Mass Movements in Great Britain

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Chapter 4

Mass-movement sites in Devonian strata

INTRODUCTION

R.G. Cooper

In Great Britain 1042 landslides are reported in Devonian strata by Jones and Lee (1994), of which almost three-quarters (75%) are of unspecified type. However, of those for which the type has been identified, the largest group is rockfalls, at 41%, followed by translational slides and debris flows at 31%. Two GCR sites have been selected in Devonian strata, **Coire Gabhail** in Scotland, and **Llyn-y-Fan Fâch** in Wales (see Figure 4.1).

Rockfalls are a commonly recognized form of mass movement, but there is usually little evidence of the processes that have brought them about. There are several possibilities, including:

(1) Frost-wedging in rock fissures: could cause a detached block to stand proud of a cliff-face to such an extent that its centre of gravity causes sliding under its own weight and eventual falling as the block becomes free of the restraint of the adjacent rock body. Dislodgement of one fragment could provide conditions for fragments above it, and possibly the whole cliff, to collapse.

(2) Weathering: could cause a decrease in cohesive strength so that a cliff-face could disintegrate into a large number of fragments.

(3) Undercutting of the cliff-face: could remove basal and lateral support.

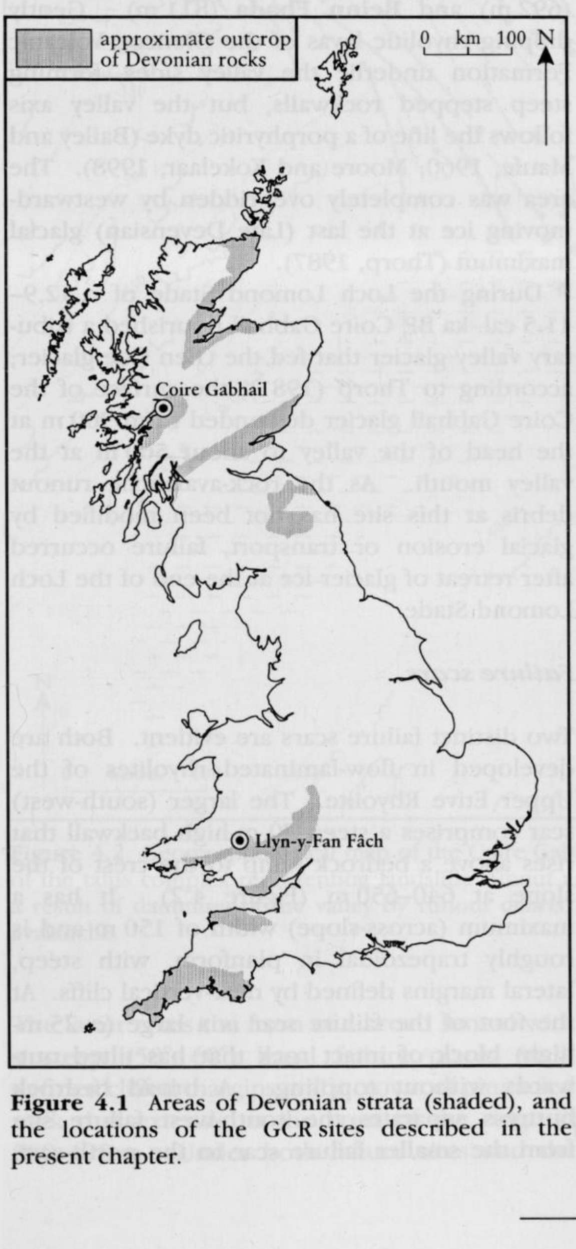

Figure 4.1 Areas of Devonian strata (shaded), and the locations of the GCR sites described in the present chapter.

Either way, the resulting form consists of a pile of fragments described as *talus*.

Where rockfall occurs on a rectilinear slope, a *talus sheet* is produced at its foot. However, where rockfall repeatedly takes place into a rock gully, the gully will constrain the path of the fallen debris, giving rise at the gully's mouth to a *talus cone*. Laterally adjacent cones may coalesce to form a sheet.

Ballantyne and Harris (1994) define a *talus slope* as a steep valley-side slope formed by the accumulation of debris at the foot of a rockwall. The term 'talus' is used to denote both the slope and its constituent material. *Scree* is taken to refer to any slope that is covered in coarse debris. Therefore 'scree' includes 'talus', which is more closely defined.

Rockfall talus characteristically may display fall-sorting, a process whereby larger fragments come to rest farther downslope than smaller fragments (Ballantyne and Harris, 1994). Grain size is commonly bi-modal, which can result in talus sliding, and sorting into garlands and stripes. This in turn imparts another characteristic to talus slopes: in profile, a talus slope lessens in steepness at its foot. This is because at the foot the talus slope consists of the largest fragments, which protrude most from whatever lies beneath the surface (usually talus from previous falls), although it can also be because the debris runs out and rests on the valley floor. This characteristic is absent from the famous talus of Wasdale, in the English Lake District, because the lower parts of these are underwater in the lake that occupies much of the glacially over-deepened valley.

Sliding may take place on a talus slope – dry and granular sliding by debris slide, or avalanche gullying. A widespread process by which debris is re-distributed on a talus slope is *debris flow*. The term is primarily used to refer to rapid downslope flow of poorly sorted debris mixed with water, but it is also used to refer to the suite of landforms produced by an individual flow. The overall effect of debris-flow activity on talus is that material is eroded from the upper part of the talus slope and deposited near or beyond the talus foot. In this way the overall gradient of the talus slope is reduced, and a long, sweeping basal concavity is produced. Debris flows disrupt the fall-sorting of undisturbed rockfall talus. Repeated flows on the same track produce a gully in the upper part of the talus slope, which is continued downslope by two parallel debris levées that mark the path of the flow and terminate downslope in one or more lobes of bouldery debris.

The GCR sites (Figure 4.1) chosen to represent some aspects of these features include the site of a very large post-glacial rockfall, **Coire Gabhail** in Glencoe, Scotland, for which the evidence is a talus of great coarseness, and **Llyn-y-Fan Fâch** in Carmarthenshire, Wales, where there are debris flows that have been investigated as a potential exemplar of an alternative explanation of the steepness of talus slopes (Statham, 1976).

COIRE GABHAIL, HIGHLAND (NN 166 557)

C.K. Ballantyne

Introduction

Of over 500 rock slope failures identified in the Scottish Highlands (Ballantyne, 1986a), relatively few involve total disintegration of the collapsed rock-mass. A double rock-avalanche site in lower Coire Gabhail, Glencoe (Figure 4.2), is of outstanding interest in several respects. First, the initial failed rock-mass completely disintegrated and accumulated as a massive talus cone resting on the valley floor. Second, a smaller talus cone partially overlaps the main cone, suggesting that a later rock-avalanche occurred after the initial failure. Third, the Coire Gabhail rock-avalanche site represents the largest failure on the Devonian volcanic rocks of the Western Highlands. Finally, the rock-avalanche debris has completely blocked the valley, forming a natural sediment trap and causing the accumulation of coarse alluvial deposits upstream (Werritty, 1997).

The larger Coire Gabhail rock-avalanche is also one of only a handful of Scottish rock slope failures to have been dated using cosmogenic isotope dating techniques.

Description

Setting

Coire Gabhail, popularly known as the 'Lost Valley', is a hanging valley on the south side of Glen Coe (Figure 4.2). The valley mouth is flanked by two truncated spurs, Geàrr Aonach (692 m) and **Beinn Fhada** (811 m). Gently dipping rhyolitic lavas of the Glencoe Volcanic Formation underlie the valley sides, forming steep stepped rockwalls, but the valley axis follows the line of a porphyritic dyke (Bailey and Maufe, 1960; Moore and Kokelaar, 1998). The area was completely over-ridden by westward-moving ice at the last (Late Devensian) glacial maximum (Thorp, 1987).

During the Loch Lomond Stade of *c.* 12.9–11.5 cal. ka BP, Coire Gabhail nourished a tributary valley glacier that fed the Glen Coe glacier; according to Thorp (1981), the surface of the Coire Gabhail glacier descended from 900 m at the head of the valley to about 560 m at the valley mouth. As the rock-avalanche runout debris at this site has not been modified by glacial erosion or transport, failure occurred after retreat of glacier ice at the end of the Loch Lomond Stade.

Failure scars

Two distinct failure scars are evident. Both are developed in flow-laminated rhyolites of the Upper Etive Rhyolite. The larger (south-west) scar comprises a steep 70 m-high backwall that rises above a bedrock ramp to the crest of the slope at 640–650 m (Figure 4.2). It has a maximum (across-slope) width of 150 m and is roughly trapezoidal in planform, with steep, lateral margins defined by near-vertical cliffs. At the foot of the failure scar is a large (*c.* 25 m-high) block of intact rock that has tilted outwards without toppling. A broad bedrock buttress separates the south-west failure site from the smaller failure scar to the north-east.

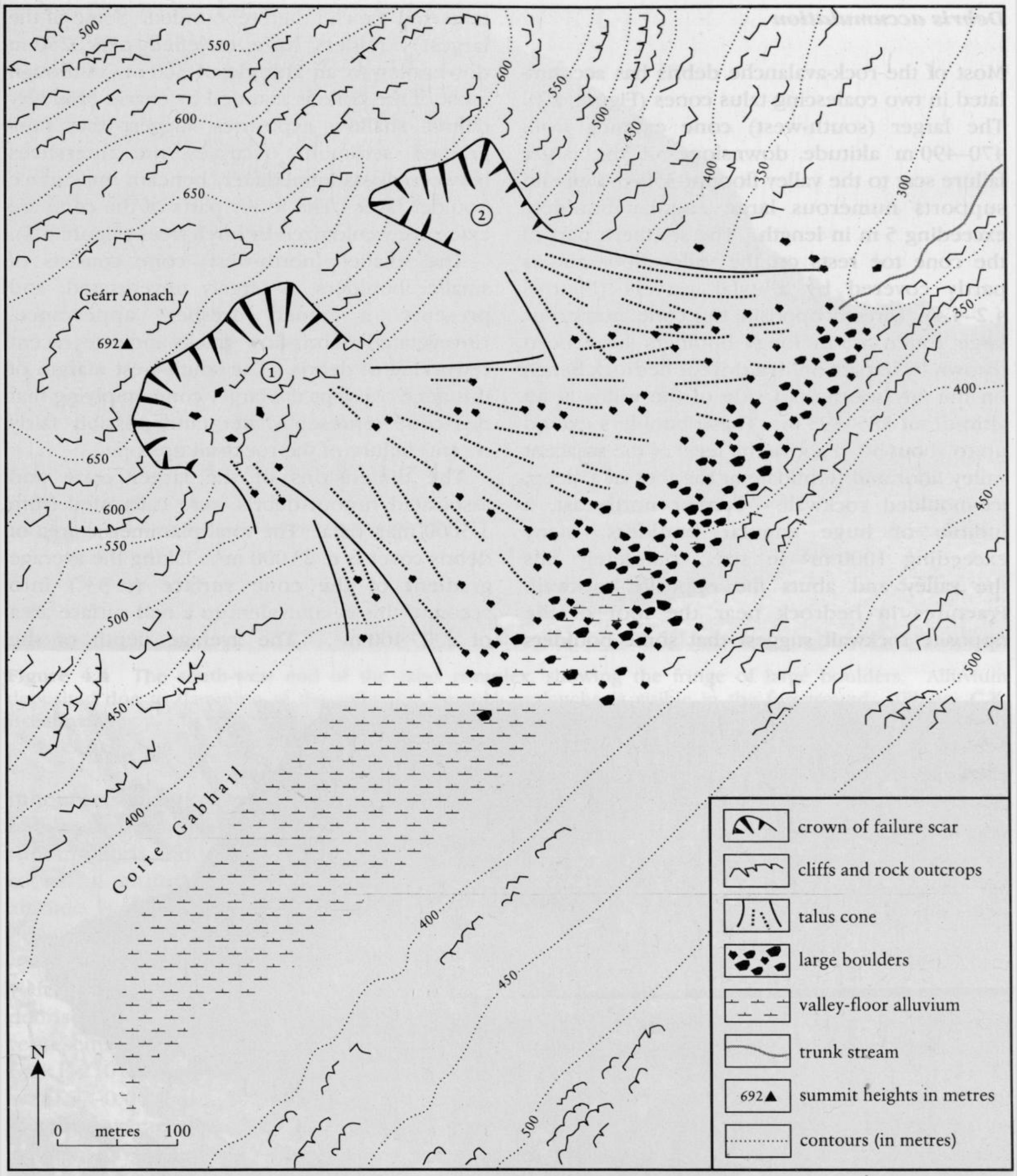

Figure 4.2 Geomorphological map of the Coire Gabhail rock-avalanches, showing the failure sites, the extent of the talus complex representing landslide runout and the area of alluvium deposited in the upper valley as a result of damming of the valley by runout debris. (1) site of initial rock-avalanche; (2) site of later rock-avalanche.

The latter takes the form of a broad funnel with a steep (50°–60°) basal failure plane, well-defined cliffed margins and a crown of pinnacled rock that extends to the crest of the ridge at 580–620 m. Under both failure sites and the adjacent rockwalls, stacked rhyolitic lava flows dip gently westwards, implying that failure was not seated on flow boundaries. The valley-side face is, however, seamed with vertical and near-vertical cooling and stress-release joints.

Debris accumulation

Most of the rock-avalanche debris has accumulated in two coalescing talus cones (Figure 4.2). The larger (south-west) cone extends from 470–490 m altitude downslope of the larger failure scar to the valley floor at 350–370 m and supports numerous large angular boulders exceeding 5 m in length. The southern part of the cone toe rests on the valley floor and is partly covered by alluvial gravels (Figures 4.2–4.4). Directly opposite this cone, numerous large (often > 5 m long) boulders have been thrown by impact on to a drift or bedrock bench on the far (south-east) side of the valley at an altitude of 385–395 m. These boulders extend up to about 30 m above the level of the adjacent valley floor and terminate at the foot of a steep, ice-moulded rockwall. Farther north-east, a jumble of huge angular boulders, many exceeding 1000 m^3 in size, completely fills the valley and abuts the opposite rockwall. Fractures in bedrock near the foot of the opposite rockwall suggest that some boulders impacted the cliff then rebounded. Some of the largest boulders have travelled over 250 m downvalley to an altitude of 300 m. Although most of the cone is mantled by coarse bouldery debris, shallow exposures suggest that fine-grained sediment occupies the interstices between clasts immediately beneath the surface boulder layer. The lower parts of the cone are extensively colonized by birch trees (Figure 4.3).

The smaller (north-west) cone consists of smaller boulders, is largely unvegetated, and presents a much fresher appearance. Unvegetated debris-flow tracks indicate recent reworking of debris. The south-west margin of this cone overlaps the larger cone, implying that this cone represents later (and possibly fairly recent) failure of the rockwall upslope.

The dimensions of the larger cone and associated runout debris were calculated from 1:5000 map data. The total planimetric area of debris cover is *c.* 23 000 m^2. Taking the average gradient of the cone surface (*c.* 33°) into account, this is equivalent to a real surface area of *c.* 27 400 m^2. The average depth of the

Figure 4.3 The complex talus accumulation formed by runout of the Coire Gabhail rock-avalanches, viewed from the south-west. The southern (left) edge of the larger talus cone is partly buried under alluvium. The conspicuous large boulder just beyond the toe of the cone rises about 10 m above the alluvium in which it is embedded. (Photo: C.K. Ballantyne.)

Figure 4.4 The south-west end of the talus complex, showing the fringe of large boulders. Alluvium deposited due to damming of the valley by the rock avalanche is visible in the foreground. (Photo: C.K. Ballantyne.)

runout accumulation, calculated by interpolating rockhead contours across the area occupied by runout debris and subtracting a grid of inferred rockhead altitudes from the debris surface altitude, is *c.* 11 m, with a maximum depth near the centre of debris accumulation of *c.* 30 m. Multiplying average depth by planimetric area yields a volume of approximately 300 000 m^3 of debris. Assuming that 20–30% of this volume represents voids, the implied volume of failed rock is 210 000–240 000 m^3, equivalent to a mass of 0.57–0.65 × 10^6 tonnes of rock for an assumed average density of 2.7 tonnes m^{-3}. The volume of the smaller cone could not be calculated, but appears to be about an order of magnitude smaller.

Interpretation

Mode of failure

The configuration of the failure scars suggests that at both failure sites thick slabs of rock failed along steep (50°–65°) basal shear planes bounded laterally by near-vertical rockwalls that represent the sites of stress-relief joints; open, near-vertical joints extend almost the full depth of the failure sites within adjacent intact rock-walls. Failure probably involved both sliding and toppling, leading to disintegration of the rock masses as they cascaded downslope. Travel of huge boulders across the valley floor and up to 250 m down the valley axis indicates that the earlier failure was characterized by extremely high energy; the large talus cone probably represents settlement of rock debris at the angle of residual shear in the final stages of movement. Only a very small component of the later failure reached the valley floor, however, with most debris accumulating in the smaller cone.

The immediate causes of the failures are unknown, but it is notable that the toes of both failure zones experienced debuttressing by retreat of glacial ice at the end of the Loch Lomond Stade, and the open near-vertical joints exposed in adjacent rockwalls imply rock-mass weakening as a result of joint extension due to deglacial stress-release. A zone of NE-trending vertical fractures followed by the Etive dyke swarm may have determined the loci of joint formation at the crown of both failures. The possible role of post-glacial seismic activity in triggering failure is difficult to assess. The larger failure scar lies roughly 1.4 km distant from two major NE-trending faults, the Ossian Fault to the north-west and Queen's Cairn Fault to the south-east, both of which originated during complex synvolcanic cauldron subsidence (Moore and Kokelaar, 1998). Neotectonic displacements due to differential glacio-isostatic adjustment along these or intervening fractures may have triggered failure, though there appears to be no evidence for post-volcanic dislocation.

Age of failure

The surface exposure age of a rock sample from the crest of a very large boulder deposited by the larger and earlier rock-avalanche has been subject to surface exposure dating by ^{36}Cl cosmogenic radionuclide assay. The provisional age obtained for this sample is 1.8 ± 0.33 ka BP, implying that failure occurred at least 9000 years after final deglaciation. The smaller and younger failure must have occurred after this date, and possibly within the last few centuries, consistent with its much fresher appearance. In common with landslide samples dated to *c.* 6.5 cal. ka BP for The Storr landslide (Ballantyne *et al.*, 1998b; see **Trotternish Escarpment** GCR site report, Chapter 6) and *c.* 4.0 cal. ka BP for the Beinn Alligin rock-avalanche (Ballantyne and Stone, 2004; see **Beinn Alligin** GCR site report, Chapter 2), this age determination implies that slope failure due to deglacial unloading and consequent stress-release has persisted well into Holocene times, and potentially may result in future failures from glacially steepened rock slopes in the Scottish Highlands.

Alluvial accumulation

According to Werritty (1997), the alluvial accumulation at Coire Gabhail is unique in Scotland, hence it is also selected as a GCR site for the Fluvial Geomorphology of Scotland GCR 'Block' (Werritty, 1997).

Following blockage of the lower valley by the earlier rock-avalanche, evacuation of coarse bedload sediment has been impeded, allowing progressive accumulation of alluvial gravels upstream. The alluvial basin is approximately 600 m long and 150 m wide, with a concave downvalley profile that is graded to the local base level created by sealing of the valley by landslide debris. The coarseness of surface gravels declines downvalley. The channel pattern is braided, but supports surface flow only following intense or long-duration rainstorms. Under normal flow conditions the river draining the valley (the Allt Coire Gabhail) sinks into the alluval deposits some distance upvalley from the rock-avalanche runout debris, and emerges near the north-east end of the boulder dam.

Conclusions

Roughly 1800 years ago, approximately 600 000 tonnes of rock failed near the mouth of Coire Gabhail, a steep-sided hanging valley cut in gently dipping rhyolites. The alignment of joints in the flank scarp of the failure zone suggests that failure was due to progressive joint extension and rock-mass weakening following debuttressing of the face as a result of glacier downwastage at the end of the Loch Lomond Stade, around 11 500 years ago. A smaller and apparently later failure occurred about 160 m north-east of the initial failure, depositing boulders as a talus cone on the flank of the debris deposited by the earlier event. This site is important for several reasons. It represents the finest example in Scotland of rock avalanches that have come to rest as massive talus cones, with debris resting at the angle of residual shear (*c.* 33°), though the high energy of the earlier failure drove boulders at least 30 m up the opposite slope and 250 m downvalley. It is also the largest rock slope failure on the Devonian volcanic rocks of the Western Highlands. This is one of the few ancient landslides in Scotland for which dating evidence is available, and the fact that failure occurred at least 9000 years after deglaciation implies that catastrophic failures due to paraglacial (glacially conditioned) stress-release persisted into late Holocene times. Finally, the site is unique in Scotland in that the rock-avalanche runout debris completely blocked the valley mouth, allowing sub-surface drainage through the runout zone but impounding coarse alluvial gravels. The alluvial floodplain that has developed upvalley as a result is without parallel in Scotland, with river runoff sinking into the alluvium near the valley head and emerging downvalley of the landslide runoff debris, except when exceptional flood events permit surface flow.

LLYN-Y-FAN FÂCH, CARMARTHENSHIRE (SN 801 215)

R.G. Cooper

Introduction

On the scarp face of the Black Mountain, facing northwards above Llyn-y-Fan Fâch, a number of deep gullies are cut through the vegetated scree surface (Figures 4.5 and 4.6). In some cases they reach the underlying bedrock for part of their length. Debris-flow activity is episodically taking place along the gullies' axes at the present time, transporting material derived from the gully sides by wash and other processes (Statham, 1976).

The Black Mountain is part of the north-facing Old Red Sandstone (Devonian) escarpment in south Wales. The scarp face above Llyn-y-Fan Fâch consists of a headwall of alternating hard sandstones and soft silty shales. The slope is about 60 m high, and stands at an overall angle of 45°–50°. Two prominent chutes/avalanche couloirs cut the upper slope of the headwall (Ellis-Gruffydd, 1972). Against the lower part of the headwall a scree-slope has accumulated, which is now well vegetated. The scree debris includes a significant proportion of coarse to fine sand-sized material, occupying the interstices between the boulders. Statham (1976) gives an average grading curve envelope for the scree material (Figure 4.7). Transport of debris down these gullies is resulting in the accumulation of debris-flow cones below the gully mouths. The cones are broadly concave

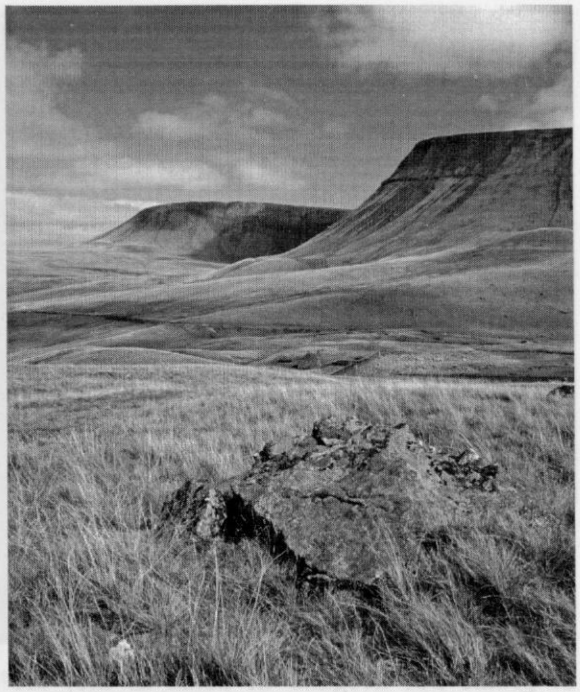

Figure 4.5 General view of the Llyn-y-Fan Fâch GCR site, showing the scarp face of the Black Mountain (Mynydd Du), and screes and gullies. (Photo: S. Campbell.)

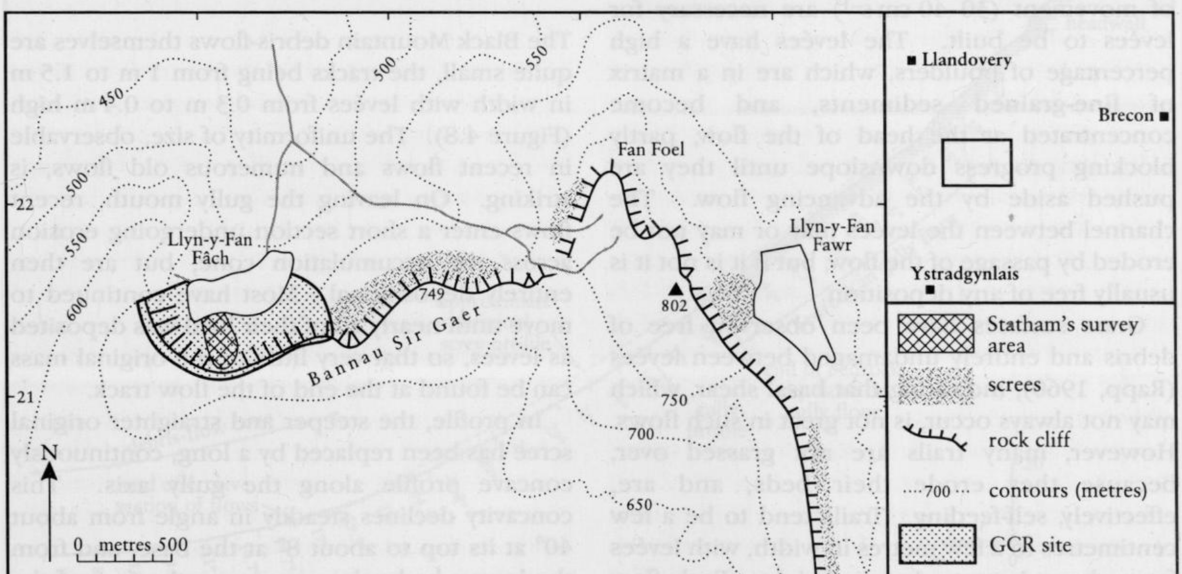

Figure 4.6 The location of the Llyn-y-Fan Fâch mass-movement site.

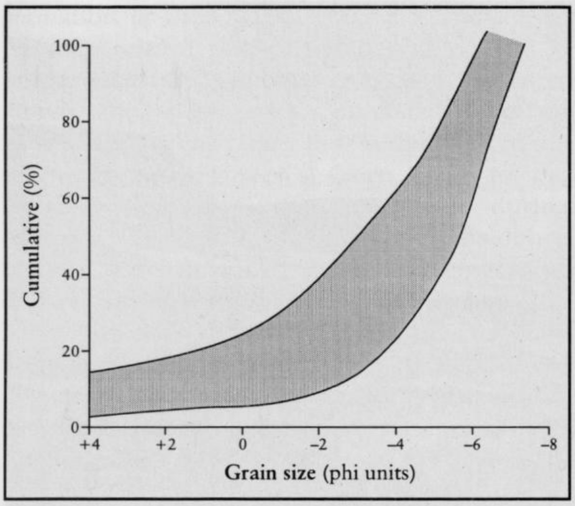

Figure 4.7 Range of sediment grain-size distribution of scree debris at Llyn-y-Fan Fâch. After Statham (1976).

in profile but irregular in detail as they are composed of a criss-cross accumulation of old debris-flow levées (Statham, 1976).

Debris flows are rapid mass movements of poorly sorted debris with a high water content. They are often associated with rockfalls from cliffs farther upslope, and tend to occur on talus that has accumulated from such rockfalls. Typically they follow a well-defined path as a fluid-like mass, leaving small linear ridges or levées on either side of their trail as they move. Sharp (1942) observed that quite rapid rates of movement (30–40 cm s^{-1}) are necessary for levées to be built. The levées have a high percentage of boulders, which are in a matrix of fine-grained sediments, and become concentrated at the head of the flow, partly blocking progress downslope until they are pushed aside by the advancing flow. The channel between the levées may or may not be eroded by passage of the flow, but if it is not it is usually free of any deposition.

Grass surfaces have been observed free of debris and entirely undamaged between levées (Rapp, 1960), indicating that basal shear, which may not always occur, is not great in such flows. However, many trails are not grassed over, because they erode their beds, and are, effectively, self-feeding. Trails tend to be a few centimetres to a few metres in width, with levées from about 1 cm to 1 m in height. Each flow generally transports anything from a few cm^3 to a few m^3 of material only. Many flows are single events, with no tendency for future flows to be concentrated along former lines, but more often flow activity is concentrated along a line to form a gully and a low-angled debris-flow cone of accumulation at its base (Johnson and Rahn, 1970). Debris flows almost always occur as a consequence of heavy rainfall or snowmelt. Mobilization may be due to the presence of a sub-surface, concentrated seepage line beneath the debris, which causes high porewater-pressure (Prior *et al.*, 1970). In the case of debris flows initiated in gullies, mobilization is a result of steady dilution of debris by water, rather than a steady increase in sediment content of a stream (Johnson and Rahn, 1970; see Iversen and Major, 1986; Addison, 1987). Movement probably begins as a slide but subsequent motion incorporates more water into the mass when it may behave as a fluid.

The Llyn-y-Fan Fâch GCR site was selected to represent well-documented debris-flows of a kind that is potentially atypical. Debris flows are a major and widespread type of mass movement in Great Britain, many occurring as relict periglacial forms (Ballantyne and Harris, 1994).

As well as being a mass-movement GCR site, the area was independently selected for the GCR for its Quaternary features of interest (Campbell and Bowen, 1989) and its fluvial geomorphology (Higgs, 1997).

Description

The Black Mountain debris-flows themselves are quite small, the tracks being from 1 m to 1.5 m in width with levées from 0.3 m to 0.4 m high (Figure 4.8). The uniformity of size, observable in recent flows and numerous old flows, is striking. On leaving the gully mouth, recent flows enter a short section undergoing erosion across the accumulation cone, but are then entirely depositional. Most have continued to move until nearly all of their load was deposited as levées, so that very little of the original mass can be found at the end of the flow track.

In profile, the steeper and straighter original scree has been replaced by a long, continuously concave profile along the gully axis. This concavity declines steadily in angle from about 40° at its top to about 8° at the base, and from the lowest bedrock exposure to the foot of the profile approximates well to a circular arc with radius of curvature about 310 m (Statham, 1976;

Figure 4.8 Detail of the gullying and mass-movement deposits at Llyn-y-Fan Fâch. (Photo: S. Campbell.)

Figure 4.9). There is no difference in curvature between the gully and the accumulation cone and so it seems reasonable to suggest that the entire form is continuous, controlled by the debris-flow process.

Debris flows are initiated in the gullies, at locations where the slope is between 27° and 37°. The gully sides attain a maximum stable angle at 43.5° when dry. Coarse debris with negligible clay- and silt-size percentages may

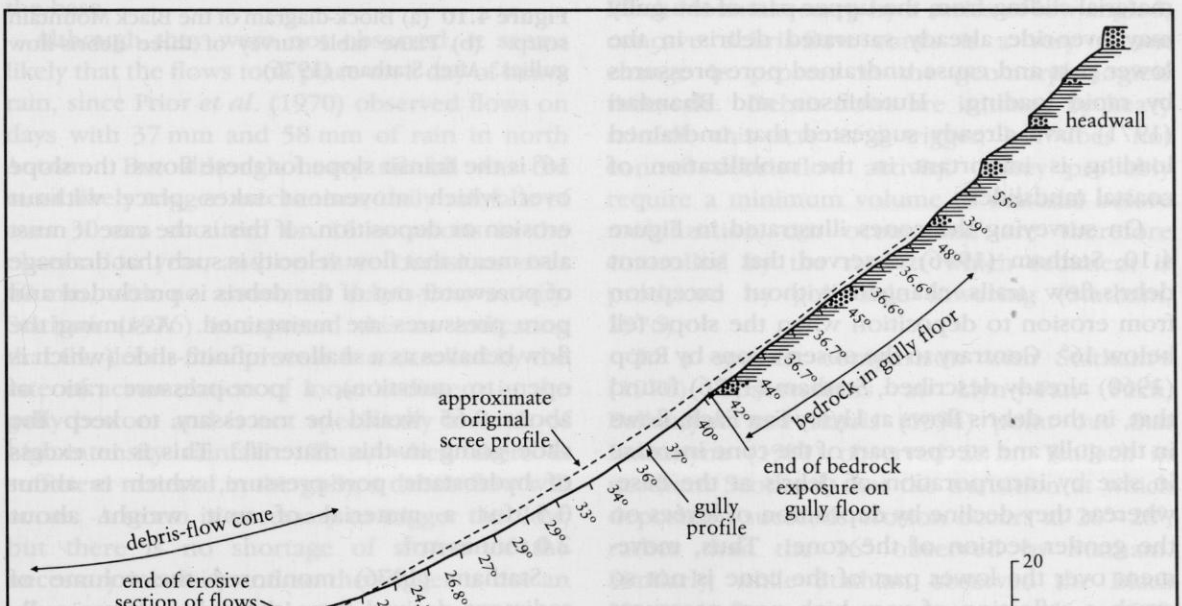

Figure 4.9 Profile of a typical gullied debris-flow cone system at Llyn-y-Fan Fâch. After Statham (1976).

be considered effectively cohesionless, and following this assumption Statham (1976) takes maximum gully-side angle as 'probably a reasonable estimate' of the lower repose angle, ϕ_r', for scree origins which involve a debris slide rather than gully flow and erosion.

Debris flows are easily formed by heavy rain and snowmelt yielding gully flow, which leads to erosion and gully-side collapse. Debris slides at the head increase the solids content, and when solids become 79% of the total by weight, the behaviour changes from mass transport to mass movement. The Statham (1976) model only considers the slide (the soil mechanics), not hydraulic or rheological models.

Assuming an infinite cohesionless slide analysis (after Skempton and DeLory, 1957) and taking ϕ_r' to be 43.5°, debris accumulating in the gully bottom due to erosion of the sides would be quite stable and would not begin to slide until the pore-pressure ratio r_u attained 0.15–0.4. For the debris to remain mobile after sliding r_u must increase steadily along the gully axis as slope angle declines. There are two mechanisms which might cause such an increase. Firstly, as suggested by Johnson and Rahn (1970), water may be added from rainfall and flow from the gully sides, which would lead to increasing water content and pore pressure as the debris moved down the gully. Secondly, accumulated material sliding from the upper part of the gully may over-ride already saturated debris in the lower part and cause undrained pore-pressures by rapid loading. Hutchinson and Bhandari (1971) have already suggested that undrained loading is important in the mobilization of coastal mudslides.

On surveying the cones illustrated in Figure 4.10, Statham (1976) observed that six recent debris-flow trails changed without exception from erosion to deposition when the slope fell below 16°. Contrary to the observations by Rapp (1960) already described, Statham (1976) found that, in the debris flows at Llyn-y-Fan Fâch, flows in the gully and steeper part of the cone increase in size by incorporation of debris at the base, whereas they decline by deposition of levées on the gentler section of the cone. Thus, movement over the lower part of the cone is not so much a reflection of very high pore-pressures but more of the inertia of the flow, as velocity declines on the lower angled section. The change from erosion to deposition implies that

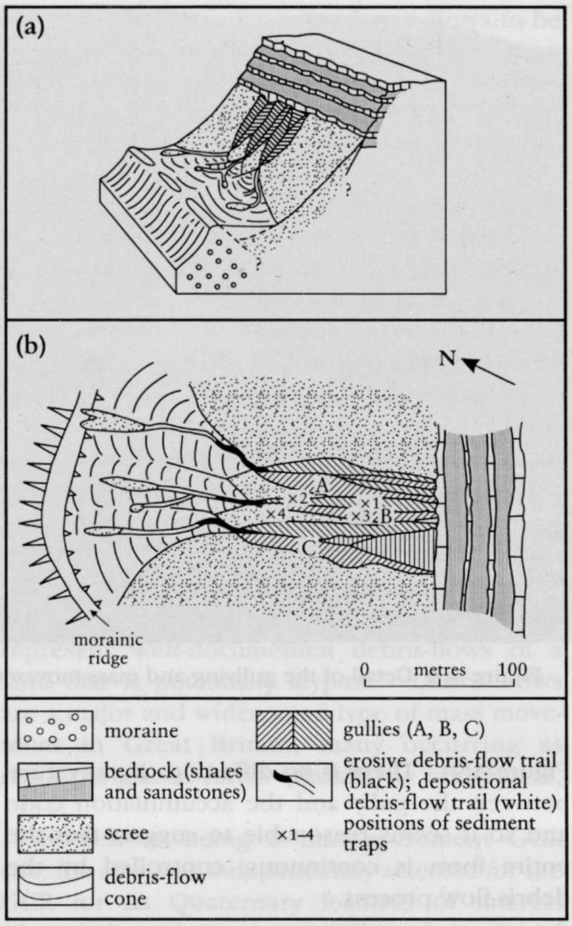

Figure 4.10 (a) Block-diagram of the Black Mountain scarp. (b) Plane table survey of three debris-flow gullies. After Statham (1976).

16° is the transit slope for these flows: the slope over which movement takes place without erosion or deposition. If this is the case it must also mean that flow velocity is such that drainage of porewater out of the debris is precluded and pore pressures are maintained. Assuming the flow behaves as a shallow infinite slide (which is open to question), a pore-pressure ratio of about 0.65 would be necessary to keep the slide going in this material. This is in excess of hydrostatic pore-pressure, which is about 0.5 for a material of unit weight about 2.0 tonnes m^{-3}.

Statham (1976) monitored the volume of sediment derived from the sides of one gully (gully B in Figure 4.10) and the volume transported to the accumulation cone of all three of the gullies in Figure 4.10. He found that the rate of lowering is much greater on the west-facing gully-sides than the east-facing sides. This has resulted in valley asymmetry in the gullies, the western slopes being both the steeper and the more unstable. Measuring the mean rate of surface lowering from the western- and eastern-facing gully-sides as 1.56 cm and 0.33 cm respectively, total yield of sediment to the gully bottom was calculated as 8.4 m^3 for the year. The volume shifted by one debris-flow from this gully was 9.8 m^3 and estimated volumes of another three flows from the three gullies (not necessarily in the observation year) were very similar, from 8.3 m^3 to 11.5 m^3. In consequence, the amount of sediment produced by surface lowering of the gully sides is roughly equal to that moved from the gully by debris-flow activity in one year. Naturally it is recognized that these quantities are only very approximate due to the methods used, and that the time-period of one year may not be sufficient to give a reliable overall picture. Nevertheless, it appears that input of sediment to the gully is balanced by output in debris flows, implying that sediment movement by other processes such as stream flow is negligible. This is supported by the fact that even in very heavy rain Statham (1976) did not observe any surface-water flow in the gullies except where bedrock was exposed in the base.

Although they were not observed, it seems likely that the flows took place on a day of heavy rain, since Prior *et al.* (1970) observed flows on days with 37 mm and 58 mm of rain in north Antrim. But although heavy rainfall was the most likely trigger mechanism, daily rainfalls of over 30 mm occurred on 16 occasions in the observation year, and on three occasions over 60 mm, with no associated debris-flow activity. Statham (1976) interpreted this as indicating that the debris-flow process is controlled by the rate of accumulation of loose sediment in the gully bottom and is not specifically a result of high-intensity rainfall. Thus, when there is sufficient material in the gully, a debris flow will occur. A storm is necessary to trigger the flow, but there is no shortage of storms of the necessary intensity and so the trigger is not an effective process control.

However, there is a substantial body of literature suggesting that availability of debris is the main control.

Interpretation

Given the occurrence of high-intensity rainfall, there seems to be very little climatic control on debris-flow activity (Statham, 1976), with a remarkable similarity of style and form of movement and of topographical situation in which flows are initiated. As the volume of the instrumented gully is 540 m^3, with removal of 8–10 m^3 per year, it cannot be more than 540–700 years old, assuming that the annual rate of removal has remained constant. Furthermore, there are no new gullies being initiated on the scarp and all the existing ones are in roughly the same state of development. It seems likely that some environmental change in the recent past was responsible for the initiation of debris-flow activity. Innes (1983) has noted similar initiation of debris flows in the recent past across the whole of upland Britain, and attributed this to environmental change, possibly deliberate fire-setting. Statham (1976) suggests that the causative environmental change at Llyn-y-Fan Fâch may have been the introduction of intensive sheep grazing in the area, resulting in damage to the vegetation surface and exposure of bare ground.

Therefore it seems that the progressive replacement of the straight scree-slopes of the Black Mountain scarp by a series of low-angled, concave debris-flow cones is a very recent change in process in the geomorphological timescale. Debris flows are initiated by heavy rainfall: this acts as a trigger, but does not control debris-flow activity. They probably require a minimum volume of material before mobilization can occur and are therefore controlled by the rate at which sediment is produced by gully-side lowering (Statham, 1976).

In a specific comparison with Statham's (1976) observations at Llyn-y-Fan Fâch, Ballantyne and Harris (1994) point out that Ballantyne (1981) observed at An Teallach in northern Scotland, that the transition at which deposition succeeds erosion occurs at 20°–28°, rather than the 16° observed by Statham. Similarly, while Statham observed the Black Mountain flows to come to rest on a slope of 8°, those at An Teallach stop on gradients of 11°–23°. They remark that these differences may reflect greater flow viscosity at An Teallach.

Conclusions

The Llyn-y-Fan Fâch GCR site is important in showing that the steeper and straighter scree-slope section of the Black Mountain scarp is being replaced by a series of low-angled, concave-upwards debris-flow cones. Rates of erosion in the debris-flow supply gullies suggest that this is a very recent change in process in the geomorphological timescale. Gully-side lowering produced about 8.4 m^3 of sediment from a monitored gully in one year and in the same year 9.8 m^3 of sediment was moved in a single debris-flow event. All of the sediment derived from the gullies is transported by debris flows, while other sediment transport processes, such as stream flow, are unimportant. Although debris flows are initiated by heavy rainfall, heavier storms occurr on other occasions, with no associated debris-flow activity. Since there is no shortage of large storms, they do not control debris-flow activity but merely act as a trigger. Debris flows probably require a minimum volume of material before mobilization can occur, and are therefore controlled by the availability of sediment, which in the Black Mountains is controlled in turn by the rate at which sediment is produced by gully-side lowering.