Quaternary of Northern England

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Chapter 7

Periglacial landforms and slope deposits of northern England

INTRODUCTION

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Periglacial climates and processes

The term 'periglacial' refers to the conditions, processes and landforms associated with cold, non-glacial environments (Harris *et al.*, 1988). The periglacial environment is diverse and includes a wide range of cold, non-glacial conditions, regardless of their proximity to a glacier either in time or space (French, 1996). Periglacial processes are dominated by frost action (frost-weathering of soils and bedrock) and by permafrost-related processes (the development of perennially frozen ground with ice segregation and the thermal contraction and thawing of such ground). Other processes include frost activity in the seasonally frozen layer, rapid mass movements (often over saturated ground) and fluvial and aeolian processes. Although there is no single periglacial climate type, periglacial processes usually dominate areas where the mean annual air temperature is less than +3°C. This commonly is subdivided further at −2°C mean annual air temperature into environments where frost action dominates (mean annual air temperature less than −2°C) and those where frost action is less dominant (between −2°C and +3°C). A combination of latitude, altitude and continentality control the location of areas affected by periglacial climates.

In Britain it is helpful to distinguish between two types of periglacial environment. First is the present-day periglacial environment, where periglacial processes are still active and periglacial landforms and sediments are currently forming. In general this is confined to mountainous and upland areas (Ballantyne, 1987). In northern England the active periglacial environment is manifested normally by small-scale periglacial processes, such as those on Cross Fell (Tufnell, 1985) and the higher summits of the Lake District (Hollingworth, 1934; Caine, 1963a, b), including Helvellyn (Warburton, 1985), Grasmoor (Caine, 1972), Skiddaw (Caine, 1972; Warburton, 1985) and High Pike (Warburton and Caine, 1999). Second is the assemblage of landforms and sediments that are no longer forming under present-day climatic conditions. These essentially are relict features that belong to periglacial climates of the past (Ballantyne, 1984).

It is important to recognize that the influence of periglacial climates on the British landscape has varied both in time and in space. Thus the limits of periglacial activity have migrated across the landscape over time during the Quaternary Period, and it is only relatively recently that the active periglacial environment has retreated into mountainous areas. On this basis, it is possible to recognize four different periglacial regions within Britain according to the intensity of landscape modification (Figure 7.1). Of these, three periglacial regions are found in northern England. First is an area occupied by glacier ice during the Loch Lomond Stadial, where periglaciation is restricted solely to high ground during Holocene times. These areas are limited in extent, and confined to isolated mountainous and upland areas of the Lake District and northern and western Pennines. Second are areas located within the limits of the Late Devensian ice sheet that were subjected to severe periglacial conditions during ice recession after *c.* 18–15 ka and again during the Loch Lomond Stadial. Much of northern England, including the Pennines, North Yorkshire, Lancashire,

Figure 7.1 The periglacial regions present in the area covered by this volume. See text for explanation.

Northumberland and the Cheshire–Shropshire Lowlands is located within this periglacial region. These areas are currently affected to some extent by Holocene periglaciation on high ground. Third are those areas probably not glaciated since the Anglian (marine Oxygen Isotope Stage 12). These areas therefore have been subjected to successive periods of prolonged severe periglacial conditions since that time and include the North York Moors, the Peak District and parts of the southern Pennines.

Partly as a result of the temporal migration of the limits of periglacial activity, a wide range of periglacial landforms and sediments is present within northern England (Table 7.1). In upland areas, such as the Cheviots, Cross Fell and the Lake District, periglacial features form major landscape elements. Examples include the Late

Table 7.1 Controls on the distribution of active periglacial phenomena and relict periglacial features on British mountains (modified from Ballantyne and Harris, 1994).

	Limiting slope angles (0°–40°)	Characteristic regolith: Open-work block deposits	Characteristic regolith: Clast-rich diamicton	Characteristic regolith: Clast-poor diamicton
Active periglacial phenomena				
Small sorted circles	≈0–8°	•		•
Small sorted stripes	≈5–25°			•
Earth hummocks	≈0–20°			•
Solifluction sheets	≈6–16°			•
Solifluction lobes	≈10–36°			•
Ploughing borders	≈6–38°			•
Talus slopes	≈6–37°	•	•	•
Debris flows	≈32–37°	•	•	
Avalanche tongues	? ≈32–36° ?	?	•	?
Deflation surfaces	≈0–11°		•	
Wind-patterned ground	≈0–8°		•	•
Turf-banked terraces	≈6–36°		•	•
Niveo-aeolian deposits	≈0–29°		•	
Sand hummocks	≈4–18°		•	
Nivation hollows	≈12–20°		•	
Relict periglacial features				
Blockfields	≈0–4°	•		
Blockslopes	≈4–40°	•		
Debris surface	≈0–3°		•	•
Debris-mantled slopes	≈3–40°		•	•
Large sorted circles	≈0–5°	•		
Elongated sorted circles	≈3–7°	•		
Large sorted stripes	≈5–18°	•		
Earth hummocks	≈0–6°			•
Hummock stripes	≈1–11°			•
Relief stripes	≈6–25°			•
Boulder sheets	≈5–17°		•	
Boulder lobes	≈10–35°		•	
Sorted sheets	≈7–15°			•
Sorted lobes	≈10–36°			•
Nonsorted sheets	≈6–17°			•
Nonsorted lobes	≈10–36°			•
Talus slopes	≈5–37°	Not applicable		
Protalus ramparts	≈0–14°	Not applicable		
Rock glaciers	≈4–15°	Not applicable		
Nivation hollows	≈9–18°		•	•

Devensian scree at Throstle Shaw, the periglacial alluvial fan at Sandbeds, the active scree slopes at Wasdale, and the protalus rampart at Dead Crags, Skiddaw. Elsewhere, at sites such as Chelford and Four Ashes, where periglacial activity is represented only in sedimentary successions and has little surface expression, the evidence is more subtle. The aim of this chapter is to outline the nature of periglacial climates and landforms, focusing in particular on the landforms and sediments that are encountered in northern England. Ballantyne and Harris (1994) provide an excellent general review of periglacial processes and landforms both globally and in Great Britain.

Permafrost characteristics

One of the distinguishing characteristics of the periglacial environment is the presence of permafrost. Permafrost generally is defined as ground in which the temperature remains below 0°C over at least two consecutive years (Ballantyne and Harris, 1994). Permafrost is not always present at the ground surface, where a seasonally thawed zone (the active layer) develops in the summer months, but normally is present only at depth. No true permafrost exists in Britain, although there is evidence that it has been widespread in the past (Seddon and Holyoak, 1985). The existence of features indicative of former permafrost conditions suggests mean annual temperatures of below –5°C (Rose, 1975; Sissons, 1977, 1979b). The landforms and sediments indicative of widespread former permafrost are therefore important palaeoclimatic indicators (Williams, 1975).

Periglacial landforms

Ice-wedge casts and polygons

The processes operating in permafrost terrain create a diverse range of landforms and sediments. Some of these processes, for example the seasonal expansion and contraction of the ice and sediment in the active layer, may also penetrate downwards into the permafrost beneath to create ice wedges. Relict examples of ice wedges, in the form of ice-wedge casts, are often present in Quaternary sedimentary successions. They have been described at sites such as Four Ashes and Chelford, as well as throughout the area covered by this volume (Worsley, 1966). They are often visible from the air in plan view as large networks of polygons.

Pingos, thufurs and thermokarst

Permafrost processes often involve the growth and subsequent melting of massive ground-ice masses within near-surface sediments. This process can create large near-surface ice and sediment blisters that rise above the surrounding topography as pingos. Melting of the ice core in pingos leaves a more or less circular depression surrounded by a low rampart (pingo scars). Examples in Britain include those of the Aberystwyth and mid-Wales area (Watson, 1971, 1972), East Anglia (Sparks *et al*., 1972) and in the Whicham Valley, Cumbria (Bryant *et al*., 1985). Bryant and Carpenter (1987) provide a detailed review of these features in Britain as a whole. Much smaller ground-ice lenses can create earth hummocks or thufurs. These features are found in the northern Pennines and possibly are forming under present-day conditions. Shallow thaw lakes on permafrost surrounding pingos often create a complex association of irregular water-filled depressions and hollows known as 'thermokarst'.

Patterned ground

Repeated freezing and thawing of the active layer in permafrost regions is responsible for a suite of small-scale landforms and sedimentary structures. The dominant process in these situations is thought to be the segregation and sorting of sediment into fines and coarser debris/clasts, after sufficient numbers of freeze–thaw cycles (Figure 7.2). The landform–sediment assemblage produced by this process includes patterned ground (circles, polygons, irregular

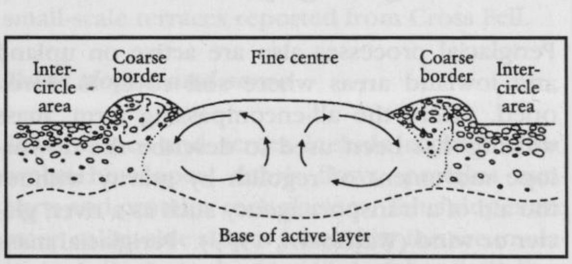

Figure 7.2 A model for sediment displacement within periglacial sorted circles (modified from Hallet and Prestrud, 1986).

networks and stripes) and cryoturbation structures such as involutions. The term 'patterned ground' refers to terrain that exhibits regular or irregular patterning in the form of circles, polygons, irregular networks or stripes (Ballantyne and Harris, 1994). Landforms of this type frequently are associated with cold environments, where ground freezing and thawing is the dominant formative mechanism. As such, they are commonly regarded as indicators of former permafrost conditions (French, 1996). The surface-patterned ground features at upland sites such as Helvellyn, Grasmoor, Cross Fell and the Stiperstones fit with this overall palaeoclimatic interpretation. Following the original classification of Washburn (1956) a distinction is commonly made between sorted patterned ground and non-sorted patterned ground. Sorted patterned ground is defined by the alteration of fine and coarse debris, whereas in non-sorted patterned ground the pattern is formed by microrelief and/or vegetation cover.

Large-scale sorted patterned ground is common on higher British mountains, where essentially it is a Late Devensian relict (Ball and Goodier, 1968; King, 1971; Ballantyne, 1984), but is less common in lowland areas. Large stone-stripes have been described from lowland sites on the chalklands of East Anglia that remained outside the limits of the Late Devensian ice sheet (Williams, 1964; Watt *et al*., 1966; Evans, 1976; Nicholson, 1976; Ballantyne and Harris, 1994) and beneath the granite tors of Dartmoor (Te Punga, 1957). Also, they have been described on the western slopes of Great Dun Fell in the northern Pennines. The slope-related transition from sorted polygons to stone stripes noted by Goudie and Piggott (1981) is consistent with observations from modern periglacial environments (Ballantyne and Harris, 1994; French, 1996).

Mass wasting and gelifluction

Periglacial processes also are active on upland and lowland areas where soil cover is developed. Here the all-encompassing term 'mass wasting' has been used to describe the downslope movement of regolith by gravity without the aid of a transport agency such as a river, glacier or wind (Washburn, 1979). Periglacial mass wasting covers both the gradual downslope movement of slope sediments as a result of repeated freezing and thawing, and the more localized sudden slope failures that occur during thawing of the active layer. The former includes frost creep and gelifluction, whereas the latter includes mudflows, debris flows and ground-ice slumps. Examples occur on Grasmoor, Skiddaw, Helvellyn and Cross Fell.

Gelifluction is the slow downslope movement of soil under conditions of seasonal freezing and thawing. This process creates numerous landforms, including solifluction sheets, lobes and terraces, as well as ploughing blocks. Solifluction operating on upland slopes during periglacial episodes previously has been invoked to explain the bulk of sediments underlying valley-floor terraces in many parts of the British Isles (e.g. Crampton and Taylor, 1967; Watson, 1970; Potts, 1971; Thomas, 1971; Harris and Wright, 1980; Douglas and Harrison, 1985).

Accumulations of periglacial slope deposits formed by mass movement are commonly known as 'head'. The term 'head' covers a wide variety of sediments, but normally these are poorly sorted mixtures of fine matrix and larger clasts. Angular clasts may dominate, although more rounded material also may be present if, for example, the head is reworked from glacial or glaciofluvial deposits. The effects of frost shattering in periglacial climates is most pronounced in upland areas, where it takes the form of mountain-top detritus, including regolith and debris-mantled slopes and plateaux, blockfields, blockslopes and tors (Ballantyne, 1994, 1999).

Nivation and cryoplanation

'Nivation' is an all-encompassing term used to describe the processes of weathering and transport that are accelerated or intensified by the presence of snow patches (Ballantyne and Harris, 1994). These processes include intensive freeze–thaw activity, enhanced chemical weathering, slopewash, transport of debris by snow creep, sediment transport across snow patches to give protalus ramparts, and accelerated solifluction through saturation of regolith downslope from melting snow (Thorn and Hall, 1980; Thorn, 1988; Ballantyne and Harris, 1994). Using modern examples in Greenland, Christiansen (1998a) concluded that the main nivation processes and landforms are backwall failure, sliding and flow, niveo-aeolian sediment transport, supra- and en-nival sediment flows, niveo-fluvial erosion, development of pronival stone pavements, accumulation of alluvial fans and basins, and pronival solifluction. Combinations of these processes are responsible for the creation of a number of erosional landforms ranging from small hollows (Nicols, 1963), through medium-size features (Nyberg, 1991; Caine, 1992; Raczkowska, 1995; Christiansen, 1996, 1998b) to nivation hollows tens of metres in size (Ballantyne, 1978; Dohrenwend, 1984; Rapp, 1984; Gullentops *et al.*, 1993).

Nivation processes operating in periglacial environments are commonly augmented by large amounts of surface runoff in subaerial fluvial streams, especially in the summer months. This type of permafrost runoff has been invoked to explain the origin of large landforms, such as the dry valleys of the Yorkshire Wolds, which were excavated by a combination of solifluction and subaerial fluvial action under periglacial conditions (Cole, 1879, 1887; Mortimer, 1885; Lewin, 1969; de Boer, 1974; Waltham *et al.*, 1997). Sediment removal and valley incision were aided by snow meltwater flowing over permafrost in the annual melt season.

Despite a long history of periglacial research in this country, however, Ballantyne and Harris (1994, p. 248) have commented that 'Relict nivation and cryoplanation landforms rank amongst the most inadequately documented of all periglacial phenomena in Great Britain'. The largest landforms attributed to nivation processes are the so-called 'nivation cirques' that have been mapped in upland Britain. Nivation cirques represent forms transitional between small-scale nivation hollows and mature glacial cirques, and these features have been identified in the Cairngorm Mountains (King, 1968), on Skye (Birks, 1973), the Cheviot Hills (Douglas and Harrison, 1985) and the Ystwyth Valley in mid-Wales (Watson, 1966). Similar processes also may have operated in northern England during the Quaternary to create large nivation hollows such as the Hole of Horcum.

Ballantyne and Harris (1994) have speculated that many of the larger features mapped in upland Britain that are often attributed to nivation processes, such as cryoplanation terraces and nivation cirques, are in fact immature glacial cirques. Nivation cirques 300 to 500 m in diameter and up to 200 m deep would take a prohibitively long time to form given the slow rates of nivation erosion (Thorn, 1976; Nyberg, 1991; Caine, 1992). The large, steep snow patches required to excavate such hollows would quickly turn to glacier ice as there is a threshold length between snow patches and glaciers of 30–70 m from backwall to toe (Ballantyne and Benn, 1994). The dimensions of these relict nivation cirques are an order of magnitude larger than active nivation hollows found in the present-day Arctic. In light of the above, it appears unlikely that nivation processes alone are sufficient to account for large-scale features that occur in northern England. These large features therefore are most probably pre-Pleistocene erosion surfaces or structural benches that merely have been modified by subsequent nivation and frost-action processes. Alternatively, they may represent immature glacial cirque forms occupied during the build up and decay of the Quaternary ice sheets.

'Cryoplanation' is the formation of near-level rock-cut platforms (known as cryoplanation or altiplanation surfaces) by frost action in periglacial conditions. Cryoplanation surfaces have been reported from various sites in the UK, mainly far to the south of the Late Devensian ice limit (Guilcher, 1950; Te Punga, 1956, Waters, 1962), but also closer to this limit in the southern Pennines (McArthur, 1981) and in Shropshire (Clark, 1994a). However, a comprehensive explanation for the processes responsible for the formation of these surfaces has never been produced. Doubts remain about the location and manner of platform initiation, the production of low-angle platform surfaces and the evacuation of debris over gentle slopes as platform growth proceeds (Czudek, 1964; Demek 1964, 1968; 1969). These doubts were sufficient to lead Budel (1982) to dismiss altogether the ability of these processes to create level bedrock surfaces. Indeed, many of these well-developed, near-level, rock-cut platforms (such as those described at Burbage Brook) may actually represent glacially eroded benches related more to the underlying solid geology than to periglacial processes. This may well be the case in the small-scale terraces reported from Cross Fell.

Talus slopes and scree

Talus slopes and screes include rockfall talus, avalanche slopes, debris flows, protalus ramparts and protalus rock glaciers. Talus slopes are steep valley-side slopes formed by the accumulation of debris at the foot of a rockwall (Ballantyne and Harris, 1994) and are a common geomorphological component of upland landscapes in northern England. The term 'talus' is used to describe both the form of the slope and its constituent material. Although the term 'scree' is often used as a synonym for talus, in modern usage this term is reserved for a slope cover of predominantly coarse debris, irrespective of location. Talus slopes are not restricted to present or former periglacial environments but occur in all areas where the products of rock weathering can accumulate at the foot of a cliff or slope.

Talus slopes commonly take one of three forms:

1. talus sheets, where sediment delivery is fairly uniform across a slope;
2. talus cones, where delivery is concentrated or funnelled down a gully;
3. coalescing talus cones, formed where talus cones intersect laterally (Ballantyne and Harris, 1994).

Both active and relict talus slopes are found in Britain. Active talus slopes are commonly found in three environments in Britain: at the foot of sea cliffs, where the products of coastal erosion naturally collect (e.g. at Flamborough Head), below structural escarpments (see for example the site report for Roman Wall Escarpments) and on the lower slopes of glacial troughs and corries in upland areas (see for example the site report for Wasdale Screes). These large talus slopes are perhaps the most spectacular and therefore it is little surprise that they have attracted the majority of attention in this country (Savigear, 1952; Tinkler, 1966; Ballantyne and Eckford, 1984, Ballantyne and Kirkbride, 1987a). Relict talus slopes also are present in lowland areas, although they are often obscured by later deposits or modern soil development. Relict talus slopes are found in many areas of Britain that have been subjected to periglacial conditions, for example at Throstle Shaw in the Lake District and at Ecton in the Peak District. These periglacial conditions persisted during the Late Devensian beyond the limit of the ice maximum, and relict talus slopes are therefore important palaeoenvironmental indicators.

Loess, coversands and ventifacts

Loess (deposits of silt particles between 2 and 64 μm, transported by aeolian processes), coversands (deposits of fine-grained sand between 64 μm to 2 mm, also transported by aeolian processes) and ventifacts (stones or pebbles that have been shaped by wind-blown sand) generally are regarded as indications of cold and arid climates. Loess and coversands are widespread in southern England outside the Late Devensian ice limits and reveal a westward decrease in particle size, consistent with an easterly wind direction (Catt, 1977a, c). An easterly wind direction is also indicated by the close similarities between the mineralogy of the silt-size fraction of Late Devensian glacial deposits and loess in eastern England, suggesting that the source of the loessic material was glaciofluvial sediment in the North Sea Basin (Catt, 1987a). Analysis of coversands in the Yorkshire Wolds, where the depth of coversand reaches 7 m in places, however, suggests westerly palaeowinds (Bateman, 1998). Thermoluminescence dating indicates a Late Devensian age for the majority of these loessic deposits (Gibbard *et al.*, 1987; Parks and Rendell, 1992), although those of north Lincolnshire have been dated to the Younger Dryas (Bateman, 1998). Widespread coversands have also been described in the Vale of York by Matthews (1970).

In contrast, within the Late Devensian ice limits, aeolian deposits are restricted to isolated pockets. Catt (1987c) has described coversands at Sewerby, east Yorkshire and ascribed these to Oxygen Isotope Sub-stage 5d to 5a (Early Devensian) age. In the southern part of the Vale of York a ground surface, including ice-wedge casts, ventifacts and thermokarst features, that developed on Ipswichian river gravels, but beneath Late Devensian glaciogenic deposits, also is probably of Early Devensian age (Bisat, 1946; Gaunt, 1970b, 1976, 1981; Gaunt *et al.* 1972, 1974; Catt, 1977c). One of the most extensive sheets of coversands is the Shirdley Hill Sand Formation in south-west Lancashire, which is of Younger Dryas or Loch Lomond Stadial age (Wilson *et al.*, 1981). Lee (1979) has described loess of a presumed Late Devensian age from Thurstaston on the Wirral. Ventifacts, including faceted and sand-blasted Bunter sandstone pebbles, of a supposed pre-Late Devensian age also have been described from Trysull in the West Midlands (Morgan, A.V., 1973) and in parts of Cheshire (Thompson and Worsley, 1967). These occurrences, in different positions within the stratigraphical record, suggest that there have been several different phases of aeolian activity.

The evolution of tors

Tors are features that are not universally regarded as periglacial in origin. Traditionally, their origin has been explained either with reference to differential weathering in temperate, humid climates or attributed to periglacial stripping. A tor can be defined as 'a residual mass of bare rock that rises conspicuously above its surroundings, is isolated by free-faces on all sides,

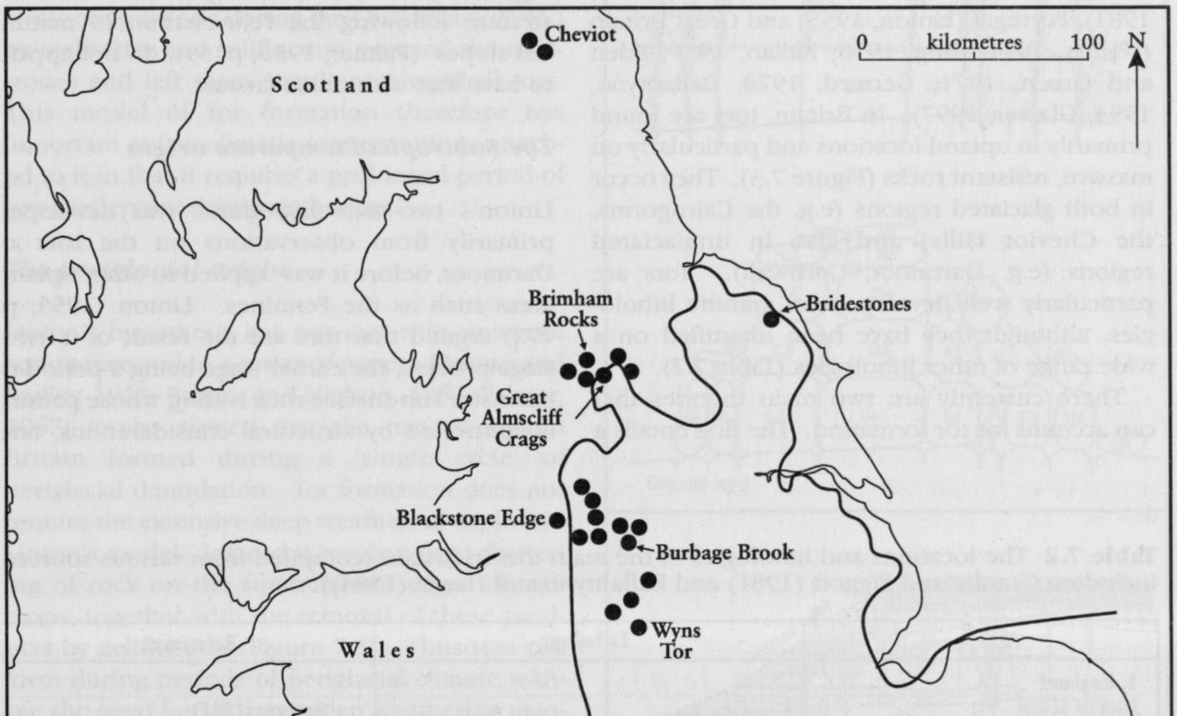

Figure 7.3 The distribution of major areas of tors in the southern British Isles, indicating the location of sites described in this chapter.

and owes its formation to differential weathering and mass wasting' (Ballantyne and Harris, 1994, p. 178). Tors have been described from all conventionally defined climatic regions across the world, including Antarctica (Selby, 1972), Australia (Caine, 1967), New Zealand (Wood, 1969), Canada (Sugden and Watts, 1977; Watts, 1981), Portugal (Linton, 1955) and Great Britain (Waters, 1954; King, 1958; Pullan, 1959; Eden and Green, 1971; Gerrard, 1978; Ballantyne, 1994; Glasser, 1997). In Britain, tors are found primarily in upland locations and particularly on massive, resistant rocks (Figure 7.3). They occur in both glaciated regions (e.g. the Cairngorms, the Cheviot Hills) and also in unglaciated regions (e.g. Dartmoor, Cornwall). Tors are particularly well developed on granitic lithologies, although they have been identified on a wide range of other lithologies (Table 7.2).

There currently are two main theories that can account for tor formation. The first entails a two-stage model of weathering and stripping (Linton, 1955) and the second requires only a single cycle of denudation under periglacial conditions (Palmer and Radley, 1961; Palmer and Nielson, 1962). Other theories, involving the action of present-day seepage moisture (Bunting, 1961) and 'the disintegration of resistant stratum following the rejuvenation of mature hill slopes' (Palmer, 1956, p. 69), do not appear to have met with much favour.

The Subtropical/temperate origin

Linton's two-stage hypothesis was developed primarily from observations on the tors of Dartmoor, before it was applied to other upland areas such as the Pennines. Linton (1955, p. 472) argued that tors are the result of 'a two-stage process, the earlier stage being a period of extensive sub-surface rock rotting whose pattern is controlled by structural considerations, and

Table 7.2 The locations and lithologies of the main tors in Britain (compiled from various sources, including Goudie and Piggott (1981) and Ballantyne and Harris (1994)).

Area	Lithology	References
1. England		
Isles of Scilly	Granite	Scourse (1987)
Dartmoor (South-west England)	Granite	Linton (1955) Palmer and Neilson (1962) Eden and Green (1971) Gerrard (1974, 1978, 1988)
Exmoor (South-west England)	Sandstone	Mottershead (1967)
Weald (South-east England)	Sandstone	Robinson and Williams (1976)
Charnwood Forest (Midlands)	Granite, microdiorite and hornstone	Ford (1967)
Tabular Hills (Yorkshire)	Silicified grits	Palmer (1956)
Derbyshire	Dolomite	Ford (1963, 1969)
	Gritstone	Palmer and Radley (1961) Linton (1964) Cunningham (1964, 1965)
Stiperstones (Shropshire)	Quartzite	Goudie and Piggott (1981) Clark (1994a)
Cheviot Hills	Granite	Common (1954) Douglas and Harrison (1985)
2. Wales		
Central Wales	Igneous rocks, grits	Potts (1971)
Pembrokeshire	Rhyolite	Linton (1955)
Preseli Hills	Dolerite	Linton (1955)
3. Scotland		
Cairngorm Mountains	Granite	Linton (1949, 1955) King (1968) Ballantyne (1994)
North-east Scotland	Granite	Linton (1955)
Ochil Hills	Andesite	Linton (1955)
Ben Loyal (Sutherland)	Syenite	Linton (1955)
Caithness	Sandstones and grits	Linton (1955)
Trotternish, Skye	Basalt	Ballantyne (1990, 1991)

the later being a period of exhumation by removal of fine-grained products of rock decay'. This model requires a long period of deep weathering of the landscape under subtropical conditions so that a weathering mantle of saprolite develops over and around the corestones beneath (Figure 7.4). Later removal of this weathering mantle during the Quaternary Period by meltwater and solifluction exposed the corestones and left them standing as residual tors. This model of tor formation therefore has important palaeoclimatic interpretations attached to it in that it requires a prolonged period of pre-Quaternary deep weathering.

The periglacial origin

Linton's hypothesis has not met with universal acceptance and in a series of papers (Palmer and Radley, 1961; Palmer and Nielson, 1962; Palmer, 1967) it was argued that the tors of upland Britain formed during a 'single cycle' of periglacial denudation. Tor formation does not require the extensive deep weathering explicit in Linton's model. Instead it involves frost shattering of rock on the summits and edges of outcrops, together with the removal of these products by gelifluction (Figure 7.5). Thus tors can form during periods of periglacial climate without the need for a former deep weathering mantle. Indeed, in the Pennines, tors are often part of a periglacial landform assemblage associated with the Millstone Grit. This includes the tors themselves, landslides, blockfields and gelifluction deposits. Evidence for widespread deep weathering generally is absent in this area, although Cunningham (1965, p. 431) in his evaluation of the south Pennine tors concluded that 'clearly the tors were weathered long before the surrounding blocks were exposed'. Cunningham believed that the Pennine tors were exhumed by Pleistocene glaciation, which decapitated some of the tors and totally erased others.

Assessment of the two models

The tors described in this chapter are clearly central to much of the larger debate surrounding tor formation in Britain. Palmer's original 'single cycle' hypothesis is based on his observations in the Pennines and he studied many of the tor sites described here. Sites such as Brimham Rocks and the Bridestones were also used as evidence by Palmer and his co-workers for the single cycle model, whereas Linton cited the Cheviot Tors, Great Almscliff Crag and the tors of Burbage Brook in support of his two-stage model. Other sites, such as the Stiperstones,

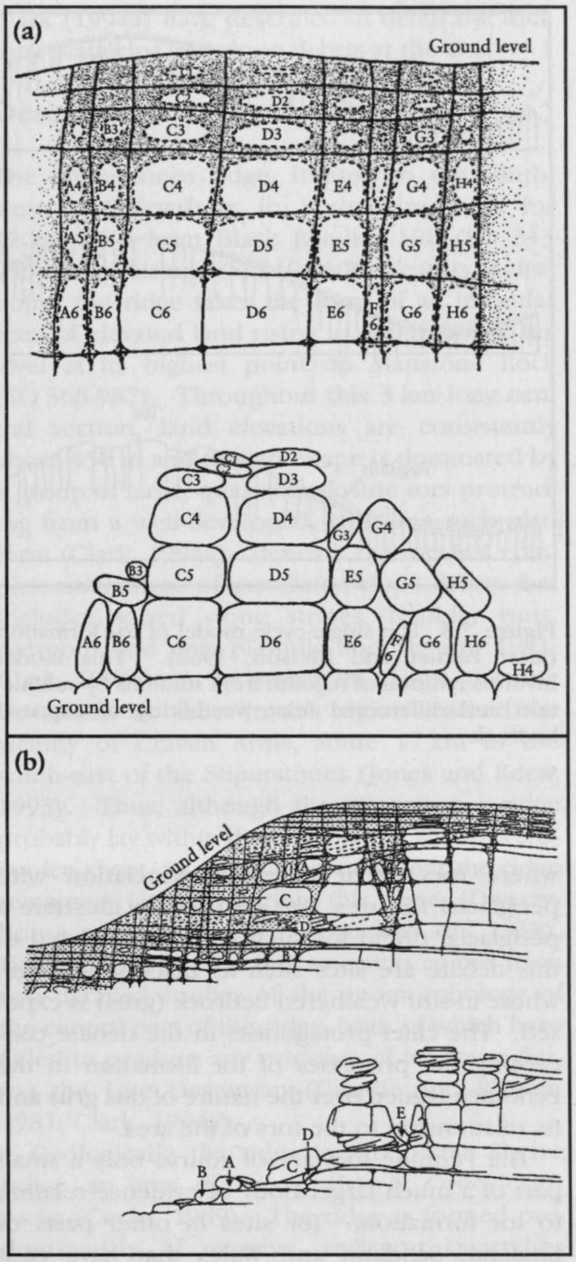

Figure 7.4 (a) The two-stage model of tor formation (modified from Linton, 1955). (b) The two-stage model for development of a gritstone edge (modified from Linton, 1964). Both models invoke preferential deep weathering of densely jointed bedrock and subsequent stripping of this weathering cover under periglacial conditions. Letters and numbers are used to indicate the pathways of individual blocks through the weathering process.

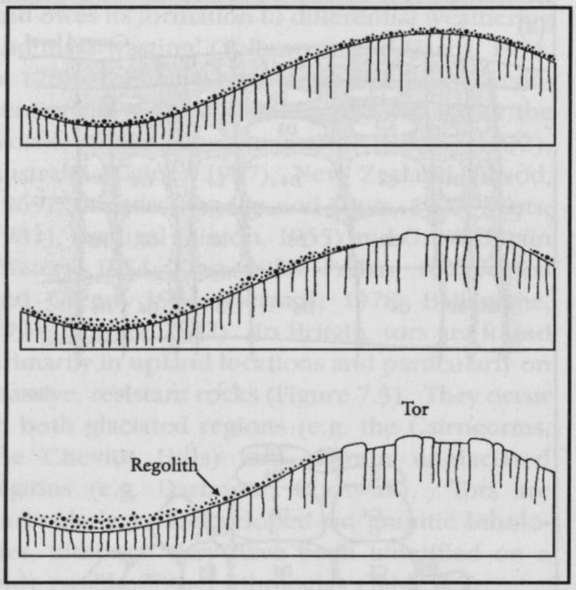

Figure 7.5 The single-cycle model of tor formation (after Palmer and Nielson, 1962). This model involves removal of regolith from summits by solifluction and differential frost weathering of exposed bedrock.

where tors occur in close association with periglacial features, can be used to illustrate a periglacial origin for such features. Related to this debate are sites such as Blackstone Edge, where in-situ weathered bedrock (grus) is exposed. The chief protagonists in the debate concerning the processes of tor formation in the Pennines argued over the nature of this grus and its relationship to the tors of the area.

The Pennine tors are of course only a small part of a much larger body of evidence relating to tor formation. Tor sites in other parts of England, Scotland and Wales that have contributed to this debate were also selected for the Geological Conservation Review. Examples of such sites elsewhere in England are Merrivale, Dartmoor and the Scilly Isles in the south-west of England (Campbell *et al*., 1998). In Scotland there are tor sites on Lochnagar and in the Cairngorm Mountains (Gordon and Sutherland, 1993). Tors occur in Wales at Preseli and Trefgarn in South Wales (Campbell and Bowen, 1989), together with Y Glyderau and Y Carneddau in North Wales (Campbell and Bowen, 1989). Each of these sites represents a different aspect of tor formation in order to encompass all the important factors such as tor location (summit versus scarp-edge), lithology (igneous versus sedimentary) and the extent of weathering and the nature of glaciation (non-glaciated versus glaciated).

Although tors may be polygenetic in origin, it is possible to make some generalizations about their formation. It may be significant that Linton's ideas were based on observations on Dartmoor, where the tors tend to be of the summit type and where deep weathering is common, whereas many of Palmer's ideas stemmed from research in the Pennines where scarp-edge tors are more common. Thus the two-stage model better fits the granite summit tors of areas such as Dartmoor, the Cairngorms and the Cheviots, whilst the single cycle model better explains the gritstone scarp-edge tors of the Pennines (Ballantyne and Harris, 1994). As a result, when first Palmer and Nielson (1962) and subsequently Linton (1964) tried to fit their own models to the other's geographically distinct field areas, flaws began to appear in both models. Ballantyne and Harris (1994, p. 180), in summarizing this debate over tor formation, make the pertinent observation that because 'the Pennine tors tend to differ in all ... respects from those of Dartmoor, it may appear that the search for a single unifying model of tor evolution is fruitless'. Their conclusion is that the evidence in each area should be evaluated independently and that the available evidence points to a polygenetic origin for tors in upland Britain.

Summary

Periglacial processes are varied, but are characterized by frost action, the freezing and thawing of permafrost and/or the active layer, rapid mass movements, nivation and aeolian processes. These processes are in turn responsible for a variety of different landforms and sediments related to factors such as the intensity of freeze–thaw cycles, the availability of moisture, the type of rock or sediment being acted upon, the extent of vegetation cover, the dominant slope angle and aspect. Of the many landforms created in the periglacial environment, tors are perhaps one of the most controversial to the extent that they are not universally regarded as periglacial in origin. The intensity of periglacial modification of the British landscape has varied over time in response to climatic changes and many of the features described in this chapter therefore have important palaeoclimatic interpretations attached to them.

STIPERSTONES (SO 367 985)

N.F. Glasser

Introduction

The Stiperstones, Shropshire, comprises a long quartzite ridge, capped by summit tors and with periglacial stone stripes and polygons developed on its slopes. This area supports one of the best assemblages of relict periglacial landforms in Britain and is an important site for studies of tor formation, patterned ground and slope development. The landform assemblage at the site includes the large summit tors, a series of well-developed blockfields on the surrounding slopes (including sorted stone stripes, polygons and nets) and rock platforms created by cryoplanation processes. The site therefore exhibits many of the key elements that have featured in papers on the development of tors and associated slope features in the British Isles.

Although the Stiperstones ridge is geographically separate from the main cluster of Pennine tors, the landforms at the site contribute much to the long-running debate concerning the origin of tors. Linton (1964) and Cunningham (1965) considered tors to represent the erosional remnants of deep chemical weathering of bedrock during Tertiary times in a subtropical climate. Palmer and Radley (1961), on the other hand, insisted that no deep rotting existed in the Pennines and that the tors represented mechanical weathering under periglacial conditions. The morphology of the Stiperstones tors and their associated slope features indicates intense frost shattering and the development of permafrost at the margins of the Late Devensian ice sheet. The evidence from the Stiperstones therefore points to a periglacial origin for these features since, although the ridge lay within the late Devensian ice limit, it escaped ice cover owing to its high elevation. A further aspect of the site is the recognition of well-developed crest-line cryoplanation platforms. The form and extent of these platforms have been used to argue that cryoplanation processes are significant in landscape change under former periglacial conditions. The site was discussed by Linton (1955). Goudie and Piggott (1981) and Clark (1994a) have described in detail the tors, rock platforms and slope debris at the site.

Description

The Stiperstones ridge, located to the south-west of Shrewsbury in Shropshire, runs for 12 km NNE from Black Rhadley Hill (SO 343 956) to Pontesbury (SJ 340 060). Near its central point, the ridge takes the form of an irregular tract of elevated land rising to 536 m above sea level at its highest point on Manstone Rock (SO 368 987). Throughout this 3 km long central section, land elevations are consistently above 450 m and the landscape is dominated by a group of large, quartzite, skyline tors protruding from a well-developed, crest-line rock platform (Clark, 1994a). Beneath the tors is a complex assemblage of associated slope debris that includes sorted stone stripes, boulder runs, polygons and nets (Goudie and Piggott, 1981; Clark, 1994a). At its maximum extent, the Late Devensian ice sheet in this region reached the vicinity of Craven Arms, some 17 km to the south-east of the Stiperstones (Jones and Keen, 1993). Thus, although the Stiperstones ridge probably lay within the limits of the Late Devensian ice sheet, its altitude ensured that the ridge remained ice-free during this time (Dwerryhouse and Miller, 1930; Pocock *et al.*, 1938). Further evidence for this assertion comes from two detailed studies of the geomorphology of the central part of the ridge, both of which have failed to produce any evidence of ice cover during the Late Devensian (Goudie and Piggott, 1981; Clark, 1994a).

Geologically, the ridge is part of the Shelve Inlier, an area of Arenig (Lower Ordovician) rocks (Cocks, 1989). The ridge is formed predominantly of massive, resistant quartzites belonging to the Stiperstones Member (Whittard, 1931). The quartzites are moderately well-bedded and usually white or light grey in colour. Colour banding is sometimes evident in the form of alternating blue-grey and fawn beds. The strata dip north of west, mainly in the range 25 to 80°. Two principal joint directions have been identified in the quartzite; one striking along the outcrop and dipping east, the other near-vertical and normal to the strike (Dines, 1958). The landform assemblage at the Stiperstones comprises three main components: the tors, rock platforms and slope debris (Figure 7.6). These are described in detail below.

Tors

Along the Stiperstones escarpment, the quartzite is weathered into a series of prominent tors (Figure 7.7). These project above the ridge to varying heights, rising to a maximum of 20 m above the surrounding ground surface at the tor known as the 'Devil's Chair'. Both Goudie and Piggott (1981) and Clark (1994a) describe the tors as 'crestal' or 'skyline' tors because they all rise steeply from the gently sloping summit of the ridge. The long axes of individual tors follow the trend of the ridge and direction of strike. Typical slope angles on the ridge summit are 5 to 13° (Goudie and Piggott (1981), although slopes on the tors themselves are much greater, between 50° and the vertical.

The tors are all composed of in-situ stacks of jointed quartzite, beneath which the slopes are

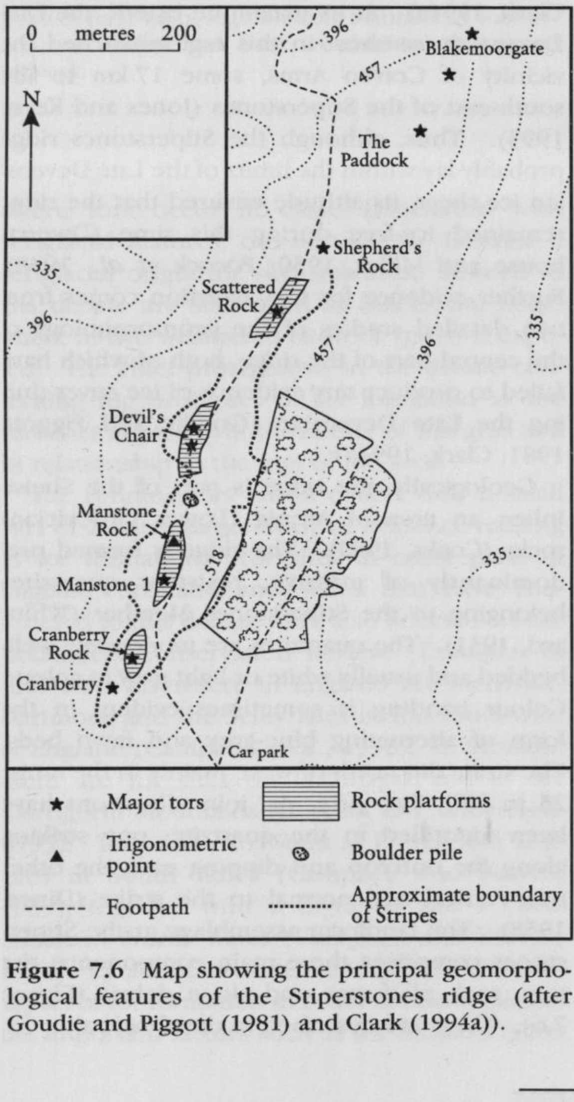

Figure 7.6 Map showing the principal geomorphological features of the Stiperstones ridge (after Goudie and Piggott (1981) and Clark (1994a)).

littered with locally derived, angular, quartzite boulders, with long axes predominantly in the range 0.2 to 0.8 m. The overall form of the tors is controlled by lithological characteristics such as bedding thickness, together with the pattern and spacing of joint planes (Clark, 1994a). The extent to which these bedding planes influences detailed form is controlled by aspect. On the western tor faces, bedding planes exert a major influence by controlling the tor ends and transverse divisions. On the eastern faces, steep longitudinal and transverse joints are more important in controlling tor morphology. Good examples of this contrast can be seen at Scattered Rock where the joints oblique to strike intersect at high angles, and to the north of Shepherd's Rock where individual rock faces reflect the influence of several joint directions and of bedding planes dipping at angles of 70–80°. Many of the tors rise directly from the bedrock ridge beneath and development of undercutting is only sporadic.

Finally, it is perhaps significant that previous studies of the Stiperstones tors (including shallow excavations on the surrounding slopes) have found no evidence of a weathering cover surrounding the tors themselves (Goudie and Piggott, 1981; Clark, 1994a).

Rock platforms

Crest-line rock platforms are common along the ridge crest of the Stiperstones (Clark, 1994a). These features form almost level rock surfaces wherever they abut the ends and sides of the tors. The lower margins of the platforms consist of either vertical or steep rock cliffs between 1 and 2 m in height, or a transition to continuous clast cover on the debris-covered slopes below. The platforms occur both singly and in series, where low rock cliffs separate individual platforms. In some places the platforms constitute the crest line of the ridge, whereas elsewhere the inner margins of the platforms are small rock plinths and piles of loose boulders up to 2 m in height. Individual platforms tend to be narrow, with maximum widths of 10 to 20 m. Clark (1994a) has argued that the form of the cliff margins, including irregular fracturing on the platform surfaces, is indicative of platform extension by retreat of the outer rock edges. Further evidence for this assertion comes from the platform surfaces, which do not appear to be frost-shattered to a great depth and generally are free of

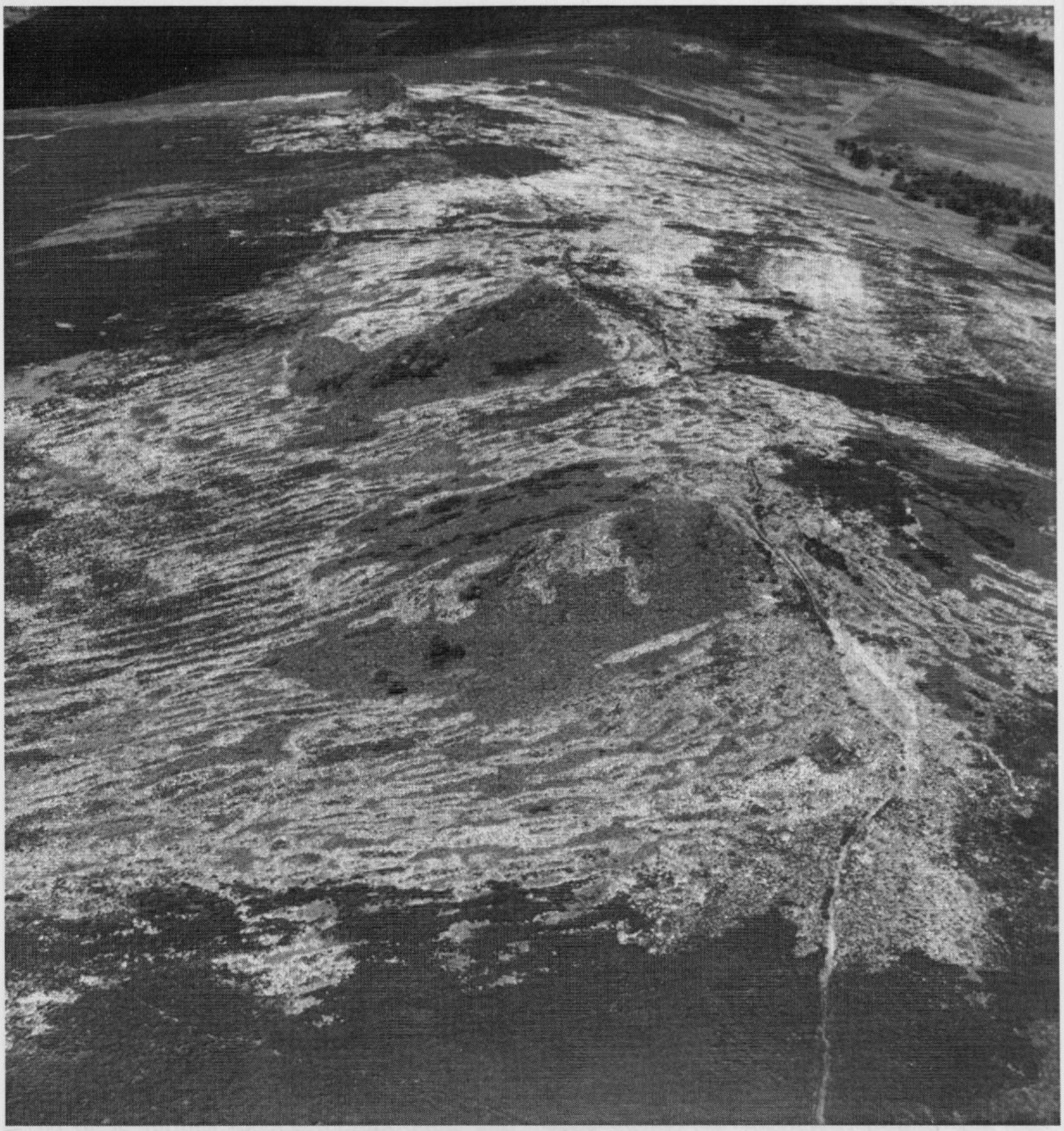

Figure 7.7 Aerial photograph of the Stiperstones Ridge (Photo: WQ 32 of Cambridge University Collection of Air Photographs. © Crown Copyright/MOD. Reproduced with the permission of the Controller of Her Majesty's Stationary Office.)

loose debris. Clark (1994a) regards the weathering along the rock edges to be the result of frost shattering under former periglacial conditions.

Slope debris

The slopes flanking the ridge are generally of a lower angle (ranging from 3 to 15°) than those on the summit (ranging from 5 to 13°). Along much of the length of the Stiperstones ridge, quartzite debris from the tors and rock platforms has moved down the ridge flanks to a level below the quartzite outcrop. Thus the slopes immediately downslope of the tors and rock platforms are littered with surface boulders to a distance of some 200 to 300 m below the crest line. Subsurface exposures are rare and the nature of this slope debris is known only from surface observations and shallow excavations east of Cranberry Rock (Clark, 1994a). This debris is clast-supported, even at depths of 1.2 m, and distinctly lacking in fine-grained material. Where fine material does occur, it is a powdery, humic clay. The excavations also demonstrate that there is a marked downward reduction in clast size. No detailed clast macro-fabric data exist for the site.

Alternating downslope stripes of vegetated ground and bare quartzite boulders cover large areas of the ridge flanks (Figure 7.7). These stripes, termed 'sorted stripe systems' have been mapped in detail by Goudie and Piggott (1981). The stripes are particularly evident in the area between Cranberry Rock and the Devil's Chair (Figure 7.8). Individual stripes attain lengths of 50 to 70 m, and in places the stripes descend 200 to 300 m down the flanks of the ridge. The unvegetated boulder stripes range in width from 0.8 to 9.8 m (mean 3.2 m), whereas the vegetated stripes range in width from 0.8 m to 9.5 m (mean 3.0 m). Clasts in the stripes are exclusively locally derived, angular quartzite and are up to 2.5 m long. According to Goudie and Piggott (1981) many of the elongate clasts in the bare boulder stripes are aligned preferentially in a downslope direction, although Clark (1994a) has noted that there are also clasts with long axes trending across slope. Furthermore, there are sites where preferential alignment of clasts is well developed, suddenly giving way locally to apparently random clast orientations. There also are sites where clasts appear to be arranged roughly concentrically around the upslope edges of larger surface boulders. Clark (1994a) has used this evidence to argue that although there are sporadic concentrations of preferentially aligned clasts, the overall impression is one of no general dominance of organized fabrics.

Close to the ridge crest, the stripes are interlinked into networks that enclose islands of vegetation, termed 'stone polygons' by Goudie and Piggott (1981). The 'polygons' are best developed on flat and low-angled ground (up to 4° maximum slope angle) and are composed of angular, quartzite boulders, centred on vegetated heather ground. They are of an irregular shape and have diameters of 7 to 9 m. They are best developed on the ridge crest, notably between Cranberry Rock and Manstone Rock, and to the south of the highest tor, the Devil's Chair. Other polygons are found on the ridge

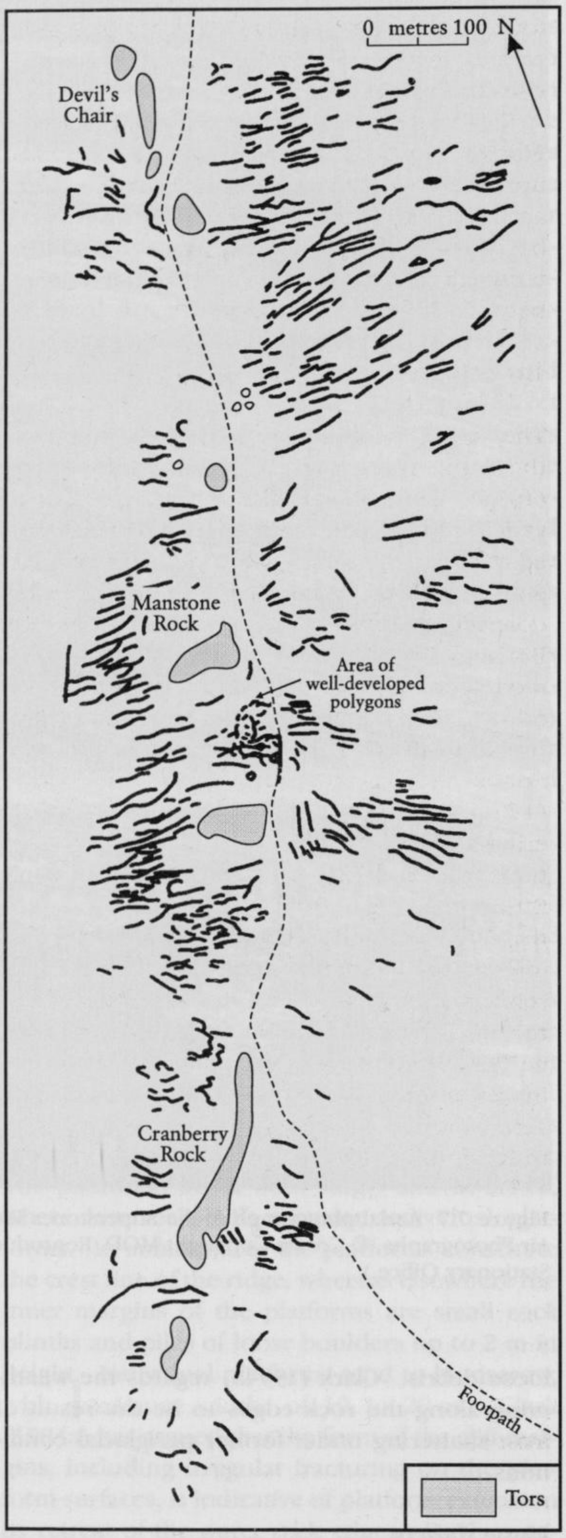

Figure 7.8 Map of the distribution of stone stripes and polygons in the area between Cranberry Rock and Devil's Chair (after Goudie, 1990).

crest around the Paddock. With increasing slope angle the polygons become more elongated in form (at slope angles of 7–10°), before giving way to sorted stone stripes as the slope angle rises to the maximum of 15°. Overall, the stone stripes are more common than the polygons. Finally, bifurcations of the bare stripes in both upslope and downslope directions are found, again creating small islands of vegetated ground between bare boulders. In a detailed study of these features, Clark (1994a) found that linkages between individual stripes are rare.

Interpretation

The geomorphological significance of the tors at the Stiperstones was first noted by Linton (1955). Linton observed that tors occurred on a variety of lithologies in the UK and he contrasted the quartzite tors at the Stiperstones with the granitic Dartmoor tors and the rhyolitic and doleritic examples in the Preselis (Campbell and Bowen, 1989). Linton proposed a two-stage model for the formation of tors, involving a period of deep chemical weathering followed by a period of stripping. Other models of tor formation also exist, including the Palmer and Radley (1961) model involving a single cycle of weathering and mass wasting under periglacial conditions, the Bunting (1961) model involving the action of contemporary seepage moisture, and the Palmer (1956) model involving 'the disintegration of resistant stratum following the rejuvenation of a mature hill-slope'. More recently, Battiau-Queney (1980, 1984) has suggested that the tors of south-west Wales formed in response to deep chemical weathering (in Tertiary times) followed by stripping in response to local uplift along older structural axes.

Of these various models, both Goudie and Piggott (1981) and Clark (1994a) considered the Palmer and Radley (1961) model to be the most applicable to the Stiperstones. The angular nature of the tors, the substantial quantities of angular debris around their bases and the associated slope features strongly indicate a periglacial origin for the landforms at the Stiperstones. No weathering cover or regolith occurs in association with the tors, and there is no evidence of the corestones or rounded boulders indicative of chemical weathering (Linton, 1955). It is likely that the form of the tors therefore dates from the periglacial conditions during the Late Devensian. At this time, the Stiperstones lay close to the ice margin. Given its altitude the ridge may even have been elevated above the ice sheet as an ice-marginal nunatak (Rowlands and Shotton, 1971). The large numbers of well-developed lichens on both the tors and the stone stripes suggest that contemporary frost action is relatively ineffective by comparison. The logical interpretation of the tors at the Stiperstones is therefore that they are relict periglacial weathering features. As there is no evidence for a former weathering cover around the tors or in the sediments downslope of the features, it seems reasonable to conclude they are not remnants of a former weathering cover, as proposed by Linton (1955) for other British tors.

The surface patterned-ground features at the Stiperstones also fit with this overall palaeoclimatic interpretation. The term 'patterned ground' refers to terrain that exhibits regular or irregular patterning in the form of circles, polygons, irregular networks or stripes (Ballantyne and Harris, 1994). Landforms of this type commonly are associated with cold environments, where ground freezing and thawing is the dominant formative mechanism. As such, they are often regarded as indicators of former permafrost conditions (French, 1996). Following the original classification of Washburn (1956), a distinction is commonly made between sorted patterned ground and non-sorted patterned ground. Sorted patterned ground is defined by the alteration of fine- and coarse-grained debris, whereas in non-sorted patterned ground the pattern is formed by microrelief and/or vegetation cover. Clearly there is evidence for both types at the Stiperstones, with the polygons and stone nets demonstrating the formation of sorted patterned ground and the alternating downslope stripes of vegetated ground and bare quartzite boulders illustrating the development of non-sorted patterned ground. The scale of the sorting and the existence of polygons 7 to 9 m in diameter suggest that they developed in association with widespread permafrost (Ballantyne and Harris, 1994). Again this is consistent with the observation that the ridge stood close to the margin of the Late Devensian ice sheet in this area. Frost sorting, downslope mass movement and frost weathering therefore would have been active on the Stiperstones ridge throughout much of the Devensian.

Large-scale, sorted patterned ground is common on higher British mountains, where essentially it is a Late Devensian relict (Ball and Goodier, 1968; King, 1971; Ballantyne, 1984). This type of patterned ground is, however, less common in lowland areas. In addition to the Stiperstones, large stone stripes have been described only from lowland sites on the chalklands of East Anglia that remained outside the limits of the Late Devensian ice sheet (Williams, 1964; Watt *et al*., 1966; Evans, 1976; Nicholson, 1976; Ballantyne and Harris, 1994) and beneath the granite tors of Dartmoor (Te Punga, 1957). The slope-related transition from sorted polygons to stone stripes noted by Goudie and Piggott (1981) is consistent with observations from modern periglacial environments (Ballantyne and Harris, 1994; French, 1996).

A final factor to consider is the relative importance of the different periglacial processes in the formation of the Stiperstones landforms. For example, Clark (1994a) has recently stressed the importance of cryoplanation (the formation of near-level rock-cut platforms by frost action in periglacial conditions) in the formation of the crest line rock platforms. Cryoplanation surfaces have been reported from other sites in the UK (Guilcher, 1950; Te Punga, 1956; Waters, 1962; Gregory, 1966; McArthur, 1981) but a comprehensive explanation for the processes responsible for the formation of these surfaces has never been produced. Doubts remain about the location and manner of platform initiation, the production of low-angle platform surfaces and the evacuation of debris over gentle slopes as platform growth proceeds (Czudek, 1964; Demek 1964, 1968, 1969). These doubts were sufficient to lead Budel (1982) to dismiss altogether the ability of these processes to create level bedrock surfaces.

Conclusions

The tors, rock platforms, patterned ground and slope deposits at the Stiperstones are among the finest in Britain. The angular quartzite tors and associated blockfields illustrate neatly the role of lithology in determining detailed morphology, and the absence of a weathering cover around the tors strongly suggests that the tors and blockfields developed concurrently. The Stiperstones tors therefore provide evidence to support the single-cycle periglacial model of tor formation advocated by Palmer and Radley (1961). Many of the tors rest on crest-line rock platforms that have been used to argue for the effectiveness of cryoplanation as a landscape modelling process in the periglacial environment. Beneath the tors are a family of patterned ground features and associated slope deposits including polygons, nets and stone stripes. The sorted polygons are amongst the largest in the UK and indicate the existence of widespread permafrost conditions. The stone stripes are illustrated by alternating downslope stripes of vegetated ground and bare quartzite boulders and are strong indicators of permafrost conditions. Although no absolute dates exist for this period of periglacial conditions it seems reasonable to assign the features at the Stiperstones to a Late Devensian age, when the site lay close to, or at, the margins of the ice sheet.

BLACKSTONE EDGE (SD 968 176)

N.F. Glasser

Introduction

Blackstone Edge, 8 km north-east of Rochdale, is one of the best exposures of weathered regolith (grus) in northern England and is important for the study of weathering processes, landscape evolution and tor formation. The site provides evidence that the grus is derived from in-situ weathering of Millstone Grit. The nature of this regolith is important because of the role ascribed to the grus by the chief protagonists in the debate concerning the processes of tor formation in the Pennines (Linton, 1955, 1964; Palmer, 1967; Palmer and Radley, 1961; Palmer and Nielson, 1962). The site exhibits some of the key elements that have featured in papers on the development of tors in the British Isles, namely the age and palaeoenvironmental significance of weathered bedrock. Exposures of grus such as that at Blackstone Edge were considered by Linton (1964) and Cunningham (1965) to represent deep chemical weathering of bedrock during Tertiary times in a subtropical climate. Palmer and Radley (1961) on the other hand insisted that no deep rotting existed in the Pennines and that the grus represents mechanical weathering under periglacial conditions. The site has been described in detail by Wilson (1980), who used a scanning electron microscope (SEM) to examine the surface texture of individual grains in the grus.

Description

The site is a disused quarry on a scarp-edge to the south of the A58 Littleborough–Ripponden road. Here a 2 m thickness of coarse sandy detritus (grus) lies beneath blanket peat. The lower surface of the grus rests directly on a platform of unweathered Millstone Grit bedrock (Figure 7.9). Wilson (1980) describes how this weathered detritus is widespread throughout the southern Pennines, although in some localities the peat rests directly on bedrock and the weathering cover is absent. Elsewhere, the peat cover is absent and the weathering cover forms the upper horizon in exposures. Wilson (1980) examined the surface texture of grains from the Millstone Grit grus at high magnification using a scanning electron microscope (SEM). In outline, the grains in the grus are predominantly angular and subangular. Mechanically produced surface features such as cleavage traces, conchoidal fractures, semi-parallel step-like fractures and arc-shaped steps are common on the quartz grains. Wilson (1980) notes that these surface features are extremely sharp and fresh in appearance, suggesting recent formation. Striations and straight or curved grooves and scratches were observed on a small proportion of the grains examined. Many of the surface features are seen to be developed on top of older crystallographically orientated etch pits, which are the product of chemical weathering.

Figure 7.9 Coarse sandy detritus (grus) resting on unweathered Millstone Grit. (Photo: N.F. Glasser).

Interpretation

The work of Wilson (1980) clearly demonstrates that evidence for glacial, subaqueous, or aeolian transport, along with high-energy chemical weathering, is absent. Comparisons of quartz grains from the Millstone Grit samples with those in the unweathered parent bedrock strongly suggest that the material is, in all cases, derived from the underlying bedrock formations. Weathering of the grus at Blackstone Edge therefore was demonstrably an in-situ event. Furthermore, the examination of surface textures on individual grains shows that weathering proceeded in two distinct phases. The first phase of weathering is indicated by the chemically produced etch pits and the second by the mechanically produced fracture patterns on the quartz grains. There is no evidence for the advanced mineral alteration indicative of prolonged weathering under a pre-Pliocene subtropical climate as Linton (1964) envisaged. This accords well with subsequent studies on weathering covers elsewhere in Britain, such as those developed on the granites of north-east Scotland (Hall, 1985, 1986a, b). These weathering covers display limited mineral alteration and they probably developed under humid temperate conditions in Pliocene and early Pleistocene times. The products of true deep chemical weathering in pre-Pliocene subtropical conditions are in fact very rare in the British Isles (Hall, 1985, 1986a, b).

Wilson's (1980) chronology for the sequence of weathering events at Blackstone Edge (chemically produced etch pits followed by mechanically produced fracture patterns) has important implications for models of landscape evolution. In an earlier experiment, Wilson (1979) proved that etch pits could be produced on the surface of quartz grains by the acid contained in peat solutes. One possible interpretation is therefore that etch pit development began only after solutes derived from the overlying peat became available. Thus the etching developed some time after peat development commenced (around 7500 ka). It therefore is possible to speculate that the Pennine grus is much younger than normally supposed. This is an important point because the mechanically produced surface features such as cleavage traces, conchoidal fractures, semi-parallel step-like fractures and arc-shaped steps are, in turn, developed upon these etched surfaces. The period of mechanical weathering therefore must post-date the period of chemical weathering and both could be Holocene in age.

The suggestion of Wilson (1980) that the weathering may be entirely Holocene is controversial because pollen studies by Tallis (1964a–d) throughout the Pennines suggested a Late Devensian age for the grus. Wilson (1980) recognizes the uncertainty of his chronology and admits that a Late Devensian to early Holocene age is also possible for the period of mechanical weathering. Elsewhere in the south Pennines, exposures of grus displaying cryoturbation structures have been reported and the weathering responsible for the formation of the grus has been attributed to Late Devensian periglacial climates (Johnson, R.H., 1967; Palmer, 1967). Whatever its exact age, the grus at Blackstone Edge is indicative of in-situ mechanical weathering and lacks direct evidence of deep chemical weathering.

Conclusions

Three important conclusions can be drawn from studies of the surface texture of quartz grains in the grus at Blackstone Edge. The first conclusion is that weathering was an in-situ event. The second conclusion is that two distinct phases of weathering are indicated by the chemically produced etch pits and the mechanically produced fracture patterns. Thirdly, the surface features developed on the quartz grains do not support the argument of Linton (1964) that the Pennine grus is the remnant of a regolith produced during a period of widespread, deep chemical weathering. Instead, the surface features suggest that mechanical weathering, presumably by macrogelivation under periglacial conditions is more likely. The grus at Blackstone Edge therefore indicates that deep rotting may have been limited in the Pennines, as first suggested by Palmer and Radley (1961). The site therefore is crucial to the debate concerning the nature of the Pennine weathering covers, landscape evolution and tor formation.

BRIMHAM ROCKS (SE 210 650)

N.F. Glasser

Introduction

The tors at Brimham Rocks, North Yorkshire, are significant for studies of rock weathering, periglacial processes and landscape evolution. The tors are a classic example of the scarp-edge tors that characterize the Millstone Grit of the northern Pennines. The Pennine tors are at the heart of a controversy concerning the nature of deep weathering and tor formation in the British Isles. There are currently two widely accepted theories that can account for tor formation. The first entails a two-stage model of weathering and stripping (Linton, 1955, 1964) and the second requires only a single cycle of denudation under periglacial conditions (Palmer and Radley, 1961; Palmer and Nielson, 1962). Most of this debate has centred on the tors of the northern Pennines, including those at Brimham Rocks, which are developed in Millstone Grit. One of the characteristics of the Pennine tors is that they commonly are developed on the edge of escarpments. Palmer and Radley (1961) termed these features 'edge tors', although they are also widely known as 'scarp-edge tors'. The tors at Brimham Rocks have been referred to by Mackintosh (1869), Ramsay (1872) and Versey (1948). Palmer and Radley (1961) used the tors here to propose a model for tor formation during the periglacial phases of the Pleistocene Epoch. Linton (1955, 1964) cited the tors at Brimham Rocks as evidence for a two-stage cycle of development involving deep weathering followed by stripping.

Description

Brimham Rocks is the collective name for a large group of tors fringing the summit of Brimham Moor. The tors are developed on the massive, pebbly, arkose sandstones of the Kinderscout Grit Group of the Millstone Grit Series. Tors occur in various locations around the summit of the moor, but are preferentially developed close to the edge of the escarpment. Many of the tors at Brimham Rocks are over 10 m high (Figure 7.10) and are located up to 100 m from the scarp edge, whereas others lie immediately in front of the scarp crest line and clearly were once part of the escarpment face. Tor morphology is strongly controlled by the density and spacing of joints in the Millstone Grit. Boulders derived from the scarp face litter the slopes in front of the escarpment.

Interpretation

The tors at Brimham Rocks have been interpreted in various ways. During the 19th century, when marine denudation was though to have created great planation surfaces across Britain, the tors were interpreted as sea stacks formed by marine erosion (Mackintosh, 1869). Subsequent to this, Ramsay (1872) and Versey (1948) attributed the form of the tors to wind-etching processes, although in the light of modern evidence this origin now seems unlikely. Palmer and Radley (1961) suggested that Brimham Rocks formed as buttresses on the scarp face as it retreated during the periglacial phases of the Pleistocene Epoch. Frost shattering along joints was considered to be the process of rock disintegration, with solifluction transporting frost-shattered debris downslope. The tors were then left as residual landforms marking the end of a 'one-cycle' phase of landform development.

Linton (1955, 1964) argued that tors are the products of a two-stage cycle of development. In this model, the Millstone Grit of the Pennines was deeply weathered during pre-Pliocene subtropical climates. During this time chemical weathering was able to penetrate to great depths in the rock so that at the onset of the Pleistocene Epoch the landscape was one of pockets of deep weathering surrounding more competent rock masses at depth. The weathered regolith was then removed by glacial and periglacial processes to leave the more resistant rock masses standing as isolated tors. Linton (1964) cited Brimham Rocks as a site that demonstrates many of the aspects of his model. Whereas Linton (1964) interpreted the boulders in front of the scarp face as rounded corestones, Palmer and Radley (1961) regarded them as subangular

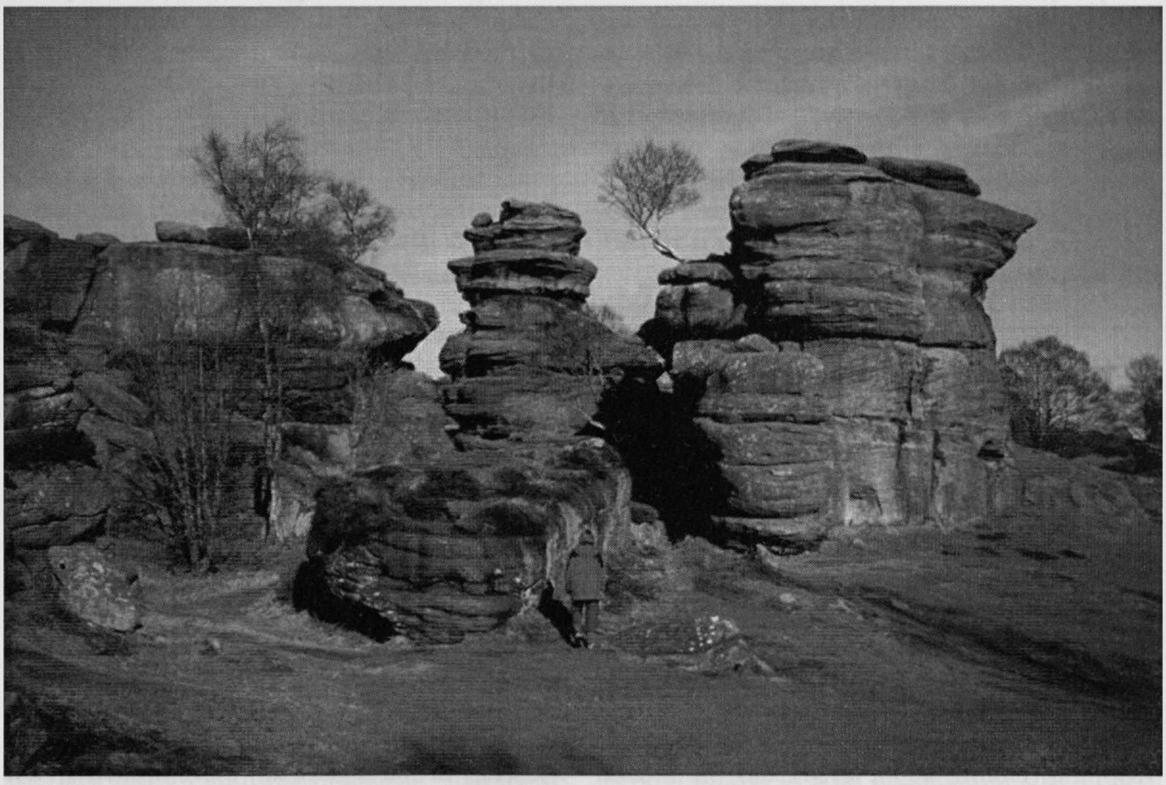

Figure 7.10 The tors at Brimham Rocks. (Photo: N.F. Glasser.)

blocks of unweathered sandstone broken from the scarp face by frost action.

The key to the debate over tor formation lies partly in the nature of the Pennine regolith. Wilson (1980) has demonstrated for a number of Pennine sites that this regolith developed from in-situ weathering of the underlying Millstone Grit. Using scanning electron microscope (SEM) to examine the surface texture of individual grains in the grus, Wilson (1980) identified two distinct phases of weathering (chemical followed by mechanical). He found no evidence, however, to support the argument of Linton (1964) that the Pennine grus is the remnant of a regolith produced during a period of widespread deep chemical weathering. Mechanical weathering, presumably by macrogelivation under periglacial conditions, is cited as the dominant weathering process (Palmer and Radley 1961). Thus it seems likely that the tors at Brimham Rocks formed as buttresses on the scarp face during scarp retreat in a periglacial climate.

Conclusions

Brimham Rocks are an excellent example of the scarp-edge tors that characterize the Millstone Grit of the northern Pennines. The precise origin of the Pennine scarp-edge tors remains unresolved, but the morphology of the features at Brimham Rocks and the lack of a widespread deep weathering cover suggest a primarily periglacial origin for these features.

BURBAGE BROOK (SK 260 815)

N.F. Glasser

Introduction

The site at Burbage Brook, contains an assemblage of sediments and landforms that are significant for studies of rock weathering, periglacial processes and landscape evolution in the Pennines. The Pennine tors are at the heart of a long-standing controversy concerning the nature of deep weathering and tor formation in the British Isles. There currently are two main theories that can account for tor formation. The first entails a two-stage model of weathering and stripping (Linton, 1955, 1964) and the second requires only a single cycle of denudation under periglacial conditions (Palmer and Radley, 1961; Palmer and Nielson, 1962). Much of this debate has focused on the tors of the Pennines, which are developed in Millstone Grit. One of the characteristics of the Pennine tors is that commonly they are developed on the edge of escarpments. Palmer and Radley (1961) termed these features 'edge tors', although they are also widely known in the literature as 'scarp-edge tors' (Cunningham, 1964, 1965). This site is noted for its particularly well-developed scarp-edge tors.

The landform–sediment assemblage at Burbage Brook includes a variety of escarpments, tors, structural benches, solifluction deposits, weathering features, blockfields, blockslopes and a Devensian Late-glacial soil. The site has been used by Linton (1964) to illustrate his theory of tor formation and by Cunningham (1965), who identified and described in detail eight tors on Mother Cap Moor as part of his discussion of the south Pennine tors. A detailed description of the geomorphology of the Burbage basin is given by Said (1969). This area of the Pennines was also used in the pioneering work on slope form and slope evolution (Young, 1961; Carson and Petley, 1970) and in attempts to establish a denudation chronology for the British uplands (Johnson and Rice, 1961; McArthur, 1970, 1971, 1977, 1981).

Description

Burbage Brook occupies an area of *c.* 3.5 km^2 of Hathersage Moor, some 10 km to the west of Sheffield. These moorland slopes form a broad valley containing the Burbage Brook. Geologically, rocks of Namurian (Upper Carboniferous) age dominate the area (the 'Millstone Grit' of Eden *et al.*, 1957). At Burbage Brook these are primarily massive sandstones, siltstones and mudstones of the Chatsworth Grit and the Middle Grit Group. A number of prominent cliff faces and scars, including those at Millstone Edge and Burbage Edge, define the upper slopes of the Burbage valley. In places, these cliffs reach heights of up to 30 m and immediately below them are accumulations of detached blocks. Large boulder fields extend for a considerable distance below the edges. On the eastern side of Burbage Brook are a series of continuous scarps and slopes. The area to the west of Burbage Brook is generally more broken up and consists of a number of tors, including those at

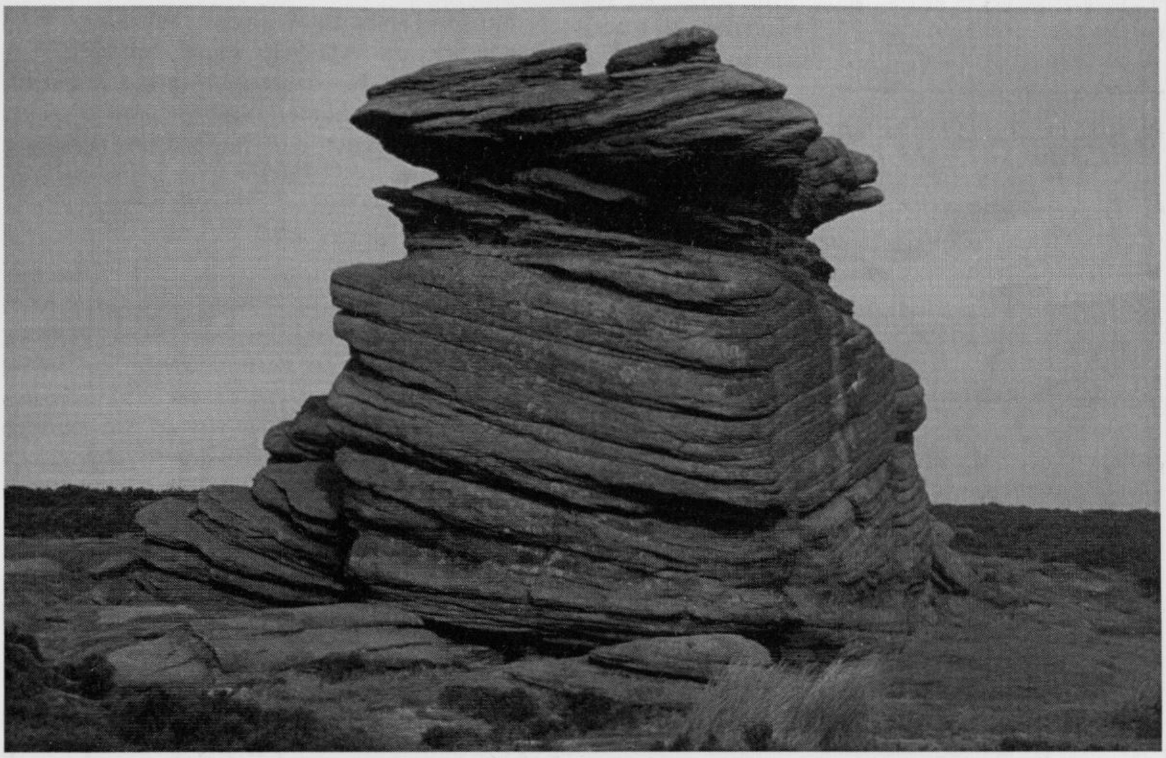

Figure 7.11 A typical tor above Burbage Brook. (Photo: N.F. Glasser.)

Higger Tor, Carl Wark, Winyards Nick and Over Owler Tor (Figure 7.11). These tors generally consist of large, isolated hills that are capped by deeply fissured and often disturbed Chatsworth Grit. Disturbance of the sandstone cap can be seen at its margins, where collapse and downslope movement is evident. Extensive blockfields are developed below the tors, especially on the slopes below Higger Tor and Earl Wark.

The tors of the Burbage area can be subdivided into two main types (Linton, 1964). These are the higher tors and the lower tors. The higher tors are located on the top of free faces, on the back slope of cuestas and on ridge summits. They exhibit evidence of deep weathering along joints and bedding planes. Large weathering pits are common on their summits. There are no free faces of corresponding height in the immediate vicinity. The lower tors are those immediately in front of free faces. They show little evidence of deep weathering along joints and bedding planes. Weathering pits are less common on their summits, but where these do occur they are shallow and widely spaced. Well-developed weathering pits have convex lips, an inner wall and floor. Some pits are crossed by open joints and cracks, possibly of periglacial origin. A second type of weathering is evident in the form of honeycomb weathering on the sloping surfaces of free faces, tor plinths and blocks. Overall, the honeycomb weathering appears to be less common than the weathering pits.

Said (1969) completed the most comprehensive study of the geomorphology of the area (Figure 7.12) in an attempt to produce a regional chronology of events. Evidence used to determine Quaternary environmental change includes sedimentary descriptions, the analysis of pollen and organic remains, slope profiles and the evolution of the edges, tor morphology, weathering pits and other weathering phenomenon, such as honeycomb weathering (Said, 1969).

Soil and vegetation currently cover most of the Burbage Brook catchment. Beneath this cover, the sedimentary evidence is exposed mainly in stream-cut sections along the course of the Burbage Brook. These show evidence for several phases of river aggradation and incision (Table 7.3). The principal facies are indicative of

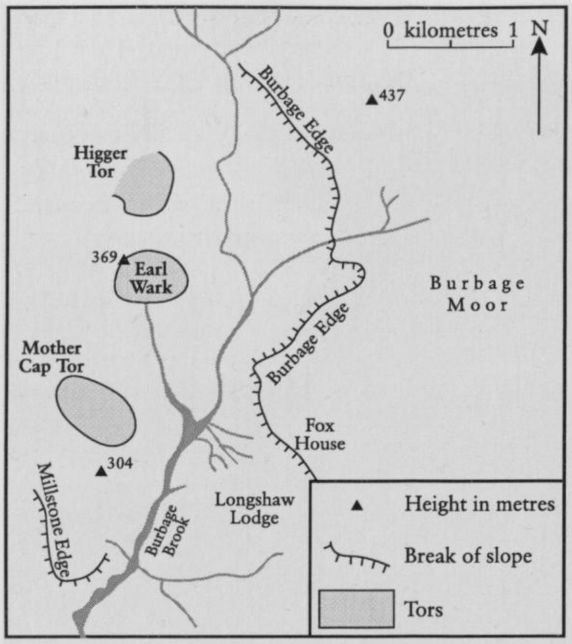

Figure 7.12 Morphological map of the Upper Burbage Basin (after Said, 1969).

alternating fluvial and periglacial deposition. Solifluction deposits (head) are widespread across the area. Above the fluvial and periglacial units are sediments indicating temperate fluvial activity and chemical weathering. Said (1969) examined nine sections along the length of Burbage Brook and in one recovered organic remains sufficiently intact to be identified.

Said (1969) also described slope profiles across the shale–grit and shale–sandstone lithologies. Convex slopes occur on highly jointed grit or sandstone lithologies. Free faces (edges) are restricted to the more massive grit outcrops, particularly near summit crests. Bedding planes and joints on the edges commonly are weathered, particularly on the flaggy sandstones. Slope profiles commonly are covered by head material, although this is deeper and better developed on concave slopes than on convex slopes. The size, location and morphology of boulder accumulations are also controlled to a certain extent by lithology. The blocks are not rounded corestones but show considerable variation in degree of roundness. They are more angular on interfluves and at the foot of free faces, but relatively more rounded at a distance. Large differences in the roundness of sandstone and grit fragments are apparent. Sandstone fragments tend to be more angular whereas the grit fragments are relatively more rounded. All shapes of grit blocks can be found, including oblate, equant, bladed and prolate. Glacial erratics are also noted in the area, but these generally are rare.

Interpretation

Said (1969) explained the different sedimentary units of the area as a function of fluctuating climate in the Late-glacial (Table 7.3). He also related changes in the activity of the Burbage Brook in terms of climate-driven changes in slope stability, vegetation cover, hydrological regime and sediment supply. The slopes of the area show widespread evidence of periglacial modification, including the deposition of head sequences, and are only very slowly modified under present-day climate. Most of the periglacial deposits therefore are regarded as relict rather than having formed under present-day conditions.

Detailed study of the different types of edges (free faces) and their morphology led Said (1969) to suggest that these were formed in the cold phase preceding the antepenultimate glaciation. There was also a phase of face development during the penultimate glaciation. Some of the free faces probably survived the last interglacial. The development of free faces was renewed in the last glacial episode and it is likely that periglacial processes shaped much of their present form at that time. The active evolution of free faces ceased with the cessation of cold conditions at the end of the Late-glacial. McArthur (1981), working in the upper Derwent basin, came to similar conclusions regarding the timing of these episodes of periglacial activity. He described 'periglacial slope planations' and argued that these valley-side benches formed through retreat of sandstone and gritstone scarps and associated removal of rock waste by solifluction. McArthur (1981) invoked river incision of 40–50 m following the initiation of scarp retreat to leave the benches in their current position. The upper Derwent basin, however, lies within the limits of the maximum glaciation and it has been suggested that this is a significant factor in their formation (Ballantyne and Harris, 1994). There appears to be strong structural control on the development of the 'periglacial slope planations' described by McArthur (1981),

Table 7.3 Chronology of the Late-glacial and Holocene events in the Burbage area (after Said, 1969).

Years	Archaeo-logical period	Climatic phase	Pollen zone	Higher areas (above 425 metres)	Lower areas	^{14}C dates (years BP)
		Sub-Atlantic	VIII		(iii) Deforestation, the deposition of washes	
AD 1000	Roman Period				(ii) Deforestation, a short phase of peat erosion and the deposition of organic mud	1730 ± 90
AD	Iron Age			Rapid growth of peat; the initiation of peat erosion	(i) Deforestation and the formation of Parson Terrace	2420 ± 90 2470 ± 80
BC 1000	Bronze Age	Sub-Boreal	VIIb	Slow-growing peat	Hazel, birch, pine woodland	
2000	Neolithic Age					
3000		Atlantic	VIIa	Degeneration of forest vegetation; the formation of peat mires on the upland	Forest vegetation	
4000						
5000		Boreal	VI	Forest vegetation (alder–birch formation)		
6000			V			
7000	Mesolithic Age	Pre-Boreal	IV	Amelioration of climate	Incision of streams in the Burbage Terrace	
8000		Younger Dryas	III	Arctic conditions:	the deposition of Burbage Head on slopes and Burbage Gravel in stream channel, and the formation of Burbage Terrace	
9000		Allerød	II	Temperate conditions:	the development of soil and vegetation; a phase of erosion; the incision of Burbage Brook in the Toad's Mouth Terrace	11 590 ± 360
10000		Older Dryas	I	Arctic conditions:	the deposition of Toad's Mouth Head on slopes and Toad's Mouth Gravel in stream channel, and the formation of Toad's Mouth Terrace	

with the lip of each bench underlain by resistant sandstone and the treads cut across shales. It therefore is equally, if not more, plausible to interpret these benches as the products of differential glacial erosion of strata of variable resistance (Ballantyne and Harris, 1994). The role of cryoplanation therefore may be limited in this area simply to scarp retreat and solifluction of regolith over the surfaces of glacially eroded benches.

Said (1969) considered the development of tors to be an integral part of the development of free faces. Whether a tor is produced in any part of a free face is a function of its lithology and structure. For example, he noted that in the Burbage area lenses of massive grit in the free faces are often left as isolated stacks or small tors. Said argued that because this area was ice covered during the antepenultimate glaciation all tors would be destroyed and that the tors in this area therefore cannot be Tertiary in origin. According to Said (1969), therefore, the active evolution of tors ceased at the end of the Late-glacial. This does not of course allow for the possibility that the tors survived beneath cold-based portions of the ice sheet, as demonstrated elsewhere for other tor landscapes (Fitzpatrick, 1963; Sugden, 1968, 1989; Hall, 1985, 1991;

Hall and Sugden, 1987; Hall and Mellor, 1988; Hall *et al.*, 1989; Ballantyne, 1994; Kleman, 1994). In contrast, both Linton (1955) and Cunningham (1965) advocated Tertiary deep weathering and argued that the tors were weathered long before the surrounding blocks were exposed.

The roundness of detached blocks was probably acquired subsequent to their detachment from the free faces and other bedrock outcrops. The difference in roundness between sandstone and grit boulders probably is lithological and holds no genetic significance. The blockfields probably formed during the penultimate and last glacial phases, but some may be much older. There is stratigraphical evidence to show that the blockfields were formed in both the Older Dryas and Younger Dryas and that their active evolution ceased at the end of the Late-glacial.

Weathering pits are currently developing on the surface of blocks, but those on the sides of blocks are relict. Pits are initiated on the surface of blocks where large quartz pebbles are removed, and develop through enlargement as a result of wetting–drying cycles, freeze–thaw and organic growth and decay. They cannot be used as evidence of Tertiary weathering because they can develop in a variety of climatic settings. Pit initiation probably began in the last interglacial but was interrupted by the onset of cold conditions during the last glacial. Their development has resumed in the Holocene. The honeycomb weathering on the sloping surfaces of free faces, tor plinths and blocks is interpreted by Said (1969) as weathering of an iron-deficient zone beneath iron crusts. Wherever the crust is punctured, for example by the removal of a quartz pebble, the inner, decomposed, iron-deficient zone is exposed and the pattern develops. The pits are probably of little palaeoclimatic significance.

Other workers have confirmed the overall conclusion of Said (1969) that the landscape of the southern Pennines owes much to relict periglacial processes. In his study of the neighbouring upper Derwent basin, for example, McArthur (1981) concluded that solifluction sheets are essentially immobile under the present-day climatic regime. Similarly, landscape elements such as benches, escarpments and regolith are periglacial in origin, but the dominant landscape-forming processes in the southern Pennines under present climatic conditions are fluvial.

Conclusions

Burbage Brook contains an assemblage of landforms and sediments typical of upland areas in the southern Pennines. It is an important site for the study of tors, edges, and the processes of rock weathering and slope formation. An important element of this site is that it has been used to construct a chronology and regional picture of events during Late-glacial and Holocene times. The evidence from Burbage Brook indicates periods of intense weathering under periglacial climates during the Older and Younger Dryas, followed by a return to fluvially dominated, landscape modification under present climatic conditions.

WYNS TOR (SK 241 603)

N.F. Glasser

Introduction

Wyns Tor, Derbyshire, is an important site for studies of rock weathering, periglacial processes and landscape evolution in the Pennines. The origin of the tor has been explained by Linton (1955) in terms of a two-stage model involving chemical weathering followed by a period of stripping under a glacial or periglacial climate. The tor is unusual for the Pennines in that it is developed not on Millstone Grit but on dolomitized limestone. Wyns Tor was selected for the Geological Conservation Review in order to represent the nature of these dolomite tors and their associated weathering cover. Ford (1963, 1969) has considered in detail the nature of the dolomite tors of this area.

Description

Wyns Tor is a large dolomite tor located immediately south of the village of Winster in Derbyshire. Although the southern Pennines are dominated by Carboniferous Limestone, towards the south-east of the limestone outcrop are patches of dolomite reflecting Permian and Triassic dolomitization of the limestone by subsurface waters (Parsons, 1922; Kent, 1957; Shirley, 1958; Ford, 1963, 1969). Tors are entirely absent on the unaltered limestone of the area, but are common on the dolomite areas (Figure 7.13). On the dolomitized limestone, the land-

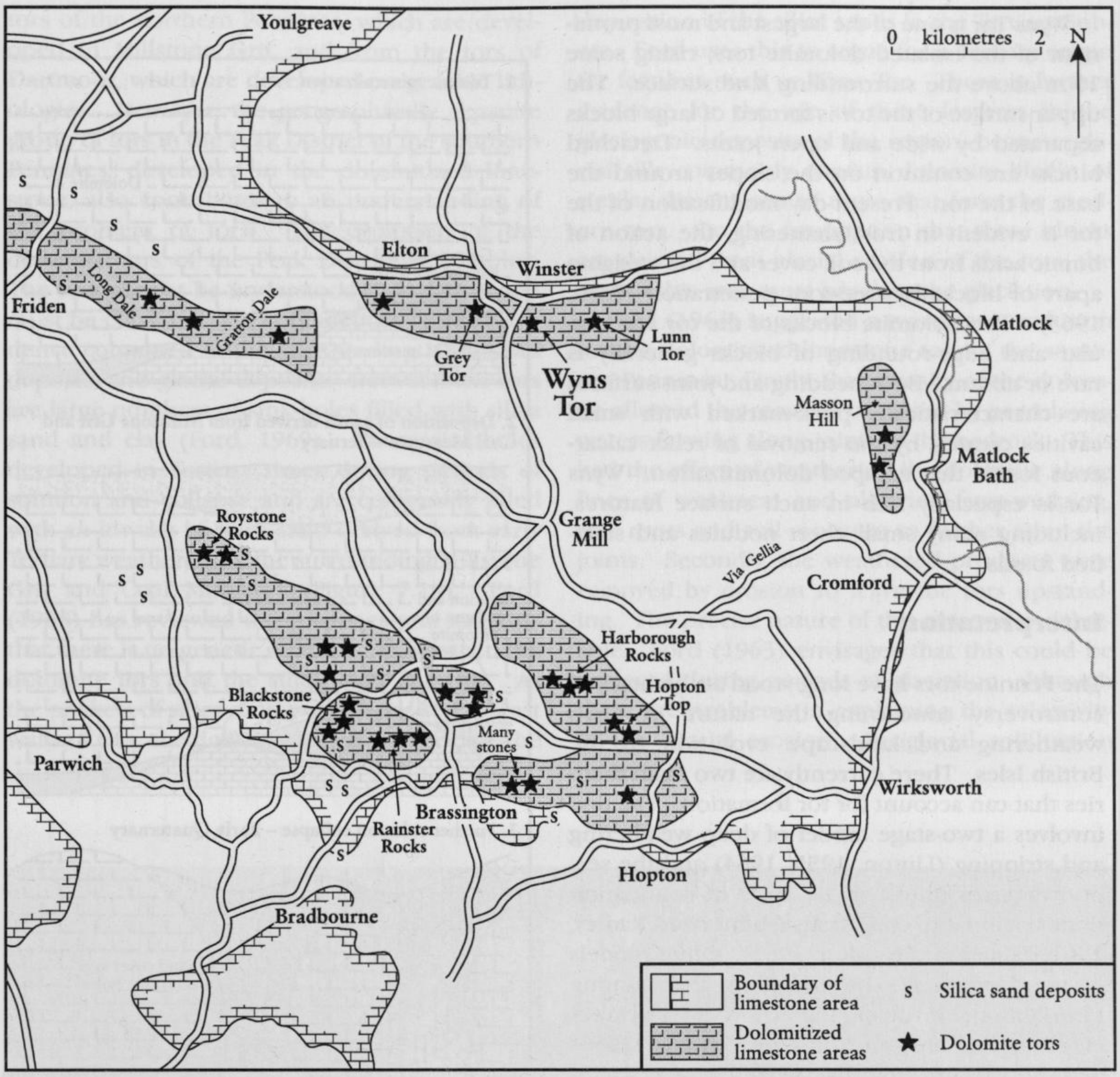

Figure 7.13 Map of the distribution of dolomite tors in the south-east of the Peak District (after Ford, 1963). Note the relationship between the distribution of the dolomitized limestone and the location of the tors.

scape is dominated by isolated tors, scattered sink-holes filled with silica sand, patches of chert gravels and soils developed on glacial till and loess (Piggott, 1962; Ford, 1969). The dolomite tors in this part of the Peak District vary from 15 m high, isolated pinnacles, such as those at Grey Tor and Wyns Tor, to much larger landscape elements such as the castellated escarpments up to 50 m high and 1 km in length at Rainster Rocks. Large, open blocks with well-developed joint systems crown all the dolomite tors and many of the lower slopes are strewn with detached boulders and blocks, sourced from the tors (Ford, 1963). Weathering of the tors in the present-day climate is evident, and temporary excavations described by Ford (1963) have highlighted the importance of vegetation in disaggregation of the dolomite. Decalcification of the dolomite is promoted by soil moisture, by the penetration of tree roots and by humic acids. The net result of this decalcification is to gradually break down the dolomite into an incoherent aggregate of dolomite crystals. In places, temporary excavations have exposed partly exhumed tors from beneath these weathering products.

Wyns Tor is one of the largest and most prominent of the isolated dolomite tors, rising some 15 m above the surrounding land surface. The upper surface of the tor is formed of large blocks separated by wide and open joints. Detached blocks are common on the slopes around the base of the tor. Present-day modification of the tor is evident in frost shattering, the action of humic acids from the soil cover and the wedging apart of blocks by tree-root penetration (Ford, 1963). The dolomite blocks of the tor are angular and edge-rounding of blocks generally is rare or absent. Both bedding and joint surfaces are characteristically pock-marked with small cavities created by the removal of relict calcareous fossils that escaped dolomitization. Wyns Tor is especially rich in such surface features, including many small chert nodules and silicified fossils.

Interpretation

The Pennine tors have long stood at the heart of controversy concerning the nature of deep weathering and landscape evolution in the British Isles. There currently are two main theories that can account for tor formation. The first involves a two-stage model of deep weathering and stripping (Linton, 1955, 1964) and the second requires only a single cycle of denudation under periglacial conditions (Palmer and Radley, 1961; Palmer and Nielson, 1962). Other models of tor formation exist, including the Bunting (1961) model involving the action of contemporary seepage moisture and the Palmer (1956) model involving 'the disintegration of resistant stratum following the rejuvenation of a mature hill-slope'. More recently, Battiau-Queney (1980, 1984) has suggested that the tors of south-west Wales formed in response to deep chemical weathering (possibly in Tertiary times) followed by stripping in response to local uplift along older structural axes. The last three of these theories are difficult to apply, however, and the first two are regarded generally as the most universally applicable.

Evidence for tor formation has been gathered primarily from two geographical areas, from the

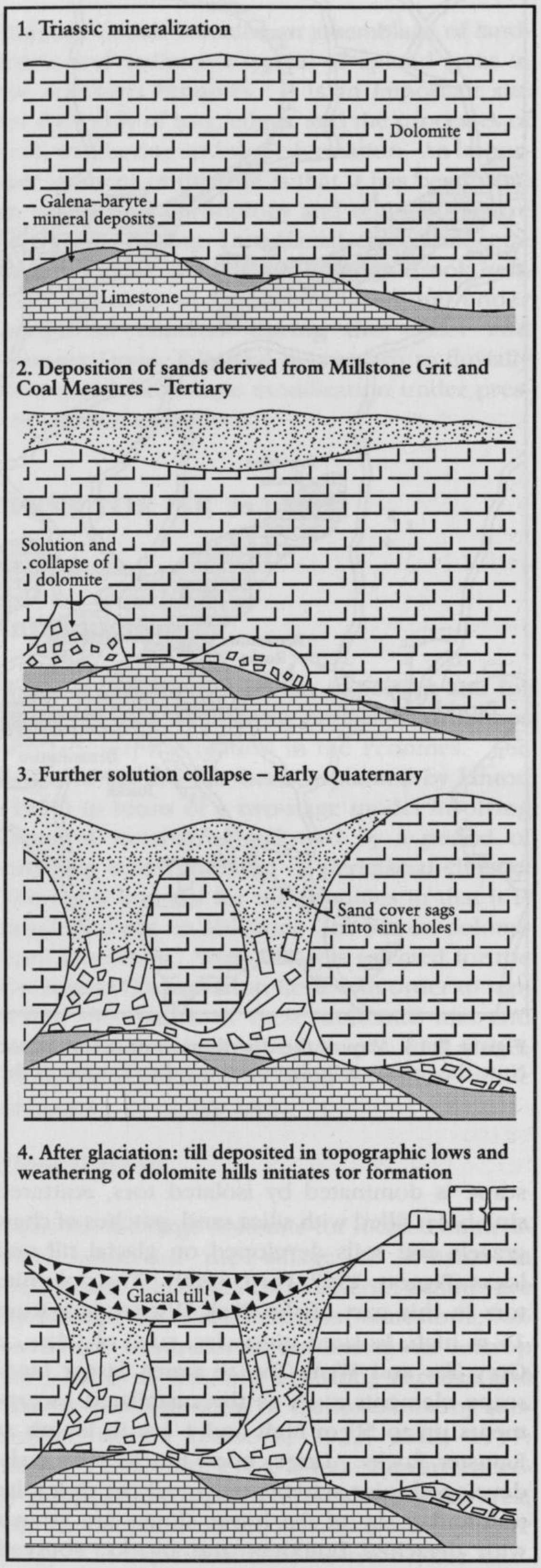

Figure 7.14▸ Simplified model showing the evolution of the sand-filled sink-holes and the dolomite tors in the Peak District (modified from Ford, 1969).

tors of the northern Pennines, which are developed in Millstone Grit, and from the tors of Dartmoor, which are developed on granitic lithologies. However, the geographically separate group of tors in the Peak District of the southern Pennines, developed in the dolomitized limestone, also contributes to an understanding of the problem of tors. The evolution of the dolomite tors of the Peak District is complex. The tors cannot be understood in isolation and must be considered in conjunction with the evidence contained in the surrounding silica sand deposits and glacial deposits. Between the tors are large numbers of sink-holes filled with silica sand and clay (Ford, 1969). These sink-holes developed in Tertiary times during periods of solution and collapse and are commonly filled with an inwash of silica sand derived from early Tertiary weathering of the surrounding Millstone Grit and Coal Measures (Figure 7.14). Ford (1963) has reviewed this evidence and suggests that there is no genetic relationship between the dolomite tors and the silica sand deposits. As the pockets of silica sand were deposited against walls of tors, the dolomite tors must pre-date the deposition of the silica sands in the Tertiary Sub-era. Ford uses this to argue for a Tertiary origin for features such as Wyns Tor. There is further evidence for the age of these features in the glaciogenic deposits of the region, because glacial tills commonly contain dolomite blocks of similar dimensions to those that form the modern tors. On the assumption that these blocks are derived from glacial erosion of the tors, the tors of this area must pre-date the glaciation.

Ford (1963) suggested a two-phase evolution for the dolomitized limestone tors of the southern Pennines. Firstly, the porosity of the dolomite allowed the removal of calcite by percolating waters flowing along joints in the bedrock. This had the effect of weathering the dolomite along lines of weakness and allowing frost-wedging, tree roots and soil moisture to further open the joints. Secondly, the weathered products were removed by erosion to leave the tors upstanding. The precise nature of this removal is debatable. Ford (1963) envisages that this could be achieved during periods of glaciation, although there are problems in explaining the selectivity of this glacial erosion. Periglacial solifluction

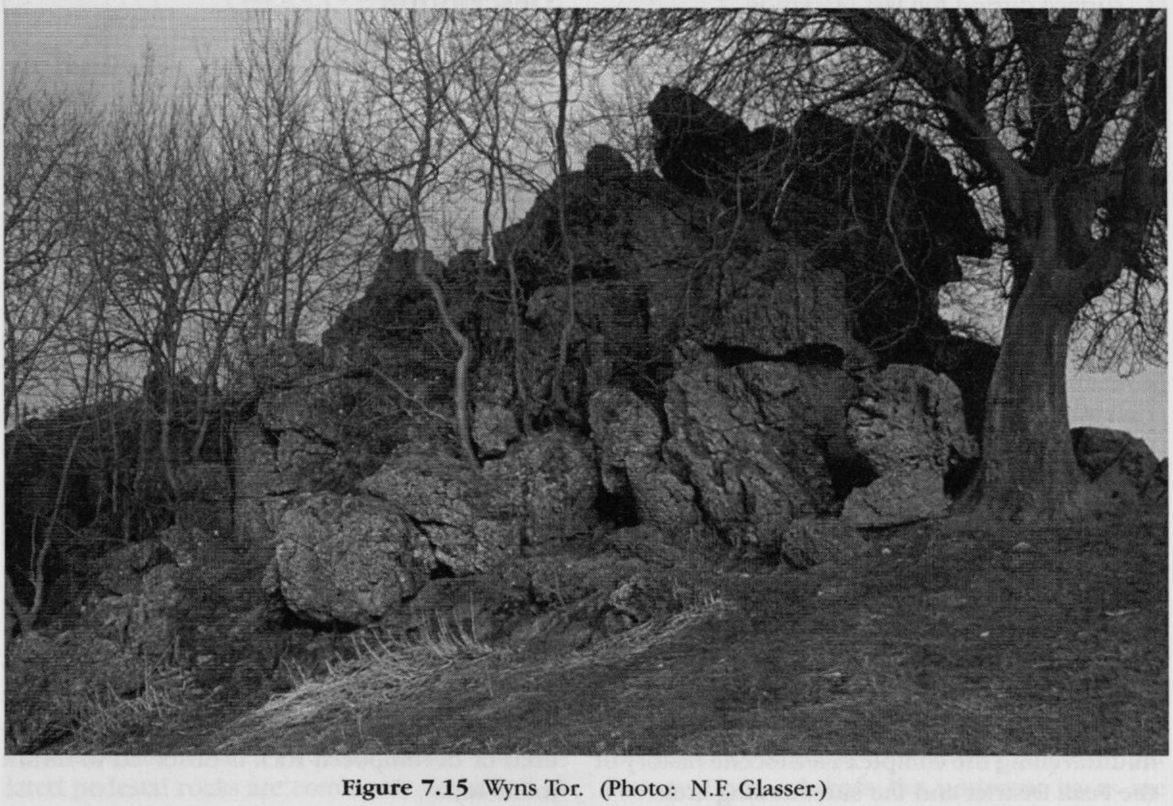

Figure 7.15 Wyns Tor. (Photo: N.F. Glasser.)

and block wedging by frost action may have assisted in the development of the tors, although there is no clear evidence for the weathering process that exposed tors such as Wyns Tor. Ford (1963) clearly prefers an age for exhumation of the tors prior to the last interglacial, whereas in the two-stage model of tor formation proposed by Linton (1955) for the Millstone Grit tors, exhumation would have been achieved during the last glaciation.

To summarize, it is helpful to list the significant features of the dolomite tors of the Peak District noted by Ford (1963). These are as follows:

1. The tors are confined entirely to the dolomite outcrop and do not occur on the Carboniferous Limestone.
2. The dolomite is highly porous and susceptible to chemical decay by the removal of calcite along bedrock joints.
3. Some of the dolomite tors may pre-date the silica sand deposits of the region, and according to Ford (1963) this suggests a Tertiary age for these features.
4. Glacial deposits overlying the silica sand contain blocks derived from the dolomite tors, implying that the tors were exposed at the surface during the last glaciation.
5. The existence of detached boulders around the base of tors such as Wyns Tor (Figure 7.15) implies that modification in the present-day climate is an ongoing process, although the rate of this modification cannot be quantified.

Conclusions

Wyns Tor is the best example of a tor developed on dolomitized limestone in Britain and the site contains important information on the nature of rock weathering, periglacial processes and landscape evolution in this part of the Pennines. The dolomite tors of the southern Pennines have a complex geological history but probably owe their gross morphology to periods of weathering in Tertiary times. Stripping of this weathering cover has proceeded intermittently under a variety of climatic regimes and continues into the present day. As such, Wyns Tor is a good exemplar of the two-stage model of tor formation proposed by Linton (1955). It is therefore a key site in unravelling the complex Pleistocene history of the Peak District and the surrounding area.

BRIDESTONES (SE 872 910)

N.F. Glasser

Introduction

The Bridestones are a collection of tor and rock-weathering forms developed in limestones on the south edge of the North Yorks Moors. The tors show considerable morphological variety, from isolated, undercut, pedestal rocks to scarp-edge tors. Together with sites such as Newtondale and the Hole of Horcum, the Bridestones are important for studies of rock weathering, periglacial processes, landscape evolution and reconstruction of the glacial history of North Yorkshire. Controversy surrounding tor formation has centred around two main groups of tors; in the Pennines and on Dartmoor. However, the tors developed in outlying areas such as Bridestones also contain important information concerning the nature of deep weathering and tor formation in the British Isles. They also make an interesting lithological contrast, being developed not on the Millstone Grit of the Pennines, but on limestones.

Description

The Bridestones are a series of isolated tors approximately 4 km north-east of the village of Levisham on the south edge of the North Yorks Moors. The tors are developed in limestones of Corallian (Upper Jurassic) age, assigned to the Passage Beds by Fox-Strangways (1892). These gently dipping rocks form conspicuous scarp slopes and this region is commonly referred to as the 'Tabular Hills'. Two lines of tors are present, known as the 'Low Bridestones' and the 'High Bridestones' (Palmer, 1956). The northern portion of both groups consists of a bare scarp face parallel to the valley axis (Figure 7.16). Joints in this rock face have commonly weathered out, leaving isolated rock masses standing as tors. These open joints reach depths of up to 2 m in places. Palmer (1956) observed the seepage of water and moisture along joints and cavities in the bedrock. Seepage was particularly pronounced at joint–bedding plane intersections and at the junction of the bare rock with the surrounding hill-slope. No deeply weathered or decomposed rock is observed to fill the joints today.

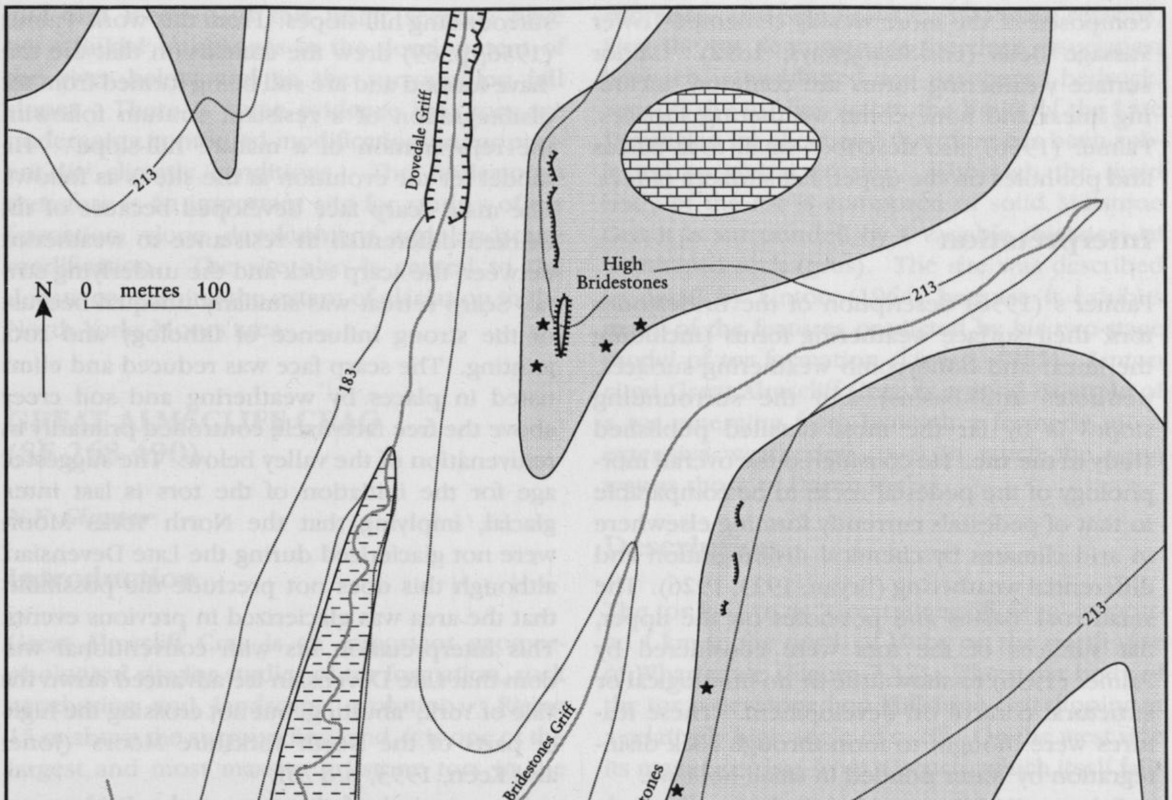

Figure 7.16 Map of the Bridestones showing relief and location of tors (after Palmer, 1956).

A wide variety of tor morphologies are encountered at the Bridestones, including both edge tors and isolated pedestal rocks. The isolated pedestal rocks are commonly described as 'mushroom-shaped' (Palmer, 1956) because they have large undercut bases. In many cases, the summits of these tors consist of the Upper Passage Beds, whereas the undercut bases are composed of the more weakly cemented Lower Passage Beds (Fox-Strangways, 1892). Minor surface weathering forms are common, including linear and honeycomb weathering surfaces. Palmer (1956) also describes small rock basins and pot-holes on the upper, flat surfaces of tors.

Interpretation

Palmer's (1956) description of the Bridestones tors, their surface weathering forms (including the linear and honeycomb weathering surfaces, pot-holes and basins) and the surrounding slopes is by far the most detailed published study of the site. He considered the overall morphology of the pedestal rocks to be comparable to that of pedestals currently forming elsewhere in arid climates by chemical disintegration and differential weathering (Bryan, 1923, 1926). The small rock basins and pot-holes on the upper, flat surfaces of the tors were considered by Palmer (1956) to show little or no lithological or structural control on development. These features were thought to form through rock disintegration by water ponded in small hollows.

Palmer (1956) considered that collectively these forms point unequivocally to rock disintegration along structural and lithological zones of weakness, with the assistance of water emerging through the rock itself. This led him (Palmer, 1956, p. 61) to conclude that 'the intimate association between all the irregularities in the form of the tors and the scarps on one hand, and the structure of the rock on the other, shows clearly that the main factor in the formation of these irregularities is the availability of water'. He invoked a complex of processes involving solution of the cemented bedrock, freeze–thaw and hydration by carbonic and humic acids to explain the physical and chemical processes responsible for rock disintegration. He considered that the processes of weathering are still active under present climatic conditions and that they depend largely on the amount of water reaching the exposed rock surfaces. Palmer (1956) concluded that deep weathering, wind abrasion or solifluction have not been significant processes in the formation of the tors.

Overall, Palmer considered that the tors of the Bridestones evolved in response to successive stages of valley-side incision and scarp-edge retreat. These stages were related primarily to fluvial dissection of the valley below and to the consequent changes in water availability on the surrounding hill-slopes. From this work, Palmer (1956, p. 69) drew the conclusion that the tors 'have formed and are still being formed from the disintegration of a resistant stratum following the rejuvenation of a mature hill-slope'. His model for tor evolution at the site is as follows. The main scarp face developed because of the marked differential in resistance to weathering between the scarp rock and the underlying strata. Scarp retreat was similarly unequal because of the strong influence of lithology and rock jointing. The scarp face was reduced and eliminated in places by weathering and soil creep above the free face itself, controlled primarily by rejuvenation of the valley below. The suggested age for the initiation of the tors is last interglacial, implying that the North Yorks Moors were not glacierized during the Late Devensian, although this does not preclude the possibility that the area was glacierized in previous events. This interpretation fits with conventional wisdom that Late Devensian ice advanced down the Vale of York, 'abutting, but not crossing the higher parts of the North Yorkshire Moors' (Jones and Keen, 1993, p. 178).

The survival of the tors at the Bridestones does not necessarily require that the area escaped glaciation. There is abundant evidence elsewhere that ice-sheet erosion can be selective in its removal of tors and former weathering covers (Fitzpatrick, 1963; Sugden, 1968, 1989; Hall, 1985, 1986a, b, 1991; Hall and Sugden, 1987; Hall and Mellor, 1988; Hall *et al.*, 1989; Ballantyne, 1994; Glasser and Hall, 1997). The survival of old landscape elements such as tors and weathering covers is related primarily to the thermal regime of the ice sheet, with preservation beneath cold-based zones of the ice sheet (Kleman, 1994; Glasser, 1995). Unfortunately, however, because former glaciological conditions in the North York Moors area are largely unknown, it is difficult to assign an age to the tors themselves. As a result, the suggested last interglacial age for their initiation remains conjectural.

Conclusions

The limestone tors at the Bridestones provide a strong contrast with the main group of gritstone Pennine tors. Their detailed morphology and surface weathering forms suggest that water seepage along joints may have played a prominent role in tor development. Changes in the rate and location of this water seepage have been linked to changes in the development of the river below and to the surrounding hill slopes. There is some evidence that tors are undergoing continued modification under present-day climatic conditions. The Bridestones therefore is an important site for studies of tor formation, slope development and landscape modification. The site also is central to the debate concerning the extent of glaciation in the North Yorks Moors area.

GREAT ALMSCLIFF CRAG (SE 268 490)

N.F. Glasser

Introduction

Great Almscliff Crag is an important geomorphological site for studies of tor formation, rock weathering and landscape evolution. Rising 15 m above the surrounding land, it is one of the largest and most massive gritstone tors in the Pennines. It is significant for two main reasons. First the tor demonstrates the close association between unweathered and weathered bedrock. Second the tor lies within the limits of the Late Devensian ice sheet and therefore has been subjected to glacial erosion. Although the main body of the tor is composed of solid Millstone Grit it is surrounded by a variable thickness of weathered rock (grus). The site was described in detail by Linton (1964) because it exhibits many of the features predicted by his two-stage model of tor formation (Linton, 1955). Linton cited Great Almscliff Crag as a good example of a tor emerging from beneath a formerly more extensive weathering cover in much the same way as those of Dartmoor.

Description

The tor known as 'Great Almscliff Crag' is located 1 km to the north of Huby on the north side of Wharfedale (Figure 7.17). The main body of the tor is developed on Millstone Grit dipping at a relatively high angle of *c.* 20°. On the west side its main face rises from a bench, which itself falls

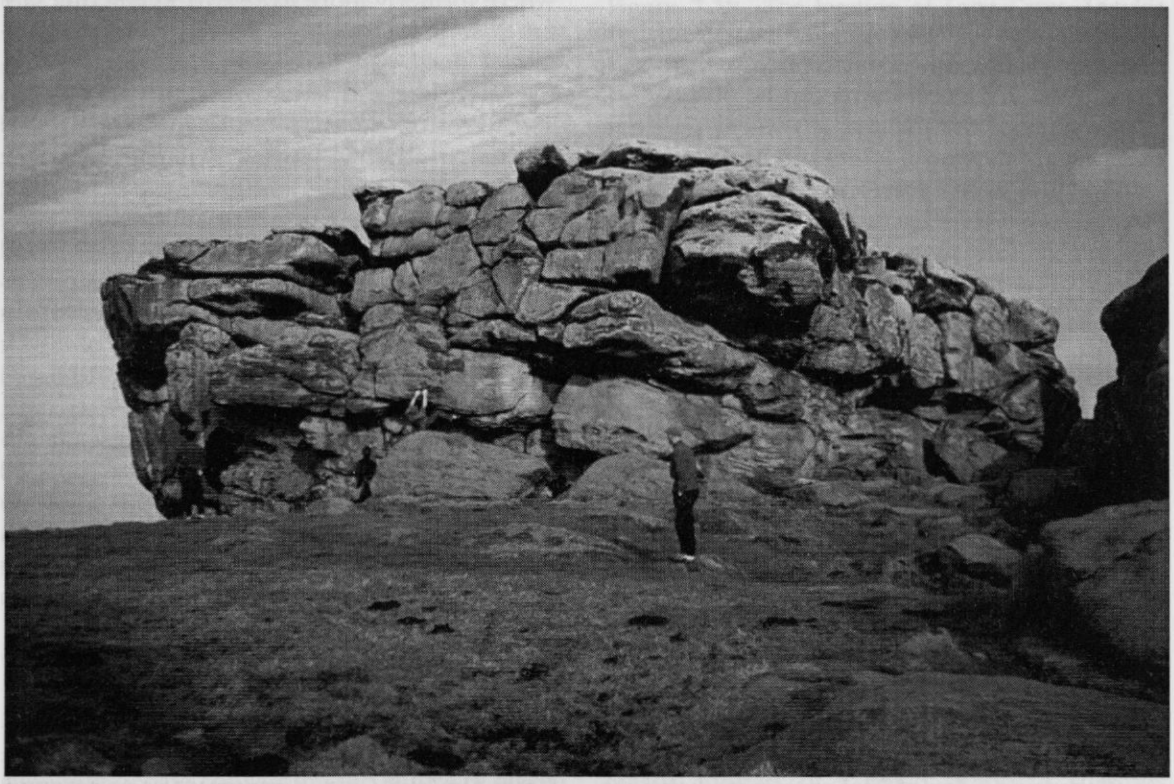

Figure 7.17 Great Almscliff Crag. (Photo: N.F. Glasser.)

away in a second, lower cliff. Linton (1964) describes how the upper part of this cliff provides evidence of sound, unweathered bedrock passing upward into more friable and weathered material. The same transition is observed laterally, with weathered bedrock towards the extremities of the tor giving way inward to solid gritstone. Bands of competent gritstone at low levels in the tor are composed of hard blocks surrounded by incoherent sand. Near the base of the tor the blocks are tessellated within the weathered grit. This is a characteristic of the deep chemical weathering described by Linton (1955), where weathering is best able to penetrate along joints and bedding planes. In places, the sandstone is represented by flags or slabs largely reduced to an ochreous sand with friable particles, resting on a bedding plane of sound rock. Flags more than 15 cm thick are fairly hard but those thinner than this are fragile and can be broken by hand (Linton, 1964). The upper layer of these flags is a limonitic sand, and the sequence is capped by podsolized hillwash (Linton, 1964).

Interpretation

Great Almscliff Crag exhibits evidence of both weathered and unweathered bedrock in close proximity. This weathering can be observed on three sides of the tor and is seen to penetrate deepest along vertical joints and horizontal bedding planes in the gritstone. The nature of this weathering is central to arguments about the formation of tors in the Pennines. In Linton's model of tor formation, Great Almscliff Crag represents a large core stone – or group of core stones – emerging from beneath a formerly more extensive cover of deeply weathered bedrock. This weathering would have taken place under subtropical conditions in pre-Pliocene times. In contrast, Palmer and Radley (1961) argued that no deep chemical weathering existed in the Pennines and that this type of regolith is the result of mechanical rather than chemical weathering under periglacial conditions. Preferential weathering along joints and bedding planes therefore is a result of frost shattering along these lines of relative weakness.

A possible clue as to the origin of this weathering is provided by the work of Wilson (1980) at Blackstone Edge. Wilson examined the surface texture of grains from the Millstone Grit grus at high magnification using a scanning electron microscope (SEM). This work clearly demonstrates that the grus at Blackstone Edge developed in two distinct phases. The first phase of weathering is indicated by chemically produced etch pits and the second by mechanically produced fracture patterns on the quartz grains. There is no evidence for the advanced mineral alteration indicative of prolonged weathering under a pre-Pliocene subtropical climate as Linton (1964) envisaged for the Pennine grus. Instead the weathering is more of the style associated with a periglacial climate. If the grus surrounding the tor at Great Almscliff Crag is of a similar origin to that at Blackstone Edge then there is no need to invoke the deep chemical weathering hypothesis of Linton (1955, 1964).

The survival of large quantities of weathered material around a tor within the Late Devensian ice limit is intriguing. Linton (1964) saw this as an indication that when the tor was glaciated it was 'still rather extensively buried in weathered material' (Linton, 1964, p. 18). Thus Linton regarded glaciation as the agent responsible for the removal of much of the regolith, exposing the central, less weathered corestones. This situation has been documented elsewhere in Britain where formerly more extensive weathering covers existed, such as those developed on the granites of north-east Scotland (Hall, 1985, 1986a, b). Here there is abundant evidence that ice-sheet erosion can be selective in its removal of former weathering covers (Fitzpatrick, 1963; Sugden, 1968, 1989; Hall, 1985, 1986a, b, 1991; Hall and Sugden, 1987; Hall and Mellor, 1988; Hall *et al.*, 1989; Ballantyne, 1994; Glasser and Hall, 1997). The survival of old landscape elements such as tors and weathering covers is related primarily to the thermal regime of the ice sheet, with preservation beneath cold-based zones of the ice sheet (Kleman, 1994; Glasser, 1995). Unfortunately, because no detailed analysis of the nature of the grus in the Great Almscliff Crag area has been undertaken, and as former glaciological conditions in the Wharfedale area are unknown, the survival of the grus remains enigmatic. It also is difficult to assign an age to the period of grus formation.

Conclusions

Great Almscliff Crag is important as an example of a large tor exhibiting evidence of both weathered and unweathered bedrock in close proximity. The weathered grus surrounding the tor may represent either the remnants of a formerly more extensive deep weathering cover that has escaped glacial erosion, or the product of periglacial frost shattering during Pleistocene times. Whichever process dominated, the tor is important for studies of tor formation, rock weathering and landscape evolution.

CHEVIOT TORS (NT 956 215–NT 967 220–NT 943 180)

S. Harrison and N.F. Glasser

Introduction

The Cheviot Hills are a large expanse of mountainous land on the English–Scottish border that support a wide range of glacial, pre-glacial and periglacial landforms, including tors, slope deposits, meltwater channels, and evidence for deep weathering. The Cheviot Tors site includes the tors at Long Crags (NT 956 215), Langlee Crags (NT 967 220) and Great Standrop (NT 943 180) (Figure 7.18). The Cheviot tors, together with the surrounding landforms and sediments provide a detailed picture of landscape evolution in the area. One of the defining characteristics of the geomorphology of the Cheviot Hills is a landform and sediment association comprising tors on interfluves and hillsides, above smoothed valley-bottom slopes that are underlain by mass-wasting deposits of Quaternary age. The tors and slope deposits of the Cheviot Hills are important elements of the landscape because of the information they provide regarding the pattern of Devensian glaciation and deglaciation in the region and the impact of the return to cold conditions during the Younger Dryas.

Descriptions of Quaternary events in the Cheviot Hills are numerous. The tors have featured in a debate concerning the extent to which the Cheviot Hills supported an independent ice cap and the extent to which the mountains were inundated by external ice derived from ice centres farther to the west. Geikie (1876), Clough (1888), Smythe (1912), Raistrick (1931b), Carruthers *et al.* (1932), Common (1954), Sissons (1964), Clark (1970, 1971) and Clapperton (1970, 1971a) have all contributed to this longstanding debate concerning the nature of glacial events in the area. The Cheviot massif also supports an excellent variety of glacial meltwater phenomena, descriptions of which are provided by Kendall and Muff (1903), Common (1953, 1957), Derbyshire (1961) and Clapperton (1966, 1968, 1971a, b), as well as a range of drift deposits comprising material derived from weathered granite. Douglas and Harrison (1985) and Harrison (1994, 1996) have described these deposits in sections at Linhope and Leech Burn. Finally, Clapperton (1967) and Awujoola (1987) have described the nature of the deep weathering on the granite bedrock.

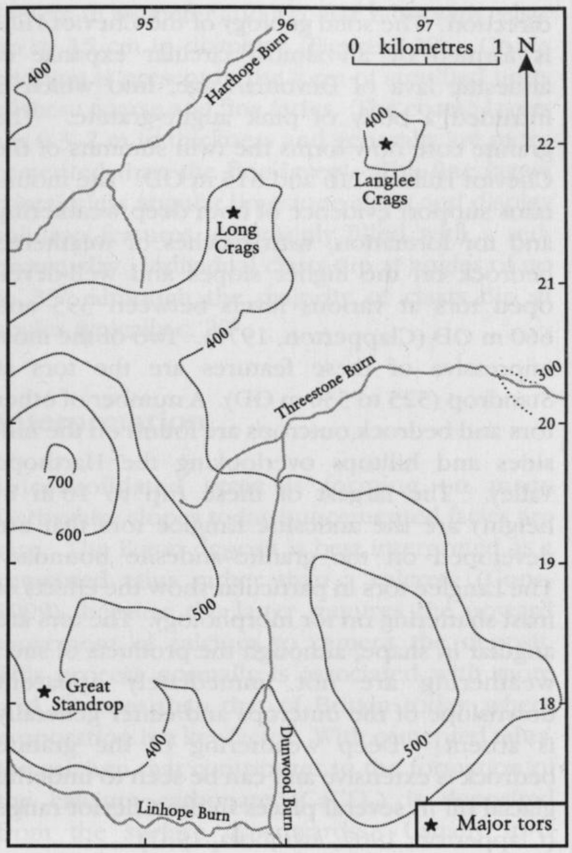

Figure 7.18 The location of Long Crags, Langlee Crags and Great Standrop. Numbers around margins refer to the UK National Grid.

Description

The Cheviot Hills form the southern limb of a 650 km² belt of high ground that girdles the Tweed drainage basin. The main body of the massif consists of a chain of broad summits aligned in a generally south-west to north-east direction. The solid geology of the Cheviot Hills is formed by an almost circular expanse of andesitic lava of Devonian age, into which is intruded a body of pink augite-granite. The granite core now forms the twin summits of the Cheviot Hills at 716 and 815 m OD. The mountains support evidence of both deep weathering and tor formation, with patches of weathered bedrock on the higher slopes and well-developed tors at various levels between 395 and 660 m OD (Clapperton, 1970). Two of the most impressive of these features are the tors at Standrop (525 to 530 m OD). A number of other tors and bedrock outcrops are found on the hillsides and hilltops overlooking the Harthope Valley. The largest of these (up to 10 m in height) are the andesitic Langlee tors that are developed on the granite–andesite boundary. The Langlee tors in particular show the effects of frost-shattering on tor morphology. The tors are angular in shape, although the products of such weathering are not immediately apparent downslope of the outcrops and clitter generally is absent. Deep weathering of the granite bedrock is extensive and can be seen to underlie glacial till in several places in the Cheviot range (Clapperton, 1967; Awujoola, 1987).

Interpretation

The Cheviot tors demonstrate exceptional preservation of tors on the higher summits of the mountains. Linton (1955) argued that tors were unlikely to survive in areas covered by glacier ice because the passage of the ice would remove such features. Clapperton (1970) maintained that the Cheviot tors should not be taken as indicators of an unglaciated area and suggested that such features could survive beneath a relatively thin and slow-flowing local ice cap. Indeed, Clapperton (1970) further argued that had the tors been exposed to severe periglacial conditions in ice-free enclaves surrounded by glacier ice, then their complete removal by mass wasting processes might be expected. Such an interpretation fits well with the geomorphological evidence for the preservation of pre-glacial landscape elements beneath glacier ice in other upland areas of Great Britain, such as the Cairngorms (Sugden, 1968; Gordon, 1993; Ballantyne, 1994). Here the preservation of tors and related phenomena has been linked to the basal thermal regime of the last ice sheet, which was warm-based and erosive in valleys but cold-based and protective over summits (Glasser, 1995). Indeed, the widespread preservation of pre-glacial landscape elements has now been identified beneath cold-based sectors of both the Laurentide ice sheet (Dyke, 1993; Kleman *et al.*, 1994) and the Scandinavian ice sheet (Kleman, 1992, 1994; Kleman *et al.*, 1992; Kleman and Borgström, 1994).

Conclusions

In England, the Cheviot Hills contain an unrivalled range of glacial, pre-glacial and periglacial landforms. The landform and sediment assemblage in the mountains includes tors, glacial deposits, reworked glacial deposits, meltwater channels, and evidence for deep weathering. The long-standing controversy over the vertical and lateral extent of glaciation in the area is important for our understanding of the dynamics and thermal regime of the last ice sheet in this area. The Standrop and Langlee tors and the associated slope deposits at Linhope and Leech Burn provide important information about the sequence and nature of geomorphological events in the Cheviot Hills following glaciation.

ECTON (SK 097 581)

N.F. Glasser and C.V. Burek

Introduction

Ecton Quarry, in Derbyshire, provides a rare exposure in Great Britain of a cemented limestone talus of presumed Late Devensian age. Talus slopes are steep valley-side slopes formed by the accumulation of debris at the foot of a rockwall (Ballantyne and Harris, 1994) and are a common geomorphological component of the Peak District landscape. The term 'talus' is used to describe both the form of the slope and its constituent material. Although the term 'scree' is often used as a synonym for talus, in modern usage this term is reserved for a slope cover of predominantly coarse debris, irrespective of location. Talus slopes are not restricted to present or former periglacial environments but occur in all areas where the products of rock weathering can accumulate at the foot of a cliff or slope. The talus at Ecton comprises stratified layers of coarse angular debris and finer, more weathered material, which, unusually for Great Britain, has been cemented by deposition of calcite in the voids between individual particles.

As described in 'Talus slopes and screes', in the introduction of this chapter, talus slopes commonly take one of three forms: talus sheets, talus cones and coalescing talus cones. Both active and relict talus slopes are found in Britain. Relict talus slopes are found in many areas of Britain subjected to periglacial conditions, for example at Gilman Point in the Marros Sands GCR site in south-west Wales (Campbell and Bowen, 1989). These periglacial conditions persisted during the Late Devensian in areas beyond the limit of the ice maximum and relict talus slopes therefore are important palaeoenvironmental indicators. Ecton Quarry is particularly important because the talus exposed there is of a cemented nature. This type of talus is unusual in Britain and is confined to limestone areas such as the Peak District. Burek (1977) has described the site in detail.

Description

Ecton Quarry is a small disused quarry on the east side of the Manifold Valley in the Peak District of Derbyshire. Lower Carboniferous (Dinantian) limestone dominates the solid geology of this area (Aitkenhead *et al.*, 1985). Ecton Quarry provides a large exposure of the Ecton Limestone (Prentice, 1951, 1952), a predominantly grey or brownish grey to dark grey, thinly bedded, bioclastic limestone. Cemented talus occurs throughout the quarry, although it is particularly evident in the upper layers. Here, the cemented talus forms a more resistant layer and therefore protrudes from the back wall of the quarry. Recent weathering has dislodged many blocks of cemented talus from the quarry walls and these now form loose blocks on the quarry floor. The maximum recorded depth of the cemented talus in the quarry is 24 m. Above the quarry walls on the modern land surface is a layer of silty drift (Burek, 1985, 1991) and a thin soil. The silty drift is a mixture of insoluble residue of limestone and loess (Piggott, 1962).

The talus consists of an accumulation of angular limestone and chert clasts held in a cemented matrix. Two facies are identified; a fine-grained talus with clasts normally in the range of 1–5 cm in diameter and a coarse talus with clasts up to 15 cm in diameter (Figure 7.19). Crude bedding is present in the form of stratified units of these coarse and fine facies. The coarse layers are 0.3–2 m in thickness and generally are more cemented than the fine layers. The fine layers contain less angular limestone clasts and display solution features, commonly filled with a red-brown clay. Individual clasts dip at angles of up to 35°, although the majority of clasts dip at angles typically *c.* 20°.

Interpretation

Unconsolidated scree is forming on many Derbyshire slopes today but cemented facies are rare. The Ecton deposit is best interpreted as a cemented talus rather than a calcrete (Cope, 1999), because the latter requires the upward movement of calcium to cement the deposit. This process normally is associated with more arid climates than that of Britain today, where evaporation is a key factor. With cemented talus, the calcium that contributes to the formation of the calcium carbonate ($CaCO_3$) is deposited from the surface downwards. Usually it is derived from solutions originating as rainwater percolating through the more finely divided and powdery scree components and dissolving minor amounts of calcium in the process. In coastal locations this cement is often influenced by sea spray and has been specifically identified as brucite ($Mg(OH)_2$), hydromagnesite ($MgCO_3OH$) and minor amounts of aragonite ($CaCO_3$) (Flinn and Pentecost, 1995). However the cement matrix at Ecton is a low-magnesium calcite.

The silty drift and soil that overlie the quarry provide information on the age of the cemented talus. These deposits are often preserved down fractures as well as on the surface (Burek, 1985, 1991). This commonly is taken to indicate that the area was not glacierized during the Late Devensian because formation of the silty drift requires both a period of chemical weathering and periglacial activity. The silty drift therefore is not present within the limits of the last glacial maximum. The loess contributes quartz and the silty drift contributes silica and clay minerals to the mineralogy available for redistribution down the profile. There is also a component of calcite present (Burek, 1991). This is then available for cementation purposes at a later date. There are other examples of cemented scree in the

Figure 7.19 Cemented talus at Ecton Quarry showing the finer-grained of the two facies. Note the angular nature of the clasts. (Photo: C.V. Burek.)

White Peak, such as those at SK 137 618 and SK 137 611, as well as partly cemented scree slopes in Lathkill Dale (Ford, 1964; Burek, 1977). Each of these has a thin layer of silty drift that has been soliflucted from above on to the talus deposit. This is important because it provides a relative age for the cementation of the talus.

Conclusions

The cemented talus accumulated under environmental conditions markedly different to those of today. Frost action was probably more important than either ice or snow cover in the formation of these deposits and limited runoff is suggested in order to allow the scree to form without contemporaneous removal. Subsequently, annual rainfall was higher than it is today, but there was neither continuous permafrost nor frozen groundwater within the top layers of material, as this would not allow percolation and subsequent deposition of calcium carbonate as a cementing agent in the voids. Discontinuous permafrost would allow the scree to accumulate and subsequent higher temperatures would encourage the calcite to precipitate and cement the deposit. A Late Devensian age is suggested for the cemented talus at Ecton because this is the last time that these environmental conditions were fulfilled.

THROSTLE SHAW (NY 237 272)

J. Boardman

Introduction

At many sites in the northern Lake District considerable thicknesses of scree, or vegetated scree, overlie glacial till, or bedrock (Boardman, 1978) (Figure 7.20). These are far better developed on the Skiddaw Group of rocks than on the Borrowdale Volcanic Group owing to the relative frost-susceptibility of the former. At high-altitude sites in the Skiddaw–Blencathra massif

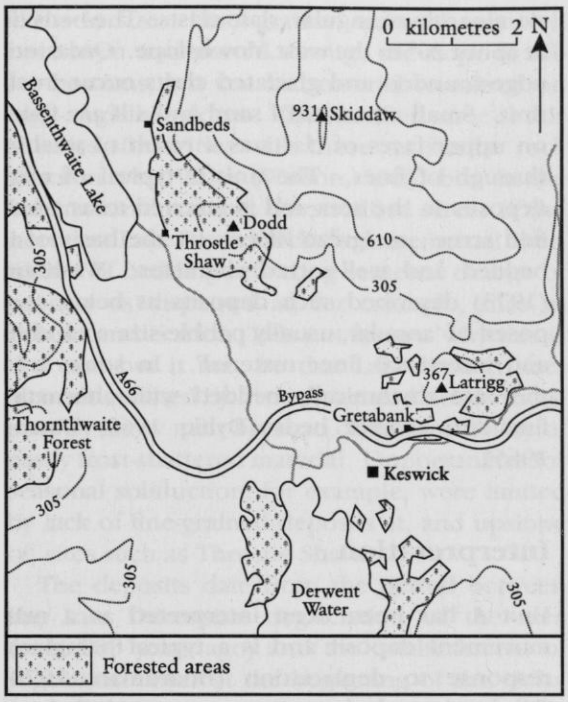

Figure 7.20 Periglacial slope deposits in the Keswick area: location of sites (after Boardman, 1985b).

(> 500 m) low-angle scree slopes are active, whereas at lower altitudes screes are vegetated and inactive under present climatic conditions. At many sites, vegetated scree overlies till of Late Devensian age and the relationship implies that scree formation post-dates deglaciation. As it is assumed that early Holocene warming and vegetation colonization would have led to cessation of scree accumulation at low-altitude sites, the screes therefore are regarded as having formed under periglacial conditions immediately post deglaciation and during the Loch Lomond Stadial climatic deterioration.

Description

Throstle Shaw (NY 237 272) is a roadside exposure, at an altitude of 100 m, at the foot of the south-west facing slope of Dodd (502 m). The deposits at Throstle Shaw are divisible into two distinct units (Figure 7.21).

Unit A is a very poorly sorted, silty and sandy diamicton with gravel clasts. The clast population is mainly angular slates but there are low

Figure 7.21 Throstle Shaw section. (Photo: J. Boardman.)

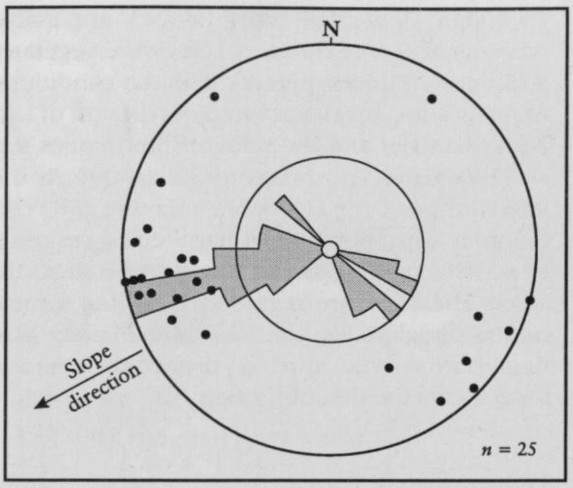

Figure 7.22 Macrofabric analysis of elongate clasts in the debris flow deposit at Throstle Shaw, Tii on Figure 7.23 (after Boardman, 1985b). *n*: sample size.

numbers of edge-rounded slates, glaciated clasts and non-slate (erratic) clasts. There is no bedding and little obvious sign of preferred orientation. However, macrofabric analysis of the larger clasts show movement to the west, suggesting downslope movement (Figure 7.22). This contrasts with a northerly direction of movement for the local tills. Stone shape is variable but there is a lack of well-developed blades and discs and a significant difference in shape compared with the overlying unit. The unit is regarded as a debris flow that contains scree, fluvial and glacial material remobilized on the slopes of Dodd.

Unit B (Figure 7.23) is a relatively well-sorted and bedded openwork gravel composed predominantly of angular, slate clasts. The beds dip at about 20° to the west, downslope. Occasional edge-rounded and glaciated clasts occur in the unit. Small amounts of sand and silt are found on upper faces of clasts as a result of washing through of fines. The unit is typical of many deposits in the area and is referred to as 'stratified scree', or 'grèze litées', on the basis of its bedded and well-sorted character. Washburn (1973) described such deposits as being composed of 'angular, usually pebble-size rock chips and interstitial finer material'. In some areas they are rhythmically bedded with alternating fine and coarser beds (Dylik, 1960; Watson, 1965).

Interpretation

Unit A has been been interpreted as a mass movement deposit and is a typical paraglacial response to deglaciation (Boardman, 1978). Till-dominated deposits on steep hill-slopes were likely to have been unstable immediately after deglaciation and downslope movement would have been triggered rapidly by snow melt in a series of catastrophic debris flows. Unit B, on the other hand, is evidence for a long period of frost action on the Skiddaw Slate hill-slopes and the consequent transport of rock chip debris to footslopes owing to the redistribution of rockfall scree by snow melt, perhaps over a permafrost table (Howarth and Bones, 1972). Distance of transport was very limited as clasts are rarely edge-rounded and during transport frost action would have continued to reduce

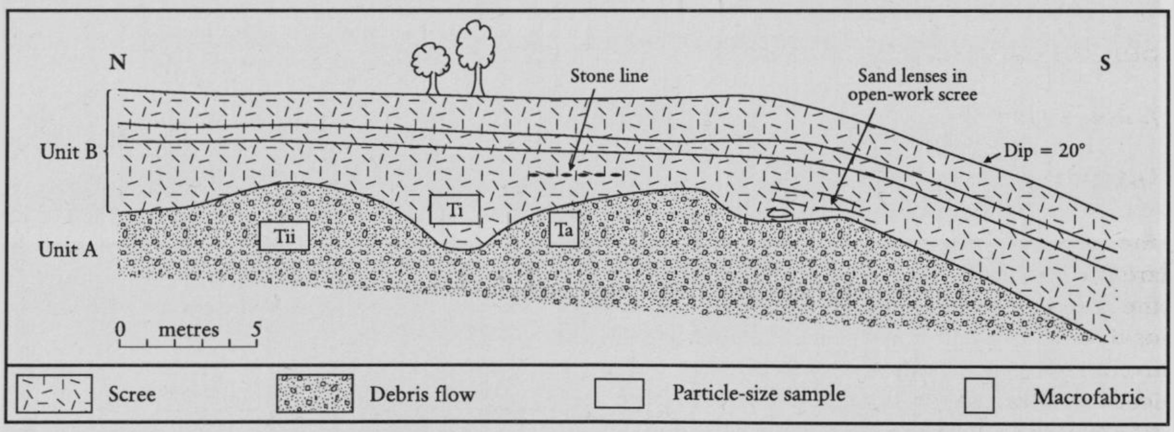

Figure 7.23 Unit B, Throstle Shaw section (after Boardman, 1985b).

clasts in size. Finer-grained material may well have been transported through the slope to valley bottoms, but coarser material remained on high-angle hill-slopes. The stratified scree at this site has similarities with those on similar rock types in west Wales (Watson, 1965) but, in the international context, the stratified screes of the northern Lake District lack the incremental alternation of bedding reflecting seasonal change of process as displayed at, for example, sites in Poland and France. It is not known why this is the case but it may be that Lake District sites were dominated by frost action and debris transport by snow melt because of the availability of easily frost-shattered material. Opportunities for seasonal solifluction, for example, were limited by lack of fine-grained deposits at, and upslope of, sites such as Throstle Shaw.

The deposits date from the period between the Late Devensian deglaciation of the area (*c.* 16 000 years ago) and the establishment of vegetation cover at the beginning of the Holocene (10 000 years ago). It is tempting to place them wholly within the Loch Lomond Stadial, for which there is an abundance of evidence of periglacial climate in northern Britain and associated frost action, snow melt and fluvial activity, but they also may have been active in the period 16 000 to 13 000 years ago.

Conclusions

Throstle Shaw provides evidence of the impact of a periglacial climate on the landscape of the northern Lake District in a period when mass movement, frost action and snow melt were potent processes in a largely devegetated landscape.

SANDBEDS FAN (NY 234 291)

J. Boardman

Introduction

Fans are sites of sediment storage. Alluvial fans are an important geomorphological feature in the Lake District landscape. They represent deposition by fluvial and mass-movement processes from hill-slope sources. They generally are located at breaks of slope, or valley widening sites where channel gradient decreases, or width increases. It is reasonable to assume that alluvial fans in the Lake District post-date the Late Devensian glaciation and potentially contain sediments accumulated over a period of ± 16 000 years. However, sediment transport is greatly enhanced under bare ground conditions and it is assumed that major growth of fans will occur in periods of periglacial climate.

The Sandbeds Fan (NY 234 291) is situated at the point where a small stream, Sandbeds Gill, debouches on to the valley bottom. The fan is at the break of slope between a steep Skiddaw Group slope and till-covered lowland, adjacent to Bassenthwaite Lake. A small quarry provides excellent exposures in the fan. The site is described by Boardman (1985b).

Description

The morphometry of the Sandbeds Fan and its catchment is shown on Figure 7.24, and exposure in the west-facing wall of the quarry is shown in Figure 7.25. The surface of the fan is totally vegetated with grass, heather and coniferous trees, and the present-day stream is confined to a deep gorge in the northern part and does not contribute sediment to the fan surface. At the base of the quarry, diamicton is exposed and macrofabric analysis shows a strong east–west orientation, suggesting post-depositional solifluction in an area where Late Devensian ice was moving to the north. At one site in the quarry, diamicton also appears to overlie fan gravels, implying solifluction (debris flow activity) during early stages of fan formation.

The sediments exposed in the quarry are predominantly gravels, which vary from fine- to very coarse-grained and discrete beds of sand are

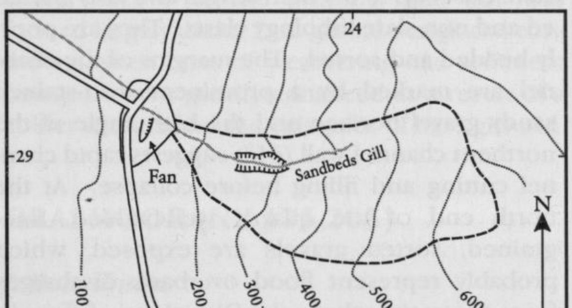

Figure 7.24 Sandbeds Fan and catchment (after Boardman, 1985b).

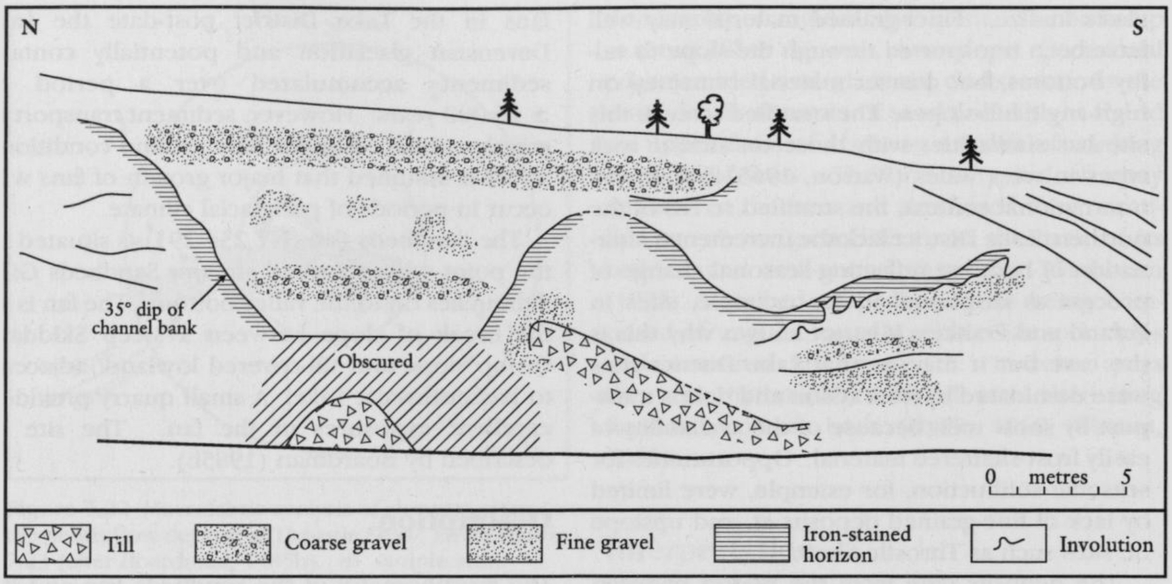

Figure 7.25 Sandbeds Fan: west-facing section in quarry (after Boardman, 1985b).

rare. The degree of sorting is variable and the amount of interstitial material in the gravels, principally sand, also is variable. A feature of many beds are the silt caps on gravel clasts. The fact that some beds lack silt caps, and underlying beds possess them, implies that the caps result from the washing in of silt during fan aggradation. All gravel units in the fan contain high proportions of edge-rounded clasts in contrast to scree deposits in the same lithology, which are predominantly angular.

In the main face of the quarry, gravels can be seen to infill a channel of about 11 m width and 7 m depth (Figure 7.25). Units of coarse blocky gravels are seen at the base and top of the channel. The coarse gravels contain occasional large slate clasts up to 0.5 m in length and also glaciated and non-slate lithology clasts. They are poorly bedded and sorted. The margins of the channel are marked by a prominent iron-stained sandy gravel horizon and the high angle of the northern channel wall (35°) suggests rapid channel cutting and filling before collapse. At the north end of the quarry, well-bedded, fine-grained, sorted gravels are exposed, which probably represent flood overbank discharges from the main channel. Directions of flow in the fan gravels are indicated by the rose diagrams in Figure 7.26 and show flow divergence from the apex of the fan.

At the southern end of the main quarry section a prominent iron-stained sandy gravel horizon occurs, which may represent the margins of a shallow channel. Below this horizon are loose, openwork gravels. The boundary between the two units is involuted, the size of the involutions being almost 1 m. These features represent disturbance within an active layer under periglacial conditions and as such imply the existence of a temporarily stable land surface on part of the fan. The marked contrast in texture between the two units, and the availability of water on the fan surface, would no doubt encourage the formation of involutions (Johnson, 1975). Burial of the involuted horizon occurred owing to flooding from the main channel, or a change in the location of channels on the fan. Involutions have not previously been reported from gravels or scree sediments in the north-eastern Lake District, although similar forms are described in the southern Lake District (Johnson, 1975).

Interpretation

The character of the fan gravels suggests that they originated largely as a result of frost shattering but were subsequently transported by fluvial processes. Present-day fluvial activity within the catchment is confined to the channel of Sandbeds Gill. Under periglacial conditions, however, drainage densities may have been high in response to seasonal snow-melt discharges.

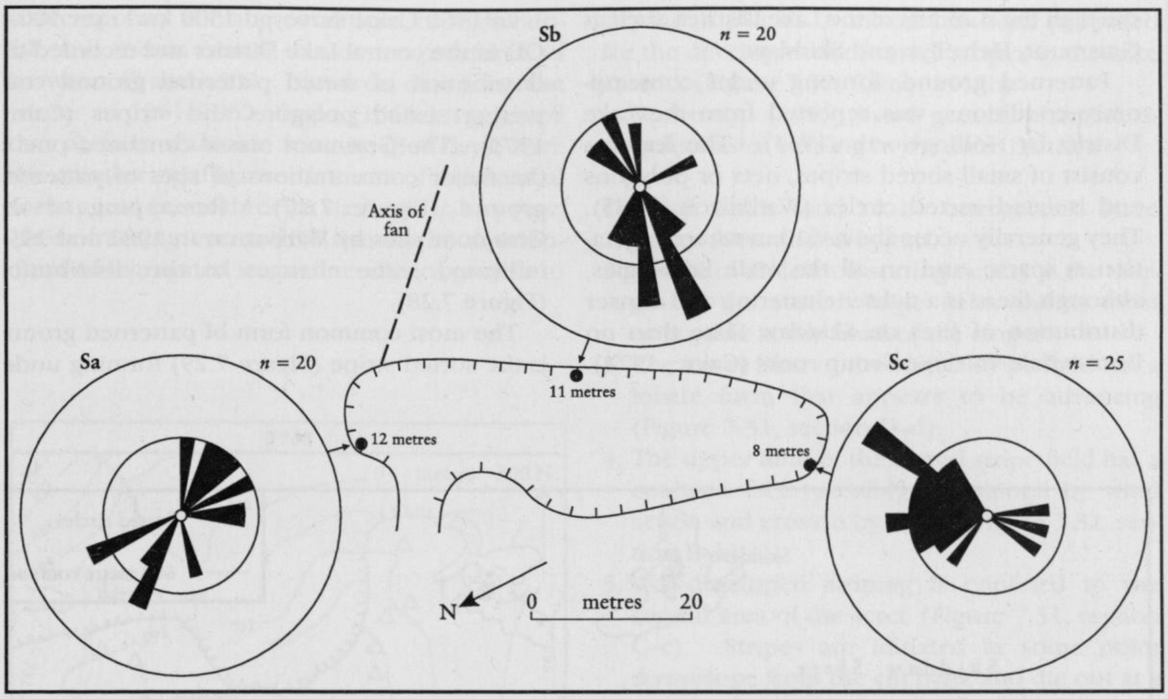

Figure 7.26 Sandbeds Fan: plan of quarry and orientation of elongate gravel clasts; *n* = sample size (after Boardman, 1985b).

Overland flow on slopes underlain by permafrost also may have been an effective means by which weathered debris was moved to channels. Rapid rates of weathering and transport would have been favoured by the absence of vegetation.

The Quaternary history of the area implies that the fan has formed since the Late Devensian glaciation and became inactive after the establishment of forest cover on the slopes at the beginning of the Holocene Epoch. The volume of the fan (150 000 m^3) represents a mean surface lowering within the catchment of 0.5 m, based on the assumption that the fan formed during the *c.* 1000 years of the Loch Lomond Stadial. This is roughly an order of magnitude higher than present-day yields from steep mountain streams in the Lake District (Newson and Leeks, 1985). The evidence for a single period of rapid aggradation is equivocal, however. The steep sides of the infilled channel imply rapid cut and fill, whereas the involuted horizon suggests a significant intervening ground-surface stability during which permafrost existed. Moreover, a well-developed soil profile would be expected beneath the fan surface if the fan had been inactive since the Loch Lomond Stadial. Soil development is, in fact, very limited, suggesting that the upper part of the fan, above the iron-stained, involuted horizon may well be of late Holocene age.

Conclusions

The Sandbeds Fan site is an excellent example of an alluvial fan. Such landforms are areas of sediment storage and as such offer the opportunity for reconstruction of environmental history during the period of accumulation. The fan is relatively inactive at the present time but was formed largely under Late-glacial periglacial conditions when rates of weathering and sediment transport were greatly enhanced as a result of a colder climate and the existence of bare ground.

GRASMOOR (NY 175 205)

J. Boardman

Introduction

Patterned ground of the type generally associated with periglacial conditions can be found on the high flat summits of the Lake District, such as Grasmoor, Helvellyn and Skiddaw.

Patterned ground, forming under contemporary conditions, was reported from the Lake District by Hollingworth (1934). The features consist of small sorted stripes, nets or polygons and isolated sorted circles (Warburton, 1985). They generally occur above 610 m where vegetation is sparse, and on all the main lithologies, although there is a tighter clustering and denser distribution of sites on Skiddaw Slate than on Borrowdale Volcanic Group rocks (Caine, 1972).

In 1960 Caine surveyed 1500 km^2 over 500 m OD in the central Lake District and recorded the distribution of sorted patterned ground comprising sorted polygons and stripes (Caine, 1972). The Grasmoor massif contained one of the major concentrations of sites of patterned ground (Figure 7.27). Remapping of the Grasmoor sites by Warburton in 1982 and 1994 indicated some changes in the distribution (Figure 7.28).

The most common form of patterned ground is the sorted stripe (Figure 7.29) forming under

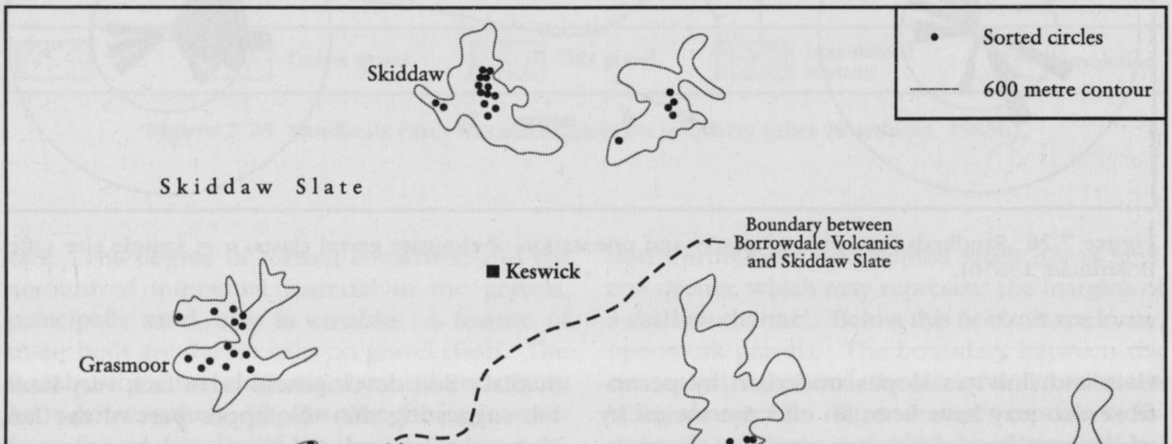

Figure 7.27 The spatial distribution of sorted patterns in the Lake District. Contour is at 600 m. The broken line shows boundary between Borrowdale Volcanics (to the south-east) and Skiddaw Slate (to the north-west). (After Caine, 1972.)

contemporary climatic conditions of high rainfall and mean annual winter temperatures of between 4° and –3°C. Based on data from the winter of 1961–1962, Caine (1963a) suggested that periods of intense freezing (3–4 days) have the greatest significance in terms of mechanical sorting and occur three to six times in any year. These freezing events give rise to the development of segregated ice and needle ice growth to depths of 30 cm. A model has been proposed for the development of sorted stripes in the Lake District (Figure 7.30; Caine, 1963b).

Warburton (1985) listed general characteristics of areas of striping in the Lake District:

1. Stripe fields tend to be unvegetated scree slopes surrounded by turf.
2. A zone of ill-defined sorting occurs at the margin of the sorted stripe field; this is most noticeable at the head of the slope.
3. The lower limit of the sorted stripe field has a lobate form that appears to be advancing (Figure 7.31, section D–d).
4. The upper limit of the sorted stripe field has a cut-bank form possibly maintained by wind action and erosion by sheep (Figure 7.31, section B–b).
5. Well-developed striping is confined to the central area of the scree (Figure 7.31, section C–c). Stripes are initiated at some point downslope from the cut bank and die out at a position just behind the frontal lobe, at which point crudely developed sorted nets may be evident. This development may be the result of locally reduced slope gradient.
6. Striping does not normally extend beyond the area of the scree (Figure 7.31), although turf remnants are sometimes observed along fine-stripe crests.
7. Sorted horizons comprise a coarse-stripe gutter with many clasts orientated vertically and some downslope and fine-stripe ridges with few clasts orientated vertically.
8. The basal layer is relatively undeformed and unsorted.
9. Bedrock is usually at a depth of 60–100 cm.

The Grasmoor sites are important in that they were the focus of work done in the 1960s by Caine (1963a, b, 1972). They have been revisited and remapped more recently by Warburton (1985, 1997) and therefore form the basis of some conclusions about changing areal distribution of patterned ground. Early work on patterned ground tended to be descriptive (e.g. Caine, 1972), although a model for patterned ground formation was also proposed (Caine, 1963b). Recently, Warburton, as well as remapping, has begun detailed recording of ground temperatures on High Pike (Warburton, 1997). The focus therefore has shifted to an attempt to understand the processes of patterned ground formation.

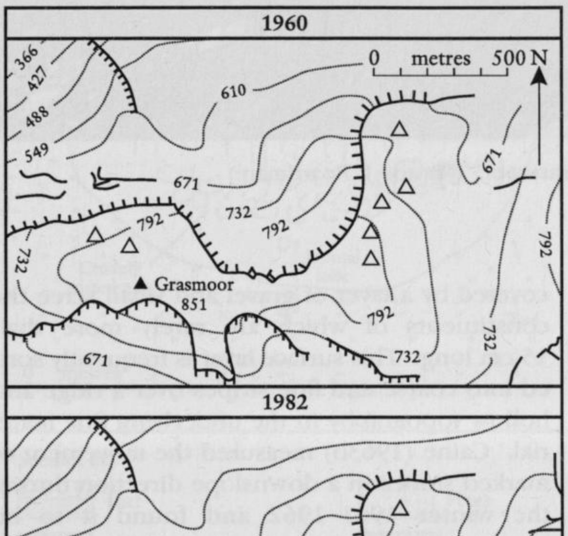

Figure 7.28 Sequential mapping of sorted stripes sites on Grasmoor between 1960 and 1994 (after Warburton, 1997).

Figure 7.29 Stone stripes on Grasmoor. (Photo: J. Boardman.)

Description

Caine (1963b) described the low-angle scree slopes on the summit area of Grasmoor as consisting of 'a generally fine soil up to 60 cm thick covered by a layer of gravel and small scree the constituents of which are rarely more than 15 cm long. This surface layer is frequently sorted into coarse and fine stripes over a ridge and hollow topography in the underlying fine material.' Caine (1963b) measured the movement of marked stones in a downslope direction during the winter 1961–1962 and found it to be between 15 and 20 cm on unvegetated slopes of about 15°. Thus the areas of patterned ground are active under contemporary conditions, with frost heave and downslope movement of clasts occurring.

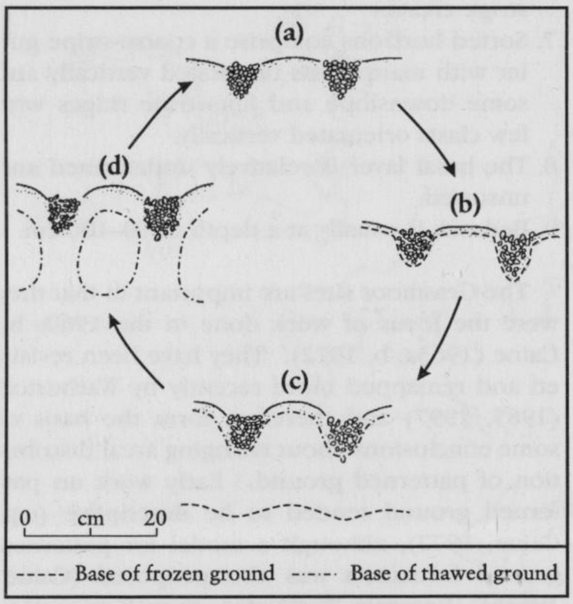

Figure 7.30 Model for the development of stone stripes (after Caine, 1963b). (a) completely thawed; (b) freezing extends downwards faster under coarse stripes than under fine; (c) thawing occurs faster under coarse stripes than under fine; (d) frozen areas remain only under fine ridges.

Warburton (1985) listed seven sites on Grasmoor where active patterned ground existed in 1982, which are grouped in an area to the north and one to the east of the summit (852 m) (Figure 7.28). The sites on Grasmoor are developed on a subdivision of the Skiddaw Group, the Mosser-Kirkstile Slates. Warburton (1985) discussed the characteristics of the Grasmoor sites in association with the lithologically similar sites on Skiddaw, and in contrast to the Helvellyn sites developed on Borrowdale Volcanic Group lithologies. Sorted stripes on Grasmoor and Skiddaw tend to be more extensive than on the Borrowdale Volcanic rocks and occur in a series of stepped benches separated by lobe fronts or turf bands. Stripe widths and depths tend to be less on Grasmoor than those at Skiddaw and Helvellyn sites. Silt and clay contents are similar to the Helvellyn sites. There is

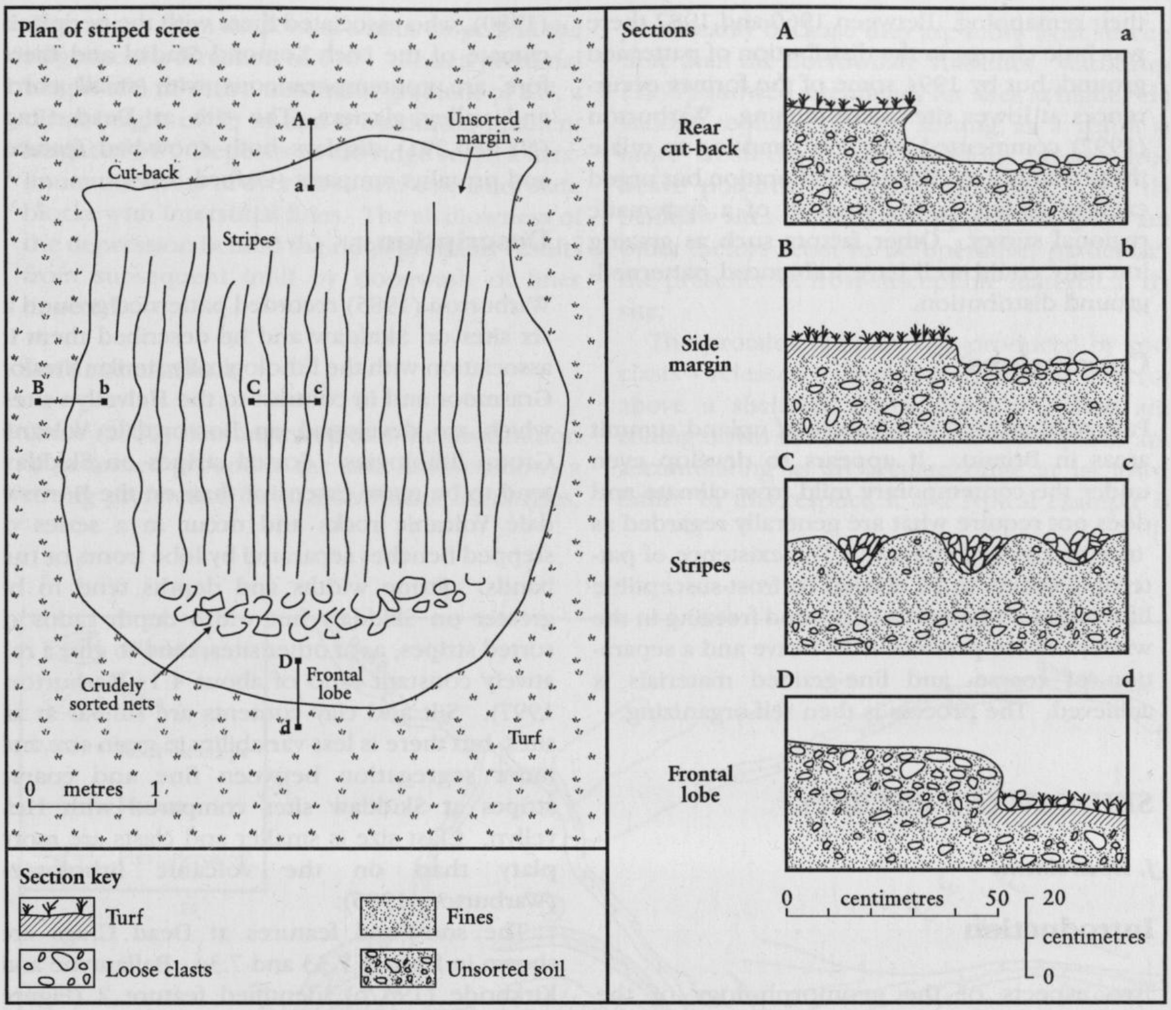

Figure 7.31 Morphology of a stone stripe field (after Warburton, 1985).

less variability in grain size and more segregation between fine and coarse stripes at Grasmoor and Skiddaw sites compared with Helvellyn. Clast size is smaller and clasts are more platy than on the volcanic lithologies (Warburton, 1985). Width–depth ratios of sorted stripes at Grasmoor, as at other sites, tend to give a relatively constant ratio of about 4:1 (Warburton, 1997).

Interpretation

Caine (1972) shows that the distribution of patterned ground in the Lake District has a strong preference for Skiddaw Slate lithologies, presumably because they are more frost susceptible than the Borrowdale Volcanics. Warburton (1997) shows a tendency for sites at higher elevations to have deeper sorting. He hypothesizes that this is because deeper sorting might be expected at higher altitudes as a result of more frost days and therefore greater frost-heave potential. However, this simple relationship also is affected by the presence of frost-susceptible material at the site. The Grasmoor sites are at relatively low altitude and show low sorting depths.

Contemporary activity of the patterned-ground area of Grasmoor was demonstrated by Caine (1963b). At other sites, formation of stripes on mining waste, and stratigraphical relations between unbroken turf cover, patterned and unpatterned soils also attest to contemporary activity (Warburton, 1997). The Grasmoor sites are particularly important as a result of their remapping. Between 1960 and 1982 there was little change in the distribution of patterned ground, but by 1994 some of the former occurrences at lower sites were missing. Warburton (1997) commented that it is tempting to relate these changes to climatic amelioration but urged caution owing to the absence of a systematic regional survey. Other factors such as grazing intensity could well have influenced patterned-ground distribution.

Conclusions

Patterned ground is a feature of upland summit areas in Britain. It appears to develop even under the contemporary mild frost climate and does not require what are generally regarded as 'true' periglacial climates. The existence of patterning seems to be related to frost-susceptible lithologies and periods of ground freezing in the winter. These produce frost heave and a separation of coarse- and fine-grained materials is achieved. The process is then self-organizing.

SKIDDAW (NY 261 287)

J. Boardman

Introduction

Two aspects of the geomorphology of the Skiddaw massif are of interest, in the context of both periglacial processes and landforms. These are the existence of currently active patterned ground, and the evidence for a more severe periglacial climate in the past in the form of snow-bed features and protalus ramparts.

Patterned ground of the type generally associated with current or past periglacial conditions can be found on the high flat summits of Skiddaw, and Caine's (1972) survey of the central Lake District recorded the distribution of sorted, patterned ground comprising polygons and stripes. Snow-bed features and protalus ramparts are of importance because they point to the former existence of perennial snow beds, and therefore a climatic regime cooler and/or wetter than the present. 'Protalus ramparts' are curving ridges of coarse debris that accumulate at the foot of steep cliffs by sliding and rolling across a snow patch (Ballantyne and Kirkbride, 1987b) (Figure 7.32). Snow-bed features were described from the Lake District by Sissons (1980), who associated them with the periglacial climate of the Loch Lomond Stadial and therefore are contemporaneous with small corrie and valley glaciers. The site at Dead Crags (NY 260 291) displays both snow-bed features and protalus ramparts (Oxford, 1994).

Description

Warburton (1985) recorded patterned ground at six sites on Skiddaw and he described them in association with the lithologically similar sites on Grasmoor and in contrast to the Helvellyn sites, which are developed on Borrowdale Volcanic Group lithologies. Sorted stripes on Skiddaw tend to be more extensive than on the Borrowdale Volcanic rocks and occur in a series of stepped benches separated by lobe fronts or turf bands. Stripe widths and depths tend to be greater on Skiddaw but width–depth ratios of sorted stripes, as at other sites, tend to give a relatively constant ratio of about 4:1 (Warburton, 1997). Silt and clay contents are similar at all sites, but there is less variability in grain size and more segregation between fine and coarse stripes at Skiddaw sites compared with Helvellyn. Clast size is smaller and clasts are more platy than on the volcanic lithologies (Warburton, 1985).

The snow-bed features at Dead Crags are shown in Figures 7.33 and 7.34. Ballantyne and Kirkbride (1987b) identified feature 2 (Figure 7.33) as a protalus rampart and describe it as an arcuate ridge, 300 m long by 40 m wide, at an altitude of 295 m. The ridge is located 25 m

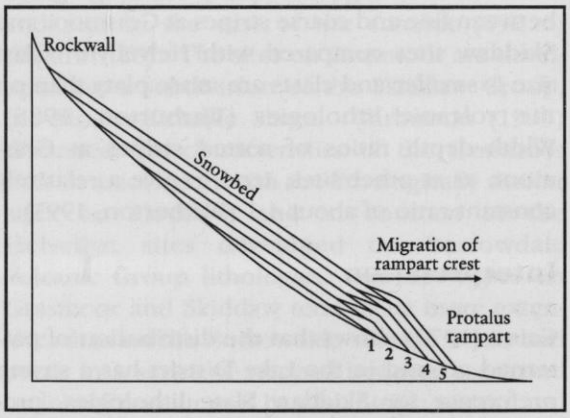

Figure 7.32 Model of protalus-rampart development at the foot of a thickening snow bed (after Ballantyne and Kirkbride, 1987b).

downslope from the foot of a talus slope and the ridge faces a direction of 040°. It has a proximal height of 3 m with a maximum gradient of 21°, a distal height of 13 m and a maximum gradient height of 34°. Deposits in the ridge reach a maximum thickness of over 10 m and consist of slate blocks with interstitial fines. The shallowness of the depression behind the ridge probably results from subsequent infill by slopewash of finer material.

Interpretation

Caine (1972) demonstrated that the distribution of patterned ground in the Lake District shows a strong preference for Skiddaw Slate lithologies, presumably because they are more frost susceptible than the Borrowdale Volcanics. Warburton (1997) showed a tendency for sites at higher elevations to have deeper sorting as a result of more frost days, and therefore greater frost heave potential at higher elevation. For the Skiddaw sites the relationship is rather poor and other factors seem to be operative, particularly the presence of frost-susceptible material at the site.

The protalus rampart was produced by rock clasts – released by frost action from an outcrop above a sheltered snow patch – sliding and rolling down the surface of the snow patch and accumulating as an arcuate ridge at its lower limit. In this respect, it is a typical example of

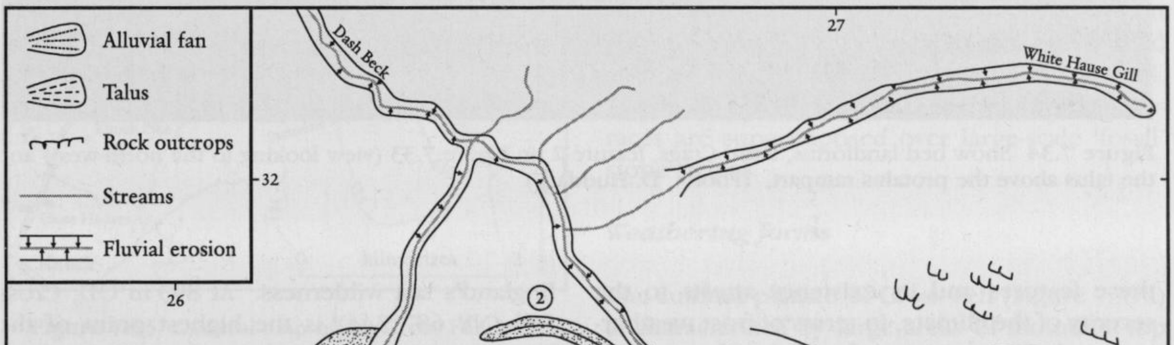

Figure 7.33 Glacial and periglacial landforms at Dead Crags (after Oxford, 1994).

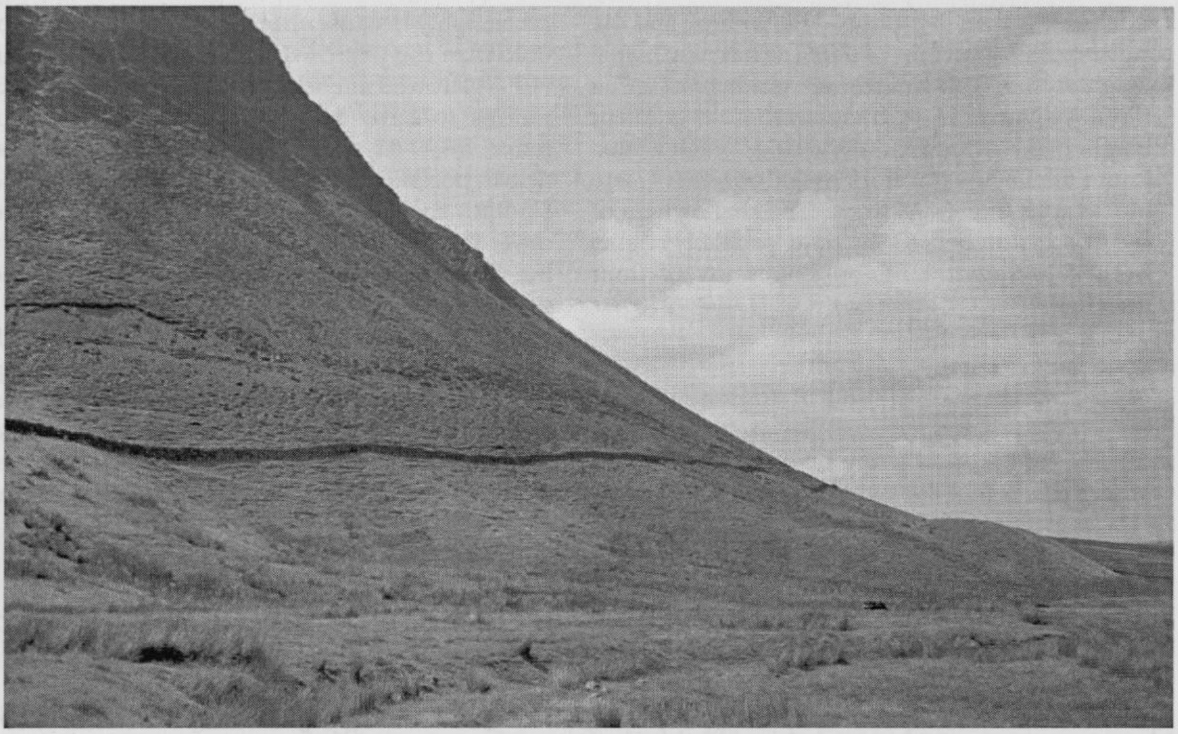

Figure 7.34 Snow bed landforms, Dead Crags, feature 2 on Figure 7.33 (view looking to the north-west) and the talus above the protalus rampart. (Photo: D. Huddart.)

these features and its existence attests to the severity of the climate, in terms of frost weathering, and the presence of a regional snowline. The feature is likely to have been developed during the Loch Lomond Stadial episode.

Conclusions

The periglacial patterned ground and snow bed and protalus rampart features of the Skiddaw area confirm the importance of periglacial processes, immediately following the Late Devensian deglaciation, during the Loch Lomond Stadial episode and, at high altitude, during the Holocene Epoch, to the development of the Lake District landscape.

CROSS FELL (NY 687 344)

W. Mitchell and D. Huddart

Introduction

The northern Pennines upland landscape of high, open and exposed plateau is, arguably, England's last wilderness. At 893 m OD, Cross Fell (NY 687 344) is the highest point of the Pennines and the coldest place in England (Manley, 1936). The site covers approximately 4 km^2 of the plateau and includes not just Cross Fell but the adjacent summits of Great Dun Fell (847 m OD) and Little Dun Fell (841 m OD) (Figure 7.35). It is bounded by a steep scarp along its western edge to the Vale of Eden, but, in contrast, the eastern slopes are more gentle and include the headwaters of the River South Tyne and River Tees. The climate generally is cool with a high number of cloudy and usually wet days, with winters characterized by negative mean monthly temperatures (Smithson, 1985). This climatic regime is similar to that of southern Iceland and allows for the present-day operation of small-scale periglacial processes (Tufnell, 1969, 1971, 1972, 1985).

Cross Fell is part of the geological unit termed the 'Alston Block' (Johnson and Dunham, 1963; Johnson and Hickling, 1970; Taylor *et al.*, 1971; Burgess and Wadge, 1974; Dunham, 1990). It is bounded to the north by the Stublick Fault and the Tyne Gap and to the south by the Lunedale Fault and the Stainmore Trough. To the west is

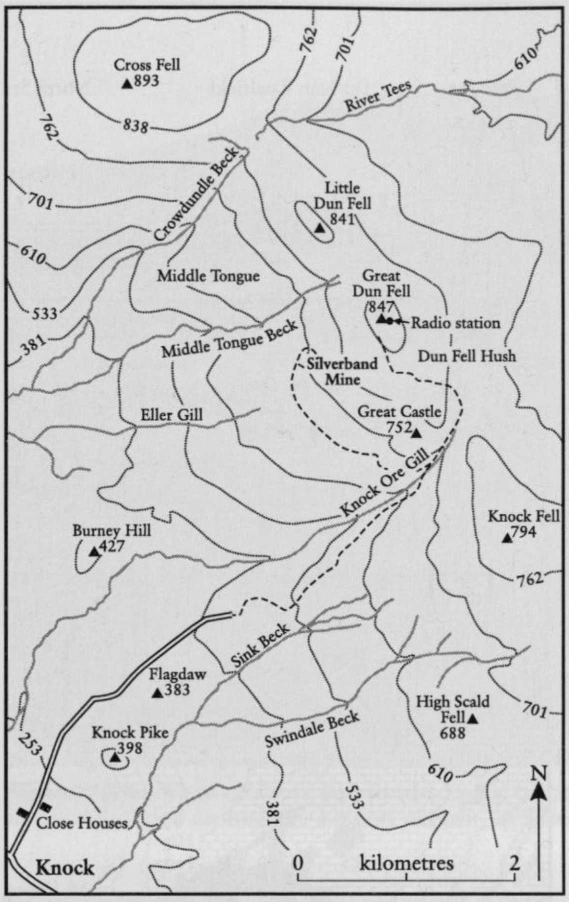

Figure 7.35 General map of the Cross Fell area.

the Pennine Fault, which has produced a major displacement and an impressive, although degraded, fault-line scarp with an estimated displacement of 600 m near Cross Fell (Figure 7.36a). The area is underlain at depth by the Weardale Granite batholith, which has caused regional deformation and the gentle easterly dip of the Alston Block (Bott, 1967). The exposed solid geology is of Carboniferous rocks, primarily sandstones, limestones and mudstones (Figure 7.36b). These lithologies alternate within definable cycles and have a direct influence on the development and style of topography and periglacial forms.

Description

Although there is little direct evidence of glaciation within the summit area of Cross Fell, a series of giant erratics, known as the 'Bullman Hills' (NY 705 375), occur 2 km to the north-east. These erratics are 100–200 m in diameter, 12–15 m high and are composed of a sequence of Great Limestone strata rafted from an outcrop about 2 km to the south, indicating ice transport northwards into the Black Burn catchment (Lunn, 1995, 1996). Drumlins also have been identified, extending down-valley from Moor House (NY 756 327) towards Cow Green and Harwood Beck.

A number of large-scale periglacial features, such as blockfields, blocky scree slopes and frost-riven cliffs, large-scale stone stripes, polygons and nivation terraces can be found in the summit area. These features are thought to be relict and associated with the severe climatic conditions of the Late-glacial period (Lowe and Walker, 1997a, b). However, the smaller scale landforms, such as gelifluction terraces, polygons, stripes, erected stones, ploughing blocks and thufur, continue to be active under present climatic conditions, because the climate is cold enough to generate frost processes (Tufnell, 1985). In places, contemporary gelifluction terraces are superimposed over large-scale 'fossil' stone stripes.

Weathering forms

The summit plateau of Cross Fell (Figure 7.37a) is characterized by large areas of mountain-top detritus formed by in-situ weathering (macrogelivation) of the underlying bedrock creating blockfields (Ballantyne and Harris, 1994). This open framework of angular boulders with no interstitial material occupies the level, or very gently sloping, ground, over an area of about 0.5 km² to the north of the summit of Cross Fell (Figure 7.37b), in association with an outcrop of coarse-grained sandstones. Smaller areas of blockfield also can be observed on the other two summit areas and excavations on the top of Great Dun Fell revealed a thick zone of in-situ weathered, Dun Fell Sandstone (Johnson and Dunham, 1963; Tufnell, 1985). Many of the surface boulders have been split by intense frost action indicating a severe climate and deep frost penetration. Block streams composed of large angular blocks of sandstone, which appear to flow from blockfields, have been reported from Knock Fell (Tufnell, 1969).

Certain of the limestone units also have been fragmented into large clasts near their outcrop. These, however, tend not to form the plateau surface but tend to form cliffs, with the under-

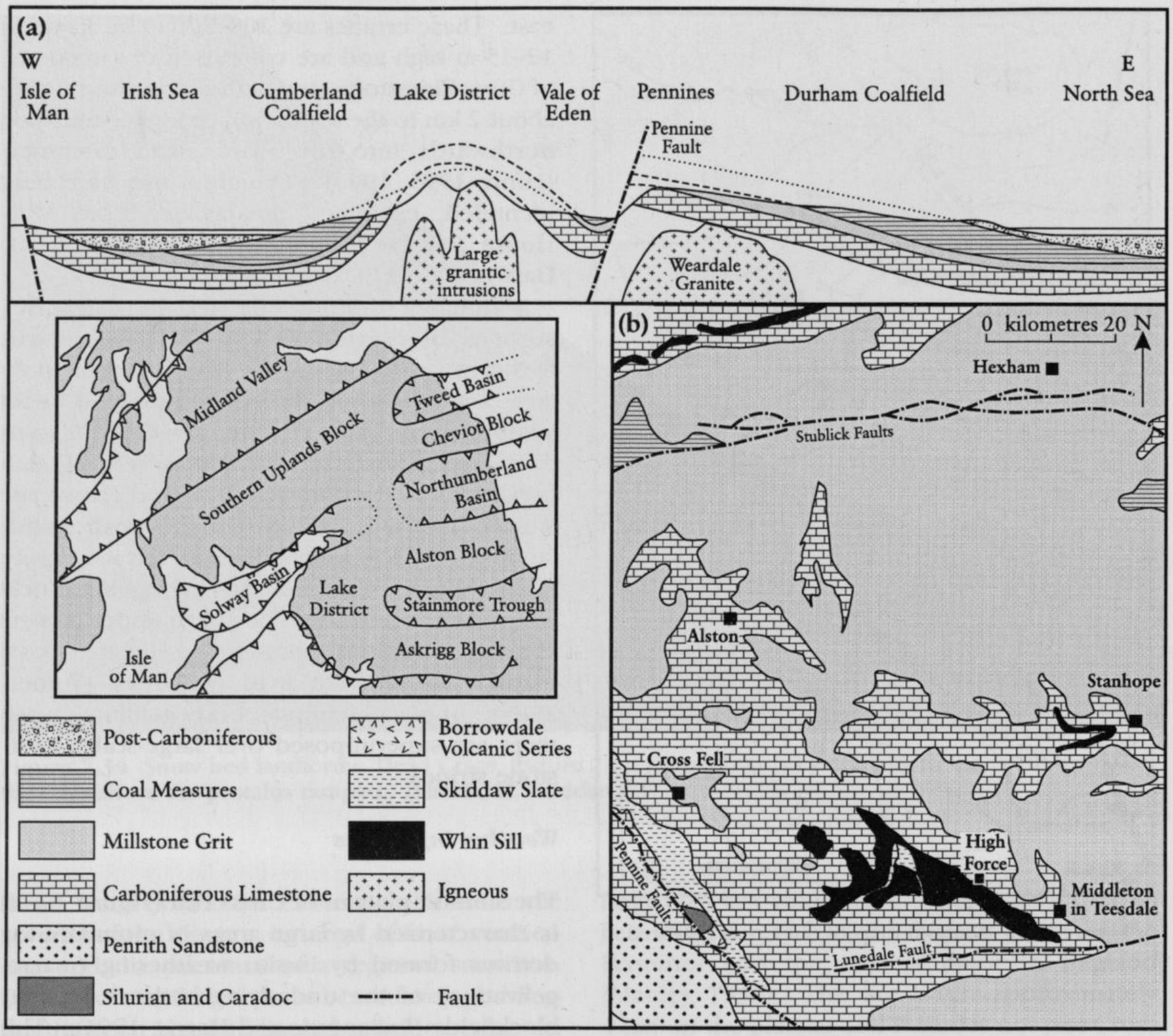

Figure 7.36 (a) Regional geological section and main structural elements of northern England (after Taylor *et al.*, 1971). (b) Geological map of the central north Pennines (after Johnson and Hickling, 1970).

lying mudstones obscured by talus slopes of limestone clasts. Talus slopes also have developed below outcrops along the northern and western flanks of Cross Fell, terminating at a break of slope at 750 m, particularly below a massive sandstone known as the 'Six Fathom Hazel' (Pounder, 1989). Lobate forms of boulders downslope of the talus slopes on the south-western side of Cross Fell indicate the operation of debris-flow processes on these slopes (Figure 7.37c).

Patterned ground

To the south and south-east of Cross Fell summit there is discernible patterned ground, particularly sorted stone polygons (Figure 7.37d) and stripes. Large-scale blocky stone stripes occur on the western slopes of Little Dun Fell (Figure 7.37f) and small active forms have been reported from the south-facing slopes and summit of Great Dun Fell (Figure 7.37e). Clasts with a vertical long axis, termed 'erect stones' by Tufnell (1969), are widespread over much of the higher ground, often in association with relict patterned ground.

A particular type of patterned ground is the vegetation covered hummocks or thufur found on many west-facing slopes at altitudes over 800 m OD (Tufnell, 1966). On Great Dun Fell they occur on slopes with a gradient of 14–17°. In upper Knock Ore Gill, between 680 and

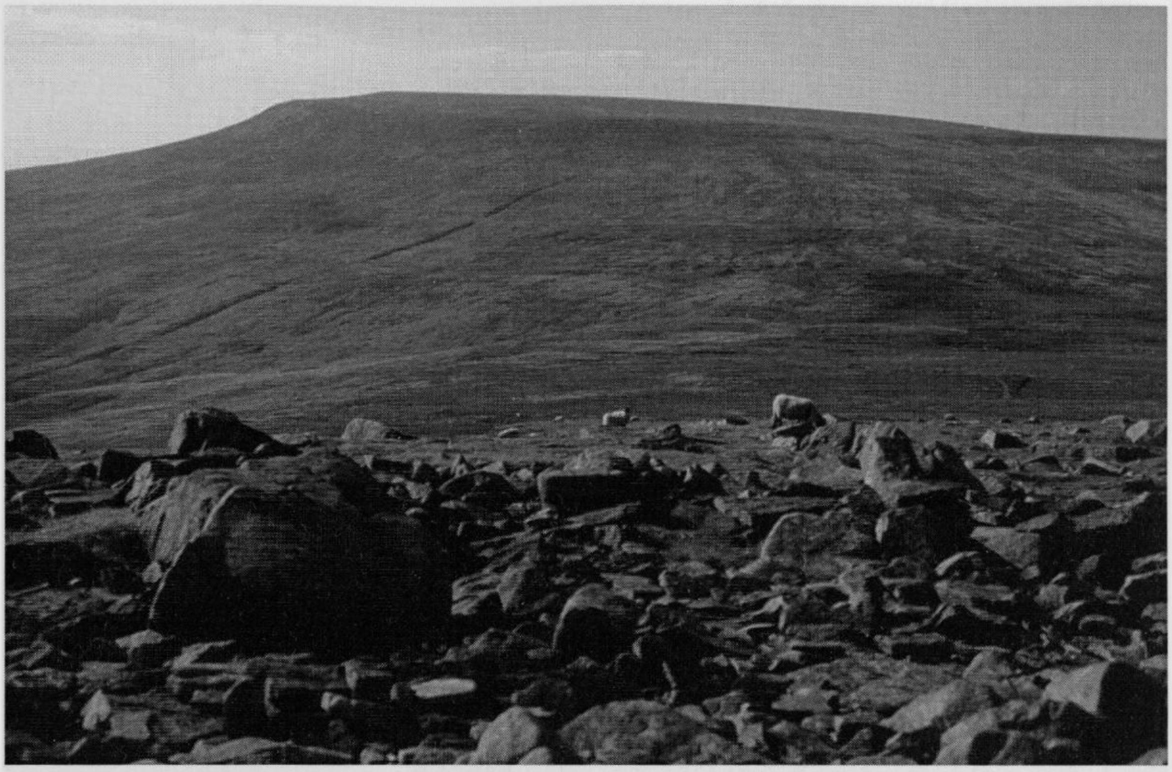

Figure 7.37a View of Cross Fell from summit of Little Dun Fell showing the general topography and talus slopes developed around the summit plateau. (Photo: W.A. Mitchell.)

Figure 7.37b Blockfield on the summit plateau of Cross Fell. (Photo: W.A. Mitchell.)

Figure 7.37c Talus slopes on the south-east flank of Cross Fell showing the lobate nature of the basal part of the talus, suggesting flow and deformation of the talus. View looking southwards towards the radio station on Great Dun Fell. (Photo: W.A. Mitchell.)

Figure 7.37d 'Fossil' polygonal patterned ground on the summit area of Cross Fell looking southwards towards Great Dun Fell. (Photo: W.A. Mitchell.)

Figure 7.37e Active, small-scale sorted circles, Great Dun Fell summit. (Photo: D. Huddart.)

Figure 7.37f 'Fossil' large-scale blocky stone stripes, western slopes of Little Dun Fell. (Photo: D. Huddart.)

Figure 7.37g Thufur field, Knock Ore Gill valley. (Photo: D. Huddart.)

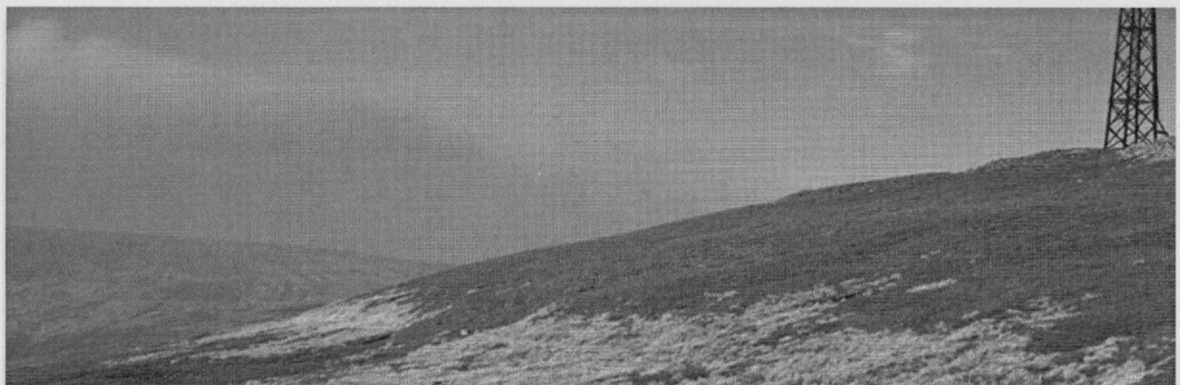

Figure 7.37h 'Fossil' altiplanation terrace, western slopes of Great Dun Fell. Note the vegetation contrast where late-lying snow banks still occur today. (Photo: D. Huddart.)

750 m OD, thufur are found in association with wet flushes on the eastern side on a 6–9° slope (Figure 7.37g). The dimensions of these landforms from a measured population of 254 indicate an average maximum height of 17.3 cm and an average maximum diameter of 36.0 cm. They are hemispherical in shape, although some are elongated as a result of slope processes and they show arching of thin vegetation layers that enclose a core of fine-grained sediment. These landforms are probably still forming today, as Tufnell (1969) measured 3–5 cm of stake movement as a result of frost heave since the late 1940s to early 1950s. They are found in association with gelifluction terraces and ploughing blocks that are moving downslope at detectable rates (Tufnell, 1972). Such hummocky microrelief may be polygenetic in origin (Tufnell, 1966; Ballantyne and Harris, 1994) and many of these forms in this part of northern England may reflect non-periglacial processes.

Solifluction

A widespread feature of upland slopes in Britain is the downslope movement of material by seasonal freezing and thawing of the upper soil layers. This may also be termed 'gelifluction', where slow saturated flow of an ice-rich soil occurs during thaw consolidation (Ballantyne and Harris, 1994) and is termed 'congelifluction' by Tufnell (1969). The presence of mudstones in the area has provided large amounts of fine-grained material to be broken down by microgelivation. Mechanical fracturing of the interbedded sandstones and limestone has provided a number of angular clasts that mix with the mud to form a widespread diamicton that generally is regarded to be the result of periglacial processes. This clay-rich diamicton has been subjected to flow and deformation under gelifluction processes, which have resulted in a number of different surface forms. Tufnell (1969, 1985) identified five different types of 'gelifluction terrace', which he divided according to whether they are convex in plan downslope (lobe) or parallel to slope (terrace), whether they are vegetated or expose sediment on the riser. Lobes are reported from Great and Little Dun Fell whereas terraces are exemplified above 730 m OD in Knock Ore Gill (Tufnell, 1985).

'Ploughing blocks' are the most widely distributed of the currently developing periglacial phenomena and have been observed down to an altitude of 450 m OD (Johnson and Dunham, 1963; Tufnell, 1969). 'Ploughing block' is a term used by Tufnell (1966) to describe large blocks on slopes that travel faster than the finer soil material. Such large clasts are able to move downslope under gravity, particularly when the soil is soft owing to high moisture content, and form distinctive mounds of ploughed material in advance of the block and a notable track or depression upslope behind the block. They indicate differential slope movement with mean annual rates of 1–5 cm a^{-1}, with a maximum movement in the spring when frozen soil is melting (Tufnell, 1969, 1972, 1985). In plan, the depressions can be niche-shaped, or elongate, depending on the differences in speed of block movement relative to that of the surrounding ground. Blocks travelling faster than adjacent parts of the slope will create an elongate depression, whereas a niche-shaped depression will form where movement just exceeds that of the surrounding slope. These landforms usually occur on grassy slopes where there is sufficient moisture; they are rarely found in areas that lack vegetation. The possible causes of ploughing block movement are summarized in Figure 7.38.

Nivation landforms

The Cross Fell area is well known for its late-lying snowdrifts and snow patches associated with distinct bedrock benches on the upper slopes of the mountains (Tufnell, 1971) and in some cases the snow is associated with semicircular hollows. They compare well with cryoplanation terraces and nivation hollows in present periglacial environments (Ballantyne and Harris, 1994) and have the following features in common:

1. they occur on the upper slopes of relatively undissected mountains that have a generally rounded form;
2. the treads of the terraces are up to 12° in slope whereas the steeper risers can vary between 20 and 35°;
3. other frost landforms are associated with them and bare ground is found on the steeper parts of the terraces, which facilitates – and is a result of – their erosion.

Some of the best terraces in the Dun Fell area are situated on the north-western slopes of the

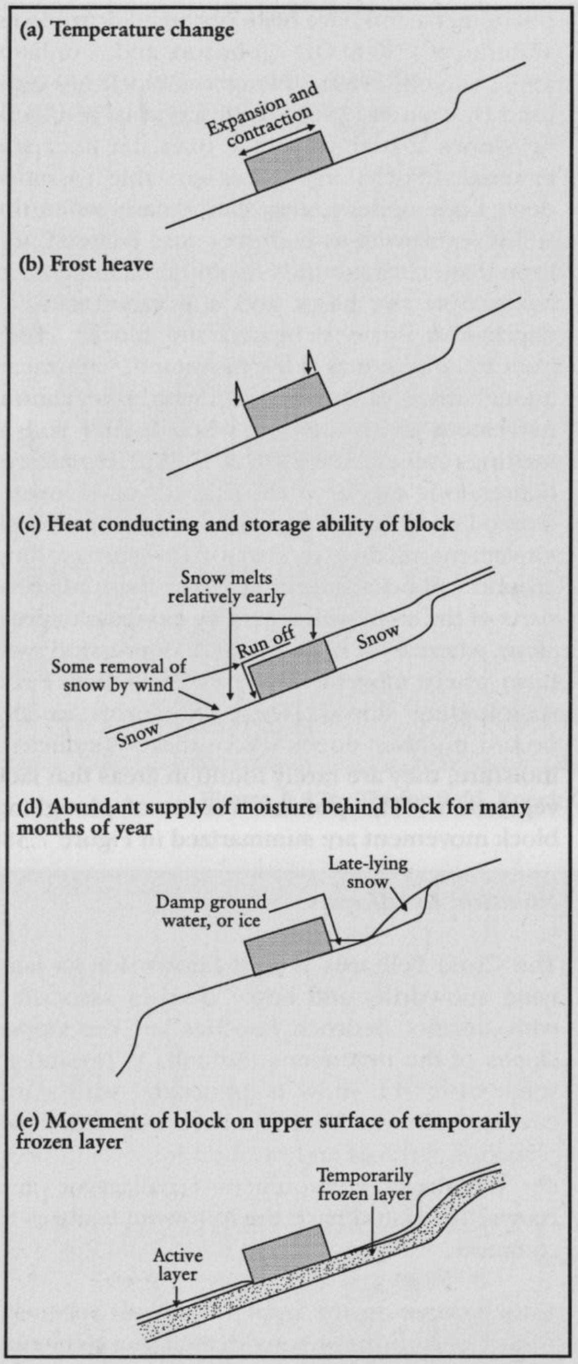

Figure 7.38 Possible causes of ploughing block movement (after Tufnell, 1966).

Upper Knock Ore Gill on ground that repeatedly experiences snow-patch formation (Figure 7.37h). This provides moisture that facilitates freeze–thaw break-up of the ground ice and needle ice, as well as the downslope movement of unconsolidated debris, water erosion and removal of the fine sediments at the base of the snow patch (Huddart, 1981d). The major controls on this nivation process are climate and topography, but rock structure with well-developed joints can promote the parallel development of frost-riven scarps. This may well be the case in the Cross Fell area, where the cyclical sequence of sedimentary rocks means that the limestone and sandstones act as cliff formers, with the shales being more easily eroded and covered by talus. It is therefore very possible that they are lithologically controlled with little periglacial modification.

Interpretation

During the last glaciation, Cross Fell acted as an area of local ice accumulation that was surrounded by ice from Scotland and the Lake District (Dwerryhouse, 1902; Raistrick, 1931b; Beaumont, 1968; Vincent, 1969; Lunn, 1996). This local ice was not thought to be very important in these original reconstructions. A similar situation in the western Pennines, however, has shown that the local ice was not just a summit ice cap but a linear ice divide, which extended along the mountains to the south of the Vale of Eden where it joined an ice divide over the Lake District (Mitchell, 1991c, 1994).

The presence of ice flow off the summit area is confirmed by drumlins to the east of the summit area in the headwaters of the River Tees, which indicate a flow direction generally south-eastwards off the summit into the Tees valley (W.A. Mitchell, unpublished data). A local origin for this ice is confirmed by the erratic content in till exposures near Moor House (NY 758 328). However, this mapping also indicates ice flow southwards from the ground around Tynehead (NY 754 360) and suggests that it was not an ice dome but a linear ice divide that extended from Cross Fell towards Burnhope Seat (NY 788 376). This interpretation would also explain the large erratic blocks of the Bullman Hills that indicate ice flow northwards off Cross Fell (Lunn, 1996). It therefore appears that during this phase the upland plateau areas were covered by local ice and not available for periglacial modification, although there is an alternative explanation.

Recent papers on other areas of the British Isles, particularly north-west Scotland (Ballantyne, 1997, 1998; Ballantyne *et al.*, 1997, 1998) and the Lake District (Lamb and Ballantyne, 1998) have identified a series of periglacial trimlines. This is defined by an upper limit on slopes of features attributable to glacial action, with the summit areas ice free and characterized by frost-shattered regolith, such as blockfields. Periglacial trimlines allow the delimitation of areas that existed as nunataks above the surface of the last ice sheet and have been shown to occur between 800 and 870 m in the Lake District (Lamb and Ballantyne, 1998). This may well have been the case with the Cross Fell summits, although it also is possible that these plateau areas were buried under cold-based ice, which allowed the preservation of the periglacial forms (cf. Rea *et al*., 1998). The height at which glacial features become apparent is therefore a reflection of a change to temperate, basal ice.

Earlier workers were convinced that Cross Fell had remained unglaciated during the last glaciation and had been used to explain the presence of a number of species of alpine plants in Upper Teesdale. The idea was taken up by Raistrick (1931b), who produced a map showing the distribution of nunataks, such as Cross Fell, based on evidence such as the lack of evidence of till above *c*. 650 m and the extensive and abundant periglacial evidence described previously. However, this idea is now discounted and the arctic–alpine plants probably migrated to these areas during the cold phase of the Late-glacial. An early description of a supposed interglacial peat on the eastern slopes of Cross Fell by Lewis (1904) has since been re-examined and discredited (Godwin and Clapham, 1951). Turner (1984) in fact demonstrates from pollen analysis of peats that woodland was present to the summit of Cross Fell in the mid-Holocene. A.G. Lunn (pers. comm., 1999) favours a complete Late Devensian ice cover, based on evidence of an ice-scoured pavement in the midst of blockfields just north of Cross Fell and a meltwater channel, presumed to be subglacial, that cuts through the main divide north of Green Fell (NY 666 365). This would imply an ice surface higher than 680 m OD. Vincent (1969) also concluded that it is difficult to account for the north-easterly ice flow in West Allendale (NY 788 500–NY 788 560) without a strong Cross Fell ice influence.

During the Late-glacial period there was no ice in the Cross Fell area, although nearby small glaciers may have formed during the Loch Lomond Stadial, such as the glacier associated with the scarp at Cronkley Scar (NY 840 294) (Wilson and Clark, 1995). There also have been landforms on the western scarp, such as the southern side of Knock Ore gully, that have been interpreted both as moraines and landslide landforms. Pounder (1989) explained this landform as a protalus rampart associated with a snow bank, but it is more reasonable to explain it as one of a series of deep-seated rotational landslides that are common in all parts of the Pennines (Mitchell, 1991b, 1996). Despite the difficulties of interpretation, it is certain that this area had a severe periglacial climate, which allowed a whole range of periglacial modification by frost, slope and snow processes during the Late-glacial, particularly the Loch Lomond Stadial (cf. Boardman, 1985b, Ballantyne and Harris, 1994).

Conclusions

The significance of periglacial processes in shaping the environment of upland Britain is clearly demonstrated by the range of associated landforms found on the highest ground of the Pennines. The Cross Fell area is important because of the range of periglacial geomorphological landforms found within such a small area. It has been one of the most actively studied periglacial landscapes in northern England.

WASDALE SCREES (NY 150 044–NY 170 060)

D. Huddart

Introduction

Wasdale Screes are a famous example of scree development (Andrews, 1961; Boardman, 1996). The screes are formed below a 180 m-high rock face below Illgill Head and extend for 2.6 km along the southern shore of Wastwater (Figure 7.39). They also are important for their rich montane flora, which includes several nationally rare species (Ratcliffe, 1960; Halliday, 1997).

Description

Wasdale Screes are formed by rockfall events from rock cliff faces on the glacially eroded, precipitous and oversteepened north face of Illgill Head. They are described as 'gully screes' because they are sourced largely from rock gullies in the cliffs. The resistant, acidic, Borrowdale Vo-

Figure 7.39 Wasdale Screes and Illgill Head from the north side of Wastwater. Note the gullies that feed debris to the screes, and the partially vegetated screes. (Photo: D. Huddart.)

lcanic andesite rocks are weathered and transported to form overlapping, fan-shaped debris accumulations from the base of gullies to below the lake level. The gullies are the result of faulting, which has resulted in shatter belts and fault breccias through the Borrowdale Volcanics, making them less resistant to weathering. Hence erosion along these lines of weakness has given rise to the major supply routes for scree. The largest boulders, up to 2 m across, occur chiefly at the base of the scree fans and there is a general fining of the grain size upslope (Figure 7.40). The average slope angle for the scree fans is *c.* 35°. This is less than the natural angle of slope for materials of this size, which is between 39 and 42°. The lower angle probably reflects the current, relatively low, rockfall input.

Measurements of slope angles reveal three main slope facets (Andrews, 1961). A relatively steep upper slope of between 35–37° decreases to 31–34° in the lower part to form a concave profile characteristic of many scree slopes (Statham, 1976). This basal concavity is not fully developed but is truncated typically 5–15 m above the lakeshore by a relatively steep basal slope of 39°, and locally up to 41–46°, where large boulder gravels accumulate. Between the main active scree fans, older, inactive scree slopes are vegetated. Slope angles on these vegetated screes are lower, at about 27°, but they too steepen at the lakeshore to 35°, and locally to 42–43°. There are considerably more fine-grained particles present between the larger gravels, and these vegetated screes possess a poorly developed soil. The vegetation of these areas and the rocky slopes above consists of patchy *Festuca–Agrostis* grassland and heather moorland (with both *Calluna vulgaris* and *Erica cinerea*). Other species present include wood sage, foxglove and goldenrod (*Solidago virgaurea*) and there are scattered trees present close to the lake and on cliff ledges, including oak, ash, holly, hawthorn and yew (Figure 7.41). In places there are former rotational slump scars

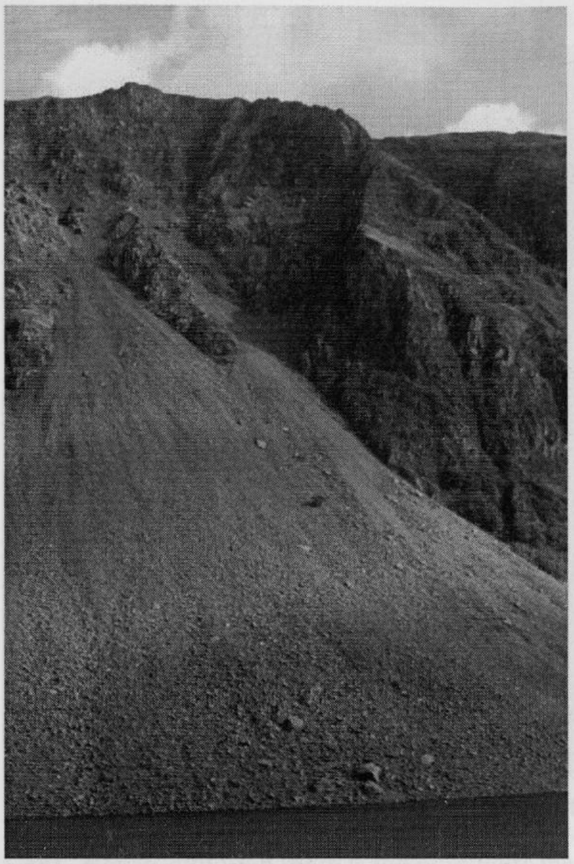

Figure 7.40 Coarse down-scree sorting on Wasdale Screes and gullies. (Photo: D. Huddart.)

that are now vegetated. In Figure 7.41 debris-flow lobes are visible where some of the finer-weathered fraction has become remobilized during periods of heavy rainfall.

Interpretation

Screes are formed from the cumulative effects of rockfall. This is a process of small-scale failures in rock entirely along planes of weakness. Rock particles released by failure then fall and roll down the scree slope under gravity in discrete events. The planes of weakness, particularly joints, are thought to be opened up by freeze–thaw processes, and in Norway when the rockfalls are plotted against time of year their greatest frequency is at times when there are continual changes of temperature about the freezing point, that is in April and October. However, Douglas (1980) found that in Antrim over 60% of all rockfall events were independent of freezing conditions. Events over 0.21 kg in size occurred seasonally, as a result of freeze–thaw, but events under 0.21 kg were observed throughout the year. Furthermore, Eichler (1981), in Ellesmere Island in the Canadian Arctic, measured rock surface temperatures as high as 39.7°C and these high temperatures, plus observations of preferential weathering on south-facing slopes, suggested that insolation weathering – and not frost shattering – was the main agent of rock breakdown.

In the case of the Wasdale screes it is unlikely that insolation weathering plays a major role in rock breakdown; water freezing in cracks in the winter is the more likely preparer of rock for subsequent rockfall. Boardman (1996), however, suggests a more radical proposal that these screes have little to do with frost action and are the result of the removal of loading by ice during deglaciation. This rapid unloading would allow expansion of the rocks, the formation of joints as pressure is released and inevitably rockfall. However, it is difficult to assess such an idea in the absence of monitoring of both rockfalls and climate, especially as the screes must have been much more active in the past. The importance of rockwall instability in accelerating rockwall collapse during and immediately after deglaciation has been demonstrated by André (1985, 1986), however, who found that the most rapid rates of rockwall retreat in Spitsbergen resulted from pressure release following glacier surges.

Screes generally are accumulation slopes, which means that the amount of sediment increases through time (although it could be argued that there will be some erosion of basal scree by lakeshore erosion, particularly during storms) and that therefore the screes should show characteristics of transit slopes, which would be a straight scree slope at the angle of repose for the debris falling. If such a slope is undercut then mass avalanching can occur, where all the particles on the scree move as a unit. The characteristics described above for the Wasdale Screes, however, suggest that although this may occur occasionally, the screes could better be described as low-rockfall input, and no, or very little output, screes. This type of scree does not possess characteristics compatible with the angle of repose model, but it does have a basal concavity, a straight slope under the angle of repose and a coarse downslope sorting. These characteristics describe the Wasdale Screes, where there is a poorly developed basal concavity, but with some undercutting.

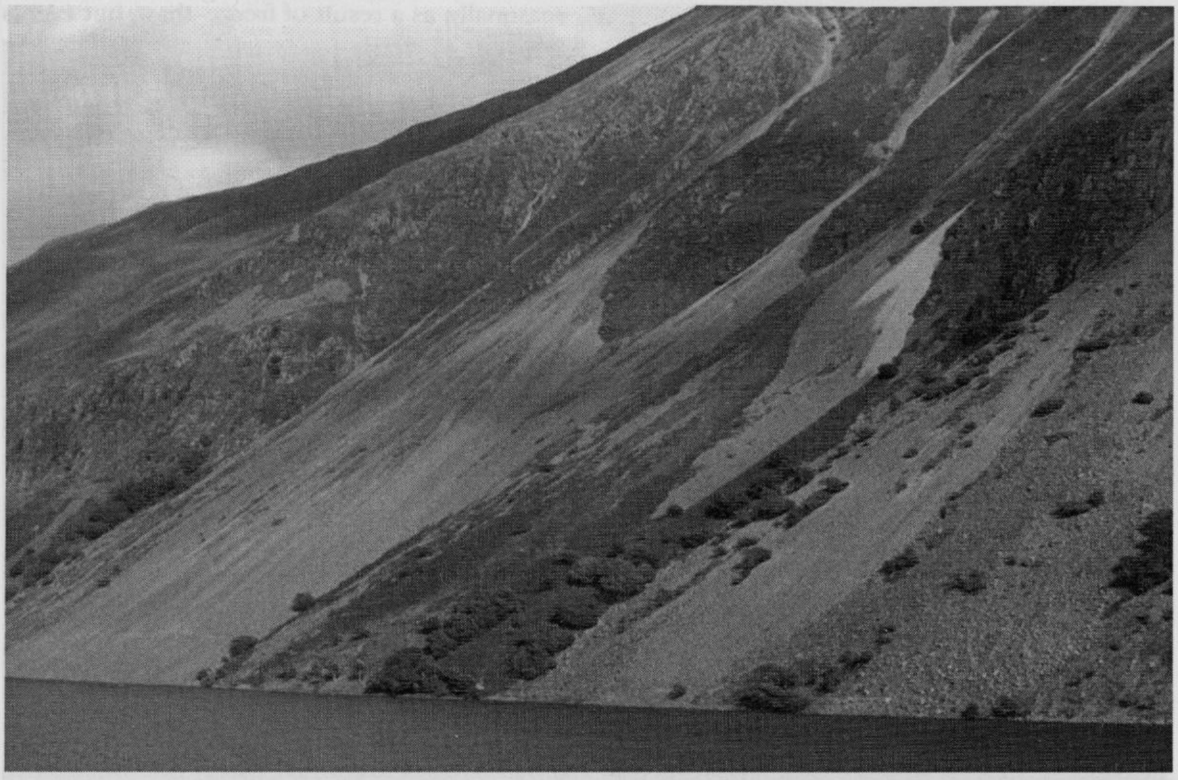

Figure 7.41 Debris-flow lobes on Wasdale Screes and vegetational colonization of the screes. (Photo: D. Huddart.)

The angle of slope is 4–11° less than the angle of repose and must be stable with respect to mass avalanching. The reason for this profile shape is because the screes are formed by one-at-a-time gravel inputs by rockfall. When these gravels hit the scree surface they have some input energy (their mass times the distance through which they have fallen). Some of this energy is transferred into downslope movement. In contrast, the angle of repose is related to the coming to rest of a sliding mass of particles, which has a velocity much less than the downslope velocity of a falling particle. Therefore the particle moves itself and others by impact downslope and keeps the slope angle lower than the repose slope. The concavity at the base of the slope is caused by faster particles running out as a tail on to the accumulation surface, although this cannot be easily seen at Wasdale because of the lake and some undercutting.

The coarse downslope sorting results from the sieve-like nature of the scree surface, which is irregular. Starting with a random or uniform size distribution on a surface, the small particles are trapped more easily by interlocking and do not move as far as the larger particles, which travel to the base of the slope by overriding the surface irregularities. Ballantyne (1985) found that fall sorting is dependent not only on the particles being trapped amongst interstices but also reflects the greater success of larger particles in overcoming frictional losses as they travel downslope. As accumulation continues, as the area of feeding rock face diminishes and the range of fall velocities becomes less, the scree slope becomes nearer the repose angle because there is less input energy to move debris downslope. The Wasdale Screes are still far from this situation and there remains a large, vertical, free face to supply further rockfall. However, the result of weathering and rockfall is a parallel stripping of rock along the rock face, which can be seen from the crags below the summit of Illgill Head.

Colder climates than today – and particularly the Loch Lomond Stadial climate and/or the Little Ice Age climate – would be a much more favourable climatic regime to provide the greater temperature range and the higher frequency of temperature oscillations that are required to cause frost shattering of the bedrock. Andrews (1961) suggested that the Wasdale Screes last received a significant debris input in the Little Ice Age of the seventeenth and eighteenth centuries. Although this possibly is correct, he did not provide any evidence to support this suggestion. However, the fact that many of the slopes in Wasdale have the characteristics of fossil, vegetated screes and even the classic, more active screes themselves can be seen to be showing the initial stages of colonization by vegetation, suggests that these screes are not as active in today's climatic regime as they have been in the past. The landforms are probably almost relict landforms that are not in equilibrium with today's processes, or as Andrews (1961) described them they are in 'a retarded state of development'.

Ballantyne and Kirkbride (1987a) suggest too that the Loch Lomond Stadial climate was particularly conducive to rockfall by frost wedging. Conditions for freeze–thaw probably were favoured by a combination of strong insolation during the spring, summer and autumn months, and much cooler air temperatures than at present. Hence, although there is the possibility that the relict Late-glacial screes might be interpreted as paraglacial landforms resulting from the rapid break-up of glacially oversteepend rock walls immediately after deglaciation, rather than periglacial landforms produced by enhanced frost action operating on the free faces during the Loch Lomond Stadial, this seems unlikely as the main cause of the screes. The relative immaturity of most Flandrian screes below steep rock walls that were occupied by glacier ice during the Loch Lomond Stadial suggests that paraglacial effects generally have been of secondary importance, and that the primary impetus for the development of screes such as the Wasdale examples during the Late-glacial was climatic (Ballantyne and Harris, 1994).

A scree surface might be thought of as an inhospitable place for plants to colonize, with free drainage, little humus and reworking of generally unstable, coarse debris slopes. Screes, however, do become slowly vegetated, as can be seen from the Wasdale examples, and the nearly ubiquitous presence of lichens on the gravels implies a low current geomorphological activity and the first stage in the colonization process. Soil development and colonization by vegetation is more rapid on finer particle sizes, but where this fine-grained sediment originates is not known. It is likely to be a combination of wind transport and the in-situ weathering of the coarser scree. As boulder gravel becomes embedded in the irregular scree surface there are quiet areas created in front of them. Bryophyte (*Rhacomitrium*) cushions colonize here, these hold water and then the parsley fern (*Cryptogramma crispa*) (Figure 7.42) comes in to provide water-holding humus (Leach, 1930). This is the most famous scree pioneer in the Lake District (Pearsall and Pennington, 1973). Other plants can then colonize from the surrounding hillsides, particularly the heath plants such as lady's bedstraw (*Galium verum*) and the grasses such as *Festuca*, *Agrostis* and *Deschampsia*. The vegetation colonizes downwards in vertical strips from the boulders and eventually becomes dominated by *Calluna* and has a typical heath composition. Similarly at the base of

Figure 7.42 The pioneer species on Wasdale Screes, the parsley fern (*Cryptogramma crispa*). (Photo: D. Huddart.)

crags away from the gullies there are locations where debris accumulates at a very slow rate and therefore vegetation can gain a hold and spread down the scree as strips. The colonized areas of these screes usually occupy depressions, because there is a piling up of debris on either side of each colonized area and hence fan-shaped gravel accumulations hold back the growth of vegetation. Typical of all active screes is the destruction of vegetation by reactivation of rockfall and by burial and the complete colonization process takes a long time. The screes also are colonized from the marginal areas, which have become stabilized. The stable scree becomes colonized by plants such as bracken (*Pteridium aquilinum*), foxglove (*Digitalis purpurea*) and bilberry (*Vaccinium uglinosum*). Eventually the screes become completely 'fossilized', as can be seen in the middle right section of Figure 7.39.

The steep and crumbling gullies, rich in calcite from the weathered iron-rich volcanic rock, provide habitats for the extremely rare shrubby cinquefoil (*Potentilla fruticosa*) on moist ledges and rock crevices. Other nationally important mountain plants include roseroot (*Sedum rosea*), mountain sorrel (*Oxyria digyna*), lesser meadow rue (*Thalictrum minus*), northern bedstraw (*Galium boreale*), stone bramble (*Rubus saxitilus*), *Alchemilla wichurae*, mossy saxifrage (*Saxifraga hypnoides*), eyebright (*Euphrasia rivularis*) and green spleenwort (*Asplenium viride*) (Ratcliffe, 1960; Pearsall and Pennington, 1973; Halliday; 1997). Several species reach their lowest elevation in Wasdale, where for example alpine lady's mantle (*Alchemilla alpina*), yellow mountain saxifrage (*Saxifraga aizoides*), alpine clubmoss (*Diphasiastrum alpinum*) and parsley fern (*Cryptogramma crispa*) reach the shore of the lake, and purple saxifrage (*Saxifraga oppositifolia*) descends to below 300 m on the slopes above.

Conclusions

Wasdale Screes show all the classic morphological and sedimentological characteristics of low-input and little-output screes, which are fed by rockfall, largely from weathering in gullies on the free face above the screes. They are largely out of equilibrium with the present climatic conditions and formed mainly in the Loch Lomond Stadial and possibly through reactivation in the Little Ice Age when the climate of this part of northern England was considerably colder than today and there would have been many temperature oscillations around 0°C. This caused frost shattering of the bedrock. The screes show various stages of vegetation colonization and there are some nationally rare, montane species.