	0.0
304-213.06/1	0.0	0.0	0.0	18.8	18.4	22.4	40.4	0.0
304-213.66/4	56.9	0.5	0.0	19.8	1.1	14.7	7.0	0.0
304-213.66/3	42.4	0.0	2.0	31.5	2.2	17.3	4.6	0.0
304-213.66/2	46.1	0.8	0.0	20.0	0.3	12.1	20.7	0.0
304-213.66/1	37.0	0.5	0.5	22.5	0.5	9.9	29.0	0.0
304-214.27/3	1.4	0.0	0.0	37.4	6.8	15.2	39.2	0.0
304-214.27/2	2.4	0.0	0.0	44.0	19.6	31.6	2.4	0.0
304-214.27/1	1.4	0.0	0.0	31.8	11.0	21.2	34.6	0.0
304-214.48/2	0.0	0.0	0.0	34.0	34.0	30.8	0.8	0.0
304-214.48/1	2.0	0.0	0.0	25.2	19.8	25.2	27.8	0.0
304-214.78/2	0.0	0.0	0.8	35.0	22.2	25.2	16.6	0.0
304-214.78/1	3.0	0.0	0.0	59.0	7.0	21.8	9.2	0.0
304-215.19/2	41.5	0.0	0.0	29.3	1.1	9.3	17.0	1.8
304-215.19/1	61.2	0.0	0.0	23.4	0.6	9.8	4.6	0.4
304-215.80/5	56.4	0.0	0.0	19.9	0.6	9.2	13.9	0.0
304-215.80/4	72.2	0.0	0.0	6.4	0.8	11.4	7.4	1.8
304-215.80/3	36.6	0.0	0.0	32.9	0.9	10.8	18.1	0.5
304-215.80/2	60.2	0.0	1.0	19.7	2.0	11.1	5.7	0.0
304-215.80/1	59.1	1.2	0.6	18.4	1.7	16.4	2.7	0.5
304-216.41/3	74.7	0.0	0.6	9.7	0.4	8.1	6.3	0.0
304-216.41/2	47.4	0.0	0.0	31.2	3.6	16.0	1.8	0.0
304-216.41/1	32.8	43.6	0.0	9.8	1.1	6.6	4.8	0.4
304-217.02/2	67.0	0.0	0.7	12.7	0.3	13.2	5.4	0.0

Core No.	Lower Palaeozoics	Carb. Lstn.	Carb. Sstn.	Sand	Silt	Clay	$CaCO_3$	$CaSO_4$
304-217.02/1	58.0	6.2	0.0	13.9	0.4	14.9	6.6	0.0
304-218.24/2	21.6	0.0	0.0	47.9	3.0	25.7	1.6	0.0
304-218.24/1	59.7	0.0	0.0	0.4	0.0	23.5	16.3	0.0
304-219.15/2	45.0	0.0	0.9	20.2	3.3	22.0	3.0	0.0
304-219.15/1	57.8	0.0	11.7	17.7	0.8	8.9	3.1	0.0

Plate 2.1

A. Sandstone horizon with an interstitial ferroan calcite cement (IC) in which the detrital grains possess pronounced clay skins (CS). (Thin section photomicrograph)

B. Detrital quartz grain (Q), partially enclosed by a thin mechanically infiltrated clay skin (CS). An isolated clay plate (P) has been partially dislocated from the grain surface. The texture of this plate is distinct from that of the surrounding authigenic matrix clay (AC). (SEM photomicrograph)

C. Clay skin (CS) enclosing detrital silt grain. Note high percentage of interstitial void (IV) and cross-connecting clay plates (meniscal bridges?). (SEM photomicrograph)

D. Detrital silt grain with clay skin. The clay plates are aligned parallel with the surface of the grain in the upper right portion of the SEM photomicrograph.

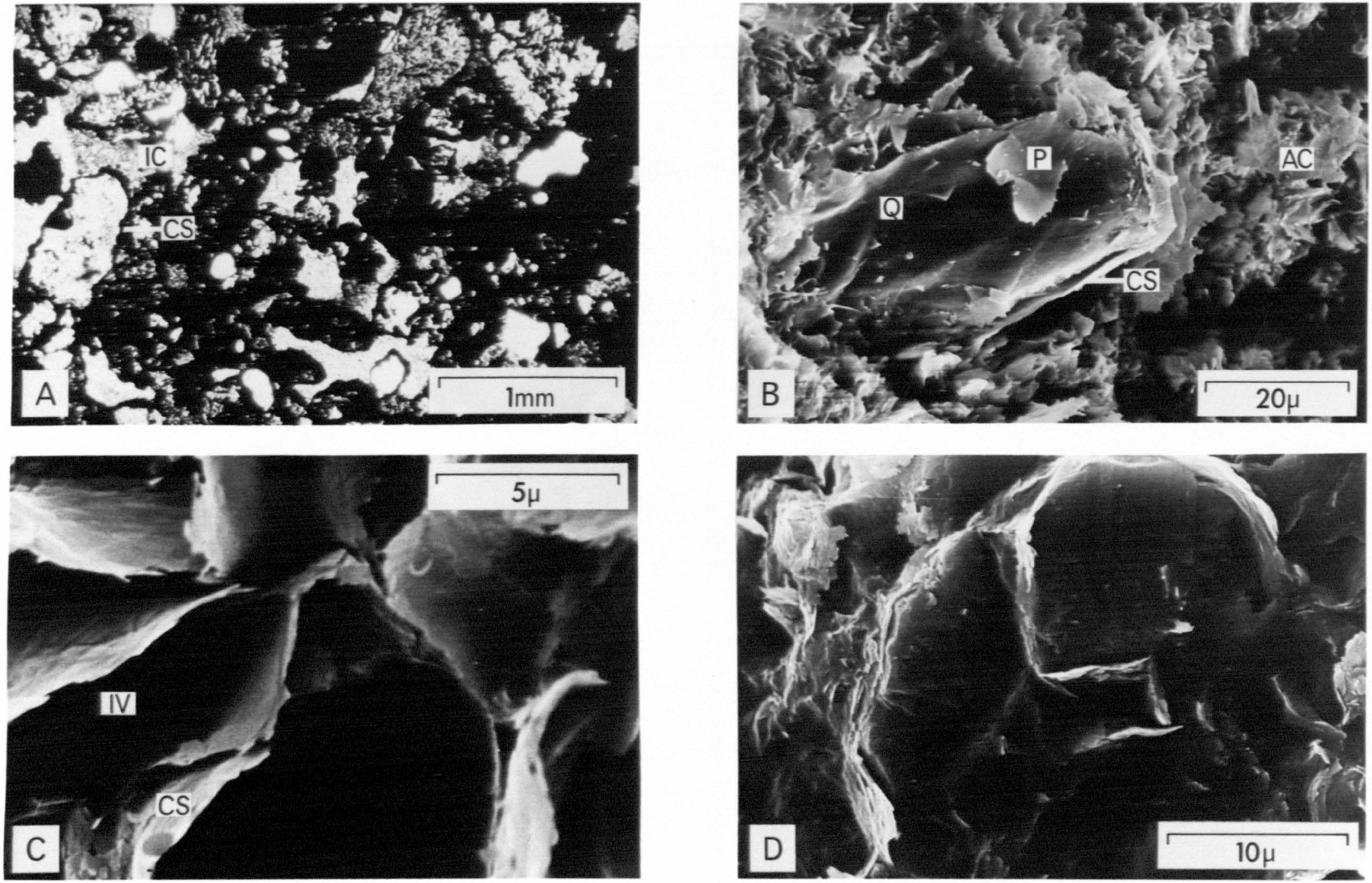

IC
CS
A
1mm
P
AC
Q
CS
B
20μ
5μ
IV
CS
C
D
10μ

Plate 2.2

A. Interstitial matrix clay (CM) and clay skin (CS) separating two detrital grains. Position of the symbols denotes location of EDS traces reproduced in B and C. (SEM photomicrograph)

B. EDS trace of interstitial matrix clay.

C. EDS trace of mechanically infiltrated clay. SEM specimen coated with gold.

D. Partially dissolved siltstone grain with well defined clay skin (CS). Dissolution voids (DV) occupy much of the area enclosed by the clay skin. (Thin section photomicrograph)

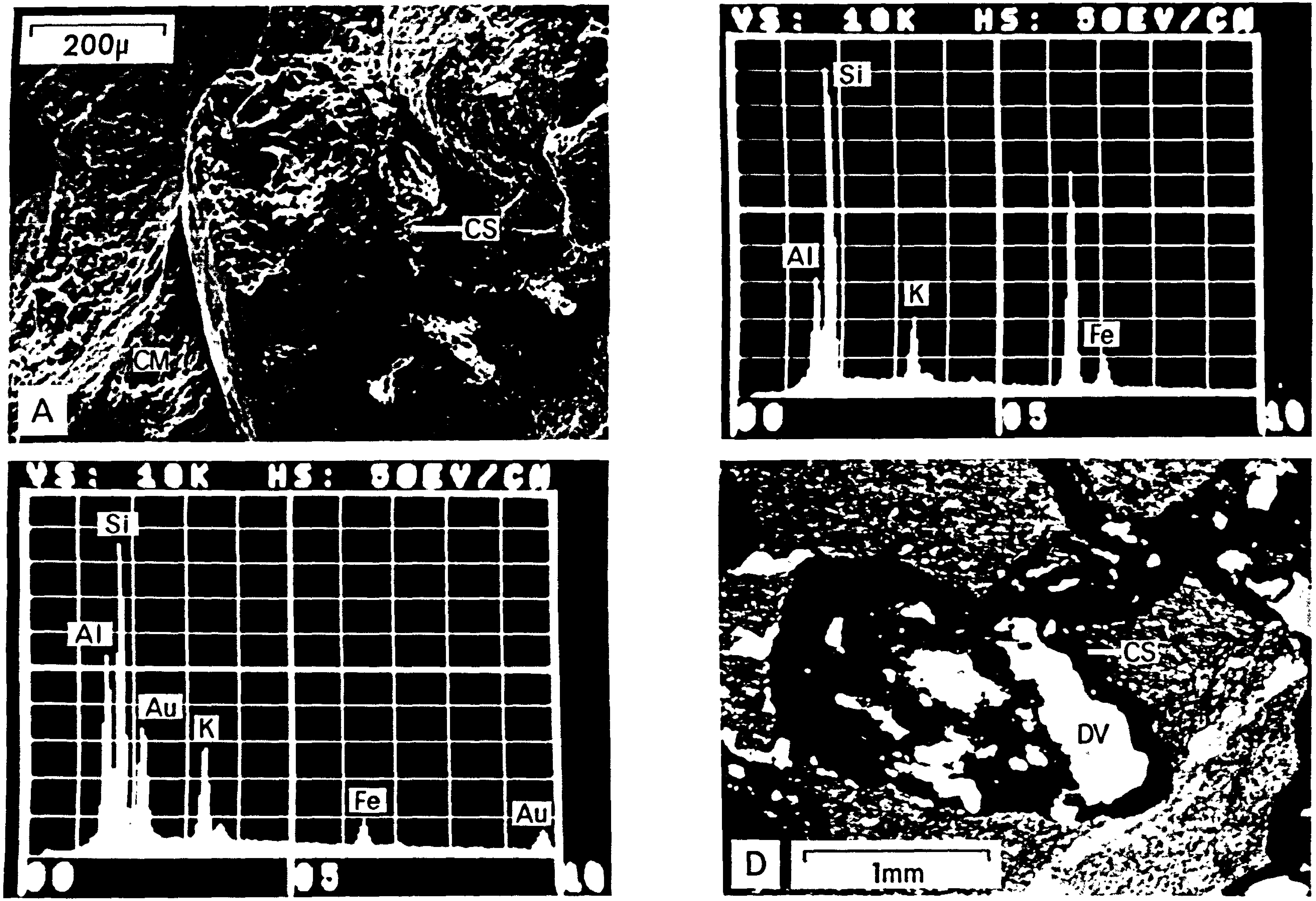

200μ
CS
CM
A
VS: 10K HS: 50EV/CM
Si
Al
K
Fe
00
05
10
VS: 10K HS: 50EV/CM
Si
Al
Au
K
Fe
Au
00
05
10
CS
DV
D
1mm

Plate 2.3

A. Relic ?feldspar grain with delicately etched needle like termination. Note similarity of this texture with feldspar grains shown in Plate 4.3A and B. (SEM photomicrograph)

B. Enlargement of part of A. (SEM photomicrograph)

C. Authigenic clay matrix. Note characteristic texture of the clay plates. (SEM photomicrograph)

D. Aggregate of authigenic clay. (SEM photomicrograph)

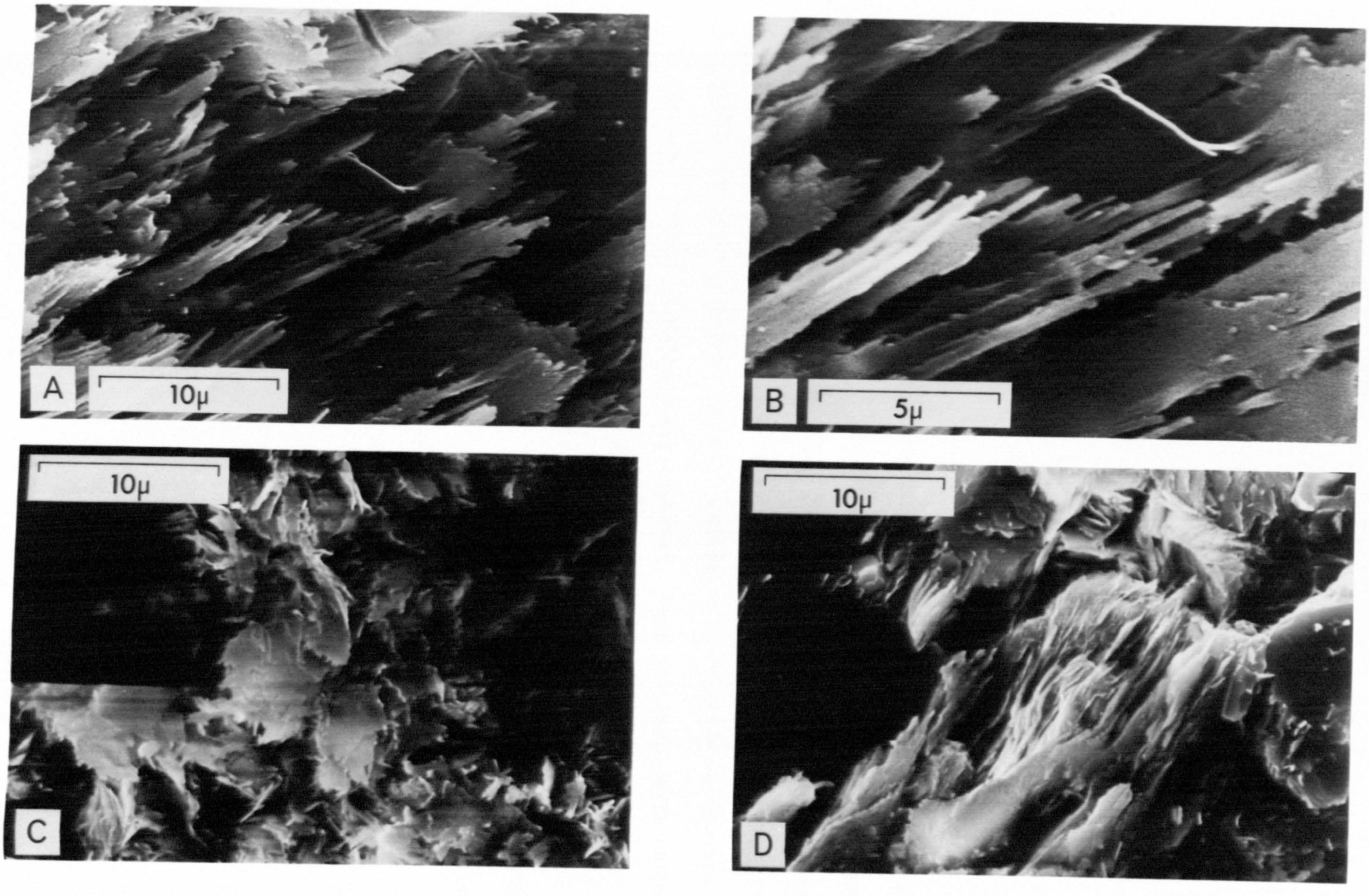

A
10μ
B
5μ
10μ
C
10μ
D

Plate 2.4

A. Preferentially replaced feldspar phenocryst within largely unaltered Borrowdale Volcanic clast. The feldspar has been completely replaced by ferroan calcite (RCC) and clay (RC). (Thin section photomicrograph)

B. Peripherally replaced Borrowdale Volcanic grain. A quartz vein (V) within the grain has been largely unaffected by replacement whereas the grain itself is extensively replaced. Note the relic silt sized grains within the replacement clay. (Thin section photomicrograph)

C. Authigenic illite crystals (AI) lining an interstitial void between Skiddaw Slate fragments. The larger Skiddaw Slate clast to the left is marginally replaced and has a diffuse margin with the interstitial matrix clay. (Thin section photomicrograph)

D. Authigenic illite. (SEM photomicrograph)

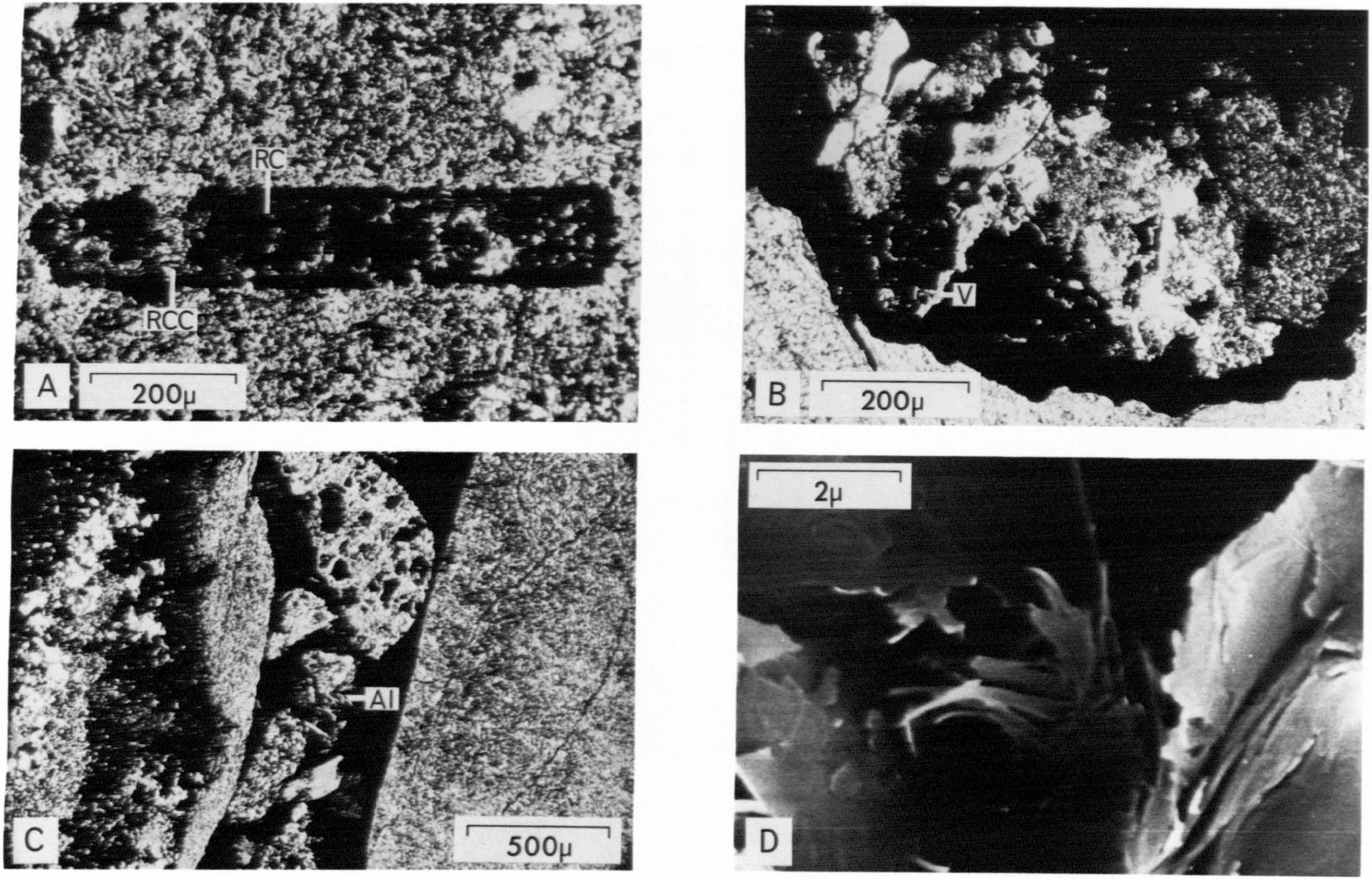

A
200μ
RC
RCC
B
200μ
V
C
500μ
Al
D
2μ

Plate 2.5

A. Authigenic illite. (SEM photomicrograph)

B. Authigenic illite. (SEM photomicrograph)

C. EDS trace of authigenic illite.

D. Detrital clast coated in authigenic illite displaying characteristic boxwork texture. (SEM photomicrograph)

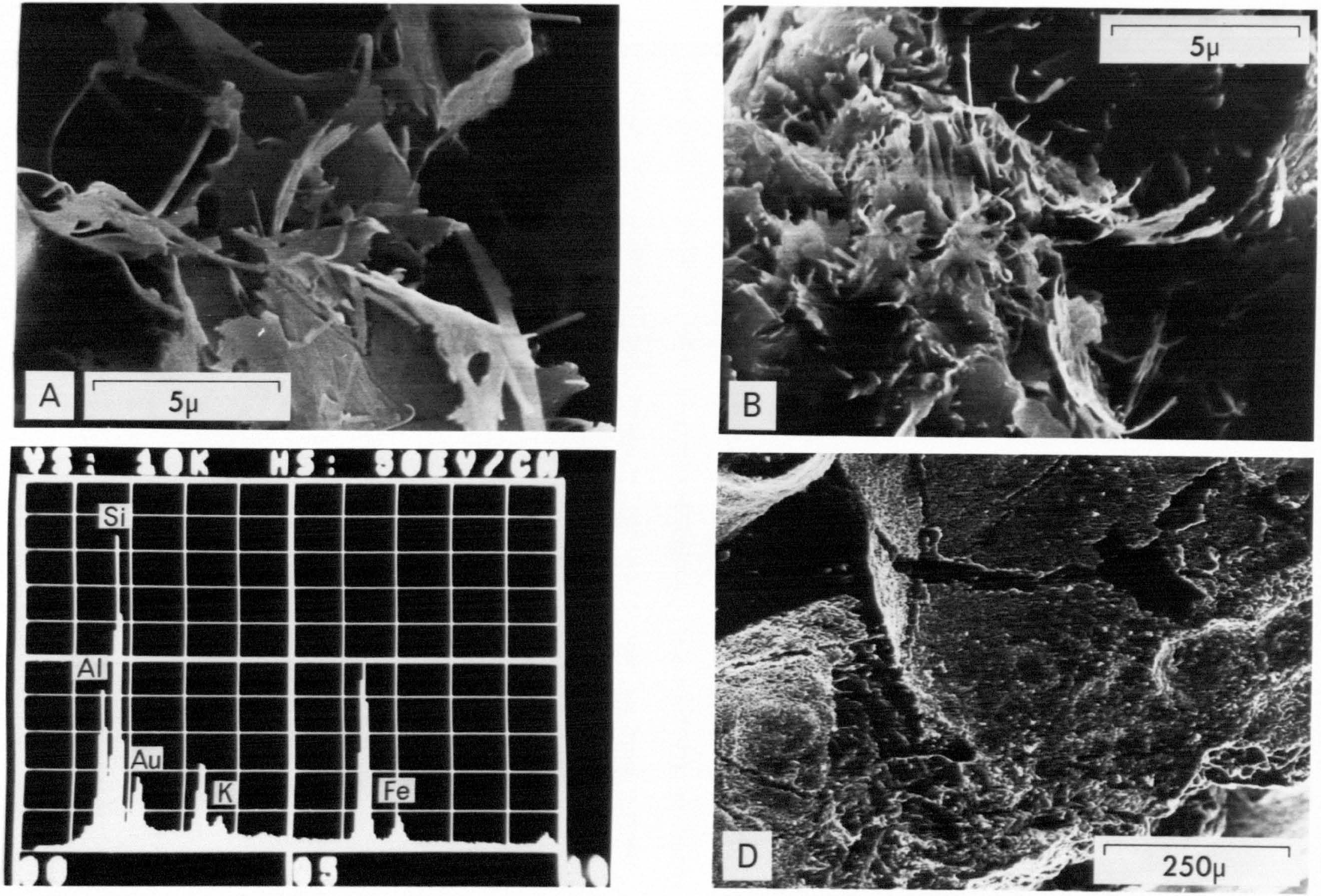

A
5μ
B
5μ
VS: 10K HS: 50EV/CM
Si
Al
Au
K
Fe
D
250μ

Plate 2.6

A. Enlargement of part of 2.5D. (SEM photomicrograph)

B. Sparry calcite cement (S) surrounding replaced or recrystallised detrital grain (G) with a well developed clay skin (CS). A poorly developed drusy fabric indicates the clay surface to have formed the initial site of calcite precipitation. The surfaces of other grains (indicated by DG), however, do not appear to have formed favourable sites for early precipitation. (Thin section photomicrograph)

C. Poikilotopic calcite cement enclosing quartz sand grain (Q) and several silt sized quartz particles. (Thin section photomicrograph)

D. Clay coating (CD) separating growth phases of calcite cement. Arrow indicates optically continuous calcite crystal. (Thin section photomicrograph)

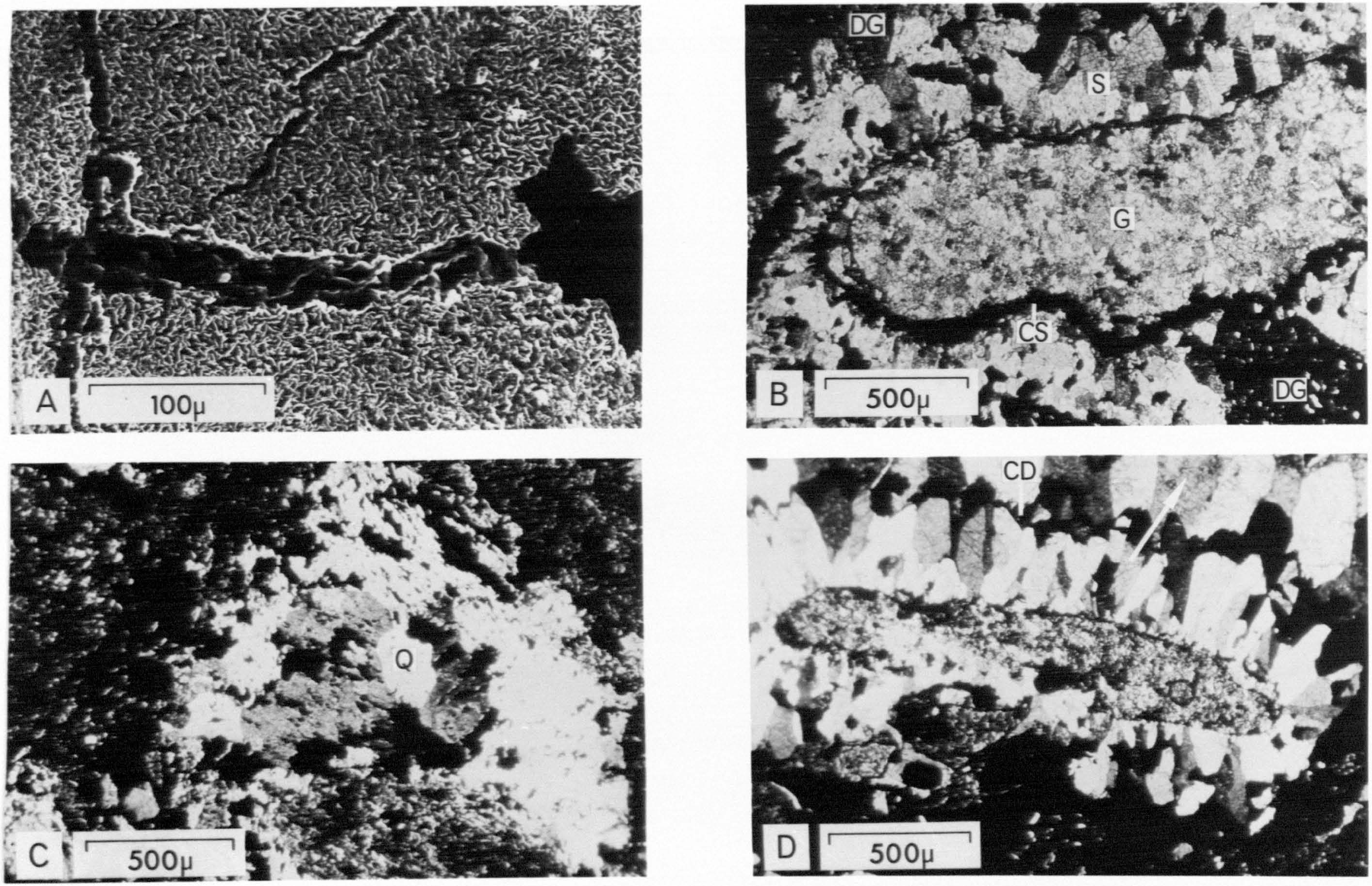

A
100μ
DG
S
G
CS
B
500μ
DG
Q
C
500μ
CD
D
500μ

Plate 2.7

A. Clay coatings separating growth phases of calcite cement around replaced sand grain (RC). (Thin section photomicrograph)

B. Dislocated clay rims within interstitial calcite cement. (Thin section photomicrograph)

C. Detached clay rims within interstitial calcite cement (IC). The grains marked RC and RCC have been replaced by clay and calcite respectively. (Thin section photomicrograph)

D. Disrupted clay rims which have been dislocated from the surface of a partially replaced grain (RG) during formation of calcite cement (C). (Thin section photomicrograph)

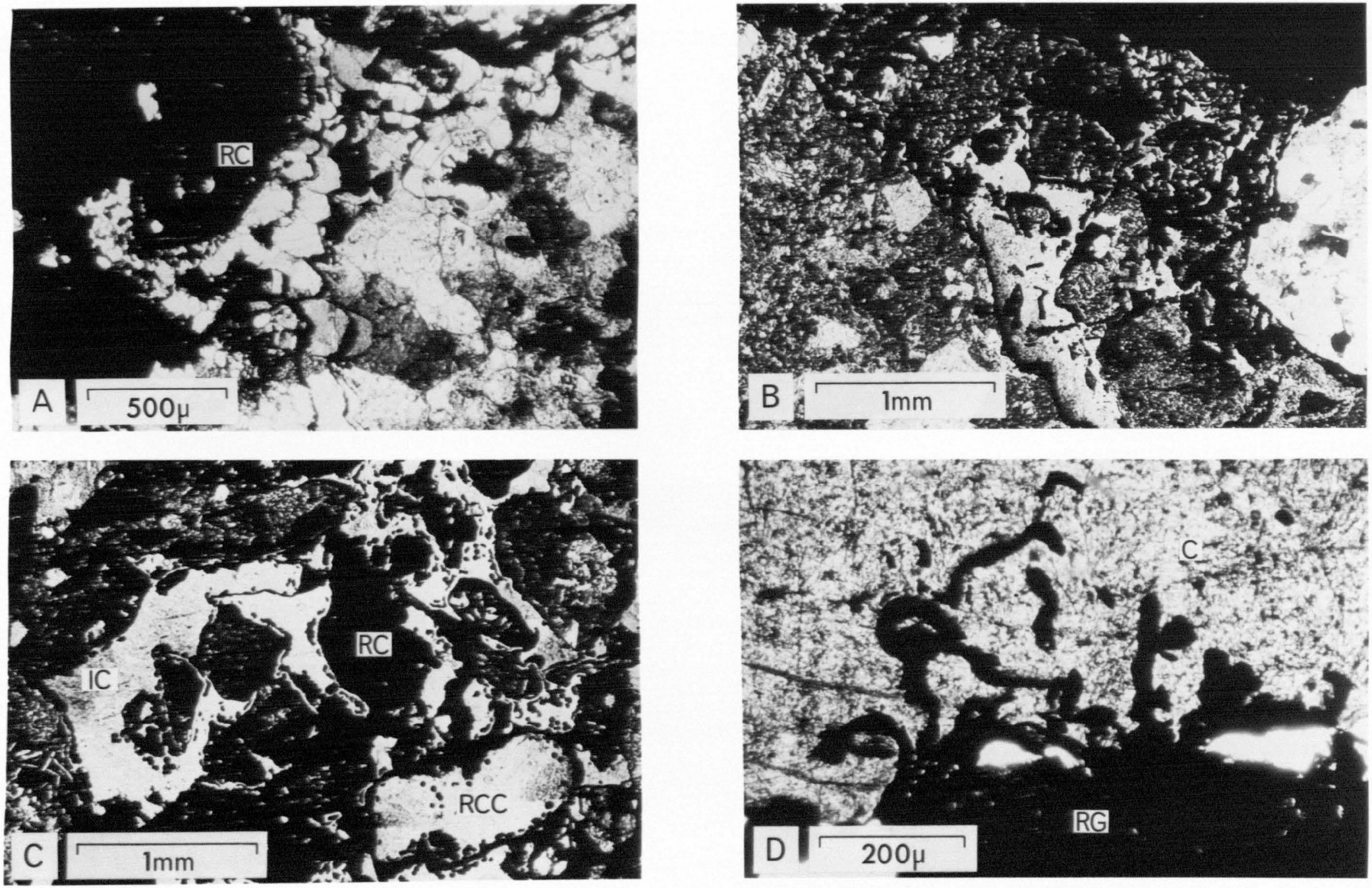

A
RC
500μ
B
1mm
C
IC
RC
RCC
1mm
D
C
RG
200μ

Plate 2.8

A. Rim clays (CS) enclosing calcite pseudomorphs after detrital grains (CCP). Larger areas of clay (RC) may be partly replacive in origin. (Thin section photomicrograph)

B. Partially dissolved and peripherally replaced Borrowdale Volcanic fragments. The grains are enclosed by thick clay rims which are probably a combination of mechanically infiltrated and replacement clay. The alteration of the larger clast (BV) is partly accomplished by an in situ chemical breakdown to clay (RC) and partly by a more complex process involving the precipitation of optically continuous calcite with clay drapes. The calcite has nucleated on the surface of the clast which is thus either the original surface, and the calcite-clay alternation an interstitial fabric, or a dissolution surface and the calcite-clay alternation a void infill with a fabric comparable to that of the interstices. The nucleation of the calcite on the surface of the grain precludes the possibility of a replacive origin. (Thin section photomicrograph)

C. Peripherally dissolved Borrowdale Volcanic clast (BV) with multiple calcite-clay infill. The original extent of the clast is indicated by arrows. The granular texture of the calcite and disseminated clay bands is distinct from that of the optically continuous calcite and clay alternations. A calcite pseudomorph (CCP) with relic clay core demonstrates an original grain contact which has been completely destroyed by diagenesis. (Thin section photomicrograph)

D. Dissolved Skiddaw Slate grain. Small remnant core (SS) is enclosed by at least three generations of authigenic calcite (C) with a granular texture. Each generation is separated by a thin irregular clay coating (CC). Small remnant quartz silt grains occur within the calcite. The original extent of the grain is indicated by a clay rim (CS) which is in contact with surrounding grains. (Thin section photomicrograph)

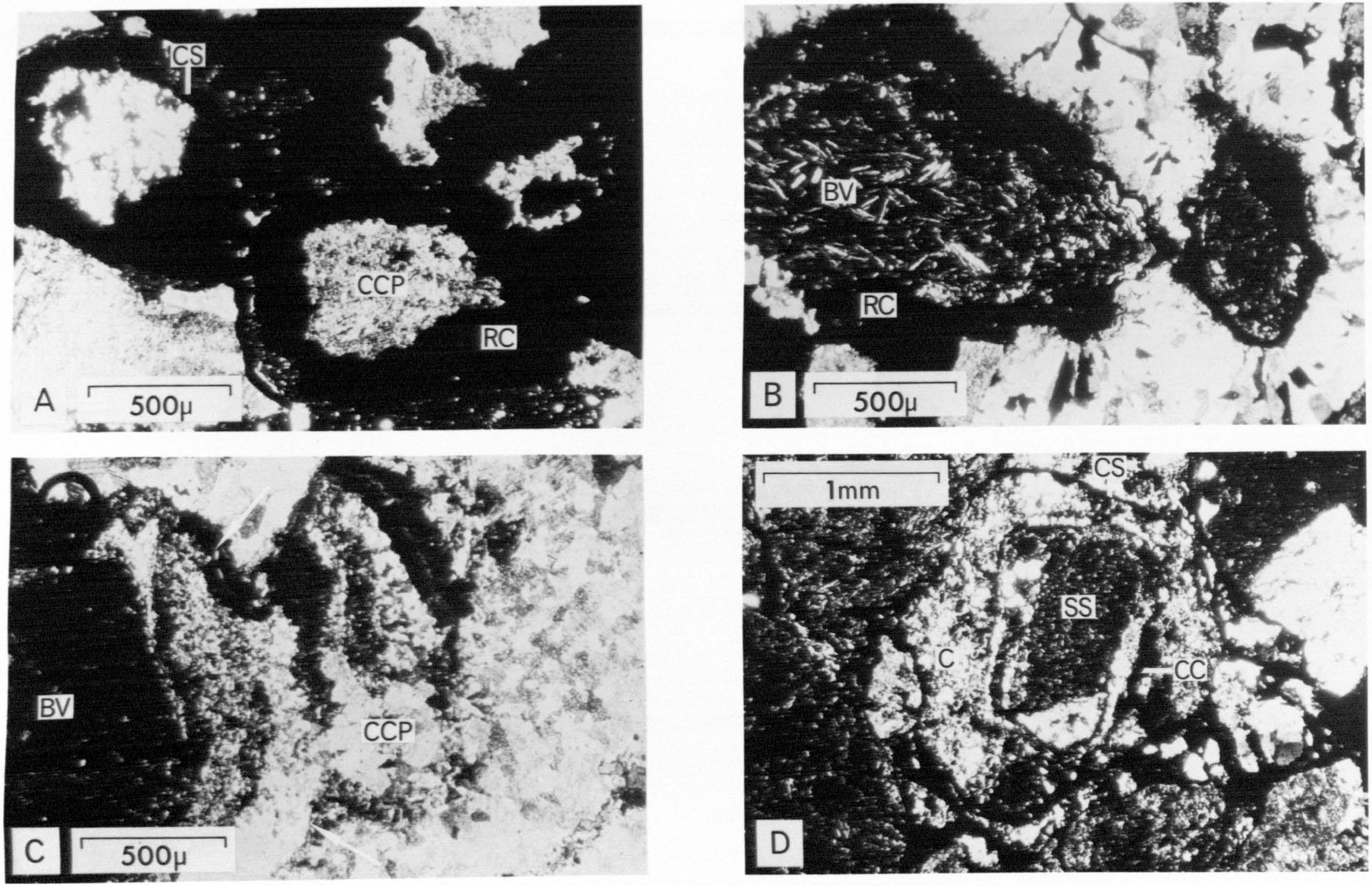

CS
CCP
RC
A
500μ
BV
RC
B
500μ
BV
CCP
C
500μ
1mm
CS
SS
C
CC
D

Plate 2.9

A. Compacted rim clay (CS) enclosing small remnant core of Skiddaw Slate clast. (Thin section photomicrograph)

B. Collapsed rim clays (CS) enclosing partial calcite infills of complete dissolution voids. The fracturing of the lower of the three examples is probably a compactional feature which has been accentuated by growth of the calcite crystals. (Thin section photomicrograph)

C. Rhomb-shaped clay pseudomorphs after calcite. (Thin section photomicrograph)

D. Selective replacement of calcite crystal. Note the stylolitic contact (SC) which is marked by a thin clay film. Replacement is terminated at this contact. (Thin section photomicrograph)

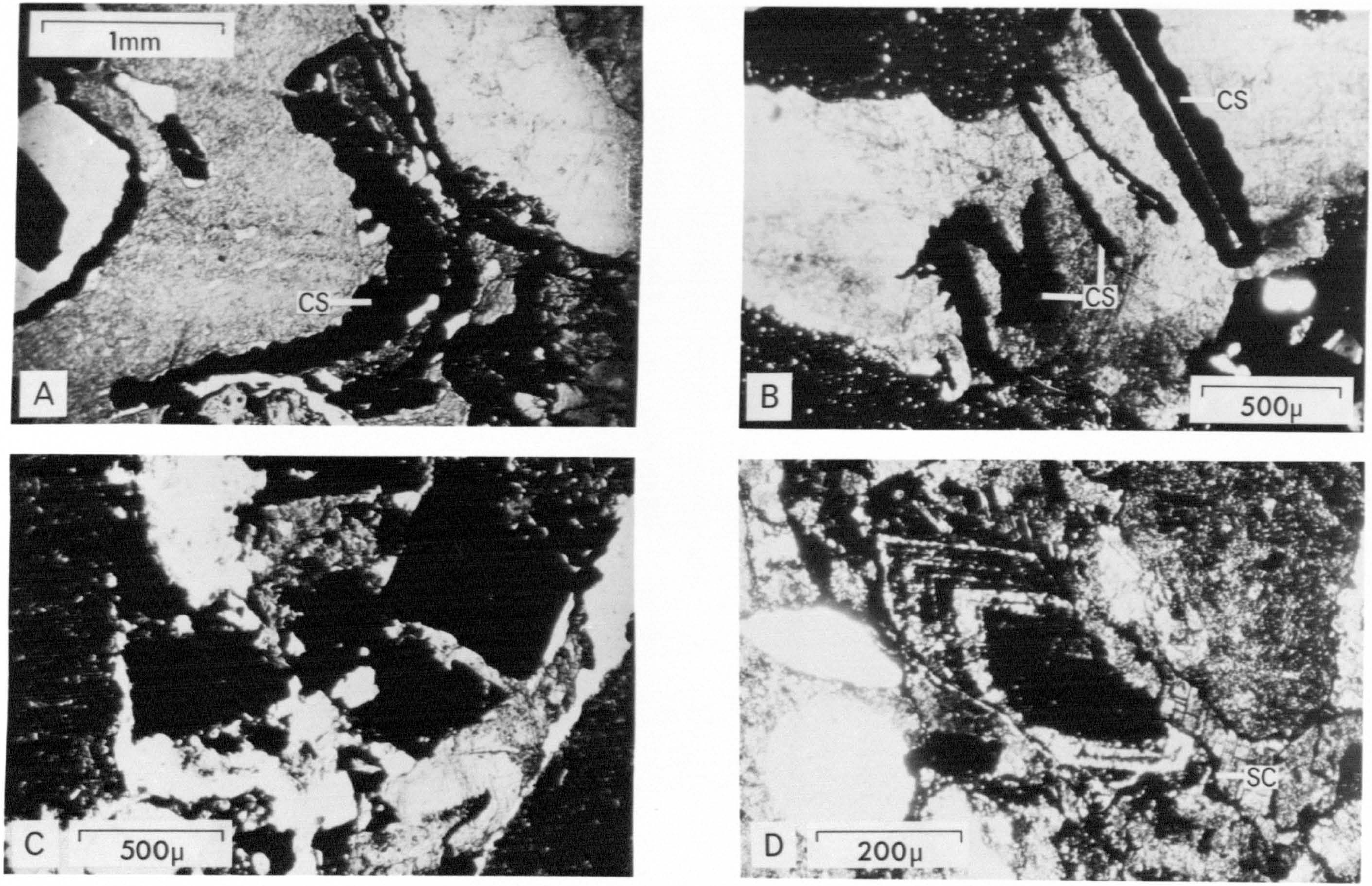

1mm
CS
A
CS
CS
B
500μ
C
500μ
SC
D
200μ

Plate 2.10

A. Botryoidal haematite. Note the radiating fibrous habit of the haematite crystals. (SEM photomicrograph)

B. Botryoidal haematite (SEM photomicrograph)

C. Enlargement of part of B. (SEM photomicrograph)

D. Botryoidal haematite. (SEM photomicrograph)

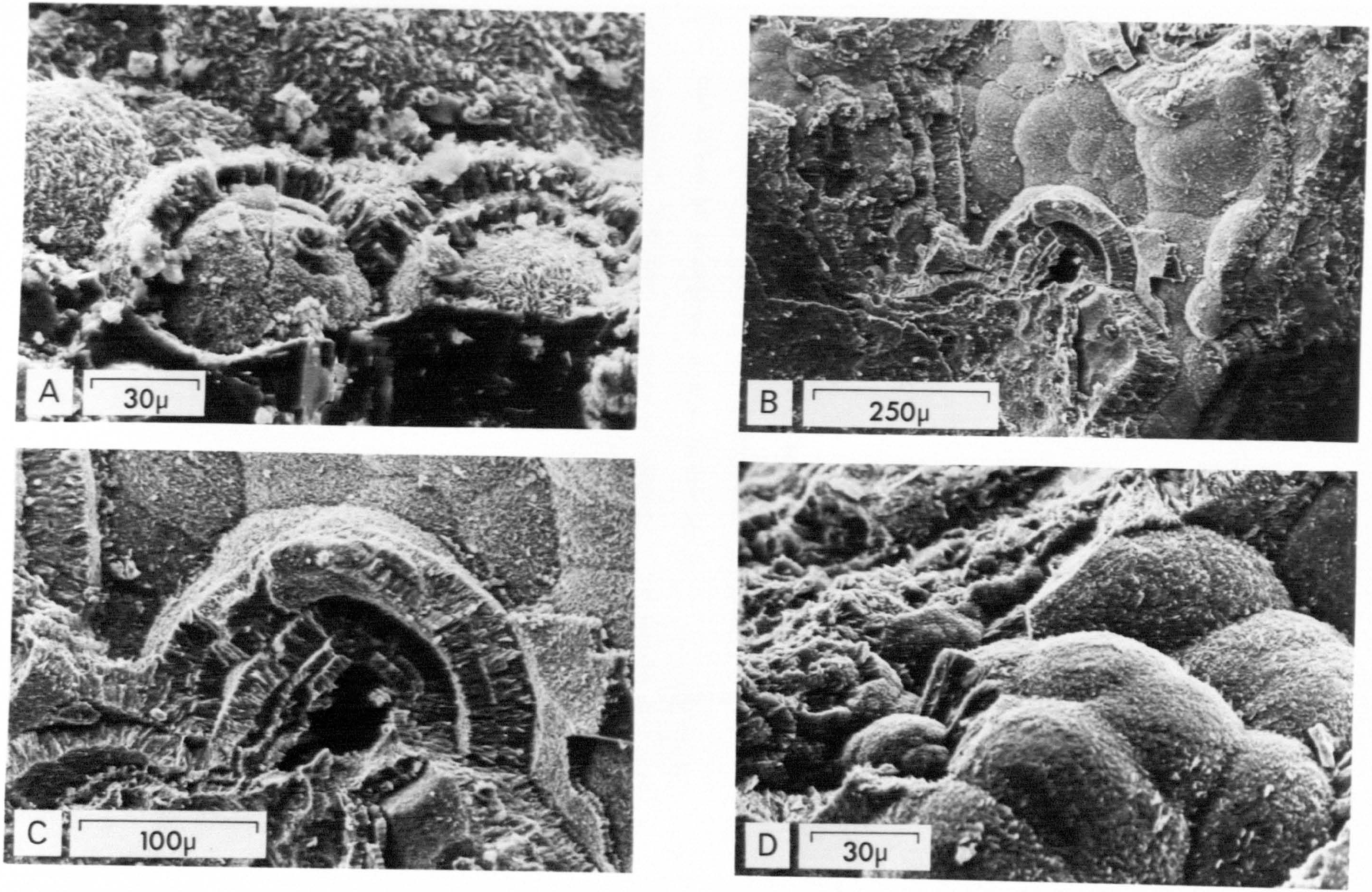

A
30µ
B
250µ
C
100µ
D
30µ

Plate 2.11

A. Hexagonal tabular crystals of haematite. (SEM photomicrograph)

B. Enlargement of part of A. (SEM photomicrograph)

C. EDS trace of hexagonal plates shown in B.

D. Hexagonal, tabular crystals of haematite replacing detrital grains. (SEM photomicrograph)

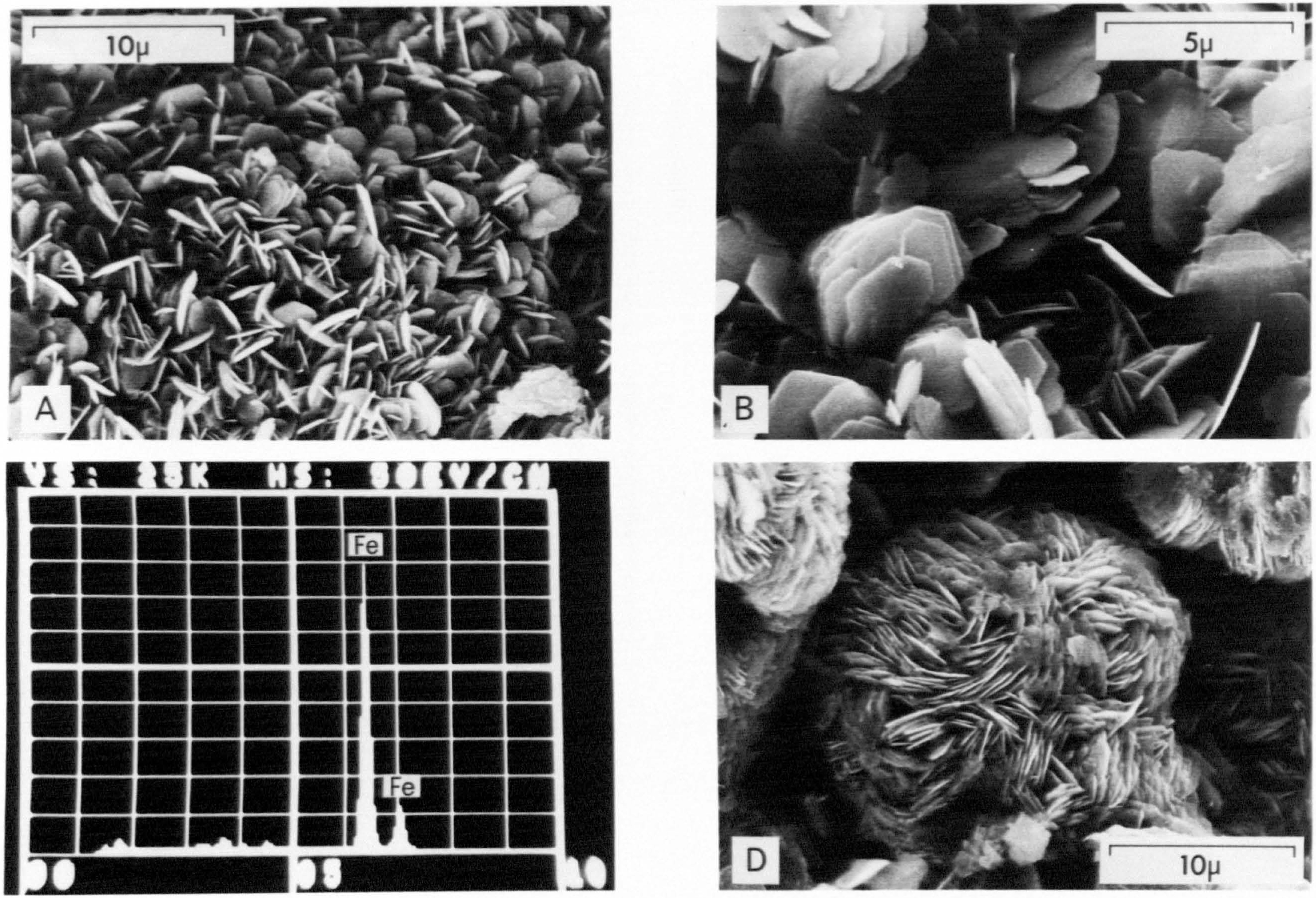

10µ
A
5µ
B
Fe
Fe
D
10µ

Plate 2.12

A. Haematite replacing detrital grains. (SEM photomicrograph)

B. Replaced detrital grain with remnant clay rim (SEM photo-micrograph)

C. Authigenic haematite. (SEM photomicrograph)

D. Enlargement of part of C. (SEM photomicrograph)

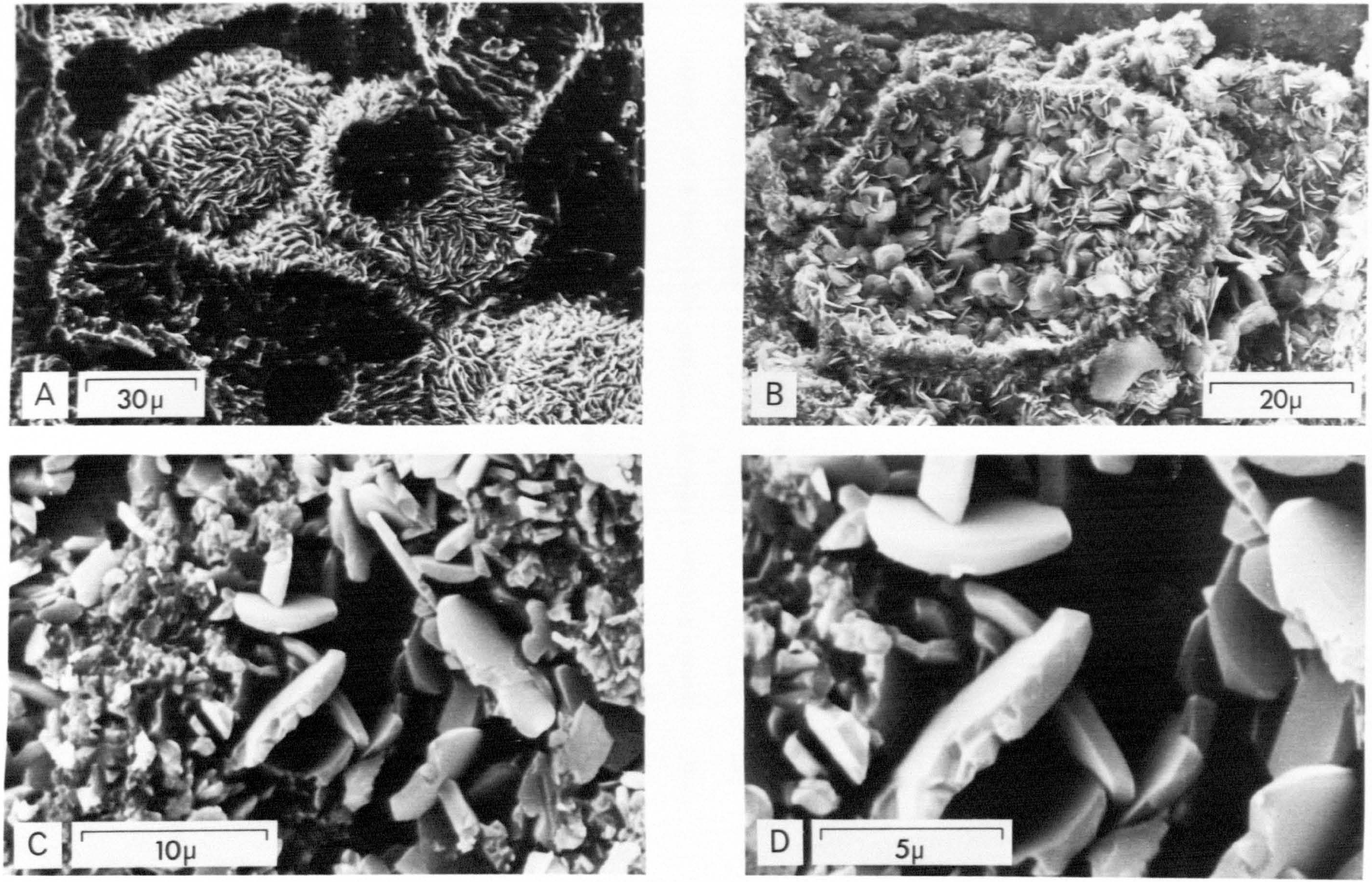

A
30μ
B
20μ
C
10μ
D
5μ

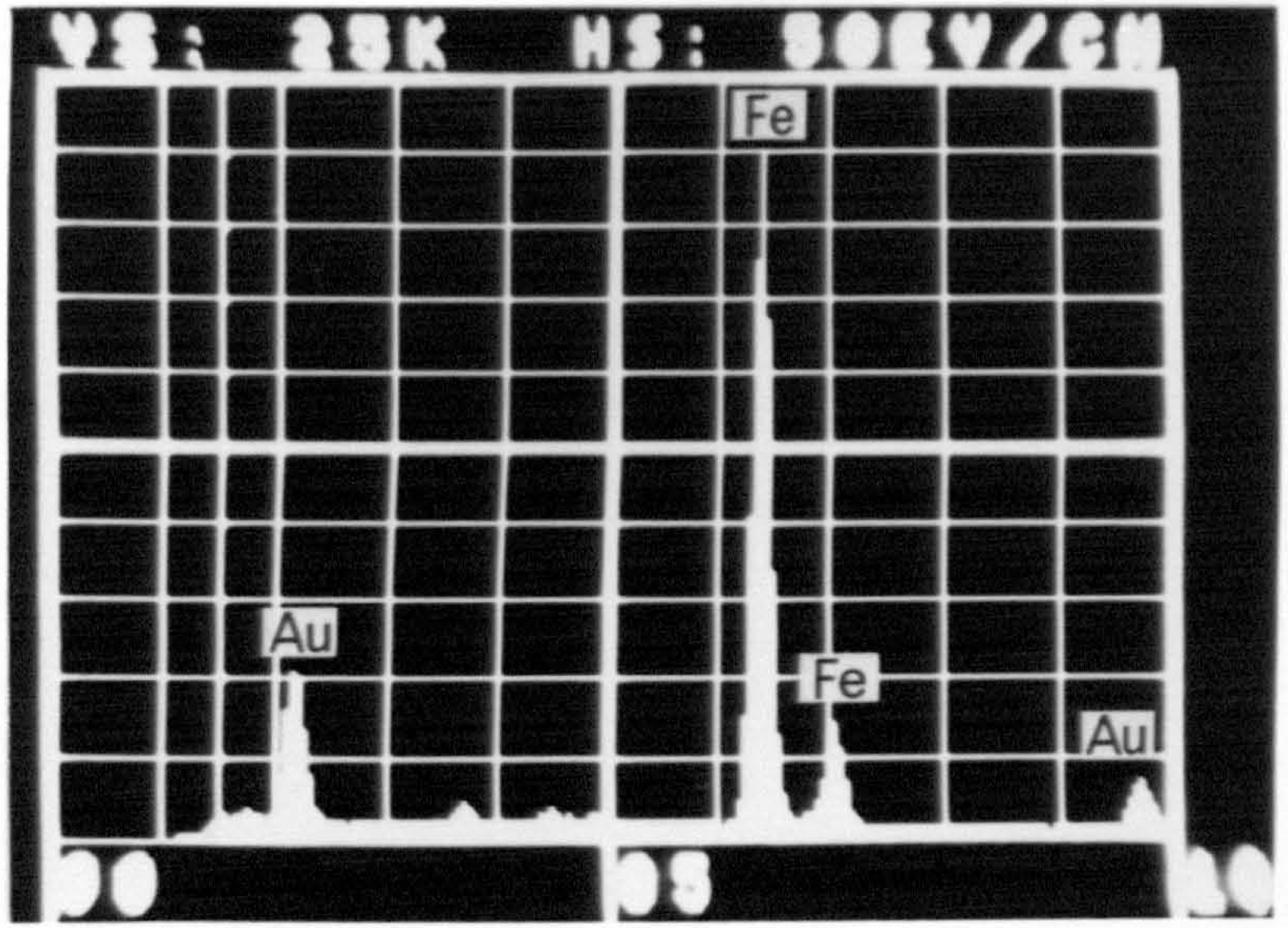
Plate 2.13

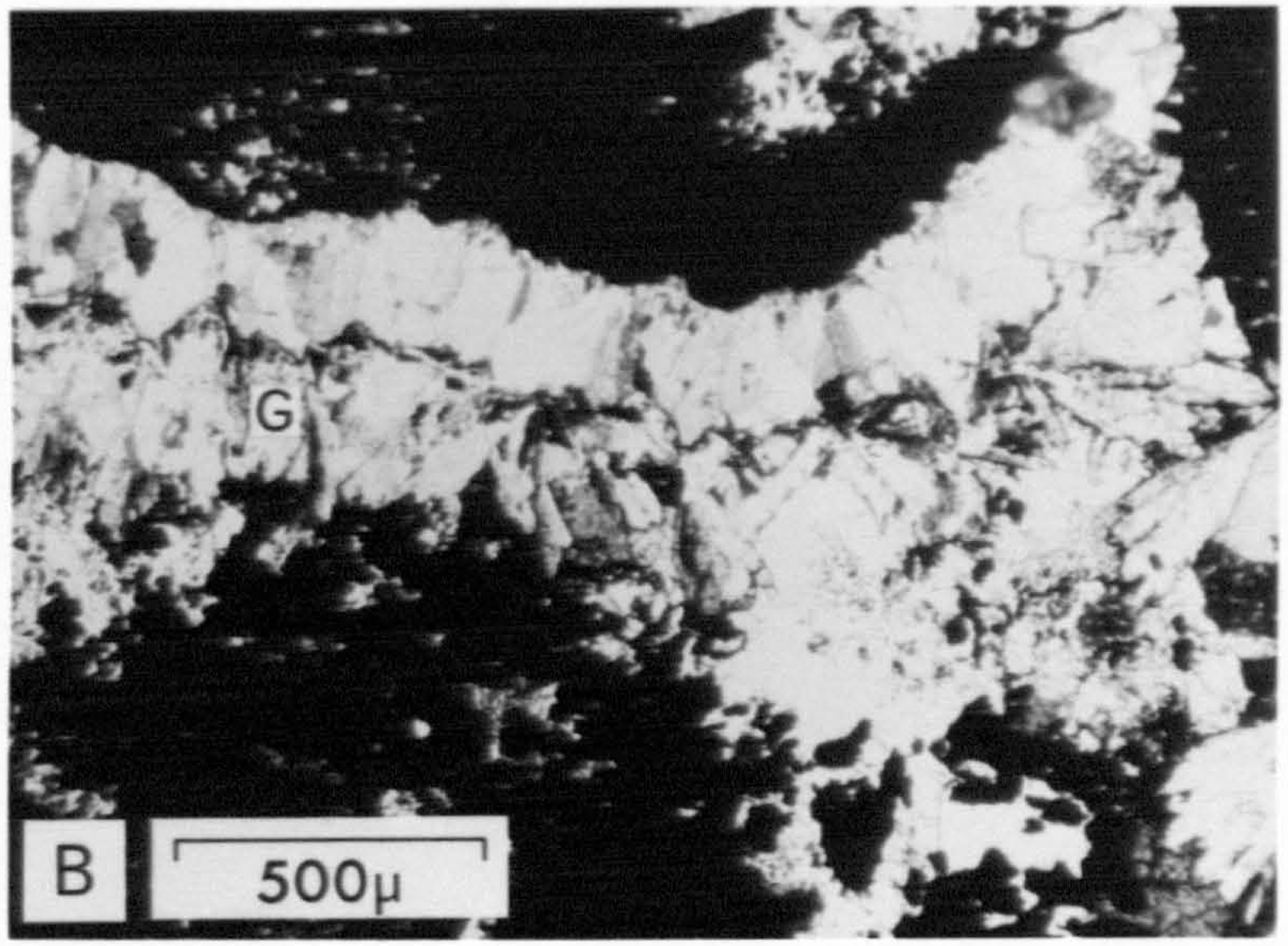
A. EDS trace of authigenic haematite shown in Plate 2.12D.

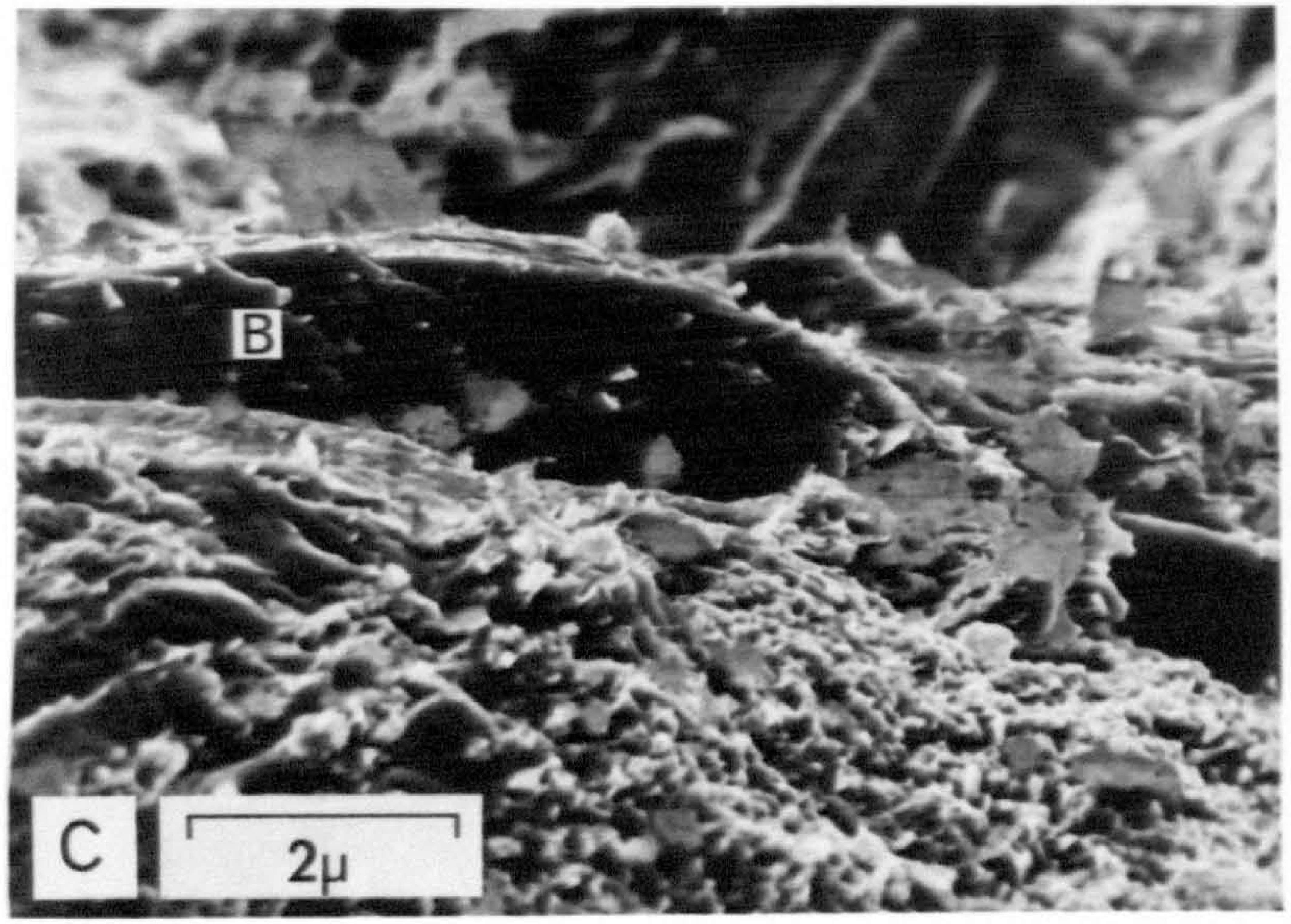
B. Interstitial gypsum cement (G). Note the nucleation of the gypsum on a thin clay coating on drusy calcite cement. (Thin section photomicrograph)

C. Barytes crystal (B) within interstitial matrix clay. (SEM photomicrograph)

D. EDS trace of barytes shown in C. The sulphur peak is masked by the gold peak recorded from the sample coating.


VS: 25K HS: 50EV/CM
Fe
Au
Fe
Au
00
05
10


G
B
500μ


B
C
2μ

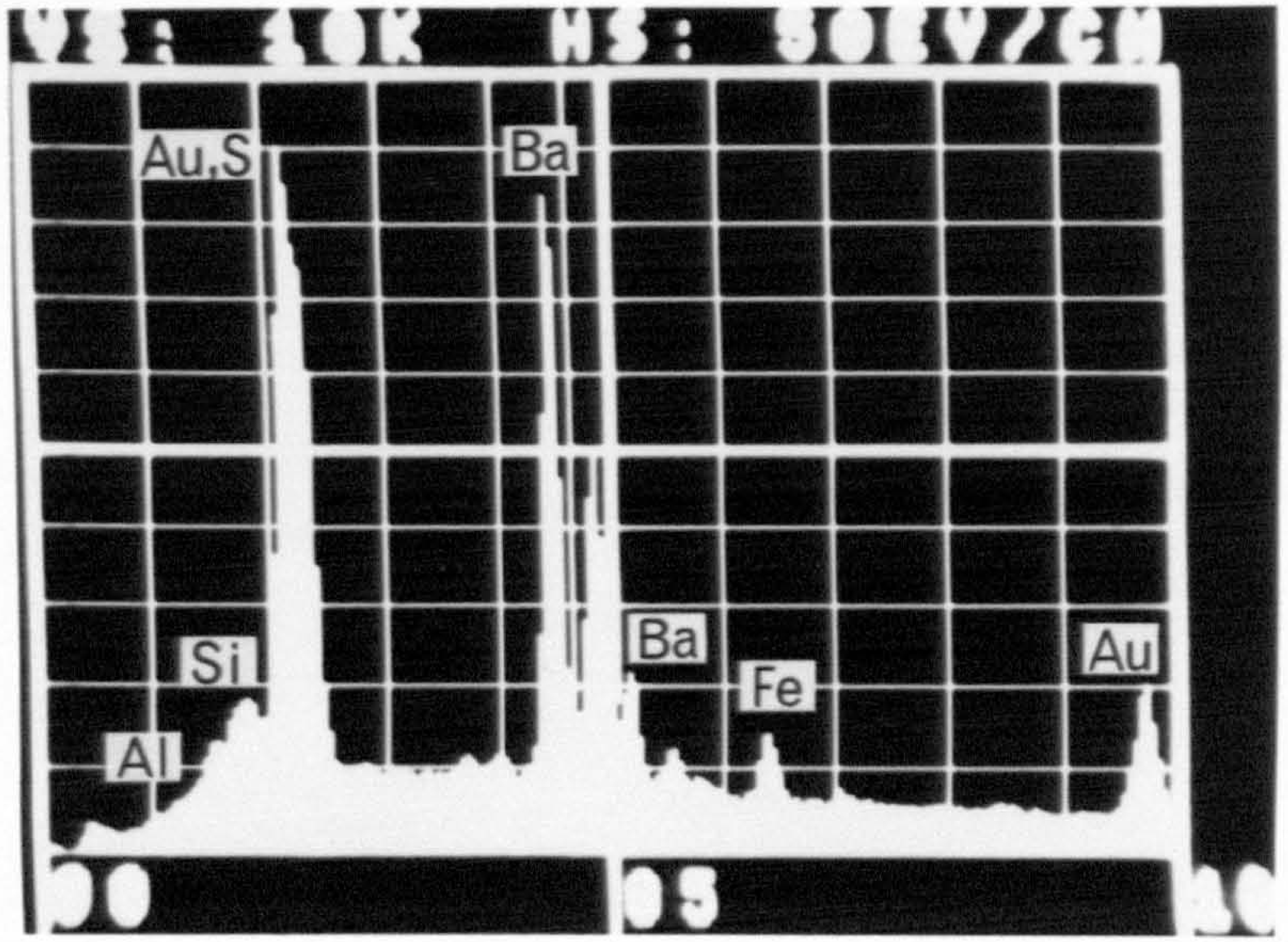

VS: 10K HS: 50EV/CM
Au,S
Ba
Si
Al
Ba
Fe
Au
00
05
10

Plate 4.1

Exposures of nodular carbonate horizons from the St. Bees Shale at North Head, St. Bees (NY949152). The location of the small waterfall in A (lower centre) marks the site of B, C and D.

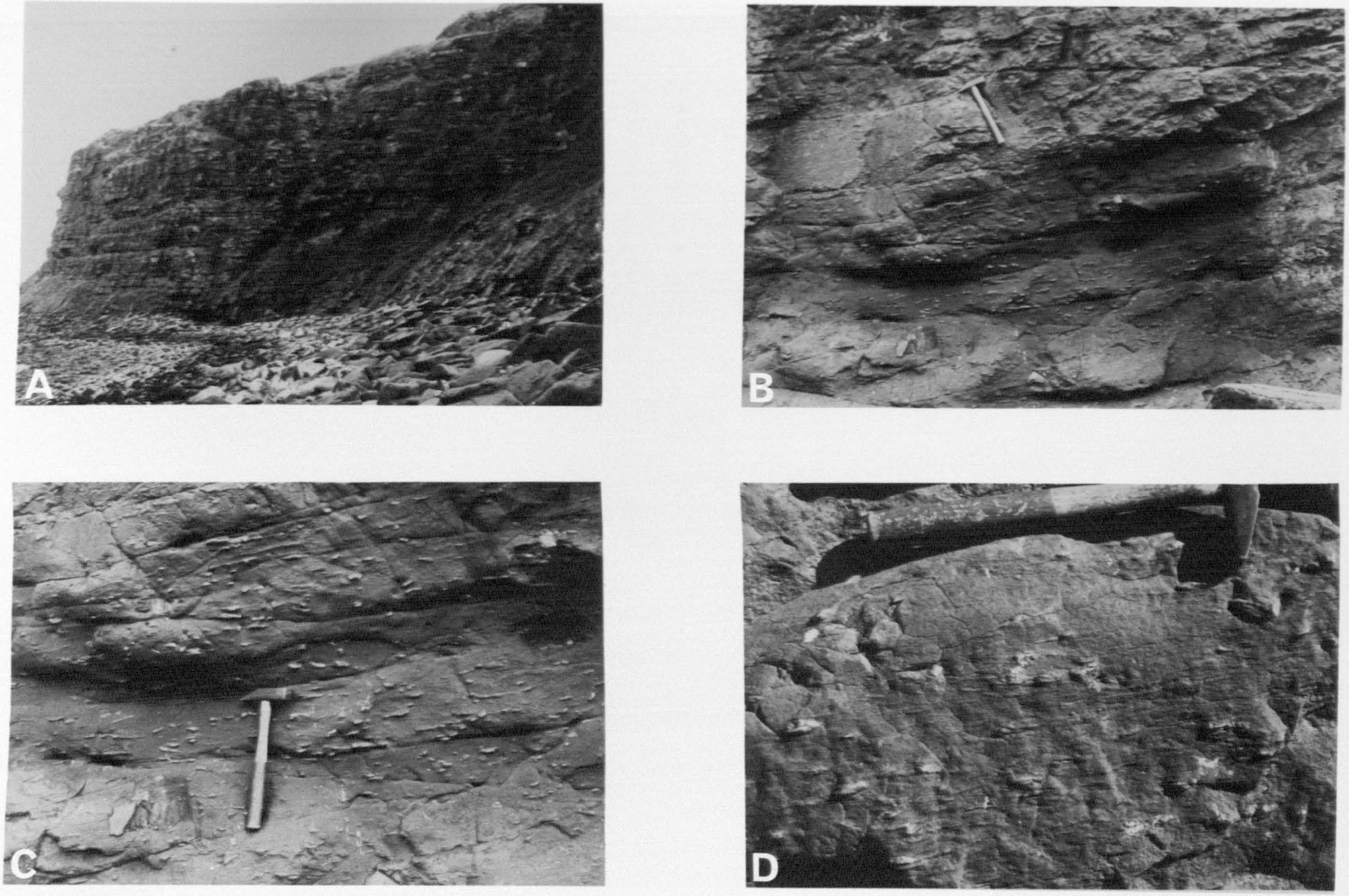


Plate 4.2

Nodular carbonate sequences from Surface Borehole (SBH) 304 (NY01770963, near Carlton).

A. Sand-silt horizon occurring from 167.9 to 165.6 metres above mining datum (AMD). The lower part of this horizon is characterised by diffuse, sequential accumulations of carbonate and the upper part by disseminated nodular carbonate.

B. Carbonate horizons from a silt-shale horizon occurring from 214.9 to 214.3 metres AMD (low core recovery).


10 CMS

Plate 4.3

Peripherally dissolved feldspar grains.

A. Peripherally dissolved feldspar grain with delicate needle-like terminations formed by dissolution. (Scanning Electron Microscope photomicrograph)

B. Enlargement of part of A. (SEM photomicrograph)

C. Extensively dissolved grain showing relict feldspar (F) and authigenic replacement clay (RC). Grain is coated in thin clay skin (CS). (SEM photomicrograph)

D. Enlargement of part of C. Dissolution voids (DV) occupy the area between the relict grain and replacement clay. (SEM photomicrograph)

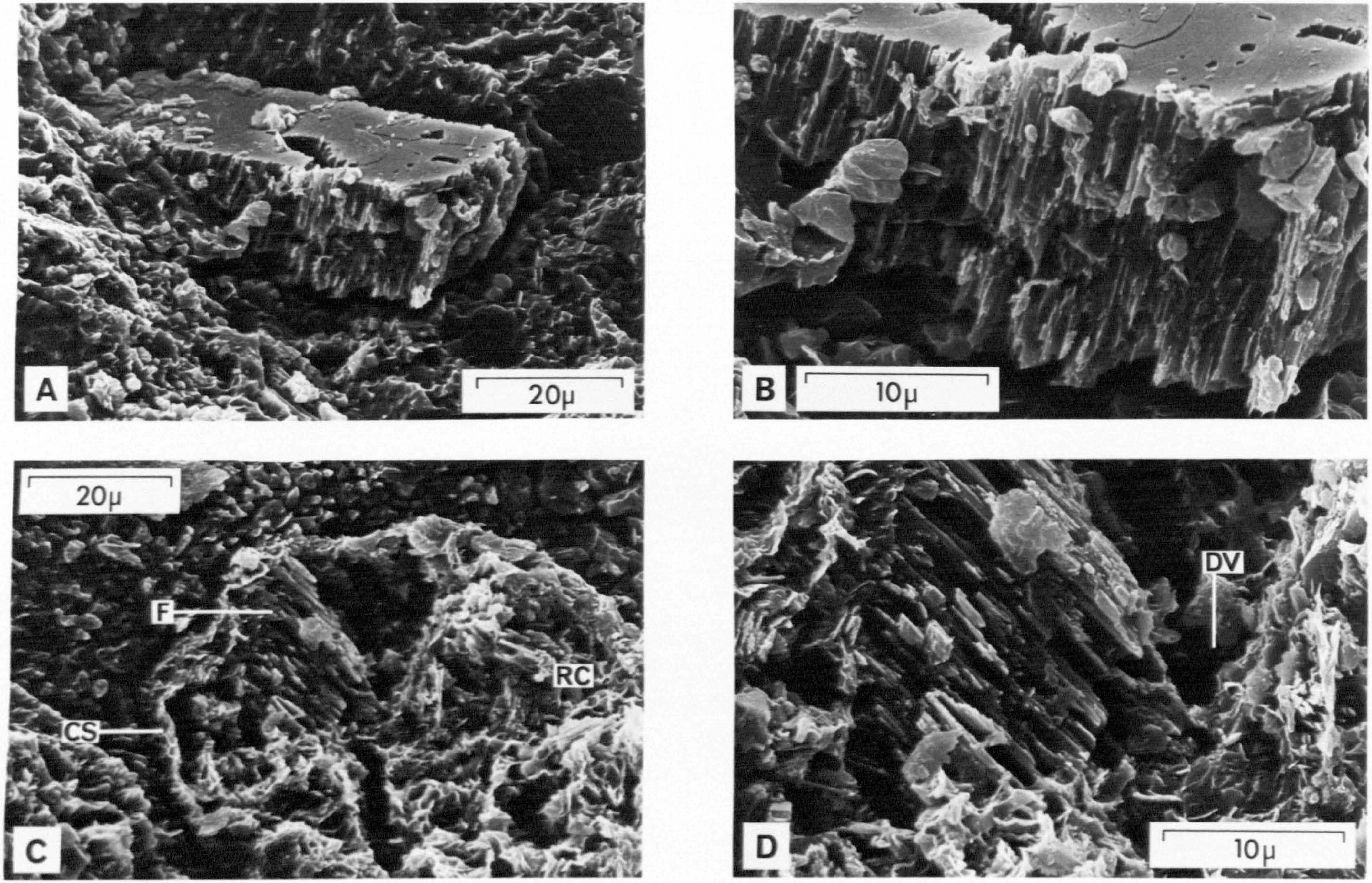

A
20μ
B
10μ
20μ
F
RC
CS
C
DV
D
10μ

Plate 4.4

A. Replaced feldspar (?) or Borrowdale Volcanic (?) grains represented by clay pseudomorphs (CP). (Thin section photomicrograph)

B. Grain, completely replaced by clay (CP). (SEM photomicrograph)

C. Interstitial carbonate cement (IC) contained within the clay matrix (CM) of the host framework sandstone. (Thin section photomicrograph)

D. Granular texture of micritic carbonate cement. (SEM photomicrograph)

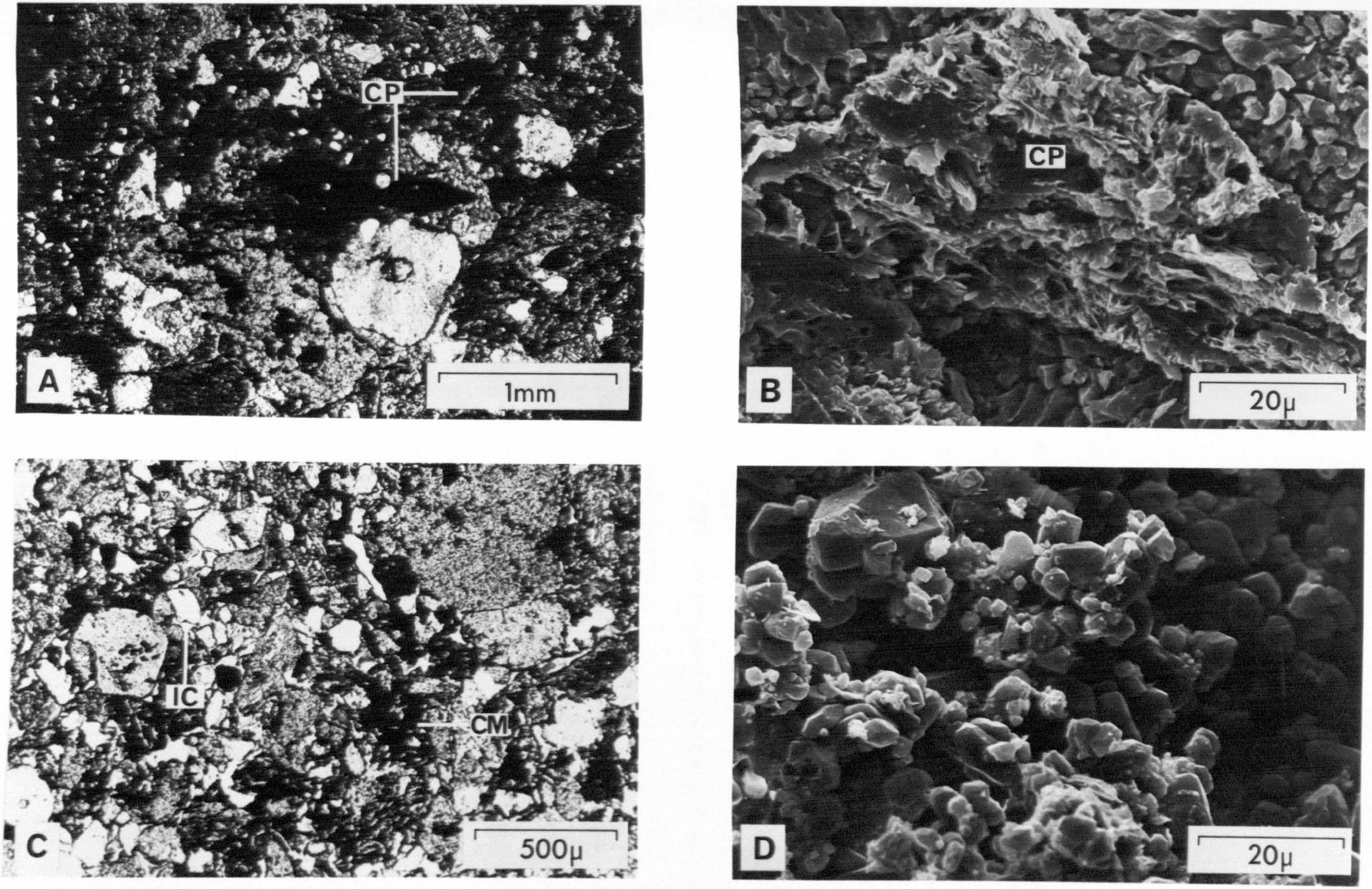

A
CP
1mm
B
CP
20μ
C
IC
CM
500μ
D
20μ

Plate 4.5

A. Diffuse carbonate nodule (CN) with indistinct lateral and vertical margins, gradational with the surrounding host sediment (HS). Note the concentration of matrix clay above, and to a lesser extent, below the nodule. (Thin section photomicrograph)

B. Micritic carbonate (M). Note the abrupt upper margin demarked by a distinct clay band. The sediment overlying the micrite contains a high proportion of interstitial carbonate. (Thin section photomicrograph)

C. Microsparitic calcite. Clay skins (CS) encircle completely replaced grains (RG). (Thin section photomicrograph)

D. Granular carbonate cement (C). Note the fresh quartz grain (Q) and replaced grain (RC). A clay skin (CS), partially enclosing a quartz grain, has been prised away from the grain surface by the growth of granular carbonate. (SEM photomicrograph)

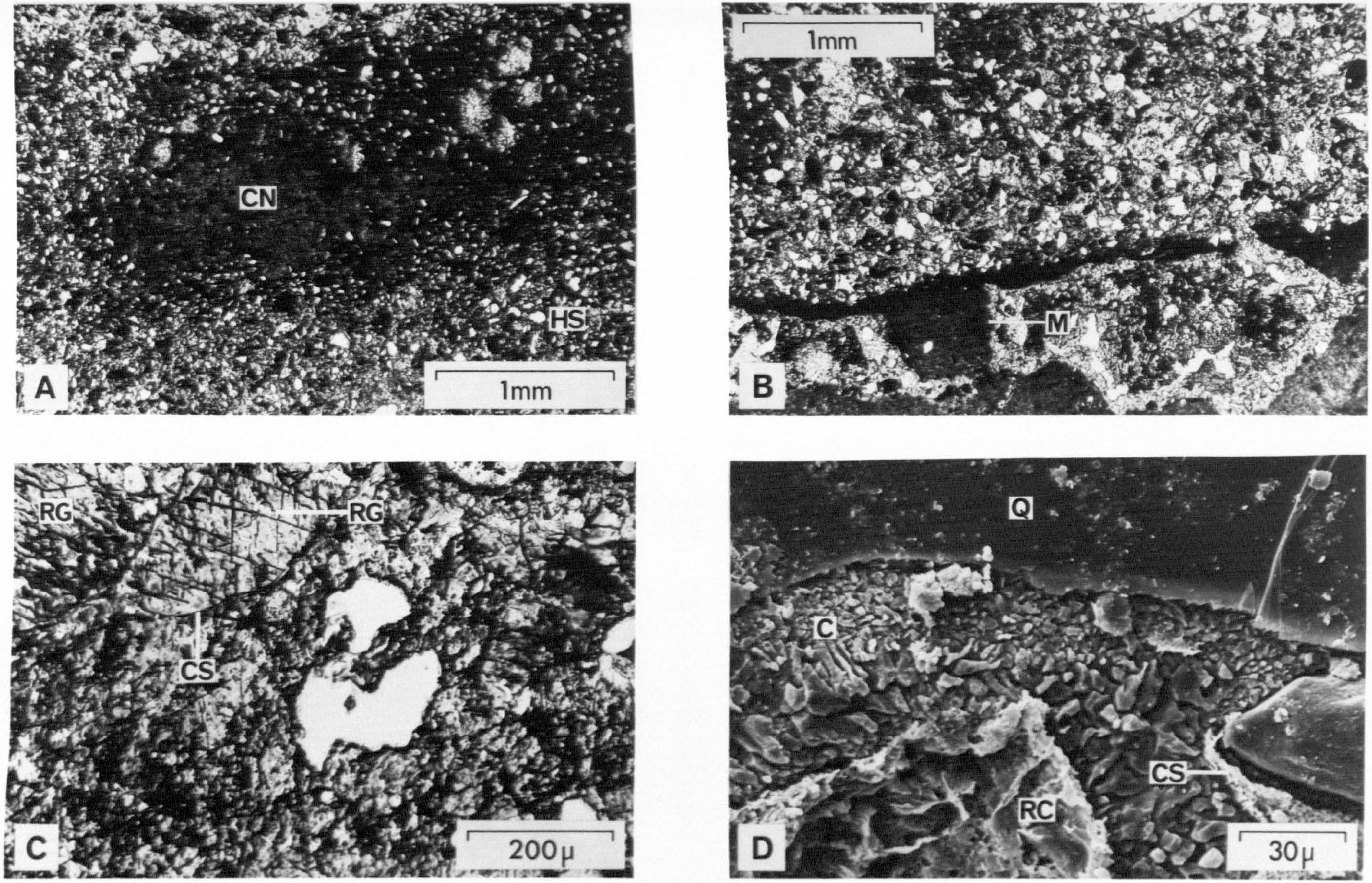

CN
HS
1mm
A
1mm
M
B
RG
RG
CS
200μ
C
Q
C
CS
RC
30μ
D

Plate 4.6

A. Elongated carbonate crystals orientated with long axes perpendicular to grains. (SEM photomicrograph)

B. Enlargement of part of A. (SEM photomicrograph)

C. Mechanically infiltrated clay skin (CS). (SEM photomicrograph)

D. Quartz grains (Q), corroded due to carbonate replacement. Note that clay also infills the corrosion indentations on grain surfaces. Photomicrograph also shows clay pseudomorphs (CP) and replaced grains (RG). (Thin section photomicrograph)

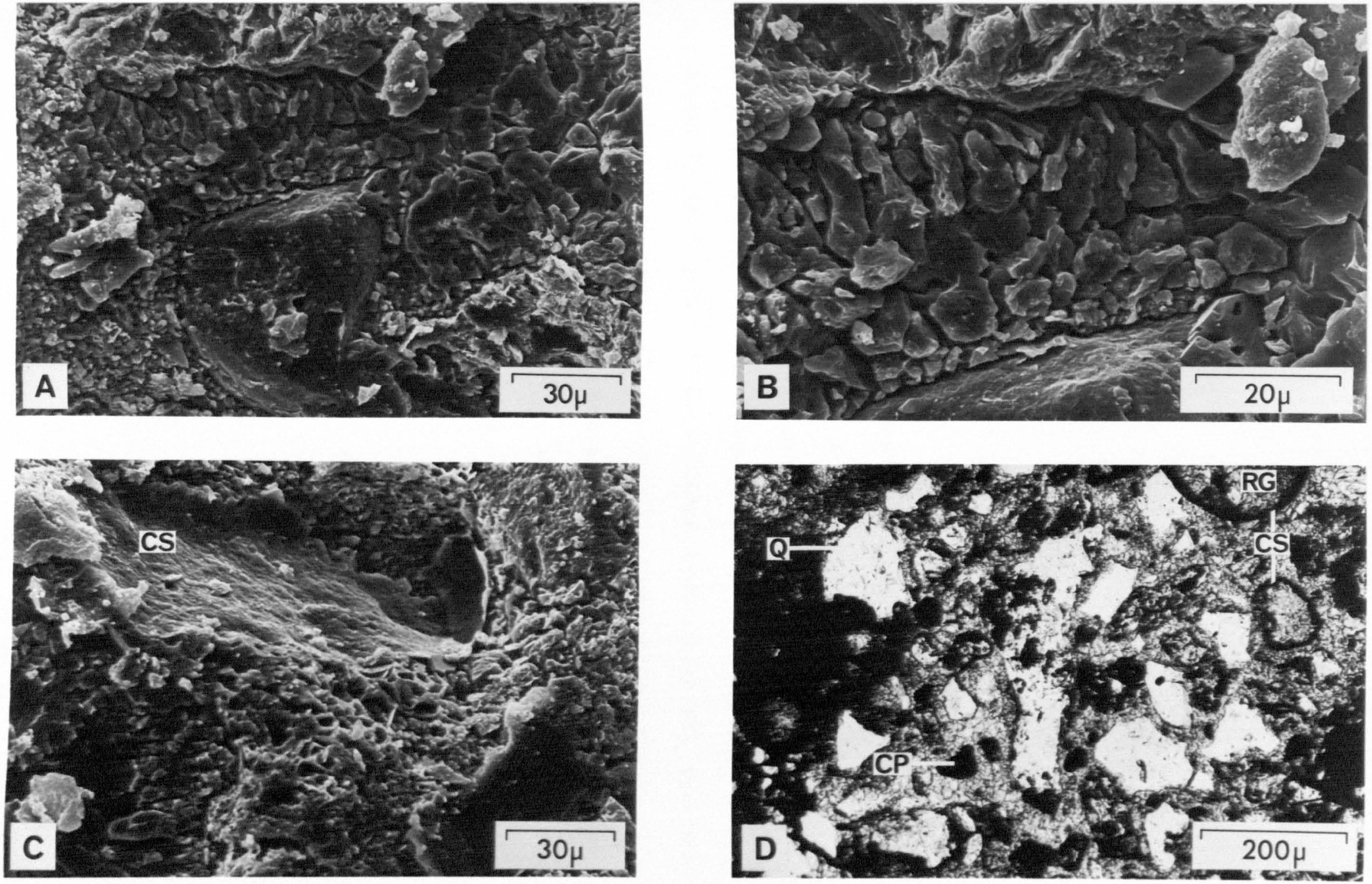

A
30μ
B
20μ
CS
C
30μ
RG
CS
Q
CP
D
200μ

Plate 4.7

Corroded quartz grains.

A. Corroded quartz grains, replaced by clay and micritic carbonate. Note that the detrital grains no longer form the framework of the sediment. (Thin section photomicrograph)

B. Enlargement of part of A. (Thin section photomicrograph)

C. Relict feldspar grain (F) and corroded quartz grain (Q), partially replaced by granular carbonate (M). (SEM photomicrograph)

D. Enlargement of part of C showing network of replacement pits (RP) and replacement carbonate (C), characteristic of corroded quartz grains. (SEM photomicrograph)

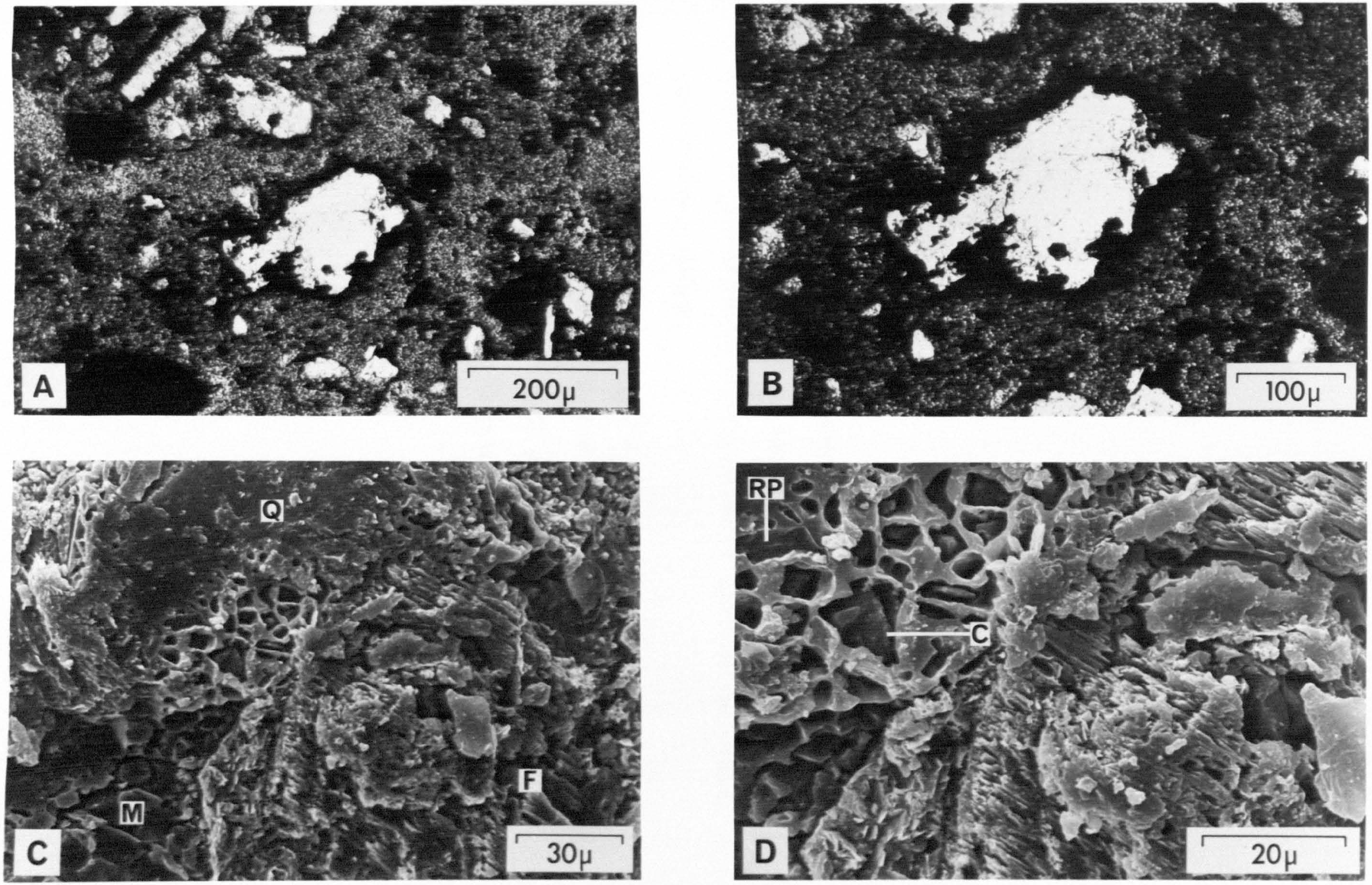

A
200μ
B
100μ
Q
M
F
C
30μ
RP
C
D
20μ

Plate 4.8

Corroded quartz grains.

A. Corroded quartz grain (Q). Note the grain above, partially enclosed by a clay skin (CS) which has been lifted from the surface of the grain by the formation of granular carbonate. (SEM photomicrograph)

B. Enlargement of part of A. (SEM photomicrograph)

C. Enlargement of replacement pits (RP) and included carbonate crystals (C), seen in B. (SEM photomicrograph)

D. Peripheral replacement of quartz grain by carbonate, but also note high proportion of clay in matrix. (SEM photomicrograph)

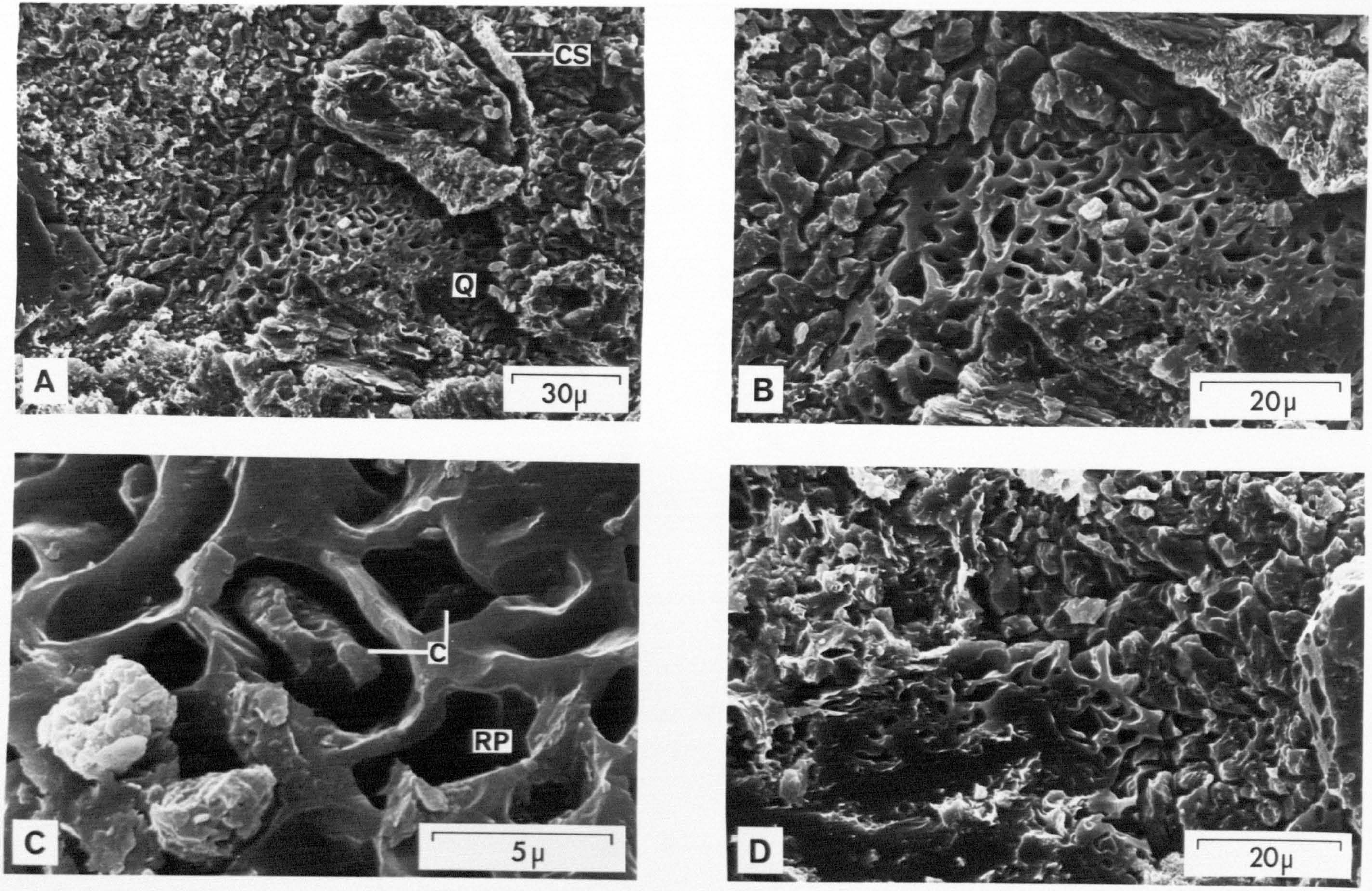

CS
Q
A
30µ
B
20µ
C
RP
C
5µ
D
20µ

Plate 4.9

Corroded quartz grains, replaced by carbonate.

A. Replaced quartz grain. (SEM photomicrograph)

B. Enlargement of part of A. (SEM photomicrograph)

C. Replaced quartz and feldspar (F) grains. (SEM photomicrograph)

D. 'Skeletal' quartz grain, showing advanced replacement. (SEM photomicrograph)

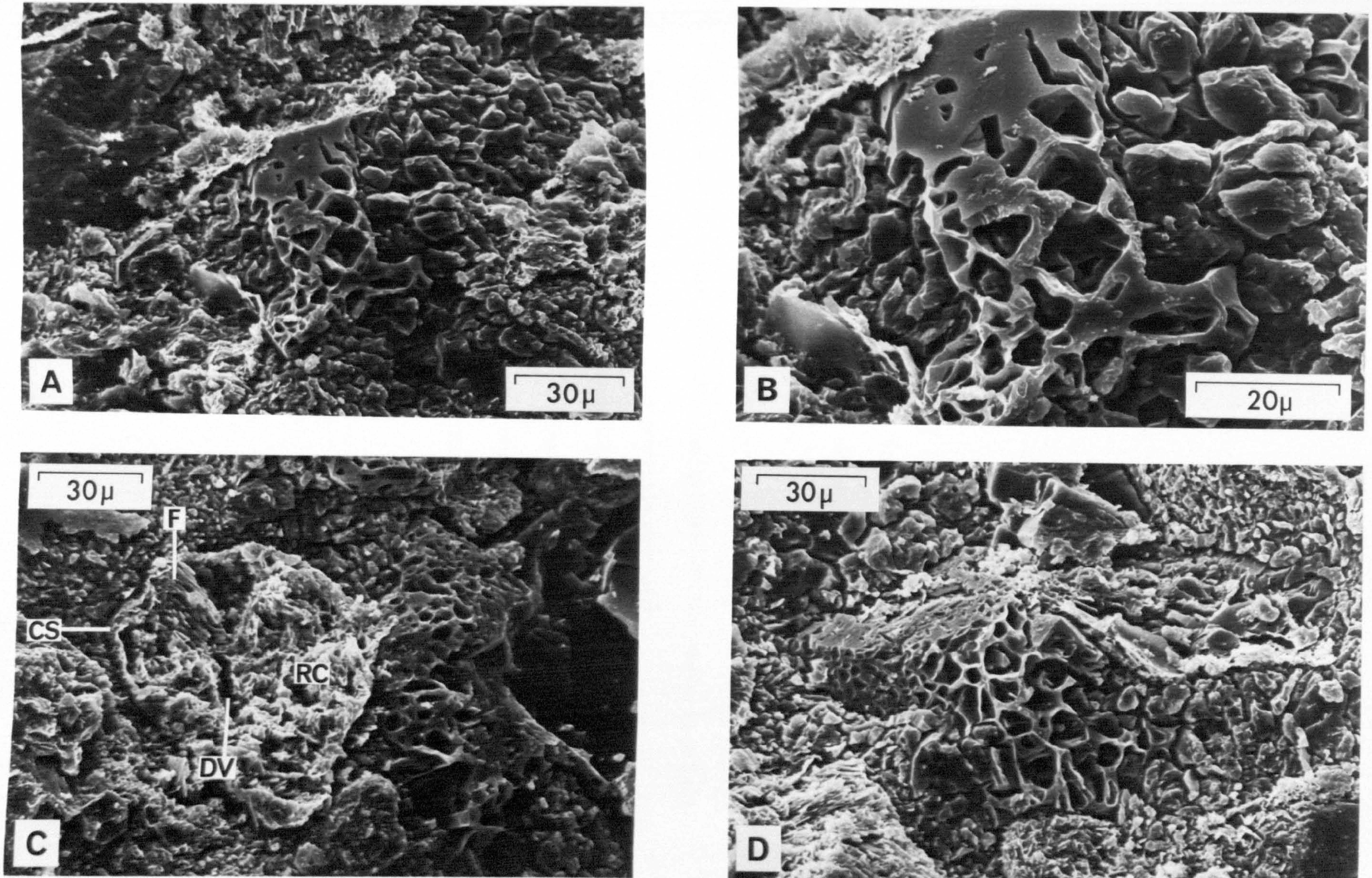

A
30µ
B
20µ
C
30µ
F
CS
RC
DV
D
30µ

Plate 4.10

Replaced quartz grains.

A. Replacement of quartz grain by a radial growth of elongated carbonate crystals. Note the localised nature of the replacement and the relative absence of corrosion indentations from the upper surface of the grain. (SEM photomicrograph)

B. Enlargement of part of A. (SEM photomicrograph)

C. Enlargement of part of B showing growth of carbonate crystals into the quartz grain. (SEM photomicrograph)

D. Lower contact of carbonate nodule (CN) accentuated by a clay band and illustrating the textural differences between the floating grains of the nodule and the framework grains of the host sediment (HS). (Thin section photomicrograph)

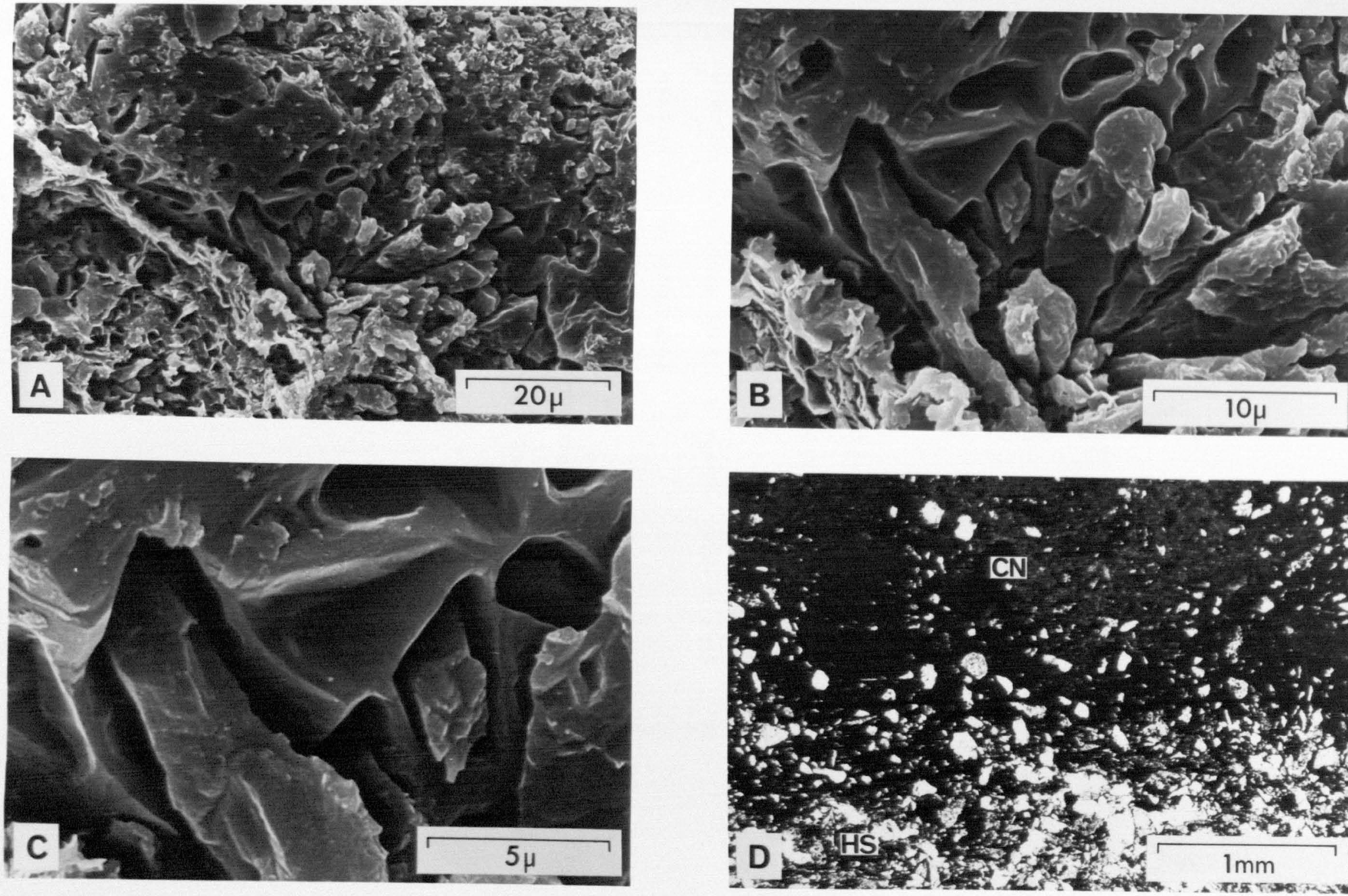

A
20μ
B
10μ
C
5μ
D
CN
HS
1mm

Plate 4.11

A. Fractured carbonate nodule (CN). Fracture is infilled with host sediment (HS). (Thin section photomicrograph)

B. Fractured carbonate nodule (CN). Fracture is lined with micrite with later infill of sparry calcite (S). (Thin section photomicrograph)

C. Distorted sedimentary laminae. Original lamination is horizontal (with respect to the photomicrograph) but, in this instance, has been domed over the carbonate nodules (CN). (Thin section photomicrograph)

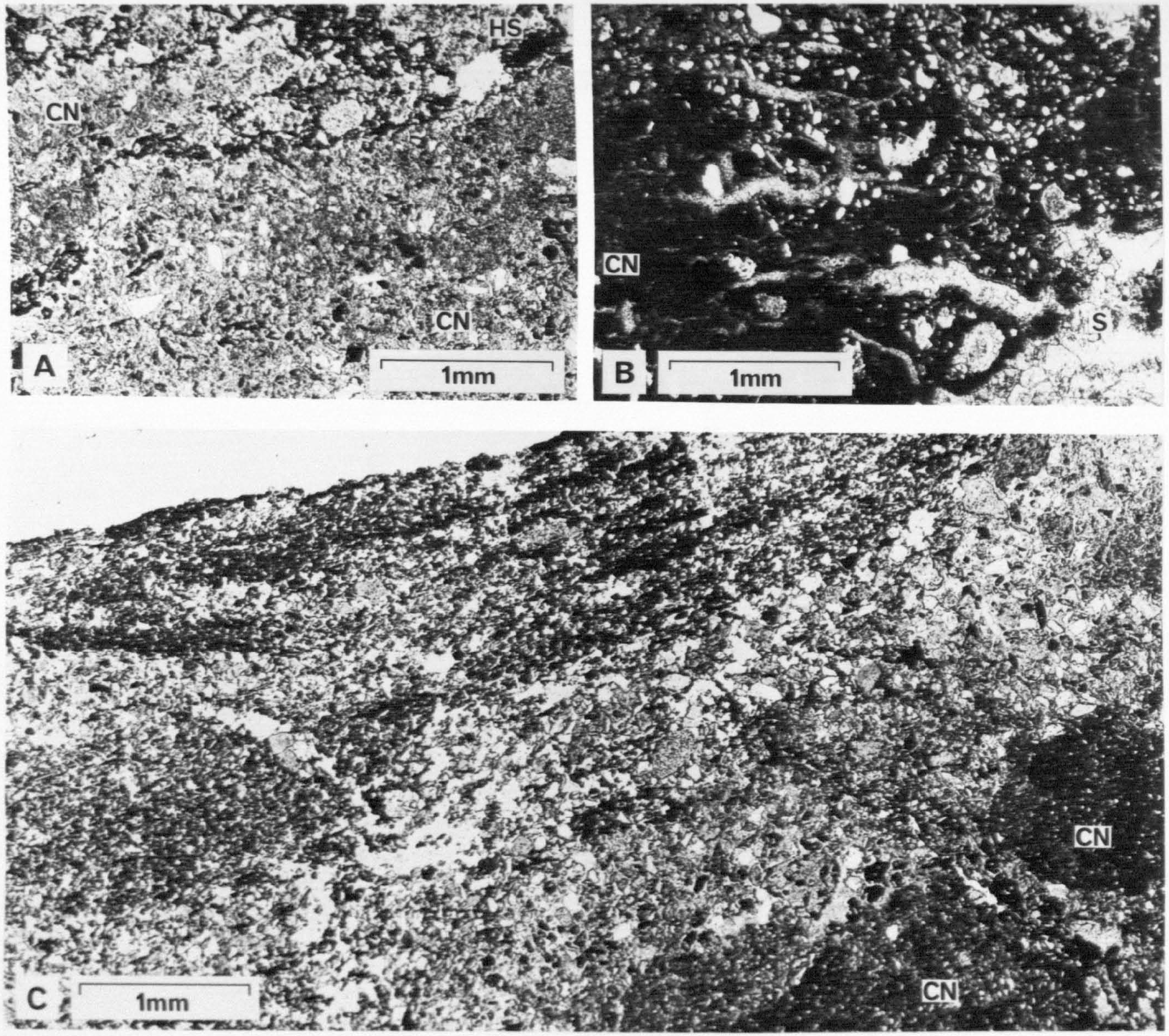

HS
CN
CN
A
1mm
CN
S
B
1mm
CN
CN
C
1mm

Plate 4.12

A. Carbonate nodule (CN) containing distorted clay band. (Thin section photomicrograph)

B. Clay lamination partly contained by carbonate nodule (CN) and partly by original sediment (HS). Approximate margin of the nodule is indicated. Note the deformation of that part of the lamination contained by the nodule. (Thin section photomicrograph)

C. Void, lined by micrite and infilled with sparite. (Thin section photomicrograph)

D. Void, lined by micrite and infilled with sparite. (Thin section photomicrograph)

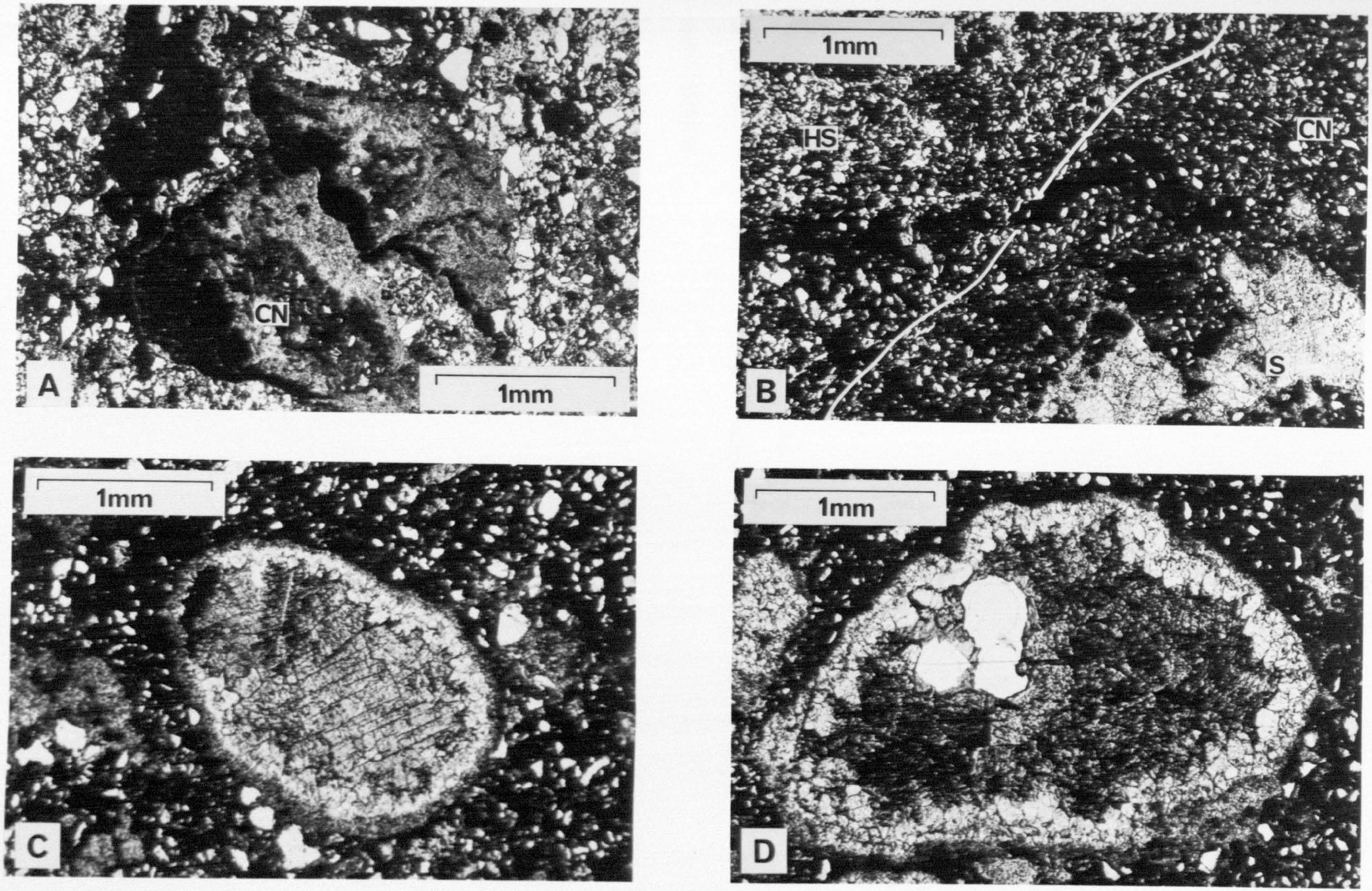

A
CN
1mm
B
1mm
HS
CN
S
C
1mm
D
1mm

Plate 4.13

Secondary quartz overgrowths. B is an enlargement of A. C and D show successive growth layers of secondary quartz. (SEM photo-micrographs)

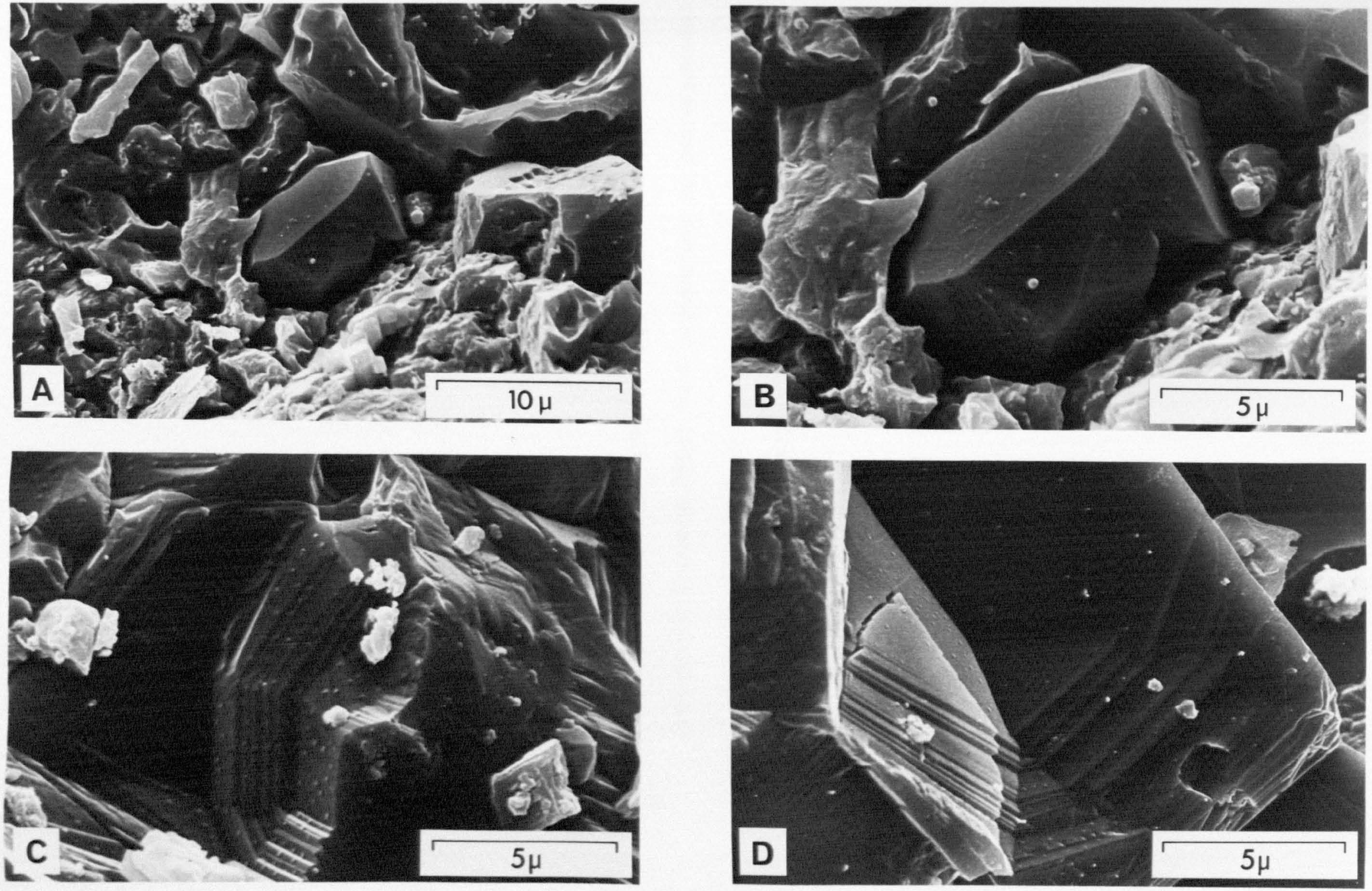

A
10 μ
B
5 μ
C
5 μ
D
5μ

Plate 4.14

Replaced secondary quartz overgrowths. B and D are enlargements of A and C respectively. Surfaces of the overgrowths are characterised by replacement pits (RP), commonly containing carbonate crystals (C). (SEM photomicrograph)

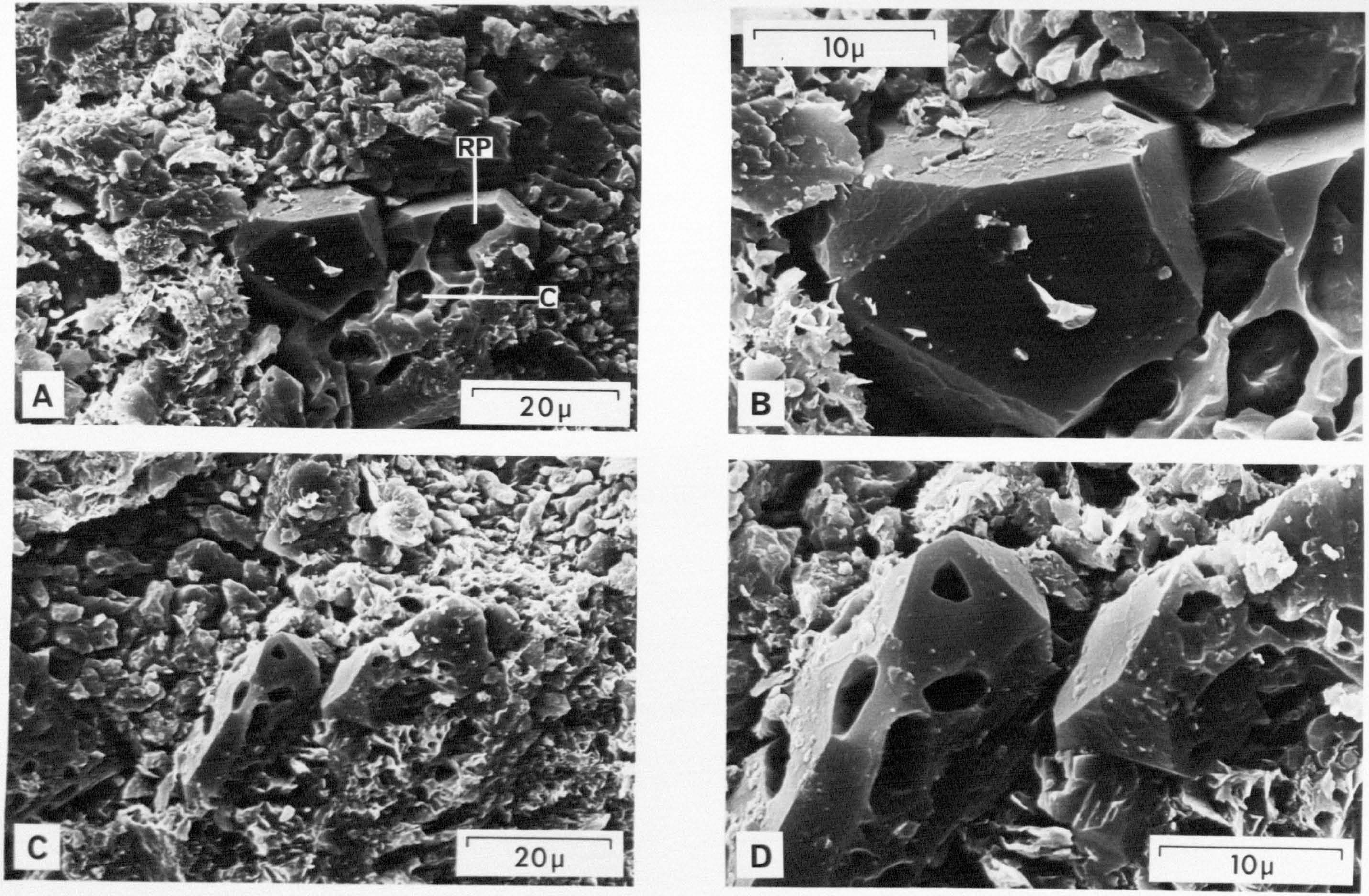

A
RP
C
20µ
B
10µ
C
20µ
D
10µ

Plate 4.15

Replaced secondary quartz overgrowths. D is an enlargement of C.

(SEM photomicrographs)

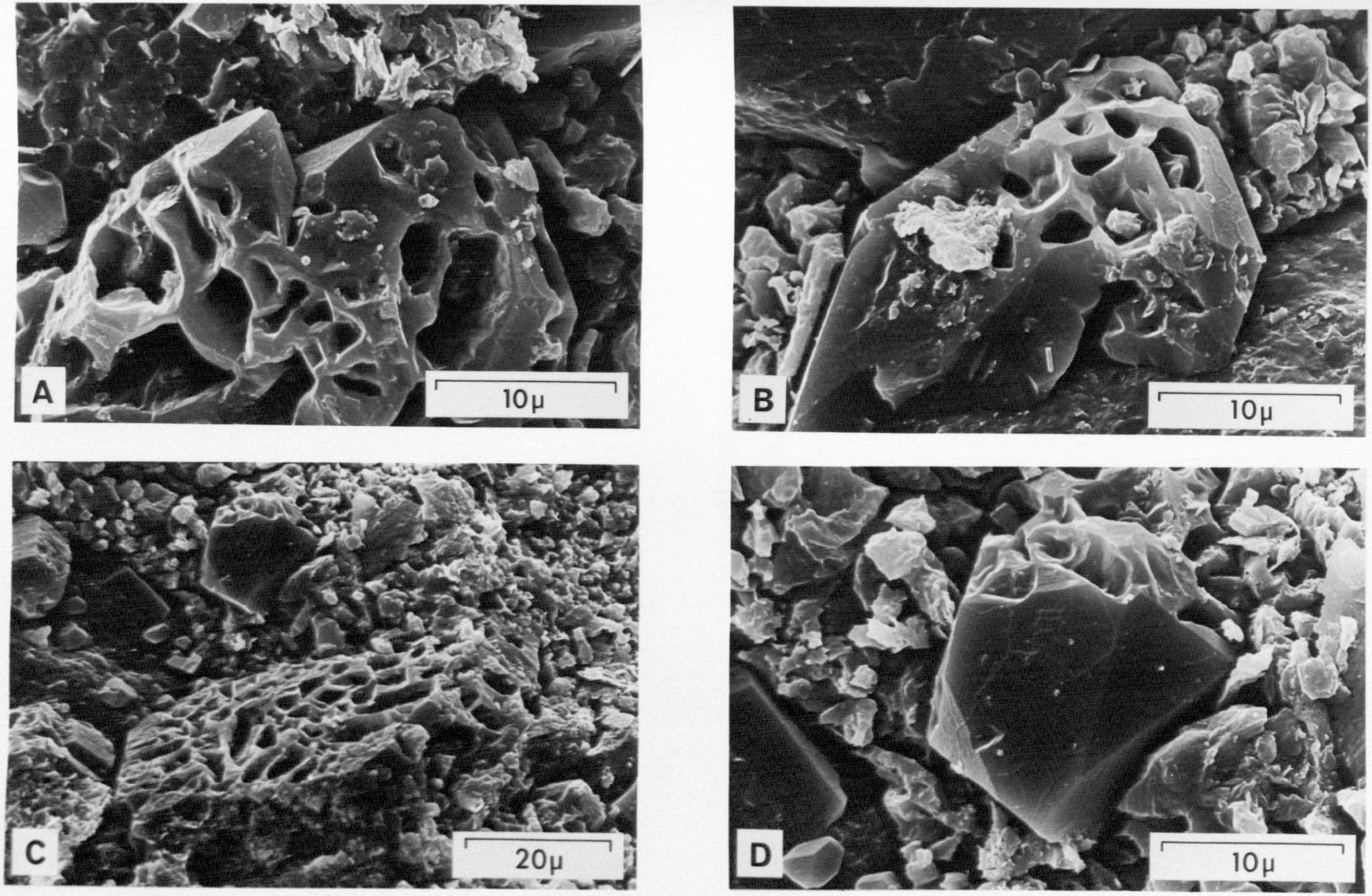

A
10μ
B
10μ
C
20μ
D
10μ

Plate 4.16

Replaced secondary quartz overgrowths. B and D are enlargements of A and C respectively. (SEM photomicrographs)

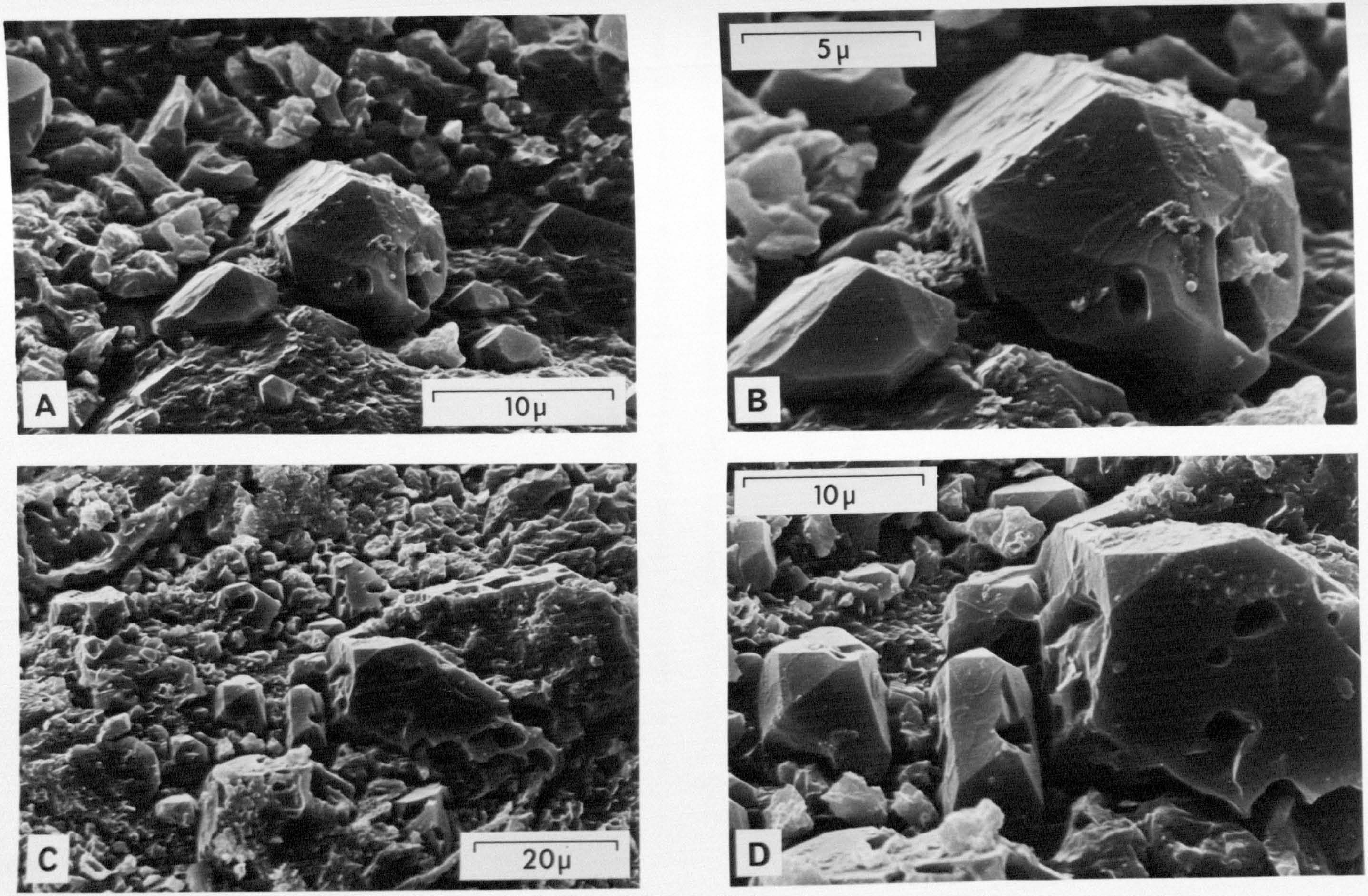

A
10μ
B
5μ
C
20μ
D
10μ

Plate 4.17

Replaced secondary quartz overgrowths. D is an enlargement of C. Photomicrographs B - D are cross-sections of grains in which the euhedral outline of the overgrowth is clearly visible. The outer margins of the grains are extensively replaced, particularly in C and D where replacement is almost exclusively confined to the overgrowth. (SEM photomicrographs)

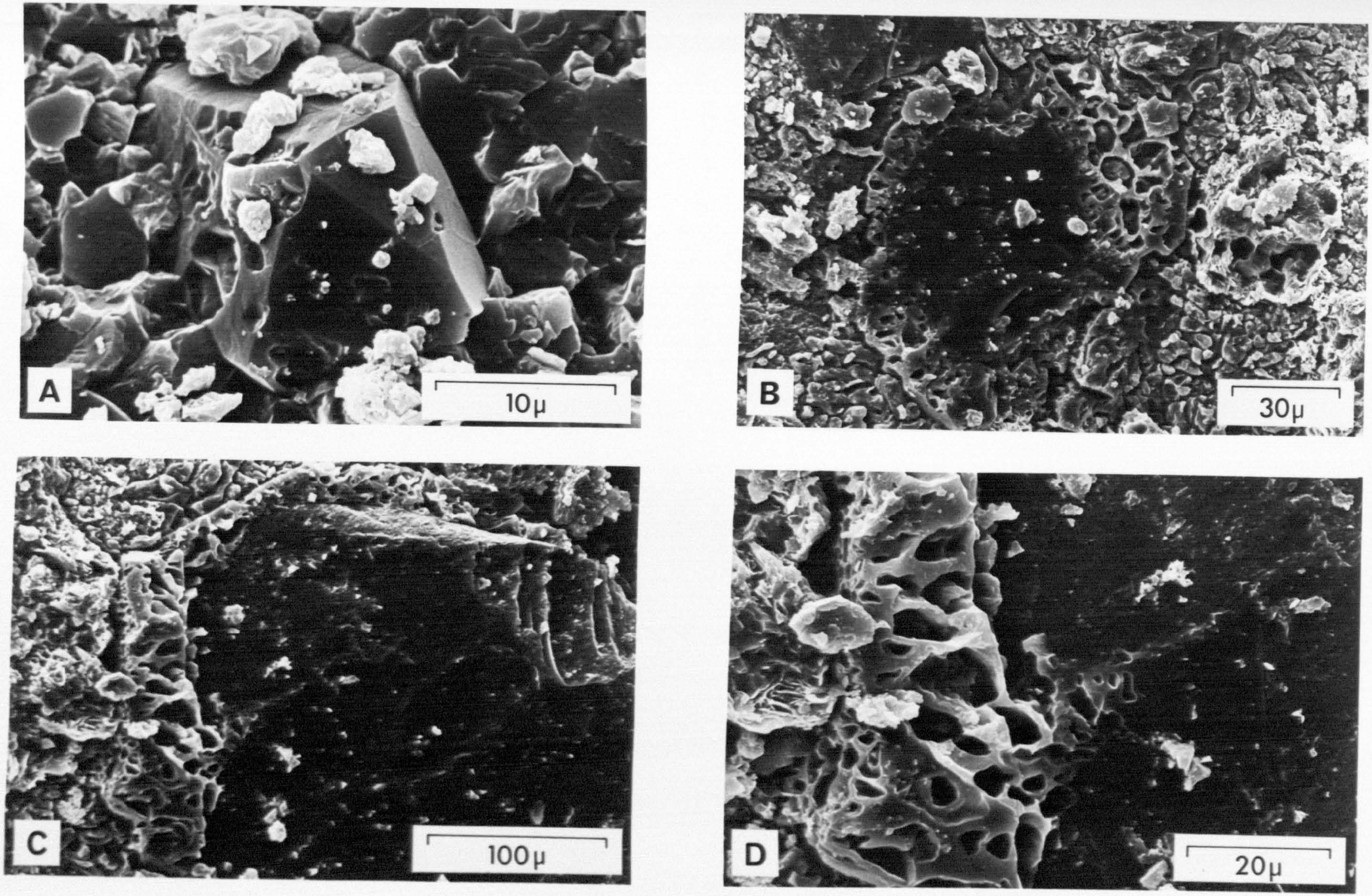

A
10μ
B
30μ
C
100μ
D
20μ

Plate 5.1

A. Smithy Wood Quarry, NY529348.

B. Baronwood Quarry, NY515428.


A
B

Plate 5.2

Whelpdale Hill Quarry, NY517338.

Plate 5.3

A. Bowscar Quarry, NY520343.

B. Bowscar Working Quarry, NY520344.


A


B

Plate 5.4

Stoneraise Quarry, NY533352.

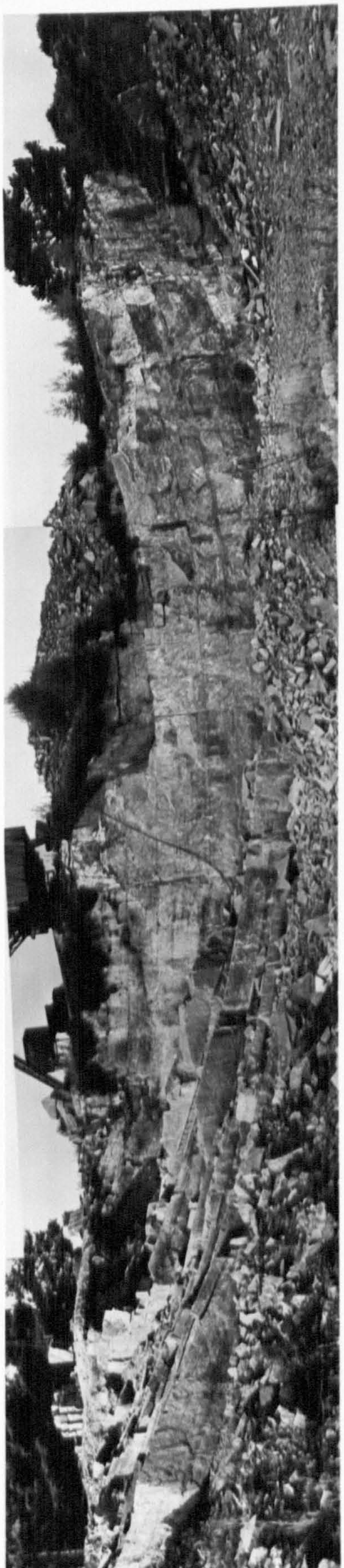


Plate 5.5

Stoneraise Quarry, NY533352.

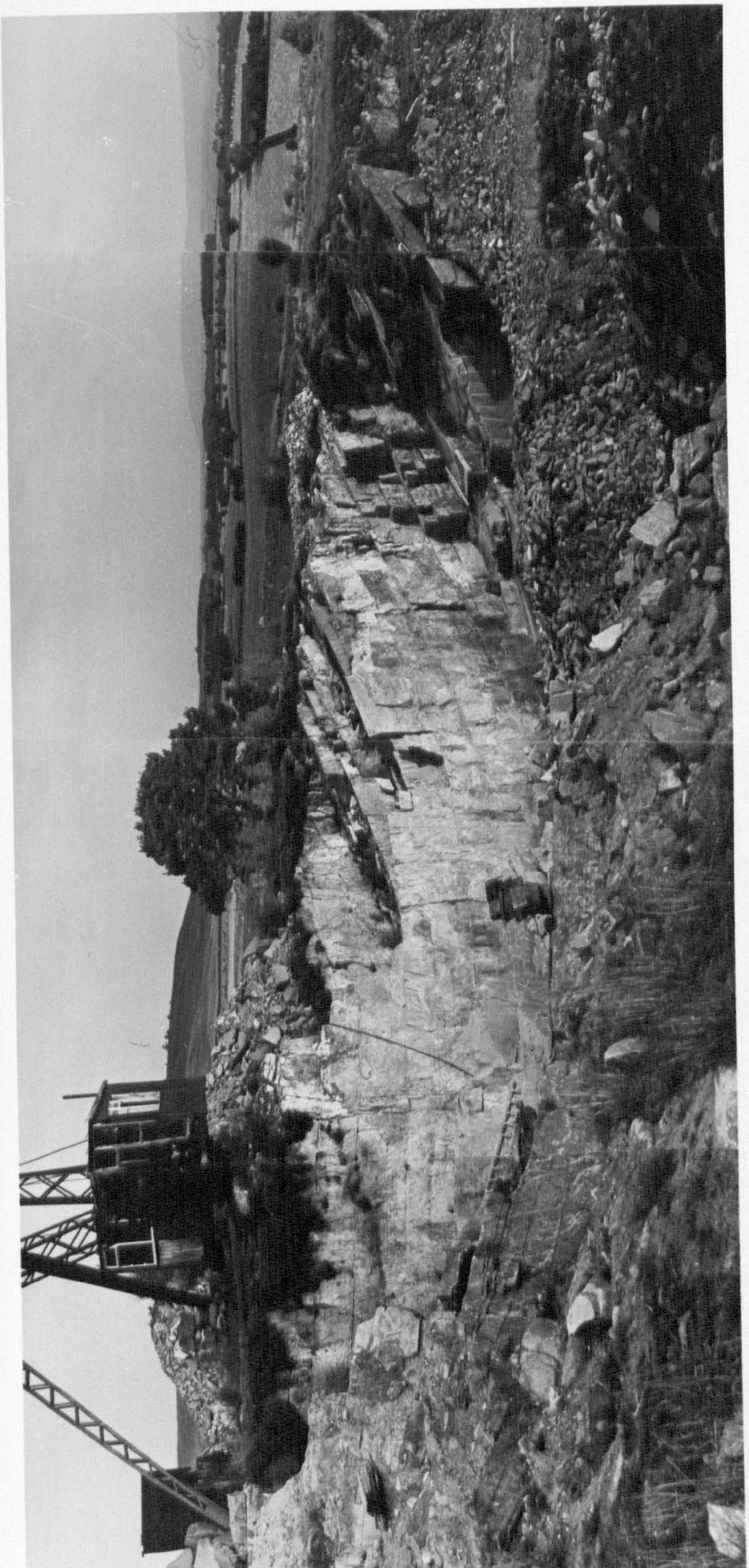


Plate 5.6

Features of the Penrith Sandstone exposed in Stoneraise Quarry, NY533352.

A. Minor variations in foreset inclination angle.

B. Primary lineation (avalanche slope lineation) on the upper surface of a cross-bedding plane. Crow-bar indicates orientation of striae.

C. As above.

D. Elongate ?scour structures on the upper surface of a cross-bedding plane. Hammer handle indicates direction of maximum depositional dip.

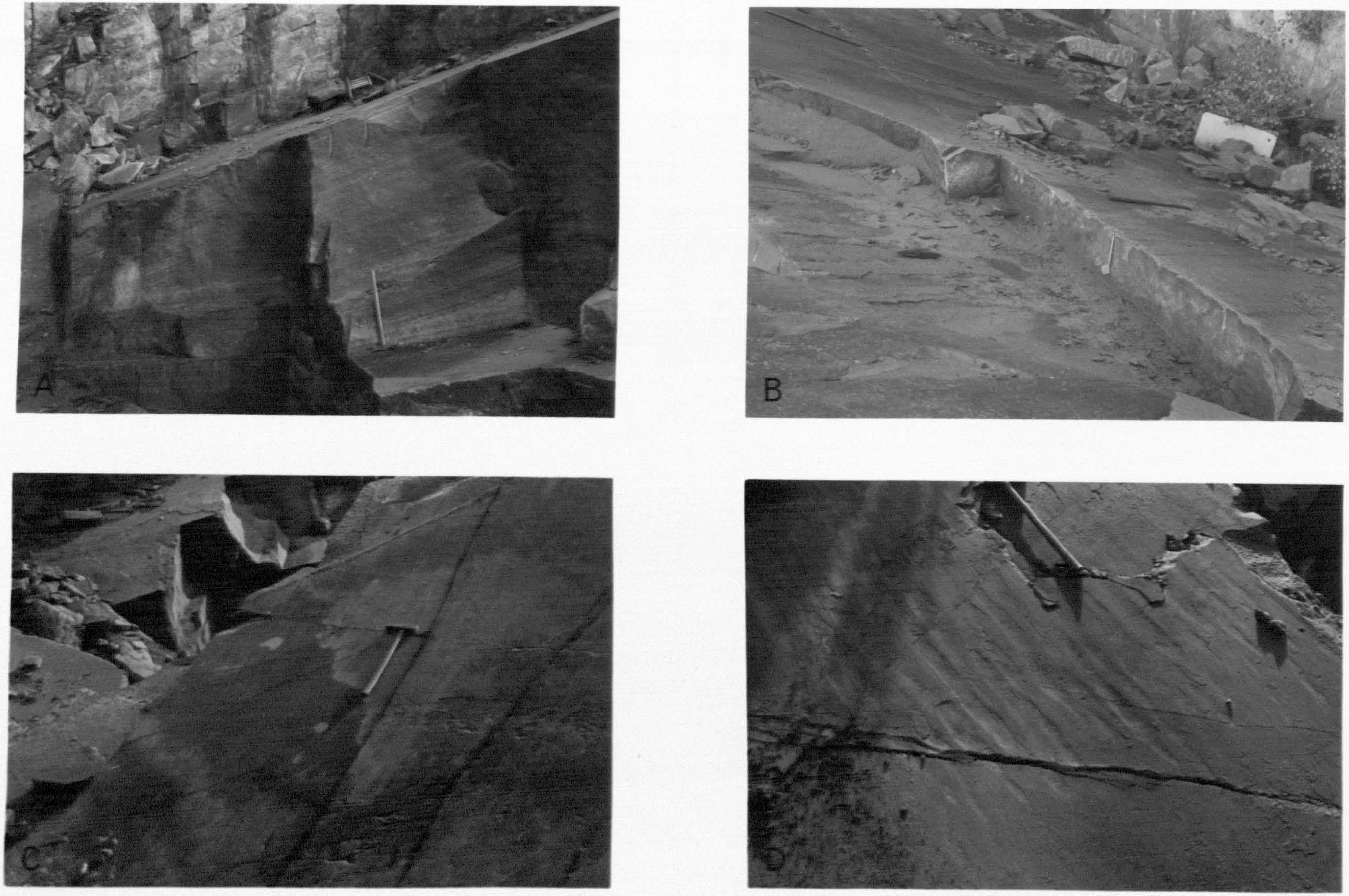

A
B
C
D

Plate 5.7

Lazonby Railway Cutting, NY544406.

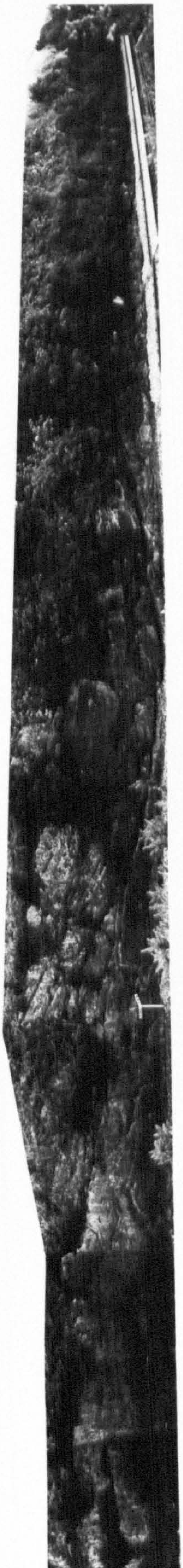
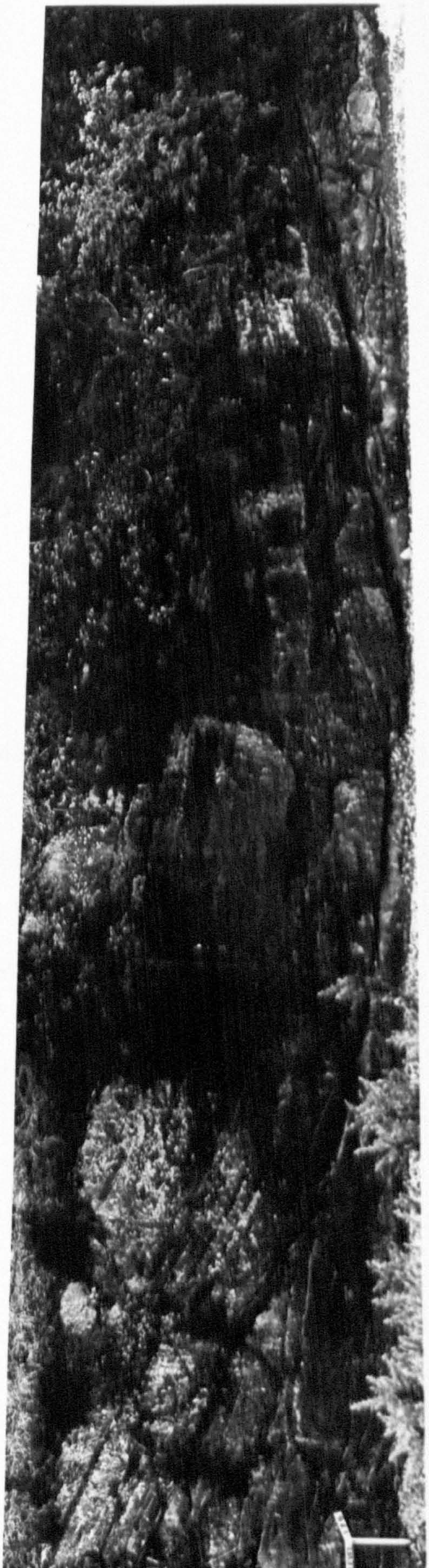



Plate 5.8

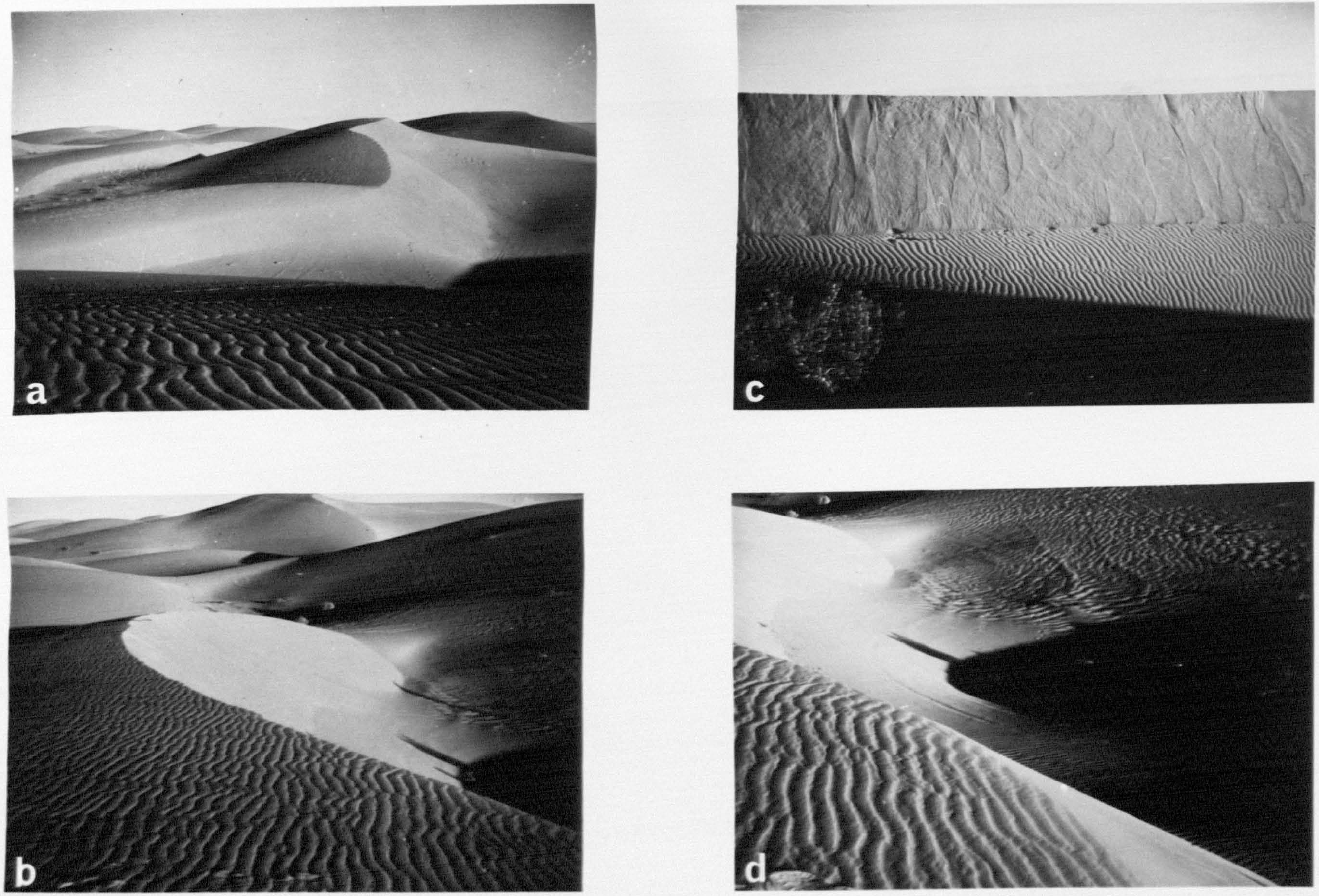
Dunes from the Algodones dune field, California (photographs supplied by B. Waugh).

A. General view of dune field.

B. Stoss slope wind ripples truncated by crescentic, asymptotically based dune slipface.

C. Angularly based slipface advancing over wind ripples formed in the lee of the dune.

D. Orientation of wind ripples formed in the lee of a dune.


a
b
c
d