Quaternary of Northern England

D. Huddart
Liverpool John Moores University,
Liverpool, UK

and

N.F. Glasser
University of Wales,
Aberystwyth, UK

With contributions from

Jim Innes
David Evans
John Boardman
Silvia Gonzalez
Richard Chiverrell
Wishart Mitchell
Andy Plater
Sarah Morriss
Cynthia Burek
Stephan Harrison
Richard Jones
Graham Wilson

GCR Editor: **G.S.P. Thomas**

JOINT NATURE CONSERVATION COMMITTEE

Chapter 5

The Devensian glacial record

INTRODUCTION

D. Huddart and N.F. Glasser

The later stages of the Quaternary Period, the Devensian cold stage and the Holocene temperate stage, are much more important in shaping the landscape of northern England than the preceding stages. During the Late Devensian, ice sheets developed in the highlands of Scotland and these advanced south along the east and west coasts of northern England to coalesce with independent ice-caps formed in the Lake District, the Cheviot Hills and parts of the Pennines. At its maximum extent, Devensian ice covered the whole of northern England, with the exception of small parts of Derbyshire. The sediments and landforms developed during this episode are described in this chapter. On retreat of the Devensian ice, periglacial processes modified much of the glacial landscape, and the landforms and deposits that resulted are dealt with in Chapter 6. At the end of the Devensian glaciation, rapid climatic fluctuation during the Late-glacial interstadial and stadial episodes caused significant vegetation changes in the region, together with much sediment accumulation in lakes, and these topics are covered in Chapter 7. The final stage of the Quaternary Period, the current Holocene temperate stage, saw a significant rise in sea level, reactivation of river systems and their modification of the glacial sediment cover, major changes in vegetation consequent on widespread climatic amelioration, and evidence of human modification of the landscape. These topics are covered in Chapter 8.

The traditional stratigraphical view is that the Devensian cold stage followed the end of the Ipswichian interglacial stage (Oxygen Isotope Stage (OIS) 5e) at approximately 115 ka and is subdivided into Early (115–50 ka), Middle (50–26 ka) and Late (26–10 ka) sub-stages (Mitchell *et al.*, 1973). The biostratigraphical evidence for the transition from the Ipswichian interglacial to the Devensian glacial, however, is scanty and, as Lowe and Walker (1984) have stated, once evidence from outside Britain is considered there seems to be a much more complex environmental sequence across the boundary than evidence in Britain provides. Thus, a climatic deterioration at 115 ka (OIS 5d) is followed by a warmer phase of either interglacial or interstadial rank about 100 ka (OIS 5c). This, in turn, is followed by a cool episode around 90 ka (OIS 5b) and an interglacial or interstadial episode about 80 ka (OIS 5a). Between 80 ka and 65 ka significant cooling may be correlated with OIS 4.

Early Devensian (115–50 ka)

Several sites in northern England yield important information related to this time period but pose problems concerning their dating and whether they provide evidence of extensive Early Devensian glaciation. Chelford has provided major biostratigraphical and palaeoclimatic evidence for an early Devensian interstadial event, known as the 'Chelford Interstadial' (Simpson and West, 1958), but its precise age is uncertain. Conventional radiocarbon dating methods have provided age estimates between 25 000 and 65 000 years BP for organic sequences within the Chelford Formation, uranium-series dates have yielded an age of 86 000 years BP (Heijnis and Vanderplicht, 1992) and thermoluminescence dates suggest absolute ages in the range 90 000–100 000 years BP (Rendell *et al.*, 1991; Rendell, 1992). The glaciogenic deposits that underlie the Chelford Formation may therefore identify an Early Devensian glaciation but equally may relate to a glacial episode preceding the Ipswichian glaciation. Similar uncertainty relates to the age of the thin peat in Mosedale on the northern fringes of the Lake District. The peat overlies the Thornsgill Till and is overlain by Late Devensian Threlkeld Till and terrace gravels. The peat has been radiocarbon dated by Boardman (1981) as >54 200 years BP and uranium-series dated at between 77 000 and 91 000 years BP, and is tentatively ascribed to an Early Devensian interstadial.

A borehole at Burland, near Nantwich (Bonney *et al.*, 1986), revealed 1 m of peat within a 9 m sequence of organic silts and clays deposited in a lake or abandoned river channel. The pollen indicates evidence for local pine–birch forest into which spruce migrated and the minimum age of these deposits is >47 200 years BP. The organic sequence is overlain by thick Devensian Till and underlain by further till correlated with the Oakwood Till at Chelford. The organic sequence therefore represents an interstadial episode of probable Chelford age, underlain by till of at least Early Devensian, but probably older age. At Four Ashes, organic sediments beneath Late Devensian till, and in contact with bedrock, have a pollen assemblage including pine, spruce and birch that is comparable with the Chelford and Burland flora. Again the insect fauna is analogous to Chelford, but overall, although the oldest stadial deposits could either pre-date or post-date the Chelford Interstadial, most of the succession is thought to be of post-Chelford age.

A number of authors have suggested an Early Devensian glaciation in eastern England (Straw, 1979c) and it has been suggested that at around 70 000 years BP there was a limited build-up of ice in Scotland and upland Cumbria (Nirex, 1997b); a suggestion made earlier by Huddart (1971b) to explain Southern Scottish erratics in the Vale of Eden and the lower tills in the sequences in the valleys of the northern fringes of the Lake District. Supporting evidence for an Early Devensian glacial comes from Norway where Baumann *et al.* (1995) provided evidence for extensive glaciations on three occasions during the Devensian. Despite this possibility Worsley (1991a) concluded that there was a paucity of convincing stratigraphical or chronological data and that on balance there is no good evidence for an Early Devensian glaciation in Britain. Bowen *et al.* (1986) also argue that such a concept remains unproven.

In the Peak District, Elder Bush Cave has produced a vertebrate fauna transitional between those of the Ipswichian and Devensian, including woolly rhinoceros, cave bear, bison, cave lion and hyaena (Bramwell, 1964; Bramwell and Shotton, 1982). Extensive cave sedimentation occurred in the Peak District during the Early Devensian (Briggs and Burek, 1985), and the Hope Terrace of the Derwent valley, formed in the Ipswichian, received a cover of soliflucted sediment in the Early Devensian (Waters and Johnson, 1958; Briggs and Burek, 1985). Robin Hood's Cave (Cresswell Crags) has produced Mousterian artefacts and Early Devensian pollen spectra from its basal stratigraphy (Campbell, 1977; Roe, 1981). At Pin Hole Cave, two separate Middle Palaeolithic industries and many vertebrate remains have been discovered in a probable Early Devensian context (Jenkinson *et al.*, 1985).

Farther north, at Stump Cross Cave in northern Yorkshire, cave sediments containing Pleistocene mammal remains that include reindeer, wolverine, wolf, red and arctic fox have been uranium-series dated to 83 000 years BP using flowstone that encased them (Sutcliffe *et al.*, 1985). Hence a cold environment existed at this period within OIS 5b, as wolverine and the animal that comprises the major part of its diet, the reindeer, currently have a circumboreal distribution that reaches the tundra. Recently more precise dates for these sediments come from flowstone enclosing the wolverine bones and these are close to 74 ka (Gilmour, pers. comm., quoted in Currant and Jacobi, 2001).

Table 5.1 The mammalian fauna from the Pin Hole Mammalian Zone, Lower Cave Earth, Pin Hole Cave, Cresswell, Derbyshire (after Currant and Jacobi, 2001).

Homo species	artefacts
Lepus timidus	mountain hare
Spermophilus major	red-cheeked suslik
Canis lupis	wolf
Vulpes vulpes	red fox
Ursus arctos	brown bear
Mustela erminea	stoat
Mustela putorius	polecat
Crocuta crocuta	spotted hyaena
Panthera leo	lion
Mammathus primigenius	woolly mammoth
Equus ferus	wild horse
Coelodonta antiquitatis	woolly rhinoceros
Megaloceros giganteus	giant deer
Rangifer tarandus	reindeer
Bison priscus	bison

Uranium-series dating of speleothems has further clarified the pattern of environmental change in this period. Gascoyne *et al.* (1983) reported low speleothem growth in north-west England between 90 000 and 45 000 years BP, which implies the existence of a non-glacial tundra-like climate. Gordon *et al.* (1989) have pointed to a clear fall in speleothem growth frequency after a peak at 95 000 years BP, with a minimum at 80 000 years BP considered to mark the boundary between the last interglacial and the last cold stage. Peaks of speleothem growth at 76 000, 57 000 and 50 000 years BP have been proposed as indicators of Devensian interstadials. It is clear though that this Early Devensian period is still not well known and climatic changes and the timing of events in northern England are known only from a few sites, with little stratigraphical continuity.

Middle Devensian (50–25 ka)

This period of the Devensian is within the limits of ^{14}C dating, yet often there is still some doubt related to age determinations. It appears that the climate was cold for much of this period, with an amelioration that gave rise to the Upton Warren Interstadial. The key site is Four Ashes, where sandy detritus peat and organic clay within sands and gravels have given ^{14}C ages of between 42 000–30 000 years BP (Morgan, A.V., 1973). However, there is evidence of a cold phase between the Chelford Interstadial and prior to 43 500 years BP, based mainly on insect evidence, with indications of a continental climate with cool summers. Between 42 500 and 38 500 years BP there were significant changes in the insect fauna, indicating a climate of reduced continentality, with warmer summers than previously. Tree growth would have been possible but there is an absence of tree pollen and no wood-eating insects. After about 38 000 years BP the numbers of northern and eastern stenothermic insects increased in a tundra landscape. Cold polar desert conditions ensued after 30 000 years BP and the Late Devensian ice advance was after this date in southern Staffordshire.

Currant and Jacobi (2001) use the Pin Hole Cave as the type locality for their Middle Devensian (OIS 3) mammalian assemblage (see Table 5.1), specifically the material from the Lower Cave Earth (Pin Hole Mammalian Zone (MAZ). There is also a rich, but hitherto unreported small mammalian fauna, collected during 1984–1989; together with birds, fish, amphibians and the preservation of contemporary pollen, both within the cave deposits (Coles, 1987) and in spotted hyaena coprolites (Lewis, pers. comm., quoted in Currant and Jacobi, 2001). The age of this faunal assemblage is well constrained by a combination of uranium-series, electron spin resonance (ESR) and ^{14}C dates that are consistent with accumulation during the interval 50–38 ka (Jacobi *et al*, 1998). Small mammals recovered from excavations in nearby Robin Hood cave (area A, south-west corner of the western chamber) in direct association with a Pin Hole MAZ fauna include *Dicrostonyx torquatus*, *Microtus oeconomus*, *Microtus gregalis* and *Arvicola terrestris*. Other sites with a similar Middle Devensian fauna include The Arch, Cresswell (also known as Lions's Mouth); Ash Tree Cave, Whitwell, Derbyshire; and the basal clay from Windy Knoll Cave, Derbyshire (Dawkins, 1877). This Pin Hole MAZ is a western extension of the characteristic later Quaternary assemblage of much of central Asia, north of the Himalayas and as such it is assumed that in Britain represents an extension of extreme continental conditions right up to the Atlantic seaboard in OIS 3 (Currant and Jacobi, 2001).

At the Oxbow opencast coal site in the Aire valley, Gaunt *et al.* (1970) described a silt, deposited in standing or quietly flowing water on a floodplain. The flora and fauna indicated a tundra environment, with some snow patches. A mammoth tusk from this sediment was dated by ^{14}C to 38 600 years BP and its age is considered applicable to the deposit it was contained in. The flora included dwarf birch, fringed sandwort, alpine cinquefoil, *Thalictrum alpinum*, *Armeria maritima*, *Salix herbacea*, *Allium schoenoprassum* and *Empetrum* species. July temperature values of around 10°C and a continental climatic regime have been suggested.

Again by means of uranium-series dates Gascoyne *et al.* (1983) have deduced overall low speleothem growth in the Craven District of north-west England from 85 000 to 35 000 years BP. For the period 44 000–34 000 years BP, a minor increase in the number of dated speleothems occurs, which may indicate episodes of milder climate. No ages were obtained for 34 000–13 000 years BP, which has been interpreted as demonstrating the existence of continuous permafrost, then ice, in this area. However, Atkinson *et al.* (1986) suggest that groundwater recharge was possible in this area at 39 000 years BP and 26 000 years BP. Intermittent growth was also suggested by dates published by Sutcliffe *et al.* (1985) of between 44 000 and 29 000 years BP from Stump Cross caverns. This means that there was discontinuous rather than continuous permafrost present in this area during this period to as late as 26 000 years BP. At Pin Hole Cave in Cresswell Crags, earlier Upper Palaeolithic artefacts have been recorded from the Upper Cave Earth (Jenkinson *et al.*, 1985) and at Robin Hood's Cave in the same area similar artefacts also have been found, and a ^{14}C date of 28 500 years BP has been obtained from bone of *Ursus arctos* (Campbell, 1977). Recent dates for spotted hyaena remains from the Cresswell area are given in Table 5.2 (Currant and Jacobi, 2001).

During the Nirex investigations (1997b) in west Cumbria, Mid-Devensian marine and lacustrine deposits were found to overlie a weathered till of possible Wolstonian age in boreholes at Drigg and Carleton. In the Drigg borehole a shelly silt of possible marine origin, below a till

Table 5.2 Radiocarbon dates (years BP) on spotted hyaena remains from the Cresswell area, Derbyshire (after Currant and Jacobi, 2001)

Robin Hood Cave	OxA-6115	22 800 ± 240
Robin Hood Cave	OxA-6114	22 980 ± 480
Church Hole	OxA-5800	24 000 ± 260
Ash Tree Cave	OxA-5798	25 660 ± 380
Church Hole	OxA-5799	26 840 ± 420
West Pin Hole (Dog Hole)	OxA-5803	29 300 ± 420
Robin Hood Cave	OxA-5802	31 050 ± 500
Pin Hole	OxA-1206	32 200 ± 1000
Robin Hood Cave	OxA-5801	33 450 ± 700
Pin Hole	OxA-1207	34 500 ± 1200
Pin Hole	OxA-4754	37 800 ± 1600
Pin Hole	OxA-1448	42 200 ± 3000

and overlying a blue-grey organic silt, yielded amino acid ratios corresponding to an age of 45 000 years BP. These deposits are important for an interpretation of Middle Devensian sea levels, but it is conceivable that they are not *in situ*, although Wingfield *et al*. (1997) suggest that as they have been located in four boreholes separated by up to 0.5 km it is unlikely they form a raft. At Carleton Hall and Hall Carleton boreholes have proved laminated muds underlying the Late Devensian Blengdale Glaciogenic Formation (Nirex, 1997b). The upper part of the laminated sequence, which contains marine microfossils, has been assigned to the Glannoventia Formation of possible OIS 3 age. The lower part is of glacio-lacustrine origin and studies of varves indicate sedimentation over a period of over 2000 years (Nirex, 1997b). This unit has been tentatively correlated with OIS 4, although it is possible that the microfossils are reworked into a glacio-lacustrine sequence of Late Devensian age. This again emphasizes the tentative nature of much of the environmental reconstructions for this period of the Early and Middle Devensian in northern England.

Late Devensian (25–10 ka)

Historical background

Northern England has played an important historical role in the development of conceptual models regarding the formation of glacial sediments in Britain. This has seen a change from a belief in the marine origin for the drifts by 19th century workers, through various types of land-based origin, and, in the late 1980s, back to a glaciomarine origin for some of the glacial sediments. Howse (1864) was an early believer in the importance of terrestrial ice-sheets and their erosive power, through his work in north-east England. Goodchild (1875, 1887), working in Edenside, stressed the complexity of the glacial deposits and considered that practically all of the glacial sediments in this area formed subglacially, or englacially, during the melting of a stagnant ice sheet. Mackintosh (1877) developed a tripartite succession for the glacial deposits in north-west England, where a lower and an upper boulder clay was separated by a middle sand. His model invoked deposition by glacial ice from the Lake District for the lower boulder clay, followed by deposition of the middle sand and the upper boulder clay by floating coastal ice. Traditionally, however, these sequences were interpreted as the product of multiple glaciations. Ice that advanced to form the lower boulder clay then retreated, deposited the sand and gravel, before readvancing to form the upper boulder clay. In contrast, Binney (1848) advocated a monoglacial view where he maintained that the glacial successions were complex and that sand and boulder clay lithologies could replace each other within any given sequence at random. Nevertheless the rigid tripartite framework involving bi-glaciation proposed by Hull (1864) gained the ascendency (Figure 5.1a). For example, Kendall (1902) suggested a tripartite division of the Cleveland area glacial deposits, the [British] Geological Survey workers in Cumbria (Smith, 1912; Trotter, 1929; Hollingworth, 1931), Durham (Smith and Francis, 1967) and Shropshire (Poole and Whiteman, 1960) described similar sequences, and in Lancashire, both Taylor (1958) and Simpson (1959) showed that there were two tills present with intervening middle sands. Later workers argued that the tripartite succession accumulated in a number of different depositional environments and did not necessarily indicate advance, retreat and subsequent readvance. Thus, Johnson (1965a) considered that the sequence belonged to one complex glaciation, developed by ice-front oscillation prior to final melting. Similarly, Thompson and Worsley (1966) considered that the tripartite succession in the Shropshire lowlands was deposited from a single stagnating ice sheet and Carruthers (1947, 1953) considered the process of glacial undermelt to be important (Figure 5.1b). In Cumbria, Trotter (1929) and Hollingworth

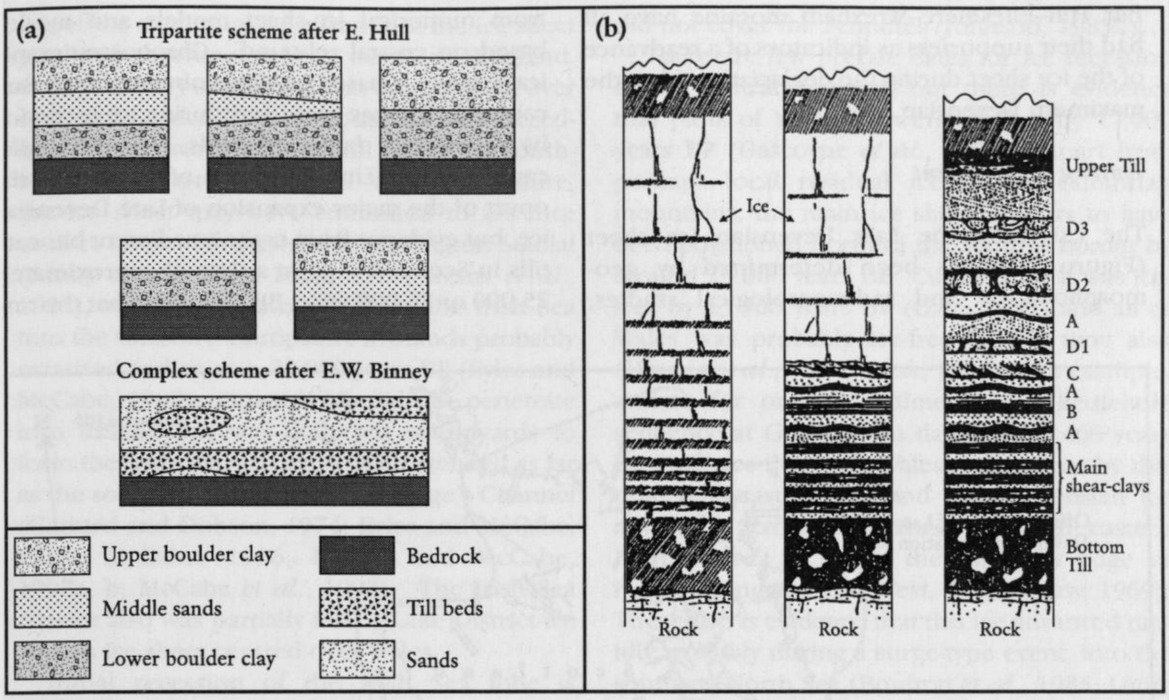


Figure 5.1 Models of glacial successions, northern England. (a) Tripartite scheme after Hull (1864) and the complex scheme after Binney (1848) (after Worsley, 1970). (b) Process of glacial undermelt from Carruthers (1953). In the left-hand diagram, ice melts by subglacial undermelting to give the sediment units (lettered) on the right, including 'bottom', 'banded' (shear-clays), and on the left englacial section, with the top 'overriding' dirts to give the Upper Till. The undermelting takes place *in situ* by liberation of sediment from the ice at the glacier base. (after Bennett and Doyle, 1994).

(1931) fitted the deposits into a model of frontal ice retreat and pro-glacial deposition.

The modern interpretation of tripartite drift sequences in northern England developed from the ideas of Boulton (1967, 1968, 1970, 1972) in Spitsbergen and subsequently in North Wales (1977), by Worsley (1967a, 1970, 1985) and G.S.P. Thomas (1989) in the Cheshire–Shropshire lowlands, and by the development of the supraglacial land-system model (Boulton and Paul, 1976; Eyles, 1983; and Paul, 1983). Hence the majority of glaciogenic sequences in northern England are now associated with the growth and decay of a single ice sheet (Evans and Arthurton, 1973). However, in the Solway lowlands, the Carlisle Plain and in coastal west and south Cumbria the sequence is regarded as more complex. Here an upper till has been recognized that was deposited by a late readvance of Scottish ice (Trotter, 1922, 1929; Trotter *et al.*, 1937; Huddart, 1970, 1971a, b, 1972, 1977, 1991, 1994). Huddart (1970, 1991, 1994, 1997) recognized an associated series of pro-glacial depositional environments, including pro-glacial lake and sandur. However, this till has also been interpreted as a glaciomarine mud drape, produced by the flocculation of clay from meltwater plumes discharging from calving, tidewater glaciers during the deglaciation of the Irish Sea basin (Eyles and McCabe, 1989, 1991). Despite having been dismissed as largely illusory by Evans and Arthurton (1973), Pennington (1978) and Thomas (1985b), the readvance concept has gained renewed support by the recognition of pro-glacial glaciotectonic deformation in west Cumbria during the Nirex investigations (e.g. Knight *et al.*, 1997; Wingfield *et al.* 1997; Akhurst *et al.* 1997; Browne *et al.*, 1997, Merritt and Auton, 2000), which was first recognized by Huddart (1970). There is still some discussion as to the number of readvances of the Irish Sea basin ice sheet and their impact on the stratigraphy and landforms. The 'Gosforth Oscillation' (Trotter *et al.*, 1937; Akhurst *et al.*, 1997; Browne *et al.*, 1997), the Low Furness Readvance (Huddart *et al.*, 1977), the Kirkham moraine of Gresswell (1967), discussed in Longworth (1985), the Delamere moraine and the

Bar Hill–Ellesmere–Wrexham moraine have all had their supporters as indicators of a readvance of the ice sheet during the deglaciation from the maximum Devensian ice limit.

Timing and extent

The extent of the Late Devensian ice sheet (Figure 5.2) has been determined by geomorphological and sedimentological studies, from numerical ice-sheet models and models based on crustal rebound. Chronostratigraphical control is based on dates obtained from surrounding ice-free areas and during ice recession by a variety of dating methods, primarily radiocarbon dates. Little is known of the date of the onset of the major expansion of Late Devensian ice, but evidence from organic sediment beneath tills in Scotland suggest a date of approximately 25 000 or 26 000 years BP. It is clear that the tim-

Figure 5.2 Limits of Anglian and Late Devensian glaciations in Great Britain (based on Eyles and Dearman, 1981; Eyles and McCabe, 1989; Boulton *et al.*, 1991).

ing of the maximum ice advance of the ice sheet was asynchronous across northern England. Sissons (1981) concluded that the ice sheet over northern Scotland was stationary or even receding at a time when ice was still advancing southwards over parts of England. In east Yorkshire, the ice sheet may have continued to advance until *c.* 18 000 BP during the Dimlington Stadial (Penny *et al.*, 1969; Rose, 1985, Evans *et al.*, 1995), whereas the advance from the Irish Sea into the Cheshire–Shropshire lowlands probably occurred earlier, at *c.* 22 000 years BP (Eyles and McCabe, 1989). Scottish ice did not penetrate into Ireland but was deflected southwards to form the Irish Sea Glacier, which extended as far as the southern entrance to St George's Channel (Garrard and Dobson, 1974; Eyles and McCabe, 1989; McCabe, 1996; Knight and McCabe, 1997a, b; McCabe *et al.*, 1998). The Irish Sea Glacier also was partially fed by Lake District ice and an ice sheet centred over Wales.

Initial recession of the Irish Sea lobe is thought to have been rapid, and by 18 000 years BP the ice margin lay across the Isle of Man, although the ^{14}C dates used in this supposition generally are thought to be too old because of the hardwater effect. It seems much more likely that the date is closer to 14 500 years BP (see McCabe *et al.*, 1998). Ice cover over Wales was dominated by a local ice cap that coalesced to the west and north with Irish Sea ice. Ice advanced across the Cheshire and Shropshire lowlands as far as the Wolverhampton area (Morgan, A.V., 1973; Bowen *et al.*, 1986; Worsley, 1991b). A more northerly end-moraine complex (the Wrexham–Ellesmere–Whitchurch moraine), believed by Boulton and Worsley (1965) to mark the maximum extent of the ice sheet in the Cheshire–Shropshire Plain, is now generally accepted to be the expression of a series of prolonged still-stands along this line (Thomas, G.S.P., 1989). The form of the Wrexham–Ellesmere–Whitchurch end-moraine complex indicates that the ice front was bilobate at this point, where it was divided by the Mid-Cheshire Ridge (Boulton and Worsley, 1965). Early workers recognized that the last ice sheet flowed out of the Irish Sea Basin in a NNW–SSE direction (Mackintosh, 1879; Morton, 1860, 1870) and this regional pattern of ice flow is now firmly established (Gresswell, 1964; Thomas, G.S.P, 1985a, 1989; Glasser and Hambrey, 1998). The eastern margin of Irish Sea ice penetration into the Cheshire–Shropshire area pushed against but did not cover the Pennines (Johnson, 1985b).

There are few precise dates for ice recession across central England, but there is evidence that parts of Yorkshire were ice-free by 17 000 years BP (Gascoyne *et al.*, 1983). Apart from perhaps local residual ice in the Cumbrian mountains, the main ice sheet appears to have been positioned north of the Scottish border by around 15 000 years BP. Cumbria itself was ice-free by 13 500 years BP (Gale, 1985) and all of Wales was probably ice-free at this time also (Haynes *et al.*, 1977; Musk, 1985). For example, a date for organic sediment in a kettlehole sequence at Glanllynnau dated to 14 468 years BP indicates that west Wales was ice-free by this time. In eastern England Late Devensian ice advanced across Holderness, through eastern Lincolnshire, as far as the northern edge of Norfolk (Suggate and West, 1959; Straw, 1969). There also is evidence that this ice advanced rapidly, possibly during a surge-type event, into the southern North Sea (Boulton *et al.*, 1985; Long, A.J. *et al.*, 1988; Eyles *et al.*, 1994; Evans *et al.*, 1995). This advance was not sustained and ice had receded from the offshore area soon after 18 000 years BP (Lambeck, 1993b).

Ice thickness

There are two contrasting views concerning ice thickness during the Last Glacial Maximum, exemplified by the 'maximum' and 'minimum' ice-sheet reconstructions of Boulton *et al.* (1977, 1985) (Figure 5.3). In their maximum model the ice thickness exceeds 1800 m and the ice sheet covers the highest peaks of all mountains in England. The geomorphological evidence of striae, erratic transport and periglacial trimlines, however, makes a convincing case that the ice sheet was much thinner than the maximum estimate. In mountainous areas such as the Lake District it therefore is entirely possible that the highest summits remained exposed as nunataks (Lamb and Ballantyne, 1998). In addition, reconstruction of the ice-sheet surface elevation based on meltwater channel altitudes in the Cheshire lowlands suggests an ice-surface elevation over the Irish Sea of *c.* 700 m (Glasser and Sambrook Smith, 1999).

Ice-sheet flow patterns and dimensions

Mitchell and Clark (1994) reviewed the evidence upon which reconstructions of the Late Devens-

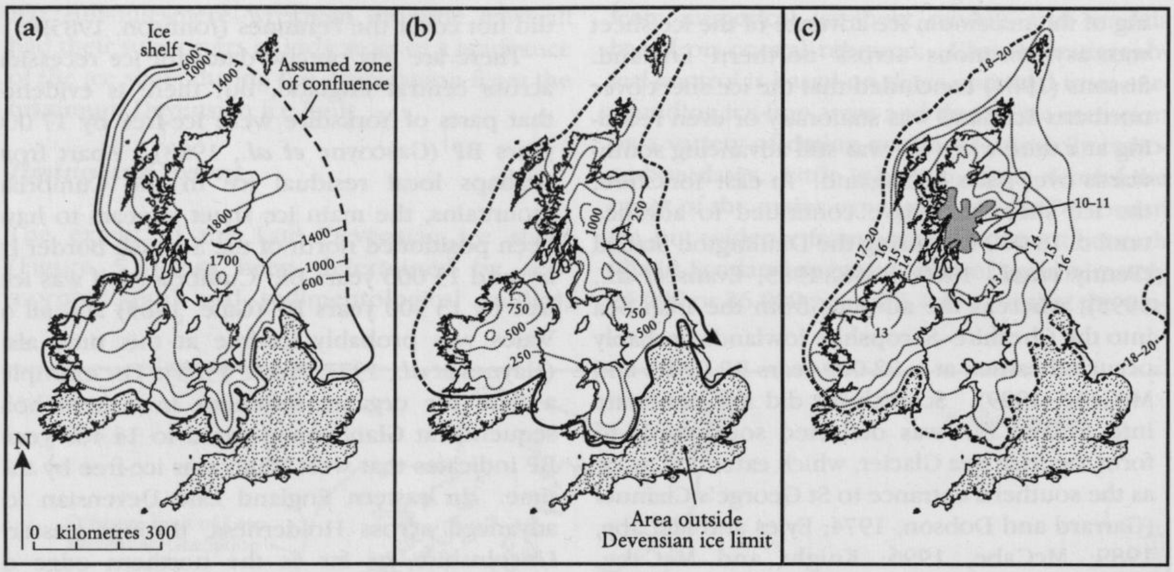


Figure 5.3 The modelled ice sheet at the time of the Last Glacial Maximum (ice thickness in metres): (a) according to Boulton *et al.* (1977); (b) according to Boulton *et al.* (1985). (c) Isochrons (ka) for recession of the last ice sheet according to Andersen (1981). The extent of the Loch Lomond (Younger Dryas) icefield (11–10 ka) is shaded.

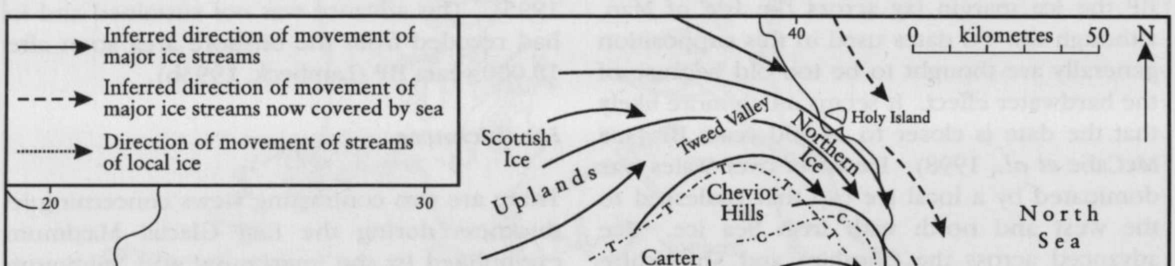


Figure 5.4a Suggested Late Devensian ice movements in northern England (after Taylor *et al.*, 1971).

ian ice sheet are based. Surprisingly little is known about detailed flow directions and dynamics of this ice sheet. Reconstructions of the entire ice sheet (Boulton *et al.*,1985; Bowen, 1991) are highly generalized. Regional reconstructions of the ice sheet in northern England (Figure 5.4a, b), such as those of Taylor *et al.* (1971), Letzer (1987), Johnson (1985a), Catt (1991a, b) and Douglas (1991), are often based on the early mapping work of the [British] Geological Survey. These early workers used striations, erratic dispersal, till distribution and drumlin distribution to identify the large-scale

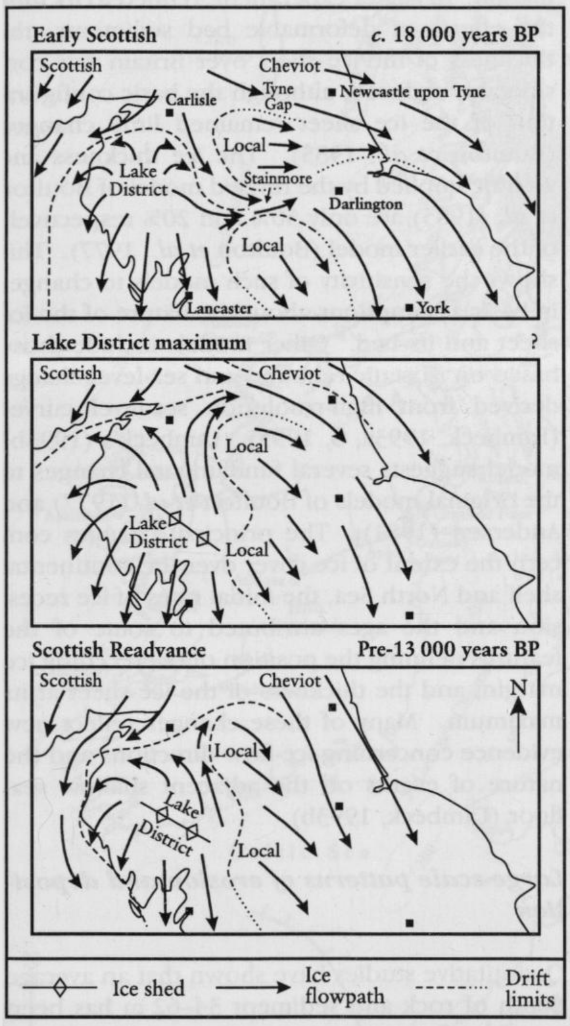


Figure 5.4b Suggested Late Devensian ice movements in northern England: generalized movements at various time periods after Letzer (1981). Note that the Early Scottish could be Early Devensian or Wolstonian and that the ice movement directions for the Scottish Readvance are incorrect (see Huddart, 1970, 1991, 1994, 1997).

patterns of ice-flow and ice-movement directions. Local centres of glaciation were identified in the north-west Yorkshire Dales, the Alston Block and the Lake District (Ward, 1873; Goodchild, 1875; Aveline and Hughes, 1888, Dakyns *et al.*, 1890, 1891). As it was generally accepted at this time that an ice centre over the Scottish Highlands dominated the last ice sheet, these areas were considered to support small ice domes that made only a minor contribution to patterns of ice dispersal. This picture was accepted and later modified by subsequent workers (Raistrick, 1926, 1933; Trotter, 1929; Hollingworth, 1931). More recent reconstructions of the ice sheet based on detailed mapping of subglacial bedforms are not always compatible with the large-scale models (Mitchell, 1994). Mitchell and Clark (1994) have argued that a major ice divide extended south-east from the central Lake District through the Howgill Fells into the adjacent Pennines. They suggested that the axis of this ice divide might have migrated during the life of the ice sheet.

Heinrich events/surge behaviour

Heinrich (H) events are the regularly occurring periods of iceberg production identified in North Atlantic Ocean sediments during the past 100 000 years (Heinrich, 1988; Andrews *et al.*, 1998). Each Heinrich event is associated with an increase in ice-rafted debris in the North Atlantic and these events have been proposed as the trigger for major climatic changes in the North Atlantic region (Broecker, 1994). Recent studies of Greenland ice cores and North Atlantic sediment cores also suggest that the last deglacial cycle (*c.* 21 000–13 000 years BP) was interrupted by a series of millennial-scale climate shifts (Dansgaard *et al.*, 1993; Bond and Lotti, 1995). These climate shifts punctuated the overall recession of the mid-latitude ice sheets and seem to be climate-driven, because discharges involve more than one ice sheet (Clark *et al.*, 1995; Bond and Lotti, 1995). McCabe (1996) first attempted to link episodes of fast ice flow ('drumlinization') in Britain to these millennial-scale discharge events. His work suggested that individual phases of drumlinization are correlated with the climate-driven circum-North Atlantic climate events. Subsequently, McCabe *et al.* (1998) identified a period of rapid bed reorganization beneath the last British ice sheet around 14 000 years BP, younger than was envisaged previously. This phase of fast ice flow involved overprinting and drumlinization of earlier transverse landforms by ice streams that fed into the Irish Sea to form tidewater calving fronts. McCabe *et al.* (1998) correlated this event with other evidence from around the North Atlantic area that suggests there was a widespread climatic event around this time. They consider this to be evidence that the British ice sheet participated in Heinrich I event (Figure 5.5). Recent results from interpretation of satellite imagery and accelerator mass spectrometry (AMS) ^{14}C dating of marine microfossils suggest that glaciogenic deposits also reflect millennial-scale oscillations of ice masses, shifts in ice-divide locations and major changes in ice-flow directions (McCabe, 1996; Knight and McCabe, 1997a, b; McCabe and Clark, 1998; McCabe *et al.*, 1998). This work on glacial bedforms is a promising avenue in reconstructing former mid-latitude ice sheets.

Basal thermal regime

Very little is known about the basal thermal regime of the Late Devensian ice sheet in northern England. No detailed calculations of ice-sheet basal thermal regime have been attempted as they have for the Late Devensian ice sheet in parts of Scotland (Gordon, 1979; Glasser, 1995). At its maximum extent, Boulton *et al.* (1977) considered the last ice sheet to have a frozen bed in its interior with a temperate margin, but provided little data on a regional scale. Clearly this thermal regime is not compatible with much of the geomorphological evidence for fast ice flow (Mitchell, 1994). Equally, little is known about how the basal thermal regime of the ice sheet changed during growth and decay. It is probable that the plateau ice fields that existed at times in areas such as the Lake District were predominantly cold-based (Rea *et al.*, 1998). G.S.P. Thomas (1989) also suggested a shift in the basal thermal regime during the Devensian ice retreat to account for cold-based supraglacial sedimentation in the south Cheshire–Shropshire lowlands and temperate-based ice in the north Cheshire–Lancashire area.

Contribution from numerical models

The most comprehensive models of the last British ice sheet are those of Andersen (1981), Boulton *et al.* (1977) and Boulton *et al.* (1985) (Figure 5.3). Andersen (1981) estimated isochrons for the recession of the ice sheet from its maximum extent at 20 000–18 000 years BP. Boulton *et al.* (1977) used geomorphological evidence for the maximum extent of ice cover, basic climatic modelling and the mechanical flow laws for ice to model the British ice sheet at its maximum. They predicted surface topography, patterns of ice-sheet movement, the direction of flow lines and the distribution of balance velocities, basal and surface temperatures for the ice sheet. Their results indicated an ice sheet with a summit height of 1800 m and velocities in the marginal area of between 150 and 500 m year^{-1}, with a central frozen bed and temperate margin. In a later experiment, refined to include the effects of deformable bed sediments, the thickness of the ice sheet over Britain was considerably reduced, although the basic configuration of the ice sheet remained little changed (Boulton *et al.*, 1985). The ice thickness and volume implied by the revised model of Boulton *et al.* (1985) are only 50% and 20% respectively of the earlier model (Boulton *et al.*, 1977). This shows the sensitivity of such models to changes in basic assumptions about the nature of the ice sheet and its bed. Other models include those based on isostatic rebound and sea-level change derived from high-resolution sea-level curves (Lambeck, 1993a, b, 1995). Lambeck's (1993b) model suggests several fundamental changes to the original models of Boulton *et al.* (1977) and Andersen (1981). The principal changes concern the extent of ice cover over the continental shelf and North Sea, the initial rates of ice recession and the ages attributed to some of the features defining the position of the receding ice margin, and the thickness of the ice sheet at its maximum. Many of these changes reflect new evidence concerning ice-flow directions and the nature of events on the adjacent shallow seafloor (Lambeck, 1993b).

Large-scale patterns of erosion and deposition

Quantitative studies have shown that an average depth of rock and sediment 34–62 m has been removed from the British landscape by glacial erosion (Glasser and Hall, 1997), rising to 125–155 m if erosion of sediment on the continental shelf is included (Clayton, 1996, 1997). Thus rates of glacial erosion locally may exceed those of non-glacial processes, although of

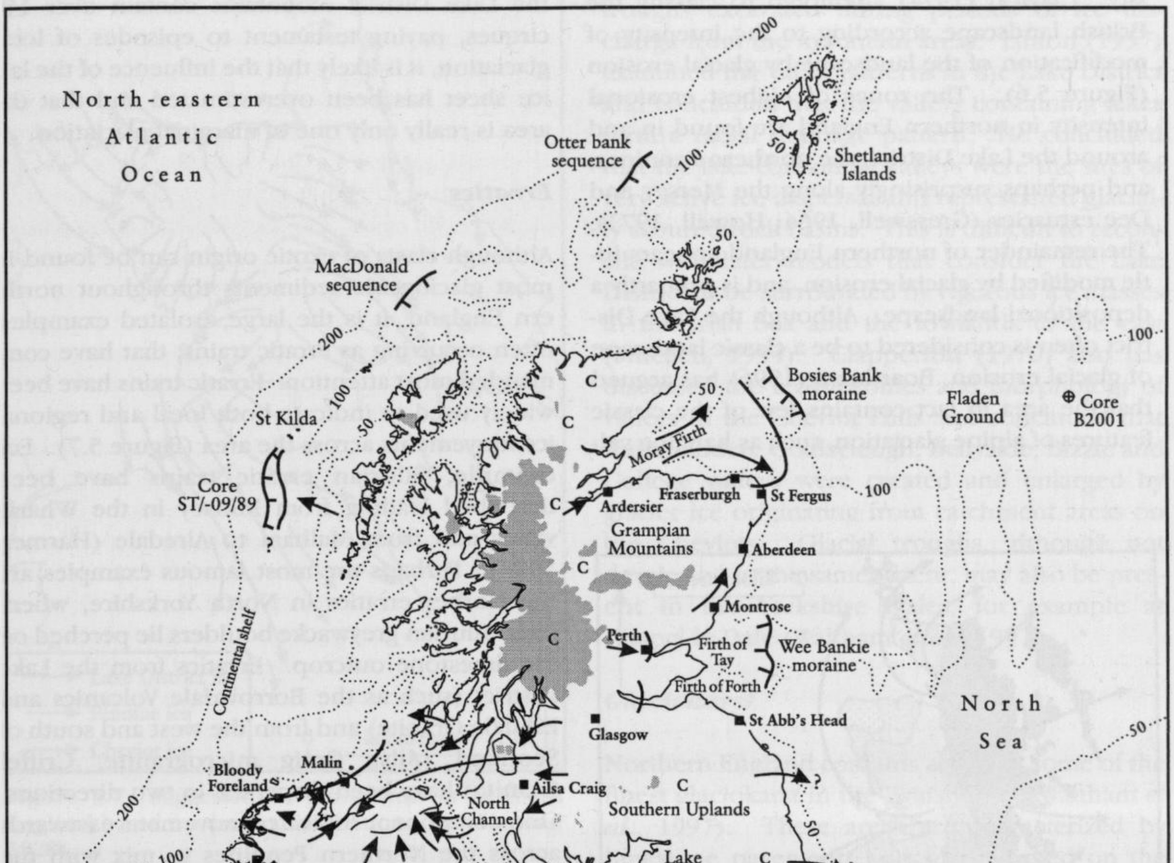


Figure 5.5 Generalized ice flows and ice limits during the Heinrich I event in northern Britain (after McCabe *et al.*, 1998). Note that the ice flows in Cumbria and east Yorkshire are considered to be incorrect by the present authors.

course there is a wide regional variation in intensity. Clayton (1974) attempted to classify the British landscape according to the intensity of modification of the landscape by glacial erosion (Figure 5.6). The zones of highest erosional intensity in northern England are found in and around the Lake District, the northern Pennines, and perhaps surprisingly along the Mersey and Dee estuaries (Gresswell, 1964; Howell, 1973). The remainder of northern England appears little modified by glacial erosion, and is primarily a depositional landscape. Although the Lake District often is considered to be a classic landscape of glacial erosion, Boardman (1996) has argued that the area in fact contains few of the classic features of alpine glaciation, such as hanging valleys, or terminal and lateral moraines. Although the Lake District mountains contain over 150 cirques, paying testament to episodes of local glaciation, it is likely that the influence of the last ice sheet has been overestimated and that the area is really only one of marginal glaciation.

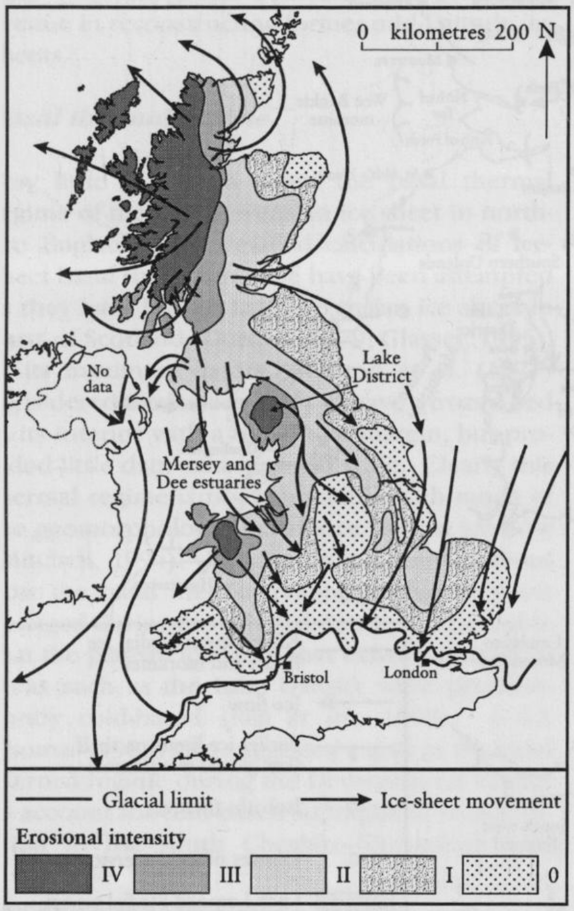


Figure 5.6 Patterns of ice-sheet movement and glacial erosion. The five zones of erosional intensity are: 0, no glacial erosion; I, limited glacial erosion; II, glacial erosion confined to major flowlines; III, widespread glacial erosion; IV, very extensive glacial erosion, no trace of the pre-glacial landscape (after Clayton, 1974).

Erratics

Although clasts of exotic origin can be found in most glaciogenic sediments throughout northern England, it is the large, isolated examples, often occurring as erratic trains, that have commanded most attention. Erratic trains have been widely used to indicate both local and regional ice movements across the area (Figure 5.7). For example, Silurian erratic trains have been described leading from Kilnsey in the Wharfe valley and from Malham to Airedale (Harmer, 1928). Perhaps the most famous examples are the Norber erratics in North Yorkshire, where large Silurian greywacke boulders lie perched on the limestone outcrop. Erratics from the Lake District (such as the Borrowdale Volcanics and Eskdale Granite) and from the west and south of Scotland (Ailsa Craig microgranite, Criffel Granite) have been dispersed in two directions. One component indicates movement eastwards across the Northern Pennines to mix with the Cheviot rock types, whereas the other spreads over Lancashire, Cheshire and into the Midlands from the Irish Sea. Scottish ice is thought to have coalesced with Lake District ice that flowed radially out of the valleys and this Scottish ice was deflected around the northern Lake District. Evidence for this is the lack of erratics in the Lake District itself (Mitchell and Clark, 1994). Mitchell and Buggie (1991) and Mitchell (1996) used erratics of the Bluecaster dolerite on the eastern flanks of the Howgill Fells, initially transported by glaciers but subsequently incorporated into drystone walls, to map the ice-flow directions (Shakesby, 1977; Letzer, 1978). The erratic density away from the Bluecaster dolerite outdrop indicates a northerly movement of the ice divide. This explains the presence of the dolerite both in the Kirkby Stephen area to the north-west and to the south-west down the Rawthey valley.

Glacial erosion in the uplands

Landforms of glacial erosion in the uplands are confined mainly to the Lake District and

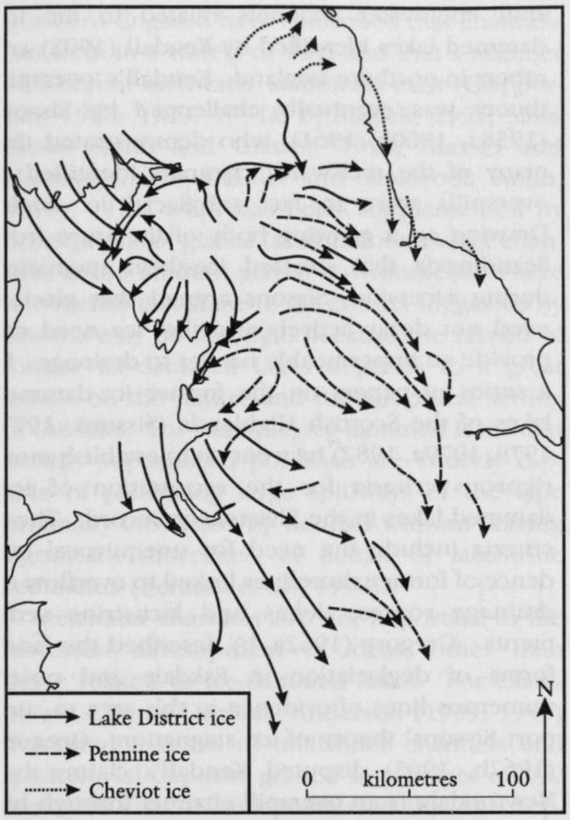


Figure 5.7 Suggested ice movements in northern England based on erratic distribution (after Harmer, 1928).

northern Pennines, although there are outlying features in areas such as the Cheviot Hills. Depending on definition, the Lake District hosts up to 158 cirques (Marr, 1916; Hay, 1934; Manley, 1959; Temple, 1965; Clough, 1977; Sissons, 1980; Evans, 1987; Evans and Cox, 1995; Rea *et al*., 1998). Many of these cirques supported small glaciers during the Loch Lomond Stadial (Younger Dryas), and during other previous episodes of marginal glaciation. Well-developed cirques are rare outside the Lake District, although cirque-like features and sites favourable for glacier growth exist in several locations in the western Pennines (Rowell and Turner, 1952; Manley, 1959; Mitchell, 1991b, d, 1996; Wilson and Clark, 1995) and in the Peak District (Johnson *et al*., 1990). Farther east, in the Cheviot Hills, two ill-defined cirques are cut into the western flank of the massif at Hen Hole and Bizzle (Clapperton, 1970). Much of the work on these features is purely descriptive in nature and the age of the cirques and their relationship to ice-sheet erosion in this area remains unknown.

A related group of features are the glacial troughs excavated during periods of ice discharge from the mountain areas. Linton (1957) examined the valley patterns in the Lake District and concluded that the valleys containing lakes form a radial drainage pattern. He concluded that the lake-containing valleys were the sites of very active ice dispersal and represented glacially scoured rock basins. This is difficult to reconcile with later models that consider the Lake District to be surrounded by vigorous ice masses in the Irish Sea and the lowlands to the east (Mitchell, 1994). Clapperton (1970) also has discussed the cross-profiles and morphology of valleys in the Cheviot Hills and concluded that the Lambden, Goldscleugh, Bellyside, Bizzle and College valleys were created and enlarged by glacier ice originating from catchment areas on the Cheviots. Glacial troughs, although not developed to the same extent, may also be present in the Yorkshire Dales, for example at Chapel-le-Dale (Waltham *et al*., 1997).

Glaciokarst

Northern England contains arguably some of the finest glaciokarst in the British Isles (Waltham *et al*., 1997). These areas are characterized by limestone pavements and scars formed on the tops and edges of the outcrops of more resistant limestone beds. Post-glacial solutional features, such as karren, fret the upper surfaces of the limestone pavements and commonly there is little post-glacial soil development. Deep gorges formed by glacial meltwater also are common. Outstanding examples of this type of landscape can be found in the Yorkshire Dales (Goldie, 1973, 1981; Waltham and Tillotson, 1989), with less extensive tracts of glaciokarst in Lancashire and Cumbria (Goldie, 1996). These areas of limestone pavement are important both geologically and biologically (Ward and Evans, 1976).

Although descriptions of glaciokarst are numerous in textbooks on geomorphology (e.g. Sweeting, 1972; Jennings, 1985; Trudgill, 1985), the subject has been surprisingly neglected in modern research literature. Most accounts of these areas talk generally about the effects of 'glacial scouring', 'glacial plucking' and of vast areas 'scraped clean by glaciers'. Descriptions are primarily qualitative and quantitative process studies are rare. There is seldom any consideration of the wider ramifications of these statements, such as the implications for the spatial variation in ice-sheet thermal regime or the glaciological conditions required to initiate widespread erosion beneath an ice sheet overlying porous bedrock. There does appear, however, to be a positive correlation between areas of the British Isles affected by Pleistocene glaciation and the location of pavements (Piggott, 1965; Williams, 1966; Clayton, 1981).

Alternative theories argue that limestone pavements are largely exhumed palaeokarst surfaces that developed beneath former soil covers and were merely exposed by glacial erosion (Vincent, 1995; Vanstone, 1998). Pavement formation therefore would be controlled by the location of palaeokarst surfaces formed during Dinantian marine transgression and recession (Ramsbottom, 1973, 1977; Wright, 1982; Adams and Horbury, 1989). Limestone pavements are therefore not exclusively Quaternary features but owe many of their attributes to the cyclic nature of Dinantian shelf sedimentation, displaying rejuvenated palaeokarstic forms (Vincent, 1995). These competing theories concerning the formation of limestone pavements remain speculative and largely untested, although clearly they have important implications for the nature and efficiency of Quaternary glacial erosion.

Glacial meltwater erosion

Meltwater behaviour beneath modern glaciers is poorly understood, so it is not surprising that the picture of meltwater flow beneath Quaternary ice masses also is incomplete. Clues to these patterns of meltwater behaviour in the Quaternary record are given by the existence of glacial meltwater channels, normally recognized by their abrupt inception and termination and lack of a significant modern catchment, and by accumulations of glaciofluvial sediment. Meltwater channels are probably one of the most-mapped glacial geomorphological features, yet their glaciological implications are rarely explored fully or fully utilized in ice-sheet reconstruction. Meltwater channels occur as isolated examples, such as Newtondale, or as part of an integrated network of channels recording ice-sheet recession, such as those that are incised into the flanks of the Pennines at Ludworth Intake (Johnson, 1965a), and Humbleton Hill in the Cheviot Hills (Clapperton, 1971b).

One of the most contentious debates in the British Quaternary concerns the origin of 'overspill' meltwater channels related to the ice-dammed lakes identified by Kendall (1902) and others in northern England. Kendall's 'overspill' theory was eventually challenged by Sissons (1958a, 1960b, 1961), who demonstrated that many of the meltwater channels identified as overspills were in fact subglacial in origin. Drawing on a growing body of literature from Scandinavia that stressed ice-sheet stagnation during recession, Sissons argued that glaciers need not decay actively and that ice need not provide an impermeable barrier to drainage. In a series of papers on the former ice-dammed lakes of the Scottish Highlands (Sissons, 1977, 1978, 1979a, 1982) he went on to establish more rigorous criteria for the recognition of ice-dammed lakes in the Pleistocene record. These criteria include the need for unequivocal evidence of former shorelines linked to overflow or drainage routes, deltas and lacustrine sediments. Gregory (1962a, b) described the landforms of deglaciation in Eskdale and noted numerous lines of evidence in this area to support Sissons' theory of ice stagnation. Gregory (1962b, 1965) disputed Kendall's claims that Newtondale is an overspill channel through his study of the glacial lakes of Eskdale, to the north of the Newtondale channel. His work showed that much of the meltwater in fact drained subglacially to the north-east. Gregory (1965) noted that of the four lines of evidence used by Kendall to support the existence of a pro-glacial lake (overflow channels, lacustrine deposits, deltas and lake shorelines) there was little of this evidence to support the existence of a pro-glacial lake in Eskdale. This led Gregory to propose that although Newtondale is undoubtedly a meltwater channel, it is unlikely that it was created purely by overflow of an Eskdale–Kildale glacier lake during the Devensian. Support for this theory comes from the gravel spread at the southern end of the channel. Catt (1977c) suggests that the size of this gravel spread is anomalously small and it is too fine-grained to have resulted from the high discharge required to cut a channel the size of Newtondale. Catt (1977c) therefore argued that although Newtondale may have been used by meltwater during the Late Devensian, it was probably cut in a pre-Devensian glaciation. There is, however, no lithostratigraphical or chronostratigraphical evidence to prove this theory.

Since the 1960s, descriptions and interpretations of other meltwater-channel systems in northern England have confirmed that channels can form in a variety of ways and that a number of different meltwater landforms exist (Clapperton, 1968; 1970, 1971a, b; Russell, 1995; Sambrook Smith and Glasser, 1998; Glasser and Hambrey, 1998; Glasser and Sambrook Smith, 1999). This work has been supplemented by descriptions of glacial lakes and meltwater channels in the modern glacial environment. There is evidence that some of the criteria suggested by Sissons may be too rigid because the record of former ice-dammed lakes depends to a great extent on the nature and topographical setting of the lake. For example, ice-dammed lakes that occupy supraglacial positions are entirely capable of producing large spillways of the type originally envisaged by Kendall without leaving significant shorelines, or deltaic or lacustrine sediments (Bennett *et al.*, 1998).

Meltwater channels also are important to the historical development of ideas, other than those related to ice-dammed lakes. For example, in County Durham, Anderson (1939, 1940) described a series of meltwater channels and glaciofluvial deposits graded to a height of 190 feet (119 m) OD and postulated that this represented the height of the Late-glacial sea level. Peel (1949, 1956) described 'up and down' channels in Northumberland at Beldon Cleugh and East Dipton and suggested for the first time a possible subglacial (as opposed to subaerial lake overflow) origin for the channels. This theme was later developed by Sissons in a series of papers (Sissons, 1958a, 1960b, 1961) and used in particular to explain the formation of many of the meltwater channels of the Tyne Gap area (Sissons, 1958b). The tunnel valleys of Merseyside described by Gresswell (1964) and Howell (1973) are likely to have formed by subglacial meltwater erosion, perhaps modifying the effects of large-scale glacial erosion in the lowlands.

Glaciofluvial sediments and landforms

There is a wide variety of glaciofluvial landforms and sediments in northern England, associated with several types of topographical and environmental locations:

1. Lowland outwash plain or sandur systems that are associated with pro-glacial deposition in front of ice sheets, often associated with terminal, or retreat stage, moraines. Examples include part of the St Bees glaciogenic sequence, the Harrington outwash (Huddart, 1970, 1991) and the Black Combe coastal plain sandur. There are relatively few sandur sequences associated with the large, mid-Cheshire moraine systems, which suggests that they built into pro-glacial lakes, although there are some examples, such as Prees Heath (Thomas, G.S.P, 1989).
2. Valley sandar in the valleys draining the Lake District and Pennines, for example the gravel sequence in the Cumbrian Derwent valley (Huddart, 1970, 1971b), aggraded as ice moved out of the upland areas. In the north-east, outwash, often terraced, and ice-contact landforms are aligned along the major river valleys of the Tyne, Aln, Coquet, Wansbeck and Wear. An example described from the Durham Derwent by Allen and Rose (1986) shows the interplay between topography and the characteristics of the melting ice sheet in controlling the geometry and location of glaciofluvial deposition. Other examples of dissected valley sandar occur in the eastern Dales valleys, such as the Tees, Gilling Beck, Swale, Wensleydale, the Aire valley below Leeds and the Ouse valley between York and Goole.
3. Relatively isolated areas of esker deposition associated with lowland areas of lodgement till deposition, such as the Bradford Kames, the eskers associated with Aqualate Mere, or the Thursby eskers (Huddart, 1970, 1973); all appear to have been associated with deposition in pro-glacial lakes.
4. Terrain that is almost entirely the result of ice-sheet stagnation and wastage, with many ice-contact glaciofluvial landforms. Some of the best developed are on the margins of the Cheviot Hills near Wooler and to the south of Cornhill in the Tweed valley. Clapperton (1971b) interpreted a complex network of eskers, kames, flat-topped terraces and kettle-holes as the result of downwasting ice in the topographically controlled basins between the Cheviot Hills and the Carboniferous cuestas. Other examples are described from the Brampton kame belt in Edenside, Cumbria, where there is evidence for topographically controlled ice-sheet stagnation during deglaciation (Huddart, 1970, 1981c, 1983). Similar landform systems are found in mid-Cheshire (Thomas, 1985a).

Glaciofluvial landforms and sediments: some examples in northern England

Pro-glacial sandur, Harrington, west Cumbria

Evidence for the advancing Scottish Readvance ice front came from the former gravel pit described in Huddart (1970, 1991) and Huddart and Tooley (1972). Here there was an increasing grain size, an increase in gravel unit thickness, a decrease in grain roundness and change in bedform from distal to more proximal sandur environments up-section. All the palaeocurrent and lithological evidence indicated deposition from the west to north-west, as compared with an ice-flow direction to the south-west indicated by the drumlinized terrain to the east (Figure 5.8a, b). Similar sequences at St Bees have been interpreted as a catastrophic jökulhlaup deposit (Merritt, 1997b) and as very proximal pro-glacial sandur sediments (Huddart, 1970). Sandur sequences associated with ice readvances also have been described from Peel Place and Newton pits, near Gosforth, west Cumbria by Auton (1997a, b).

Baronwood–Low Plains deltas and eskers

In the only major col in the Penrith Sandstone escarpment between Beacon Hill and Barrock Fell are a series of glaciofluvial landforms (Figure 5.9). A subglacial channel can be traced from the Petteril valley to Pearsgill, and to the east a series of sinuous ridges 5–10 m high can be followed through the col as far as Abbott Moss. Between this moss and Baronwood Farm is a bedrock divide, which has been breached by glacial drainage channels. To the west of these channels are a series of flat-topped deposits grading into the channels, and to the east are kettleholed, boulder-strewn sediments overlooking the Eden valley. These have been interpreted by Hollingworth (1931) as deltas prograding into a 120 m OD lake. In the 1960s, pits in the Low Plains sinuous ridges revealed the succession in Figure 5.10a, and the ridges were interpreted as a response to subglacial erosion of the bedrock and till, and infilling of subglacial tunnels. A threefold sedimentological sequence was exposed. In the ridge centre were channel gravel facies, fining upwards to cross-stratified and rippled sands. Laterally on the ridge flanks were marginal sand facies composed of cross-stratified and rippled sands. The backwater silt–clay facies infilled channels with parallel laminated silts and clays and rippled, fine-grained sand. A model of the development of the ridges is illustrated in Figure 5.10b. During stage A a subglacial river flowed under hydrostatic pressure out of the Petteril through the Baronwood col. Flow was mainly in the upper flow regime with the channels probably cut in the early summer flood period. The marginal sand facies, together with the suspension fills of many of the channels, indicate overbank flood-stage deposition. Sand would be deposited in shallow streams outside the main channel, with the silts and fine sands deposited from suspension in lagoons occupying the abandoned channels. During the development of the ridge the main channel changed position, eroded channels and the subglacial tunnel was widened. Adjacent to the ice mass in the col were ice-marginal streams, which built up several kame terraces between the downwasting ice and the valley side to the south. Similar sinuous ridges have been interpreted as subglacial eskers in Edenside, including the high-level Hallbankgate esker and megachannel, which took meltwater into the South Tyne system at the 180 m (above OD) stage, the Gilsland esker system and channels, taking drainage into the Tyne valley at the 120–180 m stages, and the Edenhall to Great Salkeld esker (Huddart, 1970, 1981c).

Ice-walled lake deposition in the Eden valley

Two major depositional levels have been noted in the Eden valley at 129 m and 99 m OD, although higher, lower and intermediate stages are known. At Baronwood there is a complex delta with three superimposed sequences of topsets, foresets and bottomsets (Figure 5.11). In stage A, ponded water accumulated in an ice-walled depression up to 129 m OD and the Low Plains subglacial river deposited a delta into this ice-walled lake. As the ice downmelted the englacial water table was lowered to 121–123 m OD, the input rivers regraded to the new level, cut channels and deposited new foresets on the stage A sediments. Much of the earlier sediment would be preserved as the downcutting rivers eroded deep gullies. With further downmelting a major change in the position of the englacial water table occurred and a large volume of meltwater escaped subglacially to the north, leaving a much lower lake level at stage C. The best evidence for lakes at 99 and 94 m OD occurs between Lazonby and Eden Lacy, where the structure, sedimentary sequence and

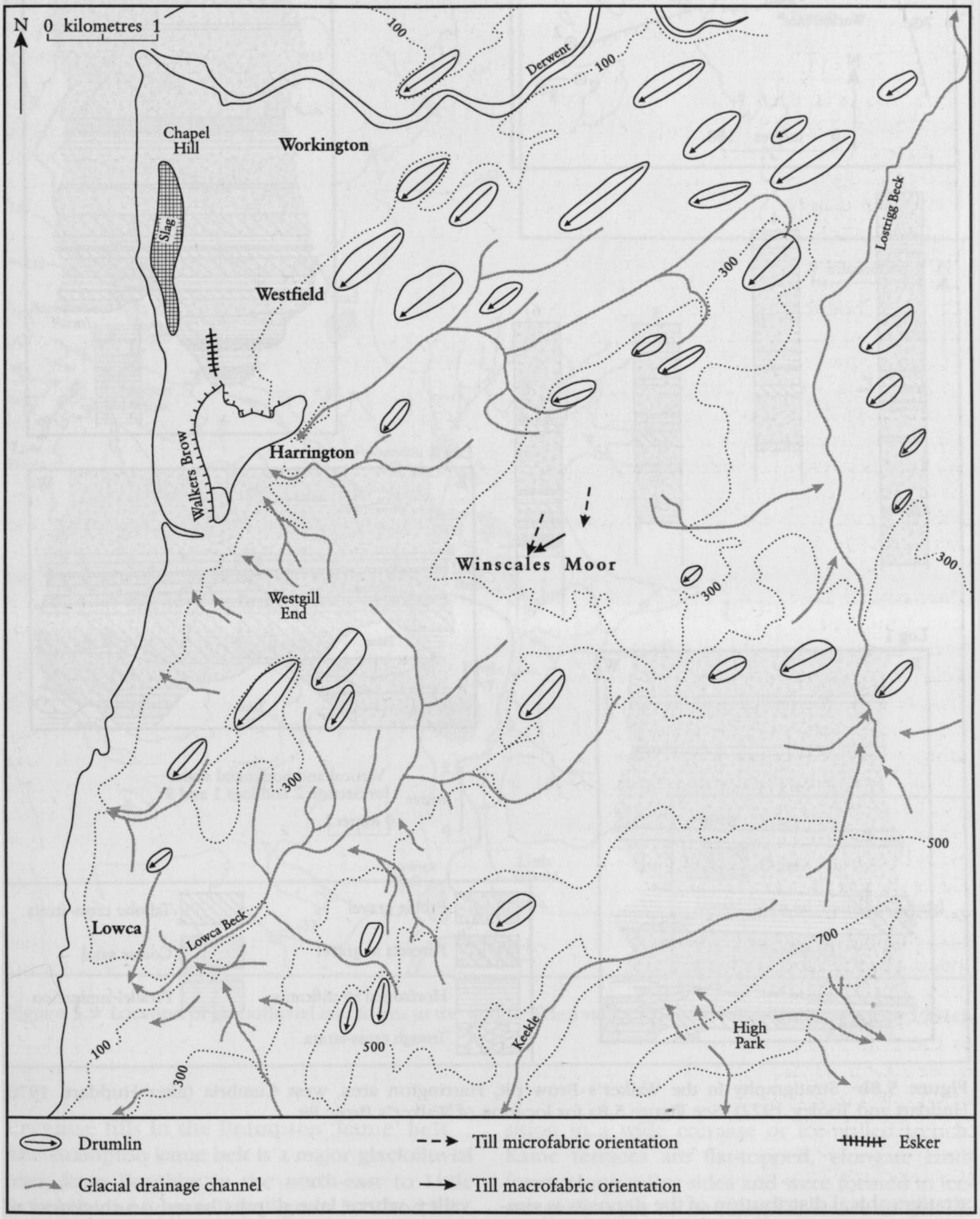


Figure 5.8a Glacial landforms in the Harrington area, west Cumbria (after Huddart, 1970; Huddart and Tooley, 1972). See Figure 5.8b for stratigraphy of the Walker's Brow pit.

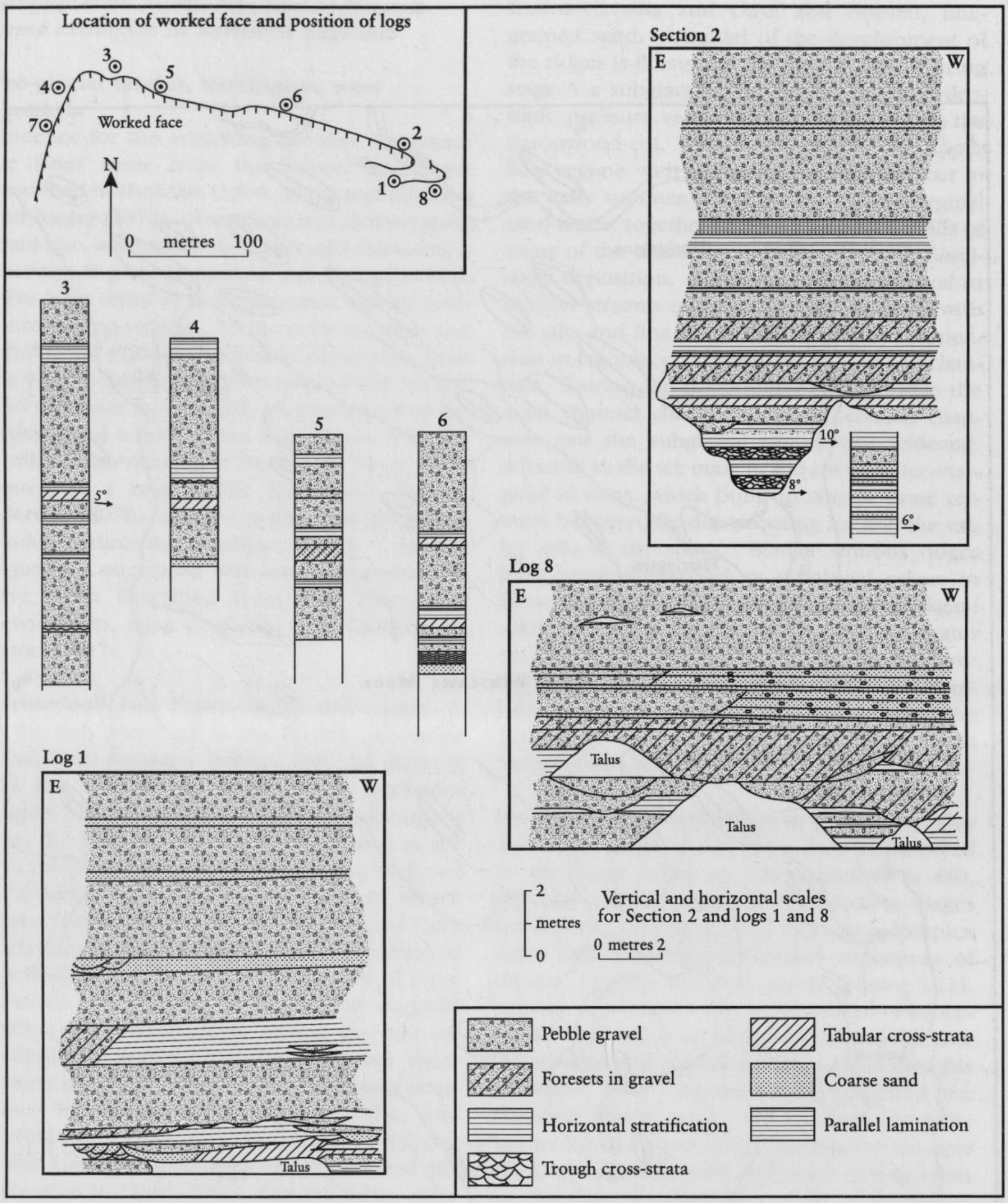


Figure 5.8b Stratigraphy in the Walker's Brow pit, Harrington area, west Cumbria (after Huddart, 1970; Huddart and Tooley, 1972). See Figure 5.8a for location of Walker's Brow Pit.

stratigraphical distribution of the deposits is similar to the Baronwood example (Huddart, 1970, 1981c). It is considered that these deposits accumulated in ice-walled lakes in downwasting stagnant ice in the middle section of the Eden valley, where lake depths based on thickness of foresets appear to have been between 4 and 8 m. Clay and silt, the deeper water sediments from these lakes, mantles much of the lower Eden valley below 90 m.

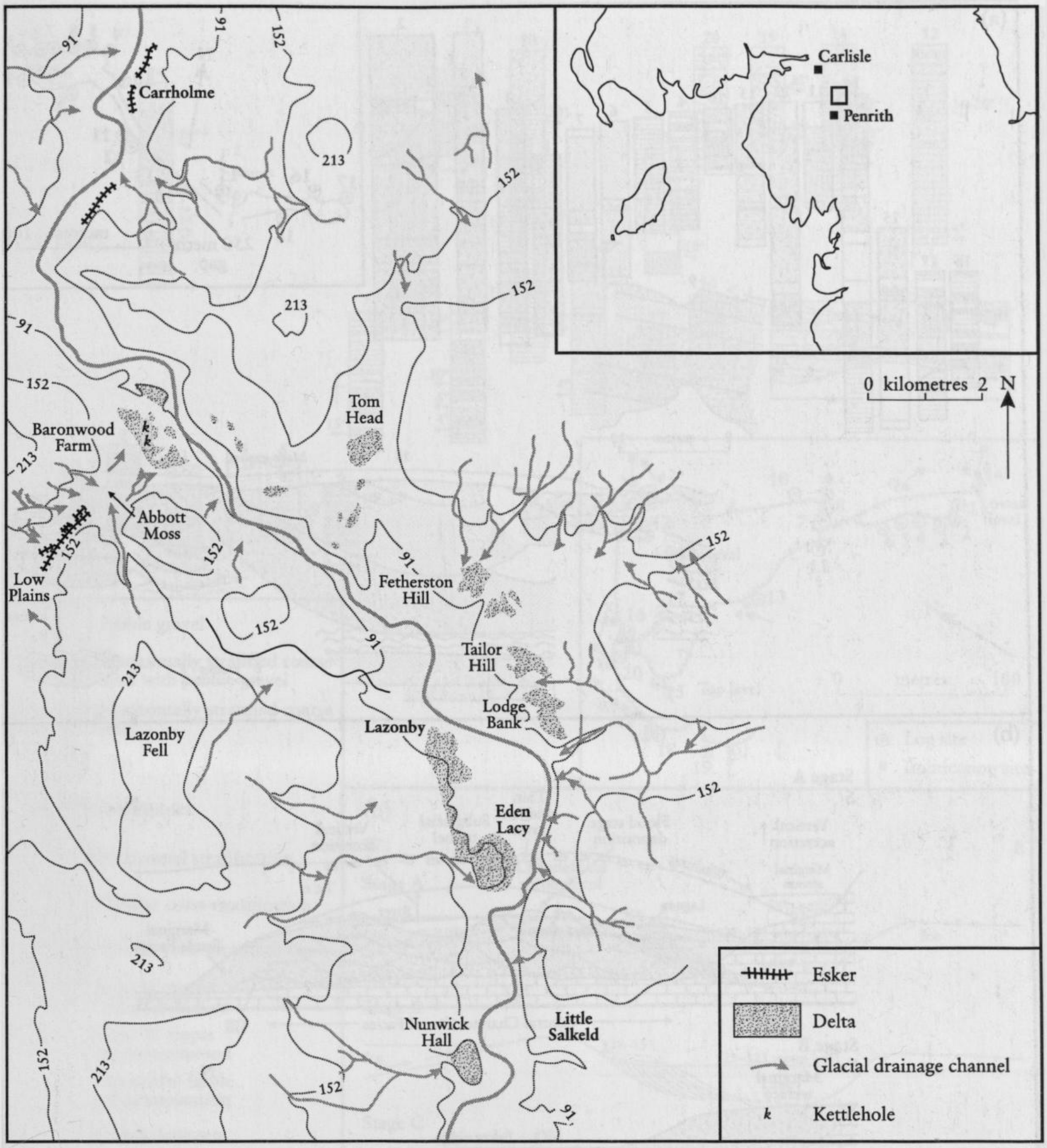


Figure 5.9 Location of glaciofluvial landforms in the middle Eden valley, Cumbria (after Huddart, 1970, 1981c).

Crevasse fills in the Brampton 'kame' belt

The Brampton kame belt is a major glaciofluvial tract from Boothby in the north-east to Little Salkeld in the Eden valley (Figures 5.12 and 5.9). In the northern part it is 5 km wide, between the 210 and 60 m contours. Flat-topped hills or kames were probably formed by meltwater filling in a small, ice-walled lake or by fluvial deposition in a wide crevasse or ice-walled trench. Kame terraces are flat-topped, elongate landforms along valley sides and were formed in ice-marginal lakes or by marginal rivers as the ice downwasted. As on the slopes above Talkin Tarn they commonly form a sequence as the ice downwasted. Glaciofluvial sediment can accumulate in englacial or supraglacial holes and

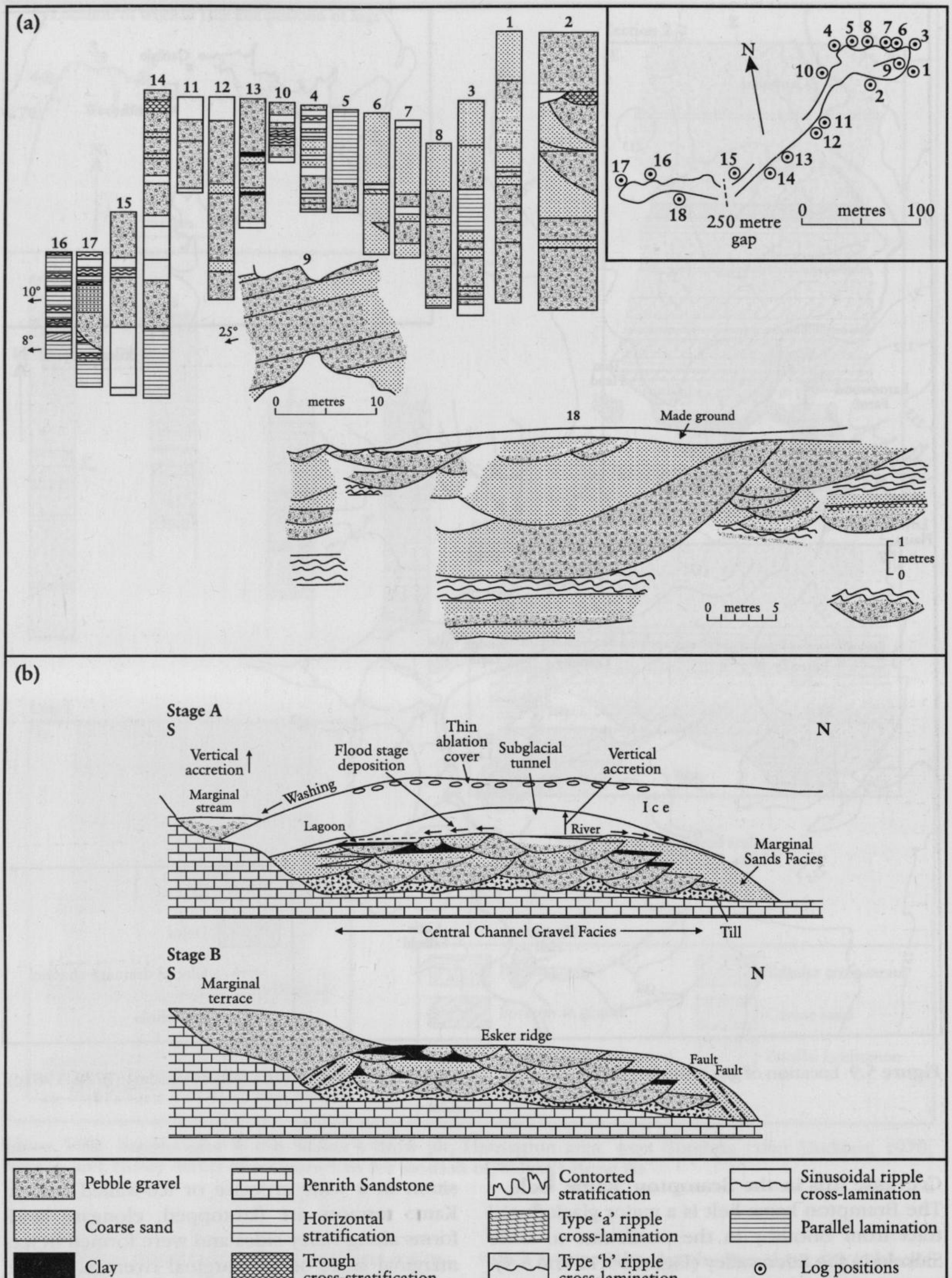


Figure 5.10 Stratigraphy and model of subglacial sedimentation at Low Plains pit, Edenside (after Huddart, 1970, 1981c): (a) stratigraphy; (b) model of subglacial sedimentation.

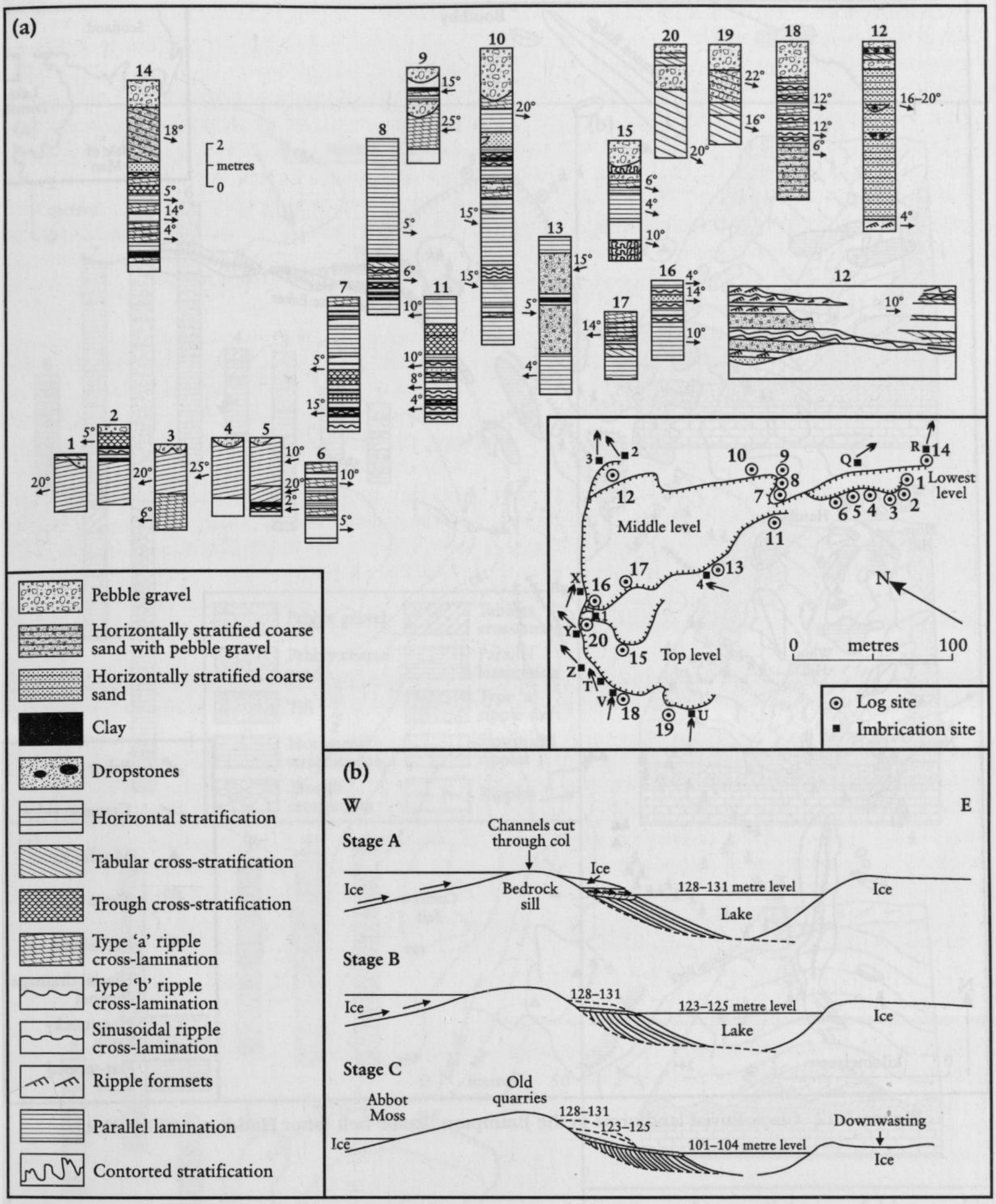


Figure 5.11 Stratigraphy and model of glacio-lacustrine sedimentation at the Baronwood pit, Edenside (after Huddart, 1970, 1981c): (a) stratigraphy; (b) model of glacio-lacustrine sedimentation.

then be let down on to the sub-ice surface, and such conical or moulin kames are common in the Brampton area, although there is no sedimentological evidence to support the depositional mechanism. Finally there are kame ridges, crevasse fills, ice-contact ridges or ice-walled, stream trenches, which are linear, short landforms, composed of fluvial sediment, and sectional evidence to reconstruct the depositional environment is available in this case. The

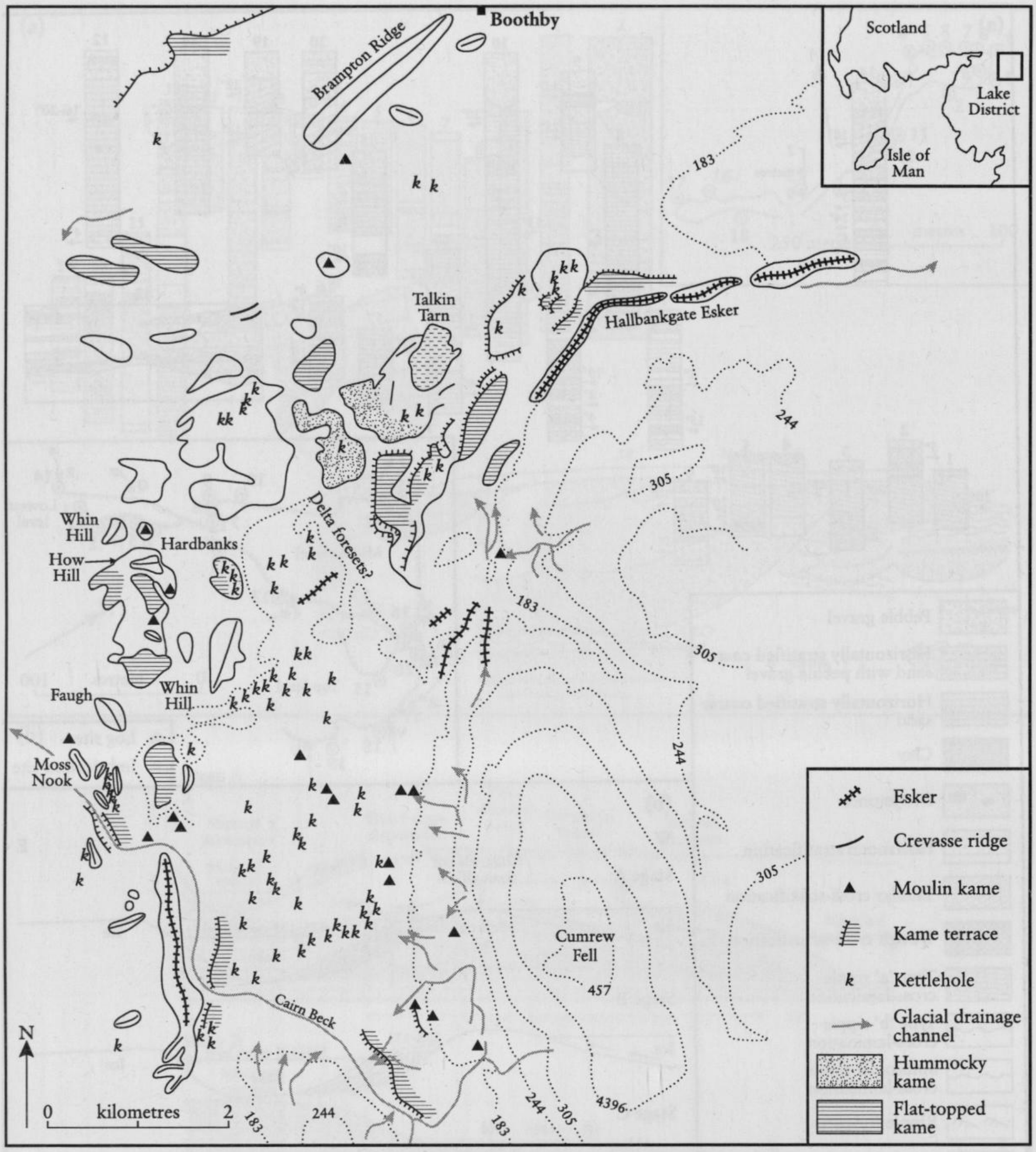


Figure 5.12 Glaciofluvial landforms in the Brampton 'kame' belt (after Huddart, 1970, 1981c).

ridges, which generally trend north–south and are situated on the western side of the kame belt, include the Brampton Ridge, Whin Hill near How Mill, Faugh and Moss Nook. They are usually between 0.5 and 1 km in length, although the Brampton Ridge is over 2 km long, and are situated between 125–127.5 m OD. They were attributed to ice frontal deposition by Trotter (1929) but this is not so. At Whin Hill, north of How Mill, a ridge 300 m long is nearly as wide as this at its maximum width and is composed of three sedimentary associations (Figure 5.13). Channel gravels and trough cross-stratification formed in the deeper part of the river, with horizontal stratification and ripples forming within the lateral margins, with occasional tabular dunes. The marginal association suggests shallow water sheet flooding, interspersed

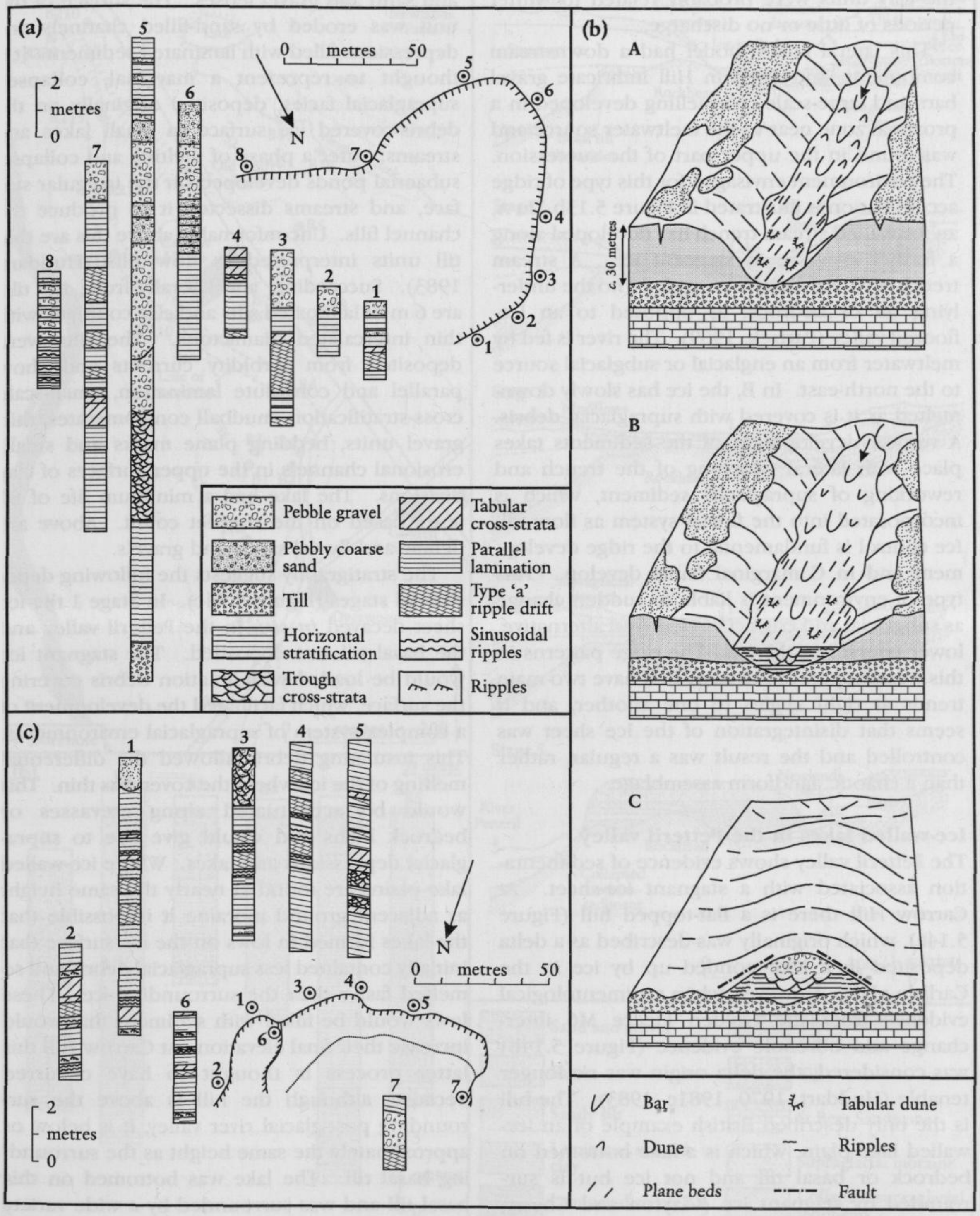


Figure 5.13 Stratigraphy in ice-walled stream trench deposition, Brampton 'kame' belt (after Huddart, 1970, 1981c): (a) stratigraphy at Whin Hill, How Mill; (b) model of deposition at Whin Hill, How Mill; (c) stratigraphy at Faugh.

with quiet water conditions and low flow, when there was minimal bedload transport and suspension sedimentation was important. Some of the clay units were probably related to winter periods of little or no discharge.

This lateral facies model had a downstream component, and at Whin Hill imbricate gravel bars and large-scale channelling developed in a proximal zone near to the meltwater source and was found in the upper part of the succession. The environment envisaged for this type of ridge accumulation is illustrated in Figure 5.13b. In A, an ice-walled stream trench has developed along a former crevasse in stagnant ice. A stream trench is where water has eroded into the underlying till or bedrock, as opposed to an ice-floored valley (Parizek, 1969). The river is fed by meltwater from an englacial or subglacial source to the north-east. In B, the ice has slowly down-melted as it is covered with supraglacial debris. A vertical development of the sediments takes place with lateral widening of the trench and reworking of supraglacial sediment, which is incorporated into the fluvial system as flow till. Ice control is fundamental to the ridge development and in C marginal faults develop. This type of environment is liable to sudden change as subglacial and englacial rivers find alternative, lower courses in the ice. The ridge patterns in this western part of the kame belt have two main trends at right angles to one another, and it seems that disintegration of the ice sheet was controlled and the result was a regular, rather than a chaotic, landform assemblage.

Ice-walled lakes in the Petteril valley

The Petteril valley shows evidence of sedimentation associated with a stagnant ice-sheet. At Carrow Hill there is a flat-topped hill (Figure 5.14a), which originally was described as a delta deposited in a lake ponded up by ice in the Carlisle plain. However, when sedimentological evidence from the Golden Fleece M6 interchange and borehole evidence (Figure 5.14b) was considered, the delta origin was no longer tenable (Huddart, 1970, 1981c, 1983). The hill is the only described British example of an ice-walled lake plain, which is a lake bottomed on bedrock or basal till and not ice but is surrounded by stagnant ice (Clayton and Cherry, 1967). As such it is an important indicator of the mode of ice decay (Huddart, 1977) and was thought to be formed in the following stages. Basal till deposited in a bedrock depression is overlain by a disturbed, heterogeneous sediment with complex grain size changes, disturbances, deformed clasts of laminated sediment and sand and gravel lenses. The surface of this unit was eroded by sand-filled channels and depressions filled with laminated sediment. It is thought to represent a marginal, collapsed supraglacial facies, deposited originally on the debris-covered ice surface in small lakes and streams. After a phase of melting and collapse, subaerial ponds developed on the irregular surface, and streams dissected it to produce the channel fills. Unconformably above this are thin till units interpreted as flow tills (Huddart, 1983). Succeeding, and laterally from, the tills are 6 m of laminated silts and clay couplets, with thin intercalated diamictons. The silts were deposited from turbidity currents and show parallel and convolute lamination, small-scale cross-stratification, mudball conglomerates, thin gravel units, bedding plane marks and small, erosional channels in the upper surfaces of clay divisions. The lake had a minimum life of 87 years based on the couplet count. Above are deltaic and fluvial sands and gravels.

The stratigraphy suggests the following depositional stages (Figure 5.14c). In Stage 1 the ice sheet decayed *in situ* in the Petteril valley and the basal till was deposited. The stagnant ice would be loaded with ablation debris covering the surface, which facilitated the development of a complex system of supraglacial environments. This insulating debris allowed the differential melting of the ice where the cover was thin. This would be accentuated along crevasses or bedrock highs and would give rise to supraglacial depressions and lakes. Where ice-walled lake plains are found at nearly the same height as adjacent ground moraine it is possible that the lakes formed in lows on the ice surface that initially contained less supraglacial debris and so melted faster than the surrounding ice. These lows would be filled with sediment that would increase their final elevation. At Carrow Hill this latter process is thought to have occurred because, although the hill is above the surrounding post-glacial river valley, it is below or approximately the same height as the surrounding basal till. The lake was bottomed on this basal till and was surrounded by a wide variety of supraglacial river and lake environments. In Stage 2, when the ice-walled lake had been initiated, meltwater streams would deposit deltas and the lake would grow both in length

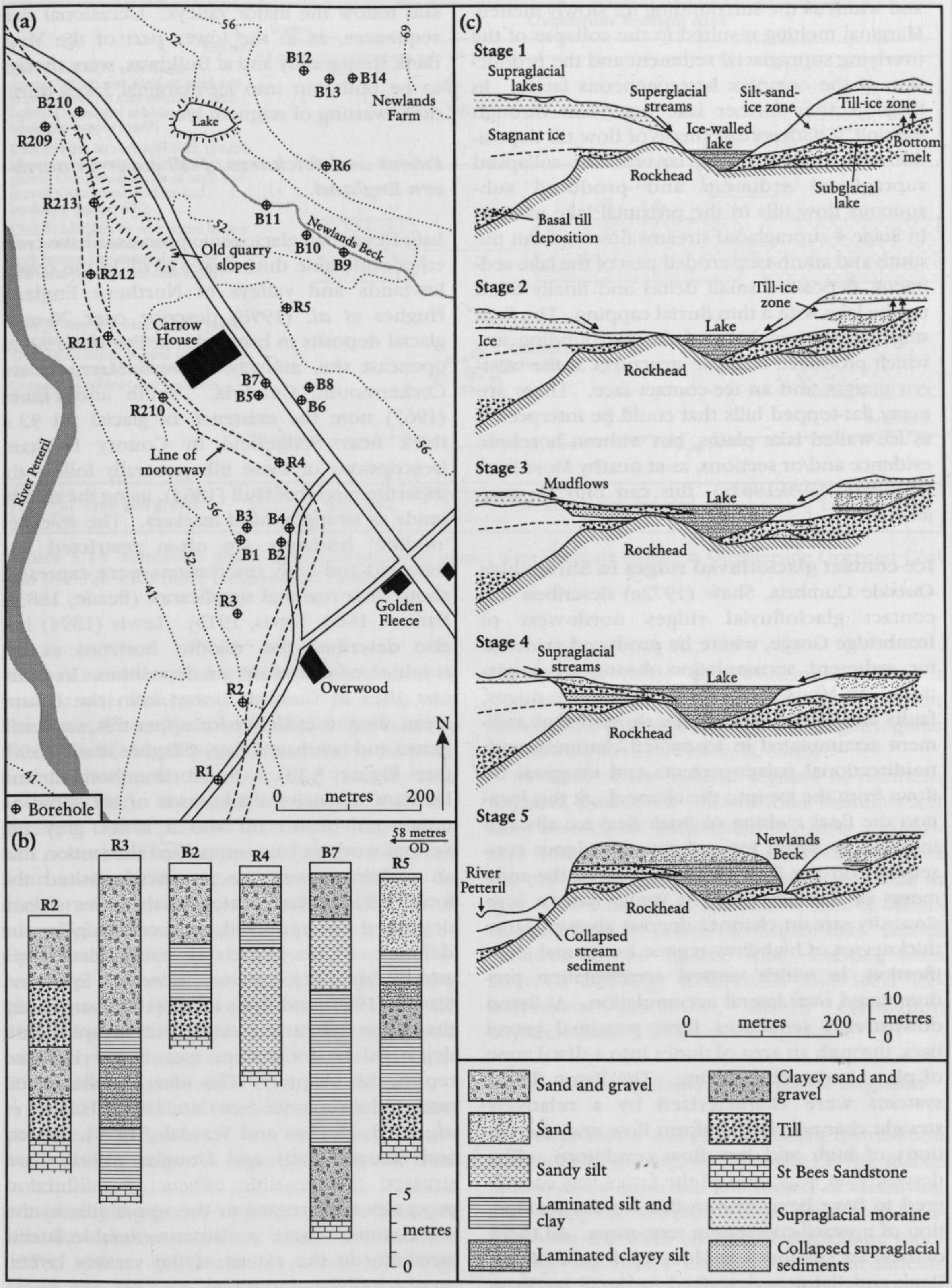


Figure 5.14 Ice-walled lake plain landforms, sediments and development at Carrow Hill, Petteril valley, Cumbria (after Huddart, 1970, 1983): (a) morphology, borehole locations and M6 motorway interchange, Carrow Hill; (b) stratigraphy of circled boreholes in (a); (c) model of ice-walled lake plain deposition.

and width as the surrounding ice slowly melted. Marginal melting resulted in the collapse of the overlying supraglacial sediment and the production of the complex heterogeneous facies. In Stage 3 after further lake extension through melting, a widespread phase of flow till deposition was initiated. This covered the collapsed supraglacial sediment and produced subaqueous flow tills in the proximal lake margin. In Stage 4 supraglacial streams flowing from the south and south-east eroded part of the lake sediment, deposited small deltas and finally filled up the lake with a thin fluvial capping. The final stage was the melting of the surrounding ice, which produced collapse structures at the western margin and an ice-contact face. There are many flat-topped hills that could be interpreted as ice-walled lake plains, but without borehole evidence and/or sections, as at nearby Moss Pool (Huddart, 1970,1981c), this can only be conjecture.

Ice-contact glaciofluvial ridges in Shropshire

Outside Cumbria, Shaw (1972a) described ice-contact glaciofluvial ridges north-west of Ironbridge Gorge, where he produced evidence for sediment accumulation during ice-stagnation. At Mousecroft Lane upstanding ridges, faults and lateral kettleholes showed that sediment accumulated in ice-walled channels, with unidirectional palaeocurrents and marginal till flows from the ice into the channel. At this location the final melting of 'Irish Sea' ice allowed invasion by Welsh ice and the subsidence consequent on the final melting involved the sediments of this advance. At Venus Bank a low-sinuosity stream channel deposit showed large thicknesses of high-flow-regime horizontal stratification, in which vertical accumulation predominated over lateral accumulation. At Ryton downstream sequences form proximal gravel bars, through an area of dunes into a distal zone of plane bed with scouring. The Ryton fluvial systems were characterized by a relatively straight channel, non-uniform flow and alternations of high and low flow conditions. The downstream migration of the facies was considered to have been responsible for the production of upward-coarsening sequences. All three of these successions – at Mousecroft Lane, Venus Bank and Ryton – described in detail by Shaw (1972a) belonged to an integrated drainage system, since their palaeocurrents are parallel to those at Norton Farm, Dorrington and Buildwas, and follow the major valleys. Occasional delta sequences, as in the lower part of the Venus Bank stratigraphy and at Buildwas, were thought to be built out into ice-marginal lakes during downwasting of stagnant ice.

Extent and thickness of till cover in northern England

Late Devensian glaciogenic processes have created considerable thicknesses of till in the coastal lowlands and valleys of Northern England. Hughes *et al.* (1998) describe over 70 m of glacial deposits in boreholes at Foxhouse South opencast site, midway between Maryport and Cockermouth, Cumbria. Smith and Francis (1967) note the existence of glacial till 92 m thick near Sedgefield in County Durham. Descriptions of these tills generally follow the tripartite model of Hull (1864), using the middle sands as stratigraphical markers. The so-called 'middle' horizons are often restricted and repeated and early reservations were expressed about their regional significance (Reade, 1883b; Barrow, 1888; Travis, 1913). Lewis (1894) had also described the 'middle' horizons as the result of subglacial stream deposition. In opencast sites in Cumbria, away from the coastal areas, there is evidence for upper tills, sand and gravel and laminated clays (Hughes *et al.*, 1998) (see Figure 5.15). In Northumberland and Durham extensive thicknesses of till show an upper red–brown till and a lower grey till. Several workers have supported the notion that an earlier western ice stream deposited the lower till and that a later northern ice stream deposited the upper till, so accounting for the different colours of the clay matrix, clast suites and till-fabric orientations. However, Eyles and Sladen (1981) and Eyles *et al.* (1982) suggested that these tills are a single but complex unit deposited by lodgement (see Sandy Bay site report, this chapter). This idea has not gained universal acceptance (see Catt, 1991a; Hughes *et al.*, 1998; Hughes and Teasdale, 1999). Anson and Sharp (1960) and Douglas (1991) have stressed the possible effects of solifluction processes with regard to the upper tills in the successions. There is also considerable lateral variability in the extent of the various layers, even at individual localities.

In the Vale of York the single Late Devensian till is a reddish-brown clay with abundant erratics of Magnesian Limestone and Triassic Sand-

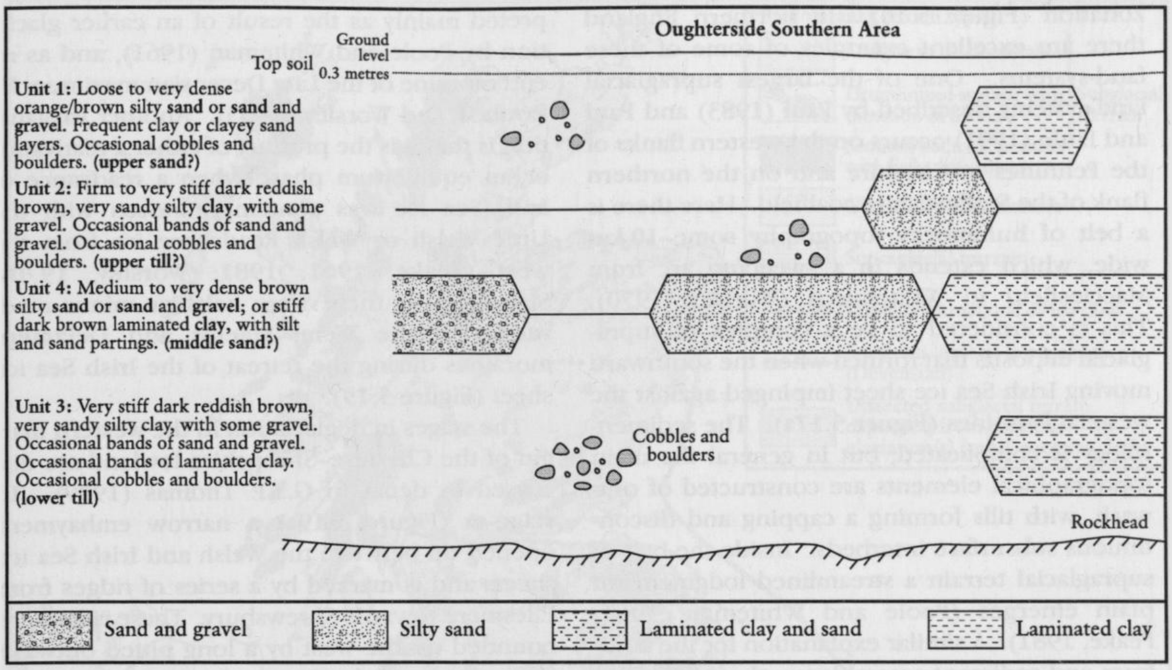


Figure 5.15 Schematic glacial lithostratigraphy in inland west Cumbria based on Oughterside Opencast Coal Site (NY 126 400) (after Hughes *et al*., 1998).

stones, as well as Lake District, Carboniferous and Lower Palaeozoic rocks. Towards the Pennines, where Yorkshire Dales ice fed into the Vale of York, the till is greyer and rich in Carboniferous Limestone, sandstone and chert erratics (Lovell, 1982). This single till contrasts with the multiple tills in parts of Cleveland, where three tills occur south of Hartlepool (James, 1982) and four at Rockcliffe Scar (Francis, 1970). In Holderness there are multiple tills, which recently have been described and interpreted by Eyles *et al*. (1994) and Evans *et al*. (1995) in terms of deformation tills deposited during glacial events shortly after 18 000 years BP. McCabe *et al*. (1998), however, have suggested a major ice readvance about 14 700–14 000 years BP (see Dimlington site report, this chapter).

In the Lancashire Plain, Longworth (1985) provided evidence of the unrepresentative nature of an extensive tripartite sequence from borehole data for the M55 motorway. Similar complex glaciogenic sequences with multiple tills have been noted in Cumbria (Huddart, 1970, 1981c, 1983) and from borehole logs along the M62 in east Lancashire (Johnson, 1985b). In the Cheshire–Shropshire lowlands, although the tripartite interpretative model had been supported by the [British] Geological Survey (e.g. Poole and Whiteman, 1961), it became apparent that the Stockport Formation embraced a complex Late Devensian glaciogenic sequence (Worsley, 1967a, b, 1985).

In most of the upland regions of northern England, such as the Cheviot Hills, Pennines and Lake District, till distribution is patchy and variable both in lithology and in thickness. Most of the tills are considered to be Late Devensian in age. This is based on the relative lack of dissection and shallow, poorly developed soils. Lateral variation in the upland tills, resulting from bedrock lithology differences and local ice-movement directions, makes correlation between areas difficult (Catt, 1991a).

Glacial land-systems in northern England

It is clear that a complex sediment distribution results from glaciogenic processes. The glacial land-systems approach is a means of classifying and mapping sediment sequences and landforms on the basis of their characteristic topography, subsurface conditions and sediments (Boulton and Paul, 1976; Boulton and Eyles, 1979; Eyles, 1983; Paul and Little, 1991). The value of such a land-systems approach is its ability to produce a large-scale geomorphological zonation (Figure 5.16). In northern England there are excellent examples of some of these land-systems. One of the largest supraglacial land-systems described by Paul (1983) and Paul and Little (1991) occurs on the western flanks of the Pennines in Cheshire and on the northern flank of the Staffordshire coalfield. Here there is a belt of hummocky topography some 10 km wide, which extends in a sweeping arc from Macclesfield to Whitchurch (Worsley, 1970). This is considered to be a complex of supraglacial deposits that formed when the southward moving Irish Sea ice sheet impinged against the western Pennines (Figure 5.17a). The sedimentology is complicated, but in general the main topographical elements are constructed of outwash, with tills forming a capping and discontinuous subsurface interbeds. Inside the belt of supraglacial terrain a streamlined lodgment till plain emerges (Poole and Whiteman, 1961; Peake, 1981). A similar explanation for the landform and sediment assemblages, the patterns of deposition and a test for the supraglacial land-system has been proposed by G.S.P. Thomas (1989; Figure 5.17b) for the western margins of the Cheshire–Shropshire lowlands. Similarly, Longworth (1985) explains the Kirkham moraine in the Fylde as a supraglacial land-system. Other glacial land-systems are readily recognized in northern England. In the Northumbrian, Cumbrian, Lancashire and Cheshire lowlands the till plain, which is drumlinized in certain areas, is part of a subglacial land-system, created at the base of the Late Devensian ice sheet. However, Johnson (1985b) provided a useful alternative land-system approach for glaciated upland areas marginal to lowland ice sheets, as along the west Pennine uplands and west Cumbria.

Glacial moraines in northern England

The maximum geographical limits of the Late Devensian ice sheet (Figure 5.18) are often not delimited by a terminal moraine, although in the southern part of the Cheshire–Shropshire lowlands ice retreat from the maximum seems to have involved pauses and perhaps readvances. These built up large terminal moraine landforms, for example the Wrexham–Ellesmere–Whitchurch hummocky moraine and the Whitchurch–Woore and Bar Hill moraine (McQuillan, 1964), separated by the Prees Heath inter-lobate outwash plain. This landform has been interpreted mainly as the result of an earlier glaciation by Poole and Whiteman (1961), and as an end moraine of the Late Devensian maximum by Boulton and Worsley (1965). Another explanation is that it is the product of either a readvance or an equilibrium phase when a readvance of Irish Sea ice was contemporaneous with the Little Welsh or Welsh Readvance ice from the west (Peake, 1961, 1981; Worsley, 1970). Subsequently there were halts or minor readvances at the Wem, Maelor and Delamere moraines during the retreat of the Irish Sea ice sheet (Figure 5.19).

The stages in deglaciation in the western margin of the Cheshire–Shropshire lowland are discussed in detail in G.S.P. Thomas (1989). In stage A (Figure 5.19a) a narrow embayment opened up between the Welsh and Irish Sea ice sheets and is marked by a series of ridges from Ellesmere towards Shrewsbury. These ridges are bounded on the west by a long pitted outwash fan running south towards the River Severn. As the Irish Sea ice continued to retreat it uncovered the escarpment south of Prees and at one stage the ice margin lay along its northern edge and looped west towards Ellesmere (stage B, Figure 5.19). From its eastern margin a number of major meltwater outlets carried sediment south via rock channels. To the south this meltwater-fed sandur drained south across an area of thin and subdued till. Retreat from this area was rapid and unaccompanied by much supraglacial sedimentation. At stage B the two ice sheets had uncoupled along much of the Welsh border, although the Irish Sea ice pressed against the foothills between Mold and Wrexham, and drainage ran within, or submarginal to, its margin near Ellesmere. As much of the sediment was deposited upon and against stagnant ice in a series of marginal troughs lying between dead-ice ridges, subsequent melt created a complex topography of till ridges and pitted and kettled outwash. During this time minor readvances generated till-cored ridges in the Ellesmere and Whitchurch areas. With further stagnation, the ice limit fell to the rear of the innermost moraine at Ellesmere and looped east and north-east, and passed Wem towards Whitchurch (stage C, Figure 5.19), where a deep, ice-marginal embayment served to deliver a sloping outwash fan towards Prees. Large flow-till volumes were deposited in the proximal parts of this fan. The toe of the fan fed a large lake dammed against the Triassic scarp. This lake had a complex

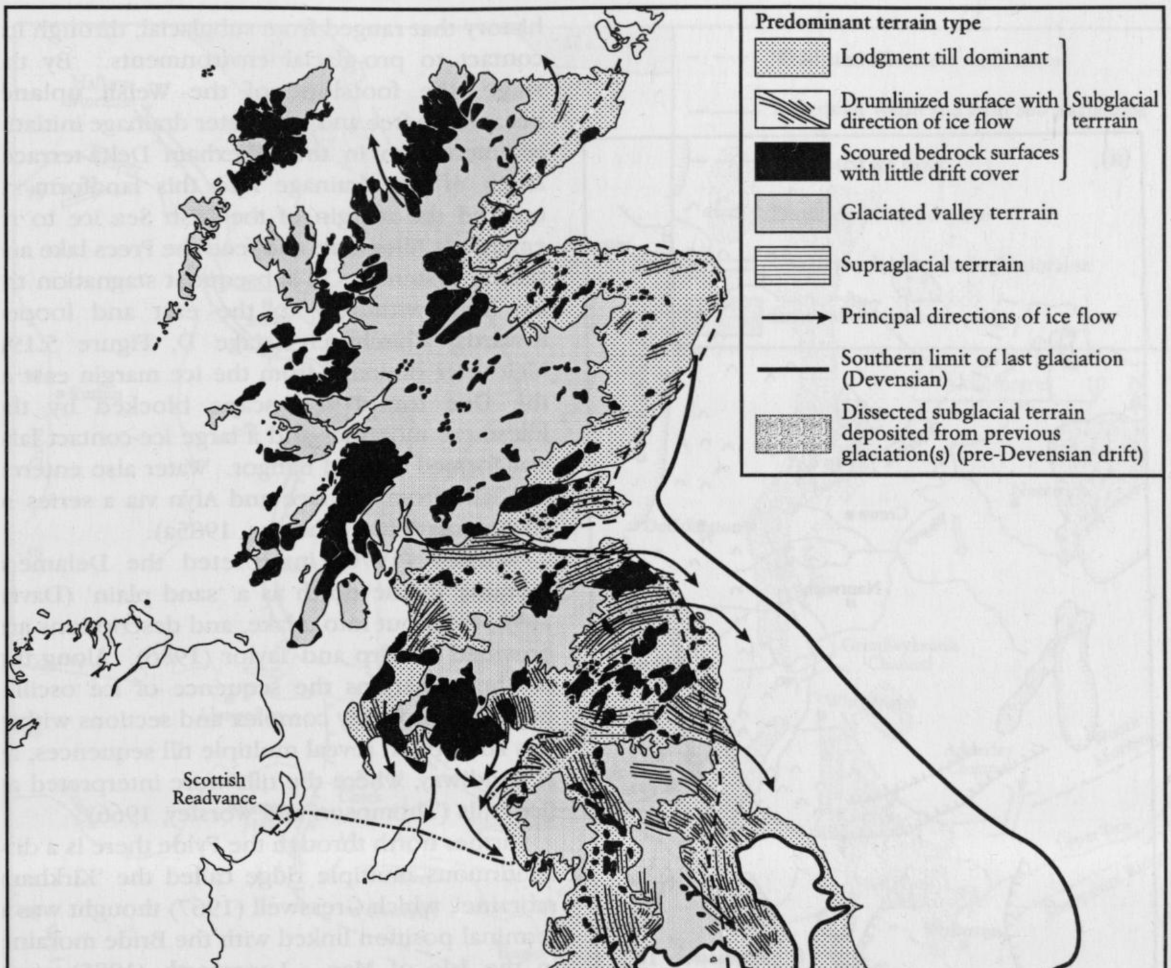
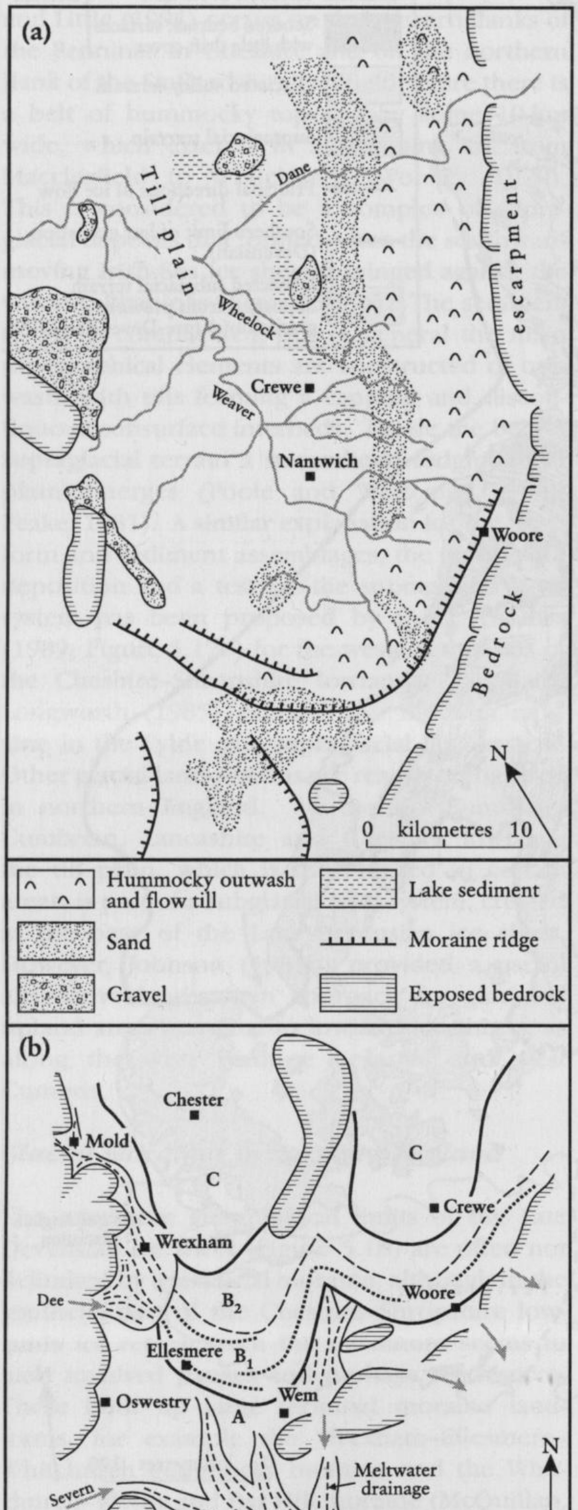


Figure 5.16 Large-scale geomorphological zonation and glacial land-system terrains and limits of glaciations (modified from Huddart, 1970; Eyles and Dearman, 1981; Cameron *et al.*, 1987).



◂**Figure 5.17** Landform–sediment assemblages in the Cheshire–Shropshire lowland: (a) Schematic diagram of a broad arcuate belt of hummocky flow till draped over outwash ridges abutting the western Pennine edge. The till plain to the north and west is composed of lodgement till and locally is obscured by glacio-lacustrine sediments and flow till. (After Paul, 1983.) (b) major sediment–landform assemblages and the relationships between former ice lobes and bedrock topography (after Thomas, G.S.P., 1989). A–C indicate moraine lines.

history that ranged from subglacial, through ice-contact to pro-glacial environments. By this stage, the footslope of the Welsh uplands became ice free and meltwater drainage initiated sedimentation in the 'Wrexham Delta-terrace'. Much of the drainage into this landform re-entered the margin of the Irish Sea ice to re-emerge at Ellesmere and feed the Prees lake and sandur systems. On subsequent stagnation the ice limit withdrew to the east and looped towards Whitchurch (stage D, Figure 5.19). Meltwater drainage from the ice margin east of the Dee found its escape blocked by the Ellesmere moraines and a large ice-contact lake was formed around Bangor. Water also entered this lake from the Dee and Alyn via a series of coalescing deltas (Thomas, 1985a).

Worsley (1970) interpreted the Delamere moraine to the north as a 'sand plain' (Davis, 1890) built out into a lake, and descriptions are provided in Earp and Taylor (1986). Along the northern margins the sequence of ice oscillations was probably complex and sections within this zone often reveal multiple till sequences, as at Sandiway, where the tills were interpreted as flow tills (Thompson and Worsley, 1966).

Farther north through the Fylde there is a discontinuous multiple ridge called the 'Kirkham moraine', which Gresswell (1967) thought was a terminal position linked with the Bride moraine in the Isle of Man. Longworth (1985) casts doubt on this interpretation, however, and invokes a supraglacial origin for its formation. In Cumbria, readvances created a moraine line associated with the Low Furness Readvance and the Gosforth Oscillation/Scottish Readvance moraine along the Black Combe coastal plain and north along the coastal zone as far as the St Bees moraine.

On the east coast, the maximum limit across the Humber estuary is marked by morainic ridges east of Winteringham and Winterton and

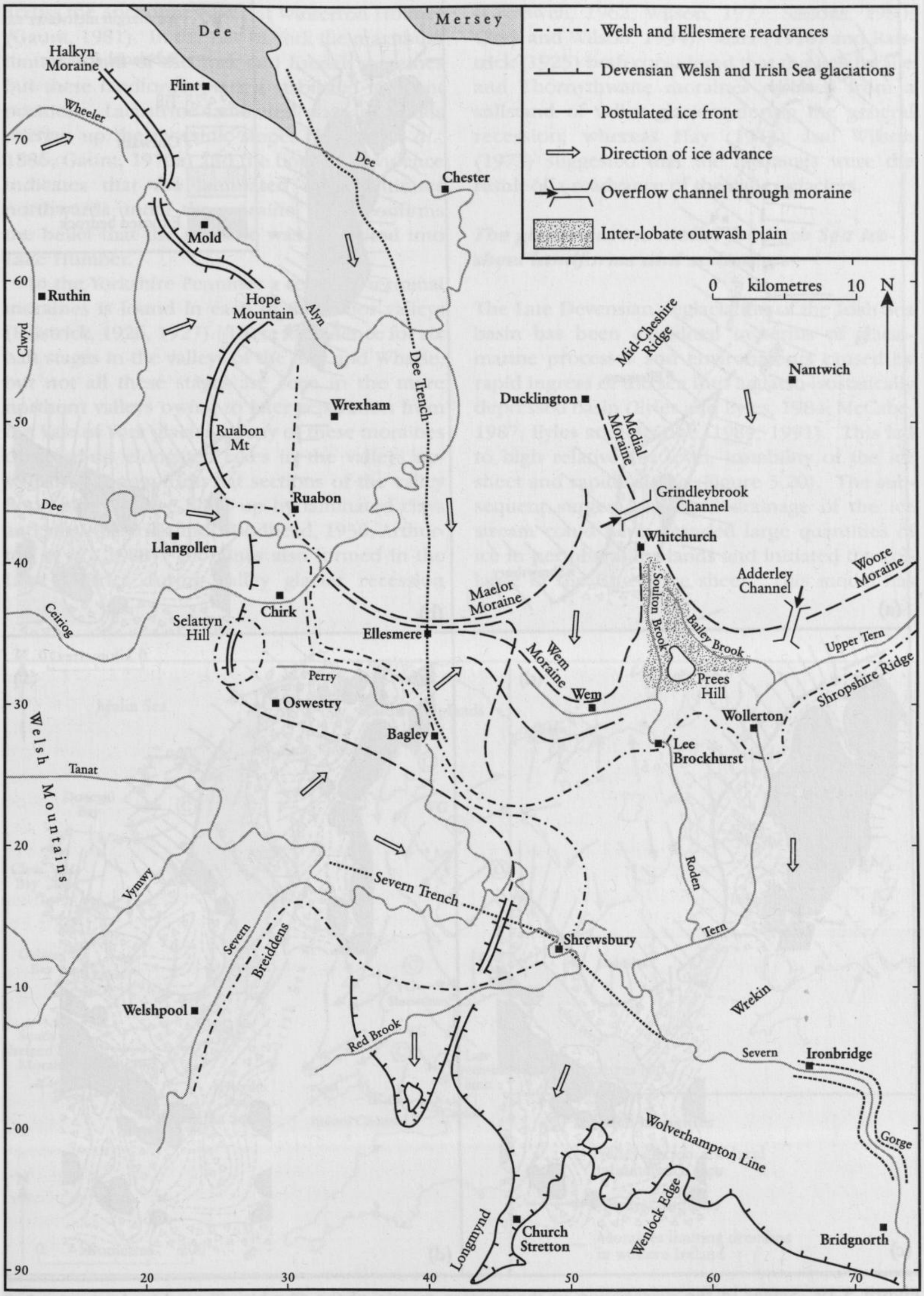


Figure 5.18 Late Devensian glaciation in the Cheshire Plain and Welsh Borders (after Peake, 1981).

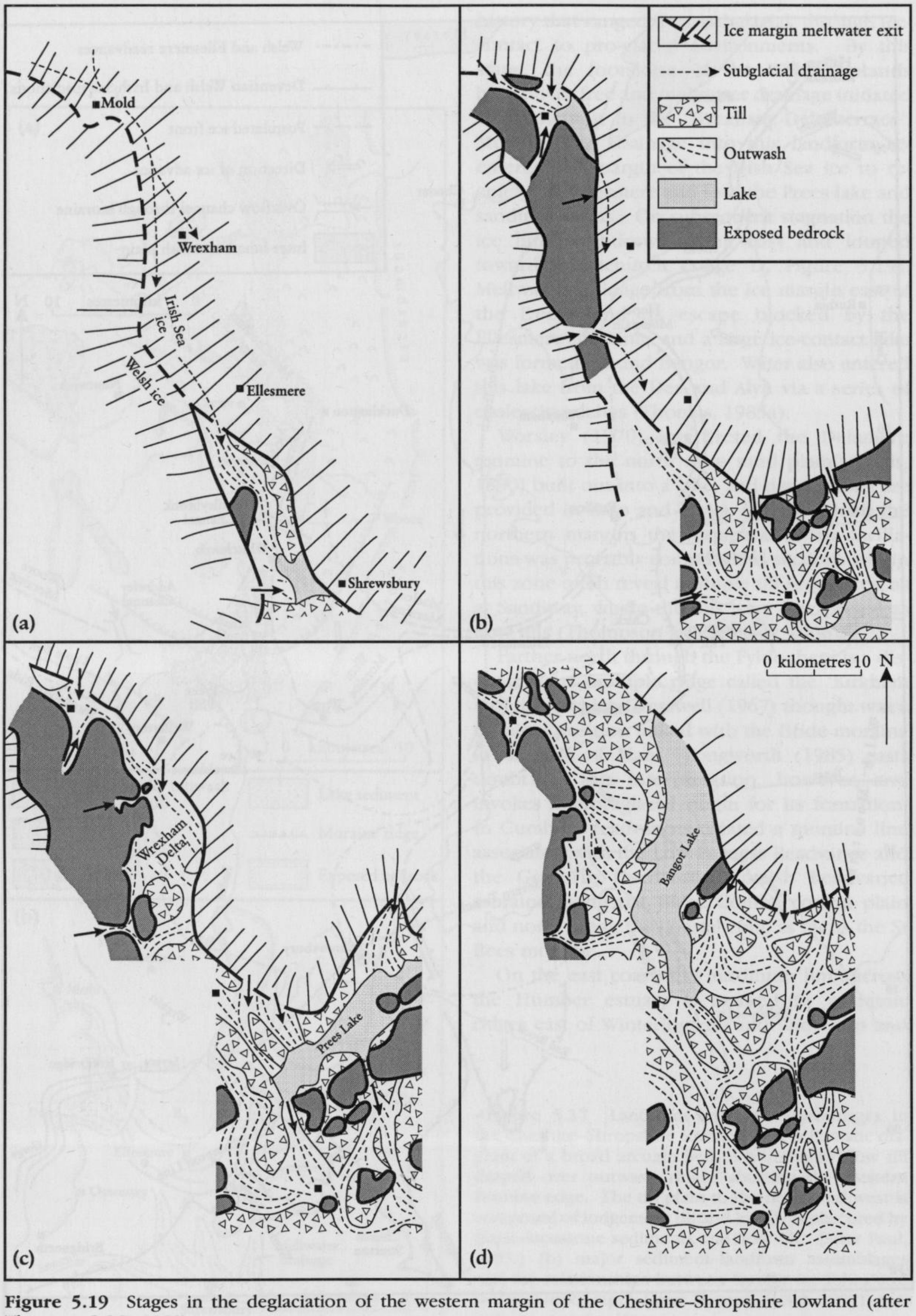


Figure 5.19 Stages in the deglaciation of the western margin of the Cheshire–Shropshire lowland (after Thomas , G.S.P., 1989).

across the Ancholme Valley at Winterton Holmes (Gaunt, 1981). In the Vale of York the maximum limit is south of the York and Escrick moraines but these landforms mark a stabilized ice-front position. Lacustrine laminated clay and sands overlap up the morainic slopes (Dakyns *et al.*, 1886; Gaunt, 1970a) and the borehole evidence indicates that the laminated clay continues northwards under the moraine. This confirms the belief that the moraine was deposited into Lake Humber.

In the Yorkshire Pennines a series of terminal moraines is found in each of the major valleys (Raistrick, 1926, 1927). There is evidence for six halt stages in the valleys of the Aire and Wharfe, but not all these stages are seen in the more northern valleys owing to later separation from the Vale of York glacier. Many of these moraines dammed-up elongated lakes in the valleys upstream and now form flat sections of the valley floor as they were filled up by laminated clays and silts (Raistrick and Woodhead, 1930; Arthurton *et al.*, 1988). Moraines also formed in the Lake District during valley glacier recession (Gresswell, 1962; Wilson, 1977; Sissons, 1980; Clark and Wilson, 1994). Marr (1916) and Raistrick (1925) both considered that the Rosthwaite and Thornythwaite moraines resulted from a stillstand of valley glaciers during the general recession, whereas Hay (1944) and Wilson (1977) suggested that the moraines were the result of a readvance of the valley glaciers.

The glaciomarine model for Irish Sea ice-sheet landforms and sediments

The Late Devensian deglaciation of the Irish Sea basin has been explained in terms of glaciomarine processes and environments caused by rapid ingress of the sea into a glacio-isostatically depressed basin (Eyles and Eyles, 1984; McCabe, 1987; Eyles and McCabe, 1989, 1991). This led to high relative sea level, instability of the ice sheet and rapid calving (Figure 5.20). The subsequent surges and rapid drainage of the ice stream could have stranded large quantities of ice in peripheral lowlands and initiated the collapse of the British ice sheet. This model has

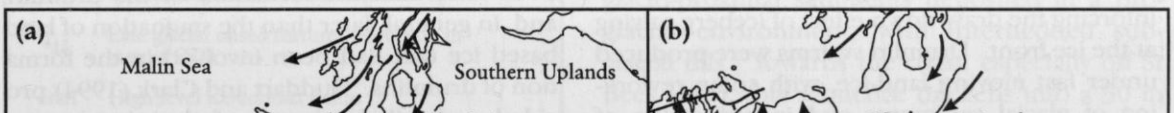


Figure 5.20 Irish Sea palaeogeography during the Late Devensian glaciation (after Eyles and McCabe, 1989): (a) shows ice flow and dispersal centres; (b) shows the final stage of disintegration with a calving front.

been severely criticized for certain parts of the Irish Sea basin; interpretations of some critical sites have been questioned and explanations in terms of land-based ice have been reasserted in some localities around the basin (e.g. Harris and McCarroll, 1990; Austin and McCarroll, 1992; McCarroll and Harris, 1992; Huddart, 1994, 1997; Huddart and Clark, 1994; Walden, 1994). Even so, the event stratigraphy proposed by Eyles and McCabe (1989) remains a powerful framework for understanding the processes and patterns of environmental change at the termination of the last glaciation.

The glaciomarine account of Irish Sea basin deglaciation begins with a rise of global sea level associated with the start of a phase of world-wide ice recession. Rising seas impinged on the southern front of the Irish Sea ice sheet. Embayments in the front of the ice shelf focused iceberg calving and concentrated the rapid southward movement of ice from the main ice mass in the basin. As sea level rose still further the marine ice-margin retreated north, stabilizing temporarily at grounding lines. The waning land-based ice around the basin produced abundant meltwater, much of which flowed at the base of land ice, so easing ice movement and reinforcing the draw-down effect of iceberg calving at the ice front. Drumlin swarms were produced under fast moving land-ice, with some reworking of glacial sediments and incorporation of material that meltwater had carried beneath the ice. Subglacial tunnel valleys were incised, particularly along the main valleys. Thick beds of marine sediments were deposited in areas where tidewater ice margins were stationary for sufficiently long intervals and morainal banks built up from debris released at the ice front. Many flat upper surfaces of sand and gravel now well above sea level have been interpreted as deltas formed where major meltwater channels at ice fronts reached the sea. Thin mud drapes were ascribed to suspension settling of fine glacial debris in the sea. At some sites around the Irish Sea basin in-situ, rather than derived and reworked, marine fossils have been recorded and demonstrate the enclosing sediments to be glaciomarine.

Gradually, the retreating ice margins and the advancing seas reached coastal Cumbria, the Solway Firth and Galloway, areas that were still incompletely recovered from glacio-isostatic crustal depression. High relative sea levels of 152 m OD were inferred for supposed marine deltas in Wasdale and Eskdale. In Dumfriesshire raised deltas were noted at Dumfries (50 m) and Dalton (70 m), but as part of a reassessment of the glaciofluvial sedimentology of the Nith and Annan basins, Huddart (1999) considered there to be no evidence of raised delta sequences. The glaciomarine explanation of deglaciation in the Cumbrian lowlands requires land-based ice moving generally westward to marine termini in the vicinity of the present coast and marine incursions. It considers that the glaciomarine landforms were contemporary, or near contemporary, with drumlin formation under mobile land ice. It proposes that the Cumbrian land-based drumlin-forming ice streams terminated at marine margins now onshore, for example, in Low Furness and north of Maryport, as evidenced by morainal banks and raised deltas (Figure 5.21). Several of the features regarded as glaciomarine by Eyles and McCabe (1989) are among those Huddart (1991, 1993, 1994, 1997) regarded as evidence for readvance of ice from the west and north-west. The readvance proposition requires the presence of active land-based ice, whereas the glaciomarine one requires sea, it requires the readvance to have been later than the formation of the drumlins and, in general, later than the stagnation of land-based ice that had been involved in the formation of drumlins. Huddart and Clark (1994) provided a detailed rebuttal of the glaciomarine models for Cumbrian landforms and sediments, including sites at Holme St Cuthbert and St Bees.

The glaciomarine interpretation requires evidence for deposition in marine environments at sites categorized as morainal banks and marine deltas. It depends on adequate evidence for contemporary ice margins at those sites and for ice cover landward of them. Evidence of late incursions of ice from the north-west and west on to the lowlands directly opposes the glaciomarine hypothesis. Ice readvance models place ice on the 'wrong' side of morainal banks and deltas (that is to the west) for the glaciomarine interpretation. It is noteworthy that in direct juxtaposition to their comment that the Scottish Readvance proposition relies on identification of high-level lacustrine deltas, Eyles and McCabe (1989) stated 'sedimentological and microfaunal evidence indicate that the deltas [their high-level glaciomarine deltas] record local marine limits', but they presented no sedimentological or microfaunal evidence from

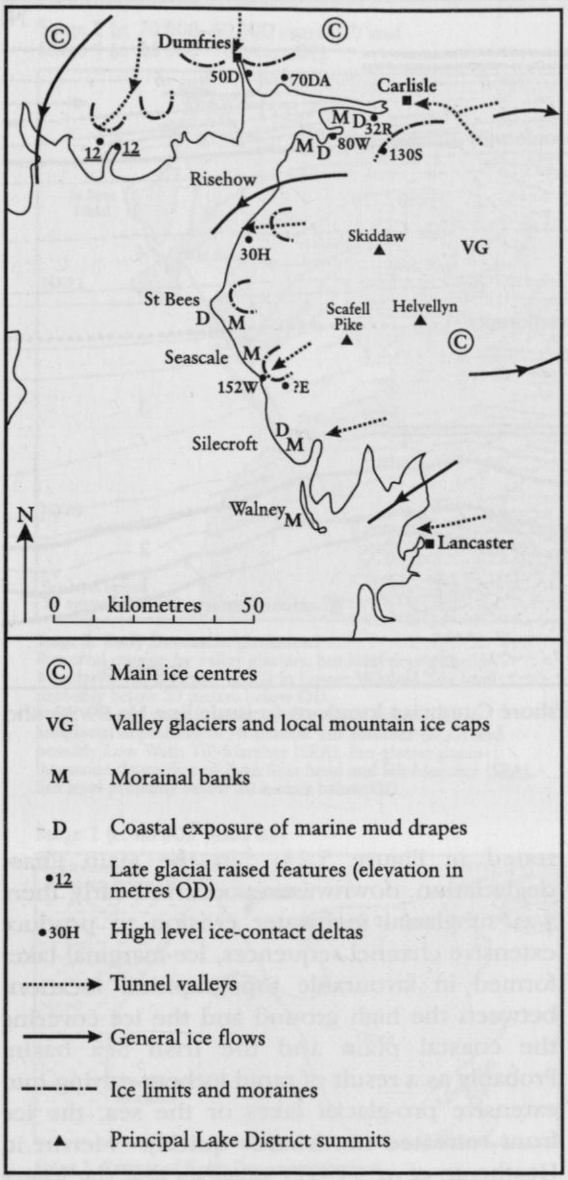


Figure 5.21 Supposed glaciomarine sites in the Cumbrian lowlands (after Eyles and McCabe, 1989; discussed in Huddart and Clark, 1994).

the Cumbrian sites. The failure to report marine fossils from supposed glaciomarine Cumbrian sediments is of major significance, especially as these fossils are not uncommon at other sites on the fringes of the Irish Sea, especially on the Irish coast (e.g. McCabe *et al.*, 1990; Haynes *et al.*, 1995). Huddart and Clark (1994) concluded that the interpretations of Eyles and McCabe (1989) have not been substantiated in Cumbria and that they have incorporated landforms and sediments from Cumbria into their model without full regard for the detailed sedimentological, palaeocurrent and lithological evidence available that suggests an alternative explanation for these glaciogenic deposits.

It is not suggested that there were no marine ice margins associated with the Late Devensian deglaciation in parts of the Irish Sea basin, but each area and sequence must be interpreted on the basis of the evidence it provides. For example, in offshore Cumbria there appears to be some evidence for glaciomarine events from an offshore seismo-stratigraphical framework and associated Nirex boreholes. The framework has been established using seismically distinctive sequences, which are bounded top and bottom by unconformities, and their correlative conformities (Eaton and Williams, 1993; Eaton, 1996; Huddart, 1997). Six major seismically distinctive sedimentary sequences (1–6 in Figure 5.22) have been recognized, with oedometer tests revealing that sequences 1–3 are overconsolidated and that sequences 4–6 are normally consolidated. Sequence 1 is characterized by tills, with subordinate sands and gravels. The rapid facies changes are interpreted as typical of glacio-proximal sediments deposited in a proglacial environment, with interbedded subglacial tills. Towards the coast, especially off St Bees Head, this sequence thickens into a 30 m wedge, which is thought to be associated with the Late Devensian Irish Sea main ice phase. Its top is marked by the W unconformity on Figure 5.22 . Sequence 2 is a thin (1–5 m) sequence that drapes Sequence 1 and locally is absent over topographical highs. It is thought to have been deposited as rain-out from floating ice in a low-energy glacio-lacustrine environment. Sequence 3 is an aggradational sequence, up to 30 m maximum, which unconformably overlies Sequence 2. It is composed of clays, silts and muds and is thought to have been deposited in a progressively deeper, perhaps distal glaciomarine environment. Above is the major, X unconformity, which separates the gently folded and overconsolidated sequences 1–3 from the overlying normally consolidated sediments of sequences 4–6. This unconformity is considered to be the result of an erosional phase caused by a major ice readvance that truncated, disturbed and overconsolidated the underlying sequences. Knight *et al.* (1997) placed a relative age of 12 000–14 000 years BP for the X unconformity caused by the Scottish Readvance and a conjectured age of

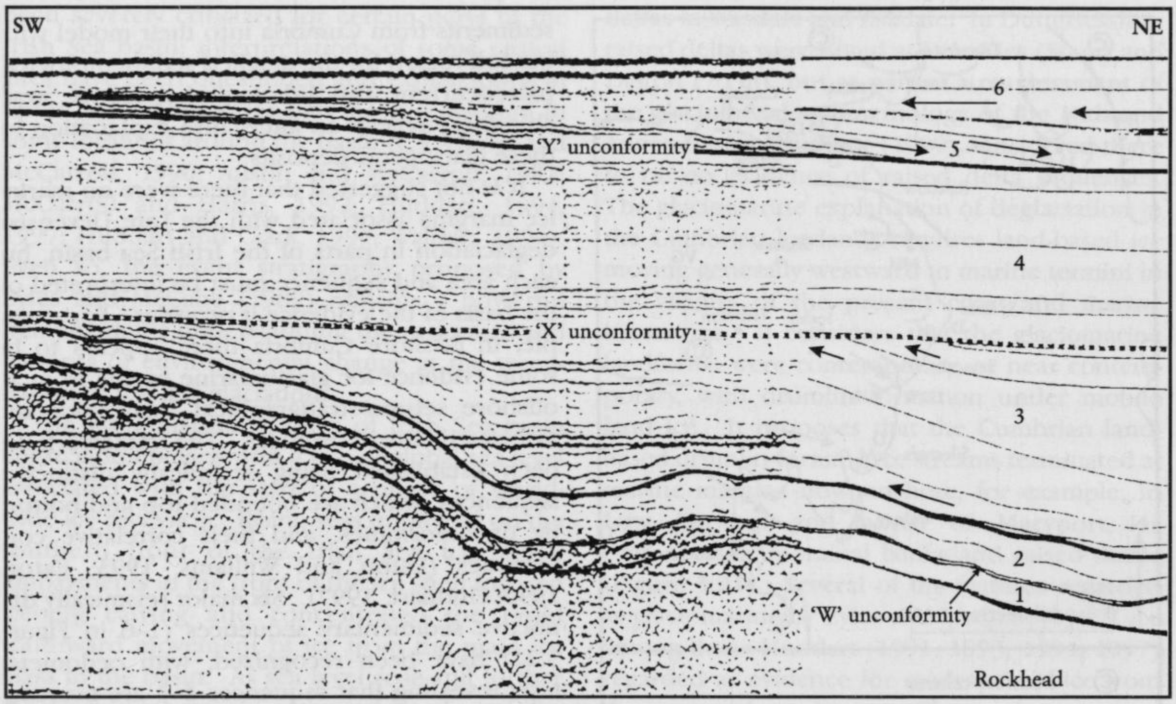
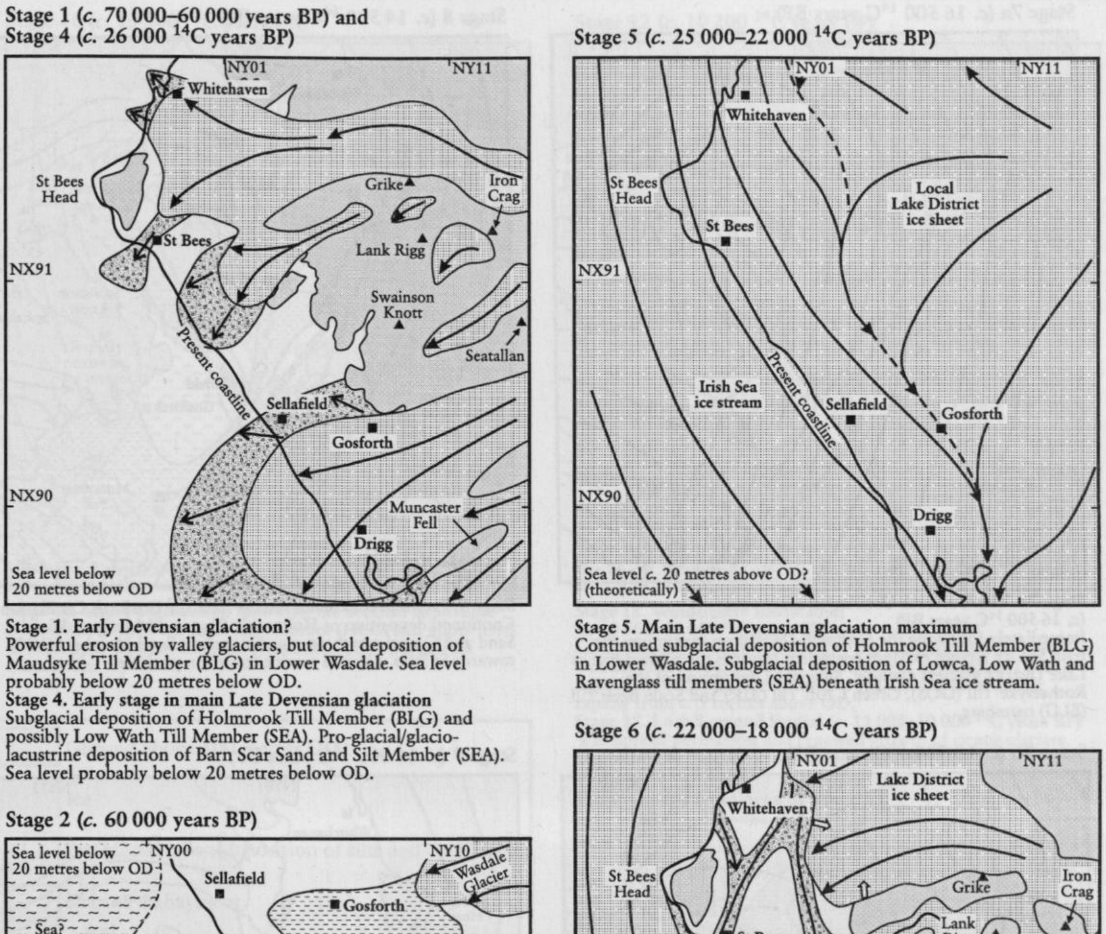


Figure 5.22 Seismic stratigraphical architecture from offshore Cumbrian locations (seismic line No 89/29, shot points 2621–2630) (after Nirex, 1997b).

16 000–18 000 years BP for the W unconformity and the Main Phase deglaciation. There is evidence for broad-amplitude folding of glaciotectonic origin within glacio-proximal sediments along the offshore coastal fringe. They are the correlatives of the onshore pro-glacially deformed sediments, as at St Bees and Drigg. These types of structures are not commonly preserved offshore as the deformed sequences preserved onshore have been removed by subglacial erosion. The compaction-related folding seen in sequences 2–3 is related to glaciotectonic processes, because it is thought to have been formed by ice loading during the Scottish Readvance.

There is thus some evidence here for glaciomarine conditions in the Irish Sea basin and it is how to relate contemporary relative sea levels with the deglaciation of the Dimlington Stadial ice sheet where most controversy still lies. The integration of the Nirex data with pre-existing data from the Lake District and Irish Sea basin has allowed a tentative model of the build up and decay of this ice sheet to be suggested as a series of cartoon sketch maps (Akhurst *et al.*, 1997; Huddart, 1997; Merritt and Auton, 1997a, b, 2000; Thorne *et al.*, 1997). These are illustrated in Figure 5.23. In the Main Phase deglaciation, downwasting occurred early, there was subglacial meltwater erosion to produce extensive channel sequences, ice-marginal lakes formed in favourable topographical locations between the high ground and the ice covering the coastal plain and the Irish Sea basin. Probably as a result of rapid iceberg calving into extensive pro-glacial lakes or the sea, the ice front retreated northwards quickly. Merritt in Heathcote *et al.* (1997) suggests that the lower reaches of valleys between Uldale and Wasdale in south-west Cumbria became inundated to a greater extent, perhaps by the sea. The ice front advanced to pond water in lower Wasdale and the Gosforth lowland. The ice front subsequently retreated actively to the north-west, a retreat that would have been punctuated by minor readvances. Following an unknown period of time, but perhaps only a few hundred years, during which the sea may have flooded the coastal lowlands to *c.* 10–15 m, the ice front readvanced a second time from the north-west, just impinging on the coast to produce the extensive zones of pro-glacial glaciotectonic thrusting that occur from St Bees southwards. Deposits of glacio-lacustrine and possibly



Stage 1. Early Devensian glaciation?
Powerful erosion by valley glaciers, but local deposition of Maudsyke Till Member (BLG) in Lower Wasdale. Sea level probably below 20 metres below OD.

Stage 4. Early stage in main Late Devensian glaciation
Subglacial deposition of Holmrook Till Member (BLG) and possibly Low Wath Till Member (SEA). Pro-glacial/glacio-lacustrine deposition of Barn Scar Sand and Silt Member (SEA). Sea level probably below 20 metres below OD.

Stage 2. Retreat of Early Devensian glaciers
Deposition of Carleton Silt Formation in a pro-glacial lake occupying Lower Wasdale.

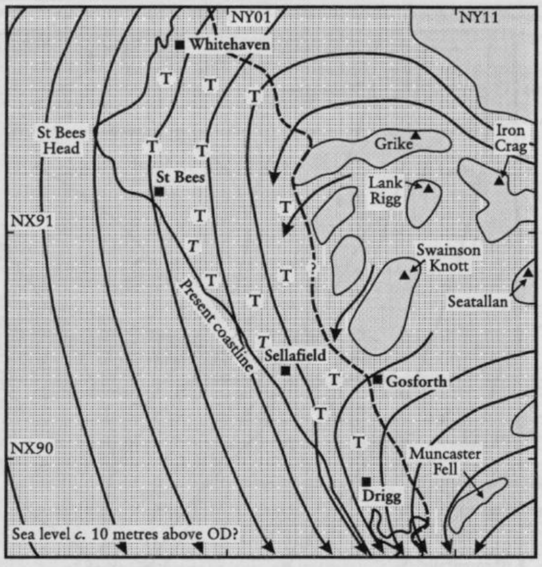
Stage 3. Middle Devensian marine inundation
Deposition of Glannoventia Formation in Lower Wasdale

Stage 5. Main Late Devensian glaciation maximum
Continued subglacial deposition of Holmrook Till Member (BLG) in Lower Wasdale. Subglacial deposition of Lowca, Low Wath and Ravenglass till members (SEA) beneath Irish Sea ice stream.

Stage 6. Early deglaciation after main Late Devensian glaciation maximum
Deposition of: glacio-lacustrine Ehen Valley Silt (SEA), St Bees Silt (SEA) and Whinneyhill Coppice Clay (AIK) members; deltaic St Bees (SEA) and Ehen Valley Sand and Gravel members (SEA); and glaciomarine Upper Western Irish Sea Formation (following deglaciation of northern Irish Sea basin).

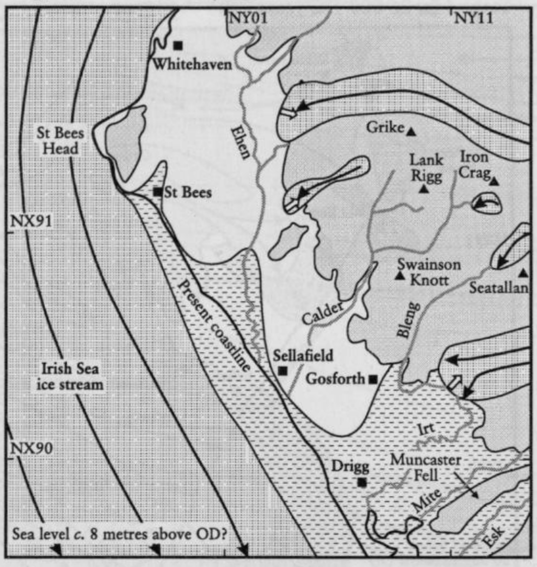
Figure 5.23 Cartoons to illustrate the suggested glacial stages during the Late Quaternary history of the Sellafield area, Cumbria. For key, see page 125. (After Nirex, 1997b; Clark and Smith, 1998.)

Stage 7a (*c.* 16 500 ^{14}C years BP)



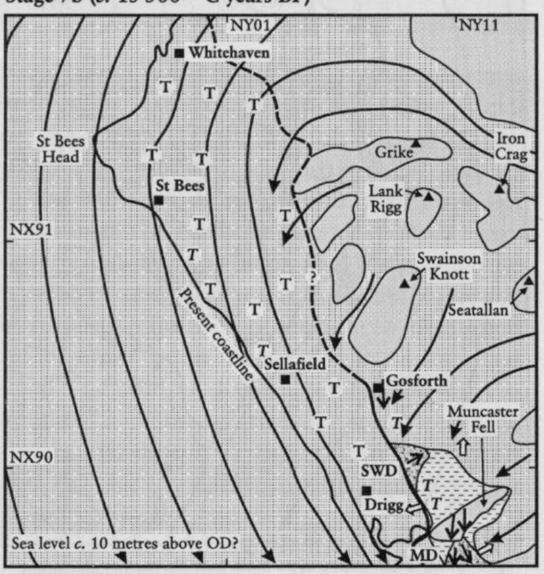
Stage 7a. The Gosforth Oscillation maximum (*c.* 16 500 ^{14}C years BP)
Immediately following the Gosforth Oscillation maximum, the Irish Sea ice stream expands eastwards towards the pecked line as Lake District glaciers dwindle. Subglacial deposition of the Rothersyke Till (GOS), Green Croft Till (AIK) and Scale Beck Till (BLG) members.

Stage 8 (*c.* 14 500 ^{14}C years BP)



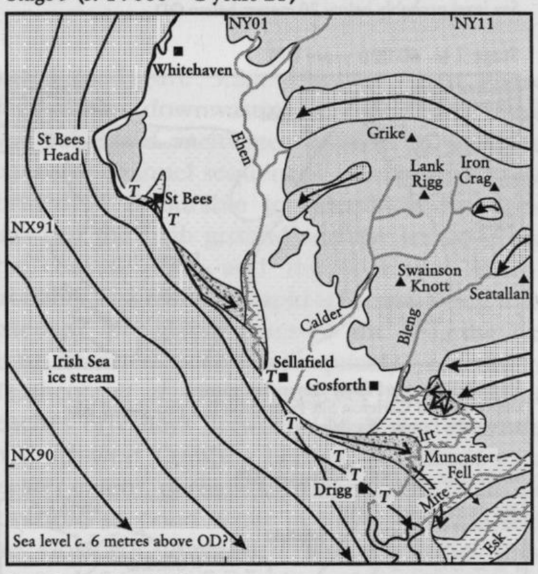
Stage 8. Probable maximum retreat after the Gosforth Oscillation
Continued deposition of Holmeside Clay and Mainsgate Wood Sand and Gravel members, also Gutterfoot Sand Member (GOS) towards St Bees. Limited marine incursion possible.

Stage 7b (*c.* 15 500 ^{14}C years BP)



Stage 7b. Early deglaciation after the Gosforth Oscillation (*c.* 15 500 ^{14}C years BP?)
Deposition of glacio-lacustrine Holmeside Clay (AIK) and deltaic Mainsgate Wood Sand and Gravel (AIK) members in Lower Wasdale following retreat of the Wasdale Glacier.

Stage 9 (*c.* 14 000 ^{14}C years BP)



Stage 9. The Scottish Readvance
Creation of St Bees push moraine. Subglacial deposition of the How Man, Fishgarth Wood and Drigg Beach till members (GOS), together with interbedded units of sand and gravel. Deposition of pro-glacial Peel Place Sand and Gravel Member (GOS) and glacio-lacustrine Drigg Moorside Silt and Meadow House Clay members (GOS). Incision of the Irt–Mite misfit valley.

Figure 5.23 (*contd*) Cartoons to illustrate the suggested glacial stages during the Late Quaternary history of the Sellafield area, Cumbria. For key, see page 125. (After Nirex, 1997b; Clark and Smith, 1998.)

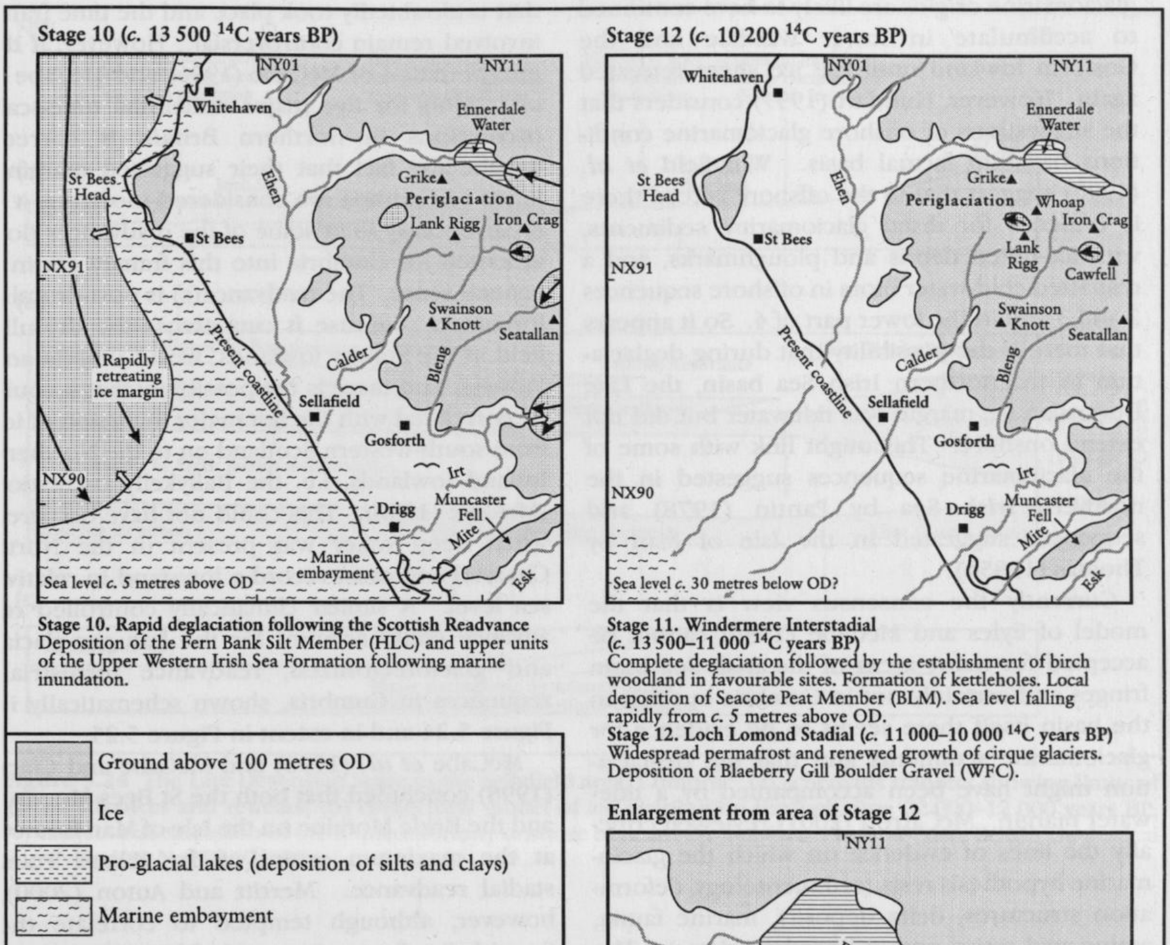


Figure 5.23 (*contd*) Cartoons to illustrate the suggested glacial stages during the Late Quaternary history of the Sellafield area, Cumbria (after Nirex, 1997b; Clark and Smith, 1998).

glaciomarine origins are likely to have continued to accumulate in lower Wasdale and the Gosforth lowland until the ice sheet retreated again. However, Huddart (1997) considers that the suggestions of onshore glaciomarine conditions have no factual basis. Wingfield *et al.* (1997) suggest that in the offshore record there is evidence for distal glaciomarine sediments, with ice-rafted debris and ploughmarks, and a restricted cold-water biota in offshore sequences 2 and 3 and in the lower part of 4. So it appears that there is the possibility that during deglaciation in the northern Irish Sea basin, the Late Devensian ice margin was tidewater but did not extend onshore. This might link with some of the glaciomarine sequences suggested in the northern Irish Sea by Pantin (1978) and sequences suggested in the Isle of Man by Thomas (1985b).

Currently the consensus view is that the model of Eyles and McCabe (1989) cannot be accepted for many areas of the Irish Sea basin fringes and certainly not in Cumbria but that in the basin itself there seems to be evidence for glaciomarine sediments and that the deglaciation might have been accompanied by a tidewater margin. McCarroll (2001) reviewed critically the lines of evidence on which the glaciomarine hypothesis rests (sedimentology, deformation structures, delta deposits, marine fauna, amino-acid ratios and radiocarbon dates). The sedimentological interpretation of many Irish Sea basin sections has been challenged, it is argued that subglacial sediments are common rather than rare and there is widespread evidence of glaciotectonism, as at St.Bees. Any delta deposits are the result of local ponding or occur where glaciers from different sources are uncoupled. They do not record past sea levels and the ad hoc theory of 'piano-key tectonics' is not required to explain the irregular pattern of altitudes. The cold water foraminifera interpreted as *in situ* he regarded as reworked from older Irish Sea sediments. Amino-acid age estimates used in support of the glaciomarine model he regarded as unreliable. However, radiocarbon dates from distinctive foraminiferal assemblages in north-east Ireland show that glaciomarine sediments do occur above present sea level, but they are restricted to low altitudes in the north-west of the basin.

As the sediment distribution and stratigraphical sequence is not that well known in the basin the sequences of retreat phases and readvances that undoubtedly took place and the time frame involved remain controversial. However, if the interpretation of McCabe (1996) and McCabe *et al.* (1998) for five major millennial time-scale oscillations in northern Britain is correct, despite the fact that their suggested marginal limits in Cumbria are considered incorrect, it is relatively easy to fit some of the readvances documented for Cumbria into this framework in a general sense. The readvance(s) is post-drumlin formation because it cuts across the drumlin field in the Solway lowlands, west Cumbria and Furness, and there is the possibility that it could be correlated with the last major incursion of ice from south-western Scotland on to the Northern Ireland lowlands, i.e. the Ballykelly oscillation (McCabe, 1996). This could not have occurred when deep water was present in the North Channel and must record a lowstand in relative sea level. A similar climatically controlled readvance could account for the late pro-glacial and glaciotectonized, readvance terrestrial-sequences in Cumbria, shown schematically in Figure 5.24 and in extent in Figure 5.25.

McCabe *et al.* (1998) and McCabe and Clark (1998) concluded that both the St Bees Moraine and the Bride Moraine on the Isle of Man formed at the maximum extent of the Killard Point stadial readvance. Merritt and Auton (2000), however, although tempted to correlate the Scottish Readvance (*sensu* Huddart 1970, 1991) with this readvance, drew attention to the fact that McCabe *et al.*'s (1998) model showed far more ice present at that time in north-west Cumbria than envisaged by Huddart (1970, 1991). Merritt and Auton (2000) suggested that their model fitted more closely with that of the Gosforth Oscillation, which involved synchronous readvances from Wasdale and the Irish Sea basin, together with drumlinization in north-west Cumbria. However, they recognized that if this correlation is made then there is little time available for a subsequent Scottish Readvance after 14 000 years BP. Hence they suggested that the Gosforth Oscillation might correlate with the Belderg Readvance recognized in western Ireland at *c.* 17 000 years BP (McCabe, 1996) or even the Dimlington Advance on the Yorkshire coast after 18 500 years BP.

In north-east England there is further controversy, because Eyles *et al.* (1994) have speculated on the role of fine-sediment trapping in extensive pro-glacial marine or freshwater bodies in promoting the southward extension of ice

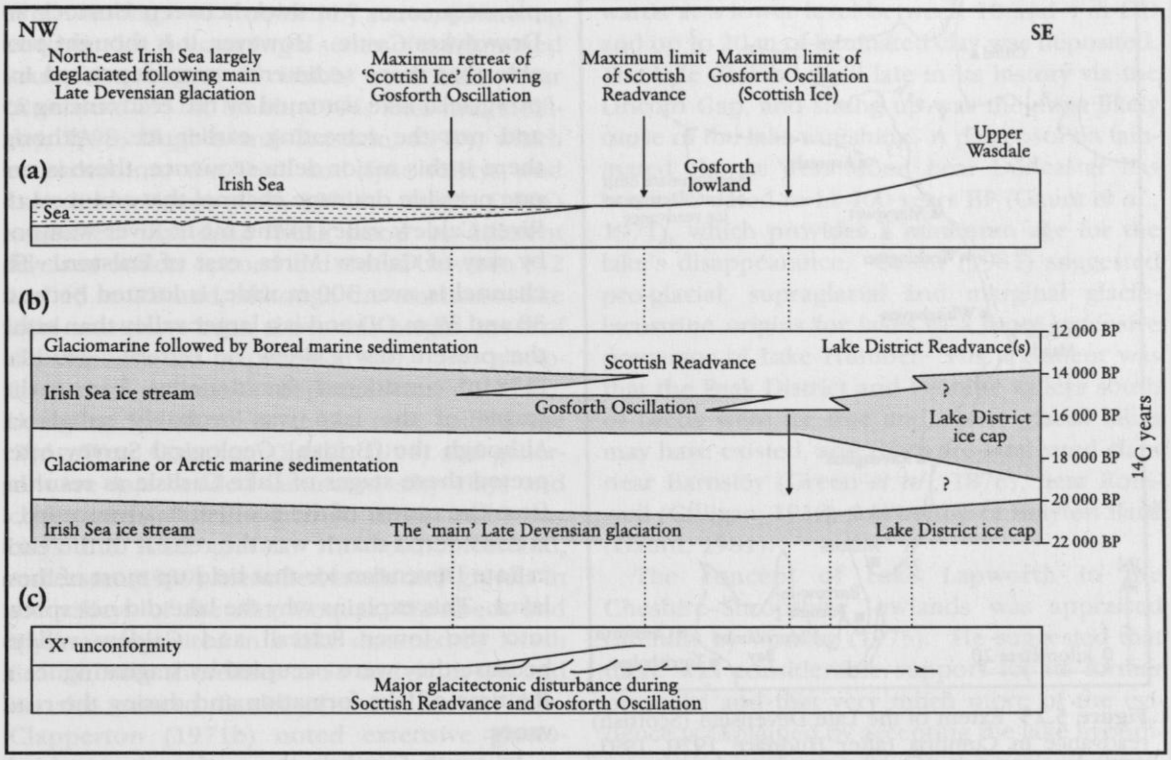


Figure 5.24 The Late Devensian Stage in the Sellafield area, Cumbria: (a) schematic transect showing limits of glacial advances and retreats; (b) conjectural model of ice distribution through time 22 000–12 000 years BP; (c) conjectural extension of the X unconformity onshore beneath the Gosforth lowland (after British Geological Survey Report No.WA/97/15C).

lobes along the east coast of England as they overran deformable sediment and surged. Teasdale and Hughes (1999) also suggest the possibility that marine influences were close, both in time and space, to the eastern coastal ice lobes. For example, the Dogger Bank Formation in the North Sea, which is thought to be the distal equivalent of the Bolders Bank Formation (clay-rich, Devensian, lodgement tills) (Balson and Jeffery, 1991), contains a dinoflagellate flora indicating shallow, open marine waters.

Glacio-lacustrine deposits in northern England

Northern England provides some of the classic glacio-lacustrine sequences in Britain. Many of these are linked to Kendall's (1902) pioneering work on ice-dammed lakes and the origin of meltwater channels (see 'Glacial meltwater erosion section', pp. 100–101). After Kendall's 'overspill' theory was challenged by Sissons (1958a, 1960b, 1961) very few ice-dammed lakes were recognized. Sissons' own work (1977, 1978, 1979a, 1982) on the classic ice-dammed lakes of the Scottish Highlands, established the criteria for unequivocal evidence: shorelines linked to overflow/drainage routes, deltas and lake sediments. Several examples are discussed below and in the site reports of Warren House Gill (Chapter 4), Holme St Cuthbert and Newtondale (this chapter). Most of the lakes are associated with ice movement from lowland areas against the topography, or retreat from such situations, and the larger examples are situated in extensive basins.

In Cumbria there is good evidence in the Carlisle plain for a lake dammed by retreating Late Devensian ice, which unlike the marginal zones of this ice sheet, maintained an active ice front (Huddart, 1970, 1981c). A suite of deltas between 66 and 30 m above OD was built out by the proto-Eden and Irthing rivers. At the earliest stage (66–60 m OD) the lake was confined to the Irthing valley (with deltas around Irthington and in the lower Cam Beck valley) and at its maximum was only 5 km by 3 km. Great thicknesses of lake-bottom sediments have been exposed

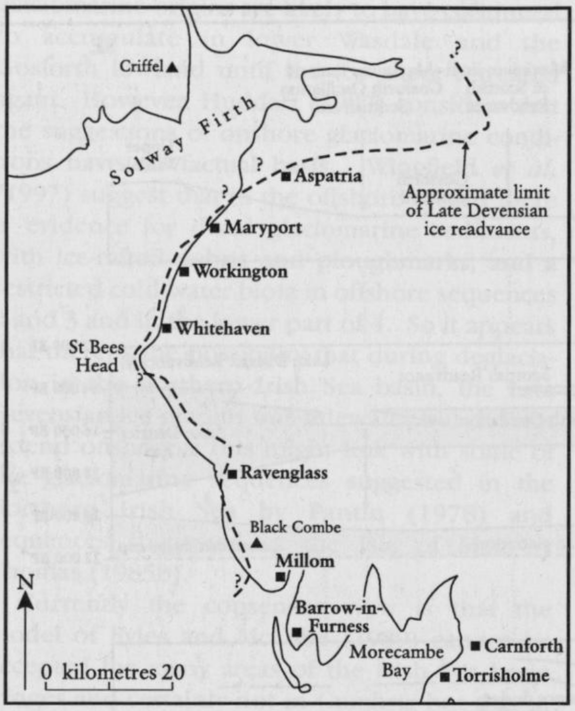


Figure 5.25 Extent of the Late Devensian (Scottish) readvance in Cumbria (after Huddart, 1970, 1991, 1994).

around Great Easby, between Castle Bank and Boothby Bank. At the latter site there is a sequence of nearshore silts and mudflows, whereas at Great Easby there are individual, 50-cm-thick, proximal varves with current structures, convolute lamination and then diamicton units (Huddart, 1970). The 54 m OD stage is recorded by deltas around the Cam Beck and around Wetheral. They indicate that the lake grew in size as the ice retreated. Below this the delta from Warwick Bridge to Aglionby indicates a lake level between 37 and 44 m OD and this stage is indicated by the flat surface on which Carlisle airport operates. Farther west a lake at 30 m OD is indicated by the Crosby Moor delta and that the delta now dissected by Collar Beck around Scotby Park and Rose Hill. The latter delta has been modified by an ice readvance that deposited a till on its slopes but did not overtop it (Huddart, 1971b).

No lower lake levels have been found to the west, but this must be because of the later readvance ice, which reworked the lake sediments. Lake-floor sediments have been found below an upper till at several locations in the Carlisle plain, east and north-east of Carlisle, for example, sequences 9 m thick between Linstock and Drawdykes Castle. However, it is thought probable that these sediments were deposited in a pro-glacial lake dammed by the readvancing ice, and not the retreating earlier ice. Although there is this major delta sequence, there is only one possible drainage channel that connects the River Caldew valley to the misfit River Wampool by way of Caldew Mires, east of Dalston. The channel is over 300 m wide, is located between 30 and 39 m OD and is a larger valley than either the present day Caldew or Petteril. Huddart (1981c) considered that drainage from earlier stages of the lake was probably subglacial. Although the [British] Geological Survey interpreted these stages of Lake Carlisle as resulting from the retreat of the Scottish Readvance ice, it is considered that it was the retreat of the earlier Late Devensian ice that held up most of these lakes. This explains why the lake did not spread into the lower Petteril and Caldew valleys, because they were occupied by stagnating ice at the time of lake formation and during the readvance.

In south Cumbria the readvance stage(s) of the Late Devensian Irish Sea ice pro-glacially dammed lakes in the St Bees valley, the lower Eden valley, lower Wasdale and the Gosforth lowlands, Eskdale, the Black Combe coast and Whicham valleys. Here the ice was advancing against the regional drainage lines. In lower Wasdale the thickness of glacial sediments proved in borehole logs is over 80 m and most seems to be glacio-lacustrine in origin (Huddart, 1970, 1981c; Huddart and Tooley, 1972). Although most of the lakes postulated by Smith (1912, 1932) in lower Wasdale, Eskdale and Miterdale were based on supposed overflow channels, deltas and beaches, these have been re-interpreted by Smith (1967) and Huddart (1970) as a product of subglacial and ice-marginal processes, as there are examples of kame terraces in lower Wasdale and subglacially engorged eskers on the flanks of Corney Fell above the major submarginal or subglacial channel systems (Huddart and Tooley, 1972). However, there is evidence in the form of lake sediments and deltas in lower Wasdale and the Gosforth lowlands that lakes formed during one of the readvance stages.

In north-east England at the close of the Late Devensian, eastward-flowing meltwater from the western Scottish, Lake District and Pennines ice, which was stagnant in the valleys and downwasting, was dammed by active, advancing northern ice, so creating a series of ice-dammed lakes, the largest of which were Lake Wear (Raistrick, 1931b; Smith, 1994; Mills and Holliday, 1998; Hughes and Teasdale, 1999), which extended into the Team and Tyne valleys, and Lake Edder Acres (Catt, 1991a). Smith (1994) suggested that Lake Wear stood at different levels as outlets opened and closed between 132 and 43 m OD and, although the maximum lake level must have been above this height, most of the deposits are below 90 m OD. The glaciolacustrine sediments deposited in this lake, between 5 and 55 m thick, were called the 'Tyne–Wear Complex' (Smith, 1994) and generally are interbedded laminated silty clays and clayey silts, with fine-grained sands and stony clays and some gravels. In Northumberland, Anderson (1939) identified sands and gravels in the valleys of the Aln, Coquet, Wansbeck and Blyth as deposited in a lake dammed by North Sea ice, with shorelines at 58 and 42 m OD and on the eastern flanks of the Cheviot Hills, Clapperton (1971b) noted extensive glaciolacustrine deposits.

Farther down the east coast there is evidence of large ice-dammed lakes, resulting from eastern ice lobes blocking the Tees estuary area, in the form of laminated clays and shoreline sands (Agar, 1954; Beaumont, 1967); Lake Pickering (see Newtondale and Hole of Horcum GCR site report) and Lake Humber. Beaumont (1967) described an Upper Tees Clay up to 100 m OD, which had very few, small stones, was up to 6 m thick and had a weak or random fabric. He thought that in the early stages of ice advance into the Tees lowlands there was ponding of subaerial drainage in a lake. Lake Pickering was dammed at both its westerly and easterly ends by ice and overflowed into the Vale of York via the Kirkham Abbey Gorge in the Howardian Hills (Catt, 1977a). Lake Humber formed when the Humber Gap was blocked by ice in the east at first and subsequently by a plug of glacial moraine and by ice fronts in the Vale of York (Lewis, 1887a, b; Gaunt, 1974, 1981). The lake developed up to 33 m OD between the ice to the north, the Pennines to the west, Lincoln Edge to the south and the Yorkshire Wolds to the east. In high-level sand and gravel deposits associated with this lake, a bone fragment from Brantingham (Gaunt, 1974), either just below or within these deposits, was dated to 21 853 years BP. During deglaciation the lake extended northwards at a lower level between 10 and 4 m OD and up to 20 m of laminated clay was deposited. The lake drained until late in its history via the Lincoln Gap, and silting up was the most likely cause of the lake vanishing. A palaeosol on laminated clay at West Moor, near Doncaster has been ^{14}C dated to 11 100 years BP (Gaunt *et al.*, 1971), which provides a minimum age for the lake's disappearance. Gaunt (1981) suggested pro-glacial, supraglacial and marginal glaciolacustrine origins for lakes in a more extensive precursor of Lake Humber. His argument was that the Peak District and Pennine valleys south of Leeds were ice free and hence glacial lakes may have existed, and there are laminated clays near Barnsley (Green *et al.*, 1878), near Rothwell (Gilligan, 1918) and on top of Brayton Barff (Gaunt, 1981).

The concept of Lake Lapworth in the Cheshire–Shropshire lowlands was appraised carefully by Worsley (1975). He suggested that there was considerable support for its former presence and that very much more of the evidence is explained by accepting the lake hypothesis than by rejecting it. On the basis of shorelines, laminated sediments, overflow channels and the nature of the glaciofluvial morphology (see Aqualate Mere site report, this chapter), Dixon (in Whitehead *et al.*, 1927) recognized Glacial Lake Newport and Wills (1924) recognized Glacial Lake Buildwas. On ice retreat these amalgamated to form the larger Lake Lapworth, which must have overflowed through the Ironbridge Gorge at some stage. Interest in Lake Lapworth was revived by Poole and Whiteman (1961), who extended the original definition of the lake to include a lake extending from Ironbridge as far north as Chorley in Lancashire. Although this concept remains extremely doubtful, Worsley (1970), in a review of Pleistocene events in the Cheshire–Shropshire lowlands, considered that the data available gave support to the glacial lake hypothesis, with the lake having the general dimensions of the classic Lake Lapworth. Shaw (1972a), however, stated that there was no evidence for a widespread lake during the retreat of the Irish Sea ice in Shropshire.

Drumlin formation

One of the most conspicuous, widespread and well-developed glacial depositional landforms, composed largely of till, are the subglacial drumlin fields, superimposed drumlins and drift tails, which largely are found in the lowlands of the region. Mitchell (1991a, b, 1994), however, describes drumlins up to 600 m in south-east Cumbria and neighbouring North Yorkshire. The main drumlin fields occur in the lower Tweed valley in Northumberland (Clapperton, 1970), in the Solway lowlands and along the coastal plain north-west of the Lake District (Huddart, 1970; King, 1976), in Edenside (Hollingworth, 1931), in the south-east Lake District around Kendal, through the Yorkshire Dales south of Settle, in the Ribble valley and the west Craven lowlands (Raistrick, 1933; Johnson, 1985b), in the southern Lake District valleys, across the Furness peninsula (Gresswell, 1962; Grieve and Hammersley, 1971; Huddart *et al.*, 1977) and into north Lancashire (Longworth, 1985). It has been suggested by Johnson (1985b) that the Morecambe Bay–Furness–Southern Lake District drumlins are located where local relief helped impede the movement of ice southwards, and according to Boulton *et al.* (1977) this would cause the sliding velocity of the basal ice to decrease to a point to which lodgement tills would be deposited and moulded into drumlins by the ice. Such changes took place where the Cumbrian valley glaciers were coalescing into larger piedmont glacier ice-streams.

The origin of drumlins, summarized in Bennett and Glasser (1996) and Benn and Evans (1998), is controversial and has given rise to numerous models, hypotheses and explanations (Menzies, 1979a, b, 1989; Menzies and Rose, 1987a, b, 1989). Any theory must be able to produce a mechanism for drumlin forms, a mechanism for promoting the unstable amplification of relief and a mechanism for quenching the unstable amplification once drumlins have reached a critical size (Hindmarsh, 1999). The following mechanisms have been proposed:

1. Drumlins are the product of subglacial deformation, produced where there is a deforming till layer that moulds itself around subglacial obstacles. This idea has been developed by Boulton (1987) and currently seems one of the more likely explanations (Boyce and Eyles, 1991; Hart, 1997; Menzies *et al.*, 1997). Hindmarsh (1999) argues the case for drumlinization and relief amplification being the consequences of the viscous deformation of till and that shock formation is a convincing explanation for some aspects of drumlin form.
2. They are the result of subglacial lodgement where till accretes by lodgement around subglacial obstacles.
3. They are erosional remnants of subglacial water sheet-floods or fluvial infills. This somewhat controversial origin involves large subglacial sheet-floods that scour the base of an ice sheet, creating cavities, similar in morphology, but much larger, to scour marks (Shaw, 1983,1989; Shaw and Kvill, 1984; Shaw and Sharpe, 1987; Shaw *et al.*, 1989). These are then infilled by fluvial sediment or subglacial till. Alternatively, subglacial meltwater erosion during the flood may dissect the bed and leave scours, which subsequently are streamlined by ice flow into drumlins. Floods in North America are thought to have been caused by the drainage of large pro-glacial and subglacial lakes (Shoemaker, 1991, 1992a, b, 1995, 1999).
4. They are the result of glaciomarine sedimentation followed by drumlinization. Here the conditions of ponded water, high meltwater pressures and high ice velocities in a surge allow a high preservation potential for the drumlins. Stratified sequences are deposited in water-filled cavities found on the lee-side of drumlins, subglacial shearing modifies the drumlin form during and following lee-side deposition, deforming the upper stratified beds, producing a till carapace and a streamlined form. Whilst most of the evidence for this hypothesis has come from an Irish context (e.g. Dardis and McCabe, 1983, 1987; Dardis, 1985, 1987; McCabe and Dardis 1989, 1994; McCabe, 1991; Dardis and Hanvey, 1994), Eyles and McCabe (1989), Eyles *et al.* (1994) and McCabe (1996) do suggest links between the extensive and sedimentologically complex drumlin bedforms that occur not only in Ireland and south-west Scotland, but in north-western England and eastern England as well, and a rhythmicity in ice-sheet activity recorded by oscillations in ice-sheet margins largely into glaciomarine environments.

Summary

Glacial events affecting northern England have left the strongest imprint on the landscape that we see today. During the Quaternary Period the area was glaciated numerous times and subjected to repeated cycles of ice-sheet growth and decay. The landform and sedimentary record contains evidence of both full ice-sheet (glacial) conditions and restricted ice-cap/cirque-glacier (stadial) events. Evidence for glaciation prior to the Late Devensian is fragmentary and most of the landforms and sediments date from this most recent glacial episode. Large quantities of rock and sediment were eroded from upland areas and transported to the lowland fringes of the ice sheets and to offshore depocentres. These changes in the landscape were accompanied by wholesale reorganization of the major drainage patterns. Palaeoglaciological reconstruction is still poorly constrained, and although parameters such as ice-sheet extent, thickness, flow patterns, thermal regime, and their change over time can be used to enhance our understanding of climate changes during the Quaternary Period, much of these data are still lacking. The GCR sites described in this chapter contribute to the resolution of the following key debates:

1. The historical development of the tripartite system and its interpretation as a monoglacial or bi-glacial sequence.
2. Stratigraphical correlation between onshore and offshore events, including the timing and extent of glaciation in the Irish Sea basin and the east of England.
3. The nature of glacial sedimentation and palaeoenvironmental history of the Irish Sea basin, and, in particular, whether this was achieved in a glacial terrestrial or glaciomarine setting.
4. The processes of drumlin formation, their use in palaeoglaciological reconstruction and their significance for glacial dynamics.
5. The historical debate concerning the nature and extent of ice-dammed lakes and overflow channels created during ice recession.
6. The origins of glaciofluvial landforms associated with active or stagnating ice sheets.

CHELFORD (SJ 824 717)

POTENTIAL GCR SITE

N.F. Glasser

Introduction

The sediments at Chelford, in Cheshire, constitute one of the key Devensian sequences in Great Britain and the site is considered as the stratotype for the Chelford Formation (Bowen, 1999). The importance of Chelford to the British Quaternary record is underlined by the fact that the site was proposed originally by the Geological Society of London Quaternary Working Group as the type locality for the last glaciation in Britain (Shotton and West, 1969), although it was later replaced by Four Ashes (Mitchell *et al.*, 1973). An organic layer within the Chelford Formation has provided biostratigraphical evidence for an Early Devensian Interstadial event known as the Chelford Interstadial (Simpson and West, 1958) and the site is now recognized both nationally and internationally as the stratotype for the 'Chelford Interstadial'. Biostratigraphical evidence for this interstadial event has been presented by Coope (1959) and by Simpson and West (1958). The importance of the site was increased further by the discovery of glaciogenic sediments (Figure 5.26) belonging to the Oakwood Formation beneath the Chelford Formation (Worsley *et al.*, 1983). The Oakwood Formation provides evidence of a stadial event prior to the Chelford Interstadial.

The precise chronostratigraphical position of the Chelford Interstadial within the British Quaternary record has been the subject of much controversy. This primarily is because the age of the organic remains that constitute the biostratigraphical horizon in the Chelford Formation lies at or beyond the limits of conventional radiocarbon dating methods and because of doubts about possible contamination of samples (Worsley, 1980). Radiocarbon dating of the Chelford organic layer was also first conducted at a time when laboratory procedures for the technique were at the pioneer stage, the limitations of the technique were not fully understood and when British Quaternary stratigraphy was not fully established. Recent advances in the dating of organic material have allowed independent dating of the Chelford organic layer. This has now confirmed that the organic material was deposited around 86 ka, well beyond the limits of conventional radiocarbon dating (Heijnis and Vanderplicht, 1992). Other technological advances, including those in thermoluminescence dating techniques that can be applied to quartz and feldspar mineral fractions in sand, also mean that the host sediment can be dated directly. Thermoluminescence dating of the Chelford Formation has yielded an absolute age for the Chelford Interstadial of 90–100 ka

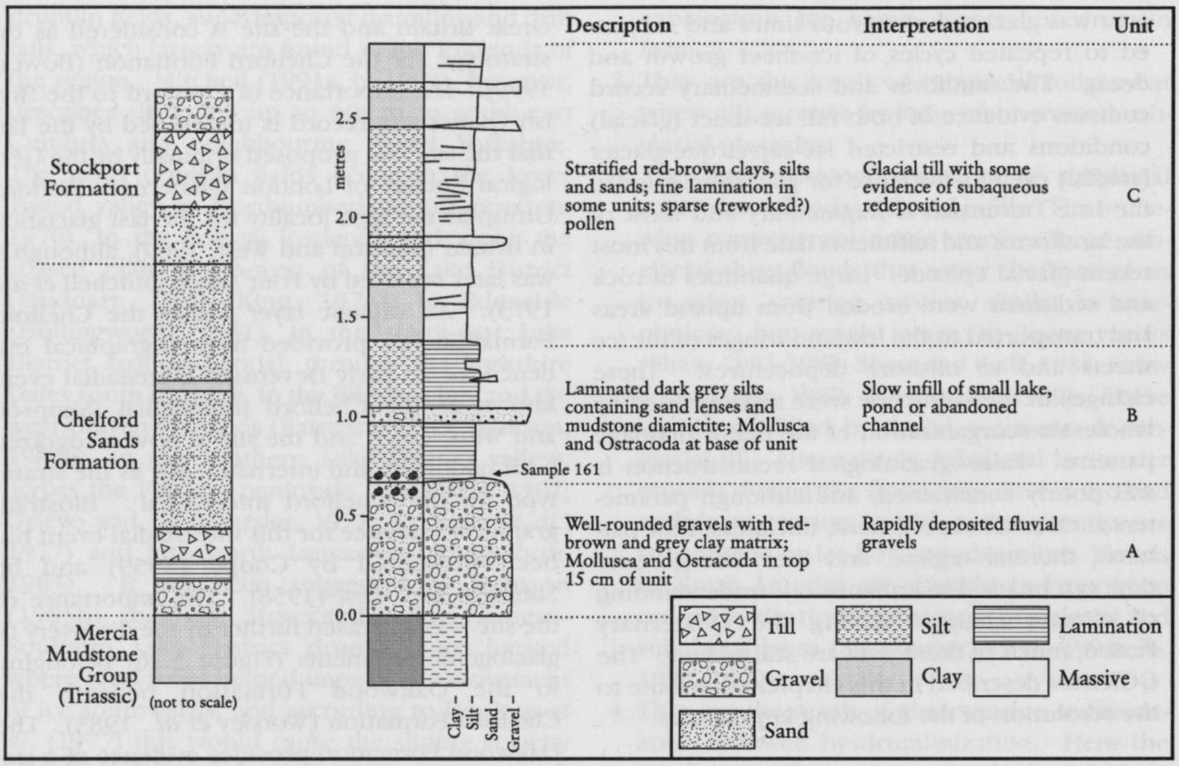


Figure 5.26 The stratigraphy recorded at Oakwood Quarry, Chelford (after Worsley *et al.*, 1983).

(Rendell *et al.*, 1991; Rendell, 1992) and generally it is now accepted that the Chelford Interstadial is of Early Devensian age. More recently, data from the site have been used in a wide ranging assessment of the palaeoclimatic conditions prevailing in north-western Europe during the last interglacial to glacial transition, assembled from palaeobotanical, coleopteran and periglacial evidence (Aalbersberg and Litt, 1998).

Description

The village of Chelford is located approximately 10 km west of Macclesfield in Cheshire. Descriptions of the Pleistocene deposits in the area come from three separate but closely spaced sand quarries south-east of the village. The original site described by Simpson and West (1958) was the working sand quarry at Farm Wood, Chelford (SJ 812 731). This quarry was worked between 1955 and 1973 but is now 'restored', all faces have been graded and no permanent exposures exist. More recent descriptions of the Chelford deposits (e.g. Worsley *et al.*, 1983; Rendell *et al.*, 1991; Rendell, 1992) are from sand quarries at Dingle Bank (SJ 807 718) and Oakwood Farm Quarry (SJ 824 717).

Lithostratigraphy

The original succession described by Simpson and West (1958) from Farm Wood Quarry is as follows:

Unit	Thickness (m)
4. Upper Sands	0–1
3. Upper Boulder Clay	0–5
2. Middle Sands (including a bed of organic mud up to 1 m thick	16–20
1. Lower Boulder Clay	not exposed by quarrying

A complete description of the Quaternary sediments down to bedrock was never obtained from the original Farm Wood Quarry owing to the nature of the working practices at the site. Partly because of this, Simpson and West (1958) considered the Chelford Formation to be a glaciofluvial deposit sandwiched between two glacial tills (an Upper Boulder Clay above and a Lower Boulder Clay below). Boulton and Worsley (1965), however, disputed the existence of this Lower Boulder Clay beneath the Chelford Formation on the basis of borehole data. Boulton and Worsley (1965) concluded that the Chelford Formation lay directly on weathered Mercia Mudstone bedrock. Subsequent exposures at the new Oakwood Quarry (Figure 5.26) in the late 1960s, however, required a re-appraisal of this conclusion when a second glaciogenic deposit (the Oakwood Formation) was recorded beneath the Chelford Sands (Worsley *et al.*, 1983). This glaciogenic deposit comprises a thin bed of red till including shale fragments, locally derived Carboniferous and Permo-Triassic sandstone and quartzite pebbles, together with igneous clasts of a presumed Lake District origin (Worsley *et al.*, 1983). The upper surface of the till has scattered ventifacts, suggesting a period of subaerial wind deflation (Thompson and Worsley, 1967). This thin, patchy till is also recognized in the area surrounding Chelford, although it is rarely more than 1.7 m thick (Evans *et al.*, 1968). The stratigraphical position of this till (informally named the 'Oakwood Till' by Worsley *et al.* (1983), and now formally known as the 'Oakwood Formation' (Bowen, 1999)), is difficult to explain as anything other than a stadial event prior to the deposition of the Chelford Formation itself. The precise timing of this glacial event is unknown (Worsley *et al.*, 1983). Finally, immediately beneath the Oakwood Till, Worsley *et al.* (1983) described a thin sequence of in-situ fossil-rich gravel and silt (the Lapwing Bed) resting directly on unweathered Mercia Mudstone bedrock. Pollen, plant macrofossils, Mollusca, Coleoptera and Ostracoda assemblages have all been recovered from this gravel and silt horizon.

Biostratigraphy

Two important biostratigraphical horizons occur at Chelford. The first of these, the organic lenses within the Chelford Formation, have yielded the abundant floral and faunal remains that represent the Chelford Interstadial. Simpson and West (1958) and Coope (1959) have published detailed accounts of this flora and fauna. The second important biostratigraphical horizon is the thin sequence of fossiliferous silt and gravel of the Lapwing Bed, immediately below the Oakwood Till. This unit lies stratigraphically well below the Chelford Formation. The biological evidence in this sequence includes pollen, plant macrofossils, Mollusca, Coleoptera and Ostracoda assemblages (Worsley *et al.*, 1983). The floral and faunal assemblages in these two horizons are described below.

Biostratigraphical evidence in the Chelford Formation

The insect fauna from the Chelford Formation consists almost entirely of Coleoptera (beetle) remains, described in detail by Coope (1959). A summary of the Chelford insect fauna is given in Table 5.3 and a full species list can be found in Coope (1959). The largest family present is Carabidae, members of which inhabit sandy, damp or marshy ground. Other important species represented include the carnivorous water beetle family Dytiscidae, the carnivorous beetle family Staphilinidae and the click beetle family Elateridae. The palaeobotany of the organic mud is provided in Simpson and West's (1958) account of the site. The vegetation is dominated by conifer and birch forest with a ground flora chiefly of *Calluna*, *Empetrum* and other ericaceous plants.

Table 5.3 Coleoptera from the Chelford Formation (data from Coope, 1959).

Family	Number of species
Carabidae	30
Dytiscidae	12
Hydrophilidae	4
Silphidae	3
Leiodidae	1
Staphilinidae	13
Elateridae	8
Helodidae	1
Byrrhidae	4
Coccinellidae	2
Scarabaeidae	2
Cerambycidae	1
Chrysomelidae	7
Curculionidae	10
Scolytidae	2

Biostratigraphical evidence for a glacial event prior to the Chelford Interstadial
The silt and gravel of the Lapwing Bed have also yielded important fossil material useful for biostratigraphy (Worsley *et al.*, 1983). Although pollen is sparse and generally poorly preserved, a pollen record from the silt has been produced. The pollen assemblage is dominated by non-arboreal species such as Cyperaceae and Gramineae, with a scarcity of tree pollen. Aquatic plants are well represented, with *Alisma*, *Lemna*, *Nuphar* and *Nymphaea* suggesting the occurrence of standing or slow-flowing water of between 0.5 and 1 m in depth. Plant macrofossils also have been obtained from the silt, and are again dominated by aquatic plants that require slow-flowing water. *Potamogeton filiformis*, *Eleocharis palustris* and *Hippuris vulgaris* grow in shallow water around 0.5 m in depth at the margins of pools. The mosses *Drepanocladus aduncus* and *Scorpidium scorpioides* are represented by abundant leafy stems, suggesting that they were growing in shallow water rather than in marshes. The molluscan fauna is also dominated by aquatic species. The abundance of *Lymnaea* spp., *Anisus leucostoma* and *Armiger crista* is indicative of shallow standing or slow-flowing water with abundant submerged vegetation. Species characteristic of large bodies of open water such as *Valvata piscinalis* are notably absent. The three species of *Pisidium* recorded all prefer muddy substrates. The terrestrial mollusca are all of species that generally live in proximity to water bodies, such as *Oxyloma pfeifferi*, *Succinea oblonga* and *Vertigo genesii*, and have modern ranges that extend into Arctic areas. The Ostracoda assemblage is dominated by species of *Candona*, *Ilyocypris* and *Limnocythere*. Their occurrence together again indicates a stagnant or slow-flowing water body, such as a stagnant vegetated pool. The insect fauna, although dominated by a total of 55 coleopteran taxa, also contains Diptera, Hymenoptera and Trichoptera. Aquatic, riparian and terrestrial communities are all present in the silt. The overall picture from the coleopteran evidence is one of a shallow, perhaps ephemeral, pool surrounded by marsh plants such as sedges and mosses. Farther away from the marsh the soil became gradually better drained with patchily developed vegetation of grass and weeds. There is no evidence that trees grew in the vicinity and a meagre mammalian fauna must have lived at this time. The absence of trees suggests an average July temperature below 10°C, and the climatic regime indicated is boreal or boreo-montane with an element of continentality. Overall, the faunal and floral assemblages indicate that the deposits accumulated in a shallow pool within a treeless, open landscape (Worsley *et al.*, 1983).

Chronostratigraphy

Determining the exact age of events at the site has proved problematic and somewhat controversial. No less than 14 separate radiocarbon dates have been obtained for the Chelford Interstadial from laboratories in the UK, the USA, Holland and Germany. Dates for the organic remains at Chelford lie across a wide range between 25 and 65 ka. As the limits of conventional radiocarbon dating lie at around 45 ka, it became apparent as far back as the late 1960s that the organic remains within the Chelford Formation lay at or beyond the limits of this dating technique. Worsley (1980) has discussed in some detail the problems associated with previous attempts to date the Chelford Interstadial using conventional radiocarbon methods. Recent advances in dating techniques, in particular in the field of thermoluminescence dating, mean that it is now possible to obtain dates directly from the quartz and feldspar mineral fractions of sands. These techniques, when applied to the Chelford Formation, have yielded an absolute age for the Chelford Interstadial of 90–100 ka (Rendell *et al.*, 1991; Rendell, 1992). Independent support for this age also comes from uranium-series dating of the Chelford organic layer itself, which has yielded an age of 86 ka (Heijnis and Vanderplicht, 1992). The lower part of the Chelford Formation probably pre-dates the Ipswichian interglacial (Oxygen Isotope Sub-stage 5e) (Bowen, 1999). The age of the interstadial pre-dating the Chelford Interstadial is less certain, but is ascribed to Oxygen Isotope Stage 6 by Bowen (1999).

Interpretation

According to Worsley *et al.* (1983) the Pleistocene succession in the Cheshire–Shropshire area consists of two basic formations. These are the Stockport Formation and the Chelford Formation. The Stockport Formation is a suite of glaciogenic deposits including till, glacio-lacustrine and glaciofluvial facies. The Stockport Formation is effectively the formal lithostratigraphical equivalent of the old tripartite sequence (Upper Boulder Clay, Middle Sands and Gravels, Lower Boulder Clay) that is recognized throughout the north-west of England and Wales (Bowen, 1999). Below the glaciogenic Stockport Formation, and separated from it by a major unconformity, is the Chelford Formation. The Chelford Formation is exposed only in deep excavations and is an exceptionally pure and well-sorted sand succession, containing in places lenses of organic peats and muds. The Chelford Formation is now known to occupy a SE–NW orientated palaeovalley, and the organic remains are confined to the fill of palaeochannels within the sands (Worsley, 1977; Worsley *et al.*, 1983). Simpson and West (1958) originally considered the Chelford Formation and associated organic remains to be part of a glaciogenic succession, although Boulton and Worsley (1965) later interpreted the sands as a fluvio-aeolian succession of alluvial fans with a dominant east to west flow direction. They regarded the organic deposits as belonging to a brief interstadial within this periglacial regime.

In his study of the organic remains within the Chelford Formation, Coope (1959) concluded from the good levels of preservation indicated that the beetles had not undergone significant post-mortem transportation. The Chelford fauna therefore can be regarded as a community of species that did in fact co-exist in the immediate vicinity. He further summarized the detailed environmental conditions that best fit the present-day requirements of the coleopteran assemblage at Chelford. The organic mud represents sedimentation in a pond choked with vegetation debris and partially overgrown by *Sphagnum*. There was little open water in the pond, which was primarily stagnant and acidic. Sporadic floods were required in order to carry in tree trunks and dead leaves. Immediately adjacent to the pond was marshy ground dominated by reeds, rushes and willow with sandy banks in places. More open sandy heath surrounded the pond with abundant birch and conifers, with numerous fallen trees and stumps. The scarcity of dung and carcass beetles suggests that there were few large animals in the area. Most of the Coleoptera have northern affinities, suggesting a colder climate than at present. Although most of the beetles found at Chelford are cosmopolitan species, Coope (1959) found that the entire faunal assemblage occurred in parts of present-day Fennoscandia. In particular, the faunal list closely matches that of southern Finland between 60° and 64°N. Average annual precipitation in this area is 600 to 650 mm. The palaeoclimatic interpretation is that of a mean annual temperature around 1°C, with temperatures during the warmest month (July) around 15°C and temperatures during the coldest month (February) around −11°C. The overall climatic reconstruction is therefore that of a continental climate with long, cold and dry winters. This is remarkably similar to the climatic reconstruction made from the palaeobotanical evidence by Simpson and West (1958). In particular, the presence of *Picea abies*, no longer native to the UK, is a clear indication of the continentality of the climate.

Mutual Climatic Range analyses of the coleopteran evidence from the Lapwing Bed (pre-dating the Chelford Interstadial) suggest that during Oxygen Isotope Stage 6 (Bowen, 1999), the mean temperature of the warmest month (July) was around 9.5°C (Worsley *et al.*, 1983). The mean temperature of the coldest months (January and February) lay somewhere between −19°C and −28°C (Worsley *et al.*, 1983). These estimates indicate an increase in continentality, with a temperature amplitude between the warmest and coldest months approaching 30°C.

Conclusions

Chelford is clearly a site of the utmost importance to British Quaternary stratigraphy. Two separate interstadial events are represented at this locality, one of which (the Chelford Interstadial) has been the subject of a concerted dating campaign. Beneath this, the Oakwood Formation contains the only reliable evidence in this part of the country for a glaciation prior to the Middle Devensian. The lithostratigraphical, chronostratigraphical and biostratigraphical evidence from Chelford is widely used in comparisons with other Quaternary sites both in the UK and in continental Europe. The volume of published material on the site, its rich and varied floral and faunal assemblages that serve as biostratigraphical reference markers, and its status as the stratotype for the Chelford Interstadial make the site of great significance to the British Quaternary record.

FOUR ASHES (SJ 914 083)

N.F. Glasser

Introduction

Four Ashes, Staffordshire, was originally chosen as the type site for the Devensian in Britain (Mitchell *et al.*, 1973) and the site has recently been formally designated as the stratotype for the Four Ashes Formation (Bowen, 1999). The name 'Devensian' is inherited from the Devenses, a British tribe formerly resident in this region. In brief, the sedimentary succession at Four Ashes shows a complex of fluvial sands and gravels (the Four Ashes Formation) overlain by the glaciogenic Stockport Formation. The total thickness of this sequence is 11 m. The sands and gravels contain lenses and beds of organic remains that have yielded important palaeoecological and palaeoclimatological information. The presence of periglacial features both within and below the gravels has also provided important palaeoenvironmental information on the sequence of events during the Devensian. A.V. Morgan (1973) has described the sequence at Four Ashes. A. Morgan (1973) has investigated the insect fauna from the site, and Andrew and West (1977) have conducted pollen analysis on the organic sediments. The palaeoecological data and radiocarbon dates from the site suggest that it is partly Ipswichian in age (Oxygen Isotope Sub-stage 5e) and partly Early and Middle Devensian (Oxygen Isotope Sub-stages 5d to Oxygen Isotope Stage 3) (Bowen, 1999). Shotton (1977) has discussed the difficulties of using type sites such as Four Ashes to delimit the stages of the Quaternary. Four Ashes is referred to in virtually every textbook dealing with the Quaternary in Britain (e.g. West, 1977; Jones and Keen, 1993; Lowe and Walker, 1997a).

Description

Four Ashes is a disused gravel pit on the outskirts of the village of Four Ashes in Staffordshire. Although at the time of writing largely overgrown, the succession in the pit has been documented on the basis of formerly more extensive exposures. The stratigraphy in the pit has been described in some detail by A.V. Morgan (1973). Very briefly, this consists of a lower unit of gravels considered by A.V. Morgan (1973) to have been deposited on a fluvial braid plain, overlain by cryoturbated gravels and till (Figure 5.27). Within the gravel succession, and at its base where it rests on Triassic sandstone bedrock, there are pockets of organic material (largely organic silts and peats). These organic sediments have provided material for radiocarbon dating and palynological analysis. Also present at a number of levels are ice-wedge casts, interpreted as the result of freeze–thaw penetration of the ground under periglacial conditions. The sedimentary succession is outlined in more detail below.

Gravel unit

A.V. Morgan (1973) described the sand and gravel unit as varying between 0.45 and 4.6 m in thickness. It consists almost entirely of Bunter-derived quartzite pebbles and rare erratics of flint, tuff, rhyolite and andesite. The sand and gravel sequence is extremely complex, with evidence of both erosion and deposition in the form of wash-out gullies, minor stratigraphical breaks, and current and graded bedding sequences. At least two layers of intraformational ice-wedge casts and involutions have been described in the gravels and the upper 2.5 m of the gravel sequence is frequently strongly involuted (Morgan, A.V., 1973).

One of the most significant features of the Four Ashes gravel unit is the presence of lenses of sandy detritus peat and organic clays. Organic material infilling bedrock hollows has yielded macrofossils of *Taxus*, *Ilex* and *Alnus*, and a pollen assemblage dominated by *Alnus* and *Quercus* (Andrew and West, 1977). Other organic sediments in contact with bedrock contain a pollen assemblage of *Pinus*, *Picea* and *Betula*. The insect fauna from these sediments was considered in detail by A. Morgan (1973), who described three separate faunal groups. The group 1 fauna was collected from detritus muds lying directly on bedrock and are composed of six species of Coleoptera associated with trees. The group 2 fauna was collected from variable contexts, including detritus peats and clays, although all were located in the lower parts of the gravel sequence. The characteristic feature of the fauna in this group is that they all contain large numbers of phytophagous beetles not associated with tree species. Group 3 faunas were also collected from variable contexts, but at all levels in the gravel sequence. The assemblages of this group are characterized by large

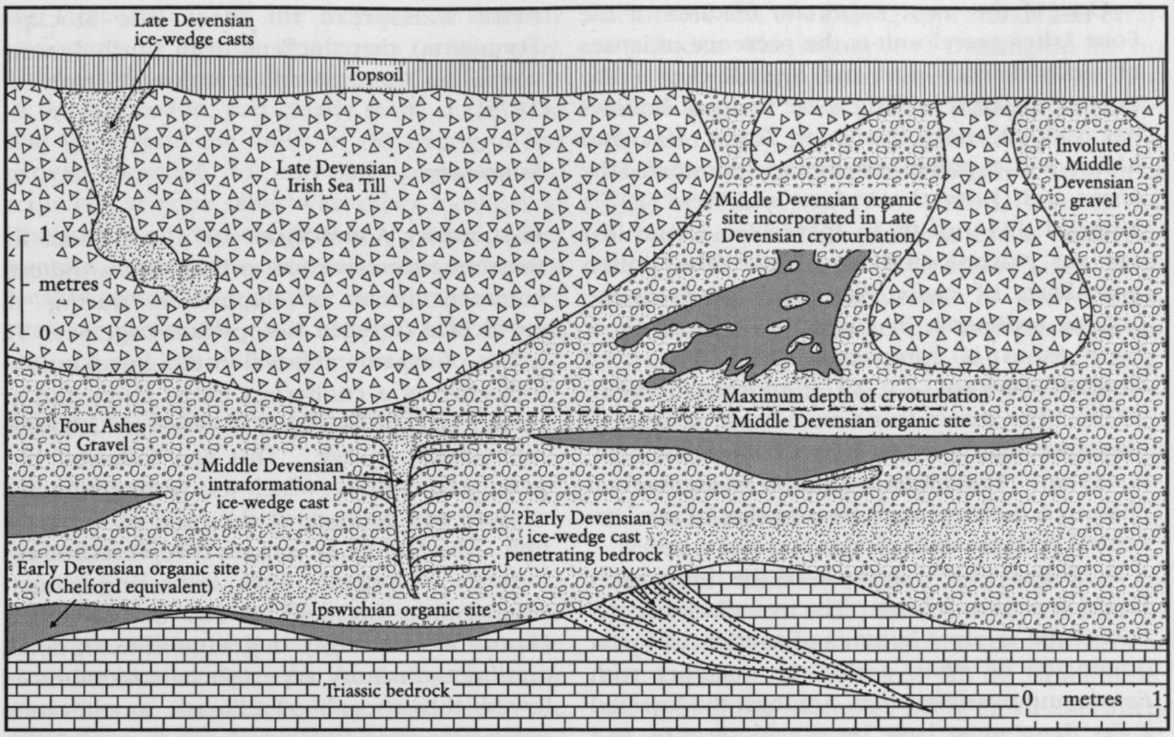


Figure 5.27 General stratigraphy at Four Ashes (after Morgan, A.V., 1973).

numbers of beetles that live on tundra or open ground.

Diamicton

A.V. Morgan (1973) describes a red diamicton unit overlying the gravels at Four Ashes. The diamicton varies from 0.05 to 2.75 m in thickness and has a high erratic content, including white granite clasts attributed to south-west Scotland. Other erratic lithologies include striated volcanic and slate clasts from the Lake District and Scotland, Eskdale granite, Ennerdale granophyre, occasional limestones, flints derived from Cretaceous chalk, and marine shells. In places, the diamicton is disturbed by periglacial features, including ice-wedge casts and involutions.

Interpretation

Gravel unit

A.V. Morgan (1973) considered the organic deposits at the base of the gravels to be Ipswichian in age. If this interpretation is correct, the Four Ashes Formation must represent deposition during the Lower and Middle Devensian. Thus the gravel unit records 40 000 years (minimum) of deposition in a sequence between 0.45 and 4.6 m thick. From this, it follows that the Four Ashes Formation was deposited by a relatively small stream system with limited discharge. The high content of Bunter-derived clasts points to a local source area. A.V. Morgan (1973, p. 241) envisaged that 'the streams transporting the Four Ashes gravel must have been choked with large quantities of gravel and sand, probably changing direction frequently, anastomosing over the underlying gravel and sands'. The implication of this description is that there was a continuous cycle of erosion and deposition throughout the Devensian. Although this sedimentary description is typical of glaciofluvial deposition, A.V. Morgan (1973) states categorically that there is no suggestion that the Four Ashes Formation is glaciofluvial in origin. The only evidence for climatic deterioration during deposition of the gravel is in the faunal record (Morgan, A., 1973) and the presence of ice-wedge casts.

One of the most important features of the Four Ashes gravel unit is the presence of lenses of sandy detritus peat and organic clays. The organic material infilling these bedrock hollows has yielded macrofossils of *Taxus*, *Ilex* and *Alnus*, and a pollen assemblage dominated by *Alnus* and *Quercus*. Andrew and West (1977) assigned this unit to an Ipswichian (Zone IIb) age on palynological grounds. The pollen assemblage of *Pinus*, *Picea* and *Betula* in the organic sediments in contact with bedrock at Four Ashes is comparable with that from Chelford (Morgan, A.V., 1973; Andrew and West, 1977). The insect fauna from these sediments also is analogous to that from Chelford (Morgan, A., 1973). The fauna overall is of continental character, perhaps analogous to that of modern central Fennoscandia (Jones and Keen, 1993).

A. Morgan (1973) interpreted the three insect faunal groups as follows. The overall interpretation of the group 1 fauna is of coniferous forest dominated by *Pinus* and *Picea*, together with *Betula* and possibly *Salix*. Amongst the forested areas there must have been acid swamps and small pools with sedges around the margins. There is very little evidence of a rich herbaceous vegetation or of bare open ground. The environment indicated by group 2 species is one with a fairly luxuriant vegetation of mosses, aquatic and semi-aquatic plants around small pools. The substrate in these swampy areas would have been rich in humus and slightly acidic. There was little open ground, some grassland and a variety of plant species. The Coleoptera in group 2 indicate that this environment remained treeless, except for some dwarf birch and willow. The overall view of the area from the group 3 fauna is of an open, barren environment. Vegetation cover was low and was composed mainly of moss and grass with various aquatic and semi-aquatic plants growing in the moister areas around small pools. All indications point to a treeless environment.

Diamicton

A.V. Morgan (1973) interpreted the red diamicton unit overlying the gravels at Four Ashes as a typical Irish Sea till. The derivation of the erratics in the till indicates source areas in the Lake District and Scotland. The implied ice-movement direction is from the north-west to the south-east into Staffordshire. Regional mapping indicates that this till is part of a much larger and more widespread till sheet (the Stockport Formation) that thickens from north to south across the Wolverhampton region (Morgan, A.V., 1973).

Summary

In summary, A.V. Morgan (1973) concludes that the sequence recorded at Four Ashes indicates commencement of deposition during the Ipswichian (Oxygen Isotope Sub-stage 5e). The lenses of organic material in the gravel unit are indicative of a coniferous forest cover and are considered to be Early Devensian (Chelford Interstadial) in age. This was followed by a periglacial episode, which was severe enough to rupture the bedrock at the base of the gravel unit. The Middle Devensian is represented by a number of organic sites indicating climatic amelioration between 42 000 and 38 000 ka (Upton Warren stage). A gradual climatic deterioration and return to periglacial conditions followed this short-lived climatic amelioration. Some time after 30 500 ka but before 13 500 ka a lobe of the Irish Sea ice sheet advanced over the Four Ashes gravel to reach a maximum position near Wolverhampton. Recession of this ice sheet was followed again by periglacial conditions, during which time polygonal patterned ground and involutions formed in the till overlying the Four Ashes gravel.

More recently, Rose (1985) has pointed out that although the biostratigraphical evidence and radiocarbon dates from Four Ashes provide important information concerning Early and Middle Devensian climates, they do not closely constrain the timing of the glacial events during the Late Devensian. Thus the sequence at Four Ashes contains little lithostratigraphical evidence of the Late Devensian glaciation. At Dimlington, on the Holderness coast, Late Devensian glaciogenic sediments are sandwiched between radiocarbon-dated organic material, and provide a more accurate timing for the maximum advance of the Late Devensian ice sheet (Rose, 1985). Rose therefore proposed Dimlington as the type site for the Late Devensian glaciation, although the Four Ashes Formation still remains the formal stratotype for the Devensian (Bowen, 1999).

Conclusions

Four Ashes is a key reference locality because it has been designated as the stratotype for part of the Devensian. The site therefore is critically important to British Quaternary stratigraphy. Although there are difficulties with dating parts of the succession, particularly those at the base that lie beyond the limits of ^{14}C dating, it has proved possible to correlate parts of the sequence at Four Ashes with other localities at Chelford and Upton Warren. Four Ashes remains a stratotype site and a key reference locality against which other British Quaternary sites are compared. It is clearly a site of both national and international importance.

DIMLINGTON (TA 390 220)

D.J.A. Evans

Introduction

The cliffs undergoing rapid erosion at Dimlington, East Riding of Yorkshire, contain valuable stratigraphical evidence for the dating of the last glaciation of the British Isles, particularly the attainment of the glacial maximum in lowland England. The stratigraphy is best exposed at Dimlington High Land where it comprises Basement Till, which rests on chalk bedrock, overlain by the Dimlington Silts with enclosed organic remains and two late Devensian tills named the 'Skipsea Till' and 'Withernsea Till' (Catt and Penny, 1966; Penny *et al*., 1969; Catt, 1987b). Radiocarbon dates on the organic remains in the Dimlington Silts provide a maximum age for the overriding of the site by the late Devensian North Sea lobe of the British Ice Sheet, which deposited the Skipsea and Withernsea tills. This has resulted in the adoption of Dimlington as the climatostratigraphical type site and the use of the term 'Dimlington Stadial' for the late Devensian glacial episode in Britain (Rose, 1985). The time period 26 000–13 000 years BP is similarly referred to as the 'Dimlington Chronozone'.

Description

The glacial deposits of the Holderness coast have received considerable attention over the last 100 years (Table 5.4; Wood and Rome, 1868; Lamplugh, 1879, 1891a; Bisat, 1932, 1939, 1940; Catt, 1963; Catt and Penny, 1966; Mitchell *et al*., 1973; Madgett and Catt, 1978; Edwards, 1981; Evans *et al*., 1995). The terminology used here is that of Madgett and Catt (1978), who identified three tills of differing appearance and erratic content. The sequence at Dimlington lies on a marine platform cut in the chalk some 30–35 m below present beach level (Lamplugh, 1919; Catt and Digby, 1988). The lowest deposit is the Basement Till, also referred to as the 'Basement Clay' by Bisat (1939, 1940; Table 5.4), which lies directly on the chalk platform and is characterized by its dominant clay content, dark grey-brown colouring, inclusions of shelly, glauconitic sand and marine clay and erratics from Scotland and Scandinavia, as well as chalk fragments. Specifically, the Basement till comprises:

1. a massive diamicton with evidence of smeared inclusions and fold structures.
2. rafts of marine sediment.

The term 'Basement Series' has been used by Catt and Penny (1966) for this complex suite of deposits. Dimlington contains the most extensive exposure of Basement Till on Holderness, the only other examples being small and infrequently exposed in the Bridlington–Flamborough area.

The Scottish and Scandinavian erratics in the Basement Till are dominant in the inclusions of marine sediment, whereas local lithologies dominate the more massive diamicton within the deposit. Numerous fold and shear structures occur within the Basement Till, often manifest as smudges of chalk or preferentially weathered stringers of sand in an otherwise massive diamicton (Eyles *et al*., 1994). Pebble fabrics indicate a glacier flow direction from the north-east (Penny and Catt, 1967), but the upper layers are often disturbed by large folds and shear structures that cross-cut the fabric orientation (Catt and Penny, 1966; Penny and Catt, 1967; Eyles *et al*., 1994). Large rafts of older sediments are often particularly well exposed in the Basement Till at Dimlington. For example, dark blue marine clays with abundant shells, similar to deposits referred to as the 'Bridlington Crag' by Reid (1885), occur as rafts and smaller inclusions (Catt and Penny, 1966; Eyles *et al*., 1994). The faunal lists compiled for the blue clays by Reid (1885), Lamplugh (1884b, 1890) and Bell (1917, 1919) indicate a mixed assemblage of arctic littoral and deep-water species, although similar rafts at Bridlington contain species of a mixture of ages, including some as old as the Pastonian (Reid and Downie, 1973). Many shell fragments also can be found in the clay matrix of the

Table 5.4 Nomenclature of the Quaternary deposits of Holderness compared to the tripartite scheme of Madgett and Catt (1978) (from Evans *et al.*, 1995).

Dimlington Silts/ Basement Till	Skipsea Till	Withernsea Till	Source
		Hessle Till Brown Till	Lamplugh (1879)
	Gravel		
	Greenish-purple Till		
Basement Till Chalk rubble			
		Upper Till	Lamplugh (1891a,b)
	Interstratified series		
		Upper Purple Clays (2 beds) Gravels Lower Purple Clays (3 beds)	Bisat (1939, 1940)
	Sand, silt and gravel		
	Upper Drab Clay Middle Drab Clay Chalk rafts Lower Drab Clay Sub Drab Clay Basement Drab Clay		
Basement Clay Sub Basement Clay			
		Hessle Till Purple Till	Catt (1963) Catt and Penny (1966)
	Drab Till		
Dimlington Silts Basement Till (Series)			
		Upper Till Unnamed Till	Mitchell *et al.* (1973)
	Gravel		
	Lower Till		
Chalk rubble Basement Till Speeton Shell Bed			
		Upper Till Series	Edwards (1981)
	Gravel		
	Lower Till Series		
Chalk rubble Basement Till Chalk rubble Speeton Shell Bed			

Basement Till where it appears as a massive diamicton. The upper surface of the Basement Till at Dimlington is described as 'discoloured, decalcified and fragmented' by Catt and Penny (1966).

A number of depressions on the upper surface of the Basement Till contain laminated silts grading upwards into sands, a sediment body termed the 'Dimlington Silts' (Bisat and Dell, 1941; Catt and Penny, 1966; Figure 5.28). The Dimlington Silts occur as discontinuous lenses truncated by the overlying Skipsea Till and contain layers of moss (Bisat and Dell, 1941; Bisat, 1948), dominated by *Pohlia wahlenbergii* var. *glacialis*, in addition to insects and freshwater ostracods indicative of cold climatic conditions (Table 5.5). The bedding of the Dimlington Silts typically parallels the basin floor, but upper beds often display post-depositional disturbance in the form of folds, which appear to record stress

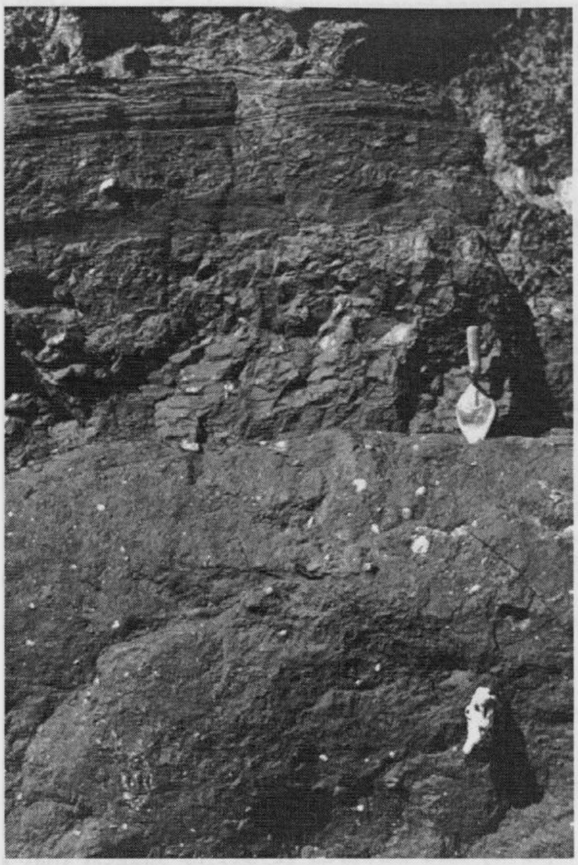


Figure 5.28 Dimlington Silts overlying Basement Till at Dimlington. (Photo: J. Rose.)

directions from the north and north-east (Catt and Penny, 1966). The alternation of sand- and silt-dominated laminae indicate changes in the sediment supply to the deep water environment in which the Dimlington Silts were deposited. Eyles *et al.* (1994) report a range of sedimentary structures from massive and laminated silty clays to interbedded silty clays and sands, including mud drapes over sand ripples (flaser bedding) and starved ripples. Radiocarbon dates of 18 500 years BP and 18 240 years BP were obtained on the moss layers by Penny *et al.* (1969), providing a maximum age for the deposition of the overlying Skipsea Till.

Truncating the Dimlington Silts and the Basement Till at Dimlington are the Skipsea and Withernsea tills. In previous classification schemes for the Holderness glacial sequence (Table 5.4) these deposits have been referred to as the 'Drab Till' and 'Purple Till' (Bisat, 1939, 1940; Catt and Penny, 1966; Catt and Madgett, 1981). The two tills are clearly differentiated by colour, erratic content, grain size and mineralogy (Madgett and Catt, 1978). Specifically, the Skipsea Till is very dark greyish brown, sandy in matrix texture, rich in garnet, hornblende, staurolite and kyanite, and is dominated by chalk and flint lithologies in addition to farther travelled igneous and metamorphic pebbles, including Scandinavian erratics. In contrast, the Withernsea Till is dark brown, possesses a clay-rich matrix, and is dominated by shale, siltstone and limestone lithologies. The textural and structural complexity of the Skipsea and Withernsea tills led Bisat (1939, 1940) to propose detailed stratigraphical subdivision of the deposits, but more recent research prefers a simple two-tier stratigraphy (Catt and Penny, 1966; Madgett and Catt, 1978; Catt, 1987b; Eyles *et al.*, 1994; Evans *et al.*, 1995). Early work on the Holderness glacial sequence identified a further more recent deposit, the Hessle Till, which Madgett and Catt (1978) later revealed to be a Holocene weathering profile in the Skipsea or Withernsea Till, depending on which deposit was at the surface.

The geographical distribution of the Withernsea Till (Figure 5.29) is far more restricted than the Skipsea Till although it is thought to extend some distance offshore (Cameron *et al.*, 1987). It extends a maximum distance of 10 km inland from the present coast but its landward margin trends offshore at Easington in the south and at Mappleton in the north (Catt and Penny, 1966; Madgett, 1975). The Withernsea and Skipsea tills are separated by a sharp erosional contact along which numerous discontinuous lenses of sands and silts and occasionally gravels lie in shallow scours on the surface of the Skipsea Till. Pebble fabrics taken from the Skipsea and Withernsea tills by Penny and Catt (1967), reveal a consistent north-east to south-west ice flow direction.

Interpretation

The abundance of marine fauna in the Basement Till led Penny (1959) to propose a marine origin for the deposit. However, the erratics, pebble fabrics and structures clearly indicate that the deposit was produced by subglacial transport (Catt and Penny, 1966; Penny and Catt, 1967) most probably as a deforming layer (Eyles *et al.*, 1994). The rafts of blue marine clay were eroded from pre-existing marine deposits by the

Table 5.5 The flora and fauna of the Dimlington Silts.

Coleoptera	
Agabus bipustulatus L.	
Aleocharinae indet.	
Amara alpina Paykull	
Amara quenseli Sch.	
Aphodius sp.	
Arpedium brachypterum Gr.	
Bembidion sp. (*lunatum* group)	
Bledius fuscipes Rye	
Byrrhus sp.	
Cercyon sp.	
Feronia blandulus Mill.	
Hydrobius sp.	
Notaris aethiops F.	
Ostracoda	
Candona neglecta Sars	
Cypridopsis vidua (Mull.)	
Cyprinotus salinus (Brady)	
Eucypris gemella Bodina	
Ilocypyris gibba (Ramdohr)	
Plants	
Daphnia ephippia	
Eleocharis palustris (L.)	
Menyanthes trifoliata (L.)	
Pohlia wahlenbergii (Web. & Mohr) *glacialis* (Schleich.)	
Potamogeton alpinus	
Potamogeton filiformis	
Trees	
Pinus (sparse pollen)	
Betula (sparse pollen)	

North Sea glacier lobe as it moved towards the Yorkshire coast. The occurrence of shells within the matrix of the Basement Till indicates that the massive diamicton has been produced by the deformation and kneading of pre-existing marine deposits. Because they occur in association with marine muds and arctic fauna, the Scottish and Scandinavian erratics in the blue marine clay rafts were most probably originally deposited on the North Sea floor by icebergs prior to the Dimlington Stadial glacier advance (Catt and Penny, 1966). Although Catt and Penny (1966) and Catt (1987b) have suggested that the Basement Till dates to a pre-Ipswichian glaciation, amino acid ratios on the incorporated shells indicate a Late Devensian age for the glacier advance (Eyles *et al*., 1994). Local erratics were incorporated into the Basement Till as it was transported on to the Yorkshire coastal lowlands from the North Sea as a subglacial deforming layer (Catt and Penny, 1966; Eyles *et al*., 1994).

The Dimlington silts are interpreted as lake deposits, which accumulated prior to the Skipsea Till glacier advance approximately 18 000 years BP. The folding in the silts and the Basement Till is interpreted as glaciotectonic in origin (Catt and Penny, 1966; Eyles *et al*., 1994), produced within a push moraine that eventually was overridden and capped by Skipsea Till. The flora and fauna of the organics enclosed within the Dimlington Silts (Table 5.5) indicate a shallow, freshwater lake surrounded by sparse vegetation and containing a restricted aquatic flora. This is typical of the environmental conditions in Britain leading up to the Last Glacial Maximum. The radiocarbon dates of 18 500 and 18 240 years BP are comparable with others used to constrain the start of the Dimlington Stadial in the region (Rose, 1985; Figure 5.30).

The clear differences in the Skipsea and Withernsea tills and the erosional contact that separates them have been used by Carruthers (1953), Catt and Penny (1966), Madgett and Catt (1978) and Edwards (1981) to suggest that a 'two-tiered' glacier was responsible for till deposition. Specifically, they suggest that superimposed tributary glaciers invaded the East Yorkshire coast during the Dimlington Stadial, each tributary carrying its own characteristic suite of erratics. The Skipsea Till was deposited by ice originating in Northumberland and southern Scotland and flowing southwards along the Yorkshire coast, whereas the Withernsea Till was deposited by a Tees valley ice stream that overrode the Skipsea Till ice and was then carried 'piggyback' style southwards to the Holderness area. Thus, the superimposition of Withernsea Till over Skipsea Till is explained by simultaneous deposition. Sedimentologically, more has been made of the stratified sediment bodies that occur between and within the Skipsea and Withernsea tills by Eyles *et al*. (1994). They suggest that the stratified sediments document periods of submergence by marine waters between onshore surges by the North Sea glacier lobe. A comparison is drawn between the glacial deposits of Dimlington and the pro-glacial sediments of the Sefstrom surging glacier in Svalbard, orginally described by Lamplugh (1911). In this depositional scenario the

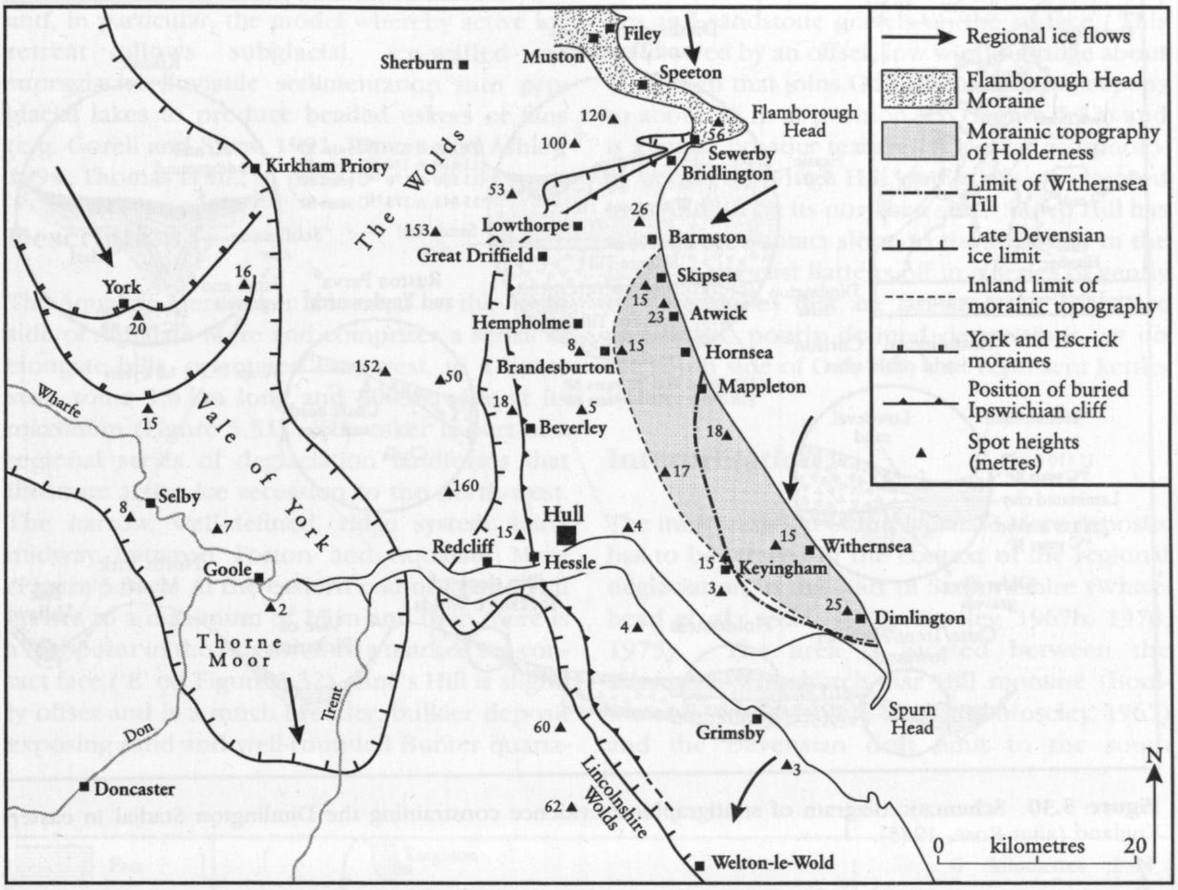


Figure 5.29 Map of Holderness showing the distribution of the Skipsea and Withernsea tills and important stratigraphical sites, based upon Eyles *et al.* (1994) and other various sources.

Basement, Skipsea and Withernsea tills are all regarded as the products of subglacial deformation of pre-existing marine or glaciomarine sediments during the Devensian glaciation. Although the occurrence of flaser bedding in the Dimlington Silts is diagnostic of tidal influences, the suggestion of a marine, rather than a lacustrine, origin for the stratified sediments in the Dimlington glacial sequence remains contentious. Nonetheless, the repeated onshore surging or regular (steady state) subglacial deformation theory (Eyles *et al.*, 1994; Evans *et al.*, 1995) appears to provide a more complete explanation of the similarities of the Skipsea and Withernsea tills than that provided by the 'piggy-back' or simultaneous deposition model.

The most recently revised correlation of Quaternary deposits in the British Isles (Bowen, 1999) has renamed the sedimentary and stratigraphical units discussed above in order to bring the nomenclature in line with standard lithostratigraphical terminology. The Basement Till is now referred to as the 'Bridlington Member', the Skipsea and Withernsea tills are called the 'Skipsea Member' and 'Withernsea Member' and the intervening sands and gravels are the 'Mill Hill Member'. The Dimlington Silts and their incorporated organic material are now labelled the 'Dimlington Bed'. A further unit, the 'Hornsea Member', has been provided for the hummocky terrain in the region, although it is not clear why this terrain is regarded as separate from and younger than the Withernsea Member. These members and the Dimlington Bed are all part of the Holderness Formation.

Conclusions

Dimlington is of considerable importance to British Quaternary stratigraphy because of its

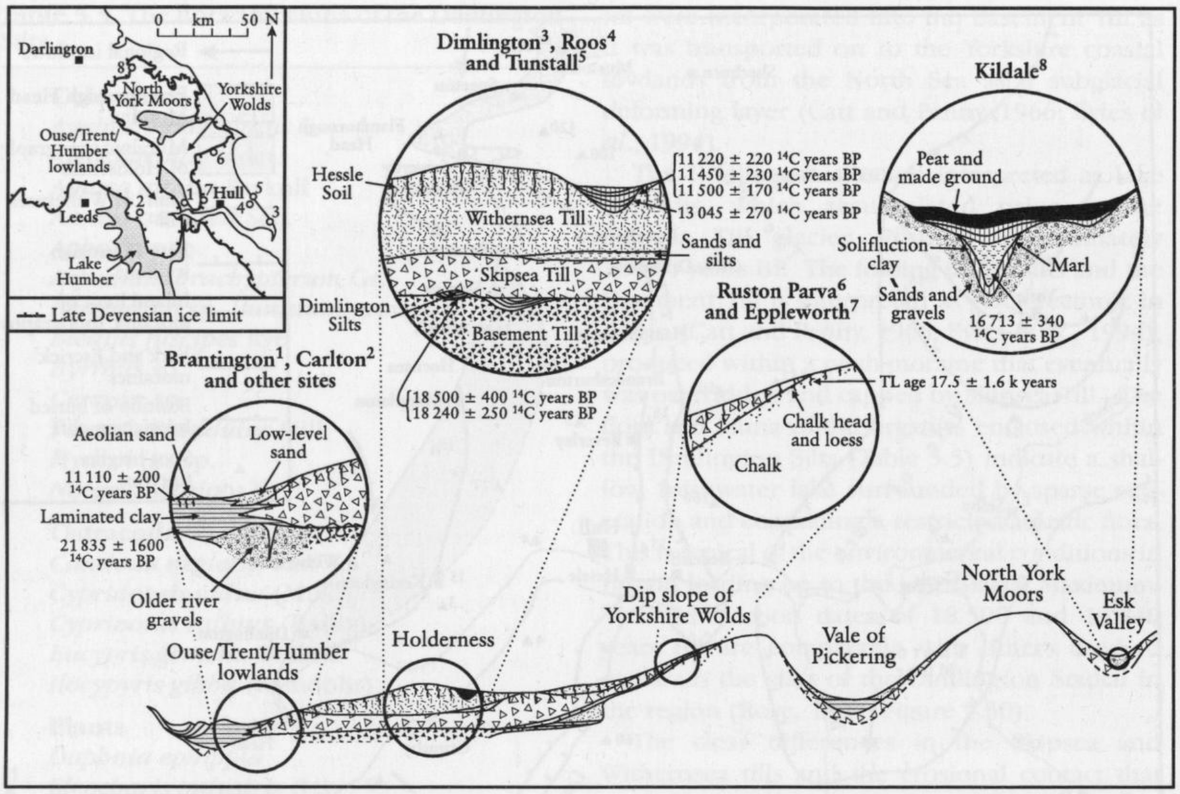


Figure 5.30 Schematic diagram of stratigraphical evidence constraining the Dimlington Stadial in eastern England (after Rose, 1985).

pre-Late Devensian and Late Devensian tills separated by radiocarbon dated organics. Owing to the clarity of the chronostratigraphical evidence at Dimlington it is used as the type site for the phase of maximum expansion of the British ice sheet during the Late Devensian, a period known as the 'Dimlington Stadial'. The climatostratigraphical episode represented by the site is named the 'Dimlington Chronozone' and covers the period 26 000–13 000 years BP (Rose, 1985). Dimlington and the nearby sites of Roos and Tunstall, the latter two containing radiocarbon dated organic materials witnessing the close of the Dimlington Stadial (Figure 5.30), provide excellent chronostratigraphical bracketing of Late Devensian glaciation. Owing to the ongoing erosion and fresh exposure of multiple tills by storm waves, this stretch of the east Yorkshire coast will continue to provide further new evidence for the reconstruction of the depositional settings and dynamics of the former southern margin of the British ice sheet in the North Sea.

AQUALATE MERE (SJ 780 205)

D. Huddart

Introduction

This site in Staffordshire, consists of an esker system formed during Late Devensian deglaciation. It is an important location because it demonstrates the close association between eskers and fan deposits and was used to develop an early model for esker sedimentation by the [British] Geological Survey (Dixon, 1922, 1926; Whitehead *et al.*, 1927, 1928). The esker, and related fan, kettleholes and ice-contact slopes, provides evidence for landform development associated with active ice-sheet marginal wastage and this region forms one of the most instructive locations in the country for demonstrating this type of esker model. There has been a recent revival of interest in the processes of esker sedimentation (e.g. Brennand, 1994; Huddart and Bennett, 1997; Thomas and Montague, 1997) and, in particular, the model whereby active ice retreat allows subglacial, ice-walled or supraglacial fluviatile sedimentation into proglacial lakes to produce beaded eskers or fans (e.g. Gorell and Shaw, 1991; Warren and Ashley, 1994; Thomas *et al.*, in press).

Description

The Aqualate Mere esker is located on the north side of Aqualate Mere and comprises a series of elongate hills, orientated east–west, in a linear strip some 1.6 km long and 300 m wide at its maximum (Figure 5.31). The esker is part of a regional series of deglaciation landforms that illustrate active ice recession to the north-west. The narrow, well-defined, ridge system starts midway between Forton and Aqualate Mere (Figure 5.31). At the eastern end of Round Hill it rises to a maximum of 15 m and then there is a low point in the ridge before a marked ice-contact face ('E' on Figure 5.32). Anc's Hill is slightly offset and is a much broader, bulkier deposit exposing sand and well-rounded Bunter quartzites and sandstone gravels on the surface. This is followed by an offset, low winding ridge about 10 m high that joins Oak Hill, where it steepens to about 35 m at position 'D' (Figure 5.32) and is a much broader feature. A similar morphology occurs on Rough Hill, and sands are exposed in an old pit on its northern side. Sheep Hill has a steep, ice-contact slope to the west and in the fields to the east flattens off in a series of gently dipping slopes that are fan-shaped. There are occasional, poorly drained depressions, as on the south side of Oak Hill, that represent kettle-holes.

Interpretation

The interpretation of the Aqualate Mere deposits has to be viewed in the context of the regional deglaciation in this part of Staffordshire (Whitehead *et al.*, 1927, 1928; Worsley, 1967b, 1970, 1975). The area is located between the Ellesmere–Whitchurch–Bar Hill moraine (Boulton and Worsley, 1965; Yates and Moseley, 1967) and the Devensian drift limit to the south

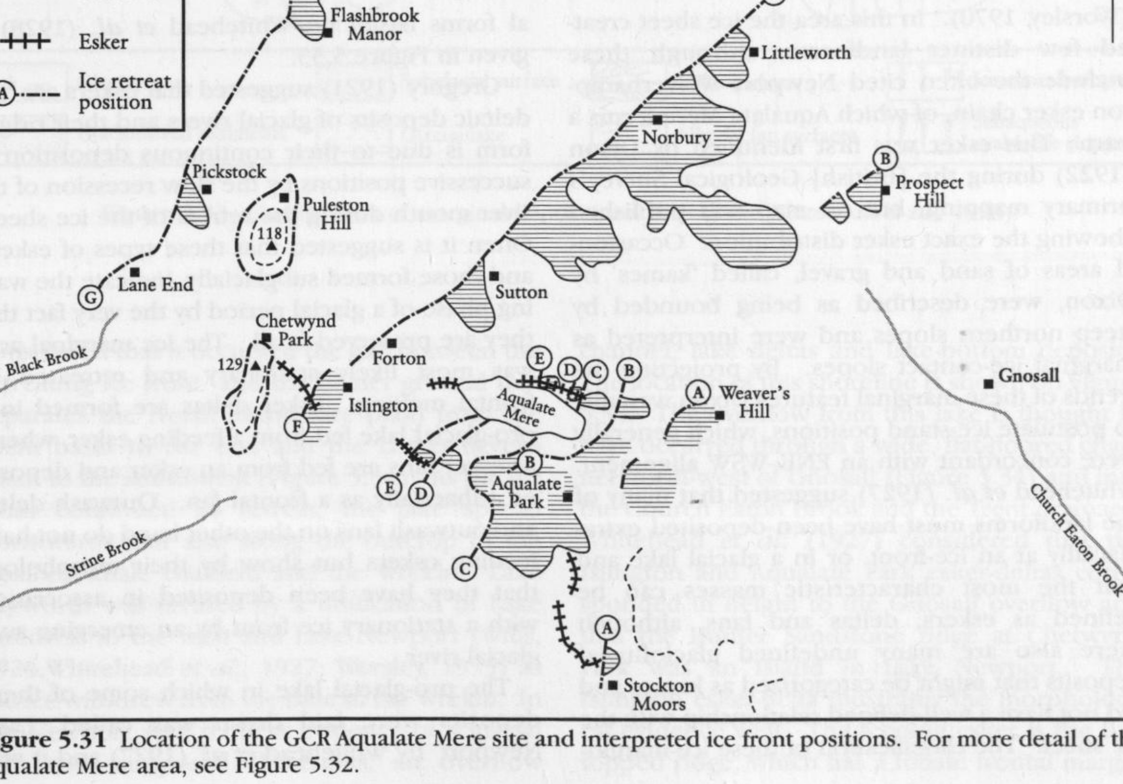


Figure 5.31 Location of the GCR Aqualate Mere site and interpreted ice front positions. For more detail of the Aqualate Mere area, see Figure 5.32.

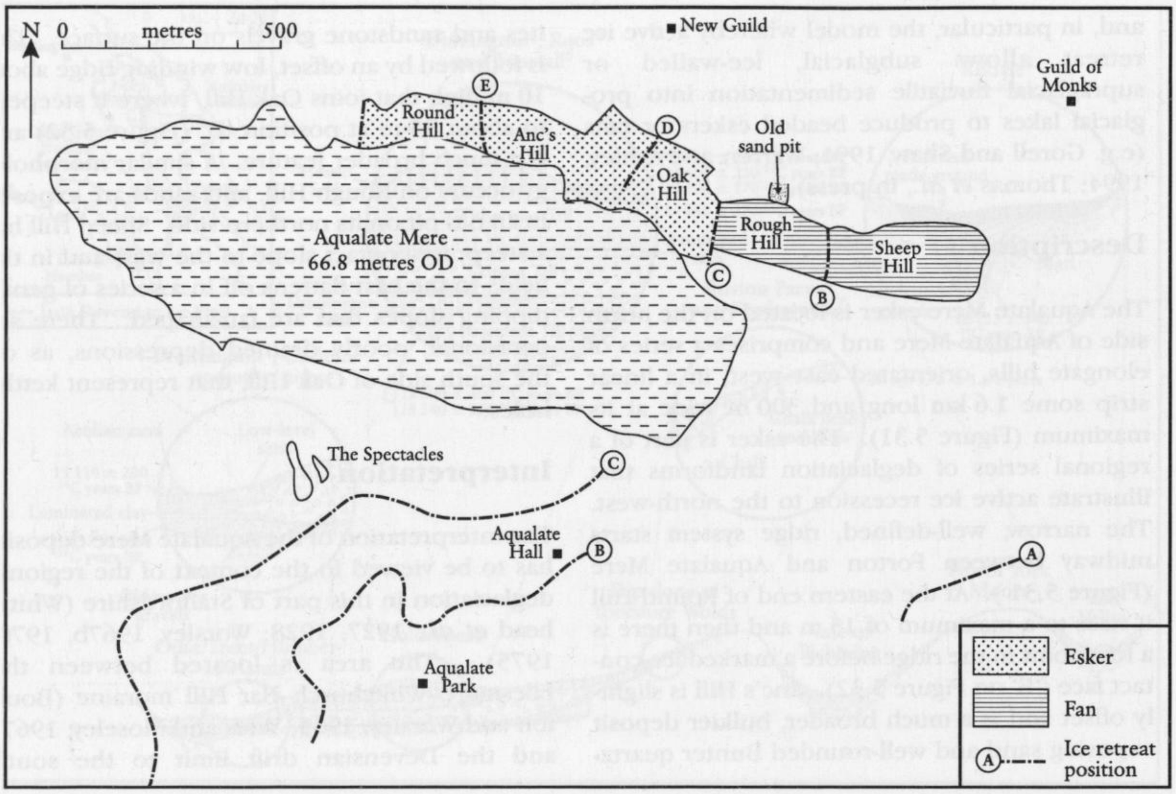


Figure 5.32 Regional distribution of glaciofluvial deposits and ice-front positions around the Aqualate Mere.

(Worsley, 1970). In this area the ice sheet created few distinct landforms, although these include the often cited Newport–Wolverhampton esker chain, of which Aqualate Mere forms a part. This esker was first identified by Dixon (1922) during the [British] Geological Survey's primary mapping, but no map was published showing the exact esker distribution. Occasional areas of sand and gravel, called 'kames' by Dixon, were described as being bounded by steep northern slopes and were interpreted as marginal ice-contact slopes. By projecting the trends of these marginal features Dixon was able to postulate ice-stand positions, which generally were concordant with an ENE–WSW alignment. Whitehead *et al*. (1927) suggested that many of the landforms must have been deposited extraglacially at an ice-front, or in a glacial lake, and that the most characteristic masses can be defined as eskers, deltas and fans, although there also are many undefined glaciofluvial deposits that might be categorized as kames and did not have a well-defined relationship with the ice sheet. The classification of these ice-marginal forms based on Whitehead *et al*. (1928) is given in Figure 5.33.

Gregory (1921) suggested that eskers are 'the deltaic deposits of glacial rivers and their ridged form is due to their continuous deposition at successive positions by the slow recession of the river mouth during the retreat of the ice sheet.' Often it is suggested that these types of eskers, and those formed subglacially, indicate the waning phase of a glacial period by the very fact that they are preserved at all. The ice marginal area was most likely stationary and retreated by frontal melting. Esker deltas are formed in a pro-glacial lake fed from a feeding esker, whereas esker fans are fed from an esker and deposited subaerially as a frontal fan. Outwash deltas and outwash fans on the other hand do not have feeding eskers but show by their morphology that they have been deposited in association with a stationary ice front by an emerging subglacial river.

The pro-glacial lake in which some of these deposits were laid down was called 'Lake Newport' by Whitehead *et al*. (1927) and it was

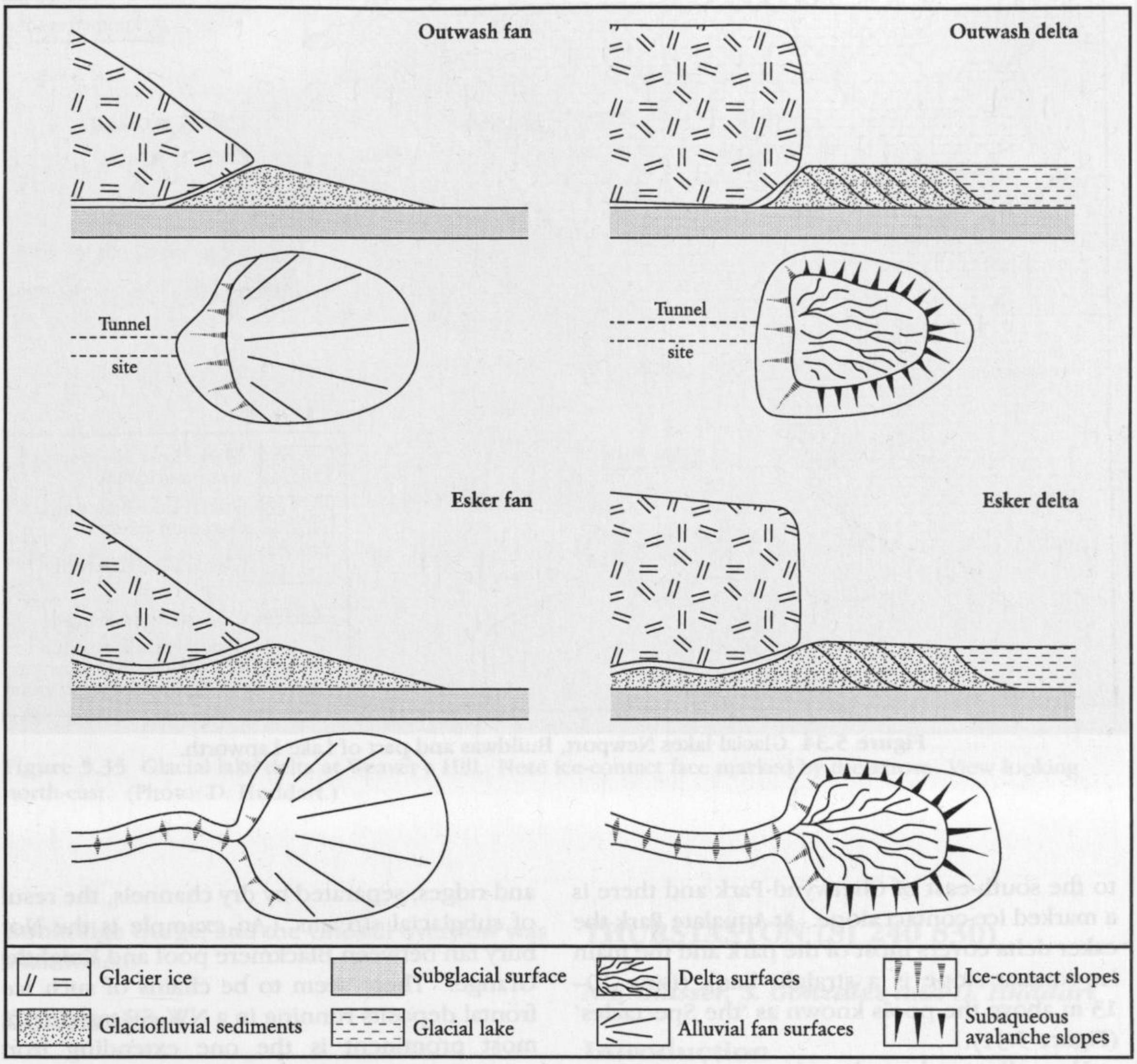


Figure 5.33 Classification of marginal depositional fans (after Whitehead *et al.*, 1928).

considered that it occupied the gap between the retreating ice front and the higher ground that separates the North Shropshire plain from the Trent basin to the east and the Lower Severn basin to the south-west (Figure 5.34). As the ice front continued its retreat, the lake spread northwards and also along the outcrop of the Coalbrookdale coalfield and the Wrekin. Lake Lapworth was created by a unification of Lake Buildwas to the west and Lake Newport (Wills, 1924;Whitehead *et al.*, 1927; Worsley, 1975) as the ice withdrew from the base of the Wrekin. In the Newport area there is evidence of a glacial lake with supposed shorelines, an overflow channel, lake deltas and lake-bottom deposits. The location of this shoreline is shown on Figure 5.34. The overflow from this lake is thought to have occurred through a wide, flat-floored channel north-west of Gnosall (Figure 5.34) and into the Church Eaton Brook and the Trent drainage. Whitehead *et al.* (1927) considered that the Islington and Aqualate Park esker-deltas corresponded in height to the Gnosall overflow and that the Bunter Sandstone ridge at Chetwynd Park was an island in Lake Newport. The Islington esker-delta illustrates the morphological evidence well. The feeding esker is the flat-topped ridge, which has a lobate frontal margin

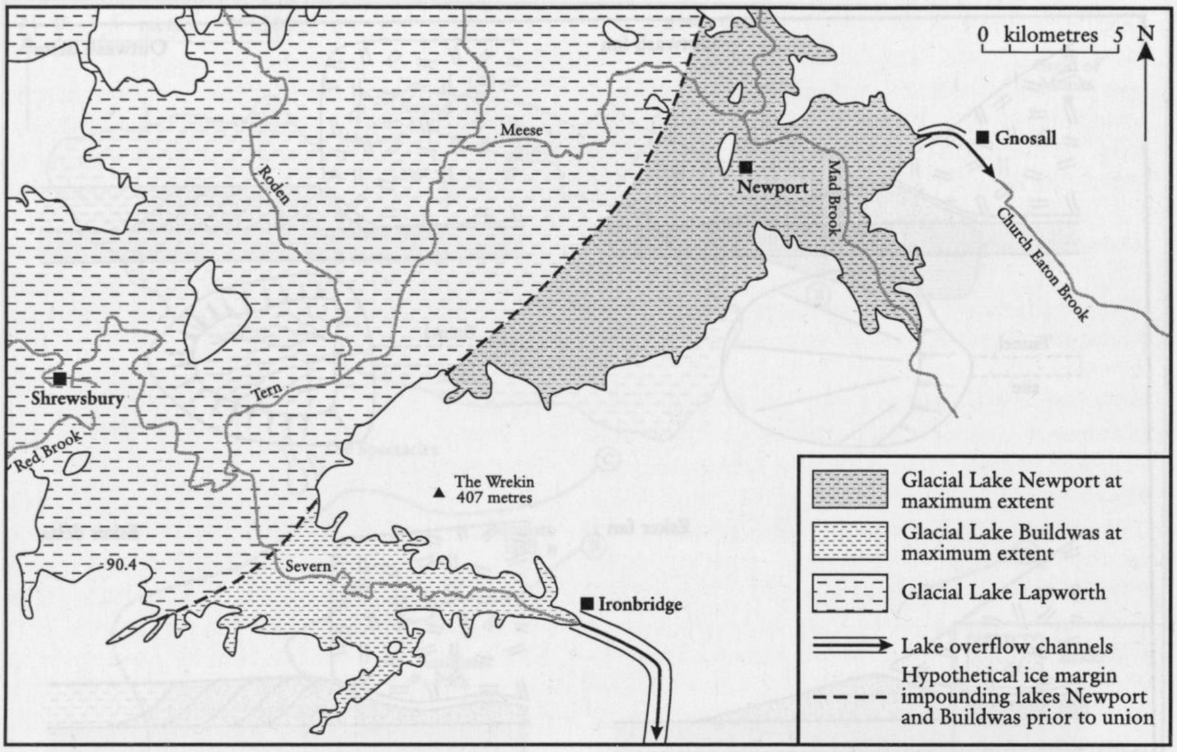


Figure 5.34 Glacial lakes Newport, Buildwas and part of Lake Lapworth.

to the south-east of Chetwynd Park and there is a marked ice-contact slope. At Aqualate Park the esker delta covers most of the park and the main ice-contact slope is a straight bank rising 10–13 m above the pools known as 'the Spectacles' (Figure 5.31).

No sedimentological evidence is available in the area to prove the environment of deposition, although at Weaver's Hill (Figure 5.35) an old sand pit exposed 15 m of rippled and parallel laminated sands, dipping at about 8° to the south-east and overlain by about 50 cm of rounded pebble gravel. The ice-contact face is to the north-west and low fans built out to the south-east. These possibly were subaqueous as no flat-top is obvious and they are similar to the fans at the south-east end of Sheep Hill. A series of ice-frontal positions marked by fans and deltas (between 'A' and 'G', Figure 5.32) indicates that the ice front was retreating to the north-west. However, there is a marked lack of continuity at any one of these frontal positions because the sediments were deposited along the subglacial river positions. Some of the fans have been eroded into a number of small mounds and ridges, separated by dry channels, the result of subglacial streams. An example is the Norbury fan between Blackmere pool and Knightley Grange. There seem to be chains of such ice-frontal deposits running in a NW–SE trend. The most prominent is the one extending from Stockton Moors–Aqualate Park–Islington to Lane End (Figure 5.32). Another chain is the one associated with this GCR site from Weaver's Hill to Forton, where the subglacial stream seems to have shifted laterally to the north-east and then develops fans at Sutton and Pickstock. A third chain runs from Prospect Hill to Norbury and Flashbrook Manor and a fourth from Littleworth to Knighton. Some of these appear to have been deposited by a groups of streams giving scattered mounds and ridges. Laminated lake-bottom deposits cover a large area drained by the Black Brook and Strine Brook and most are wholly or comparatively stoneless. It appears that Lake Newport was separated from Lake Buildwas by ice at least up to the period associated with ice front 'G' on Figure 5.32. With the creation of a Lake Lapworth (Wills, 1924; Worsley, 1975) the drainage flowed through the

Figure 5.35 Glacial lake delta at Weaver's Hill. Note ice-contact face marked by the arrow. View looking north-east. (Photo: D. Huddart.)

Ironbridge Gorge, and the Gnosall overflow was abandoned.

Conclusions

The Aqualate Mere landforms constitute an important part of the interpretation of ice-frontal deposition into pro-glacial lakes as the Irish Sea ice sheet actively retreated to the north-west. There is excellent morphological evidence for an esker chain, ice-contact slopes and probably subaqueous fans, rather than deltas, which built up pro-glacially from a subglacial river. Unfortunately sediment exposure is not available, but at nearby Weaver's Hill the sediments that are exposed are not inconsistent with this model of ice-frontal deposition into a pro-glacial lake. The regional picture is consistent with this local interpretation and it is considered that Dixon's original model for ice retreat and its association with pro-glacial lakes is likely to be correct. However, without detailed sedimentological evidence to add to the landform interpretation the model must remain unproven.

THURSTASTON (SJ 240 830)

N.F. Glasser, S. Gonzalez and D. Huddart

Introduction

The coastal section at Thurstaston on the Wirral Peninsula is the only permanent exposure in Late Devensian glaciogenic deposits for approximately 200 km along the north-west coastline of England between Cumbria and North Wales. The site therefore is important not only for the evidence it contains about glacial events in this part of the Irish Sea basin, but also for regional correlation between Cumbria, North Wales and the Cheshire–Shropshire lowlands. The section provides an opportunity to reconstruct depositional environments and to test competing glaciomarine and glacio-terrestrial theories. Thurstaston also constitutes an excellent example of the tripartite lithological sequence ('Upper Boulder Clay', 'Middle Sands' and 'Lower Boulder Clay') favoured by early workers of the [British] Geological Survey.

The glaciogenic deposits at Thurstaston were first described by Slater (1929) and other accounts include those by Brenchley (1968), Pitts (1983) and Jones (1990). More recently, the section has been considered by Glasser *et al.* (2001).

Description

The underlying bedrock on the Wirral is Triassic Sherwood Sandstone, which forms a series of ridges trending NNW–SSE, separated by intervening depressions that often are below sea level (Gresswell, 1964). Evidence from the area was used as early as the 19th century to demonstrate that during the Pleistocene Epoch a large ice sheet flowed out of the Irish Sea basin across north-west England in a NNW–SSE direction (Tiddeman, 1872; Mackintosh, 1879; Morton, 1860, 1870). As it did so, the ice sheet deposited a spread of muds, sands and gravels (De Rance, 1870; Shone, 1878; Strahan, 1886). This regional pattern of ice flow is now firmly established (Gresswell, 1964), as are the limits of the ice sheet to the south in the Cheshire–Shropshire lowlands (Boulton and Worsley, 1965; Thomas, G.S.P, 1985a, 1989). During deglaciation the ice sheet in the area was relatively thin and deglaciation was achieved under temperate conditions with abundant meltwater (Sambrook Smith and Glasser, 1998; Glasser and Sambrook Smith, 1999).

The GCR site is located around Thurstaston Steps, where coastal erosion has created a 20 m section through a low NW–SE orientated mound. The succession comprises six main lithofacies (Glasser *et al.*, 2001). These are diamicton, gravel, sand, mud, laminite and cobble pavements (Figure 5.36). In general terms the northern part of the section is dominated by sand, gravel and minor mud, whereas the diamicton dominates the southern part. In the middle of the section, near Shore Cottages, mass movements obscure the cliff section.

Diamicton

Two diamicton lithofacies are identified: a lower (clast-rich) and an upper (clast-poor) diamicton. The lower diamicton comprises a red-brown, compacted, massive, clast-rich, sandy diamicton

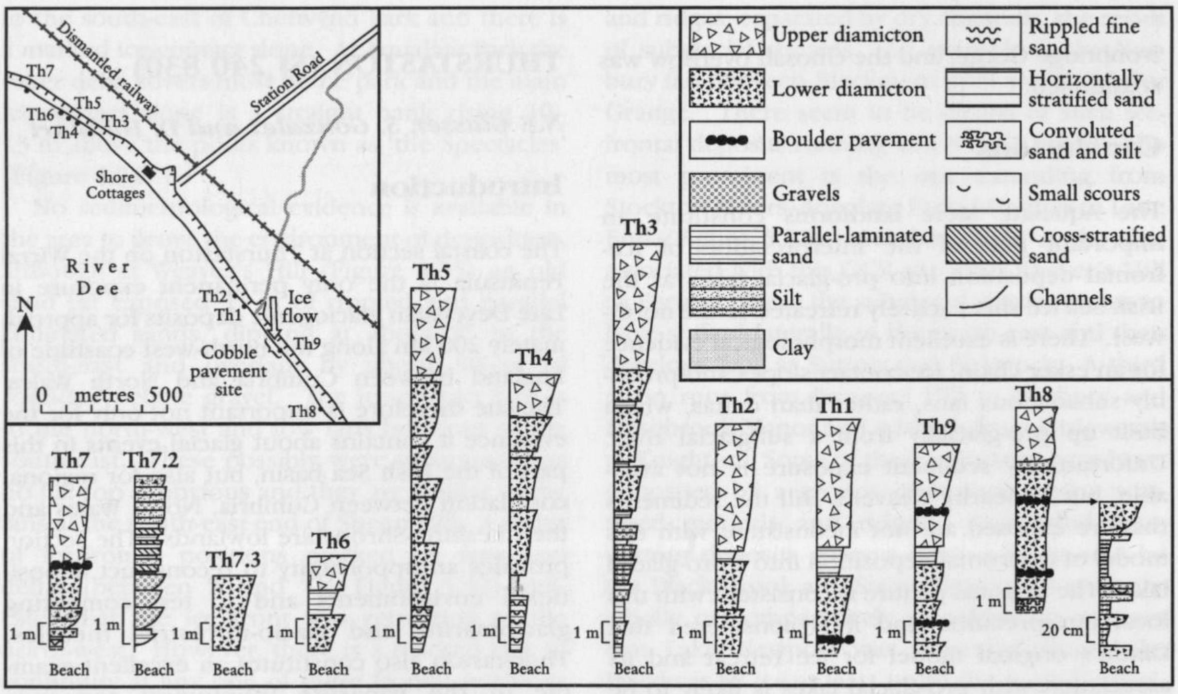


Figure 5.36 Stratigraphical logs of nine sections (TH1–TH9) at Thurstaston showing the major lithofacies identified at the site. Inset shows location of logs.

with a homogeneous texture. In places it shows a sub-horizontal structure defined by planar discontinuities (fissility). Locally derived Triassic sandstones and siltstones clasts dominate, but exotic lithologies include mudstone, limestone, quartzite, gypsum, gabbro, ignimbrite, tuff, basalt, granite and scattered shelly debris. Shear planes are present and there is evidence of poorly developed folding in the diamicton (see 'Deformation structures' section below).

The upper diamicton is a red-brown to light brown, clast-poor to very locally clast-rich, muddy diamicton. It is characterized by a relatively structureless, homogeneous texture and a uniform sediment composition. Some regular pebble-sized clast concentrations and shell fragments occur in discontinuous layers, especially near the base of the diamicton (TH2, Figure 5.36). Clast shapes range from angular to well rounded. Striated and faceted clasts are common, and the diamicton is dominated by Triassic lithologies with rare mudstone, limestone, basalt and granite clasts. There are occasional sand stringers and pods up to 3 cm thick and 30 cm across and generally a vertical joint pattern.

Sand

This lithofacies comprises well-sorted, light red-brown to yellow stratified sand consisting of layers of fine- to coarse-grained sand that generally are texturally homogeneous. The sand is composed mainly of subangular to rounded grains, with subrounded shapes dominating. In places, the sand contains isolated pods of gravel and occasional clay layers and lenses (e.g. TH4, Figure 5.36). Sedimentary structures are variable but include fining-upward sequences, parallel- and cross-laminated units, tabular cross-beds and erosional surfaces. In places, the sand fills erosional channels. Measurement of foresets in the cross-bedded layers indicates variable flow directions towards the south, south-west and north.

Mud

Mud is a minor lithofacies associated with sand horizons. Ranging from a grey clay to clay–silt–sand admixtures, it occurs as discontinuous beds a few centimetres thick within the sand. Contacts are commonly loaded or show convolute relationships. Occasional mud balls, several centimetres in diameter, are found within the sand bodies.

Laminites

This lithofacies comprises a sequence of finely laminated (millimetre scale) clay and silt, in places immediately overlying the lower diamicton. There is no obvious grading in the laminae. The average thickness of this laminated unit is 10–15 cm, although it reaches a maximum of 20 cm near site TH1.

Cobble pavements

A well-developed cobble pavement is sometimes exposed within the clast-rich (lower) diamicton towards the southern end of the section (TH9, Figure 5.36). Where cobbles are absent, there is a clear planar discontinuity in the diamicton. The pavement has a maximum thickness of 20 cm and clasts show a strong NNW–SSE preferred orientation, with planed and smoothed upper surfaces and striations. The cobbles that comprise the pavement are a variety of lithologies but are mainly igneous and Lower Palaeozoic rock-types.

Deformation structures

The clast-rich diamicton is characterized by two groups of deformation structures. First is a group of surfaces that resemble listric faults. These have cross-sectional traces around 10 m in length and are spaced over tens of metres. They flatten into a bedding-parallel orientation in the lower horizons. Nowhere are they seen to cut below the cobble pavement or penetrate upwards into the overlying units. Their curviplanar form appears concave towards the north-west. It is difficult to define the sense of displacement but vertical separation of beds in the fault is up to 4.5 m. The fault zone itself, *c.* 5 cm in thickness, is relatively indurated, probably arising from the preferential introduction of a carbonate cement within this localized zone. Occasional, approximately bedding-parallel, slickensided and slickenlined surfaces also indicate localized shearing.

The second group of structures in the diamicton is a set of roughly orthogonal joints, crossed obliquely by the listric faults. These are distributed sporadically, but are best developed towards the base of the unit. They are typically closely spaced (10–15 cm apart) in an orthogonal set, approximately at right angles to the stratification. These joints are mainly near-vertical, although they are inclined in places.

Two contrasting suites of small-scale deformation structures occur in the sand. First is a range of asymmetric load and flame structures, convolutions and attenuated isoclinal folds. These are developed on a centimetre scale and are most prominent in the laminated sands. The second type, developed in close proximity, are numerous, very sharp faults, again developed on the centimetre scale. These are a series of curvilinear, normal faults with cross-sectional traces tens of centimetres in length and, more commonly, straight, steep to vertical, very narrow faults with traces typically on the scale of centimetres.

Interpretation

Slater (1929) recognized a tripartite sequence consisting of 'Upper Boulder Clay' and 'Lower Boulder Clay' separated by sands and gravels ('Middle Sands'). Slater's tripartite sequence mirrored that reported from other locations on the eastern fringes of the Irish Sea Basin (e.g. Jehu, 1909) and is now known as the 'Stockport Formation' (Bowen, 1999). Jones (1912) studied the erratic content of the glacial deposits of the Wirral and identified a variety of clast lithologies, the majority of which are derived from the Lake District and the Southern Uplands of Scotland. This confirms that the ice in Wirral flowed out of the Irish Sea basin in a NNW–SSE direction (Morton, 1860, 1870; Mackintosh, 1879). In a more recent study of the sediments at Thurstaston, Glasser *et al.* (2001) have interpreted the section in terms of terrestrial glacial deposition beneath an Irish Sea ice sheet. They attached the following interpretations to the six lithofacies based on the sedimentary properties described above.

Diamicton

Massive diamictons are known to form in the glaciogenic environment by: (i) terrestrial lodgement and meltout of basal glacial debris (Andrews and Shimuzi, 1966; Boulton, 1970; Lindsay, 1970; Mark, 1974; Lawson, 1979a; Shaw, 1982; Haldorsen and Shaw, 1982; Dreimanis, 1989; Hart, 1994; Benn, 1995); (ii) deposition from ice at, or seawards of, a grounding line (Powell and Molnia, 1989; Powell *et al.*, 1996); (iii) from a high density of icebergs (Domack and Lawson, 1985; Dowdeswell *et al.*, 1994); (iv) from the formation of deformation tills produced by the deformation and *en masse* transport of subsole sediments as a slurry layer beneath the ice base (Boulton and Hindmarsh, 1987; Hicock and Dreimanis, 1992; Eyles *et al.*, 1994, Hart and Roberts, 1994; Evans *et al.*, 1995; Hart, 1995, 1998; Benn and Evans, 1996; Boulton, 1996; Alley *et al.*, 1997); and (v) from sediment gravity flows in both subaerial and subaquatic contexts (Boulton, 1968, 1971, 1976; Marcussen, 1973, 1975; Lawson, 1979b; 1982; van der Meer, 1987, 1993).

The clast-rich (lower) diamicton is interpreted as a basal deformation till deposited beneath temperate, terrestrially based ice. Evidence for this interpretation includes the inclusions of unlithified sediments into the diamicton, the structures indicating deformation and the weak clast macrofabric, which are similar to those described from deformation tills (Hart and Roberts, 1994; Benn and Evans, 1998). Clast-surface features, including striations and facets, and clast shape are indicative of basal transport beneath temperate, sliding glacier ice (Boulton, 1978; Bennett *et al.*, 1997). External boundary relationships, such as the presence of the cobble pavements within the diamictons are also strong evidence that it is a deformation till (Clark, 1991; Hicock, 1991; Boulton, 1996). The sub-horizontal fissility noted at Thurstaston is formed by shear along numerous discontinuous slip planes and their position has been noted at the base of the deforming layer (Boulton and Hindmarsh, 1987; Benn, 1994).

The origin of the predominately clast-poor (upper) diamicton is less clear. Possible origins include a readvance till, a glaciogenic sediment gravity flow, or a deformation till. The similarity in the lithological assemblage between the upper and lower diamictons, the lack of unequivocally pro-glacial sand and gravel between the two diamicton units and the continuity in the sediment deposition sequence suggests that it is not the product of a readvance. A second possible origin is a series of glaciogenic sediment gravity flows deposited from debris-bearing glacier ice. Evidence for this interpretation is the weakly consolidated nature of the matrix and the variable textural composition.

The nature of the clasts in this lithofacies, including the proportion of striated and faceted clasts and their degree of rounding, indicates that if this facies was formed by sediment flows then the sediment was sourced from material originally transported in the basal layers of the ice sheet. However, the thickness of this deposit (up to 6 m thick), its homogeneous nature and the absence of stacked sequences that can be differentiated on the basis of texture, basal clast concentrations, washed tops and interbedding between the diamictons and glaciofluvial sediment of exposures argue against this origin. The upper diamicton also could be interpreted as a deformation till, which built up incrementally through time. Evidence for this interpretation includes the weak clast macrofabric, which is similar to those described from other deformation tills (Hart and Roberts, 1994; Benn and Evans, 1998). Clast surface features, including striations and facets, and clast shape are also indicative of basal transport beneath temperate, sliding glacier ice (Boulton, 1978; Bennett *et al.*, 1997). External boundary relationships, such as the presence of the cobble pavements within the diamicton, are also strong evidence that it may represent a deformation till (Clark, 1991; Hicock, 1991; Boulton, 1996).

Gravel

The thickest gravel-dominated lithofacies at the base of the sequence at the north-western end of the succession probably indicates a fluvial regime characterized by major, short-lived floods (Williams and Rust, 1969; Rust, 1972, 1975; Bluck, 1974, 1979; Boothroyd and Ashley, 1975; Church and Gilbert, 1975; Hein and Walker, 1977; Macklin and Hunt, 1988; Aitken, 1998). The simplest interpretation of this lithofacies is the product of deposition in a pro-glacial braided river. The presence of Triassic sandstone and siltstone clasts in similar proportions to the two diamicton lithofacies suggests that this river may have been reworking and transporting sediment originally released directly from glacier ice. The majority of clasts fall within the rounded class, which is consistent with this interpretation. However, it is also possible the gravel lithofacies was deposited in a subglacial channel. Many of the thinner gravel and sand sequences can be demonstrated to lie between diamicton units and therefore are likely to have formed in this way.

Sand

The sand lithofacies occurs in close association with the gravel lithofacies, and the two lithofacies commonly are interbedded. These lithofacies associations are typical of a glaciofluvial environment with fluctuating discharge levels and sediment supply (Aitken, 1998). The alternation of sand with the gravel lithofacies would indicate frequent channel migration during ice recession. Sedimentary structures such as load and flame structures indicate that the sediment may have been deposited rapidly, inducing deformation and remobilization of water-saturated layers

Mud

This minor facies represents the waning flow stage of fluvial deposition or the effects of channel migration. Deformation structures in the mud are interpreted as the effect of loading of the sediment by later influxes of sand. Most mud horizons formed in ponded water.

Laminite

The clay–silt and silt–fine-sand laminite is interpreted as the product of material settling from suspension into a series of subglacial ponds on the irregular topography of the deformation till. The lateral impersistence of the laminated unit and its limited thickness indicates that these water bodies were small and that their existence was relatively short-lived.

Cobble pavements

Boulder pavements at the base of massive diamictons have previously been interpreted as evidence for subglacial deformation in a terrestrial environment (Clark, 1991; Hicock, 1991; Boulton, 1996). Similar pavements have been reported to form during stick-slip motion in a stable–deforming bed mosaic by Piotrowski and Kraus (1997) and by Fischer and Clarke (1997). Pavements also may form by current winnowing in a glaciomarine setting, with subsequent abrasion of upper surfaces by partly grounded basal ice (Eyles, 1988), a floating ice margin (Eyles and Lagoe, 1990) or floating icebergs (McCabe and Haynes, 1996). A glaciomarine origin for the cobble pavements at Thurstaston appears unlikely because of the setting of the pavement within a diamicton sequence that displays many of the properties of direct deposition from a subglacial deforming layer. The Thurstaston cobble pavements are probably products of terrestrial subglacial deformation, formed as large clasts in a deforming layer sink through the till to a level where the matrix is able to support their weight (Clark, 1991).

Deformation structures

The listric faults in the clast-rich (lower) diamicton may represent thrust faults, developed in response to the movement of ice across the area (Boyce and Eyles, 2000). The induration of the zones presumably records enhanced fluid flow along them (Maltman, 1994). The orthogonal joints with a polygonal upper surface are most suggestive of a desiccation origin, formed at a time when the diamicton was exposed subaerially. The deformation may represent the thrusting and shearing of sub-sole sediments as a result of ploughing and frictional drag at the ice–bed interface.

The ductile deformation structures in the sand lithofacies probably formed when the sediment was temporarily reduced in strength, perhaps as a result of sediment or ice overburden. Accompanying temporary variations in the density and viscosity of the different layers would prompt Raleigh–Taylor instabilities and hence the formation of the flames and lobes. During their transient loss of strength, the sands would be sensitive to even a slight slope, which, together with shear imposed by the ice or sediment, helps explain the asymmetry of many of the structures.

Conclusions

The section at Thurstaston shows evidence for sedimentation in a range of different glacial environments. The dominant lithofacies are interpreted as deformation tills and glaciofluvial and subglacial interbeds. The laminites suggest the presence of ephemeral water bodies, possibly silting ponds in a subglacial position. The cobble pavements are interpreted as an integral product of deformation till formation. Collectively, these sediments appear to record deposition by a terrestrially based Irish Sea glacier and there is little evidence for glaciomarine sedimentation at Thurstaston.

SANDY BAY (NZ 305 860)

D. Huddart

Introduction

The Devensian glacial landforms along the Northumberland coastal plain are predominantly of subglacial origin (Eyles and Sladen, 1981; Eyles *et al.*, 1982; Douglas, 1991) and there is a low-relief, extensive, drumlinized till plain extending inland from the North Sea coast at heights below 100 m OD. This plain is defined to the west of Morpeth by emerging streamlined rock highs and the rivers Tyne and Aln to the south and north. This drumlinized topography comprises the Acklington Formation (Thomas, 1999). It is composed of glaciotectonic and rafted bedrock overlain by cross-cutting lodgement till units, within and between which occur subglacially channelled gravels, sands and silts deposited in a single phase of subglacial deposition. It has been modified extensively by periglacial slope processes (Anson and Sharp, 1960) and overlies a dissected rockhead surface. Carruthers (1947, 1953) considered that the glacial sediments in this area were the result of the undermelt of a single, stratified ice sheet. Later workers recognized a tripartite stratigraphy composed of a lower, blue grey till (the Lower Boulder Clay) and an upper brown till (the Upper Boulder Clay), separated in many areas by a middle unit of gravel, sand and/or laminated clay. The entire sequence was thought to be produced by two separate glacial phases (Raistrick, 1931b; Taylor *et al.*, 1971) but with no clear evidence of either interglacial or interstadial status for the intervening glaciofluvial sediments (Woolacott, 1921).

Sandy Bay (NZ 305 860) is one of the longest and best-exposed till sections on the Northumberland coast and is situated between Spital Burn and the River Wansbeck, south of Newbiggin-by-the-Sea. It comprises a 1.8 km coastal strip of foreshore and cliff section and forms part of the Cresswell and Newbiggin Shores SSSI. The site has been used to suggest that the till originated by the lodgement process, whereby deposition occurred by continued accretionary lodging and local, pressure melt-out of dispersed englacial debris particles, or aggregated debris masses, which were being transported along flow lines at the ice–bed interface (Eyles *et al.*, 1982). The model for deposition also has been simplified to suggest deposition from one glacial phase, with the upper brown or reddened tills a result of post-glacial weathering (Eyles and Sladen, 1981; Douglas, 1991).

Description

The glacial till is best exposed between the middle and southern part of the site and the stratigraphy described by Bullerwell (1910) included a basal, coarse gravel, up to 6 m thick, which thinned to the south (Figures 5.37 and 5.38). The grain size of this deposit varied between pebble and boulder gravel with patches of finer gravel and coarse sand. It contained Cheviot rocks, granite, Carboniferous Limestone, gneissose and schistose rocks and flints, with one shell fragment noted. It was thought to be a fragment of a pre-till beach, eroded away since the original description. It has been suggested as a possible Ipswichian beach (Lunn, 1995). The overlying till is up to 4 m thick and rests in part on rockhead. This rockhead consists of either striated surfaces of Carboniferous sandstone, or of fragmentary and glaciotectonized masses of sandstone, shale, coal or mudstone that have been attenuated by glacial shear as a deformed local facies of the till. Sandstone rafts have been transported to the south over tectonized shale and mudstones that underlie the till. The clay fraction of the till in north-east England is derived almost entirely from such sources (Beaumont, 1972; Tarbet, 1973). Local sandstones predominate in the till units immediately above rockhead, with sheared out stringers and smudges of shale and coal. Farther travelled erratics from the Southern Uplands and Cheviot Hills are found higher up the sections. However, there are no major lithological differences between individual till units at this location, except that the carbonate lithologies are under-represented in the upper part of the sections as a result of post-glacial weathering (Eyles and Sladen, 1981; Eyles *et al.*, 1982). The general characteristics of the till in coastal Northumberland are summarized in Figure 5.39.

Major sub-horizontal, undulatory bedding planes divide the till stratigraphy into several distinct lensoid till units. These bedding planes are erosional disconformities and affect both the till units and the channelled glaciofluvial sediments. Erosional interfaces define multiple till units

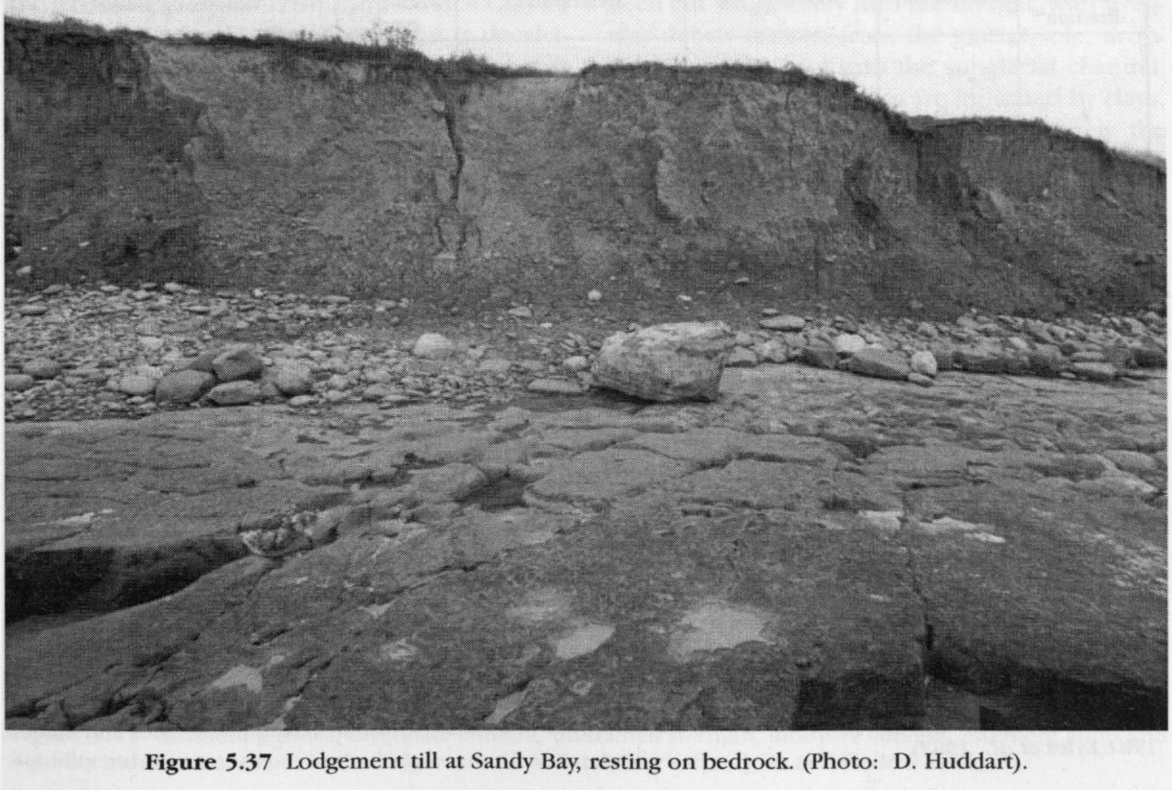


Figure 5.37 Lodgement till at Sandy Bay, resting on bedrock. (Photo: D. Huddart).

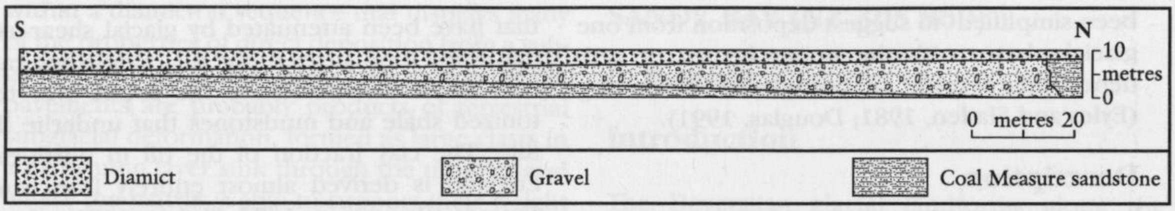


Figure 5.38 Stratigraphy of the cliff near Newbiggin (north end of Sandy Bay). After Bullerwell (1910).

over much of the exposed section, but these are not laterally continuous and fade out within a single thick till. The till units show a matrix colour that is uniform grey at depth but which becomes reddish-brown in the upper part of the section. Striated limestone clasts are common in the lower part of this grey till. Part of the exposed section is illustrated in Figure 5.40, but the section continues to the south for 500 m as a single massive till. Towards the southern end of the section there is sometimes a purple till between the grey and the reddish brown till, with a merging boundary between the units. Glaciofluvial lenses in the till sequences frequently define the base of the reddened upper till from the lower grey till below. There has been much debate as to the glacial, interglacial or interstadial status of the interbeds (Woolacott, 1921; Raistrick, 1931b; Taylor *et al.*, 1971; Land, 1974; Lunn, 1980). These glaciofluvial sediments are in two forms:

1. shoestring channel fills orientated sub-parallel to the regional direction of ice movement, where the channel bases are choked with dirty, matrix-supported gravels, lag boulders and commonly armoured till balls overlain invariably by massive, plane-bedded or ripple cross-laminated sand units or massive, irregularly laminated clays, with frequent clusters or

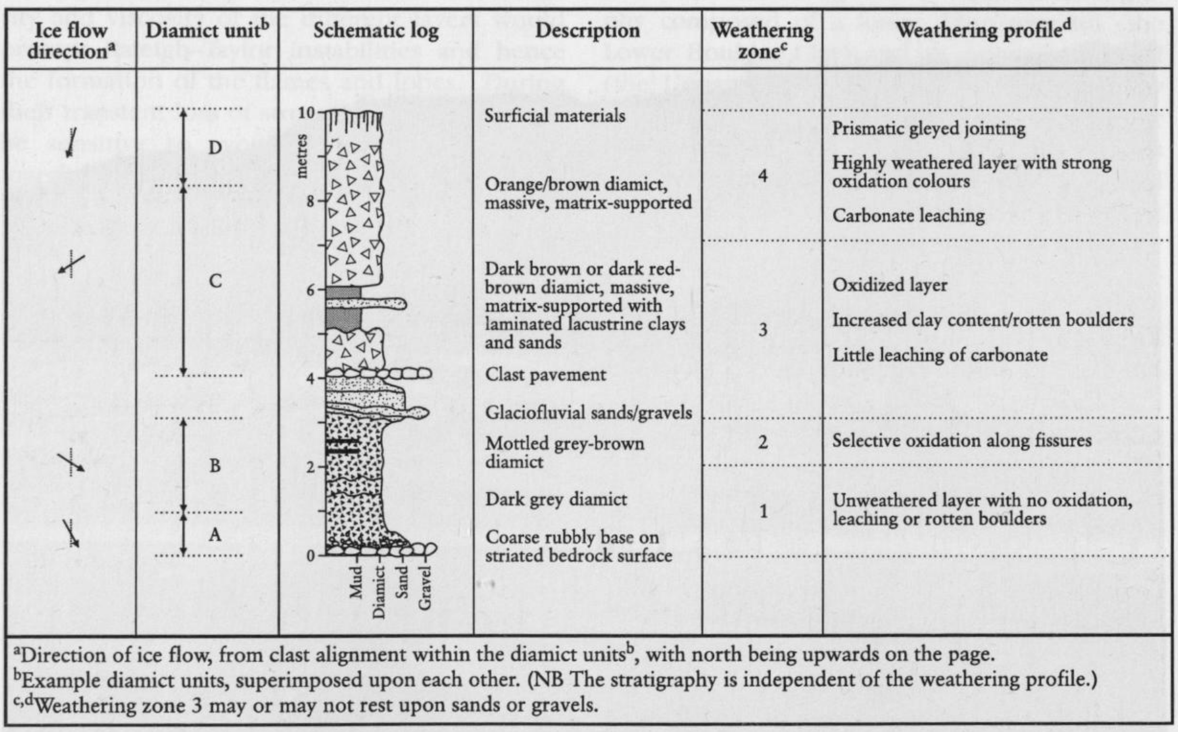


Figure 5.39 Generalized characteristics of lodgement till in coastal Northumberland (after Eyles and Sladen, 1981;Eyles *et al.*, 1982)

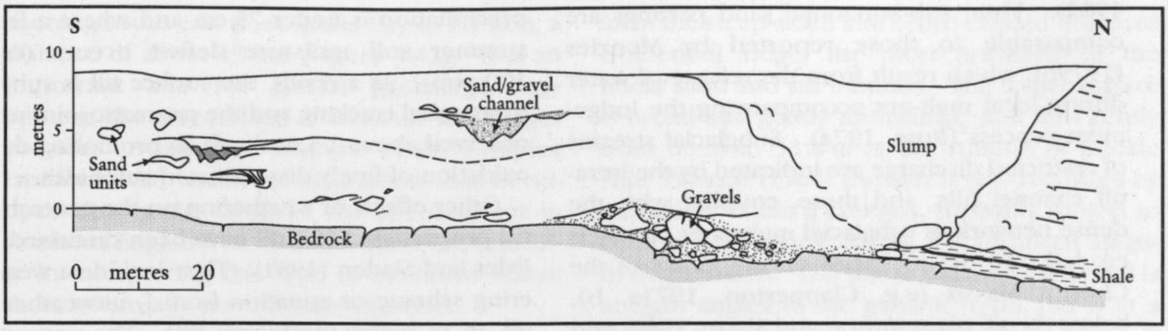


Figure 5.40 Lodgement till stratigraphy at Sandy Bay (after Eyles *et al.*, 1982).

clots of till-like debris embedded throughout;
2. units of variable geometry and restricted size from thin, under 5-cm-thick, sub-horizontal sand partings to contorted, highly deformed and attenuated masses of dirty gravels and sands, which lack any lateral connectivity.

The tills at Sandy Bay show prominent sub-horizontal bedding plane shears and well-developed vertical joint sets. The sandstone and Whin Sill dolerite boulders within the red-brown till show a weathering rind and commonly are completely rotten. Gleyed, prismatic desiccation joints disrupt the surface of the upper red till to a depth of 1.5 m and this typically forms the Prismatic Clay or Prismatic Boulder Clay of north-east England. This upper unit is decalcified, for example the $CaCO_3$ contents for this location are given in Figure 5.41, where it can be seen that below 2 m the level of $CaCO_3$ is constant. This is thought to be the primary carbonate content of the till. Above 2 m this content decreases to close to zero near ground level and clearly indicates leaching of the primary carbonate.

Interpretation

The basal gravels are likely to be either a last interglacial beach (Bullerwell, 1910; Lunn, 1995) or Devensian, glaciofluvial, probably proglacial or conceivably subglacial gravels. The till sequence at Sandy Bay is thought to result from a single complex episode of wet-based, subglacial sedimentation (Eyles *et al.*, 1982) as accretionary lodgement of debris took place. The shoestring channels also appear to have also been cut subglacially into the till bed, with irregular debris masses, from the glacier sole, dropping from the roof into the subglacial channel. Variable palaeodischarges are indicated by clays, which indicate ponding episodes, possibly the closure of the channel downstream by ice flow, or by deformation of the till bed (Eyles *et al.*,

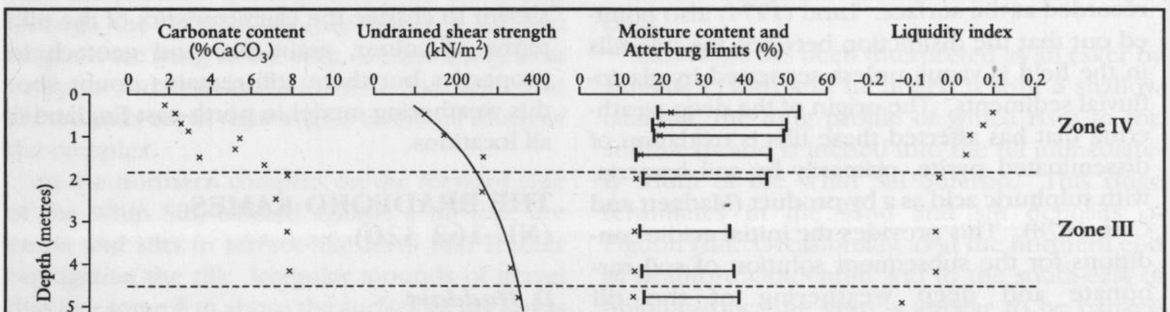


Figure 5.41 Weathering zones, carbonate content, undrained strength, moisture content, Atterberg limits and liquidity index versus depth in the lodgement till at Sandy Bay (after Eyles and Sladen, 1981).

1982). Thin, sub-horizontal sand partings are comparable to those reported by Menzies (1979b), which result from the release of water during local melt-out accompanying the lodgement process (Rose, 1974). Subglacial streams of restricted discharge are indicated by the intra-till channel fills, and these, coupled with the dense network of subglacial meltwater channels eroded into the rockhead on the flanks of the Cheviot Massif (e.g. Clapperton, 1971a, b), below the till plain (Anson and Sharp, 1960) and on the floor of the adjacent North Sea (Dingle, 1971) are evidence that these channel fills are part of an integrated and widespread subglacial network draining a wet-based ice sheet.

The reddened colour of the uppermost till unit and the presence of glaciofluvial interbeds has prompted the use of a tripartite glacial stratigraphy for Northumberland comprising lower grey tills, middle sands, gravels and clays and upper reddened tills (Mitchell *et al.*, 1973). However, Eyles and Sladen (1981) have demonstrated that along the coast the reddened colour of the uppermost till units is the result of post-glacial weathering, the depth and definition of which is directly related to the presence of glaciofluvial interbeds. The latter act to underdrain the overlying till units, which become fractured by desiccation, and the high bulk permeability thereby generated accelerates weathering processes. This is seen as a reddened, oxidized till, separated from the grey, unoxidized till below by the glaciofluvial interbeds. Along the Northumberland coast at those sites where till depths are under 4 m (e.g. Church Point, Newbiggin, to the north of Sandy Bay) a single reddened and clay-rich 'upper' till has been mapped in the past (e.g. Carruthers *et al.*, 1927). Significantly 'lower' grey tills have never been recorded at the surface. Land (1974) also pointed out that the distinction between the two tills in the field is vague unless separated by glaciofluvial sediments. The origin of the deep weathering that has affected these tills is oxidation of disseminated pyrite, primarily by soil bacteria, with sulphuric acid as a by-product (Madgett and Catt, 1978). This provides the initial acidic conditions for the subsequent solution of soil carbonate and deep weathering of the till sequences, which in Northumberland is the result partly of glacial quarrying of the sulphide-containing, Upper Carboniferous lithologies and partly it is climatically controlled because deep weathering has been reported where the annual precipitation is under 75 cm and where a large summer soil moisture deficit occurs (over 100 mm). As a result, the surface till is subject to seasonal cracking and the prismatic jointing is observed above 1.5 m depth so promoting deep oxidation of finely disseminated sulphides.

Other effects of weathering on the geotechnical properties of the till have been discussed by Eyles and Sladen (1981). They include a weathering scheme or zonation from I, unweathered till, to IV, highly weathered tills. The unweathered tills are separated from those in zone III–IV on particle size distribution, as disintegration of larger particles takes place and there is an increase in the clay and silt content; there is an increase in the natural moisture contents, increased plasticity, increased drained brittleness and a reduction in the undisturbed, undrained shear strength (Figure 5.41), although at a given moisture content the shear strength is higher for weathered till. There are also good examples of weathered Carboniferous Sandstone bedrock occurring in the cliffs to the north of the till section in Sandy Bay. However, doubts remain about this weathering model because of recent work in County Durham at some open cast sites, such as Herrington, where there is brown till, both above and below grey till (Hughes *et al*, 1998; Hughes and Teasdale, 1999; Teasdale and Hughes, 1999).

Conclusions

This site at Sandy Bay is an important location where it has been demonstrated that the till sequence is formed in one complex phase of lodgement from a wet-based ice sheet. There seems no need to invoke more than one glacial phase. Weathering of the surface till has been shown to change the characteristics of the till in terms of colour, grain size and geotechnical properties but there still remains doubt about this weathering model in north-east England for all locations.

THE BRADFORD KAMES (NU 162 320)

D. Huddart

Introduction

The Bradford kames form the northern termination of a marked linear sand and gravel zone running south-east from Budle Bay to Preston, a distance of about 13 km (Figure 5.42). It is an important site for glacial geomorphology because the assemblage of mounds and ridges provides an excellent example of the complex landform and sediment association that developed during the decay of the last ice-sheet in northern England. It also illustrates the difficulty in interpreting this type of landform where there is virtually no sediment exposure.

Description

The kames were first discussed by Gunn (1900), Smythe (1912) and Gregory (1922) and described more fully by Dinham in Carruthers *et al.* (1927) and Carruthers *et al.* (1930). Much later, Parsons (1966) placed the Bradford Kame Complex in a wider consideration of the deglaciation of this part of Northumberland. The Bradford Kame Complex is composed of a series of elongate mounds of sand and silt associated with locally sinuous, esker-like ridges composed of silt, sand and gravel (Figure 5.43). The latter have steep sides and undulating crests and their origin must be associated with that of the silt and sand mounds with which they are physically linked, as the ridges commonly terminate abruptly against the sand and silt mounds. This suggested to Parsons (1966) that they were the casts of feeder meltwater streams, which formed as subglacial or englacial eskers and that the linear silt and sand deposits formed in crevasses. Accumulation within an open crevasse was suggested originally by Dinham (in Carruthers *et al.*, 1927) to account for the formation of the esker ridge between Spindlestone (see Figure 5.42) and Hoppen and was thought to coincide with the junction between active ice advancing through the Waren Gap (see Figure 5.42) and stagnant ice lying to the east. Occasionally, as at Hoppen and Burton Goldenhill, the sand and silt mounds are at right angles to that of much of the complex.

In the northern complex on the leeward side of the Whin Sill, banked against a till rise, are sands and silts in terrace-like form that feather out against the till. Irregular mounds of gravel that rise some 6 m above the surface of the sands and silts were interpreted as moulin kames located along a zone of weakness in the stagnant ice (Parsons, 1966). Near Lidderton Hill, the sands and silts become of increasing topographical importance and to the south the deposits form the steep-sided and broad-crested Bradford Goldenhill ridge, the most northerly of the linear sand and silt mounds. The upper surface is broad and gently undulating, and falls gently from 60 m to 53.8 m at the summit of Cockle Hill. Parsons (1966) thought it was deposited by meltwater within a crevasse, probably subject to continual widening during the advanced stages of deglaciation. The linear Burton Goldenhill is at right angles to the general trend and is at a similar altitude to Bradford Goldenhill. It appears to be an ice-frontal deposit, although Parsons (1966) suggests it was formed as a crevasse filling.

To the west of Lidderton Hill, and associated with these sand and silt deposits, is the pronounced, sinuous ridge (30–50 m wide and 6–15 m high) of 'earthy' gravel, composed of local lithologies, including sandstones, shales, and Oxford and Chesterhill Dean limestones, although quartz dolerite is rare (Carruthers *et al.*, 1927). The ridge is distinguished from the surrounding land by rough scrub vegetation along the southern half and by the Long Barracks plantation in the northern part. Although there are no sections in the ridge today, arch-bedding in gravels was exposed near Goldenhill Farm (see Figure 5.42) as reported by Garwood (1893). Angular to subangular boulders, up to 1 m in diameter, are common on the crests and sides of the ridges, frequently of the locally outcropping Oxford Limestone (George *et al.*, 1976). The smaller pebbles in the gravel are generally fairly well rounded, although there are larger clasts of unabraded shale, which suggests minimal sediment transport. To back this up, striated and faceted stones have been found (Gregory, 1922; Carruthers *et al.*, 1927), which indicates subglacial sediment and little transport.

This ridge has been interpreted as an esker by Parsons (1966) and he linked it with a shallow channel, the long profile of which rises to the south-east and is incised into the till immediately south of the Whin Sill outcrop. This ridge terminates in the sand and silt deposits of Pigdon Hill. Occasionally, as at the northern end of Bradford Goldenhill and to the south-east of Spindlestone, the gravels appear to be banked against the west side of the main sand and silt deposits. This results in the more characteristic ridge morphology being divided into a series of sections generally separated from the silt and sand deposits to the east by poorly drained

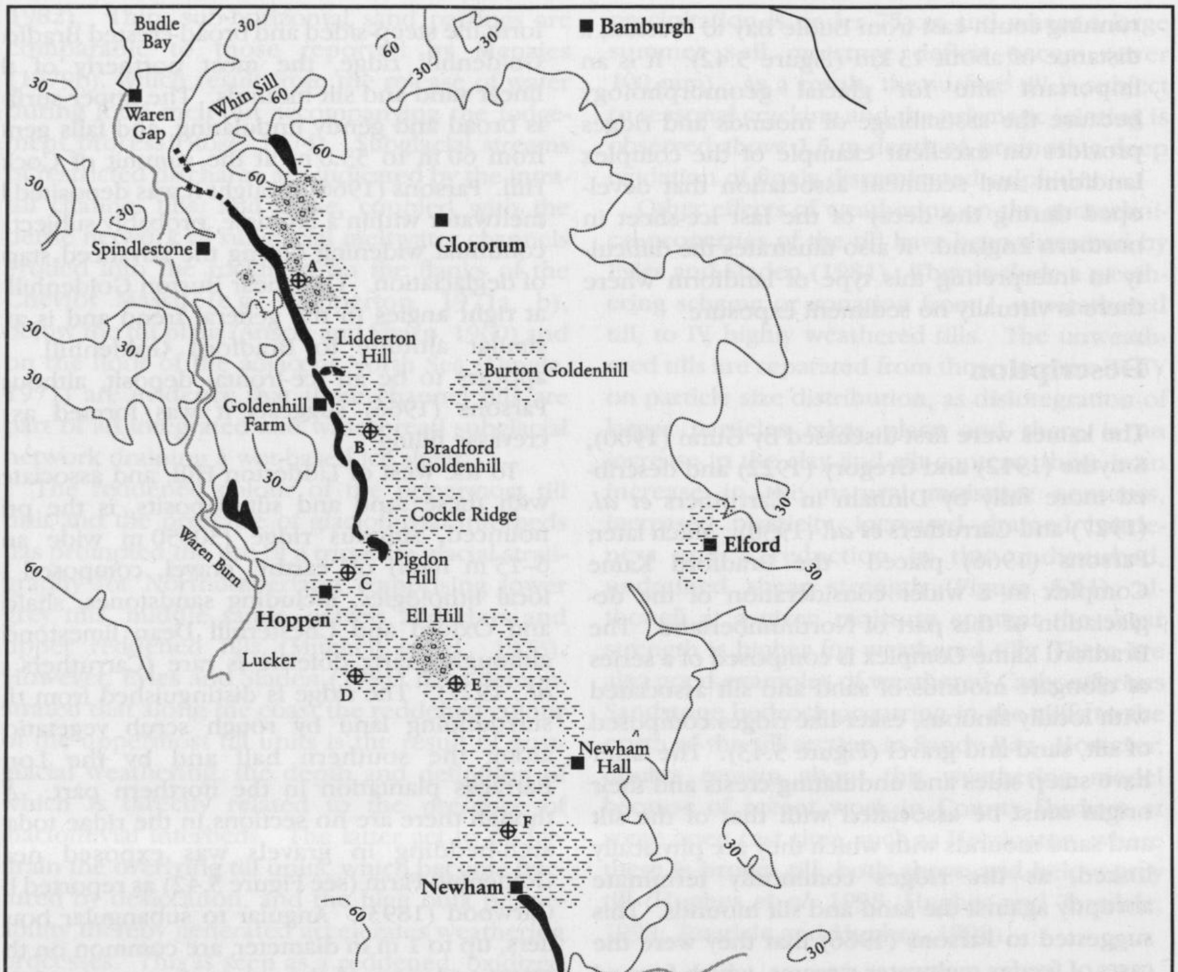


Figure 5.42 The Bradford Kames and associated deposits (from Parsons, 1966). Sections corresponding to boreholes A–I are described in Figure 5.44.

Figure 5.43 Bradford Kame topography near Long Barracks Plantation. View to the north-west, towards Budle Bay. (Photo: D. Huddart.)

hollows, once occupied by small lakes. Near Hoppen (Figure 5.42) the finer-grained deposits form a steep-sided, gently undulating, broad-topped mound, (56 m OD maximum). It has a deltaic appearance and from the south-east end of this mound a broad-topped, steep-sided ridge of silt and sand extends SSE as far as Newham. At its northern end it is 60.9 m at Ell Hill, where gravel is banked against and overlies fine-grained sediments, although the upper surface of the rest of the deposit undulates between 46.15 m and 3.8 m. South of Newham the deposits extend south-eastwards at about 30 m OD, as a sinuous and broken, irregular, esker-like ridge, composed mainly of gravel in the northern part (Gunn, 1900; Carruthers *et al.*, 1927), resting on till. The deposits rise to a maximum of 46.15 m OD at the steep-sided Horse Hill, where it is composed of sand and silt. The sand and silt deposits at Preston form a steep-sided and linear, broad-topped mound as far as Doxford Hall (see Figure 5.42), and towards and beyond Falloden Hall the deposits form a kame terrace. These deposits are not traceable beyond Christon Bank, which is thought to mark the end of the Bradford Kame Complex, although meltwater from this zone may have eroded channels at, and to the south of, Embleton.

Auguring of the sediments (Figure 5.44), reported by Parsons (1966), revealed extensive deposits of silt, laminated clay and fine-grained sand in many of the glaciofluvial deposits away from the main esker ridge and Carruthers *et al.* (1927) reported sands and clays in occasional borehole records, such as the 18.5 m of clay reported in a well at Hoppen. The auguring has been important, since much of the area underlain by these deposits was reported to be covered by boulder clay by Carruthers *et al.* (1927).

Interpretation

Gregory (1913, 1922) suggested that this kame, and kames in general, were laid down pro-glacially along the margin of the glacier. However, Dinham (in Carruthers *et al.*, 1927) thought that the greater part of the deposit suggested accumulation in a crevasse as an open channel, within a mass of clean, wasting ice and locally near its margin, although this site may have marked the site of an earlier subglacial stream. The bank on the west side of Crook

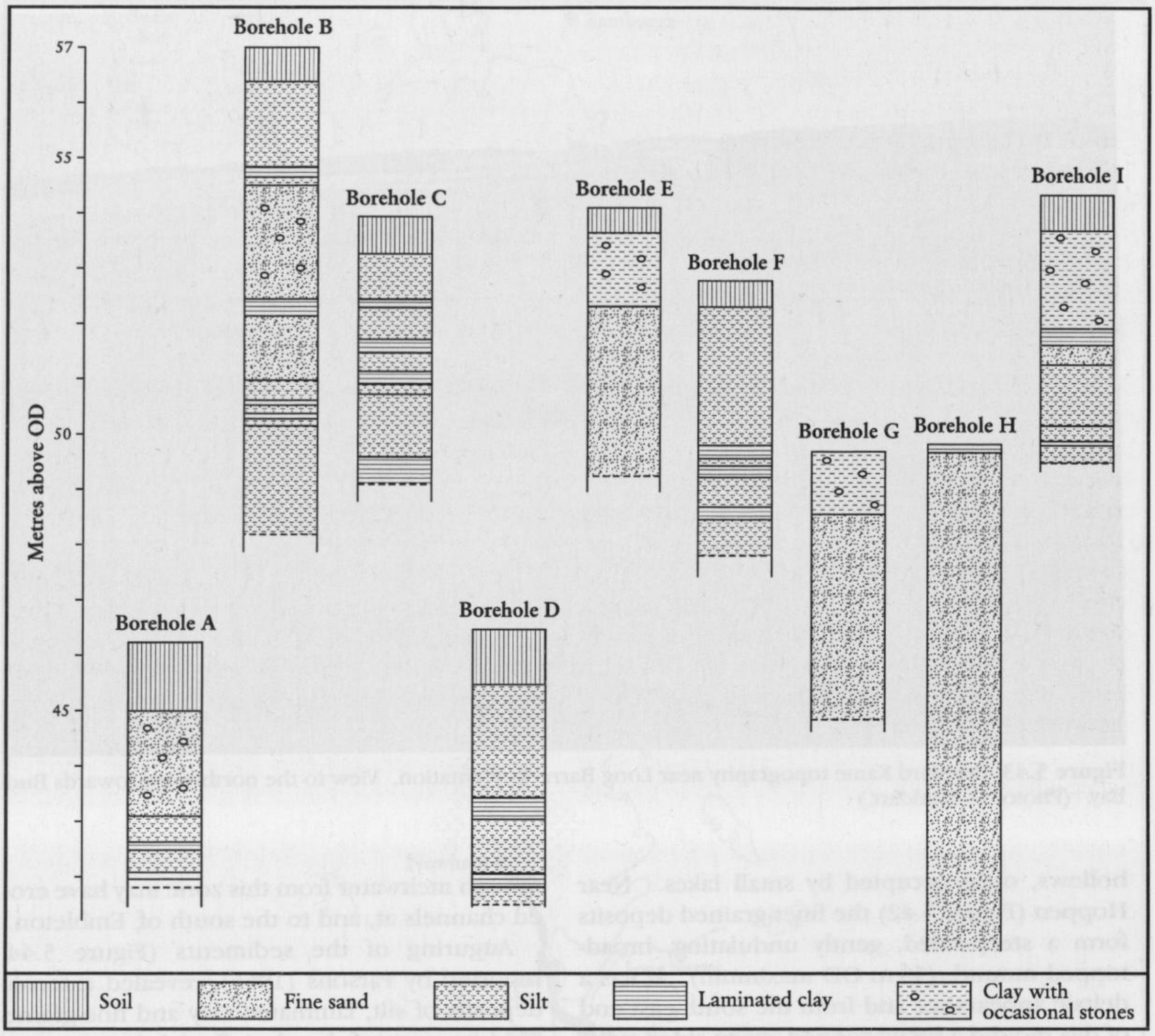


Figure 5.44 Sediments from auger holes and boreholes (locations are shown in Figure 5.42) in the Bradford Kames (adapted from Carruthers *et al.*, 1927; Parsons, 1966).

Hill–Cockle Ridge followed by the kame was supposed to mark a sudden rise of 15–27 m in the subglacial surface and a corresponding change in the ice thickness. The boundary between active and stagnant ice would be fixed and maintained along this bank for a period during which the ice margin could fluctuate and eventually retreat a considerable horizontal distance over the gradual slopes on the other side of the valley. The tensions of differential flow to be expected near such a boundary would coincide for a long time with the bank and a more or less permanent marginal crevasse might be developed here. The site of the kame was sheltered by Spindlestone Heughs on the Whin Sill ridge from the direct southerly pressure of the ice behind them and therefore deposits or crevasses would stand a better chance of preservation here. Sediment incorporated in the ice as englacial debris was released into the crevasse.

The problem with this explanation for the kame is that it does not explain the relationship with finer-grained glaciofluvial sediments, especially the laminated clays and silts, or landforms to the south and east of the main Bradford kame ridge and the fact that this type of sinuous ridge is duplicated to the south. The transition from active to stagnant ice noted above at Cockle Ridge is difficult to explain physically too, although undoubtedly the shelter of the Whin Sill ridge might be an important locating factor. Parsons (1966) interpreted the ridge system as a subglacial esker feeding meltwater and sediment into a reticulate, open crevasse system and therefore the glaciofluvial sediments were suggested to be a series of eskers and crevasse fillings in an extensive, terminal zone within decaying ice. During deglaciation Parsons (1966) believed that the ice in the northern part of this area had thinned sufficiently for the crevasse system to have penetrated to the underlying till. In parts he considered that the evidence suggests that deposition began in the south and progressed northwards, but this was believed to be a local phenomenon and that the deposits were broadly contemporaneous, probably accumulating within a decaying, terminal zone. A series of eskers and crevasse fillings extended SSE from the Whin Sill outcrop as far as Doxford Hall (see Figure 5.42). They appear to continue as a kame terrace leading south-east towards Falloden Hall, but beyond this, extensive glaciofluvial deposits are absent and the meltwater was believed to have eroded several SE-trending channels in the till west of Craster (Parsons, 1966). Douglas (1991) thought that these eskers of sand and gravel were aligned in the direction of ice flow and represented subglacial channels draining the Devensian ice sheet.

However, there recently has been renewed interest in the origin and depositional environments associated with these types of kames and eskers, and several models of sedimentation have been proposed (e.g. Brennand, 1994; Warren and Ashley, 1994; Huddart and Bennett, 1997; Thomas and Montague, 1997; Huddart, 1999; Thomas *et al.*, in press). One of these models includes sedimentation from subglacial channels into a pro-glacial lake during active ice retreat, which gives a distinctive landform and sediment association (Thomas *et al.*, in press). This model has been proposed for the Newbigging esker system in the South Medwin valley (Lanarkshire), whereas the nearby Carstairs kames have been explained by Huddart and Bennett (1997) as deposition within a supraglacial topography of ice-cored ridges, which gave a diverse range of depositional environments and a number of distinct glaciofluvial systems. Adjacent to the main ridges at Carstairs, mounds composed mainly of sand show evidence of fluvial channel systems and glacio-lacustrine, ice-contact environments, consistent with supraglacial kames. From the landform morphology and the limited sedimentary records proved by Parsons (1966), it is suggested that the Bradford Kame Complex developed in a similar series of environments to those in Lanarkshire.

1. Subglacial eskers. These occur in the Bradford kame proper between Spindlestone and Pigdon Hill and again between Newham and north of Preston. They are composed of gravel but with surface angular boulders on their crests and sides derived from supraglacial sources as the ice melted. The arch-bedding seen by Garwood (1893) is indicative of a subglacial system, although it is conceivable that these deposits could be supraglacial eskers (Huddart and Bennett, 1997).
2. Pro-glacial, glacio-lacustrine deltas. The much bulkier deposits at Burton Goldenhill, Cockle Ridge, Hoppen, Ell Hill and near Preston are steep-sided, undulating to flat-topped mounds with a gradient to the south-east. They are composed largely of sands and silts, although gravels are located at their northern, up-ice ends at the ice-contact face. They are interpreted as pro-glacial deltas that prograded into lakes as the ice retreated actively to the north. There are three such suites of landforms.
3. Other landforms. The roughly elongate sand and silt mounds best developed between north of Lidderton Hill and Cockle Ridge are thought to represent either finer-grained glaciofluvial sequences that built up as fans into the lake as breakouts from the main subglacial esker, or as a series of supraglacial kames. The finer-grained sands–silt landforms could be developed in supraglacial lakes bottomed on till.

This model invokes active retreat to the north, although between Pigdon Hill and Spindlestone in the lee of the Whin Sill there could have been stagnant ice and supraglacial environments at a late stage in deglaciation in this area. There is evidence for glaciofluvial sediment and erosion landforms in the North and South Charlton areas to the south-west of the Bradford Kame Complex, which resulted in thick sequences of glacio-lacustrine and glaciofluvial sediments in the Lower Aln valley (Parsons, 1966). To deposit such sequences of glacio-lacustrine sediments associated with the Bradford Kame there must have been a continuous ice dam to the north-east as ice retreated actively to the north. This means that the meltwater channels between Embleton and Craster, thought to be eroded by meltwater from the Bradford Kame Complex by Parsons (1966), must have been independent from that series of deposits and probably cut by subglacial erosion. It is possible that Carruthers *et al.* (1930) were correct in linking the Bradford deposits with the North and South Charlton deposits and it is considered that all these deposits and landforms best fit into a model of active ice retreat, punctuated by stillstands to build up sediments in fronting, pro-glacial lakes in topographically deeper valleys. This means that the model of crevasse fillings in an extensive, terminal zone within decaying ice, supported by Parsons (1966), is likely to be wrong for the whole complex, although a supraglacial series of landforms and deposits could be argued for the northern section close to the Whin Sill. However, although it forms a complex landform and sediment assemblage and is one of the best examples of glaciofluvial–glacio-lacustrine sedimentation in northern England, its exact mode of origin still remains to be explained completely, simply because the sediment exposure is so limited. To obtain the precision in interpretation for the depositional environments provided by Huddart and Bennett (1997), Thomas and Montague (1997) and Thomas *et al.* (in press), good sediment exposure is essential. Nevertheless, the model suggested, where the landforms are associated with an active ice retreat and subglacial meltwater inputs into pro-glacial lakes, is thought to be the most likely depositional model.

Conclusions

In the account given here, a description of the Bradford Kame Complex is presented and the various hypotheses for its formation discussed. As there are virtually no sediments exposed at the time of writing, the precise depositional environments are difficult to unravel. However, active ice retreat from south to north led to the deposition of proglacial deltas fed from subglacial eskers at three stages in the Preston area, at Ell Hill and at Cockle Ridge–Pigdon Hill–Hoppen. To the north of the latter ridges the ice probably became stagnant behind the Whin Sill ridge and a series of supraglacial environments developed. Without detailed sedimentological evidence these suggested stages remain hypothetical.

HUMBLETON HILL AND THE TROWS (NT 951 275 AND NT 963 283)

D. Huddart

Introduction

Humbleton Hill and the Trows are two separate glacial meltwater systems, which lie 4 km west of Wooler in the eastern Cheviot Hills (Northumberland). The localities illustrate some of the best examples of subglacial meltwater erosion in the north of England. The types of channel range from the simple col channel of Humbleton Hill, with its classic, 'up-and-down' or 'humped' long profile, to the complex, anastomosing channel system of the Trows, with its perched meander loops and plunge-pool channel heads. The site is an excellent locality for demonstrating typical subglacial meltwater channel characteristics that are common throughout northern England and for illustrating how the interpretation of these types of channels has changed throughout this century.

In the northern and eastern Cheviot Hills such channel systems were first associated with glacier meltwater streams by Clough (1888) but Kendall and Muff (1901, 1903) and Smythe (1912) subsequently thought that the formation of these channels in this region was associated with overflows from ice-dammed lakes. This followed from the pioneering monograph of Kendall (1902) in the North Yorkshire Moors.

In Kendall and Muff's work they examined selected channels and ascribed them to lakelets existing at the heads of the valleys, the regional pattern being produced by a sequence of lake levels controlled by massive, impermeable ice. This work was expanded by Smythe (1912), who, on the basis of the channels and related evidence, postulated a series of retreat stages of the last ice sheet in the region. However, Carruthers *et al.* (1932) and Common (1957) suggested that many of the channels in the Cheviot Hills were eroded by ice-marginal meltwater streams. Despite this early work on the origin of the meltwater channels neither of the suggested mechanisms of lake overflow or ice-marginal channels is now considered as the likely mechanism for formation. Hence these two localities illustrate well how there has been

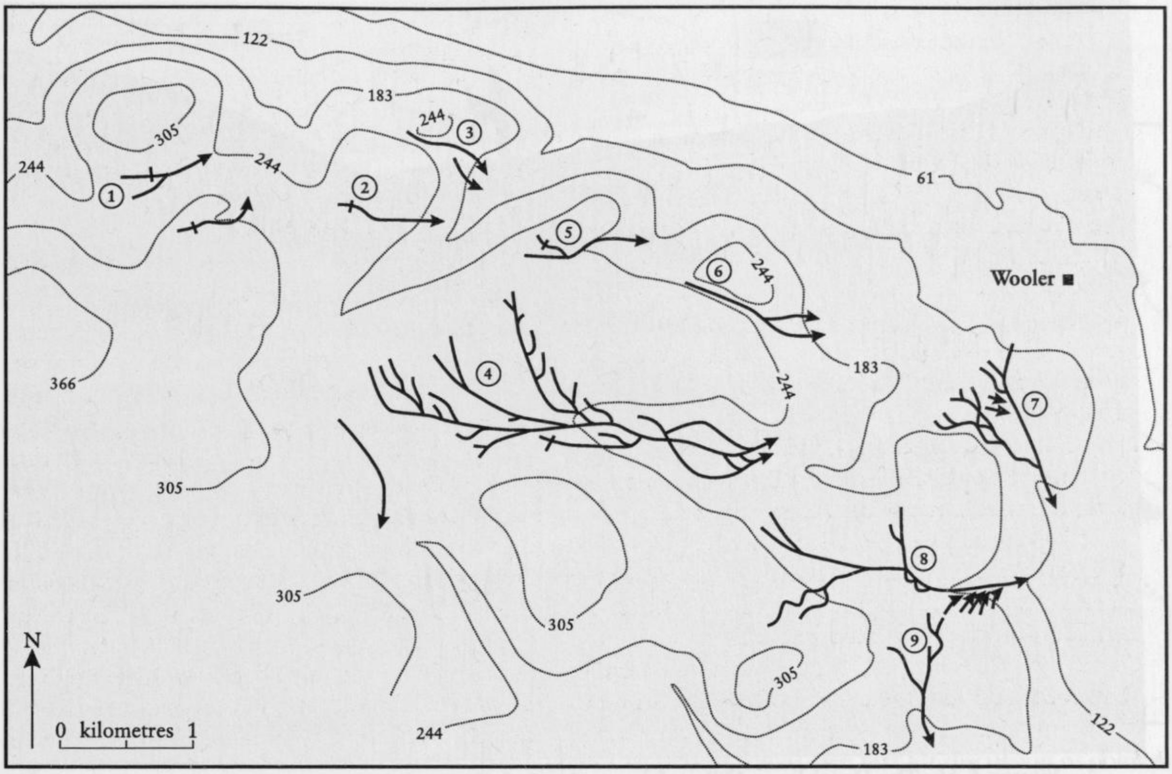


Figure 5.45 Meltwater channels in the north-east Cheviot Hills (after Clapperton, 1968). See text for details of channel/channel systems 1 to 9.

major shifts in the interpretation of the origin of meltwater channels. Both Derbyshire (1961) and Clapperton (1968) have suggested alternatives for the erosion of these channels as subglacial channels.

Description

Humbleton Hill channel

This channel was one of Derbyshire's (1961) col gullies, which breached pre-existing cols in the north and east Cheviot Hills. It is cut mainly in bedrock and occupies a relatively narrow channel cutting across the south side of Humbleton Hill (channel 6 on Figures 5.45 and 5.46). It has very steep sides and is over 30 m deep. Many of these types of channels have reversed gradients, double intakes, double outlets and interrupted long profiles. The Humbleton Hill channel shows some of these characteristics, as it rises for the first 200 m of its course and bifurcates at its lower end. Kendall and Muff (1903) had considered that the channel flowed either side of a postulated delta. However, field observations show that the final outlet of the water was by way of the Humbleton Burn valley to the west of Wooler. The slight discordance at the point of bifurcation coincides with the outcrop of a transverse dyke across the channel. There also is no field evidence for lake shorelines or a delta. Part of the channel has been infilled with post-glacial scree below the hillfort on the north side of the valley. To the west there is another channel system to the south side of Monday Clough (channel 5 in Figure 5.45), but this indicates meltwater transport to the north-east.

The Trows channel system

This is an intricate system of meltwater channels (Figures 5.47, 5.48, and system 4 in Figure 5.45), which is located in the col and slopes forming the broad head of the Humbleton Burn valley. The system consists of three main channels that join and extend part way down the small, misfit Humbleton Burn (Figure 5.47). The channel height range is between 208–320 m and the channels vary from intake furrows only 0.6 m deep to gorges over 22 m in depth. As far down-

Figure 5.46 The Humbleton Hill meltwater channel. View to the east. (Photo: D. Huddart.)

channel as point 'A' in Figure 5.47 the system is incised mainly in a microgranite bedrock, although a thin veneer of drift overlies the rock-head in places. Wide channels below 'A' have dissected a till plug, which is probably the 'earthy angular drift' of the early [British] Geological Survey work. Three shallow intakes to channel 1 (Figure 5.47) begin on the crest of Black Law ridge. The intakes to channel 2 begin on the south-eastern slope of the ridge just below its crest and three short feeders of channel 3 head on the col crest between Black Law and Grains Law. All the feeders begin as minor, shallow landforms, but each one suddenly expands into a plunge-pool section that varies in depth from 3 m in channel 2 to 18 m in channel 3. Shortly below these sections the feeders unite to form their respective main branches of the channel system. The system contains several abandoned loops at various levels above the main floor and this has partially isolated rocky knolls. Stream channels must have flowed side by side and were separated by only a narrow divide in several sections of the system, but eventually all meltwater was concentrated on the valley floor.

Interpretation

Derbyshire (1961) considered the evidence for the lake overflow hypothesis for the meltwater channel systems in the north and east Cheviot Hills and argued that lake-bottom sediments were entirely absent, that there are no undoubted shorelines and that there was a complete absence of true deltaic landforms. He therefore came to the conclusion that the lake-overflow hypothesis for the origin for the col gully channels was untenable. He suggested that the frequent reverse gradients within channels, accordant tributary development and inconsistent topographical relationships cast doubt on the then accepted ideas related to ice meltwater drainage in this area. He thought that as glacial till was found in the majority of such channels deposition of such sediment took place after erosion of the channels subglacially. He called these types of channels 'subglacial col gullies'. Clapperton (1968), however, cast doubt on some of Derbyshire's (1961) field evidence, such as the frequent reverse gradients and presence of till. The Humbleton Hill channel was used to illustrate some of these doubts as no evidence of

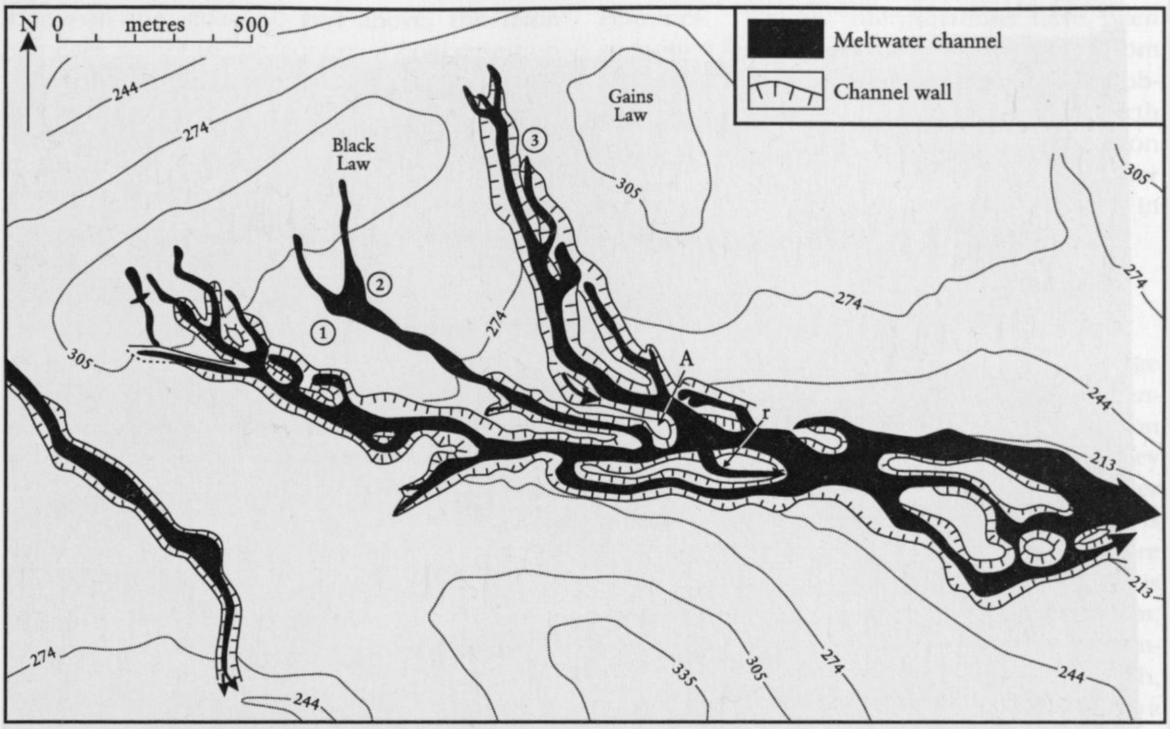


Figure 5.47 The Trows meltwater channel system (after Clapperton, 1968). See text for discussion of point A and loop r and channels 1 to 3.

till is apparent, but there is much post-glacial accumulation of slope deposits on the valley floor subsequent to its formation, so that any till deposits that might have been deposited originally would now be completely hidden. To Derbyshire (1961) the evidence from the col gullies suggested subglacial erosion, which represented the flow of meltwater under hydrostatic pressure at a level within the ice that was at, or above, the melting point. Below such channels the ice was compact and impermeable. Where col gullies occur at decreasing altitudes in the same topographical area they can be dated relatively to one another. He thought that subglacial waters in the early phases of deglaciation tended to flow outwards to areas of thinner ice and therefore the direction of flow of these channels is a good indicator of regional ice thickness.

The Trows channel system, which is an accordant system, strongly suggested to Derbyshire (1961) that the stream valleys were used as major gutters for subglacial drainage. He considered that a marginal origin for the system was difficult to reconcile with the situation of their intake only 5–40 m above the present streams. Such a thin and restricted ice mass was thought not to be sufficiently impermeable to retain drainage at the surface during the later stages of deglaciation. Derbyshire (1961) considered that this type of system represented a stage of glaciofluvial erosion marked by the formation of subglacial channel systems that showed no control by ice. The abandoned loops suggested a subglacial origin, especially the most peculiar abandoned section, which is the largest loop (point 'r' in Figure 5.47) and which climbs *c.* 12 m to a point where it resumes a course parallel to the main channel. The volume of water that cut this loop was sufficient to erode a narrow, bedrock gorge, which deepens from 6 m to 22 m along this section of the system. The main channel is here nearly 31 m deep and is rejoined after 460 m by loop r descending with a steep profile. The hydrostatic head of water pressure must have forced water out of the main channel at right angles to form loop r.

Clapperton (1968) refined the subglacial model for these channels systems by suggesting that the meltwater associated with ice downwasting over the Cheviot area became subglacial mainly by the superimposition of englacial meltwater streams in submarginal zones of the

Figure 5.48 The Trows meltwater channel system. View to the north-west. (Photo: D. Huddart.)

downwasting ice mass. As the meltstreams were aligned across the trend of spurs and valleys it was inevitable that when these stream systems became superimposed on to the underlying topography that many of the streams eroded channels across the spurs and in some valley heads. Clapperton (1968) argued persuasively that the various complexities of the channel systems needed ice during their formation and that meltwater associated with this ice must have flowed in tunnels beneath the ice. As the largest and most complex channel systems were consistently located in pre-existing cols and valley heads, and are all located in areas of broken and considerable relief, the superimposition of englacial streams explains most satisfactorily how the eroding meltwater became subglacial. Pre-existing valleys and spurs radiate from the north-east to east of the Cheviot Hills and were orientated approximately at right angles to the direction of ice movement and subsequently to that of meltwater flow upon deglaciation. If a system of englacial streams were superimposed on to a topography of short deep valleys, separated by spurs with well-developed cols in their crests, then the following events would be expected: streams that are superimposed on to the steep slopes of narrow cols and spur ends would be likely to slip laterally downwards eroding mainly in the ice at the ice–ground contact rather than into the steep hillside. In the meantime, adjacent streams located over the site of col floors would probably cut down more quickly through the ice. If connections existed between such adjacent streams and those over the col floors then the lower streams may be expected to tap the waters of those at higher levels, thereby concentrating the flow of meltwater more and more over the sites of the col floors. Eventually, incision into the col floor may be made with the combined volume of a previously spread-out englacial channel system.

The Trows channel system suggested to Clapperton (1968) that a broad shallow embayment type of valley head allowed latitude for the superimposition of englacial streams more as a complex system with many branches. Here several streams became deeply incised independently before uniting as the valley narrowed downstream. Frequent abandoned channel loops on the sides of, and above, the main branches testify to the continual concentration of meltwater into a subglacial position at the lowest available route.

The importance of such meltwater channels in interpreting the deglaciation pattern in the east Cheviot area has been discussed by Clapperton (1970), where he identified four main phases of meltwater drainage. The channels in this GCR site resulted from an ice-directed drainage in the first phase of deglaciation. In a later paper, Clapperton (1971b) discussed the location and origin of glacial meltwater phenomena in the east Cheviot Hills, which included not only the meltwater channels, but also the origin of the associated sands and gravels in the later stages of deglaciation. The orientation of meltwater channels also was used by Clapperton (1968), along with much other evidence, such as glacial striations, erratics, moraines and features of glacial erosion, to support the concept that the Cheviot massif acted as an independent ice dispersion centre during the last glacial. Glacier ice flowed radially from the core of the Cheviot massif and from the higher summits in the Cheviot Hills (Figure 5.4a).

Conclusions

The importance of this GCR site is that it shows two excellent examples of meltwater channel systems that have been interpreted in different ways over the last century. These have ranged from origins as lake overflow channels, through marginal channels to subglacial meltwater systems. It is now considered that they have been formed from the superimposition of englacial stream systems on to the underlying topography, eventually to form as subglacial channels. They are important in that they have helped us understand the mode and pattern of deglaciation and the condition of the ice during this melting. Their orientation, along with other lines of evidence, has helped establish an independent Cheviot ice cap during the last glacial.

LUDWORTH INTAKE (SJ 994 911)

N.F. Glasser

Introduction

Ludworth Intake is a prominent meltwater channel incised into the western flank of the Pennines. Various interpretations have been suggested for its origin including a spillway from a glacial lake, an ice-marginal channel and a subglacial channel. Jowett and Charlesworth (1929), Johnson (1963, 1968) and Stevenson and Gaunt (1971) have described the meltwater channel and its relationship to glacial events in the area.

Description

Ludworth Intake lies to the south of the village of Chisworth on the outskirts of Greater Manchester. The channel is incised into a low col on the watershed between the main Etherow Valley and a small tributary of the River Goyt. Under present-day climatic conditions, the channel is dry and carries no surface drainage (Figure 5.49). The main channel of Ludworth Intake is over 400 m long and attains a depth of 10 m. The channel is sinuous in plan and its floor generally falls in height from north to south, although in its most northern section it has a slight reverse or northward slope.

Johnson (1968) described the nature of the sediments mantling the sides of Ludworth Intake. These sediments were revealed by temporary sections cut in the side of the channel for a coal mine adit. This exposed a sequence of soliflucted debris layers overlying weathered shale. Johnson (1968) identified five separate layers of solifluction debris in this section. Two layers were formed of heavily comminuted clay shale, whereas the remaining three were of a much coarser texture in which shale and mudstone fragments were still identifiable. The extent of comminution in each of the debris layers was taken by Johnson (1968) to represent differing climatic regimes during deposition. Pollen grains from between the layers of finer-grained debris are consistent with a Windermere Interstadial age (Johnson, 1968).

Interpretation

Ludworth Intake was originally interpreted as a channel cut by a lake spillway for an ice-dammed lake confined to the Etherow Valley (Jowett and Charlesworth, 1929). This interpretation was inspired primarily by the pioneering work of Kendall (1902, 1903) on the lake spillways and ice-dammed lakes in the North York Moors area. Kendall identified numerous ice-dammed lakes in the North York Moors area, including a large

Figure 5.49 Photograph of part of the Ludworth Intake meltwater channel. (Photo: N.F. Glasser.)

lake (Lake Pickering) and several smaller ice-dammed lakes pinned against the northern flanks of the moors. Although other researchers accepted Kendall's ideas at the time, later challenges to his theory have meant that in many descriptions of Ludworth Intake the spillway interpretation has been largely disregarded. Johnson (1965a) divided the meltwater channels of the western Pennines into two categories; a first group of widely dispersed channels cut across watersheds and cols at high levels, and a second group of subglacial drainage channels developed on the lower slopes of the hills. Ludworth Intake clearly belongs in the former of these two categories. Both Johnson (1963, 1965a, 1968) and Stevenson and Gaunt (1971) suggested that the channel was cut by a meltwater stream flowing either laterally along the ice-sheet margin or directly under the ice-sheet edge. These meltwater streams were fed by seasonal flow directly from the ice front or from meltwater passing from one part of the ice margin to another. A subglacial or submarginal origin would best explain the reverse gradient of parts of the channel. Johnson (1985b) later developed this concept by applying a landsystems approach to the western Pennine margin in which meltwater channels at different altitudes were related to successive lowering of the margin of a lowland ice-sheet pinned against an upland area. A good analogue for this situation is provided by Derbyshire (1961) who described similar col channels in Labrador. Derbyshire argued that channel erosion takes place when the ice surface is between 120 and 150 m above a col divide. As most col channels in the Manchester area (including Ludworth Intake) occur at heights between 275 and 370 m OD, the ice surface was probably more than 460 m OD at the Pennine margin when the channels were eroded. More recent work on the meltwater channels of the Mid-Cheshire Ridge, to the west of the Pennines, including palaeohydraulic calculations and detailed ice-sheet reconstruction, has confirmed that many of these channels are indeed subglacial in origin (Sambrook Smith and Glasser, 1998; Glasser and Sambrook Smith, 1999).

Finally, Johnson (1968) attempted to establish a relative chronology for events at this site from the solifluction debris and pollen remains exposed in the temporary section. He concluded that the Ludworth Intake meltwater channel was initially cut into weathered shale before it was partially infilled by solifluction debris. The coarse basal layers of solifluction debris were interpreted as having formed immediately after deglaciation, whereas the upper coarse layers were a product of the return to a colder climatic regime during the Loch Lomond Stadial. A Windermere Interstadial age is attributed to the layers of finer-grained debris that separate these layers of coarse debris (Johnson, 1968).

Conclusions

Ludworth Intake is an exceptionally good example of an isolated col channel cut by glacial meltwater erosion, a landform type common along the western Pennine margin. Although initially it was suggested that the channel originated as a spillway from a glacial lake, it now appears more likely that these channels were eroded by meltwater flowing in an ice-marginal or subglacial position. The meltwater channel at Ludworth Intake therefore is important for interpreting the nature of events and ice-sheet thickness during the Late Devensian glaciation.

NEWTONDALE (SE 820 930) AND HOLE OF HORCUM (SE 845 935)

N.F. Glasser

Introduction

Newtondale and Hole of Horcum are both classic geomorphological features of importance to our understanding of Quaternary glacial and periglacial events in northern England. Newtondale is one of the largest and most spectacular glacial meltwater channels in Great Britain, forming a deeply incised, continuous channel for over 10 km across the North York Moors (Figure 5.50). The channel at Newtondale is also significant to the history of the development of geomorphological thought because of its role in the debate concerning the nature of ice dynamics during recession. Various origins have been proposed for the channel, including an overflow channel from a large Late Devensian pro-glacial lake and a subglacial channel pre-

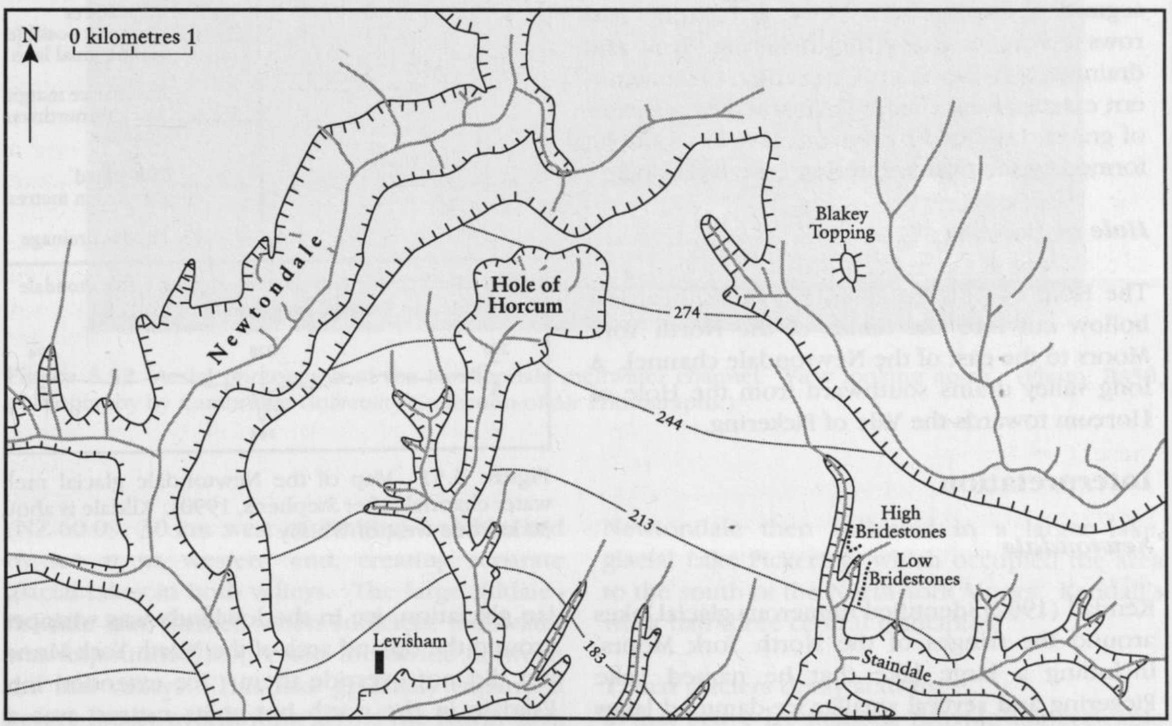


Figure 5.50 Map of the southern part of the North York Moors showing the location of Newtondale, Hole of Horcum and the Bridestones (modified from Palmer, 1956).

dating the Late Devensian. Kendall (1902, 1903) originally described the glacial lakes surrounding the North York Moors and their supposed drainage patterns. Other published accounts of the site include the description of Sewell (1904), and the re-interpretations of Smith (1932) and Gregory (1962a, b, 1965).

The nearby Hole of Horcum is a large amphitheatrical hollow cut into the North York Moors. Its origin is controversial, and mechanisms involving glacial meltwater, spring sapping and nivation have been proposed to explain its existence.

Description

Newtondale

Newtondale is a 10 km long meltwater channel crossing the main watershed of the North York Moors at 228 m OD (Figure 5.51). The channel is flat-floored and steep-sided for much of its length with a meandering channel pattern (Figure 5.52). The maximum depth of the channel is over 60 m at its head in the Eller Beck valley but the average gradient of the channel floor is low (1:1250). The altitude of the channel floor falls consistently southward along its length with no reverse (northward-facing) slope segments. At its southern end, the channel narrows before entering the Pickering Beck and draining into the Vale of Pickering. The southern extent of the channel is marked by a spread of gravel described by Kendall (1902) as a delta formed by meltwater entering Lake Pickering.

Hole of Horcum

The Hole of Horcum is a large amphitheatrical hollow cut into the flanks of the North York Moors to the east of the Newtondale channel. A long valley drains southward from the Hole of Horcum towards the Vale of Pickering.

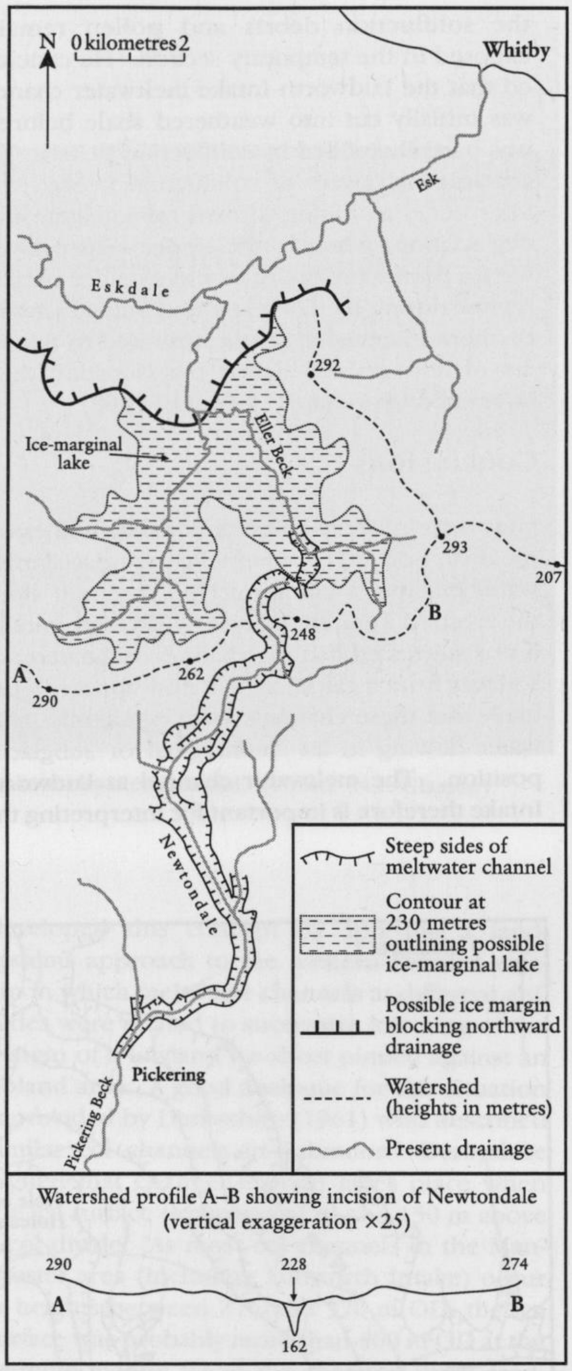


Figure 5.51 Map of the Newtondale glacial meltwater channel (after Stephens, 1990). Kildale is about 30 km due west of Whitby.

Interpretation

Newtondale

Kendall (1902) identified numerous glacial lakes around the fringes of the North York Moors, including a large lake that he named Lake Pickering and several smaller ice-dammed lakes pinned against the northern flanks of the moors. Kendall's view was that during the Late Devensian glaciation, ice in the lowlands was wrapped around the upland area of the North York Moors but did not override them. Ice extended into Eskdale in the north but at its nearest was at least 8 km from Newtondale. Eskdale was blocked by ice at its eastern end and Kildale

Figure 5.52 Aerial photograph of the Newtondale meltwater channel. View looking north. (Photo: BA30, photography by Cambridge University Collection of Air Photographs.)

(NZ 60 09, 30 km west of Whitby) was blocked by ice at its western end, creating separate glacial lakes in both valleys. The large Kildale–Eskdale lake formed when sufficient meltwater was impounded to overtop the divide between the two valleys. This lake gradually expanded before finally overflowing across the North York Moors through Newtondale at the lowest point in the Cleveland anticline. Meltwater from Newtondale then collected in a larger lake, glacial Lake Pickering, which occupied the area to the south of the North York Moors. Kendall's work had three central principles:

1. that glaciers decay actively;
2. that active ice margins provide impermeable barriers and that where ice sheets block natural lines of drainage ice-dammed lakes form;

3. that meltwater channels form on cols and spurs where there is drainage from ice-dammed lakes via overspill channels.

Kendall therefore regarded Newtondale as one such 'overflow' channel cut by meltwater draining these lakes. Overflow channels are essentially pro-glacial channels formed where water decants from ice-dammed or pro-glacial lakes over a col or low point in the topography. Spillways can be cut over a long period of time by regulated overflow water or by the catastrophic release of water during jökulhlaups. Kendall's original (1902) paper had a profound effect on British glacial geomorphology, primarily because he made a convincing argument that ice sheets were watertight and therefore could act as dams and allow the formation of lakes in valleys not actually occupied by ice.

Kendall's ice-dammed lake and overflow channel paradigm was accepted and reiterated by subsequent workers both in the immediate geographical area (Sewell, 1904; Elgee, 1908; Smith, 1932; Radge, 1939), and also in other areas of Britain (Charlesworth, 1926, 1955; Edwards, 1937; de Boer, 1945; Dean 1947, 1953; Williamson, 1952; Twidale, 1957). Subsequently, glacial lakes were envisaged to have occupied large areas of lowland Britain during deglaciation. These lakes include Lake Humber, Lake Fenland, and Lake Lapworth in north-west England (Jones and Keen, 1993). Other workers (e.g. Best, 1956) also embraced the concept of ice-dammed lakes in later accounts of the deglaciation of the North York Moors, although they began to question the relatively simple model of Kendall. Best (1956) recognized that two glaciations were required to account for the complex pattern of meltwater channels present in the area.

The 'overspill' theory eventually was challenged by Sissons (1958b, 1960b, 1961), who demonstrated that many of the meltwater channels identified as overspills were in fact subglacial in origin. Drawing on a growing body of literature from Scandinavia that stressed ice-sheet stagnation during recession, Sissons argued that glaciers need not decay actively and that ice need not provide an impermeable barrier to drainage. In a series of papers on the former ice-dammed lakes of the Scottish Highlands (Sissons, 1977, 1978, 1979a, 1982) he went on to establish more rigorous criteria for the recognition of ice-dammed lakes. These include the need for unequivocal evidence of former shorelines, deltas and lacustrine sediments.

Gregory (1962a, b) described the landforms of deglaciation in Eskdale and noted numerous lines of evidence in this area to support Sissons' theory of ice stagnation. Gregory (1962a, b, 1965) disputed Kendall's claims that Newtondale is an overspill channel through his study of the glacial lakes of Eskdale, to the north of the Newtondale channel. His work showed that much of the meltwater in fact drained subglacially to the north-east. Gregory (1965) noted that of the four lines of evidence used by Kendall to support the existence of a pro-glacial lake (overflow channels, lacustrine deposits, deltas and lake shorelines), there was little of this evidence to support the existence of a pro-glacial lake in Eskdale. This led Gregory to propose that although Newtondale is undoubtedly a meltwater channel, it is unlikely that it was created purely by overflow of an Eskdale–Kildale glacier lake during the Devensian. Support for this theory comes from the gravel spread at the southern end of the channel. Catt (1977c) suggests that the size of this gravel spread is anomalously small and fine-grained to have resulted from the high discharge required to cut a channel the size of Newtondale. Catt (1977c) therefore argued that although Newtondale may have been used by meltwater during the Late Devensian, it was probably cut in a pre-Devensian glaciation. There is, however, no lithostratigraphical or chronostratigraphical evidence to prove this theory.

Since the 1960s, descriptions and interpretations of other meltwater channel systems in Great Britain have confirmed that channels can form in a variety of ways and that a number of different meltwater landforms exist (Clapperton, 1968, 1970, 1971a, b; Russell, 1995; Sambrook Smith and Glasser, 1998; Glasser and Sambrook Smith, 1999). This work has been supplemented by descriptions of glacial lakes in modern glacial environments. There is evidence that some of the criteria suggested by Sissons may be too rigid because the record of former ice-dammed lakes depends to a great extent on the nature and topographical setting of the lake (Bennett *et al.*, 1998). For example, ice-dammed lakes that occupy supraglacial positions are entirely capable of producing large spillways of the type originally envisaged by Kendall, without leaving significant shorelines, deltas or lacustrine sediments (Bennett *et al.*, 1998). Newtondale is, however, still cited in modern textbooks as a good example of a 'spillway' (Benn and Evans, 1998).

Hole of Horcum

Several processes can be invoked to explain the large hollow at this site, including erosion by glacial meltwater, spring sapping and nivation. The first of these theories, glacial meltwater erosion, appears unlikely given the morphological dissimilarity between the Hole of Horcum landform and the adjacent Newtondale channel. Glacial meltwater may have helped to enlarge the valley downstream of the Hole of Horcum but it is difficult to see how this process could have cut the hole itself. The role of spring sapping in the evolution of the landform is perhaps more important. 'Spring sapping' is the name given to the process of erosion where groundwater flow is concentrated in more permeable zones within bedrock (Summerfield, 1991). This encourages chemical weathering or dissolution of the bedrock, which, in turn, leads to an increase in hydraulic conductivity and a further increase in the rate of water movement. This positive feedback mechanism leads to headward erosion as the spring retreats by gravitational undermining and slumping of the slope. Spring sapping is favoured in situations where a permeable lithology overlies an impermeable lithology. This process can create large amphitheatre-like valleys known as 'steepheads' or 'pocket valleys' (Ford and Williams, 1989). These can be significant landscape features, measuring several kilometres across (Small, 1965; Issar, 1983).

The last of these possible processes is that of nivation, or snow-patch erosion. 'Nivation' is an all-encompassing term now used to describe the processes of weathering and transport that are accelerated or intensified by the presence of snow patches. These processes include intensive freeze–thaw activity, enhanced chemical weathering, slopewash, transport of debris by snow creep and accelerated solifluction through saturation of regolith downslope by melting snow (Thorn and Hall, 1980; Thorn, 1988; Ballantyne and Harris, 1994). Using modern examples in Greenland, Christiansen (1998a) concludes that the main nivation processes are backwall failure, sliding and flow, niveo-aeolian sediment transport, supra- and en-nival sediment flows, niveo-fluvial erosion, development of pro-nival stone pavements, accumulation of alluvial fans and basins, and pro-nival 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 can, however, be 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 form of similar large landforms such as the dry valleys of the neighbouring Yorkshire Wolds (Cole, 1879, 1887; Mortimer, 1885; Lewin, 1969; de Boer, 1974). A particularly fine example of this is the chalk dry valley at Millington Pastures (Waltham *et al.*, 1997). Similar features occur in the chalk of southern England at the Manger (Paterson, 1977) and the Devil's Dyke (Small, 1962). These valleys may have been excavated by a combination of solifluction and subaerial fluvial action under periglacial conditions. 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 by various authors 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).

Ballantyne and Harris (1994) consider that many of the larger features mapped in upland Britain that often are attributed to nivation processes, such as cryoplanation terraces and nivation cirques, are in fact inherited features such as immature glacial cirques. Nivation cirques 300 to 500 m in diameter and up to 200 m deep would take prohibitively long time periods 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, because there is a threshold length between snow patches and glaciers of 30–70 m from backwall to toe (Ballantyne and Benn, 1994). Finally, and perhaps most conclusively, is the fact that the dimensions of these relict nivation cirques are an order of magnitude larger than active nivation hollows found in the present-day Arctic area. In light of the above, it appears unlikely that nivation processes alone are sufficient to account for a feature the size of the Hole of Horcum. These large features therefore are most probably pre-Pleistocene erosion surfaces or structural benches that have merely been modified by subsequent nivation and frost action processes. Alternatively, features such as the Hole of Horcum may represent immature glacial cirque forms occupied during the build up and decay of the Quaternary ice sheets.

Conclusions

Together, Newtondale and the Hole of Horcum are important for interpreting the Quaternary history of North Yorkshire and for the study of glacial and periglacial processes operating in Britain during the Quaternary Period. Newtondale is a spectacular glacial meltwater channel created either as a spillway from a large ice-dammed lake or in a glaciation pre-dating the Late Devensian. Interpretation of the genesis of the channel is important because of the implications for the wider Quaternary history of this part of northern England, including the existence and stability of postulated ice-dammed lakes. The origin of the Hole of Horcum also is unclear, although it appears likely that the feature is an immature glacial cirque or a nivation hollow, possibly exploiting structural weaknesses in bedrock.

ANNASIDE AND GUTTERBY BANKS (SD 085 966–SD 104 837)

D. Huddart

Introduction

This site provides important sedimentological, stratigraphical and geomorphological evidence for interpreting the glacial stratigraphy of Cumbria and has been discussed by Mackintosh (1870, 1877), Smith (1912), Huddart (1970, 1991, 1994, 1997), Huddart and Tooley (1972) and Huddart *et al.* (1977). The morainic landforms along the narrow coastal plain between Black Combe and the sea consist of a series of hummocky hills, with many enclosed depressions, and there is a definite marginal-edge limit between Coteley Bank in the south to Annaside, with an outlier around Barfield Tarn. The best sectional evidence is found between Annaside and Gutterby Banks (SD 085 966–SD 104 837).

Description

In describing the coastal plain drifts between the Duddon and St Bees, Mackintosh (1870, 1877) recognized three divisions:

1. an upper, red, loamy clay, containing few boulders, partly derived from Permian strata;
2. sand and gravel, containing pebbles and a few boulders of most of the rocks found in the clay above and below it;
3. a lower, reddish, yellowish or yellowish-brown boulder clay, which was largely derived from volcanics and Coal Measure shales.

Mackintosh thought the drifts were deposited by floating ice, and although his simplified stratigraphy was regarded as broadly correct, Smith (1912) argued that the 'lateral and vertical variations in character and the changes in level of the drifts are probably much greater than he (Mackintosh) imagined.' Smith (1912) considered that the 'Middle Sands and Gravels' were formed during the waning of the ice sheet that deposited the 'Lower Boulder Clay', preceding a slight advance that introduced the 'Upper Boulder Clay'. Huddart (1997) considered that this landform and sediment assemblage marked the limit of a readvance of Irish Sea ice on to the coastal lowland after the main Late Devensian glaciation. Eyles and McCabe (1989), however, considered the assemblage to be a glaciomarine morainal bank.

The stratigraphy, mapped by Huddart (1970) is shown in Figure 5.53, and the following units were identified.

8. Raised coastal sediments found only in Selker Bay (Huddart and Tooley, 1972)
7. Kettlehole sediments, including laminated silts, sands and gravels and peats

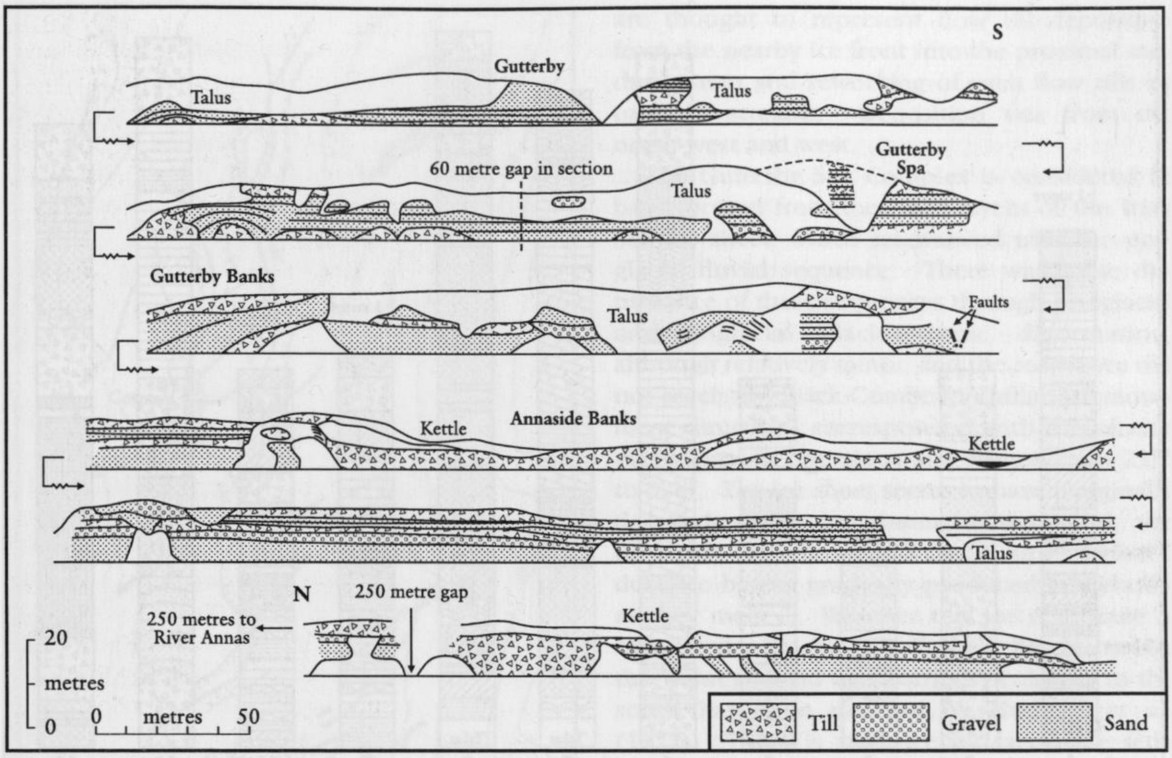


Figure 5.53 The stratigraphy along the Black Combe coast (after Huddart, 1970, 1991).

6. Till: the Coteley Member
5. Sand and gravel: the Coteley Bank Member
4. Till: the Gutterby Spa Member
3. Gravel: the Annaside Member
2. Sand: the Annaside Member
1. Till: the Selker Member (Huddart, 1970; Thomas, 1999).

The basal unit, the Selker Till (unit 1, above), shows 2 m of dark reddish brown, sandy, pebbly till with a sand/silt matrix ratio of 0.55 (Huddart, 1971a), much higher than the values obtained from the till units higher in the stratigraphical sequence. This high ratio, plus the high pebble content, are the reasons for correlating this unit with the Lowca Till at St Bees. The Annaside Member (units 2 and 3, above) displays an upward succession from fine-grained sands, through coarse sands to gravels and this sequence is fairly constant in thickness and height throughout the length of the coastal sections. Log 9 (Figure 5.54) shows a typical coarsening upwards succession. At beach level parallel laminated silt is succeeded by thin units of small-scale, cross-stratified and parallel laminated fine-grained sand and silt and thicker units of horizontally stratified coarse sand. Above are units of large-scale cross-stratification and horizontal stratification succeeded by 2 m of horizontally stratified coarse sand and 3 m of pebble gravel. This is capped by 2 m of red till. Similar glaciofluvial sequences are illustrated in logs 1 and 10, with low-flow-regime indicators predominating. Soft-sediment deformation is characteristic of the low-flow-regime sequences and contorted stratification, load casts, flame structures and sandy clasts within silts indicate the importance of grain size and density contrasts during suspension sedimentation. The thickest sequence of gravels is found at Annaside Bank (Log 8, Figure 5.54), where 12 m occur. The gravels are generally imbricate and the palaeocurrent pattern shows a dominant transport direction from the north-west.

Huddart (1970) grouped units 4, 5 and 6 into the Gutterby Spa Complex. The lowest unit, the Gutterby Spa Member, is composed of a tripartite sequence consisting of till–sandy clay–till, with no obvious break in sedimentation. The individual units vary in thickness in different locations and at the northern end of Annaside Bank the sandy clay is missing and till composes

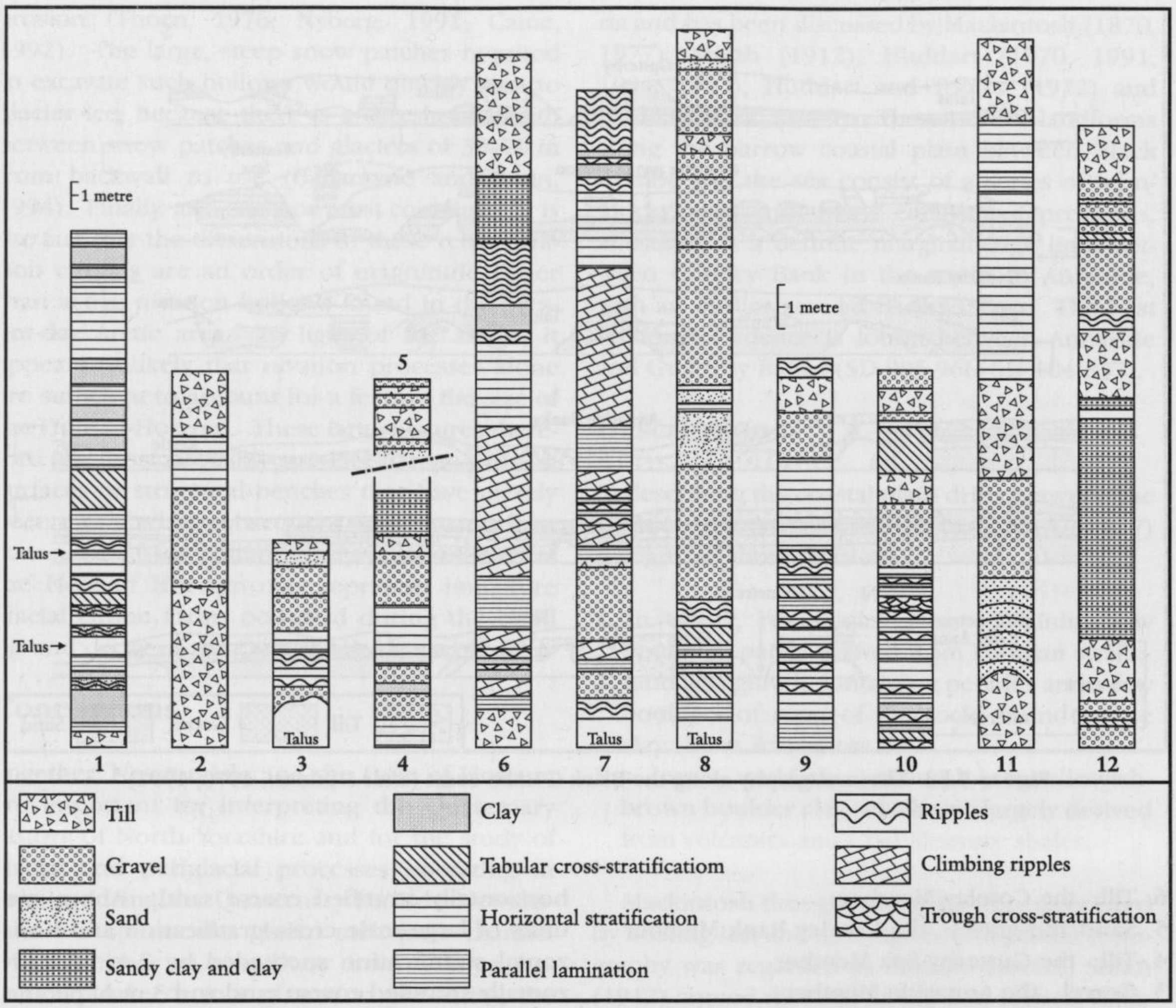


Figure 5.54 Sedimentary logs, Black Combe coast (after Huddart, 1970, 1991).

the whole section. The type site for the sequence is at Gutterby Spa and the succession is shown in log 12 (Figure 5.54). Above granule gravel at the base of the succession is a reddish brown, sandy till unit with a few pebble gravel clasts and many sand inclusions and sandy channels. There is no disturbance at the top of the unit where the till merges upwards into the sandy clay. The sandy clay attains its greatest development at this site, where it is 8.5 m thick and includes a 2.2 m sequence of intermixed sand and silt and parallel laminated sand and clay. Concretions are common and tend to form at the base of clay bands. The upper till in this tripartite succession is reddish brown, very sandy, with few pebble clasts and has comminuted shell fragments (Smith, 1912).

Interpretation

The sediments along this coastal plain have been correlated with those from St Bees, but here they are relatively undeformed. The lowest unit, Selker Till, is interpreted as the basal till of the Late Devensian main glaciation advance and is equivalent to the Lowca Till at St Bees and the Ravenglass Till (Huddart, 1970). During advance, subglacial meltwater activity eroded drainage channels across the bedrock floor and deposited the subglacially engorged eskers on the lower slopes of Corney, Bootle and Little Fells (Smith, 1967) (Figure 5.55). The Irish Sea ice sheet then retreated and readvanced, with pro-glacial, braided rivers producing a thick, widespread sandur sequence. The lower part of

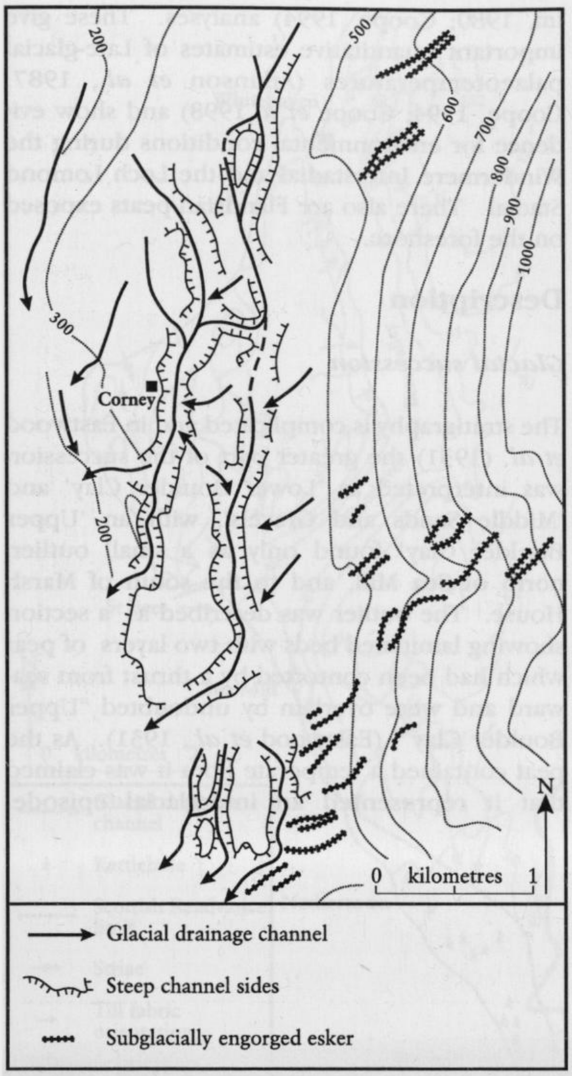


Figure 5.55 Glacial meltwater channels and subglacially engorged eskers on Corney Fell (after Smith, 1967; Huddart and Tooley, 1972).

the Annaside Sands and Gravels indicates deposition in the lower flow regime and probably represents distal sandur and/or marginal glaciolacustrine deposition, with deltas prograding into the dammed lakes. Evidence from dropstones and the general sediment sequences is indicative of such environments, although large-scale foresets, as noted at Holme St Cuthbert, are not evident. That the ice sheet was advancing is indicated by the vertical increase in grain size and the change to upper-flow-regime indicators. Proximal, pro-glacial, longitudinal bar deposition is indicated by the imbricate, pebble gravel units, and the thin till units and clay balls are thought to represent flow till deposition from the nearby ice front into the proximal sandur system and reworking of such flow tills by outwash streams. Deposition was from the north-west and west.

The Gutterby Spa Complex is considered to have formed from the basal layers of the Irish Sea ice sheet, which readvanced over the pro-glacial fluvial sequence. There was some disturbance of these sequences through pro-glacial and englacial glaciotectonic deformation, although relatively minor, and the readvance did not reach the Black Combe foothills. Ice movement directions corresponded with the sandur palaeocurrent directions and ranged from 255° to 328°. The ice sheet seems to have marginally decayed at its maximum extent and the till–sandy clay–till melted out *in situ*. Isolated dead-ice blocks gradually produced kettleholes as they melted. Whether this ice readvance is correlated with the Low Furness Readvance on the Walney Island and Furness peninsula to the south has been debated by Huddart *et al.* (1977). There is evidence for lacustrine sedimentation in the Whicham Valley to the south (Bryant *et al.*, 1985; Clark and Smith, 1998) and it seems likely that this same ice sheet blocked off that valley to create a pro-glacial lake.

Conclusions

The stratigraphical succession and landform assemblage along the Black Combe coastal plain indicate a moraine that marks an ice-marginal position of the Scottish Readvance phase of the Irish Sea ice sheet. There is no evidence for glaciomarine morainal banks or glaciomarine mud drape sedimentation as discussed by Eyles and McCabe (1989), but it is possible that this moraine could be linked to either a surge of the Irish Sea ice sheet, or as part of the Heinrich I phase of McCabe *et al.* (1998), but with all the evidence pointing to deposition from the north-west and west and not from the Lake District as he implies.

ST BEES (NX 965 113)

D. Huddart

Introduction

This site in Cumbria is of considerable importance for interpreting Late Devensian glacial events and Late-glacial environmental conditions in north-west England. The major interest is the St Bees moraine (NX 965 113) (Figure 5.56), a series of low ridges and hummocks situated at the south-west end of the fault-guided, Whitehaven–St Bees glacial valley. The moraine extends from Gutterfoot in the north-west to Sea Mill in the south-east (Figure 5.57). Sections in the moraine have been described by Eastwood *et al.* (1931), Huddart (1970, 1972, 1991, 1994), Huddart and Tooley, (1972), Huddart *et al.* (1977), Tooley (1977), Eyles and McCabe (1989), Walden *et al.*, (1994) and Merritt and Auton (1997a, b). The moraine was considered by Huddart (1970, 1991, 1994) to mark the terminal position of the Late Devensian Scottish Readvance episode, by Evans and Arthurton (1973) and Thomas (1985b) as a response to local ice-marginal oscillation during the retreat of the main Devensian glaciation, and by Eyles and McCabe (1989) as a glaciomarine morainic bank.

On its surface the moraine has a series of organic filled kettleholes that have been interpreted using pollen (Walker, 1956, 1966b) and coleopteran (Pearson, 1962; Coope and Joachim, 1980; Coope, 1994) analyses. These give important quantitative estimates of Late-glacial palaeotemperatures (Atkinson *et al.*, 1987; Coope, 1994; Coope *et al*,1998) and show evidence for environmental conditions during the Windermere Interstadial and the Loch Lomond Stadial. There also are Flandrian peats exposed on the foreshore.

Description

Glacial succession

The stratigraphy is complicated and in Eastwood *et al.* (1931) the greater part of the succession was interpreted as 'Lower Boulder Clay' and 'Middle Sands and Gravels', with an 'Upper Boulder Clay' found only as a small outlier, north of Sea Mill, and to the south of Marsh House. The outlier was described as 'a section showing laminated beds with two layers of peat which had been contorted by a thrust from seaward and were overlain by undoubted "Upper Boulder Clay"' (Eastwood *et al.*, 1931). As the peat contained a temperate flora it was claimed that it represented an interglacial episode.

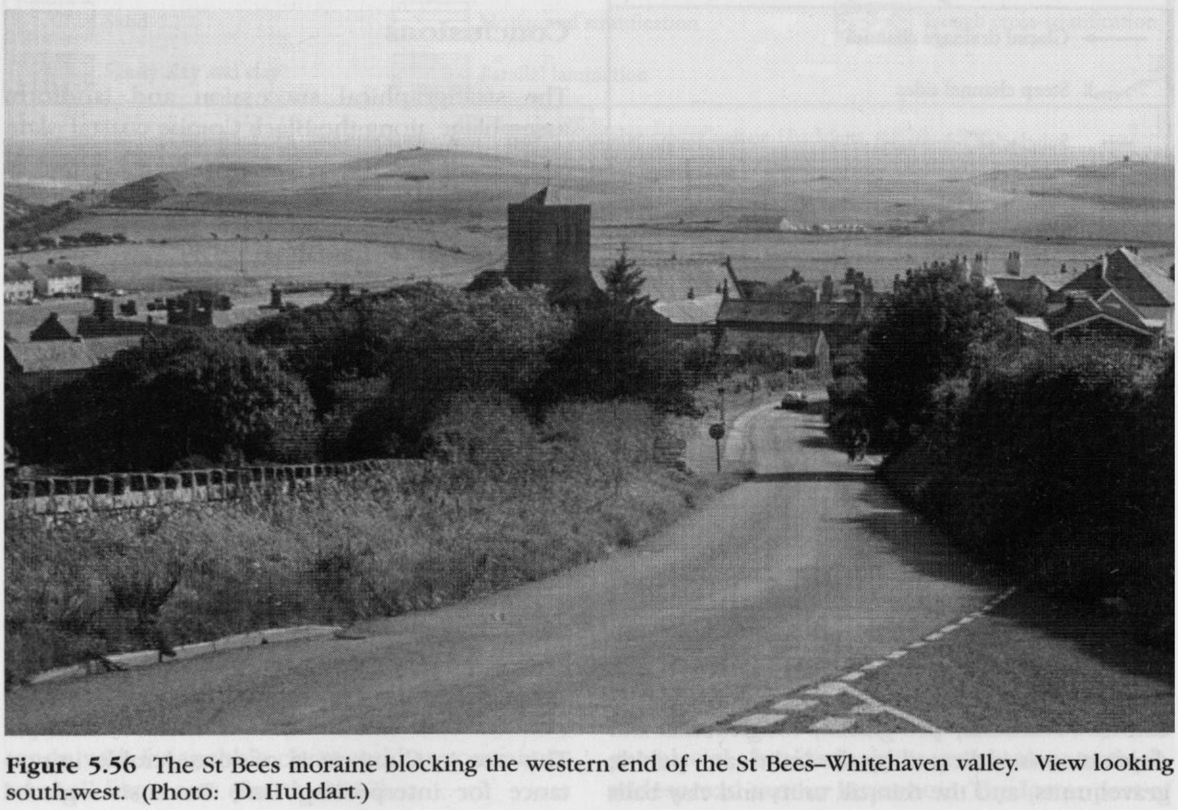


Figure 5.56 The St Bees moraine blocking the western end of the St Bees–Whitehaven valley. View looking south-west. (Photo: D. Huddart.)

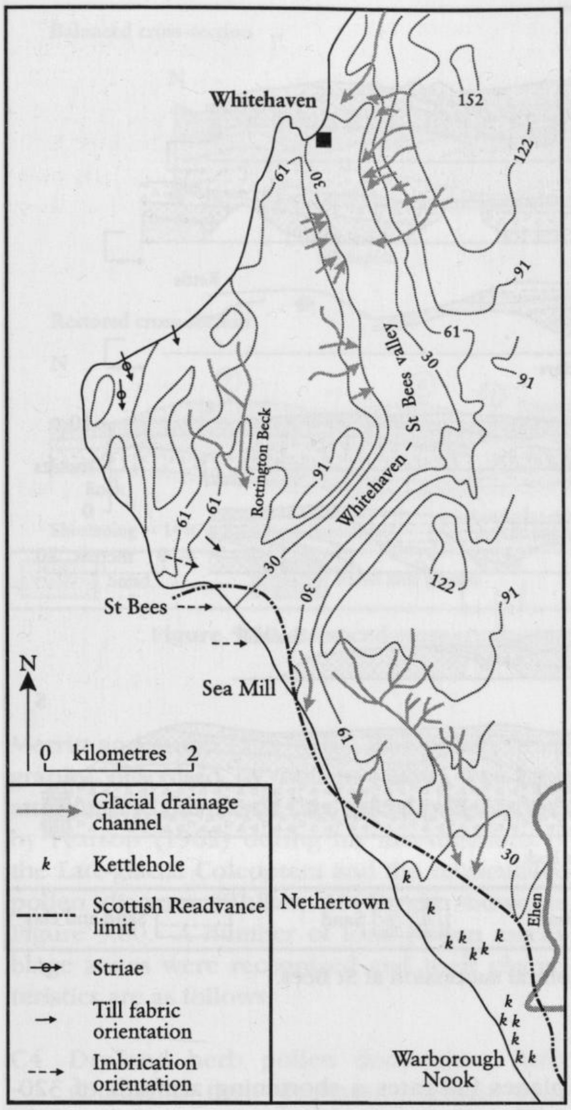


Figure 5.57 Location and morphology of the St Bees area (after Huddart 1994).

Unfortunately this critical section was eroded by the sea in 1924, but in the discussion of Walker's (1956) paper, Hollingworth stated that 'although Chandler's recognition of the temperate character of the flora of the peat-bearing sands between two boulder clays in Eastwood *et al*. (1931) appeared to establish the interglacial age of the deposits … more than a little doubt was felt by some of Smith's colleagues as to whether the "Upper Boulder Clay" was *in situ* and not a reworked deposit.' These doubts seemed justified as Walker (1956, 1966b) and Pearson (1962) showed that the organic deposit was of Devensian Late-glacial, rather than interglacial, age.

The stratigraphy is illustrated in Figure 5.58 and is derived from Huddart (1970, 1994) and Merritt and Auton (1997a, b). The lowest stratigraphical unit, the Lowca Till is between 4 and 10 m thick and is a stiff, pebbly and sandy diamicton with occasional sand lenses. It has a well-developed fabric showing resultant orientations between 257° and 310°. It contains angular to subrounded clasts of the local, Triassic, red sandstone up to 1 m diameter, with some granite, Borrowdale Volcanic lithologies, slate, Ennerdale granophyre, Coal Measure sandstones and mudstones, limestone and ironstone. Its lower boundary is on red sandstone north of the beach and its upper boundary is erosional, overlain by gravel.

The St Bees Silts and Clays are exposed in the core of a large fold and are at least 5 m thick. They consist of repetitions of grey-blue silty clays, brown-red clay and a parallel laminated, yellowish-brown silt or fine sand division. The unit has an overall horizontal bedding and is laminated but there is much disturbance of the original structure, with small faults, brecciated zones and contorted stratification, as well as the large-scale folding. A record from a borehole nearby suggests that this unit lies on the Lowca Till. The upper boundary is gradational with the overlying St Bees Sands and Gravels.

The St Bees Sands and Gravels consist of alternations of parallel laminated fine and medium sand and small-scale, cross-stratified sands, with thin intrastratal contorted stratification units. Towards the top the sands pass upwards into thin, pebble gravel units, alternating with parallel laminated, fine and medium sands and horizontally stratified, coarse sand. These, in turn, pass upwards into gravel. The thickest development of the gravel is at the south end of the main cliff section, where cobble and boulder gravel form the whole succession. These gravels show imbrication in places that suggests a west–east palaeoflow.

The St Bees Till is between 1.5 and 10.0 m thick and is present along most of the main cliff. It is a reddish brown, matrix-supported, sandy clay, relatively pebble-free (Huddart, 1971a). It has included sand lenses and clasts and tends to break up along horizontal, sandy fissility planes. Microfabric analyses show a resultant orientation from 275°. The till grades upwards into a sandy clay, between 2 and 8 m thick and

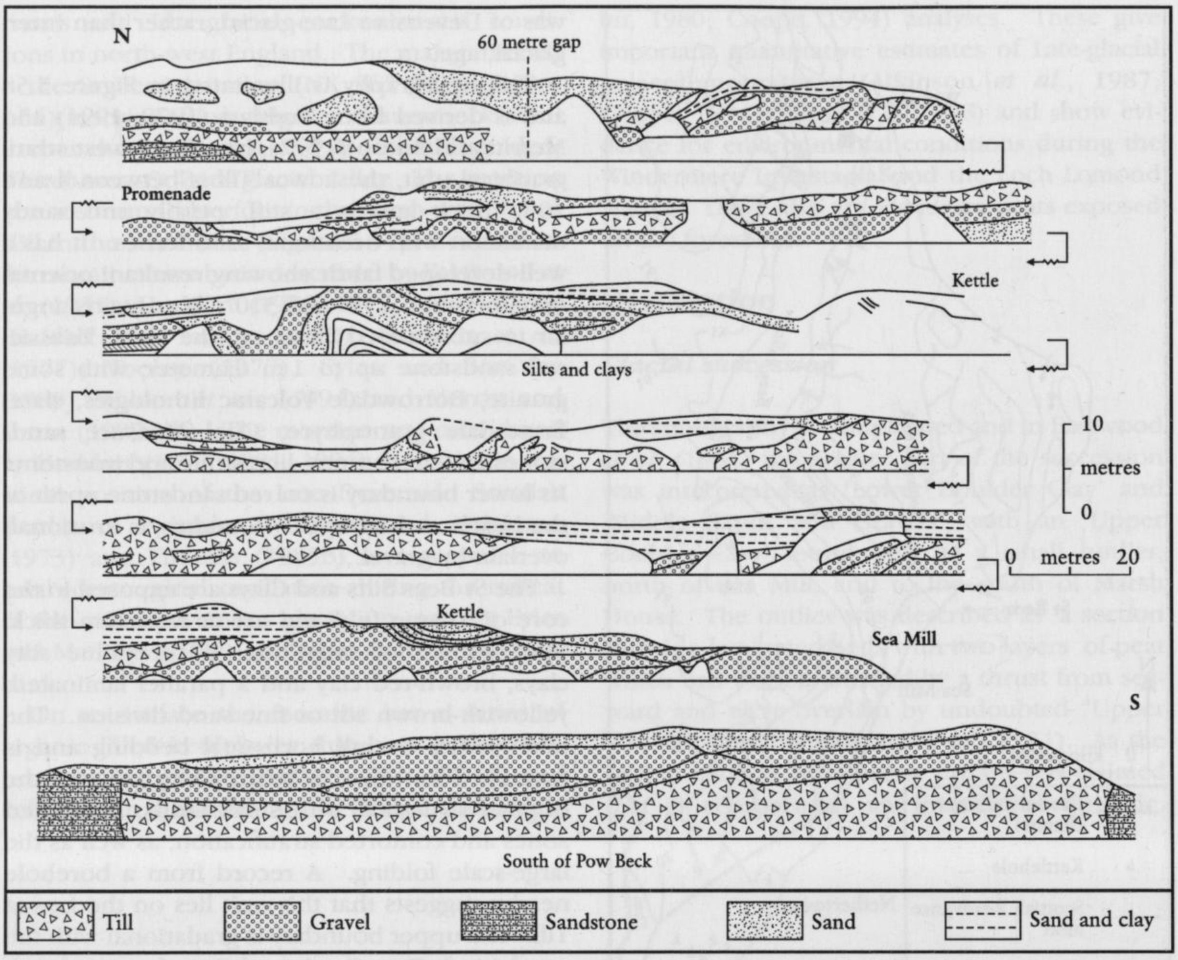


Figure 5.58 The lithostratigraphical succession at St Bees.

gradually becomes pebble-free. The unit is disturbed and displays load casts, contorted stratification, flame and shear structures and faults.

The succession was revised by Merritt and Auton (1997a, b), who identified three units. The Gutterfoot Sand member, equivalent to the sandy clay above the St Bees Till; The How Man Till member, equivalent to the St Bees Till, and the Peckmill Sand Member equivalent to the St Bees Sands and Silts.

The sequence displays considerable large- and small-scale glaciotectonic deformation, especially in the northern part of the section. Deformation structures include conjugate extensional faults, vertical faults, thrust faults and folds (Eaton, 1997). Many structures are thrust-related and show slickensides on bedding surfaces, indicating that they were formed by a flexural slip process resulting from buckling due to compression. Directional data from bedding planes indicates a shortening azimuth of 320–140°, with thrusts showing a general dip to the north-west and a transport direction to the south-east. Seismic refraction data acquired by Nirex (1997b) have allowed a balanced cross-section (Figure 5.59) for the more deformed northern part of the section to be constructed (Eaton, 1997). The shortening is 14%, or approximately 80 m of cumulative thrust transport. It includes two areas of backthrusts to the south and intimately associated with the thrusting process is extensional faulting, resulting in numerous minor, conjugate normal faults.

Late-glacial succession

In kettleholes on the surface of the moraine, sequences display more than 4 m of organic clay, sand, peat and detritus mud. The peat was described as the 'Seacote Peat Member' by

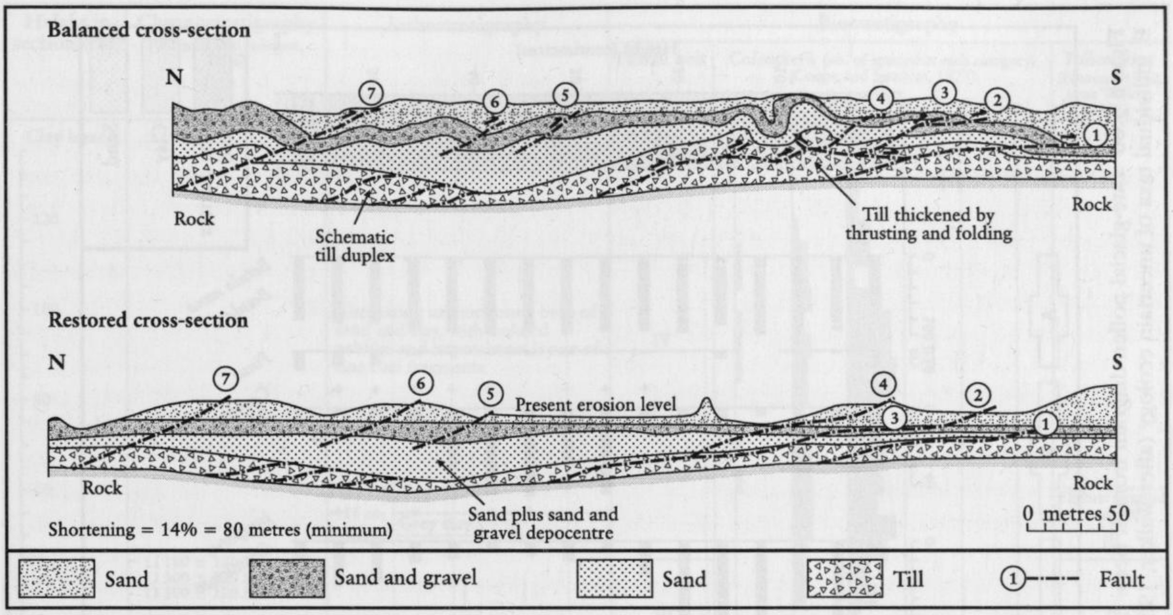


Figure 5.59 Balanced cross-section for the St Bees cliff section (after Eaton, 1997).

Merritt and Auton (1997a, b). The pollen stratigraphy, described by Walker (1956) was later attributed to the classic Late-glacial pollen zones by Pearson (1962) during his investigations of the Late-glacial Coleoptera and the recalculated pollen diagram (Walker, 1966b) is shown in Figure 5.60. A number of local pollen assemblage zones were recognized and their characteristics are as follows.

C4 Dry-land herb pollen dominant. *Salix* abundant, then declines. *Betula* overwhelmingly dominant over *Pinus*. Large but declining frequencies of *Betula nana*. Gramineae more abundant than Cyperaceae and *Artemisia*, Caryophyllaceae frequent throughout.

C5 Pollen of dry-land herbs remains dominant over trees, the frequency of *Betula* increases to 20% of the pollen sum and *Salix* is abundant during the latter half. *Artemisia* and Caryophyllaceae values are low. *Empetrum* present.

C6 *Salix* pollen more abundant and *Juniperus* first recorded. *Helianthemum* consistently present.

C7 Pollen of herbaceous taxa rise in frequency to dominance over trees. *Pinus* frequencies low but continuous. *Helianthemum*, Compositae, Caryophyllaceae and *Empetrum* more frequent than hitherto.

C8 Tree and shrub pollen rare. Frequencies of Gramineae, Caryophyllaceae, Compositae and *Empetrum* are all higher than in C7 and *Artemisia* and *Rumex* are frequent.

In order to investigate the apparent discrepancy between the climatic inferences drawn from the Coleoptera and the classic interpretation based on pollen analysis, the insect fauna was re-examined by Coope and Joachim (1980). The long and complicated coleopteran sequence can be condensed by grouping the species into seven biogeographical categories and the occurrence of species in each of these categories in the stratigraphical succession is shown in Figure 5.61, where the thermally sensitive species occur in an orderly sequence, with the most thermophilous at the base and the cold-adapted species towards the top. Many ^{14}C dates have been obtained from this sequence (Godwin and Willis, 1959; Walker 1966b; Coope and Joachim, 1980), but some are thought to be erroneous and only 4 out of 11 were accepted by the authors.

The Pow Beck Peat Member (Merritt and Auton, 1997a, b) comprises Flandrian peats exposed occasionally on the foreshore at Seamill, as described by Pickering (1878) and Kendall (1881), and to the south-west of the Seacote Hotel, as described by Eastwood *et al.* (1931). A piston-sample of peat at 2.1–2.2 m in

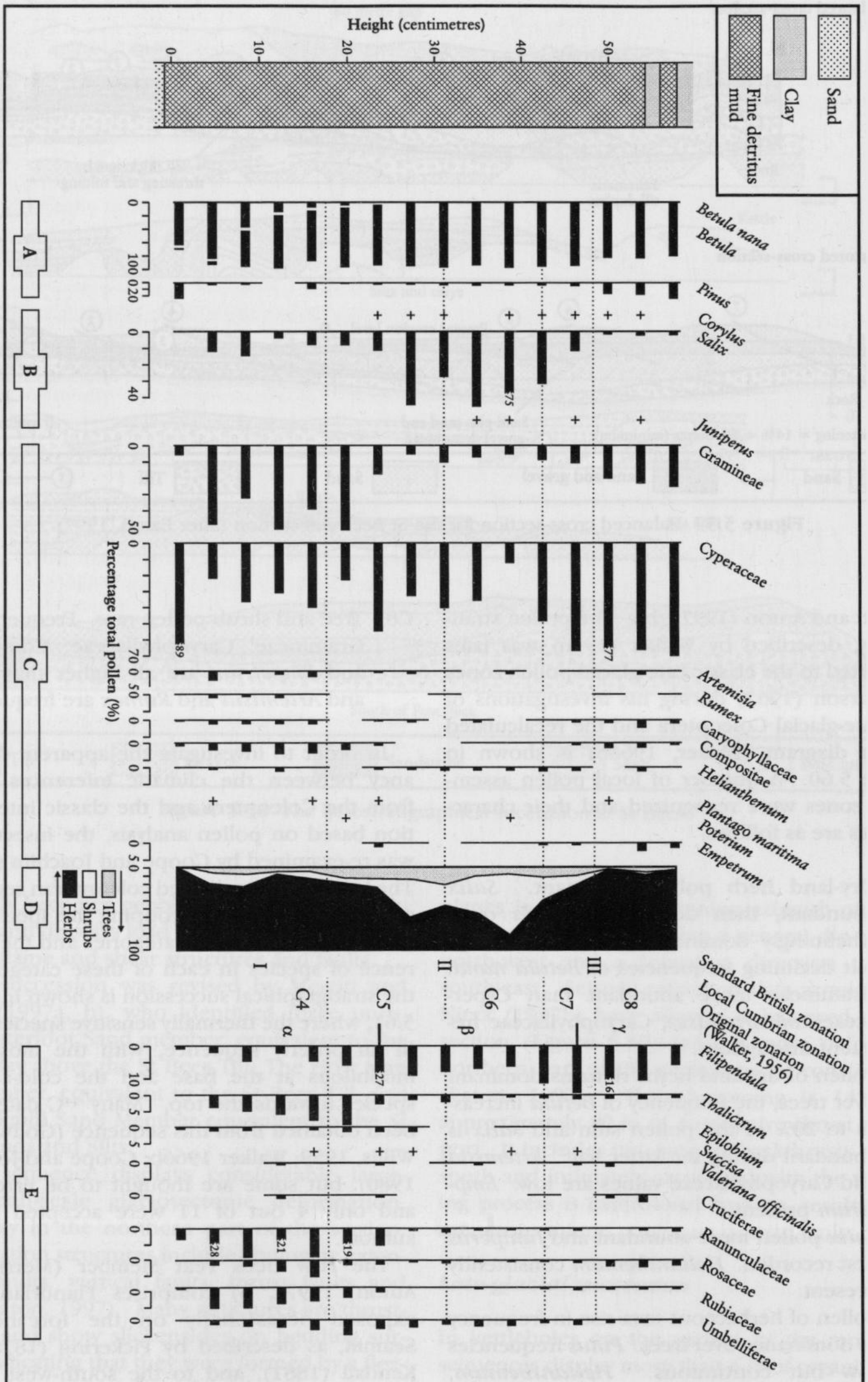


Figure 5.60 Late-glacial pollen diagram from St Bees: A, individual trees; B, individual shrubs; C, individual land herbs; D, summary curves for land flora; E, individual taxa of uncertain ecology (after Walker, 1966b).

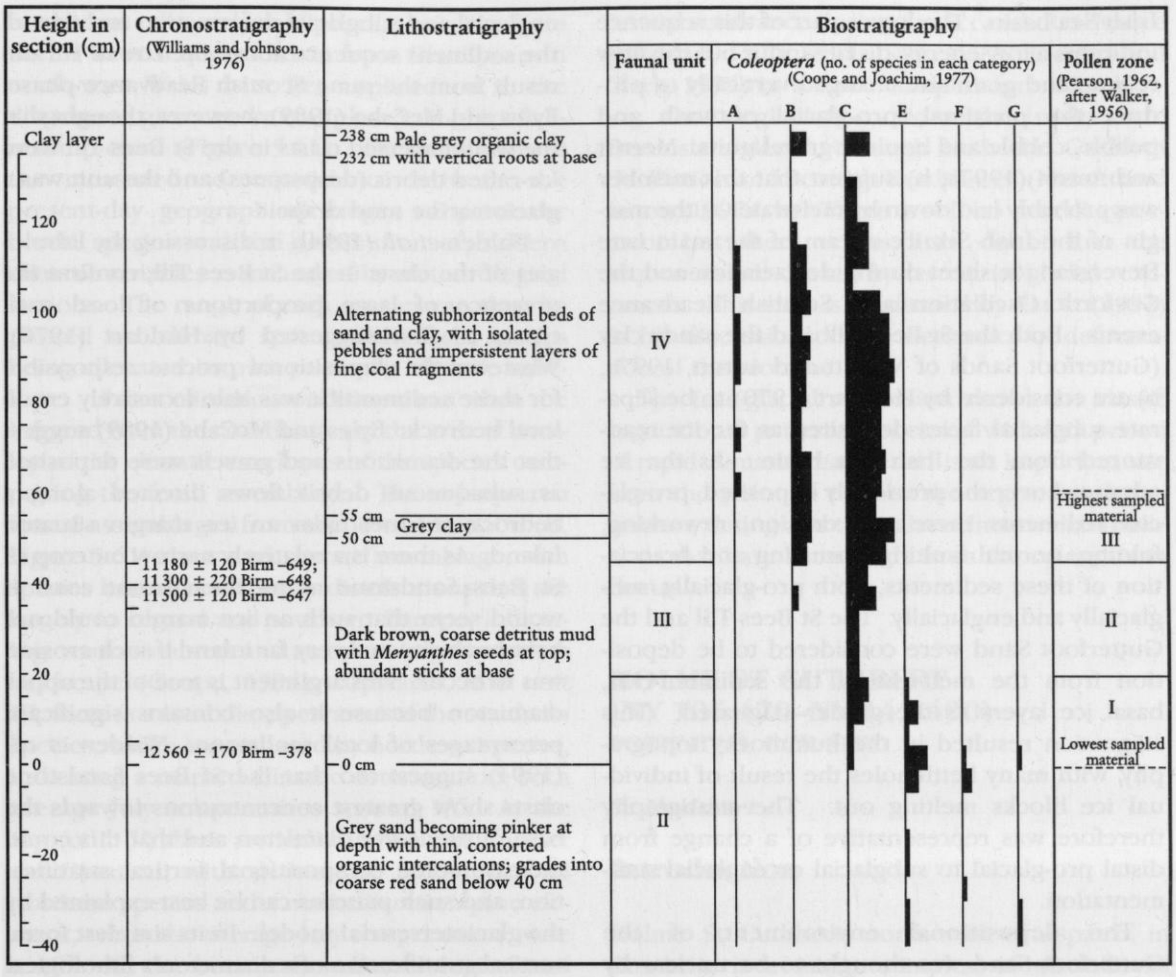


Figure 5.61 Synoptic diagram of chronostratigraphy, lithostratigraphy and coleopteran biostratigraphy at St Bees (after Cope and Joachim, 1980). A, boreal and montane species whose normal range is above the tree-line; B, boreal and montane species whose normal range also includes the upper part of the coniferous forest; C, widespread species whose normal range is north of central Britain; D, cosmopolitan species; E, widespread species whose normal range is south of central Britain; F, southern species whose northern limit of distribution just reaches, or fails to reach, southern England; G, southern European species.

Borehole QBH19 at St Bees gave a ^{14}C date of 7360 years BP, confirming a mid-Flandrian age (Nirex, 1997b).

Interpretation

Glacial succession

The Lowca Till was interpreted by Huddart (1970) as a basal, lodgement till of the main Late Devensian glaciation. This ice was responsible for striae on St Bees Head and the drumlin belt from Edenside to just south of St Bees. During the decay of this ice sheet, meltwater erosion produced the subglacial chutes and channels around St Bees Head and the Whitehaven–St Bees valley. The St Bees Silts and Clays were considered by Huddart (1970) to be glacio-lacustrine deposition in a pro-glacial lake as regional drainage was blocked off by the Scottish Re-advance of Irish Sea ice at the western end of this valley. Merritt and Auton (1997a, b) considered these deposits to be the bottomsets/toesets of a delta formed by meltwater flowing from the valley during deglaciation of the main Late Devensian ice sheet and Eyles and McCabe (1989) considered them part of a glaciomarine sequence. However, there is no fauna in these deposits to substantiate a glaciomarine origin.

The St Bees Sands and Gravels were considered by Huddart (1970) to represent a gradual ice advance towards the St Bees area from the

Irish Sea basin. The lower part of this sequence indicates a low energy, distal sandur but the flow regime and grain size changed vertically to produce the proximal, pro-glacial outwash and pebble, cobble and boulder gravel units. Merritt and Auton (1997a, b) suggest that this member was probably laid down by meltwater at the margin of the Irish Sea ice-stream of the main Late Devensian ice sheet during deglaciation and the Gosforth Oscillation and Scottish Readvance events. Both the St Bees Till and the sandy clay (Gutterfoot Sands of Merritt and Auton, 1997a, b) are considered by Huddart (1970) to be separate subglacial facies deposited as the ice readvanced from the Irish Sea basin. As the ice advanced over the previously deposited, pro-glacial sediments there was erosion, reworking, folding, normal faulting, thrusting and brecciation of these sediments, both pro-glacially, subglacially and englacially. The St Bees Till and the Gutterfoot Sand were considered to be deposition from the melt-out of the sediment-rich, basal ice layers as the glacier stagnated. This stagnation resulted in the hummocky topography, with many kettleholes the result of individual ice blocks melting out. The stratigraphy therefore was representative of a change from distal pro-glacial to subglacial or englacial sedimentation.

The depositional environment of the Gutterfoot Sand was thought to be unclear by Merritt and Auton (1997a, b) as the primary sedimentary structures had been destroyed by glaciotectonism. However, they did consider that deposition possibly occurred shortly before the Scottish Readvance in pro-glacial lakes. It is possible that the terminal moraine marks one of the limits of McCabe *et al.*'s (1998) Heinrich I events in the northern Irish Sea basin and/or represents a surge position of the ice sheet. Merritt and Auton (1997a, b) considered that it was possible that the bed-parallel shearing in the deposits resulted from subglacial deformation, but if this was so they considered that the overriding must have occurred before the Scottish Readvance because they considered that the whole sequence was pro-glacially tectonized during the Scottish Readvance. They considered the St Bees Till to be a deformation till formed mainly of sediment derived from the Irish Sea basin and redeposited beneath the Irish Sea ice-stream during the Gosforth Oscillation. Huddart (1997) nevertheless considers that this pro-glacial deformation phase can be followed by englacial and subglacial deformation and that all the sediment sequence above the Lowca Till can result from the same Scottish Readvance phase. Eyles and McCabe (1989), however, thought that the well-dispersed clasts in the St Bees Till were ice-rafted debris (dropstones) and the unit was a glaciomarine mud drape.

Walden *et al.* (1994), in discussing the lithologies of the clasts in the St Bees Till, confirm the presence of large proportions of local rock types, as first suggested by Huddart (1970). Whatever the depositional process responsible for these sediments it was able to actively erode local bedrock. Eyles and McCabe (1989) suggest that the diamictons and gravels were deposited as subaqueous debris flows directed along a bedrock channel from an ice margin situated inland. As there is a relatively narrow outcrop of St Bees Sandstone along the current coast it would seem that such an ice margin could not have been situated very far inland if such erosion was to occur. This argument is true of the upper diamicton because it also contains significant percentages of local sandstone. Walden *et al.* (1994) suggest too that the St Bees Sandstone clasts show greatest concentrations towards the base of the lower diamicton and that this could indicate some compositional vertical stratification, and such patterns can be best explained by the glacioterrestrial model. In its simplest form, vertical stratification of a diamicton's lithological composition could be expected in response to variations in the sequence of up-ice outcrop geology, and it is less easy to argue how compositional stratification could be achieved in subaqueous debris flows.

The How Mill Till was considered by Merritt and Auton (1997a, b) to be laid down during the Scottish Readvance by the Irish Sea ice-stream, but it was considered to be part of the St Bees Till by Huddart. The Peckmill Sands were considered by Merritt and Auton (1997a, b) to have accumulated as debris flows and sheetwash derived from the Irish Sea ice-stream during its retreat following the Scottish Readvance, and it has been identified along the coast between St Bees and Ravenglass.

Late-glacial succession

The Late-glacial succession has allowed Coope (1994) to quantify the thermal climate in terms of the mean July temperature using the Mutual Climatic Range method (Atkinson *et al.*, 1987), but because the early cold phase is missing from the St Bees sequence, data from Glanllynau in north-west Wales have been spliced on to the diagram to give a more complete Late-glacial climatic picture (Figure 5.61). The method uses the measured correspondence between the present-day geographical ranges of particular beetle species and the distribution of modern climatic variables as measured by meteorological stations. The range of each species can then be plotted on climate space, with each species being allocated its own particular climatic envelope. Palaeoclimatic conditions can then be inferred from the coordinates of the area of overlap of the climatic envelopes of the species comprising the fossil assemblage. The coleopteran faunal assemblages from the lowest horizon (II, Figure 5.61) show a temperate assemblage, with species having present-day distributions that are largely in central and southern Europe, indicating warm summers and an open, largely treeless landscape (Coope, 1994). Upwards, into unit III, there is a total disappearance of the southern species, which are replaced by northern species, indicating a marked climatic deterioration, with average July temperatures around 13°. At the top (unit IV) there are indications of still further deterioration, with an increase in numbers of northern species and a marked arctic/alpine element, when the average July temperatures were *c.* 10°. All the botanical, sedimentological and oceanic evidence points to a climate of glacial severity and in a short time, as the great thickness of these sediments at St Bees testifies to the importance of active periglacial slope processes that filled the sedimentary basin. It was noted initially at St Bees that the rise of the tree birch pollen curve is correlated with the first major fall in temperature after the thermal maximum of the Late-glacial Interstadial (Coope and Joachim, 1980), but subsequently this has been found to be a widespread phenomenon in Britain.

Conclusions

St Bees is an important Devensian site in northern England as it has prompted much controversy regarding its stratigraphical interpretation and the consequent interpretation of Late Devensian glacial retreat conditions. These interpretations range from glaciomarine (Eyles and McCabe, 1989), minor terrestrial readvance during the main Devensian retreat stage (Evans and Arthurton, 1973; Thomas 1985b), major terrestrial readvance during the Scottish Readvance (Huddart, 1970, 1991, 1994, 1997; Huddart and Clark, 1994) to a more complex formation during the deglaciation from the main Late Devensian glacial, through both the Gosforth Oscillation and Scottish Readvances (Merritt and Auton, 1997a, b). There is still no overall agreement but there seems little doubt that the sequences are caused by terrestrial glacial readvances. This view is confirmed by Eaton (1997) in his discussion of the mechanics of deformation of the moraine.

The Late-glacial succession shows the classic vegetational patterns associated with this period of rapid climatic change and is an important site from the point of view of the vegetational succession, dating and the palaeoclimatic extrapolations that have been made from the coleopteran faunas.

HOLME ST CUTHBERT (NY 105 470–NY 145 490)

POTENTIAL GCR SITE

D. Huddart

Introduction

Holme St Cuthbert, north-west of Aspatria in Cumbria, is a major complex of glaciofluvial landforms at a height of 30–48 m OD. It has been interpreted as an esker-fed deltaic complex deposited into a pro-glacial lake ponded between an ice front trending NE–SW and the drumlin country to the east. It has been suggested as marking one of the terminal positions of the Late Devensian Scottish Readvance ice sheet, which advanced on to the Cumbrian lowland (Huddart, 1970, 1991, 1993, 1994; 1997; Huddart and Tooley, 1972; Huddart *et al.*, 1977; Huddart and Clark, 1994). However, it is a controversial location because it also has been interpreted as marking a glaciomarine morainal bank deposited from Lake District ice into a high sea level during deglaciation of the Irish Sea ice sheet (Eyles and McCabe, 1989).

Description

The landforms around Holme St Cuthbert are shown in Figure 5.62 and several well-defined morphological zones can be recognized. In the south-east is a series of NE–SW-trending drum-

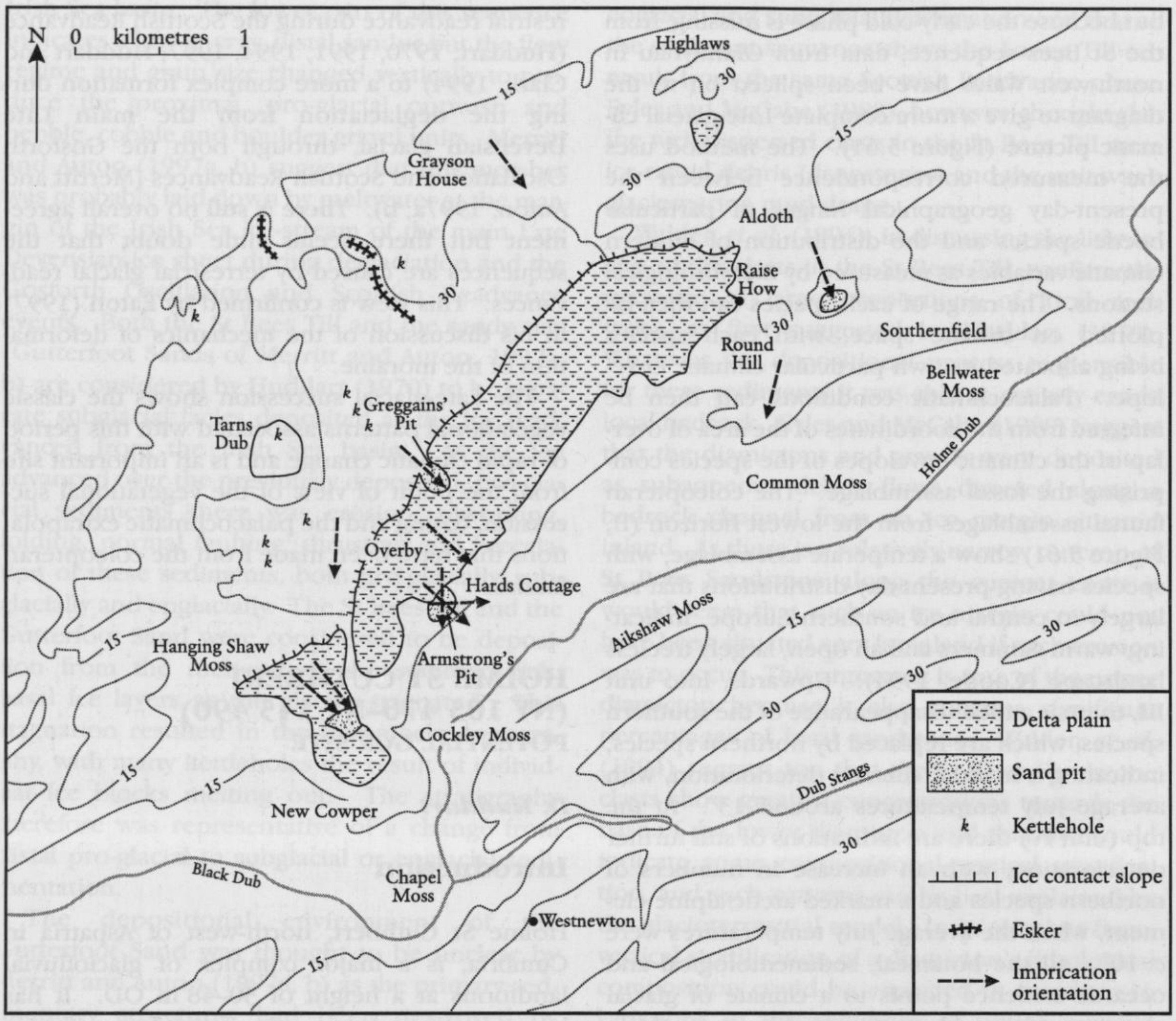


Figure 5.62 Pro-glacial lacustrine sediments and landforms in the Holme St Cuthbert area (after Huddart, 1991).

lins. Immediately adjacent to, and surrounding these drumlins north-east of Westnewton are the low-lying, poorly drained Bellwater, Common, Cockley and Chapel Mosses. Overlooking this area of low relief is a steep, east-facing scarp that runs for 4.8 km between Round Hill in the north to New Cowper in the south. This scarp forms the eastern extremity of a flat-topped surface that extends for 0.8 km and rises from 42.7 to 48.8 m OD in an east–west direction, and which is bounded to the west by a scarp slope running from Raise How to Hangingshaw Moss. Both scarps are crenellated and lobed. Around New Cowper there is evidence for a lower level between 30.5 and 36.6 m OD. West of the flat-topped zone is an area containing many undrained depressions and rising above the general surface are several NW–SE-trending linear ridges.

The evidence for the pro-glacial lake hypothesis for the origin of this complex was obtained from several sand and gravel pits. The sedimentary facies can be divided into topsets and foresets, both associated with a fluvial–deltaic system. The topset facies were best represented in Greggains' Pit (Figure 5.63) by 12–15 m of pebble gravel and pebbly, coarse sand, although there is also medium-grained sand infilling channels. The transverse-to-current section reveals large erosional channels, imbricate, gravel bars and occasional disturbed stratification. Palaeocurrent data indicate a constant transport direction from the north-west. The grain size increases from the stratigraphical base to the top of the

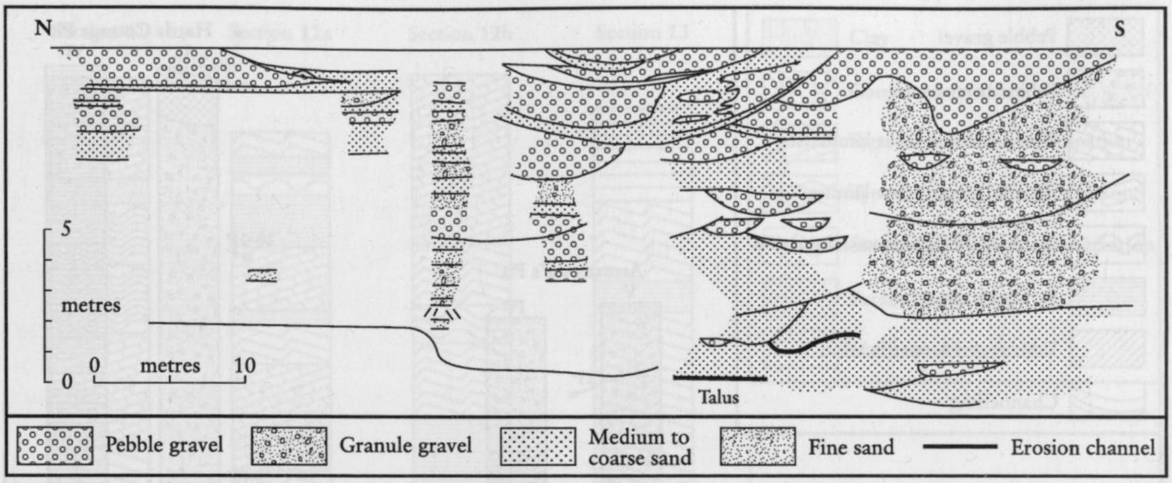


Figure 5.63 Topset facies, Greggains' Pit, Holme St Cuthbert (after Huddart, 1972).

pit. The lowest topset sediments consist of horizontally stratified, coarse and medium sands, with occasional small, erosional scours which pass up vertically into pebbly, coarse sand, with occasional pebble gravel channels and finally into imbricate, pebble gravel units and pebble gravel channels. The individual channel fills vary from solely pebble gravel to a gradation from a basal pebble gravel, through coarse- and medium-grained sand to rippled, fine-grained sand.

The foreset facies is preserved both below the topsets in Greggains' Pit and laterally to the east in Armstrong's Pit. The facies can be divided into foresets proper, which are large-scale, cross-stratified sets, dipping at between 14 and 30°, through toesets to bottomsets, which dip at 2–5°. There is a continuum from foresets to bottomsets deposits. These beds are characterized by:

1. the individual beds decreasing in angle in the down-dip sedimentary transport direction from 21° nearest the sediment supply to 2–3° in the distal bottomsets;
2. the sediment grain size generally decreasing in the down-dip direction from pebbly, coarse sands in the proximal zone to silty sand and silts in the distal zone;
3. a total foreset height at about 30 m;
4. a predominance of ripple drift, small-scale, cross-lamination.

The best development of foresets proper is located in Armstrong's and Hards Cottage Pits (Figure 5.64). Palaeocurrent data from these pebbly foresets range from 89 to 174°. In the Hards Cottage Pit the bottomsets are dominated by parallel laminated fine sands and silts, with subsidiary sinusoidal and type 'b' ripple cross-lamination and occasional sets of solitary, tabular cross-stratification. There is a lack of type 'a' ripple cross-lamination, which indicates that the depositional environment was dominated mainly by suspension, with occasional phases of solitary sand-wave movement across the bottomset surface. In Armstrong's Pit the dominant stratification type is type 'a' ripple cross-lamination, which indicates that bedload transport was more important in this part of the delta.

Figure 5.65 shows log profiles in bottomsets from Greggains' Pit and reveals cycles of both increasing and decreasing bedload movement relative to suspension fall-out in the changes in sedimentary structure. These cycles end with parallel lamination and/or clay, which indicates only suspension sedimentation. If these represent a temporal cycle and the clay bands possibly indicate a winter period of no sediment transport, then there are records of at least nine years sedimentation in logs 11 and 12. If an annual cycle is assumed then an average of 1.4 m, with a range between 0.5 and 4.2 m, was deposited per year.

The lithological composition of the pebble gravels is dominated by high percentages of Criffel granite and other southern Scottish erratics, such as greywackes and Lower Calciferous Sandstone conglomerate from the Kircudbrightshire coast (Huddart, 1970; Huddart and Tooley, 1972).

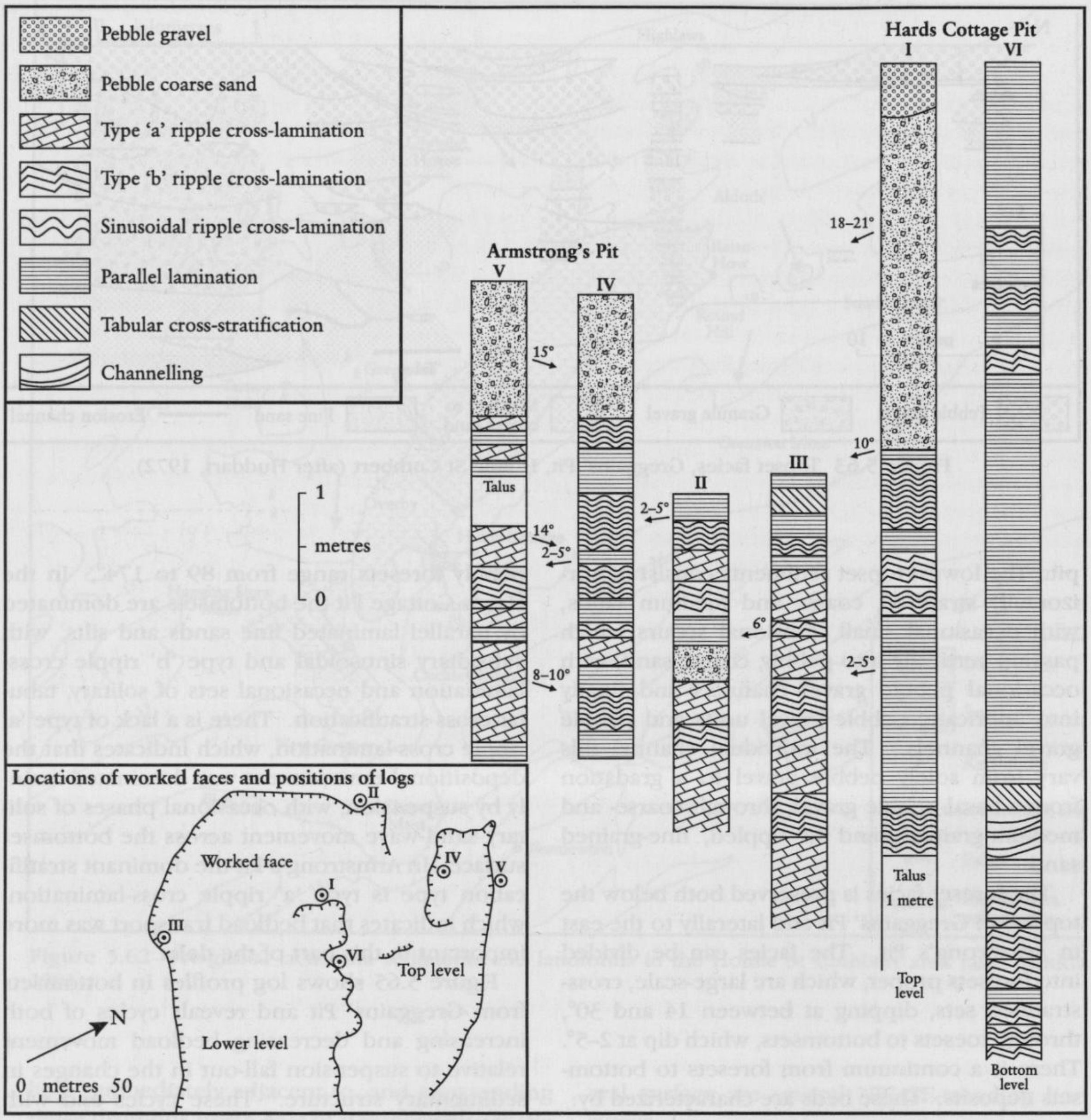


Figure 5.64 Foreset and bottomset facies, Armstrong's Pit and Hards Cottage Pit (after Huddart, 1972).

Interpretation

These sediments and landforms are thought to represent glaciofluvial sedimentation prograding into a pro-glacial lake formed in a depression between the drumlin belt to the east and ice to the west and south. Within the foreset and bottomset zones in the delta complex the following processes operated. Firstly, gravitational sliding on the foreset slope to give proximal, pebbly coarse sands and secondly, pulsatory density underflows caused by the high sediment concentration that flowed along the lower delta slopes. These deposited the various types of ripple drift, from small-scale current ripple trains in combination with high suspension rain-out and the isolated, large-scale cross-stratified sets from sand waves. Occasional scouring was associated with higher flow speeds at the current head. Some underflows may have been triggered by slumping on the foreset slope, but most were initiated by the high sediment concentration.

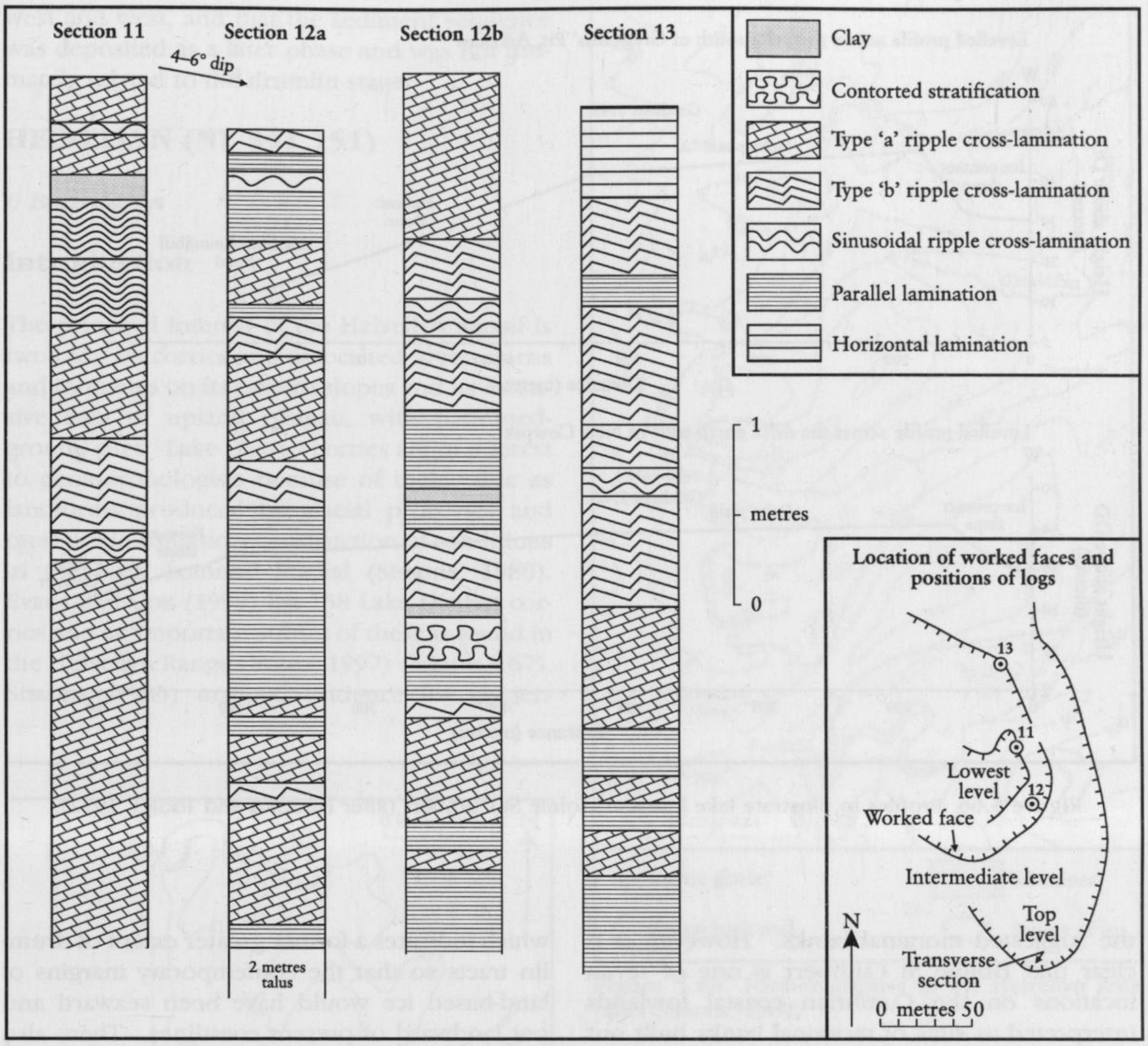


Figure 5.65 Bottomset facies, Greggains' Pit (after Huddart 1972).

Thirdly, there was continuous sedimentation from suspension, which resulted in parallel laminated sands and silts and which also contributed to the formation of the ripple-drift sequences.

Two lake surfaces have been levelled at 42.7 m and *c.* 30.5 m OD, based on the transition from topset to foresets. The later, lower lake stage is not extensive and has been located only in the southern part of the delta complex at New Cowper. Profiles shown in Figure 5.66 depict the two main stages, which overflowed in the initial lake stage through a channel east of Westnewton and probably at the later stage through the Black Dub. The ice front that held up this pro-glacial lake clearly post-dated the main Late Devensian glaciation, and the ice-marginal position of the later readvance of Scottish ice, based on the palaeocurrent evidence and the lithologies, has been correlated with ice-marginal positions in the Carlisle Plain, Harrington, St Bees and Annaside–Gutterby Banks (Huddart, 1970 1991, 1994; Huddart and Tooley, 1972; Huddart *et al.*, 1977).

The Holme St Cuthbert deposit is probably one of the three sites marked near the Solway shore by Eyles and McCabe (1989), from the Risehow area to near Carlisle, but the exact locations are not given in their text. Nor does their text discuss alignments, altitudes and limits of

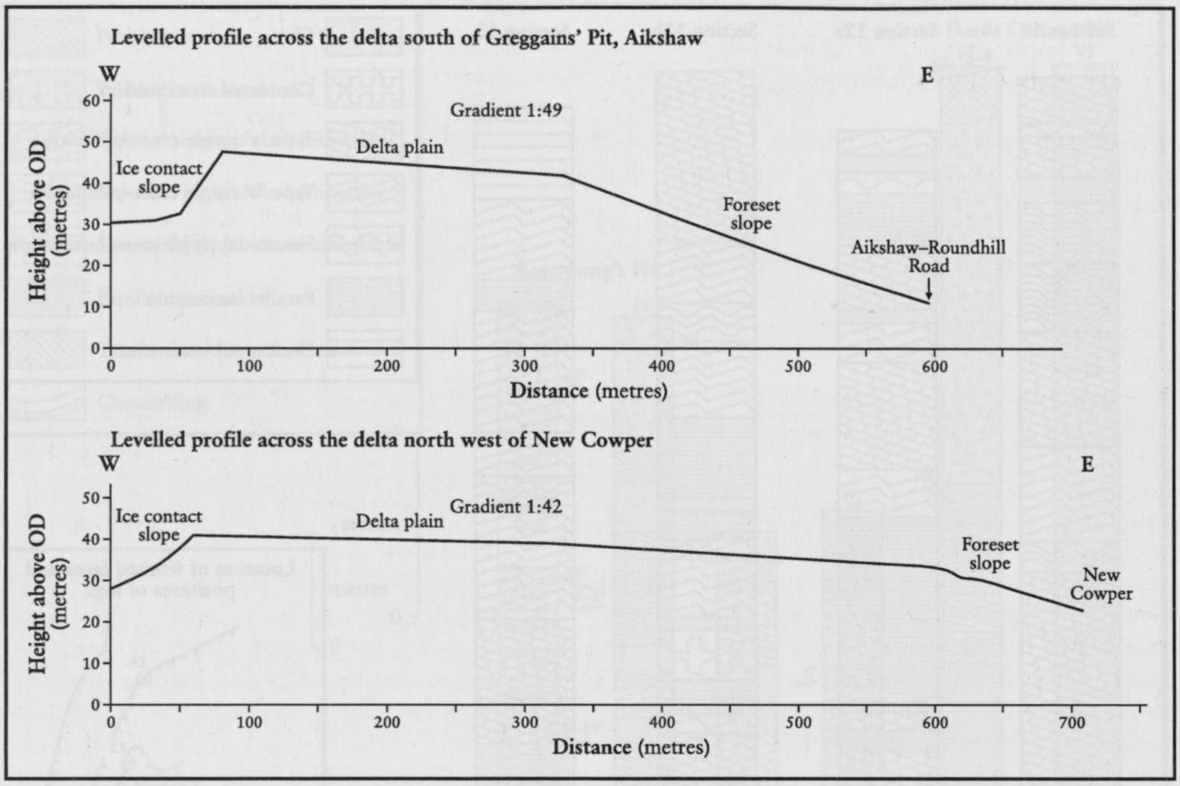


Figure 5.66 Profiles to illustrate lake levels at Holme St Cuthbert (after Huddart and Tooley, 1972).

the suggested morainal banks. However, it is clear that Holme St Cuthbert is one of seven locations on the Cumbrian coastal lowlands interpreted as sites of morainal banks built out as tidewater sediment accumulations at the retreating margins of glaciers (e.g. Powell, 1981; Huddart and Peacock, 1989). Each complex may represent a temporary halt of the margin because they are typically associated with bedrock highs and relatively shallow water (Eyles and McCabe, 1989). The Holme St Cuthbert deposit supposedly marks the middle morainal bank in the Solway lowlands marking a temporarily stable marine terminus of the inland ice. Eyles and McCabe (1989) thought that such banks were related closely to the presence of drumlins in adjacent areas behind (i.e. upstream of) the banks. They presented no data for the sedimentology of this deposit, nor were relationships in Cumbria between morainal banks, ice streams and drumlin formation discussed. It should be noted that on the coasts south from Mawbray and on the Furness peninsula, drumlins have been been eroded partially or wholly, which indicates a former greater extent of drumlin tracts so that the contemporary margins of land-based ice would have been seaward and not landward of present coastlines. There also are no marine macro- or microfossils, as are often found in unequivocal morainal banks (Huddart and Peacock, 1989). Ice marginal positions demonstrated here also have been ignored by McCabe *et al.* (1998) in their discussion of the Heinrich I event in the northern Irish Sea basin.

Conclusions

The sedimentary sequences and landforms at Holme St Cuthbert have been interpreted as esker-fed, ice-contact 'Gilbert-type' deltas into a pro-glacial lake from Scottish ice to the west and north-west. This ice ponded up water to the east in the Holme Dub valley to a depth of approximately 30 m. The conclusion from this evidence is that the sedimentary environment was not a morainal bank but a pro-glacial lake, that ice was not present to the east (inland) but to the north-west and west, and that the sediment sequence was deposited as a later phase and was not intimately related to the drumlin stage.

HELVELLYN (NY 341 151)

J. Boardman

Introduction

The principal interest of the Helvellyn massif is twofold: its corries and associated arêtes, tarns and moraines on its eastern slopes and its extensive area of upland plateau, with patterned-ground sites. Lake District corries are of interest to geomorphologists because of their value as landforms produced by glacial processes and their use in climatic reconstruction of conditions in the Loch Lomond Stadial (Sissons, 1980). Evans and Cox (1995) list 158 Lake District corries and an important subset of these is found in the Helvellyn Range (Evans, 1997) (Figure 5.67). Sissons (1980) mapped evidence for glaciers during the Loch Lomond Stadial and his reconstruction for the Helvellyn corries is shown in Figure 5.68.

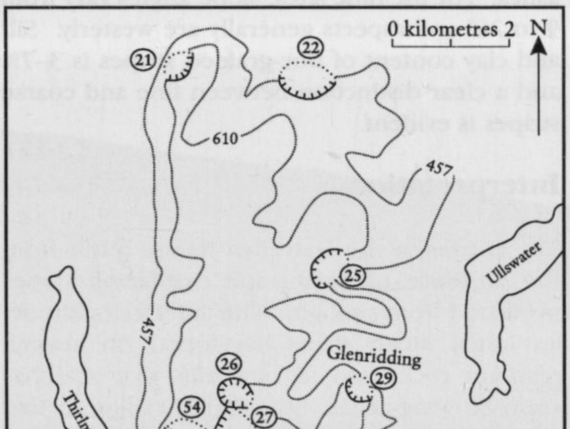


Figure 5.67 Corries of the Helvellyn and Fairfield Ranges (after Evans, 1997).

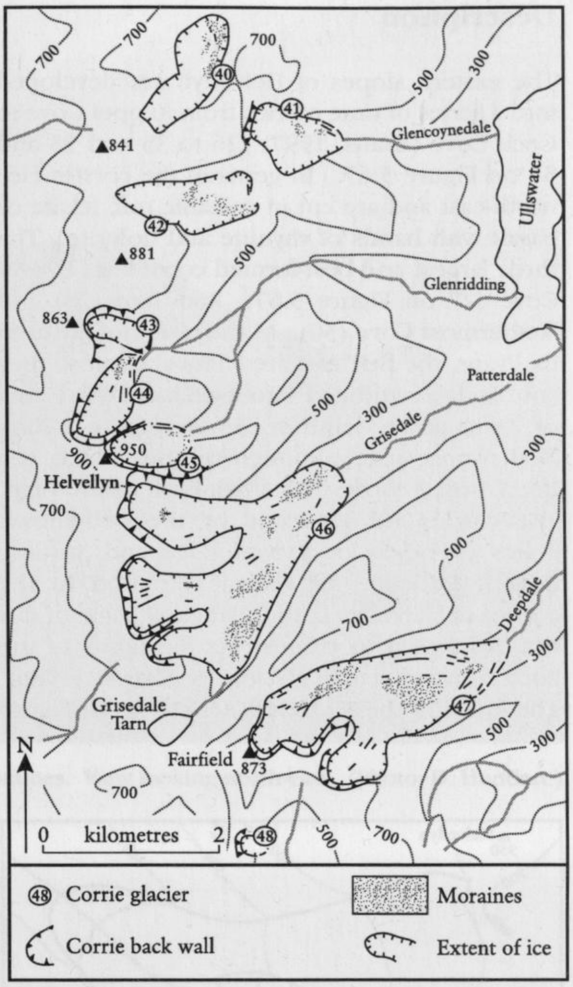


Figure 5.68 Former glaciers in the Helvellyn area (after Sissons, 1980).

Patterned ground of the type generally associated with former periglacial conditions can be found on many high flat summits of the Lake District, such as Grasmoor, Helvellyn and Skiddaw. The Helvellyn patterned-ground sites differ from others in this review in that they are developed on Borrowdale Volcanic Group lithologies. Patterned ground, forming under contemporary conditions, was reported from the Lake District by Hollingworth (1934) and active periglacial phenomena consisting of small sorted stripes, nets or polygons and isolated sorted circles are described by Warburton (1985). They generally occur above 610 m, where vegetation is sparse, and on all the main lithologies.

Description

The eastern slopes of Helvellyn are developed into a series of nine corries from Keppel Cove to Cock Cove (Evans, 1997) (26 to 33 and 53 and 54 on Figure 5.67. In general, the corries face north-east and are cut in volcanic tuff, felsite or basalt with bands of rhyolite and dolerite. The three largest and best-formed corries are Brown Cove (27 on Figure 5.67), Red Tarn (28) and Nethermost Cove (30). In the terminology used by Evans, the first two are 'classic' because they contain lakes and Red Tarn itself has a water area of 9.8 ha and a depth of 25.6 m (Figure 5.70a). Nethermost Cove, although without a lake, is a deep, broad corrie with a headwall 280 m high. The corries are separated by the well-known arêtes or ridges of Swirral Edge and Striding Edge. Evidence for recent glaciation in the corries of Helvellyn is abundant and some of this has been used to reconstruct the limits of the Loch Lomond Stadial glaciers by Sissons (1980). The limits of these glaciers are shown in Figure 5.68 and a reconstruction of the former limits of the Grisedale glacier in Figure 5.69.

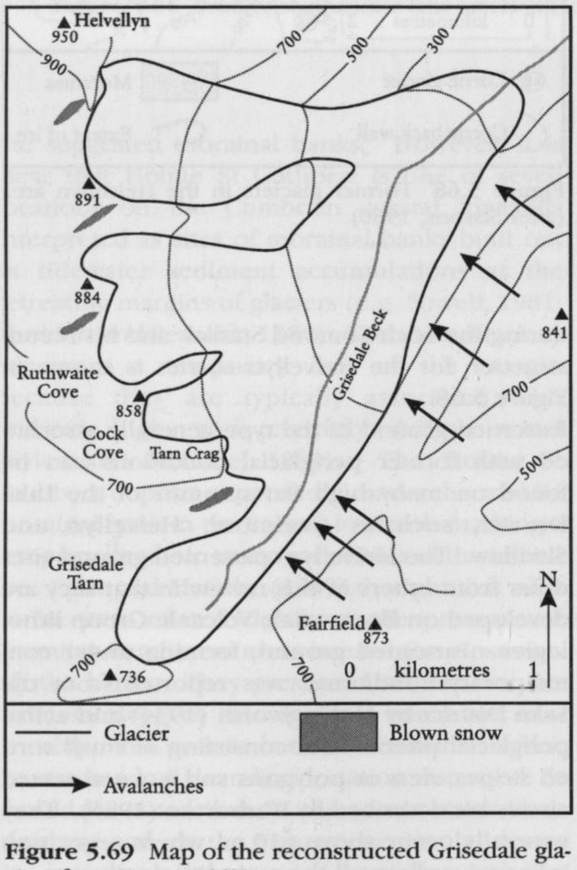


Figure 5.69 Map of the reconstructed Grisedale glacier (from Evans, 1997).

Generalized descriptions of a range of periglacial phenomena, both active and relict, present on the Helvellyn summit plateau and slopes are given by Hay (1936, 1937). He is not specific with regard to locations but describes the phenomena as:

1. stone stripes
2. stone polygons; on Raise in the Helvellyn Range
3. disintegrated vegetation due to wind
4. solifluction terraces (Figure 5.70b)
5. frost-degraded turf patterns
6. gliding blocks
7. soliflucted scree (Figure 5.70c)
8. summit blockfields and vertical stones (Figure 5.70d).

Warburton (1985) describes nine patterned-ground sites on the Helvellyn plateau. These range from three on Raise in the north to several near the summit of Helvellyn itself. All are underlain by Borrowdale Volcanic lavas and ashes. For the nine sites, slope angles vary from 9 to 20° and aspects generally are westerly. Silt and clay content of fine-grained stripes is 3–7% and a clear distinction between fine and coarse stripes is evident.

Interpretation

The corries in the Helvellyn Range result from the presence of north- and east-facing slopes protected from sunlight, with sufficient altitude, an initial steep slope developed on strong, resistant rocks and flat summits that acted as snow-blowing areas. Evans (1997) suggests that the palaeoglaciation level in the Helvellyn Range was 700 to 800 m but that factors other than altitude played a significant part in where corries developed. The development of corries represents glaciation of a restricted character, but which presumably was repeated many times during the Quaternary Period. It is likely that at the beginning and end of major glacial events corries were occupied by ice. It also is likely that discrete glacial events of a short-lived character, such as the Loch Lomond Stadial, contributed significantly to the development of these corries.

Caine (1972) demonstrated that the distribution of patterned ground in the Lake District shows a strong preference for Skiddaw Slate

Figure 5.70a Red Tarn, overdeepened lake basin and moraines. View looking north-east. (Photo: D. Huddart.)

Figure 5.70b Solifluction terrace, Helvellyn. (Photo: D. Huddart.)

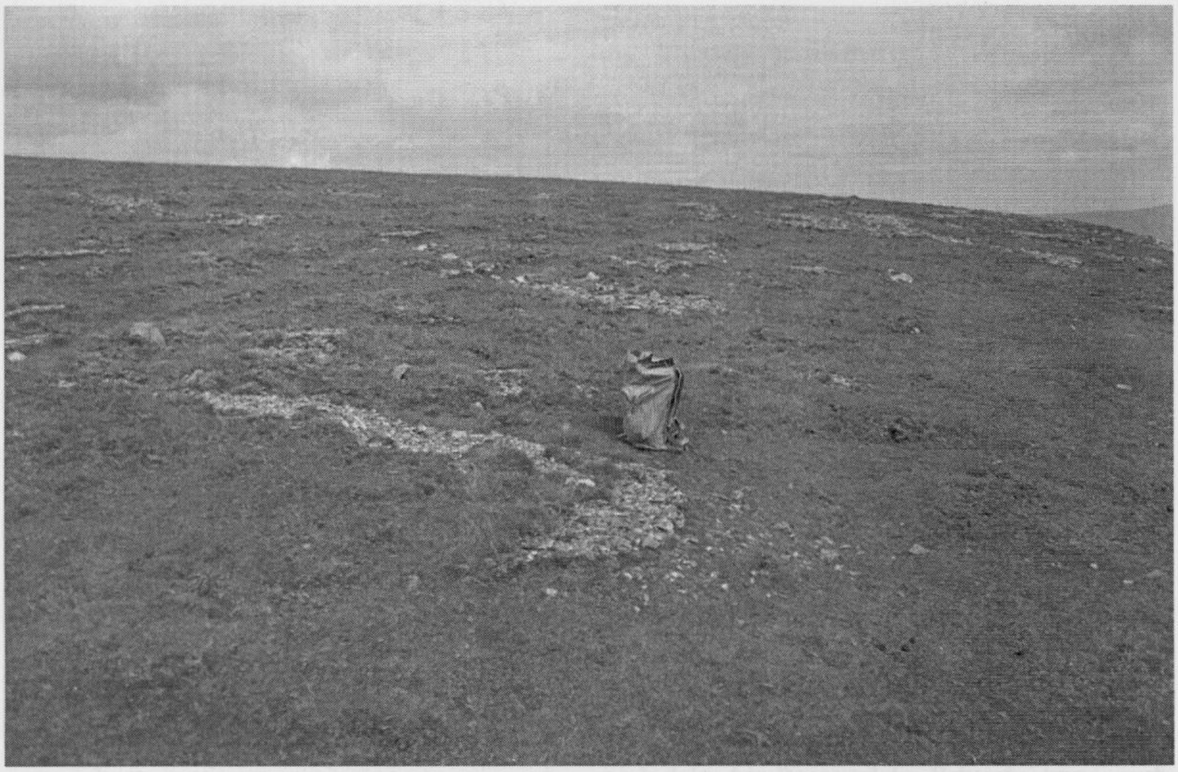


Figure 5.70c Solifluction lobes, Helvellyn. (Photo: D. Huddart.)

Figure 5.70d Vertical stones, Nethermost Pike. (Photo: D. Huddart.)

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; perhaps because at higher altitudes there will be more frost days and therefore greater frost-heave potential. However, the simple relationship also is affected by the presence of frost-susceptible material at the site. Widths and depths of patterned ground on Helvellyn are in general greater than at sites on Skiddaw Slates and this may be a reflection of higher altitude (Warburton, 1997).

Conclusions

Unambiguous glacial phenomena such as corries and arêtes are extremely well displayed in the Helvellyn Range. They are the result of long periods of glaciation of a restricted character with ice confined to corries and upland valley heads. Depositional features in and around the corries allow the extent of Loch Lomond Stadial glaciation to be established. This has led to important conclusions regarding conditions during this short period of climatic deterioration.

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 generally are 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.

ROMAN WALL (NY 715 667–NY 810 712)

N.F. Glasser

Introduction

The Roman Wall site is centred on a *c.* 10 km long section of Hadrian's Wall in Northumberland. The entire wall stretches for some 118 km across northern England from Wallsend, near the mouth of the Tyne in the east, to Bowness-on-Solway in the west. It is probably the most famous of the Roman walls in Britain, and its archaeological and cultural significance has been described in numerous publications (e.g. Breeze and Dobson, 1978; Johnson, 1989; Richards, 1993; Crow, 1995). The wall was built for defensive purposes and for much of its length it follows natural topographical barriers resulting from the underlying geology. This is particularly the case in its central portion around Roman Wall (NY 715 667–NY 810 712), where a large quartz-dolerite sill (the Whin Sill) provides steep north-facing cliffs that show the influence of geological structure on the landforms produced by glacial erosion. A particular highlight of the site is the development of a cuesta (scarp and dip) landscape. This is particularly well developed in the basaltic sills and dykes associated with the Whin Sill, with shallow rock basins developed in the less resistant sedimentary rocks (mainly limestones) between the sills. A suite of complimentary cuesta landforms of various magnitudes also occurs to both the north and south of the Whin Sill itself, in association with the outcrops of limestone and sandstone.

Johnson (1997) provides an overview of the solid geology along the entire route taken by Hadrian's Wall and numerous publications exist on the nature of the quartz-dolerite Whin Sill (Holmes and Harwood, 1928; Johnson, 1959; Dunham and Kaye, 1965; Fitch and Miller, 1967; Thorpe and McDonald, 1985). Surprisingly, however, there is little reference to the cuesta landscape in the published literature. A detailed record of Holocene environmental change is preserved in the shallow peat-filled depressions developed in the less resistant sedimentary rocks between the sills (Manning *et al.*, 1997).

Description

Solid geology

The solid geology upon which the wall rests in its central portion is dominated by Lower Carboniferous (Dinantian) sediments into which are intruded a series of quartz-dolerite sills (Figure 5.71). These quartz-dolerite sills are known collectively as the 'Whin Sill' or 'Great Whin Sill'. The central sector of the wall is an upland region of rugged scarps with well-exposed geological strata that contrasts with the more subdued and drift-dominated landscape to both the east and west (Johnson, 1997). The Whin Sill is the largest of a number of basaltic sills and a suite of dykes of Late Carboniferous age emplaced into the surrounding Carboniferous sedimentary rocks of north-east England

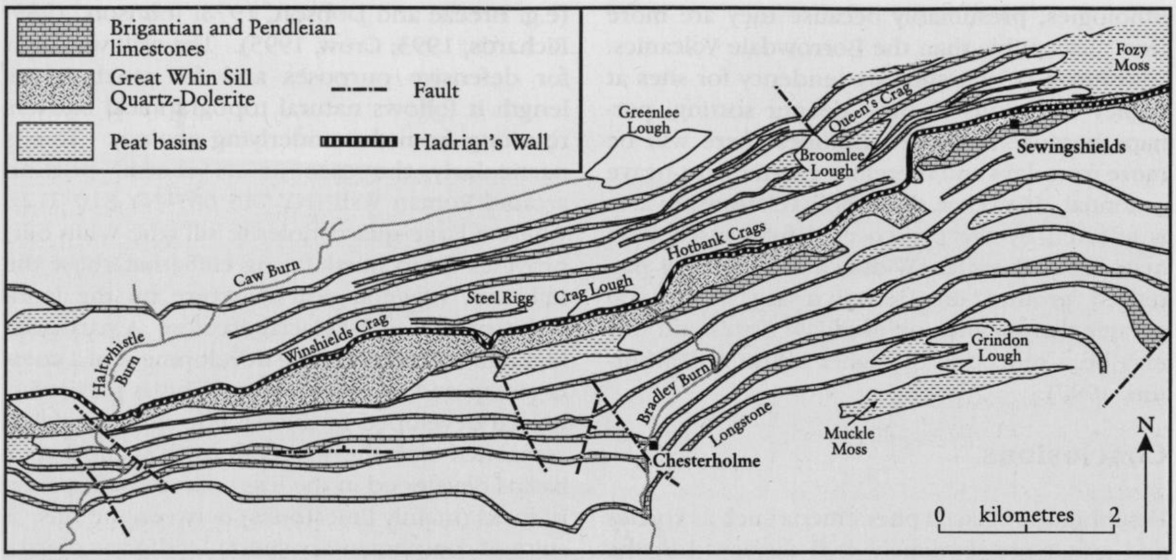


Figure 5.71 Geological map of Hadrian's Wall in the vicinity of the Roman Wall GCR site (after Johnson, 1997). Note how the route of the wall follows the Whin Sill and the general ENE–WSW grain of the landscape.

(Thorpe and Macdonald, 1985). The individual sills that together comprise the Whin Sill complex are composed primarily of quartz dolerites with thin, chilled upper and lower margins. The sills were injected at successively higher horizons in the sedimentary rocks when traced from east to west across the region (Johnson, 1959). The age and mechanism of emplacement of the sill are considered in detail by Holmes and Harwood (1928), Johnson (1959), Dunham and Kaye (1965), Fitch and Miller (1967) and Thorpe and McDonald (1985). Deep weathering of the dolerite sill to a depth of 15 m has been reported around the Middleton-in-Teesdale area (Hornung and Hatton, 1974). Occurrences of deep weathering are limited along the Roman Wall, being confined to shallow isolated pockets and to small patches of spheroidal weathering in the Bamburgh area (Hornung and Hatton, 1974).

Although much of the Whin Sill complex is not visible at the surface, and across the rest of northern England is encountered only in deep boreholes (Ridd *et al*., 1970), the sill is particularly well-exposed in the vicinity of Roman Wall. At this point, approximately 2 km to the north of Haltwhistle in Northumberland, the Whin Sill forms a spectacular north-facing escarpment in a series of cliffs up to 20 m high. On the southern side of the sill, where the dip of the strata decreases, the slopes are more gentle and bedding-parallel. The route of Hadrian's Wall takes advantage of this natural topographical feature and follows the crest of the escarpment at this point.

The cuesta landscape

The term 'cuesta' can be applied to any large-scale asymmetric landform. However, the term generally is reserved for landforms whose development is structurally controlled. In many cases, cuestas are created in situations where there is a gently inclined surface parallel to the dip of bedding planes and an escarpment or scarp face that is steeply inclined in the opposite direction to the dip slope and cuts across bedding planes. 'Cuesta' is Spanish for a flank or slope, and geomorphologists have used the term to describe asymmetric landforms in a variety of climatic settings (see e.g. Dayan, 1993; Tippett and Kamp, 1995). A landscape dominated by a series of cuestas is developed in the Whin Sill and its host sedimentary rocks along much of the length of Hadrian's Wall, especially in its central portion. The cuestas are particularly well developed where glacial erosion has exploited structural lines in the bedrock and accentuated the natural variability in the durability of the bedrock.

As a result, high north-facing cliffs and gentler south-facing dip slopes dominate the landscape wherever the Whin Sill has been subjected to glacial erosion. Farther away from the wall itself, limestone and sandstone strata around King's Crag and Queen's Crag (NY 795 704) also define a more subdued cuesta landscape, where differences in overall relief are less than at the wall. Structural control is still obvious in the form of scarp and dip contrasts, minor cliffs and rock basins. The most famous of the rock basins is the large Crag Lough (NY 765 681), where a lake occupies an ice-scoured bedrock hollow immediately beneath the plucked north face of the Whin Sill escarpment. Farther west, near Winshields Crag (NY 740 675) is the highest point of the wall at 375 m OD. Good examples of glacial meltwater channels are incised into the Whin Sill at elevations of up to *c.* 320 m OD at the top of Green Slack (NY 742 675). Here a large flat-floored channel runs south-east down the dip slope of the escarpment. Other examples of meltwater channels at Lodhams Slack (NY 738 672), at *c.* 300 m OD, also cut across the escarpment in the same orientation.

Interpretation

Johnson (1997) suggests that around Roman Wall the striking scarp and dip landforms arise from a combination of three factors. Firstly, the Carboniferous succession comprises a repeated or cyclic sequence of limestone, mudstone, sandstone and coal. This produces an alternation of hard and soft beds of different susceptibility to weathering and erosion. The quartz-dolerite Whin Sill, intruded into this succession in the Late Carboniferous is of varying thickness and transgresses from one horizon to another. By comparison with the limestone, mudstone, sandstone and coal, the Whin Sill is very durable and therefore stands proud of these other rocks. Secondly, uplift associated with the Cheviot pluton to the north resulted in doming of the sediments surrounding the igneous centre. This imparted a significant southerly dip to the strata in the Roman Wall area. Thirdly, that Late Devensian ice movement in this area was predominantly west to east, parallel to the strike of the beds. Evidence for this easterly ice flow across the area comes from the movement eastwards of Lake District erratics (Johnson, 1952). This ice movement direction is partly responsible for accentuating the scarp and dip topography.

In addition to this landform interest, the cuesta landscape also provides insight into the processes of glacial erosion operating beneath the Late Devensian ice sheet. Ice-sheet erosion can be achieved by the physical processes of abrasion and plucking, and by meltwater erosion (both chemical and mechanical) (Bennett and Glasser, 1996). The resulting landforms are a complex amalgam relating both to the processes operating within the ice sheet itself and to the nature of its bed. Variables such as topography (Glasser, 1995), ice-sheet thermal regime (Gordon, 1979; Kleman, 1994), preglacial relief and weathering (Klimaszewski, 1964; Lindström, 1988), ice dynamics (Nye and Martin, 1968; Andrews, 1972), scale (Glasser and Warren, 1990), the primary nature of the substrate (Boulton, 1979) and bedrock lithology (Gordon, 1981; Glasser *et al.*, 1998) are all important in this context. Variables external to the ice-sheet system, such as the duration of glaciation, also are important (Porter, 1989). The cuesta landscape exemplified at Roman Wall is primarily the result of glacial quarrying by rock fracture and entrainment that acts to produce rock steps, cliffs and plucked bedrock surfaces. Previous research has shown that at this scale glacial quarrying is most effective in situations where a cavity forms between the ice and its bed, usually in the lee of a bedrock obstacle (Glasser and Warren, 1990; Sugden *et al.*, 1992). Rock breakdown can then occur as a result of frost action (Walder and Hallet, 1985, 1986) and fluctuations in subglacial water pressure (Hooke, 1991; Iverson, 1991). The situation at Roman Wall is somewhat unusual because the landforms are orientated with the prominent cliff proximal to ice flow rather than distal. As the formation of cavities is favoured under thin ice with high velocities and abundant meltwater (Sugden *et al.*, 1992), this provides insight into the nature of the Late Devensian ice sheet. This fits well with other geomorphological evidence in the region (e.g. Common, 1953, 1957; Derbyshire, 1961; Clapperton, 1966, 1968, 1971a, b).

Finally, it should be noted that erosion of the escarpment could play an important role in former ice-sheet dynamics and in determining the distribution of debris within the Late Devensian ice sheet (Benn and Evans, 1998). For example, compressive flow is likely to encourage transportation of debris to the ice surface wherever an ice sheet flows against a topographical obstacle such as an escarpment (Straw, 1968a; Worsley, 1969; Eyles and Menzies, 1983; Paul, 1983; Sharpe, 1988). This control on sediment supply has important implications for landform development further afield, especially during ice recession.

Conclusions

The Whin Sill and surrounding sedimentary rocks demonstrate the strong influence of geological structure on the development of glacial erosional landforms. The steeply dipping scarp face on the north side forms a prominent escarpment, whereas a gentle dip slope is developed to the south, creating a cuesta landscape. Less resistant sedimentary strata also have been exploited by glacial erosion to form minor rock basins in the intervening linear depressions. The development of these landforms can be related directly to geological structure, ice movement direction and glacial erosional processes. In total, the site provides a striking example of the influence of geological structure on glacial landform development and provides insight into the nature of subglacial processes operating beneath the Late Devensian ice sheet.

NORBER ERRATICS (SD 765 700)

D. Huddart

Introduction

The Norber ridge (Figure 5.72) and associated Robin Proctor's Scar and Nappa Scars, in North Yorkshire, have become a classic area for glacial geomorphology as they demonstrate an assemblage of perched glacial erratic blocks, resting on limestone pavement and pedestals, which may have important repercussions in demonstrating the amount and rate of post-glacial weathering on limestone. They have been described many times, by Phillips (1827, 1855), Howson (1850), Mackintosh (1883), Hughes (1886), Kendall and Wroot (1924), Brumhead (1979), Waltham (1987) and Waltham *et al.* (1997).

Description

The principal interest in the site is the spectacular erratic train from the west side of Crummack Dale, which contains hundreds of large, angular erratics of Austwick Formation sandstones and siltstones (of Silurian Wenlock age) from outcrops to the north-west (Arthurton *et al.*, 1988). These erratics occur on various scales up to boulders with long axes up to 4 m and intermediate axes up to 2 m (Figure 5.73). Many of the erratics are perched on the underlying limestone pavements, but the latter are poorly developed and the erratics usually are scattered over thinly drift-covered limestone grassland surfaces. Occasional limestone boulders are perched on pedestals of limestone and they usually can be recognized by their 'cockly' or crinkled weathering caused by the direct atmospheric action of rainwater. It is clear that the erratics have been transported glacially by ice moving from the north, and transported across the limestone outcrop for more than 1 km and more than 120 m above their source outcrop. At Norber, the erratic boulders stand on pedestals of limestone up to 30–50 cm above the surrounding pavements and it always has been assumed that erratic-bearing pedestals demonstrate the amount of surface lowering of the limestone since the erratics were deposited.

The area overall is a classic example of glaciokarst where the initial stage of pavement formation was by glacial scour to give a relatively planed surface (Sweeting, 1966; Trudgill 1985). Although much of the limestone pavement shows no evidence of glacial scour, this occasionally can be seen beneath erratic blocks, where the limestone surface is smoothed, polished, furrowed and striated, as at Norber (Hughes, 1886).

Interpretation

The subaerial dissolution of limestone, or corrosion by direct rainfall, has been suggested as an important erosional process (Arthurton *et al.*, 1988) and its amount can be estimated by measuring the height of erratic-bearing limestone pillars standing above the surrounding pavement levels. If the age of emplacement of the erratics is known then the rate of subsequent dissolution of limestone can be deduced. Thus, if one assumes that the pedestals of limestone have been protected from corrosion by direct rainfall whilst the surrounding limestone surface has been lowered by subaerial and subsoil solution then as the pedestals are mostly 40–50 cm high, the mean rate of lowering of the exposed surface has been about 30–40 mm ka^{-1} since the Devensian ice retreat (Sweeting, 1966). Very

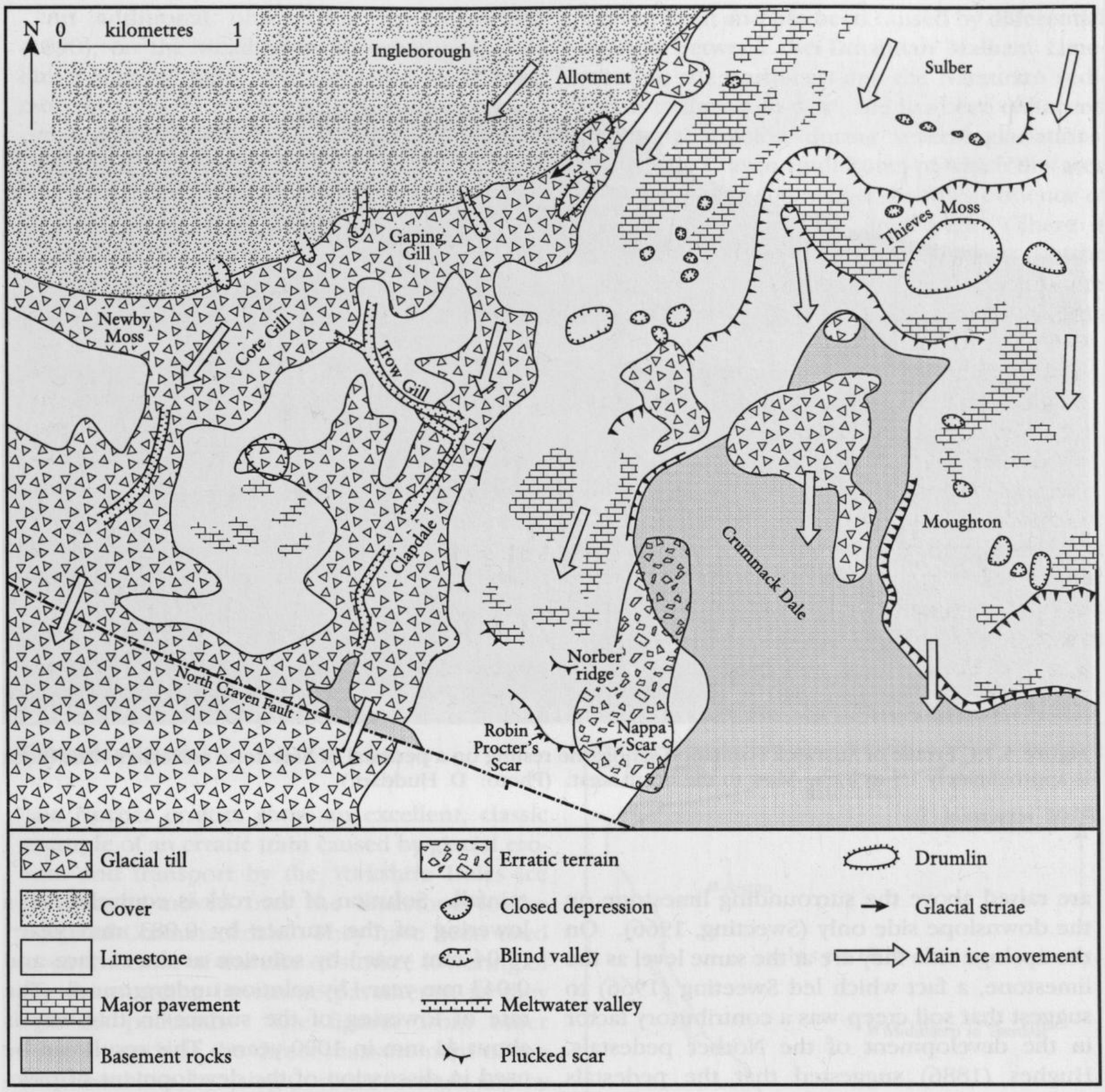


Figure 5.72 Geomorphological map of the southern sector of Ingleborough (after Waltham 1990).

little quantitative work, however, has been attempted to test this assumption and there appear to be several problems related to its accuracy as a technique.

Perched blocks of limestone, which are common on the north-west side of Ingleborough at Scar Close and on Harry Hallam's Moss, show that the clint surfaces have been lowered by about 50 cm since their formation (Sweeting, 1966). However, many are probably what Corbel (1957) has called 'pseudo-erratics', that is residuals left by stripping of the beds by solution as described by Sweeting (1966), and on Scar Close there is a height variation of 20 cm within a relatively small area, which probably indicates a process other than a subaerial one in their formation. It also could be envisaged that erratics on pedestals could be uncovered by chance as a glacial drift cover was stripped away from the limestone pavements, and hence their exposure to subaerial processes could be of different post-glacial ages.

At Norber, conditions for solution under prevailing peaty drainage are less favourable than for those at, for example, Twistleton Scars and Scar Close in Chapel-le-Dale, and many erratics

Figure 5.73 Erratic of Austwick Formation sandstone resting on a pedestal of limestone at Norber. The block is approximately 1.5 m high. View to the north-west. (Photo: D. Huddart.)

are raised above the surrounding limestone on the downslope side only (Sweeting, 1966). On the upslope side they are at the same level as the limestone, a fact which led Sweeting (1966) to suggest that soil creep was a contributory factor in the development of the Norber pedestals. Hughes (1886) suggested that the pedestals seem to be gradually perishing and that there must be a natural limit to their height. Using the umbrella analogy, he thought that an umbrella held too high will shelter only from perfectly vertical rain so when the pedestal has attained such a height that the rain can beat in under the boulder, the pedestal gets eroded at its base. This occurs more on one side than on another according to the prevailing winds. Eventually the boulder topples and the result of this can be seen at Norber.

As Corbel (1959) had devised a formula to calculate the number of millimetres of limestone per 1000 years that was removed by unaided solution, it was applied to the figures for north-west Yorkshire by Sweeting (1966) at the present rates of limestone solution and with the present rainfall. Solution of the rock is equivalent to a lowering of the surface by 0.083 mm year^{-1} (0.04 mm year^{-1} by solution at the surface and 0.043 mm year^{-1} by solution underground). The rate of lowering of the surface in this way is about 41 mm in 1000 years. This result can be used in discussion of the development of pavements and the location of perched erratic blocks. Sweeting (1966) suggested that the erratics may be regarded as dating from the Late Devensian period about 12 000 years ago, although 14 000 or 15 000 years seems a more likely figure. Assuming the rate of dissolution of the limestone at the surface to have remained much the same over time as it is now, the amount of lowering of the surface would be about 49 cm. This result is of the same order of magnitude as the height of the pedestals that support the perched erratic boulders. However, the rate of surface solution might not have remained the same because we know that climate, including both rainfall and temperature, which both affect the rate of solution, have varied though the post-glacial period.

An additional observation by Goodchild (1890), on the weathering of limestones, is of direct interest here. He used the weathering rate on local monuments and gravestones to estimate the rate of weathering of the northern England limestones at about 48 mm in 1000 years. On this basis the weathering of the limestone since the last glacial would be about 60 cm, a figure not too dissimilar to that already quoted. Goodchild regarded his figure as a lowest estimate and thought that 80 mm in 1000 years might be nearer the truth. Nevertheless, Sweeting (1966) suggested that Goodchild's findings remained in reasonable accordance with the Corbel formula.

Direct measurements of the erosion rate of the limestone using micro-erosion measurements over three years by Nicholson (1990) on the pavement close to Alum Pot beck in Ribblesdale show an erosion rate of 0.33 m 1000^{-1} years. This agrees well with the rate of 0.4 m 1000 years^{-1} as an average since deglaciation estimated using the Norber erratics and pedestals by Sweeting (1966).

Conclusions

The Norber erratics show an excellent, classic example of an erratic train caused by glacial erosion and transport by the Yorkshire Dales ice sheet as it moved over the limestone to the north of Crummackdale. They have been used as an indicator of the rate of surface lowering of the surrounding limestone pavements, as they rest on pedestals. The figures from other sources, such as the direct measurement of the solutional loss, by using Corbel's formula and by a comparison of the weathering of limestone gravestones and monuments all suggest an approximate rate of surface lowering. However, some doubts remain as to the exact formation of the erratics and their pedestals and their real value in measuring the rate of surface limestone lowering.

GIGGLESWICK SCAR (SD 803 655)

D. Huddart

Introduction

Giggleswick Scar is a classic example of a fault-line scarp. It follows the line of the South Craven Fault and has been caused by differential erosion between the Dinantian Malham Limestone to the north-east and the Namurian sediments to the south-west, and has been enhanced by glacial erosion during several glaciations. The whole Craven Fault Zone, of which this area forms a part, is complex and shows evidence of long-continued tectonic movement. There is additional geomorphological interest in the drift-covered terraces, screes, dry valley systems and intermittent spring and small caves associated with the scarp.

Description

Fault-line scarp

The major structural units of the northern Pennines, the Alston and Askrigg Blocks, are divided by the Stainmore Syncline (Hudson, 1933, 1938), and are defined by complex, fault-line scarps forming prominent features in the

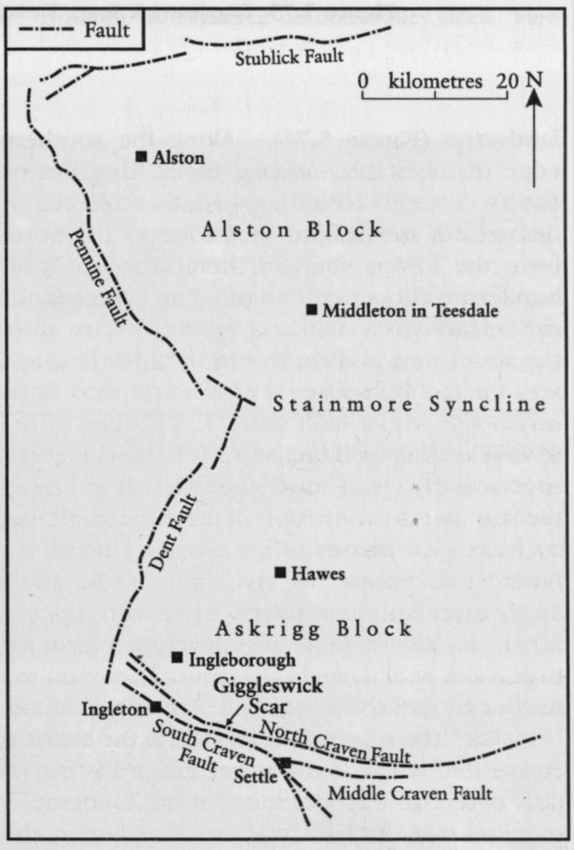


Figure 5.74 Structural units in the northern Pennines and the position of the Craven faults.

Figure 5.75 Giggleswick Scar. View looking south-east. (Photo: D. Huddart.)

landscape (Figure 5.74). Along the southern edge of the stable, Askrigg Block, the 'Craven Faults System' (Phillips, 1836) collectively demarcates the upland limestone to the north from the lower, younger, non-carbonate, Carboniferous rocks to the south. The Craven faults run mainly WNW–ESE and the throws are all to the south-west and south. In the Ingleborough area the fault-line scarp is a broad-stepped slope across the whole fault zone, but farther to the south-east this fault-line scarp is best seen at the spectacularly clean-cut Giggleswick Scar. Here, the scar marks the erosion of the actual fault line by successive phases of ice erosion during the Pleistocene Epoch, to give one of the most impressive fault-line scarps in Britain (Figure 5.75). It extends in its best developed form for over 2 km, with a height differential between the north-east and south-west sides of approximately 100 m. The scar itself is formed of the Malham Formation, whereas the lower ground is glacial drift covered. The freshness of the topography suggests that the fault was active as late as the Late Tertiary (Brumhead, 1979; Nicholson, 1990) and it is possible that it is still active today.

Glacial landforms

The scars are formed by ice plucking the limestone so that differential erosion has produced a series of terraces, most of which are covered by thin, glacial drift. The drumlin zone south and south-east of Ingleborough indicates ice moving south-eastwards towards Airedale from Kingsdale, Chapel-le-Dale, Crummackdale and down Ribblesdale (Tiddeman, 1872, Arthurton *et al.*, 1988). At the western end of the scar, near Scar Top, a shallow, dry, valley system runs across the scar to two feeder entry points which coalesce into a large depression. A series of similar shallow valleys cut the surface of the scar to the south-east. One such narrow valley, about 25 m in width, can be seen running from Kinsey cave for about 75 m to the front face of the scar, where it abruptly terminates (Figure 5.76a). Blocky scree is found on both sides of this valley and the most likely explanation for the dry valleys is a subglacial channel, probably a subglacial chute, which ran down the scar under the ice and flowed back into the ice where it suddenly terminates. Extensive aprons of blocky scree

Figure 5.76a Dry valley system leading from Kinsey cave. Note the blocky scree on both valley sides. View looking north-east. (Photo: D. Huddart.)

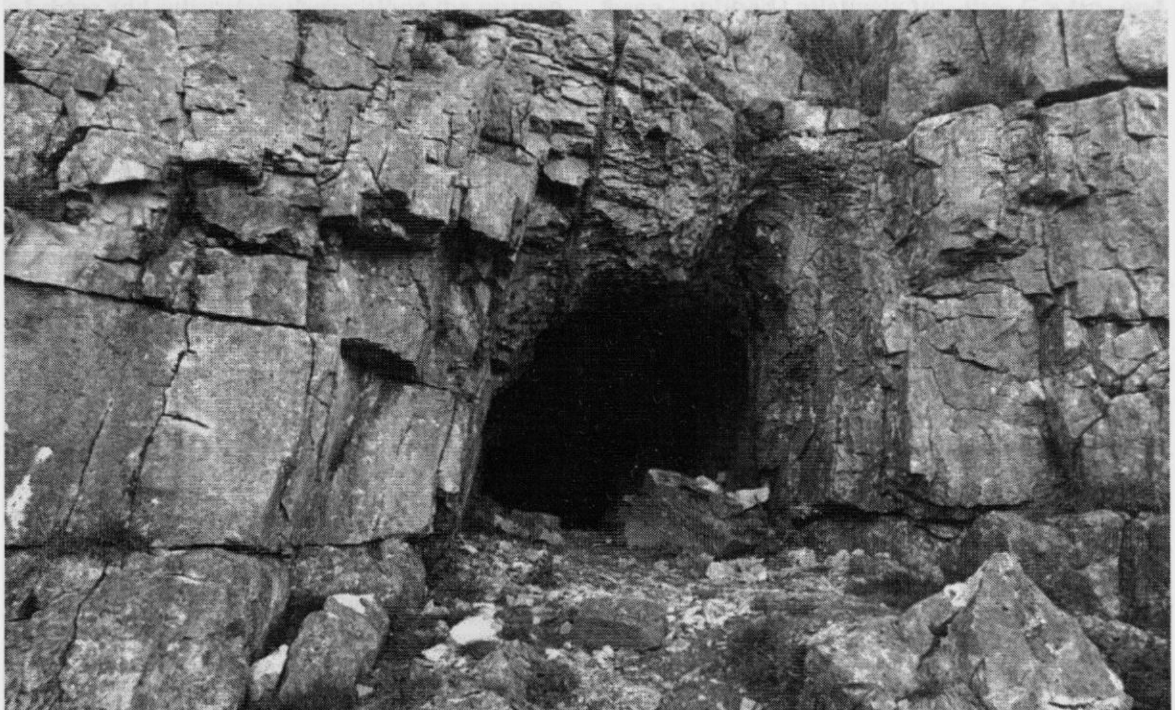


Figure 5.76b Spider Cave entrance, a phreatic tube. (Photo: D. Huddart.)

below the individual scars form along the length of Giggleswick Scar.

Caves

Brook *et al.* (1982) describe a series of small caves situated on Giggleswick Scar (Figure 5.77), which are important in terms of archaeology but also indicate an earlier, phreatic drainage system that has been totally disrupted by glacial erosion. For example, Spider Cave is a 21 m phreatic tube that is now connected to the surface along a grike (Figure 5.76b). Kinsey Cave, just to the north, has an 8 m wide overhanging entrance 2.4 m high, is 23 m long and has a phreatic roof showing large scallops and anastomoses. It has important sediment fills, flowstone on the walls and terminates in a calcited blockfall.

Archaeologically, King (1974) described these caves and others, such as Kelco and Sewell's Cave, as south-facing, human rock shelters but that they had incomplete early excavation reports. Excavations from 1925 provided evidence of Romano-British and Late Pleistocene occupation (Jackson and Mattinson, 1932; Thornber, 1959). The remains of domestic animals used for food and Romano-British material were found in the upper layers and in the lower layers the remains of reindeer, bison, bear and lynx were located. A fragment of worked rod of reindeer antler was excavated from the lower layer, which had a tapering squared section, the corners being rounded and longitudinally it had been scored with paired lines. Some Creswellian points were also excavated. In Sewell's Cave one complete and one incomplete Roman sword were excavated, together with a piece of possible plate mail, six human skulls and the skull of a young pig, but because of disturbance it was not possible to say if the material was Romano-British or earlier. Sir Arthur Keith in his report suggested that the cave had been used as a burial place, with new burials disturbing the older ones (Raistrick, 1936). At Greater and Lesser Kelco caves excavations from as early as 1850 and later from 1928 to 1933 yielded numerous Romano-British finds, along with the remains of domestic food animals. Both caves also produced Neolithic pottery and human skulls, possibly from the Bronze age.

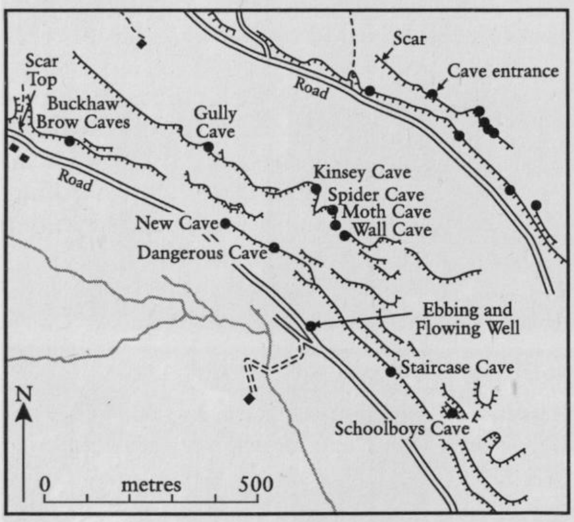


Figure 5.77 Caves in the Giggleswick Scar area (after Brook *et al.*, 1982).

Intermittent spring

Sweeting (1972) suggests that the best known intermittent spring in Britain is the Ebbing and Flowing Well at the base of the Giggleswick Scar (Figure 5.78). This well is more a reciprocating spring, which is one that does not dry up between the flows but runs continuously, with irregular discharge. All observers of the Ebbing Flowing Well have commented on its irregular flow and Stevens (1964) has collected information about the well, such as the rises and falls noted by Swainson in 1796 (quoted in Housman, 1800) and Sweeting in 1947. It now ebbs and flows only during periods of moderate rainfall. In dry weather it flows without interruption and after heavy rain it floods like others in the same area. It appears to be less active than in the past and it has been suggested that in time all ebbing and flowing springs become normal karst springs (Sweeting, 1972).

Interpretation

Fault-line scarp

Although Giggleswick Scar is situated on the line of the South Craven Fault it is part of a complex tectonic area, the evolution of which is controversial (see details in Tiddeman, 1889, 1891; Hudson, 1930, 1933, 1944; Wager, 1931; Hudson and Mitchell, 1937; O'Connor *et al.*, 1974; Johnson, 1967; Tegerdine *et al.*, 1981; Leeder, 1982; Arthurton, 1984). The controversy lies in when the faulting was initiated and its influence

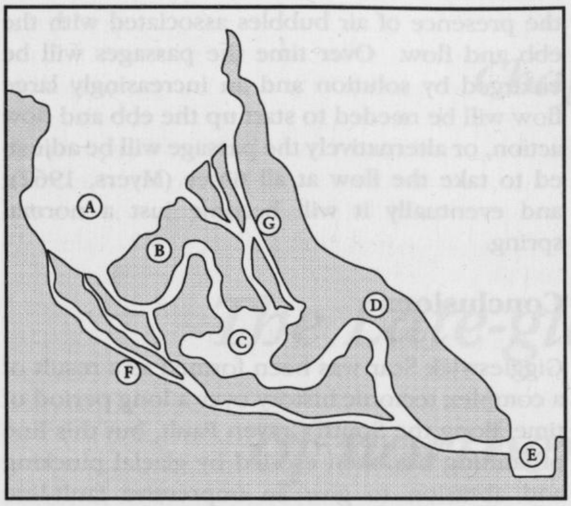


Figure 5.78 Intermittent spring at Giggleswick Scar (after Howson, 1850). A–G: see text.

on subsequent sedimentation (Arthurton *et al.*, 1988).

Recent movement has occurred on the South Craven Fault system, as Versey (1948) recorded a seismic event known as the 'Skipton (or Settle) earthquake', which occurred in 1944. He regarded the epicentre as in all probability lying along the Gargrave fault. Damage was caused to a bridge and subsurface drains along the unspecified 'mapped line of the fault'. There has been repeated occurrence of small earthquakes along the edge of the Askrigg Block that bear witness to differential movement still taking place along these fault systems.

The results of this complex fault movement are important for the landscape because Wager (1931) demonstrated that the direction and density of the joints in the Dinantian limestones were related to the major faulting along the Craven Fault system, although the mechanism for their formation is still in dispute (Moseley and Ahmed, 1967; Doughty, 1968). Wager (1931) demonstrated the direction of jointing in the Malham area and interpreted the changes near the faults as an indication that the North Craven Fault was sinistral and the Mid-Craven Fault dextral. Away from the faults the joint trends are NW–SE and NE–SW. As the North Craven Fault is approached from the north the joints tend to swing anticlockwise so that those in the north-west sector swing to a more east–west trend. Less marked but nevertheless the same is the anticlockwise deflection found close to the south of the fault. In contrast, the joints close to the Mid-Craven Fault show a clockwise deflection and joints in the north-west sector tend to swing to a more northerly direction. Between the two faults there are a number of cross-faults, with most trending north-west to south-east. Their throw is small, sometimes negligible and they are probably wrench faults formed by the stresses and strains set up in the lateral movement of the North Craven and Mid-Craven faults. These small faults can be picked out by glacial drainage as weaknesses and it is possible that many of the dry valley systems could be initiated along these lines. They certainly have been demonstrated to be important in the development of vertical pitches in caves (Waltham *et al.*, 1997) and some of the collapse dolines, such as Great Douk Pot, are on such faults. The density of jointing close to faults also gives a very different type of limestone pavement morphology (Sweeting, 1966; Trudgill, 1985; Waltham *et al.*, 1997).

Cave development

On a more major scale, rejuvenation of the Askrigg Block along the line of the Craven Fault System has influenced the relief and the caves markedly. This is because the Askrigg Block has been uplifted relatively to the Craven and Lonsdale lowlands since at least Tertiary times. This has allowed waters to descend more rapidly into the limestone and to form deeper underground channels, vertical potholes and shafts and to produce the waterfalls and gorge systems in an area. The caves associated with Giggleswick Scar do not show this verticality and are relict phreatic tubes, which formed when the water table was much higher and therefore must be of considerable antiquity. The cave systems have now been exposed by glacial erosion on the scars and form rock shelters that have been used by humans from the late Palaeolithic period onwards.

Reciprocating spring

Different explanations have been given for the intermittent spring that tried to account for the reciprocation, its irregularity and the influence of wet and dry seasons. A solution on the principle of the double syphon was given by Thomas Hargraves of Settle (Howson, 1850) and is illustrated in Figure 5.78. Chamber A is connected to a further chamber C by a duct (B). This chamber is connected to the resurgence at E by a further duct (D). Above and below duct B are a series of further ducts (F and G). Ducts B and D each form a syphon, where B draws off the water from the basin A and fills the smaller basin C until it runs over at D. As D is wider than B it soon empties the basin C and then the stream ceases until C is filled again, thus causing the reciprocation. The irregularity of the reciprocation is caused by the fact that B draws off water from A faster than it is supplied by the spring and therefore A becomes empty and no reciprocation takes place until it is filled again to the height of syphon B. At which time the fullness of A causes a powerful flow and before the well goes down to its average flow, a series of flows takes place and the interval between each flux and reflux increases, until A is emptied again. In dry weather there is no reciprocation because the water is insufficient to fill B and it escapes through the duct F below. After much rain the basin C is supplied by too much water from B and the duct G operates. Sweeting (1972) suggested that the irregularity could be the result of the presence of air bubbles associated with the ebb and flow. Over time the passages will be enlarged by solution and an increasingly large flow will be needed to start up the ebb and flow action, or alternatively the passage will be adjusted to take the flow at all times (Myers, 1962), and eventually it will become just a normal spring.

Conclusions

Giggleswick Scar was been formed as a result of a complex tectonic history over a long period of time along the South Craven Fault, but this line of faulting has been eroded by glacial plucking and abrasion to give an impressive fault-line scarp. A series of small caves, originally forming beneath the water table in the phreatic zone, which was disrupted on a major scale by these glacials, now form rock shelters in which there are important archaeological and palaeontological remains. Minor glacial landforms, such as scars, drift-covered terraces and dry valley systems are located on Giggleswick Scar and there is an important reciprocating spring.