KARST PALAEOENVIRONMENTS: A RECONSTRUCTION WITH PARTICULAR REFERENCE TO THE MORECAMBE BAY AREA

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DECLARATION

I hereby declare that the work presented in this thesis is my own, unless otherwise stated in the text, and that the thesis has been composed and written by myself.

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ABSTRACT

At least four phases of karst development can be recognised in the Morecambe Bay area of northwest England : an intra-Carboniferous phase resulting from intermittent sub-aerial erosion associated with temporary falls in the level of the Lower Carboniferous sea; a Permo-Carboniferous phase probably characterised by streams from adjacent areas of impermeable rock sinking at the limestone boundary and forming swallet-type cave systems; an ?Oligo-Miocene phase of interstratal karstification; and a Late Quaternary phase of karst development. The caves of the Late Quaternary phase contain deposits of Late and Post-glacial age, although palaeomagnetic evidence suggests that certain of the deposits may be somewhat older. Thus, the bulk of these caves may be of at least last interglacial age.

It is possible to reconstruct the former pattern of drainage through the now-abandoned caves and to show that the caves probably date from more than one phase of hydrological development. Nevertheless, the results of earlier attempts to date former episodes of hydrological development by matching the altitudes of caves with those of former base levels in the area are shown to be invalid.

A semi-quantitative picture of the palaeohydrology of certain karst drainage systems in the area is provided using evidence obtained from bedform flow-features and hydraulically-transported deposits in the caves. By means of sedimentological analysis, it is also possible to reconstruct the hydraulic conditions under which former flows occurred.

The surface topography of the area, previously regarded as largely a response to episodes of planation associated with falling sea levels, is studied. No evidence, neither of a morphological nor of a sedimentological nature, is found to support the idea of marine planation. Finally, some of the characteristic surface karst features of the area are described, including large-scale closed depressions, cemented screes, and features regarded by earlier workers as poljes.

Notation

- A = catchment area (L^2)
- a = cross-sectional area of channel flow (L^2)
- B_L = channel-bed roughness constant
- C = Chézy coefficient ($L^{\frac{1}{2}}\ T^{-1}$)
- $C/\sqrt{g}$ = dimensionless Chézy coefficient
- D = particle diameter (L)
- D crit = critical test statistic in the Kolmogorov-Smirnov test
- D max = maximum difference between observed and expected values in the Kolmogorov-Smirnov test.
- D_n = particle diameter at which n% of the grain-size distribution (by weight) is coarser (L).
- D test = calculated test statistic in the Kolmogorov-Smirnov test.
- d = flow depth (L)
- F = Froude number
- f = Darcy-Weisbach friction factor
- g = gravitational acceleration (9.8 m s^{-2}) ($L\ T^{-2}$)
- H = a channel dimension (L)
- K_n = constant n
- L = scallop wavelength (L)
- Lc = length of channel within a single meander wave (L)
- L w = Meander wavelength (L)
- Md = median grain-size (D_{50}) (L)
- P = power of stream flow per unit area of bed ($M\ L^{-2}\ T^{-3}$)
- Pd = index of roundness (Wadell, 1932; 1933)
- Q = discharge ($L^3\ T^{-1}$)
- Qb = bankfull discharge ($L^2\ T^{-1}$)
- QDa = quartile deviation ($((D_{25} - D_{75})/2)$) (L)
- Qma = mean annual flood ($L^2\ T^{-1}$)
- Q mean = mean discharge ($L^2\ T^{-1}$)
- Q max = high-stage discharge ($L^3\ T^{-1}$)

Q_{mm}	=	mean maximum monthly discharge ($L^3\ T^{-1}$)
Q unit	=	discharge per unit width of channel ($L^3\ T^{-1}$)
Q_{40}	=	discharge of 40% exceedance probability per annum ($L^3\ T^{-1}$)
Re	=	channel Reynolds number
Re_L	=	mean flow-velocity stable-scallop Reynolds number
Re*	=	bed shear-velocity stable-scallop Reynolds number
S	=	slope of the channel energy grade line
Sc	=	Schmidt number
Se	=	standard error
Sh	=	Sherwood number
So	=	grain-size sorting ($\sqrt{D_{25}/D_{75}}$) (Trask, 1930; 1932)
t	=	temperature
u	=	flow velocity ($L\ T^{-1}$)
u_{max}	=	maximum flow velocity ($L\ T^{-1}$)
u_y	=	flow velocity at distance y from the channel bed ($L\ T^{-1}$)
$\bar{u}$	=	mean flow velocity ($L\ T^{-1}$)
u'	=	upward vertical flow-velocity fluctuations near the channel bed ($L\ T^{-1}$)
u*	=	bed shear-velocity ($L\ T^{-1}$)
Va max	=	maximum coefficient of variation of slope profile segments in Best Units analysis (Young, 1971)
Vc max	=	maximum coefficient of variation of slope profile elements in Best Units analysis (Young, 1971)
W	=	channel width (L).
$\bar{x}$	=	mean grain-size ($2^{-(25\phi + 50\phi + 75\phi)/3}$ mm) (L)
α	=	statistical-test significance level
η	=	diffusivity of solute ions in solution ($L^2\ T^{-1}$)
θ	=	dimensionless boundary shear stress ($\tau/(\rho s-\rho f)gD_{50}$)
μ	=	fluid dynamic viscosity (1.3×10^{-3} N s^{-1} m^{-2} for pure water at 10^{o}C($M\ L^{-1}\ T^{-1}$)

ν	=	fluid kinematic viscosity (1.3×10^{-6} m^2 s^{-1} for pure water at 10°C) (L^2 T^{-1}).
ρf	=	fluid density (1000 kg m^{-3} for pure water at 10°C) (M L^{-3}).
ρs	=	sediment density (2.65×10^3 kg m^{-3} for quartz) (M L^{-3}).
σ'	=	sample standard deviation.
τ	=	mean boundary shear stress (M L^{-1} T^{-2}).
τ_{crit}	=	critical erosion boundary shear stress (M L^{-1} T^{-2}).
ω	=	particle settling velocity in a fluid (L T^{-1}).

CONTENTS

I : INTRODUCTION

1. INTRODUCTION

The study of karst palaeoenvironments may be regarded as including both an investigation of the conditions under which now-relict karst landscapes developed and a consideration of the development of specific components of the karst environment. The especial value of karst for this type of environmental reconstruction lies mainly in its ability to preserve evidence of past phases of environmental history. Two aspects of karst morphology, in particular, facilitate the preservation of palaeoenvironmental records. Firstly, the characteristic system of karst drainage through underground fissures and conduits tends to be preserved when it is abandoned, rather than being subject to erosion and reworking as in the case of surface drainage features. Secondly, these conduits, and, to a lesser extent, surface closed depressions, tend to act as sediment traps within which long sequences of deposits may be preserved as indicators of past environmental conditions.

The present study considers both these factors in an investigation of environmental change in the karst area surrounding Morecambe Bay in northwest England (Fig. 1.1). The karst area itself is composed of an almost continuous belt of Lower Carboniferous Limestone fringing the southern edge of the Lake District and extending from Millom in the west to the Lune Valley in the east. The limestone is traversed by a set of N-S faults with the result that the area is characterised by a series of upstanding limestone blocks (Fig. 1.1). Within the Low Furness and Millom districts, most of the karst is blanketed by glacial deposits,

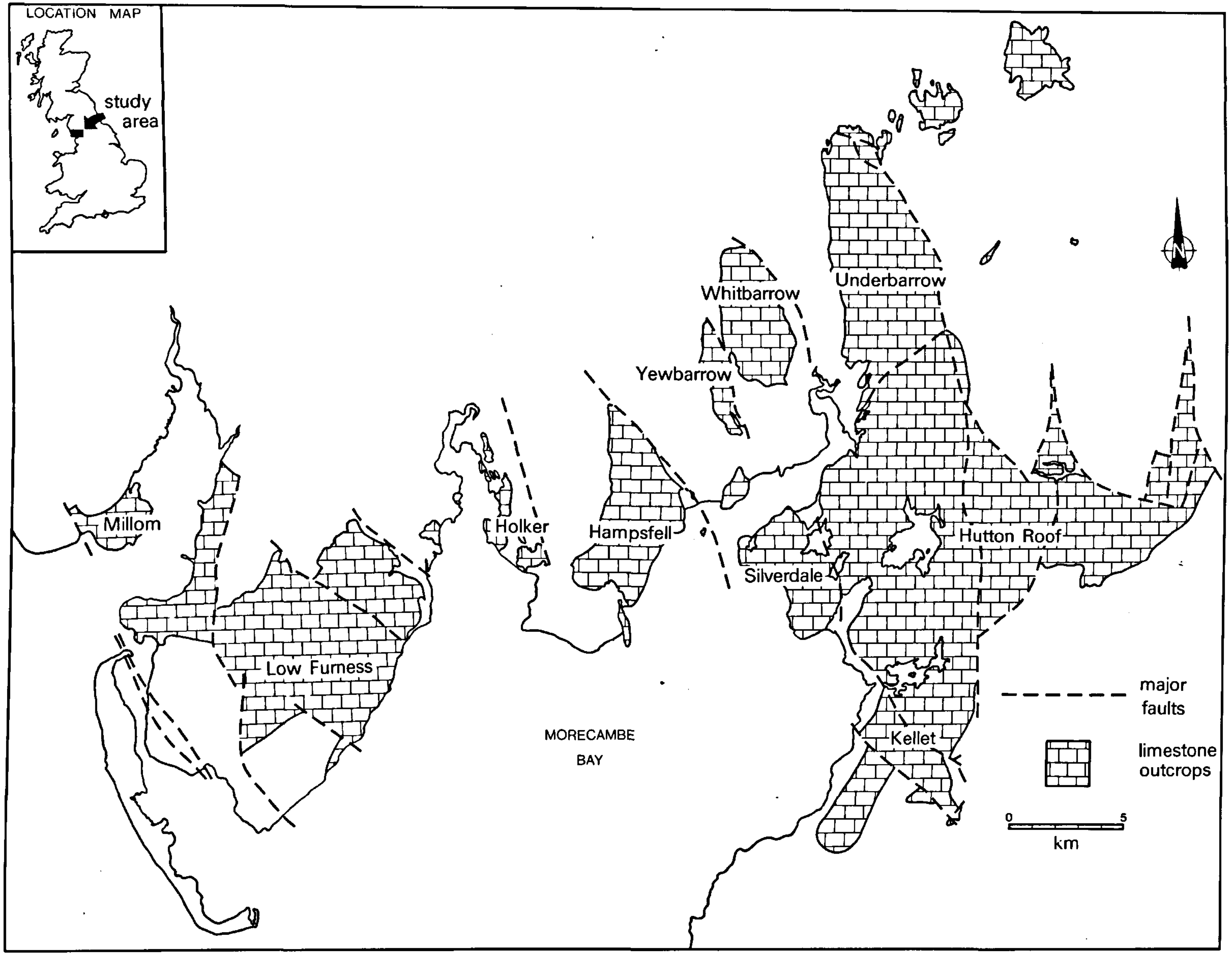


Fig. 1.1 The Morecambe Bay karst

and the same is true of the Kellet area. Elsewhere, however, much of the limestone is exposed, revealing the most distinctive landscape features of the whole area (Plate 1.1).

Evidence of past phases of karstification may be found in every karst area in the British Isles, but rarely is the history of landscape development as lengthy or as complex as that found in the Morecambe Bay area. There is a long record of cave investigation in the area, beginning with the antiquarians and early tourists of the late eighteenth and nineteenth centuries (Ashmead, 1973). However, with the exception of Ashmead, who studied the chronology of cave development, the area has been largely ignored by modern investigators. Studies in two other fields are, nevertheless, of particular relevance: the haematite deposits of the area have provided a fruitful source of information on the palaeokarst features with which they are intimately associated, whilst the mosses of the area have proved to be valuable sites for the palynological investigation of Post-glacial environmental change. In general, however, little palaeoenvironmental research has been undertaken, and yet, the Morecambe Bay area, as Sweeting (1974, 78) has pointed out, ".... could be a significant one in the history of climatic change in Britain". The present study applies a variety of interpretative techniques to particular facets of the karst palaeoenvironment of the Morecambe Bay area, and undertakes a clarification of the complex environmental history of the area as a whole.

Plate 1.1 Whitbarrow, one of the fault-bound limestone blocks characteristic of the Morecambe Bay karst

II : PRE-QUATERNARY ENVIRONMENTAL HISTORY

2. TECTONIC HISTORY AND STRUCTURE

2.1 Introduction

Two major processes have been involved in the structural development of northwest England. On a local scale, there has been almost continuous isostatic adjustment of the massifs and basins which characterise the area. On a continental scale, on the other hand, the more important process has been the intermittent orogenic phases resulting from plate movements.

2.2 Isostatic adjustment of massifs and basins

In structural terms, northwest England is composed of three main massif units (Fig. 2.1). Of these, one is centred beneath the Lake District (Bott, 1974) and one beneath the Alston Block (Bott and Masson-Smith, 1957; Dunham, Dunham, Hodge and Johnson, 1965), whilst geophysical evidence suggests that a batholith-like body lies beneath the southern part of the Askrigg Block (Bott, 1967; Myers and Wardell, 1967). The three major massifs are separated from each other by basin structures, which Ashton (1971, 15) termed the Northumberland Trough, the Penrith Trough, the Barnard Castle Trough, and the Kendal Depression (Fig. 2.1). There is also a major basin unit within the Irish Sea which seems to comprise two distinct basins, the Solway-Peel Basin and the East Irish Sea Basin, separated by a further block, the Man Ridge (Bott, 1968; Bott and Young, 1970; Hallam, 1972; 1973; Hall and Smythe, 1973; Dobson, 1977) (Fig. 2.1).

The low density granitic masses of the blocks were intruded either during the Precambrian, in the case of the Askrigg Block, or during the post-Caledonian maximum in early Devonian times, in the case of the Lake District and the Alston

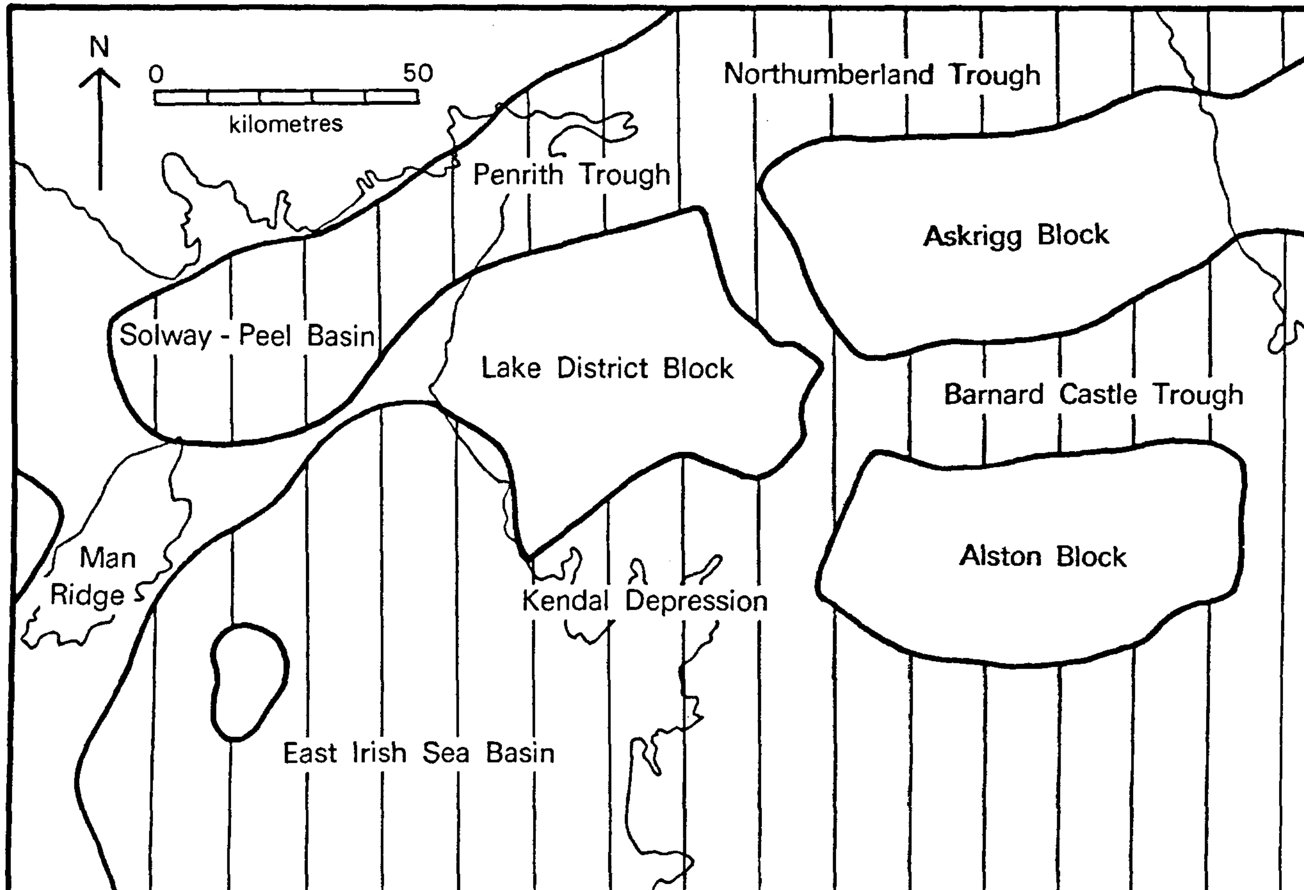


Fig. 2.1 The basement structure of northwest England (see text for sources of information)

Block. These areas have therefore been characterised by isostatic uplift from at least Devonian times to the present day. During this period they have continued to be positive areas, often of high topography (Johnson, 1967, 179-180). The basins are generally regarded as being similarly long-lived phenomena. Bott (1968, 113-114) and Dobson (1977, 14) considered that much of the present topography of the northern Irish Sea depends directly or indirectly on crustal structures of Caledonian or earlier age, and that the present coastline generally lies close to the edge of the Irish Sea basin units. By contrast, Hallam (1972, 175; 1973, 301-302) felt that although the Irish Sea basins were initiated soon after the Hercynian, there is no good reason for invoking subsequent, almost continuous, subsidence, nor to seek a direct relationship between more recent large-scale structures and those existing in the past.

2.3 Orogenic movements

Superimposed on the isostatic movements detailed in 2.2, and to some extent initiating movement along the massif and basin boundaries, were large-scale orogenic movements.

The earliest tectonic phase whose effects can be traced in northwest England is the Caledonian. The Caledonian period is taken to include all significant tectonic movements from the start of the Cambrian until the end of the Devonian. By the end of the Devonian, movements attributable to the Hercynian period had begun, and the diminishing pulses of the Caledonian are equally well regarded as the anticipatory pulses of the Hercynian.

Moseley (1972) recognised three main periods of tectonic unrest during the Caledonian phase in northwest England. During the first, the Pre-Borrowdale Volcanic phase, folding occurred along N-S and NW-SE axes. The second phase, the Pre-Caradoc, saw renewed movements along the same trend lines. During the third phase, the End-Silurian, similar movements occurred as in the early stages, but the movements regarded by Moseley (1972, 582) as constituting the principal Caledonian movements involved folding along ENE-WSW axes and more localised E-W cleavage. The resultant Caledonian structures are summarised by Moseley (1972, 587) as predominantly grouped into three orientations: NW-SE (dextral wrench), NE-SW (cleavage), N-S (sinistral wrench). These orientations are also reflected in the massif and basin structure of the Caledonian basement (Fig. 2.1).

In northern England, several phases of deformation have been recognised between Lower Carboniferous and Permian times, which have been attributed to the Hercynian Orogeny. In Cumbria, however, the only important phase appears to have been the end-Carboniferous Saalian, and it is reasonably certain that the major Hercynian folding and faulting belongs largely to this period (Moseley, 1972, 580-581; Soper and Moseley, 1978, 64). Nevertheless, localised, small-scale disturbances can be recognised as having occurred during Lower Carboniferous times; west of Kendal, for example, the Thysanophyllum Band of the Martin Limestone is pinched up into sharp anticlines, whilst the beds immediately above and below remain horizontal and undisturbed. Lateral displacement of this type is common in the area, as demonstrated by the slickensides constantly observed on the surface of bedding planes (Garwood, 1912, 518). Minor earth movements may also be indicated by the karstification of the upper surfaces of some of the Lower Carboniferous limestones (see 3.2 and Eastwood, Dixon, Hollingworth and Smith, 1931).

The Caledonian basement exerted a strong influence on the orientation and style of Hercynian tectonics. The general E-W Hercynian stress over northern England resulted in the re-activation of earlier N-S trend lines (Moseley, 1972). Superimposed upon this were the effects of the Alston and Askrigg Blocks and the Lake District massif. The buckling together of the Pennine and Lake District blocks meant that the block margins were deeply fractured, again mainly along N-S trend lines (Moseley, 1972, 581).

The effect of these distributions in the Morecambe Bay area was to develop a system of N-S monoclinal folds: the Hutton Roof monocline, the Hutton monocline-Kendal fault, and the Silverdale monocline, the latter probably traceable to the north along the eastern side of Whitbarrow (Garwood, 1912, 508; Moseley, 1972, 584) (Fig. 2.2 and Fig. 2.3). In the northern part of the area, the characteristic N-S fault lines and general easterly dip are possibly the northward continuation of these fold belts.

Moseley (1972, 587) regarded the Hercynian faults in the area as falling into two groups. Firstly, there are the N-S faults associated with the monoclines. These are complex fault belts generally having a westerly downthrow, which are associated with the steep monoclinal limbs of the folds. Secondly, there are the NW-SE and NE-SW faults which form a

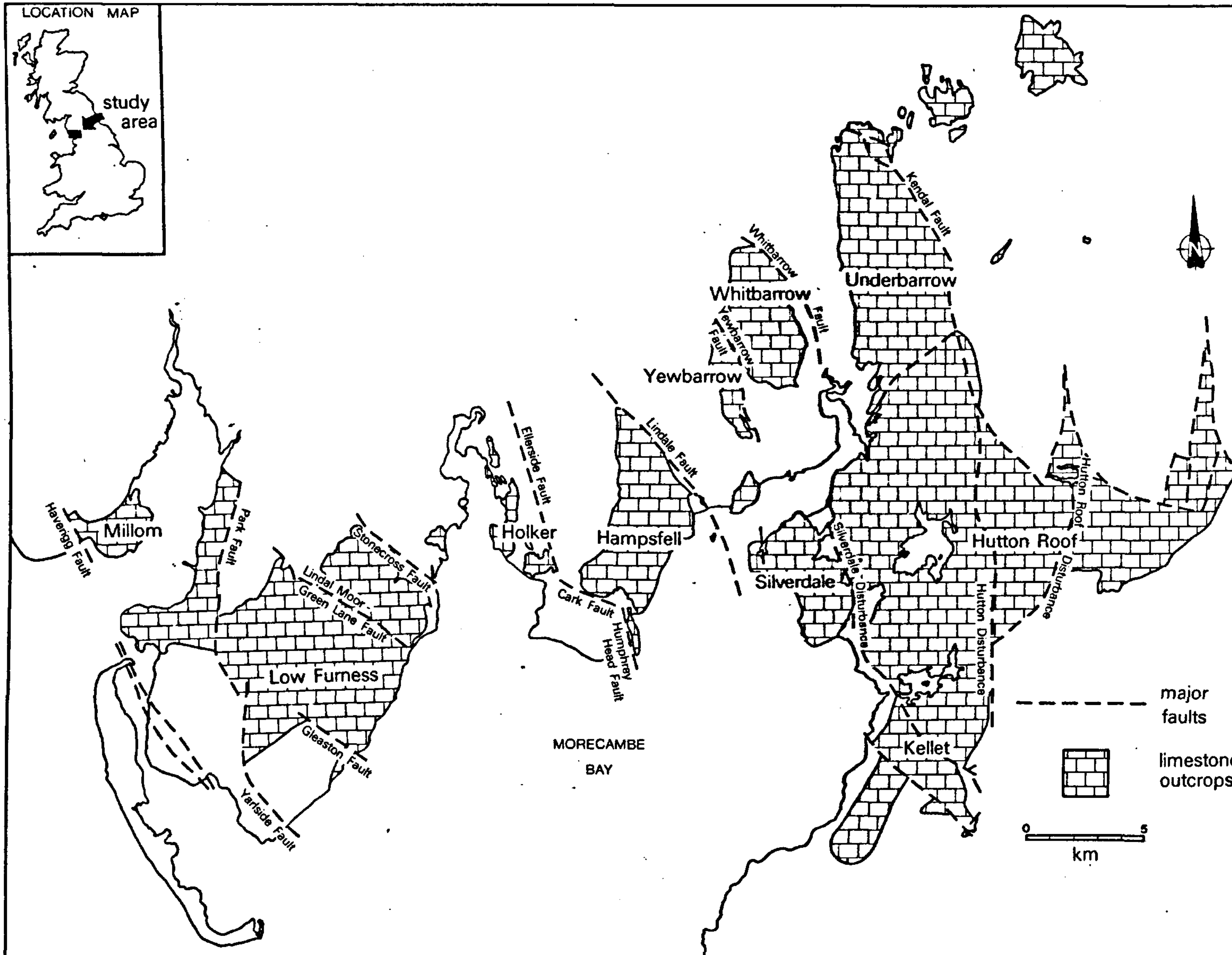


Fig. 2.2 Major faults affecting the Morecambe Bay karst

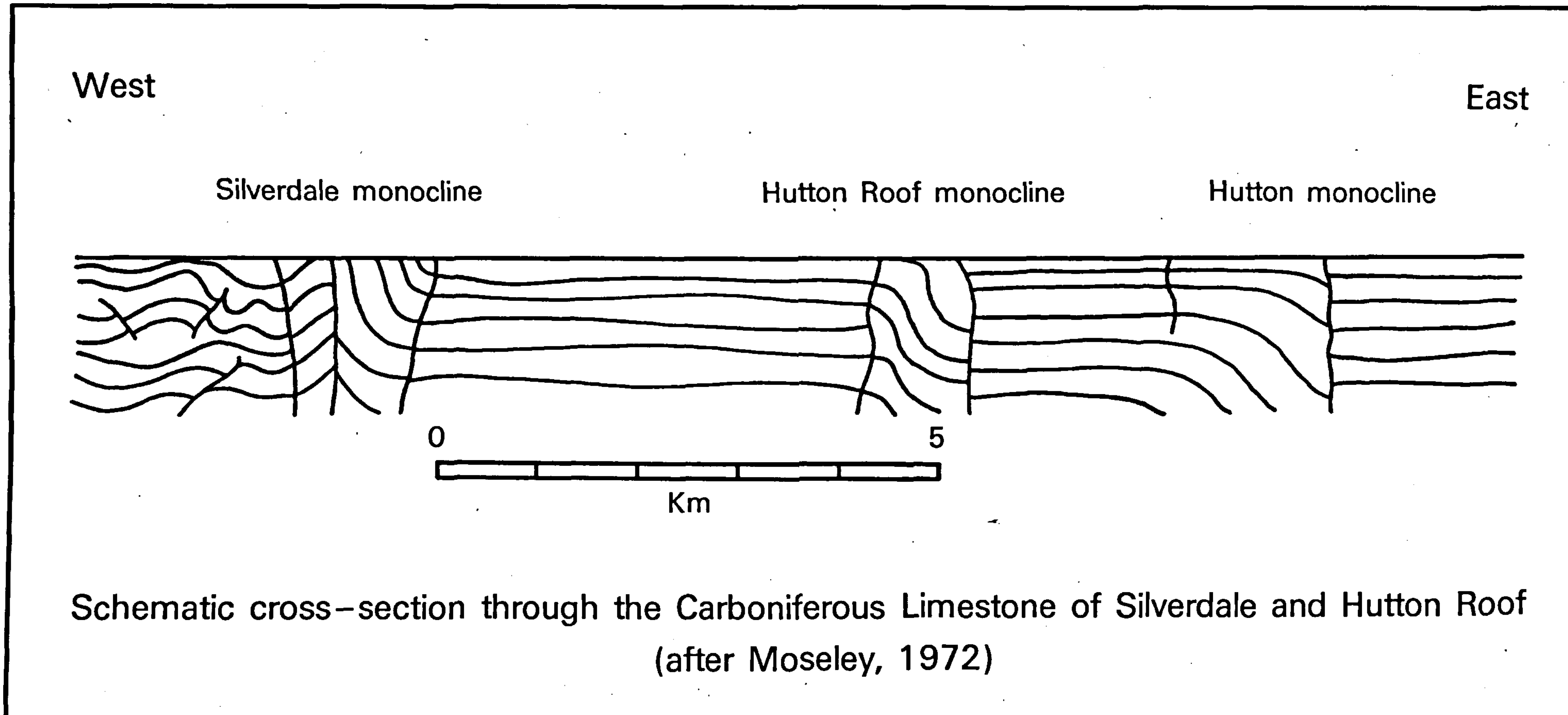


Fig. 2.3 Schematic cross-section through the Carboniferous Limestone of Silverdale and Hutton Roof

conjugate system and which have their origin in wrench-type structures in response to a regional E-W stress. The parallel trend of these structures with those of the Caledonian is apparent. Rose and Dunham (1977, 66) suggested that the characteristic westerly and southwesterly downthrow of the faults pointed to a foundering of the Lower Palaeozoic basement towards the Irish Sea and Morecambe Bay which began in Carboniferous times and continued intermittently until late in the Tertiary. This effect is consistent with the tensional situation arising from the opening of the Atlantic.

The last orogenic phase to have influenced northwest England is customarily termed the Alpine. In common with previous phases, its effect cannot be regarded as distinct. George (1974, 114-119) has demonstrated a continuity of pulsed tectonism from the Hercynian to the culmination of the Alpine in the Miocene. The most spectacular evidence of this is the post-Hercynian Cenomanian transgression. Indeed, the slowly sagging basin of the North Sea is regarded by some as evidence for continuing deformation (Clarke, 1973; George, 1974, 117).

It is generally accepted that a horst and graben type topography was initiated, or re-initiated, in the northwest England-Irish Sea area during the Tertiary. This crustal movement has been linked by Taylor and Smalley (1969) and by Hallam (1972) to north Atlantic splitting, and specifically to the separation of Greenland from Europe at around 60M B.P. On the other hand, Hall and Smythe (1973, 88-89) considered that it was impossible to correlate specific structural features of Britain's continental margin with particular episodes of north Atlantic growth.

On the present land mass, the principal effects of the Alpine Orogeny were the reactivation of existing faults and the uplift of the Lake District, Alston and Askrigg Blocks. The NW-SE faults were particularly active, having displacements of over 1000 m in some situations. Moseley (1972, 582) again stressed that all Tertiary structures owe their orientation and development to earlier structures.

2.4 Structures in the Carboniferous Limestone of Morecambe Bay

2.4.1 Folds

The folds of the Morecambe Bay karst are generally monoclinal in form with the east-facing limb vertical and the west-facing horizontal. The fold belts are orientated N-S, the most important structures being the Hutton Roof monocline, the Hutton monocline and the Silverdale monocline (see 2.3) (Fig. 2.3).

2.4.2 Faults

The three dominant fault trends of N-S, NW-SE and NE-SW all have their counterparts in structures in the underlying Lower Palaeozoic rocks (see 2.3). The N-S faults are associated with the vertical limbs of the monoclinal structures. Where these faults continue to the north of the area they frequently constitute important topographic features. Thus, Yewbarrow, Whitbarrow and Underbarrow form a series of scars across the district resulting from the downthrow to the west of the N-S trending faults (Fig. 2.2). Similarly, throughout the whole of the area, the N-S faults have controlled the development of both closed depressions and polje-like features (see 8.3.2 and 8.4.3).

The NW-SE trending faults are by far the most numerous (Fig. 2.2). They are traceable right across the Morecambe Bay area, forming a regular pattern of normal faults dipping at angles of between 50^{o} and 70^{o} (Moseley, 1973, 101). As in the case of the N-S faults, the NW-SE faults often form significant landscape features, in particular along the fault scarp forming the southwest edge of Warton Crag. The NW-SE faults also control the distribution of sops and other ore bodies in the Furness area (see 4.2) and the location of closed depressions in the Silverdale area (see 8.4.3).

The NE-SW faults have essentially the same characteristics as the NW-SE faults (Moseley, 1973, 101), but in the Morecambe Bay area are only locally significant and do not form conspicuous landscape features.

2.4.3 Joints

Joints provide the main control on the hydraulic conductivity of crystalline limestone and hence act as the locus of solutional activity within the rock. Joints therefore exert a significant influence on landform development in karst areas, both in terms of surface and subsurface karst forms.

Moseley and his co-workers observed three predominant joint sets in the area (Moseley and Ahmed, 1967, 63-65; Ahmed, 1968). These sets are orientated N-S, NW-SE and ENE-WSW, although more recent work by Moseley (1972, 587-588; 1973, 103) suggested the latter orientation to be NE-SW. Ahmed (1968) noted the presence of an E-W joint set, but this was thought to be only locally developed. Other minor sets were found to be common, but were not persistent from one locality to another.

The most common joint set was found to be that orientated NW-SE. It is present in all areas and, according to Ahmed (1968), consists of long joints, often traceable for over 100 m. The NE-SW set is also represented on most outcrops, but as a subordinate set to that orientated NW-SE.

Although the principal joint trends have their counterparts in fault orientations, they are not of the same origin. Most of the faults have hades of the order of 20^{o}, whereas the joints are generally normal to the bedding (Moseley and Ahmed, 1967, 73-74). Moseley and Ahmed (1967, 74) suggested that the joints were already present at the time of important movements of the normal faults.

As part of a study of the influence of structure on landforms in the area (see 8.4.3), a considerable number of joint orientations were measured to complement those of Moseley and his co-workers. The actual selection of sites for joint measurement was largely governed by the existence of suitable exposures of bedrock, with the bulk of the measurements being made in those areas where particular landforms were studied (see 8.4) (Fig. 2.4). In total, over 1000 measurements were made at 24 sites located in eight districts mainly within the Silverdale area (see 11.1 for methods of sampling and analysis).

The detailed picture of jointing within specific areas which these measurements provide (Fig.2.5) does not agree in all respects with the pattern proposed by Moseley. At only three of the sample sites are there significant NE-SW and ENE-WSW orientations, and at two of these sites, Wood Well 2 and Newbiggin 3, the peaks merely constitute subsidiary components

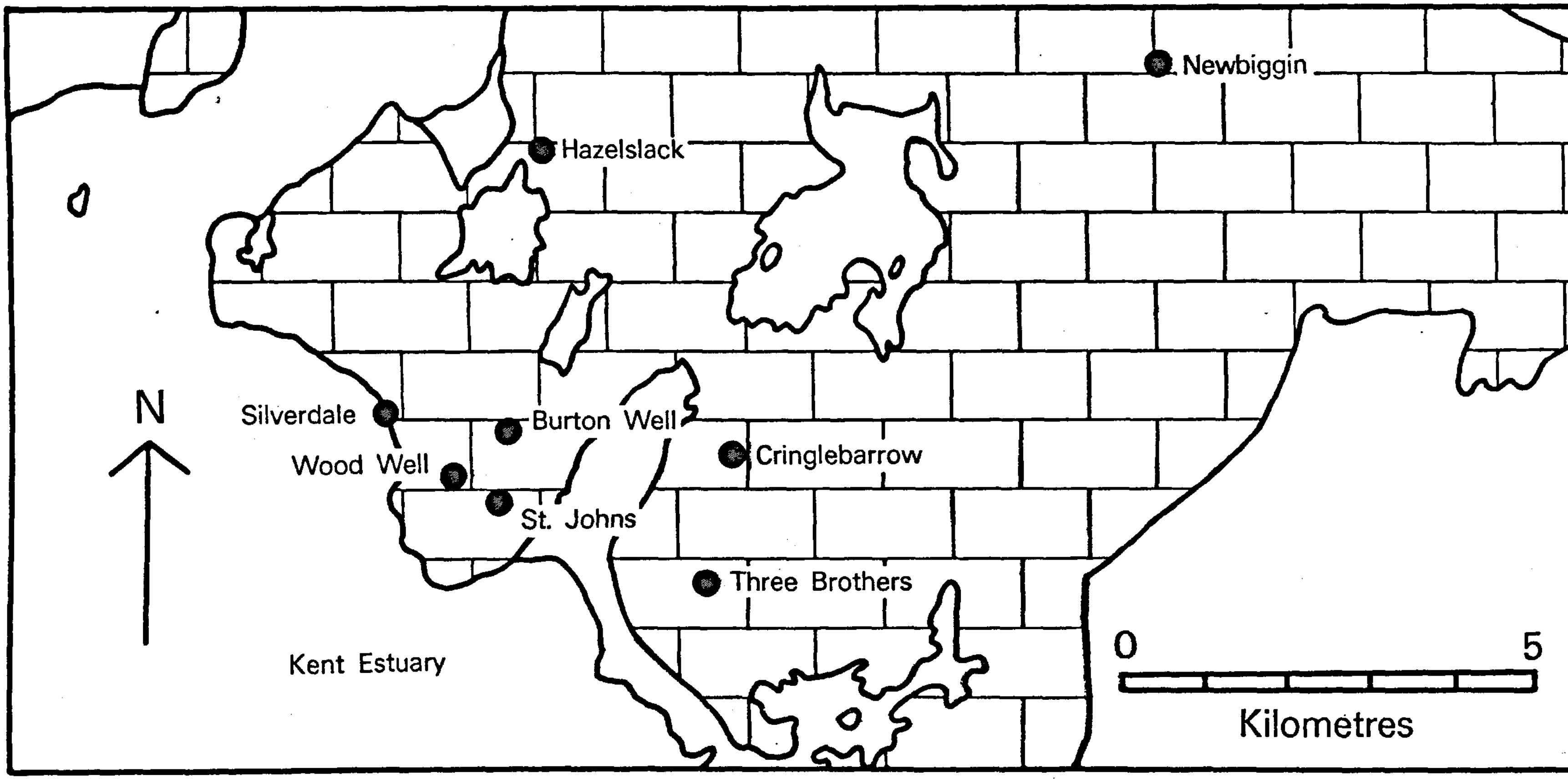
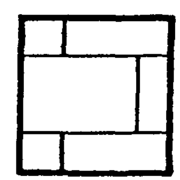


Fig. 2.4 Sites of joint - orientation measurement



limestone outcrops

site of joint - orientation measurement

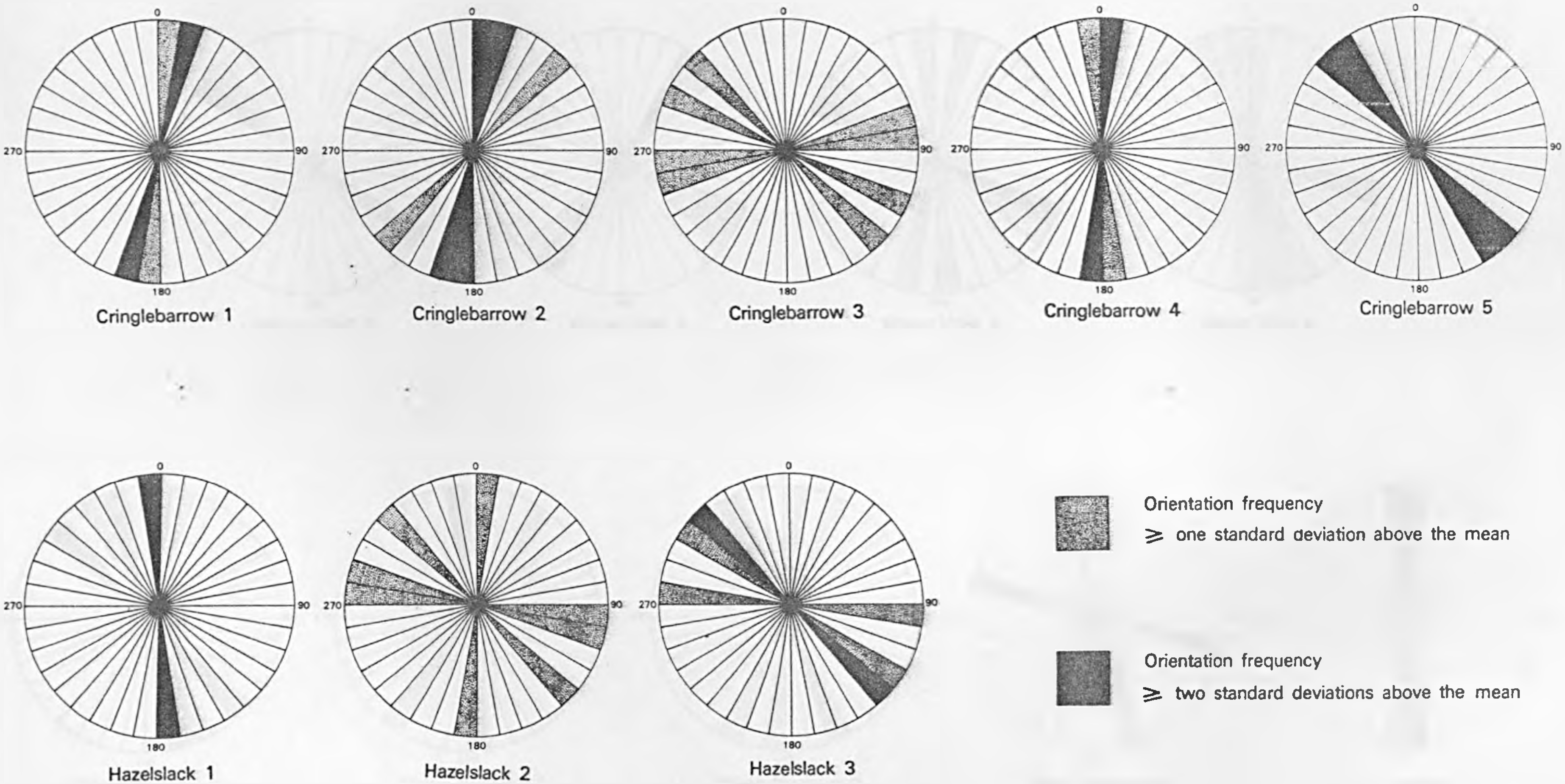
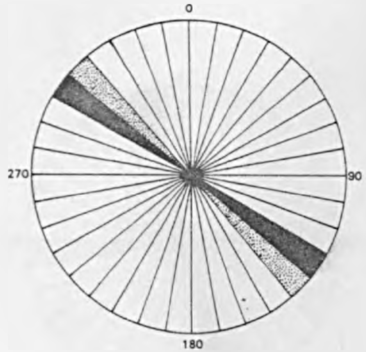
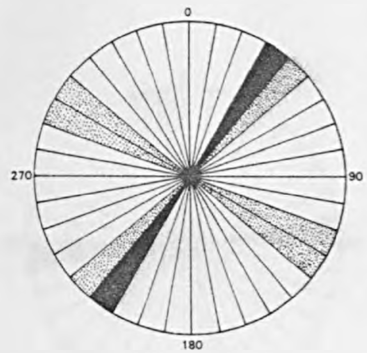


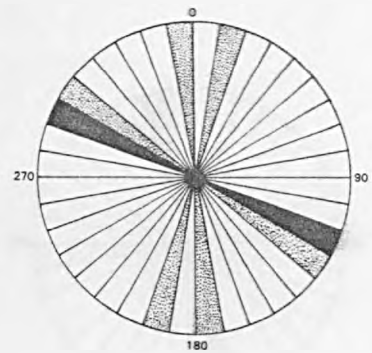
Fig. 2.5 Significant joint orientations at sites in the Silverdale and Hutton Roof areas



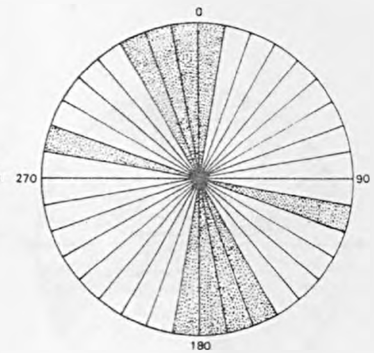
Wood Well 1



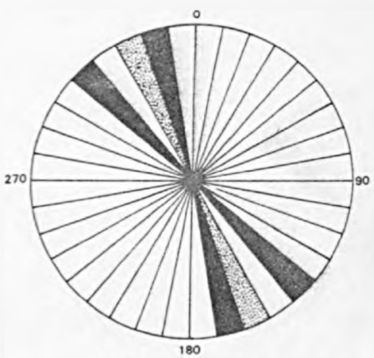
Wood Well 2



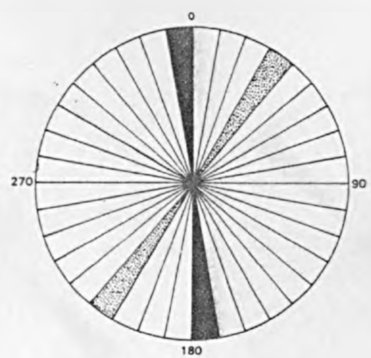
Wood Well 3



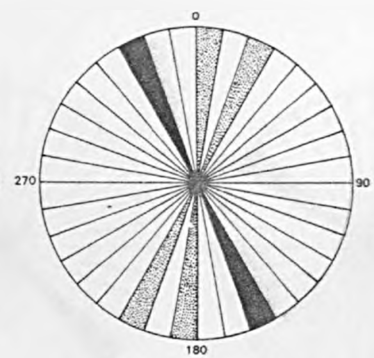
Wood Well 4



Three Brothers 1



Three Brothers 2



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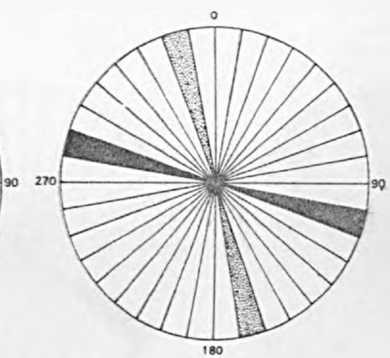

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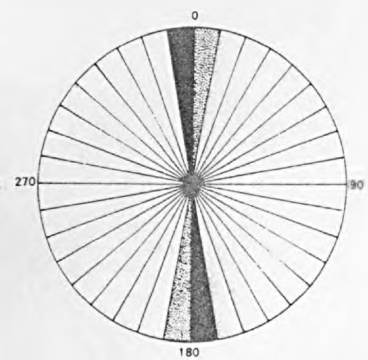
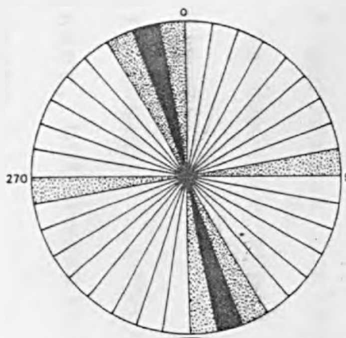
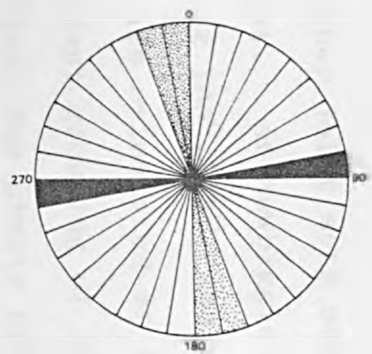


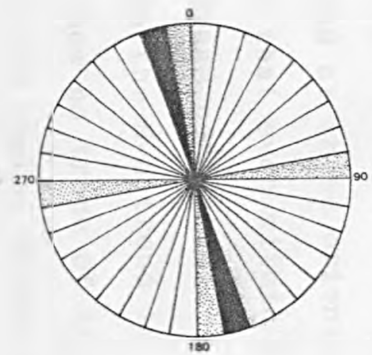
Three Brothers 5



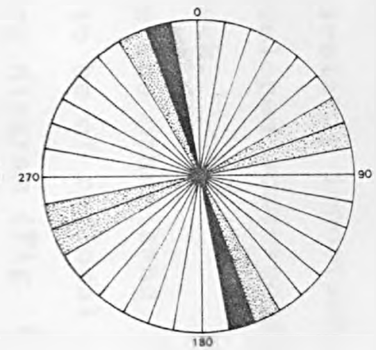
Newbiggin: aerial photograph analysis



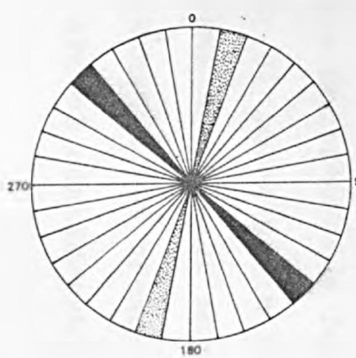
Newbiggin 1



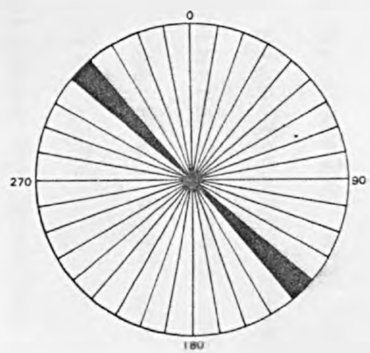
Newbiggin 2



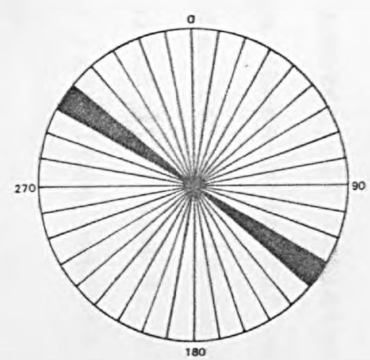
Newbiggin 3



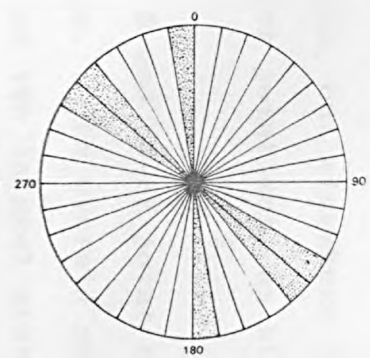
St. John's 1



St. John's 2



Burton Well



Silverdale Cove

of more significant joint sets orientated at 30^{o}-39^{o} and 60^{o}-79^{o} respectively.

The locally significant E-W set noted by Ahmed (1968) is particularly important at Newbiggin, but elsewhere it is only developed occasionally as a minor joint set, as at Hazelslack.

On the other hand, both the N-S and NW-SE joint sets appear to be of regional significance. The overall peak frequency diagram (Fig. 2.6) shows the two largest peaks to be developed at these orientations. However, it is interesting to note that even within a small area these orientations are by no means persistent. At Cringlebarrow, for instance, although the N-S trend is clear at sites 1, 2 and 4, sites 3 and 5 show no trace of it, yet these two sites exhibit a clear NW-SE trend absent at the other sites.

Other joint sets not mentioned by Moseley are locally important. At Newbiggin there is a significant cluster of joints at 160^{o}-169^{o}, and at Three Brothers the bulk of the joints are orientated at 150^{o}-189^{o}. At Wood Well the most significant grouping of joints is at 110^{o}-129^{o}.

Thus, the simple pattern established by Moseley for the whole of northwest England may require modification for detailed studies in smaller areas. The study of jointing in the Silverdale area shows the NE-SW and ENE-WSW trends to be generally absent. Overall, the N-S and NW-SE trends are predominant, but locally other sets are of considerably greater significance.

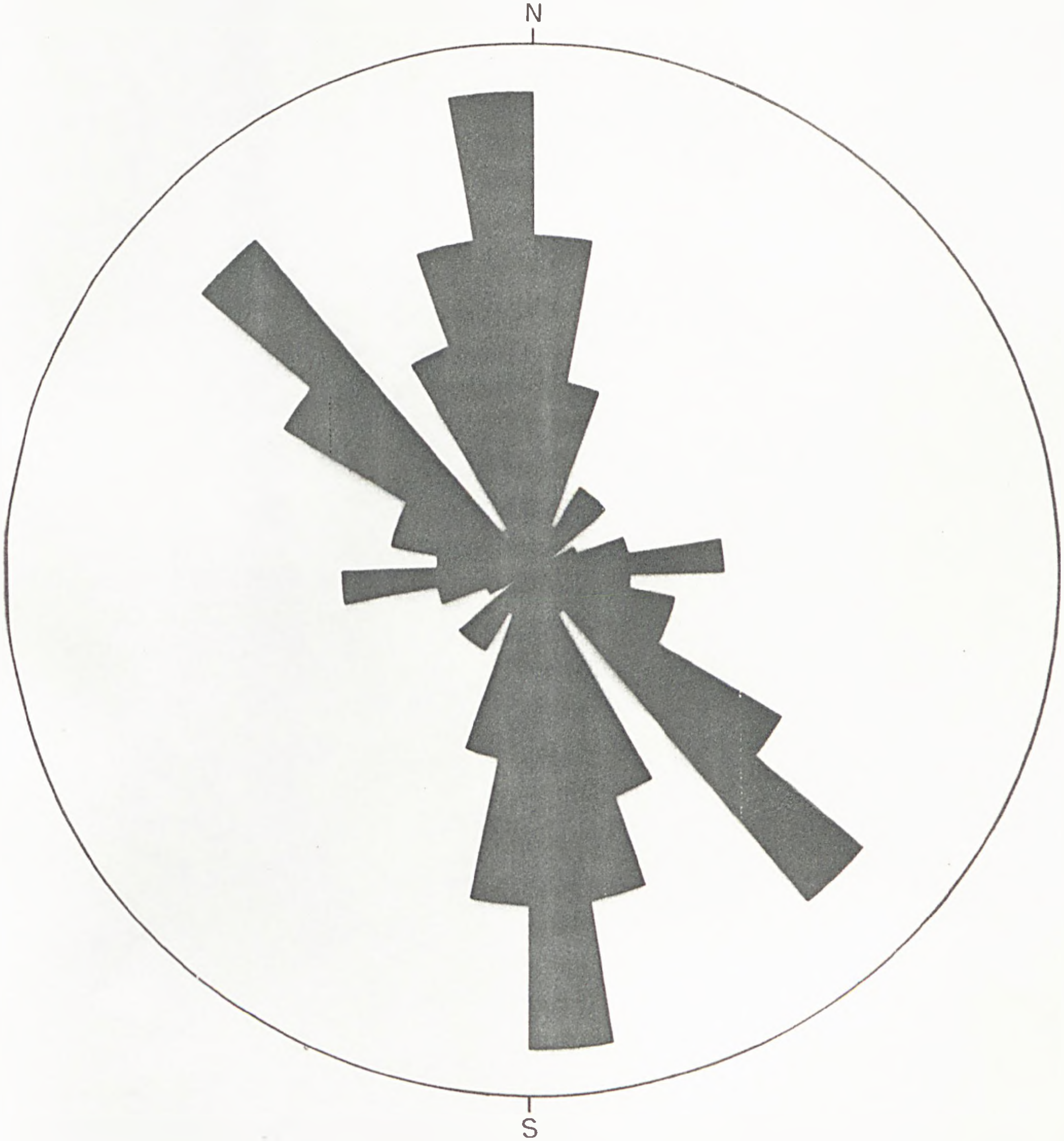


Fig. 2.6 Overall frequency of significant joint orientations

3. PRE-QUATERNARY KARSTIFICATION

3.1 The deposition of the limestone

The sedimentation of limestone in the Morecambe Bay area during Carboniferous times followed a hiatus in deposition extending back as far as the Late Silurian. No deposits of Devonian age are known in the area and the Devonian period is regarded as one of considerable erosion following the peak of Caledonian orogenic activity during Late Silurian times. The resultant pre-Carboniferous erosional platform was made up of structurally controlled massif and basin units. These units seem to have formed elements in the landscape of northwest England which have persisted from Caledonian times to the present day (see 2.2). Thus, the late Devonian landscape was one of moderate relief with a broad pattern of high and low ground reflecting the underlying block and basin structure (Taylor et al, 1971, 38-39).

The pattern of Early Carboniferous sedimentation was largely determined by the topography of the sub-Carboniferous platform. The oldest beds are therefore situated in the fault-controlled basins. Throughout most of the Morecambe Bay area sedimentation probably first occurred during Courceyan times with the deposition of the Basement Beds (Rose and Dunham, 1977, 33). These reach a thickness of 240 m in Low Furness (Rose and Dunham, 1977, 27), but are largely absent in Cartmel and only thinly developed in the Kendal area.

The Basement Beds consist predominantly of conglomerates, sandstones and shales, with occasional thin and impure limestones and gypsums. The conglomerates are believed to represent torrential deposits derived from the Cumbrian block and laid down in shallow waters near the head of an estuary; the gypsum beds represent marginal lagoons, and the limestones more marine conditions (Rose and Dunham, 1977, 27-8).

The limestones become predominant towards the top of the Basement Beds, indicating the onset of marine conditions (Nicholas, 1968). The increased dominance of limestone is regarded by Ramsbottom (1973, 569-572) as representing the early transgressive phase of a major transgression-regression cycle of deposition. Such cycles have been detected by Ramsbottom throughout British Lower Carboniferous deposits and are regarded as evidence of eustatic changes in sea level. In total, four major transgression-regression cycles have been proposed, followed by two phases of minor cyclicity (Table 3.1). The transgressions are thought to be marked by marine bioclastic limestones, and the regressions by algal, dolomitic and fine grained limestones, shales and sandstones. Emergence is indicated in shelf areas by breccia beds, unconformities, non-sequences and karstic horizons. George (1978), however, disputed the evidence for these cycles. He claimed that certain faunal and lithological changes within British Lower Carboniferous beds are inconsistent with Ramsbottom's model. Moreover, the palaeogeographical context of some of the deposits contradicts a cyclic interpretation of sedimentation. Furthermore, depositional thicknesses of shallow water limestones of around 1000 m in the case of Morecambe Bay were seen as inconsistent with sedimentation as a result of eustatic change. Where the limestones are of shallow water origin, almost all

RADIOMETRIC AGE (m.y.)[1]	STAGE[2]	MAJOR CYCLES[3]	LITHOLOGICAL DIVISION[4]
325±5			
334±17	Brigantian	Sixth group of minor cycles	Gleaston Group
328±6 338±4	Asbian	Fifth group of minor cycles	Urswick Limestone
	Holkerian	Regression	Park Limestone
		Transgression	
345±5	Arundian	Regression	Dalton Beds
347±5		Transgression	Red Hill Beds
	Chadian	Regression	Martin Limestone
		Transgression	
		Regression	
		Transgression	
359±5 355±12 360±5	Courceyan		Basement Beds

Table 3.1 The chronology, chronostratigraphy, depositional cycles and lithostratigraphy of the Lower Carboniferous rocks in the Morecambe Bay area

1 After George *et al.*, 1976, 75-77.

2 After George *et al.*, 1976, 75-77.

3 After Ramsbottom, 1973.

4 After Dunham and Rose, 1941; Nicholas, 1968; Rose and Dunham, 1977.

the beds in a cumulative thickness represent an advance of the sea in relation to any selected datum. Absolute sea level changes of this magnitude appear unlikely without the influence of tectonic subsidence, allowing the accumulation of great sedimentary thicknesses within the structural basins of the area. The complementary positive tracts, by contrast, support only relatively thin Lower Carboniferous cappings, denoting either uplift or relatively minor subsidence. Eustatic effects can hardly be regarded as being of the same scale. Instead, George (1978, 247) proposed eustatic pulses superimposed on the overall trend of sea level change, giving rise to local transgressive and regressive phases.

3.2 Geomorphic processes during the Lower Carboniferous

Deposition during the Lower Carboniferous was occasionally interrupted by exposure of the limestone to subaerial processes as a result of relative falls in sea level. Only rarely were the effects of such falls more than locally developed. Nevertheless, features such as shale bands and palaeokarstic surfaces are important indicators of palaeogeomorphological processes.

Shale bands occur in many of the limestone formations of the Morecambe Bay area, including the Basement Beds, the Martin Limestone, the Dalton Beds, the Urswick Limestone and the Gleaston Formation. Most of these shales are probably not subaerial features, but terrestrial deposits laid down in continental margin shallow waters. This latter theory is supported by the calcareous nature of the shales within the Martin Limestone and Dalton Beds (Rose and Dunham, 1977, 28-29). Waltham (1971a, 290), in fact, suggested that, even where shale bands occur above karstification surfaces, the shales do not necessarily represent phases of subaerial deposition. However, Walkden (in Mitchell 1978, 175) regarded certain of the shales found within the Asbian rocks around the Lake District as palaeosols derived from non-local volcanic-rich atmospheric dusts accumulated during emergent phases. Similarly, Mitchell (1978, 175) found evidence of numerous phases of emergence, karstification and soil formation during the deposition of Asbian rocks in Cumbria, giving rise to shale bands above palaeokarst surfaces; but no evidence seems to have been found of such features in the Urswick Limestone of Morecambe Bay.

The palaeokarst surfaces found within the limestone sequence are more definite evidence of the effect of subaerial processes during the Lower Carboniferous. Only one reference has previously been made to these features in the Morecambe Bay area, that by Nicholas (1968), who noted a karstic surface developed on top of the "Breccia Bed" underlying the Red Hill Limestone at Hazelhurst Point (SD335800). However, further examples have been found during the present investigation at Barker Scar (SD33307827), where a number of mammillated surfaces with an overall relief of 3-4 cm occur near the top of the Dalton Beds (Plate 3.1). The form of these features is similar to that of deckenkarren (Sweeting, 1972, 94-95). The absence of noticeable gryke development is presumably the result of erosion prior to the formation of joints in the rock, possibly even in largely unlithified carbonate sediment (Walkden, 1972, 181). The surfaces are

Plate 3.1 Palaeokarst surfaces near the top of the Dalton Beds, Barker Scar (SD33307827) (lens cap is 3.5 cm in diameter)

exactly mirrored by the limestone bed above and there can be no doubt that they predate the overlying limestone.

Emergence and exposure of the limestone to subaerial processes can also be inferred from the presence of desiccation polygons in a muddy bed of the Martin Limestone associated with the Algal Layer (Garwood, 1912, 505-507) and exposed in Meathop Quarry (SD436793) (Day, 1976; Rose and Dunham, 1977, 33). Thus, both the karst surfaces and the dessication cracks indicate phases of emergence close to the Chadian-Arundian and the Arundian-Holkerian boundaries, coinciding almost excactly with the regression maxima predicted by Ramsbottom's (1973) major cycle model (Table 3.1).

Ashmead regarded phases of emergence during the later part of the Lower Carboniferous as likely to have initiated periods of widespread speleogenesis. He cited Swantley Pot (SD52256792), found on the reef knolls of the Kellet area, as having been formed in Asbian-Brigantian (D_1-D_2) times (1974a, 207); and some of the ore-filled caves on Warton Crag (SD4873) as probably initiated during the Early Namurian (1969a, 202). However, although Namurian sediments rest directly on D_1 reef limestones at Swantley (Hudson, 1936, 345), there is no reason to believe erosion was more than locally significant. Swantley Pot is choked with boulders of Namurian sandstone, but there is no evidence of deposition having occurred in the cave during Namurian times. Indeed, as Holland (1967, 88) pointed out, the boulders seem to have been recently emplaced to seal the entrance. Similarly, the fill in the natural cavities on Warton Crag seems to date from post-Carboniferous times (see 4.5).

3.3 Upper Carboniferous environmental history

The only Upper Carboniferous strata preserved in the Morecambe Bay area are the Namurian sandstones found at Ings Point (SD479724) and exposed between the River Lune and the River Keer (Aveline, Hughes and Tiddeman, 1872, 24-25); and the Namurian mudstones, sandstones and conglomerates found in boreholes in the southern parts of Low Furness and Cartmel (Rose and Dunham, 1977, 32-33). These beds appear to have been laid down in a deep, muddy sea (Rose and Dunham, 1977, 33). It is likely that considerable thicknesses of these strata were deposited in the Morecambe Bay area, a possibility which is borne out by the known depths of Roosecote Mudstones encountered in borings in Furness and Cartmel. These beds alone are often over 400 m thick, despite having been deposited wholly within the earliest, Pendleian, stage of the Namurian (Rose and Dunham, 1977, 33). The absence of the remaining Upper Carbonate beds from the area is more likely to have been the result of erosion than non-deposition, outliers such as the Ingleton~~ian~~, Stainmore and Midgeholme coalfields bearing witness to the probable former continuity of Upper Carboniferous strata in the Morecambe Bay area (Taylor *et al*, 1971, 63; Taylor, 1978, 180).

The higher Namurian beds found elsewhere in north-west England seem to be characteristic of deltaic conditions (Taylor *et al*, 1971, 56), which became lagoon or swamp-like during the Westphalian, and it is likely that similar conditions prevailed in the Morecambe Bay area during the later Carboniferous.

3.4 Permo-Carboniferous environmental history

3.4.1 The Hercynian Orogeny

At the end of the Carboniferous, the region was subjected to the increasingly violent tectonic effects of the Hercynian Orogeny (see 2.3), whose effect in the Morecambe Bay area was to impart an overall easterly dip to the beds and to cause faulting along a number of N-S axes. These movements appear to have initiated a phase of considerable erosion, as evidenced by the almost total absence of Upper Carboniferous beds in the area. Consequently, the succeeding Permo-Triassic deposits were laid down throughout most of Low Furness on an eroded Carboniferous surface (Binney, 1847, 428; Dunham and Rose, 1949, 17; Rose and Dunham, 1977, 59). There is little reason to doubt that a similar phase of erosion occurred throughout the rest of the Morecambe Bay area, although, as a result of the paucity of Permo-Triassic remnants, only one unconformable junction is known, that at Flookburgh (SD3675) in the Cartmel peninsula (Sedgwick, 1836, 389).

Ashmead (1974a, 207) has proposed that, by the end of the Carboniferous, the principal feature of the area was a limestone plateau extending from the N-S Ellerside fault in the west to the Silverdale monocline in the east (Fig. 2.2), where the limestone dipped beneath a cover of Namurian rocks. However, such a simple topographic picture is unlikely to reflect the reality of geomorphology at the end of the Carboniferous. The Hercynian Orogeny exploited structural weaknesses in the Caledonian basement which were subsequently reactivated during the Tertiary, giving rise to the present structural pattern of the area (see 2.3). It is therefore not unreasonable to suppose that the N-S fault lines of the Alpine Orogeny were also active during the Hercynian, and that features such as the Ellerside, Yewbarrow, Whitbarrow and Kendal faults (Fig. 2.2) formed important structural elements in the area as they do today. The resultant topography would have been similar to that of the present, characterised by a series of west-facing scarp slopes and east-dipping plateau-like surfaces.

3.4.2 Permo-Carboniferous karstification

During the Permo-Carboniferous transition, the environment of Britain changed from that of the tropical conditions of the Westphalian (Schwarzbach, 1963, 133; Bennison and Wright, 1969, 223) to that of the hot, dry conditions of the Permian. This was partly the result of the gradual northward movement of Britain from about 5^{o}N during the Upper Carboniferous (Turner and Tarling, 1975, 484) to about 10^{o}N by the Early Permian (Nairn, 1964, 555; Van der Voo and French, 1974, 109), and partly the result of the increased continentality of Permian Britain (Schwarzbach, 1961, 265).

With the exposure of the limestone, conditions were favourable for a phase of karstification, at least during the early part of the period. It is likely that the limestone exposed in the Morecambe Bay was adjoined in part by areas of topographically higher, impermeable beds. To the north these would have comprised the upstanding mass of the Lake District block, whilst to the south and southeast would have been the retreating Upper Carboniferous cover which had formerly capped the Morecambe Bay limestone. Outliers of these Upper Carboniferous beds exist today to the south and southeast of the area (see 3.3), suggesting their retreat southwards and their preservation in the basinal areas of deeper deposition. At the same time it is likely that the higher ground in the Morecambe Bay area was capped by remnants of the former Upper Carboniferous cover, as suggested by borehole logs showing the existence of Namurian remnants on the sub-Permian surface (Rose and Dunham, 1977, 124-153).

A number of workers have proposed that the sops of Low Furness represent mineralised karst depressions and caves formed during this period, but, as shown in 4.3, the evidence does not support this theory. Nevertheless, as has been demonstrated, both climatic and hydrological conditions were favourable for karstification, at least during the early part of the period.

Unfortunately, little evidence of such a phase of karstification exists in the Morecambe Bay area. Remnants of the pre-Permian surface survive only where they have remained buried by Permo-Triassic beds. According to Rose and Dunham (1977, 59), the pre-Permian surface was probably an uneven pediment with low ridges and hollows scattered across it. However, a study of the logs of boreholes in the area shows little to support their interpretation and reveals features not noted by earlier workers.

There are a number of potential problems involved in the use of borehole records to reconstruct the form of eroded surfaces in the Morecambe Bay area. Firstly, the effects of post-Triassic earth movements make it difficult to reconstruct the whole of the surface. However, within certain areas, the Permo-Triassic beds show no sign of faulting and these areas may be used for a study of the pre-Permian surface. Secondly, the effect of Hercynian tectonic activity on the Carboniferous basement may be obscured by the subsequent cover of Permo-Triassic rocks. Nevertheless, Rose and Dunham (1977, 69) point out that "the age of the main movement of several of the major faults is demonstrably post-Triassic and there is no reason to suppose that the age of many of the smaller faults seen only to affect Lower Carboniferous rocks is not the same". Finally, allowance must be made for any tectonic dip of the Carboniferous basement. Throughout Low Furness this is generally 10-15^{o} to the southwest.

Only one area meets the conditions necessary to allow the use of borehole records to reconstruct the pre-Permian surface: that around Sandscale (SD1973). Here the eroded Carboniferous surface has a relative relief of at least 80 m, and probably even more once allowance is made for the southwesterly dip of the rock (Fig. 3.1).

3.4.3 Pre-Permian red staining

At some stage before the earliest Permian deposition in the area, but after, or possibly associated with, the late phases of pre-Permian landscape development, there occurred a phase of red staining of the eroded Carboniferous surface. This staining, by iron oxides, including haematite,

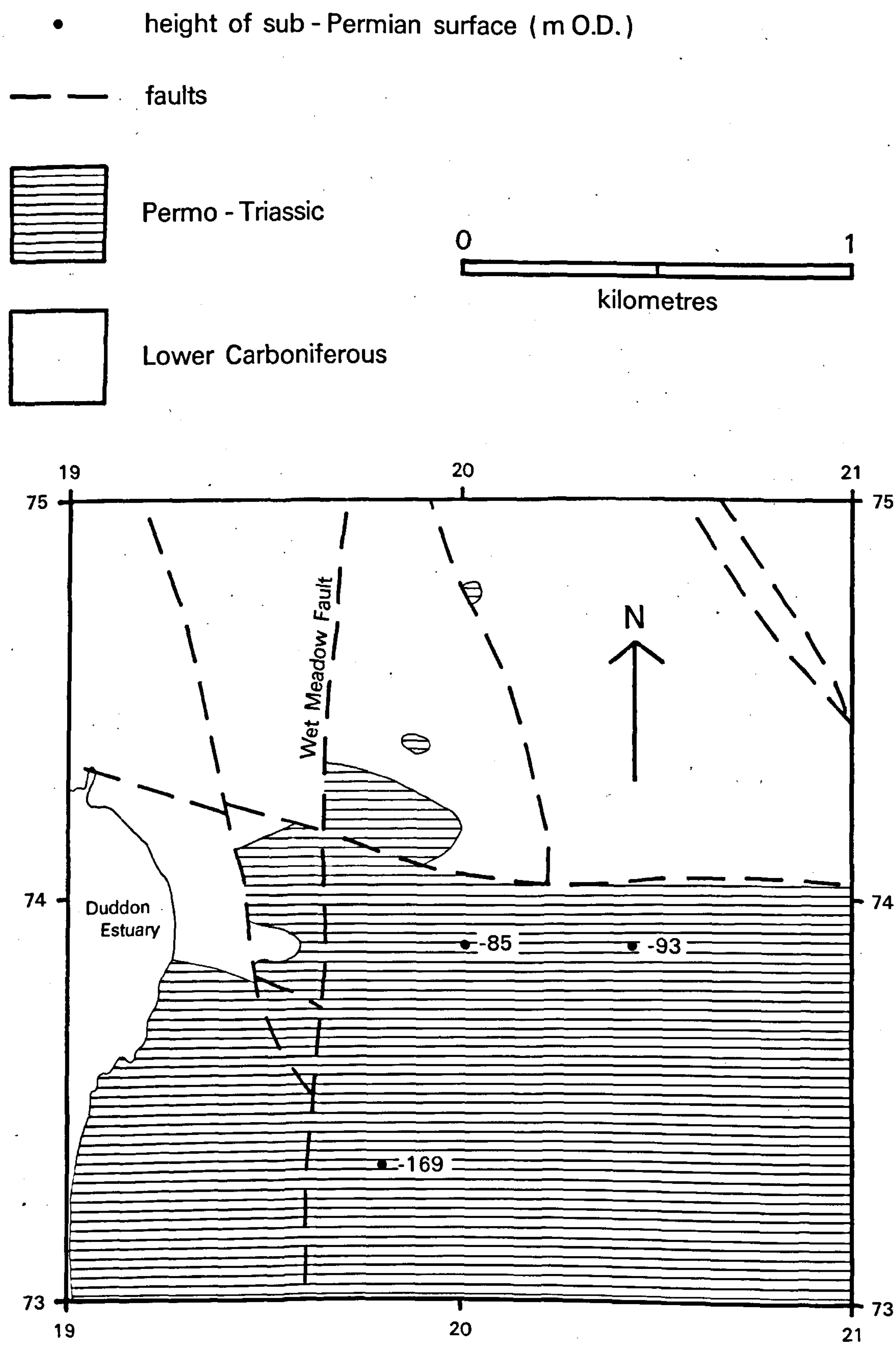


Fig.3.1 The sub - Permian surface in Low Furness

was widespread both within the area and throughout much of Cumbria (Trotter, 1939; Rose and Dunham, 1977, 59). That this staining occurred no later than the Permo-Carboniferous boundary is evidenced by:

(i) The white gypsum beds and grey Hilton Plant Beds of Upper Permian age, which overlie the stained surface in Edenside and the Carlisle Basin, show no sign of the percolation of ferric iron (Totter, 1939, 415).

(ii) The Early Triassic St. Bees Sandstone, which overlies the stained surface in West Cumberland, shows no sign of leaching (Trotter, 1945, 69).

(iii) The occurrence of stained and unstained rounded pebbles, derived from the same parent beds, found in the Permian Basal Breccia (Trotter, 1939, 410). These pebbles were probably derived from the Carboniferous stained zone, thereby providing a pre-Permian date for the staining. This process also explains the apparently anomalous occurrence of haematite pebbles in Permo-Triassic sediments (see 4.3), taken by some workers as providing evidence for a pre-Permian age for haematite mineralisation.

Goodchild (1881-82, 117; 1889, 63) considered the staining to be the result of downward percolation of waters from the overlying Permo-Triassic beds, but Totter's evidence renders this explanation impossible. Trotter (1939, 415) favoured instead weathering under desert conditions. Red staining under desert conditions is not a widespread phenomenon. Nevertheless, the existence of red beds in association with evaporites and aeolian deposits, or with abundant desiccation marks and salt casts, provides definite evidence of their formation under hot, dry conditions (Van Houten, 1964, 657, 659).

Rose and Dunham (1977, 90) argued against Trotter's hypothesis on the basis of Dunham's (1952) findings that haematisation in Britain is not characteristic of arid conditions. As an alternative, they suggested the staining to be the result of laterisation. This implies specific environmental conditions in the area, in particular, warm temperatures. Rather less clear, however, are the moisture conditions under which laterisation occurs. The precipitation of sesquioxides necessitates alternating moisture conditions, and this has come to be regarded as synonymous with marked wet and dry seasons. Nevertheless, a climatic control is not strictly necessary, for similar conditions may be provided by fluctuations in the height of the water table. On the other hand, there is also considerable evidence for laterite formation under permanently moist atmospheric conditions, whilst the study of individual laterite constituents suggests moist conditions to be general favourable to laterisation (McFarlane, 1976, 45).

Thus, suitable environments for both the proposed staining processes existed in the area during the Permo-Carboniferous transition. The warm, moist Westphalian environment provided the necessary conditions for laterisation, given that the traditional view of the incompatibility of forest vegetation and laterite genesis has been shown to be unsound (McFarlane, 1976, 46-52); whilst the progressively drier conditions heralding the start of the Permian would have favoured staining under arid conditions.

The stratigraphic position of the stained beds suggests that staining occurred relatively late in the period, after considerable erosion of Carboniferous beds had taken place and just prior to the deposition of the Permian Basal Breccia. On this evidence it seems likely that the climate tended to be dry rather than moist, and that the staining is the result of desert oxidation processes, but without definite knowledge of pre-Permian environmental conditions this genesis cannot be confirmed.

3.5 Permo-Triassic environmental history

The arid or semi-arid conditions of immediately pre-Permian times would have resulted in the cessation of karst processes and the fossilisation of the karst landscape, which was subsequently buried beneath the accumulation of Permo-Triassic beds. The earliest Permo-Triassic deposits consist of thin, intermittent breccias, which were subsequently submerged by the advancing Upper Permian sea which laid down a thin sequence of dark mudstones and siltstones. The environment soon became more saline and the overlying St. Bees Shales seem to have formed in a playa lake or on a marginal sabkha surface. There is a gradual upward transition into the St. Bees Sandstone, of probable largely fluviatile origin. This is succeeded finally by a sequence of red mudstones containing thick beds of rock salt. These beds are almost certainly of Scythian-Anisian-Ladinian age (Rose and Dunham, 1977, 59).

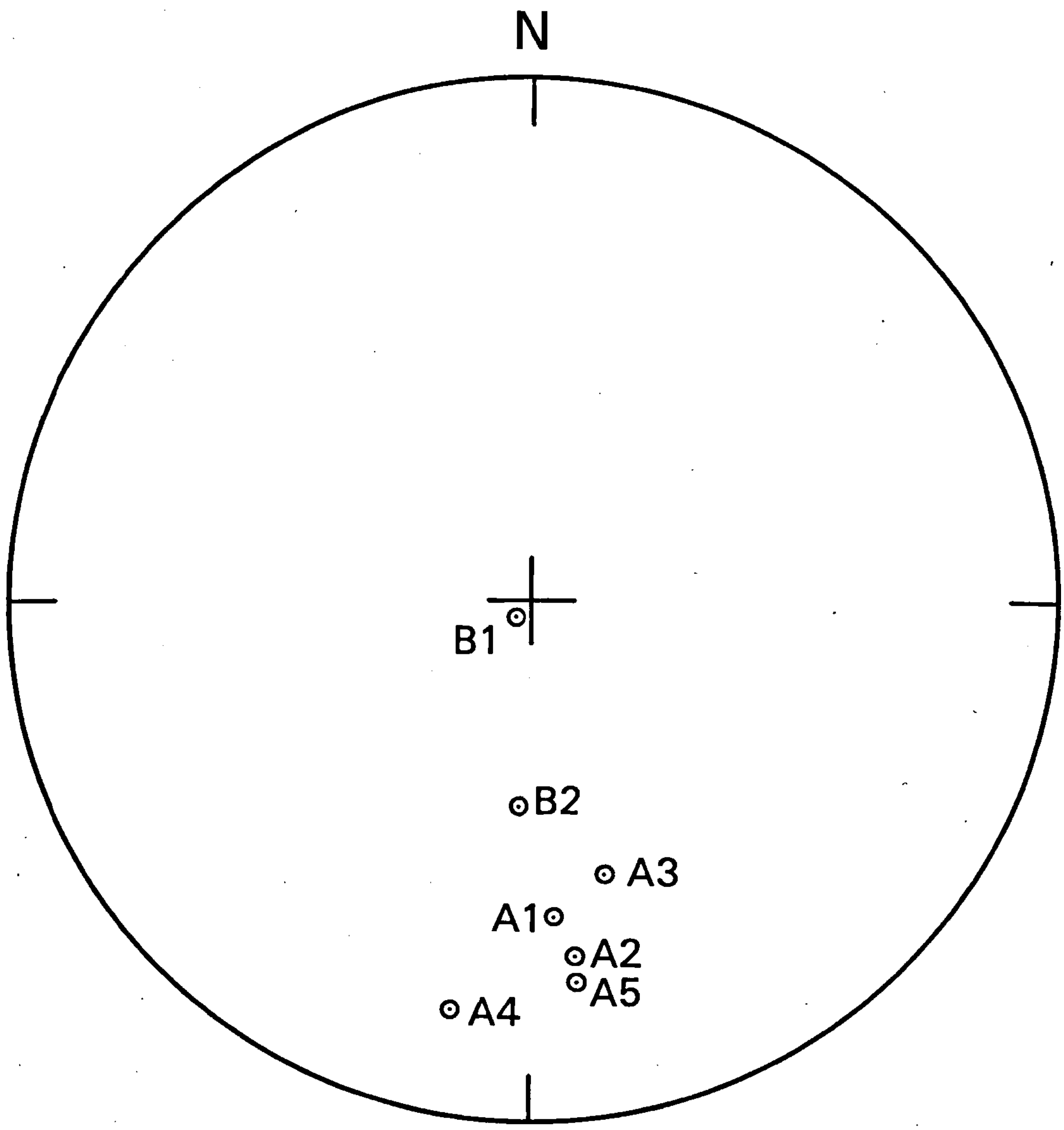
3.6 Post-Triassic karstification

With the exception of unconsolidated Quaternary deposits, the youngest beds found in the Morecambe Bay area are the Kirkham Mudstones of mid-Triassic age (Rose and Dunham, 1977, 61). Between this time and the Late Pleistocene, there is a major gap in the geological record, during which time environmental conditions can only be inferred from events in the rest of the British Isles. At some stage after the Triassic, an extensive interstratal karst developed in the Morecambe Bay area, caves being infilled both by collapse of the overlying beds and by haematite mineralisation (see Chapter 4). This phase of karstification has been tentatively assigned to the Oligo-Miocene.

A further phase of karst development may be indicated by deposits found in Grizedale Wood Drainage Level (SD48257409), a mine level on the west side of Warton Crag in the Silverdale district. Approximately mid-way along its length, the level intersects a natural cavity which is infilled by unlithified, laminated, red and grey silty beds overlying a breccia comprised of similar material (Plate 3.2). There is some evidence of post-depositional settling of the laminated beds at the western end of the deposit, indicated by reverse faulting, but the remainder of the deposit appears to be *in situ*. Palaeomagnetic analysis of the undisturbed deposits (see 11.2) clearly shows that the beds were laid down during a period of reversed polarity (Fig. 3.2).

Assuming the ancient magnetic field to have been of an axial dipole type, these deposits indicate a virtual geomagnetic pole (VGP) located in the region of

Plate 3.2 Red and grey beds infilling a natural cavity in Grizedale Wood Drainage Level (SD48257409)



Laminated beds, Grizedale Wood Drainage Level, Lancashire, England (SD 48257409)
Direction of natural remanent magnetisation
(Lambert-Schmidt net)
Southern exposure: Samples A1 – A5
Northern exposure: Samples B1 – B2

Fig. 3.2 Direction of natural remanent magnetisation: laminated beds, Grizedale Wood Drainage Level (SD48257409)

the Gulf of Guinea off the coast of West Africa (Table 3.2). For most of the last 700,000 years, the earth's magnetic field has been normally magnetised. However, there is evidence of a number of polarity excursions during this period, lasting perhaps 10^2 to 10^3 years, during which the VGP may have reached latitudes more than 135^{o} from the pole. However, with the exception of the Blake magnetic excursion, which occurred approximately 100,000 B.P., none of these excursions exhibit VGPs similar to those indicated by the Grizedale Wood Drainage Level deposits. During the Blake excursion, the VGP passed through West Africa during the transition from normal to reversed polarity (Denham, 1976). Nevertheless, there is no clear evidence to suggest that the Blake excursion was the result of motion of the dipolar field, and it is possible that the Blake palaeomagnetic record simply reflects a regional nondipolar fluctuation which was not experienced over other parts of the earth. However, within the limits of present knowledge of the behaviour of the earth's magnetic field over time, the Blake excursion provides a minimum age for the Grizedale Wood Drainage Level deposits and for the cavity that they occupy.

The laminated nature of the upper beds in Grizedale Wood Drainage Level suggests deposition under quiet-water conditions, although the underlying beds appear more indicative of torrential deposition. The alternating red and grey colour of the beds is probably the result of post-depositional processes, for scanning electron microscope analysis shows that it is impossible to differentiate the beds on either mineralogical or granulometric grounds (Plates 3.3 and 3.4). It is more likely that the grey colouration is the result of

Sample	Declination	Inclination	Palaeo-latitude	Palaeo-longitude
A1	175.5°	-51.2°	3.9°S	1.0°
A2	172.9°	-51.2°	3.8°S	3.3°
A3	163.1°	-45.6°	7.5°S	12.4°
A4	190.9°	-69.3°	17.4°N	350.3°
A5	173.1°	-62.8°	8.6°N	2.2°

Table 3.2 Orientation of natural remanent magnetisation and location of virtual geomagnetic poles: Section A, Grizedale Wood Drainage Level.

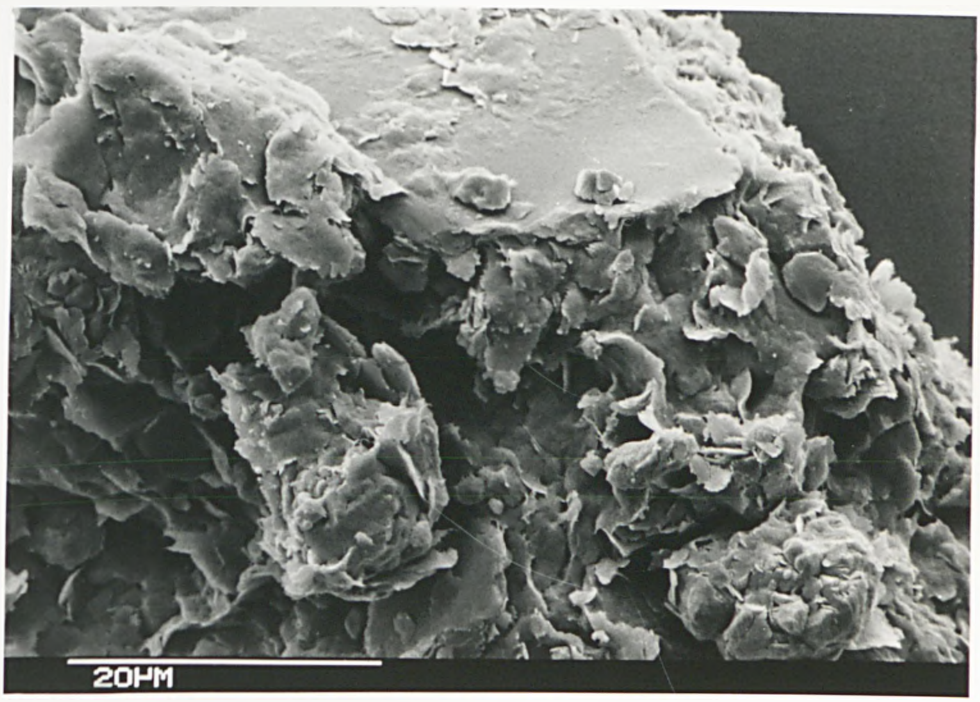


Plate 3.3 Scanning electron photomicrograph: grey beds,

Grizedale Wood Drainage Level (SD48257409)

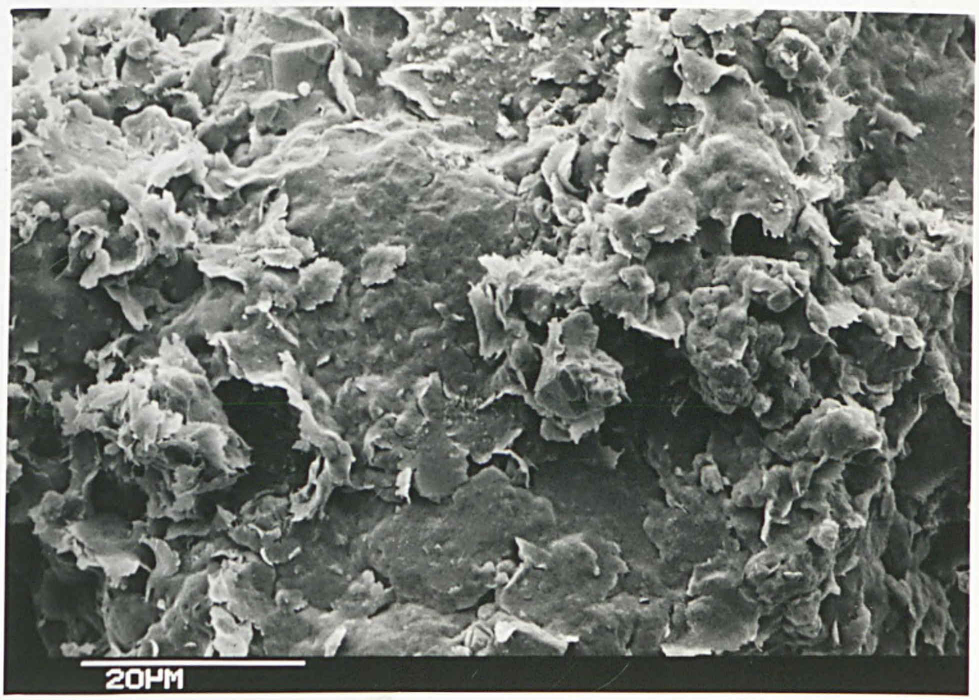


Plate 3.4 Scanning electron photomicrograph: red beds, Grizedale Wood Drainage Level (SD48257409)

secondary reduction of the ferric oxides which give the red beds their characteristic colour. It is possible that this occurred in the cave under anaerobic conditions in standing water bodies. Thus, each of the grey beds within the laminated sequence may reflect a temporary phase of non-deposition within a static water environment.

Scanning electron microscope analysis shows that the deposits contain an abundance of clay minerals. Sediment particles are coated with a smooth textured and platy clay which is almost certainly detrital (Waugh, pers. comm.) (Plates 3.3 and 3.4). The most likely source of the deposits would therefore seem to have been the surface glacial materials of the area.

Further phases of karstification in the area during the Tertiary have been proposed by a number of workers. Corbel (1957, 289) compared the isolated limestone hills which emerge from alluvial plains in the area to a tropical karst landscape. On morphological evidence alone, he regarded the karst as relict from a warm humid phase of the Tertiary. In fact, the isolated nature of the limestone hills stems from structural control (see 2.4.2) and there is little evidence for the past working of tropical tower-foot processes around the base of the hills (see 7.2.3).

Sweeting (1970, 239; 1972, 301; 1973, 105-106) supported the hypothesis of karstification under relatively warm climatic conditions, probably in the Tertiary. Her evidence included the existence of apparently solutionally-eroded slopes beneath the fossil screes of the area, surface deposits of reddish soil or clay of a <u>terra rossa</u> type, and rock pendants in the caves. However, the rock slopes beneath the fossil screes of the area can be shown to be glacially rather than solutionally modified (see 8.5); the red deposits of the area are undoubtedly glacial, derived from the red Permo-Triassic beds of the region (Rose and Dunham, 1977, 119-122); and the rock pendants in the caves seem to be the result of differential erosion under phreatic conditions, rather than the result of climatic control. Thus, although a phase of Late Tertiary karstification is likely, possibly as a result of denudation associated with the culmination of Alpine tectonic activity (see 2.3), this hypothesis can be regarded as no more than conjectural until definite supporting evidence can be found.

3.7 Conclusions

At least three phases of pre-Quaternary karstification can be recognised in the Morecambe Bay area. The earliest of these is associated with the regression maxima of the Lower Carboniferous seas, during which unlithified limestones were temporarily exposed to subaerial processes. It is likely that karstification of this nature occurred on a number of occasions during the Lower Carboniferous, although at present it can only be shown to have taken place at the Chadian-Arundian and the Arundian-Holkerian boundaries. There is no support, however, for the widespread phase of Late Dinantian speleogenesis proposed by Ashmead (1969a, 202; 1974a, 207).

The next phase of karstification occurred during the Permo-Carboniferous when erosion, possibly associated with

Hercynian tectonic activity, exposed the limestone to sub-aerial processes. There is evidence for the development of a landscape of considerable relief, contrasting markedly with the picture of an uneven pre-Permian pediment proposed by Rose and Dunham (1977, 69) and with the picture of a limestone plateau proposed by Ashmead (1974a, 207).

The increasingly arid conditions which heralded the onset of the Permian fossilised the karst landscape and resulted in a phase of red staining which appears to have affected the sub-Permian surface throughout the whole of Cumbria.

Following this, it is difficult to establish a clear chronology of karstification. There appears to have been a phase of interstratal karstification during the ?Oligo-Miocene (see 4.4.3), whilst reversely magnetised deposits in a natural cavity may provide evidence for another ancient phase of karstification. However, the evidence which earlier workers advanced in support of a phase of Tertiary tropical karstification has been shown to be unreliable and the Tertiary cannot be regarded with certainty as having given rise to surface karst forms in the area.

4. MINERALISATION AND KARST DEVELOPMENT

4.1 Introduction

At some stage after the Namurian, the Morecambe Bay area seems to have experienced a phase of considerable karstification. The evidence for this lies in the extensive exhumed karst landscape found in Low Furness, along with similar features in the Silverdale area.

The exhumed karst is almost invariably associated with deposits of haematite. The source of the haematite itself has been the subject of considerable debate (for useful summaries see Nicholas, 1968; and Rose and Dunham, 1977, 89), but most workers accept that at least some ore deposition took place in solutionally-eroded features in the Carboniferous Limestone known as sops. However, the processes and date of the formation and infilling of these features remain controversial.

4.2 The sops

The sops, as the basin-like, ore-filled features are locally known, are a unique feature of Low Furness. They are circular to oval in plan, narrowing downwards into deep pockets (Smith, 1924) (Fig. 4.1). They are developed only in limestone, although in some cases, where the sop is developed along a fault, one side is composed of a different lithology. By far the most favourable lithological division of the limestone for sop development is the Red Hill Beds (Rose and Dunham, 1977, 88). Sops are also common in the upper parts of the Martin Limestone and the lower parts of the Dalton Beds, although elsewhere within these divisions shale beds become relatively more important (Nicholas, 1968)

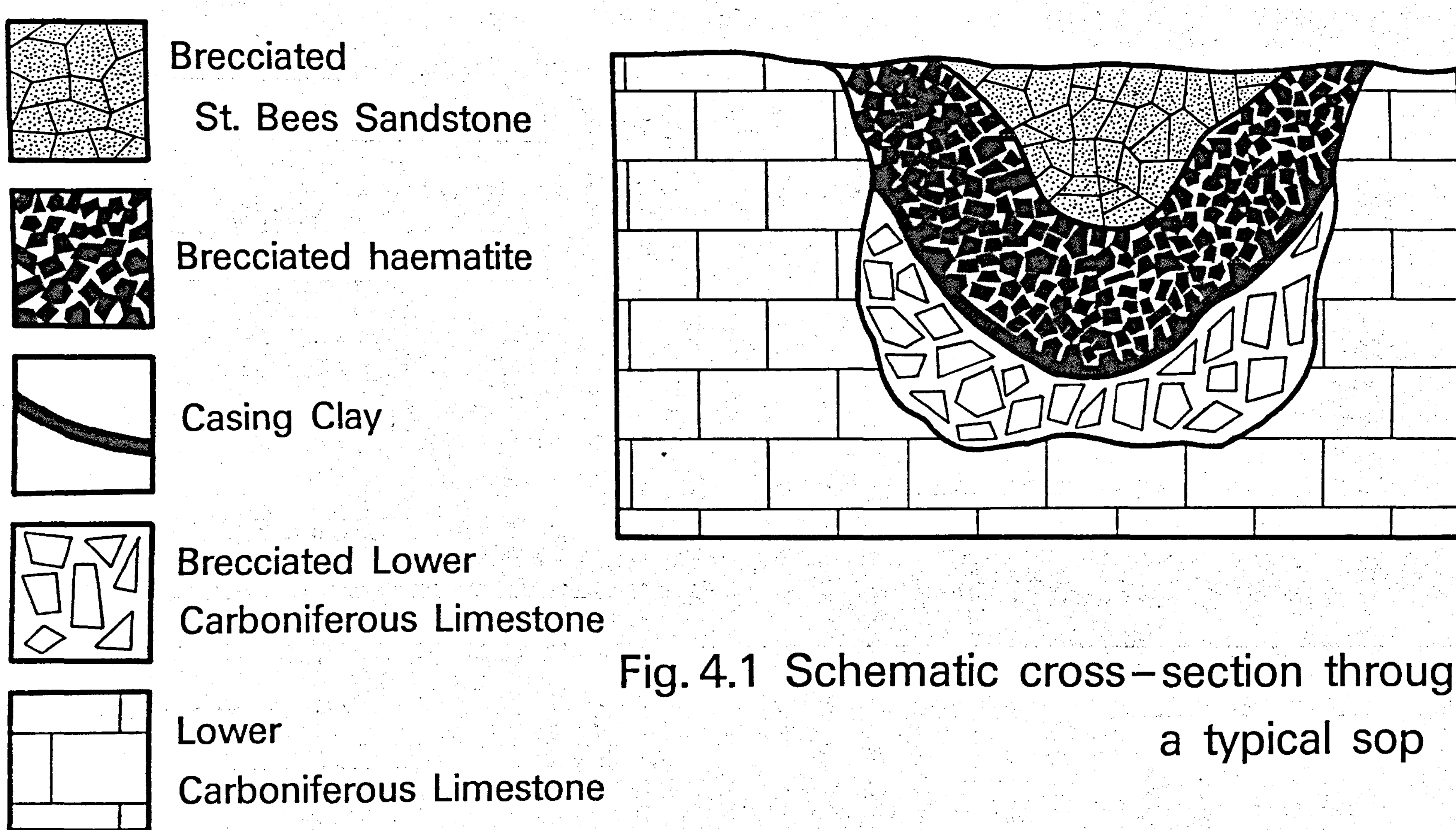


Fig. 4.1 Schematic cross-section through a typical sop

and sop development is restricted. Around Roanhead and Park, sops are also developed within the Park Limestone (Rose and Dunham, 1977, 105).

These stratigraphic restrictions are interesting in view of the apparently karstic nature of the sops. The Martin Limestone, Red Hill Beds and Dalton Limestone provide favourable conditions for karstification, in that they are well-bedded and, in the case of the Red Hill Beds, coarse grained (Nicholas, 1968). The Park Limestone, however, does not support distinctive karst forms in the region. It is poorly-bedded and its closely spaced horizontal and vertical joints do not weather into grykes, yet it supports sop development. On the other hand, the Urswick Limestone is a massively-bedded and jointed relatively pure limestone, which constitutes the most important karst lithology in the region, yet no sops are developed within it. The likely explanation for this anomaly is that the Urswick Limestone was not present in those areas where sop development was concentrated (for example, Hodbarrow and Dalton-Lindale), perhaps as a result of erosion prior to sop development (see 3.4). The absence of sops in the Basement Beds and the Gleaston Group is the result of the largely non-calcareous nature of these divisions.

The distribution of many of the sops, as well as that of ore bodies in veins and flats, is structurally controlled. The system of Hercynian NW-SE faults, and less commonly joints, provides the dominant structural influence (Rose and Dunham, 1977, 87-88). As in the case of the sops, some of the vein ore bodies may have infilled pre-existing caverns aligned along structural discontinuities.

The largest sop, Park (SD2175), is 420 m long, 20-180 m wide, and reaches a maximum depth of 180 m. Like the other large sops, it seems to have formed by the union of two adjacent hollows. From these largest bodies there is a gradation of size down to sops only 30 m wide and less than 30 m deep (Rose and Dunham, 1977, 80).

The sops possess a characteristic depositional sequence, although not all members of this sequence are necessarily found in each case (Fig. 4.1). The lowest bed consists of a limestone rubble of waterworn blocks up to a metre across. These have a white crust, apparently the result of partial solution (Dunham and Rose, 1949, 29). The blocks are randomly orientated and seem to result from collapse. Generally, a clay is found mixed with the rubble. It was described by Rose and Dunham (1977, 80) as "either a dark brown cave earth variety (probably the insoluble residue of the dissolved limestone), or a pale grey, white and pink clay known as 'hunger', or a nearly black manganiferous 'black muck'". According to Smith (1924, 39), the 'hunger' is partly composed of weathered shale like that interbedded with the limestone; an analysis of the 'black muck' is given by Kendall (1881-82, 225). At Nigel (SD202755) and at Anty Cross (SD229736), some blocks of sandstone resembling St. Bees Sandstone were also found (Smith, 1924, 39), while at Tytup (SD237755), the basal bed consisted of clay.

The second layer, known as the 'casing', consists of bright red clay containing angular fragments of haematite. It is between 10 cm and 1 m thick, and separates the basal limestone breccia from the ore. It is heavily slickensided

(Rose and Dunham, 1977, 80).

The next layer, the ore, comprises a mass of broken fragments and blocks of iron oxide minerals, with quartz and occasional manganese oxides, set in a red powdery haematite matrix or enclosed in 'black muck' (Rose and Dunham, 1977, 80). Lower Carboniferous shales are occasionally found embedded in the ore (Smith, 1924, 39). In the largest sops, a central core of sand exists. This is made up of large broken blocks indistinguishable from the St. Bees Sandstone (Rose and Dunham, 1977, 80).

It should be pointed out that the sops are not exclusively associated with haematite deposits and mineralisation, and that at least two hollows contain sand but no ore (Rose and Dunham, 1977, 80).

4.3 Summary of previous research on the sops of Low Furness.

Possibly the earliest reference to a post-Dinantian phase of karstification in the area was made by Binney (1847, 443-444). He claimed that fissures in the limestone had been opened by solution and infilled with haematite some time after the deposition of the limestone, and before the deposition of the St. Bees Sandstone (of Upper Permian-Lower Triassic age). In a subsequent paper (1867, 57-59) he provided details of fossils of Westphalian age, converted into haematite, found within the sops. He suggested that these had either been washed in with the waters that were the source of the iron, or that they had fallen into the sops when the haematite fill was in a soft state. Either alternative he regarded as clear evidence of a Westphalian date for the ore deposition.

Kendall (1875, 282), however, argued that under the conditions envisaged by Binney, considerable amounts of clastic material would also have been deposited with the iron. Kendall claimed that no such material existed, although descriptions of the sops noted the occurrence of quartz, black muck and red clay within and around the ore bodies (Rose and Dunham, 1977, 80). However, Kendall (1875, 270-271) also pointed out that considerable uncertainty was attached to the exact location of the fossils within the ore. He suspected that their actual source was the Upper Carboniferous beds which form the roof of some of the deposits. It is also possible that the fossils collapsed into the sops at some stage after deposition, to be converted later to haematite by metasomatic replacement.

Subsequent examination of specimens of the fossils by Kendall (1893, 86-87) led him to conclude that they were nothing more than concentric pieces of kidney ore, although it was impossible to say whether Binney's fossils were of the same nature. Binney's thesis was, however, convincingly demolished by Kendall's (1893, 283-284) observation that haematite deposits are not displaced by intersecting Hercynian faults, which would have been the case had they been of Westphalian age.

A similar theory to that of Binney was advanced by Phillips and Barker (1858), who regarded faults and joints in the limestone as having been opened by water action to form depressions and caves in which haematite was subsequently deposited. The date of deposition was regarded as probably Permian, although a Triassic age was also suggested. Neither date was supported by any evidence.

Harkness and Murchison (1864, 152-153) proposed a phase of karstification in pre-Permian times, during which caves were developed and joints opened up allowing haematite to be deposited. The infilled fissures were overlain by "crab rock", which they regarded as a Lower Permian breccia, thereby providing a minimum date for the phase of karstification. However, Kendall (1875, 280-282) subsequently demonstrated that the "crab rock" was of probable Pleistocene age, and this was later confirmed by Dunham and Rose (1949, 13).

Brockbank (1867, 60-61) considered that the "clefts and cavities" in which the haematite is found were the result of a combination of tectonic and subaerial processes. On the evidence of the Westphalian fossils found in the haematite by Binney (1867), Brockbank concluded that ore deposition took place during the Carboniferous. He also referred to pebbles of haematite, presumably reworked, found in Permo-Triassic deposits at Rougholme Point, Humphrey Head and at St. Bees, which provide a minimum date for infilling of the sops. Later, Brockbank (1875, 290-292) proposed that the haematite was deposited in Early Permian seas into fissures on the denuded Carboniferous surface. Later deposition would have occurred in caves subsequently developed in the limestone.

Würzburger (1872, 138) pointed out that some haematite was deposited in pre-existing hollows in the limestone, but gave no indication as to when or how the hollows might have formed.

Perhaps the most important early work on the sops of Furness was in a series of papers by Kendall. In the earliest of these, published in 1875, Kendall suggested that at least part of the haematite deposition in Furness took place in solutionally opened fissures and caves (1875, 256-258, 266-272). He regarded the source of the haematite fill as downward percolating waters from the iron-rich sandstones and shales of the overlying Westphalian Coal Measures (1875, 278-279, 283). Given that the Westphalian beds had been eroded from the area by Permian times (see 3.4), this suggests that the haematite had been deposited by early Permian times at the latest. Further support for such an age is provided by the existence of rounded pebbles of haematite within the Permian Basal Breccia at Whitehaven (Kendall, 1875, 283), and by the non-displacement of ore bodies by faults which shift Westphalian beds (Kendall, 1893, 284). However, those later faults which dislocate the St. Bees Sandstone are regarded as having also displaced the ore bodies. By implication, therefore, the ore must pre-date this second movement (Kendall, 1893, 285).

Kendall's proposed date for haematite deposition has been the focus of considerable criticism. It was pointed out by Aitken (in Kendall, 1875, 303-304) that the shales within the Westphalian beds would have rendered them impermeable to downward percolation, and that sufficient water could not have made its way down faults in the shales to the underlying limestone.

More substantive was the criticism of Kendall's explanation of the faulting associated with the ore bodies. Smith (1919, 32; 1924, 40) found it impossible to accept that ore bodies, supposedly deposited in pre-Permian times, had been faulted by movements that displaced Triassic beds. He could find no certain case, with the exception of the Baybarrow fault (Smith, 1924, 40), in which an ore body had been subjected to faulting since its formation, except to a very minor degree. Where flats of ore occurred in the same bed of limestone at different levels on either side of a fault, Smith felt that the most that could be said was that the bed was particularly susceptible to metasomatic alteration. He concluded that deposition must have occurred after the post-Triassic fault movements.

Kendall (1920a, 59-60; 1920b, 284) countered this criticism by providing geological sections which demonstrated that, along certain faults, the displacement of the ore body associated with the Carboniferous Limestone, was greater than that of the Permian Basal Breccia, thus implying a second movement along these faults in post-Permian times. Unfortunately, these displacements were determined from borehole records on the assumption that the junction between successive beds was a plane surface. Yet the Basal Breccia, at least, was deposited on an uneven surface (see 3.4.2).

Smith's (1919; 1924) interpretation of the faults was supported by Trotter (1945, 68-69). The ore bodies are not brecciated against post-Triassic faults and they cross such faults without displacement. Trotter also reviewed the evidence for slickensided surfaces within the ore bodies. In some cases, these may indicate slight minor adjustment since deposition; in others, the haematite actually coats slickensided limestone surfaces.

The evidence of pebbles of haematite within the Permo-Triassic deposits of the area (Brockbank, 1867, 61; Kendall, 1875, 283) has posed a difficult problem for workers seeking to establish a later date for mineralisation. Goodchild (1881-82, 118; 1889, 64) considered that the pebbles were probably of limestone, post-depositionally altered to haematite, but Kendall (1920a, 60) pointed out the existence of adjacent pieces of pure haematite and pure limestone in the Permian Basal Breccia at Whitehaven which it is impossible to imagine were post-depositionally converted. The problem was convincingly resolved by Trotter (1939, 410; 1945, 68) who considered that both altered and non-altered pebbles could have been derived from the pre-Permian zone of arid or lateritic staining (see 3.4.3). Finally, Trotter (1945, 68) referred to the evidence for mineralisation of both the Basal Breccia (the matrix as well as the pebbles) and the St. Bees Sandstone as confirmation of the post Permo-Triassic age of the ore formation.

In a subsequent paper, Kendall (1881-82) modified his views. After comparing the Furness haematite with similar deposits in Cumberland and after a close study of the stratigraphical relations within the sops and the small-scale structure of the deposits, he no longer supported the idea of deposition in caverns. Instead, he regarded the haematite as a replacement deposit resulting from the upward movement of iron-bearing solutions associated with a phase of Hercynian igneous activity (1881-82, 230-236). Any associated karst features were developed subsequently, adjacent to the ore bodies. Kendall believed that this would have occurred by the movement of aggressive waters through the ore bodies (1881-82, 236), although it could equally have been by the concentration of groundwater flow along the impermeable boundaries of the haematite bodies. The resultant cavities were later infilled by sand and clay (1881-82, 236-37). Kendall supported this thesis by contrasting the lack of clastic material in the ore with the sand and clay deposits surrounding the ore bodies.

A number of flaws exist in Kendall's argument. Firstly, he failed to explain the limestone breakdown found at the bottom of the sops. This is not in situ limestone, dissolved by percolating waters for, if it were, it would display a characteristic fabric (see 4.2). It is a collapse feature, implying a pre-existing cavity. Secondly, in contrast to Kendall's view, considerable quantities of clastic material are found mixed with the ore (see 4.2). Thirdly, in the largest sops, large blocks of probable St. Bees Sandstone formerly existed within the ore. It is impossible to imagine these being transported by percolation waters, and again some form of mass movement into an open space is implied.

The first worker to make explicit the concept of karstification as a significant process in sop formation was Smith (1919; 1920; 1924). He suggested that, some time before the Permo-Triassic, the Carboniferous Limestone was exposed as a result of denudation, facilitating the development of dolines and caves. These features experienced occasional collapse, with the result that they were lined with a layer of angular limestone blocks, and as they became abandoned they were filled by inwashed material. During the Permo-Triassic, this landscape was buried by magnesian limestones, sandstones, etc. At some later date, however, the area was uplifted, faulted and exposed to renewed denudation. Presumably this took place during the Alpine orogeny (see 2.3). Smith considered that one result of this uplift was the initiation of the downward movement of iron-bearing waters, whose source was presumably the Permo-Triassic beds, although the processes involved are unclear. These waters attacked the limestone breccia in the caves and dolines. As the breccia was altered, the newly formed ore packed together with the result that the overlying Permo-Triassic beds began to sag into the incipient sops.

Smith (1919, 31) noted "jumbled masses" of Permo-Triassic sands and inwashed clays within the sops, which he visualised as the result of collapse consequent upon mineralisation. Kendall (1920b, 283; 1921, 146-147, 149-150), however, disputed Smith's observations and claimed that the sands and clays existed as distinct masses within the ore, and that they could not have gained such positions as a result of collapse. Although this may be the case, it does not materially alter Smith's hypothesis, for Kendall proposed that the sands and clays were introduced into voids in the ore by downward-percolating waters. The source of these materials was regarded as the overlying drift and the sandstone outcrop to the south of the area, deposition within the ore presumably taking place during Quaternary times.

In response, Smith (1924, 42-43) admitted that some of the sand and clay intermixed with the ore was derived from drift deposits, but he maintained that the majority of the sands were derived from the Triassic beds. He cast doubt on the transport of sands into the ore bodies through sub-surface conduits from the present outcrops of the St. Bees Sandstone to the south, and demonstrated the distinction between the sands within the sops and the glaciofluvial sands of the area.

Kendall (1921, 148-149) suggested that, had the sops formed immediately beneath the Permo-Triassic beds, they would by now have been destroyed by subaerial processes. Smith (1924, 42) accepted that denudation could have taken place, but he pointed out that the existing sops do not necessarily represent the whole of the initial form. If 120 m of the Park sop had been removed, for instance, it would still leave a sop over 60 m deep. Furthermore, the low quality ore within some sops suggests that erosion did indeed take place and that only the lowest deposits remain in a number of cases. Finally, Kendall's objections are invalidated, given interstratal karstification, as proposed by Dunham and Rose (1949) and detailed below.

Perhaps more relevant was Kendall's (1921, 146) observation that many of the blocks of probable St. Bees Sandstone within the ore are vertically orientated and hence in positions which they could not have adopted as a result of collapse. Dunham and Rose (1949, 32), on the other hand, implied that the St. Bees Sandstone within the sops did not occupy its initial depositional position and that the breccia which it forms must be the result of collapse. They disagreed with Kendall in that they proposed that, given Smith's suppositions, the St. Bees Sandstone would have been deposited directly into the depressions. The existence of the breccia led them to conclude that this could not have occurred.

They therefore proposed alternative theses whereby interstratal karst developed beneath a St. Bees Sandstone cover. Such a process is supported by the existence of sops only where the St. Bees Sandstone rested directly on the Carboniferous Limestone. Nothing comparable to the Permian Brockram, Magnesian Limestone or St. Bees Shale has been found in the sops, and no sops have been found in numerous borings through the St. Bees Shale. Dunham and Rose therefore took the view that the sops could only have formed where overlying strata allowed the downward movement of water (1949, 29-32). Similarly, in west Cumbria, Trotter et al (1937, 70-72) concluded that haematites are restricted to areas where permeable beds rest directly on Carboniferous Limestone.

Given the existence of karstic cavities, Dunham and Rose provided two explanations of the stratigraphy of the sops. The first involves a phase of cave breakdown to provide the basal limestone breccia, followed by the introduction into the void of iron from mineralised waters and, finally, collapse of the St. Bees Sandstone. It should be pointed out that the last two stages of this sequence were not stated explicitly. The model explains neither the unaltered nature of the basal limestone breccia nor the existence of the casing layer between the haematite and the basal breccia (although both the casing and the "clay" matrix of the basal breccia may be the result of the transport of fines in suspension by percolation waters).

According to the second explanation, the karstic voids would have been mineralised prior to the formation of the limestone breccia. The subsequent brecciation may have been the result either of the concentration of groundwater flow through the ore body or of effluent mineralising waters, though again this was not made explicit. The phase of collapse would have let down the haematite, providing a void for the ultimate collapse of the St. Bees Sandstone. The casing and breccia matrix could have been formed as outlined above. This model is similar to the diagrammatic interpretation provided by Rose and Dunham (1977, 91) (Fig. 4.2), with the exception that mineralisation occurs not only of the void, but also of the surrounding bedrock, and all the limestone through which the mineralising waters percolated is replaced.

One point Dunham and Rose (1949) failed to recognise in their criticism of Smith's theory is that collapse of the St. Bees Sandstone was thought to have taken place even after its deposition within the depressions. Nevertheless, the process by which this was considered to have occurred is invalidated because of its dependence on the mineralisation of the basal limestone breccia. Smith stated explicity (1919; 1920, 17; 1924, 39) that the mineralising solution attacked the basal limestone breccia, but a study of the deposits shows the breccia to be unaltered (apart from minor solution and reprecipitation): it is above this bed, and separated from it by the red clay casing, that the ore has been deposited.

4.4 Implications of the sops for palaeoenvironmental interpretation

4.4.1 The sequence of sop development

Two main models have been advanced to explain sop development: buried surface karst and interstratal karst. The buried surface karst model has been shown to be intenable, for it requires mineralisation of the sop fill to take place

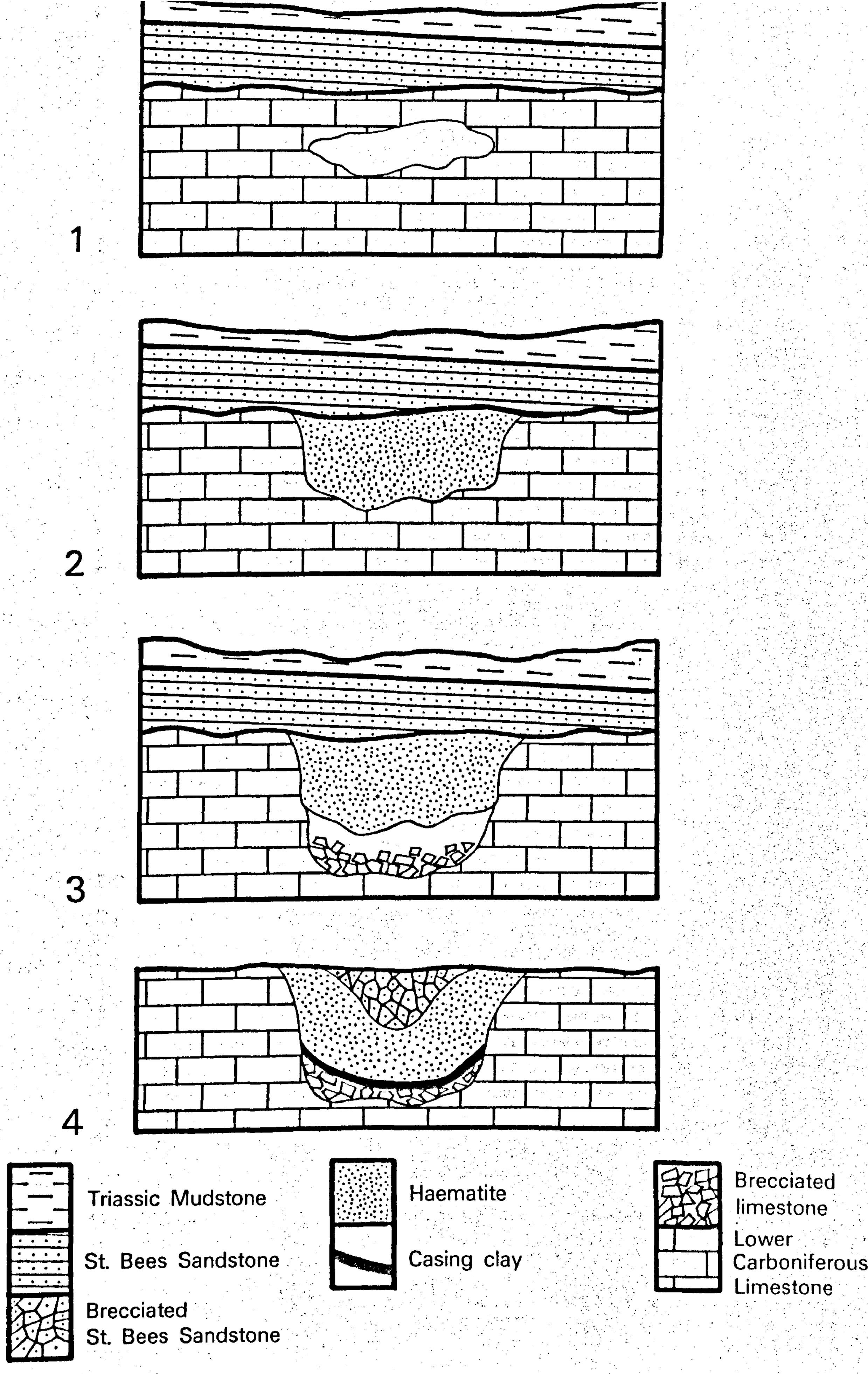


Fig. 4.2 Hypothesised stages in the formation of sops (partly after Dunham and Rose, 1949)

without any replacement of the basal limestone breccia. There seems to be no evidence for mineralisation of the breccia, and given the conditions implied by the model it is difficult to understand how this would have failed to occur. On the other hand, the processes envisaged by the interstratal karst model of Rose and Dunham (1977, 91) appear likely, once certain points have been made explicit.

4.4.2 The environment of sop formation

The application by Rose and Dunham (1977, 87) of modern analytical techniques to the ore deposits has resolved much of the controversy surrounding the processes of ore deposition. Analysis of fluid inclusions within calcites intimately associated with the haematite has provided an indication of temperature conditions during the closing stages of mineralisation. Temperatures of the order of 84-110^{o}C appear to rule out the possibility of mineralisation from downward-percolating meteoric waters.

Salinity measurements from fluid inclusions in samples from Hodbarrow sop (SD1778) have given values four to six times that of sea water. The calcites were thus deposited from hypersaline brines, and it is reasonable to assume that this also applied to the iron oxides and the quartz.

On the basis of these findings, Rose and Dunham (1977, 90-92) proposed a mechanism of mineralisation whereby hypersaline fluids were driven up-dip along the permeable St. Bees Sandstone beds from a heat source in the northeast Irish Sea basin. Leaching of the St. Bees Sandstone would have provided the necessary quantities of iron, the iron having been initially derived from the iron-rich Carboniferous surface (see 3.4.3). When the warm, ferriferous brines reached places in west and south Cumbria where they had free access to existing fractures, to replaceable limestones and to karst features, they are believed to have descended into the limestones to form the ore deposits of the present day.

Recent work by Shepherd (1974) in the west Cumberland ore-field has provided evidence for an alternative mineralisation mechanism. Study of the geochemistry of both metasomatic and vein deposits has revealed simple and homogeneous chemical assemblages, with minor variations explicable in terms of wallrock-ore fluid interactions. Shepherd suggested that this indicated a single source, and he favoured the granitic basement rocks of the region. This is in agreement with studies which suggest epigenetic metalliferous mineralisation in central and northern England to be related to hydrothermal convection maintained by high heat production in granitic basement rocks (Brown et al, 1980).

The model proposed by Shepherd involved the convective circulation of hot saline fluids within the granite, leaching iron and redepositing it at higher levels. Firman (1978, 239) queried the post-Carboniferous date of ore deposition, rather than the Caledonian date to be expected given the Caledonian age of the granitic intrusions. Nevertheless, he acknowledged that conditions may not have been favourable for haematite formation until the (?)Kimmerian uplift allowed oxidising waters draining a lateritic subsoil (see 3.4.3) to mix with iron-rich brines rising from the granitic basement.

Unfortunately, there is no evidence in the sops for the upward movement of mineralising waters (see 4.3) and the model also fails to explain the location of sops only beneath those Permo-Triassic beds of a permeable nature. It may be possible to reconcile Shepherd's model with that of Rose and Dunham if the hot mineralising waters leached iron from the Permo-Triassic beds and redeposited it by downward movement into the limestone. The convective heat source might have been the same as that envisaged by Rose and Dunham, whilst the movement of the waters up-dip by tectonic pressures may explain the date of mineralisation (see 4.4.3).

The hydrological conditions under which interstratal karsts develop are little known. Even presuming the sops to have been formed by the solutional action of hot mineralising waters, a low-level hydrological outlet must have existed within the limestone. Without that, groundwater movement would have been slow, waters would have become rapidly saturated and karstification would not have occurred on the same scale. That an outflow level did exist at considerable depth is supported by the vertical range of mineralisation. Rose and Dunham (1977, 88) noted that most of the ore bodies either terminated, or show conclusive signs of so doing, at around -180 m O.D., even where structural and lithological controls still appear favourable to further development. At Park, the sop was bottomed at -187 m, whilst the Rita and Nigel sops were found to extend to -196 m and -183 m respectively. The Stank South Vein terminates at -192 m and the North Vein at -172 m. The Lowfield section of the Lindal Moor Vein was worked to a depth of -180 m, where it was found to be dying out. In the Moorbank South section at Hodbarrow, working continued to about -150 m with the orebody showing signs of pinching out; and at the Whicham Mine the base of the orebody was reached at about the same depth.

An outflow level at a depth of perhaps -200 m O.D. implies a considerable relief during the period of sop formation. It also suggests the existence of a relatively low sea level, possibly associated with Alpine uplift.

One interesting feature of the sops is that they all appear to have experienced a phase of breakdown at approximately the same time. A possible explanation of this is a rapid regional lowering of groundwater levels causing a resultant fall in pressure. This explanation has been proposed by Sweeting (1950, 75) for northwest Yorkshire and by Brink and Partridge (1965, 25-33) for South Africa, although in the Morecambe Bay area it seems difficult to reconcile this with the low groundwater levels necessary for sop formation.

4.4.3 The date of sop development

It is likely that the phases of sop mineralisation and karstification were roughly contemporaneous; or at least that mineralisation reinitiated karstic development. As shown in 4.3, the mineralisation post-dates the period of post-Triassic faulting. Furthermore, the infilling of the sops must have pre-dated the removal of the Permo-Triassic cover from the area. In central and northern England, a mid-Tertiary age is usually assigned to faults that can be shown to displace Mesozoic strata (Owen, 1976, 144). This coincides approximately with the culmination of Alpine tectonic activity in Miocene times. However, the possibility of earlier fracturing must not be ignored, for there is a continuous history of pulsed tectonism in northern England from Hercynian times to the present day (see 2.3).

It does not appear unreasonable to suggest that the karstification which initiated the sops was the result of this tectonic activity. Fracturing could have provided fissures for groundwater movement, uplift could have resulted in greater hydraulic gradients, and increased denudation could have enabled the circulation of meteoric waters through the limestone. Furthermore, tectonic pressures could have driven ferriferous formation waters up-dip from the Irish Sea to the Morecambe Bay area, as proposed by Rose and Dunham (1977, 90-91).

Thus, if a provisional Oligocene-Miocene age is accepted for the faulting, along with a similar age for sop formation, a Late Palaeogene-Early Neogene date can be cautiously assigned to the Morecambe Bay interstratal karst. Such a date is broadly comparable to those of karst development in Derbyshire (Walsh et al, 1972) and the South Downs (Curry et al, 1979, 50-51)[1]. It is not implied that the Tertiary must be regarded as an unusually active period of karstification in Britain. On the contrary, the evidence points to karstification taking place almost continuously in one place or another in the British Isles since the Carboniferous (Table 4.1).

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[1]A Tertiary age has also been proposed on the basis of less reliable evidence for karst features in northwest Yorkshire, northeast Wales and south Wales (see Table 4.1).

Date of karstification	Location	Reference	Comments
Upper Pliocene-Lower Pleistocene	Derbyshire	Dawkins (1903)	Fissure fill contained fauna regarded as of Upper Pliocene-Lower Pleistocene age. Re-interpreted as of Lower Pleistocene age by Spencer and Melville (1971).
Pliocene	Durham	Trechman(1915; 1919) Reid (1920)	Fissure fill contained flora and fauna of Pliocene age overlain by glacial material. Nature of deposits suggested that Pliocene material had been reworked by ice. West (1968, 297-298) reinterpreted the organic deposits as of Lower Pleistocene age.
Pliocene	Kent	Curry *et al* (1979,50-51)	Chalk solution pipes infilled by Lenham Beds which include marine mollusca of Pliocene age with some reworked Miocene forms. According to Worssam (1963, 85), pipes have developed beneath the Lenham Beds, which have collapsed into them.
Upper Miocene-Lower Pliocene	Derbyshire	Walsh *et al* (1972)	Closed depression fill contained flora of Upper Miocene-Lower Pliocene age.
Upper Miocene-Lower Pliocene	Clwyd	Walsh and Brown(1971)	Lithology of fill and altitude of closed depressions regarded as similar to those closed depressions of Mio-Pliocene age found in Derbyshire.
Upper Miocene-Lower Pliocene	Powys	Thomas (1973)	Interstratal karstification; no firm evidence of date.
Middle Tertiary-Present	Southeast Dyfed-Southwest Powys	Thomas (1963)	Interstratal karstification dated on evidence of extrapolated rates of subsidence and relation of karst features to "Miocene" erosion surfaces.

Middle Tertiary	Powys	Thomas (1954a)	Interstratal karstification; no firm evidence of date.
Upper Eocene	Tipperary	Watts (1957; 1962)	Closed depression fill contained flora of Tertiary (?Upper Eocene) age.
Lower Palaeogene	Hertfordshire	Kirkaldy(1950); Thorez et al (1971)	Chalk solution pipes and closed depression infilled by Thanet Beds and Reading Beds of Lower Palaeogene age. Pipes considered to have developed beneath the Thanet Beds, but the depression contains nearly horizontal Palaeogene deposits.
Tertiary	North Humberside	Versey (1937)	Chalk solution pipes infilled by sand regarded as similar to material of "Tertiary" age. Likely that pipes post-date deposition of sand. One depression reinvestigated by Bray et al (1981); shown to be not necessarily of karstic origin and fill to be of penultimate glacial age.
Tertiary	Tipperary	Wynne (1857)	Lignite found in pipeclays infilling pockets in Lower Carboniferous Limestone. Presence of lignite not confirmed during subsequent investigation by Bishopp and McCluskey (1948).
Tertiary	Cork	Murphy (1966)	Gravity measurements indicate closed depressions infilled by unconsolidated material. Depressions assigned Tertiary age by comparison with depressions of "Tertiary" age in Powys, Derbyshire, etc.
Upper Cretaceous	Kerry	Walsh (1966)	Dated on evidence of lithology and texture of fissure fill.
Upper Triassic	Glamorgan, Gloucestershire, Somerset	Robinson (1957); Marshall and Whiteside (1980)	Cave and fissure fill contained flora and fauna of Upper Triassic age.

Triassic	Southwest Dyfed	Dixon (1921)	Lithology of depression fill regarded as identical to that of Upper Triassic age in Glamorgan. Depressions regarded by Thomas (1971) as tectonically formed.
Triassic	Somerset	Halstead and Nicoll (1971)	Cave and fissure fill contained fauna of Triassic age.
Permo-Triassic	Derbyshire	T.D.Ford (1969); Worley and Ford (1977)	Mineralisation of Permo-Triassic age infilled pre-existing caves and led to the solution, infilling and collapse of other karst features.

Table 4.1 The evidence for karstification in the British Isles during the Mesozoic and early Cenozoic

4.5 Mineralisation and palaeokarst in the limestone to the east of Low Furness

Although considerable evidence of mineralisation exists in the limestone to the east of Low Furness, minerals are rarely present in exploitable quantities (Holland, 1967). Moreover, the bulk of the mineralisation is in the form of veins developed along joints. The absence of large-scale mineralisation of palaeokarst features may be the result of a number of factors. It is possible that the large-scale movement of mineralising waters proposed by Rose and Dunham (1977, 90-92) was prevented by the more dissected geological structure of the area. Impermeable Permian beds might have prevented the downward percolation of these waters; alternatively, end-Carboniferous erosion might not have removed impermeable Upper Carboniferous beds, as it did in Low Furness (see 3.4.2). The absence of Permo-Triassic beds within the area implies more denudation than in Low Furness; hence, the overlying Permo-Triassic beds might have been removed by the time of mineralisation. On the other hand, such erosion might have removed any trace of mineralised palaeokarst features.

Work by C.M. Moseley (1969), however, suggested that some mineralised palaeokarst features do exist in the Warton Crag area (SD4873). This proposal is based on a number of lines of evidence. The faces of many worked-out ore cavities show evidence of solutional pocketing (1969, 13) and open spaces exist above ores found in cave passages (1969, 19), implying that the ores were deposited in pre-existing solutional cavities. However, these features do not exclude the possibility of local solution around the edges of the ore bodies.

Moseley (1969, 19) also pointed out that some ores are horizontally bedded and consist of water-lain arenaceous deposits which have been post-depositionally mineralised (1969, 20). This observation provides more convincing evidence of pre-mineralisation karstification.

4.6 Conclusions

The sops of Low Furness, along with similar features in the rest of the Morecambe Bay area, appear to be interstratal karst features which have been mineralised by downward-moving, warm, ferriferous brines, and which have subsequently collapsed to let down the overlying St. Bees Sandstone. The vertical range of the karst features suggests a former outflow level for both karst and mineralising waters at approximately -200 m O.D., implying the existence of a considerable relief during the period of sop formation. On the grounds of the relation of the sops to post-Triassic tectonic features, the date of sop infilling is cautiously assigned to the Late Palaeogene-Early Neogene.

III : QUATERNARY ENVIRONMENTAL HISTORY

5. CAVE DEPOSITS IN PALAEOENVIRONMENTAL RECONSTRUCTION

5.1 Introduction

By their very nature, caves tend to act as sediment traps for a wide variety of deposits; some may also provide shelter for man and animals or act as pitfalls for animals. The materials deposited in the cave environment are protected from the erosive effects of many subaerial processes. Thus, caves frequently exhibit long and complex depositional sequences which are partly a reflection of external environmental conditions and partly a reflection of the local cave environment.

The significance of cave sediments in environmental reconstruction has long been recognised, systematic study of cave deposits beginning with Pengelly's excavation of Kent's Cavern in Devon between 1865 and 1880. Much of this work has, however, been archaeological in nature, concentrating on the interpretation of cultural artifacts and faunal remains. For this reason the majority of the work has involved the study of entrance facies deposits (for example, Brain, 1958; Sutcliffe, 1960). Neverthelesss, since the 1950s, an increasing proportion of work has involved the study of clastic and chemical cave-sediments. This approach has mainly followed two lines. Firstly, there has been the application of granulometric and mineralogical techniques in an attempt to discover the origin and depositional environment of clastic cave-sediments (for example, Frank, 1969; Campbell, 1977). Secondly, there has been the study of carbonate precipitates in caves, most recently using isotopic methods such as ^{14}C (Broecker and Olson, 1959), $^{18}O/^{16}O$ (Labeyrie, Duplessy, Delibrias and Letolle, 1967) and $^{230}Th/^{234}U$ (Ford, Thompson and Schwarz, 1972) in order to establish cave palaeotemperatures and provide a geochronometric framework for cave sediment studies.

A significant proportion of recent work has also involved the study of deposits found deep inside cave systems, enabling the reconstruction of the environmental history of karst drainage routes (for example, Bull, 1976; Atkinson, Harmon, Smart and Waltham, 1978). This approach has been taken a step further by a small number of workers who, not content merely to describe sediments and assign them to particular environments, have investigated the processes of sediment transport and deposition, concentrating particularly on fluvial cave-sediments (Burkhardt, 1958; Collier and Flint, 1964; Renault, 1968; White and White, 1968; Reams, 1972).

In the present study, both entrance facies and hydraulically-transported cave sediments are investigated. A variety of interpretative techniques are used, including the study of sediment grain-size to establish environment of deposition, the study of pebble petrography to establish sediment provenance, and the study of faunal and cultural remains (see 5.2). Sediments from all known caves in the Morecambe Bay karst have been studied. Detailed analysis has been restricted to three caves: Kirkhead Cave (see 5.3) and Dog Holes (see 5.4), both of which contain significant faunal and cultural remains, and Fissure Cave (see 7.3) which contains a complex sequence of hydraulically-transported deposits.

5.2 The determination of the depositional environment of cave sediments: methodology

5.2.1 Grain-size study of clastic sediments (see 11.3 for sampling and laboratory procedure)

Since Udden's pioneer work of 1898 and 1914, there have been many attempts to use grain-size parameters to characterise environments of deposition (see Folk's review, 1966). However, with few exceptions (for example, Friedman, 1961; 1967), these efforts have had less than impressive results and have been generally inapplicable beyond the study area concerned. Moreover, they have been unsuccessful in differentiating more subtle differences in grain-size distributions such as those between fluvial and quiet-water deposits. However, in a recent series of papers, Buller and McManus (1972a, b, c; 1973a, b, c; 1974; 1975) have established a series of environmental envelopes for non-biogenic sediments based upon simple grain-size parameters. Using data from several thousand samples from a wide range of sedimentary environments, they found that plots of median grain size (D_{50}=Md) against quartile deviation ((D_{25}-D_{75})/2 = QDa) provided distinct groupings of sedimentary types. They concluded that, for water-borne and airborne sediments, these parameters reflected the energy conditions and fluid viscosity of the environment (1972a, 16), whilst for glacially derived sediments they reflected the effects of population addition and subtraction (1973a, 141-143).

The distinct environmental envelopes which can be derived by using this approach and the ability to distinguish by its use subtle changes in environment of deposition made it an obvious choice as an analytical method in the present study. Buller and McManus' envelopes of glacial and freshwater deposition (1972a, 15; 1973a, 138) were replotted as a basis for sediment analysis. The quiet-water envelope was redrawn to exclude the marine quiet-water environment. Onto this graph were plotted the Md and QDa values for each sediment analysed (Fig. 5.1).

5.2.2 Petrographic study

Rock fragments are perhaps the best criterion for use in interpreting the provenance of a sediment (Humphries, 1978, 595). This is mainly because, unlike particles of smaller size, they do not tend to survive continued reworking from environment to environment. Within each sample, therefore, particles of pebble size and larger ($\geq$4 mm) were analysed and the frequency of occurrence of petrographic types calculated on a weight basis (Fig. 5.2).

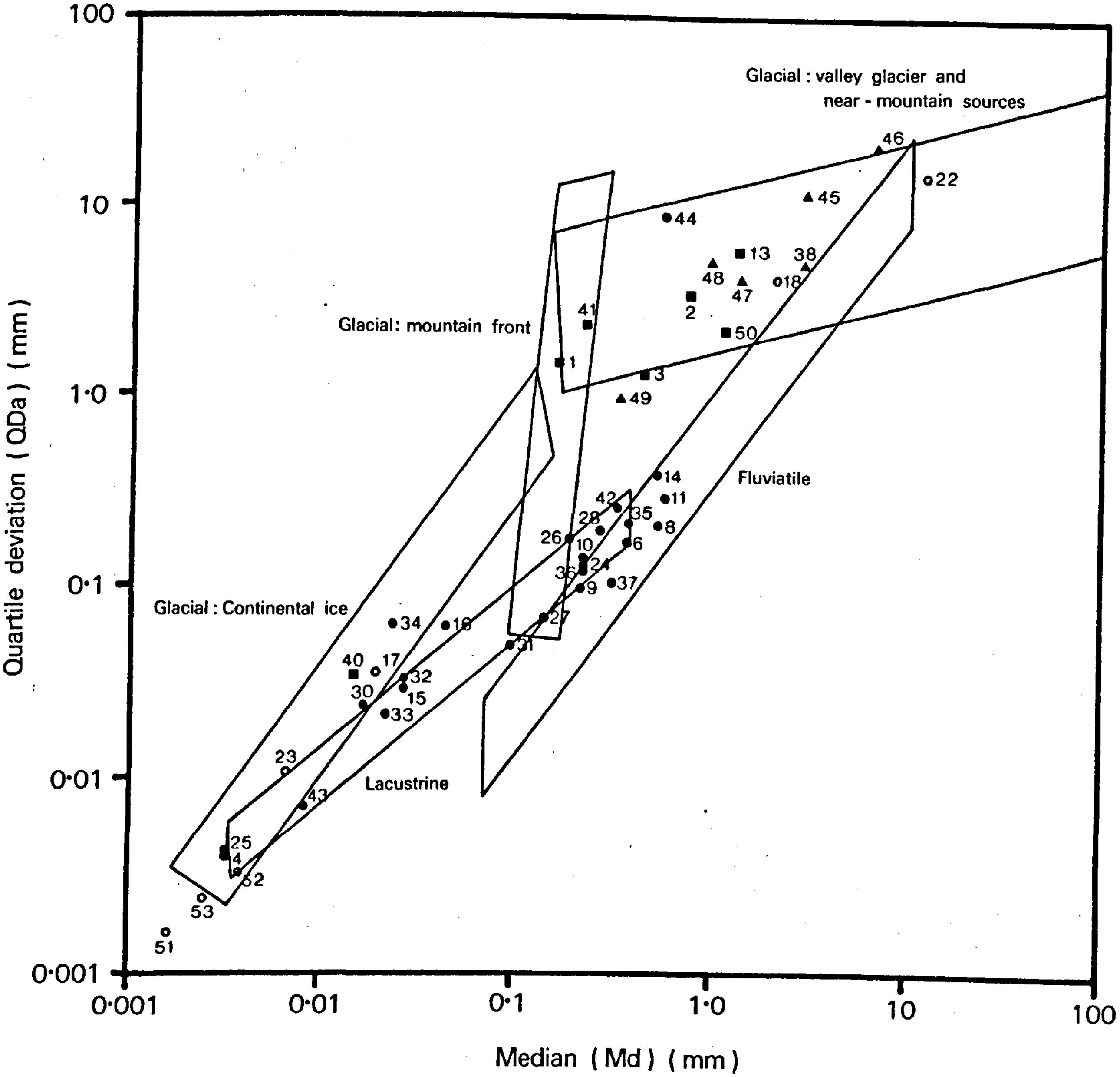
5.2.3 Faunal remains

Molluscan and mammallian remains were found within certain of the sediments investigated. In those cases where it could be certain that the remains were in situ, the ecology of the species was used to establish environmental conditions at the time of sediment deposition.

5.3 Kirkhead Cave

5.3.1 Introduction

Kirkhead Cave (SD39097562) consists of a single large chamber, phreatic in origin and partly infilled by sediments (Fig. 5.3). The cave stands at an altitude of 38 m O.D. and



- Sediments from Fissure Cave
- Sediments from Kirkhead Cave
- Sediments from other caves in the Morecambe Bay karst
- Surface sediments from the Morecambe Bay karst

Fig. 5.1 QDa - Md values of sediments from the Morecambe Bay karst (environmental envelopes after Buller and McManus, 1972a; 1973a)

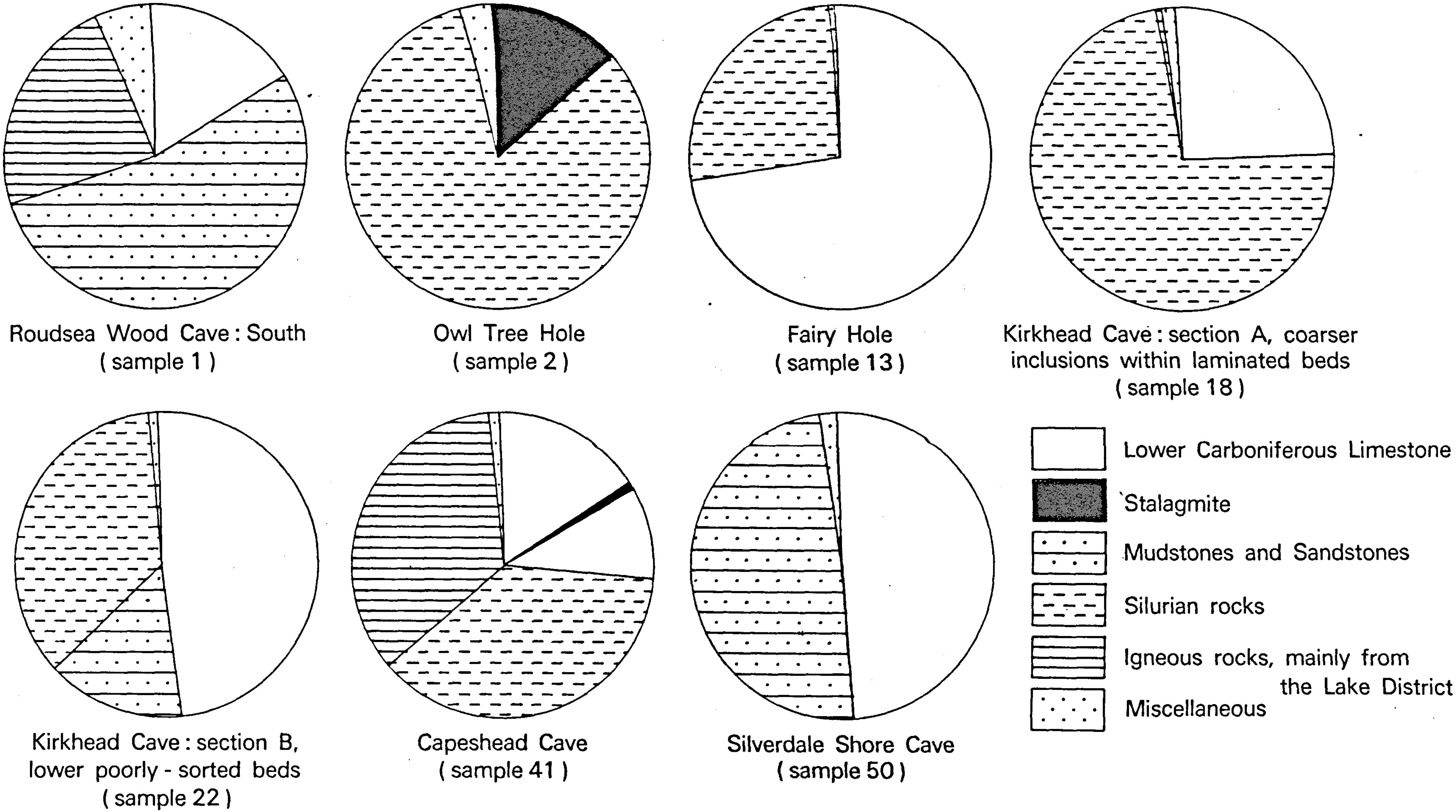


Fig. 5.2 The frequency of occurrence (by weight) of petrographic types within the pebble and cobble fraction ($\geqslant$4 mm) of cave sediments in the Morecambe Bay area

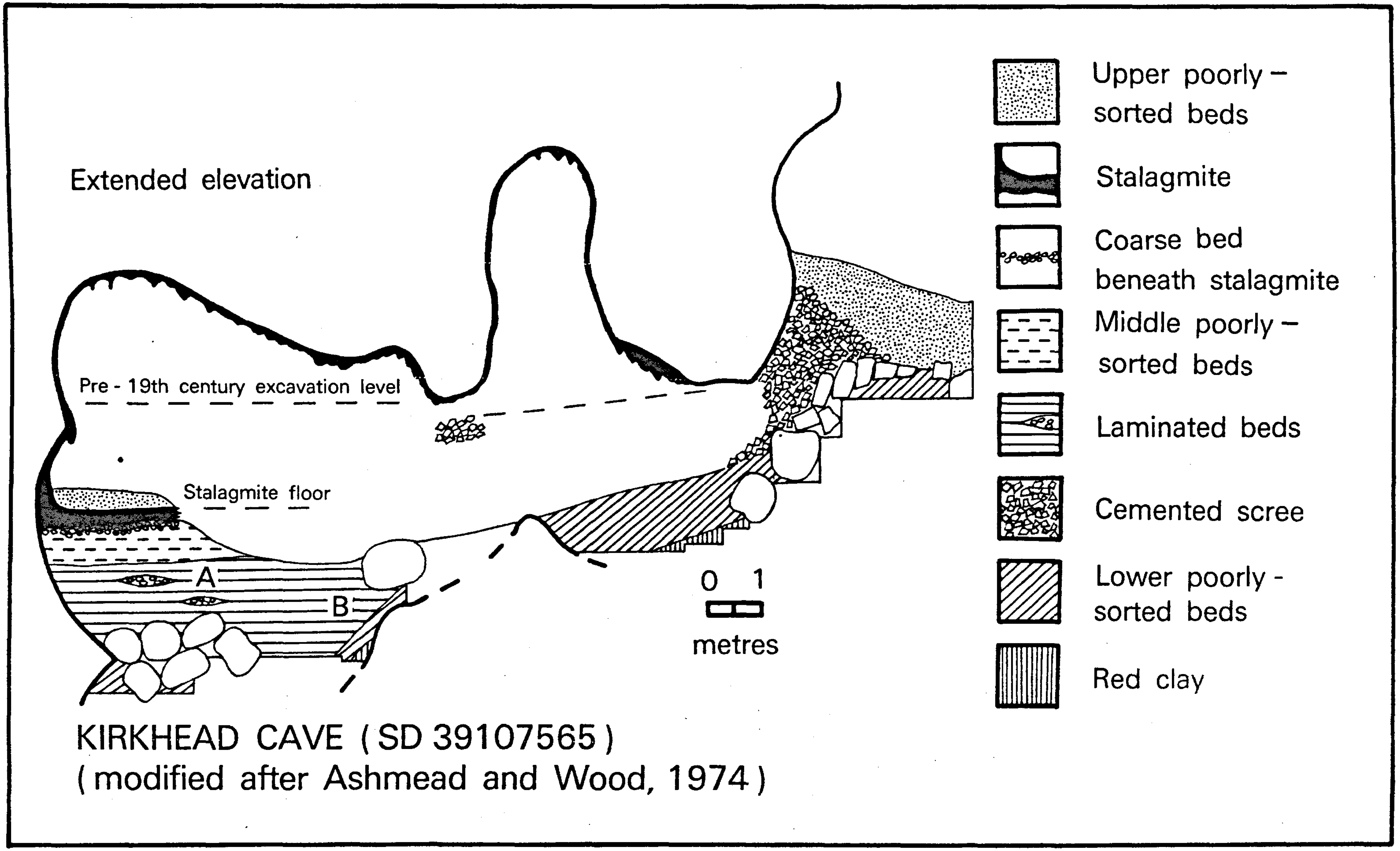


Fig. 5.3 Kirkhead Cave

is situated in a cliff on the west side of Kirkhead Hill, a headland jutting out into Morecambe Bay. Kirkhead Cave has been regarded both as a sea cave (Ashmead, 1969a, 204) and as having been enlarged by marine erosion (Ashmead, 1974a, 223-224; Ashmead and Wood, 1974, 30). There have also been attempts to relate it to an apparent marine notch found at the cave entrace and traceable along Kirkhead Hill (Ashmead, 1969a, 204; Wood, Ashmead and Mellars, 1969, 21). In fact, the concave break of slope at the entrance occurs where the top of the scree slope abuts against the free face in which the cave is found. There is no morphological or sedimentological evidence either outside the cave or in the area as a whole for the existence of a sea level at this height (see 8.2).

Kirkhead Cave constitutes an abandoned high-level phreatic system similar to those found elsewhere in the Morecambe Bay area (see 7.5.2). The cave displays no conclusive evidence of a vadose phase, although this may be largely because of the considerable depth of sediment within the cave.

5.3.2 Interpretation of the cave deposits

The lowest known bed in the cave has been described by Ashmead and Wood (1974, 28, 32) as "a stiff red clay ... of glacial origin". This bed was found resting directly on bedrock, but as the bed is no longer exposed it is uncertain whether or not lower horizons existed. The glacial deposits of the area characteristically have a reddish matrix and it is not unreasonable to interpret this horizon as derived from surface glacial deposits.

The overlying beds contain rather coarser material. These were originally regarded as river gravels including, where the deposit thickens at the back of the cave, rounded and polished cobbles interpreted as the result of fluvial transport (Wood, Ashmead and Mellars, 1969, 22). Later. however, Ashmead and Wood (1974, 32) regarded the deposit as a glaciofluvial gravel.

Analysis of the sediment (sample 22) shows it to be very poorly sorted (S_o= 10.2). The pebble fraction contains a large proportion of erratic material from the Lake District, including fragments of Ennerdale Granophyre (Fig. 5.2). The sediment falls clearly within the near-mountain source glacial envelope of the QDa-Md diagram (Fig. 5.1). It is therefore concluded that the deposit consists of little-modified glacial material, either deposited directly into the cave or slumped and washed in. That the latter mechanisms were important is indicated by the existence of rounded ($P_d \sim 0.7$) pebbles and cobbles near the back of the cave, a position they would have achieved by rolling. These cobbles consist largely of either Silurian rocks, derived from the north of the area, or limestone.

The next horizon, quantitatively the most important in the cave, consists of a series of laminated beds (samples 51, 52 and 53) (Plate 5.1) with occasional inclusions of coarser material (sample 18). At section A, these beds are approximately 50 cm thick, whilst at section B, they reach a thickness of approximately 110 cm and rest directly on bedrock. The deposits are horizontally or near-horizontally bedded, each lamination being approximately 1 mm thick. The beds were initially interpreted by

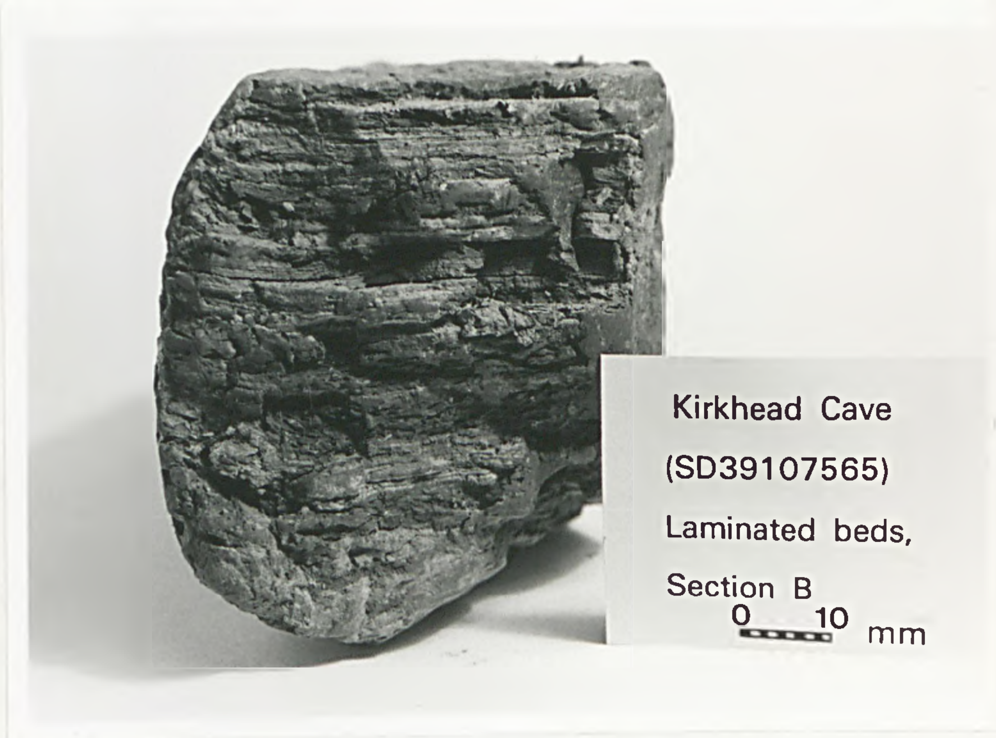


Plate 5.1 Laminated beds, section B, Kirkhead Cave (SD39107565)

Wood, Ashmead and Mellars (1969, 22) as glacial meltwater sediments laid down in a static water body inside the cave, although a later paper merely referred to the periodic washing of sands and gravels into the cave as having formed the laminated sediments (Ashmead and Wood, 1974, 32).

Analysis of samples from three consectutive laminae (samples 51, 52 and 53) shows each bed to be either clayey or muddy in texture and to have a similar grain-size distribution (Fig. 5.4). The deposits plot just outside the lacustrine and continental ice environment envelopes of the QDa-Md diagram (Fig. 5.1).

The interpretation of the mechanism and periodicity of deposition of laminated beds has long been a problem for sedimentologists. Many examples of such beds have been found in caves (see, for example, Sweeting, 1950, 73; Reams, 1968; Bull, 1980), but in almost all these cases the laminae exhibit coarse-fine couplets, hence the suggested processes of deposition are not necessarily applicable to the Kirkhead Cave situation.

In the case of the Kirkhead deposits, a number of depositional mechanisms may be proposed. Firstly, the deposits may have been laid down by sedimentation in a static water body. The absence of underlying or overlying hydraulically-lain deposits makes it unlikely that the source of such a water body was groundwater flow into the cave. Alternatively, therefore, the cave may have been flooded by glacial meltwater. Secondly, the deposits may represent seasonal or storm-related slope wash into the cave. Under these conditions, however, the beds would not be expected to be horizontal. Finally, a

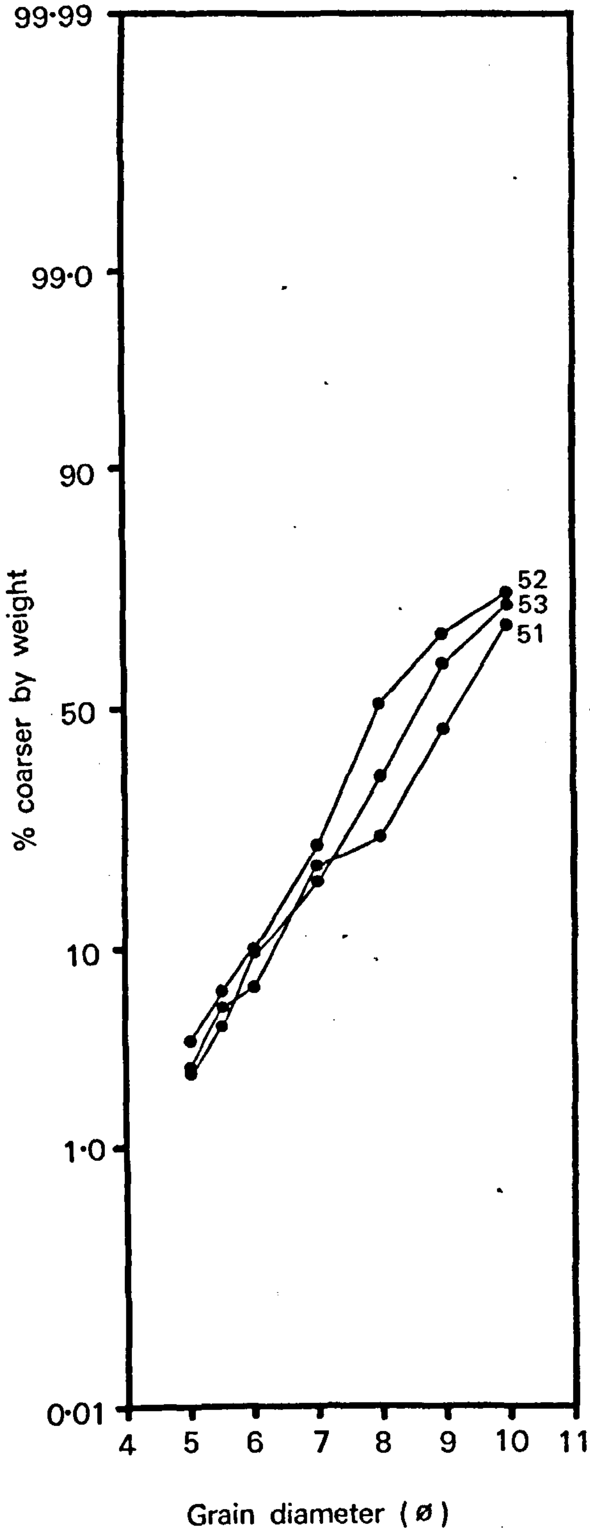


Fig. 5.4 Grain - size distribution curves of samples from consecutive beds, laminated beds, Section B, Kirkhead Cave

sediment "rain" from the ground surface via fissures into the cave, as proposed by Bull (1980), could have given rise to laminated beds, although under these circumstances, evidence of drip pit development within the laminated beds might be expected.

Above the laminated deposits is a very poorly sorted (So = 3.36), muddy-textured bed approximately 60 cm thick and containing occasional rounded granules of limestone and erratics from the Lake District (sample 23). As is found in the case of similar deposits in Fissure Cave (see 7.3.2), this material falls within the continental-ice environmental envelope of the QDa-Md diagram (Fig. 5.1), and, as with those sediments, it is suggested that the deposit was formed by the truncation of the coarser fraction of a near-mountain source glacial deposit. This truncation could have been achieved by selective transport of the finer fraction of the glacial materials found outside the cave, possibly by slope wash and wind, or even by trampling. The influence of man could well have been important, for it is within this bed that the earliest evidence of man's occupation of the cave occurs. Within the bottom 15 cm of the deposit have been found 10 flint implements, with a further 11 implements found in the same bed at the cave entrance (Wood, Ashmead and Mellars, 1969, 21; Ashmead and Wood, 1974, 27-28). According to Mellars (in Wood, Ashmead and Mellars, 1969, 23), typologically the flints could belong in either a Palaeolithic or a Mesolithic context. However, the shape and technique of manufacture of one of the tools is strongly suggestive of a Palaeolithic origin. Campbell (1977, 124, 168-169, 183) also proposed a

Later Upper Palaeolithic age for the implements. At the same level as the flints was discovered a large vertebra, possibly of an ox (Wood, Ashmead and Mellars, 1969, 21), whilst further evidence of habitation, including a whetstone and charcoal fragments, was found at the top of the bed (Ashmead and Wood, 1974, 27).

With the exception of the open-air site at Flixton, Yorkshire, where two artifacts have been found (Moore, 1954), Kirkhead Cave is the most northerly Upper Palaeolithic site in Britain. It is likely that conditions were too severe at this time for winter habitation and that the cave, along with the other Later Upper Palaeolithic cave sites in Craven, would have been used for summer exploitation of horse and reindeer in the surrounding district (Campbell, 1977, 169).

The Upper Palaeolithic is classically regarded as contemporaneous with the last glacial in Britain. In many areas, however, particularly in the north where environmental conditions were less favourable, Later Upper Palaeolithic cultures lingered into the Post-glacial. Thus, since reliable ^{14}C dates of Later Upper Palaeolithic/?Mesolithic sites occur as recently as 7602 ± 140 B.P. (Campbell, 1977, Table 4), the Kirkhead implements may be cautiously assigned to the period between the final retreat of the ice from the area at about 14000 B.P. (Pennington, 1978) and approximately 7600 B.P. This conclusion is supported by the ^{14}C date of $\sim$10700 B.P. obtained from bone taken from the same level as the flints in the cave (Mellars, pers. comm.).

The deposits in which the artifacts are found are overlain by a stalagmite floor, which is up to 30 cm thick in those places where it has infilled depressions in the underlying bed. The stalagmite is frequently porous and crumbly. It consists of a number of distinct bands, obviously representing phases of accretion. According to earlier investigators, the stalagmite formerly covered the whole of the cave floor, although remnants can now be found only around the cave walls. For the stalagmite to accumulate, it is likely that the cave was temporarily abandoned by man; possibly because continuing percolation made the cave an unfavourable environment. Nevertheless, both a flint flake (Morris, 1865-66, 359; 1866, 170) and several pieces of charcoal (Bolton, 1864, 252) have been reported as being found embedded within the stalagmite. Molluscan remains have also been found within the stalagmite (Wood, Ashmead and Mellars, 1969, 22).

The deposition of the stalagmite also resulted in the cementation of the limestone scree deposits which are found in the cave, since in some places the stalagmite floor and the scree cement can be seen to be continuous (Ashmead and Wood, 1974, 30). The scree appears to have been banked up against the cliff outside the cave and to have formed a detrital fan inside the cave (Fig. 5.3). The scree must have been deposited prior to the phase of stalagmite precipitation, but it is difficult to clarify its stratigraphic relationship with lower beds in the cave. Elsewhere in the area, similar scree deposits have been tentatively interpreted as the result of Late-glacial periglacial activity (see 8.5).

Owing to the activity of excavators during the nineteenth century, only traces remain of the beds which formerly overlay the stalagmite floor in the cave. However, reports all agree that approximately 2 m of "brownish-red indurated clay", containing angular fragments of Lower Carboniferous Limestone and rounded pebbles and small cobbles of Upper Ireleth Slate and Coniston Flags, overlay the stalagmite (Barrie, 1864; Bolton, 1864; 1869, 166-168; Morris, 1865-66; 1866). Analysis of material from the remaining parts of this bed (sample 17) shows it to be an extremely poorly sorted (S_0 = 5.46) silty mud, falling within the continental-ice envelope of the QDa-Md diagram (Fig. 5.1). The deposition of this sediment is explicable in terms of the same processes as those which deposited the bed immediately below the stalagmite.

It is difficult to establish a stratigraphy of the finds made in the upper deposit from the reports of the early excavators. A portion of deer antler was found resting upon and partly cemented to the stalagmite floor (Wood, Ashmead and Mellars, 1969, 21, 22), whilst within a stalagmite boss near the cave entrance were found a charcoal layer, a human skull and a few animal bones, including those of pig, boar and red deer. Near the base of the upper bed were found two bone implements, with further bone implements at a depth of about a metre in the deposit. Throughout the remainder of the deposit were found numerous mammalian and avian bones: horse, dog, rat, badger, ox, wild goose, wild cat, wild boar, fox, goat, pig and red and roe deer. Many of these bones had been split to extract the marrow. In association with the bones were found bone implements, fragments of crude pottery, human bones, sticks burnt at one end, and bronze ornaments and implements, including a flat axe of Early Bronze age (Ashmead and Wood, 1974, 24). A trial trench dug by Ashmead and Wood (1974, 27) in the northwest corner of the cave entrance yielded a flint of Late Neolithic/Early Bronze Age and a sherd of very coarse, blackened pottery, but the stratigraphic relationship of these finds to those of the early excavators is unclear. At the top of the deposit, just beneath the original floor of the cave, was found a Roman coin dating from the time of the Emperor Domitian (A.D. 81-96). A few centimetres lower were found a portion of an iron axe, a hammer and a knife blade, all apparently of Roman type (Morris, 1866, 168).

Overlying the cemented scree on the hillslope outside the cave is a Brown Calcareous Soil (Hall and Folland, 1970, 68), developed beneath a deciduous woodland cover. The soil almost certainly post-dates the scree cementation phase and is likely to be penecontemporaneous with the upper deposit inside the cave. Similar scree/carbonate precipitate/soil sequences have been found elsewhere in the area (see 8.5), as well as in southern England, where the soil has been interpreted as the result of hillwash due to tillage or overgrazing during the Neolithic to Iron Age and, exceptionally, later periods (Evans, 1978, 97-100). In the upper part of the soil was found modern pottery, glass fragments and the bones of birds and small rodents (Ashmead and Wood, 1974, 28).

5.3.3 Date of deposition of the stalagmite bed

The stalagmite bed in Kirkhead Cave is found in a similar stratigraphic position to that in almost every other cave in the Morecambe Bay karst (see 5.5.1 and 7.3.3) and hence may be of considerable importance as a chronological datum. Frank (1975) reviewed the extensive literature on the stratigraphic location of stalagmite beds, the processes of stalagmite deposition, and isotopic palaeotemperature and dating studies of stalagmite. He concluded that "a relatively wet and not too cold environment" was the optimum for the development of stalagmite, although there were numerous qualifications to this generality. $^{230}Th/^{234}U$ dating work by Gascoyne (1977) supported the conclusion that stalagmite forms most rapidly during warm periods. However, Gascoyne warned that cave temperatures may not necessarily reflect surface conditions. Finally, work by Atkinson, Harmon, Smart and Waltham (1978) suggested that the deposition of stalagmite in British caves occurred preferentially during non-glacial periods.

The deposition of stalagmite presupposes percolation of water into the cave. At present, percolation due to cave drips and flow down the cave walls is minimal. As there is no reason to believe that any major change has taken place in the hydrology of the overlying rock since the end of the last glaciation, this suggests that stalagmite deposition can be related to a phase of wetter, and possibly warmer, conditions during the Post-glacial. The stalagmite in Kirkhead Cave overlies cultural deposits of probable Late-glacial/early Post-glacial age, and is itself overlain by cultural deposits of Late Neolithic/Early Bronze Age. Organic material associated with the underlying cultural deposits has yielded a ^{14}C date of $\sim$10700 B.P. This suggests an early Middle-Flandrian age for the stalagmite, i.e.

approximately 8000-4000 B.P. This period includes that of the Atlantic (7120-5100 B.P. according to Shotton, 1977, 26), which seems to have provided optimal conditions for stalagmite development, for it appears to have been generally the wettest phase of the Post-glacial, as well as the warmest, although Osborne (1976) very cautiously suggested that temperatures in Britain remained constant from about 9500 B.P. to about 5500 B.P. The Atlantic also seems to have been the peak period of tufa precipitation in Britain (Evans, 1975, 76), whilst stalagmitic and tufaceous beds in similar stratigraphic positions to that in Kirkhead Cave have been assigned a Mesolithic date in caves in the Mendips (Tratman, 1975, 373).

5.3.4 Conclusions

The deposits in Kirkhead Cave represent an almost complete sequence of sedimentation from the Late-glacial to the present day. The lowest beds in the cave can be regarded, on granulometric and lithological grounds, as derived from glacial materials. These are overlain by laminated deposits which probably reflect some form of periodic sediment input. Above these beds, within deposits possibly derived from slope wash into the cave, is found evidence of Later Upper Palaeolithic occupation, which has been dated at ~10700 B.P. These finds make Kirkhead Cave the most northerly Palaeolithic site of any significance in Britain. An overlying stalagmite floor has been cautiously assigned to the Atlantic. This is itself overlain by a bed containing further evidence of human occupation, with Bronze Age remains overlain by Romano-British and modern deposits. This bed also includes artifacts dating back to the Late Neolithic/Early Bronze Age.

5.4 Dog Holes

5.4.1 Introduction

Dog Holes (SD48407295) is an abandoned phreatic cave developed at an altitude of approximately 55 m O.D. beneath a structural bench on the southwest side of Warton Crag. As in the case of Kirkhead Cave, Dog Holes displays no conclusive evidence of a vadose phase, although this may be largely because of the considerable depth of fill found in the passages. Excavations made by Jackson in various parts of the cave during the period 1907-1912 revealed evidence of faunal and cultural remains having implications for the environmental history of the area as a whole.

5.4.2 Interpretation of the cave deposits

Excavations were made both in the Bone Chamber and in the vicinity of the main shaft (Fig. 5.5). In the Bone Chamber, the lowest deposit found was a sterile, greyish "gravelly clay" (Jackson, 1910c, 217). Unfortunately, this deposit is no longer exposed in the cave and it is difficult to obtain an indication of the conditions prevailing during its deposition. Both in the Bone Chamber and throughout the rest of the cave, the overlying deposits consist of ill-sorted materials similar to those found in other caves in the Morecambe Bay area (see 5.5.1). Jackson described these deposits as having a brownish (1910b, 86) or reddish (1910c, 218) clayey matrix within which could be found rounded pebbles of limestone, slate and other erratics (1909, 4; 1910b, 86), as well as angular blocks of limestone (1910b, 86), the latter probably the result of cave breakdown. With increasing

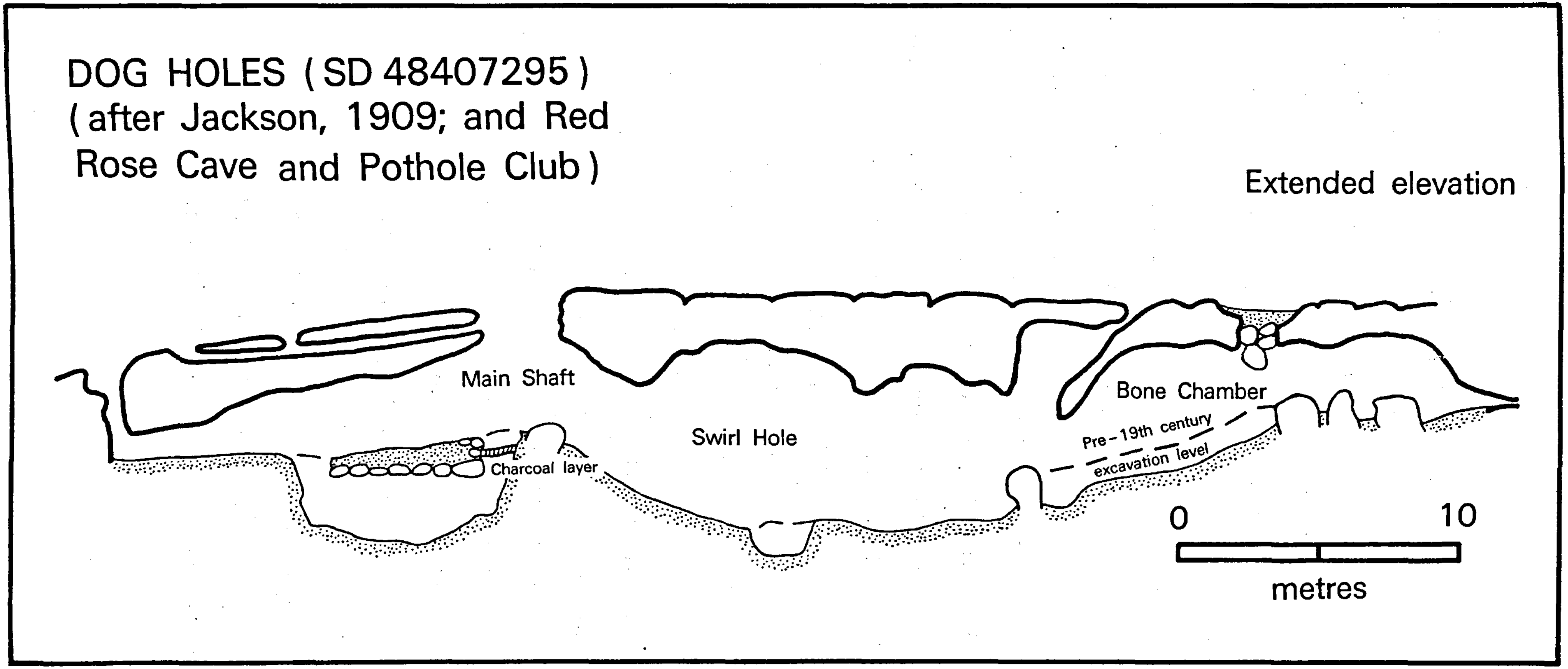


Fig. 5.5 Dog Holes

depth, however, the clayey matrix contained fewer coarse fragments (1909, 5). This deposit is almost certainly derived from surface glacial materials. The increased abundance of coarse material in the upper part of the deposit may be the result of the enlargement of the openings between the cave and the surface at some stage. There is clear evidence that such an enlargement occurred, possibly towards the end of the Iron Age, with the collapse of the cave roof in the main shaft (see below).

At the lower end of the Bone Chamber, these deposits reached a maximum thickness of approximately 1.5 m, and in the lowest part of these beds a considerable find of mollusca was made (Jackson, 1910c). With the exception of *Cecilioides acicula*, *Vallonia excentrica*, *Lymaea peregra* and *L. truncatula*, the ecology of all the species found is indicative of a woodland habitat (Evans, 1972). Of the anomalous species, *L. peregra* and *L. truncatula* are freshwater catholic and slum species respectively (Evans, 1972, 199, 349), whilst *V. excentrica* is virtually unknown in wooded or shaded habitats. However, only one example of each of these species has been discovered and it is possible that they have been derived from surface sediments or were carried into the cave by other animals. On the other hand, *C. acicula*, which is common in cultivated areas (Evans, 1972, 168), is plentiful throughout the deposit (Jackson, 1910c, 234). This species may be regarded as postdating the deposit as a whole, for *C. acicula* is a burrowing species typically found at depth in soils (Evans, 1972, 168).

Both because of the sampling method used by Jackson and because of the possible derivation of these deposits from surface sediments, it is impossible to interpret the mollusca as constituting one or more assemblages which can provide evidence of either chronology or environment. Nevertheless, almost all the species can be referred to both the Ipswichian and the Flandrian (Evans, 1972). Within the Flandrian, most of the species found in Dog Holes were present in this country by the late Boreal or early Atlantic. Three species, however, do not fit this picture. C. acicula and Helix aspersa are recent introductions to Britain. H. aspersa was almost certainly only introduced to Britain during the first century A.D. (Evans, 1972, 175-176). Discus ruderatus, on the other hand, although present in earlier interglacials, appears to have become extinct at the start of the Atlantic period, having arrived in this country around 9000 B.P. (Evans, 1972, 183-185). As might be expected, D. ruderatus is present only in the lowest part of the deposit in Dog Holes (Jackson, 1910c, 218), whilst H. aspersa only occurs near the top (Jackson, 1910c, 233). Nevertheless, this distribution may be fortuitous in view of the highly mixed nature of the vertebrate remains found within the same deposit (see below).

Also discovered in the Bone Chamber were remains of vertebrate fauna now extinct in this country (Jackson, 1910d; 1910e; 1912). Lemmus lemmus is now found mostly in mountain birch woods and in the zone immediately above the tree line (Kurtén, 1968, 220), whilst Dicrostonyx torquatus now inhabits the treeless tundra (Kurtén, 1968, 221). Both species occur in typical Devensian cold faunal assemblages in this country

(Stuart, 1974). Microtus ratticeps is found in England in interglacial deposits, although there is also evidence for its occurrence in the Devensian (Kurtēn, 1968, 218). However, as with L. lemmus and D. torquatus, there is no evidence for its existence in this country after the end of the last glacial (Stuart, 1974). Unfortunately, the present distribution of M. ratticeps is relict, so it is difficult to infer a specific habitat from its presence. All the remaining vertebrate fauna in the Bone Chamber either exist in this country at the present day or are associated with cultural remains in the cave.

The extinct vertebrate fauna would have existed most recently in the area during the Late-glacial. That the faunal remains can be referred to approximately this period is supported by the association of extinct fauna with the remains of recent vertebrates, ".... the Lemming occurring with the Field Vole; the Northern Vole with the Mole and Bat" (Jackson, 1910e, 331). One explanation for the intermixed nature of the deposit is burrowing, for which Jackson (1910e, 331) stated that there was no evidence, certainly in the lower part of the deposit. Alternatively, and more probably, the fossils may have been derived, along with the clastic deposits, from surface sediments of the area.

In the upper part of this deposit in the Bone Chamber were found a number of cultural remains. At a depth of approximately 60 cm, there occurred a bed of charcoal and burnt bones. Associated with the charcoal was an antler weaving comb of probable Neolithic age (Jackson, 1909, 13-14; 1913a, 104-105). Nearby were found bones, some of which were split, teeth, and other objects (Jackson, 1909, 6); and at a similar depth there occurred a chert flake comparable to those found in Neolithic sites in east Lancashire (Jackson, 1909, 12). Pottery fragments found within and above this layer were assigned to the first century A.D. (Jackson, 1909, 9). 10 cm above the charcoal bed, a bronze object of probable Celtic age was found, and, 37 cm above the charcoal, a knife blade, of undoubted Roman origin, was discovered (Jackson, 1909, 12-13). Within the same chamber were also found apparent burials of either Bronze or, more likely, Neolithic age (Jackson, 1909, 16, 19; 1913a, 123; 1913b, 57; 1915, 73). Numerous animal remains were discovered in association with these finds but they give little indication of their date of deposition, since they consist of forms usually found together in rubbish heaps of Neolithic to Romano-British age (Jackson, 1909, 19; 1913a, 123).

The remaining excavations in Dog Holes were made in the area of the main shaft (Fig. 5.5). The shaft, which opens onto the overlying limestone pavement, was formed as a result of block collapse, allowing debris from the surface to accumulate in a mound beneath the entrance (Jackson, 1909, 3). In the top part of the mound were found a number of large fallen limestone blocks, in the spaces beneath which were discovered the bones of sheep, fallow deer, Celtic shorthorn (Bos longifrons), and several human bones. From their fractured condition, the ox and human bones were assumed to have been deposited prior to the episode of roof collapse (Jackson, 1910a, 61). Similar remains were found, along with a hammer stone, between the limestone blocks and the south wall of the passage (Jackson, 1910a, 61). Immediately in front of the blocks, within the debris mound, were discovered a variety of artifacts, including a bronze scale pan and beam, a blue and red enamelled bronze pendant, several small iron objects, bone awls and pins, small fragments of pottery, and flakes of black chert and flint (Jackson, 1910a, 62, 72; 1913a, 100). The scale was assigned to the fourth or fifth century A.D., whilst the pendant and the iron objects were regarded as of Roman origin (Jackson 1910a, 71-72).

Several patches of charcoal were found at a depth of approximately 60 cm beneath one of the collapsed limestone blocks. In association with these were found cockle shells, fragments of iron, iron nails, a strip of lead, a bone awl and bone pins, and a fragment of ornamented bronze (Jackson, 1913a, 100). Elsewhere within the debris mound, at approximately the same depth, was found a distinct charcoal layer 3-5 cm thick, containing split bones, animal remains and burnt stones (Jackson, 1909, 8), as well as several types of Late Celtic pottery (Jackson, 1913a, 100). The animal remains are of similar species to those found in association with the charcoal layer in the Bone Chamber (Jackson, 1910a, 61).

Below the charcoal layer, the character of the deposit changes. The clayey material characteristic of the upper part of the deposit is replaced by "limestone gravel and boulders" (Jackson, 1909, 8), which includes occasional remains of horse, ox, dog, sheep and fragments of human skulls (Jackson, 1909, 8; 1913a, 100). This bed may be equivalent to the greyish "gravelly clay" found in the same stratigraphic position in the Bone Chamber.

5.4.3 Conclusions

The oldest deposits in the cave which can be regarded as in situ, and to which a date can be assigned, are the Neolithic artifacts in the Bone Chamber. The older faunal remains could easily have been derived from surface deposits and cannot be used as an indicator of age stratigraphy. Above the Neolithic remains occurs a rough sequence of Iron Age and then Romano-British deposits, with a similar sequence in the main shaft including post-Roman artifacts of the fourth or fifth century A.D.

It is difficult to correlate the deposits found in Dog Holes with those found in Kirkhead Cave and elsewhere. The main reason for this is the absence, in Dog Holes, of the stalagmite bed which forms such a prominent feature of the stratigraphy of other caves. It is possible that Jackson's excavations failed to reach the stalagmite, which was at a depth of 2 m in Kirkhead Cave, and this is borne out by the fact that the lowest beds found by Jackson, his "limestone gravel and boulders" and greyish "gravelly clay", contain the reamins of species common in the area today.

5.5 Other caves

5.5.1 The sequence of development

The morphology and sedimentary infill of every known cave in the Morecambe Bay karst has been studied. Almost without exception, the caves display evidence of the following sequence of conditions:

I phreatic cave development

II flow abandonment

III clastic infill

IV stalagmite deposition

Episode I must have consisted of at least two phases: an earlier phreatic phase (Ia), and a later vadose phase as the cave drained (Ib). The morphological effect of phase Ib is not always clear, although in many cases this may be because the sedimentary infill hides any evidence of vadose trenching. This sequence, often with a final episode of fill excavation (phase V), can be recognised in the following caves:

(i) Holker: Roudsea Wood Cave (north) (SD332826) (see 7.2)
Capeshead Cave (SD33337814) (see 7.4.2.1)

(ii) Hampsfell: Grand Arch (SD39037388)
Kirkhead Cave (SD39097562) (see 5.3)

(iii) Whitbarrow: Pool Bank Cave (SD43378773) (see 7.5.3)
Lyth Valley Cave (SD451897) (see 7.5.3)

(iv) Underbarrow: Helsfell Cave (SD500938)
Levens Cave (SD48438572)

(v) Silverdale: Arnside Cave (SD44737710)
Wall End Cave (SD45447572)
Cove Hole (SD45557561)
Fissure Cave (SD45557560) (see 7.3)
Silverdale Shore Cave (SD45627549)
Badger Hole (SD48217283)
Millhead Big Cave (SD49797149)

(vi) Other areas: Dunald Mill Hole (SD51586763) (see 7.5.4)
Owl Tree Hole (SD561777) (see below)

The sedimentary fill of episode III generally consists of ill-sorted deposits, containing frequent erratic materials (Fig. 5.2) and a coarse fraction whose particles vary in shape from angular to rounded. Samples of the clastic fill from the following caves were analysed (see 11.3). (For detailed analyses of sediments from Kirkhead Cave and Fissure Cave see 5.3.2 and 7.3 respectively):

(i) Capeshead Cave (sample 41)

(ii) Helsfell Cave (sample 3)

(iii) Wall End Cave (sample 40)

(iv) Silverdale Shore Cave (sample 50)

(v) Owl Tree Hole (sample 2)

In all cases, these samples fall within the glacial environmental envelopes of the QDa-Md diagram (Fig. 5.1), suggesting that the cave fill is derived from the surface glacial deposits of the area. This hypothesis is supported by petrographic analysis of the pebble fraction of the deposits (Fig. 5.2). In all cases, the deposits contain a significant proportion of material derived from outcrops to the north of the area and found in association only in glacially-transported deposits.

Analysis of samples of the surface glacial deposits of the area (samples 45, 46, 47, 48 and 49; Water Resources Board, 1970) shows that, as expected, they fall either within or close to the near-mountain scource glacial environmental envelope of the QDa-Md diagram (Fig. 5.1). The deposits in Capeshead Cave, Helsfell Cave, Silverdale Shore Cave and Owl Tree Hole are indistinguishable from these glacial materials on granulometric grounds, and it is suggested that these sediments were deposited in the caves by a variety of slumping, rolling and inwashing processes from glacial materials outside the caves. The possibility of direct deposition by ice can be ruled out on account of the faunal and cultural remains within the cave deposits (see below). These processes of deposition, previously invoked in explanation of beds in a similar stratigraphic position in Kirkhead Cave (see 5.3.2), did not markedly alter the character of the original deposit. In the case of Wall End Cave, however, the cave fill falls within the continental-ice environmental envelope of the QDa-Md diagram (Fig. 5.1). As with similar deposits in Kirkhead Cave (see 5.3.2), it is suggested that the deposit was modified by the truncation of the coarse fraction of a near-mountain source glacial deposit. This truncation could have been achieved by selective transport of the finer fraction of the deposit from outside the cave, possibly by slope wash, by wind or by biogenic transport.

In most of the caves of the area, these clastic beds contain abundant faunal and cultural remains. Although the fossils could have been emplaced by biogenic burrowing, the beds are often sealed by stalagmite deposits and there is generally no evidence of post-depositional disturbance. During the present study, faunal remains have been found in the beds underlying the stalagmite in Capeshead Cave and the Grand Arch. In Capeshead Cave, the majority of the fossils were either of frog or toad; there was also a slow worm scale and the vertebra of a small fish. The only mammalian species present was Clethrionomys glareolus (bank vole), represented by some very juvenile teeth. The assemblage, although rather limited, suggests a Post-glacial temperate environment and is probably quite recent in origin (Currant, pers. comm.).

In the Grand Arch, the clastic deposits underlying the stalagmite within the main phreatic inlet to the cave include remains of the land snails Oxychilus cellarius, Discus rotundatus and Retinella nitudula, and the rodents Microtus agrestis (field vole) and Sorex sp. (shrew). The snails are common in British cave deposits (Evans, 1972, 308-310), but are restricted to inactive caves and are characteristic neither of marine nor freshwater conditions. A similar, inactive, cave environment may be inferred from the rodent remains.

Capeshead Cave (Morris, 1865-66, 359), Helsfell Cave (MacPherson, 1892), Arnside Cave (Jackson, 1910b, 85) and Badger Hole (Jackson, 1910a, 64-65) were all excavated during the nineteenth and early twentieth centuries. In all four cases, the beds beneath the stalagmite were found to contain similar faunal assemblages, mainly consisting of wild and domestic species found in the area throughout most of the Post-glacial. Even the now-extinct Ursus arctos and Canis lupus, found in Helsfell and Arnside Caves, were common in England in historical times. In Capeshead Cave and Helsfell Cave, there was also evidence of human occupation and burial, although, in the absence of the finds themselves, no date can be confidently assigned to these remains.

There is no evidence of man in northern England prior to the Late-glacial, and it is considered by Evans (1975, 17-20) that this provides a true reflection of his distribution, certainly in the case of Lower Palaeolithic man. The evidence of human occupation in Capeshead Cave and Helsfell Cave would therefore tend to confirm the post-last glacial age of the clastic beds underlying the stalagmite.

In Badger Hole, the stalagmite is overlain by clastic deposits, whilst in Roudsea Wood Cave (south) (SD333825), Merlewood Cave (SD41157892), Haverbrack Bank Cave (SD48298025) and Fairy Hole (SD49697296), excavations have not yet exposed a stalagmite layer. In those cases where cultural artifacts and faunal remains have been found within these deposits, they tend to confirm the expected post-Atlantic age of the beds. Thus, in Merlewood Cave, Cowper (1893) found fragments of red and black pottery, of possible Romano-British origin, and seven Northumbrian coins dating from the ninth century A.D. In Haverbrack Bank Cave, Benson and Bland (1963) interpreted the deposits as representing a sequence from Romano-British to Mediaeval times, although at least 3 m of cave fill still remains to be excavated. Jackson (1910a) recorded Romano-British remains in the entrace fill of Fairy Hole, along with many shells of the land snail _Helix aspersa_, introduced to Britain during the first century A.D. Finally, in Badger Hole, where the upper clastic beds can be shown to overlie the stalagmite, Jackson (1910a, 64-65) discovered a faunal assemblage containing domestic and wild forms found in the area today, along with a knife handle and bone awl, for which no date was suggested.

Analysis of samples of the upper clastic beds from Roudsea Wood Cave (south) (sample 1) and Fairy Hole (sample 13) shows the deposits to be very similar to the glacial material from which they were almost certainly derived (Fig- 5.1). In both cases, the pebble fraction of the samples contains a significant proportion of material derived from outcrops to the north of the area (Fig. 5.2). As in the case of the beds underlying the stalagmite, these upper clastic beds have probably been transported into the caves by a variety of processes.

The final episode in the caves, that of excavation (phase V), can be explained most simply in terms of the reinitiation of vadose activity. It is equally possible, however, that in some caves the excavation was the result of early archaeological work. This may have been the case in Capeshead Cave, Helsfell Cave and Fairy Hole.

In some of the caves, it is possible to infer a more detailed sequence than that given above. In Owl Tree Hole, a joint-aligned, phreatic rift-passage is infilled by breakdown and by glacially-derived fill (sample 2). However, within the fill can be found fragments of stalagmite, suggesting the reworking of a previous phase of stalagmite deposition. Later phases of entrenchment and stalagmite deposition can be identified from the stratigraphy of the deposits (Fig. 5.6). Thus, a more complex sequence of conditions may be proposed:

I	development of the rift passage under phreatic conditions
IIa	flow abandonment
IIb	block breakdown } sequence uncertain
IIc	stalagmite deposition } sequence uncertain
IIIa	clastic infill, reworking stalagmite of IIc.
?IIIb	trenching of infill by stream
IV	stalagmite deposition on cave floor and walls
V	removal of stalagmite floor and some clastic fill
IV	stalagmite deposition.

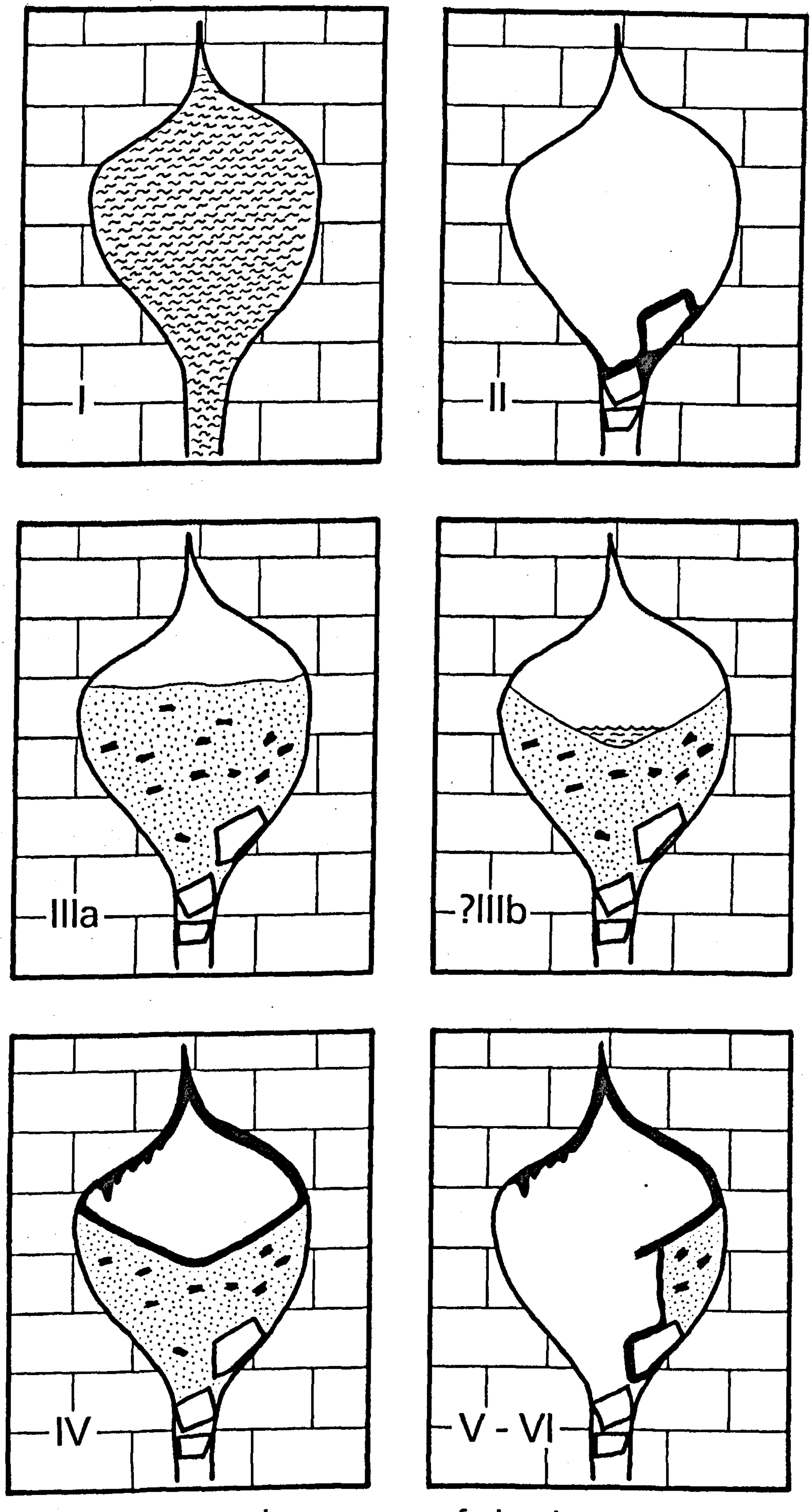


Fig. 5.6 Proposed sequence of development of Owl Tree Hole

5.5.2 The age of the caves

The majority of the abandoned caves of the Morecambe Bay area accumulated considerable depths of clastic infill during the Post-glacial. In order to establish a more accurate date of deposition, it may be possible to see the stalagmite bed, which is frequently found associated with the clastic deposits, as a chronostratigraphic indicator. Owing to its near-entrance situation, the deposition of the stalagmite is likely to have been controlled by external environmental conditions, rather than by specific conditions inside each cave. It is therefore probable that stalagmite deposition took place contemporaneously throughout the area. In Kirkhead Cave, the stalagmite is almost certainly of Atlantic age (see 5.3.3). If this date is a region-wide one, then the depositional sequence in most of the caves in the area may be interpreted as representing the occurrence of deposition throughout most of the Post-glacial. Since in no case do either the clastic sediments or the faunal remains indicate the existence of active fluvial conditions in the caves, it would appear that episode II of cave development was of, at the latest, last interglacial age, assuming negligible groundwater flow to have taken place during glacial periods. Consequently, the caves themselves must be of no younger than last interglacial age.

6. APPROACHES TO THE STUDY OF KARST PALAEOHYDROLOGY

6.1 Introduction

Although numerous studies have been made of the palaeohydrology of non-karstic aquifers (see, for example, Issar, Bein and Michaeli, 1972), such studies have generally adopted the approach of groundwater dating by radiometric means. By contrast, detailed studies of the hydrological development of karst aquifers are commonplace, mainly because karst aquifers retain physical evidence of former flow routes in the form of flow conduits. The morphology of these features has been interpreted both in terms of overall drainage system development and in terms of flow conditions in specific passages (see, for example, Ford, 1964; Waltham, 1974). With few exceptions, however, other approaches have been neglected. Nevertheless, the potential of karst aquifers for palaeohydrological study has been demonstrated by a number of workers. Ford, for example, has used potholes in the beds of cave streams to indicate the sequence of fluvial activity (1965a), and has applied the Manning equation to estimate flow velocity along phreatic bore passages (1965b). Brook (1971) has proposed that, where there is a reasonable concordance of the apex altitudes of phreatic roof domes along a passage, this may be taken to indicate the former height of hydrostatic head within the passage. Numerous workers have considered the relationship of scallops and flutes to former flow conditions in the passage (see 6.5), although the relationships have rarely been applied to field situations. Finally, a handful of studies have attempted, albeit only in a preliminary fashion, to reconstruct former flow conditions from the evidence of hydraulically-transported cave sediments (Burkhardt, 1958; Renault, 1968;

White and White, 1968).

In the present study, it is only possible to study the development of drainage on a regional scale in a very general manner, for, in the case of most of the caves, only fragments of the former passages remain. Nevertheless, it is possible to reconstruct the former hydraulic environment within specific passages by the study of hydraulically-transported cave sediments (see 6.2 and 6.3), and by the study of cave meanders (see 6.4) and cave scallops (see 6.5). On the basis of this work, it may also be possible to give some indication of the palaeocatchment area contributing to the passages under consideration (see 6.6).

6.2 The processes of fluvial erosion, transport and deposition, and their application to the interpretation of the flow conditions under which fluvial sediments are deposited

6.2.1 Introduction

The sedimentary sequence within any fluvial system is a product of the history of the fluvial processes operating within the stream. An understanding of these processes and their influence on stream load should therefore enable the reconstruction of the palaeoflow conditions in the stream. This depends, however, on a number of conditions being fulfilled:

(i) That sediment of a wide range of grain-sizes is available to be transported and deposited by all flows occurring within the stream. Flows may occur which are competent to transport and deposit material, but sediment of the necessary size may not be available. Hence, stream deposits may be partly a reflection of sediment availability.

(ii) That erosional or non-depositional unconformities do not occur. The sedimentary sequence may be incomplete due to erosion of beds by subsequent highly competent flows or to non-deposition by certain flows.

(iii) That a complete understanding of the processes of fluvial erosion, transport and deposition exists.

6.2.2 Fluvial erosion and transport

6.2.2.1 Bedload

As stream competence increases, the initial movement of particles larger than ~0.2 mm will be as bedload (Sundborg, 1967, 338). Smaller particles are generally transported immediately as suspended load. Thus, in the case of relatively coarse particles, it is necessary to consider only the influence of bedload transport. A large number of equations have been derived to model bedload transport (e.g. Du Boys, Chang, Meyer-Peter, Shields), none of which are entirely satisfactory under natural conditions. There are many reasons for this. Existing transport formulae are based largely on the results of flume studies employing uniformly sized and shaped sediment. Moreover, most equations take into account the entire cross-section of the stream and, consequently, employ one-dimensional parameters (e.g. energy slope and mean stream velocity) and assume steady flow. Armoured beds and non-uniform flow, especially spiral motion in curved channels, present complexities which existing formulae are unable to handle successfully. Nevertheless, it has been found that, in the case of relatively coarse sediments, the critical velocity of erosion is a function of grain-size. More precisely, Sundborg (1956, 167-168) stated that resistance to erosion is a function of the immersed weight of a particle (itself a function of grain size, and fluid and particle density) and its angle of repose.

Much of the work on bedload transport has been summarised by Hjulström (1935, 292-320), Menard (1950) and Sundborg (1956, 169-191) (Fig. 6.1), Sundborg, in particular, taking into consideration the influence of particle shape and density, channel form and roughness, and the effect of suspended sediment on bedload transport. All these workers use stream velocity as an indicator of erosional competence. It should be pointed out, however, that velocity is of indirect influence on sediment motion, and that bed velocity, drag force, or the vertical lift induced by the velocity gradient exercise more direct control. Nevertheless, the advantages of using stream velocity as an indicator are twofold. Firstly, velocity is of more relevance in palaeohydrological studies, and secondly, the important syntheses of work on the processes of fluvial erosion and deposition all adopt stream velocity as an independent variable.

Fig. 6.1 summarises the results of Hjulström (1935), Menard (1950) and Sundborg (1956). Hjulström's curves are for uniform, loosely packed particles of density 2.6-2.7 g cm^{-3}. Flow depths are assumed to be >1 m and stream flow is given in terms of mean velocity (1935, 293-296). Menard's curves are for mean velocities and flow depths <1 m (1950, 149-150). Sundborg's curves provide velocities at varying heights above the stream bed for particles of density 2.65 g cm^{-3}.

Although new bedload transport equations appear almost annually, they all have the same general form (see Raudkivi, 1967, 103) and it is considered inappropriate to attempt to

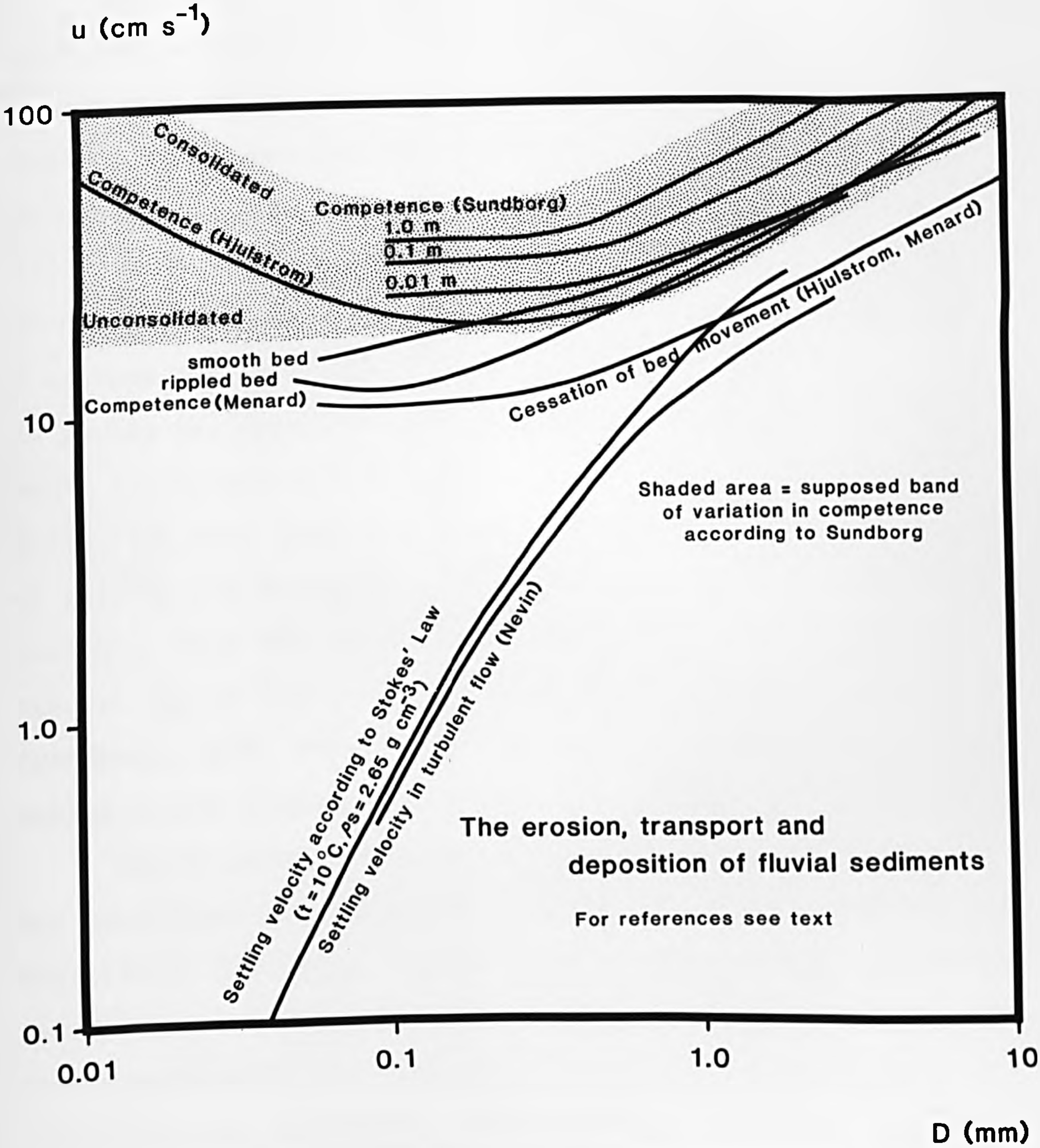


Fig. 6.1 The erosion, transport and deposition of fluvial sediments as a function of flow velocity and grain-size

summarise their findings here.

6.2.2.2 Suspended load

As stream competence increases, coarse particles may begin to be transported by saltation and, eventually, may become part of the suspended load. However, particles finer than ~0.2 mm, once eroded, generally go directly into suspension. Unfortunately, the relationship between grain-size and erodibility which exists for coarse particles only holds when the forces resisting separation of the particles are frictional, i.e. when the particles exhibit no cohesive properties. This is partly the result of gravitational and buoyancy forces being negligible for fine particles. Apart from the fluid force, the most important forces acting on a particle are those of gravity and buoyancy. A particle's equilibrium becomes unstable when the resultant of these three forces passes through one of the points at which the particle is supported (Sundborg, 1956, 168). More important, however, are the cohesive forces binding fine particles together (see Grim, 1962).

The diverse literature on the erosion of cohesive beds has been summarised by Enger, Smerdon and Masch (1968) and by Graf (1971, 322-347). While it is recognised that the scour resistance of a cohesive sediment depends in some fashion on the properties of the sediment, the complete significance of such parameters as density, moisture content, percentage clay content and shear strength is not yet fully known. However, as the physical properties causing cohesion in fine materials are largely dependent on the dominant type of clay mineral in the material, a close correlation between grain-size and erodibility should not be expected. Thus, any empirical relationship between grain-size and erodibility is only applicable within relatively small regions where clays have a similar geological origin. Nevertheless, Graf's review showed that the critical shear stress at which erosion occurs may be a direct function of the plasticity index and the arrangement of particles, and an inverse function of moisture content and void ratio. Both particle arrangement and plasticity index are strongly related to the particle-size distribution of cohesive materials (Dos Santos, 1953), while moisture content and void ratio will themselves be a function of particle-size distribution. Thus, Hjulström's (1935, 293-300) summary of early work on the critical erosion velocities of cohesive sediments as a plot of critical velocity against grain-size (Fig. 6.1), and Sundborg's (1956, 178) assumption that the magnitude of cohesive forces is likely to vary inversely with particle-size, may not be as unrealistic as they might at first appear. Thus, Sundborg assumed:

$$\tau_{crit} = K_1 \, 1/D \qquad 6.1$$

From von Kármán's universal velocity profile law for turbulent flow across a smooth boundary, Sundborg (1956, 167) related the boundary shear stress to flow velocity in the following fashion:

$$\tau = K_2 \, u_{max}^{1.75} \qquad 6.2$$

combining equations 6.1 and 6.2, Sundborg derived a relationship between the boundary shear stress at which entrainment occurs and flow velocity:

$$K_2 \, u_{max}^{1.75} = K_1 \, 1/D$$

$$\log u_{max} = -4/7 \, (\log D) - K_3 \qquad 6.3$$

Equation 6.3 provides a relationship between grain-size of cohesive particles and critical erosion velocity which plots as a line of slope -4/7 on log axes (Fig. 6.1). Despite the assumptions involved in the initial premise of a direct relationship between critical boundary shear stress and grain-size, the derived curve has a similar slope to that of the empirical relationships found by Hjulström (1935, 293-300) for fluvial erosion and by Bagnold (1941) for aeolian erosion. Thus, Sundborg (1956, 179) used his theoretically established relationship to provide an expression of the variation of critical velocity with grain-size in cohesive, consolidated beds. For cohesive, unconsolidated beds he arbitrarily drew a line of zero slope in view of the total absence of data, although he admitted (1956, 180) that, under these conditions, critical velocity might decrease with particle-size (Fig. 6.1).

6.2.3 Fluvial deposition

6.2.3.1 Bedload

Hjulström (1935, 320-323) published a plot of mean flow velocities at which bed movement ceased ("lowest transport velocities") against grain-size. This was based mainly on the experimental work of Schaffernak (1922) who studied particles no smaller than 5 mm. Hjulström concluded that depositional velocities were approximately $^2/_3$ those at which erosion took place (Fig. 6.1). Flume studies on the cessation of bed movement by Menard (1950, 151-152) extended Hjulström's curve to particles of 59 μm, finding a similar relationship (Fig. 6.1).

6.2.3.2 Suspended load

The settling velocities of particles in suspension have been studied by Rubey (1933). For particles smaller than 0.14 mm, settling velocity is chiefly a function of fluid viscosity and may be calculated by the use of Stokes' Law. For larger particles, however, additional factors must be taken into account. During the settling of larger particles, a drag force develops which tends to decrease the settling velocity. The drag force originates because larger particles falling at greater velocities develop a low pressure zone behind them as they settle. Turbulence results, taking energy from the system and decreasing the settling velocity. Turbulence resulting from the nonspherical shape of particles has similar consequences (Zeigler and Gill, 1959, cited in Galehouse, 1971, 78).

On the other hand, two factors tend to increase the settling velocity of coarser particles (Cook, 1969, 781-782). Firstly, groups of sand grains tend to fall as a unit until the grains become dispersed, which requires settling through a distance of several centimetres. Secondly, smaller grains tend to be entrained by larger grains and accelerated. Thus, settling velocity is not simply a function of particle-size, but also a function of particle interaction, as characterised by the sorting and skewness of the total sample.

Rubey (1933) hypothesised that, for particles larger than 1.55 mm, settling velocity is chiefly a function of the impact of the fluid on the falling particle, and so may be calculated from the Impact Law. Assuming a gradual change from impact to viscosity effects for particles in the range 1.55-0.14 mm, Rubey derived a general expression for the settling velocity of a wide range of particle-sizes. Comparison with known settling velocities of particles ranging from 12.85 mm to 1.52 µm (see also Zeigler and Gill, 1959) showed the equation to accord closely with the published data, certainly for particles of density 2.65 g cm^{-3}. Thus, particles settle out only if their settling velocity is greater than the velocity of upward turbulence in the flow. Turbulence, however, depends not only on the horizontal stream velocity, but also on bed roughness, eddies within the flow, etc. Thus, settling velocity cannot be regarded as a true indicator of the actual velocity of stream flow during deposition of suspended material.

Despite this, experimental work by Nevin (1946, 657-661) on the horizontal flow velocities at which particles could be kept in suspension demonstrated that, for particles of between $\sim$0.4 mm and 0.09 mm, depositional velocities agreed closely with those expected under Stokes' Law (Fig. 6.1). This suggests that Stokes' Law may be used, at least as a first approximation, to establish the horizontal flow velocities at which particles may be deposited from suspension.

Under natural conditions, however, concentrations of sediment in suspension modify the apparent viscosity of the water-sediment mix with a resultant decrease in the fall velocity of particles (Simons, Richardson and Hauschild, 1963). In the fluvial systems investigated in the present study, the suspended sediment concentration which existed is an unknown quantity. It is reasonable to postulate *a priori* at least a moderate amount of suspended load on the grounds of the glacial origin of the fluvial deposits and the presence of rock flour in the parent sediment (see 7.3.2). There are other cogent reasons for this premise. Firstly, fine-member beds, presumed to have been deposited from the suspended load, occur frequently. Secondly, even the coarse-member beds contain a large proportion of suspended-grade material (Fig. 7.13) from which a high suspended sediment concentration can be inferred.

6.2.4 The effect of non-uniform sediments on fluvial erosion, transport and deposition

Most studies of sediment erosion, transport and deposition in fluvial systems have been based on observations of the behaviour of uniform sediments. But, in natural streams, sediments are uniform in neither size, shape nor density. The effect of particle density on fluvial sediment processes has been studied by Ljunggren and Sundborg (1968) and by White and Williams (1967), whilst the implications of particle shape for settling velocity have been considered by Wadell (1934). However, little work has been done on the effects of sediment-size mixtures.

6.2.4.1 Erosion and transport of sediment-size mixtures

Sundborg (1956, 185-188) summarised earlier work on the erosion of grain-size mixtures. He found that the effects of particle-size distribution on the process of entrainment could be summarised as follows:

(i) Coarse mixed material (>6-8 mm): flow is completely rough and turbulence extends down between superficial particles. Thus, the smaller particles are most easily entrained.

(ii) Mixed material (0.3 to 6-8 mm): the main flow is rough or transitional, but turbulence does not extend fully into the interstices between particles, thus particles of 0.3 to 0.8-1.0 mm are shielded and are within the laminar sublayer. Hence, particles of approximately 2 to 4 mm are most easily entrained. Work by Ashcroft (1980) on mixtures of 0.2 to 0.7 mm demonstrated that the addition of particles of 0.6 to 1.2 mm caused an increase in critical shear velocity of 40% on average. However, the addition instead of particles of 1.2 to 2.4 mm caused no sheltering of the finer particles and hence no increase in critical shear velocity.

(iii) Fine mixed material (<0.3 mm): flow is smooth and entrainment depends largely on the content of cohesive material in the mixture. The coarsest grains tend to be moved most easily. After entrainment, the mixture is transported mainly in suspension with the coarsest particles moved as bedload.

Thus, although particles of 0.2-0.5 mm are most easily eroded when sediments are of uniform size (Fig. 6.1), particles of $\sim$1.6 mm tend to be least stable in grain-size mixtures.

In view of the difficulties involved in applying equations derived from observation of uniform sediments to natural grain-size mixtures, numerous proposals have been put forward to allow natural sediments to be characterised by a single size-parameter in order to enable transport equations to be used. Hjulström (1935, 300) suggested the use of the modal grain-size of particles in movement to allow the application of his curves to grain-size mixtures. Bagnold (1966, 18) argued that, in order to take into account the effect of sediment-size on bed friction, the physical effects of the arithmetic mean grain-size are more relevant than those of the D_{50} or modal grain-size. Sundborg (1956, 183), on the other hand, regarded his curves as applicable only to uniform or well-sorted materials.

Unfortunately, as demonstrated by Sundborg's work above, any representative grain-size is likely to be so only for rough, transitional or smooth flow. In attempting to overcome this difficulty, Jopling (1966, 8-9) firstly defined the critical erosion velocity as identifiable by the general movement of the sediment mix (providing the mixture is not too poorly sorted). Reviewing the literature, he then found that for moderately well-sorted sands, the critical erosion velocity was approximately that predicted by Sundborg's method if the sediment was characterised by its D_{50} value. However, for less well-sorted mixtures, the smaller particles had a binding effect, and so higher critical velocities were required than those predicted on the basis of D_{50}. As a first approximation, Jopling concluded that the Sundborg critical velocity would have to be increased "by perhaps 20 percent(?) for poorly sorted mixes".

6.2.4.2 Deposition of sediment-size mixtures

Little work has been published on the effects of grain-size mixtures on particle deposition, and it is not easy to characterise the size of a mixed sediment in order to enable an approximate determination of its velocity of deposition. Hjulström (1935, 321) reported the findings of Kramer, who observed that grain-size mixtures have a critical influence on the velocity at which bedload ceases to move. However, it is difficult to discover a systematic pattern of influence from Kramer's results. Even less is known about the effect of sediment mixtures on the depositional velocities of suspended sediments.

6.2.5 Application to the present study

In order to establish values of critical erosion velocity in the present study, it was decided to utilise the relationships between critical velocity and grain-size derived by Sundborg. Sundborg's curves take into account a wide range of flow depths and particle densities. Moreover, Jopling (1966, 8-9) has demonstrated the applicability of Sundborg's results to poorly-sorted sediments. Jopling's recommendation that sediments be characterised by their D_{50} value is followed and, in view of the poorly-sorted nature of many of the sediments investigated (Fig. 7.13), critical velocities calculated by the Sundborg method are increased by 20%. Following Saunderson and Jopling (1980, 177-178), the surface velocity given by Sundborg's curve is converted to an approximate mean flow velocity by multiplying by 0.8.

On the grounds of simplicity and in the absence of a better alternative, it was decided to use the D_{50} value as the characteristic grain-size in the estimation of depositional velocities. For bedload deposition, the combined curves of Hjulström (1935) and Menard (1950) were adopted (Fig. 6.1). For suspended load, on the basis of Nevin's (1946) work, depositional velocity was regarded as equivalent to the settling velocity given by Stokes' Law (Fig. 6.1). The problem of the influence of suspended sediment on fall velocities was ignored in view of the difficulty of estimating its influence and because of the likely errors already involved in calculating depositional velocity.

6.3 The derivation of palaeohydraulic information from grain-size studies of fluvial sediments

It was first suggested by Visher (1969) that cumulative grain-size distribution curves may exhibit features which reflect specific mechanisms of transport. By plotting these curves on log-probability axes, distinctive straight-line segments can be found. Visher interpreted these segments as corresponding to truncated log-normal distributions of grains moved by different transport mechanisms. The grain-sizes at break points between segments and the percentage of the total distribution within each segment were deemed to be indicative of particular depositional environments.

Visher distinguished subpopulations produced by three transport mechanisms: surface creep, saltation and suspension. These appear analogous to Einstein's (1950) bedload transport, suspended load transport and wash load transport respectively. For fluvial sediments, Visher studied several hundred modern and ancient deposits. He concluded that a distinctive break point occurred between 2.75 ϕ and 3.5 ϕ (0.15 - 0.09 mm) which was interpreted as corresponding to the junction of saltation and suspension populations. He maintained that, if present, the surface-creep population would be coarser than 1 ϕ (0.50 mm).

Visher's work is open to a number of criticisms. He assumed each sediment subpopulation to be composed of a truncated log-normal distribution. Other workers, however, maintain that many cumulative curves are composed of overlapping log-normal distributions (Tanner, 1958; 1959; 1964; Fuller, 1961; Walger, 1962; 1965; Spencer, 1963). Middleton (1976, 406) showed that occasionally this may give rise to significant differences in the interpretation of the location of the break point. However, the interpretation of overlapping distributions demands high precision grain-size data. Owing to this, and as a result of other potential sources of error in subsequent calculations, the assumption of truncated distributions is followed in the present work.

Implicit in Middleton's (1976) study is a criticism of Visher's (1969) dismissal of bedload transport as a contributor to fluvial grain-size distributions. Middleton (1976, 407-409) interpreted fluvial sediments as being composed of a combination of bedload and intermittent suspension load, the latter including material transported in true suspension and in true saltation. This interpretation appears more reasonable since, assuming surface creep to be the initial movement that occurs as bed shear-velocity increases, any fluvial deposit must consist of a greater or lesser proportion of bedload. On the other hand, shear velocity may never reach the stage where a wash load can exist. Moreover, once a wash load exists, it is held almost continuously in suspension (Middleton, 1976, 407) and is not deposited along with the remainder of the sediment load.

Finally, Jackson (1978, 25-26) demonstrated that Visher's simplistic assertion of characteristic break points for specific environments does not hold, and that grain-size curves from contrasting environments may resemble the ideal fluvial plot, whilst fluvial curves may have break points which do not correspond to those predicted by the ideal model.

Nevertheless, Middleton (1976, 406-407, 423) showed that the basic principle of Visher's work applies. He confirmed that breaks between straight-line segments of grain-size distributions are related to hydraulic sorting rather than to source materials or breakage. He also argued that the characteristic two-segment form of fluvial grain-size plots reflects the bedload and saltated load components of transport (Middleton, 1976, 408). Unlike Visher, however, he recognised that the position of the break point might vary, dependent on hydraulic conditions. By determining the boundary between the bedload and the intermittent suspension load, he felt that it should be possible to estimate the dominant shear velocity flow.

The coarsest particles taken into suspension are those whose settling velocity (ω) is approximately equal to the root mean square of the upward vertical velocity fluctuations near the bed (u') (Middleton, 1976, 409).

$$\omega \sim \sqrt{\overline{(u')^2}} \qquad 6.4$$

It is not possible to write an exact form of equation 6.4 because the relationship is a statistical one and because it is not known exactly what the relationship is between the measured root mean square (which includes downward as well as upward components) and the root mean square of upward components only.

According to Bagnold (1973), $\sqrt{\overline{(u')^2}}$ reaches values of approximately 1.2 times that of the shear velocity (u_*). Middleton (1976, 410) cites further evidence in support of this. Thus,

$$\frac{\sqrt{\overline{(u')^2}}}{u_*} \leqslant 1.2 \qquad 6.5$$

It follows from equations 6.4 and 6.5 that the coarsest particles in suspension will approximately obey

$$\frac{\omega}{u*} = 1 \qquad 6.6$$

That this equation has the correct form is supported by experimental evidence (Middleton, 1976, 410).

Thus, assuming that the grain-size at the break point is equal to the coarsest size in suspension, the settling velocity of particles of this size can be found (assuming spherical particles of density 2.65 g cm^{-3} and distilled water at $10^{o}C$) from the tables in Gibbs, Mathews and Link (1971) and hence u* calculated. For sand-size materials, values of u* greater than approximately 20 cm s^{-1} rapidly remove all sand, leaving only coarser materials, whilst a minimum value of approximately 1.4 cm s^{-1} is required to move sand at all (Middleton, 1976, 424).

Once u* has been calculated, other palaeohydraulic parameters can be computed:

(i) The Chezy coefficient (C)

$$C = \sqrt{g}\,\frac{\bar{u}}{u*}$$

(ii) The Darcy-Weisbach resistance coefficient (f)

$$f = 8/\left|\frac{\bar{u}}{u*}\right|^2$$

(iii) Discharge per unit channel width (Q unit)

$$Q\,unit = d\,\bar{u}$$

(iv) Slope of the energy grade line (S)

$$S = (u*)^2/\,g\,d$$

(v) Reynolds number (Re)

$$Re = (\rho_f\,\bar{u}\,d)\,/\,\mu$$

(vi) Froude number (F)

$$F = \bar{u}/\sqrt{g\ d}$$

(vii) Mean boundary shear stress (τ)

$$\tau = (u*)^2 \ \rho_f$$

(viii) Power of stream flow per unit area of bed (P) (Colby, 1964, 24)

$$P = \bar{u}\ \tau$$

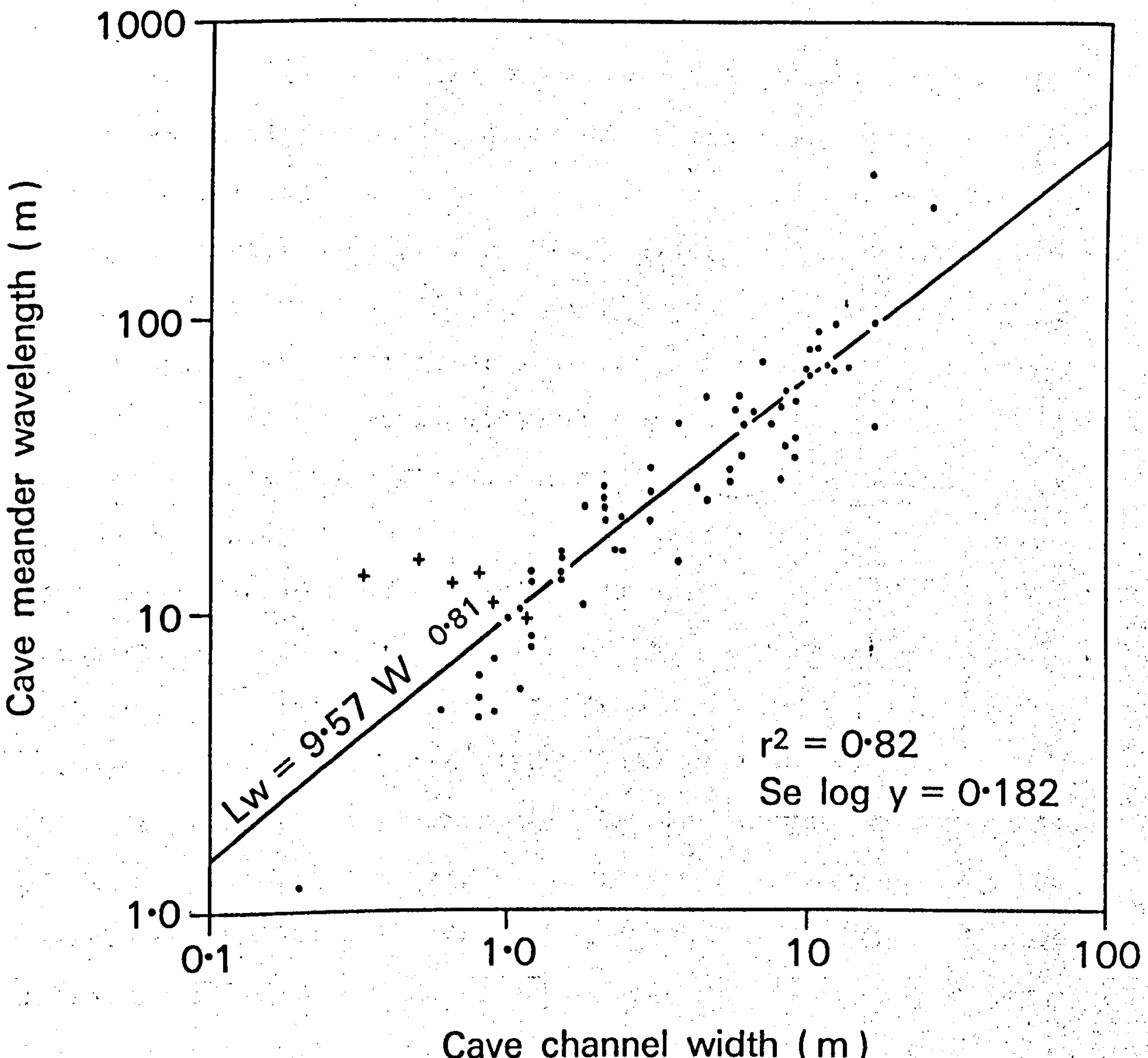
6.4 Cave meanders[1] as palaeohydrological indicators

6.4.1 Cave meander morphology

Several studies have been made of cave meander form (Deike, 1967; Wheeler, 1967; Ongley, 1968; Deike and White, 1969; Hanna and High, 1970; High, 1970b; Baker, 1973; Kuniansky, 1974; Smart, 1977). High (1970b) has pointed out that measurements made by Deike and White (1969) in three Irish caves are suspect as a result of the use of inaccurate cave surveys. High surveyed a short length of one of these caves and this was subsequently resurveyed fully by Smart (1977), whose data will be used in analyses in the present study. The data from the other Irish caves studied by Deike and White (1969) have been discarded.

Combining the data from these studies, a relationship of the form $Lw = 9.57\ W^{0.81}$ was found between meander wavelength and channel width ($r^2 = 0.82$ and Se log y = 0.182) (Fig. 6.2). Although a variety of methods of measuring these parameters

[1]The use of the term meander implies no strict definition of sinuosity (see Dury, 1969, 421).



• Data from Ongley (1968), Deike and White (1969) and Baker (1973)

+ Data from Smart (1977)

Fig. 6.2 The relationship between cave meander wavelength and channel width have been adopted by previous workers, it is considered that the available data provide an acceptable picture of the relationship between Lw and W, and one which appears to apply for caves in different limestone lithologies and under a variety of climatic conditions. On the basis of measurements from six caves, however, Smart (1977) proposed a relationship of the form $Lw = 64\ W^{-0.54}$. As Fig. 6.2 illustrates, Smart's data tend to cluster above the calculated regression line of the present work, although not exclusively so. It is considered that the difference between Smart's data and that of the remaining workers is at least partly due to the method of measurement adopted. Smart measured the parameters with considerable accuracy in the field; by contrast, most previous workers have obtained their data from pre-existing cave surveys. Deike and White (1969, 233-236) and Baker (1973, 703), for example, were only able to make a general measurement of cave passage width, whereas Smart (1977, 66) was able to measure stream channel width. On the basis of this difference alone, most previous workers would have overestimated W for any given value of Lw, with the result that their values would plot below those found by Smart.

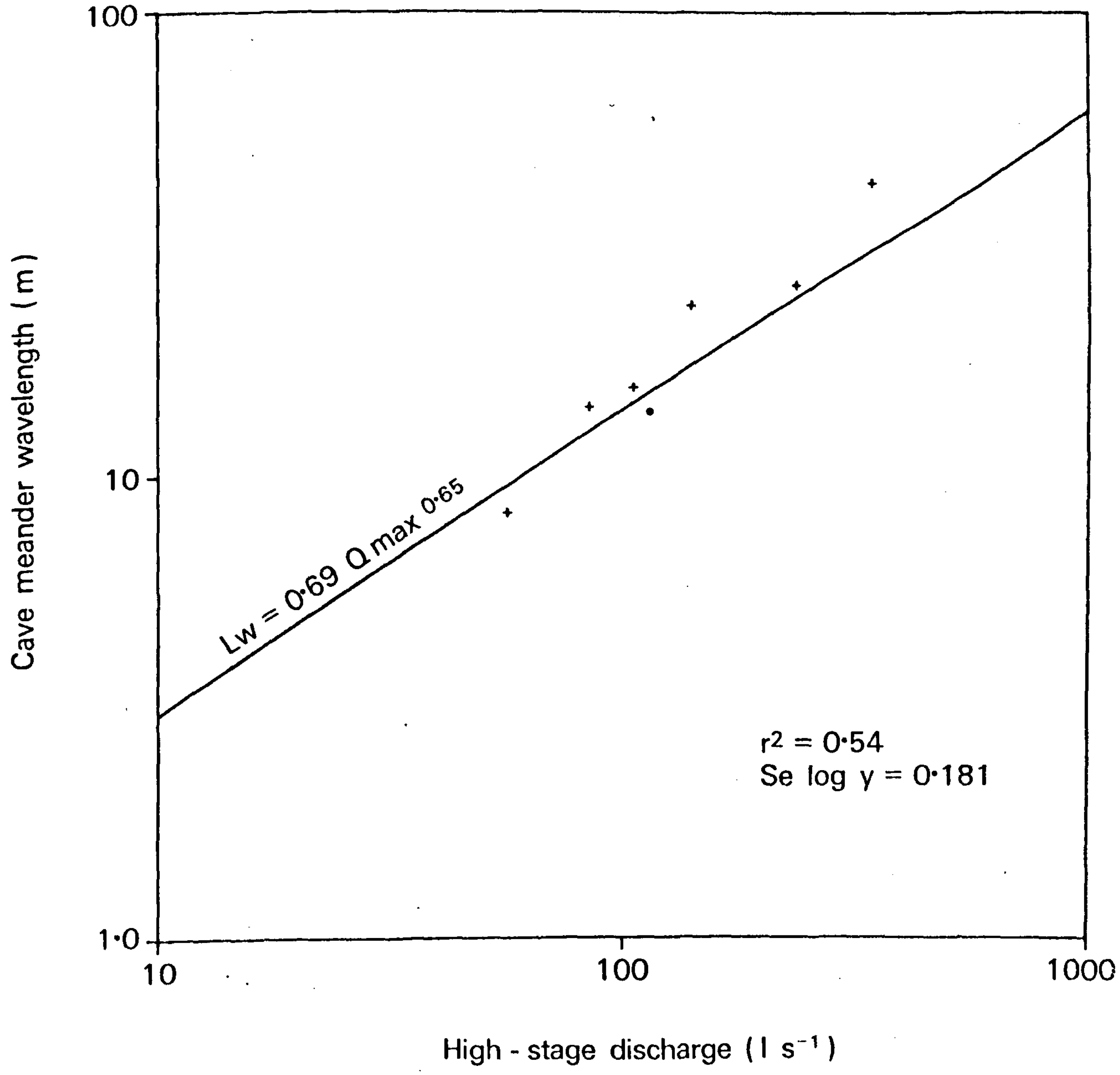
6.4.2 The use of meanders as palaeohydrological indicators

Studies of meanders in surface streams have established relationships with channel slope and the amount and character of sediment load (Gregory and Walling, 1973, 249-253). More particularly, close relationships have been shown to exist between meander morphology and stream discharge. The critical discharge has been regarded as Qb by a number of workers, for example, Inglis (1947), Leopold and Wolman (1957, 58) and Dury (1965, 5). Carlston (1965, 880), however, has shown that, for 14 rivers in the eastern United States, scatter was minimised to a standard error of 11.8% by plotting Lw as a function of Q_{40}. He concluded that Lw was controlled by a range of flows of 10-40% exceedance probability per annum. The use of exceedance probability as a measure of discharge was criticised by Ackers and Charlton (1970, 237) as an unsuitable parameter in situations where the flow cycle includes a drought period. More importantly, using the same variables as Carlston, Schumm (1967) found an order of magnitude scatter for 36 stable alluvial rivers in Australia and the United States. He suggested that the relationship derived by Carlston was partly the result of sampling only one type of fluvial system.

In an effort to resolve the problem, Ackers and Charlton (1970) used a wide range of new and pre-existing data to evaluate the influence of four possible controlling discharges on meander morphology: Qmean, Qmm, Qma and Qb. They showed that Carlston's preference for a dominant discharge approximating to Qmean gave a considerable scatter of points and a relatively low explanation in comparison with Qb.

However, these results are not necessarily directly applicable to the reconstruction of flow conditions in fossil meandering cave passages, for it has been suggested that the characteristics of the channel bed may modify the morphological relationships characteristic of free meanders in alluvial channels. Hack (1965), for example, observed a four times increase in meander wavelength as rivers crossed from alluvial to bedrock reaches. This observation was supported by Chang and Toebes (1970), and by Tinker (1971), who noted that meander wavelength was higher for bedrock channels in general. Tinkler (1972) and Kennedy (1972) suggested that meanders are characteristic of the discharge most effective in eroding the channel banks, the greater strength of bedrock requiring more energy for erosion than alluvial meanders. Hence the channel-forming discharges would be greater, resulting in larger meander wavelengths and a longer controlling flood recurrence interval.

Thus, in the case of cave meanders, it is inadequate to apply relationships derived from alluvial environments. Unfortunately, only a limited amount of data relating cave meander form to discharge exists. The existing information is summarised in Fig. 6.3. This consists of measurements of high stage flows, probably of maximum annual flood, from caves in East Central New York (Baker, 1973) and similar measurements of summer floods along Shaft Gallery in Poulnagollum, County Clare (Smith, High and Nicholson, 1969, 109). The calculated regression line has the form $Lw = 0.69\ Q_{max}^{0.65}$ (where Lw is in m and Q_{max} is in $l\ s^{-1}$) ($r^2 = 0.54$ and Se log y = 0.181). The discharge term of the regression equation is a measure of high stage flow, possibly comparable to Qb or Qma. Thus, the calculated regression equation may be cautiously used to reconstruct the quasi-equilibrium flow conditions along abandoned meandering cave passages.



+ Data from Baker (1973)
• Data from Smith, High and Nicholson (1969) and Smart (1977)

Fig. 6.3 The relationship between cave meander wavelength and high - stage discharge

6.5 Cave scallops as palaeohydrological indicators

6.5.1 Introduction

Scallops are the eroded, ripple-like forms frequently found on the walls and roofs of caves (Plate 6.1). They are generally regarded as the result of the adjustment of the form of the cave passage to turbulent flow conditions. This adjustment seems to occur solely as a result of solution, for many scalloped cave walls have insoluble material standing out from the surface (Bretz, 1942, 731; Coleman, 1949, 58), and highly friable, non-calcareous beds may be found unscalloped adjacent to areas of scalloped limestone (Goodchild and Ford, 1971, 54). Furthermore, many workers have simulated the development of scallops solely by solution under flume conditions (see 6.5.2 and 6.5.3).

There has been considerable debate concerning the hydraulic implications of scallop geometry. Two main schools of thought exist: the defect school, which considers the location and size of scallops to be fixed by bed inhomogeneities; and the passive bed school, which regards scallop form as a function of fluid flow conditions.

6.5.2 The defect school

The earliest proponent of this school was Coleman (1949, 58-59), who regarded scallops as forming initially around bed inhomogeneities which indicated localised flow vortices. The inhomogeneities were thereby enlarged, mainly by abrasion. Once formed, the scallops stabilised the location of the vortices, presumably leading to a maintenance of scallop position and size. Yeh (in Davies, 1963) also proposed a

Plate 6.1 Scallops on the roof of Capeshead Cave (SD33337814)

a defect origin. However, he considered that the initial vortex would continue to erode the bed, forming larger and larger scallops. Thus, scallop size could be regarded as age-dependent, although Yeh failed to consider the implications of overlapping scallops. Rudnicki (1960) provided experimental evidence that bed irregularities determined scallop location. T.D. Ford (1964, 12) argued that limestone lithology was important in the location of cave scallops. Ollier and Tratman (1969, 81-83) thought that scallops grew from separate small depressions under turbulent flow conditions which caused differential rates of solution. They considered that, once the scallops had coalesced, their form did not change.

The major proponent of the defect school, however, has been Allen. In his earliest paper (Allen, 1971a), he studied the effects of turbulent flow in a flume on a soluble bed of calcium sulphate. On an initially plane bed, he observed scallops to develop from bed defects of greater than a critical size. In a subsequent paper, Allen (1972) confirmed his findings on the defect origin of scallops and concluded (1972, 7) that, provided no secondary processes intervene, scallop spacing is controlled by the initial distribution of defects on the bed. Thus, flow velocity is only significant in that it determines the critical defect size which enables scallops to develop. However, he also noted the results of some secondary effects which occurred once a degree of scallop maturity existed (1972, 18). Some of these effects acted to produce new solutional marks, hence reducing mean scallop wavelength; others acted to increase mean scallop wavelength by enlarging existing scallops. Allen concluded that his experiments had been conducted over a period inadequate to determine whether wavelength eventually became independent of time.

6.5.3 The passive bed school

The earliest proponent of this school was Bretz (1942, 731) who maintained that scallop form was determined by the characteristics of the flow rather than by the nature of the bed. In following this approach, most subsequent workers have proposed an inverse relationship between scallop size and flow velocity. This was first suggested by Glennie (1963) and supported by Eyre (1963), who observed that scallops in Easegill became smaller as flow accelerated, eventually becoming non-existent. Eyre (1964) later noted that scallops developed on a bulge on a cave wall in Gaping Hill were smaller on the upstream, accelerated flow, side than on the downstream, decelerated flow, side. This was subsequently confirmed quantitatively by Goodchild (1969, 106). Moore and Nicholas (1964, 11) provided a plot of the relationship between scallop wavelength and mean flow velocity having the form $L = 56\ \bar{u}^{-0.75}$ (in c.g.s. units). On the basis of measurements of scallops in the Nakimu Caves, British Columbia, Goodchild (1969, 88-91) found scallop wavelength to be inversely related to channel slope. Since flow velocity is a direct function of slope, Goodchild concluded that scallop wavelength was an inverse function of flow velocity. Finally, from a study of flow over a soluble calcium sulphate bed in a flume, Rudnicki (1960) found that the number of scallops per unit area increased in faster waters.

Other workers, however, have proposed a direct relationship between scallop wavelength and flow velocity. Thus, T.D. Ford (1964) claimed that large scallops are produced by turbulent, sediment-laden waters, while small scallops are produced by solution in slowly moving waters.

The first rigorous study of solutionally developed bedforms was that of Curl (1966). Although this was restricted to flutes, Curl regarded both scallops and flutes as the result of the interaction of fluid flow and rate of solution of a soluble surface. As the bedforms develop, the flow pattern and rate of removal of material are modified. If the form and processes can eventually come into equilibrium so that form is no longer modified, but propogates unchanged into and along the bed, then stable forms will result.

Curl (1966, 126) simplified the problem by considering only certain parameters to be significant:

(i) Flow conditions: These were initially characterised by $\bar{u}$ and H. Blumberg (1970) subsequently used u_{max} in the vicinity of the wall, and Goodchild and Ford (1971) used velocity at 3 cm above the bed. Blumberg and Curl (1974) finally characterised flow velocity in terms of $u*$.

(ii) Fluid properties: According to Curl (1966, 126) the only ones which can be involved are ρ_f and μ.

(iii) Rock properties: Curl regarded rock solubility as only important with respect to rate of propagation. However, he acknowledged that the diffusivity of solute ions in solution (η) might affect the pattern of local solution rates through the concentration boundary layer (1966, 126-127). The initiation of scallops from surface irregularities was recognised (Blumberg and Curl, 1974, 735), but it was maintained that the initial distribution of irregularities has no effect on the final equilibrium dimension of bedforms.

Thus (Curl, 1966, 127),

$$L = f(\bar{u}, H, \rho_f, \mu, \eta) \tag{6.7}$$

The choice of these parameters was criticised by Allen (1972, 5), who suggested a more complex relationship existed. Goodchild (1969, 54-59) had, in fact, taken into account the effect of flow acceleration on the wavelength of scallops developed under flume conditions, but it was found not to be of significance.

By the application of dimensional analysis to equation 6.7, Curl (1966, 127) derived the following relationship:

$$\rho_f \bar{u} L/\mu = f(L/H, \mu/\eta \rho_f) \tag{6.8}$$

The length ratio, L/H, is generally small and may be discounted. $\mu/\eta\rho_f$, the Schmidt number (Sc), which relates the diffusivities of momentum and matter, controls the relative thickness of the mass and momentum transfer layer. Curl (1966, 127) concluded that Sc should not be important when it is large and the concentration boundary layer is extremely thin. Hence, if $\rho_f \bar{u} L/\mu$ is only weakly dependent on Sc, then $\rho_f \bar{u} L/\mu$ must be nearly constant. Thus,

$$\rho_f \bar{u} L/\mu = Re_L \tag{6.9}$$

where Re_L is the stable flute Reynolds number.

However, Wigley (1972) used data from Goodchild and Ford (1971) to demonstrate a high probability that mean scallop length depends significantly on Sc. Following Wigley, Blumberg and Curl (1974, 739), in their reiteration of Curl's (1966) earlier work, noted a possible dependence of Re_L on Sc. However, under flume conditions, they produced flutes at 27.6°C and 44.5°C, with flow velocities adjusted to give nearly identical values of Re_L, but with values of Sc differing by a factor of two. In both cases similar stable flute patterns resulted. They concluded that the Schmidt number does not have a dominant influence on the flow marking stability pattern (Blumberg and Curl, 1974, 743).

Assuming Curl's argument to be correct, then $\bar{u}$ may be determined from a knowledge of L, Re_L, and the fluid properties. To this end, Curl (1966, 128) measured these variables in two contrasting environments where flutes occurred, in an active cave stream and in an active ice cave. He obtained Re_L values of 23500 and 21600 respectively, supporting his thesis of the existence of a stable flute Reynolds number. The data used to obtain these values are in many ways inadequate, however. It was assumed that the bedforms were in phase with existing flow conditions and that the two environments operated similarly. The velocity of the cave stream was only estimated and, most importantly, only two values of Re_L were obtained.

Further work was done by Goodchild (1969), who studied the forms generated on a soluble calcium sulphate bed in a flume. This was subsequently also reported by Goodchild and Ford (1971). Goodchild (1969, 54-59, 145-148) took into account three variables: flow velocity, fluid viscosity and acceleration of flow. He found an almost linear inverse relationship between mean scallop wavelength and velocity. An estimate of stable scallop Reynolds number made on the basis of mean scallop length and velocity measured in mid-stream 3 cm above the channel bed gave a value of 11476, although it was noted that the use of mean velocity would result in a higher value of Re_L. However, no indication is given as to whether stable bedforms had been allowed to develop, although a run time of 100 hours is implied by Goodchild and Ford (1971, 55).

In the same paper, however, Goodchild and Ford (1971, 59-61) cited the example of the Bonnechère Caves, Ontario where scallops of quite different wavelengths have developed on distinct limestone beds, which must have been subjected to similar hydrodynamic conditions. They argued that this represented a "serious disagreement" with their flume results. Yet in another paper, Ford (1971b, 33-34) revealed that scallop length in the Bonnechère Caves appeared to be controlled by the distribution of insoluble matter through each limestone bed, closely spaced insoluble fragments restricting the mean length of scallops and vice versa. Under these conditions, scallops are obviously unable to attain equilibrium with the flow. The Bonnechere scallops may therefore be regarded as a special case rather than evidence against the equilibrium development of scallops.

Curl subsequently extended his earlier work to deal with the development of scallops (Blumberg and Curl, 1974; Curl, 1974). Scallops were regarded as developing initially from irregularities in the channel bed (Blumberg and Curl, 1974, 735). However, it was considered that the distribution of irregularities has no effect on the final distribution of bedforms and that, with continued dissolution, an equilibrium pattern of scallops is able to develop. This pattern was regarded as stable in the sense that local mass transfer rate distributions remain consistent with the geometry, although the equilibrium scalloped surface itself is probably only stable in the statistical sense.

In this later work, Curl modified his original equations, replacing the mean channel velocity by the bed shear velocity. Under turbulent flow near a wall, the velocity profile depends primarily on the wall roughness and the bed shear velocity (cf. Prandtl's universal velocity distribution law). Thus, for uniform flows, the characteristic scallop dimension (L) will be a direct function of u*, rather than an indirect function of $\bar{u}$ as previously assumed. Hence Re_L may be replaced by Re*, which ought to provide a better basis for the analysis of scallop dimensions.

$$Re* = \rho_f \, u* \, L/\mu \qquad 6.10$$

Having obtained a characteristic scallop dimension (L) and knowing Re*, it should be possible to calculate u. Good approximations of velocity profiles near rough walls are given by Prandtl's universal velocity distribution law:

$$u_y \, / \, u* = 2.5 \log_e (y/L) + B_L \qquad 6.11$$

$\bar{u}$ may be found by integrating equation 6.11 from y = 0 to y = d/2, where d may be either the diameter of a circular conduit or the width between the two parallel walls of a channel. Hence,

for a circular conduit

$$\bar{u} = u* (2.5 (\log_e (d/2L) -3/2) + B_L) \qquad 6.12$$

for a parallel walled conduit

$$\bar{u} = u* (2.5 (\log_e (d/2L) -1) + B_L) \qquad 6.13$$

Blumberg and Curl (1974, 741-743) confirmed Curl's original measured value of Re_L by flume experiments (9 runs giving flutes and 2 runs giving scallops), which gave values of Re_L ranging between 20500 and 25600 (20500 and 21000 for scallops alone). To obtain corresponding values of Re*, the Prandtl equations may be multiplied by $\rho_f L/\mu$. Thus, for a circular conduit,

$$Re_L = Re* (2.5 (\log_e (d/2L) -3/2) + B_L) \qquad 6.14$$

The values of Re* thus obtained ranged between 1920 and 3560 (2120 and 2320 for scallops alone). Blumberg and Curl also found from their experiments values of B_L of 7.2-11.7 (8.8 and 9.9 for scallops), giving a mean value of 9.4. Unfortunately, it is unclear from their report how long the flume runs lasted and whether bedform stability was achieved.

In order to investigate Curl's concept of a stable scallop Reynolds number, Allen (1971b, 332-333) substituted data from his flume experiments into equation 6.14, obtaining values ranging between 15740 and 85150. Allen concluded that scallop geometry is not dependent on flow conditions. However, a plot of Re_L against relative scallop age derived from Allen's data shows that, with increasing age, Re_L approaches a constant value of ~16500 (Allen, 1971b, 334). This is somewhat less than the value obtained by Curl. However, Blumberg and Curl (1974, 750) noted that if Re_L is calculated using the Sauter-mean wavelength ($\Sigma L^3/\Sigma L^2$), as they recommend, then a value of ~25000 is obtained, which agrees closely with their results.

Allen (1971b, 251-253) also pointed out in his argument against the passive bed school that spatially-periodic kinematic structures within the flow occur only during the laminar-turbulent transition phase (Re = 1260-3000). Under steady laminar flow conditions there are no kinematic structures of dimensions smaller than the boundary layer itself, whilst under quasi-steady turbulent flow conditions, kinematic structures are short-lived and spatially random. But it is precisely at laminar-turbulent transition values of Re* that scallops appear to develop, suggesting that spatially-periodic flow structures are indeed significant in their formation.

Thus, if Re* and B_L are known from experimental work, if d and L can be measured in the cave, and if values of μ and ρ_f can be assumed, then Curl's method appears a suitable approach to the calculation of u.

Recent work by Thomas (1979) has supported Curl's conclusions. Thomas presented evidence from a wide range of environments (e.g. metal pipes, boiler tubes, ice) for scalloping as the result of the imprint of eddy patterns inherent in turbulent flow. His data spanned five orders of magnitude of scallop wavelength and from these he derived a relationship of the form

$$L = 10^3 \nu / u_* \qquad 6.15$$

giving an Re* value of 1000.

6.4.4 The application of scallops to the determination of palaeohydraulic conditions

The practical application of the relationship between scallop size and flow conditions lies in deducing palaeoflow velocities in hydrologically inactive cave systems. The choice of a dimension to characterise scallop size is to some extent rather arbitrary. Most workers have measured scallop wavelength along the maximum length parallel to the direction of flow, taking the mean of a number of values.

However, Curl noted that bed inhomogeneities may result in the development of small scallops. In addition, small scallops appear to develop around the rims of larger scallops. Consequently, he recommended that the Sauter mean ($\Sigma L^3/\Sigma L^2$) be used in order to suppress the importance of smaller features when characterising scallop assemblages (Blumberg and Curl, 1974, 743; Curl, 1974, 3).

In order to apply the methods detailed in 6.5.3 to the analysis of scallops in caves, a variety of criteria must be met (Curl, 1974, 4). The cave passage where the scallops are measured must be of regular cross-section and must be sufficiently long and straight for almost fully developed flow to be established. Assuming the scallops were formed under phreatic conditions, it has been shown by Goodchild (1969, 106-108) that it makes no significant difference whether scallops are measured on the cave walls, roof or floor. Finally, the passage cross-section should be unchanging for some distance either side of the point of measurement. In order for equilibrium scallops to develop, uniform flow conditions must prevail, at least throughout the period of final development of the scallop pattern. In any specific case this is unlikely to be true. However, as it is likely that scallops develop in equilibrium with certain flows (see below), the problem of non-uniform flow may be resolved satisfactorily. Finally, it should also be noted that a variety of other factors may modify scallop patterns. These include close fracturing and insoluble inclusions in the limestone, a heavy bedload, and the deposition of clays.

In order that scallops and flutes may be used as palaeohydrological indicators, it is important that the relationship between bedform dimension and the range of flows naturally occurring within a stream be understood.

It is clear that solution is the only process at work in the formation of scallops and flutes (see 6.5.1). The rate of solution in a pipe under turbulent flow conditions has been shown by Wigley (1971) to be determined by the Sherwood number (Sh). This is a known function of the Schmidt and Reynolds numbers of the flow (Monteith, 1973, 135).

$$Sh = f(\rho_f u L/\mu, \mu/\eta \rho_f) \qquad 6.16$$

Assuming fluid properties to be constant, the rate of solution in a pipe is therefore a direct function of flow velocity and channel size, and an inverse function of the diffusivity of solute ions in solution. The relationship between flow velocity and the solution rate of limestone has been demonstrated by Kaye (1957, 37-40), by Weyl (1958) and by Newson (1970, 79-81), whilst micro-erosion meter measurements made in vadose streamway caves by High (1970a, 79, 104-110) show the rate of lowering of the cave bed by solution to be directly related to stream velocity. These results support Curl's (1974, 4) observation that scallops develop most rapidly at high velocities.

In general, and certainly under phreatic conditions, higher velocities imply higher discharges. Thus, one might expect more solution to take place under higher stage flows. This is borne out by the results of Newson's (1970, 74-75) work on the weight loss by solution of limestone tablets in cave streams, which suggested that 56-60% more solution occurred during a flood event of 60-110 year return period (Hanwell, 1969, 211) than during a similar period of "normal flow". Rather different results were obtained by High (1970a, 76-77), who used micro-erosion meters to study active vadose streamway caves in County Clare. Measurements in Cullaun 1 over a period of three years showed nearly constant rates of streambed lowering, despite the occurrence during this period of a >50 year flood. However, it is likely that any effect of the flood remained undetected due to the long time interval at which bed lowering was sampled, up to 200 days at the time of the flood.

That scallop development also occurs rapidly at high discharges has been shown by Doehring and Vierbuchen (1971), who studied the effects of a >1000 year flood in Cave Springs Cave, Virginia. The flood resulted in the development of fresh scallops, although it is likely that these were a re-development of existing features, and it is difficult to know from the report whether the new scallops were in phase with the flood conditions.

Similarly, Smart (1977, 57, 114) measured scallops of smaller wavelength above the level of vadose streams compared with those below. The higher scallops may be interpreted as having developed in response to higher stage flows of higher flow velocity, whilst the lower scallops are modified by the more frequently occurring lower stage flows.

From a palaeohydrological point of view it is important to know whether scallops exist in a state of quasi-equilibrium with a range of flows, whether there is a dominant scallop-forming flow in terms of magnitude-frequency considerations, or whether scallops are continually modified to be in phase with the most recent flow. Laboratory simulation of scallop formation has demonstrated that scallops do not immediately develop in phase with flows, even under steady flow conditions over highly soluble beds (see 6.5.3). This suggests, firstly, that it is unlikely that scallops will simply develop in phase with the most recent flow, and, secondly, that the dominant flow or range of flows is likely to reflect some balance between the most frequently occurring flow and the flows that perform most solutional work, in other words, a balance between the modal flow and the rarer, large-scale event. Unfortunately, data do not exist to enable the determination of the dominant scallop-forming flow(s). Therefore, as a first approximation, it is proposed that scallop size be regarded as related to the maximum annual flood, thereby enabling a direct comparison to be made with palaeodischarge values established from the study of cave meanders. Although the considerable drawbacks of this approach are appreciated, it is argued that this method may be used to establish at least the order of magnitude of cave palaeodischarge.

6.6 The estimation of palaeocatchment area using palaeodischarge values in karst areas

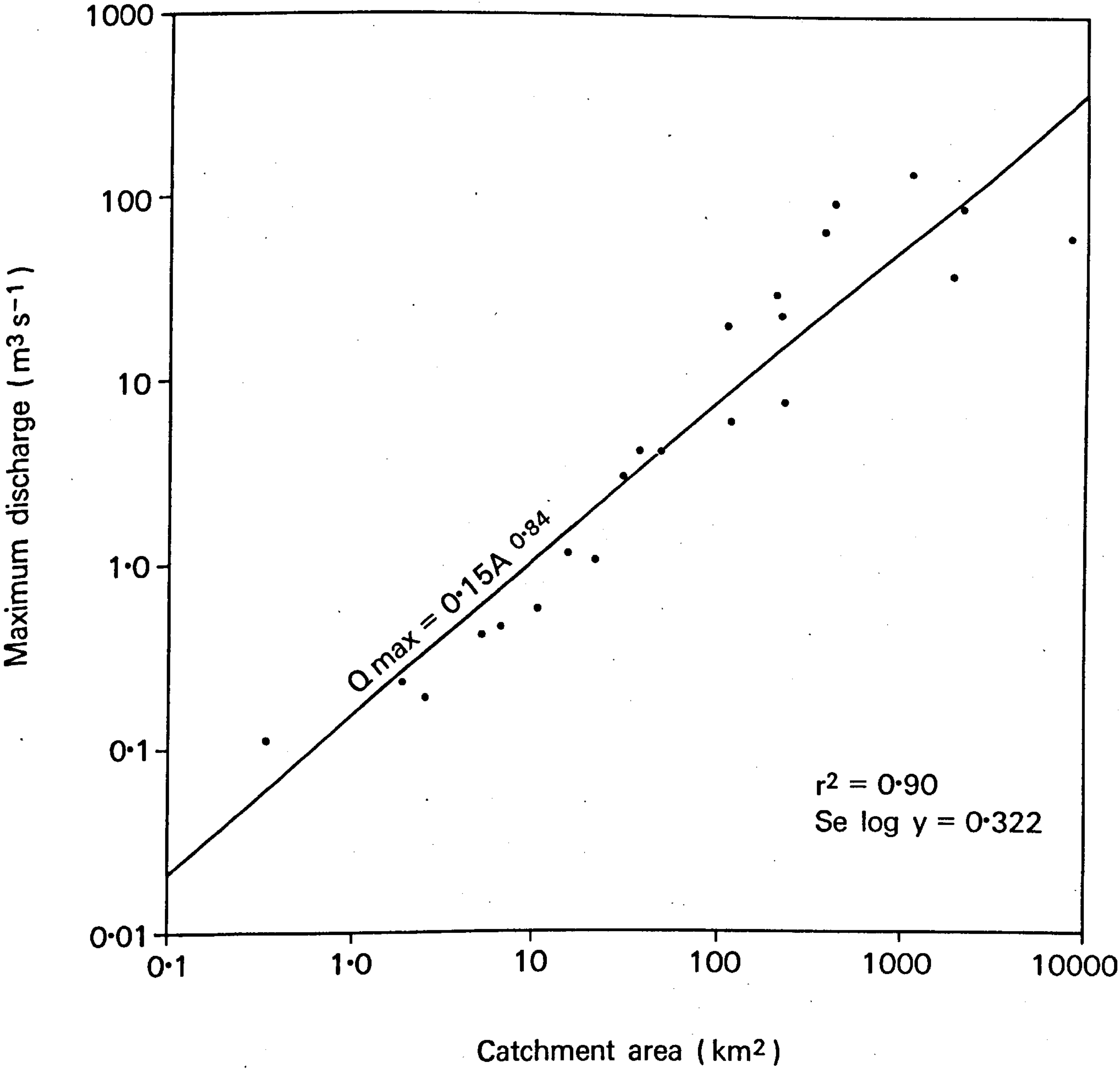
The present recharge of the drift-free areas of Carboniferous Limestone in the Morecambe Bay area has been estimated as 420 mm yr^{-1} by Wadge (1966), giving a relationship of Qmean = 0.133A (where Qmean is in $m^3\ s^{-1}$ and A is in km^2). Assuming that all percolation within a given catchment resurges at the point of discharge measurement, this relationship may be used to estimate the contributing area of active caves and springs in the Morecambe Bay karst. Applying the relationship to palaeodischarge values would permit the reconstruction of palaeocatchment areas within the region. However, this approach has two disadvantages. Firstly, present percolation rates cannot be justifiably extrapolated back to earlier environmental conditions. Secondly, the bulk of the palaeodischarge figures obtained are of Qmax rather than Qmean.

In order to obtain an indication of the variation of Qmax as a function of catchment area, data from a variety of temperate humid karst catchments (Table 6.1) were plotted, and a relationship of the form $Qmax = 0.15A^{0.84}$ ($r^2 = 0.90$, Se log y = 0.322) obtained (Fig. 6.4). Given the fact that discharge is a function of the volume of percolation taking place over a catchment and is only an indirect function of the catchment area itself, the strength of this relationship is unexpected. However, with regard to the similarity of climatic conditions under which these catchments are found, it is perhaps justifiable to regard variations in the magnitude and frequency of percolation input as relatively insignificant in comparison to the total volume of percolation input as a function of catchment area.

What is perhaps more interesting is that a simple value of Qmax should provide such a strong relationship, even in catchments where discharge has been measured over a relatively short period of time. It might be expected that higher magnitude, lower frequency floods would result in a greater spread of discharge values on Fig. 6.4. A possible explanation

Location	Catchment area (km^2)	Q max (m^3 s^{-1})	Q type	Climate (Trewartha's modification of Koppen)	Source
Orangeville Rise, Upper Lost River, Indiana	119	5.7	maximum 1972-73	Daf	Bassett, 1976
Elk River Springs, West Virginia	238	7.16	high stage	Dbf	Medville, 1977
Silver Spring, Florida	1900	36.5	maximum recorded over number of years	Caf	Faulkner, 1976
Ombla Spring, Yugoslavia	2100	88	maximum recorded over number of years	Caf	Milanovic, 1976
Ljubljanica Karst Basin, Slovenia	1109	131.7	maximum 1972-75	Caf	Zibrik *et al*, 1976
Ashwick Lower Spring, Mendip	2.6	0.19	maximum 1966-67	Cbf	Drew, 1967 and 1975
Cheddar Spring, Mendip	38.0	4.04	maximum 1970	Cbf	Atkinson, 1971; Drew, 1975
Rodney Stoke Spring, Mendip	6.63	0.46	" "	Cbf	" "
Wookey Hole Spring, Mendip	30.7	2.89	" "	Cbf	" "
Becker's Cave Spring, New York	0.34	0.11	mean 27.3.67-2.4.67	Dbf	Baker, 1973
Beaverdam Spring, New York	1.94	0.23	" "	Dbf	" "
Spider Cave Spring, New York	5.18	0.42	" "	Dbf	" "
Howe Cave Spring, New York	10.36	0.57	" "	Dbf	" "
Pitcher Farm Spring, New York	15.54	1.13	" "	Dbf	" "
Doc Shaul's Spring, New York	22.01	1.05	" "	Dbf	" "
Ras-el-Ain, Iraq	8100	59.5	maximum recorded over number of years	Csb	Burdon and Safadi, 1963
Achhabal, Kashmir	50	4	mean annual maximum	Caw	Coward, Waltham and Bowser, 1972
Spring Creek, Pennsylvania	225.8	21.7	2.3 year flood	Dbf	White and Reich, 1970
Spring Creek, Pennsylvania	375.5	63.2	" "	Dbf	" "
Little Lehigh Creek, Pennsylvania	209.3	27.9	" "	Dbf	" "
Monocacy Creek, Pennsylvania	115.3	19.3	" "	Dbf	" "
Kishacoquillas Creek, Pennsylvania	424.8	90.1	" "	Dbf	" "

Table 6.1 Catchment area and maximum discharge for selected temperate humid karst catchments.



Source of data: see Table 6.1

Fig. 6.4 The relationship between maximum discharge and catchment area for selected temperate - humid karst catchments is that karst aquifers possess the capacity to dampen flood peaks so that no flood occurs beyond a certain level. This explanation is supported by an investigation of a number of long-term karst spring hydrographs (see, for example, Burdon and Safadi, 1963, and Bassett, 1976). It is suggested that this is the result of discharge along karst conduits being a function of the smallest phreatic cross-section within the conduit and the greatest hydraulic gradient that can develop at that point. Although the hydraulic gradient will increase as flow backs up behind the constriction, it is considered that beyond a certain stage, water will tend to pass into bank storage in the fracture zone around the conduits with the result that hydraulic gradient will rarely increase above a certain value. Thus, it is possible that karst conduits develop in an approximate state of equilibrium with discharges of a given magnitude and frequency, here termed Qmax, and that flows of greater magnitude, and hence greater erosive potential, occur with insufficient frequency to affect materially passage form. This is borne out by Doehring and Vierbuchen's (1971) observations of the effects of a >1000 year flood in Cave Springs Cave, Virginia. Despite its catastrophic nature, this flow only effected such minor changes in passage morphology as the reworking of pre-existing solutional scallops.

Given the strong relationship found between Qmax and catchment area, albeit only within areas of temperate humid climate, it is proposed that this relationship be used to establish approximate values for palaeocatchment areas from the palaeodischarge data obtained in the present study. Such an approach assumes, firstly, that the drainage systems of the

Morecambe Bay area developed under environmental conditions similar to those of the catchments in Fig. 6.4. Assuming groundwater flow to be negligible under glacial and peri-glacial conditions, this does not seem an unreasonable assumption. Secondly, it is assumed that the data set of Fig. 6.4 is representative of temperate humid karst areas generally.