Mass Movements in Great Britain

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Chapter 5

Mass-movement sites in Carboniferous strata

INTRODUCTION

R.G. Cooper

Rocks of Carboniferous age are widespread in Great Britain (Figure 5.1); for example, they give rise to their highest hills in the Pennines. In this area there has been a large amount of mass movement, the largest and most interesting sites being found in the southern part, the 'Southern Pennines', broadly co-inciding with the Peak District National Park. Several reviews of landsliding in this general area have been produced (Cross, 1987; Doornkamp, 1990; see Figure 5.2).

The Carboniferous strata of Great Britain have a high density of recorded landslides (9.0 per 10 km^2). The Namurian sandstones ('Millstone Grit') account for 42% of these, the Westphalian coal-bearing strata ('Coal Measures') account for 33% and the Dinantian limestones ('Carboniferous Limestone') for the remaining 25% (Jones and Lee, 1994). In the national review of landsliding (Jones and Lee, 1994), 57% of landslides in Carboniferous strata were recorded as 'unspecified' in type. Of those with specified types, 25% were recorded as single rotational slips, 34% as 'complex', and 22% as translational.

Two sites have been selected from the Lower Carboniferous Dinantian strata for the mass-movements GCR network. Both are in the Carboniferous Limestone Series. They are **Eglwyseg Scarp** (Creigiau Eglwyseg) near Llangollen in North Wales, and **Hob's House**, at Monsal Dale in the Peak District, Derbyshire.

Five sites have been selected from the Upper Carboniferous Namurian strata. All are in the

Figure 5.1 Areas of Carboniferous strata (shaded) and the locations of the GCR sites described in the present chapter.

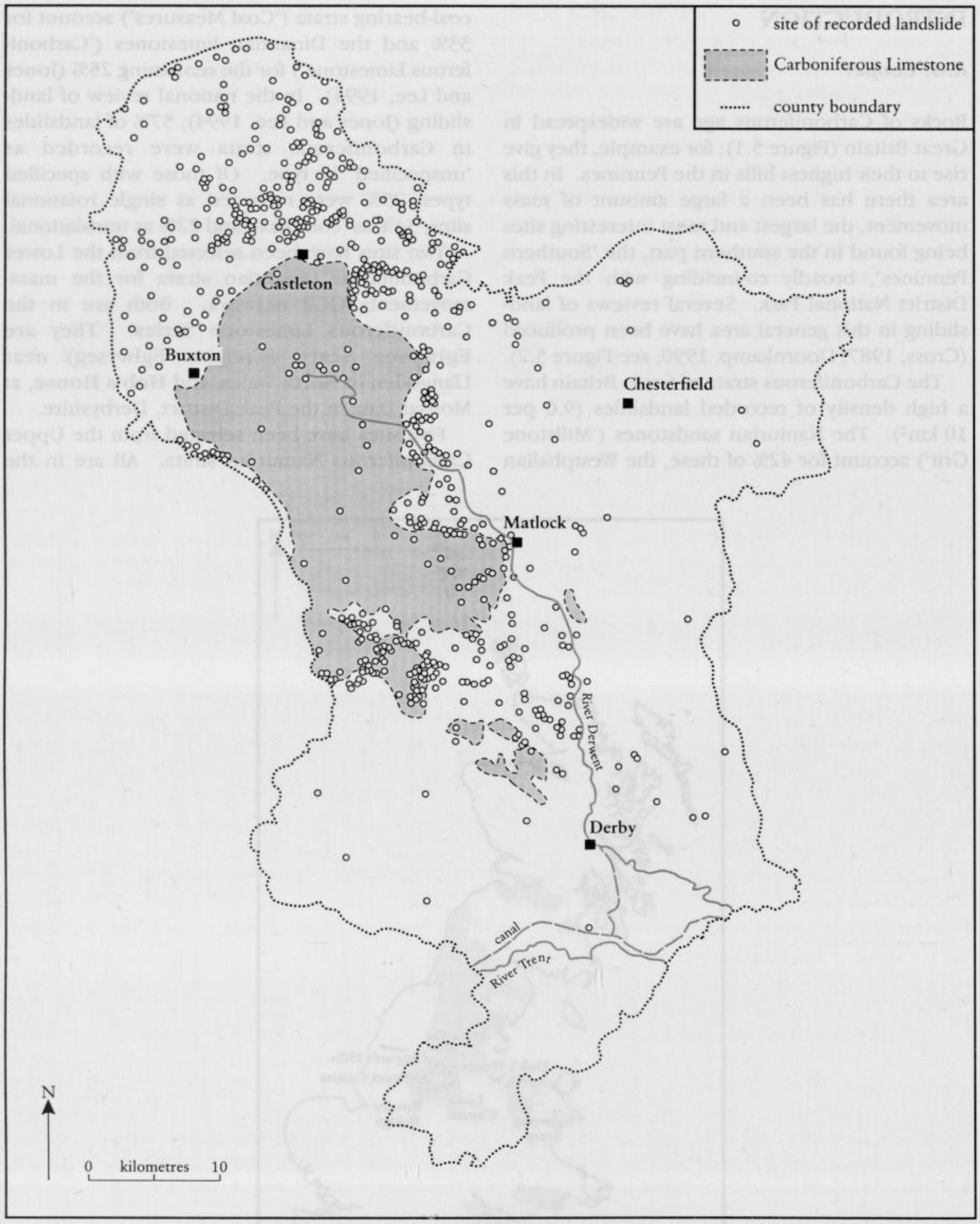


Figure 5.2 The distribution of recorded landslides in Derbyshire. After Geomorphological Services Ltd (1988); from Doornkamp (1990).

sandstones and mudstones of the 'Millstone Grit' series. They are **Alport Castles**, **Canyards Hills**, **Lud's Church**, **Mam Tor**, and **Rowlee Bridge**. Of these, all lie within the Peak District National Park, Lud's Church being in Staffordshire and the rest in Derbyshire.

For convenience, the GCR sites selected are discussed here in two sections, covering the Lower Carboniferous strata and the Upper Carboniferous strata respectively.

MASS-MOVEMENT SITES IN LOWER CARBONIFEROUS STRATA

EGLWYSEG SCARP (CREIGIAU EGLWYSEG), CLWYD (SJ 235 432–SJ 235 480)

R.G. Cooper

Introduction

Eglwyseg Scarp extends 8 km northwards from Castell Dinas Brân near Llangollen (see Figures 5.3 and 5.4). It faces westwards and is divided into a series of buttresses by deep gully dissection. The site has frequently been quoted as a type example for escarpments and screes, but it differs markedly in both form and history from many others in upland Britain.

Tinkler (1966) identified two distinct types of depositional slope below the free face. He used the term 'clitter' to describe a thin veneer of coarse rock fragments on a slope, the form of which is controlled by underlying structure, and reserved the term 'scree' for loose fragments in an accumulation of sufficient depth for the angle of repose to be determined by the physical characteristics of the fragments themselves, as distinct from its being determined by whatever lies beneath the accumulation of fragments.

The outcrop of the Lower Grey and Brown Limestone co-incides with that part of the escarpment below the most significant free face (Figure 5.5). It is characteristically composed of limestones 0.62–0.9 m thick, with intercalated shale beds (5–15 cm), and forms a stepped lower bedrock slope. The Middle White Limestone is lithologically distinct and forms the free faces. It is composed of three massive beds about 7.5 m, 13.5 m and 6 m thick respectively, separated by narrow shale beds that are locally absent. Lateral variations in thickness are considerable, with a general thinning towards the south. Above this are several low and degraded scarps in the Upper Grey Limestone and Sandy Limestone.

Description

The escarpment seems to have been initiated at a time when the River Eglwyseg joined the River Dee near Castell Dinas Brân, at about 300 m

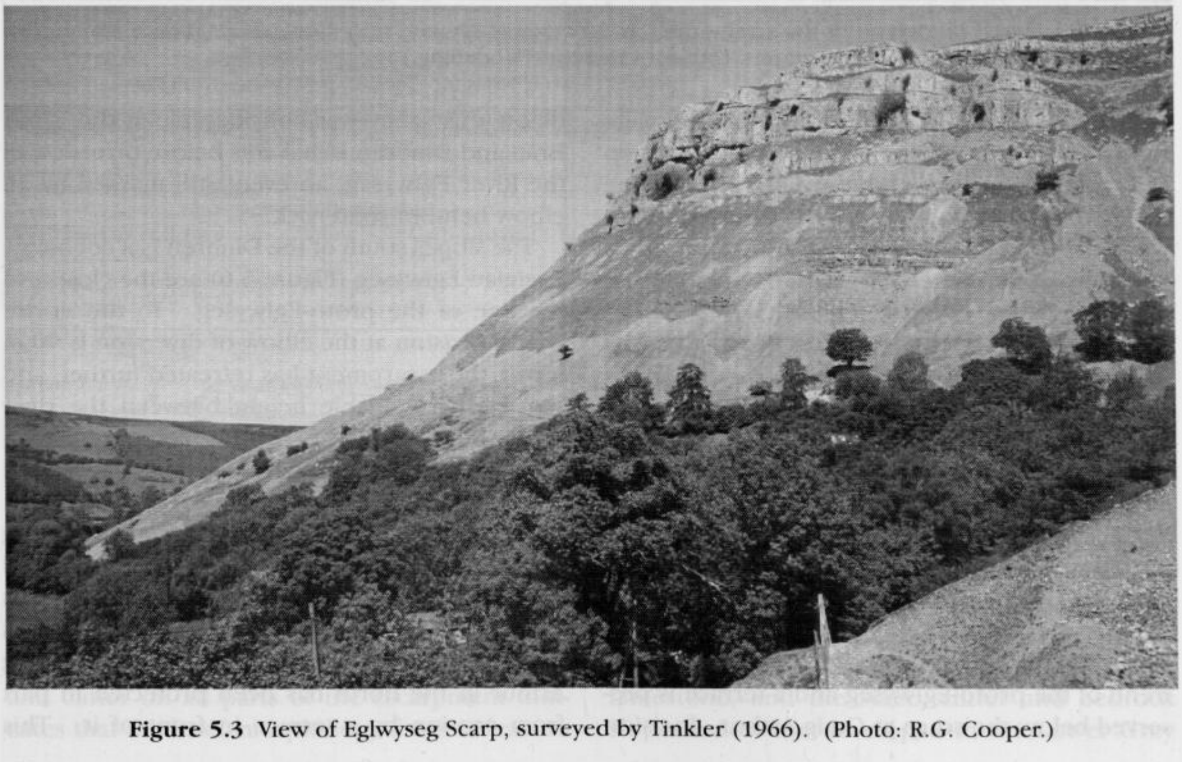


Figure 5.3 View of Eglwyseg Scarp, surveyed by Tinkler (1966). (Photo: R.G. Cooper.)

Figure 5.4 Aerial phtograph of the scree-slopes at Eglwyseg Mountain, near Llangollen. (Photo: Cambridge University Collection of Air Photographs, Unit for Landscape Modelling.)

above OD. Local slopes indicate this drainage trend, and a terrace is preserved below the scarp at Craig Arthur. Some time in Early Pleistocene times the River Eglwyseg was diverted westwards, so that slopes south of the Dinbren Isaf col have since developed without a river to facilitate transportation or erosion (Tinkler, 1966). They appear to have declined, aided by the southward thinning of the Middle White Limestone. An extensive mantle of Devensian till covers the uplands and is also found in the deeper fissures on the scarps, on the inter-scarp ledges and on the main slopes below. During deglaciation, meltwater and periglacial activity re-deposited some of this as head, and upper deposits are common along the foot of the main escarpment (Figure 5.6).

Local slopes indicate the former drainage trend of the proto-Eglwyseg and a terrace is preserved below the scarp at Craig Arthur. Incision below this valley was considerable at the Dinas Brân and Dinbren Uchaf cols before diversion of the River Eglwyseg, an event still marked by an elbow bend in solid rock.

The slopes south of the Dinbren Isaf col below Creigiau Eglwyseg (Figure 5.6) are the closest to the line of the proto-Eglwyseg. To the north, where incision at the elbow of diversion is 90 m lower, the escarpment has retreated further, and the greater available height between the river and the slope crest (270 m compared with 150 m at Trefor Rocks) permits greater horizontal retreat of the upper cliff before complete decline. The latter stage is approached at Eglwyseg Mountain and Creigiau Eglwyseg. Pinfold Buttress, which shows least sign of decline, is opposite the elbow of diversion. At Trefor Rocks the slopes are in a degraded state, while Craig Arthur in the north has been protected in part from erosion by a terrace in front of it. This

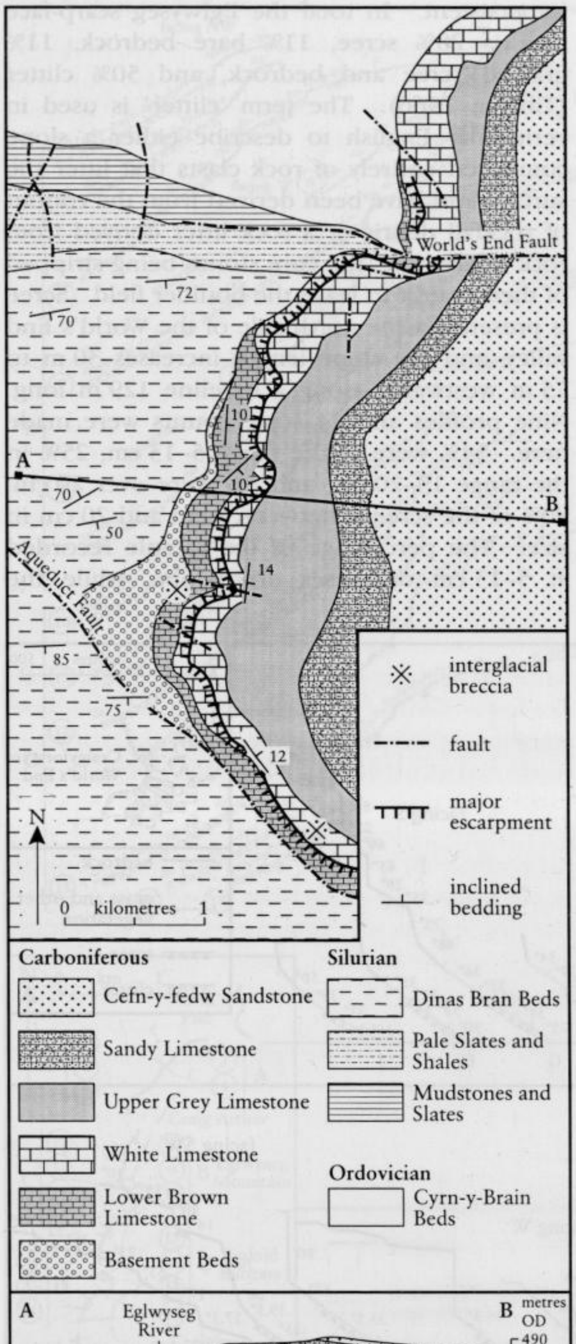


Figure 5.5 The geology of the Eglwyseg Valley, North Wales. After Tinkler (1966).

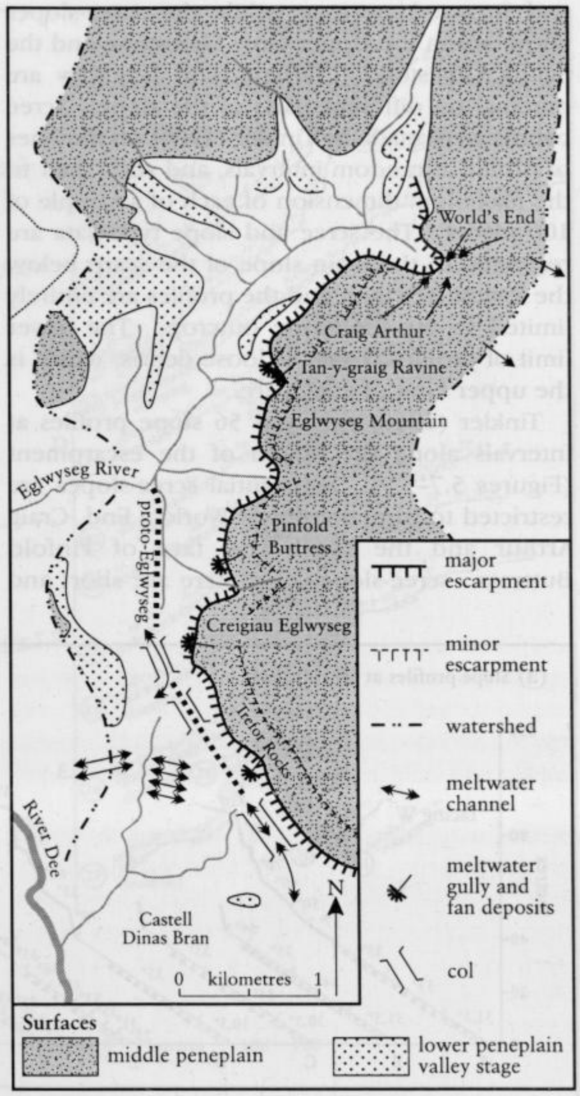


Figure 5.6 The geomorphology of the Eglwyseg Valley, North Wales. After Tinkler (1966).

buttress lies to the front of the general line of the escarpment. The presence of till below scree and clitter on all parts of the escarpment indicates that the basic morphology dates at least to the last (Ipswichian) interglacial, with only minor modification since then. The debris cover is therefore a shallow mantle on a fossil bedrock form partly buried by till. Its accumulation can have had little influence on the form of the bedrock slope below the free face or upon the free face itself (Tinkler, 1966). Jointing is variable, but deep cracks in the free face can produce huge tabular blocks that become embedded in the scree. Normally, fine surface cracks and joints in the free face have given rise to scree debris up to 30 cm maximum dimension.

The tallest free face is always in the White Limestone, and the scree, clitter and bedrock slopes below are developed in the Lower Grey and Brown Limestone, while the scree-slopes above are in the Upper Grey Limestone and the Sandy Limestone. Variations in lithology are minimal on different parts of the slope. Scree counts were made by Tinkler (1966) on the lines of profile at random intervals, and sizes refer to the maximum dimension of each of a sample of 100 pieces. The scree and slope type-data are restricted to the main slope of the scarp below the lowest free face, and the profiles are entirely limited to the limestone outcrop. The lower limit of profiles is that of loose debris, which is the upper limit of enclosure.

Tinkler (1966) surveyed 56 slope profiles at intervals along the length of the escarpment (Figures 5.7–5.9). Substantial scree-slopes are restricted to three localities: World's End, Craig Arthur and the south-west face of Pinfold Buttress. scree-slopes elsewhere are short and impersistent. In total the Eglwyseg scarp-face area is 28% scree, 11% bare bedrock, 11% grassed scree and bedrock, and 50% clitter (Tinkler, 1966). The term 'clitter' is used in vernacular English to describe either a slope composed entirely of rock clasts that litter the surface and have been derived from the runout of rockfall debris or a rock litter derived from rock weathering, the core stones being stripped of their matrix to leave the boulder field. Scree is present on the north side of the World's End valley, and the slope length increases 30 m to 75 m westwards along a baseline 129 m long. Nine profiles and 24 scree counts were made here. 54% were in the range 5–13 cm, 25% in the range 13–20 cm, and 3% were over 20 cm. 80% of the scree is between 5 cm and 20 cm in size. The percentage of the sample recorded at 5–13 cm decreases downslope, while the

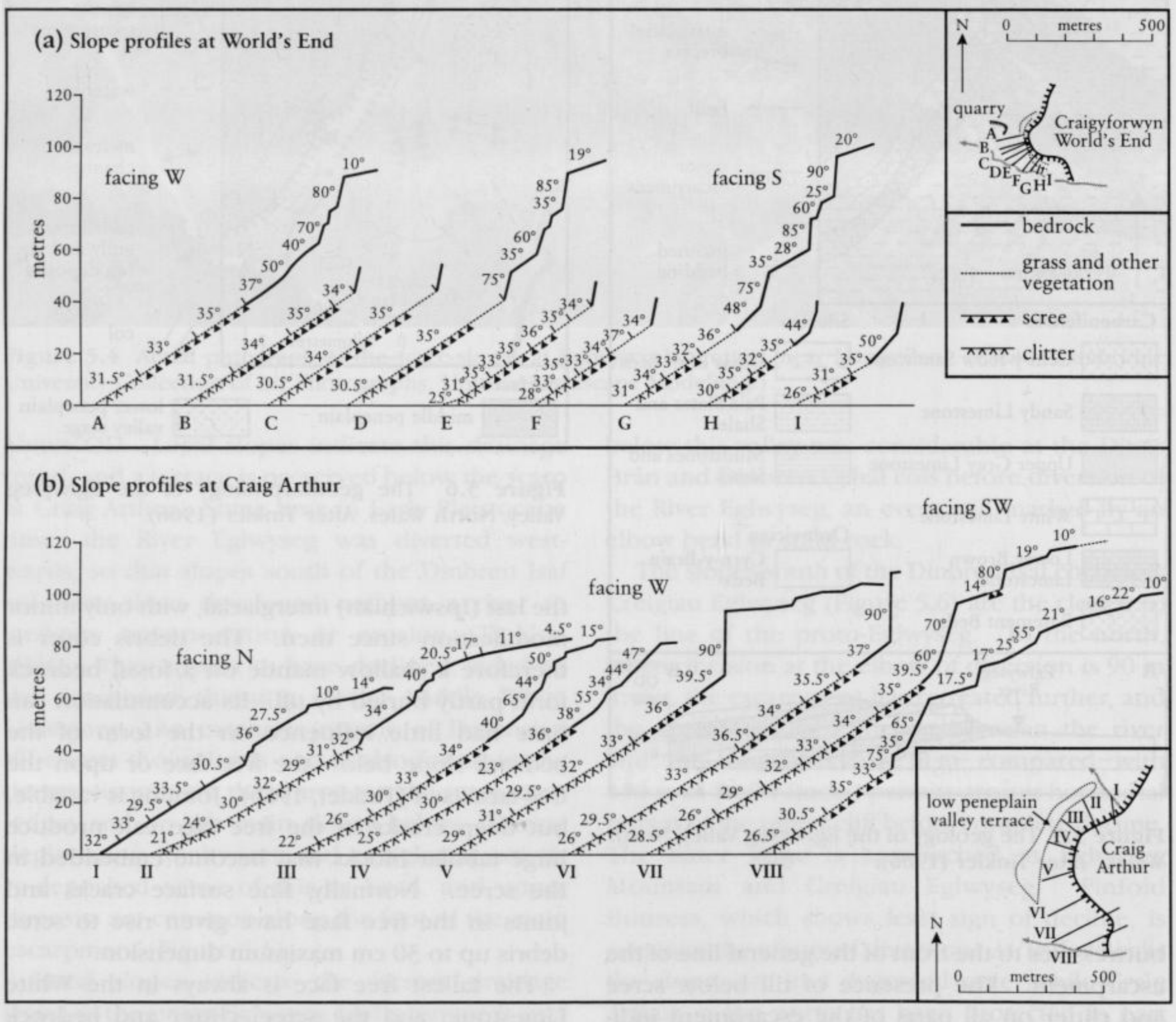


Figure 5.7 Slope profiles of (a) World's End, and (b) Craig Arthur. After Tinkler (1966).

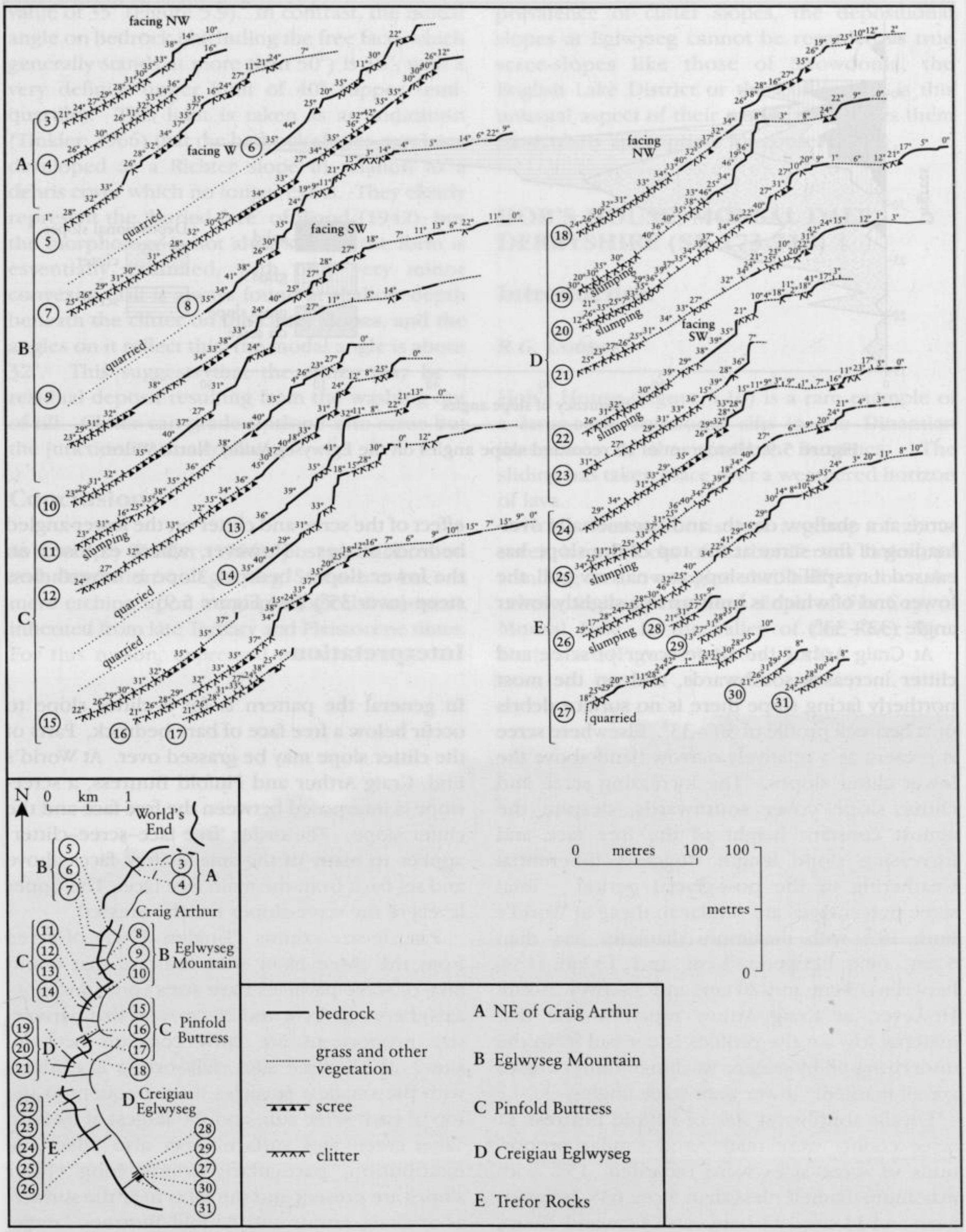


Figure 5.8 Slope profiles on Eglwyseg Scarp, North Wales, surveyed by Tinkler (1966).

percentage recorded at 13–20 cm increases downslope. Scree over 20 cm is limited to the bottom of the profile but is significantly related to slope position at the 1% level. Scree below 5 cm is barely represented in most of the samples. Coarser scree is present below the fine

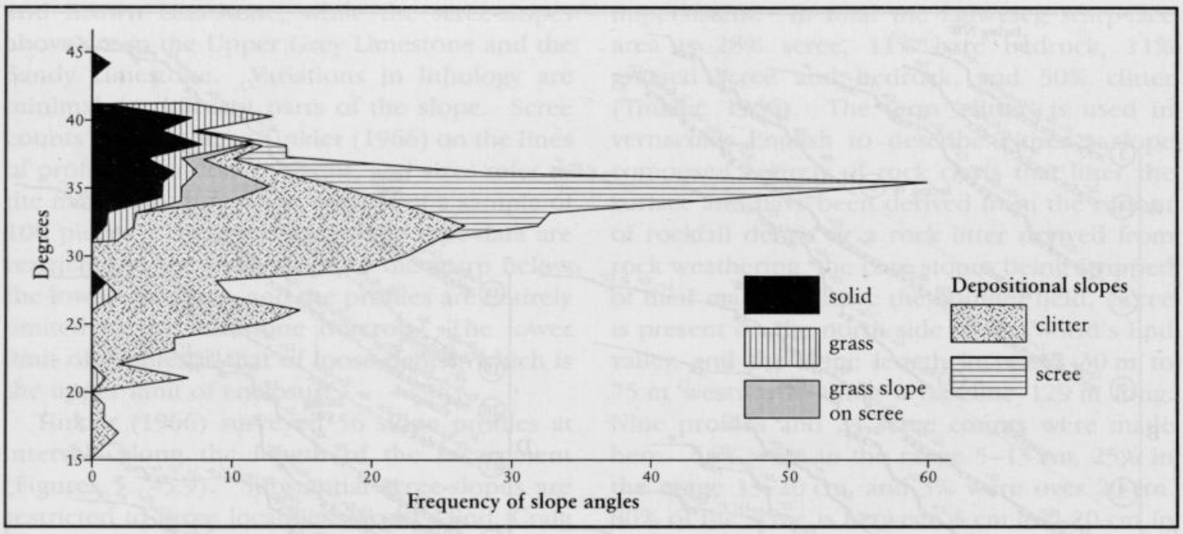


Figure 5.9 Histogram of all recorded slope angles on the Eglwyseg Valley, North Wales.

scree at a shallow depth, and occasionally overloading of fine scree at the top of the slope has caused it to spill downslope as a narrow trail, the lower end of which is built up at a slightly lower angle (32°–33°).

At Craig Arthur the slope cover of scree and clitter increases southwards, and on the most northerly facing slope there is no surface debris on a bedrock profile of 30°–33°. Elsewhere scree is present as a relatively narrow band above the lower clitter slopes. The increasing scree and clitter slope cover southwards, despite the almost constant height of the free face and increasing slope length, suggests differential weathering in the post-glacial period. Total scree percentages are similar to those at World's End: 18% with maximum diameter less than 5 cm, 66% between 5 cm and 13 cm, 13% between 13 cm and 20 cm, and 3% over 20 cm. However, at Craig Arthur much of the fine material low on the profiles is derived from the underlying till by surface washing. Clitter angles are all markedly lower than scree angles.

On the south-west side of Pinfold Buttress 11 scree counts were made, and similar proportions of scree sizes were recorded: 15% with maximum diameter less than 5 cm, 63% between 5 cm and 13 cm, 19% between 13 cm and 20 cm, and 3% over 20 cm. The fairly constant proportions at three different sites seem to highlight the constant lithology of the free face.

The distribution of bedrock angles is significantly higher than the distributions of scree or clitter, and this partly depends on the masking effect of the scree and clitter on the lower-angled bedrock slopes. However, where exposed on the lower slopes, bedrock slope is nevertheless steep (over 35°) (see Figure 5.9).

Interpretation

In general the pattern is for a clitter slope to occur below a free face of bare bedrock. Parts of the clitter slope may be grassed over. At World's End, Craig Arthur and Pinfold Buttress, a scree-slope is interposed between the free face and the clitter slope. The order 'free face–scree–clitter' applies to many of the smaller free faces above and set back from the main free face. The upper levels of the scree-slopes may be grassed.

Particle-size counts (Tinkler, 1966) of scree from the three main sites indicate that about 80% of scree particles have sizes (presumably *b*-axis) between 5 cm and 20 cm, and that particle-size proportions are fairly constant between sites. At all three sites, fall-sorting is evident, with the smallest particles most frequent at the top of each scree run, and the largest at its foot. Talus creep and surface wash also affect the distribution, particularly where long clitter slopes are present and the till is near the surface, as at Craig Arthur and Pinfold Buttress. Scree-slopes have only developed where there is a substantial free face. They are currently active, and only stabilize where a thin soil covers the uppermost part, as at World's End.

The range of slope angles recorded on the screes is only 6°–7°, clustering around a modal value of 35° (Figure 5.9). In contrast, the modal angle on bedrock (excluding the free face, which generally stands at more than 50°) is 38°, with a very definite upper limit of 40° (upper semi-quartile). This limit is taken as an indication (Tinkler, 1966) that the bedrock slopes may have developed as a Richter slope in relation to a debris cover which no longer exists. They clearly represent the 'buried face' of Wood (1942), but the morphology is not always clear: the form is essentially exhumed, with only very minor convexity. Till is always found at shallow depth beneath the clitter on the clitter slopes, and the angles on it reflect this: the modal angle is about 32°. This suggests that the clitter may be a residual deposit resulting from the washing out of till. Clitter can grade upslope into scree but the junction is usually sharp.

Conclusions

As noted by Tinkler (1966), post-glacial erosion and deposition at Eglwyseg Scarp has been a mere etching upon a morphological framework inherited from late Tertiary and Pleistocene times. For this reason, expressed mainly through the prevalence of clitter slopes, the depositional slopes at Eglwyseg cannot be regarded as true scree-slopes like those of Snowdonia, the English Lake District or the Cuillins. It is this unusual aspect of their nature that makes them particularly appropriate for conservation.

HOB'S HOUSE, MONSAL DALE, DERBYSHIRE (SK 173 710)

Introduction

R.G. Cooper

Hob's House (Figure 5.10) is a rare example of a large-scale rotational slip in the Dinantian limestones of the southern Pennines. The sliding has taken place over a weathered horizon of lava.

Hob's House consists of a group of about seven large blocks of Carboniferous limestone standing on a low-angled shelf halfway down the otherwise steep northern slope of Fin Cop, at Monsal Dale, in the valley of the River Wye, Derbyshire (Figures 5.10 and 5.11).

Figure 5.10 The backscar and transported blocks of the Hob's House landslide. (Photo: S. Graham, English Nature/Natural England.)

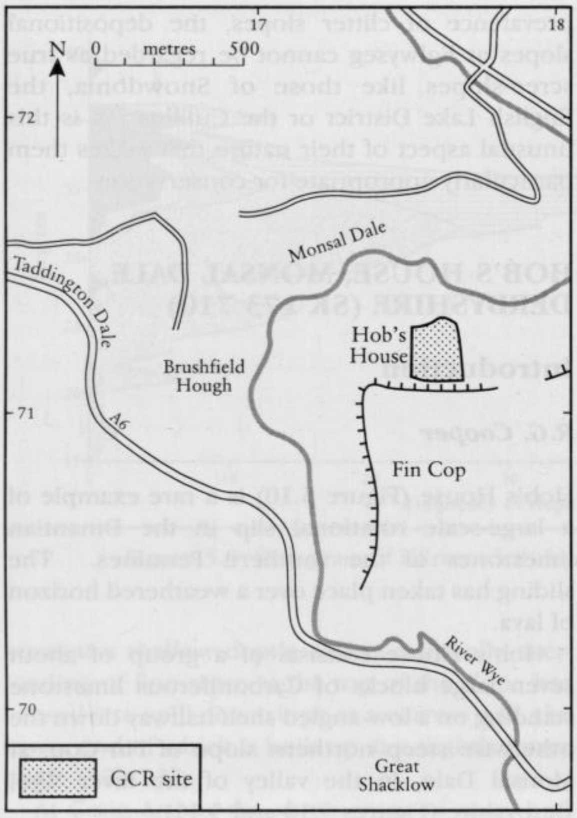


Figure 5.11 Location of Hob's House GCR site.

The Hob's House mass-movement site at Monsal Dale should not be confused with Hob Hurst's House, a Bronze Age round barrow on Baslow Moor 10 km to the east, which was excavated by Thomas Bateman in 1853 (Bateman, 1861, pp. 87–88).

Description

The blocks at Hob's House are in the dark litho-facies of the Monsal Dale Limestones (Dinantian, Lower Carboniferous). The slope is about 330 m long and stands at an overall angle of about 35° (Figure 5.12). The vertical-sided blocks, standing on an approximately horizontal boulder-strewn shelf 65 m wide, are up to 7 m high and 20 m long and broad. They are backed by a 12 m-high vertical cliff-face in the limestone bedrock, above which the slope continues 150 m to the crest of the hill, at about 40°. Below the shelf the slope runs down 130 m to the river, at an angle of about 20°. The cliff contains an enterable fissure which has been penetrated for 20 m, indicating that it is in a shattered condition, and has itself been engaged

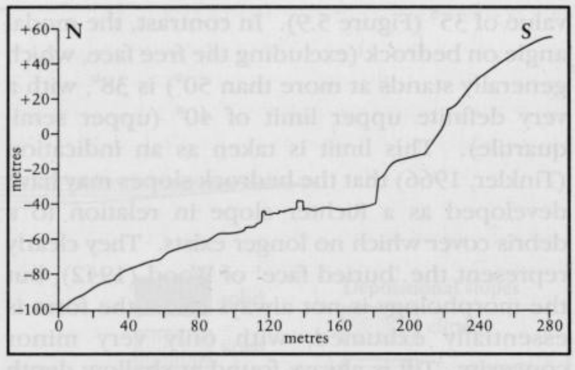


Figure 5.12 Slope profile at Hob's House.

in some mass movement (Figure 5.13). Coarse limestone scree mantles the shelf between the blocks and the cliff, and around and between the blocks. On either side, and uphill and downhill from these features, the slope is grassed over.

Some indication of the degree of displacement can be gained from the stratigraphical levels. Aitkenhead *et al.* (1985) point out that

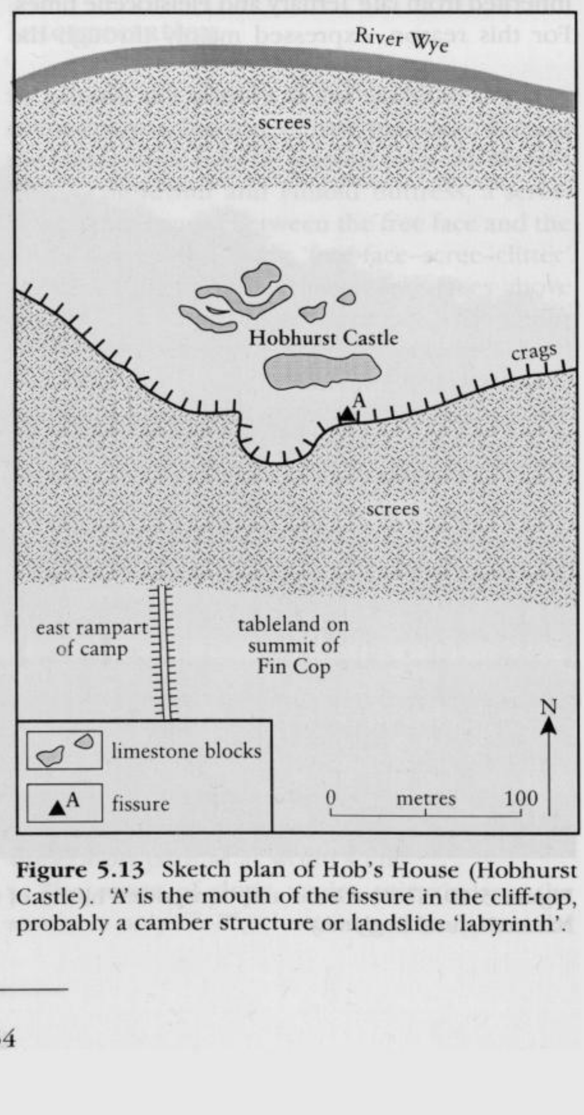


Figure 5.13 Sketch plan of Hob's House (Hobhurst Castle). 'A' is the mouth of the fissure in the cliff-top, probably a camber structure or landslide 'labyrinth'.

the 51.7 m section at Hob's House contains the Hob's House Coral Band, which is 0.4 m thick. Brown (1973) identified the cliff, and the 'towers' as containing his 'Hob's House Coral', and that this coral has dropped 15 m in the towers compared with the cliff. Furthermore, the coral is at different heights in the different towers.

Interpretation

The drop in altitude suggests that the blocks lie within the upper zone of a large-scale rotational slip. Aitkenhead *et al.* (1985) remark that landslips are not widespread on the Dinantian limestone outcrop in Derbyshire. They cite Hob's House as a rare example, where rotational slipping has occurred due to movement on softer-weathered igneous rocks, in particular the Shacklow Wood Lava. Such lavas are discussed by Ford (1977): characteristically, in the Derbyshire outcrop, their original minerals (from basalt, tuffs and dolerites) have broken down under chemical attack, usually resulting in clay minerals. The process forms a soft, green clay at the top of the chemically altered lavas, which are known locally as 'toadstones'.

Conclusions

At Hob's House an unusual set of circumstances has led to a landslip which has resulted in several block-shaped limestone 'towers' of large size having become separated from the cliff behind them. Their situation, halfway down the slope, is also unusual.

MASS-MOVEMENT SITES IN UPPER CARBONIFEROUS STRATA

ALPORT CASTLES, DERBYSHIRE (SK 142 914)

R.G. Cooper and D. Jarman

Introduction

The Alport Castles are a massive landslide complex, the most prominent elements of an extensive landslip complex affecting at least 0.85 km^2 along the eastern (west-facing) side of Alport Dale near Ladybower in the Peak District (Figure 5.14). Alport Dale is one of several valleys incised in the sandstone plateau of the Dark Peak, here by up to 200 m. Landslipping is very extensive in this part of the Pennines (cf. Johnson, 1965; Stevenson and Gaunt, 1971; Johnson and Walthall, 1979), and the geological controls on its incidence are particularly evident here. This site is notable for its array of distinct slip-masses and slumps in varying stages of intactness, attitude and distance travelled, extending in places to the River Alport.

Description

Landslipping on the steep valley-sides in the 'Millstone Grit' areas of the Pennines commonly occurs where competent sandstones overlie less competent shales. Here, the River Alport rises on the Bleaklow plateau of Shale Grit (Kinderscoutian R1 stage of the Namurian, Upper Carboniferous). About 1.5 km above this site, it begins to cut down through the Mam Tor sandstones (Kinderscoutian). It eventually reaches the underlying Edale Shales (Alportian (H2) stage of the Namurian Period). These are predominantly mudstones, though sandstones occur, and include exceptionally weak pyritic shales studied at **Mam Tor** (Vear and Curtis, 1981). Immediately the river enters them, the valley-sides 'become covered with huge landslips formed of masses of the Shale Grit which have slid down from the hilltops above' (Green *et al.*, 1887; Figure 5.14). The weakness of these Edale

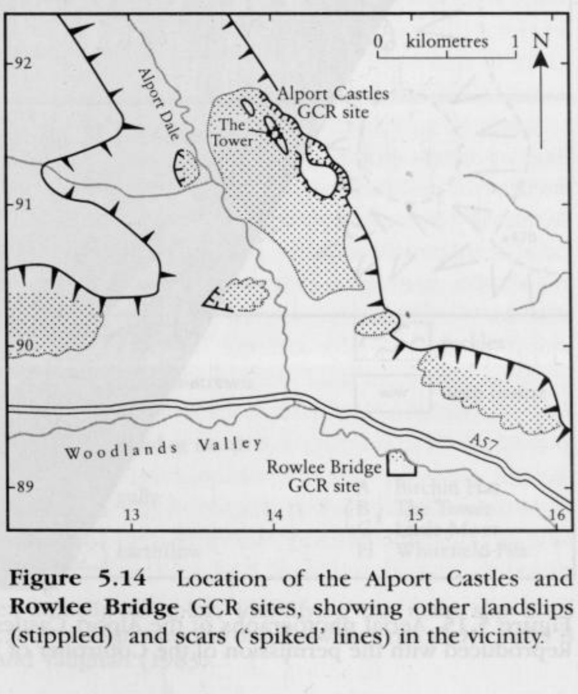


Figure 5.14 Location of the Alport Castles and **Rowlee Bridge** GCR sites, showing other landslips (stippled) and scars ('spiked' lines) in the vicinity.

Shales is evidenced by deep crumpling revealed in dam trenches in the adjacent valleys (Thompson, 1949) and at **Rowlee Bridge** nearby. The strata are here nearly level, with a slight eastward tilt.

The Alport Castles site is one of the largest landslip complexes in the district (Figure 5.15), and has been studied by Johnson and Vaughan (1983). It divides into two sectors (Figure 5.16): a main (northern) sector, where landslipping encroaches into the plateau rim along a bold craggy scar, and where the prominent 'Castles' are located; and a secondary (southern) sector, where the source scar is a less significant feature running across the upper valley-side.

Northern sector

Three main rock-masses have broken away from the rim of Alport Moor (Units A–C), leaving a scar which above Unit B reaches 68 m to the narrow boulder-filled trench floor. The scar here comprises a 30 m vertical sandstone crag above a talus slope and the rift-trench, which is of unknown original depth (Figure 5.17). These units are substantially intact, and prominent

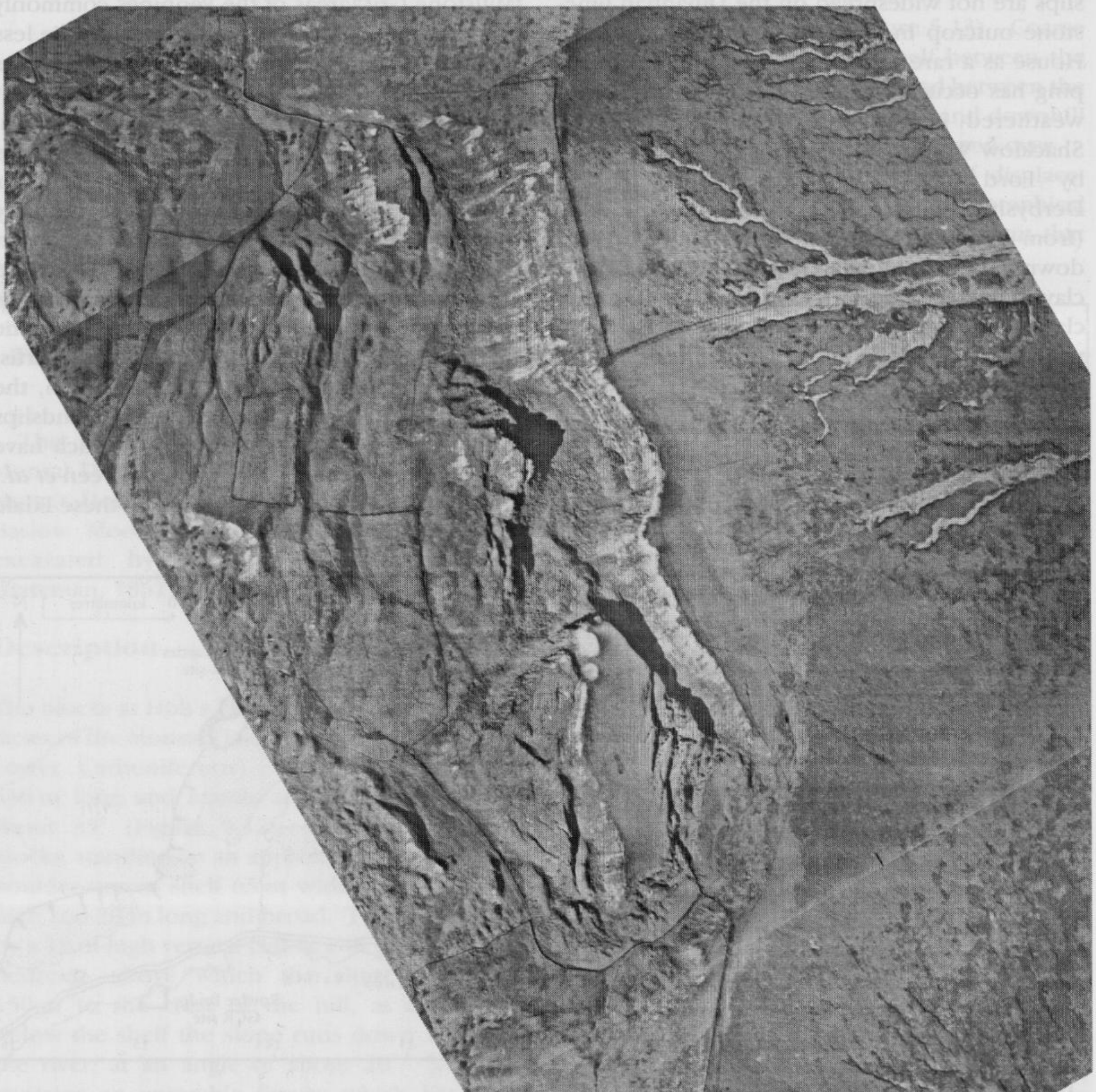


Figure 5.15 Aerial photographs of the Alport Castles landslip complex. (Photos: © Crown Copyright/MOD. Reproduced with the permission of the Controller of Her Majesty's Stationary Office.)

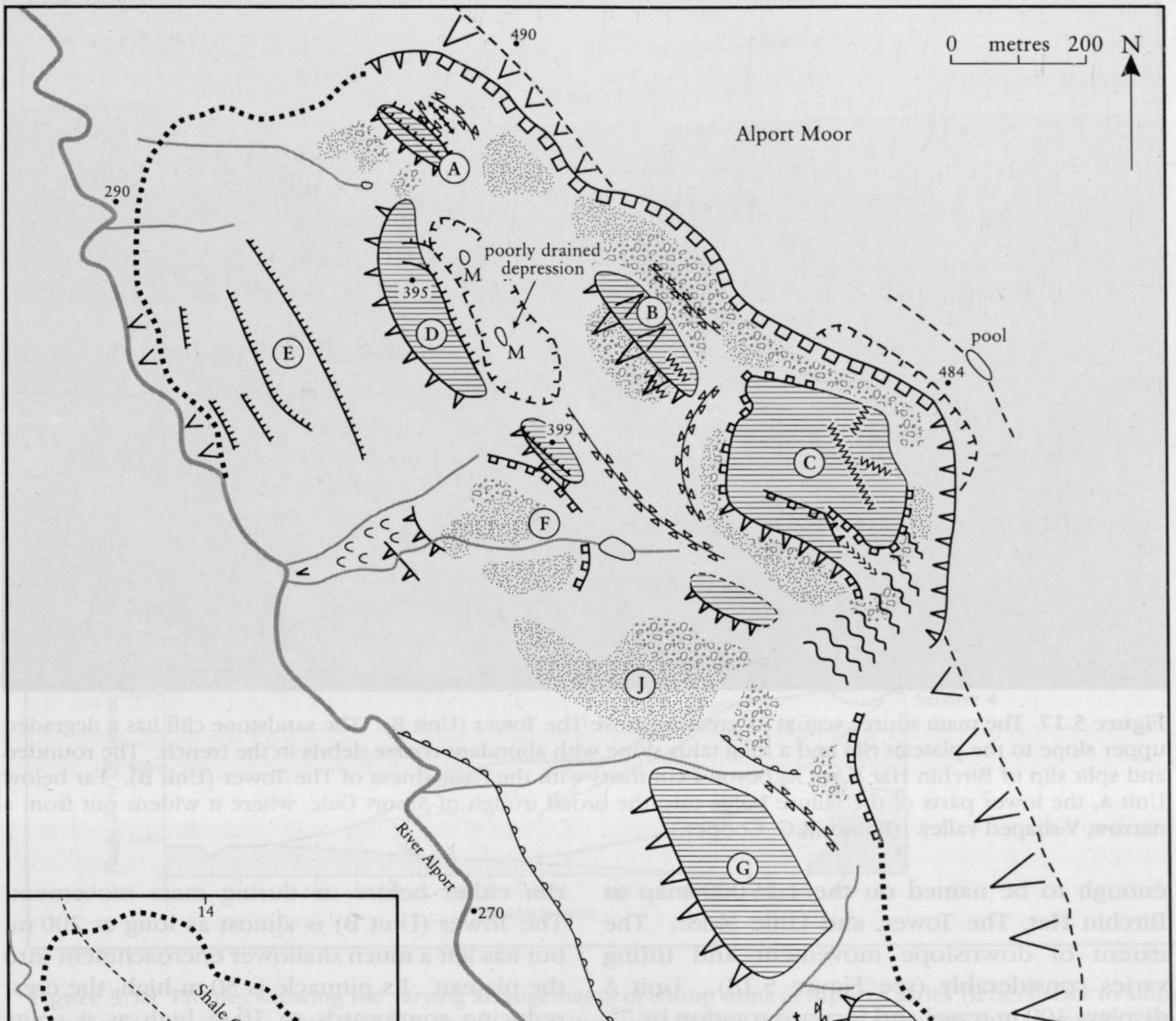


Figure 5.16 Morphological map of the Alport Castles landslip complex, identifying the main slip units described in the text, and indicative geology. The source scar transgresses the original valley rim above Units B and C, but daylights below it elsewhere. After Johnson and Vaughan (1983).

Figure 5.17 The main source scar at its greatest above The Tower (Unit B). The sandstone cliff has a degraded upper slope to the plateau rim and a long talus slope with abundant coarse debris in the trench. The rounded and split slip of Birchin Hat (Unit A) beyond contrasts with the ruggedness of The Tower (Unit B). Far below Unit A, the lower parts of the failure bulge into the broad trough of Alport Dale, where it widens out from a narrow, V-shaped valley. (Photo: R.G. Cooper.)

enough to be named on the 1:25 000 map as Birchin Hat, The Tower, and Little Moor. The extent of downslope movement and tilting varies considerably (see Figure 5.18). Unit A displays 100 m travel and forward rotation by 7°; Unit B displays 130 m travel and considerable backward rotation to leave a sharp crest; while Unit C has only descended 5 m, with a slight valley-ward tilt.

Little Moor (Unit C) is much the largest, about 300 m long (valley-parallel), leaving a deep wedge-shaped bay encroaching into the plateau by some 200 m from the inferred original valley rim. It has moved out by about 70 m, creating a trench 18 m deep which contains several back-tilted slices of rock. Some of these appear to be small rotational slips off the scar, with others produced by backward rotation from the rear of the slipped mass. Open tension fissures up to 8 m deep have begun to dissect the mass, which on its southern side has disintegrated into low chaotic ruckles below a degraded grassy return scarp. A ridge halfway down the steep outer slope appears to be formed by forward rotation, and with similarly tilted blocks on the outer face of Unit B suggests tensional stress on the plateau rim either before or during mass movement. The Tower (Unit B) is almost as long at 200 m, but has left a much shallower encroachment into the plateau. Its pinnacle is 30 m high, the crest reducing southwards to 10 m high as it splits into triple fins with 3 m deep trenches. Birchin Hat (Unit A) is only 100 m long, with a more rounded character; it has split to give 3–5 m anti-scarps (Figure 5.17). The source scar is here well below the valley rim, so that there is a steep grassy slope above the crag.

The rim is essentially intact except above Little Moor, where a 1.5 m furrow runs up to 15 m behind the plateau rim, a proto-slip extending 16 m into the plateau has dropped down by 0.5 m, and a lineament with linear pool lies 45 m in. These indications of incipient or latent failure extend over a length of 300 m.

Below the three 'castellations' a broad swale crosses most of the northern sector. Now partly infilled, it is partly drained by small streams, but below The Tower there is a large depression with ponds. This depression is impounded by another slipped mass, Unit D, which is about 300 m long. Its summit is of the Shale Grit, which constitutes the rim, and it presents an

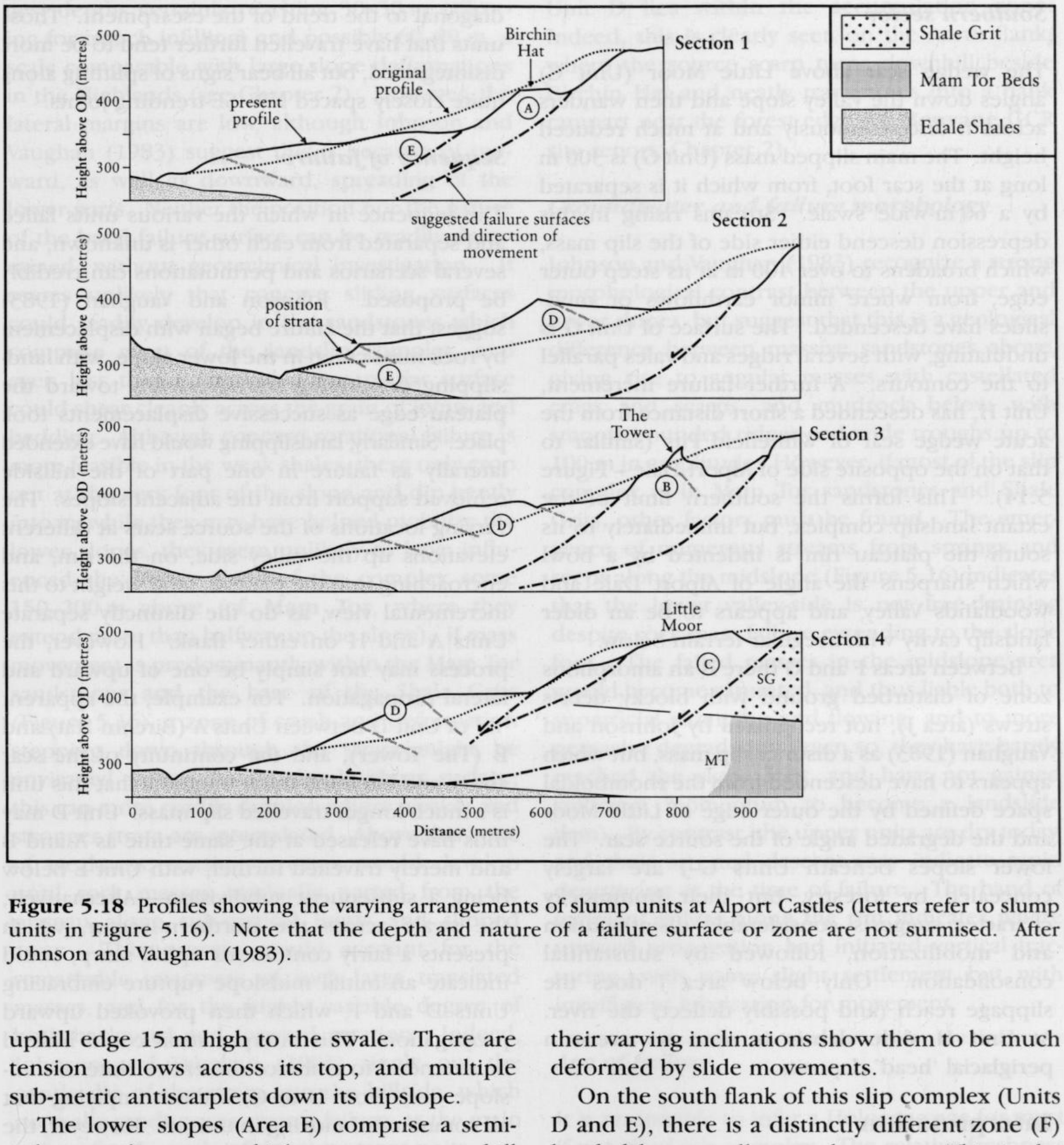


Figure 5.18 Profiles showing the varying arrangements of slump units at Alport Castles (letters refer to slump units in Figure 5.16). Note that the depth and nature of a failure surface or zone are not surmised. After Johnson and Vaughan (1983).

uphill edge 15 m high to the swale. There are tension hollows across its top, and multiple sub-metric antiscarplets down its dipslope.

The lower slopes (Area E) comprise a semi-coherent slip mass which presents an uphill edge 5 m high to a shallow transverse depression. Its surface is generally smooth, with several 1 m antiscarps. It bulges to narrow the valley appreciably, but although the slide toe steepens into a 15–25 m bluff above the river, which meanders in a flood plain upstream of this site, the river does not appear to have been significantly dammed or displaced. Conversely, fluvial erosion of the toe has caused minor undercutting without removing much of the material or re-activating sliding. Mam Tor strata are exposed on the surface and in the river cuts, and while individual elements appear intact, their varying inclinations show them to be much deformed by slide movements.

On the south flank of this slip complex (Units D and E), there is a distinctly different zone (F) headed by a small secondary scar. This is the product of more recent slumping, where saturation by impeded drainage has led to disintegration and in one part of the lower slope an 'earthflow'. The scar reveals Mam Tor beds which dip steeply backwards, suggesting that prior failure of the midslope had predisposed it to more complete collapse. The slump reaches the river, where erosion has produced an 8 m cliff in stiff structureless debris containing < 300 mm sandstone clasts, which contrast with the very weak thin flaggy mudstones revealed in the 'slipped bedrock' cuts, immediately upstream.

Southern sector

The wedge scar above Little Moor (Unit C) angles down the valley slope and then wanders across it discontinuously and at much reduced height. The main slipped mass (Unit G) is 300 m long at the scar foot, from which it is separated by a 60 m-wide swale. Streams rising in this depression descend either side of the slip mass, which broadens to over 400 m at its steep outer edge, from where minor earthflows or mud-slides have descended. The surface of Unit G is undulating, with several ridges and vales parallel to the contours. A further failure increment, Unit H, has descended a short distance from the acute wedge scar of Whitefield Pits (similar to that on the opposite side of Alport Dale – Figure 5.14). This forms the southern limit of the extant landslip complex, but immediately to its south the plateau rim is indented by a bowl which sharpens the angle of Alport Dale and Woodlands Valley, and appears to be an older landslip cavity with irregular terrain below.

Between areas F and G there is an amorphous zone of disturbed ground with blocky debris strews (area J), not recognized by Johnson and Vaughan (1983) as a distinct slip mass, but which appears to have descended from the rhomboidal space defined by the outer edge of Little Moor and the degraded angle of the source scar. The lower slopes beneath Units G–J are largely concealed by forestry, but their hummocky character suggests considerable disintegration and mobilization, followed by substantial consolidation. Only below area J does the slippage reach (and possibly deflect) the river. At Unit H slide debris is seen to over-run periglacial 'head' deposits.

Interpretation

The extensive landslipping here clearly reflects a combination of deep valley incision and geology, with weak strata exposed in the valley floor at the base of the failed slope. The preservation of very large masses of former plateau in varying degrees of intactness reflects the competence of the upper sandstone strata and the gradual nature of their translation. Little Moor (Unit C) is one of the largest known individual slipped masses, with an area of 0.06 km^2 that despite deep fissuring is substantially coherent. Its rhomboidal shape probably relates to a source cavity controlled by near-vertical joint orientations diagonal to the trend of the escarpment. Those units that have travelled further tend to be more disintegrated, but all bear signs of splitting along quite closely spaced NW–SE-trending joints.

Sequence of failure

The sequence in which the various units failed and separated from each other is unknown, and several scenarios and permutations can credibly be proposed. Johnson and Vaughan (1983) suggest that the failure began with displacement by rock-mass creep in the lower slope, with land-slipping developing retrogressively toward the plateau edge as successive displacements took place. Similarly, landslipping would have extended laterally as failure in one part of the hillside removed support from the adjacent slopes. The varying locations of the source scars at different elevations up the valley side, on the rim, and encroaching into the plateau, lend weight to this incremental view, as do the distinctly separate Units A and H on either flank. However, the process may not simply be one of upward and lateral propagation. For example, the apparent 'fit' of Unit D between Units A (Birchin Hat) and B (The Tower), and the continuity of the scar above and between them, suggests that this unit is a much longer travelled slip mass. Unit D may thus have released at the same time as A and B and merely travelled further, with Unit E below being a subsequent subdivision. Alternatively, the swale across the northern sector, which presents a fairly continuous uphill scarp, could indicate an initial midslope rupture embracing Units D and F, which then provoked upward propagation. This interpretation could indeed be extended to embrace the entire suite of mid-slope units from D to G, without implying that all commenced moving in unison. However, the greater degradation of the most enigmatic area J might suggest that failure originated here, with the outer face of Little Moor either being intact plateau rim at that time, or part of a whole central sector (Units C/F/J) which failed at depth *en masse*, with the lower parts becoming more disaggregated, breaking away, and slipping and slumping to the slope foot.

Depth and mode of failure

The depth to which failure extends is equally unclear. The size of the coherent masses, and the heights of the scar and the trench walls, suggest depths certainly reaching 30–50 m (allowing for trench infilling) and possibly 60–80 m, a scale comparable with large slope deformations in the Highlands (see Chapter 2). However, the lateral margins are low, although Johnson and Vaughan (1983) suggest this is because of outward, as well as downward, spreading of the lower parts. Neither the position nor the nature of the basal failure surface can be readily determined, without geotechnical investigation. It seems unlikely that concave sliding surfaces could readily develop in the sandstones which comprise most of the landslip complex, and even less than a through-going planar surface could shear cleanly across the grain of joints and bedding. Although concave rotational failure is more feasible in the weak shales, these only crop out at the very foot of the slope and dip gently into it; while they may have helped mobilize the lower slopes, they seem unlikely to have influenced the higher parts of the complex some 150–200 m above (cf. **Mam Tor**, where they extend more than halfway up the slope). If mass movement is predominantly within the Mam Tor sandstones and the base of the Shale Grits (Figure 5.16), a zone of crush and deformation stepping down through the strata might be envisaged rather than a simple shear surface; this can more readily develop where weaker and stronger strata are intercalated. Above this zone of weakness, tension stresses would develop until rock masses gradually parted from the plateau along sub-vertical joints and slipped away. This process would account for the remarkable intactness of such large translated masses, and for the highly variable degree of both backward and forward rotation. Indeed, Johnson and Vaughan (1983) single out the gentle dip of the strata into the hillside, which normally predisposes against failure, as the main reason for the scale of the movement units.

They also divide the landslip complex into zones of depletion and accumulation, following Varnes (1978), whereby the latter zone stands proud at a higher level than the original (pre-failure) ground surface (Figure 5.18). This can arise either by debris over-running the intact lower slope, or by the landslip mass bulging out under compression. The latter must apply here, if failure has propagated from the slope foot, leaving no original ground surface in place; the antiscarped character of Area E attests to such compression. Johnson and Vaughan (1983) place the transition along the 'swale', such that Unit D lies within the accumulation zone. Indeed, this is clearly seen on the north flank, where the source scarp turns downhill beside Birchin Hat and neatly transmutes into a flank rampart near the forest edge (cf. **Benvane** GCR site report, Chapter 2).

Groundwater and failure morphology

Johnson and Vaughan (1983) recognize a strong morphological contrast between the upper and lower slopes, but suggest that this is a geological difference between massive sandstones above, giving rise to angular masses with castellated crags and scarps, and mudrock below, with smooth rounded ridges and wide troughs up to 100 m in amplitude. However, if most of the slip complex is in Mam Tor sandstones and Shale Grits, other factors must be found. The emergence of numerous streams from springs and seeps along the midslope (Figure 5.16) indicates that the lower valley-side is not free-draining despite rock-mass failure extending to the slope foot. The failed masses in the midslope area would become saturated, and thus liable both to superficial slumping and flowing, and to more pervasive degradation (even so, they have barely reached the slope foot, and have not gained sufficient momentum to become a landslide dam). By contrast, the upper units are dry today, and their arrested descent may indicate rapid dewatering at the time of failure. The band of incipient failure along the rim indicates where upward propagation had initiated vertical fracturing, with some slight settlement but with insufficient lubrication for movement.

Age of failure

It is reasonable to infer a Holocene age for most if not all of this complex. The relative freshness of much of the upper morphology implies lack of periglacial attrition, although the top 5–8 m of the scarp above The Tower is a battered grass slope in thinner or deep-weathered strata (an unusual hazard requiring fencing). The over-riding of periglacial head by Unit H has been noted, and pollen from a small peat lens in the slide toe suggests that the flows are not more than 8300 years old (Johnson and Vaughan, 1983). However, this need not preclude a history of landslipping here and in the vicinity earlier in the Quaternary. Alport Dale has a fluvially incised character in its upper reaches on the Bleaklow moors, but widens and straightens at the slide locus; the extent to which erosion by local glacier ice has played any part in slope destabilization merits further exploration in the Pennines.

Conclusions

Alport Castles is one of the largest landslip complexes on the sedimentary lithologies of inland Britain. It is particularly remarkable for the size and relative intactness of its individual movement units, some of which are striking and well-known landscape features. It clearly displays geological controls on both its location and its topography. The depth to which failure extends, the nature of the translation surface or zone, and the sequence of evolution are largely unknown, and Alport Castles presents excellent opportunities for further research of wider relevance in the Pennines. The model of upward propagation, after rupturing in weak strata exposed by valley incision, has been applied here and may account for the freshness of the uppermost units and the boldness of the main scar, which attains an exceptional 60 m plus in height. The scale of encroachment into the plateau by up to 200 m at Little Moor, with signs of further incipient extension, exemplifies the contribution of bedrock mass movement to valley widening, with local rates of scarp retreat vastly in excess of those yielded by all other slope processes. This is far from being an isolated case (cf. **Beinn Fhada**, Chapter 2; **Trotternish Escarpment**, Chapter 6), and represents an extensive suite of such slope failures in the vicinity and in similar lithological contexts across the Dark Peak and farther north in the Pennines.

CANYARDS HILLS, SHEFFIELD (SK 250 948)

R.G. Cooper

Introduction

The Canyards Hills GCR site is an area of irregularly ridged ground downslope from a 10 m-high vertical scar, in the Ewden Valley, south of Broomhead Reservoir, near Bradfield in South Yorkshire. The ridges are the complex physiographical expression of a large landslip, 1 km long from west to east, extending downslope from the scar for at least 0.4 km (Figure 5.19).

Canyards Hills is in the upper part of the Millstone Grit succession (Namurian, Upper Carboniferous), but higher in the succession than at **Mam Tor** or **Alport Castles**. The site is formed in Beacon Hill Flags and the Huddersfield White Rock, with a thick series of shales in between (Elliott, 1979). South-east of Wigtwizzle (Wightwizzle) the Huddersfield White Rock forms a gently sloping plateau with a steep northern scarp face. The north-easterly dip is causing the rock to slip over the underlying shales, and great masses of slipped material cover the slopes below (Figure 5.20). The western part of the plateau has been reduced by this process of denudation to a tongue of high ground only 180 m wide (Bromehead *et al.*, 1933). The Huddersfield White Rock exposed in the scar consists of massive well-bedded and open-jointed sandstones less than 30 m thick, dipping 6° north-eastwards, i.e. towards the river. Overall, the slope is concave in profile. The shale outcrop occupies the longest downslope segment of the slope in the western part of the area, and the profile is most concave there also. The landslide area also occupies the longest segment of the slope in the west, but does not reach the river until the centre and east (Bass, 1954).

Hunter (1869, writing in 1819) described the landslide as 'The Canyers, a range of conical hills

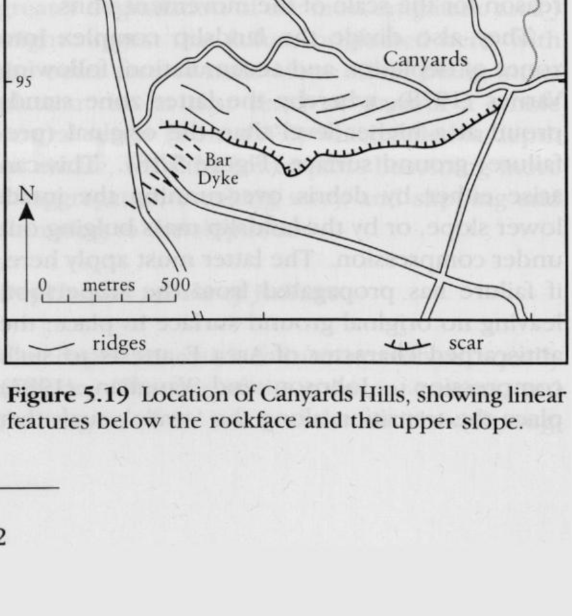


Figure 5.19 Location of Canyards Hills, showing linear features below the rockface and the upper slope.

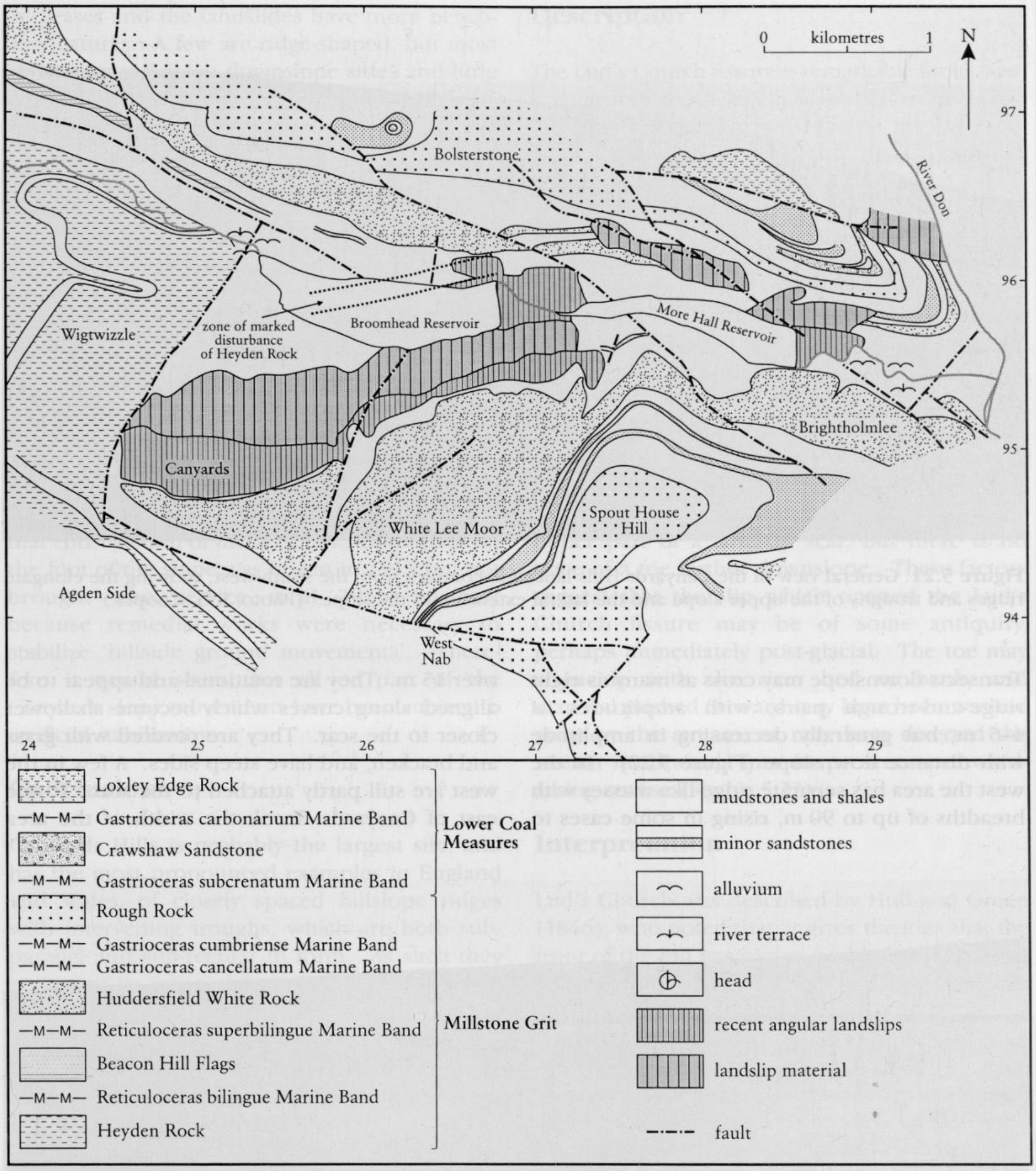


Figure 5.20 Geological map of the setting of the Canyard Hills landslide complex south of the Broomhead Reservoir.

stretching about a mile', while Hepworth (1954) remarks, 'Canyard Hills were formerly called 'Kenhere' or Kenyer Hills'. These variations of name (now fixed by the Ordnance Survey as 'Canyards') probably result from the remote location, but the landslip certainly does not consist of 'conical' hills.

Description

The site is chiefly remarkable for the very large number of irregular ridges running along the slope approximately parallel to the cliff-face (Figure 5.21). They enclose numerous poorly drained and often marshy elongate troughs.

Figure 5.21 General view of the Canyards Hills landslide complex from the south-west, showing the elongate ridges and troughs of the upper slope and the lateral extension of the slope. (Photo: R.G. Cooper.)

Transects downslope may cross as many as eight ridge-and-trough pairs, with amplitudes of 4–5 m, but generally decreasing in amplitude with distance downslope (Figure 5.22). In the west the area has separate ridge-like masses with breadths of up to 90 m, rising in some cases to over 15 m. They are rotational and appear to be aligned along curves which become shallower closer to the scar. They are covered with grass and bracken, and have steep sides. A few in the west are still partly attached to the scar. To the east of Canyards Brook the width of the area

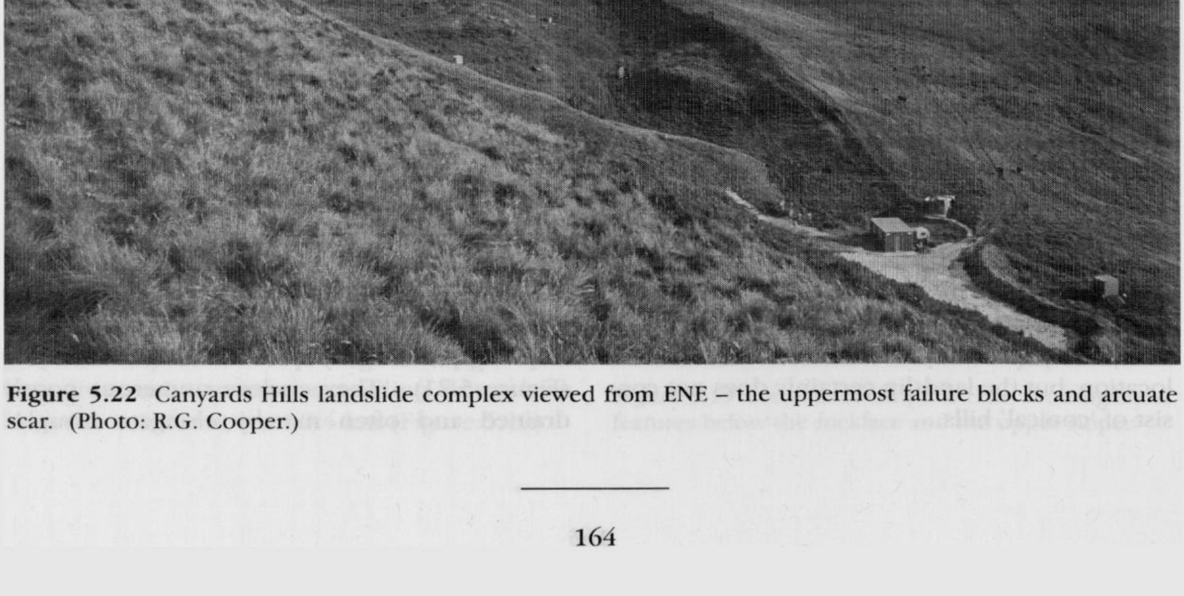


Figure 5.22 Canyards Hills landslide complex viewed from ENE – the uppermost failure blocks and arcuate scar. (Photo: R.G. Cooper.)

decreases and the landslides have more bench-like features. A few are ridge-shaped, but most have wide and steep downslope sides and little or no backslope (Bass, 1954).

Interpretation

The features are thought to result from major break-up of the sliding mass during a single movement, as there is little to suggest that a succession of upper slides has taken place causing gradual cliff recession. The ridged physiography is probably due to break-up along the lines of fairly closely spaced pre-existing joints in the Huddersfield White Rock. This physiography is similar to that developed in shales, Shale Grit, siltstones and sandy shales of the Namurian at Bretton Clough, 18 km to the south (Boggett, 1989). Wood (1949) records that construction of the Broomhead Reservoir at the foot of the slope was begun in 1913 but not brought into service until 1936, in part because remedial works were necessary to stabilize 'hillside ground movements'. Therefore it would seem that all or part of the Canyards slope movements have been subject to artificial stabilization.

Conclusions

Canyards Hills is probably the largest site, and has the most pronounced examples in England and Wales, of closely spaced hillslope ridges with intervening troughs, which are both sub-parallel and sub-regular in form. As such they represent an unusual form of lateral extension failure caused by retrogressive unloading along a weak Namurian shale layer. The associated ridge-trough form is due to the coherence of the Huddersfield White Rock and the joint spacing.

LUD'S CHURCH, NORTH STAFFORDSHIRE (SJ 987 656)

R.G. Cooper

Introduction

Lud's Church is a vertical fissure in Roaches Grit (Namurian R_2b; C.M. Jones, 1980). It lies on a north-facing slope in Back Forest, overlooking the River Dane, in the Staffordshire moorlands 10 km to the north of Leek (Figure 5.23).

Description

The Lud's Church fissure is remarkable for its size: it measures about 165 m from end to end, and including all its side passages its length totals 220 m. For much of its length it is 4–5 m wide, and up to 18 m deep (Figures 5.23 and 5.24).

Associated with the fissure are hillside trenches and their associated intervening ridges ('ridge-and-trough' features) and a curious tor known as 'Castle Cliff Rocks'. This is sited 70 m from the fissure, and at a similar position about halfway down the slope. It rises 4 m above the surrounding soil surface. Its location on the slope is unusual in the Pennines (Palmer and Radley, 1961) and probably relates to its position on the surface of a slipped mass. There is a short steeper section upslope from Lud's Church, which could be interpreted as the degraded upper part of a landslip scar, but there is no trace of a toe farther downslope. These factors suggest that the slip which opened the Lud's Church fissure may be of some antiquity, perhaps immediately post-glacial. The toe may co-incide with the river bed or bank, and material pushed forward may have been washed away by the stream, or may have diverted the stream northwards into its present northward-arcing course (Figure 5.23).

Interpretation

Lud's Church was described by Hull and Green (1866), who noted that 'it gives the idea that the front of the hill has parted bodily from the main mass, and slipped a little forward, leaving this fissure along the line of fracture', (if this is correct the fissure is a tension crack or 'gull' marking the backscar of a landslide). More recently Millward and Robinson (1975) ascribed its origins to post-glacial incision of the River Dane. Lud's Church appears to have been formed as a result of the detachment of a large sliding mass as it began to move valley-ward over a possibly irregularly shaped slip-plane. The possible backface scar on the slope profile suggests that the Lud's Church fissure may be *within* the slipped mass, both of its walls having moved with the main slipped mass, followed or accompanied by a more surficial movement (at least 18 m thick) as the fissure opened. Cooper *et al.* (1977) noted the presence within it of fissures that are roofed-over by fallen boulders, forming covered tunnels up to 12 m deep.

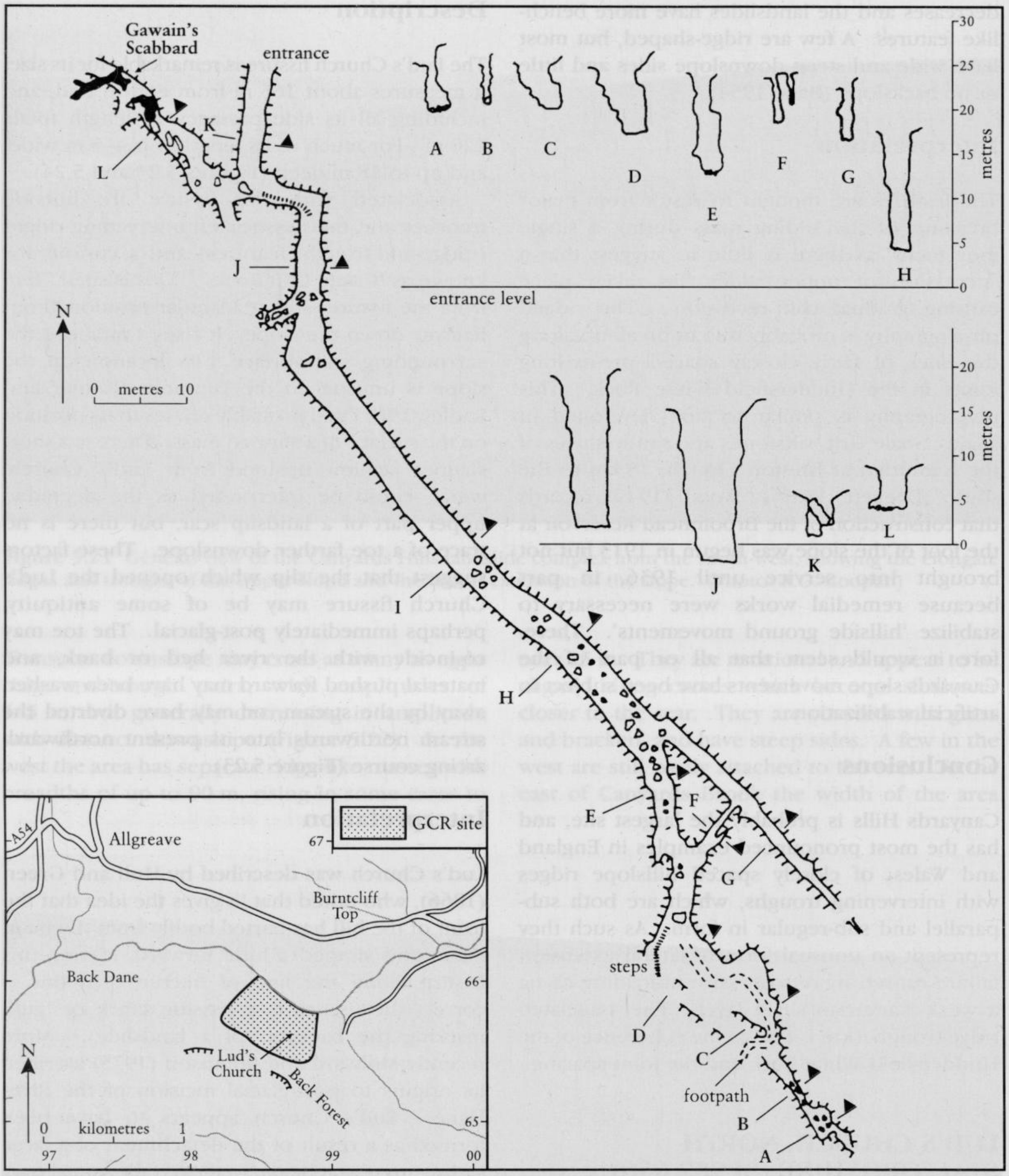


Figure 5.23 The location and general morphology of Lud's Church, Staffordshire.

Aitkenhead *et al.* (1985) describe it as 'a spectacular example in sandstone, of bedding-plane slip, which is common in major sand-stone units and in mudstone-with-sandstone sequences.' However, there is no feature lower down the same slope corresponding to the lower end of the mass which has slipped on the bedding plane. While Aitkenhead *et al.* (1985) seem to describe an essentially translational movement, it is uncertain whether the main movement, which opened the Lud's Church fissure, was translational, or whether material within a rotationally upper mass involving the whole of the slope from crest to stream, underwent a small

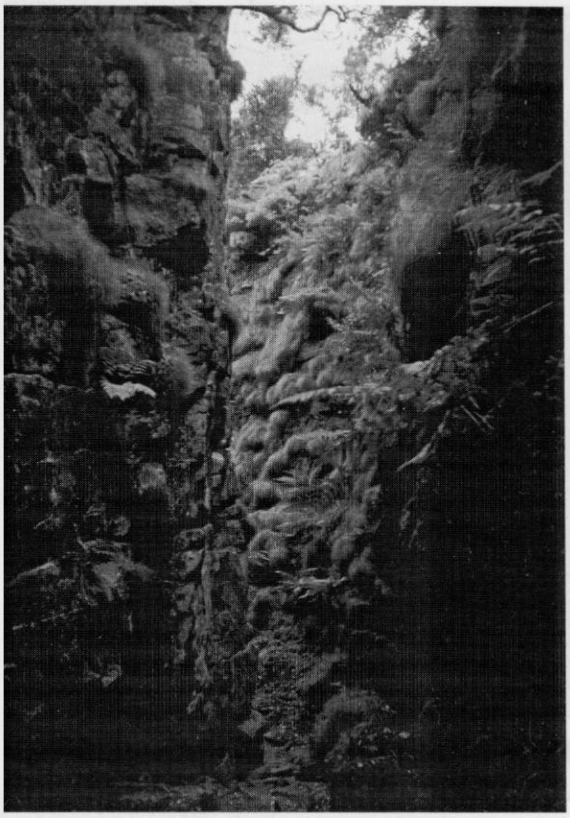


Figure 5.24 View along the 'labyrinth' of the Lud's Church fissure. (Photo: R.G. Cooper.)

translational movement equal to the width of Lud's Church.

Elliott's (1977) meticulous identification of Lud's Church with the 'Green Chapel' of the medieval alliterative poem *Sir Gawain and the Green Knight* (Tolkein and Gordon, 1967; Stone, 1974) may provide some indication concerning its age. The unknown *Gawain*-poet wrote in a north Midlands dialect which has been identified as late 14th century. It may be concluded, tentatively, that the Lud's Church fissure was both open, and wide enough to allow axe-swinging men to fight in, more than six hundred years ago.

Conclusions

Lud's Church is of educational importance because of the unique opportunity it provides to walk on a reasonably easy footpath through the interior of a large-scale 'detaching' landslide, and examine it from within. It is by far the largest such fissure within a landslipped mass in Great Britain, and may be the best example of a 'rock labyrinth' or 'lattice' structure formed by unloading.

MAM TOR, DERBYSHIRE (SK 130 836)

R.G. Cooper and D. Jarman

Introduction

Mam Tor is the locus for one of the most conspicuous and active landslides in the inland sedimentary strata of Britain. It has long been well known as the 'shivering mountain'. The head of the Hope Valley in the Peak District (Figure 5.25) rises steeply to the little plateau top of Mam Tor (517 m), the side of which has sheared away leaving a bold rock scar 80 m high above talus and the top of the slip mass (Figure 5.26). Failure probably occurred as a single event in about 3600 BP, an unusually recent date for inland areas. It is inferred that the initial slide came to rest at a marginally unstable angle, rendering it liable to re-activation after periods of heavy rain. It is believed to have extended downslope by over 500 m at very slow and gradually reducing rates of intermittent creep, but whereas almost all similar failures in the Pennines are now inactive, Mam Tor is exceptional in being predicted to continue moving for a long time to come. This reflects the unusually weak pyritic shales that both underlie the slide and comprise a proportion of the debris.

It is unusual for an inland slide in Britain to interfere with important communication routes. Here, the original direct road between Manchester and Sheffield (A625) became so affected by repeated slippage (Figure 5.27) that it was closed in 1979, necessitating unprecedentedly lengthy detours for trunk traffic. This has prompted a series of detailed geotechnical studies including Lounsbury (1962), Brown (1966, 1977), Stevenson and Gaunt (1971), Lant (1973), Lupini (1980), Al-Dabbagh (1985), and notably Skempton *et al.* (1989), making Mam Tor probably the best understood mass movement of natural origin in inland Britain. It provides an invaluable point of reference for interpreting similar and contrasting sites. It also has the benefit of a guide for visitors interested in its geology (Cripps and Hird, 1992), while Derbyshire County Council has placed an informative board at the foot of the closed road.

Although the Mam Tor slide is of only medium extent, affecting 0.35 km² (Figures 5.28 and 5.29), it is of wider significance as part of a cluster of four major extant landslips which collectively

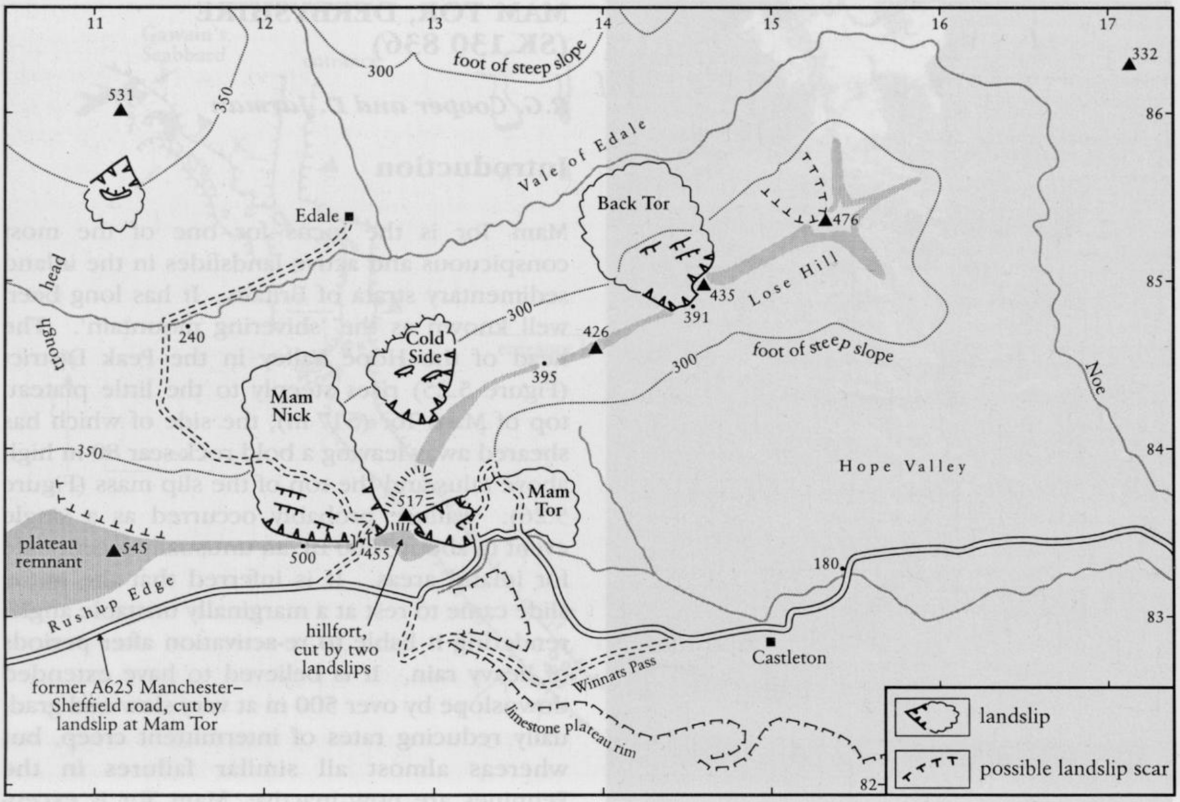


Figure 5.25 Location of the Mam Tor landslide, showing other major landslides also encroaching into Mam Tor hillfort and Rushup–Lose Hill ridge, and the former trunk road severed by it.

Figure 5.26 The Mam Tor landslide scar from the top of the upper slump sector. (Photo: M. Murphy, English Nature/Natural England.)

Figure 5.27 Oblique aerial photograph of the Mam Tor area, showing the old road which was closed in 1979 owing to repeated slippage. The scar of the landslip is clearly visible, as well as a slumped mass. (Photo: © National Trust/High Peak.)

have encroached into the Rushup Edge–Mam Tor–Lose Hill ridge (Figure 5.25). The summit of Mam Tor is ringed by the earthworks of a hill-fort, except for two sections where landslip scars cut into the hill from opposite sides, adding an archaeological dimension to site interpretation.

Description

As with many landslip locations in the 'Millstone Grit' areas of the Pennines, the slopes of Mam Tor are predisposed to mass movement by virtue of successions of weak strata underlying more competent rocks. Here, the top 100 m are of Mam Tor Beds in which micaceous sandstones alternate with siltstone and shale (Spears and Amin, 1981; Figure 5.30). Beneath them, the Edale Shales are hard mudstones with occasional bands of siltstone and ironstone. Pyrite occurs at several horizons, generally as scattered crystalline aggregations. Below some metres of weathered material, the mudstones are weakened by fissuring to about 10 m depth, probably resulting from stress-release after valley erosion, accentuated by Pleistocene permafrost action (Skempton *et al.*, 1989; cf. **Rowlee Bridge** GCR site report, this chapter). The strata dip NNE at about 8°.

The scar itself stands at an average angle of 45° (range 40°–51°) with a crest at 510 m above OD, and a free-face height of 80 m, entirely in Mam Tor Beds (Figure 5.29). It is an asymmetrical rectilinear wedge, but with little evidence of strong joint control. Talus extends about 30 m down from the scar foot at an average slope of 23° (range 18°–28°). The partially evacuated cavity has minor rockfalls and slumps within it. The debris mass has extended to 1000 m long at an average gradient of 12°; it is gently convex and attains 450 m wide, although the scar itself is only about 200 m wide.

Skempton *et al.* (1989) divide the landslide debris into three sectors (Figure 5.28).

(A) Upper 'slump' sector

The upper slump sector is that part of the initial slide mass that has travelled a relatively short distance, and remains largely where it first came to rest before subsequent extension of its toe. It extends from about 370 m above OD down to

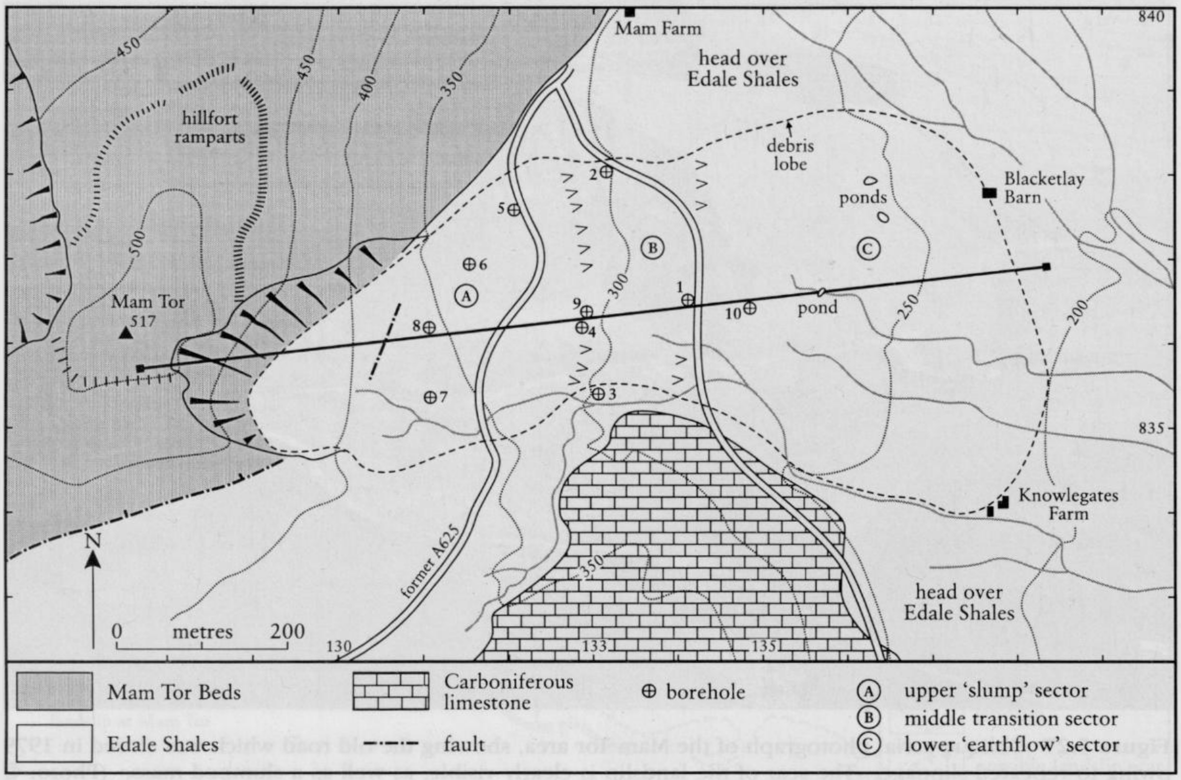


Figure 5.28 Schematic plan of the Mam Tor landslide, showing sectors, geology, borehole locations, and the former trunk road. The line running almost west–east is the line-of-section shown in Figure 5.29. After Skempton *et al.* (1989).

about 310 m above OD. Geological markers indicate that the slump mass has moved about 160 m (Figure 5.31), although the actual distance from the centre of the cavity to the lower contour is over 400 m (Figure 5.29). The slump has a very irregular surface, initially rather flat-topped then steepening where traversed by the upper leg of the former road. Borehole 8

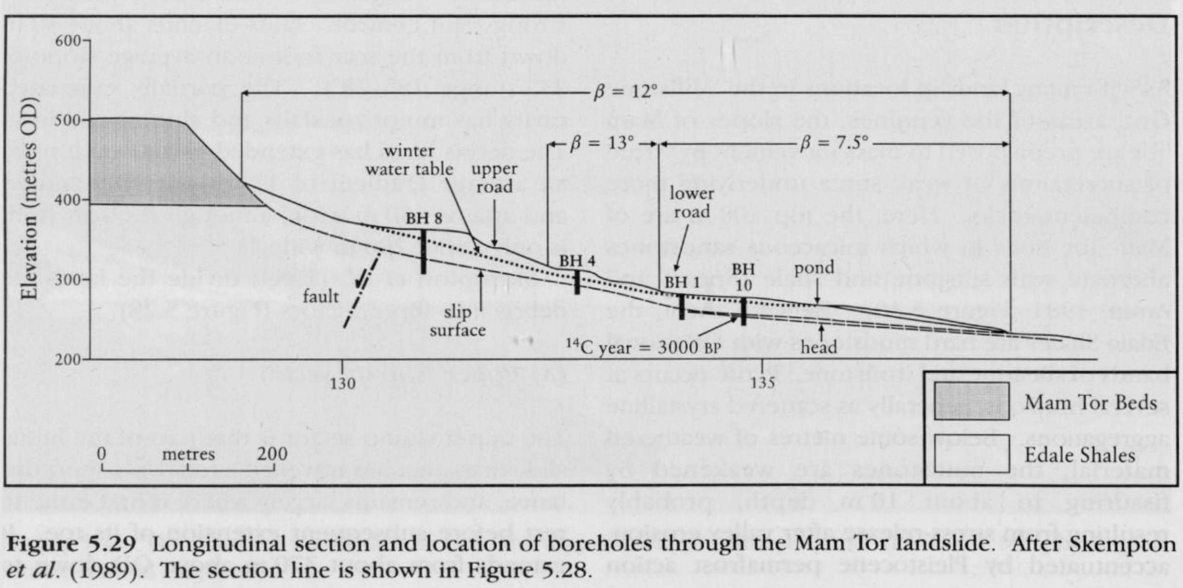
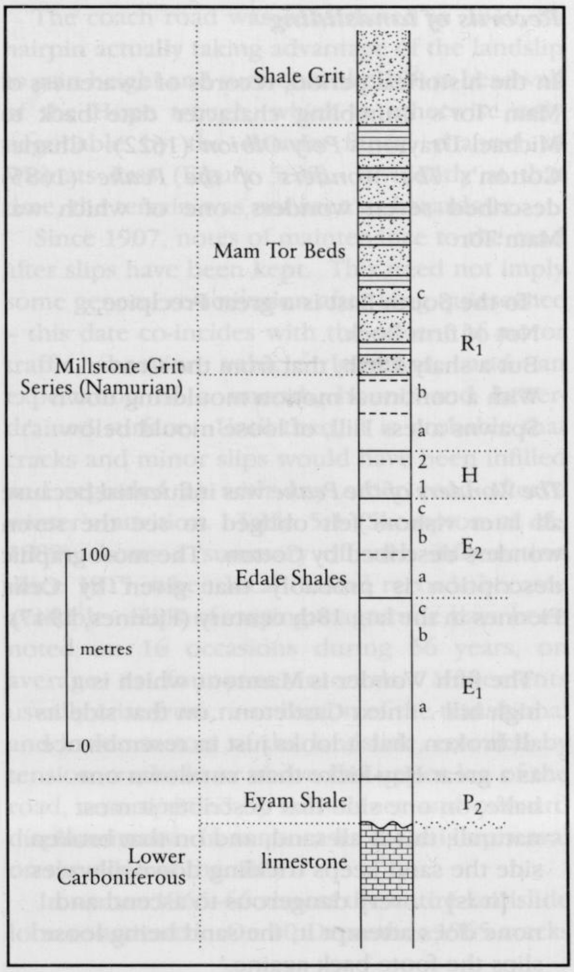


Figure 5.29 Longitudinal section and location of boreholes through the Mam Tor landslide. After Skempton *et al.* (1989). The section line is shown in Figure 5.28.



◂**Figure 5.30** Stratigraphical section at Mam Tor. After Skempton *et al.* (1989).

(Figures 5.28 and 5.29) proved a slide thickness of 32 m here, with two polished and striated slip-surfaces, 10 cm apart, near the base of 2.6 m of brecciated clay. This shear zone cuts through unweathered Edale mudstone, of which at least 8 m has been removed (in addition to weathered material and superficial deposits). Landslide debris includes pieces of mudstone from the upper Edale Shales (Zone R_{1a}) in the base, and 10 m above there are blocks of micaceous sandstone from the Mam Tor Beds, back-tilted at 40°–45°. Though broken and distorted, and missing a portion of Zone R_{1b}, the rocks are recognizably in their correct sequence, and show a displacement along the slip-surface here of about 160 m. Similar features were found in several other boreholes (Skempton *et al.*, 1989).

(B) Middle transition sector

The middle transition sector extends for about 150 m down to the lower leg of the road at about 280 m above OD. The terrain is less irregular but still quite steep, ranging between 9°–16°, and with transverse compression ridges. The debris is generally less than 20 m thick

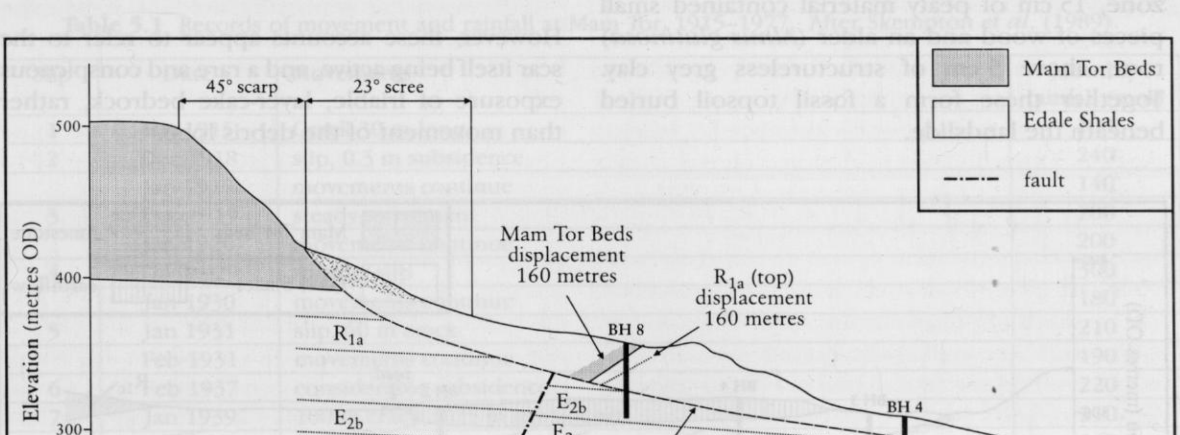


Figure 5.31 Geological and morphological details of the scar and the upper slump sector of the Mam Tor landslide, showing the locations of boreholes 4 and 8 (BH 4 and BH 8). After Skempton *et al.* (1989).

(Figure 5.32), and becomes very thin at the road. Borehole 4 proved a slide thickness of 11.8 m to a 1.8 m shear zone of brecciated clay with a thin layer of intensely sheared clay at its base on or just below the top of the weathered mudstone, which was disrupted for a further 2.2 m. This indicates that by this point no intact bedrock was incorporated in the slide.

(C) Lower 'earthflow' sector

The lower earthflow sector extends for about 420 m, down to a present lowest point of 220 m above OD. The slope varies between 6°–9°, and the deposit is generally less than 10 m thick. This part of the landslide has moved in almost pure translation by sliding on or just below the original ground surface, which is here little disturbed. The flow stands above the adjacent ground with clearly defined flanks and toe. The surface is hummocky, with transverse ridges in the upper parts, and contrasts strongly with the smooth fields alongside. The slide tongue has spread laterally, widening to 450 m compared with 250 m in the upper sector. Ponds and marshes indicate high winter groundwater levels, while there are perennial surface streams.

Near the head of this 'earthflow' Borehole 10 proved a slide thickness of 19.3 m, with a slip-surface 10 cm above the base of 1.2 m of brecciated clay. Immediately beneath this shear zone, 15 cm of peaty material contained small pieces of wood and an alder (*Alnus glutinosa*) root, above 5 cm of structureless grey clay. Together these form a fossil topsoil buried beneath the landslide.

Records of landsliding

In the historical period, records of awareness of Mam Tor's crumbling character date back to Michael Drayton's *Poly-Olbion* (1622). Charles Cotton's *The Wonders of the Peake* (1685) described seven wonders, one of which was Mam Tor:

> 'To the South-East is a great Precipice,
> Not of firm Rock...
> But a shaly Earth, that from the Crown
> With a continual motion mouldring down
> Spawns a less Hill, of loose mould below...'

The Wonders of the Peake was influential because all later visitors felt obliged to see the seven wonders described by Cotton. The most graphic description is probably that given by Celia Fiennes in the late 18th century (Fiennes, 1947):

> 'The fifth Wonder is Mamtour which is a high hill ... next Castleton...on that side its all broken that it looks just in resemblance as a great Hay-Ricke thats cut down one halfe, on one side that describes it most naturall, this is all sand, and on that broken side the sand keeps trickling down allwayes ... [it is] ... very dangerous to ascend and none does attempt it, the sand being loose slips the foote back againe.'

However, these accounts appear to refer to the scar itself being active, and a rare and conspicuous exposure of friable, layer-cake bedrock, rather than movement of the debris lobe.

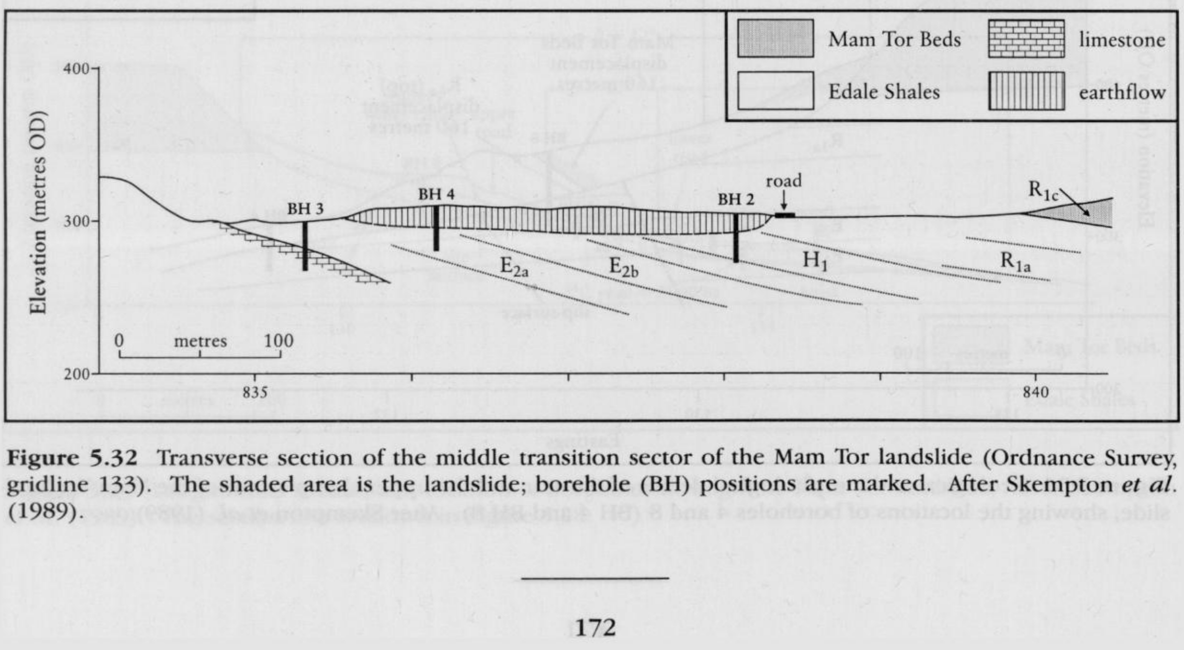


Figure 5.32 Transverse section of the middle transition sector of the Mam Tor landslide (Ordnance Survey, gridline 133). The shaded area is the landslide; borehole (BH) positions are marked. After Skempton *et al.* (1989).

The coach road was constructed in 1810, its hairpin actually taking advantage of the landslip to gain height and surmount the 100 m headwall of the Hope trough, which is otherwise only negotiable by the narrow fluvial channel of Winnats Pass (Figure 5.25). Evidently at that time, the terrain was not seen as hazardous.

Since 1907, notes of maintenance to the road after slips have been kept. This need not imply some general re-activation after long quiescence – this date co-incides with the advent of motor traffic, heavier vehicle loadings, and an expectation of a smooth, bound and better-drained surface. Until then, it is probable that cracks and minor slips would have been infilled and regraded as with any other road after a winter's attrition. Table 5.1 (Skempton *et al.*, 1989) shows a summary of such information since 1915, when local rainfall records became available. Slips of varying magnitude have been noted on 16 occasions during 66 years, on average at four-year intervals. Movements usually arise from re-activation of the transitional and lower sectors of the landslide, revealed by tension cracks in or above the upper leg of the road, accompanied by subsidence and outward displacements. In some cases there is upheaval on the edge of the slide.

In winter 1965–66 almost the entire landslide lobe re-activated. On 10 December 1965 cracks appeared following 120 mm of rain in six days. Abrupt movements were noticed on 18, 23 and 29 December, in each case within a day of further rainy spells. By mid-January, when movement had practically ceased, the total displacement in the upper sector amounted to 0.7 m, and shear displacements of about 0.4 m were observed where the road crosses the flanks of the slide (Brown, 1966). Most of this movement would have taken place during the last 20 days of December at an average rate of around 30 mm per day. The upper road subsided by as much as 1.5 m in places and a local 'confined' slip developed over a short width below the road. Activity renewed in February, mainly in response to 100 mm of rain in 10 days towards the end of the month. The rate at that period was about 15 mm per day and diminished almost to zero by mid-March (Figure 5.33).

Movements at the lower road are less than at the upper road, an observation consistent with the existence of compression ridges; indeed, it remains open for access to Mam Farm. Forward movements of the toe are therefore smaller than those at the upper road, though on a long timescale the difference cannot be great, and clear proof of advance at the toe in recent times is provided by slide debris encroaching on Blacketlay barn (see Figure 5.28).

Table 5.1 Records of movement and rainfall at Mam Tor, 1915–1977. After Skempton *et al.* (1989).

Slip	Date	Movements	Monthly rainfall (mm)
1	Jan 1915	Crack 30 m long	200
2	Dec 1918	slip, 0.3 m subsidence	240
	Jan 1919	movements continue	140
3	Dec 1919	steady movement	280
	Jan 1920	movements continue	200
4	Dec 1929	serious slip	300
	Jan 1930	movements continue	180
5	Jan 1931	slip, 60 m crack	210
	Feb 1931	movements continue	190
6	Feb 1937	considerable subsidence	220
7	Jan 1939	100 m crack, 0.25 m subsidence	210
8	Oct 1942	30 m crack, 0.1 m subsidence	160
9	Feb 1946	extensive slip	240
10	Nov 1946	new movements	230
11	Feb 1948	subsidence on 200 m length (preceded by 280 mm rain in Jan)	100
12	Dec 1949	slip (no details)	230
13	Jan 1952	large slip (preceded by 400 mm rain in November and December)	150
14	Dec 1965	serious slip, 0.7 m displacement	320
15	Feb 1966	renewed movement, 0.3 m displacement (preceded by 385 mm rain in December and January)	190
16	Feb 1977	large slip; 0.4 m subsidence (average)	230

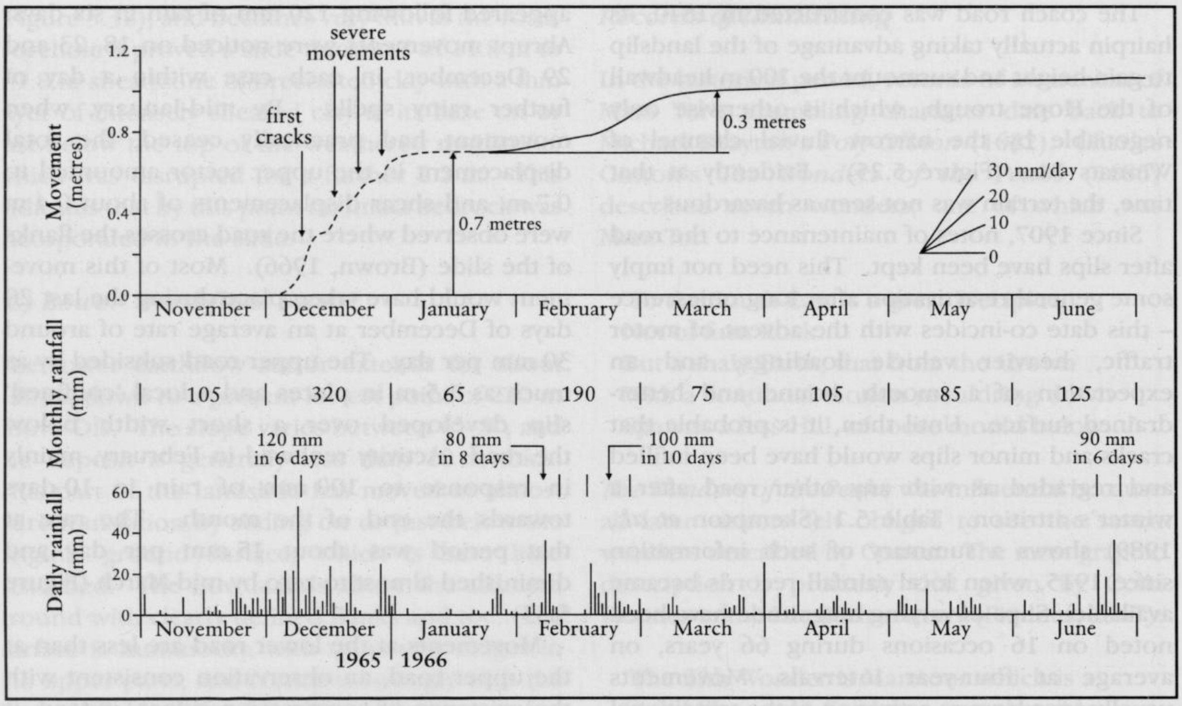


Figure 5.33 The relationship between rainfall and landslide movements, winter 1956–1966. Recorded by Brown (1966), updated by Skempton *et al*. (1989).

From observations in 1918, 1939, 1965, 1966 and 1977 the average displacement of the upper road when a re-activation slip occurs is about 0.3 m, allowing for the fact that not all slips involve the full width. With a return period of four years this is equivalent to 7.5 m per century, and because the toe of the landslip will have moved by a rather smaller amount the present rate of advance is likely to be about 7 m per century.

Rainfall and groundwater

As is commonly reported in Britain and elsewhere, landslide mobilization can relate closely to groundwater availability, both seasonally and in relation to peak rainfall events. The unusually detailed records for Mam Tor exemplify this relationship. Remobilization events are most frequent in December–February, while rainfall is heaviest from October to February (Figure 5.34). Records from local rainfall stations were analysed by Skempton *et al*. (1989) in terms of the return periods of rainfall amounts over 3-day, 6-day, 10-day, and 1-month periods. Comparing these with the situation at Mam Tor in December 1965 shows that instability was almost certain to occur at this time, and in the case of the February 1966 situation (Figure 5.33) when the 10-day rainfall was around 100 mm, the analysis showed that for every ten such events about five may be expected to result in a slip. When no slip occurs, this is likely to be because winter groundwater has been lower than average.

Observations from piezometers installed at or near the shear zone for short periods in 1977 and 1978 give an indication of the seasonal response of the landslide's groundwater levels to rainfall (Figures 5.35 and 5.36). In winter months, when the soil moisture deficit is effectively zero, the greater part of the rainfall not lost as runoff penetrates to augment groundwater. Under such conditions the 'storm response' in groundwater level is more or less directly proportional to the rainfall. While the ratio may vary locally with permeability, slope angle, depth to water table and intensity of rainfall, in uniform strata the response, seasonal as well as short term, is practically the same at different depths. Thus for winter rainstorms capable of causing substantial movements, the corresponding transient groundwater rise in the lower sector is about 0.5 m. In the upper sector,

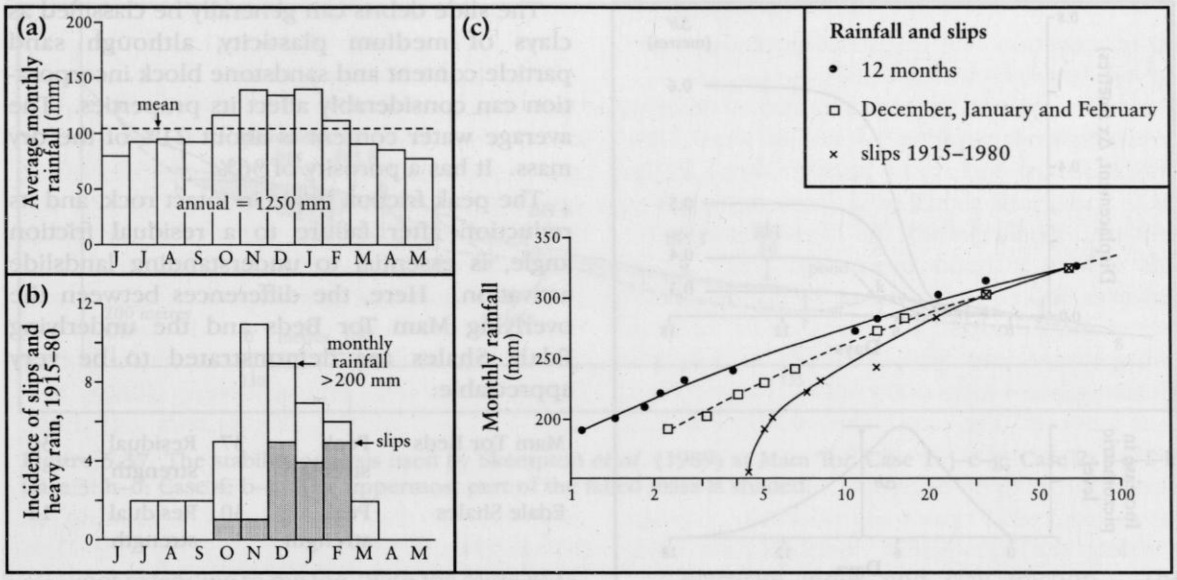


Figure 5.34 The relationship between average monthly rainfall and the incidence of heavy rain (1915–1980) and the return periods of monthly rainfall on an annual and winter basis. After Skempton *et al.* (1989).

permeability is higher, where broken sandstone exists in the debris, and winter groundwater is at a greater depth; the average slope is steeper, but runoff from the scarp face will contribute a throughflow component. Therefore storm response in this part of the landslide is not very different from that in the lower sector, and in any case it is unlikely, even as an upper limit, to exceed the seasonal response of 0.7 m measured in a borehole in the upper sector.

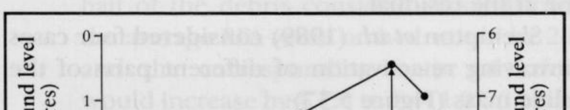


Figure 5.35 Piezometric levels recorded at Mam Tor during 1977 and 1978. After Skempton *et al.* (1989).

The chemistry of the groundwater at Mam Tor has been studied, with results of great significance. This is probably the classic discovery of the importance of the role of pyrite weathering reactions in lowering the residual shear-strength of the mobile horizon of a large landslide. Oxidation of pyrite (FeS_2) within the mudstones forms sulphuric acid, which liberates iron, calcium and other elements into solution. Chemical analyses have shown this process to be operating in the landslide (Vear and Curtis, 1981) because water issuing from springs and seeping off the lower sector in winter is acidic with an ion concentration much in excess of that in runoff from adjacent slopes. Steward and Cripps (1983) have shown that the residual shear-strength of the pyritic Edale Shale near Castleton is sensitive to modification of both mineralogical and porewater composition. At Mam Tor, weathering solutions penetrate deeply into the landslide and may attack fresh shale below the slip-surface, thus reducing its strength. Over a period of years, this would then reduce the factor of safety to a value close to unity so that other destabilizing effects could initiate a major failure event. Of course, the effect of this leaching on the strength of the landslide materials is likely to be small in the short term; strength and other geotechnical properties determined by Skempton *et al.* (1989) are those resulting from at least 3000

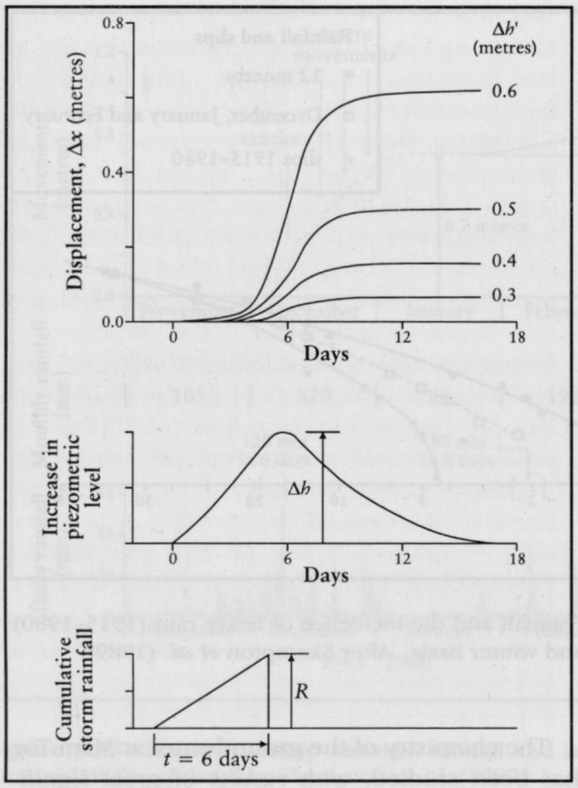


Figure 5.36 Displacements at Mam Tor caused by winter rainstorms. Note: movement begins at a piezometric level that is lower than the level at which movement ceases. This may imply that there has been a strength gain that may be of chemical origin in which movement releases cations. After Skempton *et al.* (1989).

years of debris-lobe activity. The importance of geochemical research here is in demonstrating that residual shear-strength is dynamic over the long term, and that there will be small seasonal variations in strength.

Geotechnical analyses

Skempton *et al.* (1989) carried out detailed geotechnical analyses to establish the index properties of the slide debris, its residual strength and stability, and the mechanics of storm-response movements, leading to a comprehensive back-analysis of the whole sequence of past and contemporary movement. Note that what follows is a much-simplified account, extracting key findings relevant to the Mam Tor failure; reference should be made to the original study and standard engineering geology texts.

The slide debris can generally be classified as clays of medium plasticity, although sand particle content and sandstone block incorporation can considerably affect its properties. The average water content is about 21% of the dry mass. It has a porosity of 36%.

The peak friction angle in intact rock, and its reduction after failure to a residual friction angle, is essential to understanding landslide activation. Here, the differences between the overlying Mam Tor Beds and the underlying Edale Shales are demonstrated to be very appreciable:

Mam Tor Beds	Peak strength	37	Residual strength	30
Edale Shales	Peak strength	30	Residual strength	14

The post-failure drop in strength therefore amounts to about 30% in sandstone and 60% in mudstone. At Mam Tor, where slip-surface testing was not feasible, estimates were obtained from the index properties. The value of 14°–15° deduced for the shear zone compares well with test results of slip-surfaces developed in compacted mudstone at other sites.

Stability analyses demonstrate that in high winter groundwater conditions, as studied in February 1978, the landslide is close to limiting equilibrium, with the factor of safety at 1.0, because it can be substantially re-activated by a transient rise in water level of 0.5 m. Moreover, because large displacements have occurred in the past, the strength along a slip-surface must be at the residual.

Skempton *et al.* (1989) considered four cases involving re-activation of different parts of the slide mass (Figure 5.37):

Case 1: Re-activation of the whole transitional and lower sectors of the landslide, below tension cracks at the upper road (between points j–e–g), and sliding on a slip-surface passing through slide debris and along the basal shear zone. The best result was obtained with residual friction angles (RFA) of 18° (in slide debris) and 14° (shear zone) when the calculated value of factor of safety (*F*) is 1.02.

Case 2: As a variant of case 1 the slip-surface is assumed to thrust upward through the slide debris at f–k. This simulates a 'confined' slip

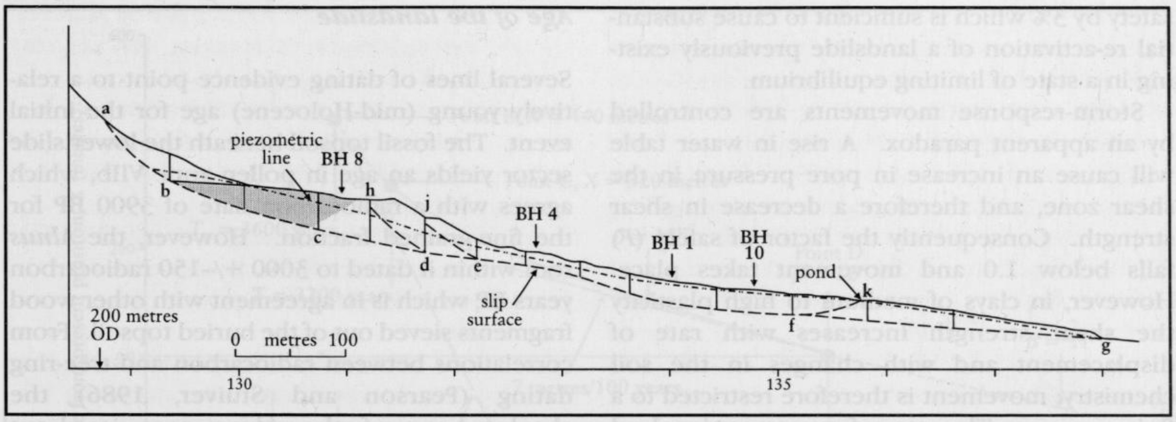


Figure 5.37 The stability analysis used by Skempton *et al.* (1989) at Mam Tor. Case 1: j–e–g; Case 2: j–e–f–k; Case 3: h–d; Case 4: b–c. The uppermost part of the failed mass is shaded.

not extending to the toe. With the same RFAs, $F = 1.00$, confirming that these two modes of failure have very similar probabilities.

Case 3: In the slip of February 1977 cracks additional to those at the upper road were seen above it, as at point h. On the slip-surface h–d a higher RFA value may be taken because the slide debris here probably includes a considerable proportion of sandstone, as observed in nearby Borehole 6. A reasonable assumption is that about one-third of the debris is sandstone, with RFA of 30°, and the rest is clayey debris with RFA of 18°. The average residual angle on h–d is then 22°, and, still using an RFA of 14° in the basal shear zone, $F = 0.99$. Had it been assumed, as perhaps an upper limit, that half of the debris consisted of sandstone, the average RFA value on h–d would be 24° and the corresponding factor of safety would increase by 1.7% to $F = 1.02$ (i.e. the higher the sandstone content, the less prone to re-activation).

Case 4: Exceptionally heavy rain in December 1965 led to re-activation of practically the entire landslide up to and including the talus. The talus is granular material with little if any clay fraction, in which the RFA can be taken as 30°. The uppermost part of the failed mass (toned in Figure 5.37) will contain a rather high proportion of sand, derived from the overlying predominantly sandstone debris in this upper part of the upper sector. The RFA in sandy clays is very dependent on relatively small changes in plasticity index and clay fraction. The procedure in this case was therefore to determine, by back-analysis, an RFA value in the shear zone b–c. This gave a factor of safety not less than 1.0, with an RFA of 14° in the slip below point c. The result is an RFA value lying between 23°, which gives $F = 1.0$ exactly, and 24° which gives $F = 1.02$. An RFA of 24° corresponds to a clay fraction of approximately 20%. This is an acceptable result, as compared to 35% in the shear zone further down the landslide where the debris consists chiefly or entirely of degraded mudstone and clay matrix.

These four cases demonstrate that no difficulty exists in showing that the landslide as a whole, and various parts of it, are delicately balanced in a state close to limiting equilibrium with groundwater level at about the normal winter maximum.

Skempton *et al.* (1989) conclude their stability analyses by examining the effects of variations in the parameters and assumptions used, including lateral confining pressures and their interaction with internal forces. A change of RFA in the basal shear zone of 1° leads to an increase or decrease in F by 5–7%. Changes of this magnitude they describe as not admissible, indicating that 14° is the correct value for the RFA within narrow limits. Finally, a change in groundwater level of 0.5 m results in a change in factor of safety of 2–3%. Although this only affects the RFA in the basal shear zone by ± 0.4°, such a rise in groundwater level would reduce the factor of safety by 3% which is sufficient to cause substantial re-activation of a landslide previously existing in a state of limiting equilibrium.

Storm-response movements are controlled by an apparent paradox. A rise in water table will cause an increase in pore pressure in the shear zone, and therefore a decrease in shear strength. Consequently the factor of safety (F) falls below 1.0 and movement takes place. However, in clays of medium to high plasticity the shear strength increases with rate of displacement and with changes in the soil chemistry; movement is therefore restricted to a finite amount. The rates of movement involved are sufficiently low (about 100 mm per day) for inertial forces to be negligible. Thus although a rise in water table causes an initial drop in factor of safety, the consequent remobilization leads to an increase in strength and cessation of movement. This is why displacements are chiefly concentrated within a few days of peak rainfall events, and why they become self-limiting regardless of continued high water-table conditions (in other words, movement does not continue indefinitely, or accelerate into a mudflow).

The total advance of the slide mass at Mam Tor by these episodic storm-response movements is currently less than 10 m in a century. This has produced a very small change in overall geometry of the slide mass, and therefore a correspondingly small change in static factor of safety under normal winter groundwater conditions. Consequently the process can be repeated many times without any considerable change in parameters. Nevertheless, on a long timescale the cumulative effect must be to bring the slide mass into a more stable configuration: F becomes marginally greater and a larger water-table rise is required to produce a given displacement. The return period for re-activation is therefore longer and the movement per century is smaller. Eventually, a state will be reached in which F is sufficiently high for the landslide to remain stable under the heaviest winter rainstorms, and this may be defined as the condition of permanent equilibrium, under present climatic and geomorphological conditions.

Interpretation

Although the nature and behaviour of the Mam Tor slide can be described in unusual detail, there still remain matters of interpretation.

Age of the landslide

Several lines of dating evidence point to a relatively young (mid-Holocene) age for the initial event. The fossil topsoil beneath the lower slide sector yields an age in pollen zone VIIb, which agrees with a radiocarbon date of 3900 BP for the fine-grained fraction. However, the *Alnus* root within it dated to 3000 +/–150 radiocarbon years BP, which is in agreement with other wood fragments sieved out of the buried topsoil. From correlations between radiocarbon and tree-ring dating (Pearson and Stuiver, 1986) the absolute age of the *Alnus* root is about 3200 ± 200 calendar years BP.

Skempton *et al.* (1989) estimated the date of inception of the landslide by reasoning as follows (Figure 5.38):

If at a time T in the past the toe of the lower sector was at a distance X from its present position, any curve relating X and T would have to satisfy these conditions:

(a) At the toe, $X = 0$, $T = 0$ and dX/dT is the present rate of movement (7 m per century).

(b) At the borehole containing the dated Alder root, $X = 320$ m and $T = 3200$ years.

By extrapolation to point B, where $X = 440$ m, the time T_1 to the initiation of lower sector movement can be found. For the time-displacement curve shown in Figure 5.38, $T_1 = 3600$ years, and dX/dT at that time is 0.5 m per annum. The initial slip, by comparison, would have been a sudden event. Therefore T_1 is the estimated date of origin of the Mam Tor landslide. This is about 3600 ± 400 calendar years BP, which agrees with the upper limit of the radiocarbon dates described above.

However the radiocarbon dates relate to material beneath the lower slide toe, and Figure 5.38 relates to re-activation of an initial slide which may previously have come to rest. Climatic changes to wetter conditions after 4000 BP could have contributed to re-activation, as could consolidation of the debris to elevating water tables. The possibility of an earlier initial event cannot readily be ruled out.

The archaeological evidence is also equivocal. The rampart is generally agreed to date from the Iron Age (flourished *c.* 2500 years BP), although remains of Bronze Age (*c.* 4000 years BP)

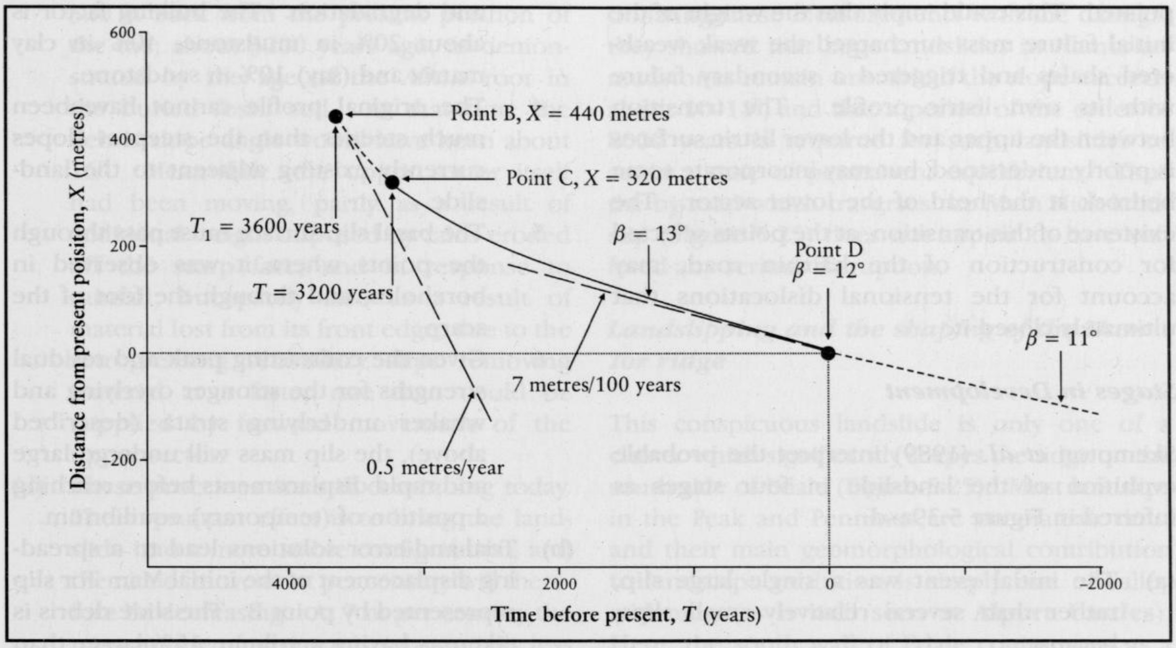


Figure 5.38 Method of determining age of Mam Tor landslide by projecting back from current configuration and rate of movement (points B–D as Figure 5.39). $T = 0$ corresponds to 1950 AD After Skempton *et al*. (1989).

dwellings have been found within it. Both the Mam Tor landslide and the Mam Nick landslide on the opposite side interrupt the rampart (Figure 5.25). Archaeological opinion is that the rampart was originally continuous, although both scars are so steep as to make construction of a rampart superfluous. Although the rampart appears to have been breached by a subsequent landslide, this could simply be the product of the later attrition which has created the large talus bank. The Mam Nick scar is however grassy, and a sample at the toe gives an age of 5900 radiocarbon years BP, suggesting that its flank scarp might have been incorporated in the defences.

Failure character and geometry

Mam Tor is described by Skempton *et al*. (1989) as a 'massive example of a slump-earthflow'. It is certainly a 'composite landslip' (WP/WLI, 1993). Following Hutchinson (1988; see Chapter 1), it can be classified as H4 (Landslides breaking down into mudslides or flows at the toe) and D3 (Compound failures – markedly non-circular, with listric or bi-planar slip-surfaces).

Although borehole evidence indicates relatively limited initial displacement, this may relate to particular back-tilted masses originating near the base of the scar cavity. It is possible that material released near the rim travelled further, over-riding the basal material to form the forward part of the initial toe. Further investigation of the rapidity of the initial movement, its degree of disintegration, and its trajectory is merited; Mam Tor has some of the topographical characteristics of **Beinn Alligin** (see GCR site report, Chapter 2), also in near-horizontal sandstones, if in miniature. The role of faulting in facilitating failure here might also be examined, along with proximity to the formerly mined zone of mineralization in the adjacent limestone.

The actual failure surface has been described as a 'concave-upwards curved slip-surface' evidenced by the back-tilted strata (Skempton *et al*., 1989). Figure 5.31 clarifies that the upper sector is a listric failure, i.e. a spoon-shaped surface that is here essentially bi-planar (scar plane and basal plane) linked by a curve. In detail, this curve may take place within a shear zone rather than as a discrete smooth concavity, given that in this sector it is in unweathered bedrock. From borehole evidence a marked convexo-concave failure surface step-down in the transition zone to the lower sector is interpolated. This could imply that the weight of the initial failure mass surcharged the weak weathered shales and triggered a secondary failure with its own listric profile. The transition between the upper and the lower listric surfaces is poorly understood, but may incorporate some bedrock at the head of the lower sector. The existence of this transition, at the points selected for construction of the hairpin road, may account for the tensional dislocations that ultimately closed it.

Stages in Development

Skempton *et al*. (1989) interpret the probable evolution of the landslide in four stages as inferred in Figure 5.39a–d:

(a) The initial event was a single large slip, rather than several relatively small slips, because:

1. About 520 m from the present toe, in the transitional zone, the slip-surface is at a very shallow depth below the original ground level. Almost all of the slide debris east of this point, including the lower sector, must therefore have derived from material to the west.
2. For the same reason the initial slip or slips must have been to the west of this point, i.e. in the upper sector.
3. The volume of the initial slip or slips must equal that of the slide debris, after allowing for expansion due to softening and degradation. The bulking factor is about 20% in mudstone, 40% in clay matrix and (say) 10% in sandstone.
4. The original profile cannot have been much steeper than the steepest slopes currently existing adjacent to the landslide.
5. The basal slip-surface must pass through the points where it was observed in boreholes, and through the foot of the scarp.
6. Given the contrasting peak and residual strengths for the stronger overlying and weaker underlying strata (described above), the slip mass will undergo large and rapid displacements before reaching a position of (temporary) equilibrium.

(b) Trial-and-error solutions lead to a spreading displacement of the initial Mam Tor slip represented by point B. The slide debris is taken as having a volume 15% larger than that of the initial slip, to allow for bulking without a substantial increase in water content.

(c) As a result of degradation and softening, secondary slips occur in the lower part of the mass of the upper sector, leading to the development of a lower sector. Comparative studies (see below) indicate that the rate of advance of the lower sector would initially have been far greater, by roughly one order of magnitude, than the average figure in recent times. The advancing lower sector reached point C,

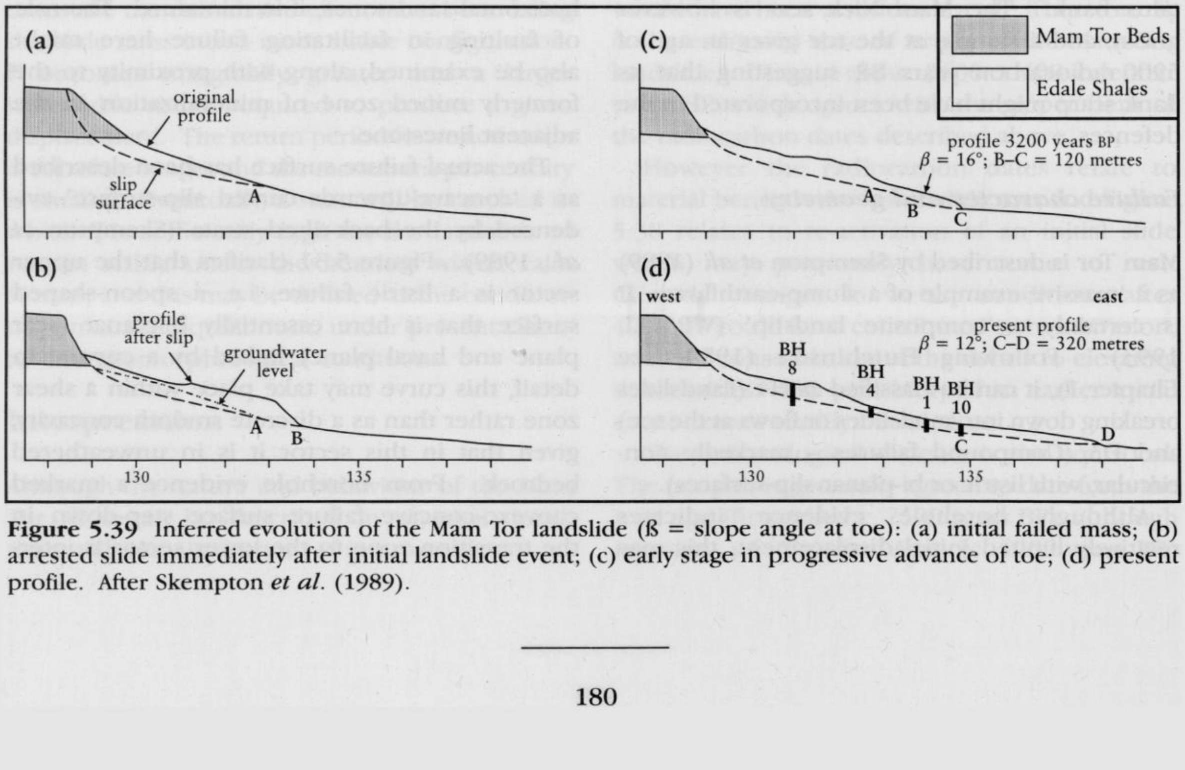


Figure 5.39 Inferred evolution of the Mam Tor landslide (ß = slope angle at toe): (a) initial failed mass; (b) arrested slide immediately after initial landslide event; (c) early stage in progressive advance of toe; (d) present profile. After Skempton *et al.* (1989).

320 m back from the present position of the toe, about 3200 years ago, as demonstrated by the age of the *Alnus* root in the buried fossil soil. At that time the debris slope angle would have been about 16°. Meanwhile, the upper sector itself had been moving, partly as a result of additional weight imposed by talus eroded off the scarp face, and in response to rainfall, but partly also as a result of material lost from its front edge, due to the retrogressing secondary slips removing material at a faster rate than could be supplied by forward movement of the upper sector.

(d) These processes are still continuing today. Their resultant effect is to bring the landslide into a more stable configuration, and the rate of movement per century will therefore be decreasing. At 7 m per century the present rate of advance is substantially less than the average of 10 m per century for the past three millenia. However it is clear that a state of permanent equilibrium has not yet been reached, and the present shape of the landslide Figure 5.39d, with a debris slope angle of 12°, is simply the latest stage in a development that will continue for a very long time.

Comparisons

The southern Pennines contain many other large landslides in Namurian strata. These have been described by Johnson and co-workers (Johnson 1965, Franks and Johnson 1964, Johnson and Walthall 1979, Tallis and Johnson 1980; cf. the **Alport Castles** and **Canyard Hills** GCR sites). Pollen analysis and radiocarbon dating show that some of these are older than Mam Tor, and, unlike Mam Tor, apparently stable (as at Alport Castles). Skempton *et al.* (1989) made map measurements of the of the debris slope angle at four other landslips to compare with Mam Tor:

Coombes Tor	9.5°	Rough Rock cap over shales
Millstone Rocks	12.5°	Kinderscout grit over Grindslow Shale
Didsburk Intake	11°	Kinderscout grit over Grindslow Shale
Mam Nick	10°	Mam Tor Beds over Edale Shales
Mam Tor	12°	Mam Tor Beds over Edale Shales

Collating these measurements with the datings, they showed that large landslides in Namurian mudstones remain unstable if the slope exceeds about 10–11°, and that a period of the order of 8000 years is required for such landslides to attain a state of permanent equilibrium. Thus the by-road which traverses the Mam Nick landslip (Figure 5.25) does not appear to have suffered any serious disruption.

Landslipping and the shaping of the Mam Tor ridge

This conspicuous landslide is only one of a cluster which significantly shapes the ridge on the south side of Edale (Figure 5.25). Most landslips in the Peak and Pennines are on plateau rims, and their main geomorphological contribution to landscape evolution is simply one of valley widening (as is well seen at **Alport Castles**). Here, the south wall of Edale commences as a plateau, but from the summit of Rushup Edge eastwards for 4 km to its terminus at Lose Hill it has a well-defined crest. Only the summit of Mam Tor itself broadens out, as a residual of the former plateau ridge.

The Mam Tor slide is the only one on the southern aspect of this ridge. On the north side, three major extant slips occur, although the terrain suggests that earlier events have embayed the ridge, the failed material having been evacuated by subsequent valley glaciers:

(a) Mam Nick (0.60 km^2): this landslide has twice the extent of the Mam Tor event, and narrows the crest of Rushup Edge to a half-arête for 500 m (Figure 5.40). The source configuration is an obtuse wedge. In its south-east corner, it breaks through the ridge to create the 'Nick' followed by the minor road over to Edale. Here it has lowered the ridge by about 40 m, and truncates the west flank of Mam Tor, including its hillfort rampart (Figure 5.41). The main headscarp is of steep grass approximately 25 m high, with a clutch of short-travel sharp-crested slip masses having bold anti-scarps 3–5 m high. Beneath these, the apron of the main failed mass is crossed by the road, below which an amorphous 'earth-flow' extends at a lesser gradient for 500 m down to an 8–15 m-high toe bank above Greenhill Farm (Figure 5.40). This two-tier configuration is very similar to Mam Tor.

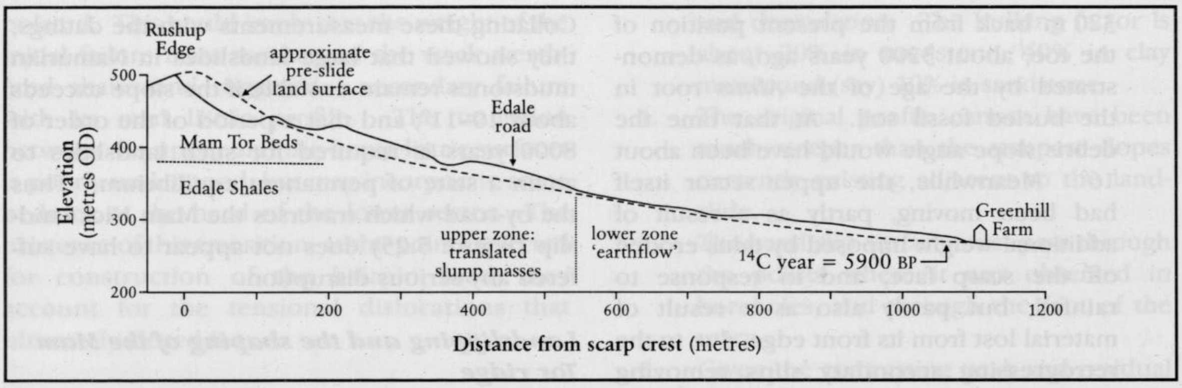


Figure 5.40 Landslide profiles at Mam Nick, where ß = 10°. After Skempton *et al.* (1989).

Figure 5.41 Mam Nick landslide source cavity and upper slump zone, from Rushup Edge looking east. Note the far flank scar interrupting Mam Tor hillfort rampart, and the headscarp narrowing the former plateau ridge to a crest. Edale road passes through Mam Nick and descends across the slump, which has grassy slip-masses presenting uphill-facing scarps. (Photo: D. Jarman.)

(b) Cold Side (0.25 km^2): the grassy source scars are exposed just below the crest north-east of Mam Tor (cf. **Alport Castles**). They are up to 32 m in height, in a double-wedge obtuse splay. The main slip mass has a striking antiscarp 8 m high impounding a pond, while the toe is a steep rampart 10–15 m high.

(c) Back Tor (0.50 km^2): this dramatic landslide bites right through the ridge west of Lose Hill for 200 m, lowering it by up to approximately 50 m (Figure 5.42). The main failed mass has slumped almost to the floor of Edale, possibly deflecting the river slightly. A more recent increment on the east side yields impenetrable antiscarped terrain colonized by Backtor Wood, below a 60 m sandstone crag comparable in scale with that on Mam Tor.

Figure 5.42 Back Tor landside from the east. The 60 m main crag is a source scar comparable to Mam Tor. The intervening ridge has been lowered by some 50 m by virtue of the slide surface here exposed behind the crest. (Photo: D. Jarman.)

Whereas these three slips are metastable, and the Mam Tor slide is evolving towards that condition, future re-activation in response to fluvial or glacial valley incision will tend to see these and other slope failures coalesce, first eliminating Mam Tor as a separate hill, and then reducing the whole ridge to a rump.

Conclusions

The landslide at Mam Tor is not unusually large or articulated, by comparison with (for example) the **Alport Castles** GCR site. In many ways, it represents the typical Pennine slump-flow where competent rocks overlie weaker strata that are less permeable and prone to deformation. However, Mam Tor is important for the high degree of knowledge of its recent evolution, including quantitative assessment of the relationships between movements, ground-water levels, storm response, and properties of the clay/mudstone (notably its geochemistry) of which the slide is largely composed. This is because it was crossed by a trunk road that eventually had to be closed because of continuing and irremediable slippage – the only such case affecting a major transport artery in inland Britain.

Mam Tor is unusual in still being an active landslide, with spasmodic advances associated with peak rainfall and raised water-table conditions. The current recession rate is 7 m per century, which, while gradually diminishing, has no foreseeable end-date. The nexus of factors that sustain instability are finely balanced, and this site is of great comparative importance. Mam Tor is also a conspicuous and readily accessible site, with a bold scar, and the 'shivering mountain' has long been known as a 'Wonder of the Peak'. It is of considerable interest to archaeologists, since this and the Mam Nick slide interrupt an Iron Age hillfort rampart. It is also important in studies of land-scape evolution, since this and several even larger sites have made substantial inroads into the ridge separating Edale from the Hope Valley. It is thought to be a relatively young feature, with an inferred mid-Holocene date of initial failure at around 3600 BP, although an older date remains possible.

ROWLEE BRIDGE, ASHOP VALLEY, DERBYSHIRE (SK 150 894)

R.G. Cooper

Introduction

On the right (southern) bank of the River Ashop, 130 m upstream from Rowlee Bridge (Figures 5.13 and 5.43) and 2 km before the east-flowing river enters the Ladybower Reservoir in Derbyshire, there is an exposure, 3 m high, of the highest Edale Shales (Alportian (H_2)) with the lowest Mam Tor Beds (Kinderscoutian (R_1)) of Namurian, (Upper Carboniferous) age.

Description

The exposure shows sharp, symmetrical, straight-limbed folds. Similar but less well-exposed structures exist in the Edale Shales and Mam Tor Beds in Edale, the Derwent valley and its tributary valleys including Abbey Brook and Ouzelden Clough, and in the Hope valley. Trenches excavated for the foundations of the Howden and Derwent dams exposed a single large fold in each case, decreasing in magnitude with depth (Lapworth, 1911; Sandeman, 1918; Fearnsides *et al.*, 1932) (Figure 5.44). Photographs of the trenches excavated during Sandeman's (pre-1910) construction of these dams and exhibited by Thompson in 1949 show that at the Howden Dam this fold is a 'crumple' in the form of a 'double V' fault in the strata. This extends to at least 15 m below the ground surface, but reduces with depth. At the Derwent Dam there is a similar crumple, but the crumpling is less complex: it takes the form of a simple 'V'; and at the bottom of the trench (about 15 m deep) the movement had largely died out (Thompson, 1949).

A more detailed investigation was undertaken on construction of the Ladybower Dam in the 1930s and 1940s, with a large number of vertical borings across the valley along the line of the dam (Hill, 1949). These revealed the presence of an enormous crumple extending down from the valley bottom for at least 58 m vertically (Figure 5.44). Thompson (1949) commented that the crumples occur throughout the Derwent and Ashop valleys.

Figure 5.43 Valley-bulge structures exposed in the Rowlee Bridge section, in the Ashop Valley, Derbyshire. The sharp, symmetrical folds are one of the most remarkable examples of compressional folding ever recorded. The folds are due to the extrusion of clays and ductile flow in bedded strata, often called valley-bulging. (Photo: R.G. Cooper.)

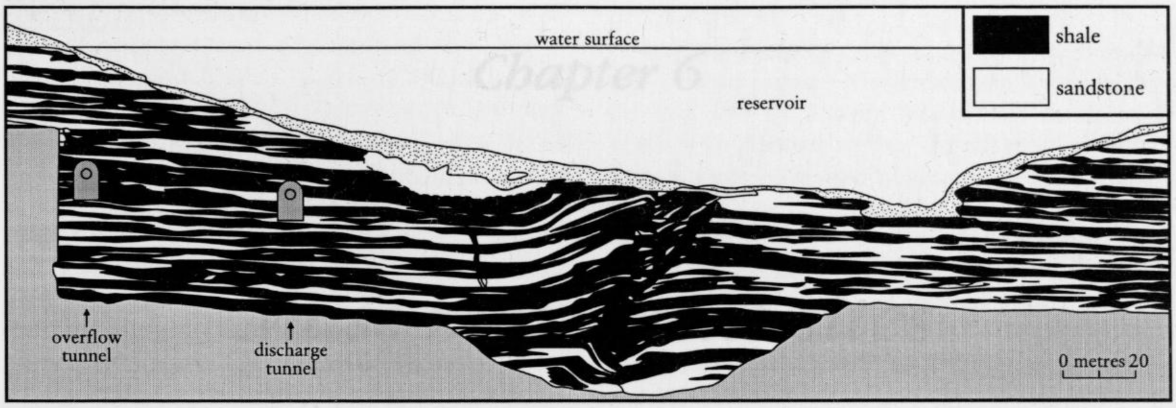


Figure 5.44 Section through valley-floor crumple (according to Thompson (1949)), at Ladybower Reservoir. After Hill (1949).

Interpretation

It may be inferred, therefore, that a similar crumple lies beneath the Ashop valley at Rowlee Bridge, in which case the folding observed at the site represents accommodation of the surface strata to a larger event that took place at greater depth.

Thompson's (1949) explanation for the crumples was that since the strata in the hillsides and at the bottom of the valleys consists of a succession of beds of shale and sandstone, the weight of the hillsides had compressed the shale and had caused it to flow towards the valleys, where the stress had been relieved by crumpling. In the case of the Ladybower Dam site the process had gone a little further, and the fold had become an overthrust fault. The reduction of the structure with depth distinguishes it from an ordinary geological fault. Stevenson and Gaunt (1971) came to broadly similar conclusions.

These structures have many similarities to those described in Northamptonshire by Hollingworth *et al.* (1944) and by Hollingworth and Taylor (1951); they are believed to have resulted from the pressure exerted on predominantly argillaceous strata by superimposed beds inducing lateral movements or 'squeezing out' of the argillaceous strata into the adjacent load-free valley areas, producing compression folds and thrusts (Stevenson and Gaunt, 1971). Movements caused in this way would be possible wherever valley-deepening exposed the top of any substantial thickness of shale, but would be greatly facilitated by thawing of ground-ice following a glacial or periglacial phase.

Conclusions

The site is at some risk from normal fluvial erosion processes. The River Ashop runs at the foot of the exposure itself, and clearly it is eroding the exposure. This would not matter but for the position of the exposure on a short promontory 3 m high, standing in the floodplain of the river. The promontory is a salient from what appears to be a river terrace, but it is only 5 m across, and when the river has eroded 5 m into it, the promontory and the site will have ceased to exist. The River Ashop does not, however, experience continuous natural flow levels at this point; about 200 m upvalley a variable proportion of its flow is from time-to-time taken into a concrete channel and thence by tunnel to the Rivelin valley as part of the water supply for Sheffield (Wood, 1949). This may be responsible for the site's survival thus far.

In the absence of a clear exposure associating valley-bulging with cambers and gulls in the Northampton ironstone field (the good ones have all been backfilled), the Rowlee Bridge exposure provides the clearest example of valley-bulge structures currently accessible in Great Britain.