7. THE PALAEOHYDROLOGY OF THE MORECAMBE BAY KARST

7.1 Introduction

In common with many other karst areas, the Morecambe Bay area has been the focus of a considerable amount of palaeohydrological research. This work has tended to concentrate on two particular aspects of cave development: firstly, the problem of the "phreatic network" caves, variously regarded as Tertiary relicts or products of polje-margin processes; and secondly, the question of the development of the regional drainage system.

The present study attempts to resolve the question of "phreatic network" cave formation (see 7.2) and applies the methods detailed in Chapter 6 to specific cave systems in an effort to understand the role of these caves in the overall development of the drainage system (see 7.3 and 7.4). Finally, an attempt is made to synthesise these findings into a model of the palaeohydrological development of the whole region (see 7.5).

7.2 The "phreatic network" caves

7.2.1 Introduction

Certain of the caves within the Morecambe Bay area have been regarded as unique in terms of their method of formation. These caves, termed by Ashmead (1969a, 207; 1974a, 213) "phreatic network" caves, have been considered for some time as the only true "water table" caves in the British Isles (Waltham, 1974, 89-91); that is, caves which have been formed by solutional action at the water table and which tend to be nearly horizontal despite the inclination of the rock. They have also been seen as evidence for phases of karstification in Northern England dating back as far as the Tertiary (Sweeting, 1970, 239).

To date, five groups of caves have been placed in the "phreatic network" category (Fig. 7.1):

(i) Roudsea Wood: northern system (SD332826) (Fig. 7.2), southern system (SD333825) (Fig. 7.3).

(ii) Nichols Moss: Shale Cave (SD434827) (Fig. 7.4)
Fairy Cave (SD434827) (Fig. 7.4)
Yew Tree Cave (SD434826) (Fig. 7.4)
Fissure Cave (SD434826)

(iii) Gilpin Bank Caves (SD4687)

(iv) Levens Cave (SD484857)

(v) Hale Moss: Brackenthwaite Cave (SD498772) (Fig. 7.5)
Hazel Grove Cave (SD499771) (Fig. 7.6)
Hale Moss Cave (SD502777) (Fig. 7.7)

7.2.2 Previous Work

Two contrasting hypotheses have been advanced to explain the existence of the "phreatic network" caves, each having wide implications for the chronology of karst development both in the area and in northern England as a whole. Firstly, according to Sweeting (1970, 239) these caves correspond to the <u>fusshöhlen</u>, or foot caves, found in tropical karst areas. She considered it likely that the caves were developed under warmer conditions than those of the present, either during a previous interglacial or, more probably, during the Tertiary. The steep-sided limestone hills, around which the caves are found, were also assigned to the Tertiary by Corbel (1957, 289), who considered them to be

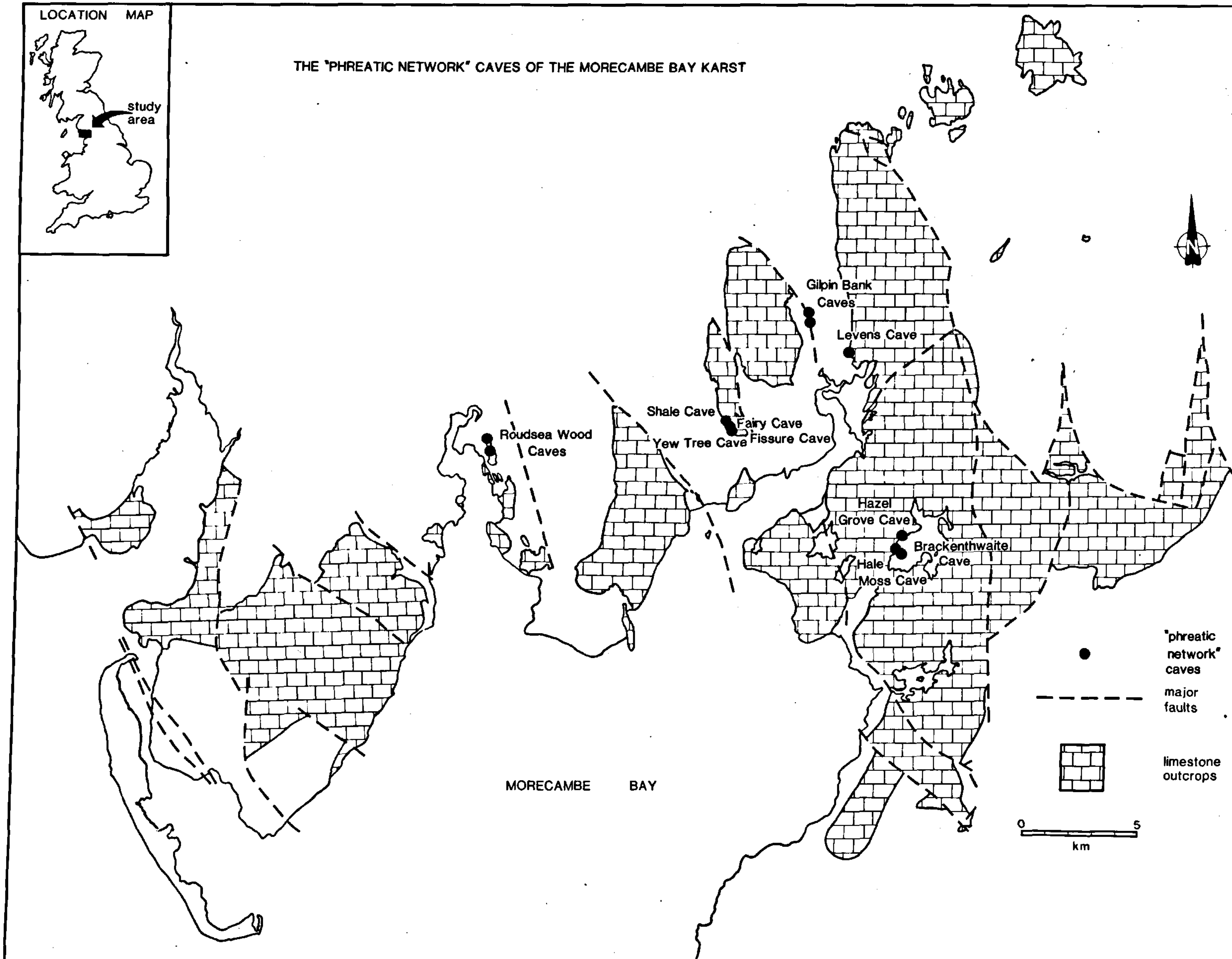


Fig. 7.1 The "phreatic network" caves of the Morecambe Bay karst

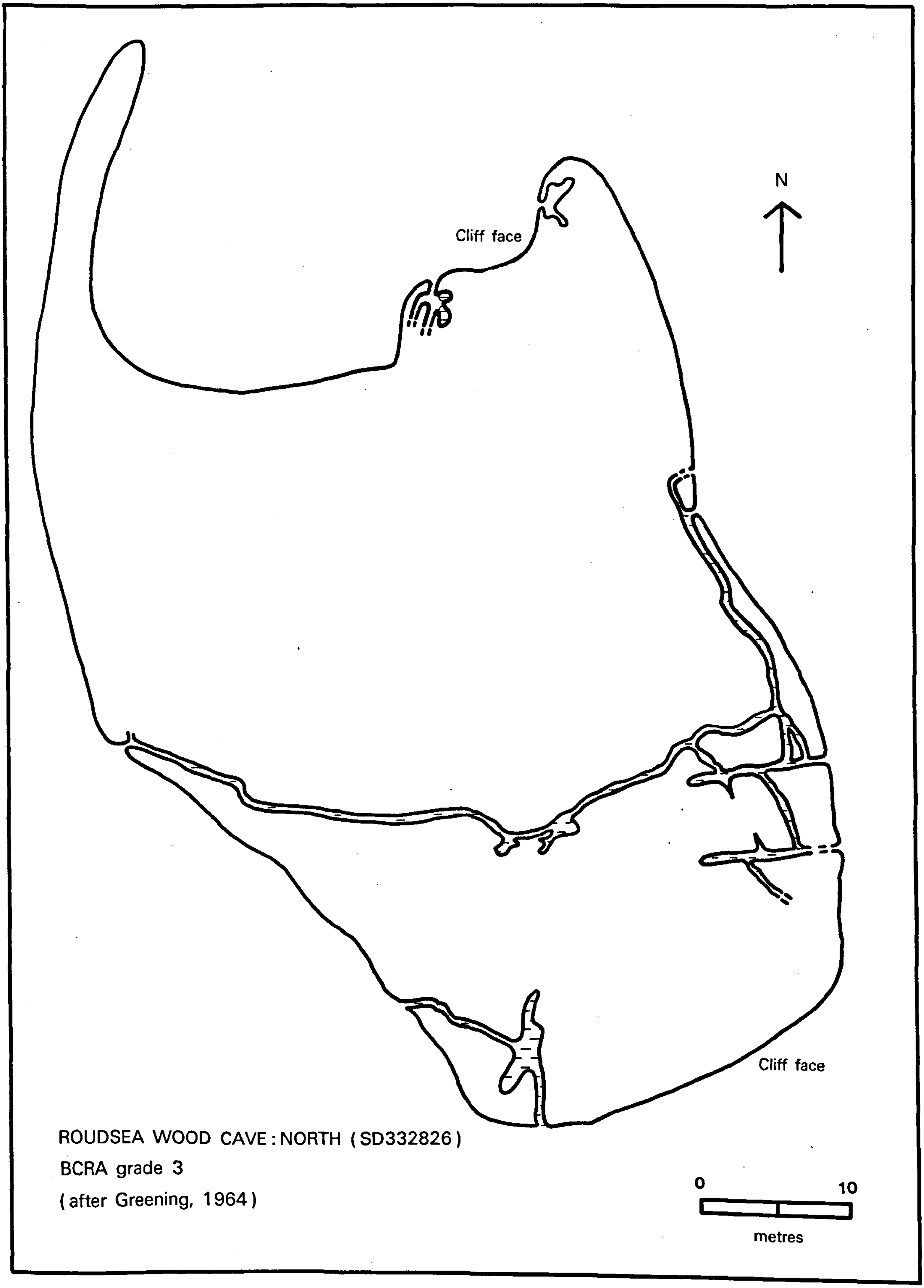


Fig. 7.2 Roudsea Wood Cave: North

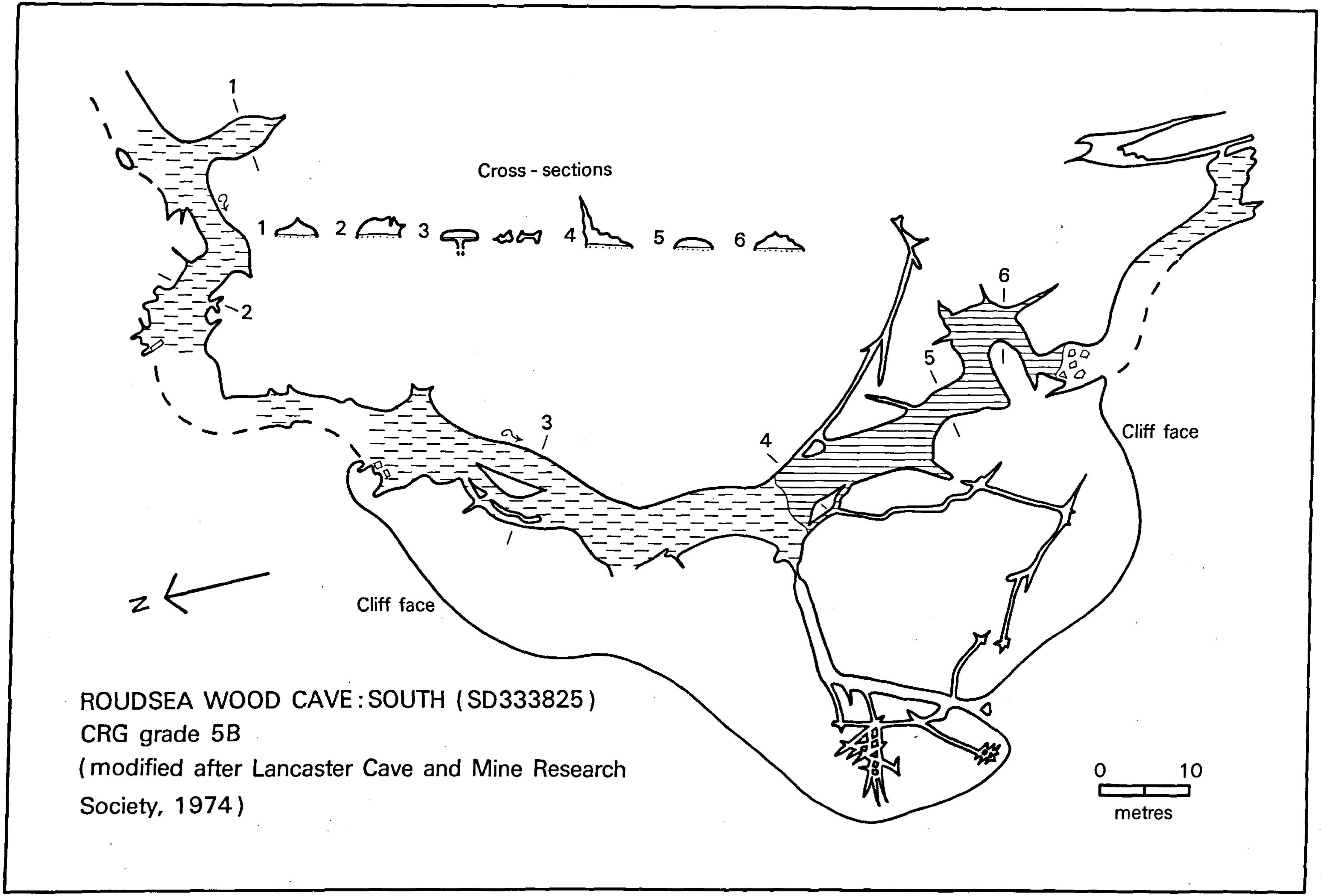


Fig. 7.3 Roudsea Wood Cave: South

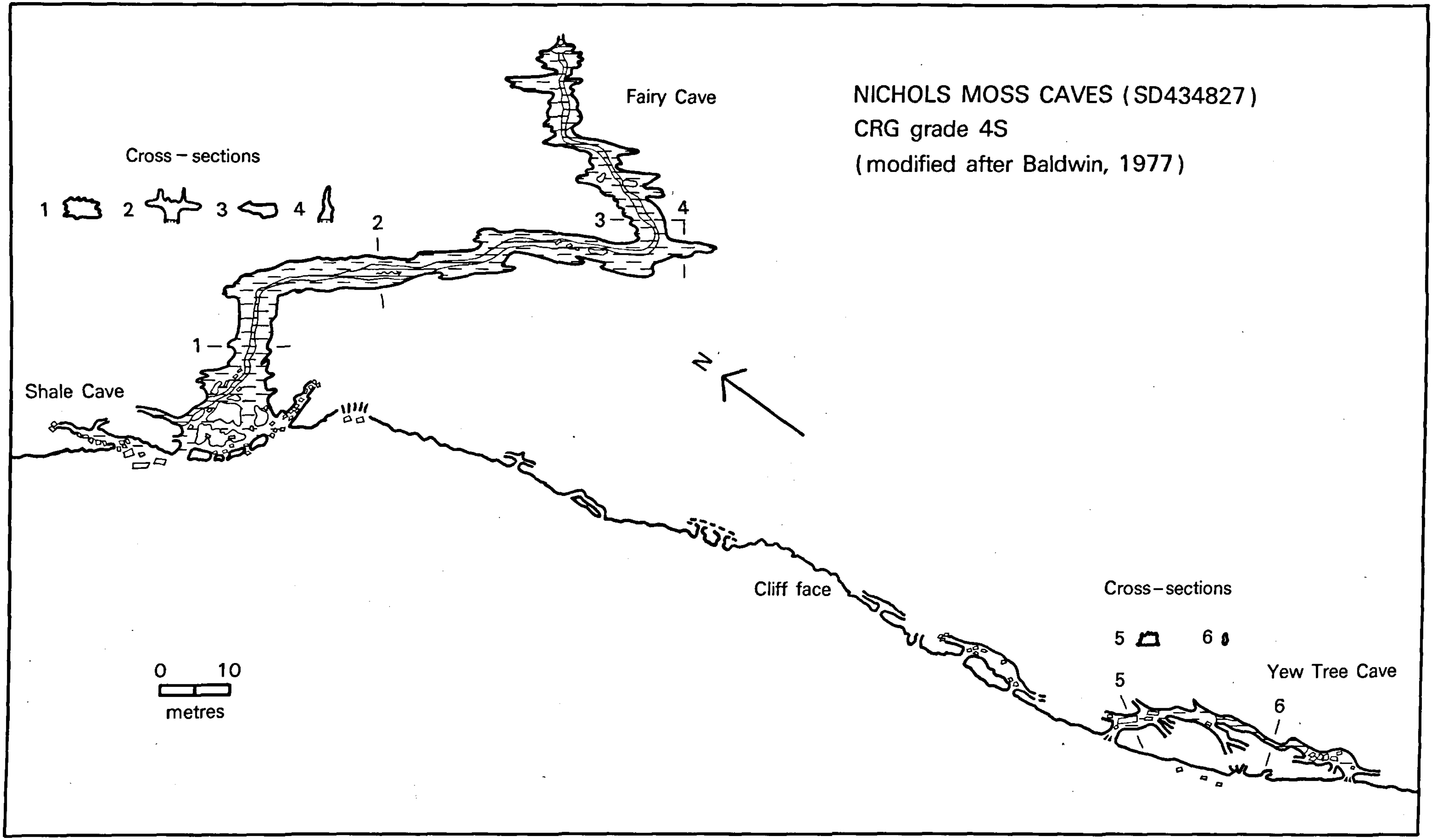


Fig. 7.4 Nichols Moss Caves

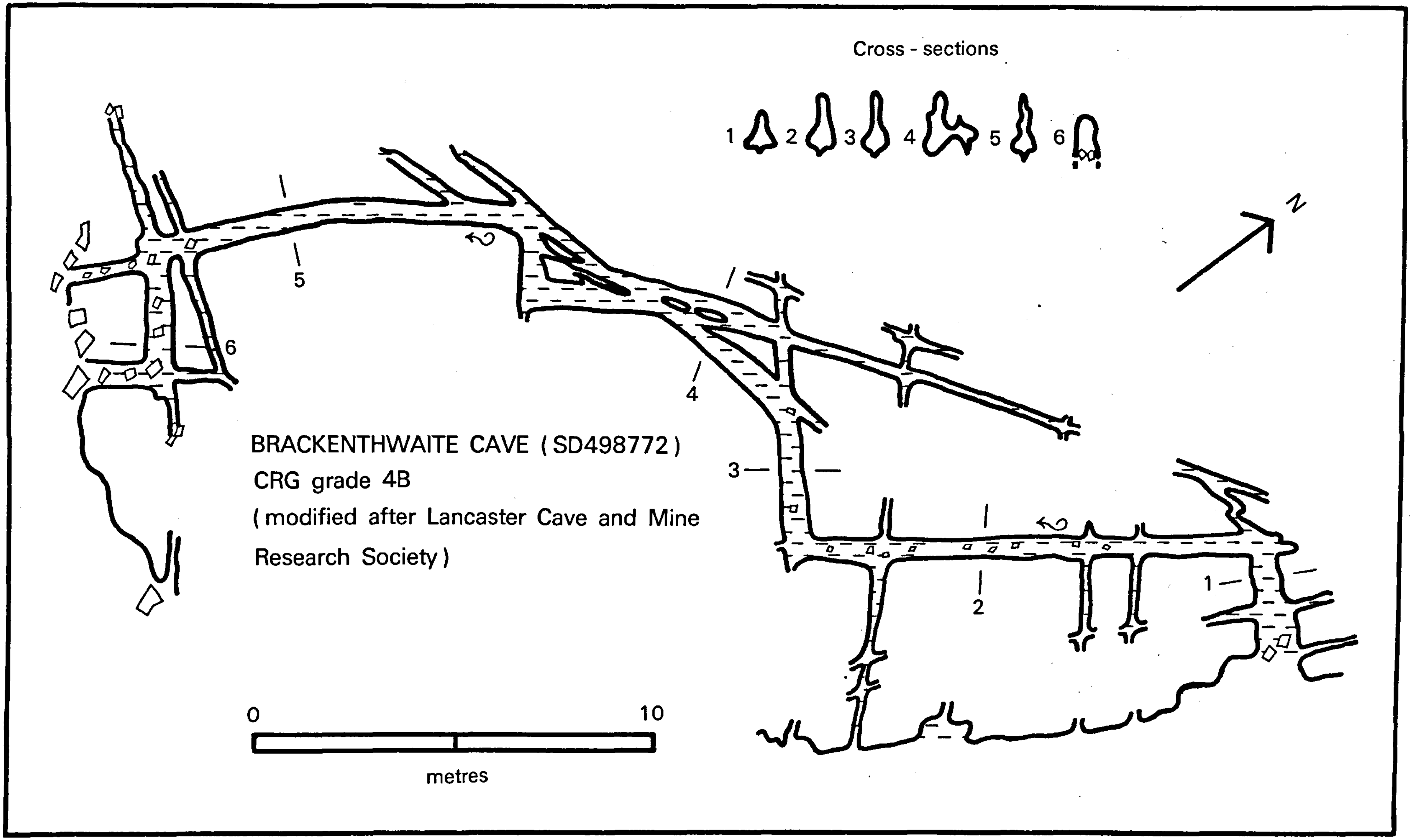


Fig. 7.5 Brackenthwaite Cave

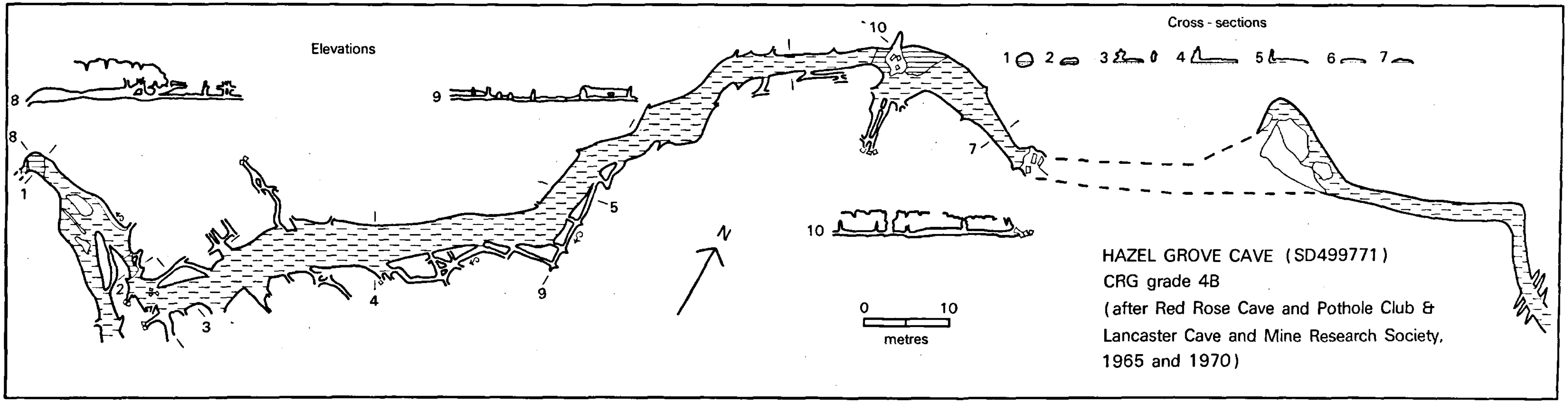


Fig. 7.6 Hazel Grove Cave

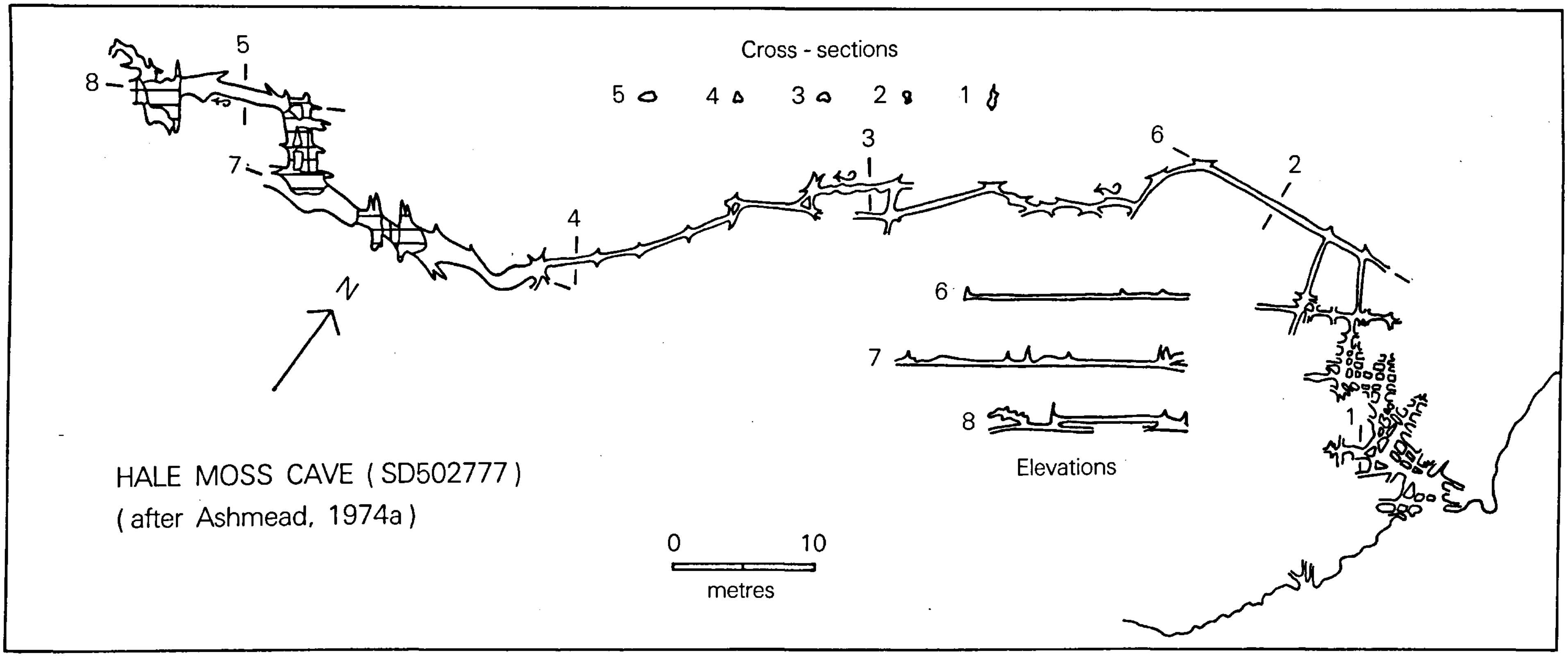


Fig. 7.7 Hale Moss Cave

fossil karst towers similar to those found in tropical regions today. Since the same processes as form karst towers are generally regarded as also having formed tower foot caves, this provides indirect support for Sweeting's thesis.

Secondly, Ashmead (1966; 1969a; 1974a, b, c), and subsequently Baldwin (1977), advanced the hypothesis that the caves were formed by solutional attack from the relatively static estuarine and lacustrine water bodies thought to have existed in the mosses of the area (see 8.3.2) and on the fringes of which all the known "phreatic network" systems are found. According to Ashmead (1966, 19; 1969a, 207), these waters were often peaty and, as a result of their high organic acid content, very aggressive. This meant that the limestone surrounding the mosses was solutionally fretted along the fractures exposed at the limestone-water interface, resulting in a tight network of intersecting solution fissures. These networks diminish in intensity of development with distance from the water edge, corresponding to the reduced aggressivity of the waters which resulted from contact with the limestone mass (Ashmead, 1974a, 217-218; 1974b, 22). The caves are almost totally phreatic in nature and the absence of any vadose development was regarded by Ashmead (1974a, 214) as supporting the idea of formation by static water bodies. The caves are horizontally developed, despite their occurrence in beds dipping at up to 25^{o}, as at Roudsea Wood. Furthermore, the vertical development of each system is restricted to a height range of 2-3 m (Ashmead, 1974a, 213; 1974b, 19-20; 1974c, 57). These features were taken by Ashmead to indicate the influence of water level control and the absence of structural constraints upon cave formation.

Ashmead (1974a, 214) suggested that cave development took place during interglacial phases of high sea level as well as during the Late and Post-glacial period. He argued that the "network" caves originated and evolved by processes integral with the solutional widening of the mosses which they fringe (see 8.3.4). It is likely that the mosses were considerably modified by the action of ice erosion during successive glacials and that the solutionally dissected moss edges would have been removed by glacial scour (see 8.2.4). It is therefore probable that the processes invoked by Ashmead to explain the "network" caves are post-Devensian in age. Ashmead himself supported this theory in the case of Roudsea Wood Caves which he regarded as having been formed by a proglacial lake dammed in the Leven Estuary (1974a, 218; 1974b, 22). With the Late Flandrian drainage of the mosses, the caves have been abandoned and now lie hydrologically inactive along the moss edges.

7.2.3 The tropical karst hypothesis: discussion

The _fusshöhlen_ of contemporary tropical karst areas are characterised by a horizontal, notch-like morphology and a generally smoothly rounded form. The notches tend to be several hundred metres in length along the base of the limestone slopes; they are commonly two to three metres high and vary in depth from one to fifteen metres. Only occasionally do connecting passages extend further into the hills (Paton, 1964; Wilford and Wall, 1965, 52-54; Jennings, 1971, 191;

1976, 96; McDonald, 1976, 85).

Fusshöhlen have been interpreted as the result of a variety of processes at work at the base of karst towers. These include:

(i) lateral fluvial solution and corrosion;

(ii) solutional surface wash;

(iii) spring sapping at the limestone-alluvium contact;

(iv) solutional undercutting by swamp waters, by ephemeral lakes at the base of the towers, and by soil moisture; and

(v) marine erosion

(Lehmann, 1953; Paton, 1964; Wilford, 1964; Wilford and Wall, 1965; Sweeting, 1972; McDonald, 1975; 1976; 1979; Jennings, 1976).

Apart from their situation along the base of the limestone hills of the area, the "phreatic network" caves of the Morecambe Bay karst have little in common with the foot caves of tropical karst. They have discrete entrances rather than a notch-like form, the majority of the caves extend into the hills, and there is no evidence of solution notches anywhere on the cliffs surrounding the mosses (Plate 7.1).

7.2.4 The static water hypothesis: discussion

Perhaps the two most significant characteristics of the "phreatic network" caves are their horizontal nature and their limited vertical extent. These features were regarded by Ashmead as indicative of formation at or near to the water table. Yet the shallow depth of the water bodies in the mosses (see 8.3) would, of itself, have confined

Plate 7.1 "Phreatic network" cave entrances, Gilpin Bank Caves (SD4687)

solutional attack to a narrow vertical range. More importantly, in many cases the horizontal nature of the caves can be shown to be the result of structural constraints.

As may be seen from Fig. 7.8, the majority of the cave systems are either strike-aligned or are in areas of low structural dip. The strike-aligned passages occasionally deviate in orientation somewhat from that of the strike, particularly in the case of Fairy Cave. It is contended that this is the result of joint-control on passage location and hence the preferential development of the joint most nearly aligned along the strike. The orientation diagrams also show an element of dip-control on passage alignment. In most cases these dip-controlled passages are very short and have either been captured by hydrologically more efficient strike passages or act simply as strike passage feeders. In the case of Roudsea Wood Cave (north), however, the apparently dip-controlled element consists of a single passage approximately 50 m long which is almost horizontal for the whole of its length (Fig. 7.2). This anomaly will be considered in more detail below.

The strike-aligned nature of the majority of these systems explains their horizontal to sub-horizontal morphology. However, it fails to explain why the systems are generally aligned parallel to the edges of the limestone masses and why the embryonic drainage systems failed to take advantage of the greater hydraulic gradient provided by drainage into the mosses. In the main, the parallel nature of the caves to the edge of the limestone mass is simply a result of the similar orientation of the strike of the rock.

THE "PHREATIC NETWORK" CAVES : PASSAGE ORIENTATION AND STRUCTURAL RELATIONSHIP

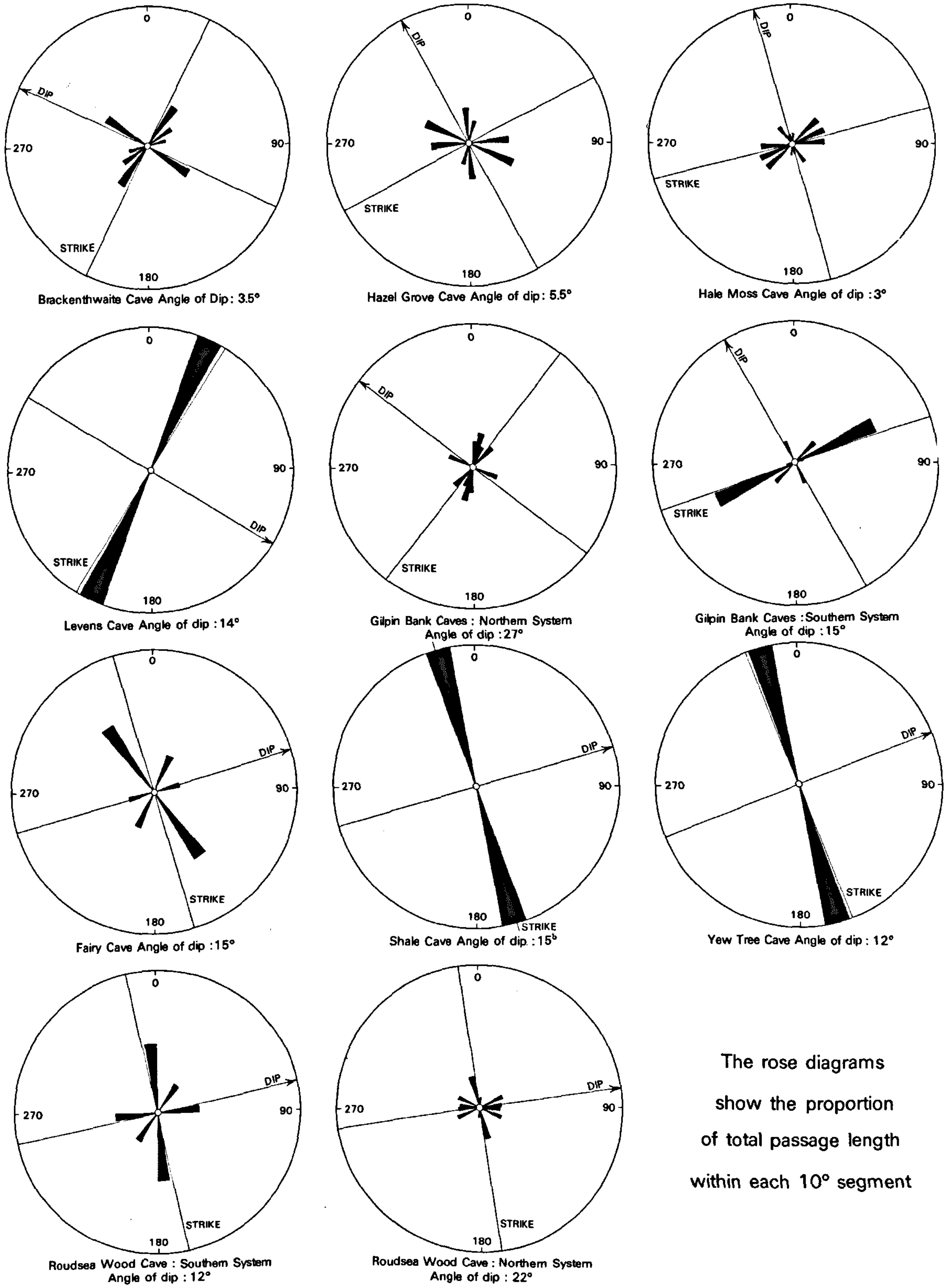


Fig. 7.8 The "phreatic network" caves: passage orientation and structural relationship

It can be seen that in those cases where the strike is not parallel to the edge of the limestone, as at Gilpin Bank, the caves still follow the strike. A structural explanation may also be proposed for the absence of drainage directly into the mosses. In all cases this would have necessitated groundwater flow updip. It is likely that, when these systems developed, the limestone scarp fronts had not retreated to their present position. Under these conditions, the up-dip scarp front would have been unable to attract groundwater drainage, whilst the dip-front would not have been exposed and able to attract down-dip drainage. Given the geomorphic conditions, drainage along the strike is to be expected (Ford, 1971a, 88).

All the remaining "phreatic network" caves, Brackenthwaite Cave, Hale Moss Cave and Hazel Grove Cave, are in areas of low structural dip (Fig. 7.8). Ford (1971a, 91-92) has shown that under these structural conditions, the predominant water intake from the surface must be by joints. These do not usually have sufficient vertical extent to form a main flow route, and if they combine via bedding planes with other joints to reach lower levels they form highly inefficient conduits. On the other hand, in a flat-lying limestone mass, joints and bedding planes are generally continuous to the edge of that mass and therefore to spring points; they also provide straighter, more efficient flow routes. Thus, they will tend to capture any vertical conduits to result in a cave system characterised by discrete joint inlets and sub-horizontal main conduits.

This model can be applied to those "phreatic network" caves found in areas of low dip. In the case of Brackenthwaite Cave and Hale Moss Cave, the passages are developed along joints, either within a single bed or continuous through a number of beds, the joints acting as both inlet and outlet for a flow route. In the case of Hazel Grove Cave, the main passage is developed along a bedding plane, with joint rifts providing hydrological links with the surface.

It is therefore contended that, in most cases, the horizontal nature and limited vertical extent of the "phreatic network" caves provide little evidence for cave formation at the water table as, in general, cave morphology is explicable in terms of structural control. The preferential location of caves at the foot of the hills of the area is partly a result of the frequency with which bedrock cliffs occur in these situations, thereby preferentially exposing cave entrances, and partly a result of the concentration of groundwater flow in resurgence caves at the base of hills.

Nevertheless, as has been mentioned above, Roudsea Wood Cave (north) does not fit the simple picture of structural control. The majority of its passages are strike-aligned and contain a sequence of clastic fill overlain by stalagmite characteristic of caves of the area (see 5.5.1). However, one passage cuts parallel to the dip through the limestone knoll in which the cave is situated. The rock dips at 22^{o} and yet the passage is almost horizontal. The passage form is influenced by the bedding planes exposed in its roof, but these are of little influence on its overall morphology. The passage is developed parallel to known faults within and between adjacent limestone knolls, but no indication of faulting can be seen around the cave. The Martin Limestone in which the cave is developed on the west side of the knoll is extremely closely bedded and jointed, and this may partly account for the apparent absence of structural control on the passage. However, as a result of the dip of the rock, the eastern part of the passage is developed in the overlying Red Hill Beds which are rather more massively bedded and jointed. The only reasonable explanation for the development of the passage involves some external control, for example, a water table or a high sea level; but it is difficult to explain why only one passage in the immediate area should have developed in such a fashion.

In order to compare the altitude of the "phreatic network" caves with that of the controlling water bodies (sea level in the case of those caves in former estuarine situations), wherever possible the altitude of the caves was found by levelling (Table 7.1). In all cases, it can be seen that the caves are developed at heights considerably above the maximum 5.78 m O.D. reached by the mean high-water mark of spring tides during the Post-glacial (see 10.2.1). It was not possible to level Gilpin Bank Caves and Roudsea Wood Caves. However, borings by Gresswell (1958, 90) in the Lyth Valley indicate that marine deposits occur at least 1 m below the present level of the moss at Gilpin Bank, at and above which level the caves are found. At Roudsea Wood Caves, on the other hand, given suitable conditions, flooding may still occur up to and beyond the base of the limestone

Cave	Altitude (m O.D.)	
Levens Cave	8.14	(entrance floor)
	7.89	(entrance floor)
Brackenthwaite Cave	24.76	(entrance floor)
Hale Moss Cave	26.02	(entrance floor 1)
	25.32	(entrance floor 2)
Hazel Grove Cave	27.05	(entrance floor)
Fairy Cave	8.9-11.3	(entrance floor)[1]
Shale Cave	8.9-11.3	(entrance floor)[1]
Yew Tree Cave	9.4-11.8	(cave floor)[1]

Table 7.1 Altitude of the "phreatic network" caves

1 These altitudes levelled from the surface of Nichols Moss, the altitude of which is 8.4-10.8 m O.D. adjacent to the caves (Smith, 1959, 108).

knolls. Nature Conservancy Council records indicate that this has occurred at least eight times in the last 20 years.

7.2.5 The "phreatic network" caves as karst drainage features

There is considerable evidence that most of the "phreatic network" caves are merely fragments of typical karst drainage systems. Some of the passages demonstrate the characteristic circular cross-sections of phreatic tubes formed by water flowing under hydrostatic pressure (Plate 7.1). Elsewhere, in Gilpin Bank Caves for example, phreatically formed aven-like features are developed up to 1.5 m above the general ceiling level of the passages. Such features would require considerable hydrostatic head to enable them to develop and are thus incompatible with the thesis of formation by a static water body.

Other evidence supporting the idea of the caves as typical drainage conduits comes from the presence of bedform erosional features, in particular scallops and meanders, which indicate relatively high flows not expected under static water conditions.

Scallops are diagnostic of solution under a turbulent flow regime. They may be used as indicators of palaeoflow velocities, and hence discharge along cave conduits (see 6.5). Within the "phreatic network" caves, distinct scallop assemblages have been found in Fairy Cave, from which velocities of 10-23 cm s^{-1} and discharges of 0.3-0.6 m^3 s^{-1} can be inferred, and in Roudsea Wood Cave (south), from which velocities of 4-9 cm s^{-1} and discharges of 0.1-0.2 m^3 s^{-1} can be inferred (see 7.4.2).

Other "phreatic network" caves in which scallop-like features have been found are Hale Moss Cave, Brackenthwaite Cave and Hazel Grove Cave (Table 7.2) (Figs. 7.5, 7.6 and 7.7).

Hale Moss Cave : main passage	1-2
Hale Moss Cave : entrance series	~7
Brackenthwaite Cave : main passage	15-30

Table 7.2 Wavelength of scallops (cm) in Hale Moss Cave and Brackenthwaite Cave

In these cases, the scallops are neither distinct nor numerous, and are not in situations where they could be used to reconstruct palaeoflow conditions, but they do give some indication of former flows.

As with scallops, cave meanders may be used to indicate palaeohydrological conditions (see 6.4). Within the "phreatic network" caves, meandering conduit forms occur in Yew Tree Cave, where the fossil meanders indicate a former high-state discharge of $7.5 {+6.8 \atop -3.5}$ $l\,s^{-1}$ (see 7.4.3).

Nevertheless, a number of the caves include distinctive elements not typical of karst conduits. These caves, Brackenthwaite Cave and the entrance series of Hale Moss Cave, are characterised by a dense grid-iron pattern of passages in which almost every joint has been opened by solution. Also noticeable is the location of such networks along the edges of the limestone mass.

According to Palmer (1975, 59), the development of almost all maze caves (sic) is the result of either (i) temporal variations in head and discharge as water moves through a developing limestone drainage system, preventing a fixed passage configuration from developing with respect to flow; or (ii) diffuse, aggressive recharge entering uniformly into all fractures either from an adjacent insoluble formation or from the overlying land surface.

The first condition appears unreasonable as a general explanation. Furthermore, the maze-like components of the caves show little evidence of the scallops, breakdown and localised sediment accumulations regarded by Palmer (1975, 65) as characteristic of such flood networks. Networks formed as a result of this process are rudimentary and of relatively irregular pattern (Palmer, 1975, 65), in contrast to the well-developed networks of the Morecambe Bay systems. Finally, this mechanism provides no explanation for the proximity of the networks to the moss edges.

The second mechanism proposed by Palmer appears more applicable to the Morecambe Bay situation. Although no adjacent insoluble formations occur near the caves, there is considerable evidence of recharge entering the cave uniformly from the ground surface. In many places within the networks, the roof rifts are choked by roots and surface soil, whilst evidence of percolation is provided by the drip pits and linear flutings which form whenever the cave floor has been undisturbed for a few months. However, it is difficult to explain why this process should have resulted only in the development of networks adjacent to the moss edges.

There is evidence that Hale Moss, around which all the true network caves are located, was formerly infilled to a greater extent than at present by peaty sediments (Munn

Rankin, 1910, 256). It is possible that the peat acted as a barrier to water flowing out of the limestone mass, resulting in the build up of head within the cave system and the movement of water from the main passage into adjacent fractures in a form of bank storage. This would have enabled the opening up of a dense network of fractures adjacent to the moss. Alternatively, the networks could be the result of some external process: meltwater from adjacent decaying ice bodies as suggested by Sweeting (1974, 76), solution at the air-rock-soil interface, or the action of adjacent static water bodies as proposed by Ashmead.

The development of tropical karst foot caves at the base of limestone hills by solution at the air-rock-soil interface has been suggested by both Wilford and Wall (1965, 66) and Jennings (1976, 96). However, observation of exhumed tower foot-slopes in tropical areas suggests that this process gives rise to a notch-like morphology (Jennings, 1976, 96), rather than the Morecambe Bay-type network system.

The existence of a static water body in Hale Moss, which would have enabled the networks to form after the fashion proposed by Ashmead, has been supported by Munn Rankin (1910, 155; 1911, 256-257). Munn Rankin found exposures of lacustrine silts, overlain by 0.5 m of shell marl, almost wholly composed of freshwater snails, itself overlain by peat. Unfortunately, Munn Rankin provided no information as to the height of the lacustrine sediments, and exposures of the deposits have been subsequently destroyed by cultivation so it is impossible to reconstruct the former water level of the lake.

Despite the former existence of a water body, the mechanism of cave formation proposed by Ashmead has been shown to give rise to notch-like cave forms in tropical karst and marine situations (Paton, 1964, 145; Wilford, 1964, 9; Trudgill, 1976). The external water body thesis would therefore seem to be inapplicable as an explanation for the true network caves of the Morecambe Bay area.

7.2.6 Conclusions

The "phreatic network" caves of the Morecambe Bay karst have been variously regarded as foot caves resulting from the working of tropical karst processes and water table caves whose form is the result of control by external water bodies. With the exception of Roudsea Wood Cave (north), the origin of which remains enigmatic, there is little evidence to support either of these hypotheses. The caves themselves appear to fall into two distinct categories. Firstly, the true network systems found around Hale Moss, for the development of which a number of hypotheses have been advanced. Secondly, those caves which represent fragments of abandoned karst drainage networks. These caves, including Levens Cave, Roudsea Wood Cave (south) and the caves surrounding Nichols Moss, were obviously placed in the "phreatic network" category solely as a result of their proximity to the moss edges. However, their horizontal nature and limited vertical extent can be explained in terms of structural control; they are located at altitudes above those of the controlling static water bodies, and they contain clear evidence of former periods of relatively high discharge.

7.3 The palaeohydrology of Fissure Cave

7.3.1 Introduction

Fissure Cave (SD45557560) is located within the fossil marine cliffline of Silverdale Cove. Although only 15 m long, the cave contains complex sedimentological sequences, consisting mainly of hydraulically-transported deposits. Equally important, the depositional sequences within the cave can be related, on stratigraphic grounds, to those in other caves in the Morecambe Bay area (see 5.5.1).

By the application of techniques previously only used in surface environments, it is possible to reconstruct a detailed picture of fluctuating environmental conditions within the cave, and to obtain previously unavailable information on the palaeohydraulic environment of caves.

7.3.2 The clastic sediments of Fissure Cave

The majority of the clastic sediments found in Fissure Cave fall within the lacustrine and fluviatile environmental envelopes of the QDa-Md diagram (Fig. 5.1). This interpretation of their origin is borne out by a study of the grain-size curves of the sediments, which closely resemble those of typical hydraulically-transported materials (Fig. 7.9). A small number of the deposits, however, fall either within the glacial envelopes of the QDa-Md diagram, or in transitional positions between the glacial and hydraulic envelopes (Fig. 5.1). It is suggested that the apparently anomalous origin of these deposits is the result of the derivation of all the clastic sediments within Fissure Cave from surface glacial materials. Within typical cave systems,

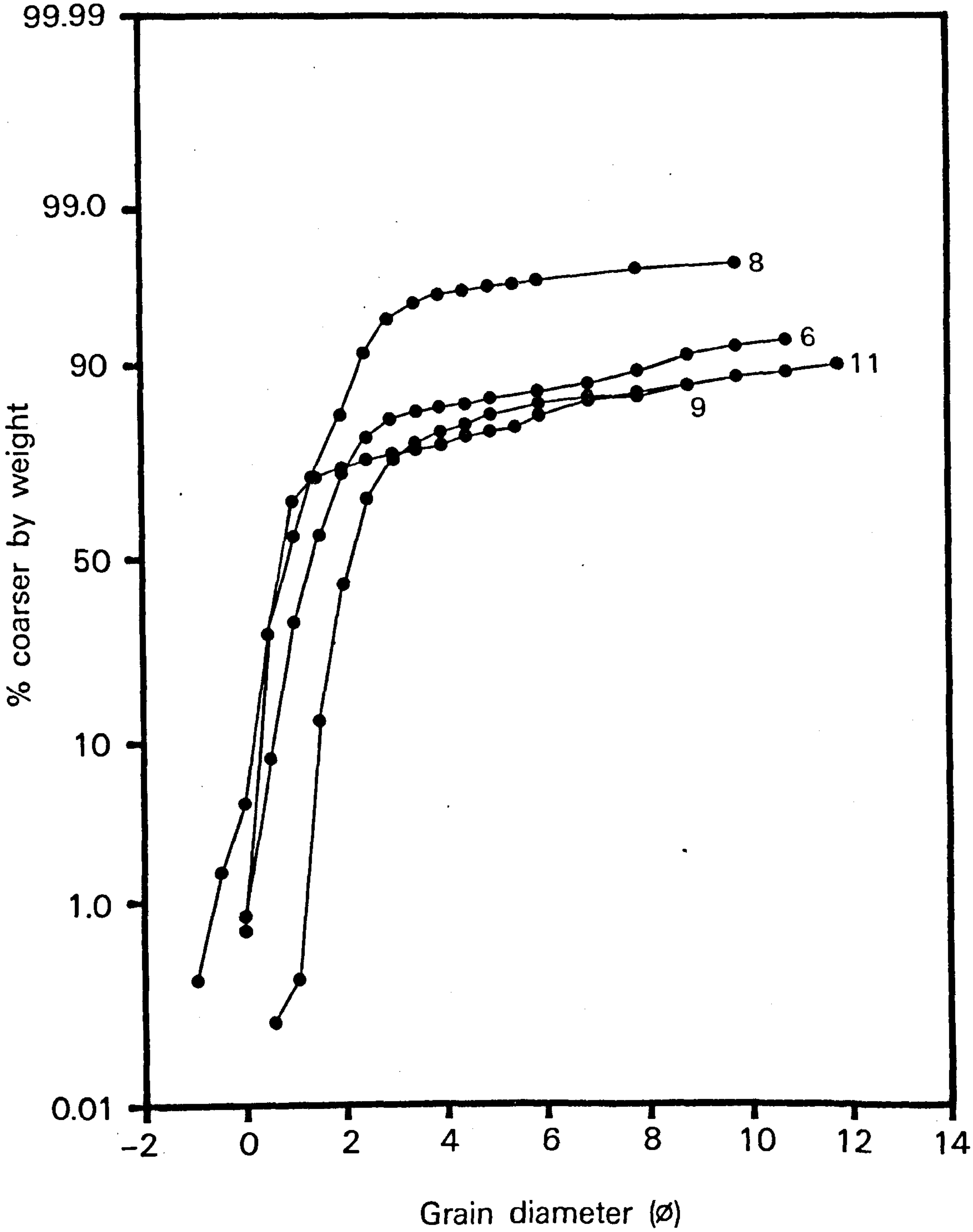


Fig. 7.9 Typical grain-size distribution curves of fluvially transported sediments from Fissure Cave

the bulk of the clastic, non-calcareous deposits are derived from surface sources, often via major hydrological inlets or swallets (see, for example, Davies and Chao, 1959; Bull, 1976). Analysis of surface deposits in the Morecambe Bay area (samples 45, 46, 47, 48 and 49; Water Resources Board, 1970) shows them to fall within the near-mountain source glacial environmental envelope of the QDa-Md diagram (Fig. 5.1). Similarly, field study of the deposits found within the closed depressions of the Silverdale area shows that the depression fill is typically poorly-sorted material containing a wide variety of particle sizes (see 8.4.4), whilst analysis of samples of depression fill from St. John's (sample 38) and swallet fill from Fissure Cave (sample 44) shows the sediments to be indistinguishable from the glacial deposits of the area (Fig. 5.1). The derivation of the closed depression and swallet fill from glacial deposits is also indicated by petrographic analysis. A proportion of the pebble and cobble fraction ($\geq$4 mm) of these deposits is erratic material only found in association in sediments transported by ice from the Lake District (Fig. 7.10).

It is therefore probable that the sediments from which the deposits in Fissure Cave are derived had a similar granulometric distribution to the sediments found on the surface and in the closed depressions of the area. This being so, the non-hydraulically transported deposits in Fissure Cave can be seen to fall into two categories (Fig. 5.1):

(i) Deposits which cannot be differentiated from those found on the surface, i.e. near-mountain source glacial deposits (sample 44).

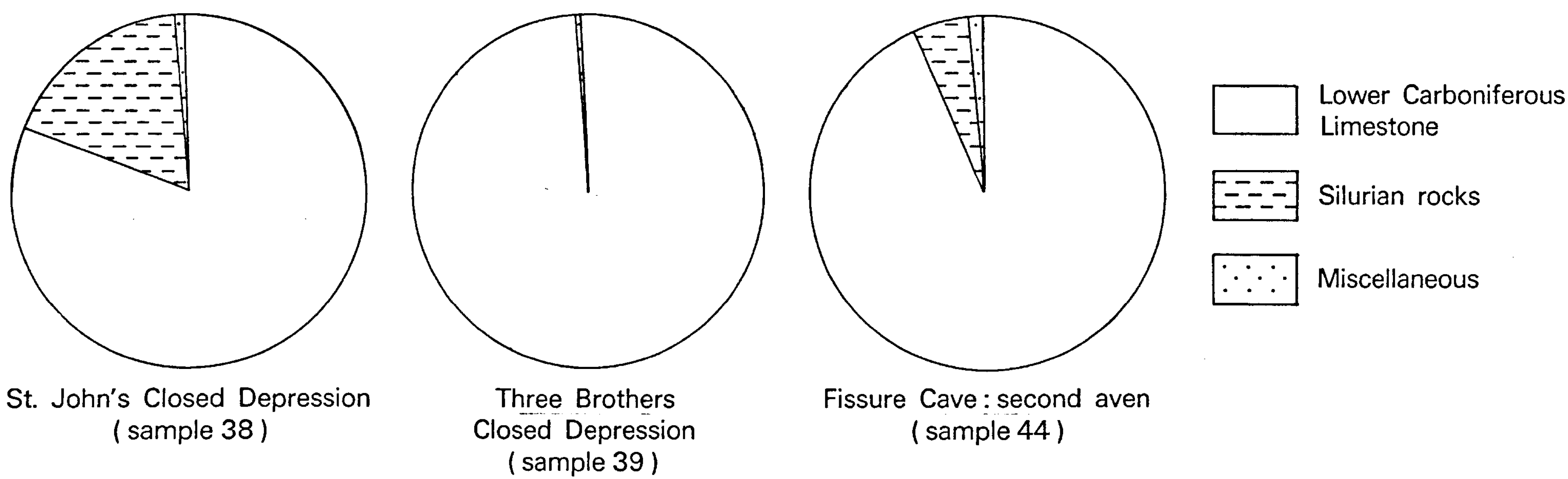


Fig. 7.10 The frequency of occurrence (by weight) of petrographic types within the pebble and cobble fraction ($\geqslant$ 4 mm) of closed depression and swallet fill in the Morecambe Bay area

(ii) Deposits falling within or near the continental ice environmental envelope (samples 16, 30 and 34).

The sediments of category (i) are found within the second aven and form the basal component of the depositional sequences in the cave. The height of these basal deposits decreases further into the cave and away from the second aven. Moreover, the morphology and infill of the second aven are such that a direct surface connection must once have existed through it. Thus, it seems reasonable to postulate that unmodified glacial material entered the cave via the aven to form a cone of material almost filling the cave and blocking both the present entrance and the second aven (Fig. 7.11).

The sediments of category (ii) are unlikely to have been deposited from continental ice, given the location of the area adjacent to the Lake District, a major mountain-ice source. However, it is maintained that a similar grain-size distribution could be achieved by truncating the coarser fraction of a near-mountain source glacial deposit, thereby reducing both Md and QDa, without any reworking of the sediment. This is what would occur if a surface deposit were to be moved through a small inlet into a cave system.

In some cases (e.g. samples 16 and 30), the deposits are transitional between the continental ice and the lacustrine environmental envelopes. Examination of the grain-size curves of these deposits shows that they have forms typical of hydraulically-deposited sediments (Fig. 7.12), but with less steep straight-line coarse-fraction segments. This

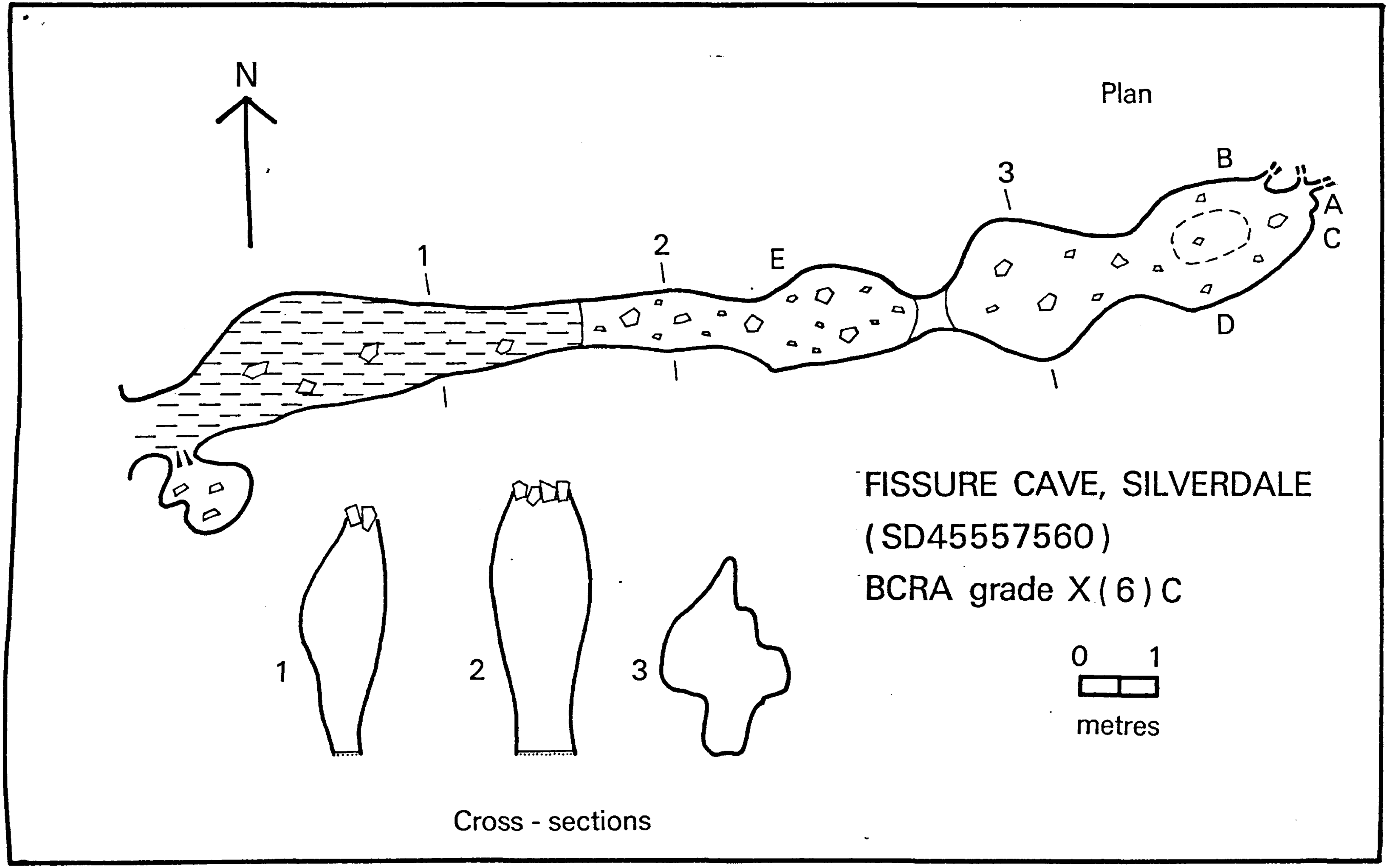


Fig. 7.11a Fissure Cave: plan and extended elevation

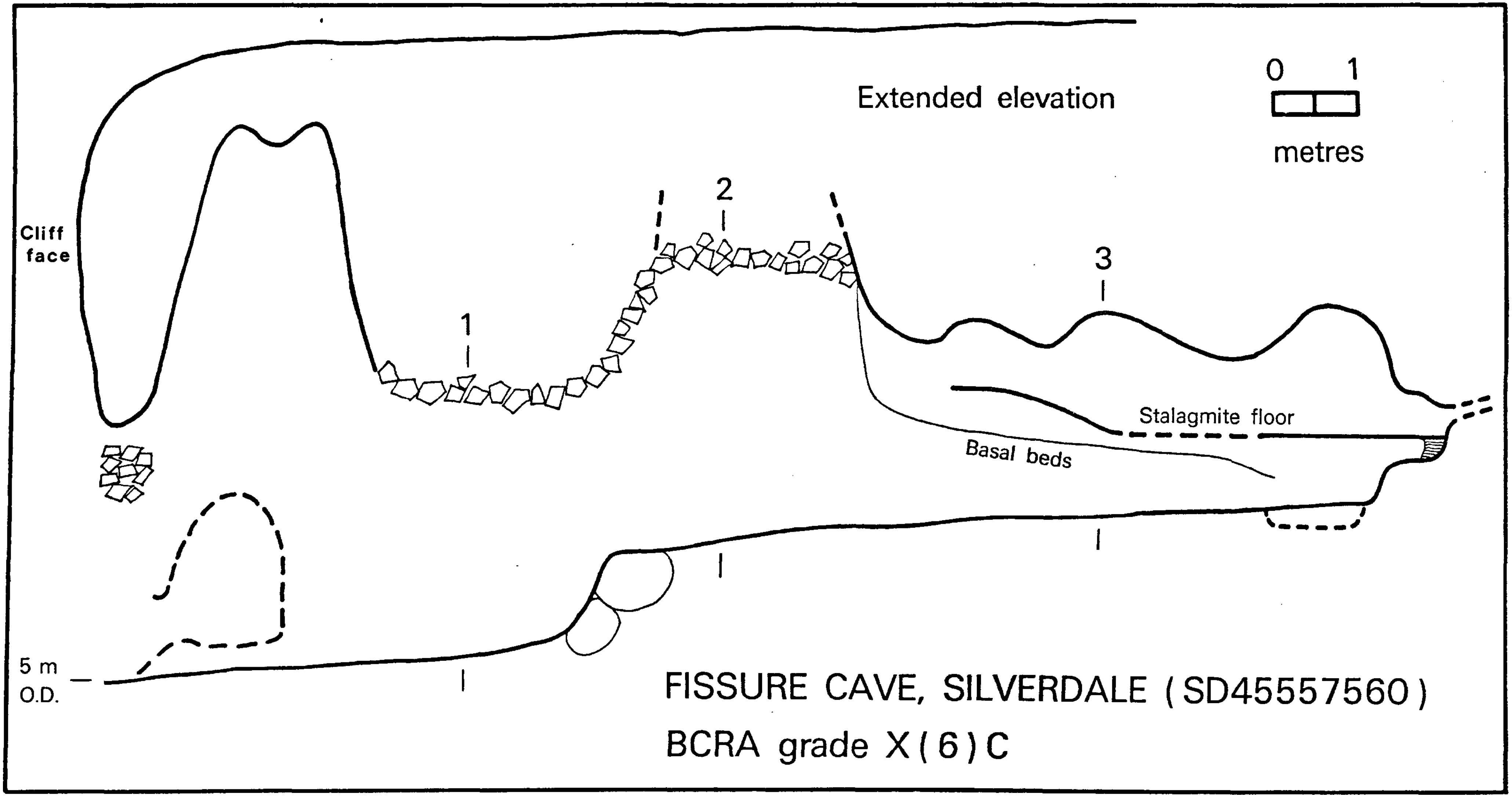


Fig. 7.11b Fissure Cave: extended elevation

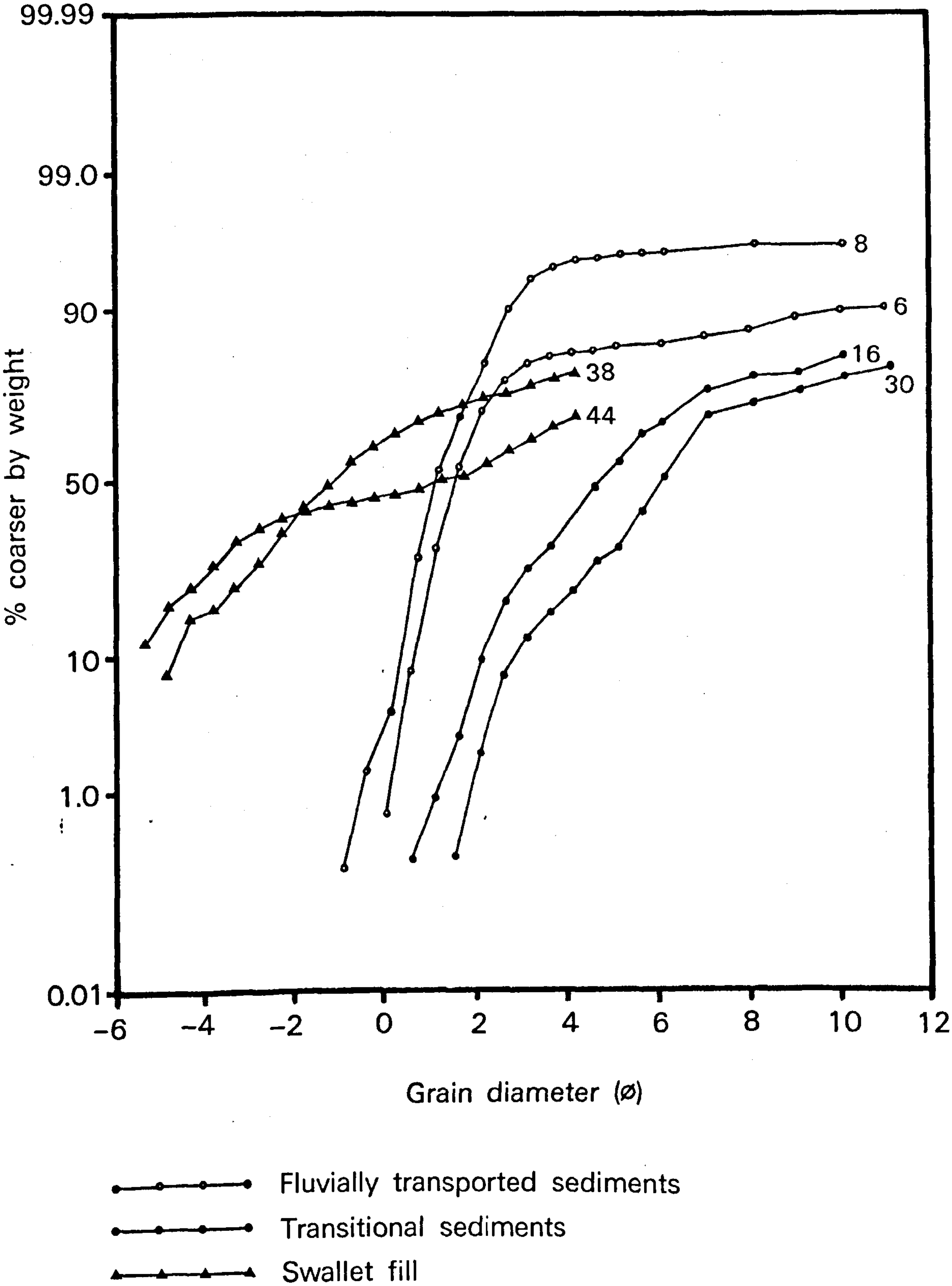


Fig. 7.12 Grain-size distribution curves of swallet fill, fluvially transported sediments and transitional sediments from Fissure Cave

characteristic indicates that, although the initial unsorted glacial material has been considerably reworked, the process has not advanced sufficiently to reduce significantly the fine fraction of the deposits. The same is true of all the sediments clustered in this area on the QDa-Md diagram (samples 15, 32 and 33). All group on the "glacial" side of the lacustrine envelope, indicating incomplete reworking. It is suggested that this is the result of their deposition in less-active water conditions with less energy available to rework the sample, especially over the short distances of flow involved. It is also possible that some of these deposits may have been transported by percolation waters as proposed in 7.3.3.

Thus, the clastic sediments in Fissure Cave represent a range of types between the two end-members of the sequence: unmodified glacial material and fluvial or lacustrine sediments. Most of the sediments in Fissure Cave have been reworked to such an extent that they fall clearly within the hydraulic envelopes of the QDa-Md diagram (Fig. 5.1). Nevertheless, the samples cluster significantly on the "glacial" side of these envelopes. Moreover, a study of individual grain-size curves shows that, although the sediments have forms typical of fluvial and lacustrine materials, they still include a high percentage of clays, suggesting incomplete reworking (see Fig. 7.13).

Evidence from the work of Buller and McManus (1973a, 143-144) shows QDa-Md analysis to be highly sensitive to the effects of hydraulic modification on sediments. They have found that within a few tens of metres of the snouts of

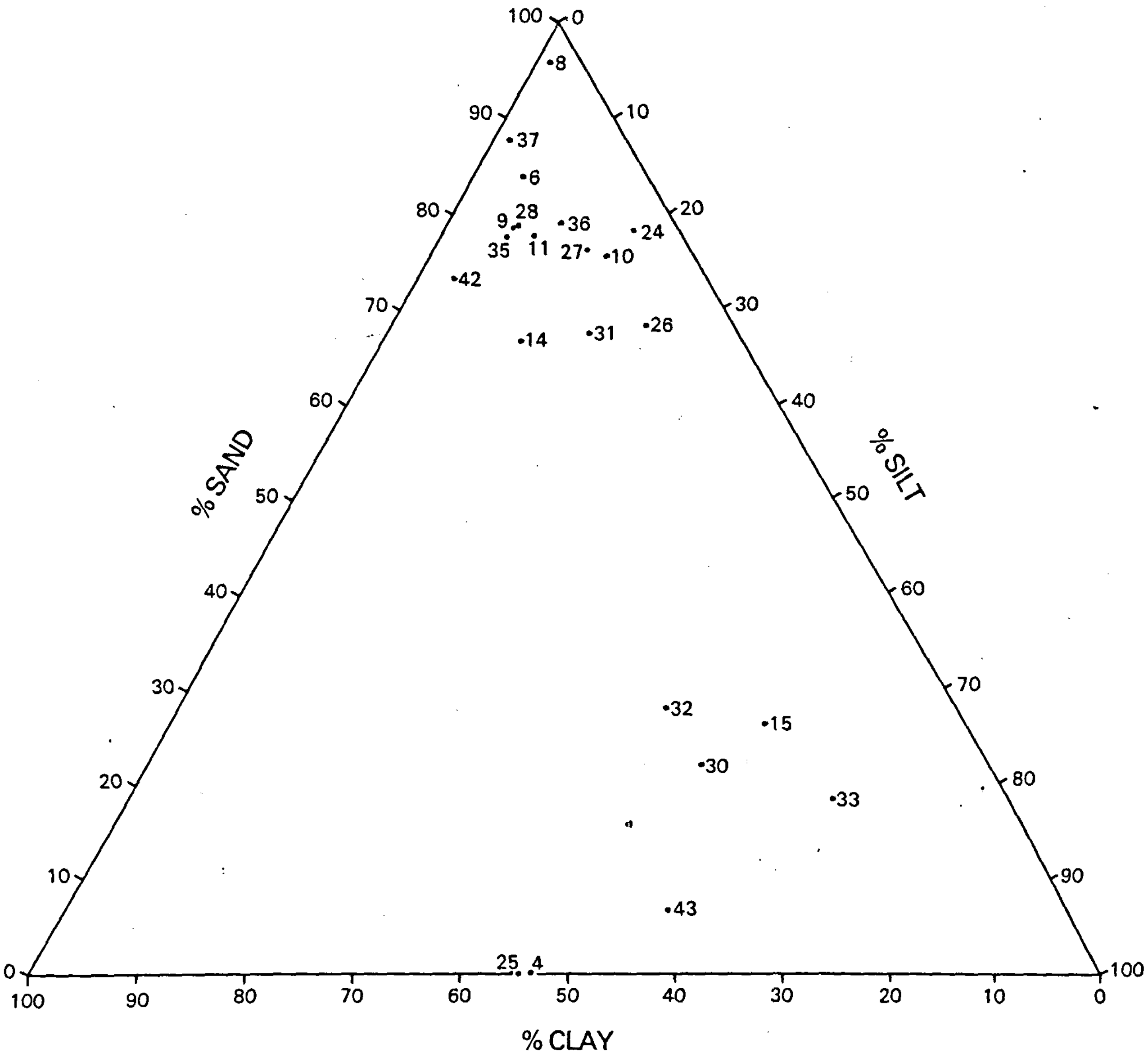


Fig. 7.13 The hydraulically - transported sediments of Fissure Cave : textural analysis

Alpine glaciers, tills are reworked by outwash streams into typical fluvial grain-distributions. The work in Fissure Cave bears this out. Fissure Cave is a short system, probably no more than a few hundred metres long, allowing for small phreatic inlet systems, yet materials derived from unsorted till on the surface are reworked into characteristically fluvial or lacustrine grain-size distributions within that distance.

7.3.3 Environmental interpretation of the deposits found in Fissure Cave

The basal deposits found in Fissure Cave consist of generally unsorted materials which appear to have formed a detrital cone within the cave (see 7.3.2) (Fig. 7.11). These deposits are indistinguishable from surface glacial materials, suggesting that little modification occurred during their movement into the cave. No sign of any sedimentary fabric is discernible within the deposits, which were probably laid down as the result of relatively rapid mass movement.

There is no reason to believe the basal deposits to be derived from anything other than the surface deposits of the last glacial phase in the area. It is difficult to be more specific as to the date of deposition, although apparently overlying the deposits in the roof of the second aven are angular, point-cemented limestone blocks. Elsewhere in the area, similar deposits are interpreted as being of periglacial origin and of probable Late-glacial age (see 8.5). Notwithstanding this, deposition of the basal beds must have preceded the phase of stream activity in the cave which is indicated by the overlying deposits. Under present climatic conditions no stream flow occurs in the cave, and as there is no evidence to suggest a diversion of the former flow, it is likely that the fluvial deposits in the cave were laid down under climatic conditions moister than those prevailing at present. A considerable body of evidence points to the Atlantic period (7120-5100 B.P. according to Shotton, 1977, 26) as the time of highest precipitation during the Post-glacial (see Evans, 1975, 74-77 and references therein); and so the phase of stream activity may be cautiously assigned to this period, with the deposition of the basal beds occurring earlier.

Five sedimentary exposures were studied within the cave (Figs. 7.11 and 7.14), of which four were intensively sampled (Table 7.3), the fifth (section C) being too highly cemented to permit sampling. All the exposures are capped by the remains of a stalagmite floor, which may be used to establish some degree of correspondence between deposits in the cave. However, only in two of the exposures (D and E) are the basal beds exposed at the bottom of the section. The remaining sections are further into the cave where the basal beds, if they exist, would be expected to be below the level of the present cave floor. That this is so is suggested by the nature of the cave floor in the final chamber, which comprises clays, silts, sands and erratic pebbles and cobbles, at least partly disturbed by human activity.

In exposures D and E, the basal beds are overlain by sandy muds (sample 34) which fall into category (ii) of

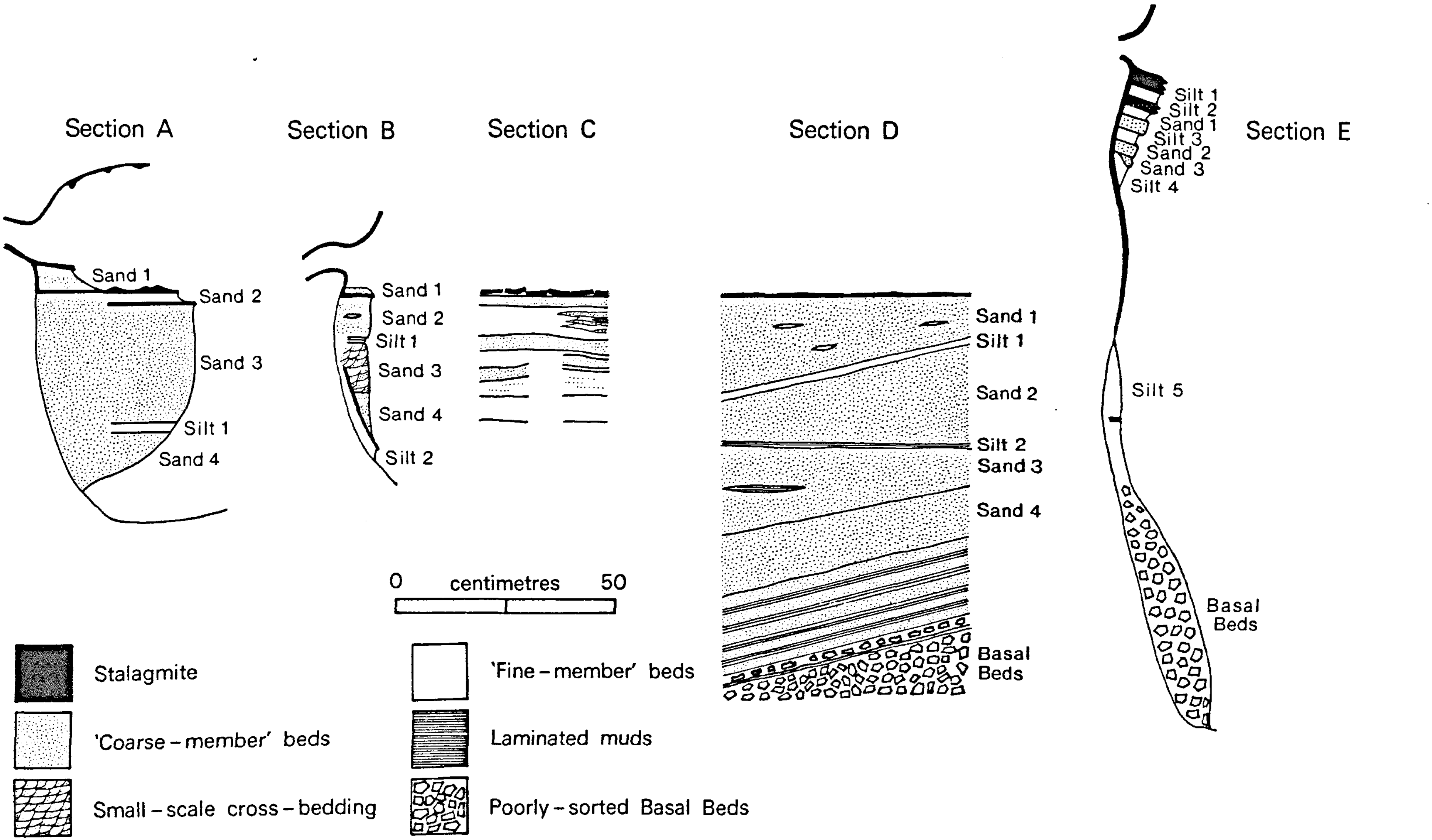


Fig. 7.14 The stratigraphy of sections A–E in Fissure Cave

Section	Bed	Sample Number	D_{50} (mm)	Mean bed thickness (cm)	Mean critical erosion velocity ($cm\ s^{-1}$)	Depositional velocity ($cm\ s^{-1}$)
A	Sand 1	8	0.54	6	29	13
A	Sand 2	36	0.22	4	25	2.4
A	Sand 3	24	0.22	35	31	11
A	Silt 1	43	0.0084	2	?	5×10^{-3}
A	Sand 4	42	0.33	10	29	12
B	Sand 1					
B	Sand 2	10	0.22	10	29	11
B	Silt 1	25	0.0034	2	?	8×10^{-4}
B	Sand 3	6	0.38	10	30	12
B	Sand 3	35	0.38	10	30	12
B	Sand 4	37	0.31	5	26	11
B	Silt 2	16	0.044		not fluvial	
D	Sand 1	26	0.19	15	31	11
D	Silt 1	4	0.0034	2	?	8×10^{-4}
D	Sand 2	11	0.57	15	35	14
D	Silt 2			1		
D	Sand 3	14	0.50	7	29	13
D	Sand 4	9	0.22	10	29	11
E	Silt 1	15	0.027	3	?	5×10^{-2}
E	Silt 2	30	0.017	2	?	2×10^{-2}
E	Sand 1	27	0.14	3	24	1.1
E	Silt 3	32	0.027	3	?	5×10^{-2}
E	Sand 2	31	0.095	3	?	0.57
E	Sand 3	28	0.27	2	24	11
E	Silt 4	33	0.022	5	?	3×10^{-2}
E	Silt 5	34	0.024		not fluvial	

Table 7.3 Fissure Cave : stratigraphy and calculated erosional and depositional velocities.

7.3.2. This suggests either very minor reworking of the basal beds or deposition from the same glacial source accompanied by truncation of the coarse fraction of the grain-size distribution. The close relationship between these beds and the basal deposits is borne out by the lateral transition of Silt 5 in section E to beds containing rounded, erratic pebbles typical of the basal beds. Similarly, in section D, the cemented sandy mud is overlain by a further bed including erratic pebbles in a sandy mud matrix.

During this period there appears to have been considerable percolation into the cave. This is supported by a number of lines of evidence. Within section E, the basal beds are highly cemented and thin layers of stalagmite are interbedded with the sandy muds of Silt 5. Elsewhere, in sections B and E, for example (Fig. 7.14), the cave walls are coated with a veneer of flowstone which predates the later phase of cave fill. All these features suggest that carbonate-rich waters percolated into the cave via fractures in the roof.

Other evidence for percolation is the existence of clastic beds (sample 16) coating the cave wall and underlying the flowstone veneer in section B (Fig. 7.14). These deposits could only have achieved this position by sedimentation from waters percolating down the cave wall and carrying material in suspension. Reference to Fig. 5.1 shows these deposits to be partially-reworked hydraulically and, as far as can be inferred, their grain-size distribution is similar to that of materials transported by slope-wash processes (Ellison, 1945).

Similar processes seem to have been at work in the second aven. Given the direct contact of the aven with the ground surface, a considerable amount of percolation might have been expected. This is evidenced by pockets of finer material left on the walls of the aven and among the coarser, unsorted material.

All the deposits in stratigraphically higher positions in the cave have been hydraulically-transported (Fig. 5.1), implying a considerable change in environmental conditions within the cave. It is possible that permafrost had prevented flow in the cave at an earlier stage, as Ford (1963) suggested was the case for caves in Mendip, but the presence of the percolation deposits mentioned above makes this unlikely. More probably, the initiation of fluvial activity was the result either of the clearing of the conduit feeders, choked by glacial deposits, or of a significant increase in precipitation over the catchment.

The immediate source of the fluvial deposits in the cave seems to have been the phreatic inlets at the end of the cave. This theory is supported by the existence of sedimentary structures in the final chamber which indicate flow out of the cave (see below). There is no accumulation of deposits around the other phreatic inlets in the cave and there are no clastic deposits underlying the flowstone veneer within these inlets, suggesting that they performed, at most, a minor role in sediment input.

Consideration of the hydraulic deposits as a whole reveals a significant change in mean grain-size along the cave passage, deposits tending to be finer with distance from the phreatic inlet (Fig. 7.15). Rarely are coarse-member beds, to use Allen's (1965a, 229) terminology, found beyond the final chamber, implying a progressive decline in the competence of the transporting stream. This is in agreement with the evidence in the cave as, at the commencement of stream flow, the cave would have been effectively dammed by the detrital cone in the second aven, allowing only slow flow of water out of the cave and near-static water conditions within the cave. As a result, sediment-laden flow, previously confined within the phreatic inlet system and flowing under pressure, would have expanded and decelerated upon entering the standing water body, depositing its coarse load and carrying finer material in suspension to be deposited further along the cave. Even without a standing water body, flow would be decelerated due to the increased friction from the greater wetted perimeter.

It is frequently difficult to correlate beds between sections, even over distances as short as 1 m, probably as the result of the complex nature of the sedimentological processes at work in the cave and the small scale of the resultant sedimentary structures and bodies. Within the final chamber, the most complete depositional sequence is that of section D. As described above, the bottom of this sequence consists of the basal beds overlain by reworked or percolation-water borne deposits. From the evidence of the height of the remaining basal beds, it would appear that a basin-like feature existed in the final chamber prior to the initiation of fluvial activity (Figs. 7.11 and 7.14). This is borne out by the overlying deposits in section D, which

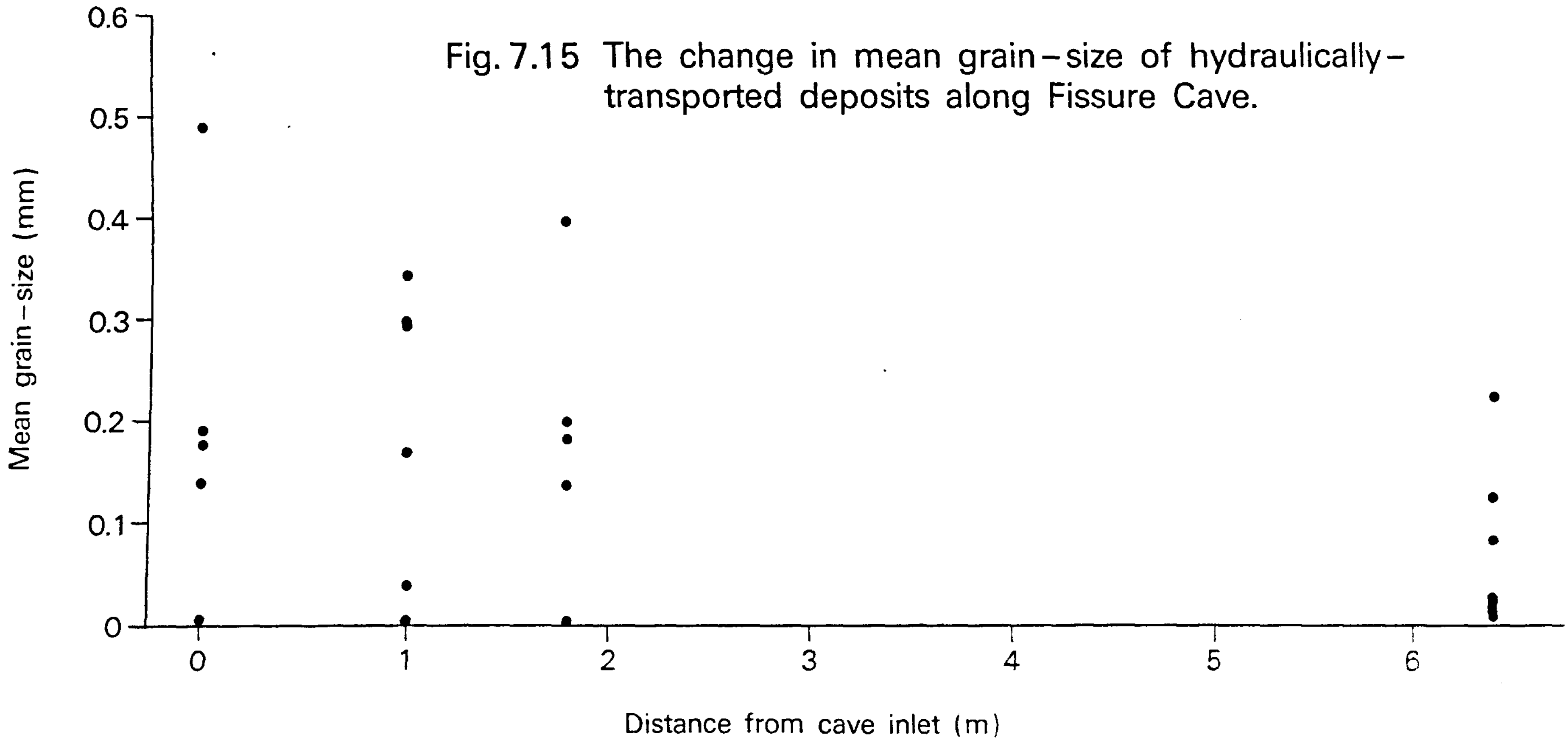


Fig. 7.15 The change in mean grain-size of hydraulically-transported deposits along Fissure Cave.

consist of alternating beds of hard-cemented silts and sands dipping into the cave with an apparent dip of 18°. It is suggested that these sediments accumulated by parallel accretion on the basal surface, a process shown by Reams (1968, 62-64) to be the result of the input of fine sediment into standing water. It is likely that at this stage the cave was still dammed by the debris cone in the second aven.

The overlying beds in section D, Sand 4 and Sand 3 (samples 9 and 14), are of fluvial origin, indicating the replacement of the quiescent environment of the earlier depositional phase by more active conditions. This change is likely to have been the result of the partial removal of the blockage in the second aven, allowing more rapid through-flow of water. The morphology of these beds indicates an infilling of the basin which had previously existed in the final chamber. That the stream depositing these beds was insufficiently large to fill the cave passage is suggested by the lens of bedded muds found within Sand 3. Such horizontally laminated silts and muds are generally interpreted as the product of suspended sediment fallout from slow-moving waters on overbank areas (Collinson, 1978, 48; Taylor and Woodyer, 1978, 263). It is possible to infer from this that at least one episode of channel migration occurred during this phase. It is also reasonable to assume, applying the classic deltaic model of fluvial deposition in static water bodies (Gilbert, 1885), that coarser materials were deposited contemporaneously at the exit from the phreatic feeder and that finer materials were carried in suspension further down the cave. However, no definite evidence of any such deposits remains.

This active phase appears to have been succeeded in section D by a period of waning stream power indicated by the deposition of a thin layer of horizontally laminated muds. This quiescent phase was succeeded by a further period of fluvial activity represented by the accumulation of Sand 2, from which a depositional velocity of 14 cm s^{-1} can be inferred. This bed, and all those above it in the section, can be correlated by a count-from-top procedure with similar beds at similar heights in sections A and B (Fig. 7.14). Thus, both Sand 4 in section A and Sands 3 and 4 in section B can be related to the same depositional event as Sand 2 in section D. All these deposits (samples 6, 11, 35, 27 and 42) fall within the fluvial environmental envelope of the QDa-Md diagram (Fig. 5.1). They indicate depositional velocities of the order of 11-14 cm s^{-1}.

The similarity of the deposits representing this fluvial phase lends credence to the theory that they constitute parts of the same bed. It is likely that the whole bed was laid down in a phase of relatively rapid depositional activity, forming a fan within the final chamber. This is in agreement with the argument of Friend and Moody-Stuart (1972, 36-37) that single-set sandstone deposits of one grain size indicate a short-lived episode of bedload deposition, probably the result of a single flood.

Sand 3 in section B exhibits cross-bedding structures (Plate 7.2). These dip at 30-34^{o} out of the cave above what appears to be an erosional contact plane with Sand 4. The cross-beds provide unequivocal evidence for flow out of the

Plate 7.2 Fissure Cave, section B. The marker is resting on the remains of the stalagmite floor. Beneath the thin, pale-coloured horizon of Silt 1 are the cross-beds of Sand 3.

cave. By using Southard's (1971, 906) depth-velocity-bedform relationship for 0.45 mm sands, since this is closest to the D_{50} value of Sand 3 of 0.38 mm, and knowing the depositional velocity of Sand 3 to be 12 cm s^{-1}, it can be demonstrated that these features are ripples rather than dunes. Southard (1971) found that, at this velocity, ripples tend to develop at flow depths of 9-30 cm, although it is implicit in his work that ripple development may not be confined solely to these depths. For clastic beds of this sort, probably deposited within a single phase, Friend and Moody-Stuart (1972, 36-40) regarded the thickness of the sediment body as an approximate measure of channel depth. In the case of Sand 3 in section B, mean bed thickness is 10 cm, which provides a reasonable match with flow depth as predicated by Southard's plots. Assuming that flow occurred across the whole width of the cave (130 cm at section B), then a flow discharge of $\sim$16 l s^{-1} may be calculated. During the same depositional phase, particles of up to at least 0.57 mm (D_{50}) were transported into the cave (Sand 2 of section D). This indicates a minimum velocity of flow into the cave of 35 cm s^{-1}. The effective cross-sectional area of the phreatic inlet into the cave is $\sim$200 cm^2 (see below), giving an approximate discharge into the cave of 7 l s^{-1}. By the principle of continuity, this should be equal to the discharge at section B. Taking into account the fact that larger particles were no doubt transported into the cave during this phase and deposited near the phreatic inlet, the critical velocity, and hence discharge into the cave, could be considerably greater than estimated.

Nevertheless, even allowing for the errors of the method, there is insufficient agreement between the values to uphold the initial assumption of flow across the whole width of the cave during this phase. This conclusion is supported by the depositional sequence at the same height in section C, in which there is a rapidly alternating succession of muds and fine sands (Fig. 7.14). The muds are horizontally laminated, each lamination having a modal thickness of 0.3-0.4 mm, suggesting periodic deposition of suspended sediment on the interchannel area. The sands represent channel deposits (Allen, 1965b, 127). One part of the section exhibits a complex of mud and fine sand lenses. The muds probably again represent interchannel deposits, whilst sharp-sided sandy beds, wedging out laterally and interbedded with siltstones and mudstones, have been interpreted elsewhere as the product of episodic decelerating flows (Collinson, 1978, 48).

In an alternating sequence of this type, changes from episodes of major suspended load accumulation to episodes of bedload sedimentation must result from lateral movements of the stream channel complex (Friend and Moody-Stuart, 1972, 37-39), thereby explaining the rapid lateral transition between depositional sequences in sections C and D.

It is possible that the sediment-laden flow of this phase temporarily choked the phreatic inlet, for in sections A, B and D the fluvial deposits are succeeded by fine-member beds, suggesting a reduction in both flow competence and discharge throughout the cave. These fine-member beds, which form Silt 1 in sections A, B and D respectively, exhibit a remarkable degree of similarity. Inspection of the environmental envelopes of Fig. 5.1 shows that samples 4, 25 and 43 cluster together at the low energy end of the lacustrine envelope. Although samples 4 and 25 lie just outside the lacustrine envelope, inspection of their grain-size curves shows them to have a well-sorted, typically-hydraulic form (Fig. 7.16). In all three sections these beds are 1-2 cm thick and, in the case of sections A and B, the beds may be traced along the cave wall between the sections. Furthermore, these fine-member deposits all represent a hiatus between two episodes of coarse-member deposition. As would be expected, the flows transporting these materials into the cave deposited rather coarser materials near the phreatic inlet and rather finer materials with distance into the cave. Hence, the D_{50} value of sample 43 of bed A is 8.4 μ m, compared with 3.4 μ m for the corresponding beds in section B and D.

This quiet phase of deposition appears to have been succeeded by a similar phase to the one preceding it. An episode of generally finer coarse-member deposition is indicated in sections A, B and D by Sands 3, 2 and 1 respectively (samples 24, 10 and 26), whilst a similar bed occurs in the same stratigraphic situation and at a similar height in exposure C (Fig. 7.14). Critical velocities of approximately 30 cm s^{-1} can be inferred from these beds. Using bed thickness as an indicator of flow depth, it can be seen that flow depth decreased with distance from the phreatic inlet

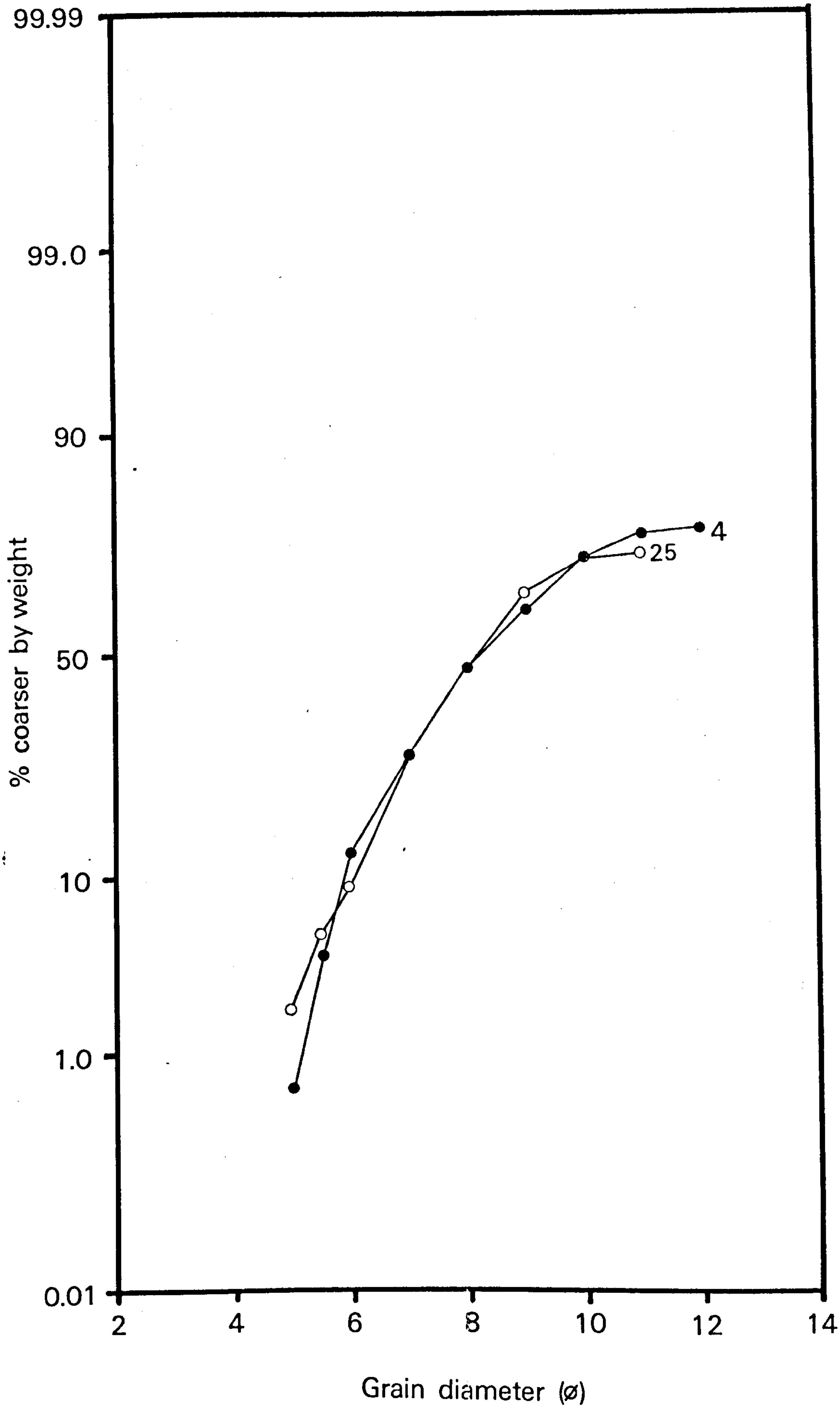


Fig. 7.16 Grain-size distribution curves of Silt 1 (section B) and Silt 1 (section D)

(Table 7.3). This is as would be expected on the basis of the increased cave width. Given a flow depth of 35 cm at section A and a depositional velocity of 11 cm s^{-1}, a discharge into the cave of 29 l s^{-1} can be calculated. At section B, values of 15 cm for flow depth, 125 cm for channel width and 11 cm s^{-1} for depositional velocity give a discharge of 21 l s^{-1}. By the principle of continuity these discharge values should be equal. There is an acceptable level of agreement, well within the errors of the assumptions, suggesting that, at least during this stage, flow was adequate to cover the whole floor of the cave.

In contrast to those of the previous fluvial episode, these deposits are not totally of a homogeneous coarse-member facies. Both Sand 2 in section B and Sand 1 in section D include a number of silt lenses, which may be indicative of channel migration and overbank deposition, as well as mud nodules which may be reworked from the underlying bed.

Immediately after this stage there appears to have been a significant change in the cave environment, with the temporary cessation of flow and the initiation of a phase of stalagmite deposition. The stalagmite appears to have formed a false floor on top of the cave fill throughout the whole of the cave. Above this, a thin veneer of stalagmite coats the roof and walls of the cave, with occasional formations in the roof. The deposition of stalagmite presupposes percolation of water into the cave. At present, percolation into the cave is minimal and, since 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, it seems reasonable to assume that the stalagmite deposition can be related to a phase of wetter, and possibly warmer, conditions during the Post-glacial (see 5.3.3). Stalagmite beds in identical stratigraphic locations elsewhere in the area have been cautiously assigned to the Atlantic stage of the Post-glacial (see 5.3.3 and 5.5.1) and this is in agreement with the age suggested for the fluvial deposits with which the stalagmite is associated in Fissure Cave (see above).

Any attempt to correlate events in the final chamber with those further along the cave at section E is made difficult by the absence of intervening sedimentary exposures. This difficulty is compounded by the gap in the depositional sequence of section E between Silt 5, representing the initial phase of deposition in the cave, described above, and Silt 4. Silt 4 (sample 33) falls clearly into the lacustrine environmental envelope of the QDa-Md diagram (Fig. 5.1). The most acceptable explanation of this bed is that it constitutes the wash-load deposition phase associated with the reduction in flow competence as water entered the cave. On this basis, the horizon can be compared to deltaic bottomset beds.

It is difficult to link Silt 4 to specific events further into the cave. However, on height considerations alone, it is likely that the deposit is related to a later phase of sedimentation in the final chamber. The depositional velocity of Silt 4 was 0.03 cm s^{-1}, which suggests that deposition of suspended material was taking place as water trapped within the cave seeped slowly around the obstruction formed by the debris cove in the second aven.

The lens-shaped coarse-member deposit of Sand 3 (sample 28) probably represents the channel fill of an intermittent flow moving across the underlying muds of Silt 4. Assuming a flow depth of 2 cm, the approximate thickness of the lens, the critical velocity would have been around 24 cm s^{-1}, more than adequate to erode the underlying beds, which were presumably still in a relatively unconsolidated state.

Concave-based, lenticular bodies of this sort are generally thought to be the product of low-sinuosity streams (Moody-Stuart, 1966, 1110-1111; Allen, 1970, 140). The sands within the lens were probably deposited <u>in situ</u> as velocity in the channel fell to about 11 cm s^{-1}, leaving the channel fill dominated by traction and saltation load materials. The channel was ultimately abandoned and buried by the fine-grained, quiet-water deposits of Sand 2 and Silt 3 (samples 31 and 32), which include occasional thin horizontal beds and lenses of mud. These may be either wash-load deposits, as represented by Silt 4, or overbank deposits. The latter appears the most likely explanation in view of the fact that horizontally laminated muds are generally interpreted as the product of suspended sediment fallout in overbank areas (see above). The cycle Sand 3 - Sand 2 - Silt 3 forms a typical upward-fining sequence (Allen, 1965a), which suggests gradual migration of the channel away from the section.

This sequence is repeated in the overlying beds, Sand 1 and Silt 2 (samples 27 and 30). Sand 1 reflects the renewal of stream conditions, the deposits indicating a critical velocity almost identical to that of Sand 3. This is possibly the result of the lateral migration of the stream channel. The channel deposits themselves contain occasional mud lenses and laminations, which may be interpreted as overbank deposits. The overlying bed, Silt 2, may either indicate channel migration and wash-load deposition in the interchannel zone, or deposition from percolation waters as in the case of sample 16 of section B. In either case, deposition was followed by total cessation of flow and the deposition of a stalagmite bed. As in other parts of the cave, the stalagmite appears to have been deposited as a sheet over the cave infill. Within the small chamber at the site of section E, the remains of the stalagmite floor dip into the cave passage, suggesting a degree of incision by the final stream to have flowed in the passage.

This stalagmite bed can be correlated with the phase of stalagmite deposition which occurred throughout the rest of the cave. In some parts of the cave, however, this deposition was intermittently interrupted by episodes of clastic deposition. In section A, the stalagmite is overlain by Sand 2, a uniform, coarse-member deposit (sample 36). Similarly, in section E, the stalagmite is overlain by Silt 1 (sample 15), indicating a phase of quiet-water deposition. It is possible that Silt 1 constituted the suspended load component of the sediment inputs which gave rise to Sand 2 in section A, for the two sets of deposits are at similar heights within the cave (Fig. 7.14). Applying Friend and Moody-Stuart's (1972, 36-40) measure of bed thickness as an indicator of the flow depth of rapidly deposited fluvial sediments, Sand 2 in section A has an approximate flow depth of 4 cm and a channel width of 75 cm. Assuming a mean depositional velocity of 11 cm s^{-1} for Sand 2, discharge into the cave would have been ~3 l s^{-1}. Since the same flow must have been competent to transport the materials of Sand 2 into the cave, given a mean critical velocity of 25 cm s^{-1}, by application of the principle of continuity the effective cross-sectional area of the phreatic inlet tube can be calculated as ~130 cm^2.

Renewed percolation at section A resulted in the cementation of the sandy deposits of Sand 2 and the development of a flowstone floor penecontemporaneously with that in the rest of the cave.

The final phase of clastic deposition in section A, that of Sand 1, appears in many respects to be similar to that which deposited Sand 2. That this deposition occurred after a considerable hiatus may be seen from the highly developed nature of the underlying flowstone floor (Plate 7.3). Both the intermittent nature of the flow and the fan-like morphology of the deposit suggest deposition in a single phase, perhaps as the result of storm flow. Applying the same methods as for Sand 2, a discharge of ~6 l s^{-1} into the cave and an effective phreatic inlet cross-section of ~200 cm^2 are indicated. Similar sand deposits can be found overlying the flowstone floor in section B.

Some percolation must have continued in the cave, since the upper sand deposits in both sections A and B have been cemented. In neither case has the precipitation proceeded sufficiently to deposit a further flowstone bed. It is worth noting, however, that in many parts of the cave

Plate 7.3 Fissure Cave, section A. The marker is resting on the well-developed middle stalagmite bed. Behind the marker is the fan-like form of Sand 1.

(sections B, C and E) the flowstone floor has been cracked and recemented. Although the shattering of stalagmite beds has often been interpreted as the result of frost action (for example, Sutcliffe, 1960, 15), other mechanisms, such as the undermining and slumping of the underlying beds, are possible.

The final episode of sediment removal in the cave remains somewhat enigmatic. The immediate explanation involves a phase of vadose incision, but this would have necessitated highly competent flows to remove the basal deposits in the second aven. Moreover, any flow of this magnitude would have left considerable evidence of trenching in section A, where an uneroded flowstone floor lies directly across the line of flow. An alternative partial explanation is that marine erosion helped remove some of the material in the second aven. The maximum mean high-water mark of spring tides reached during the Post-glacial was 5.78 m O.D. (see 10.2.1), at which height water would have penetrated to just beyond the first aven, perhaps thereby removing some of the debris cone, but not providing a mechanism for the clearance of the rest of the cave.

The only other plausible explanation is that the cave has been partly excavated by man, as there is evidence of the use of explosives in some parts of the cave. This might have been the work of miners prospecting for haematite since a worked vein, Red Rake Mine, is located only a few hundred metres from Fissure Cave, and there is considerable evidence of trial excavations at several locations along the cliffs of Silverdale Shore.

7.3.4. The palaeohydraulic environment as determined from fluvial channel-deposits in Fissure Cave

The palaeohydraulic parameters detailed in 6.3 were calculated for those sediments in Fissure Cave dominated by bedload materials (i.e. Allen's (1965b, 127) channel deposits). In the first instance, bedload sediments were defined as those of $D_{50} > \sim 0.2$ mm since, according to Sundborg (1967, 238), particles of smaller than this tend to go immediately into suspension once eroded. It was subsequently found that, in all cases, u* fell within the range 20.0-1.4 cm s^{-1} regarded by Middleton (1976, 424) as necessary for the bedload transport of sand. A complete list of derived parameters is given in Table 7.4.

Very few measurements have been made even of the basic hydraulic parameters of cave streams. Observations by White and White (1970, 44) suggest that velocities of < 30 cm s^{-1} and flow depths of the order of "tenths of meters" (sic) are typical. These values are similar to those derived from the Fissure Cave sediments, although it might be expected that flows of greater magnitude would be found in larger caves.

Friction factors, such as those of Chézy and Darcy-Weisbach, express the resistance to flow of a fluid by a solid boundary. In alluvial channels, frictional losses depend on (i) the grain-size of the sediment, (ii) the nature and size of the bedforms, and (iii) the channel form. Two friction factors, C and f, were calculated. The Darcy-Weisbach friction factor is recommended by the Task Force on Friction Factors in Open Channels (1963) in slight

Section	Sample Bed	Sample Number	Break point (mm)	ω (cm s^{-1})	u* (cm s^{-1})	D_{50} (mm)	Mean bed thickness ≡ flow depth (d) (cm)	$\bar{u}$ (cm s^{-1})	$C/\sqrt{g}$	f	S	Re	F	P (W m^{-2})	τ (N m^{-2})	θ	Q_{unit} (m^3 s^{-1})
A	Sand 1	8	0.18	1.7	1.7	0.54	6	29	17.1	0.027	4.9×10^{-4}	13385	0.38	0.08	0.29	0.03	0.017
A	Sand 3	24	0.20	2.1	2.1	0.22	35	31	14.8	0.037	1.3×10^{-4}	83462	0.17	0.14	0.44	0.12	0.109
A	Sand 4	42	0.27	3.2	3.2	0.33	10	29	9.1	0.097	1.0×10^{-3}	22308	0.29	0.30	1.02	0.19	0.029
B	Sand 2	10	0.16	1.4	1.4	0.22	10	29	20.7	0.019	2.0×10^{-4}	22308	0.29	0.06	0.20	0.06	0.029
B	Sand 3	6	0.23	2.6	2.6	0.38	10	30	11.5	0.060	6.9×10^{-4}	23077	0.30	0.20	0.68	0.11	0.030
B	Sand 3	35	0.29	3.6	3.6	0.38	10	30	8.3	0.115	1.3×10^{-3}	23077	0.30	0.39	1.30	0.21	0.030
B	Sand 4	37	0.22	2.4	2.4	0.31	5	26	10.8	0.068	1.2×10^{-3}	10000	0.37	0.15	0.58	0.12	0.013
D	Sand 1	26	0.18	1.7	1.7	0.19	15	31	18.2	0.024	2.0×10^{-4}	35769	0.26	0.09	0.29	0.09	0.047
D	Sand 2	11	0.50	7.2	7.2	0.57	15	35	4.9	0.339	3.5×10^{-3}	40385	0.29	1.81	5.18	0.56	0.053
D	Sand 3	14	0.57	8.3	8.3	0.50	7	29	3.5	0.655	1.0×10^{-2}	15615	0.35	2.00	6.89	0.85	0.020
D	Sand 4	9	0.18	1.7	1.7	0.22	10	29	17.1	0.027	2.9×10^{-4}	22308	0.29	0.08	0.29	0.08	0.029
E	Sand 3	28	0.16	1.4	1.4	0.27	2	24	17.1	0.027	1.0×10^{-3}	3692	0.54	0.05	0.20	0.05	0.005

Table 7.4 Fissure Cave fluvial channel deposits : palaeohydraulic parameters.

preference to the Chēzy coefficient. Considering, therefore, only f, it can be shown that a strong relationship exists between frictional resistance and the grain-size of the sediments (characterised by their D_{50} value) in Fissure Cave (Fig. 7.17). It should be pointed out, however, that f and D_{50} are not entirely independent, both being derived from the initial grain-size distribution of the sediment.

Bagnold (1966) has published a list of friction factors for present day rivers. These show that 95% of f-values lie between 0.01 and 0.15, whilst the results of flume studies by Guy, Simons and Richardson (1966) indicate that all friction factors lie between these values. With the exception of samples 11 and 14, the results from Fissure Cave all fall within this range.

The slope of the energy grade line may be regarded as approximately equal to the slope of the stream bed, although this is only truly the case under conditions of uniform flow. In all cases, the computed slope is less than that of the cave itself (5.4×10^{-2}), suggesting an adjustment of the channel slope to different conditions of sediment and water discharge. This is in accord with the picture of sediment aggradation in a static water body proposed in 7.3.3.

The Reynolds number expresses the dimensionless ratio of inertial to viscous forces in the stream and serves as a criterion to distinguish between laminar and turbulent flow. As expected, calculated values of Re are all in excess of 2000, indicating that all the sediments were laid down by fully turbulent flows.

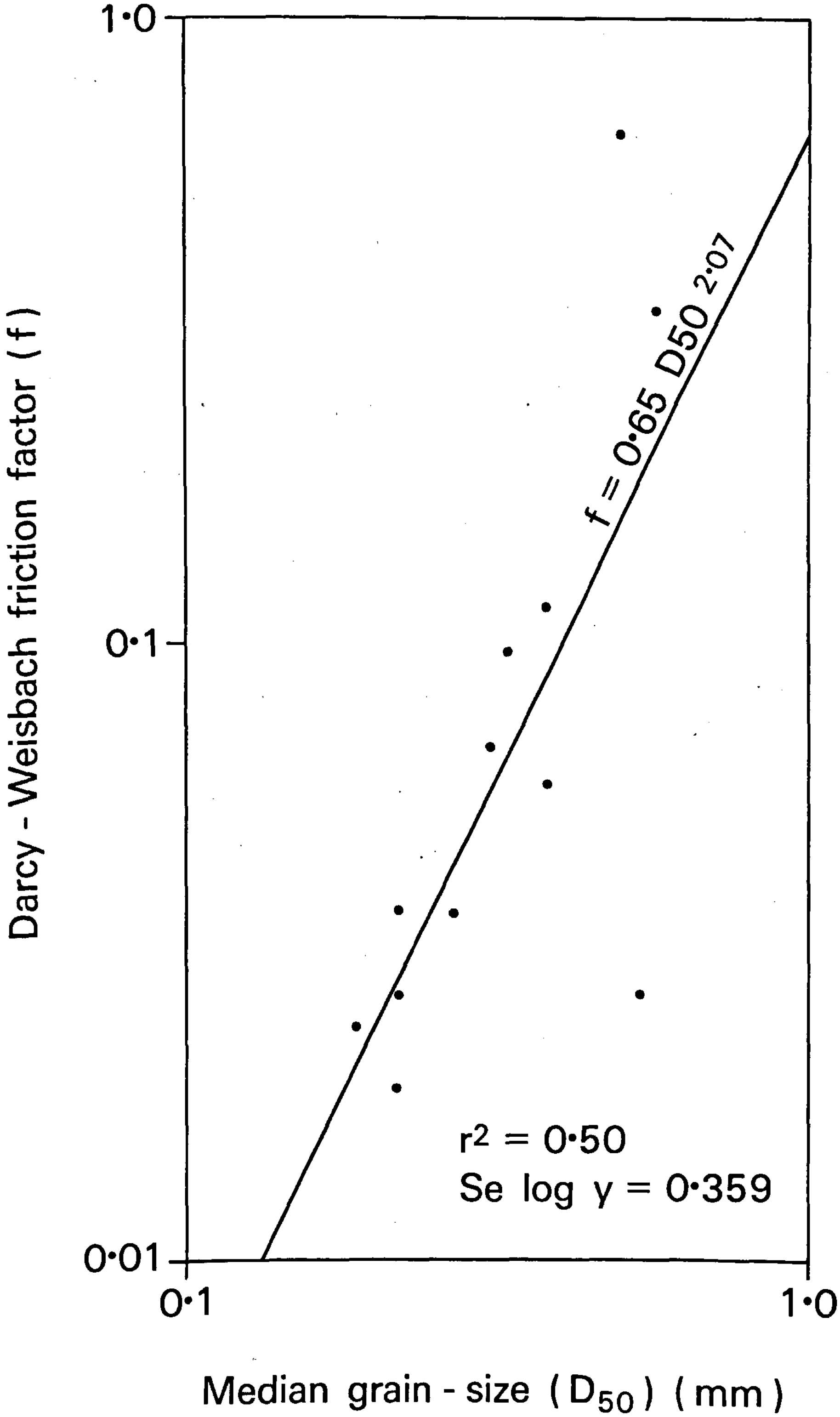


Fig. 7.17 The relationship between the Darcy - Weisbach friction factor and median grain - size for fluvial channel deposits in Fissure cave

The Froude number indicates the ratio between flow velocity and the velocity with which waves can move over the water surface, i.e. the ratio between inertial and gravity forces in the fluid. Thus, the Froude number distinguishes supercritical and subcritical flow according to whether F is greater or less than unity. In all cases, the bedload sediments in the cave were transported by subcritical flows, indicating transport under a subcritical-turbulent flow regime. This is in accord with the general observations of White and White (1970, 44) that most cave streams are characterised by lower-turbulent flow regimes.

The boundary shear stress is the retarding resistance at the channel bed acting against the direction of flow. In open channels this stress tends to be non-uniform on account of the shape of the channel cross-section and the presence of secondary flows within the channel. Thus, the boundary resistance may be written in terms of mean shear stress, even though the shear stress distribution is unknown (Henderson, 1966, 88-90).

To compare τ with values obtained from other rivers, shear stress may be recalculated as a dimensionless parameter (θ) :

$$\theta = \tau/(\rho_s-\rho_f)\ g\ D_{50} \quad \text{(Bagnold, 1966, 9)}$$

Values of θ for 115 rivers, mainly large alluvial streams, are listed by Bagnold (1966, 30-33). Mean $\bar{\theta}$ is 2.24 ($\sigma' = 4.58$) but the range of values lies between 0.14 and 30. As expected, the calculated values of θ for Fissure Cave lie at the bottom end of this range. Saunderson and Jopling (1980) studied a similar-scale environment to that of Fissure Cave, a micro-delta within an esker. They obtained values more comparable to those in Fissure Cave : $\tau = 4.50$ N m^{-2}, giving $\theta = 1.33$.

Stream power can be defined as the rate at which a stream loses energy. It is equal to the product of the weight of water in a reach and the loss of energy head per unit time (Colby, 1964, 24). Stream power has been discussed as a measure of sediment discharge by Cook (1935) and by Bagnold (1960) but, as Cook indicated, its relationship to sediment discharge is not simple. On the basis of the stream power categories devised by Friend and Moody-Stuart (1972, 45-47), the Fissure Cave sediments fall below the sand rivers class (median power ~ 2 W m^{-2}).

Considerable caution must be exercised in the use of these derived palaeohydraulic values, for it is clear that the errors which exist in the original parameters of $\bar{u}$, u* and d can only be compounded by using them to derive further values (see 6.2, 6.3 and 7.3.3). Nevertheless, the values so obtained fall within the expected ranges of a system of this scale, suggesting that the errors resulting from the initial assumptions and simplifications of the method are not significantly high. Furthermore, since so little is known of either karst or small-stream hydraulics, the data provide a basis for comparison with further work.

7.3.5 Conclusions

As in the case of many of the caves in the Morecambe Bay area, none of the sediments in Fissure Cave appear to pre-date the last glacial phase. Even the lowest depositional member in the cave, the poorly-sorted basal beds, seems to have been derived from Devensian glacial deposits. These beds entered the cave via the second aven to form a detrital cone in the cave passage. Overlying the basal beds are deposits which indicate a phase of considerable percolation into the cave followed by a long period of fluvial activity. During the early part of the fluvial period, flow into the cave was probably dammed-up behind the detrital cone in the second aven. This may have resulted in the development of a static water body in the cave, with the result that sediment-laden flow deposited coarser materials at the exit from the phreatic feeder and carried finer materials in suspension further into the cave.

The cave appears to have experienced at least three episodes of alternately quiet and active flow conditions. These were succeeded by a phase of stalagmite deposition, itself punctuated by further episodes of fluvial activity. The channel deposits which characterise the active fluvial phases seem to have been laid down as the result of single flood events, indicating a flashy flow regime, characterised by highly-peaked, short-duration flood hydrographs. The magnitude of these floods was of approximately 5-30 $l\ s^{-1}$, although rarely were flows adequate to cover the whole floor of the cave. Instead, there is evidence of frequent lateral shifts of channels and phases of overbank deposition. The remarkable similarity of these floods, both in terms of their magnitude and their hydraulic behaviour, is probably at least partly a reflection of the controlling influence of the geometry of the phreatic inlet system.

The hydraulically-transported deposits in Fissure Cave appear to have been derived solely from surface glacial deposits. It is therefore likely that a considerable amount of fine material was carried into the cave as suspended load, resulting in highly turbid flows, and this is borne out by evidence in the cave. Nevertheless, the ability of streamflow to rework materials rapidly is demonstrated by the fact that, within a few hundred metres of transport along the cave, the unsorted glacial deposits had been sorted into characteristic fluvial grain-size distributions.

7.4 Palaeohydrological reconstruction from the evidence of bedform flow-features in caves

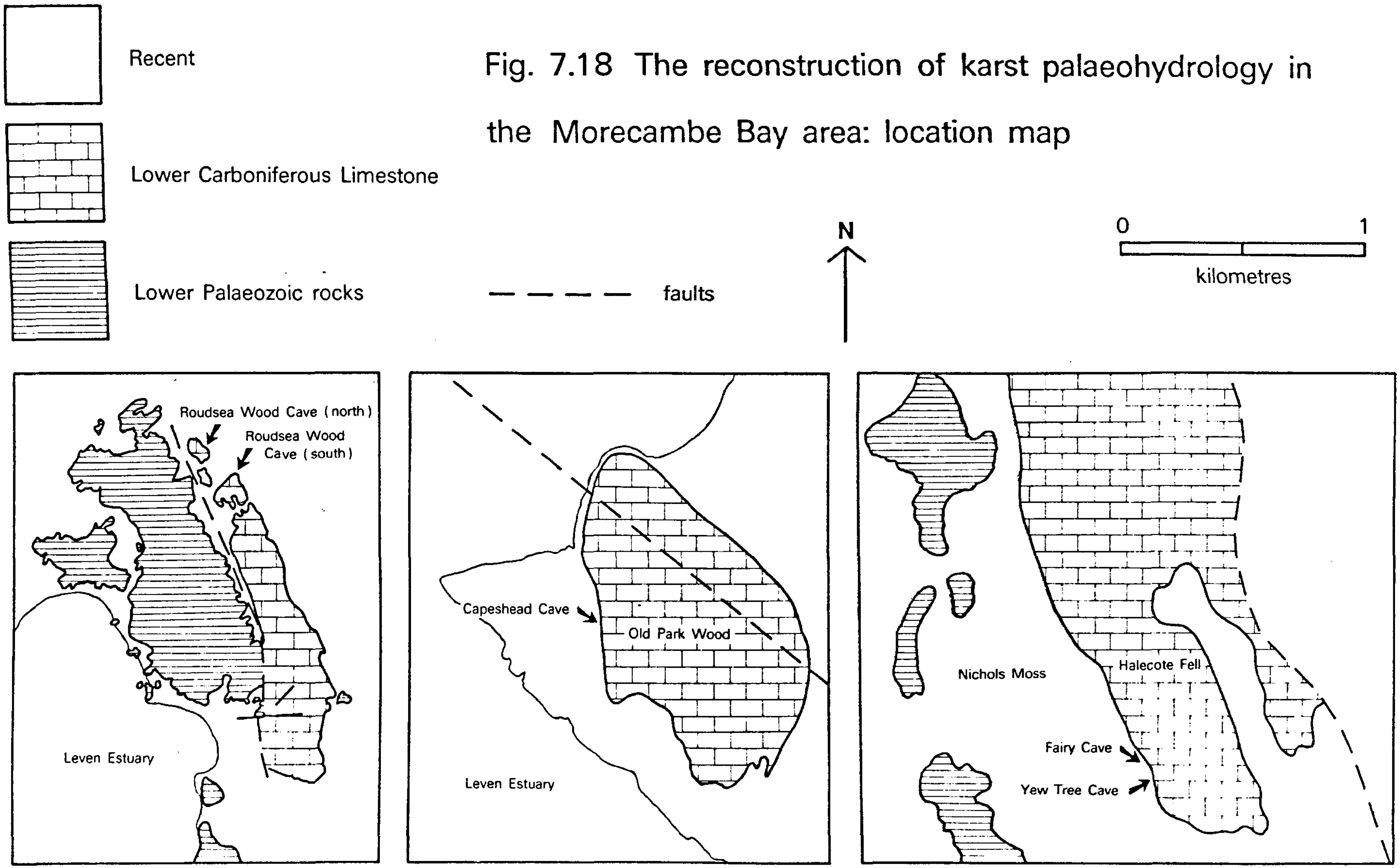
7.4.1 Introduction

Using the methods detailed in 6.4, 6.5 and 6.6, which describe the reconstruction of cave palaeohydrology from bedform flow-features, the former discharge and catchment area of particular drainage routes may be calculated. These methods are applied to all caves in the Morecambe Bay area in which suitable bedform assemblages are found (Fig. 7.18).

7.4.2 Scallops

Scallop assemblages were measured in three caves in the area: Roudsea Wood Cave (south), Capeshead Cave and Fairy Cave (Fig. 7.18). In each of these cases, Curl's site criteria of a regularly-shaped cave cross-section, unchanging for some distance along a straight section of passage (see 6.4.4) were fulfilled.

Fig. 7.18 The reconstruction of karst palaeohydrology in the Morecambe Bay area: location map



The maximum scallop length parallel to the direction of flow was measured for at least 25 scallops. These were measured in the central part of the cave roof in order to minimise the possible influence of boundary effects at the corners of passage cross-sections. The scallops thus measured had been formed under phreatic conditions, hence computed velocities could be used to calculate discharge.

Blumberg and Curl (1974, 743) recommended that the Sauter mean be used to characterise average scallop size. This measure suppresses the importance of smaller, anomalous scallops (see 6.4.4). In all cases, however, it was found that scallop wavelengths were approximately normally distributed and a simple measure of mean length was considered to be adequate.

Mean flow velocities were calculated from the scallop data using the methods of Curl (1974) and Blumberg and Curl (1974), and applying Prandtl's equation for velocity profiles near rough walls (see 6.4.3). B_L was assumed to be 9.4 and u* was calculated assuming Re* values of 1000 and 2220 to take into account the findings of both Blumberg and Curl (1974) and Thomas (1979) (see 6.4.3).

7.4.2.1 Roudsea Wood Cave (south)

$L = 24.0$ cm ($\sigma' = 6.4$ cm)

$a = 2.4\ m^2$

d = width between parallel walls

= 0.70 m

If $Re^* = 1000$, $u^* = 0.5$ cm s^{-1}

$\bar{u} = 4.2$ cm s^{-1}

$Q = 0.1$ m^3 s^{-1}

If $Re^* = 2220$, $u^* = 1.2$ cm s^{-1}

$\bar{u} = 9.4$ cm s^{-1}

$Q = 0.2$ m^3 s^{-1}

Roudsea Wood Cave (south) (SD333825) forms a now-dissected passage which is likely to have once been of considerable horizontal extent (Fig. 7.3). A passage of this size would have formed a significant component of the drainage system of the Holker block.

Scallops were measured in the cave at cross-section 3 where the passage is straight for more than 30 m, over which length the cross-section approximates to a flattened arch (Fig. 7.3). The only disadvantage to this site is the existence of a side-passage which by-passes the main flow route. Nevertheless, the scallop data enable a minimum discharge value to be obtained.

In order to establish the form of the cave floor beneath the peaty infill, a series of probes were put down. By this means a trench at least 1.25 m deep was discovered in the centre of the passage (Fig. 7.3). This was presumed to be the result of later vadose incision of a pre-existing phreatic conduit. It was considered unlikely that such a T-shaped form would have developed under phreatic conditions without erosion acting along a joint in the passage floor. Inspection of the cave roof revealed no corresponding joint and it was concluded that the trench was the result of vadose flow.

The scallops indicate a high-stage discharge of 0.1-0.2 $m^3 s^{-1}$ along the cave (as a result of the passage bifurcation this is probably a minimum value), suggesting a palaeocatchment area of approximately 0.6-1.4 km^2. By contrast, the area of the detached limestone hillock in which the present cave is found is approximately 0.01 km^2. It is obvious that the cave must once have drained a far larger area and, hence, that it must predate the present dissection of the Holker block. The scallops in the cave indicate a palaeoflow direction from the east. It is therefore likely that Roudsea Wood Cave (south) drained the limestone area that must once have existed between the present outcrop and the Ellerside fault to the east. It seems most reasonable to attribute the removal of this limestone to glacial scour by ice from the valleys of the south central Lake District. On this basis, the Roudsea Wood drainage system is of at least last interglacial age.

7.4.2.2 Capeshead Cave

$L = 32.4$ cm ($\sigma' = 11.9$ cm)

$a = 6.0\ m^2$

d = width between parallel walls

$= 3.20$ m

If $Re^* = 1000$, $u^* = 0.4\ cm\ s^{-1}$

$\bar{u} = 4.4\ cm\ s^{-1}$

$Q = 0.3\ m^3\ s^{-1}$

If $Re^* = 2220$, $u^* = 0.9\ cm\ s^{-1}$

$\bar{u} = 9.7\ cm\ s^{-1}$

$Q = 0.6\ cm\ s^{-1}$

Capeshead Cave (SD33337814) is the only known cave found in the detached block of limestone known as Old Park Wood (Fig. 7.18). The cave consists of a single passage approximately 25 m in length, of near-uniform cross-section, which closes down to a sediment choked phreatic inlet. Traces of scalloping and stalagmite deposition along the cliff in which the cave is found indicate that the cave formerly extended at least 10 m beyond the present entrance.

The scallops (Plate 6.1) indicate relatively slow-moving flow and a high-stage discharge of 0.3-0.6 $m^3 s^{-1}$, suggesting a palaeocatchment area of approximately 2-5 km^2. This is larger than the present area of Old Park Wood, which covers about 1 km^2. Nevertheless, given the expected errors of the calculations, the two areas are reasonably similar and it is possible to conclude that Capeshead Cave probably formerly functioned as the drainage outlet for the whole of Old Park Wood. This conclusion is supported by the absence of further resurgence caves around the edge of the block, and the presence of only two small springs at the base of the cliffs at the southeast corner of the block.

Capeshead Cave appears to be of at least last interglacial age (see 5.5). Thus, it is possible that the catchment area of Old Park Wood has been reduced by glacial scour during the Late Pleistocene, partly explaining the apparent overestimation of palaeocatchment area by the methods given above. It is also possible that glacial overdeepening of the adjacent Leven Estuary has resulted in the lowering of the resurgences level of the catchment, with the result that present groundwater resurges in the estuary, as Brown (1973) has shown to be the case in the Silverdale area.

7.4.2.3 Fairy Cave (Nichols Moss)

$L = 14.2$ cm ($\sigma' = 4.4$ cm)

$a = 2.8\ m^2$

d = width between parallel walls

$= 1.50$ m

If $Re^* = 1000$, $u^* = 0.9\ cm\ s^{-1}$

$\bar{u} = 10.2\ cm\ s^{-1}$

$Q = 0.3\ m^3\ s^{-1}$

If $Re^* = 2220$, $u^* = 2.0\ cm\ s^{-1}$

$\bar{u} = 22.5\ cm\ s^{-1}$

$Q = 0.6\ m^3\ s^{-1}$

Fairy Cave (SD434827) consists of a single abandoned passage located at the foot of Halecote Fell (Fig. 7.4). The phreatic-vadose sequence indicated by the cave morphology was succeeded by an episode of sediment infilling. The fill was subsequently incised by stream erosion, possibly by the present stream which now flows into the cave, leaving terrace-like banks of deposits on either side of the stream. In the stream bed, the bedrock floor of the cave has been re-exposed.

The size of the cave passage suggests that Fairy Cave was a conduit of more than local significance in the hydrology of Halecote Fell. However, as all waters flowing from Halecote Fell and resurging in Nichols Moss must flow up-dip, this means that Fairy Cave sumps after 120 m in what is probably part of a series of structurally-controlled phreatic loops described by the cave (Fig. 7.4).

Although scallops were measured in the central part of a long, straight passage of near unchanging cross-section (cross-section 2 of Fig. 7.4), the cave is by no means ideal for the determination of flow velocity, for it exhibits a far from simple passage cross-section. Nevertheless, it was considered that the scallops would give at least an indication of former flow conditions in the cave.

As in the case of Capeshead Cave, the calculated high-stage discharge indicates a palaeocatchment area of 2-5 km^2. This seems an overestimation when compared with the $\sim$0.5 km^2 of limestone which constitutes the southern end of Halecote Fell. Nevertheless, given the expected errors of the calculations, the two areas are reasonably similar, supporting the contention that Fairy Cave played a major role in the former drainage of the Halecote Fell area.

7.4.3 Cave meanders

The only definite evidence of hydrologically-formed cave meanders in the Morecambe Bay area comes from Yew Tree Cave. Yew Tree Cave (SD434826) is found in the low cliff-line which comprises the western edge of Halecote Fell (Fig. 7.18). The cave is aligned sub-parallel to the cliff and consists of a sinuous, abandoned vadose passage developed along the strike of the rock. At each end of this passage slab breakdown (Davies, 1951) has occurred, probably along the former extension of the sinuous passage (Fig. 7.4).

A possible control on the plan form of meandering passages is joint alignment. Yew Tree Cave appears to have been formed by waters flowing along the strike of the rock which subsequently cut down from the initial opening of the bedding plane through the underlying beds. Hence, the initial line of weakness became less significant and no longer controlled cave alignment. This thesis is supported by the flat bedding plane roofs found along the passage (Fig. 7.4), rather than the roof rifts associated with joint-aligned passages. In certain parts of the passage, higher-level meander remnants can be found. Beneath these, the later meanders have incised a lower-level route on a different course, indicating the probable absence of joint-control, at least in the alignment of the lower sinuosities. Furthermore, as Table 7.5 demonstrates, the measured meander parameters of wavelength, channel length and channel width, as well as the indices of sinuosity and wavelength-width ratio, are similar for each meander, suggesting hydraulic rather than structural control of meander form.

A horizontal survey of the meandering section of Yew Tree Cave was made to B.C.R.A. Grade 5 (Ellis, 1976, 2-7) (Fig. 7.19). The passage oscillations were studied using the methods developed by Ongley (1968) for dealing with asymmetrical meanders. The cave thalweg, defined as the central line of the channel, was divided into segments of equal length. An interval of 0.25 m was chosen, following Speight's (1965, 13) suggestion that the interval be equal to the mean unvegetated channel width. This was found to be sufficiently sensitive to changes in thalweg direction. Positive and negative values were ascribed to each segment, depending on its orientation with respect to the previous

Minus-plus oscillations (metres)

Oscillation	Lw	Lc	W	Lc/Lw	Lw/W
1	2.75	3.25	0.30	1.18	9.17
2	2.05	2.25	0.25	1.10	8.20
3	2.42	2.50	0.30	1.03	8.07
Mean	2.41	2.67	0.28	1.10	8.48

Plus-minus oscillations (metres)

Oscillation	Lw	Lc	W	Lc/Lw	Lw/W
1	2.90	3.00	0.30	1.03	9.67
2	2.50	2.75	0.25	1.10	10.00
Mean	2.70	2.88	0.28	1.07	9.84

Overall mean values (metres)

Oscillation	Lw	Lc	W	Lc/Lw	Lw/W
	2.56	2.78	0.28	1.09	9.16

Table 7.5 Meander morphometry: Yew Tree Cave

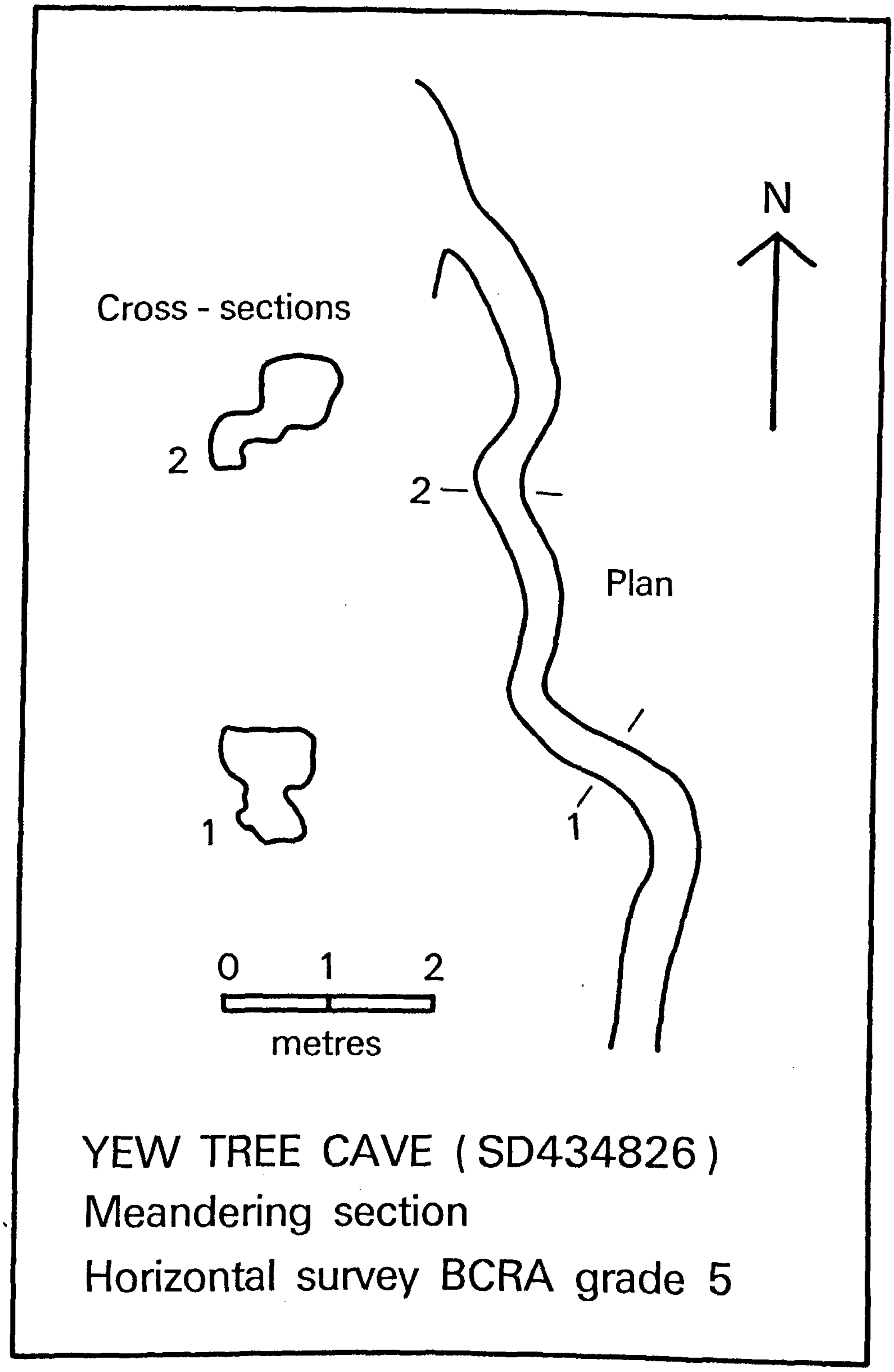


Fig. 7.19 The meandering section of Yew Tree Cave

segment, positive values being assigned to segments angling to the right and negative to those angling to the left. Two complete and successive series of segments with positive and negative angles comprise one wave. Once the thalweg has been divided in this fashion, it is possible to measure, for each oscillation, wavelength (Lw), channel length (Lc), channel width (W), sinuosity (Lc/Lw) and wavelength-width ratio (Lw/W).

To check whether the order of positive and negative segments affected the result, the same measurements were made dividing the thalweg on the basis of a negative followed by a positive component. Similar values were found and a mean set of values taken (Table 7.5).

The results of the analysis were compared with the results of all other studies of cave meander morphology (see 6.4.1). The values of W and Lw obtained in Yew Tree Cave were within one standard deviation of the calculated regression line, suggesting a close agreement with the morphological relationship found elsewhere.

Substituting the mean meander wavelength found in Yew Tree Cave into the relationship derived in 6.4.2, a high-stage discharge for the stream occupying the cave of $7.5 ^{+6.8}_{-3.5}$ l s^{-1} may be inferred. From this, a palaeo-catchment area of ~ 0.03 km^2 may be calculated, suggesting that Yew Tree Cave was of no more than local significance in the drainage of Halecote Fell, certainly when compared to the role played by nearby Fairy Cave (see 7.4.2.3).

7.5 The regional palaeohydrology of the Morecambe Bay karst

7.5.1 Previous work

The regional palaeohydrology of the Morecambe Bay karst has been studied in detail by Ashmead (for example, 1969a; 1974a; 1974c), who regarded cave development in the area as having been directly controlled by the heights of former water tables in the limestone. These water tables were considered to have been closely related to the heights of former regional base levels identified by Parry (1958; 1960a) throughout the whole of the southwestern Lake District (see 8.2). On the basis of this relationship, Ashmead was able to provide a relative chronology for hydrological development in the area.

Ford (1971a, 91-92), however, has demonstrated that levels of cave development can be directly related to base levels only in situations where the limestone is flat-lying (that is, where the dip is less than 5°) or where the fissure density of the rock is so great that a high hydraulic conductivity results and a water table type cave can develop. In the Morecambe Bay region neither of these criteria holds over a wide area. The limestone is structurally complex and in few places is its dip as low as 5°, whilst the rock is relatively massively bedded and jointed, beds less than 0.3 m thick being abundant only in the Martin Limestone lithological division (Nicholas, 1968).

Yet, according to Ashmead (1969a; 1974c), the caves of the Morecambe Bay area show a close relationship with former water tables in the limestone. In order to test the strength of this relationship, the altitude of near-horizontal cave passages within the area (i.e. those assumed to be related to former water tables (Sweeting, 1950, 74-75)) was compared with the expected altitudes at which caves would be found assuming an even distribution of caves over the area. All known caves occurring at heights at or above 15 m O.D. in the Morecambe Bay karst east of Low Furness were included in the study (Holland, 1967; Lancaster Cave and Mine Research Society Records). No weighting was given to passages of different lengths. This was felt to be reasonable given the generally small and discrete nature of the cave within the Morecambe Bay area.

The caves were grouped into altitudinal class divisions of five metres to preserve the variations in cave frequency with height which might have been masked by greater grouping. An interval of five metres also had the advantage of being less than that between any of Parry's erosion levels. The expected distribution was found by measuring the proportion of the limestone outcrop of the region found within each altitudinal class and then calculating the number of caves expected within each class on the basis of an identical number of caves per unit area of limestone (Fig. 7.20).

In order to compare the observed and expected distributions, the Kolmogorov-Smirnov one-sample test was applied to the data. The hypothesis of a significant difference between the two distributions was rejected, even at the 80% confidence level, suggesting that the distribution of caves with height conforms closely to that expected, given an even distribution of caves over the area, and demonstrating

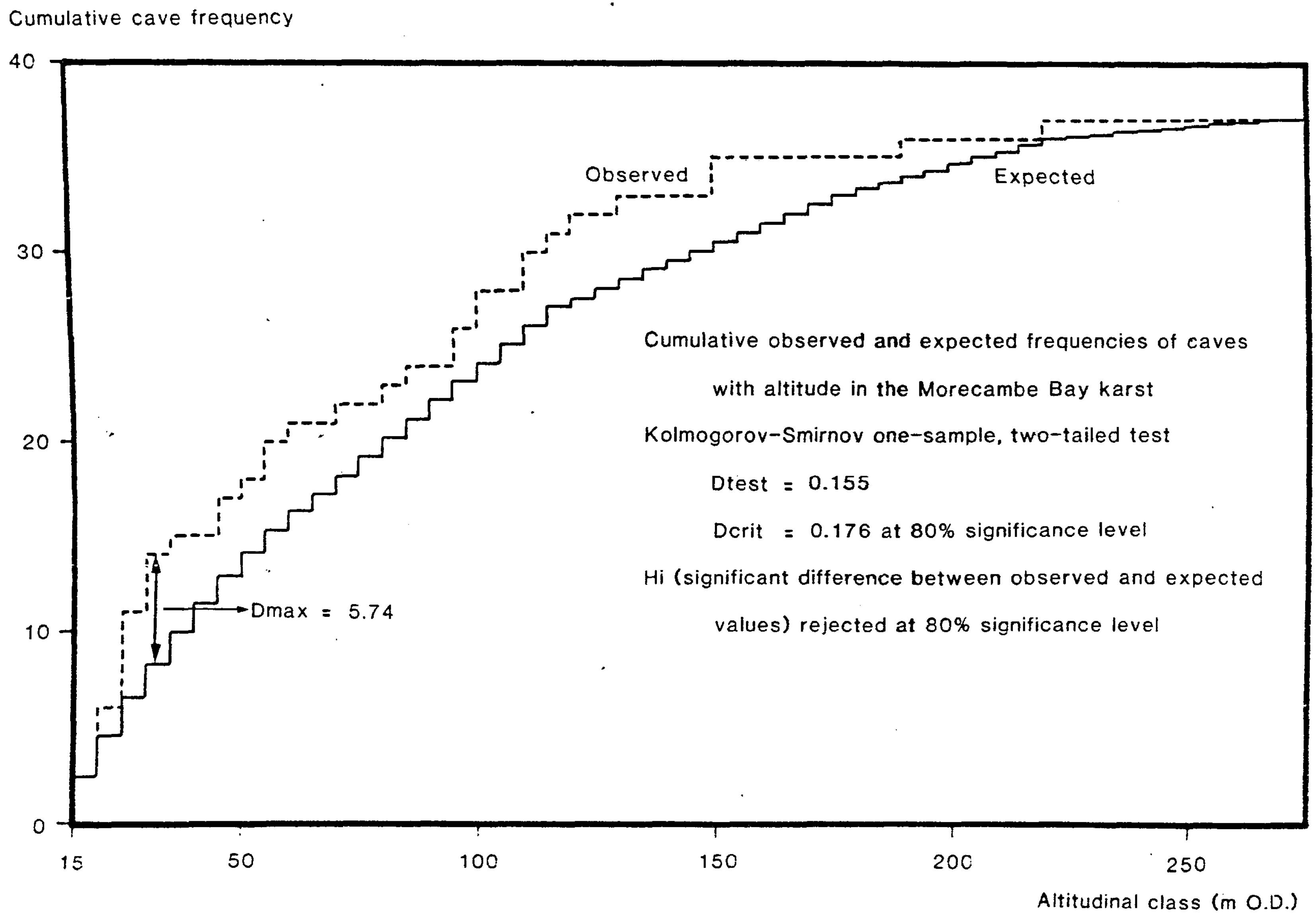


Fig. 7.20 Cumulative observed and expected frequencies of caves with altitude in the Morecambe Bay karst

the lack of a significant height control on the pattern of cave occurrence.

These results suggest that Ashmead's interpretation of the hydrological development of the area should be revised, and that drainage history cannot be easily related to the sequence of base level changes proposed by Parry. There are a number of possible explanations for this. At no stage have the resurgence levels of each of the limestone blocks necessarily been the same. Furthermore, the fact that karst development has occurred over a long period of time in the Morecambe Bay area has no doubt resulted in a complex superimposition of drainage systems. Finally, the results of work in areas of flat-lying limestones have tended to present an oversimplified picture of the effects of height control on cave development; the phreatic loops characteristic of dipping limestones, such as those found in the Morecambe Bay area, present a more complex pattern of cave levels than that found elsewhere.

7.5.2 The palaeohydrological development of the Morecambe Bay karst

It is apparent, even from a cursory study, that the Morecambe Bay karst displays considerable evidence of drainage development under hydrological conditions which differ from those of the present. Former drainage routes, often of regional significance, now lie abandoned (see 7.4), whilst elsewhere only fragments of once-extensive systems remain. The caves of the area, both active and abandoned, may be classified into three categories: vadose inlets, phreatic passages, and resurgence caves (Table 7.6). In most cases, the phreatic passages have also experienced a later vadose phase, but frequently this cannot be identified with certainty because of the presence of cave fill. Such a classification is necessarily oversimplified since it is based mainly on morphology and since any one cave system may exhibit all three genetic components. Nevertheless, the classification may be a useful aid to the study of cave development in the area. Thus, by considering the distribution of the various cave types in relation to the structure (*sensu lato*) of the area, it may be possible to advance a simple model to explain drainage development in the region.

The contemporary karst landscape consists principally of a series of fault-bound, cuesta-like limestone blocks (Fig. 1.1). Any impermeable cover which formerly existed on these blocks has been removed and, as a result, there is little integration of surface flow. It is difficult to estimate the date of removal of the cover, although Sweeting (1974, 74) believed that the limestones may have been fully exposed by the beginning of the Quaternary. Nevertheless, there is little direct evidence to support this view. The hydrological effect of an impermeable cover on the limestone blocks would have been to concentrate surface flow and encourage the development of vadose inlet systems at the fringe of the impermeable beds. The existence of a number of vadose inlets within the Morecambe Bay area (Table 7.6) suggests that the removal of such an impermeable cover may not have taken place until relatively recently. On the other hand, it is possible that such vadose feeders

Abandoned "phreatic" caves	Active "phreatic" caves	Abandoned resurgence caves	Active resurgence caves	Abandoned vadose inlets	Active vadose inlets
Allithwaite Cave	Hawesbridge Cave	Badger Hole	Beck Head	Crag House Pots	Burton Well Cave
Arnside Cave	Heron Corn Mill Cave	Barrow Scout Cave 1	Cunswick Scar	Dunald Mill Flute	Dunald Mill Hole
Backlane Quarry Cave	Otter Holes	Barrow Scout Cave 2	resurgence	Haverbrack Bank	Lupton East
Broca Hill Caves		Capeshead Cave	Helsington	Pot	Swallets
Cow Close Cave		Creek Cave	resurgence	Owl Tree Hole	White Beck Sinks
Cow Close Pot		Fairy Caves	and many other	Swantley Pot	
Crag Foot Mine:natural cavities		(Humphrey Head)	active springs	Wakebarrow Pot	
Dog Holes		Fairy Cave	in the area		
Dunald Mill Cave		(Nichols Moss)			
Fairy Hole		Fissure Cave			
Fissure Cave (Nichols Moss)		(Silverdale)			
Hale Moss Cave		Grand Arch			
Harry Hest Hole		Kirkhead Cave			
Hazel Grove Cave		Levens Cave			
Headech Cave		Lyth Valley Cave			
High Roads Cave 1		Pool Bank Cave			
High Roads Cave 2		Shale Cave			
Helsfell Cave		Silverdale Shore			
Joe Hole		Cave			
Kirkhead Cliff Caves		Wall End Cave			
Merlewood Cave					
Midnight Hole					
Millhead Caves					
Over Kellet Caves					
Paper Mill Cave					
Quarry Wood Cave					
Roudsea Wood Cave : south					
Scar Top Cave					
Shatter Cave					
Unnamed Cave					
Whitbarrow Cave					
Withers Lane Cave					
Yew Tree Cave					

Table 7.6 Genetic classification of the caves of the Morecambe Bay karst.

have been recently formed adjacent to areas of drift cover.

The bulk of the known caves of the Morecambe Bay karst consist of fragments of abandoned phreatic passages (Table 7.6). The size of many of these passages indicates that they were of considerable hydrological importance and suggest a far greater degree of subsurface flow concentration than is apparent in the area today. The sedimentary fill and fragmentary nature of these cave systems makes it difficult to make any deductions about the palaeohydrology of the area. Each passage must have connected upstream to one or more concentrated or diffuse flow feeders, and downstream to a rising. However, it is difficult to infer any relationship between known cave fragments, although Jackson (1910a, 64), and subsequently Ashmead (1969a, 204-207), regarded Badger Hole and the Barrow Scout Caves on the escarpment of Warton Crag (SD4872) as outlets for the Dog Holes system.

At the base of the blocks of the area, the limestone is either underlain by impermeable Silurian rocks or overlain by Recent sediments, resulting in the concentration of groundwater flow at risings. At higher levels, possible former resurgence caves lie perched above the present resurgence level, probably as the result of glacial incision in the valleys. It is likely that many more such systems lie buried beneath the drift which is banked against the foot of the hills. Characteristically, the abandoned resurgence caves are short and close down quickly to inaccessible phreatic inlet tubes. This phenomenon may be explicable in terms of headward enlargement of cave passages back from their resurgence points, as suggested by both Rhodes and Sinacori (1941) and Glennie (1952, 63-65).

On the basis of present knowledge, it is only possible to advance a simple model to account for the palaeohydrology of the Morecambe Bay karst. At least two phases of hydrological development can be recognised: a contemporary drainage system occasionally utilising older lines of drainage, and a fossil drainage system. On the evidence of cave fill, the fossil drainage system appears to date from at least the last interglacial phase (see 5.5). It is likely, however, that many phase of drainage development are represented amongst these older caves, and partly because of this, it is difficult to piece together the known drainage fragments into a related system.

The structural and hydrological similarity of most of the limestone blocks of the Morecambe Bay area makes it necessary only to describe the hydrogeology and palaeohydrology of one in detail, and therefore Whitbarrow has been selected as representative of the general picture. Nonetheless, the atypical nature of the Kellet area is such that it requires separate consideration, for the hydrological picture presented by Kellet has more in common with that of northwest Yorkshire than that of the rest of the Morecambe Bay karst.

7.5.3 The palaeohydrology of Whitbarrow (Fig. 7.21)

The present hydrology of Whitbarrow is reasonably well-defined and the area provides good examples of the type of vadose inlet and resurgence cave characteristic of the

Morecambe Bay region as a whole. On the summit surface are at least three internally-draining closed depressions, as well as one active and one abandoned vadose inlet. The absence of any cover of impermeable rocks discourages surface flow integration and even the stream feeding Wakebarrow Pot, the active vadose inlet, is almost certainly underfit. The bulk of the groundwater recharge is thus by diffuse flow. However, a considerable amount of flow integration must occur within the limestone, for groundwater reappears at discrete risings around the base of the hill where drift is banked up against the limestone. To the north of Whitbarrow, where the limestone is only thinly veneered by drift, groundwater rises through the cover of unconsolidated deposits under hydrostatic pressure.

Similar integration of subsurface flow must have occurred in the past, for abandoned resurgence caves, for example, Pool Bank Cave and Lyth Valley Cave, can be found at various points around the base of the hill. These caves are typically 5 to 10 m above the present resurgence level and, in all cases, are located above existing active springs. This suggests that the contemporary drainage system utilises pre-existing underground flow routes. This is supported by the fact that, during high-stage flows, the old resurgence caves sometimes function as overflows (for example, Lyth Valley Cave; Dobey, 1966). As in the rest of the area, the lowering of the resurgence level is most likely to have been the result of glacial incision in the adjacent valleys. This would suggest the abandoned resurgence caves to be of pre-last glacial age, a thesis which is borne out by the spring

vadose inlet

abandoned phreatic cave

closed depression

subsurface hydrological link

major fault

Recent

Upper Carboniferous Sandstone

Gleaston Formation

Lower Carboniferous Limestone

KEY: Figs. 7.21 and 7.22

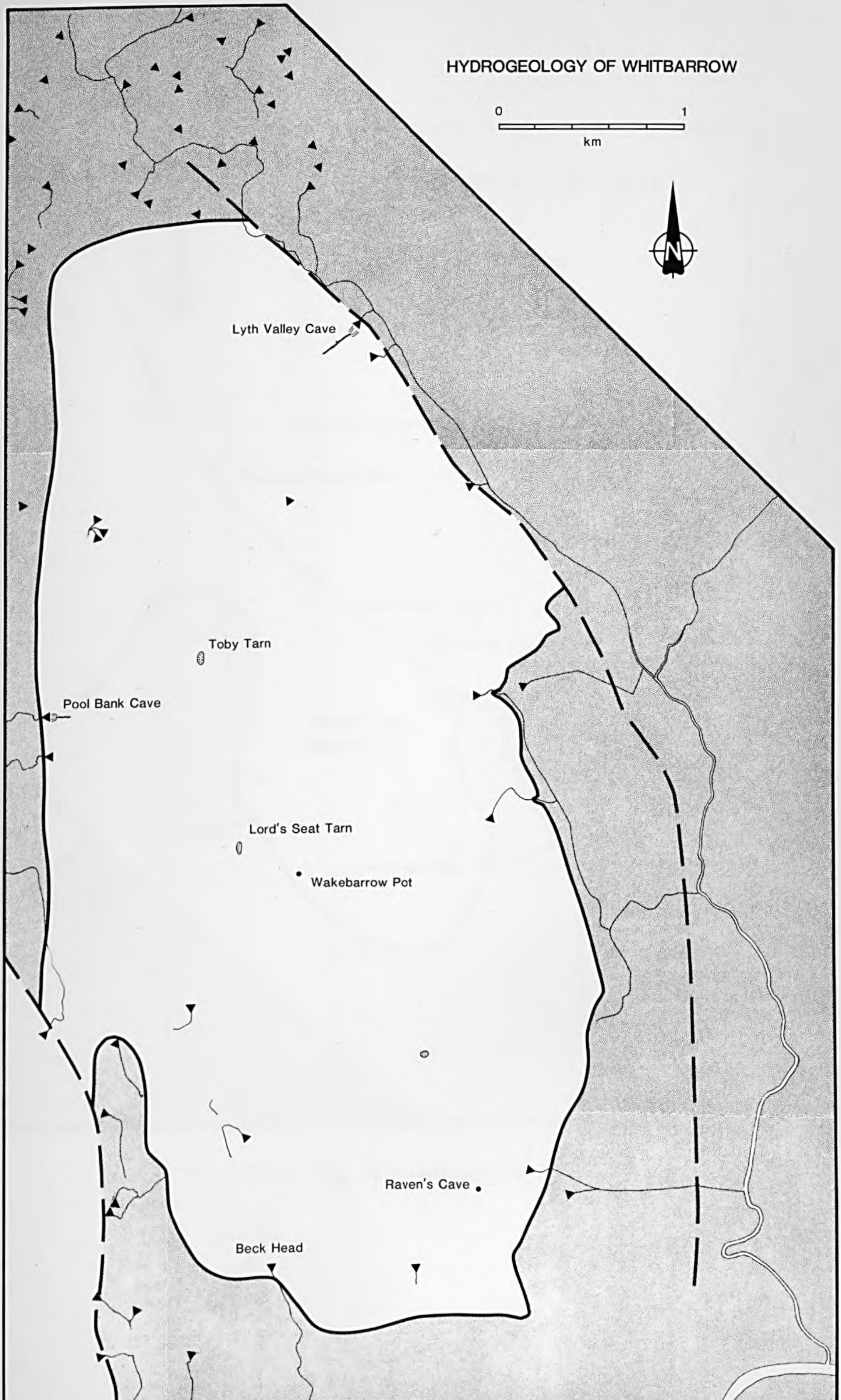


Fig. 7.21 The hydrogeology of Whitbarrow

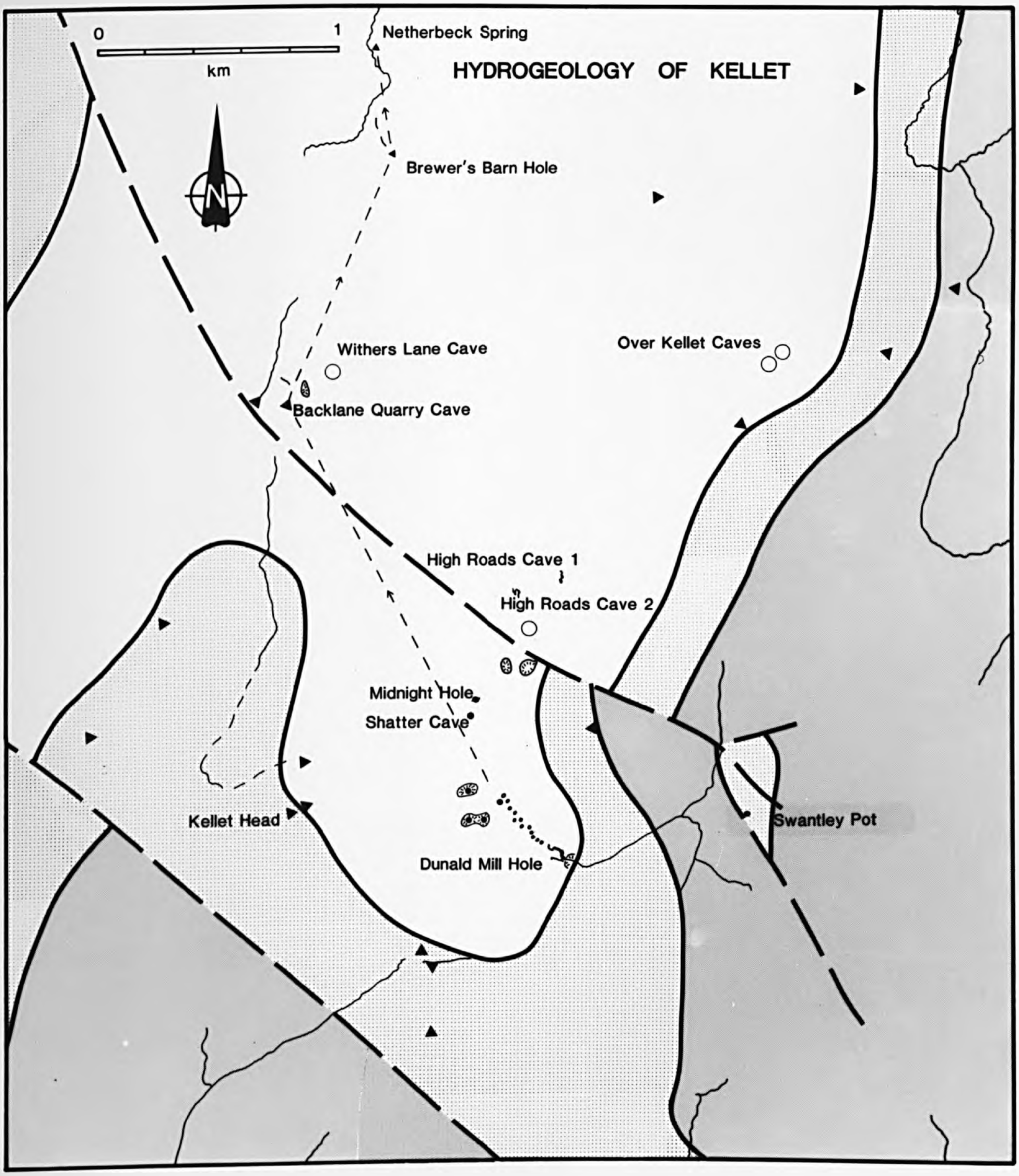


Fig. 7.22 The hydrogeology of Kellet

sedimentological evidence in the caves (see 5.5).

7.5.4 The palaeohydrology of Kellet (Fig. 7.22)

Kellet differs from the rest of the Morecambe Bay area mainly in terms of its structure. In contrast to the simple, cuesta-like limestone blocks of the rest of the region, the rocks of Kellet are highly folded and fault-dissected. The area consists basically of a planed synclinal structure, of which the most important result, from a hydrological point of view, is the fringe of impermeable lithologies surrounding the limestone. In general, the limestone constitutes the higher ground in the area. This means that the limestone-Gleaston Formation boundary tends to form a resurgence line for karst groundwaters, with a less significant grouping of springs occurring around the basal outcrops of the Namurian Sandstone.

At Dunald Mill Hole, the general hydrological situation is reversed and the limestone forms an engulfment point for waters flowing from the adjacent impermeable rocks, the stream sinking soon after meeting the limestone. The entrance to the cave forms an impressive blind valley, which has resulted, in part, from progressive collapse of the cave passage.

Numerous remnants of an abandoned karst drainage network can be found in the area, and many of the caves which constitute this network provide evidence of a phreatic phase of development, followed by a period of vadose incision (for example, High Roads Caves 1 and 2, Dunald Mill Cave, Midnight Hole, Shatter Cave; Baldwin, 1969a; 1969b). Elsewhere, however, as at Withers Lane Cave and the Over Kellet Caves, there is conclusive evidence only of phreatic development. Some of these abandoned caves (for example, Midnight Hole, Shatter Cave and High Roads Cave 1) also possess fossil vadose inlets connecting up to the present land surface, and similar vadose inlets, occasionally in association with surface closed depressions, can be found elsewhere in the area.

From a study of the altitudes of the abandoned phreatic cave remnants, Ashmead (1969b; 1974a, 222) concluded that they were developed at a general base level of approximately 80 m O.D. He took this base level to be related to the 88 m sea level proposed by Parry (1958; 1960) (see 8.2.1). Thus, taking into account the deposits in the caves, which he considered to be of last glacial age, and the dissected nature of the cave systems, thought to be the result of glacial erosion, Ashmead assigned a probable last-interglacial age to the Kellet caves.

The altitudes of the phreatic components of the caves of the Kellet area are given in Table 7.7. On the evidence of so few caves it is difficult to justify their relationship to a specific base level, particularly bearing in mind that, in Kellet, the dip of the limestone varies generally between 10^{o} and 20^{o}. Ford (1971a) has shown that, under these circumstances, the height of any one part of a phreatic system cannot be directly related to that of a controlling base level. Furthermore, the existence of known active phreatic systems at $\sim$55 m O.D. in Dunald Mill Cave (Smith, 1890, 7) and at $\sim$86 m O.D. in High Roads Cave 2

Cave	Altitude (m O.D.)	
Withers Lane Cave	30	(Baldwin, unpub.)
Over Kellet Caves	85	(Holland, 1967)
High Roads Cave 1	91	(Baldwin, 1969a)
High Roads Cave 2	101	(Ashmead, 1974a)
	91	(Baldwin, 1969a)
Midnight Hole	85	(Baldwin, 1969b)
Dunald Mill Hole	73	(Smith, 1890)

Table 7.7 Altitude of the phreatic remnants of the caves of the Kellet area

(Baldwin, 1969a) provides evidence of continuing cave development at similar heights to that proposed by Ashmead for ?last interglacial speleogenesis.

As a result of quarrying activity, many of the caves mentioned above no longer exist. However, the major engulfment system of Dunald Mill Hole-Dunald Mill Cave is able to provide considerable information as to the palaeo-hydrological development of Kellet. According to Ashmead (1964, 13; 1969a, 207; 1974a, 221-223; 1974c, 51), the development of Dunald Mill Hole has progressed through a number of phases. The earliest, phreatic, phase was regarded as related to a base level developed at 80 m O.D. throughout the area. This phase was succeeded by two vadose episodes, the first related to a base level at 76 m O.D. and the second to a base level at 61 m O.D. The cave was subsequently infilled by glacially-derived materials, which were later removed by the present, underfit, stream.

The present work proposes an alternative, more complex, sequence of events. The Dunald Mill system comprises two separate caves, linked only by a short, hydrologically important, bedding plane crawl (Fig. 7.23). The smaller, upper system, Dunald Mill Cave, is developed approximately parallel to Dunald Mill Hole. It is suggested that the upper system formed an earlier drainage route whose waters were subsequently captured by the development of the lower system.

The upper system, Dunald Mill Cave, demonstrates the characteristic sequence of development of all the caves

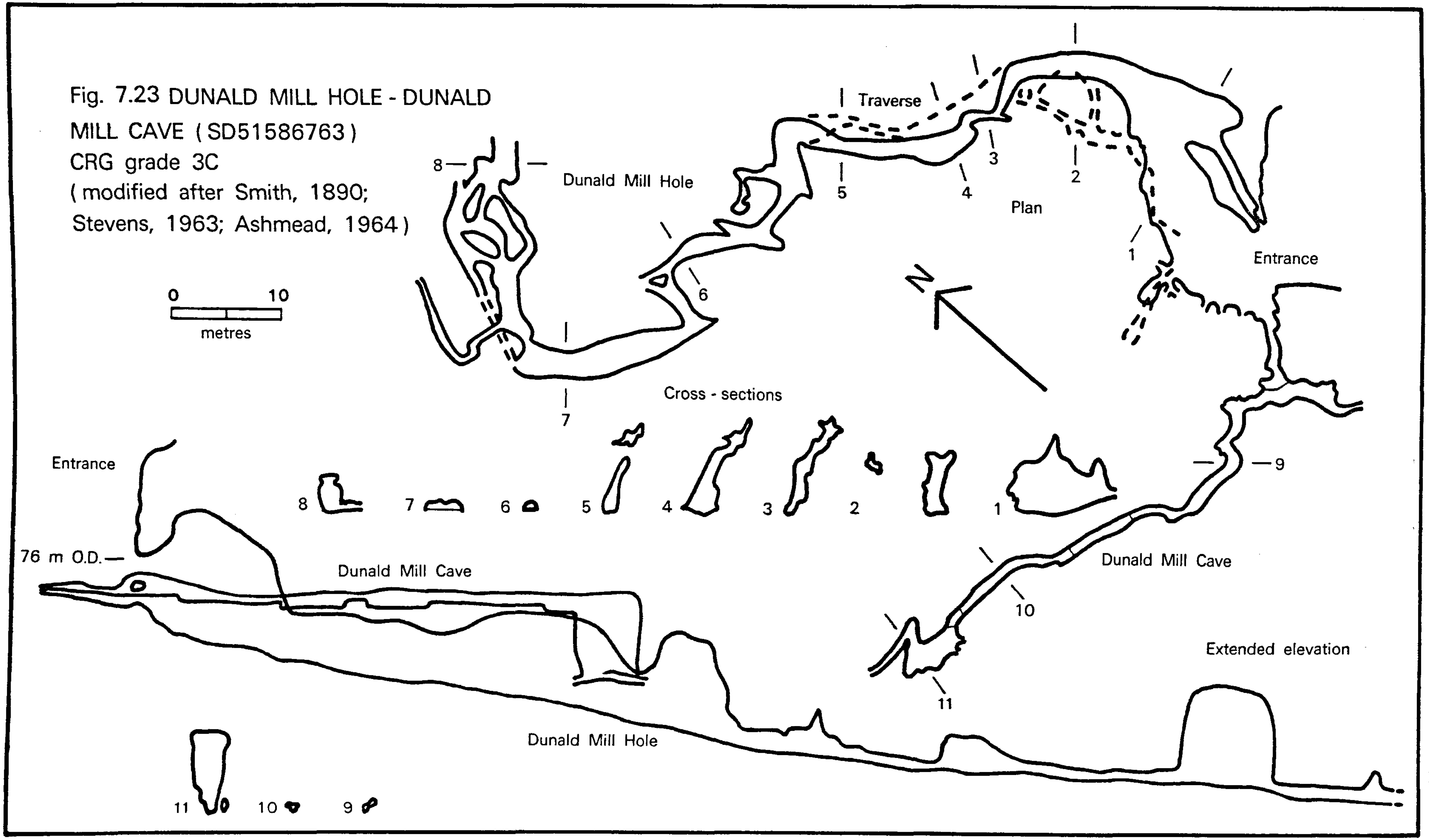


Fig. 7.23 DUNALD MILL HOLE - DUNALD MILL CAVE (SD51586763)
CRG grade 3C
(modified after Smith, 1890; Stevens, 1963; Ashmead, 1964)
0
10
metres
Dunald Mill Hole
Traverse
Plan
Entrance
Cross - sections
1
2
3
4
5
6
7
8
9
10
11
Dunald Mill Cave
Extended elevation
76 m O.D.
Entrance
Dunald Mill Cave
Dunald Mill Hole

in the Kellet area. The initial phreatic-vadose succession, which may be recognised in the T-section passage form, was followed by a phase of infill, during which ill-sorted clastic deposits, consisting of rounded erratic pebbles and cobbles in a matrix of sands, silts and clays, were deposited in the vadose trench. It is possible that these deposits choked the outlet to the cave, for up to a metre of laminated fine-member beds are found in the pitch at the end of the cave, suggesting the existence of ponded conditions in the shaft.

This phase was succeeded by a quiescent period, during which percolation waters deposited stalagmite in the cave. The stalagmite coats the roof and walls of the passage, as well as forming a false floor over the clastic deposits. It is possible that two such phases of infilling and stalagmite deposition occurred, for the remains of a second, higher stalagmite floor overlying clastic beds can be seen in the streamway. Each of these phases was followed by a period of vadose incision. Thus, the present stream has partially eroded the lower stalagmite floor, whilst a former, larger stream at the foot of the shaft has cut through the laminated beds which presumably once filled this chamber.

The lower system, Dunald Mill Hole, contains evidence of a high-level, abandoned phreatic passage, the Traverse, at approximately the same level as Dunald Mill Cave (Fig. 7.23). This passage, possibly part of an early anastomosing phase, has been left inactive by the vadose incision of the main streamway. Ashmead regarded these features as indicative of a phreatic-vadose-vadose sequence in the cave. There is, however, no need to invoke any hydrological change to account for the present form of Dunald Mill Hole. Any inlet cave of this type will experience a progressive downcutting of the entrance by the inflowing stream, resulting in the extension of vadose conditions along the cave and the progressive collapse of the cave entrance. The importance of this process in Dunald Mill Hole is evidenced by the gorge-like form of the inlet valley and the massive collapsed blocks of the first chamber.

The cave contains a large amount of clastic sediment, which forms considerable banks in parts of the cave. Initially, this must have consisted of the poorly-sorted deposits characteristic of other caves in the Morecambe Bay area, but, in the active streamway, at least, the fine fraction of the deposit has been removed, leaving only gravel-grade material. There is some stalagmite in the roof of the first chamber and in the high-level phreatic section, but the stratigraphic relationship of the stalagmite to the clastic sediments can only be inferred from the relative height of the deposits in the cave.

It is difficult to assign a chronology to the development of the Dunald Mill system. The depositional sequence in Dunald Mill Cave, and possibly in Dunald Mill Hole, is identical to that found elsewhere in the Morecambe Bay area, where it is cautiously interpreted as indicating a pre-last glacial age for the caves (see 5.5.2). The morphology of the Dunald Mill system, however, suggests a relatively complex pattern of development, perhaps taking place over a long time-span. The approximately parallel nature of Dunald Mill Hole and Dunald Mill Cave suggests that they once performed a similar hydrological role. The upper system may have been left abandoned by the development of the lower system. Alternatively, since the upper phreatic passages in Dunald Mill Hole are at a similar height to Dunald Mill Cave, the upper passages could have formed part of the same proto-Dunald Mill system which was abandoned by the incision of the active streamway.

The present hydrology of the Dunald Mill system has been studied by Smith (1890), who showed that at least part of the water from the caves resurges at Netherbeck Spring, and that, during high-stage flows, both Backlane Quarry Cave and Brewer's Barn Hole act as overflows for the system (Fig. 7.22).

7.5.5 Conclusions

Previous models of drainage development in the Morecambe Bay area, which sought to relate cave development to the heights of former regional base levels, have been shown to be unsoundly based. In their place, a simple model of regional hydrological development has been advanced, founded upon the assumption that similar arrangements of subsurface drainage existed during earlier phases of hydrological development as exist in the area today. The contemporary drainage system consists of concentrated and diffuse flow inlets feeding phreatic passages which, in turn, link up with springs and resurgence caves concentrated around the base of the limestone blocks. The abandoned subsurface drainage elements must therefore represent at least one earlier phase of hydrological development, of pre-last glacial age. It is more likely, however, that the abandoned passages represent more than one episode of drainage history, the relicts of which have been superimposed to provide the complex pattern of fossil passages found in the area today.

8. SURFACE KARST

8.1 Introduction

The majority of the surface karst features of the Morecambe Bay area are relict forms. Partly this is because much of the relief is the product of successive episodes of glaciation during the Quaternary, and partly it is because karst processes are not active over much of the area at present. With the exception of the coast and allogenic rivers such as the River Kent and Poaka Beck in Furness, there is little concentration of water at the surface and hence little surface concentration of solutional activity.

Previous work on the surface karst of the area has attempted to relate the development of the karst landscape to the erosional history of the region as a whole. The evidence for such an erosional chronology is reviewed in 8.2. The more important elements of the karst landscape are considered in 8.3 and 8.4, in particular the mosses of the area, which have been regarded as examples of true karst poljes (see 8.3), and the previously undescribed large closed depressions of the region (see 8.4). Finally, a less well known karstic phenomenon, cemented screes, is discussed (see 8.5). Cemented screes, which are a frequently occurring yet little described feature of glaciokarst landscapes, form an important stratigraphic element within the Morecambe Bay area.

8.2 Regional geomorphology

8.2.1 Previous work

The regional geomorphology of the Morecambe Bay karst has been interpreted by Parry (1958; 1960a) as comprising a series of erosion surface remnants formed by the processes of planation and marginal trimming resulting from a long and complex sequence of regional base level changes. The highest surface found in the karst area is the 213-290 m O.D. (700-950 ft) "partial peneplain". Parry assigned to this surface a Mio-Pliocene age after correlating it with similar relicts in Wales (Brown, 1952). The 210 m O.D. (690 ft) surface is ascribed to the Plio-Pleistocene marine transgression. Below this, minor stillstands are marked by notches and platforms at 174 m O.D. (570 ft), 146 m O.D. (480 ft) and 131 m O.D. (430 ft). There are remnants of a widespread marine surface at 116 m O.D. (380 ft) and a rather more restricted surface at 101 m O.D. (330 ft). A further stage is recorded at 88 m O.D. (290 ft) but it is so extensively covered by drift that it is difficult to interpret.

Parry believed that sea level fluctuations during the Late Pleistocene resulted in stillstands at 46 m O.D. (150 ft), 30 m O.D. (100 ft) and 15 m O.D. (50 ft). Benches associated with the latter two levels are cut into the drift of the Main Glaciation and are therefore regarded as of interstadial or postglacial origin, whilst the 46 m surface is overlain by drift of the Main Glaciation and is considered to be of interstadial or interglacial origin.

Ashmead (1969a; 1974a; 1974c) subsequently attempted to relate these erosion surface remnants to altitudinal zones of landform development in the area. He cited numerous examples of wave-cut notches, marine benches and sea caves as direct evidence of marine stillstands; and regarded dolines, poljes and cave systems, all apparently constrained within well-defined height ranges, as being associated with former water tables in the limestone.

It has already been demonstrated (see 7.5.1) that the evidence for altitudinal control of cave development in the area is insubstantial. Similarly, many of the wave-cut features mentioned by Ashmead cannot be found on the ground and no evidence of them is shown on the large-scale topographic maps of the area. These include the 46 m bench on the western side of Arnside Knott (SD4577) (Ashmead, 1974c, 43) and the "marine notches" at 15 m O.D. on Humphrey Head (SD3973) (Ashmead, 1974a, 224). In those cases where notches and benches can be identified they are often structurally controlled and have a form incompatible with a wave-cut origin. Thus, "quarrying of frost scree breccia" on Whitbarrow (SD460852) was said to have "revealed a well-developed (sea) notch at 30 m O.D." (Ashmead, 1974a, 223). Yet inspection of this feature reveals that it is a glacially striated bedding plane dipping at approximately 10° to the east, rather than seaward to the south (Plates 8.1 and 8.2). Furthermore, the "marine platform" said to be developed at 40-45 m O.D. on the west side of Kirkhead Hill (SD3975) can be shown to be a structural bench, bound on its landward side by a fault scarp cliffline.

All the features interpreted by Ashmead (1969a, 204; 1974c, 61; 1974a, 223-224), and subsequently by Tooley (1977, 6; 1979, 146), as sea caves contain many of the diagnostic features of solutional erosion under phreatic conditions, such as wall and ceiling pockets (Bretz, 1942). This is true both of the high level caves, such as Whitbarrow Cave (SD451845) at 110 m O.D. and Harry Hest Hole (SD494728) at 111 m O.D., and

Plate 8.1 Bedrock bench at White Scar, Whitbarrow (SD460852), exposed due to quarrying of the overlying cemented scree. Looking east.

Plate 8.2 Glacial Striations on the bedrock bench at White Scar, Whitbarrow (SD460852) (see Plate 8.1)

the caves at lower altitudes. The latter include the Grand Arch (SD390738) and the Fairy Caves (SD390739) at 34 and 29 m O.D. respectively on Humphrey Head; Kirkhead Cave (SD392755) at 38 m O.D.; and, along the southwest fault scarp of Warton Crag (SD482728), Badger Hole at 29 m O.D. and the Barrow Scout Caves at 15 m and 23 m O.D.

Tooley (1977, 6) suggested that those caves found between 12 m and 29 m O.D. were formed by high sea levels during the period 25000-19000 B.P., although a Hoxnian or earlier age was subsequently proposed by him for these and higher features (Tooley, 1979, 146). Ashmead, on the other hand, claimed that the sea caves occur at distinct levels of 46 m O.D., 30 m O.D. and 15 m O.D., and that the caves are the result of erosion by the Late Pleistocene sea levels proposed by Parry at these altitudes. More accurate measurements of cave altitudes indicates, however, that the caves do not occur at heights coincident with those proposed for marine still-stands and do not easily fit into a picture of formation by Late Pleistocene high sea levels.

The Grand Arch was regarded as a particular example of a sea cave with a blowhole extending for 12 m to a wave-cut platform at 45 m O.D. Yet the cave entrance faces north-north-west, despite both the westerly aspects of the marine cliff on which it is situated and the prevailing fetch from the southwest. The "blowhole" itself shows a remarkable morphological similarity to the potholes and associated dolines characteristic of karst areas (Plate 8.3).

Finally, as has been shown in 5.3 and 5.5, none of the caves contain any evidence of the marine deposits that might

Plate 8.3 The entrance to the Grand Arch, with its associated doline, on Humphrey Head (SD390738)

be expected, given their proposed origin.

8.2.2 Study of the erosion surface remnants

It was decided to re-examine the ideas of Parry and Ashmead in the light of these findings. Hitherto, erosion levels have been studied by the recognition of erosion surface remnants, yet such remnants frequently occupy only a small proportion of the land surface area; moreover, for some time the whole basis of such studies has been under attack (Chorley 1965a, 34; 1965b, 148-149). Even where the more stringent technique of trend surface analysis has been used (e.g., King, 1969; Rodda, 1970), it has been suggested that serious abuse of the method has occurred (Tarrant, 1970; Unwin and Lewin, 1971). Further problems arise from the proposed existence in the Morecambe Bay area of a series of marginal topographic trimmings rather than the widespread surfaces necessary for the application of trend surface techniques.

In order to overcome the difficulties of studying such a series of erosion surface remnants, it was proposed that precisely and accurately measured slope profiles be used as an objective means of data collection (see 11.4.3). Such profiles provide a continuous record of landscape morphology suitable for quantitative analysis.

Hampsfell and Underbarrow, two of the cuesta-like limestone blocks which fringe Morecambe Bay, were chosen for the study (Fig. 1.1). They are mainly free of the thick drift cover which plasters the Low Furness and Kellet blocks; they are structurally simple; and they are sufficiently high to allow a study of the higher altitude erosion surfaces.

More importantly, they are in what Parry would have regarded as former "estuarine" or "coastal" sites and are considered by him to include remnants of almost all the erosion surfaces which cut across the limestone.

As can be seen from Fig. 8.1, practical difficulties of access rendered the random sampling of profiles impossible. Nevertheless, as the choice of profiles was governed by the absence of obstacles along the survey line rather than by subjective selection, it is believed that the sampling method does not preclude statistical treatment of the results.

Fourteen profiles were surveyed from the crest to the foot of the limestone blocks. The location of almost all these profiles on the scarp slopes of the blocks provided two advantages:

i) the generally steeper scarp slopes allowed a greater altitudinal range to be covered more quickly;

ii) the scarp slopes are in locations which would have been more susceptible to marine erosion so that erosion level remnants are more likely to have survived.

8.2.3 Analysis and results

Young's (1971) system of Best Units was adopted as an objective means of analysis of the slope data collected in the field. This system splits the profile into segments and elements, each of which possesses certain properties of form, so that coefficients of variation do not exceed specified values. Of the coefficients of variation suggested by Young, the following were found to divide the profiles into units with a good standard of internal uniformity and were also compatible with the relatively long surveyed unit length:

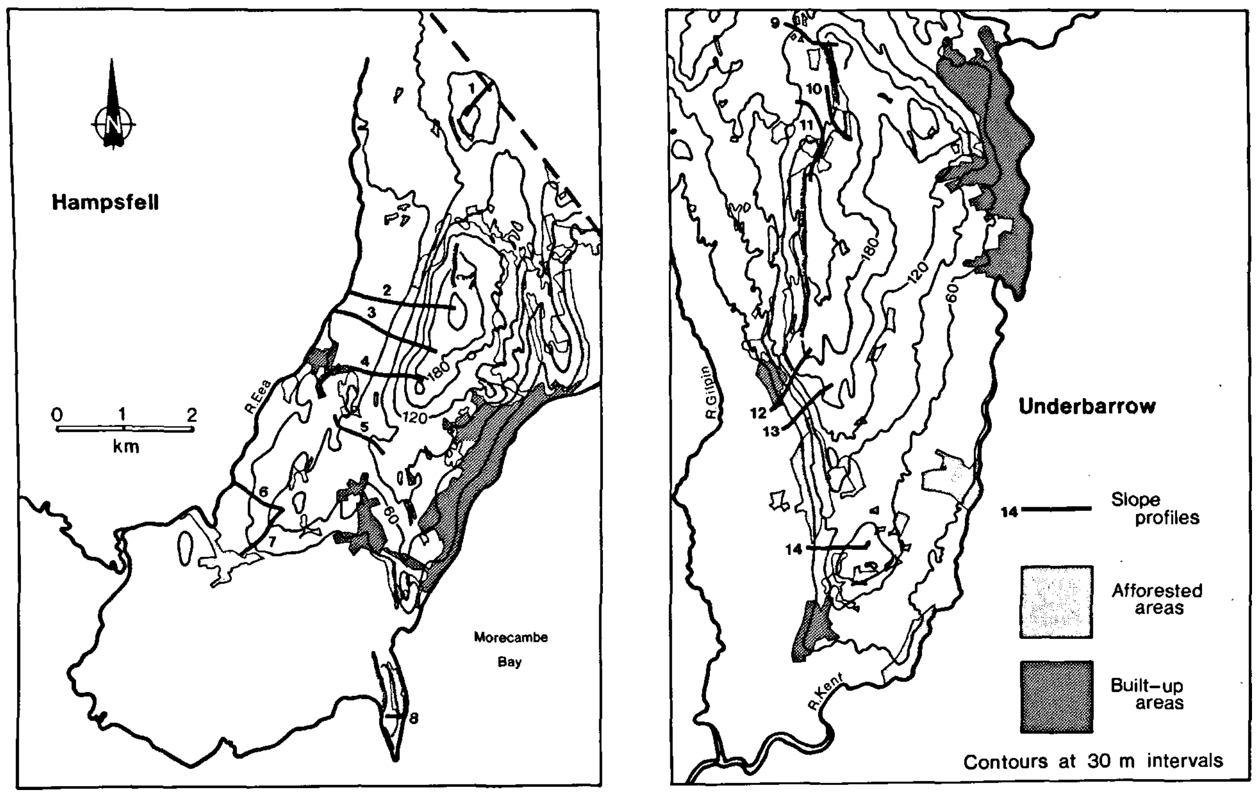


Fig. 8.1 The location of the surveyed slope profiles

segments	Va	max	25%
elements	Vc	max	50%

Young's program is designed so that when slopes have angles of 2^{o} and less the coefficient of variation is altered to give units with a higher degree of internal homogeneity. However, in a landscape such as that under consideration it was felt to be more useful to follow Savigear (1967) and to consider all slopes of 2^{o} and less as horizontal.

It was decided to test whether the concave breaks of slope regarded by Parry as approximating to the base of former sea cliffs (1958, 51-2) could be found at the altitudes at which erosion-level notches should exist on the surveyed profiles. Analysis was restricted to those parts of the profile at and above 15 m O.D. This is just below the height of the lowest erosional facet examined in detail by Parry (1958, 95). The altitudes of all concave breaks of slope found by Best Units analysis was noted, whether they occurred in solid or superficial materials. This was justified by the fact that many of the lower level surface remnants identified by Parry are cut into the drift and that elsewhere it seems likely that the surface form reflects bedrock changes which have been but thinly veneered by superficial deposits.

Occasionally, the Best Units method of analysis appears to choose apparently arbitrary points of slope change. Nevertheless, no attempt has been made to remove such points from the analysis as this would have made it difficult to select breaks of slope without establishing specific criteria for each case. Moreover, the interpretation of a feature as being arbitrarily selected may in itself be a subjective opinion based on an inability to perceive subtle changes in slope form.

The altitudinal ranges at which "inner marginal breaks of slope" occurred were noted (Table 8.1) and the number of concave breaks of slope on the profiles which fell within these ranges was found. This was compared with the expected number of concave breaks of slope that might be found assuming the altitude of the "inner marginal breaks of slope" to have had no influence on their location, and the chi-square test applied (Table 8.2). It was found that even if the critical value (α) were only 0.25 (i.e. 75% confidence level), the null hypothesis of no difference between observed and expected values must still be accepted. A significant difference not having been established, it can be concluded that the erosion levels postulated by Parry do not have a significant effect on slope morphology, at least within the areas studied.

8.2.4 Discussion

Within the Morecambe Bay karst area, Parry claimed to be able to recognise the remnants of 11 erosion surfaces, which have also been traced throughout the whole of the southwestern Lake District. This implies an average occurrence of one surface every 20 m. The heights attributed to these surfaces, however, are those of the inshore margins of the marine platforms and they ignore the height ranges over which Parry was able to identify surface remnants. If the height ranges are considered, an interesting pattern emerges. It can be seen from Fig. 8.2 that a considerable amount of overlap exists for the altitudinal ranges at which surface

Erosion surface	Altitudinal Range at which "inner marginal breaks of slope" occur	
	Feet	Metres
Monk Coniston	800-1000	243.8-304.8
Winder	660-690	201.2-210.3[1]
Winder: near former "headlands"[2]	650-670	198.1-204.2
Knittleton	550-570	167.6-173.7
Whitriggs	450-500	137.2-152.4
Hannah Moor	430	131.1
Tomlin	380-390	115.8-118.9
Green Haume	330	100.6
Coneyside	270-290	82.3- 88.4
	140-160	42.7- 48.8
	90-120	27.4- 36.6
	60- 67	18.3- 20.4

1. Parry (1958, 76) also notes the existence of an "inner marginal break of slope" beyond this range at 213.4 m.

2. All measured profiles fall into the "headland" area delimited by Parry (1958, 79-80).

Table 8.1 The altitude of "inner marginal breaks of slope" in the Morecambe Bay area according to Parry (1958).

	Observed	Expected
Concave breaks of slope coincident with Parry's "inner marginal breaks of slope"	28	23.4
Concave breaks of slope not coincident with Parry's "inner marginal breaks of slope"	51	55.6

$\chi^2 = 1.28$

$\chi^2 \quad (0.25; 1) = 1.32$

$\therefore$ H_1 (significant difference between observed and expected values) rejected at 75% significance level.

Table 8.2 Comparison of observed and expected altitudes of "inner marginal breaks of slope" using the chi-square test.

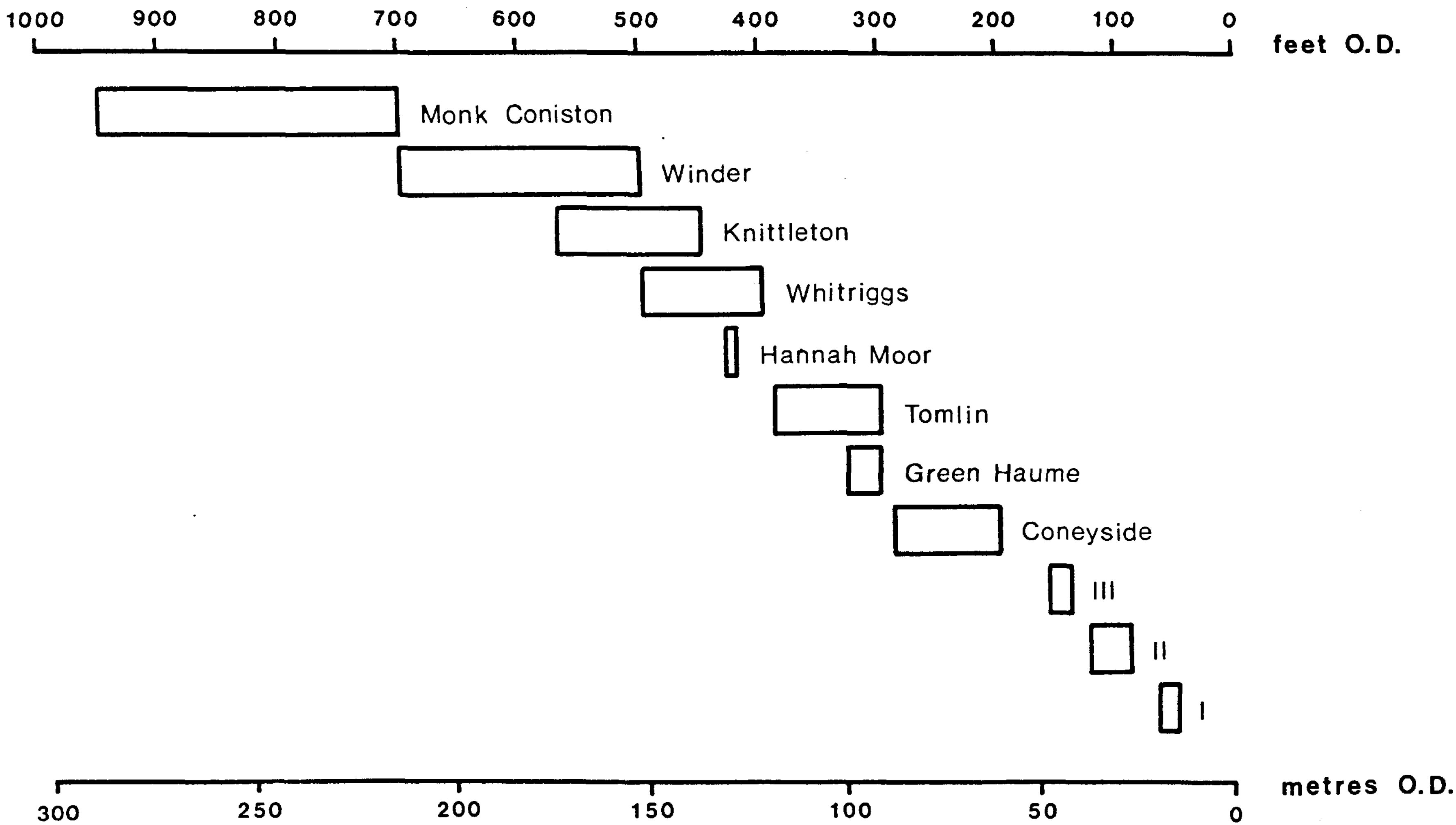


Fig. 8.2 The heights at which erosion surfaces are recognised in the southwestern Lake District, England (source of data: Parry, 1958)

remnants can be found and, where there is no overlap, surfaces are differentiated by only a few metres. Having recognised this, it becomes difficult to see how Parry could have failed to have discovered surface remnants corresponding to his hypothetical erosion levels. This is revealed in his means of analysis. The Knittleton Surface, for example, whose inner marginal break of slope is at 570 ft O.D., was identified by a series of "well-marked benches between 500 ft and 570 ft ... sufficiently numerous to warrant their treatment as a separate erosional phase" (1958, 80). Similarly, the Whitriggs Surface immediately below it was identified by "benches at between 400 ft and 480 ft" (1958, 82).

A further point arising from a study of the altitudinal ranges of erosion surface remnants is the possibility that Parry's work is the result of the transformation of uniform into non-uniform distributions as suggested by Hodgson, Rayner and Catt (1974) in their criticism of Sparks' (1949) study of the denudation chronology of the South Downs. They recognised a striking 50 ft periodicity in the height distribution of Sparks' surface which they saw as a possible consequence of the preliminary selection of likely "flats" from the Ordnance Survey 1:63360 maps, which are contoured at 50 ft intervals.

Much of Parry's analysis was similarly cartographical and it may be that this influenced the selection of the 50 ft interval which characterises the height ranges (and especially the lower height limits) of his erosion surfaces (Table 8.3). The main exceptions to the 50 ft periodicity of the data are the Hannah Moor Surface, which at 420-430 ft exists merely as a sub-unit of the Whitriggs Surface, and the three lower

Erosion Surface	Altitudinal range (ft)
Monk Coniston	700 - 950
Winder	500 - 700
Knittleton	450 - 570
Whitriggs	400 - 500
Hannah Moor	420 - 430
Tomlin	300 - 390
Green Haume	300 - 330
Coneyside	200 - 290
	140 - 160
	90 - 125
	50 - 67

Table 8.3 The altitudes at which "flats" can be found in the Morecambe Bay area according to Parry (1958).

surfaces, whose altitudinal ranges are based remarkably closely on the lowest three contours found on Ordnance Survey 1:63360 maps: 50 ft, 100 ft and 150 ft.

The apparent absence of base level controls on hill-slope morphology within the Morecambe Bay karst has important implications for geomorphological and palaeohydrological studies in the area. It has been shown that the evidence for altitudinal control of landforms is, at best, weak, and now it would also appear that there is little evidence, either of a morphological or sedimentological nature, for the process by which changes in the controlling water level were thought to have occurred.

It is likely that the step-like nature of slopes in the area (Plate 8.4), which possibly encouraged Parry in his belief in the existence of wave-cut platforms and notches, is the result of structural control (sensu lato). Certainly this seems to be true for other glacio-karst areas, where attempts to relate slope form to structural controls have been common since Goodchild (1875, 72-74) first suggested that, where ice is known to have moved along outcrops of bedded limestone, the result was a form of terracing caused by the alternation of hard and soft beds.

Since then, numerous other controls have been suggested which could have given rise to similar relief. These include the frequency of bedding and jointing in the limestone (Douglas, 1909, 541-44; Sweeting, 1966, 180; Williams, 1966, 162-63) and lithological variations between limestone beds (Parry, 1960b, 15-16; Rapp, 1960; Sweeting, 1966, 180; Williams, 1966, 162). It has also been suggested that the

Plate 8.4 *Schichttreppenkarst* on Whitbarrow (SD4584)

bedrock may be weakened along certain structural lines prior to glaciation and the weakened strata subsequently removed by glacial scour. This may take place as a result of subaerial erosion or subsoil solution along joints and bedding planes (Bögli, 1964, 67; Clayton, 1966, 366, 370, 380), or due to the presence of shale interbeds which protect the underlying rock from solutional attack whilst facilitating the dissection of the overlying strata (Moisley, 1954, 37-39).

According to Johnson, Tallis and Pearson (1972, 535), ice probably covered the Morecambe Bay karst as late as 18000 B.P. This has left considerable evidence of scouring on all the interfluve areas, yet glacial scour seems to have failed to result in slopes whose form is systematically related to any one aspect of structural control. Specific changes in hillslope morphology are structurally governed. For example, the platey, somewhat siliceous limestone of the Red Hill Beds forms cliff-like outcrops along the length of the Underbarrow scarp slope, despite the fact that thereabouts these beds are only 7-9 m thick (Garwood, 1912, 519). However, no single controlling element can be identified as being predominant. It seems most likely that choice of specific beds for scour is the result of a complex of controls whose end result is in most cases an apparently arbitrary selection, as concluded by Williams (1966, 161) in County Clare.

8.2.5 Conclusions

The results obtained suggest that slope form, at least within the localities studied, does not conform to that expected on the basis of Parry's work, which itself is open to criticism as a result of possible crucial flaws in data collection technique. No simple interpretation of slope morphology is possible. The area possesses considerable similarities with other upland glaciokarst areas (for example, Craven and County Clare in the British Isles, and the Märenberg Plateau in Switzerland), although its more steeply dipping beds may have resulted in some modification of the typical picture of schichttreppenkarst. The results also suggest that existing interpretations of the geomorphological development of the area should be revised and that landscape history cannot be easily related to the simple sequence of base level changes proposed by Parry.

8.3 The "poljes" of the Morecambe Bay karst

8.3.1 Introduction

It was first suggested by Ashmead (1969a, 207; 1974a, 212-213; 1974c, 49-50, 57) that the mosses of the Morecambe Bay area might be regarded as true karst poljes. At least 9 of the mosses of the area might qualify to be so classified (Fig. 8.3):

Winster Valley (SD4284)
Witherslack (SD4485)
Lyth Valley (SD4787)
New Barns Bay (SD4477)
Silverdale Moss (SD4678)
Leighton Beck (SD4977)
Hawes Water (SD4776)
Burton Moss (SD5076)
Leighton Moss (SD4875)

Although there have been several attempts to define the polje (see Gams, 1978 for a review of definitions), no single feature may be regarded as characteristic of all poljes.

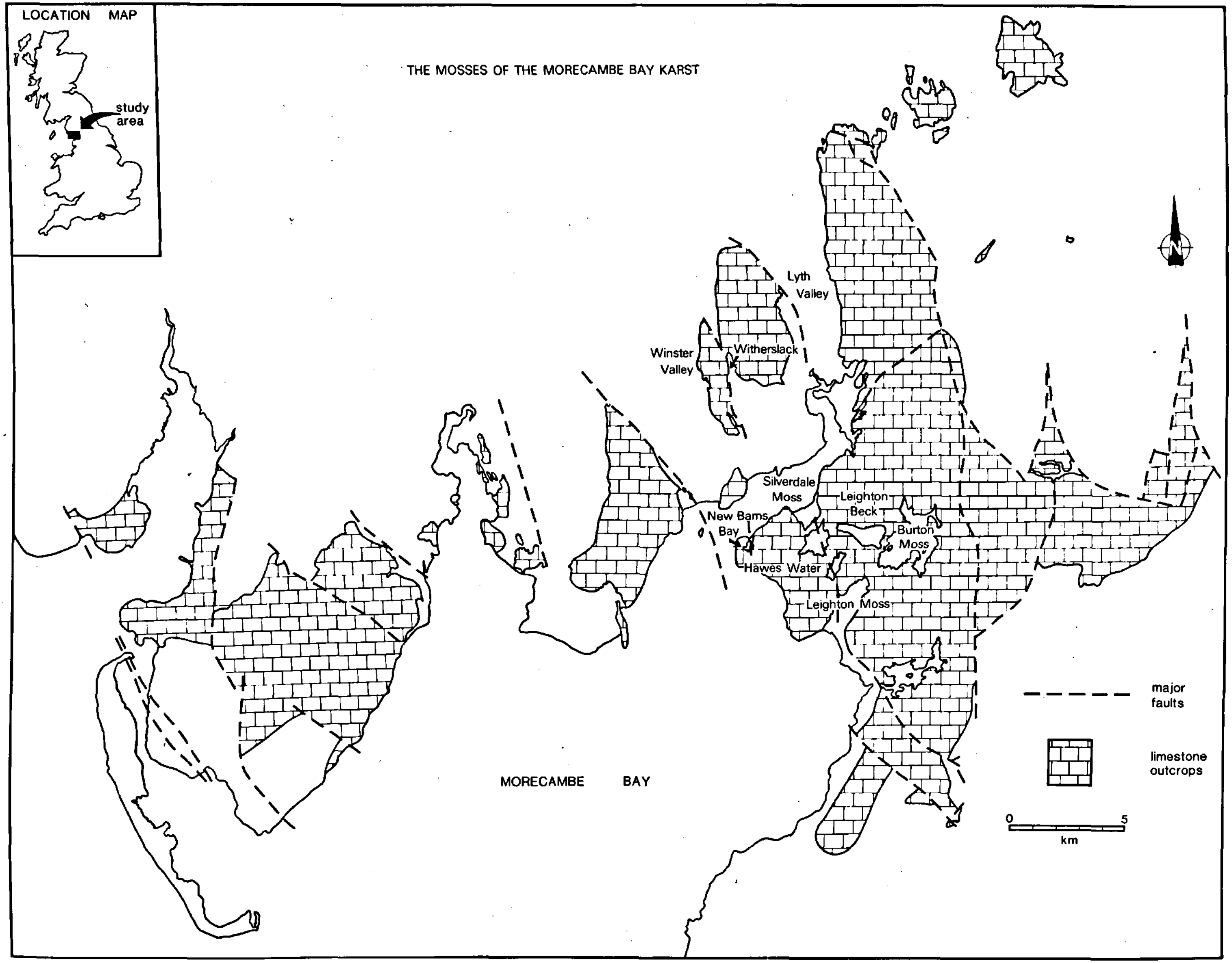


Fig. 8.3 The mosses of the Morecambe Bay karst

Nevertheless, three groups of characteristics: morphological, hydrological and process-related, appear to be almost ubiquitous, and the mosses of the Morecambe Bay area may be classified in terms of the presence or absence of these.

8.3.2 Morphology

Characteristically, poljes, are large, flat-bottomed depressions, whose sides, not necessarily of limestone, rise steeply from the polje floor. The polje floor is typically composed of impermeable materials. Occasionally, isolated, steep-sided residuals of limestone, known as hums, remain within the polje. Many workers (for example, Sweeting, 1972, 200) have also stressed the significance of structural control upon polje location and morphology.

The Morecambe Bay mosses range in size from 0.33 km^2 to 16 km^2. They are floored by estuarine and lacustrine sediments (Munn Rankin, 1910; 1911; Erdtmann, 1928; Gresswell, 1958; Smith, 1958; 1959; Oldfield, 1960a; 1960b; Tooley, 1969; Huddart, Tooley and Carter, 1977) and surrounded on at least two sides by low limestone cliffs. Hum-like features occur at Gilpin Bank (SD4687) in the Lyth Valley (Plate 8.5) and Hazel Grove (SD500771) in Burton Moss. In all cases the mosses are either strike-aligned, as in the case of the Winster Valley, or are developed along major tectonic axes (Fig. 8.3). The Silverdale monocline and its northward extension have been particularly important in this respect, for Leighton Moss, Hawes Water, Silverdale Moss and the Lyth Valley are all developed along this line.

Plate 8.5 Hum-like features in the Lyth Valley at Gilpin Bank (SD4687)

8.3.3 Hydrology

According to Sweeting (1972, 193-194), the poljes of the Classical and Dinaric karst are drained by at least one river. However, they are also subject to frequent seasonal inundation, during which times much of the drainage may be through ponors in the polje floor. The floods are partly the result of seasonal high flows in the rivers and partly the result of inflow from marginal springs. During rising flood stages, the ponors may also function as springs, hence becoming estavelles.

By contrast, Gams' (1978) definition of the hydrological characteristics of poljes is both simpler and more strict. The drainage must consist of a sinking inflowing river and karstic outflows, and the imbalance between these two components must result in the intermittent inundation of the polje floor.

In general, the Morecambe Bay mosses are characterised by marginal springs whose waters either flow directly out to sea, or which drain to allogenic rivers flowing across the mosses. Only in the case of Witherslack is outflow totally karstic in nature. On the other hand, it is apparent that all the mosses, with the exception of Witherslack, would be subject to seasonal inundation were it not for the artificial drainage to which they are subjected. However, this flooding is a function of the low-lying nature of the mosses, rather than the inefficiency of the karstic outflow system as in the case of true karst poljes.

8.3.4 Process

The fundamental process of polje development is one of solutional planation of the polje floor and sides. As poljes tend to act as catchments for non-calcareous material, Roglič (1939) envisaged planation occurring as a result of the concentration of solution at the polje floor and sides beneath the unconsolidated fill. Jennings (1971, 142) also pointed out the possible influence of solution by static water bodies at the polje margins.

Although Ashmead regarded the "phreatic network" caves which fringe the mosses of the Morecambe Bay area as evidence of the work of lateral solution, the present study shows that, with the possible exception of the network caves around Burton Moss, there is little support for this view (see 7.2). Nor is there evidence of planed limestone surfaces beneath the mosses of the area. The only survey of bedrock morphology beneath the mosses was that of Gresswell (1958, 102-103) in the Lyth Valley, where seismic surveys showed that the valley had obviously been subjected to glacial over-deepening, the bedrock occurring at depths of over 60 m below the moss surface.

8.3.5 Conclusions

Although the mosses of the Morecambe Bay area have the form of poljes, they do not display the hydrological characteristics of true karst poljes. Furthermore, it is difficult to be certain whether any solutional planation processes are at work in the mosses, no confirmatory evidence having been found. Cvijič (1960), however, has described features, both in the Dinaric karst and elsewhere, which are fully comparable with poljes, except that they have external surface drainage along a narrow valley. He has termed these features "open poljes", a description which could be applied to certain of the features of the Morecambe Bay area, in particular Hawes Water, Burton Moss and the Winster Valley. Given Cvijič's definition, there is no necessity for karst processes to operate in the polje. Hence, since numerous workers, including Roglič (1965) and Sweeting (1972, 207), have regarded poljes as essentially non-karstic forms, perhaps Cvijič's definition is acceptable and applicable to the features of the Morecambe Bay area.

8.4 The closed depressions of the Morecambe Bay karst

8.4.1 Introduction

The closed depression has long been regarded as the fundamental unit of karst relief. Within the Morecambe Bay area these features are limited in number in comparison with other karst areas (Fig. 8.4), but they display a wide range of forms, the largest of which are comparable to any found elsewhere in the British Isles. Despite this, the large closed depressions of the area have rarely been mentioned in the literature. In order to remedy this, the following depressions, comprising all the large (>100 m diameter) closed depressions in the Silverdale area, were selected for study:

Hazelslack (SD476785)
Deepdale (SD495786)
Burton Well (SD470753)
Cringlebarrow (SD497749)
Wood Well (SD463747)
St. John's (SD468745)
Three Brothers (SD494733)

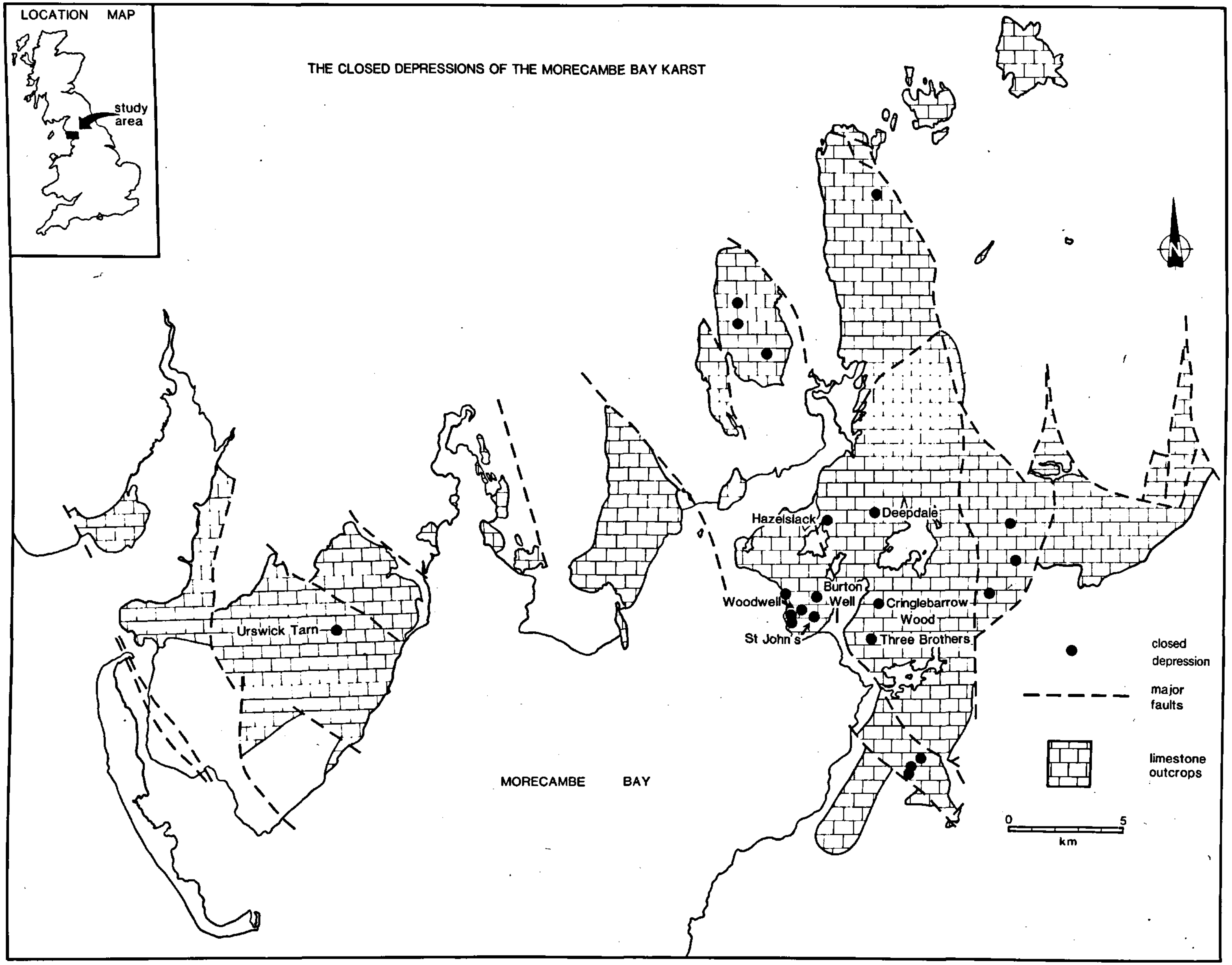


Fig. 8.4 The closed depressions of the Morecambe Bay karst

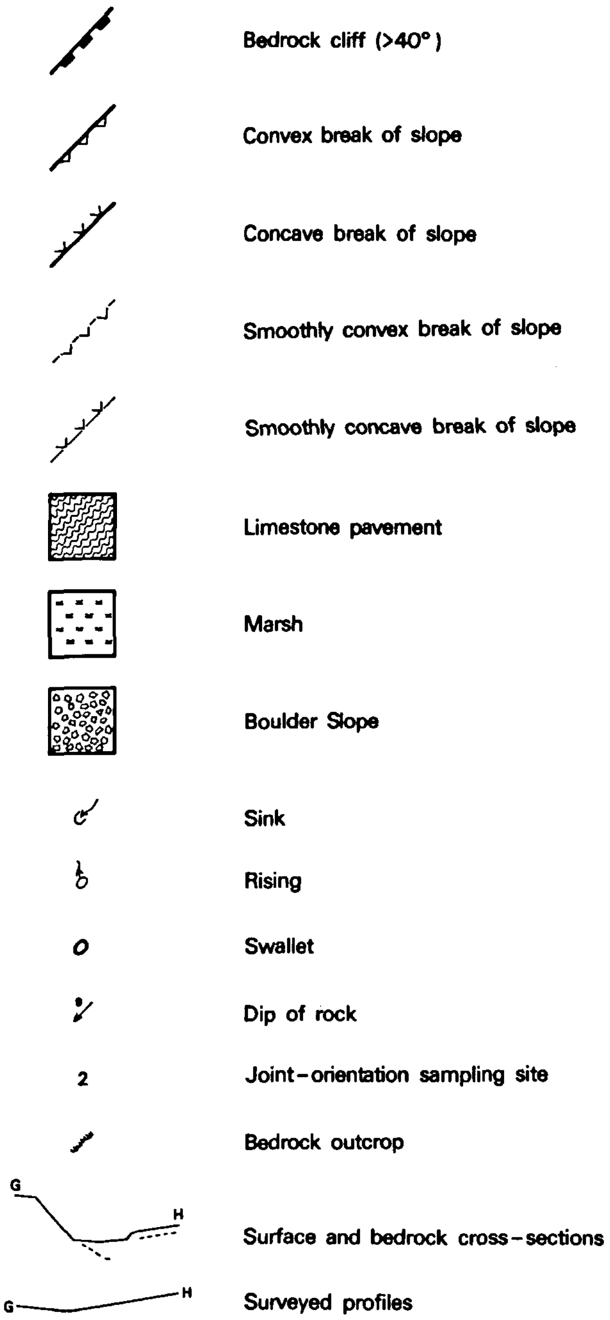
8.4.2 Field study

For each depression, long and cross-profiles were surveyed (see 11.4.3), morphological maps drawn (see 11.4.4), structural and joint orientation measurements made on all bedrock exposures (see 11.1.1) and the bedrock form investigated by both augering and geophysical methods (see 11.5). The results of the field study are shown in Fig. 2.5 and Figs. 8.5-8.11.

8.4.3 Structural control on closed depression form and location

Despite the detailed study of joints within the closed depressions, overall closed depression form does not appear to be strongly related to joint orientation. Only in the case of Burton Well is the distinct joint alignment peak at 120-129^{o} reflected in the orientation of the depression. At Three Brothers, the N-S trend of the depression is mirrored to some extent by joint orientations, but even so there is a wide scatter of significant orientations between 150^{o} and 190^{o}.

In a wider context, however, all the depressions have a structural control on their form and location. This may be seen in the preferential alignment of depressions along the strike of the rock. Burton Well, Wood Well, St. John's and Three Brothers may all be regarded as "strike hollows" (Williams, 1968) whose cross-sectional form is essentially the result of structural control. These features are probably the result of a combination of glacial and karstic processes. The characteristic scarp and bench form of the strike hollows is likely to have been the result of glacial



KEY: Figs. 8.5–8.11

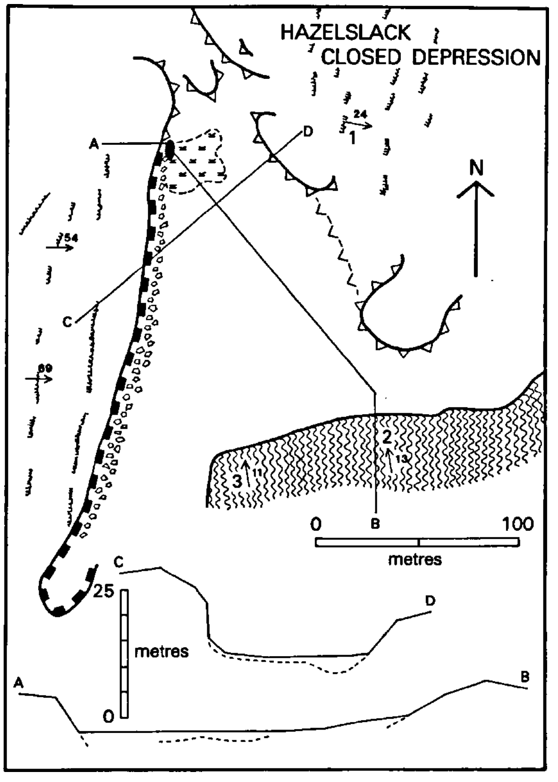


Fig. 8.5 Hazelslack closed depression

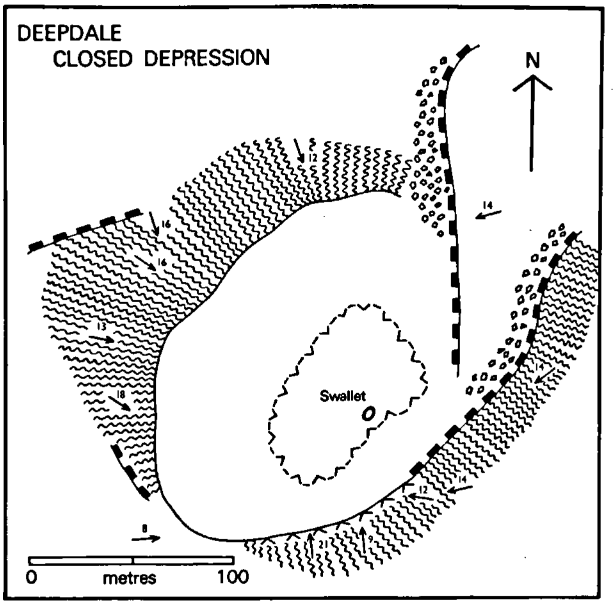


Fig. 8.6 Deepdale closed depression

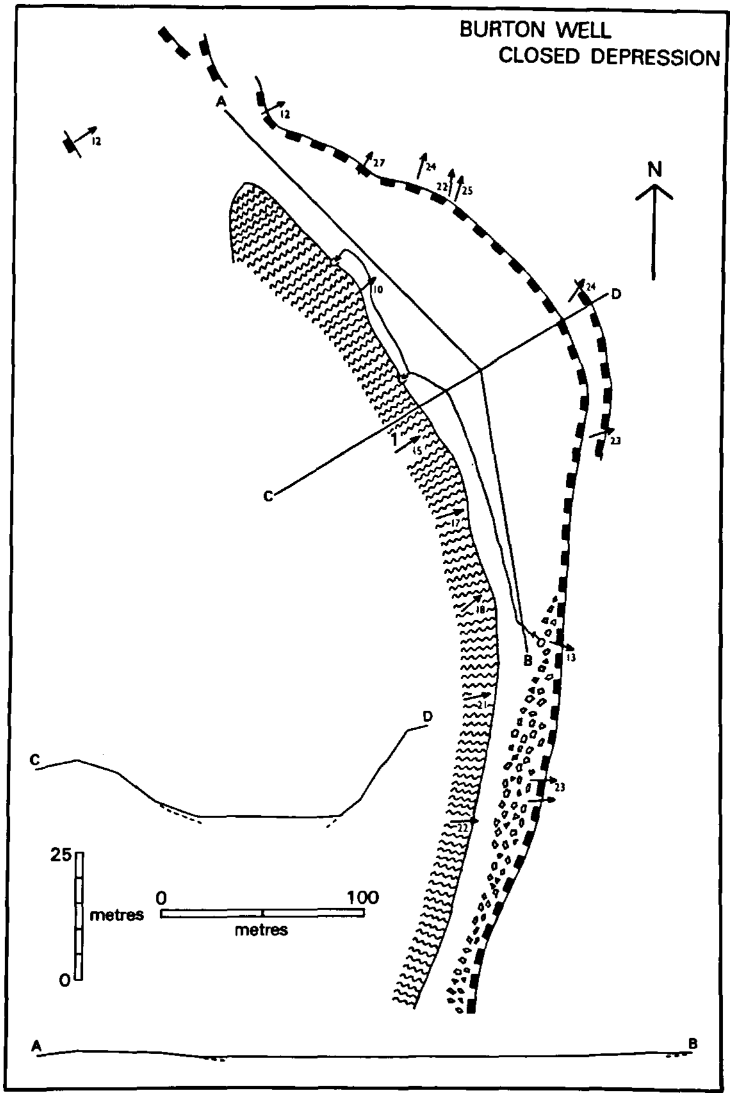


Fig. 8.7 Burton Well closed depression

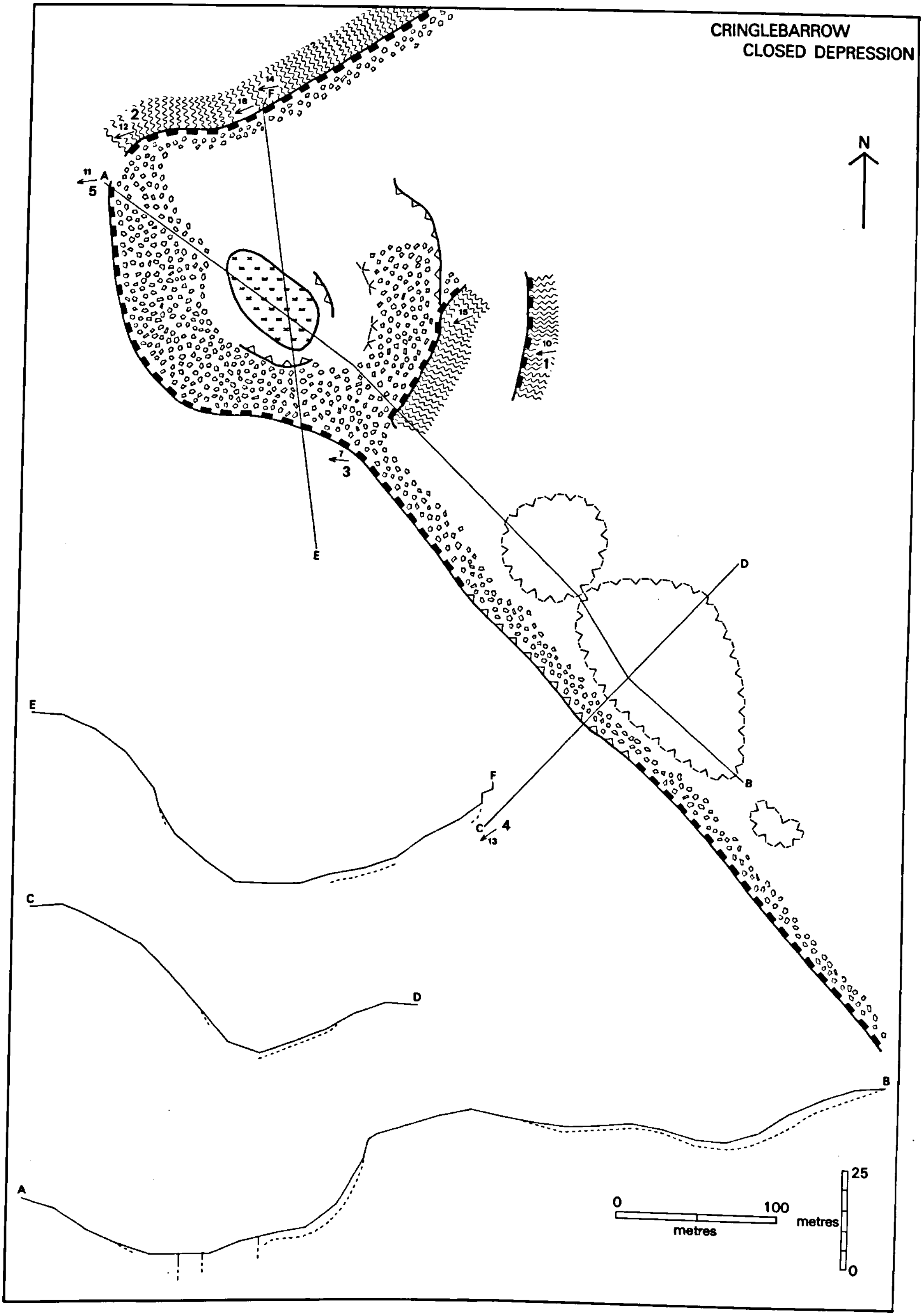


Fig. 8.8 Cringlebarrow closed depression

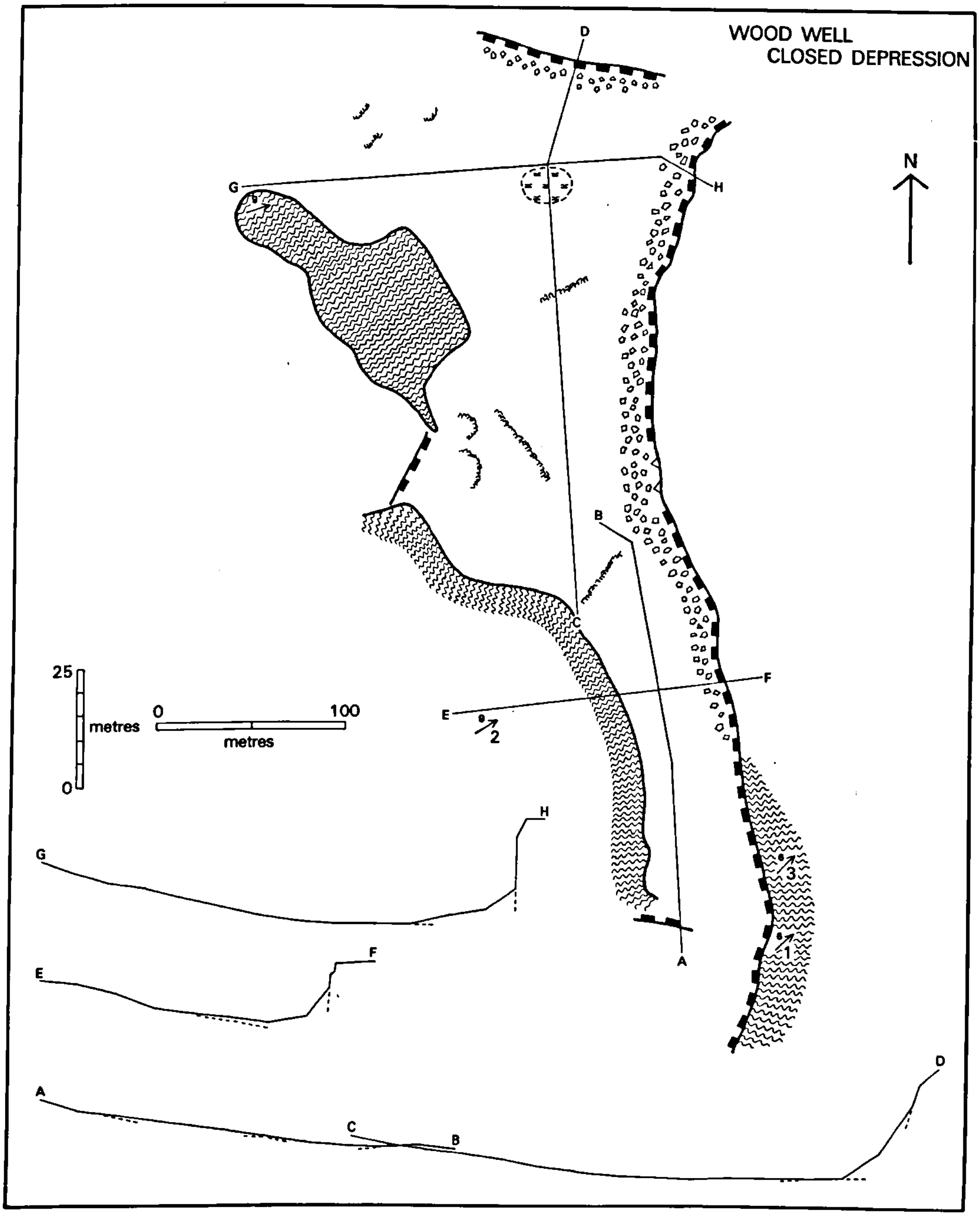


Fig. 8.9 Wood Well closed depression

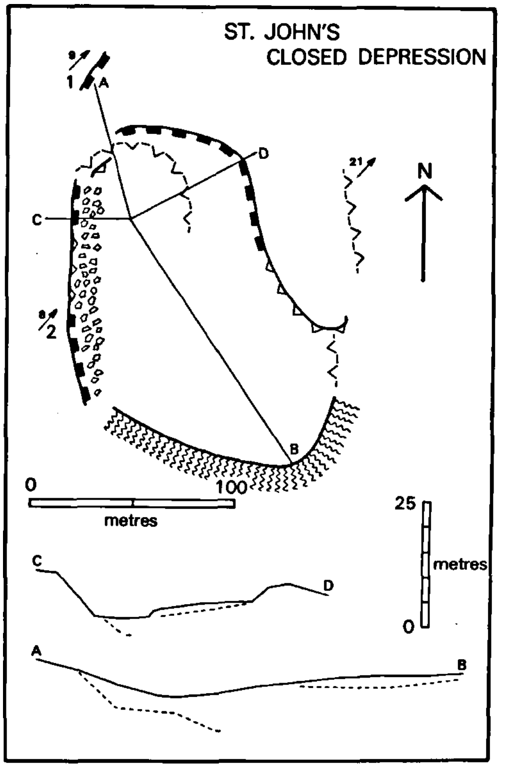


Fig. 8.10 St. John's closed depression

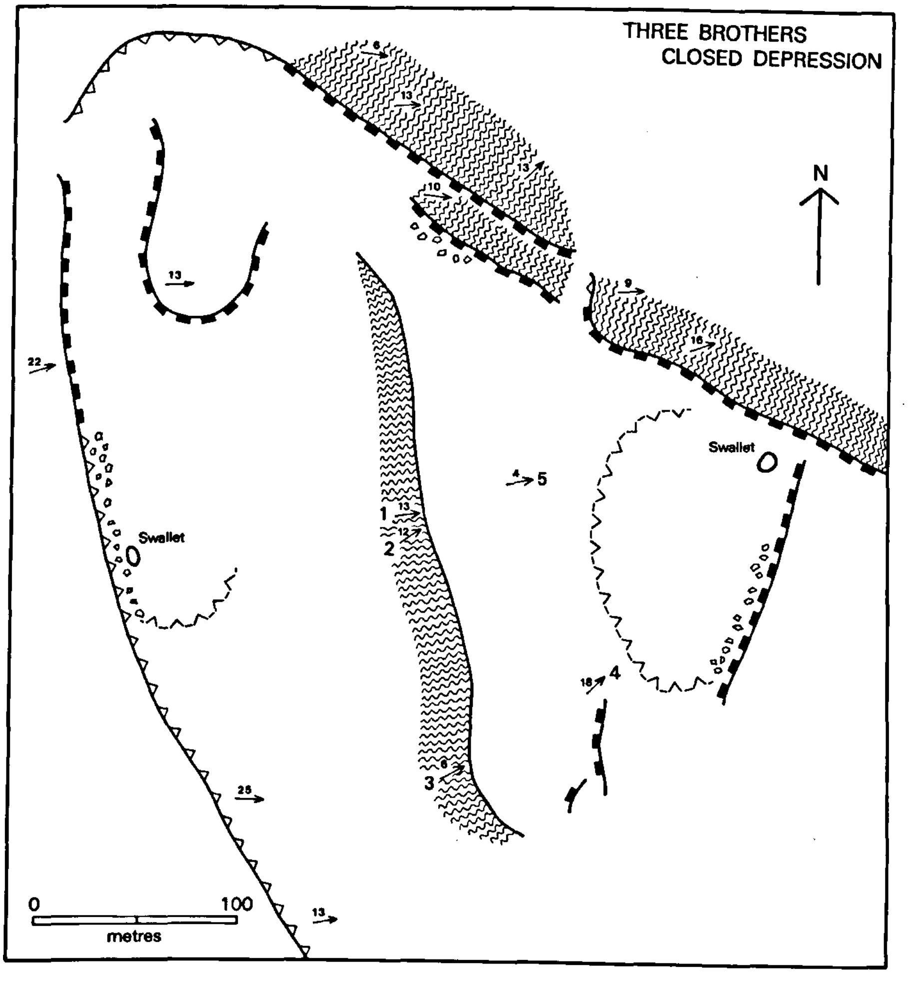


Fig. 8.11 Three Brothers closed depression

scour, as proposed for similar features elsewhere in the Morecambe Bay district (see 8.2.4). Once formed, the strike hollow would act as a focus of drainage, encouraging solution at the bottom of the depression and the development of karstic outflow. In most of these features, the basic structural form has been only little modified by subsequent karstic processes, although at Three Brothers solution seems to have resulted in the development of a depression of considerable depth.

Of the remaining depressions, both Cringlebarrow and Hazelslack appear to be fault-controlled features. Hazelslack is developed along the line of the Silverdale monocline, with vertical beds to the west and gently dipping beds to the east. At Cringlebarrow, evidence of faulting is less conclusive. Nevertheless, the cross-sectional asymmetry of the whole depression complex and its NW-SE orientation parallel to at least three other faults further south on Warton Crag, including the major Warton Crag fault, suggest a NW-SE fault with its downthrow to the northeast (Fig. 8.8).

Faults can have a variety of effects on subsurface drainage in limestone (see Waltham, 1971b, Ford and Worley, 1977) most of which encourage drainage either along or adjacent to the fault. Thus, solution, cavity formation and collapse take place preferentially along fault lines, the surface expression of these processes being features such as Hazelslack and Cringlebarrow closed depressions.

The final depression studied, Deepdale, differs from the others in that its form and location appear to be closely related to the dip of the rock, the depression being developed within a small synclinal basin (Fig. 8.6). The syncline presumably encouraged internal drainage with the consequent development of a closed depression as a result of the concentration of solution.

8.4.4 Interpretation of seismic survey results

In all cases, the seismic survey results indicate a simple two-layer situation in the depressions, an upper layer having seismic wave velocities of 100-700 m s^{-1} and a lower layer having velocities of 1300-3500 m s^{-1}. The interpretation of this pattern as unconsolidated fill overlying bedrock was corroborated by the results of augering at Burton Well, Cringlebarrow and Wood Well.

Typically, the upper layer consists of unsorted material ranging in size from boulders to clays. Analysis of a sample of the upper layer from St. John's (sample 38) showed it to fall within the near-mountain source glacier environmental envelope of the QDa-Md diagram (Fig. 5.1), whilst the range of seismic velocities indicates that the fill consists of unconsolidated superficial deposits (Kesel, 1976, 95).

Earlier studies of seismic wave velocities within the Carboniferous Limestone of the area suggest a range of values of 3000-6200 m s^{-1} (Terrasearch Ltd., 1968): generally higher than those found in the present study. However, the higher values are of buried and unweathered rock, whereas it is likely that the bedrock within the depressions is both weathered and solutionally fretted along joints, the resultant grykes being infilled by lower velocity overburden. These conclusions are supported by comparative data on seismic wave velocities in weathered limestone of 2000-3500 m s^{-1} and in solid limestone of 3500-5000 m s^{-1} (Bison Instruments Ltd., 1971, 2).

The bedrock form of the depressions tends to support the structural interpretation of 8.4.3. Thus, the strike hollows of St. John's, Wood Well and Burton Well all exhibit only a veneer of fill overlying a form which is in essence structurally controlled. On the other hand, the main depression of Cringlebarrow shows considerable evidence of solutional development.

8.4.5 Conclusions

Although a number of the large closed depressions of the Morecambe Bay area, in particular the strike hollows, have not been formed solely by karstic processes, all the depressions exhibit karstic drainage and have been modified to a greater or lesser extent by solutional processes. These features may be numbered amongst the largest and most spectacular of their kind in the British Isles. Cringlebarrow, in particular, with a diameter of ~250 m and a depth of 30-40 m is comparable to the largest depressions of the Malham Tarn area (Moisley, 1954).

8.5 The cemented screes of the Morecambe Bay karst

8.5.1 Introduction

Cemented screes (Plate 8.6) have been found at a number of locations in the Morecambe Bay area: Kent's Bank (SD392755), Buckhouse Scar (SD449846), White Scar (SD460852), Arnside Knott (SD457772), Hawes Water (SD476768) and Barrow Scout (SD482728). In every case the cemented screes are

Plate 8.6 Cemented Screes at Buckhouse Scar (SD449846)

found below limestone free-faces, at or near the foot of the limestone hills of the area. It is likely that other deposits occur in similar situations elsewhere, but the presence of an overlying soil cover means that this cannot be stated with certainty.

The only previous description of the cemented screes is that of Sweeting (1970, 239; 1972, 301; 1973, 106; 1974, 78) who interpreted the screes as of glacial or periglacial origin and Pleistocene age, and the underlying "smoothed and potholed" bedrock surface "associated with reddish soil or clay" as representative of an earlier phase of karstification. In one paper (1970, 239) it was suggested that the bedrock surface may have dated from "the great Interglacial in the Pleistocene or more likely a warmer phase during the Tertiary".

8.5.2 Description

The screes consist almost totally of angular to sub-angular limestone fragments ranging in size from approximately 1 to 15 cm diameter. Occasionally, as at Buckhouse Scar, subrounded pebbles of erratic origin are included within the scree, most probably having been derived from upslope glacial deposits.

The scree displays no apparent fabric, although beds of different sized material can sometimes be seen to overlie each other. At Buckhouse Scar, for example, a bed of median particle size ~5 cm overlies one of median particle size ~1 cm, the interface dipping at approximately the angle of the underlying slope.

Within the screes the particles are in point contact with one another. In all cases the screes have been subjected to a post-depositional phase of carbonate cementation which has frequently led to the particles being coated with flowstone and the point cementation of the scree. In some cases the carbonate precipitation has also led to the development of a delicate, fibrous-like precipitate on the underside of the particles. Although the cemented screes often form an openwork structure, elsewhere the interstices have been infilled by a variety of materials, including glacially derived deposits, as at White Scar, and soft calcareous cement, as at Arnside Knott and Buckhouse Scar.

At all the sites, the cemented scree either is now or was formerly overlain by the Brown Calcareous Soils (Hall and Folland, 1970, 68) characteristic of the area. Generally the soil is only thinly developed, although it characteristically supports the growth of Taxus and Betula above the deposits of cemented scree.

8.5.3 Interpretation

The cemented screes clearly post-date the last glacial episode in the area, for the exposures at Buckhouse Scar, White Scar and Arnside Knott all overlie either striated limestone bedrock, till or both. The existence of striations on the underlying bedrock (Plates 8.1 and 8.2) makes Sweeting's (1970, 239) suggested Tertiary karstic origin of the bedrock surface (see 8.4.1) highly unlikely.

The location of all the deposits at the foot of slopes beneath limestone free-faces suggests that they are gravitationally derived from bedrock exposures. The breakdown of limestone cliffs in this fashion may be the result of a variety of wedging effects and the climatic significance of scree must be interpreted with caution. Nevertheless, given the stratigraphic relationship of the screes to evidence of glacial action, it is highly likely that their formation was the result of frost wedging under periglacial conditions. This is supported by the laboratory work of Guillien and Lautridou (1970) which indicates that frost wedging of certain limestones can produce the same kind of material as found in the Pleistocene _grézes litèes_ of Charente. A climatically controlled origin of the cemented scree is also suggested by the low rate of scree production in the area under present conditions (see also Sweeting, 1966, 182-183 for present rates of scree formation in north west Yorkshire).

Hence, the most active phase of scree formation is likely to have been during the Late-glacial period of periglacial activity. This phase must have been completed by the time of deposition of the overlying soil and the cementation of the screes. The process of cementation demands percolation of water through the deposit, and, although this is not incompatible with periglacial conditions, the fact that cementation occurred subsequent to scree accumulation suggests that the cementation process was associated with an amelioration of climate. It is arguable that the cementation phase post-dated the deposition of the overlying soil and this possibility is indicated by the presence of a layer of brown clay-silt overlain by calcite on some of the scree fragments at Arnside Knott, and by continuing percolation and calcite precipitation at White Scar. Elsewhere, however, there is evidence that the overlying soil has been prevented from infilling the scree interstices by the prior cementation of the particles.

Although it is tempting to try to relate the deposition of the overlying soil and the cementation phase to an amelioration in climate, there is little evidence for this. The soil is equally likely to owe its existence to slope wash resulting from cultivation and grazing on the deeper soils of the relatively flat hilltops of the region. This appears to be the case in southern England where similar scree/carbonate precipitate/soil sequences can be found, and where the soil contains artifacts ranging from the Neolithic to Iron Age and, exceptionally, later periods (Evans, 1978, 97-100).

8.5.4 Conclusions

Cemented screes are rarely described features of periglacial and glaciokarst areas (see Judson, 1949; Prentice and Morris, 1959; Burek, 1977, 93-95), although it is clear that they are ubiquitous in such areas. Apart from their significance as a morphological feature, the cemented screes of the Morecambe Bay area constitute a useful stratigraphic horizon which may be used to relate depositional sequences inside caves to those on the surface (see 5.3 and 7.3.3).

9. DISCUSSION

9.1 The reconstruction of karst palaeoenvironments : methodology and philosophy

Karst regions possess two unique advantages over other landscapes from the point of view of palaeoenvironmental reconstruction. Firstly, they may preserve the fine detail of past hydrological conditions in the form of caves eroded in the bedrock. Thus, details of former meanders, scallops, cut-offs and so on remain etched in the rock as the cave stream cuts down. By contrast, surface streams continually rework their channels so that only fragments of earlier systems are preserved to enable us to infer past hydrological conditions. Secondly, once formed, these caves and their surface counterparts, closed depressions, may function as sediment traps, preserving sedimentological evidence of underground and surface events away from the disturbing influences of subaerial and biological activity.

Despite these very considerable advantages, very little use has been made of karst environments in palaeoenvironmental studies. Partly this has been because of the lack of a general appreciation of the advantages of karst, and partly it is because of the natural barriers to investigation which caves, in particular, present. The Morecambe Bay study has therefore attempted to capitalise on these advantages and to investigate the various approaches which might be of value in a study of this nature. In the course of the study the drawbacks of many of the techniques and some of the disadvantages of karst landscapes for palaeoenvironmental work have also been recognised and will be discussed below.

The ultimate aim of the study has been to draw together information from a variety of sources, distributed over space and time, in an effort to establish both a chronology of karstification in the Morecambe Bay area and a reconstruction of the environment during each phase of karstification. Given the timescale under consideration and the diversity of environments studied, it is clear that no single technique could alone provide the information that was required. A wide range of methods were therefore adopted in an attempt to obtain as much information as possible from the available evidence. The major source of palaeoenvironmental information is that derived from the sedimentological record. However, since karstic sediments are extremely diverse in character, a wide range of sedimentological techniques had to be assembled to deal with them. It was therefore found necessary to investigate their macro and microscopic physical nature, their mineralogy and petrography, their magnetic properties, their sedimentary structure, their faunal content, any cultural artifacts they might contain and their extent as determined by the use of geophysical methods. The other main source of palaeoenvironmental information is that derived from fossil landforms, which may provide relatively precise information about past environments. The various sources of such information included fossil screes, interstratal and buried karst forms, caves and palaeokarst surfaces.

In applying these techniques, it is inevitable that as one moves towards the present, more evidence of past environments is available and a wider range of techniques can be applied. To some extent, therefore, reconstructions become less uncertain and chronologies more exact. Nevertheless, certain problems do arise. In the case of the Morecambe Bay area, the sources of more recent environmental history are so diverse and form such discrete entities that it becomes difficult to piece together the complex fragments of evidence into a single picture. This contrasts markedly with the elegance, simplicity and, doubtless, incompleteness of the reconstructions derived from the few strands of evidence available as one moves back in time.

Despite the wide range of methods adopted in the study, the work hinges on the application of two concepts, those of stratigraphy and uniformitarianism. The stratigraphic approach may be considered first. This provides the fundamental chronological basis of the work, whether establishing the age of palaeokarst surfaces, the sequence of development of sops, or the date of cave fills. One of the major drawbacks of this method is illustrated in its application to cave fills. By their very nature, depositional sequences in caves can rarely be easily correlated with depositional sequences elsewhere. This can only be achieved either where chronostratigraphic correlations can be established, or where cave deposits can be laterally correlated with surface deposits, themselves traceable between caves. In no case in the Morecambe Bay area has it yet been possible to achieve either of these correlations. This has meant that lithostratigraphic and biostratigraphic approaches have had to be adopted. Although both lithostratigraphy and biostratigraphy can be successfully correlated in almost every cave in the Morecambe Bay area, these approaches face two major difficulties. The first is that of chronology. The traditional means of dating, by means of pollen, mollusca, etc., are hopelessly inadequate and have a very low time-resolution. Almost exclusively, therefore, the burden of dating and correlation falls on radiometric and geomagnetic methods. Unfortunately, these techniques are to a large extent dependent on finding suitable material to date, and this has posed considerable problems in the Morecambe Bay study. Moreover, even when available, the reliability of such geochronometric dates is open to question. In this study, both geomagnetic and radiometric results have been utilised, but in each case the results are subject to difficulties either of error or multiple interpretation. The result of this is that the entire sequence of Late Quaternary cave fills is dated by reference to the sedimentological succession in Kirkhead Cave, which has itself been dated on the evidence of a succession of cultural artifacts and a single ^{14}C date, itself possibly subject to hard water error. It is true that upper parts of the depositional sequences in other caves have been dated on cultural grounds, but it is essentially on the evidence of Kirkhead Cave that ages have been assigned to lithologically and biologically similar sequences in other caves in the area. This leads to the second difficulty. The existence of a lithological and biological sequence in a cave is no evidence of chronology, particularly since, given a succession of glacial and interglacial environments, one would expect similar lithological and biological sequences to recur over the time. It is both simple and attractive to date a sequence of this sort by "count from the top" methods, particularly, as in the case of the Morecambe Bay area, where this allows sequences to be matched and where the few chronological markers that are available sit neatly in sequence. Nevertheless, such an approach is riddled with assumptions and is intellectually unexacting. Even worse is the technique of force-fitting a sequence into preconceived chronological pigeonholes. Perhaps the best that can be achieved therefore, is the development of floating chronologies which can be slotted into sequence and which are at least amenable to changes in local, regional and national chronological knowledge.

Fewer of these difficulties of dating and correlation between sites exist for earlier phases of karstification, mainly because both the degree of temporal resolution and the number of sites are significantly reduced. In these cases, however, stratigraphic reconstruction is reliant on laterally-limited information from borehole logs, from rare exposures in the bedrock record, or from stratigraphic reports, particularly from mines, which can no longer be confirmed by inspection. Moreover, with older phases of karstification, the problems of chronology becomes more intractable. The ?Oligo-Miocene phase of karstification, for example, is dated merely on the basis of the relationship of mineralised karst features to faults regarded as of Alpine age.

The second major conceptual approach used in the study is that of uniformitarianism. Although the concept of uniformitarianism is invaluable in any chronological study, it is not always easy to apply, particularly since many of the features considered, such as palaeokarst surfaces and sops, have no modern analogues. In these circumstances, palaeoenvironmental reconstruction must be at least partly based on speculation. But even in more favourable circumstances, the application of a knowledge of modern processes to an understanding of processes working in the past is not necessarily straightforward. Firstly, the untestable assumption must be made that uniformitarianism is a reasonable concept; that is, that present processes worked in the same way in the past. Secondly, the concept assumes that a full understanding of contemporary processes is available. This is most certainly not the case. Even where processes have been studied from physical first principles, as in the case of scallop development, the only conditions which have been considered are those of steady uniform flow in simple systems, and it is questionable whether the results of such studies may be applied to the transient and non-uniform conditions of reality. In other cases, the understanding of process is based simply on empirical relationships, as in the case of the statistical relationship found between cave-meander form and discharge. Thus, considerable doubt must be case on the results of the magnitude of past processes from the application of relationships derived from poorly-understood contemporary processes.

9.2 The integration and correlation of phases of palaeokarst development in the Morecambe Bay area with those in the rest of the British Isles

9.2.1 Introduction

With the exception of work by T.D.Ford (1964; 1977) in Derbyshire, there have been no previous attempts to provide an integrated picture of karst development over time in any part of Britain. Yet, as shown in Table 4.1, there is considerable evidence of karstic development in Britain throughout the Mesozoic and Cenozoic. Furthermore, although karstification seems to have taken place almost continuously in one place or another in the British Isles since the Lower Carboniferous, there are certain periods during which the country seems to have experienced widespread karstification, suggesting a picture of an almost nationwide karst landscape. To a large extent, these periods coincided with the major periods of karstification in the Morecambe Bay area. It is instructive, therefore, to contrast the Morecambe Bay experience with the situation elsewhere in the country during each phase of karstification, and hence to build up a picture of the karst landscape in Britain during each karst episode.

9.2.2 Lower Carboniferous karstification

Those phases of Lower Carboniferous karstification which have been identified in the Morecambe Bay area from the evidence of palaeokarst surfaces can be closely related to the regression maxima of the Lower Carboniferous sea proposed by Ramsbottom (1973). Since fluctuations in Lower Carboniferous sea level seem to have occurred contemporaneously over the whole of Britain, it might be expected that similar episodes of karstification would have occurred at the same times throughout the country. Palaeokarst surfaces have been reported from the Lower Carboniferous Limestone of north and east Cumbria (Mitchell, 1978), Yorkshire, Lancashire, Mendip and Fife (Walkden, 1974), but it is only in Wales and Derbyshire that adequate stratigraphic information is available to allow the date of karstification to be determined. In north Wales, Power and Somerville (1975) and Somerville (1979a; 1979b) found palaeokarst surfaces in limestones of D_1 (Asbian) and D_2 (Brigantian) age, which they considered to correspond to the regression maxima of Ramsbottom's (1973) minor cycles (Table 3.1) In Anglesey, Baughen and Walsh (1980) reported sandstone-filled solution pipes developed in limestones of D_1 (Asbian), D_{2a}

and D_{2b} (Brigantian) age. Atkinson (1980) subsequently suggested, however, that these features had been formed by a process of mixing solution under a thin cover of sand at the sea bed. In south Wales, Thomas (1953) interpreted marl-filled solution pipes in D_1 (Asbian) limestones as the result of two episodes of subaerial solution, presumably also corresponding to minor-cycle regression maxima. In Derbyshire, Walkden (1974) found that palaeokarst surfaces were best developed in middle and upper D_1 (Asbian) limestones which had been deposited in a shelf environment; the surfaces were unknown in areas of basinal limestone. Similar features were reported from the same area by Stevenson and Gaunt (1971) in the Upper Monsal Dale Beds of D_2 (Brigantian) age.

In many areas of the country, an end-Lower Carboniferous phase of karstification also seems to have occurred, although there is no evidence for this in the Morecambe Bay area. Thus, in Derbyshire (T.D.Ford, 1964, 53; 1977, 54-55) and Monmouthshire (Dixon, 1909), beds of Namurian Sandstone unconformably overlie a solutionally-eroded karst surface. However, similar features elsewhere in south Wales have been attributed to the subaquatic introduction of quartzitic material into unconsolidated calcareous sediment (Owen and Jones, 1961).

9.2.3 Permo-Carboniferous karstification

The only traces of the landscape thought to have developed in the Morecambe Bay area as the result of karstification during the Permo-Carboniferous occur in the Low Furness area. Elsewhere, exhumation and exposure to subaerial processes has taken place and any discussion of the former nature and extent of the landscape can be no more than speculative. Nor can comparisons easily be made with other British karst areas. The occurrence of limestone pebbles in Permo-Triassic deposits in Derbyshire suggests that the limestone was exposed to subaerial processes at this time, possibly, as in the case of Morecambe Bay, due to tectonic uplift and erosion of the overlying impermeable cover (T.D.Ford, 1964, 54-55). Elsewhere in Britain, however, such a landscape either failed to develop, since in most areas end-Carboniferous erosion was inadequate to remove the cover of the Upper Carboniferous rocks, or has been subsequently eroded.

9.2.4 Tertiary karstification

The evidence for Tertiary karstification in the Morecambe Bay area is restricted to Low Furness. Elsewhere in the area any conclusive evidence of Tertiary landforms has been removed by subsequent denudation, although mineralised karst features in the Silverdale area might well represent fragments of the same karst phase.

The Tertiary karst of the Morecambe Bay area is particularly important since it appears to have developed contemporaneously with many other karst features throughout the British Isles (see Table 4.1). This suggests that uplift and accelerated erosion associated with the gathering climax of the Alpine tectonic storm resulted in an increased fissure density in the limestone and a greater hydraulic gradient. Thus, the limestone was exposed to subaerial processes and the circulation of meteoric waters through the rock was encouraged.

More importantly, these karst features provide almost the sole evidence of Upper Tertiary environmental conditions in Britain. Elsewhere, indubitable Upper Tertiary deposits are restricted to the Suffolk Bone Beds of East Anglia, the Coralline Crag of southeast England, and the St. Erth Beds of southwest England (Curry _et al_., 1979, 50-51). With these exceptions, all the features previously placed in the Upper Tertiary have been so assigned on the basis of either unreliable evidence or dubious reasoning. These include deposits given an Upper Tertiary age simply on the grounds of their weathered condition or their "preglacial" stratigraphic position (Walsh _et al_., 1972, 543-548); and landforms regarded as the result of Upper Tertiary geomorphic processes, particularly marine and subaerial planation. "Erosion surfaces", considered to have resulted from such planation have been described both in south Cumbria, where three surfaces of Mio-Pliocene age have been recognised (Parry, 1958; 1960), and elsewhere in Britain (see, for example, Wooldridge and Linton, 1955; Brown, 1960; Sissons, 1960). The methodology by which these surfaces were identified has been largely discredited (Chorley 1965a; 1965b, 148-149) and wherever they have been reinvestigated it has been found difficult either to identify or to date them (see, for example, Hodgson, Rainer and Catt, 1974; Jones, 1980; and 8.2).

9.2.5 Quaternary karstification

Every karst area in the country shows evidence of karstification having taken place during the Quaternary. In the Morecambe Bay area, at least two phases of cave and karst development can be recognised. The earlier of these occurred after approximately 100 m of valley incision, possibly by glacial action. Surface flow concentration seems to have taken place on the impermeable rocks overlying the limestone prior to flow sinking into the aquifer. During this phase, resurgence levels were approximately 5 to 10 m above present valley levels.

The later phase of karst development is that which can be seen at the present, valley levels having been further lowered and the impermeable capping to the limestone removed, in both cases probably largely by glacial action. Thus, present aquifer recharge is mainly by diffuse flow, and groundwater resurges either at valley level of, in the case of those aquifers fringing Morecambe Bay, below the level of mean high-water mark of spring tides.

Similar sequences of development can be identified elsewhere in Britain. In northwest Yorkshire, two phases of valley lowering, the first of up to 75 m and the second of 7 to 20 m, have been recognised, each having led to rejuvenation of the groundwater drainage system. In both cases it is suggested that valley incision was the result of glacial erosion. $^{230}Th/^{234}U$ dates on stalagmite in the caves indicate that the earlier phase of incision had been completed by 400,000 B.P. (Atkinson, Harmon, Smart and Waltham, 1978).

In Mendip, fossil resurgence caves occur up to 70 m above present resurgence levels, and at Cheddar at least three former resurgence levels can be recognised. The highest of these seems to have been active prior to 360,000 B.P. Subsequent valley incision resulting in the lowering of resurgence levels is thought to have been the result of both fluvial erosion and torrential erosion under periglacial conditions (Atkinson, Harmon, Smart and Waltham, 1978).

In Derbyshire, no cave deposits have yet been dated radiometrically, although it is clear from the Lower Pleistocene fauna found in Victory Quarry Fissure, Dove Holes (Spencer and Melville, 1974) that vadose inlet caves were active in the area at least 600 000 years ago. Much of the sequence of cave development can therefore only be inferred from geomorphic evidence. In the Castleton area, T.D.Ford (1966; 1977, 339-342) proposed a sequence of valley incision and retreat of the impermeable cover from the limestone. This resulted in the progressive lowering of resurgence levels and the progressive abandonment of vadose inlet caves.

Cave systems thus seem to have followed broadly the same sequence of development in all the main karst areas of Britain during the Quaternary. Nevertheless, any attempt to relate phases of cave development, cave infill and resurgence lowering from region to region is fraught with difficulties. Partly, this is because no adequate geochronometric control is yet available for events in each area. Perhaps more importantly, however, it is clear that the same processes were not necessarily at work contemporaneously in each area. Northwest Yorkshire experienced the full brunt of successive glacial events, whereas Mendip appears never to have been glaciated, and Derbyshire formed an unglaciated enclave at least during the last glacial. On this basis, the experience of the Morecambe Bay area can perhaps be most closely related to that of northeast Yorkshire. Nevertheless, it is probably taking the available evidence too far to suggest that phases of valley incision and resurgence lowering have a one to one relationship between the two areas, despite the similarity between the vertical ranges of the two phases of valley incision and resurgence lowering recognised in each area.

10. CONCLUSIONS

The aims of the present study have been twofold; firstly, to reconstruct the history of karstification in the Morecambe Bay area, and secondly, to investigate the particular environments under which karstification has occurred. More than any others, karst landscapes exhibit a high degree of structural control (sensu lato). Any investigation of this sort must, of necessity, therefore, take account of the influence of both structural geology and lithology on past and present landform development. At the smallest scale, for example, joints and bedding planes act as the loci of solution and provide the main control on hydraulic conductivity within the rock. Nevertheless, in the Morecambe Bay area, these features exert only minor influence on geomorphology. Instead, folds and, in particular, faults provide the primary control on the form of the landscape, influencing the location and morphology of major karst features such as sops, poljes and closed depressions, as well as the overall distribution of high and low ground in the area.

Lithological variations are also of significance. The Lower Carboniferous Limestone in the area consists of seven distinctive lithological units whose distribution controls to a remarkable degree the extent and style of karstification around Morecambe Bay. Neither the largely non-calcareous units, such as the Basement Beds and the Gleaston Group, nor the poorly-bedded and thinly-jointed Park Limestone generally support karst development. On the other hand, the massively bedded and jointed, and relatively pure Urswick Limestone constitutes the most important karst lithology in the area.

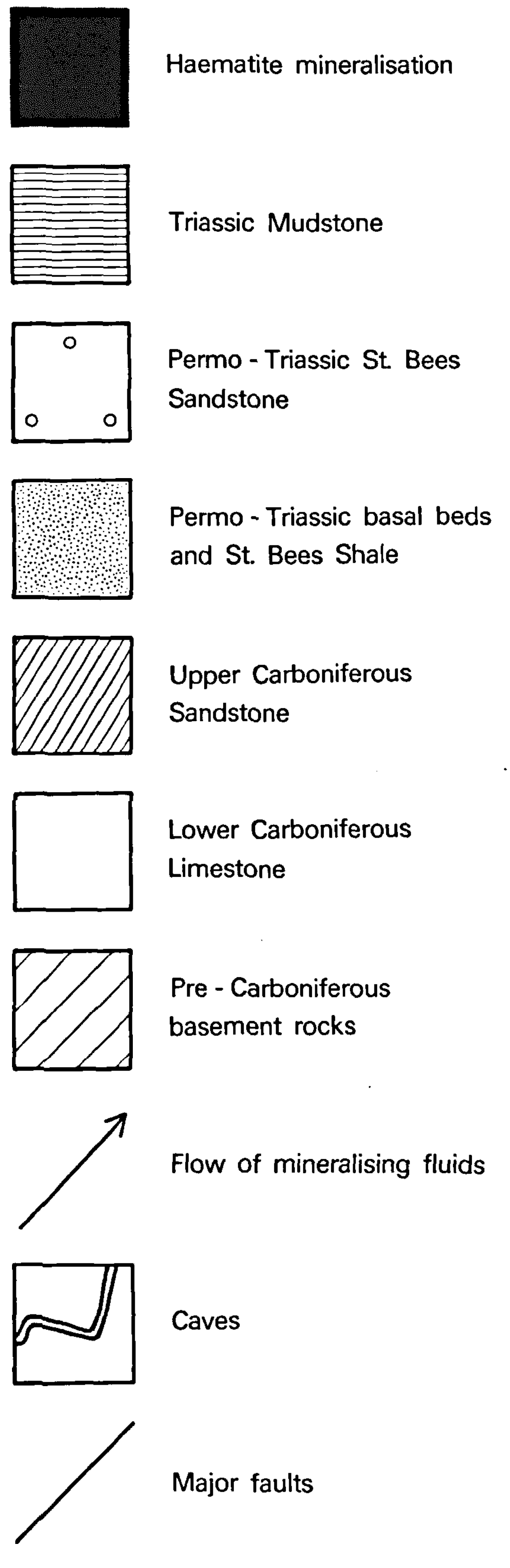
Having established the structural basis of karstification, it is possible to recognise at least four phases of karst development in the area: Lower Carboniferous, Permo-Carboniferous, ?Oligo-Miocene and Late Quaternary. Other phases of karstification may well have occurred, but owing to the existence of considerable gaps in the geological record, in particularly between the Triassic and the Late Quaternary, the evidence of much of the landscape history of the area is missing.

During the Lower Carboniferous, karstification took place as the result of the subaerial erosion of unlithified carbonate sediment during periods of temporary falls in sea level. The resultant palaeokarst surfaces can be related to the regression maxima of Ramsbottom's (1973) major cycle model of British Lower Carboniferous sedimentation and hence to similar features in similar stratigraphic positions elsewhere in British Lower Carboniferous Limestones. Despite this, there is no evidence to support Ashmead's (1969a; 1974a) case for a widespread end-Lower Carboniferous phase of subaerial karst and cave development in the area.

At the end of the Carboniferous, the region was faulted and uplifted by the increasingly violent tectonic effects of the Hercynian Orogeny. These movements initiated a phase of considerable erosion with the result that the limestone was once again exposed to subaerial conditions. Thus, during the Permo-Carboniferous transition, the re-exposed karst was adjoined to the north, south and southeast by areas of topographically higher, impermeable beds, whilst within the karst area the hills were capped by remnants of impermeable Upper Carboniferous Sandstone. It therefore seems likely that streams from these impermeable areas sank at the limestone boundary to form swallet-type cave systems and that a karst landscape similar to that of contemporary northwest Yorkshire developed (Fig.10.1).

Earlier workers proposed that during this period the area formed either a simple limestone plateau (Ashmead, 1974a, 207) or an uneven pediment (Rose and Dunham, 1977, 59). This seems difficult to reconcile with the complex pattern of relief to have been expected as a result of the recent Hercynian Orogeny. Moreover, it is clear from borehole records that relative relief in the area exceeded 80 m at this time.

With the onset of Permo-Triassic deposition, the Permo-Carboniferous karst landscape was buried beneath accumulations of Permo-Triassic beds and no further karstification appears to have occurred in the area until the Tertiary. Previous workers regarded the Tertiary as a period of considerable karstification. They envisaged the large-scale exposure of limestone under tropical humid conditions and the development of tower karst, füsshohlen, terra rossa soils, etc. (Corbel, 1957; Sweeting, 1970; 1972; 1973). The present study has found no evidence to support this thesis. Indeed, although the limestone to the east of the area may have been exposed to subaerial processes during the Tertiary, the evidence of mineralisation suggests that, as with the limestone to the west of the area, the karst remained buried beneath Permo-Triassic beds until at least the Miocene. Nevertheless, it does seem that during this period there developed in the area certain unique features, locally known as sops, which were the result of interstratal karstification in association with downward-moving mineralising fluids (Fig.10.2). On the basis of the relationship of these features to structures of Alpine age, the phase of karstification which gave rise to them has been cautiously assigned to



KEY: Figs. 10.1 - 10.3

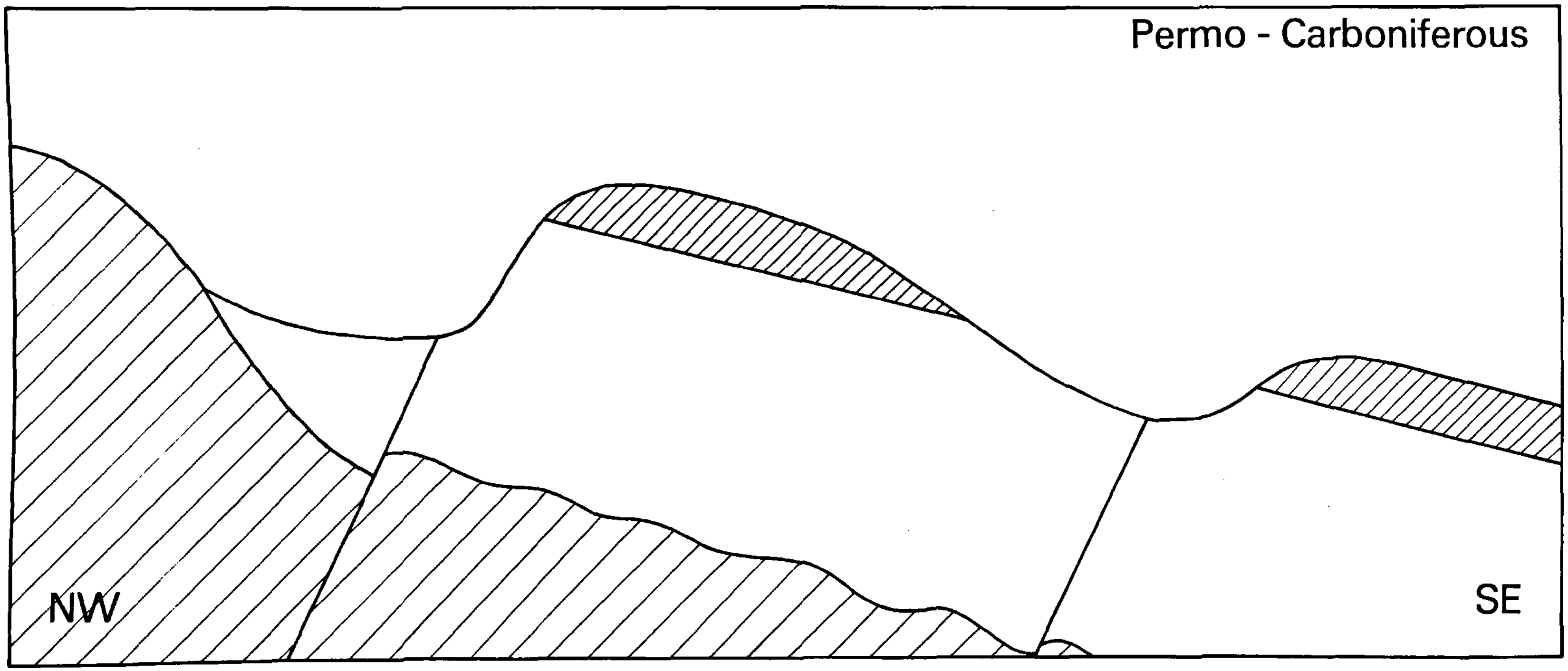


Fig. 10.1 Schematic cross - section through the Morecambe Bay area during the Permo - Carboniferous phase of karstification

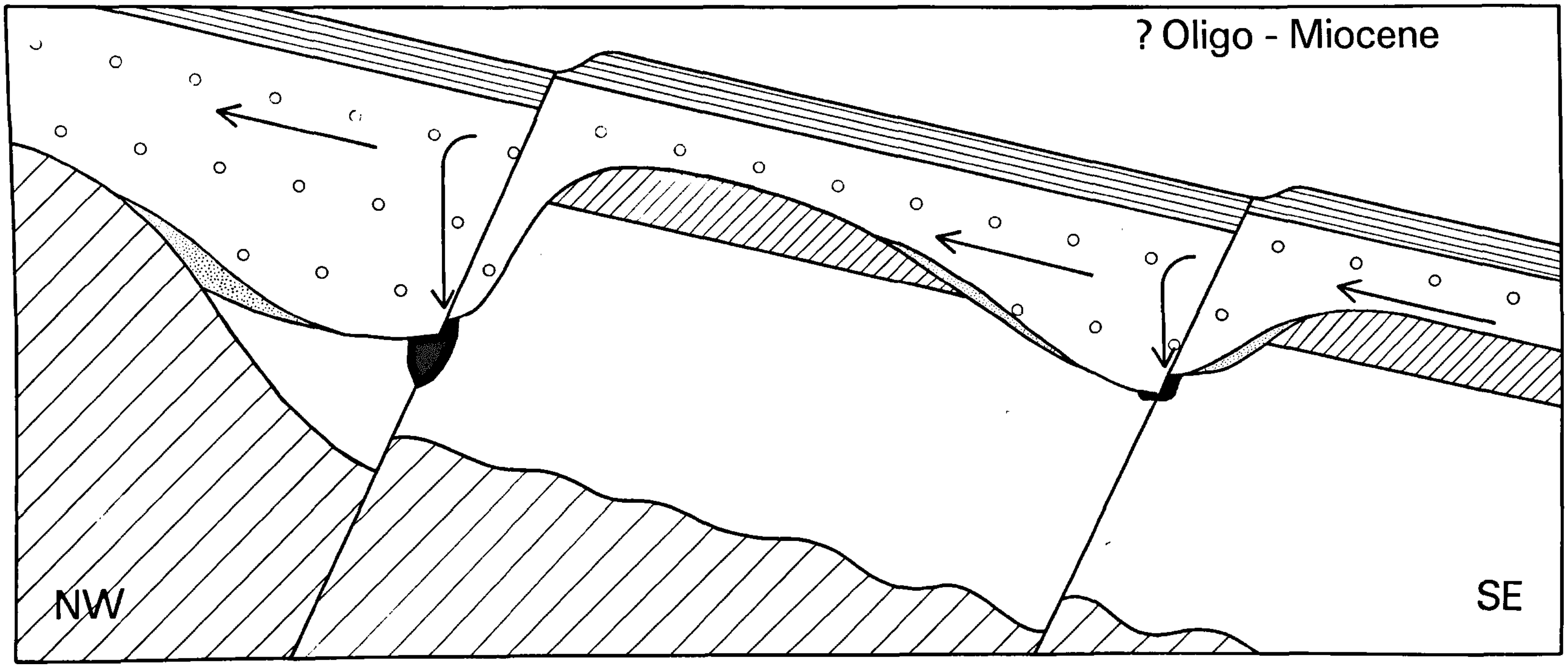


Fig. 10.2 Schematic cross - section through the Morecambe Bay area during the ? Oligo - Miocene phase of karstification

the Oligo-Miocene.

With the end of the Alpine Orogeny, the area assumed a form similar to that of the present day, with upstanding fault-bound limestone blocks separated by extensive low-lying tracts. Previous workers have attempted to fit the pattern of cave and karst development on these blocks into a model of falling base-level control. Nevertheless, the present study has shown that there is evidence neither of a preferential height-distribution of karst features in the area nor of the existence of the high sea levels thought to have controlled base level. Instead a simple pattern of groundwater drainage seems to have developed, with streams sinking around the fringes of the impermeable capping on the limestone blocks and resurging at the base of the hills. Successive glaciations appear to have overdeepened the valleys and removed the impermeable capping from the area, leaving the earlier drainage system perched above present ground-water flows (Fig.10.3). Thus, the contemporary drainage system consists of diffuse flow recharge feeding low-level resurgence systems.

Both on sedimentological and morphological grounds, the abandoned cave systems can be shown to be of at least last inter-glacial age. These cave systems include those features prev-iously termed "phreatic network" caves and regarded as either Tertiary-relict füsshohlen or the product of polje-margin processes. The present study, however, has shown that the "phreatic network" caves are, in fact, either fragments of typical abandoned karst conduits of similar age to other cave systems in the Morecambe Bay area, or true network systems.

Owing to the absence of topographically-higher imper-meable rocks adjacent to the limestone in the area, there is

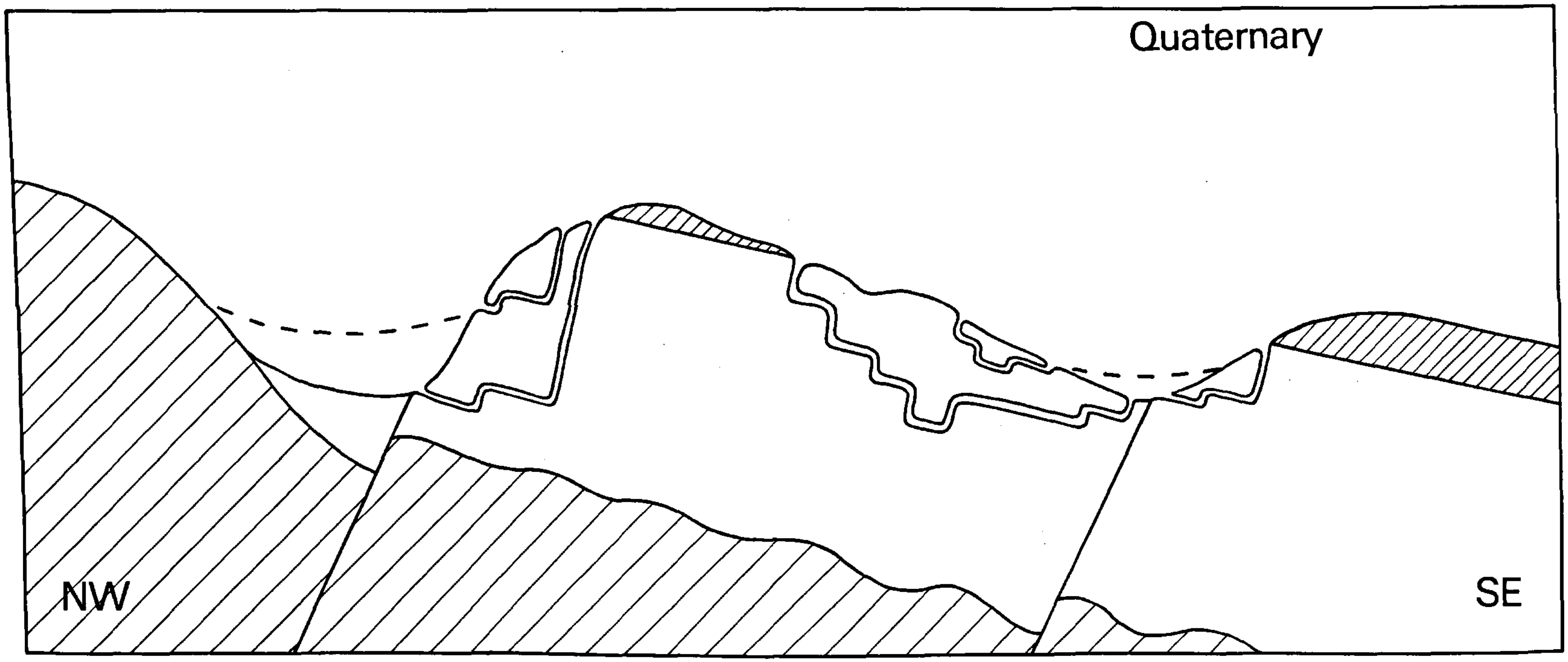


Fig. 10.3 Schematic cross - section through the Morecambe Bay area during the Quaternary phase of karstification

little contemporary concentration of water flow at the surface and hence little concentration of solutional activity. Thus, most of the surface karst features in the area are relict. The study has described certain of these features, including cemented screes, a widespread yet rarely investigated feature of glacio-karst, and karstic closed depressions. Both the true karst depressions of the area and the features which have been previously regarded as poljes are comparable in size to the largest similar features found elsewhere in the British Isles.

Apart from recording the history and environment of karstification in a qualitative manner, the study has also attempted to quantify karst palaeohydrology. Although numerous studies have been made of the history of karst drainage systems, few have attempted anything beyond a reconstruction of the chronology of events. Yet caves, in particular, contain much evidence of the hydraulic conditions under which former flows occurred. Thus, the dimensions of scallops and cave meanders have been used to infer past conditions of flow, and hydraulically-transported cave sediments have provided a useful record of palaeohydraulic conditions. This study is the first which has applied these methods of palaeohydrological analysis to the investigation of a particular area. More especially, the detailed study of sediments from Fissure Cave has provided the first semi-quantitative picture of the palaeohydraulic behaviour of karst drainage systems.

The study of cave sediments has also revealed chronological information of more than regional significance. Kirkhead Cave has been shown to contain an almost complete sequence of deposits dating from the Late-glacial to the present day, as well as possessing evidence of the most northerly Upper Palaeolithic industry of any significance in Britain. Similarly, Grizedale Wood Drainage Level has been shown to contain reversely magnetised sediments, whose age is as yet unclear, but which are very likely to prove the oldest known Quaternary deposits yet found in the whole of Cumbria.

The present study has therefore demonstrated the significance of the Morecambe Bay area in terms of its value in both palaeoenvironmental reconstruction and karst study. The area is one which has been neglected for too long and which deserves to be set alongside the classic karst areas of the British Isles, both with regard to its landscape value and to its environmental history.

V : APPENDICES

11. APPENDIX 1: QUATERNARY SEA LEVEL CHANGE

11.1 Pre-Flandrian sea level change

Although much is known of sea level history in Morecambe Bay during the Flandrian (see 10.2.1), the sea level curve cannot yet be extended further back in time with any confidence. The classical picture of sea level change during the Quaternary is one of generally falling sea levels upon which glacio-eustatic fluctuations are superimposed (see, for example, P. Evans, 1971), and it was on the basis of this model that Parry (1958; 1960a) proposed that the Morecambe Bay area had been planed by a series of successively lower sea levels. However, the present study has found no evidence, neither of a morphological nor a sedimentological nature, for such planation (see 8.2).

Citing evidence of sea caves and wave-cut notches at altitudes of between 12 m and 29 m O.D., Tooley (1977, 6) proposed the existence of high sea levels during the period 25000-19000 B.P. He dated these features by correlation with coastal forms and marine deposits found at similar altitudes on the Isle of Man. Sea levels of such altitude at that time are difficult to reconcile with the glacio-eustatic model of sea level change. Evidence from oxygen isotope analysis of deep-sea cores (Shackleton and Opdyke, 1973, 45-46), as well as from other sources (see below), suggests that sea levels at that time would have been >100 m below those of the present. Thomas (1976, 321) has tentatively explained the Isle of Man evidence in terms of isostatic depression, a theory which is more consistent with present knowledge. Subsequently, Tooley (1979, 146) proposed a

Hoxnian or earlier age for the sea caves and wave-cut notches of the Morecambe Bay area, although he provided no evidence to support this view. More detailed study of the caves and notches, however, has revealed no indication of a marine origin (see 8.2.1).

Reviews of the evidence of Quaternary sea level change in the Irish Sea by Mitchell (1972; 1977) suggest that, although there may have been an overall drop in sea level, it was on nothing like the scale proposed by the classical model, which involves a fall of as much as 200 m through the Quaternary. It appears, instead, that each transgression returned to approximately the same level of 5-20 m O.D. in successive interglacials, certainly during the Late Pleistocene. This theory is broadly supported by the deep-sea core evidence, from which changes in global ice volumes and hence glacio-eustatic sea level changes may be inferred (Shackleton and Opdyke, 1973). It is to one of these interglacial stages that the only pre-Devensian marine deposit known in the region has been ascribed. At Wigton, in the Solway lowlands, fossiliferous marine clay has been found beneath thick glacial deposits in a borehole, and has been provisionally assigned to the Ipswichian, although it is possible that the clay is not <u>in situ</u> (Eastwood <u>et al</u>, 1968, 219, 227; Huddart, Tooley and Carter, 1977, 123).

It is reasonable to assume that, during the cold stages of the Quaternary, sea levels were lower than at present. Computation of world ice volumes during each cold phase, and hence the amount of oceanic water taken up as ice, has enabled the approximate calculation of cold-stage sea levels. During the penultimate glacial phase, sea level may have fallen by ~140-160 m (Don, Farrand and Ewing, 1962), whilst during the last glacial phase, sea level may have fallen by ~105-135 m (Cotton, 1962; Donn, Farrand and Ewing, 1962; Van Straaten, 1965).

As a result of lower sea levels during glacial periods, Howell (1965) believed that rivers were able to overdeepen their valleys and cut channels across the floor of the Irish Sea. From his studies of buried channel form, he concluded that the valleys were graded to base levels at depths of approximately 30, 60 and 90 m below O.D. Subsequently he proposed that incision may not have been so great as he formerly believed and that the valleys may have been locally overdeepened by subglacial waters (Howell, 1973; 1978, pers. comm.). Nevertheless, there is evidence that fluvial erosion has been active on the present floor of the Irish Sea, at least in the Morecambe Bay area. Whittow (1970, 187-188) was the first to point out that the submarine contours of the northeastern Irish Sea may indicate the former extension of a fluvial drainage net across the continental shelf (Fig. 10.1). It is possible that the submarine channels are explicable in terms of either glacial scour along ice ways or erosion by subglacial waters, but this is supported neither by the approximately E-W alignment of the channels nor by their apparently dendritic pattern. The channels can be traced out to sea to depths of at least 75 m below O.D. Since it is unlikely that the channels were eroded by estuarine waters flowing across the continental shelf, due to the lower density of fresh water in

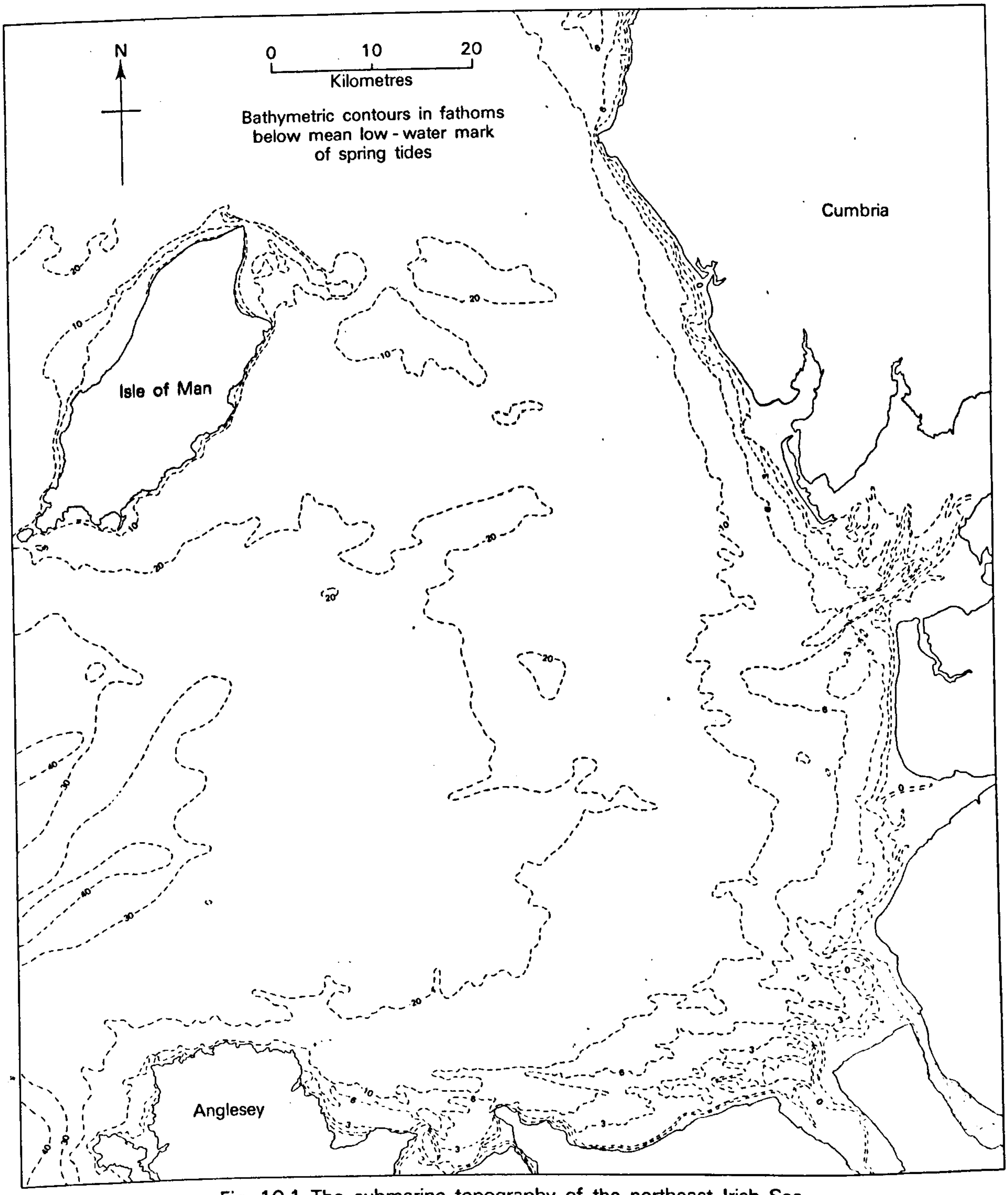


Fig. 10.1 The submarine topography of the northeast Irish Sea

comparison with sea water, it must be concluded that sea level stood at least 75 m lower than at present at some stage during the Quaternary. In order for rivers to have existed on the present floor of the northeastern Irish Sea, conditions must have been sufficiently cold to have resulted in low sea levels, yet the ice front must have been some distance north of the area.

Pantin (1978) has shown that during the Late Devensian the northeastern Irish Sea was probably occupied by a pro-glacial lagoon in which bedded sands were deposited on the till of the retreating Late Devensian ice sheet. At this time the sea probably lay to the west of the Isle of Man. The bedded sands show no evidence of fluvial conditions and they are unconformably overlain by lower-middle to upper Flandrian marine sediments. This suggests that a marine transgression, eroding the bedded sands as it advanced across the continental shelf, took place before surface drainage could be established on the floor of the Irish Sea.

However, evidence from within the Bay suggests the existence of a more complex picture. The floor of the Bay consists of a network of bedrock channels overlain by a sequence of till, varved clays, and two sequences of sands, silts and clays (Water Resources Board, 1970). The upper sand, silt and clay complex contains marine molluscan and mammalian fossils (Kendall, 1900; Reade, 1904; Water Resources Board, 1970). Underlying it, and separating it from the lower sand, silt and clay complex, are discontinuous peat horizons at altitudes of -11.13 m to -16.40 m O.D. (Kendall, 1900; Reade, 1904; Water Resources Board, 1970), suggesting a marine transgressive phase (see 10.2.1). ^{14}C dates from the peat range from 9270 ± 200 B.P. to 7725 ± 95 B.P. (Tooley, 1974). The lower clastic complex infills valleys cut in the varved clays. The sands, silts and clays have been interpreted as of freshwater origin (Water Resources Board, 1970). Thus, a sequence of glacial deposition, pro-glacial ponding, fluvial erosion and, possibly, deposition, followed by terrestrial conditions preceding a marine transgression is indicated. This may be related to the similar sequence in the northeastern Irish Sea, bearing in mind that an earlier marine transgression in the deeper parts of the Irish Sea basin would have prevented the occurrence of fluvial and terrestrial conditions. Assuming similar conditions to have prevailed in previous late-glacial periods, the erosion of channels across the continental shelf may be cautiously regarded as largely an early-glacial phenomena.

10.2.1 Sea level change during the Flandrian

The marine transgression consequent upon the world-wide melting of the ice of the last glaciation is the best substantiated of all the marine events in the Morecambe Bay area. The evidence for Flandrian sea level change has been reviewed by Huddart, Tooley and Carter (1977), who deduced the existence of five discrete transgressions and regressions during the period (Table 10.1). The height of each sea level represents the height of the mean high-water mark of spring tides (MHWMST) on the basis of Tooley's (1969; 1974, 31) contention that minerogenic sedimentation is replaced by biogenic sedimentation above the MHWMST. However, three possible sources of error do not seem to have been taken into account in establishing this sea level chronology. Firstly, there is the effect of consolidation on the depositional height of the marine deposits. In their study of the Flandrian marine deposits of the Somerset Levels, Kidson and Heyworth (1976, 226) concluded that consolidation and compaction had resulted in considerable downward displacement of deposits. Secondly, the effect of peat shrinkage has been overlooked. The bulk of the Morecambe Bay marine deposits occur in association with peats, most of which have now been drained for agricultural purposes. Morgan (1969, 249) has shown that the drainage of peat mosses can result in a vertical shrinkage of 4 m in $\sim$100 years. Finally, no allowance is made for the surge effect as tides enter constricted estuarine areas, with its consequent impact on the height of the MHWMST. Yet many of the Morecambe Bay marine sites, for example, the Kent Estuary, the Leven Estuary and Rusland Pool, are within such constricted areas. Thus, it may not be reasonable to level each marine deposit to $\pm$0.5 cm and to regard deposits at marginally different altitudes as providing evidence of discrete transgressions or regressions as proposed by Tooley and his co-workers.

10.2.2 The geomorphic effects of the Flandrian transgression

Apart from depositing the extensive tracts of marine sediments which occupy both the Bay and the adjoining coastal areas up to heights of about 5-6 m O.D., the Post-glacial rise in sea level may have affected the geomorphology of the area in several ways. According to Millward and Robinson (1972, 22-23), the fossil cliffline which is such a feature of the Bay (Plate 10.1), was eroded during the peak of the Flandrian transgression, which they regarded as having occurred approximately 6000 years ago, although they also admitted that the feature may have been initially developed during a previous erosional phase. Study of tidal records from Barrow-in-Furness, extending back to 1920, suggests a contemporary MHWMST of 5.06 m O.D., only 0.72 m below the maximum reached during the Post-glacial. Yet present spring tides only just reach the foot of the fossil cliffs, achieving little in the way of erosion. It is difficult to believe that tides even 0.72 m higher would be adequate to erode near-vertical limestone cliffs which reach heights of up to 10 m. On the other hand, it is possible that the lack of contemporary erosion along the cliffs may be the result of a shift of the channels within the Bay. Until about 1850, for example, the Kent channel was close inshore at Silverdale, enabling boats to moor there. After this date the channel moved and the area silted up. Early in the present century, the Kent moved towards Silverdale again and pleasure steamers from Morecambe were able to land passengers. The channel stayed well in-shore until about 1920, since when the salt-marshes have extended out into the Bay (Mitchell, 1977, 43). Nevertheless, it is likely that the fossil sea cliff is a compound feature, for a refraction seismic transect (see 11.5)

Plate 10.1 The "fossil" marine cliffline at Silverdale (SD4575)

normal to the cliffline at Silverdale Shore (SD455756) clearly shows a buried wave-cut abrasion platform backed by the fossil cliffline developed at a height of ~3 m O.D. (Fig. 10.2). This platform may have been developed during an earlier part of the Post-glacial or, more probably, during an earlier interglacial when marine erosion may have been more active.

Oldfield (1960b, 108) has recorded horizontal notches cut into the base of the cliffs all around the Silverdale peninsula. He regarded these as wave-cut features, although they are not being actively eroded at the present time and are considered to date from the maximum of the Post-glacial marine transgression. Tooley (1977, 5) provisionally assigned the same notches to the Ipswichian, although he subsequently claimed that they may reflect, in part, relatively higher sea levels from 3500-800 B.P., as well as being an inheritance from Ipswichian sea levels 8-13 m above those of the present (Tooley, 1978, 146-147).

The notches which Oldfield and Tooley described are not as distinct a feature as they would infer. Where they do occur, there is little to suggest that the notches are not continuous to a level below that of the salt-marsh. Furthermore, the notches are best developed where perennial pools occur along the base of the cliffs. These pools are replenished by high tides and this fact, in association with the probable high organic acid content of waters in contact with the salt-marsh, means that the waters are able to erode the base of the cliffs solutionally and develop

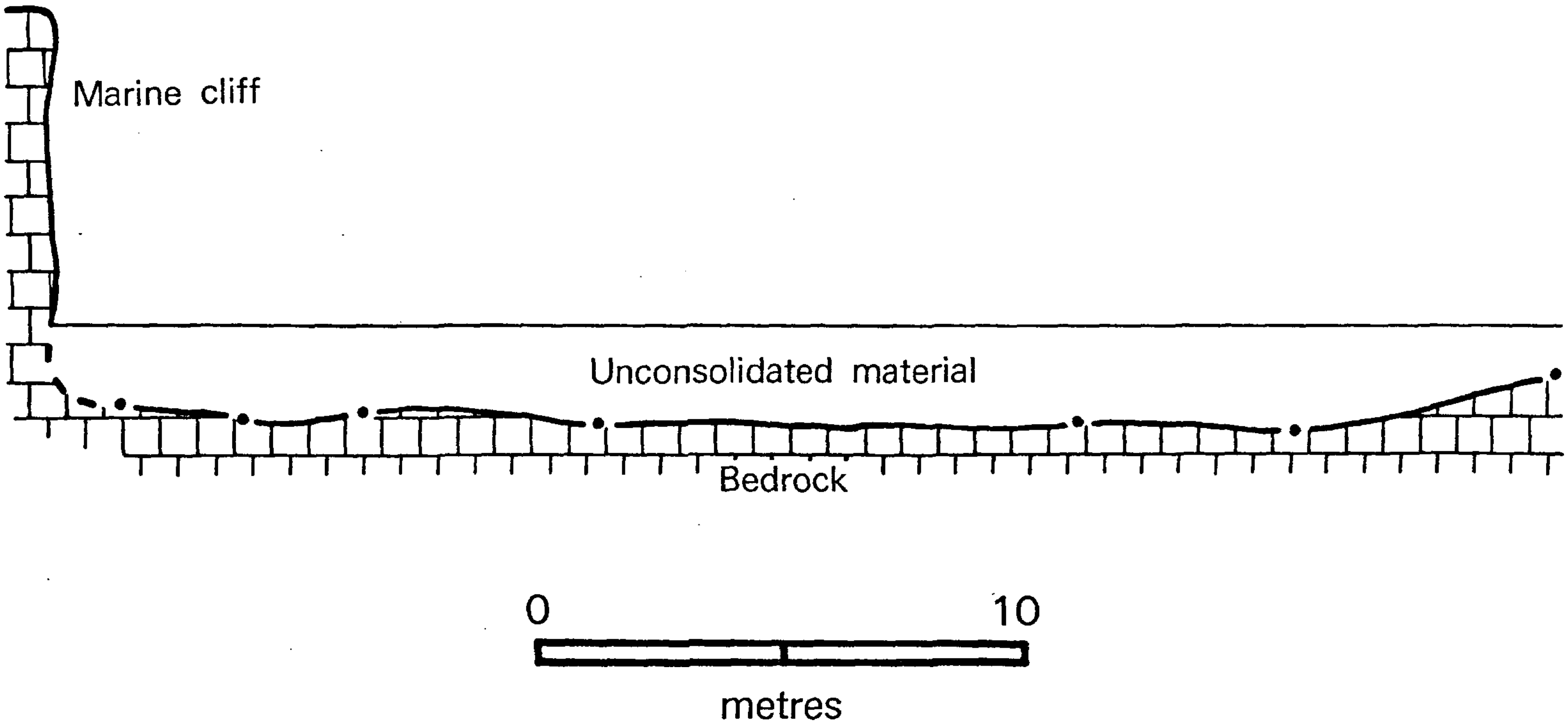


Fig. 10.2 Depth to bedrock determined from a seismic refraction transect at Silverdale Shore (SD455756)

a notch similar to that which exists.

Holland (1967, 67-69) has reported the existence of "beach" deposits in the caves found along the cliffs at Silverdale. Although he assigned no date to these deposits, his reference to them as being part of the "25 foot beach" implies a link with the Post-glacial 25 foot beach of Scotland. In Silverdale Shore Cave (SD45557560) the deposits are found at heights up to 9.36 m O.D., 3.58 m above the maximum MHWMST during the Flandrian. This is considerably above the height both of contemporary storm-beaches and of known fossil storm-beaches (see below). Analysis of the petrography and grain-size distribution of the "beach" deposits from Fissure Cave, Silverdale Shore Cave and Wall End Cave was carried out (see 5.5.1 and 7.3.2). With the exception of the fluviatile and lacustrine deposits found in Fissure Cave, all the deposits fell within the glacial environmental envelope of the QDa-Md diagram (Fig. 5.1).

Finally, at a number of locations around the Bay, in particular around Low Furness and Silverdale, are found fossil storm-beach deposits. These occur at heights of 6.7-7.9 m O.D. around Low Furness (Rose and Dunham, 1977, 122) and up to 0.7 m above MHWMST at Jenny Brown's Point, Silverdale (SD45967355). The Silverdale deposits consist mainly of rounded limestone pebbles. The clasts are in point or face contact, while the interstices are filled by inwashed fines. The deposits underlie a vegetation cover and no longer appear to be undergoing deposition. Instead they are being undercut by contemporary high tides, leaving the derived pebbles banked up against the deposits (Plate 10.2).

Plate 10.2 "Fossil" storm beach deposits at

Jenny Brown's Point, Silverdale (SD45967355)

12. APPENDIX 2: LABORATORY AND FIELD METHODS

11.1 The study of joint orientations

11.1.1 Measurement

A wide variety of methods of measuring joint orientations has been proposed. Many of these techniques involve a considerable subjective element. Chapman and Rioux (1958, 121), for example, suggested recording the mean orientation of 20 parallel joints at selected stations, whilst Weaver (1975) proposed the measurement of dominant joint sets only. In areas of complex jointing, however, these methods may neglect critical joint trends, as well as being statistically invalid.

In an attempt to obtain impartial samples, Muller (1933) recommended the measurement of all joints at the sample site, all data being collected by one observer. Muller noted the problems of inadvertently sampling long fractures many times, and the possibility that fractures at the weathered surface may differ from those at depth. These problems were not so important in the Morecambe Bay study, however, for the data were intended to test the degree of structural control on landforms. Thus, the relative length of fractures was of significance, and the orientation of fractures at the surface was of greatest interest for the study. This is not withstanding the observations of Appleby (1942, cited in Pincus, 1951, 92) which suggest that fractures in dynamited exposures correspond well with those of normal exposures.

Spencer (1959, 475) recommended beginning measurement at an arbitrary point and measuring all fractures in the vicinity of that point until 100-120 observations had been made. A similar method was adopted by Firman (1960, 320).

Unfortunately, despite their objectivity, these methods fail to indicate the concentration and relative length of the fractures. As these parameters were important to the study, a rather different method of measurement was finally adopted. A predetermined area was selected, generally of 2.5 m square, within which all joint orientations were recorded. This area was increased if the number of joints measured was inadequate to fulfil statistical requirements. In order to obtain an idea of the relative length of joints, and because of the sinuous nature of many of the joints, fractures were measured between each point of intersection with other fractures. Preliminary investigation supported the conclusion of Moseley and Ahmed (1967, 73-74) that the joints are generally orientated normal to the bedding, so a single measurement of dip at each site was sufficient to orientate the joints in three dimensions. In fact, it was subsequently found that two-dimensional analysis was adequate for the study of joint trends.

In order to check the reliability of the sampling and measuring procedures, two adjacent sites on the same bedding plane were measured at Newbiggin Crags (SD549792). Analysis of the two sample populations by the methods given in 11.1.2 showed that both sites had significant data peaks at the same orientations of 80-89° and 160-179° (Fig. 2.5), thereby confirming the reliability of the methods. Interestingly, analysis of joints at the same location, but at the next bedding plane in the stratigraphic sequence, indicated significant joint peaks differing from the others by about 10° (Fig. 2.5), underlining the importance of measuring joint orientations strictly at the point of interest.

In order to observe whether ground measurements of joints differed from those obtained from aerial photographs, upon which Moseley based much of his analysis, the results of measurements at Newbiggin 1 and 2 were compared with those of the same site obtained from 1:10000 vertical aerial photographs. Using the same methods of analysis, almost identical results were obtained, confirming the comparability of the two methods of data collection (Fig. 2.5).

11.1.2 Analysis

Orientated data have been frequently found not to be amenable to standard statistical methods requiring linear data, whilst even those tests specifically designed for orientated data usually assume the data to be both unimodal and directional (see, for example, Gaile and Burt, 1980). Unfortunately, joint-orientation data is both axial (i.e. non-directional) and, usually, polymodal. To date, the only method of analysis capable of dealing with such distributions is that developed by Tanner (1955, 2473-2475).

Tanner's method involves the division of the compass into a number of equal sized sectors and the calculation of measured orientation frequency within each sector. Tanner proposed that those sectors with an orientation frequency one standard deviation above the mean represented prominent orientation modes. This represents a significance level of 67%, which would be considered low when applied to standard statistical-significance tests. However, High and Picard (1971, 37) suggested that a high level of significance should not be demanded for geological polymodal-orientation data and that the 67% level is both acceptable and indicative of a preferred direction. It is interesting to note, moreover, that when applied to axial data, acceptable levels of significance may be reduced because of the bimodal nature of each peak.

Implicit in this technique is the assumption that the frequency of measured orientations within each sector is derived from an initial population approximating to a gaussian distribution. The sample populations of the Morecambe Bay data tend to be positively skewed. A closer approximation to a normal distribution could be achieved by increasing the size of the sector division. However, it was considered that any increase of the 10^{o} interval selected would have led to results too crude to allow differentiation of subtle changes in joint orientation.

11.2 Palaeomagnetic analysis of deposits in Grizedale Wood Drainage Level (SD48257409)

11.2.1 Sampling

Five orientated samples were taken from the undisturbed laminated beds in the upper part of the exposure on the southern side of the mine level (section A). In order to meet Thompson's (1977, 52) criterion that palaeomagnetic directions should be confirmed by investigating two separate sections in the same locality, two further samples were taken from a similar smaller exposure a few metres away on the north side of the level (section B).

11.2.2 Measurement

One of the samples from section A was selected at random and the stability of the remanent magnetism examined by progressive stepwise demagnetisation in alternating magnetic fields, up to a maximum field strength of 1000 Oe. The direction and intensity of magnetism after each demagnetisation step was measured using a Digico spinner magnetometer. This procedure indicates the optimum level of demagnetisation which will remove any secondary components of remanence, acquired since deposition, thus leaving the primary magnetisation, whose direction is close to that of the ancient field. From a study of the results (Figs. 11.1 and 11.2), it was decided that neither intensity nor direction of magnetisation changed significantly with demagnetisation and that measurements of natural remanent magnetism could be regarded as indicative of that of the ancient field. The direction of the natural remanent magnetism of the remaining smaples was therefore measured using the Digico spinner magnetometer and the results plotted on Fig. 3.2.

11.3 Sediment sampling and analysis

11.3.1 Sampling

Open sections were firstly recorded by sketching and photography. The faces of sections were then cleaned by scraping or trimming along the lines of visible strata to prevent contamination and a vertical sequence of samples taken from each bed. Bulk samples from closed depressions

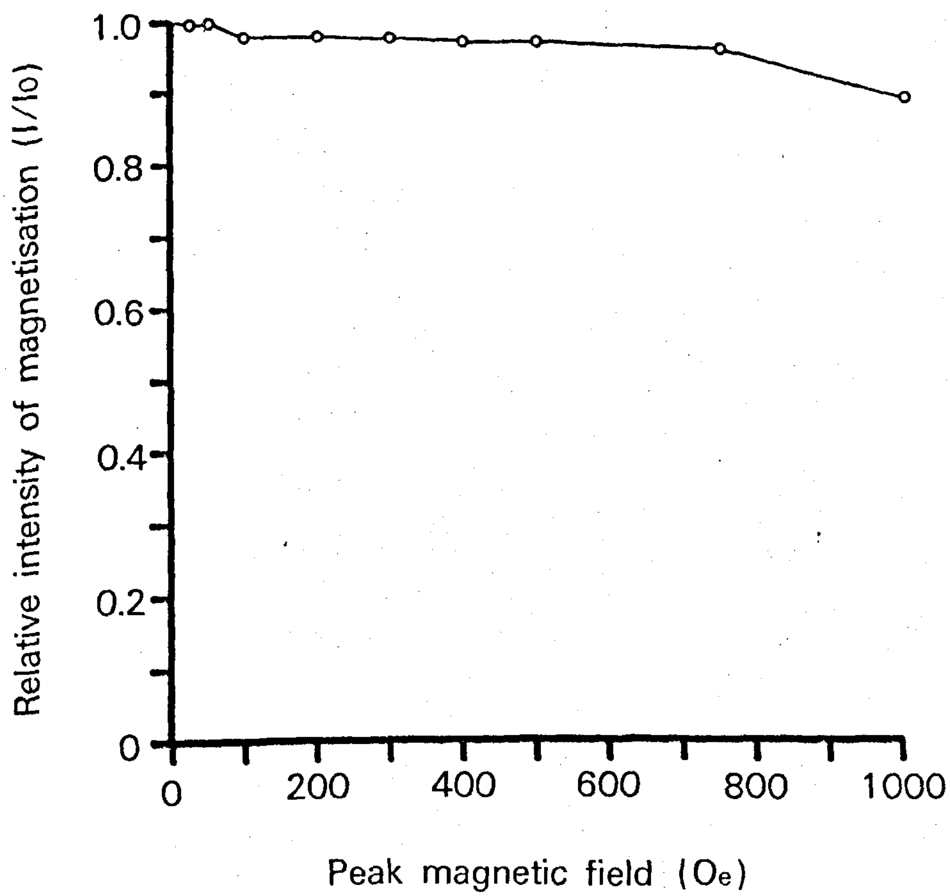


Fig. 11.1 Progressive demagnetisation by alternating magnetic fields of sample A1, Grizedale Wood Drainage Level

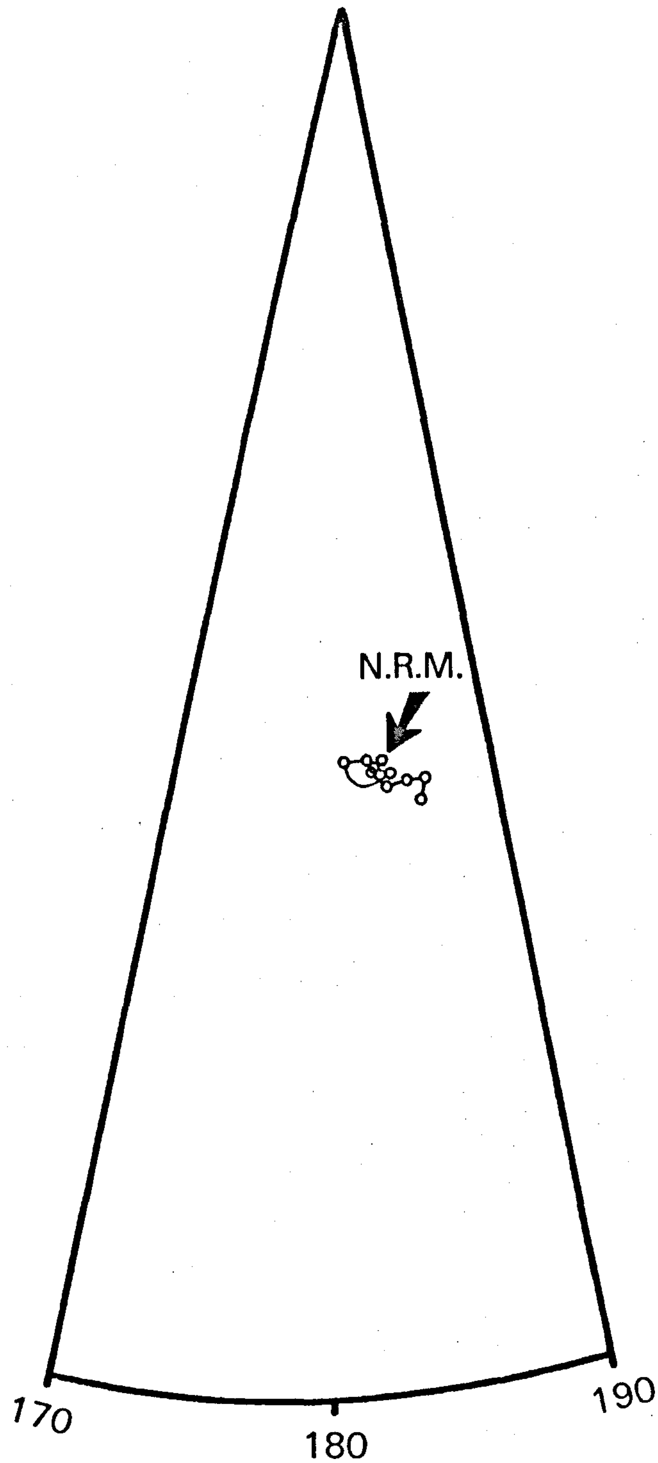


Fig. 11.2 The effect of progressive demagnetisation by alternating magnetic fields on the direction of magnetisation of sample A1, Grizedale Wood Drainage Level (Lambert–Schmidt net)

were taken from beneath the organic horizon in pits. In all cases, the minimum sample weight was dependent on the maximum clast size, as determined by Mace (1964).

11.3.2 Sediment analysis

Each sample was air-dried and weighed to 0.001 g, and then disaggregated by using a pestle and mortar, and a brush to clean the gravel and cobble fraction. For those samples containing particles of ≥ 4 mm, the coarser fraction was separated by sieving through a 4 mm (-2 ϕ) sieve. The retained fraction was then sieved by hand through sieves at ½ ϕ intervals, weighing the material retained on each sieve to 0.001 g and retaining it for petrographic and shape analysis.

The <4 mm fraction was split by using a riffle box to obtain a subsample of 60-70 g. The subsample was weighed to 0.001 g and then wet-sieved through a 63 μm (4 ϕ) sieve into a pan using a wash bottle containing 0.5% w/v $(Na\,Po_3)_6$ (Calgon) solution to act as a dispersant (American Society for Testing Materials, 1963, 208; Ingram, 1971, 60). The screen of the sieve had been previously wet on both sides with the dispersant. After wet-sieving the subsample, the contents of the pan were transferred into a 1000 ml cylinder.

The 63 μm sieve and the retained sediment fraction were dried at 30^{o}C and the dried, disaggregated sample poured into the top of a stack of sieves ranging in size from 2.83 mm (-1.5 ϕ) to 63 μm (4 ϕ) at ½ ϕ intervals. The sieve stack, which had been previously calibrated by sieving standard samples, was agitated in an Endecott mechanical shaker for 10 mins (Ingram, 1971, 64) and the fraction retained on each sieve weighed to 0.001 g. Any material of <63 μm was transferred into the 1000 ml cylinder.

The suspension in the 1000 ml cylinder was made up to the 1000 ml mark by adding dispersant. The pipette method of sedimentation analysis was then followed, using the procedure outlined by Galehouse (1971, 79-87). A constant-temperature bath at 24°C was used and samples were taken of particles finer than 4.0 ϕ, 4.5 ϕ, 5.0 ϕ, 5.5 ϕ, 6.0 ϕ and at 1 ϕ intervals thereafter, up to 11 ϕ. The oven-dried pipette samples were allowed to cool in the laboratory for 24 h to accumulate ambient moisture before weighing to 0.001 g (Folk, 1968, 38; Galehouse, 1971, 85).

11.3.3 Data analysis

For the total sample, % weight coarser than each size fraction was calculated, following the United States practice. The data were plotted as a cumulative curve of $-\log_2$ grain-size (ϕ) against % weight coarser, using a probability-percentage ordinate scale to allow curve extrapolation and interpolation.

Grain-size summary statistics were calculated on the basis of graphically interpolated percentile values. Owing to the large proportion of fine material in some of the sediments studied, it was often only possible to obtain percentile values up to D_{75}. Apart from the median (D_{50}) and quartile deviation $((D_{25}-D_{75})/2)$ (see 5.2.1), two other parameters were calculated:

(i) Mean $(\bar{x}) = 2^{-(\phi 25 + \phi 50 + \phi 75)/3}$ mm (Briggs, 1977, 79)

and

(ii) Sorting (So) = $\sqrt{D_{25}/D_{75}}$ (Trask, 1930; 1932).

Wentworth's (1922) grain-size nomenclature was adopted for verbal size-scales, and Folk's (1954) categories were used as a means of describing sediment texture.

11.4 Surveying

11.4.1 Levelling

"Phreatic network" cave altitudes were obtained by levelling, using Tilting-type and Dumpy levels. Wherever possible, traverses were tied into local bench marks. Any corrections due to misclosure were distributed evenly throughout the traverse.

11.4.2 Cave surveying

In the relatively large and accessible caves along the coast at Silverdale, it was possible to survey using levelling methods rather than the usual techniques of compass and clinometer survey. Using a fixed target and the leapfrog method, station position error is reduced to a minimum, while vertical differences between stations are more accurately measured. Detail survey, plotting and drawing followed the methods recommended by Ellis (1976). Direction was measured using a hand-held liquid-filled prismatic compass (W.D. Mark III) and distance was measured using a Fibron tape.

11.4.3 Profile surveying

Slope profiles and closed depression profiles were surveyed using an Abney level, tape and ranging poles, angle measurements being made to the nearest $\frac{1}{2}^{o}$. The level was checked by paired readings at the beginning of each profile, a correction being applied whenever necessary. Profiles were measured along the line of true slope. In the case of the slope profiles, at the top of the slope, profiles were continued perpendicular to the hill crest, as suggested by Pitty (1966); whilst at the foot of the slope, profiles were terminated where a non-slope landform, such as a flood plain or wave-cut platform, had become clearly established. Closed depression profiles, on the other hand, were surveyed along the line of the depression long axis, passing through the lowest point of the depression. Transects were also surveyed perpendicular to the long axis, again passing through the lowest point of the depression. Both sets of transects were surveyed from lip to lip of the depression.

For all profile surveys, a standard measured length of 20 m was adopted, both to maintain objectivity and in view of the overall length of profile being measured. This is the maximum length suggested by Young (1972, 146) in order to provide a close representation of ground form. The main disadvantage of a fixed measured length is that breaks of slope, visibly apparent to the surveyor, are recorded with deliberate inaccuracy. As it was important that such breaks of slope be recorded, a shorter measured length was occasionally used where the slope was judged to change significantly, for example, at the tops of the cliffs and at the lowest points of depressions.

11.4.4 Morphological mapping

Morphological mapping of the closed depressions was accomplished using the methods proposed by Savigear (1965) for slope morphology, and the mapping symbols devised by the I.T.C. for the representation of geomorphic features (Verstappen and Zuidan, 1968).

11.5 Survey of buried bedrock surfaces

11.5.1 Introduction

The usual approach to the survey of buried bedrock surfaces involves the use of excavation pits or augering methods. In the present study, neither of these methods proved successful due to the rocky nature of the superficial materials. In these circumstances, geophysical surveying methods are able to provide a continuous record of buried bedrock form, as well as an indication of the nature of the overlying materials.

Initial results from augering suggested that a simple two-layer situation, unconsolidated fill over bedrock, existed in all the locations where surveys were required. Given this situation, seismic survey methods have numerous advantages, as they provide more detailed, precise and unambiguous information than any other method of indirect subsurface investigation. On shallow interfaces such as those surveyed in the present study, it is necessary to adopt refraction seismic techniques, for in these circumstances waves reflected from shallow bodies return to the recorder so rapidly that the amplifier is unable to recover from the severe overloading to which it is also subjected by the direct wave.

11.5.2 Surveying method

A portable Bison Model 1550 Signal Enhancement seismograph was used to carry out the seismic surveys, seismic waves being generated by a 7 kg sledgehammer and a steel striking plate. Traverses were run along existing surface survey lines. Each survey leg was 20-40 m long with a geophone at each end. Reverse profiles were run along each leg with seismic waves generated at 5 m intervals between the geophones and the time of arrival of the first wave to each geophone recorded to 0.01 ms. In order to obtain a good indication of seismic wave velocity in the shallow upper layer, seismic waves were also generated at distances of 2.5 m, and occasionally 1.25 m from each geophone.

11.5.2 Interpretation of the results

For each survey leg, seismic wave travel time was plotted against distance from each geophone. In a simple two-layer situation where the interface is a velocity increase, the resultant curve consists of two straight-line segments. Points plotting above or below the line are the result of either observational error or, more likely, irregularities in the bedrock surface. Analysis of the results followed the methods of Mota (1954) to give depth to bedrock beneath each geophone, and those of Hagedoorn (1959) to give variations in velocity and depth of the refracting layer along the line of the traverse.

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