Chapter 5 The Devensian glacial record

Introduction

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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 (5026 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 interstadi-als. 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 (50–2 5 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 BE For the period 44 000–34 000 years BP, a minor increase in the number of dated speleo-thems 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 BE 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 Uenkinson *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 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.

(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

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 sub-glacially, 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 5.1)a). 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 5.1)b). In Cumbria, Trotter (1929) and Hollingworth (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 timing 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 stillstands 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 BR 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 Devensian 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 5.4)a, 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 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', 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 sea-floor (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 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.

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 out-drop 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 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 'over spill' 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 sub-glacially 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 sub-glacial 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 northeast, 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, Gilding 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 kettleholes 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 deglac-iation (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 5.8)a, 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 5.10)a, 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 5.10)b. 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 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.

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 ((Figure 5.12) and (Figure 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 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 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 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 5.13)b. 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 downmelted 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 5.14)a), 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 5.14)b) 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 5.14)c). 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 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 Sandstones, 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 supra-glacial deposits that formed when the southward moving Irish Sea ice sheet impinged against the western Pennines ((Figure 5.17)a). The sedimentology is complicated, but in general the main topographical elements are constructed of out-wash, 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 5.17)b) 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 interlobate outwash plain. This landform has been inter preted 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 5.19)a) 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 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 reentered the margin of the Irish Sea ice to reemerge 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 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; Arthur-ton *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 Rais-trick (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 glaciomarineprocesses 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 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 worldwide 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 drawdown 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 glaciomarineinterpretation. 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 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 overcon-solidated 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 pro-glacial environment, with interbedded sub-glacial 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 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 giaciomarine 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 preexisting 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 illus trated 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 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. McCarron (2001) reviewed critically the lines of evidence on which the glaciomarinehypothesis 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 northwest 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 Kinard 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 northwest 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 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 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 exam ple, 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 glacio-lacustrine 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 glacio-lacustrine 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 sub-aerial 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 north wards 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 glacio-lacustrine 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 sub-glacial obstacles.
3. They are erosional remnants of subglacial water sheet-floods or fluvial infills. This somewhat controversial origin involves large sub-glacial 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 sub-glacial 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 glaciomarinesetting.
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.

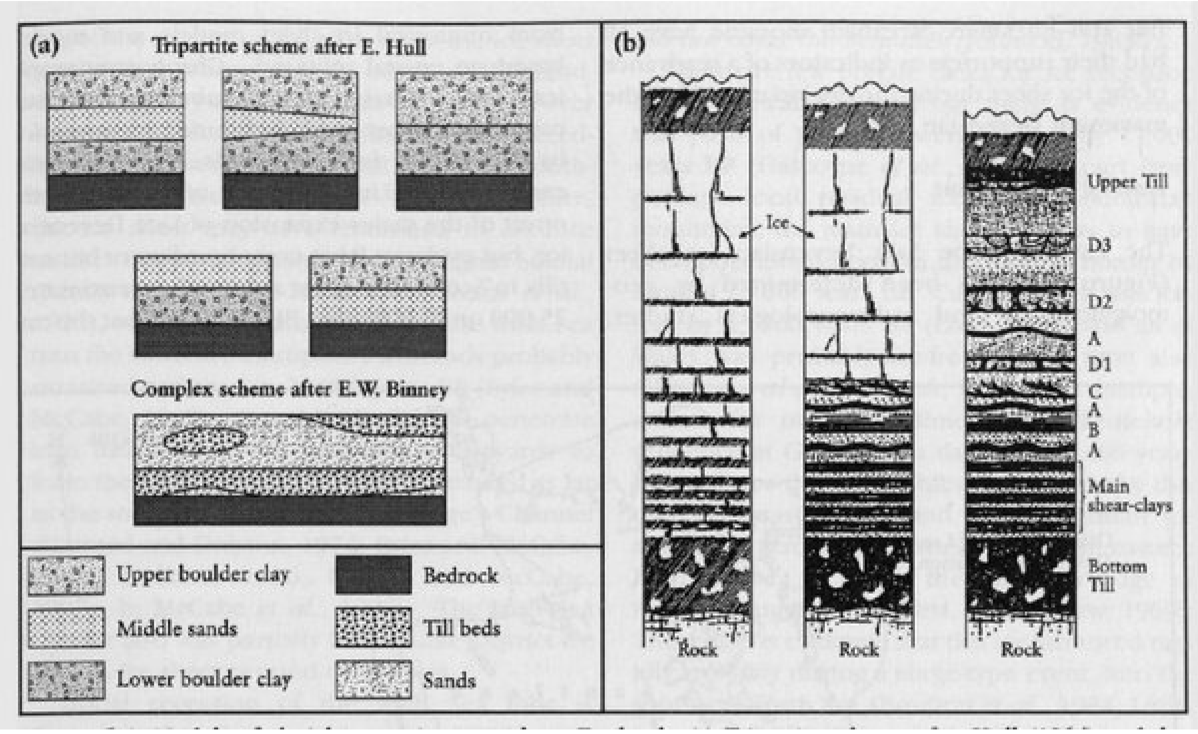
References

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

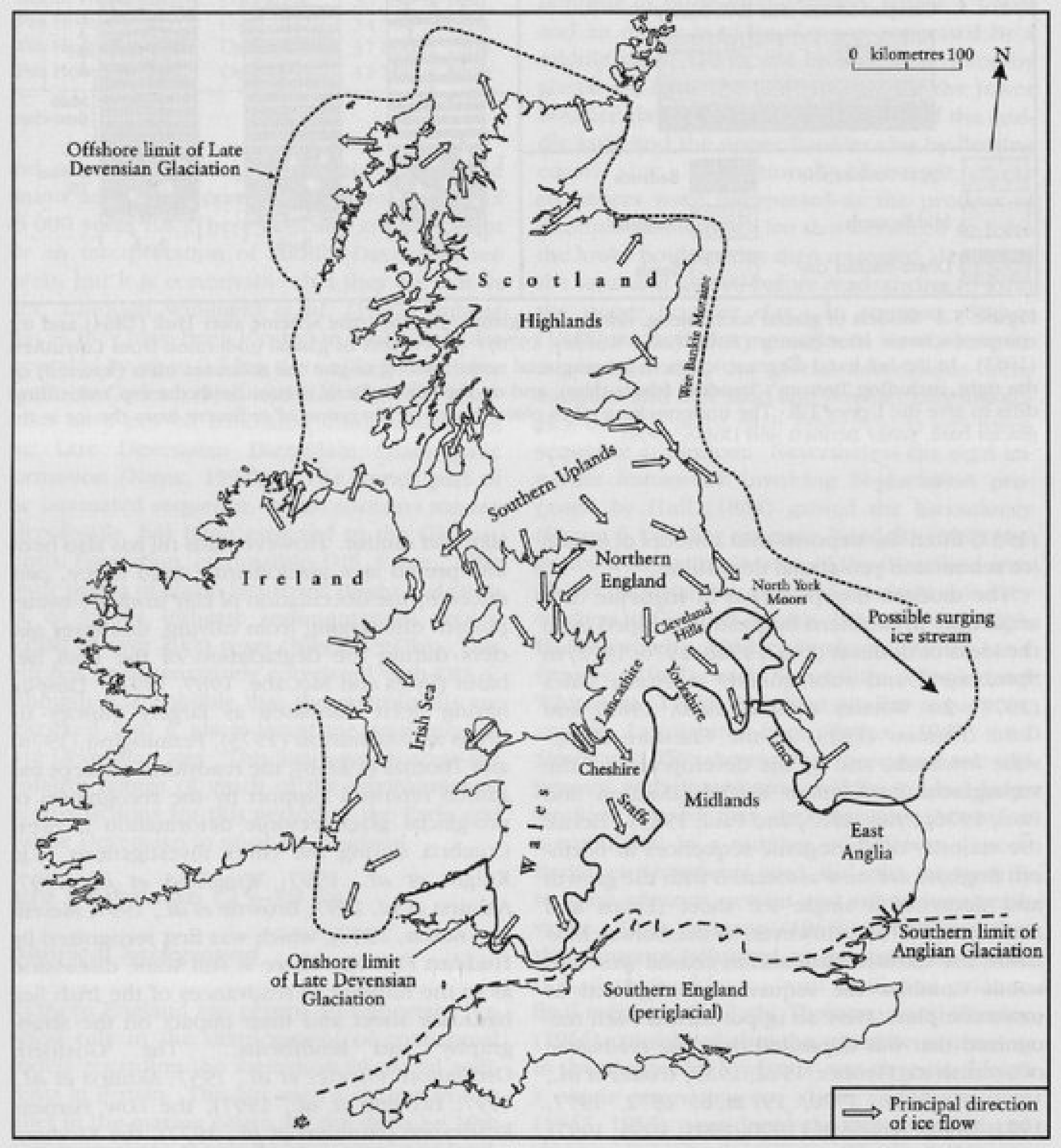
(Table 5.1) The mammalian fauna from the Pin Hole Mammalian Zone, Lower Cave Earth, Pin Hole Cave, Cresswell, Derbyshire (after Currant and Jacobi, 2001).

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

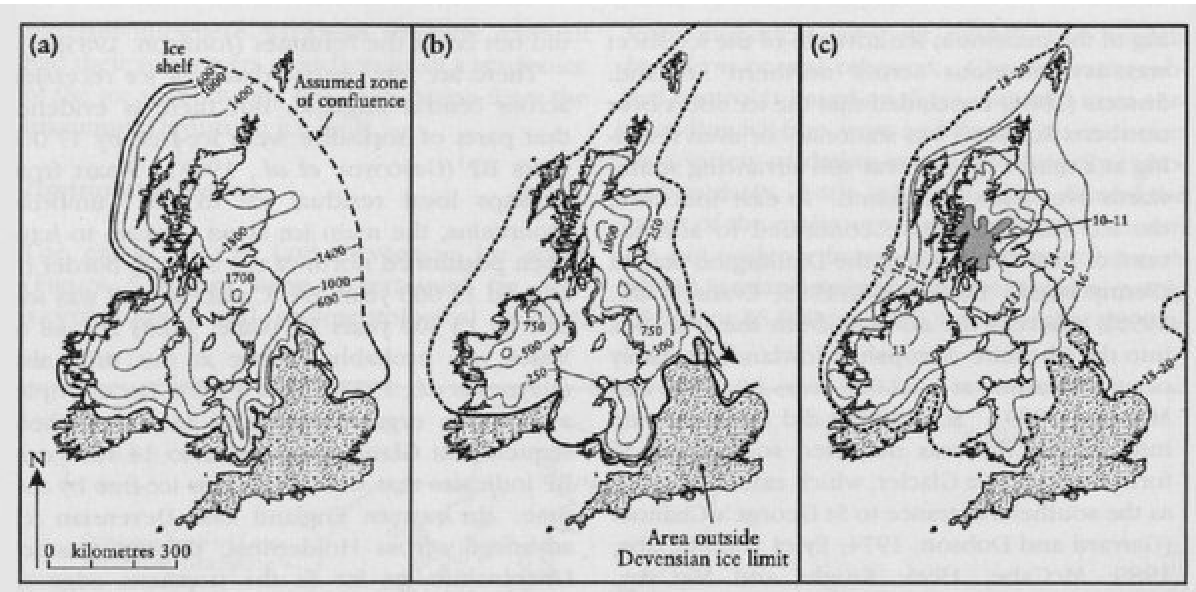
(Table 5.2) Radiocarbon dates (years BP) on spotted hyaena remains from the Cresswell area, Derbyshire (after Currant and Jacobi, 2001)



(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' dins 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).

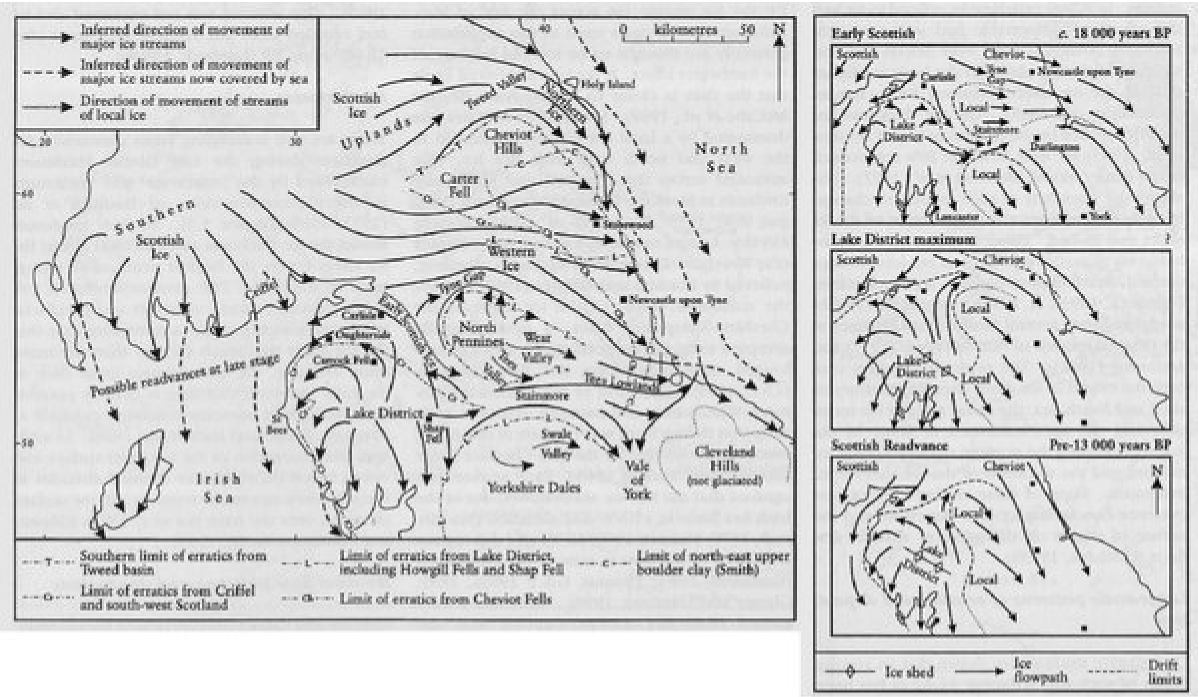


(Figure 5.2) Limits of Anglian and Late Devensian glaciations in Great Britain (based on Eyles and Dearman, 1981; Eyles and McCabe, 1989; Boulton et al., 1991).

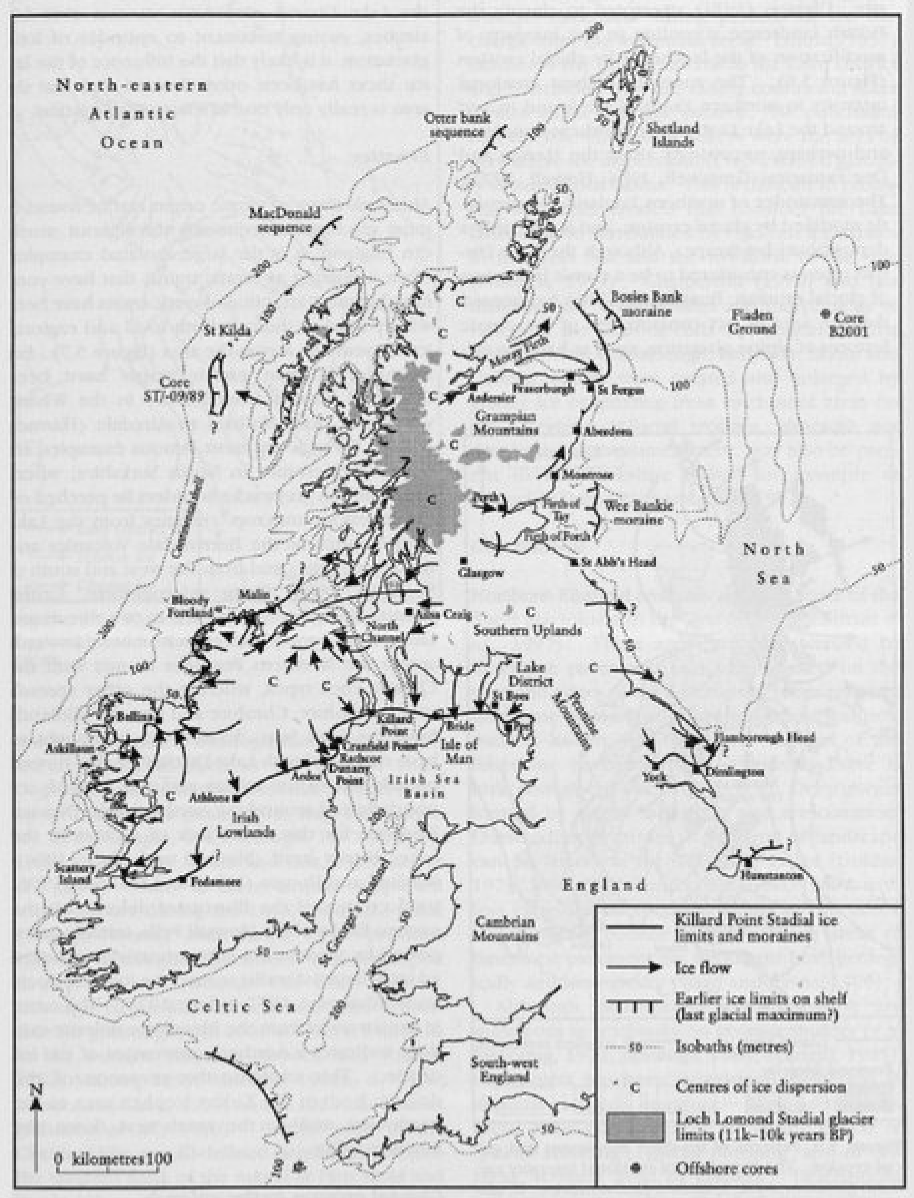


(Figure 5.3) The modelled ice sheet at the time of the Last Glacial Maximum (ice thickness in metres): (a) according to Boulton et al. (1977); (b) according to Boulton et al. (1985). (c) Isochrons (ka) for recession of the last ice sheet according

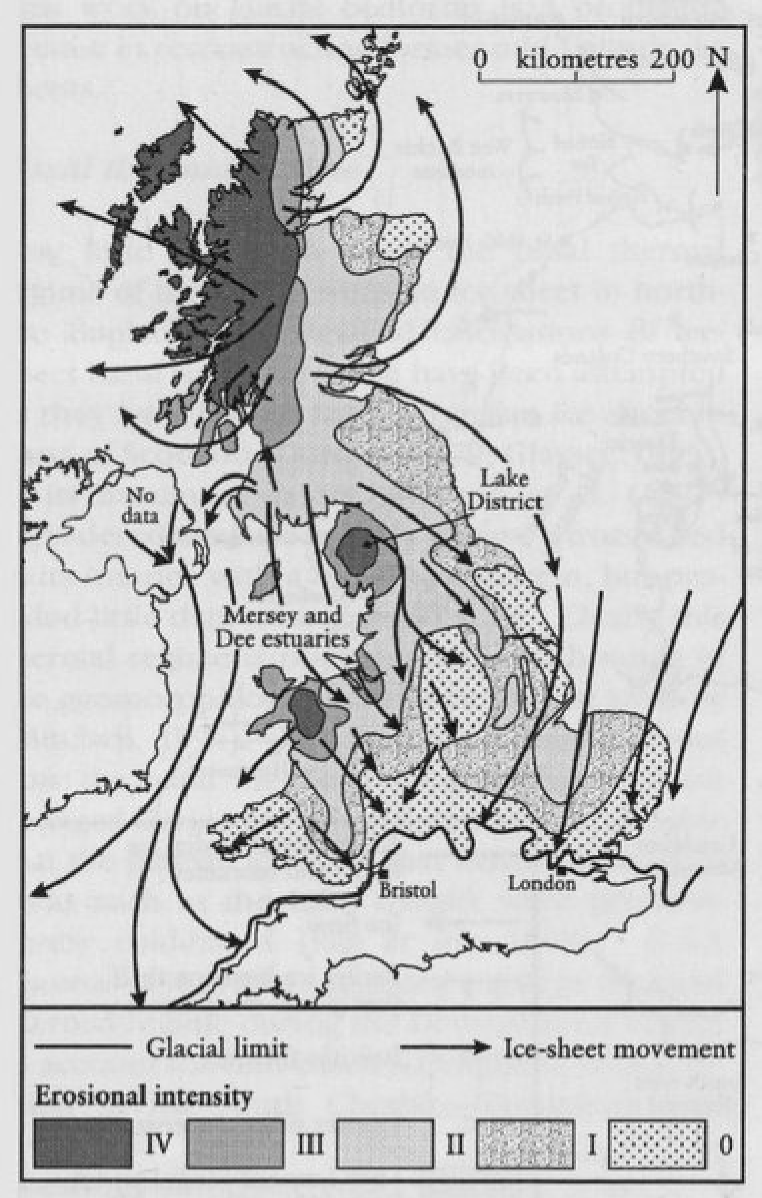
to Andersen (1981). The extent of the Loch Lomond (Younger Dryas) icefield (11–10 ka) is shaded.



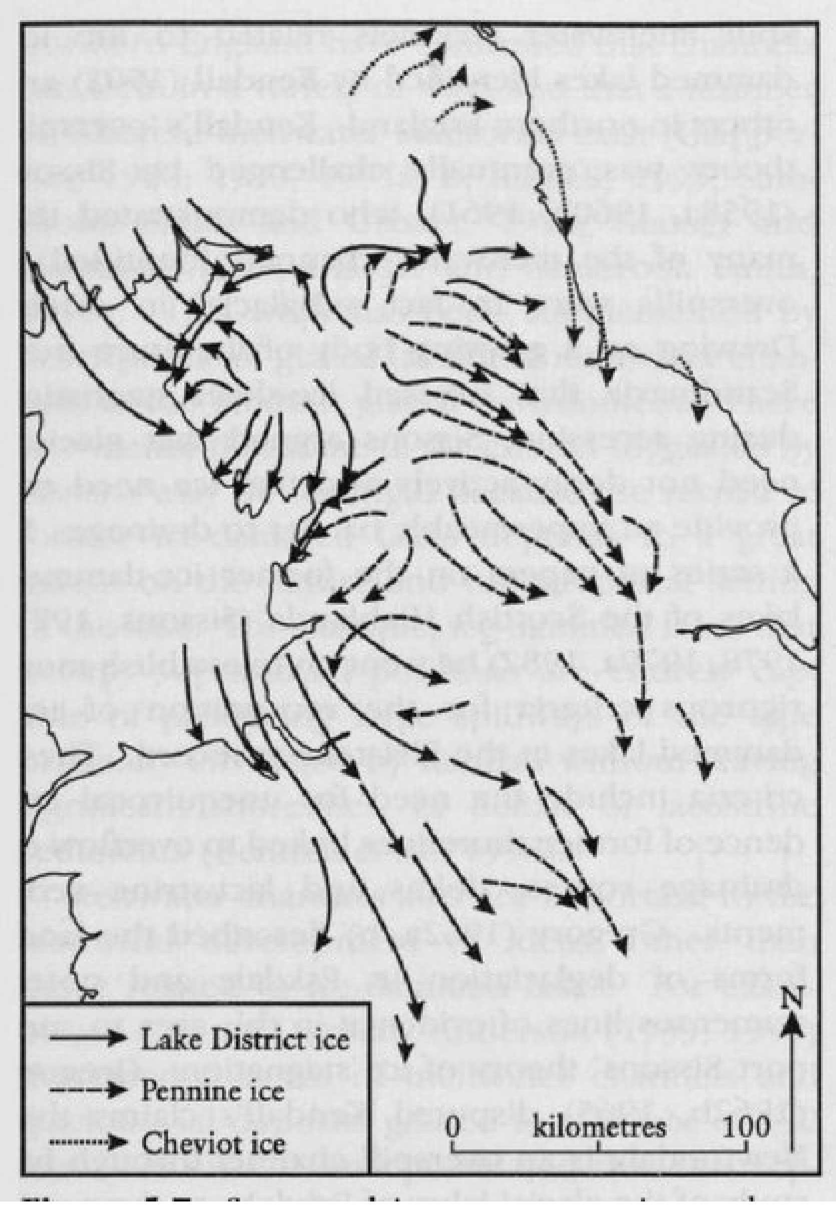
(Figure 5.4)a. Suggested Late Devensian ice movements in northern England (after Taylor et al., 1971). b. Suggested Late Devensian ice movements in northern England: generalized movements at various time periods after Letzer (1981). Note that the Early Scottish could be Early Devensian or Wolstonian and that the ice movement directions for the Scottish Readvance are incorrect (see Huddart, 1970, 1991, 1994, 1997).



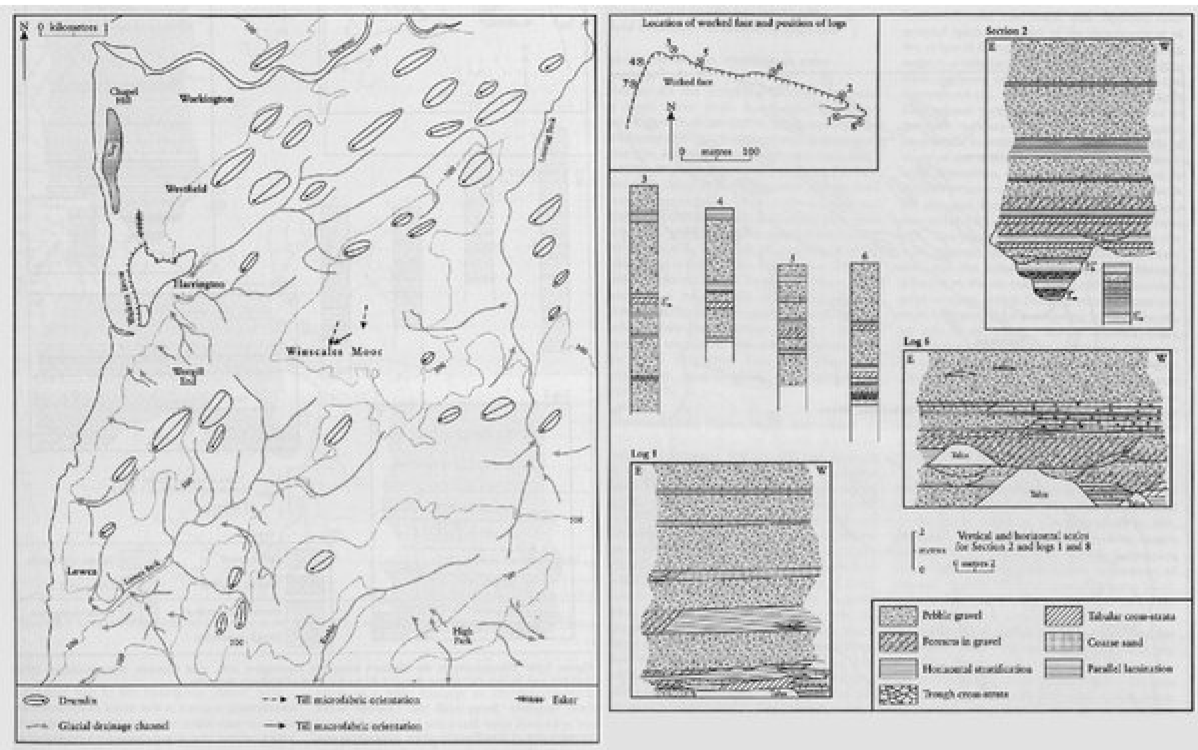
(Figure 5.5) Generalized ice flows and ice limits during the Heinrich I event in northern Britain (after McCabe et al., 1998). Note that the ice flows in Cumbria and east Yorkshire are considered to be incorrect by the present authors.



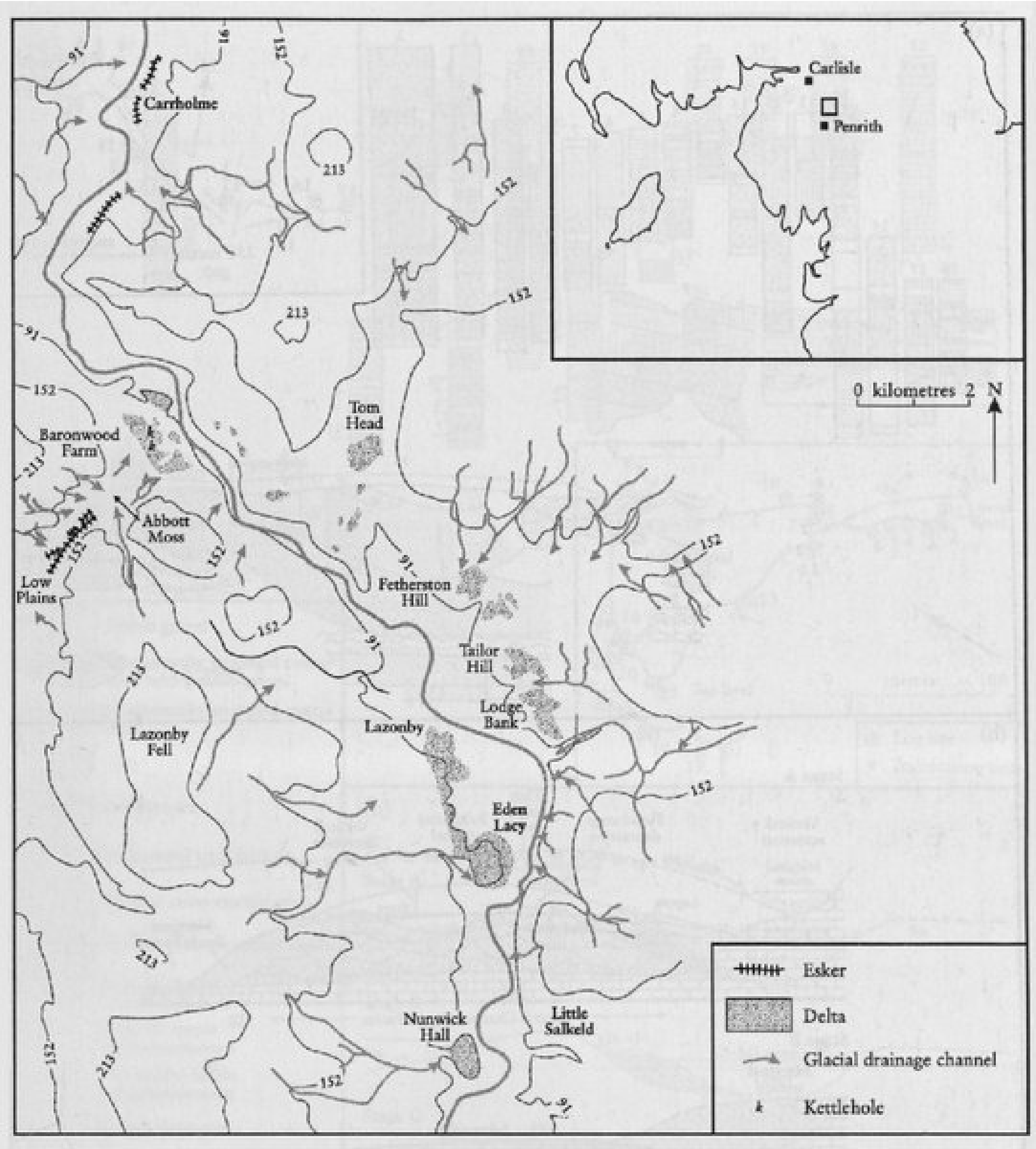
(Figure 5.6) Patterns of ice-sheet movement and glacial erosion. The five zones of erosional intensity are: 0, no glacial erosion; I, limited glacial erosion; II, glacial erosion confined to major flowlines; III, widespread glacial erosion; IV, very extensive glacial erosion, no trace of the pre-glacial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



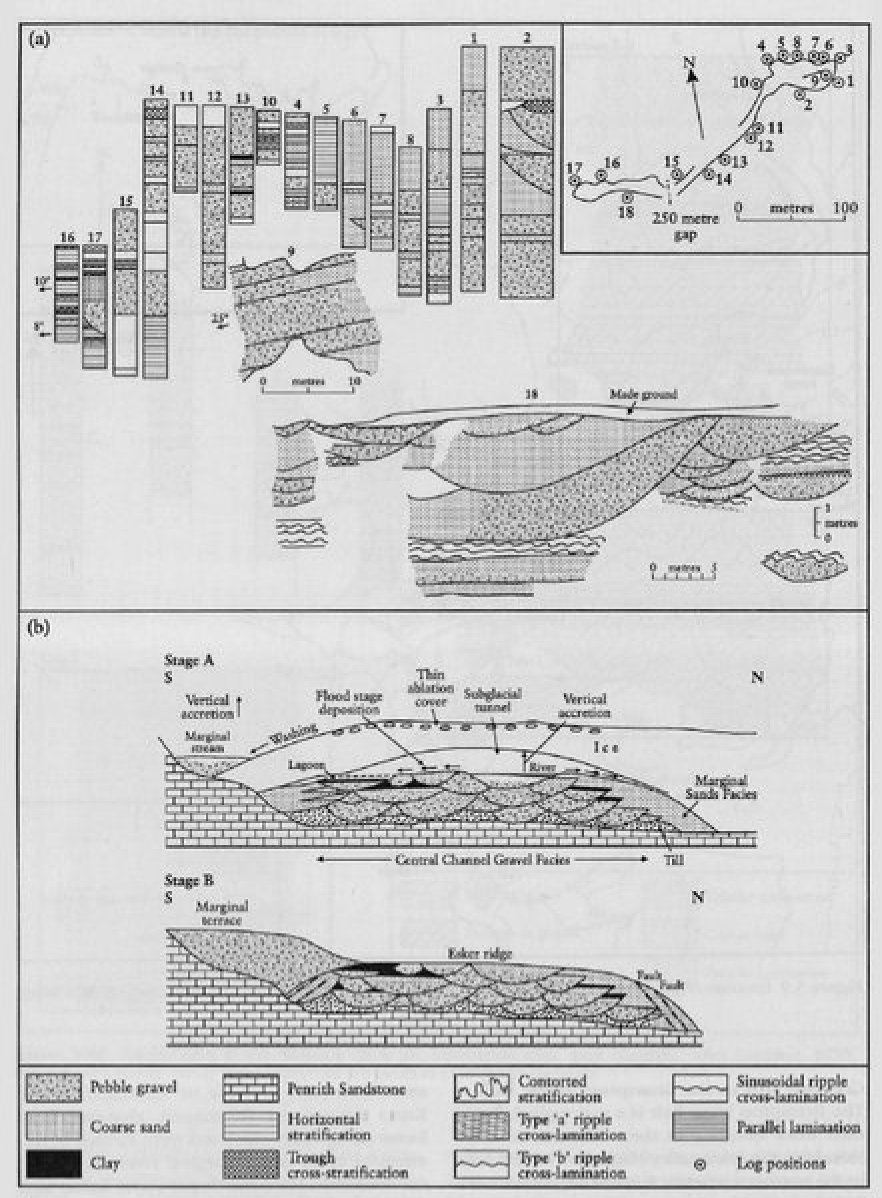
(Figure 5.7) Suggested ice movements in northern England based on erratic distribution (after Harmer, 1928).



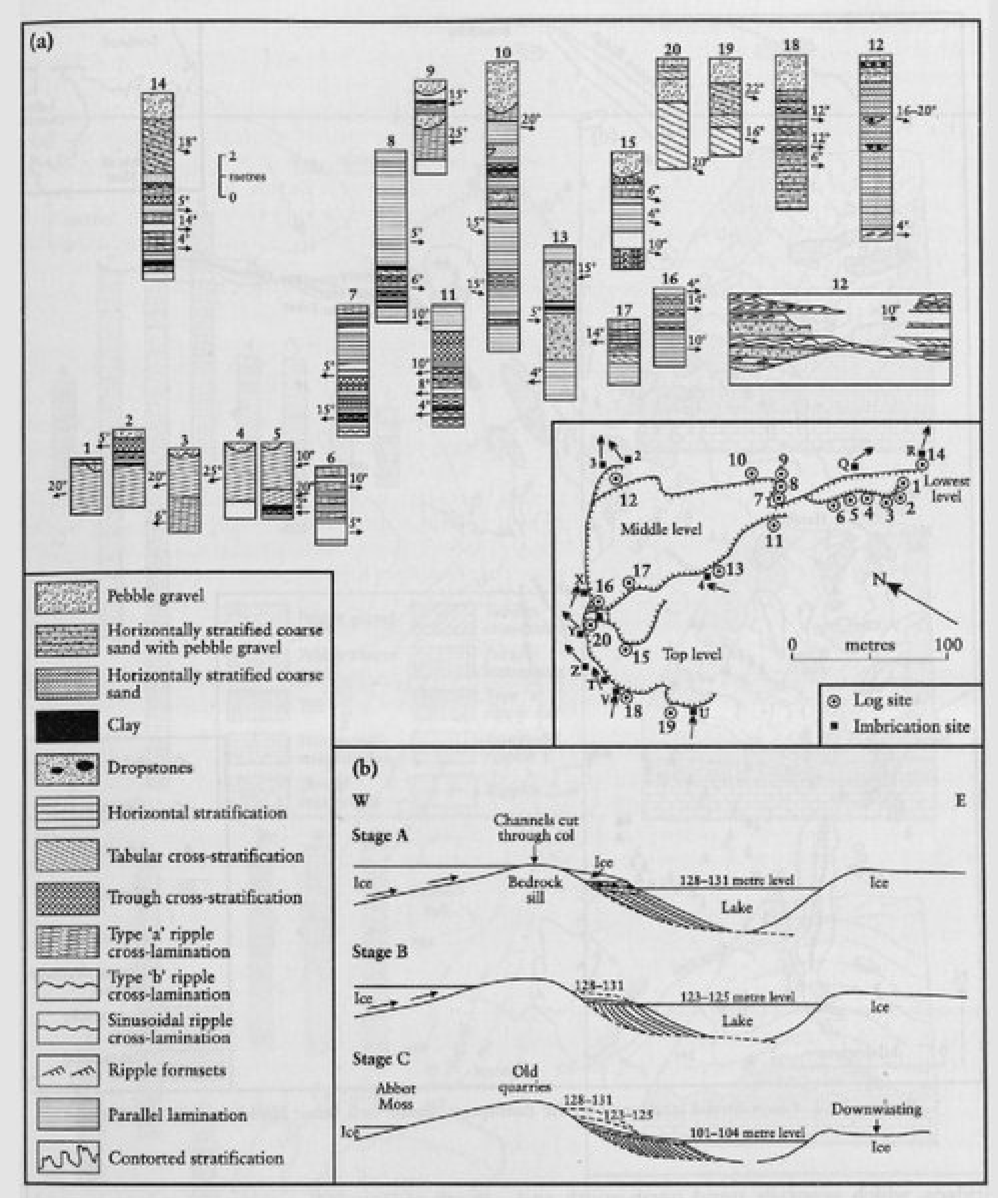
(Figure 5.8) a. Glacial landforms in the Harrington area, west Cumbria (after Huddart, 1970; Huddart and Tooley, 1972). See (Figure 5.8)b for stratigraphy of the Walker's Brow pit. b. Stratigraphy in the Walker's Brow pit, Harrington area, west Cumbria (after Huddart, 1970; Huddart and Tooley, 1972).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



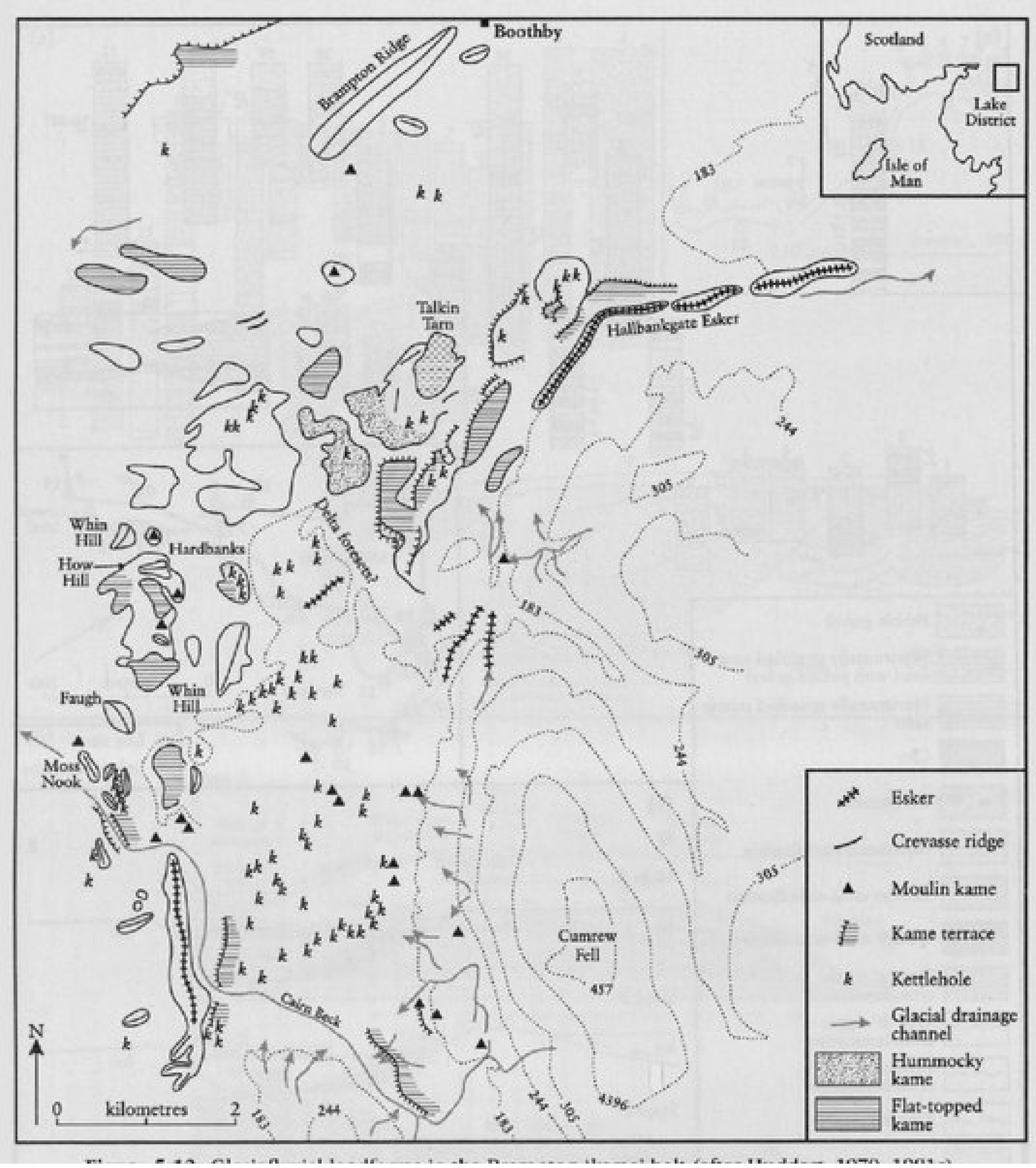
(Figure 5.9) Location of glaciofluvial landforms in the middle Eden valley, Cumbria (after Huddart, 1970, 1981c).



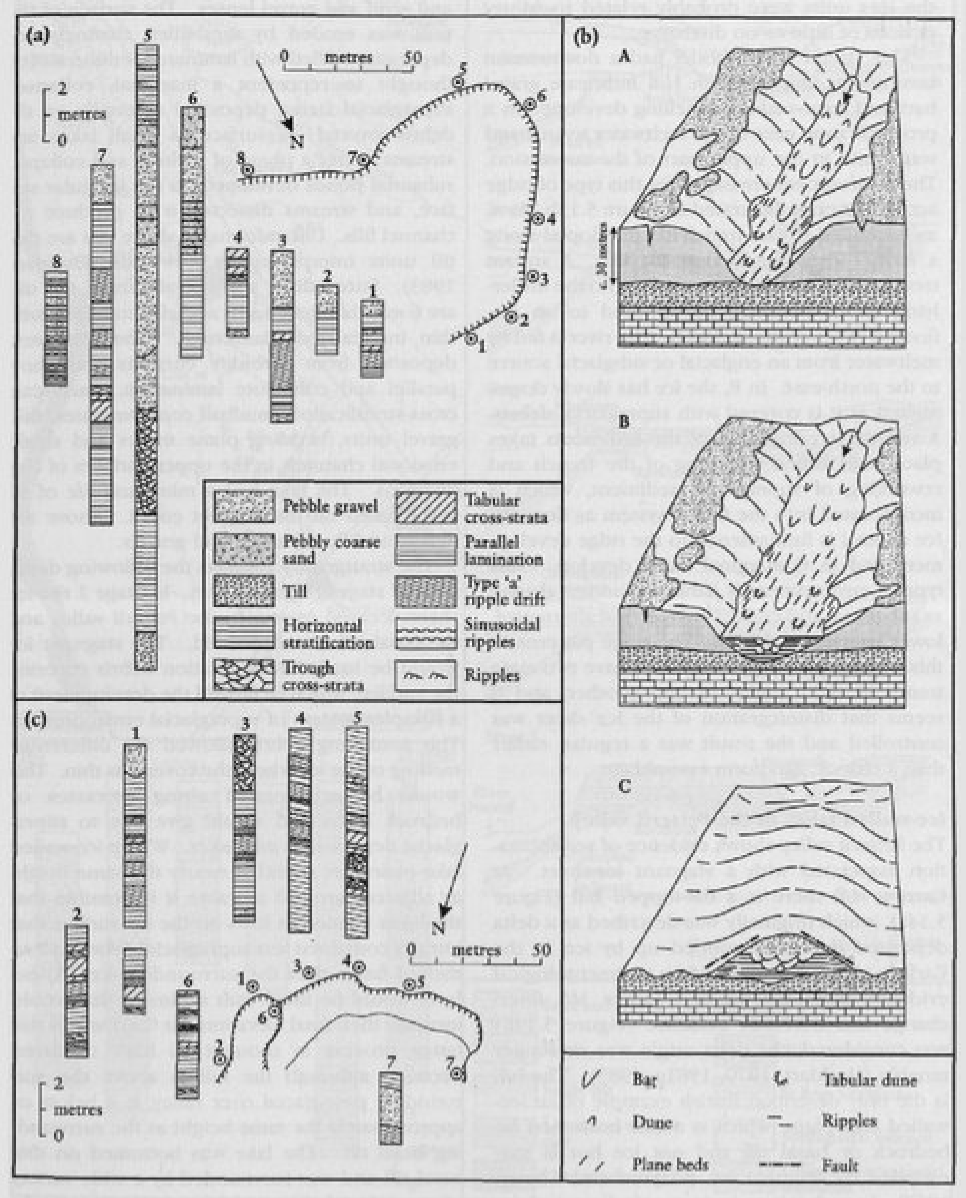
(Figure 5.10) Stratigraphy and model of subglacial sedimentation at Low Plains pit, Edenside (after Huddart, 1970, 1981c): (a) stratigraphy; (b) model of subglacial sedimentation.



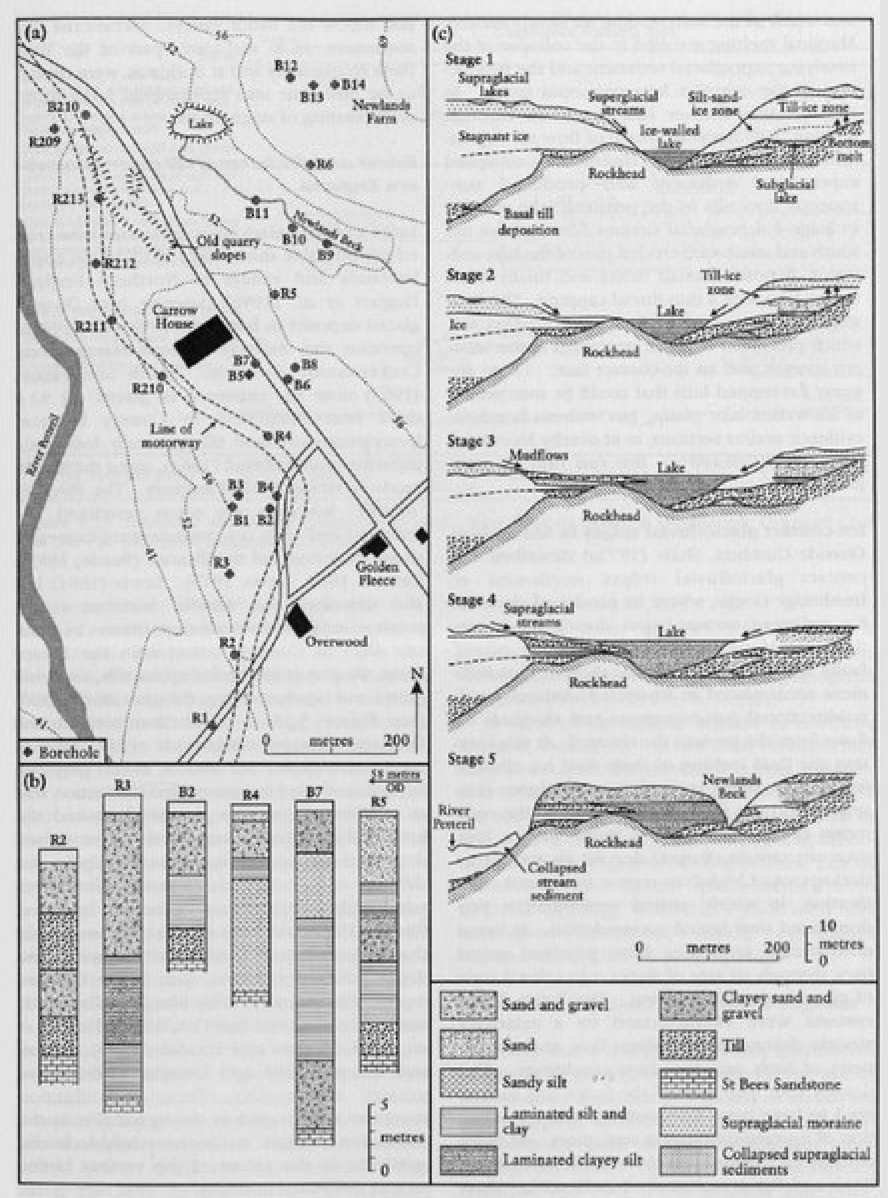
(Figure 5.11) Stratigraphy and model of glacio-lacustrine sedimentation at the Baronwood pit, Edenside (after Huddart, 1970, 1981c): (a) stratigraphy; (b) model of glacio-lacustrine sedimentation.



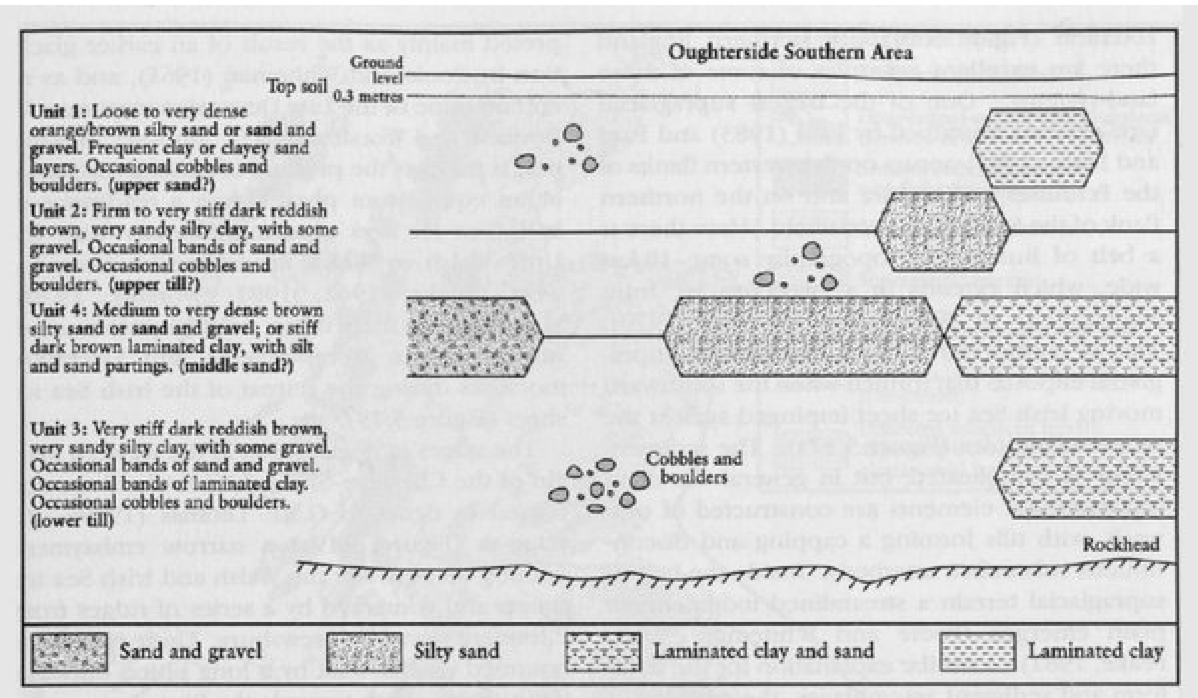
(Figure 5.12) Glaciofluvial landforms in the Brampton 'kame' belt (after Huddart, 1970, 1981c).



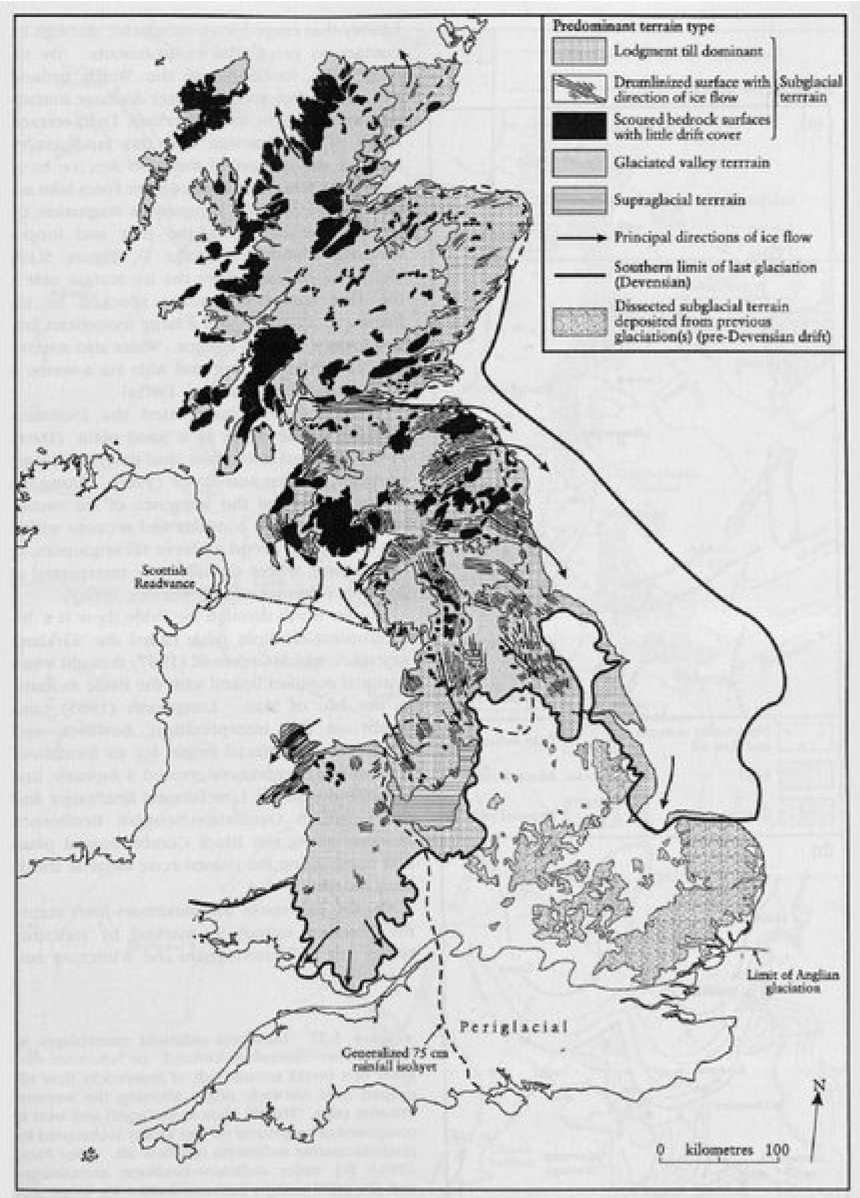
(Figure 5.13) Stratigraphy in ice-walled stream trench deposition, Brampton 'kame' belt (after Huddart, 1970, 1981c): (a) stratigraphy at Whin Hill, How Mill; (b) model of deposition at Whin Hill, How Mill; (c) stratigraphy at Faugh.



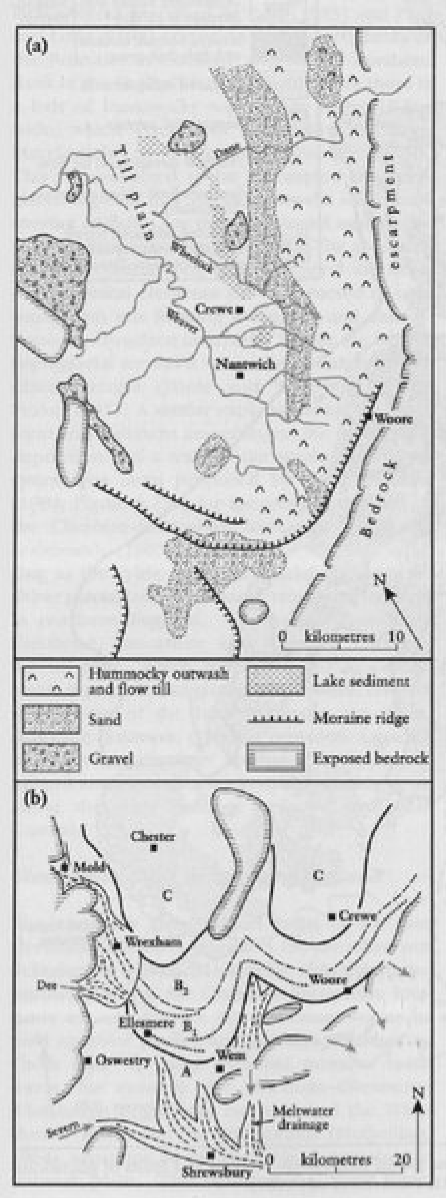
(Figure 5.14) Ice-walled lake plain landforms, sediments and development at Carrow Hill, Petteril valley, Cumbria (after Huddart, 1970, 1983): (a) morphology, borehole locations and M6 motorway interchange, Carrow Hill; (b) stratigraphy of circled boreholes in (a); (c) model of ice-walled lake plain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



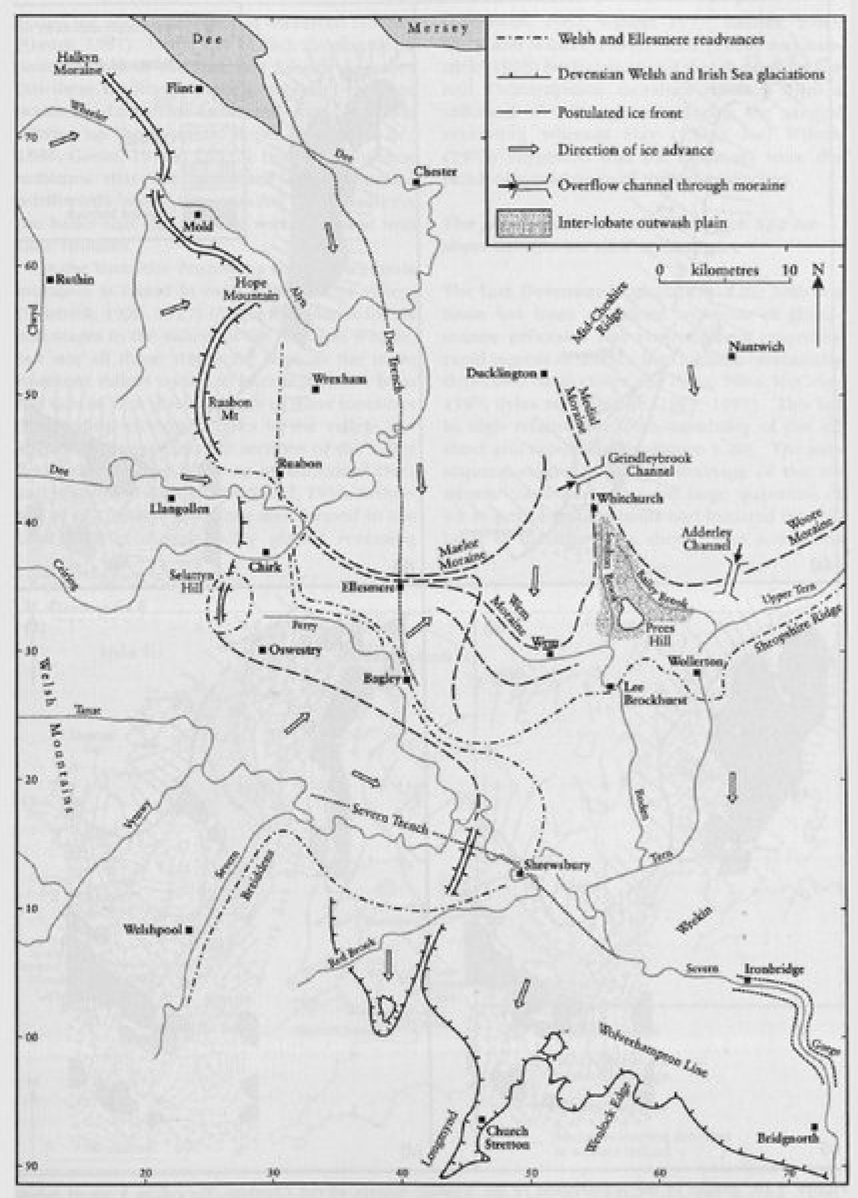
(Figure 5.15) Schematic glacial lithostratigraphy in inland west Cumbria based on Oughterside Opencast Coal Site [NY 126 400] (after Hughes et al., 1998).



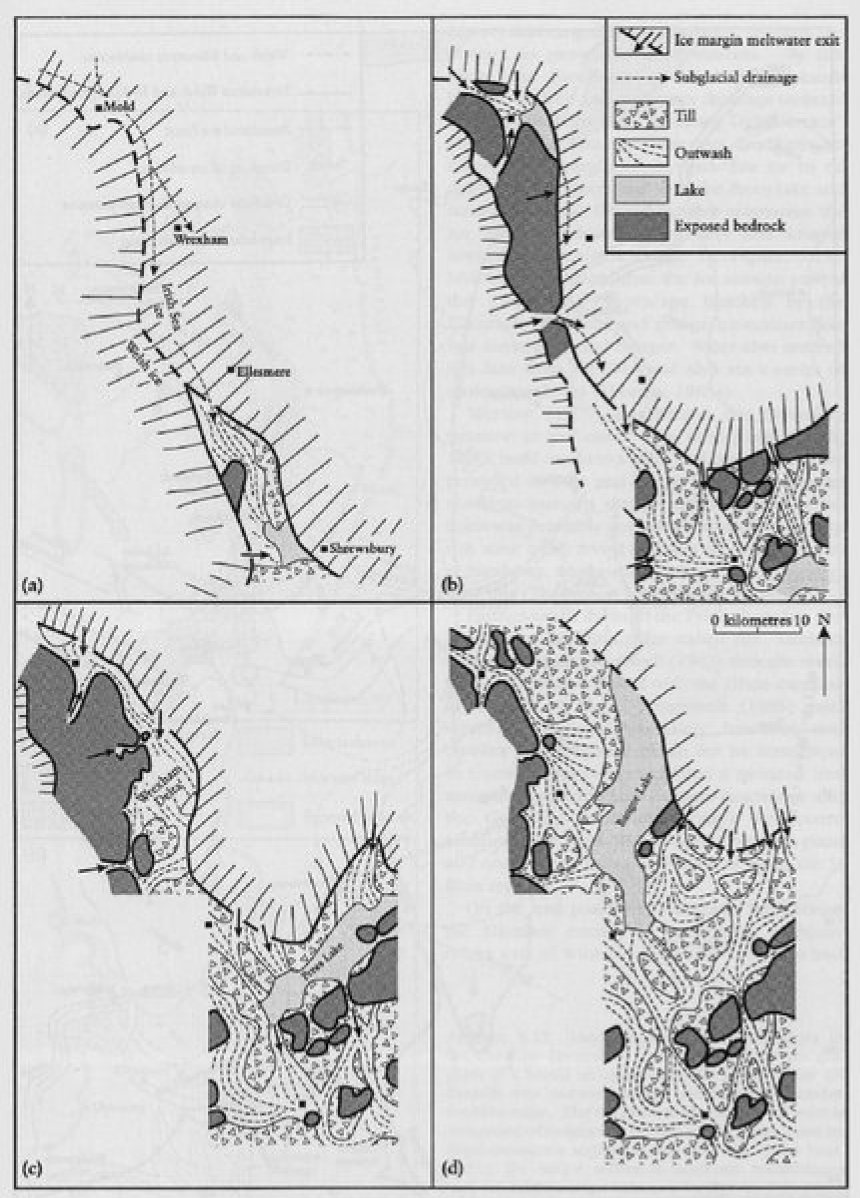
(Figure 5.16) Large-scale geomorphological zonation and glacial land-system terrains and limits of glaciations (modified from Huddart, 1970; Eyles and Dearman, 1981; Cameron et al., 1987).



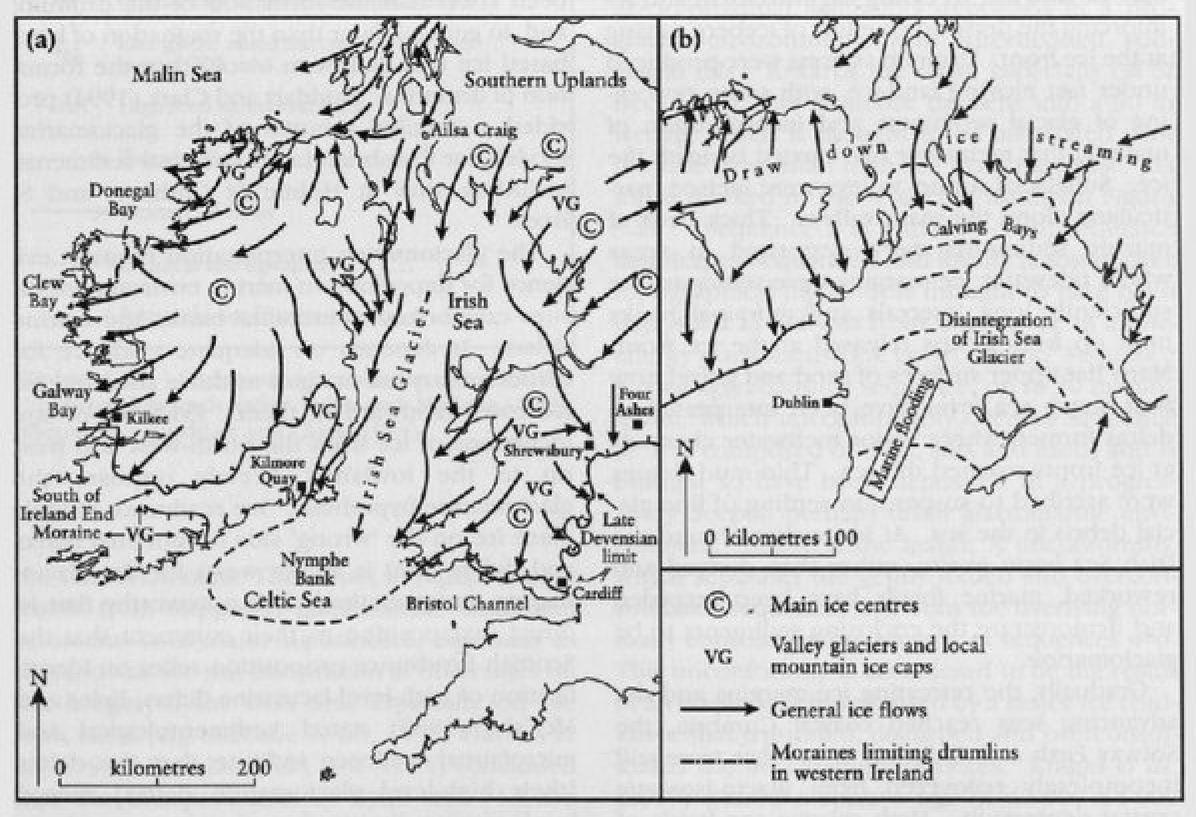
(Figure 5.17) Landform-sediment assemblages in the Cheshire–Shropshire lowland: (a) Schematic diagram of a broad arcuate belt of hummocky flow till draped over outwash ridges abutting the western Pennine edge. The till plain to the north and west is composed of lodgement till and locally is obscured 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.



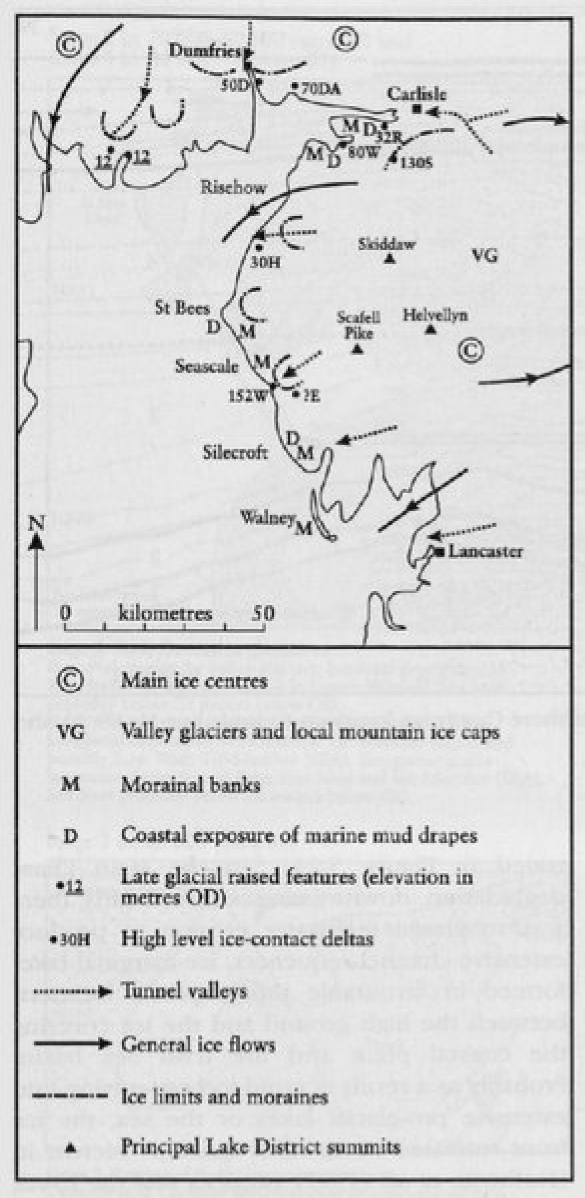
(Figure 5.18) Late Devensian glaciation in the Cheshire Plain and Welsh Borders (after Peake, 1981).



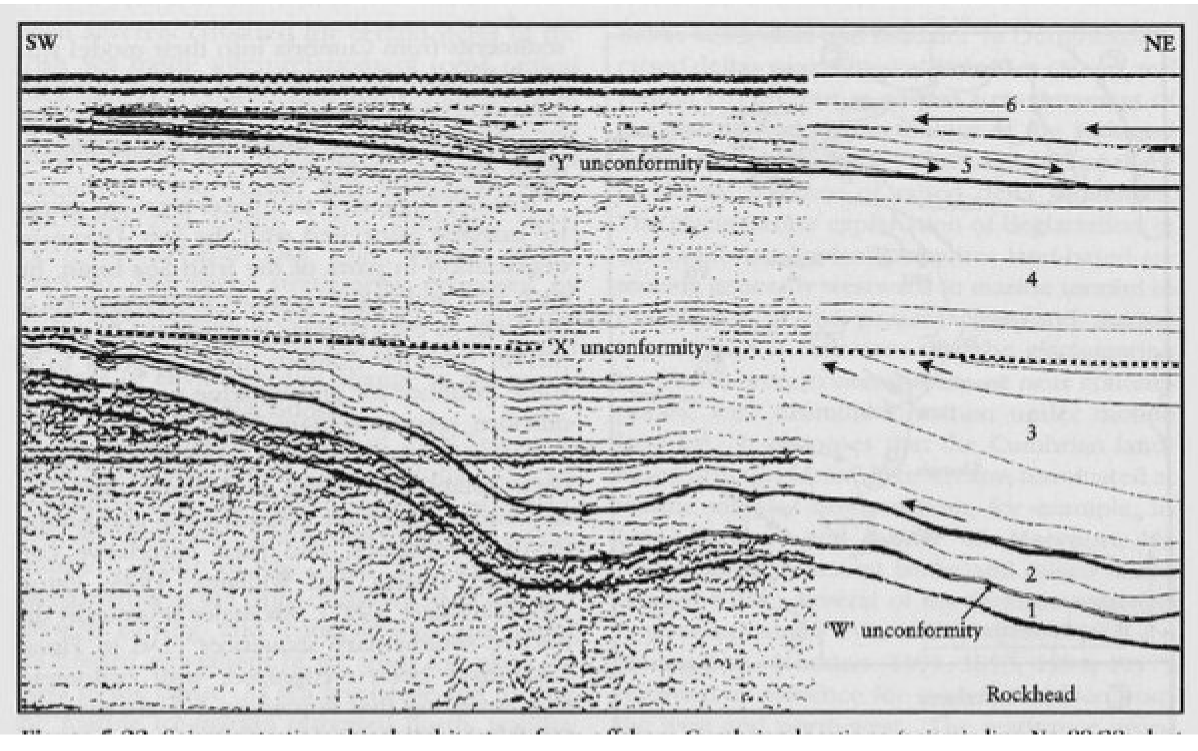
(Figure 5.19) Stages in the deglaciation of the western margin of the Cheshire–Shropshire lowland (after Thomas , G.S.P., 1989).



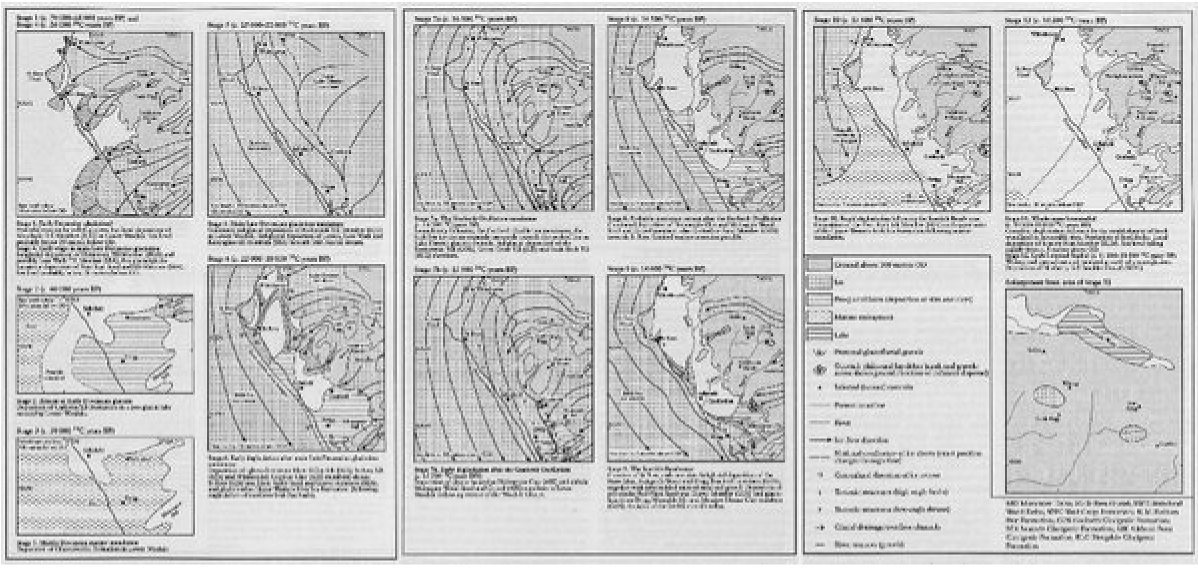
(Figure 5.20) Irish Sea palaeogeography during the Late Devensian glaciation (after Eyles and McCabe, 1989): (a) shows ice flow and dispersal centres; (b) shows the final stage of disintegration with a calving front.



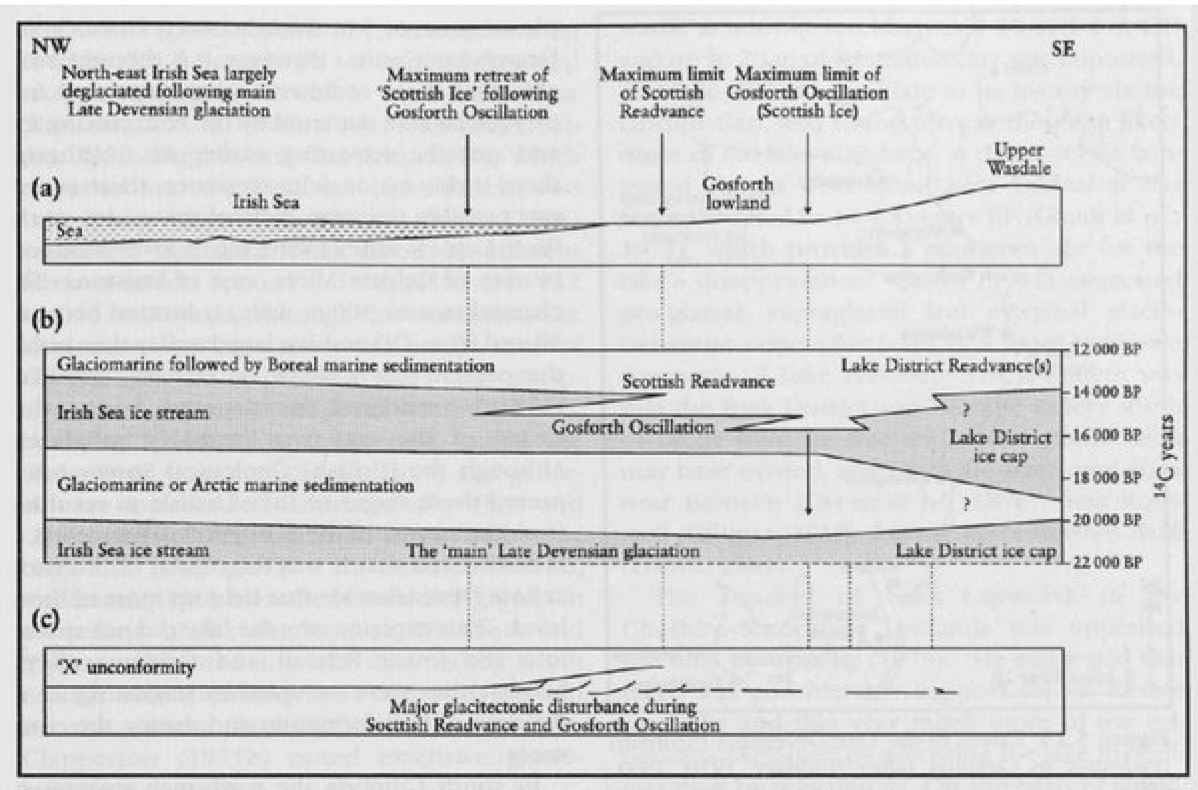
(Figure 5.21) Supposed glaciomarine sites in the Cumbrian lowlands (after Eyles and McCabe, 1989; discussed in Huddart and Clark, 1994).



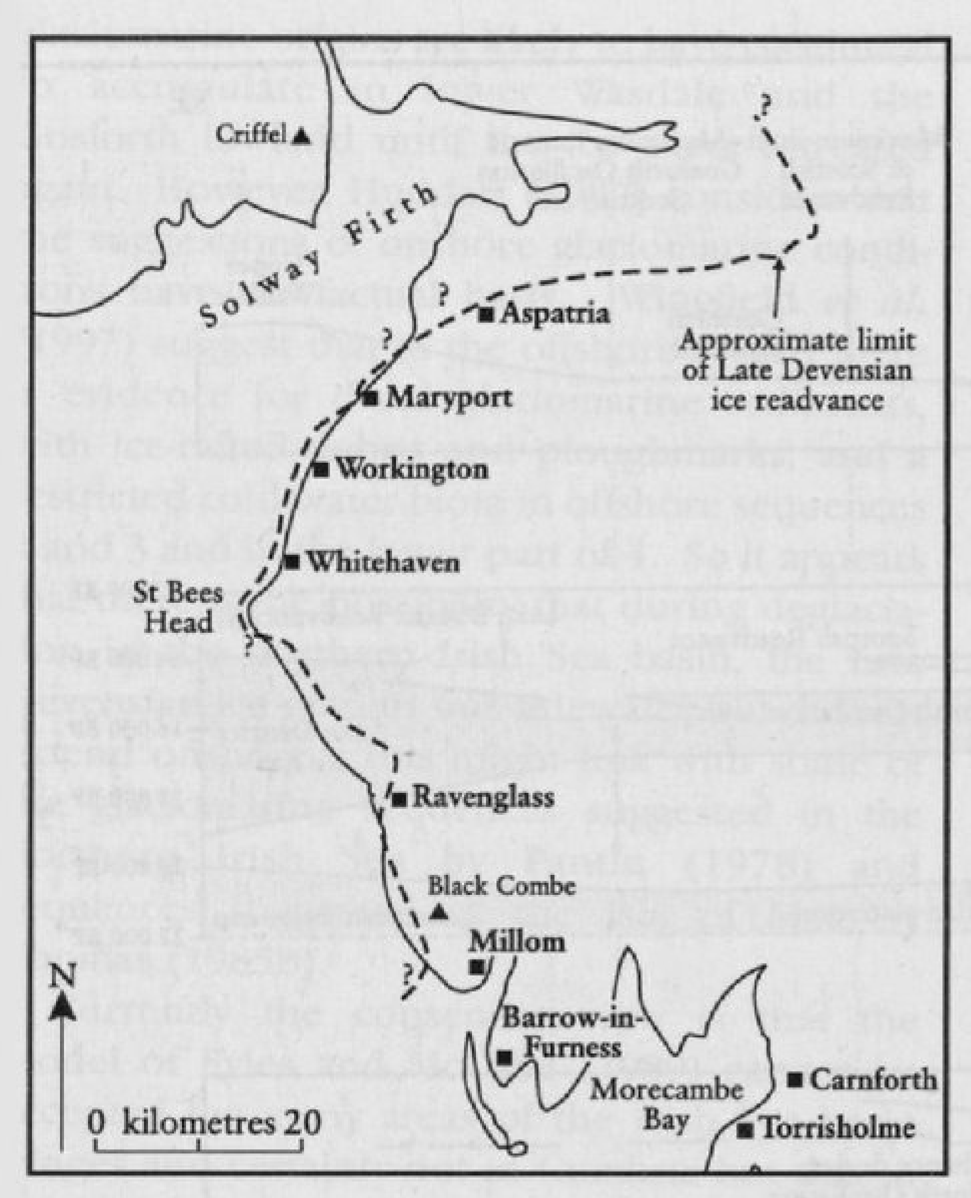
(Figure 5.22) Seismic stratigraphical architecture from offshore Cumbrian locations (seismic line No 89/29, shot points 2621–2630) (after Nirex, 1997b).



(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.)



(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).



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