# *Quaternary of Northern England*

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# Chapter 6

# The Late-glacial record of northern England

## INTRODUCTION

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As discussed in the previous chapter, northern England was heavily glaciated by successive ice advances during the Devensian and at the maximum ice-sheet expansion, ice cover extended to the whole of the region except the North York Moors, the Peak District uplands and parts of southern Yorkshire, Humberside and the Yorkshire Wolds (Johnson, 1985b; Worsley, 1985; Catt, 1991a, b; Douglas, 1991; Huddart, 1991, 1994). Other upland summits may have formed nunataks, although these areas had their own local ice centres. The areas that escaped actual ice cover must have been subject to extreme periglacial processes and ice-marginal environments (Watson, 1977). Climatic warming at the Last Termination (Björk *et al.*, 1998), at the end of the Devensian period, initiated rapid deglaciation and the creation of a complex suite of erosional and depositional landforms associated with ice wastage and glaciofluvial processes described in Chapter 5. These provided a wide range of depositional environments, which accumulated sediments preserving litho- and biostratigraphical evidence during the Late-glacial period, a phase of rapid climatic fluctuations and environmental changes between the end of full glacial conditions and the establishment of the stable, full interglacial conditions of the Flandrian. The Late-glacial deposits of northern England have been classified by Thomas (1999).

This chapter examines the lithofacies and biostratigraphical evidence for this extended period of environmental transition, broadly between 15 000 and 10 000 radiocarbon years BP. Within this broad time frame the timing, character and rate of Late-glacial environmental changes will have differed markedly across northern England (Watts, 1980), responding to local environmental factors and stimuli. The history of these environmental changes has been reconstructed using biostratigraphical proxies that have contrasting environmental sensitivities, such as pollen and insect analyses, and so have time-transgressive responses to climate change. Although chronostratigraphical divisions and subdivisions are used here to describe the history of the Late-glacial, the event stratigraphy terminology proposed by Björk *et al.* (1998) based on the GRIP ice-core oxygen isotope record is also used for inter-site correlation.

Litho- and biostratigraphical evidence for Late-glacial climatic change has long been recognized in northern England (Pennington, 1943, 1947; Blackburn, 1952). Lacustrine sediment sequences have formed in depressions in the glaciogenic diamicton and outwash sand and gravel deposits left across much of the landscape upon deglaciation, such as in kettle features formed by melting dead-ice blocks, in meltwater channels, or simply in hollows in the undulating morainic till plain. The ubiquitous Late-glacial succession is a tripartite division comprising two clastic units separated by a more organic sequence, with the clastic units interpreted as forming under severe, cold climate conditions interrupted by the organic sequence representing sedimentation under more temperate, interstadial environments. Sites displaying this tripartite sequence are present throughout the region, except in a few areas at high altitude, where they have been destroyed by local glacier advance under the extreme cold conditions of the Younger Dryas Stadial phase at the end of the Late-glacial period. Late-glacial sites are even preserved offshore, or in the intertidal zone, where they have been submerged by Flandrian sea-level rise. For example, at Rossall Beach in Lancashire, Late-glacial kettlehole deposits exposed in the intertidal zone contain a Late-glacial type pollen-assemblage dominated by non-tree taxa, and are dated to 12 320 years BP (Tooley, 1985).

The lithological evidence of a Late-glacial temperate interlude has been confirmed by many pollen analyses of land vegetation changes (cf. Pennington, 1977), by insect studies (Ashworth, 1972; Coope, 1977; Coope and Joachim, 1980; Walker *et al.*, 1993; Lowe and Walker, 1997b; Hughes *et al.*, 2000) and by a range of other proxies, notably plant macrofossils (Bush, 1993; Hughes *et al.*, 2000), molluscs and ostracods (Keen *et al.*, 1984; Thew and Woodall, 1984), or chemical and loss-on-ignition studies (Pennington, 1970). The temperate Windermere Interstadial in northern England (Coope and Pennington, 1977) was regarded as equivalent to the Bølling–Allerød Interstadial complex of north-west Europe, and the post-interstadial cold phase, in north-west Europe termed the 'Younger Dryas' (Lowe *et al.*, 1995a), was designated the Loch Lomond Stadial in Britain. Dating by radiocarbon of the biozones (pollen zones I to III) associated with these climatic phases allowed them to be regarded as Late-glacial chronozones, although their boundaries were clearly not synchronous everywhere (Walker *et al.*, 1994). Radiocarbon chronostratigraphical control on the Late-glacial in northern England has, until recently, been of coarse resolution, even though a few sites had many dates at close intervals, such as Blelham Bog (Pennington, 1975a) and Low Wray Bay (Pennington, 1977) in Cumbria, which produced a coherent series and provided a good basis for understanding Late-glacial chronology. Most sites have had very few dates or none at all. Many Late-glacial radiocarbon datings on bulk sediments may have suffered from the problems inherent in the resolution of the method, as a result of possible 'hard water' error, by the incorporation of inwashed older material into the limnic deposits and as a result of local geochemical peculiarities (Lowe, 1991), as shown by the initial poor results obtained from Gransmoor in Holderness and Whitrig Bog by the Scottish Border (Walker *et al.*, 1993; Lowe *et al.*, 1995b; Mayle *et al.*, 1999). Although a series of accelerator mass spectrometry (AMS) $^{14}C$ dates on terrestrial macrofossils has largely resolved the dating of the Gransmoor site (Lowe *et al.*, 1995b), the Late-glacial radiocarbon record from northern England as a whole must be regarded as imprecise and of coarse resolution, and individual dates require critical evaluation.

Recent high-resolution study of ice-core and proxy climate records (Björk *et al.*, 1998; Mayle *et al.*, 1999) has made it clear that, although the simple threefold division of the Late-glacial is valid at the broad scale, the climatic and environmental changes during this period were much more complex, with many finer-scale oscillations. Several minor climatic reversals of varying duration and intensity occurred during the three major Late-glacial stadial and interstadial periods, operating at least at a millennial scale and even of shorter duration (Lowe *et al.*, 1994a; Björk *et al.*, 1998). Thus two brief colder phases can be recognized within the Late-glacial interstadial phase (Björk *et al.*, 1998), as well as a major cold oscillation against the warming trend of the pre-interstadial phase (McCabe *et al.*, 1998). The post-interstadial 'Younger Dryas' Loch Lomond Stadial is also capable of subdivision (Walker, 1995). Lowe *et al.* (1994a, 1999) have proposed a provisional event stratigraphy for the British Isles, which uses time-equivalent markers based upon fossil-insect data and ice-core records (Atkinson *et al.*, 1987; Walker *et al.*, 1993; Lowe *et al.*, 1995b; Lowe and Walker, 1997b). Two of the key sites from which high-precision data have been recovered to compose this classification scheme are directly relevant to the northern England region; Whitrig Bog, just beyond the northern English border (Mayle *et al.*, 1999) and Gransmoor in Holderness (Walker *et al.*, 1993; Lowe *et al.*, 1995b), both of which have high-resolution AMS radiocarbon chronologies. As precise correlation between the British records and the Greenland ice-cores has not yet been accomplished (Lowe *et al.*, 1994a, 1999) the timescale used in this overview remains in radiocarbon years BP, and so is easily comparable with the radiocarbon dating archive from the region, although Lowe *et al.* (1999) recently have published a calibrated radiocarbon timescale for the period. The following climate shifts have been proposed as regionally equivalent events in northern England (Lowe *et al.*, 1994a; McCabe *et al.*, 1998), although as the dates shown for their transitions remain approximate, correlations with GRIP event terminology (Björk *et al.*, 1998) are also shown in parenthesis, and used in the text: 15.0 to *c.* 14.5 ka (GS-2b), a warming trend after the Devensian Termination; *c.* 14.5 to 14.0 ka (GS-2a), significant cooling correlating with meltwater Heinrich Event 1 and prompting limited ice advance in the north of northern England; 14.0 to 13.0 ka (later GS-2a), gradual warming and ice retreat; 13.0 to 12.4 ka (GI-1e), abrupt warming and steep thermal rise; 12.4 to 12.0 ka (GI-1d), a cooling phase; 12.0 to 11.4 ka (GI-1c), a warming phase; 11.4 to 11.2 ka, (GI-1b), a brief cooling phase; 11.2 to 10.9 ka (GI-1a), a warming phase; 10.9 to 10.5 ka (GS-1), a very cold phase prompting mountain glacier advance; 10.5 to 10.0 ka (also GS-1), a cold and arid phase with indications of warming towards the end; 10.0 ka onwards (Flandrian Interglacial), abrupt warming and steep thermal rise. The timing, rate and character of environmental responses to these climatic oscillations will have varied considerably across the region because of local factors, but this provisional event chronology provides a framework within which the Late-glacial evidence from sites in northern England can be compared and discussed.

## The pre-interstadial Late-glacial 15–13 ka (GS-2)

Deglaciation of northern England after the Dimlington Stadial was time-transgressive and locally complex but rapid, and it seems probable that ice removal had been accomplished across almost all of the region during the first half of this period, although small glaciers may have lingered in the highest uplands. The depositional signals of the initial post-deglaciation period are abundant landslips, debris flows and mudslides owing to failure of unstable slopes under extreme periglacial conditions (Mitchell, 1991a). For example, major landslipping took place on the scarp slope of the North York Moors at Gormire Lake during the early Late-glacial (Blackham *et al.*, 1981). Non-fossiliferous coarsely laminated clays beneath Late-glacial sequences in Cumbria are evidence of snow-melt conditions from wasting glaciers probably early in this period, but no secure deglaciation chronology is available. Evidence for ice re-advance in the northern Irish Sea basin has caused McCabe *et al.* (1998) to propose a cold phase beginning about 14.5 ka instigated by the Heinrich Event 1 meltwater pulse, which may have reversed the effects of deglaciation in north-west England for several centuries. Evidence for a Late-glacial readvance in the west and north Cumbrian lowlands (Huddart, 1970, 1991, 1994, 1997) during post-Dimlington Stadial ice wastage may support this, although its chronology remains conjectural. Ruddiman *et al.* (1977) have proposed significant warming in the eastern North Atlantic area between 14 and 13 ka (later GS-2a), which began a gradual amelioration of climate leading up to the interstadial proper, allowing the start of soil stabilization and pioneer vegetation. The earliest radiocarbon dates for this process come from small enclosed kettlehole sites in Cumbria. At Blelham Bog (Pennington, 1970; Pennington and Bonny, 1970) the lowest organic muds have duplicate radiocarbon dates of 14330 ± 230 years BP and 14280 ± 230 years BP. These are very early but are supported by a consistent series of dates above them that span the Late-glacial as a whole (Pennington and Bonny, 1970).

Another long series of dates on the Late-glacial sequence from nearby Low Wray Bay (Pennington, 1977, 1996) is less coherent, but has basal dates of similar very early age, 13 938 ± 210 years BP to 14557 ± 280 years BP and 14 623 ± 360 years BP. Pennington (1977) suggested that these lower dates at Low Wray Bay may well be too old by about 500 years owing to hard-water error and analogous dates from Blelham Bog may be so too, although Blelham Bog is in a non-calcareous catchment. The dates' correction would bring the earliest organic sedimentation into line with the proposed start of warmer conditions around 14 ka, which is perhaps a more plausible age that would produce very coherent matching series of dates from these two key sites for the whole of the Late-glacial. These very early organic facies may reflect the more oceanic conditions of western Britain in contrast to the continued continental conditions to the east.

Although ice cover was less complete in east Yorkshire at the glacial maximum, there are no comparably early dates on deglaciated terrain after 15 ka. A radiocarbon date from a kettlehole at Kildale Hall in Cleveland of 16 713 ± 340 years BP is on moss fragments beneath marl and resting upon silty clay over glaciofluvial gravels. If correct, this date suggests very early deglaciation indeed, but the high probability of hard-water error at this site makes this early date questionable. More reliable dates for first post-deglaciation organic deposition in the east of the region cluster around 13 ka, perhaps because of a more continental climate. More research is required to test this apparent west–east dichotomy.

Stabilization of the environment after deglaciation would have been a slow process. The climatic evidence is for some warming from 15 ka and then more significantly after 14 ka (Ruddiman *et al.*, 1977; Lowe *et al.*, 1994a) and this is supported by evidence from caves in west Yorkshire (Gascoyne *et al.*, 1983; Atkinson *et al.*, 1986) that suggests the start of speleothem deposition, and thus unfreezing of groundwater, around 15 ka and increased speleothem deposition from about 14 ka BP throughout the Late-glacial. Some considerable delay would have occurred in the response of ecological systems to such limited temperature increases. Comparatively cold, arid conditions still existed (Atkinson *et al.*, 1987), which are reflected in the biological data. A transition from barren polar desert to-snow bed and sedge-tundra biota as solifluction ceased allowed the colonization of raw skeletal soils by Gramineae, Cyperaceae and *Salix herbacea*. Lack of radiocarbon dating control for this early stage in Late-glacial history makes its recognition insecure in many cases, but a significant number of sites with deposits almost certainly of this pre-interstadial period (event GS-2a) have been described in northern

England. Apart from the limited organic deposition recorded in the Windermere area, in all cases sediments are of fine-grained silts and clays, often laminated, with a very low organic content owing to the low humus status of the soils. Star Carr in the Vale of Pickering (Day, 1996) is a typical example, where the pioneer grass–sedge tundra flora is joined by open ground taxa *Rumex*, *Artemisia* and *Thalictrum* as conditions improve and soils stabilize. As with most pollen records some *Betula* and *Pinus* are present from the beginning, but this is almost certainly the result of long-distance transport, with any local *Betula* growth being *B. nana*. The very open, broken ground flora of this early phase diversified as the period progressed, with increases in herb taxa such as *Helianthemum* indicating a more closed, stable grassland type on base-rich soils. High *Helianthemum* frequencies are characteristic of this period (Pennington, 1977) at many sites throughout northern England (Blackburn, 1952; Beales, 1980; Hunt *et al.*, 1984). Community succession and increasingly congenial climate later in event GS-2a, after about 13.8 ka, prompted the patchy spread of low shrubs *Empetrum*, *Juniperus* and *Hippophäe*, but plant cover was still mainly grass and tall herb steppe, with low pollen accumulation rates indicating low vegetation productivity. These pre-interstadial pollen assemblages (Pennington and Lishman, 1971) are characterized by low taxa diversity and low pollen productivity. Shrub cover thickened as the interstadial approached and in places tree *Betula* seems to have been present in the later pre-interstadial phase. North of Morecambe Bay seems to have been an area of early tree birch colonization (Johnson *et al.*, 1972), and Pennington (1981) has reported *Betula pubescens* fruits and catkin scales in sediments from Windermere older than 13 ka. Some altitudinal differentiation in vegetation type is apparent in the later pre-interstadial phase (GS-2a), with upland sites such as Sty Head Tarn in Cumbria (Pennington, 1996) continuing to have a very open pioneer steppe-tundra flora with *Artemisia* and *Salix herbacea*, but in the lowlands a *Rumex*–Gramineae association became important prior to the expansion of low shrub taxa, as at Bleham Bog and Low Wray Bay (Pennington, 1970, 1977), Green Lane, Furness (Johnson *et al.*, 1972), in the Eden Valley at Moorthwaite Moss and Abbott Moss (Walker, 1966b), and elsewhere in northern England (e.g. Beales, 1980).

In this and all subsequent Late-glacial phases pollen of several thermophilous forest taxa, notably *Corylus* and *Alnus*, has been found at many sites throughout northern England. A localized presence of these taxa in the Late-glacial cannot be ruled out altogether, and is supported by reports of *Alnus* macrofossils from the late interstadial at Willow Garth in east Yorkshire (Bush and Hall, 1987) and from Hawes Water in Cumbria (Oldfield, 1960a, b). A strong case against the presence of such thermophilous trees has been made by Tallantire (1992), however, and the recovery of pollen of *Corylus*, *Alnus* and other forest trees from the till beneath the Late Glacial sequence at Skipsea Withow Mere (Hunt *et al.*, 1984) suggests that reworking may be responsible for these enigmatic pollen records, if not long-distance transport as routinely accepted for *Pinus* pollen. Incorporation of till-derived mineral material into sediment profiles continued throughout the Late-glacial, shown by the low organic content of most interstadial limnic deposits.

It is probable that during rapid deglaciation in the early part of this phase, meltwater drainage created both erosional and depositional environments. Erosional channels produced by ice-marginal meltwater are common on the flanks of the Pennines and on the North York Moors and some contain Late-glacial sediments (Yates and Moseley, 1958; Johnson *et al.*, 1970; Jones 1978; Glasser and Sambrook Smith, 1999). Major results of deglaciation include pro-glacial lakes such as Lake Humber in the southern Vale of York, and impounded waters in this and similar lakes, such as in the Vale of Pickering and the lower Tees, laid down great thicknesses of laminated clays in the earlier Late-glacial (Gaunt, 1981). Although an organic buried soil at West Moor rests upon Lake Humber clays and its radiocarbon date of 11 100 ± 200 years BP provides a minimum date for the final silting up of this lake (Gaunt *et al.*, 1971), it remains conjectural as to how long Lake Humber and similar water bodies (Plater *et al.*, 2000b) lasted into the Late-glacial and further research is required to clarify the history of this important sedimentary environment.

## The earlier Late-glacial interstadial 13–12 ka (GI-1e and GI-1d)

A very abrupt rise in temperature marks the start of the Late-glacial Interstadial in the Greenland ice-core data at 13 ka (Björk *et al.*, 1998; Lowe *et al.*, 1999) and the following climatic warm phase comprises event GI-1e. This sharp thermal rise is reflected very clearly in the lithofacies and biostratigraphical data from sites in northern England. The earliest organic accumulation in several lake basins, evidence of soil stabilization, vegetation of the catchment and increased biological production within the lake, has radiocarbon dates that match this chronology very closely. At the well-dated site of Blelham Bog in Cumbria, at a point interpolated as 13 000 years BP (Pennington, 1975a), open-ground herb frequencies fall and *Juniperus* rises sharply. The total pollen influx rate increases abruptly (Pennington, 1973; Pennington and Bonny, 1970). The same sequence of events occurs at Blea Tarn and at nearby Low Wray Bay at a dated level of 13 185 ± 170 years BP. Dates for initial deposition of organic sediments in the east of the region of 13 045 ± 270 years BP from The Bog, Roos (Beckett, 1981) and 13 042 ± 140 years BP from Seamer Carrs (Jones, 1976a) are closely comparable. At the King's Pool, Stafford the date on the basal contact of organic deposits was rather earlier at 13 490 ± 375 years BP (Bartley and Morgan, 1990), perhaps because of its more southerly location, although the larger standard deviation makes the age more difficult to interpret. Other sites in northern England preserve thin organic deposits below the main interstadial organic unit, which, although undated, are very probably of analogous date around 13 ka and which show the increase in shrub and tree birch pollen associated with event GI-1e elsewhere. These include Skipsea Bail Mere in Holderness (Flenley, 1984, 1987) and Hawes Water in Lonsdale (Oldfield, 1960a, b). Dating imprecision makes correlation with the Bølling Interstadial of north-west Europe unwise (Pennington, 1975a), but these organic layers and early dates demonstrate the existence of two discrete temperate phases in the Late-glacial, as shown by the GRIP ice-core and coleopteran record (Björk *et al.*, 1998).

Evidence from beetle (Lowe *et al.*, 1999) and molluscan (Thew and Woodall, 1984) data support the ice-core evidence (Mayle *et al.*, 1999) in indicating that during this initial interstadial temperate phase, event GI-1e, temperatures were easily the highest of the whole Late-glacial. A gradual decline in temperature then commenced, which continued until the steep Flandrian thermal rise at *c.* 10 ka, with cold events and thermal recoveries superimposed upon this trend. Although the later warm event GI-1c coincided with maximum vegetation development and organic sedimentation, this was the result not of optimum climate but of time lags in ecological systems such as soil maturation, moisture regimes and plant migration rates (Pennington, 1986). The early event GI-1e had much the more benign climate and the rapid thermal rise at 13 ka stimulated expansion of a rich vegetation cover. Even where the increase in organic sedimentation is insufficient to be visible in the profile, great increases in pollen accumulation rate occur, as at Blelham Bog (Pennington, 1975a), as a result of this change. Herbaceous steppe-tundra and tall-herb associations on raw soils, with taxa such as *Helianthemum*, *Rumex* and *Thalictrum*, as well as sedge and grassland, were characteristic, and formed the initial interstadial vegetation phase, into which *Juniperus* spread quickly. *Juniperus* and *Betula* were present from the start of the phase and began to expand. In places, such as Tadcaster (Bartley, 1962), tree birches must have increased quickly, but in many Late-glacial diagrams a phase of very high *Juniperus* pollen frequencies and concentrations up to about 12.5 ka is the distinctive feature (Bellamy *et al.*, 1966), indicating abundant shrub juniper across much of the landscape. Organic accumulation in the form of coarse detritus mud began at St Bees in west Cumbria at 12 560 ± 170 years BP (Coope and Joachim, 1980), although the coleopteran curve indicates a decline in temperatures from the 13 ka maximum. By 12.5 ka, even as temperature was declining, tree *Betula* copses had become established in most areas in a succession to birch–juniper park-tundra, although in many less favourable areas *Juniperus* dominance continued. Where data are available from high altitude, as in the Lake District (Pennington, 1964, 1973, 1996), a stable grassy herbaceous tundra vegetation remained unchanged throughout the entire interstadial. Elsewhere, successional changes towards greater shrub–woodland cover were time-transgressive across the region.

The development of increasingly dense *Juniperus* and *Betula* parkland in event GI-1e was interrupted by a period of colder climate for a few centuries before about 12 ka. This cold phase, event GI-1d, is recorded distinctly in several northern England pollen diagrams as a vegetation reversion, in which *Betula* and/or *Juniper* temporarily decline sharply and are replaced by herbaceous tundra-type communities, before recovering their former abundance with the return of temperate conditions around 12 ka. At sites where *Juniperus* remained most abundant, such as Blea Tarn in Cumbria (Pennington and Lishman, 1971; Pennington, 1973), it is that taxon which falls sharply. In lowland Yorkshire at sites, such as Tadcaster (Bartley, 1962) or Gransmoor (Walker *et al.*, 1993), dominant *Betula* was the most diminished, forming a distinctive double *Betula* peak in the early to mid-interstadial, a feature recorded in the Yorkshire region also at Kildale Hall (Jones, 1977a, b), Seamer Carrs (Jones, 1976a), Skipsea Bail Mere (Flenley, 1984), Skipsea Withow Mere (Hunt *et al.*, 1984), The Bog, Roos (Beckett, 1981), Thorpe Bulmer (Bartley *et al.*, 1976), Dishforth Bog (Giles, 1992) and in the Vale of Pickering (Walker and Godwin, 1954; Day, 1996). It also has been observed at altitude in the Howgill Fells at Garths (Gunson, 1991). This oscillation also has been recognized less distinctly in the pollen record from west of the Pennines, at Hawes Water (Oldfield, 1960a, b), Witherslack Hall (Smith, 1958c) and Blea Tarn, Blelham Bog and Windermere (Pennington, 1973, 1977, 1981) for example. Its occurrence at Blea Tarn, however, has been confirmed by numerical principal component analyses (Pennington and Sackin, 1975). It possibly also has been recorded at Crose Mere in Shropshire (Beales, 1980). It may well be present at other sites, but unidentified owing to lower resolution study. Although the accumulated body of evidence is compelling, even with detailed data care must be exercised in establishing that the pollen fluctuations reflect real climatic reversion rather than effects of vegetation community succession (Watts, 1970; Tipping, 1991a). Some authors have equated this period of low-magnitude climatic deterioration with the Older Dryas cold phase of Europe, but dating control is imprecise and allows no secure correlation. At some sites such as Skipsea Withow Mere in Holderness (Gilbertson *et al.*, 1984, 1987), erosion of catchment mineral soils, a process that had all but ceased under increasingly developed vegetation cover, became reactivated. The molluscan evidence from Skipsea Withow Mere (Thew and Woodall, 1984) supports the pollen evidence for a real climatic deterioration in this phase.

The presence of human populations in northern England in this early interstadial phase before 12 ka is proven by cut marks on mountain hare bones from Derbyshire caves, with AMS dates that cluster between 12 and 12.5 ka (Housley, 1991). Dates at these sites on aurochs and mammoth are from the same time period, as are dates on mammoth bones from a kettlehole at Condover in Shropshire (Lister, 1991) with conforming pollen and beetle evidence (Coope and Lister, 1987). All of these fauna are indicative of the pre-forest, or at most very open park-tundra environments, that existed in the earlier interstadial, events GI-1e and GI-1d. Faunal remains of the more woodland-adapted elk have been recovered mainly from sediments of the succeeding event GI-1c (Blackburn, 1952), suggesting a more developed park woodland cover during this later period.

## The later Late-glacial interstadial 12–10.9 ka (GI-1c to GI-1a)

Although temperatures did not attain the high levels reached after 13 ka BP in the Late-glacial thermal maximum early in event GI-1e, the return of warmer climate after *c.* 12 ka in event GI-1c initiated the maximum expansion and development of vegetation communities during the interstadial across most of northern England. The chronological parameters of event GI-1c allow its broad correlation with the main phase of the British Late-glacial Windermere Interstadial (cf. Allerød) and pollen zone II. The expansion of tree *Betula* woodland is the diagnostic vegetation change of this phase, at least in lowland areas, as time lags in ecosystem development and soil maturation ceased to restrain plant successions (Pennington, 1986; Walker *et al.*, 1993). The pollen record from the many interstadial profiles available in northern England shows, however, that as with the Late-glacial generally, vegetation developments in this period were spatially distinct and time transgressive. The maximum tree birch values at most sites are recorded during this period. Tree *Betula* macrofossils occur at some sites, as at Poulton-le-Fylde (Hallam *et al.*, 1973) and in north Cumbria (Walker, 1966b), proving local tree birch growth. Beckett (1981) reported *Betula* pollen expansion between 12 and 11.5 ka in Holderness at The Bog, Roos, and in west Lancashire at Poulton-le-Fylde (Hallam *et al.*, 1973) the *Betula* maximum is dated between 12 200 ± 160 and 11 665 ± 140 BP. Anomalously early areas of *Betula* expansion did exist, perhaps in sheltered or edaphically conducive locations, as in the southern Lake District where birch appears to have become established around 12 500 years BP (Pennington, 1981, 1986) and formed locally dense woods. Significant renewed increases in *Betula* frequencies and concentration also occur in this area at Blelham Tarn, Blea Tarn and Low Wray Bay around 12 ka (Pennington, 1996). In most cases, however, a distinct north–south gradient is apparent in the maximum *Betula* percentages achieved in GI-1c. Lowland sites in south and east Yorkshire show a full development of birch-woods, although perhaps quite open, with *Betula* frequencies up to 80% of total land pollen at The Bog, Roos, 70% at Bingley Bog (Keen *et al*., 1988) and similar high figures at Tadcaster (Bartley, 1962) and Gransmoor (Walker *et al*., 1993). In the Tees lowlands, Blackburn (1952) noted that although birch percentages pointed to locally dense stands of the tree, the persistence of *Helianthemum* and other arctic–alpine herbs throughout her pollen zone II suggested an open park-tundra rather than closed woodland. This accords with data at nearby Kildale Hall (Jones, 1977b) and the findings of Bartley *et al*. (1976) in south Durham, where at Thorpe Bulmer the birch peak is much lower than in Yorkshire sites and indicates a very open park-tundra vegetation with extensive shrub and grassy areas. Thorpe Bulmer is the northernmost site in eastern England where a distinct GI-1c *Betula* maximum can be recognized and may well have been near the northerly limit of birch-wood expansion during this period. Farther north, in north Durham and Northumberland, *Juniperus* persists and forms characteristic shrub–heath vegetation with *Empetrum* and grass–sedge associations. *Betula* is very poorly represented even in the coastal lowlands at Bradford Kames (Bartley, 1966), Cranberry Bog (Turner and Kershaw, 1973) and Broomhouse Farm (Shennan *et al*., 2000a), and particularly so at higher altitudes such as Longlee Moor (Bartley, 1966) and at Din Moss (Hibbert and Switsur, 1976), where it reaches barely 15% after organic inception at 12 250 ± 250 years BP. This latitudinal gradient is also present in the west, as Crose Mere in Shropshire (Beales, 1980) and the King's Pool in Staffordshire after 12 070 ± 220 years BP (Bartley and Morgan, 1990) show *Betula* frequencies up to 50% and 30%, respectively, indicating patchy open woodland, whereas at sites to the north in lowland Lancashire and Cheshire *Betula* values are uniformly rather less, below 20% at Moss Lake, Liverpool (Godwin, 1959) and Bag Mere and Chat Moss (Birks, 1965a). Recent mossland survey in Cheshire has identified more sites that conform to this pattern (Leah *et al*., 1997), including a full interstadial sequence at White Moss below a dated level of 10 715 ± 65 years BP. *Betula* frequencies are as low in more northerly lowlands, as at St Bees on the Cumbrian coast (Walker, 1966b), throughout this phase. A simple latitudinal control is insufficient to explain the variability in *Betula* maxima, however, as several sites in Scotland exhibit high interstadial *Betula* percentages (Tipping, 1991b), and more subtle environmental factors must have operated (Pennington 1981, 1986; Walker *et al*., 1993). Superimposed upon this north–south gradient were the effects of local topographical and edaphic controls, which were in places critical, so that at Willow Garth upon the chalk upland of the Yorkshire Wolds (Bush, 1993, Bush and Flenley, 1987) conditions remained open, with a grass–sedge tundra environment including some local *Juniperus* but with very little *Betula*. Similarly in the limestone areas of the Pennine upland at Malham Tarn Moss (Piggott and Piggott, 1963) tree *Betula* may not have been present at all and *Juniperus* and tundra herbs maintained a very open plant cover. Deposits from elsewhere in the Pennines (Walker, 1955a) and from higher altitude tarns in the Lake District, such as Sty Head Tarn (Pennington, 1996), confirm this pattern for the uplands during this period.

The lithostratigraphical signature of the GI-1c event after 12 ka at virtually all sites is an increase in the organic content of sediments through increased aquatic and terrestrial vegetation cover and maturation of soils. In lowland areas, true peats, often composed of common fen mosses (Dickson, 1973), formed in wet hollows (Morgan, A.V., 1973; Baxter, 1983; Bush, 1993; Lillie and Gearey, 2000; Hughes *et al*., 2000). In most lakes, however, detrital organic gyttjas, which contain a very high proportion of mineral material, accumulated, reflecting the still open nature of the vegetation cover on the by then stable catchment soils, with park tundra, or at best open woodland, being the climax community achieved in most locations. Most 'loss on ignition' studies on interstadial organic sediments have revealed their very low organic content (Pennington, 1964, 1970; Lowe *et al*., 1999). Where very open and mainly herbaceous or low shrub vegetation persisted, as at altitude or upon thin calcareous soils, a poorly organic gyttja, a calcareous marl or even only a slightly organic silt usually represents the interstadial deposit. This is demonstrated at Willow Garth in the Yorkshire Wolds (Bush, 1993), at Malham Tarn Moss in the Pennines (Piggott and Piggott, 1963) and at high altitude sites in the Lake District (Pennington, 1964, 1996). An exception seems to be Din Moss in the Cheviot Hills (Hibbert and Switsur, 1976), where a fine detritus mud with wood fragments persists throughout the whole Late Devensian, despite tree and shrub pollen percentages never exceeding 30%. Many lowland sites, such as at Chat Moss (Birks, 1965a) on the Mersey valley terraces where *Betula* values remain low in phase GI-1c and only a slightly organic clay mud accumulated at this time, also reflect this low organic input. Even in the areas with higher tree *Betula* records on the Lake District periphery (Smith, 1958c; Franks and Pennington, 1961) or in south Yorkshire (Bartley, 1962) a poorly organic clay with organic detritus is present rather than a highly organic gyttja. By contrast interstadial sediments at Church Moss, Cheshire are highly organic, herbaceous and moss peats (Hughes *et al.*, 2000) formed in a base-rich fen from 12 450 ± 60 years BP onwards and probably maintained through gradual subsidence of underlying salt-bearing strata.

Artefactual evidence for human presence and activity during this woodland interstadial phase between *c.* 12 and *c.* 11 ka is recorded at several sites across northern England. Notable are Poulton-le-Fylde, associated with elk bones (Hallam *et al.*, 1973), Victoria Cave, Settle (Wymer, 1981), Skipsea Withow Mere (Gilbertson, 1984a) and Gransmoor (Sheldrick *et al.*, 1997) in Holderness, Porth-y-Waen in Shropshire, with a radiocarbon date of 11 390 ± 120 years BP (Hedges *et al.*, 1990) and at Flixton in the Vale of Pickering (Schadla-Hall, 1987a, b), whereas at nearby Star Carr, temporary *Betula* pollen decline and high charcoal frequencies indicate burning of the open *Betula* woodland, perhaps by local human occupants (Day, 1996). The role of fire in altering vegetation patterns and deflecting woodland development, whether as a result of human activity or natural ignition, could have been significant under open park-tundra conditions and warrants further consideration. Charcoal is often recorded from Late-glacial sediments (Pennington, 1977; Day, 1996; Leah *et al.*, 1997), although the effects of burning would have been localized.

A climatic oscillation near the end of the interstadial is clearly manifest in the GRIP ice-core record as events GI-1b and GI-1a (Lowe *et al.*, 1999; Mayle *et al.*, 1999). It was of short duration, only a few centuries, and may not have been of very high intensity, for its litho- or biostratigraphical signature is not well defined in most northern England profiles. Many sites do show gradually more open late interstadial vegetation and a continuing cooling trend as the Younger Dryas Stadial approaches, but few record the vegetation fluctuations of a brief cold event (GI-1b), followed by a recovery in climate (GI-1a). These have been recognized in east Yorkshire at Gransmoor and Star Carr by a *Betula* pollen oscillation, its fall accompanied by an increase in tundra herbs followed by *Betula* recovery and increases in *Juniperus* and other thermophilous taxa frequencies (Walker *et al.*, 1993; Day, 1996). At Gransmoor, insect data show a significant cooling in event GI-1b, confirming the pollen evidence. Mayle *et al.* (1999) suggest that the cooling event at the start of GI-1b had a more marked effect on vegetation and catchment soils than even the later very cold event of the Younger Dryas (GS-1). High-resolution data are required to observe the effects of this late interstadial cooling event. A few pollen profiles from north-west England may also show indications of this oscillation, as at Blelham Bog (Pennington, 1975a), where a brief phase of reduced *Betula* frequencies and concentrations, with increases in several tundra herbs such as *Rumex*, *Artemisia*, Cyperaceae and Gramineae, occurs late in pollen zone II. Analogous changes occur at Blea Tarn (Pennington 1996), where uniquely the phase includes significant soil erosion. Unusually high *Empetrum* values at this site may indicate locally unstable late interstadial soils. Similar pollen features occur at nearby Helton Tarn and Witherslack Hall (Smith, 1958c), whereas other sites have less clear phases of reduced thermophilous taxa between 12 and 11 ka, which may reflect this event, as at Broomhouse Farm in Northumberland (Shennan *et al.*, 2000a) where temporary replacement of *Juniperus* by Cyperaceae occurs. Higher resolution research is required to investigate this GRIP-recognized climatic oscillation event, which seems to be manifest in sedimentary records around the North Atlantic margins (Levesque *et al.*, 1993). It may be identifiable in the integrated insect, pollen and macrofossil records from Church Moss in Cheshire (Hughes *et al.*, 2000).

## The post-interstadial Late-glacial 10.9–10 ka (GS-1)

The return of severe cold climate conditions in the last millennium of the Late-glacial, the Younger Dryas/Loch Lomond Stadial (event GS-1), brought major environmental change throughout northern England. The coleopteran, chironomid and ice-core records indicate a rapid switch to this cold event as temperatures fell by several degrees centigrade (Lowe *et al.*,1994a, b, 1995a, b, 1999; Lowe and Walker, 1997b; Mayle *et al.*, 1999). The renewed cold and an increase in effective precipitation in the first part of the event induced re-establishment of cirque and small valley glaciers in the highest parts of the Lake District and Pennines (Gray and Coxon, 1991; Mitchell 1991b). As after initial deglaciation, periglacial activity and the destabilization of land surfaces was widespread and, in the uplands, landslips, debris flows and mudslides were common during event GS-1 owing to failure of devegetated slopes (Mitchell, 1991b). Strong vegetation reversion occurred with a return to a very thin and patchy steppe-tundra herbaceous flora dominated by Gramineae and Cyperaceae, with a wide range of cold-tolerant herbs of which *Artemisia* is the most diagnostic (Pennington, 1977), particularly in the more arid phases of this cold event. However, there is still much intersite variation in pollen assemblages and thus in palaeoenvironments, even across areas as restricted as Cumbria (Pennington, 1996). Ruderal taxa such as Chenopodiaceae, Caryophyllaceae and *Thalictrum* are prominent among an arctic–alpine pollen suite of herbs, which is clarified by detailed macrofossil records from key sites such as Skipsea Withow Mere (Hunt *et al.*, 1984) and Willow Garth (Bush, 1993). *Juniperus* and tree *Betula* did survive in the region, mainly in sheltered lowland areas such as south-east Cumbria, as some diagrams (Smith, 1958a; Oldfield, 1960a, b) show persistent pollen frequencies, although much of the latter may well be *B. nana*. Tree *Betula* wood with dates of 10.7 and 10.4 ka from clastic stadial sediment at Skipsea Withow Mere (Hunt *et al.*, 1984) suggest local survival. The later part of the Younger Dryas event is marked by *Rumex* expansion, which continues through *Empetrum* and *Juniperus* succession into the *Betula* woodland of the early Flandrian. In northern areas such as Northumberland and Durham a clear *Juniperus* peak occurs at the transition to the Flandrian (Bartley, 1966) but in more southerly areas rapid *Betula* expansion prevents this (Bartley, 1962).

In lake catchments and lowland river systems the characteristic geological signal of the GS-1 cold event is clastic facies deposition. Erosion of soils in all basin catchments makes this clastic layer virtually ubiquitous in limnic sequences, usually a silty clay with low pollen concentration but sometimes a coarser-grained sand or even a gravel unit, as at Church Stretton in Shropshire (Osborne, 1972). Often the microfossil content of destroyed interstadial soils is redeposited in these clastic facies, which form the upper clastic unit of the classic Late-glacial tripartite lithostratigraphy. Many unpenetrated clastic units that underlie early Flandrian organic sequences, as in the meres and mires of the Cheshire–Shropshire plain (Tallis, 1973b; Reynolds, 1979), will be of this age. At Church Moss in Cheshire, however, organic sedimentation persisted as bryophyte fen peats throughout the Loch Lomond Stadial, although with greatly reduced loss-on-ignition values (Hughes *et al.*, 2000). Loch Lomond Stadial depositional facies in lowland river systems are mainly sands laid down as levées or alluvium within mobile braided channels (Gaunt *et al.*, 1971; Gaunt, 1981, 1994; Dinnin, 1997a, b). Many basal peats within palaeochannel fluvial sequences in the River Hull and the Humberhead Levels (Dinnin, 1997a, b; Lillie and Gearey, 2000) valleys have Late-glacial pollen assemblages typical of event GS-1 or late GI-1 type, and aggradation and floodplain sedimentation appears to have started prior to the Flandrian in these areas. The cold, arid and windy conditions of the Loch Lomond Stadial GS-1 event led to the erosion and aeolian redistribution of glaciofluvial and levée sands, creating extensive coversand formations in the southern Vale of York and south-west Lancashire (Gaunt *et al.*, 1971; Bateman, 1995; Wilson *et al.*, 1981). The Shirdley Hill Sands of south-west Lancashire have been shown by lithostratigraphy (Godwin, 1959) to have a Loch Lomond Stadial context and at Clieves Hills the sand overlies a peat, with a stadial tundra pollen assemblage, that is radiocarbon dated to 10 455 ± 100 years BP (Tooley, 1978a; Innes, 1986; Innes *et al.*, 1989). Baxter (1983) reported similar ages for peats below

Shirdley Hill Sand in the Mersey lowlands. In the Vale of York at Cawood a peat above levée sand and sealed by coversand is dated to 10 469 ± 60 years BP (Jones and Gaunt, 1976). Peat dates within coversand of 10 700 ± 190 years BP near York (Matthews, 1970) and 10 550 ± 250 years BP with an associated cold climate insect fauna at Messingham (Buckland, 1982) give intermediate ages for aeolian activity, and a peaty soil sealed by the sands at West Moor (Gaunt *et al.*, 1971) provides a maximum date for the start of sand blowing of 11 100 ± 200 years BP. In the Vale of Pickering at Seamer Carrs (Schadla-Hall, 1987a, b) a blown sand layer sealed by early Flandrian peat showed the effects of frost action and it overlay a peat containing Upper Palaeolithic flint artefacts. Dates on the buried peat ranged between 10.2 and 11.3 ka, indicating a late Loch Lomond Stadial age for sand emplacement. These dates conform with a date of 10 413 ± 210 years BP for organic muds with a Loch Lomond Stadial pollen assemblage at nearby Flixton that contained *Equus* bones and flint artefacts (Godwin and Willis, 1959). Clastic sediment accumulation under Loch Lomond Stadial conditions is also illustrated by silty loam sediments in Kirkhead Cave in south Cumbria, which contain molluscan and pollen records suggesting stadial pollen zone III deposition for their lower levels, which included flint artefacts and a *Megaloceros* antler boss dated 10 700 ± 200 years BP (Gale and Hunt, 1985).

These coversand dates conform very closely with the age parameters of inwashed clastic units of stadial GS-1 event age provided by the many dated limnic sequences in northern England. At Routh Quarry in the Hull valley (Lillie and Gearey, 2000) peat composed of the arctic moss *Homalothecium nitens* formed between 11 260 ± 75 years BP and 10 740 ± 75 years BP, with the latter date being a maximum age for a Loch Lomond Stadial clay unit that rests upon the peat. At The Bog, Roos (Beckett, 1981) the stadial clay unit formed between 11 220 ± 220 years BP and 10 120 ± 180 years BP, and at St Bees in Cumbria (Coope and Joachim, 1980) clastic stadial deposition began soon after 11 180 ± 120 years BP. Typical dates of 10 828 ± 185 years BP and 10 318 ± 215 years BP from Scaleby Moss in north Cumbria for the start and end of the Loch Lomond Stadial (pollen zone III) were among the first to be reported, as part of a long radiocarbon-dated standard pollen diagram (Godwin *et al.*, 1957). A range of dates from northern England cluster around 11 ka for the start of stadial clastic deposition, for example, 10 900 ± 80 years BP at Broomhouse Farm in north Northumberland (Shennan *et al.*, 2000a), 10 700 ± 70 years BP at Willow Garth in east Yorkshire (Bush, 1993) and 10 715 ± 65 years BP at White Moss in Cheshire (Leah *et al.*, 1997). Dates for the termination of the Loch Lomond Stadial are also similar throughout the region before 10 ka, for example, 10 340 ± 200 years BP at Din Moss at the Scottish Border (Hibbert and Switsur, 1976), 10 150 ± 80 years BP at Gransmoor in Holderness (Lowe *et al.*, 1995b), 10 350 ± 200 years BP at Kildale Hall in north Yorkshire (Keen *et al.*, 1984) and 10 310 ± 210 years BP at Crose Mere in Shropshire (Beales, 1980). A later date of 10 070 ± 190 years BP from Weelhead Moss in Upper Teesdale in the north Pennines (Turner *et al.*, 1973) suggests a small altitudinal time lag for the end of Late-glacial stadial environments, but in general the timing of this last Late-glacial event (GS-1) seems closely comparable throughout northern England and conforms with the ages of *c.* 11 ka to *c.* 10 ka interpolated from the radiocarbon age curve at the type site of Blelham Bog in Cumbria (Pennington, 1977).

## LOW WRAY BAY (WINDERMERE) (NY 377 013)

***D. Huddart***

### Introduction

Low Wray Bay, Cumbria (Figure 6.1) is the type locality for the Windermere Interstadial (Coope and Pennington, 1977) which occurred after the Late Devensian ice sheet retreated (Dimlington Stadial). The site is situated on the western side of Lake Windermere (NY 377 013), which is 17 km long, 14.8 km² in area and reaches a maximum depth of 64 m. This lake basin contains a thick sequence of lacustrine sediments in which are recorded the later stages of Devensian deglaciation, the Late-glacial period and the Flandrian succession. The importance of the site stems from the detail in which this sequence is known, with biological, geochemical and geophysical investigations being carried out since the late 1930s. At Low Wray Bay the thick lacustrine sequence has provided vital biostratigraphical, lithostratigraphical, chronostratigraphical and geochemical evidence for the rapid

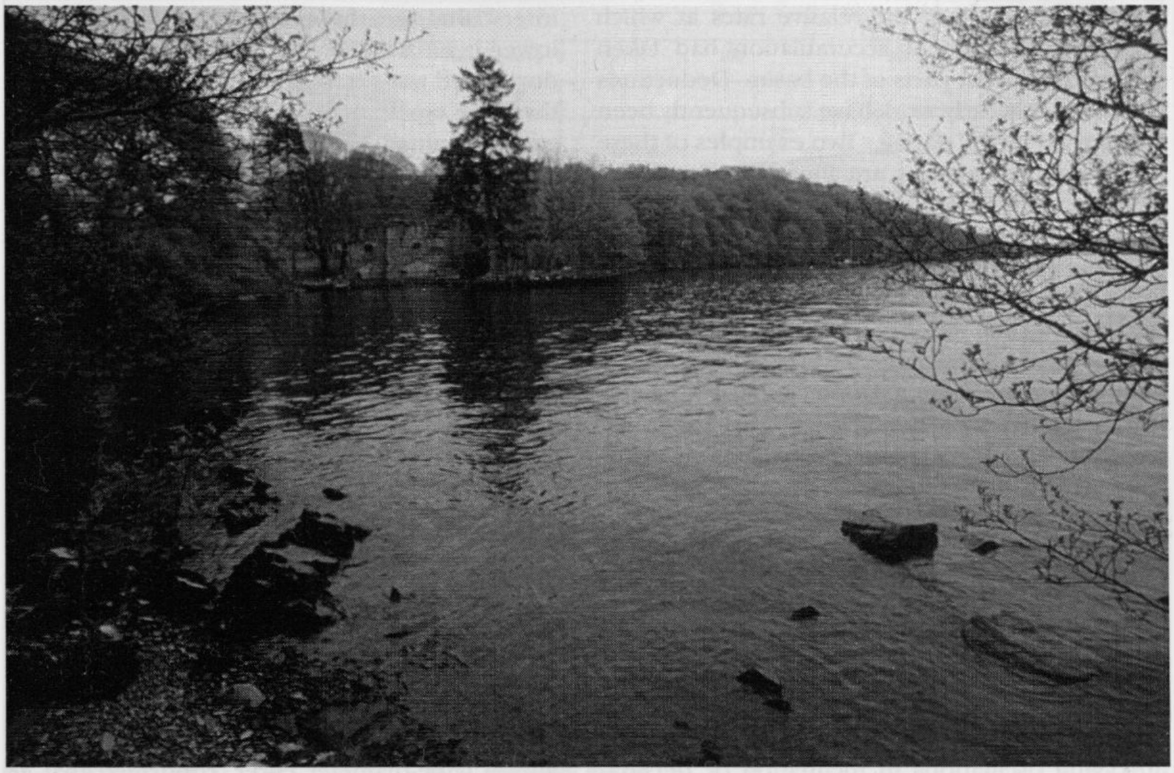


**Figure 6.1** Low Wray Bay, Windermere. View looking north. (Photo: D. Huddart.)

sequence of environmental changes during the past 15 000 years. The biostratigraphical record obtained from this location is superior to other British sites and has allowed the Windermere Interstadial to be defined on the basis of both lithostratigraphy and biostratigraphy and the sediments also have been radiocarbon-dated in detail. This has helped to define this interstadial in age, allowed international correlation and raised a number of important issues with regard to the biostratigraphy.

## Description

The Admiralty Chart (1937) of Lake Windermere was prepared from results of an early echo-sounding survey, which showed a discontinuity in the unconsolidated sediments overlying bedrock. A programme of sediment coring was initiated (Jenkin and Mortimer, 1938) to determine whether this discontinuity represented the interface between glacial clay and post-glacial mud. The results were positive and since this early work more detailed profiles from 'sparker' and 'pinger' seismic profiling have been published (Howell, 1971). These surveys have shown that the North Basin contains 21 m of sediment and the South Basin up to 40 m. Over the submerged ridge with islands, which separates the two deep basins, the water is shallow and the glacial clays are here covered by less than 1 m of post-glacial mud. The stratigraphy above the bedrock showed up to 30-cm-diameter cobble gravels and the sediments above these gravels have been sampled by Mackereth corers (Mackereth, 1958). These penetrate to the gravels in shallower water but not in the central parts of the lake, where the sediments are much deeper.

The brown, organic post-glacial muds, which reach a thickness of 5–6 m in the North Basin and *c.* 3 m in the South Basin, represent the deposits of the past 10 000 years and have been shown to be clearly stratified, although in their lower parts they include the occasional slump comparable with the slump structures in the underlying clays (Smith, A.J., 1959; Mackereth, 1966). The pattern of accumulation of these brown muds in the North Basin was first established by pollen analysis of ten cores (Pennington, 1947), where the application of a pollen chronology showed the relative rates at which post-glacial sediment accumulation had taken place in different parts of the basin. Deductions based on this early work have subsequently been confirmed by $^{14}C$ dating. Two examples of these early pollen diagrams are illustrated in Figure 6.2 from Low Wray Bay. Plant macrofossils were also analysed by Pennington (1947) from the detritus silt and an example is given in Figure 6.3. The post-glacial brown mud from the central parts of the South Basin provided the cores on which Mackereth (1971) carried out his pioneering work on the palaeomagnetism of lacustrine sediments. The curve for the horizontal component of remanent magnetization, in a core from which nine $^{14}C$ dates were obtained, showed regular swings in direction identified with past oscillations. This work has been continued and expanded by Thompson (1975) and Turner and Thompson (1981) and the 'master curve' from Windermere has been used as a basis for comparison for observations on other lake sequences. The secular variations in the magnetic field have been short term and local, involving oscillations in inclination or declination of 20–30° over 2000–3000 years.

Below the brown Flandrian muds there is in all parts of the lake a thick deposit of laminated pink-grey clay in which the paired laminations (couplets) are composed of a fine-grained pink layer and a coarser, greyish-pink layer in which graded bedding is visible (Pennington, 1947). These laminated clays contain a sparse pollen content characterized by *Artemisia* (Pennington, 1980). About 50 cm below the upper boundary of this laminated clay, its lower boundary is formed by a discontinuity. Although the laminated clay is similar over most of the lake floor, everywhere containing about 400 couplets, the underlying sediment below the discontinuity is markedly different in shallower water, in water depths of 10 m or less. In the shallow, littoral areas an organic interstadial deposit is present, but this rapidly wedges out with increasing water depth, and inside the 15 m water-depth contour only a thin layer of grey silt underlies the laminated clay (Pennington, 1947).

This grey silt in the North Basin is coarse and non-fossiliferous, but in the central parts of the South Basin it contains pollen and on both pollen and $^{14}C$ evidence it has been shown to be correlative with the thicker (*c.* 50 cm) and organic interstadial deposits of the marginal bays (Pennington, 1947). Below these sediments of interstadial age, there is in all parts of the lake a lower laminated clay, in which there is a rapid downward transition from laminated couplets of the same type as in the upper laminated clay to coarser-grained sediment in which the couplets are much thicker and show a stronger colour and particle-size contrast between the two layers of the couplets. In the shallower water areas this deposit is under 50 cm thick and passes down into gravel but outside of these areas there is a great thickness of laminated clay, within which there are thick layers of sand and silt, resembling the turbidites described from Lake Brienz, where the catchment is 18% glaciated (Sturm and Matter, 1978).

The site of Jenkin and Mortimer's (1938) original coring in Low Wray Bay provided the first evidence in Britain for the Late-glacial Interstadial, then called the 'Allerød', which had then just been recognized in Ireland (Jessen and Farrington, 1938; Pennington, 1943). The first supporting $^{14}C$ date (Q294) was obtained by Godwin (1960a) but subsequent work on this profile, and a series of ten $^{14}C$ dates obtained from a wide-diameter corer, confirmed that an interstadial beginning much earlier than the continental Allerød was represented (Pennington, 1970, 1977). The Late-glacial pollen diagram is illustrated in Figure 6.4 and a transverse section of Windermere in the latitude of Low Wray Bay showing a section through the deposits is illustrated in Figure 6.5

The earliest organic deposits in Low Wray Bay, which contain pollen spectra and annual pollen deposition rates indicative of treeless tundra, have been dated to 14 557 ± 280 years BP and 14 623 ± 360 years BP. Prior to these dates seasonal snow melt from the valley glacier led to the deposition of the Lower Laminated Clay, from which no organic remains have been found. In these varved clays there is an upward change in thickness from the lowest varves immediately overlying a stony clay, which average 0.5–1.0 cm, to the upper varves, which may be only 0.2 mm (Pennington, 1943). It can be interpreted as a change from an ice-proximal to an ice-distal position as the valley glacier retreated towards its source during deglaciation. Pennington (1943) also reported that a count of the annual laminations in the Upper Laminated Clay gave 377 as the total, but if the grey clay–silt layer containing *Oxyria digyna* at the base of the laminated clay is included, 400 to 500 years would be a reasonable estimate for that period.

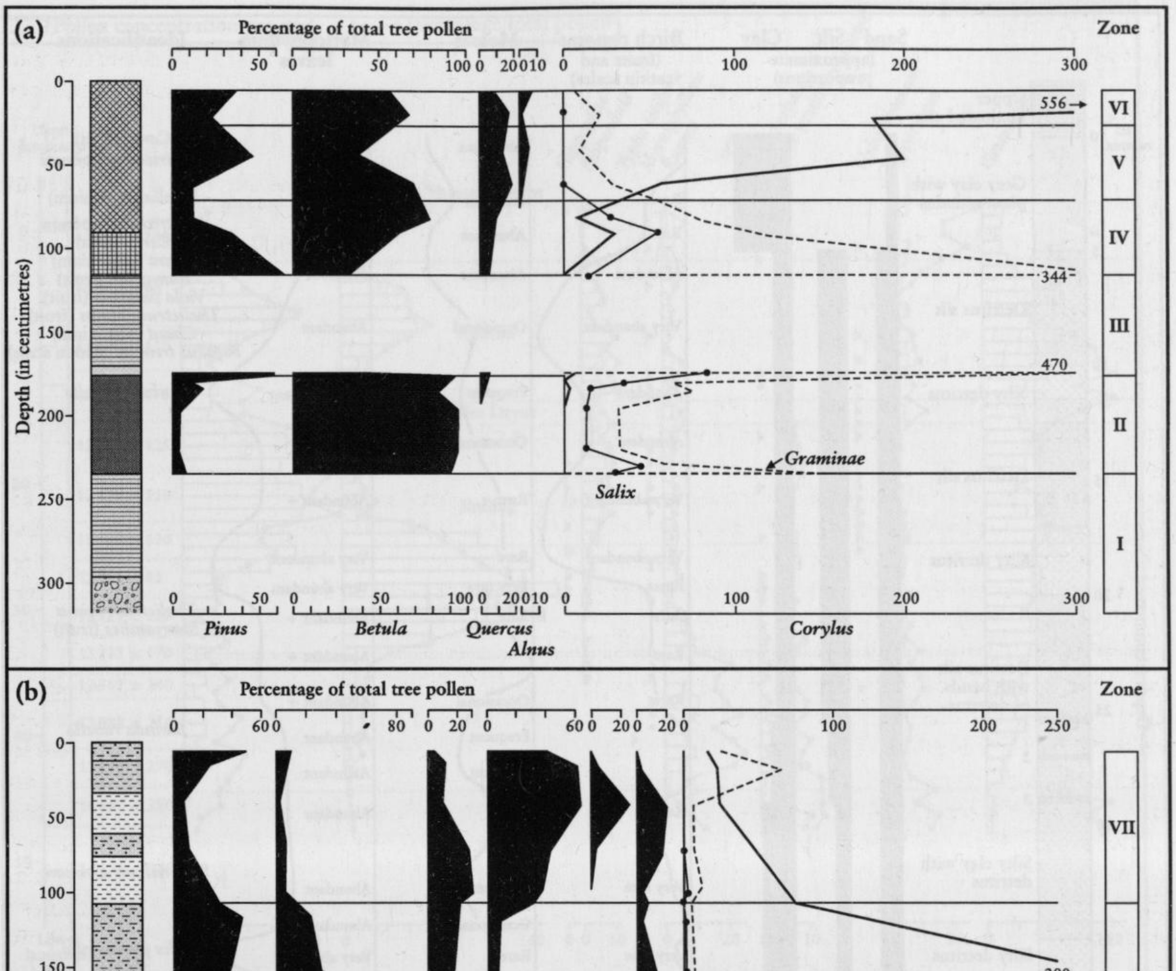



**Figure 6.2** Examples of pollen chronology from the North Basin, Windermere at different locations (see Figure 6.5) in Low Wray Bay: (a) core 1; (b) core 2 (after Pennington, 1970).

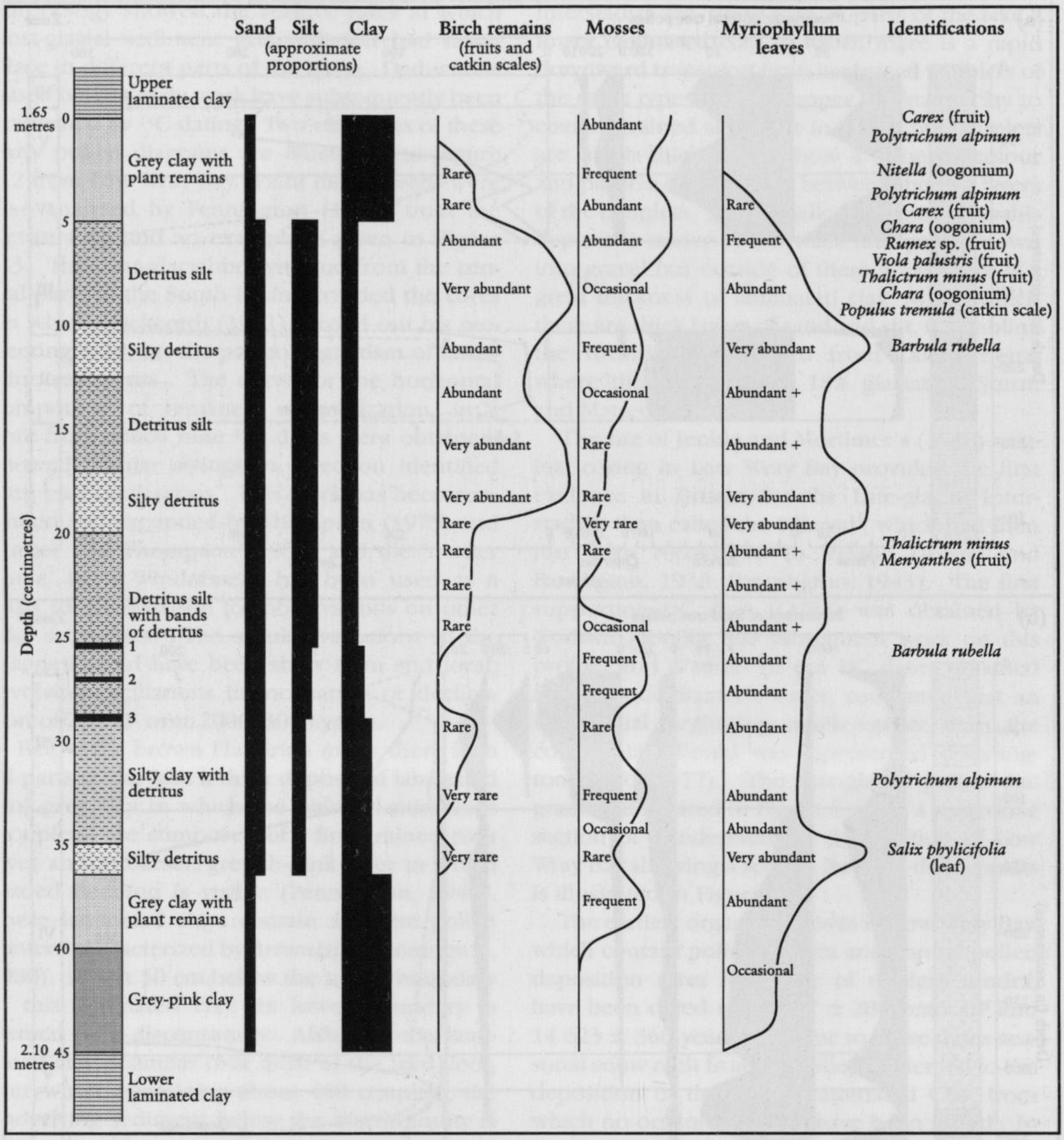



**Figure 6.3** Distribution of plant macrofossils in the detritus silt in core 1, Low Wray Bay (after Pennington, 1970).

The following synthesis of the pollen assemblage zones for Windermere is taken from Pennington (1977). Before *c.* 14 000 years BP there is a *Salix herbacea*–Cyperaceae–*Lycopodium selago* zone, with such low pollen concentration that annual deposition rates cannot have exceeded 100 grains per cm² per year, a figure low even for tundra (Davis *et al.*, 1973). Chionophilous (snow bed) and/or fjell field plant communities existed in a periglacial environment, between *c.* 14 000–13 000 years BP. Figure 6.4 shows a pollen zone dominated by 20% *Rumex* and 25% grass pollen, and pollen concentration is ten times that of the underlying zone. Within this zone appear many taxa suggestive of a climatic amelioration, for example, *Filipendula*, *Hippophäe* and *Typha latifolia*, together with at the base of this zone a significant

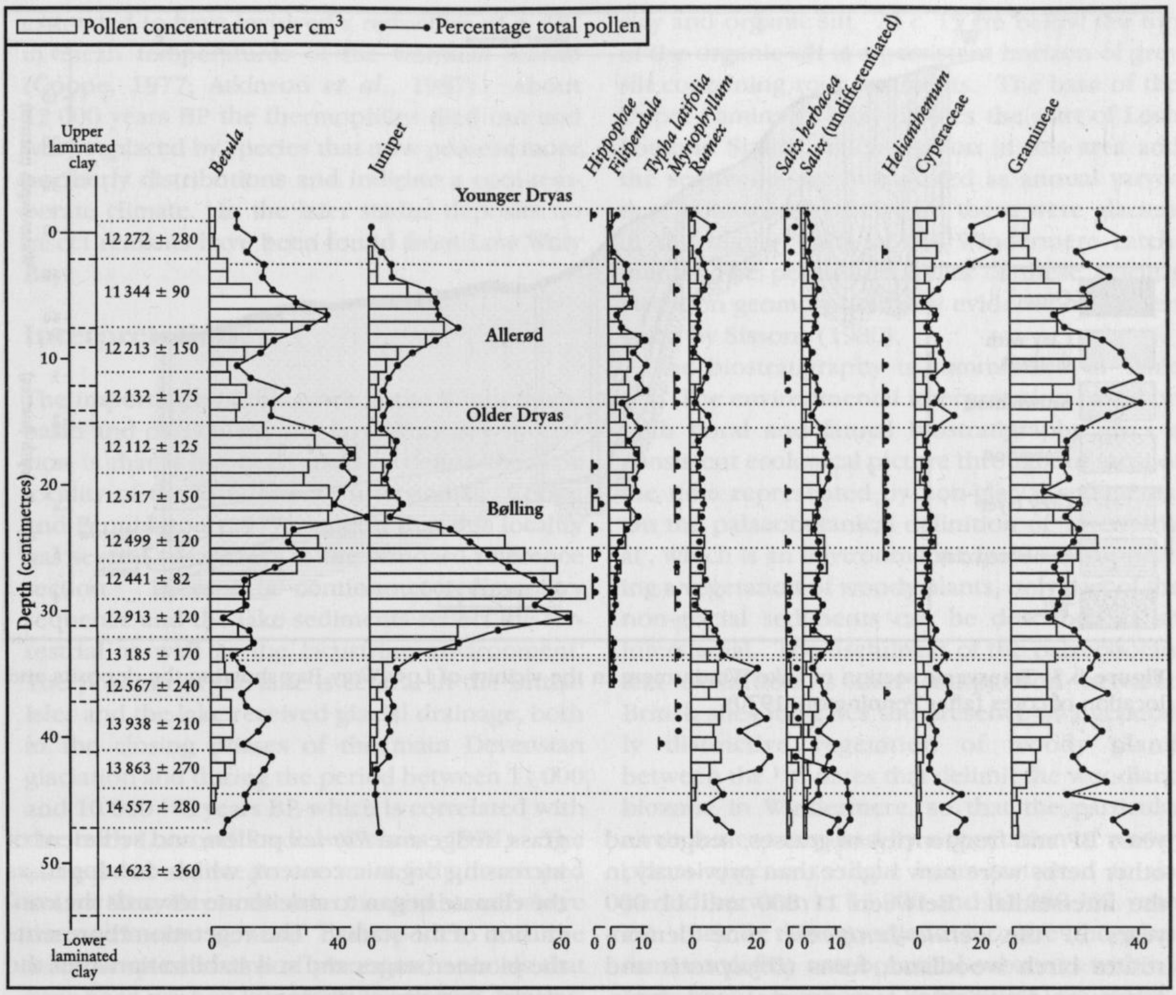



**Figure 6.4** Late-glacial percentage pollen diagrams from Low Wray Bay (after Pennington, 1970).

increase in percentage and absolute values for juniper and tree birch pollen. Through this zone *Salix herbacea* declines, Cyperaceae values remain low and *Lycopodium selago* disappears. Some undoubted *Betula nana* pollen is present, together with that of *Empetrum*, *Thalictrum* and Rubiaceae, whilst there is a rich herb assemblage, including *Helianthemum* and *Artemisia*. This climatic amelioration was unaccompanied by a significant expansion of woody plants. The next two pollen assemblage zones illustrate the Windermere Interstadial. Between 13 000 and 12 500 years BP the *Juniperus* zone shows a steep percentage and absolute rise in juniper dated to 13 000 years BP, and the Betula zone between 12 500 and 12 000 years BP shows the maximum percentages of birch (30% of the pollen record or more, corresponding to 800–1000 grains per cm² per year, which indicates, according to Birks (1973), a canopy cover of at least 50% and exceeds by a factor of four the present deposition rate from park-tundra). The characteristic Late-glacial assemblage of *Helianthemum*, *Hippophäe* and *Artemisia* reaches its maximum development in this zone and, although it has not been noted in modern vegetation, it must represent a community of unshaded habitats. A local factor probably continued to inhibit the growth of tree birches on some terrain types so that there was a mosaic of woodland and open areas with unstable soils. From 12 000 to 11 800 BP there is a narrow zone of reduced pollen deposition of the more thermophilous taxa, which is dominated by *Betula–Rumex*. It represents a partial regression towards the composition of deposits pre-13 000

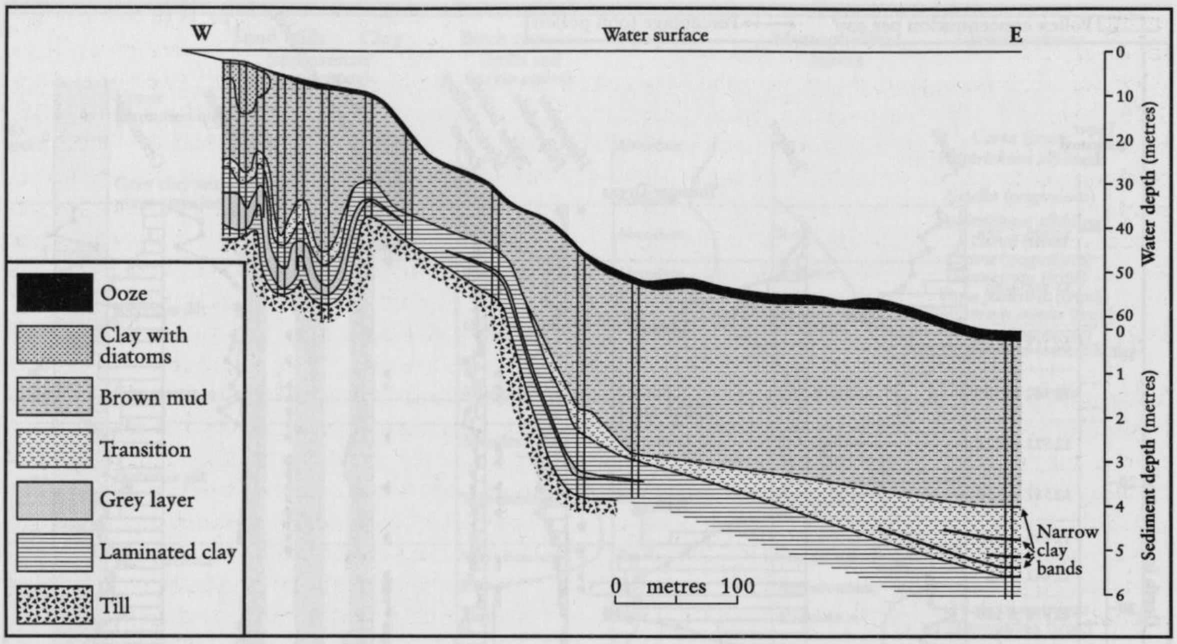



**Figure 6.5** Transverse section of Lake Windermere in the vicinity of Low Wray Bay showing the deposits and location of cores (after Pennington, 1970).

years BP and frequencies of grasses, sedges and other herbs were now higher than previously in the interstadial. Between 11 800 and 11 000 years BP the *Betula–Juniperus* zone demonstrates birch woodland, ferns (*Dryopteris* and other, unidentifiable, fern spores) and grass, sedge and *Empetrum* pollen. The inference is of more open woodland than the earlier woodland episode. Disintegration of the plant cover, the most likely cause being thought to be increased snowfall, is marked by a Cyperaceae–*Selaginella* local subzone, where there were reduced frequencies of birch and juniper. Between 11 000 and *c.* 10 500 years BP there were three pollen assemblage zones recognized: *Rumex–Artemisia*, *Artemisia–Rumex* and *Artemisia*–Caryophyllaceae, which have been dated to the first half of the Younger Dryas period, during which time the varves in Windermere were being deposited from meltwater of the corrie glaciers farther up-valley. The *Artemisia* pollen zones coincide with a distinct lithostratigraphical unit of lower organic content, indicative of periglacial soil movement and identifiable geochemically with the unweathered products of intense periglacial erosion. From about 10 500 to 10 000 years BP there is a herb-dominated pollen assemblage zone that contains much grass, sedge and *Rumex* pollen, and sediment of increasing organic content, which developed as the climate began to ameliorate towards the conclusion of the stadial. The vegetation represents the pioneer stages and soil stabilization after the glacial–periglacial phase.

The interpretation from pollen analysis was confirmed by study of the coleopteran remains (Coope, 1977), which showed that sediments dating from *c.* 14 000 years BP already contained a temperate insect fauna. No fossils, however, have yet been found from the Lower Laminated Clays, but even in the lowest interstadial sediment there are none of the obligate arctic–alpine species that were a characteristic element of the pre-interstadial faunas elsewhere in Britain. Yet species that are characteristic of the present-day Lake District are present and they include a suite of species that are either absent from the more northerly parts of Europe or are very rare there. These relatively southern, thermophilous species are not present in the upper part of the interstadial sequence and in their place there is the first incoming of the more northern species, heralding the truly arctic faunas of the Loch Lomond Stadial. Both the rise to, and fall from, a thermal maximum early in the interstadial appear to have been rapid and the decline is estimated to have involved a reduction of *c.* 3°C in mean temperatures of the warmest month (Coope, 1977; Atkinson *et al.*, 1987). About 12 000 years BP the thermophiles died out and were replaced by species that now possess more northerly distributions and indicate a cool-temperate climate. In the later stadial deposits no insect remains have been found from Low Wray Bay.

## Interpretation

The importance of this work in the Windermere basin and particularly the Low Wray Bay succession is that it has been used to define the type locality of the Windermere Interstadial. Coope and Pennington (1977) suggest that this locality has several advantages as the standard reference section. There is a continuous sedimentary sequence and the lake sediments reflect the terrestrial as well as the lacustrine environment. The position of the lake is central in the British Isles and the lake received glacial drainage, both in the closing phases of the main Devensian glaciation and during the period between 11 000 and 10 000 $^{14}C$ years BP, which is correlated with the Loch Lomond Stadial (Sissons, 1974). The interstadial sediments are most fully developed in Low Wray Bay in the lake marginal area, where the sediments contain both pollen and the macroscopic remains of plants and animals that have been washed into the lake by small streams.

Detailed analyses of the section at Low Wray Bay show that the lower boundary of the interstadial sediments may be drawn at different horizons, depending on the criteria that are adopted. Table 6.1 shows the position of the Windermere Interstadial according to lithostratigraphical, palaeobotanical or palaeoentomological criteria, from which it is possible to have alternative views based on a climatic interpretation. In terms of lithostratigraphy the upper boundary of the Lower Laminated Clay (Pennington, 1943) is interpreted as representative of the termination of the main Devensian glaciation, where there is a climatic change from a glacial to a non-glacial environment and a change to the grey organic layer. The junction has been dated to 14 623 ± 360 years BP. Above the Lower Laminated Clay the lowest sediments are almost entirely minerogenic with very sparse fragments of plant and animal tissue. The change to biogenic sediment is gradual and there is no clearly discernible boundary between clay and organic silt. At *c.* 15 cm below the top of the organic silt is a persistent horizon of grey silt containing rock fragments. The base of the Upper Laminated Clay defines the start of Loch Lomond Stadial sedimentation in this area and the sediments are interpreted as annual varves deposited in the lake when there were glaciers in the higher parts of the Windermere catchment. The postulated limits of these glaciers based on geomorphological evidence have been given by Sissons (1980).

The biostratigraphy is summarized in Table 6.1. The environmental interpretation based on both floral and faunal biostratigraphy gives a consistent ecological picture throughout most of the time represented by non-glacial sediments. On the palaeobotanical definition of 'interstadial', which is an environment capable of supporting a vegetation of woody plants, only part of the non-glacial sediments can be described as an interstadial. The usefulness of the palaeobotanical definition is that comparison between British sites indicates the presence of a generally distinctive vegetation of woody plants between the $^{14}C$ dates that delimit the woodland biozone in Windermere, so that the particular environment required by a vegetation of woody plants appears to have been restricted to the period between *c.* 13 000 and 11 000 $^{14}C$ years BP. Below this woodland biozone the plant remains define two botanical subzones within a herbaceous biozone. Pollen data equate the lower of these with present alpine vegetation in northern Europe. The first pollen grains do not occur until 13 863 ± 270 years BP. In the upper biozone the assemblage of herbs identified includes species that do not grow in the Arctic today and both pollen and macroscopic evidence shows that tree birches were present but did not expand into woodland, which is interpreted as indicative of conditions not generally favourable to trees. Only within the herbaceous biozone is there any discrepancy between the interpretations based on botanical and entomological data. Throughout the two subdivisions the fauna show a continuous development but even at the base there are no alpine species present. A number of insect species occur that, although widespread in Europe today, are absent from arctic areas and the alpine zones of mountains. This suggests that at least during the *Rumex*–Gramineae pollen zone, temperate climatic conditions prevailed. There are several comparable localities, often with much richer

**Table 6.1** Correlation table for the lithostratigraphy, biostratigraphy and chronostratigraphy of the Late Devensian deposits at Low Wray Bay, Windermere (after Coope and Pennington, 1977). p.a.z.: pollen assemblage zone.

| Lithology | Fossils: Plants | Fossils: Animals – *Coleoptera* | Fossils: Animals – Others | $^{14}$C dates | Interpretation: Environment | Interpretation: Climate | Duration of interstadial climate: WP | Duration of interstadial climate: GRC |
|---|---|---|---|---|---|---|---|---|
| Rock fragments increase<br>Upper laminated clay | *Artemisia* p.a.z.<br>Pollen very sparse | Sparse fauna with characteristic arctic–alpine species | | | Seasonal melt from mountain glaciers in Lake District corries | Very cold | | |
| Becoming less organic<br>Maximum carbon<br>Organic detritus silt | Tree birch macroscopics<br>Cyperaceae<br>*Betula–Juniperus* p.a.z. | Woodland species present<br>Fauna less rich, without 'southern' species | Cladocera decline to 0 | (11 000)<br>SRR-669 11 344 ± 90 | Increased snowfall, declining temperatures, increased soil erosion<br>Re-establishment of *Betula–Juniperus* woods | Cool temperate, becoming colder towards the top | ↕ | ↕ |
| Paler silt with rock fragments | *Betula–Rumex* p.a.z. | *Helophorus glacialis* | Cladocera reduced | SRR-670 12 213 ± 150 | Increased soil erosion, woodland reduced | Sudden cooling | | |
| Maximum carbon<br>Organic detritus silt | *Betula* p.a.z.<br>*Juniperus* p.a.z. | Sparse fauna at base, becoming richer towards the top; species present that are absent from arctic Europe | Trichoptera larvae<br>Chironomidae larvae<br>*Cristatella* statoblasts | SRR-672 12 112 ± 125<br>SRR-673 12 517 ± 150<br>SRR-675 12 441 ± 82<br>SRR-676 12 913 ± 120<br>(13 000) | Maximum environmental diversity of Windermere; interstadial (woodlands established) | Temperate throughout | | |
| Somewhat organic silt<br>Carbon content increasing | *Rumex–*Gramineae p.a.z.<br>Tree birches present | | Cladocera suddenly increased | SRR-677 13 185 ± 170<br>SRR-679 13 938* ± 210<br>SRR-680 13 863* ± 270 | Phase of pioneering flora and fauna, vegetation rich in herbs, including species which do not grow in the arctic today; although tree birches are present, no development of woodland | Plant evidence inconclusive on temperature, but some factor unfavourable to trees<br>Temperate throughout on insect evidence | | |
| Slightly organic silt | *Salix herbacea–*Cyperaceae–*Lycopodium selago* p.a.z.<br>Moss stems | Very sparse fauna, no alpine species | | SRR-681 14 557 ± 280<br>SRR-682 14 623 ± 360 | Plants compare with Middle Alpine zone of north-west Europe<br>Insect fauna lacks any alpine species and suggests early pioneer fauna | Plant evidence suggests alpine conditions<br>Insects inconclusive, but no evidence to support alpine conditions | | |
| Clay | No plant fossils | | | | | | | |
| Lower laminated clay | | | | | Seasonal melt from mountain glaciers | Very cold | | |

insect assemblages than at Windermere, now known from Britain, always in the same palynological context, where climatic warming is indicated by the insect data well before the start of the woodland biozone.

The first thermophilous insect assemblages have been dated to 13 938 ± 210 years BP. The difference between the biostratigraphies probably reflects the quicker response time of insect communities to climatic warming and has important implications for other stadials or interstadials that have been defined only on the basis of the pollen record.

Using modern examples from Greenland and southern Norway, Pennington (1973, 1977, 1980) suggested that the botanical evidence indicated that the early pollen assemblages from *c.* 14 000 to 13 000 years BP can be correlated with the high-altitude and pioneer plant communities of presently glaciated parts of southern Norway that have an oceanic periglacial climate. The arrival of tree birches before 13 000 years BP suggests that by that time the climate was as favourable as today in the inland parts of south-west Greenland but some factor, such as dryness and a short growing season, prevented their expansion to form continuous woodland. During that part of the Windermere Interstadial most favourable to plants and when woodland grew, temperatures must have exceeded those of south-west Greenland today. In the uplands, such as at Blea Tarn, where dwarf shrub heath with a varying proportion of sedges formed the interstadial vegetation, conditions appear to have been within the range now found on the outer coasts of West Greenland from *c.* 70°N southwards. During the cold period of the Younger Dryas the vegetation – and hence probably the climate – was different from that of the early period and less oceanic. The closest parallel found seems to be the continental low arctic (periglacial) vegetation of central West Greenland and comparisons suggest a highly differentiated pattern of snow cover and corresponding vegetation mosaic.

Why juniper increased before birch in the early part of the Windermere Interstadial needs some discussion, as there is evidence for tree birches already being present before this phase. Godwin (1975) in discussing the ecology of juniper, gives the evidence for its present status as a dominant above the present tree-line in northern Britain, a habitat which suggests that in some respects it is less exacting than birch. The data presented by Birks (1973) on modern pollen rain within juniper communities shows, however, that the high percentages found within this pollen assemblage zone can be matched only in tall dense thickets of *Juniperus communis* species and not by the more prostrate forms found at higher altitudes. The increase is clearly a response to climatic amelioration but it is still not clear why juniper and not birch expanded first.

The chemical investigations reported by Mackereth (1966) add detail to the environmental changes discussed so far. The changes in carbon content could be related either to a fluctuating rate of production and deposition of organic matter, or to a fluctuating rate of erosion of mineral material. The data on carbon content alone offer no means of distinguishing the relative importance of these processes but it is considered that the maximum accumulation of biogenic sediments takes place during ecologically stable interglacial climates and that minimum organic matter concentrations are found under glacial conditions, when all the soils have been destroyed by glacial erosion and deposits consist of rock flour and minerogenic sediments. This argument can be carried a stage further so that in periods of high erosion rate soil material rich in K, Na and Mg is rapidly transferred to lake basins, whereas in periods of low erosion rate these elements are rapidly removed by leaching from the catchment surface. The material that is transferred to the lakes is relatively poor in K, Na and Mg. Thus elements such as these which exhibit recurrent trends have been interpreted as an indication of erosional activity within catchments. Substantial percentages of Na and K in the mineral sediments have been interpreted as a reflection of intensive erosion of mineral soil and a low leaching level because these elements are soluble and so usually susceptible to removal in downward-percolating water. Small percentages of these elements in minerogenic sediments have been linked to reduced erosion and increased leaching. For Windermere there are relatively high concentrations of erosion indicators in both the pre-interstadial and post-interstadial deposits. There is a minimum concentration in interstadial deposits, where the maximum values for total carbon show maximum humus accumulation in the maturing soil profiles. These relationships were first shown in a series of lakes in the Lake District, including the South Basin of Windermere.

The original suggestion by Pennington (1943) that the great increase in the numbers of the diatom *Asterionella* in the uppermost 20 cm of sediment represented a change in lake biology consequent on the growth of lake-side villages and the increasing input of sewage and sewage effluents from the mid-nineteenth century, has been substantiated by radionuclide dating of these recent sediments (Pennington, 1973).

## Conclusions

The published analyses for the Low Wray Bay sequence provide a particularly full data set for environmental reconstruction, including stratigraphy and diatoms (Pennington, 1943), pollen analyses and macroscopic plant remains (Pennington, 1947, 1970, 1977, 1981; Godwin, 1960a), geochemistry (Pennington and Lishman, 1971; Pennington, 1977) and insect remains (Coope, 1977). From this evidence it has become the national type site for the Late-glacial Windermere Interstadial and it has illustrated many difficulties in defining chronostratigraphy on the basis of different types of biostratigraphy. It is also the site of the 'master curve' for the changing magnetic declination.

# BLELHAM BOG (NY 366 006)

*D. Huddart*

## Introduction

Blelham Bog, Cumbria (Figure 6.6) is an important site for providing biostratigraphical, lithostratigraphical and chronostratigraphical evidence for the major environmental changes that took place in northern England during the Late-glacial. It has been investigated in great detail by Harmsworth (1968), Evans (1970), Oldfield (1970), Pennington and Bonny (1970) and Pennington (1973). Its palaeoenvironmental record is significant for studies of the regional differentiation of the Late-glacial vegetation, lithostratigraphy and climate in Britain and north-west Europe. The site also is important because it shows evidence for both Flandrian and historical environmental change (Oldfield, 1970).

## Description

The site is situated 3 km south of Ambleside and 1 km west of Windermere. The geology of the Blelham basin is composed of Upper Silurian

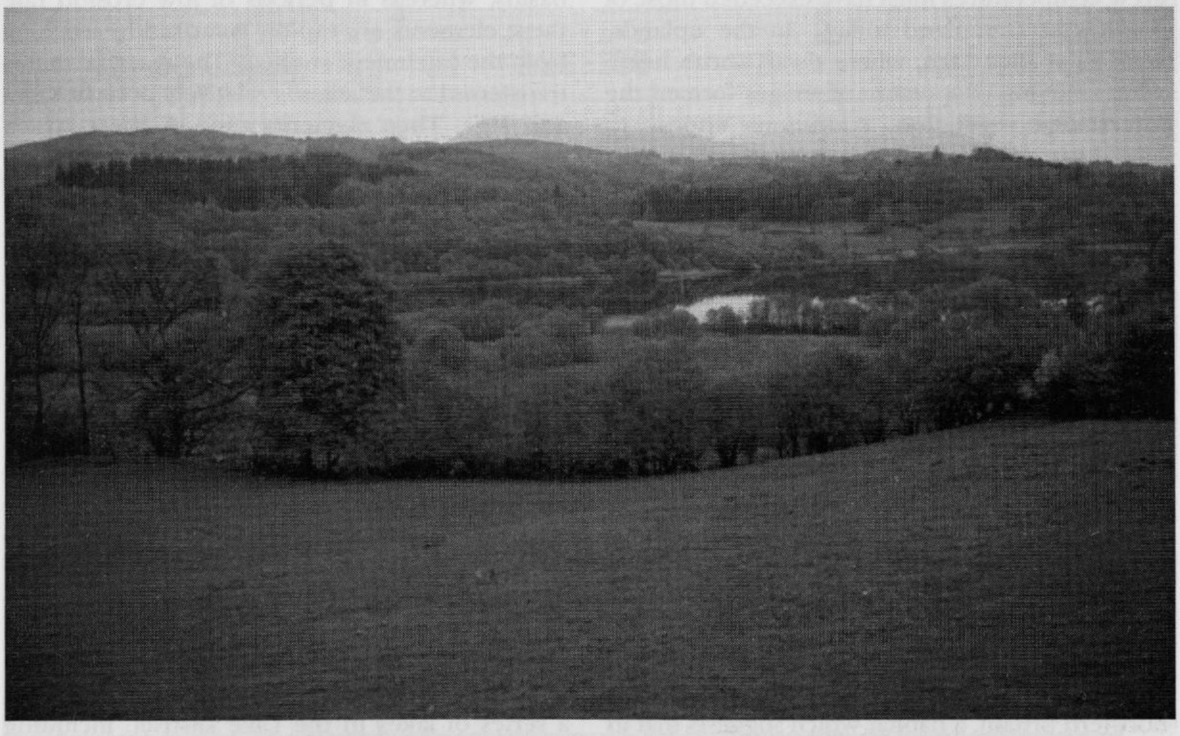


**Figure 6.6** Blelham Bog, looking to the north-east. (Photo: D. Huddart.)

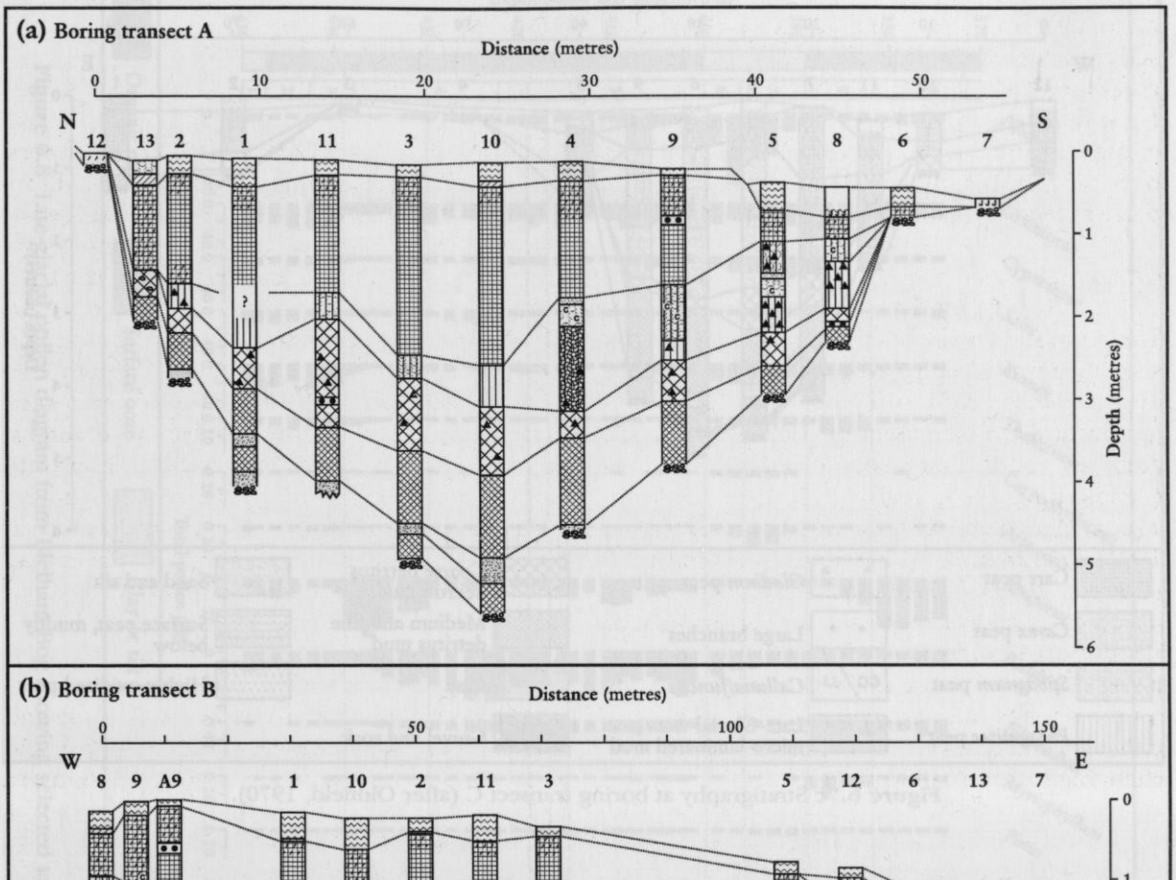



**Figure 6.7a, b and d** Stratigraphy at boring transects A, B, and D (after Oldfield, 1970). (See Figure 6.7c overleaf for boring transect C and key.)

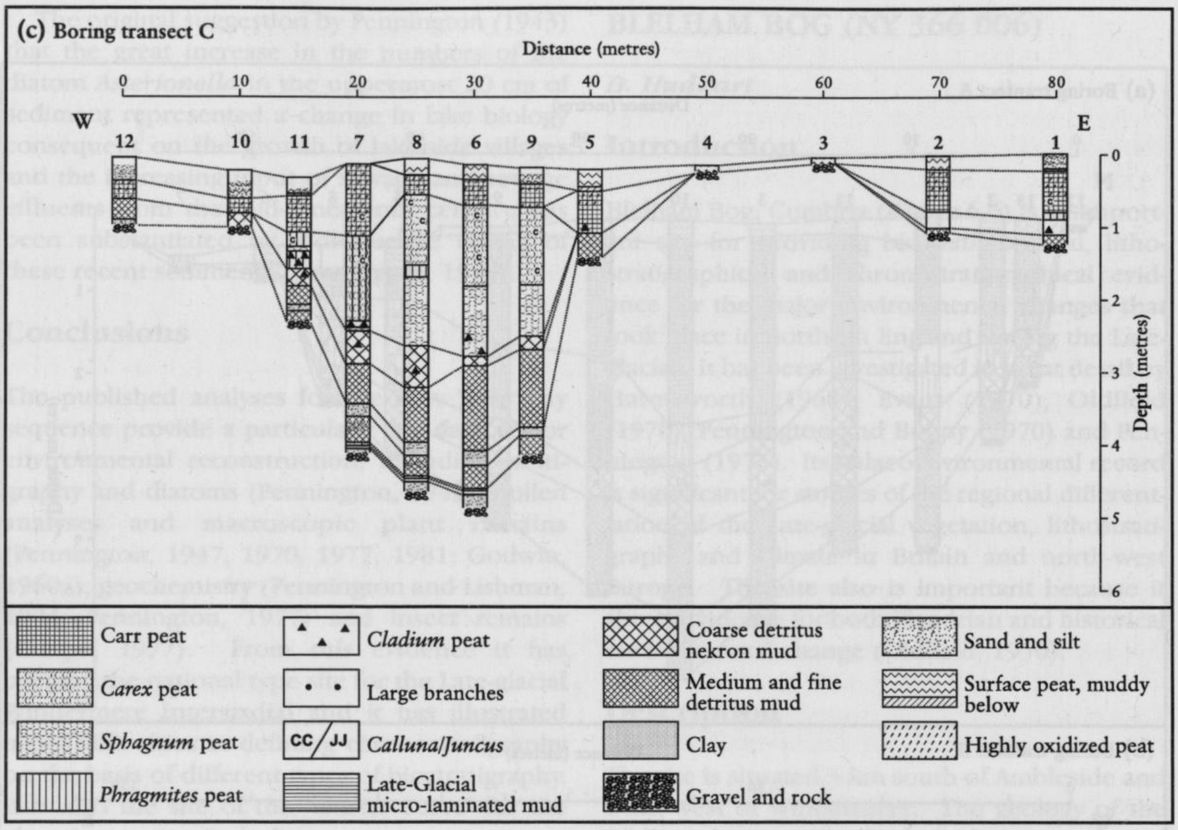



**Figure 6.7c** Stratigraphy at boring transect C (after Oldfield, 1970).

flags and slates of the Coniston Series, which are overlain by glacial drift in which several kettleholes were formed during Late Devensian deglaciation. Blelham Bog is developed in one such kettlehole, in gravels laid down marginal to ice stagnation in the Windermere basin (Hollingworth, 1951). Blelham Bog joins Blelham Tarn at its eastern end, but is separated from it on the south by a wooded and rocky knoll. Currently the bog is relatively flat but Oldfield's (1970) core transects show that the underlying topography comprises two kettlehole basins (Figure 6.7a–d).

### *Kettlehole stratigraphy*

The largest of the kettleholes occupies the eastern, unwooded part of the bog and extends westwards under the causeway. The detailed stratigraphy from four boring transects in this kettlehole are shown in Figure 6.7. A generalized lithostratigraphy at the western kettlehole (intersected by transects A and B) shows:

7. surface rootlet peats
6. carr peats
5. fen-swamp peats
4. reed-swamp peats
3. coarser shallow-water lake muds
2. clays alternating with greenish, open-water nekron muds
1. basal gravels

The sequence in transect C is similar, with one important exception. At C the basal deposits do not show a simple threefold division (clay–mud–clay) but instead a microlaminated organic mud occurs between the lowest clay–mud boundary. Radiocarbon dates put this layer at *c*. 14 300 years BP (Godwin and Willis, 1964; Pennington, 1973, 1975a). The Late-glacial pollen diagram is shown in Figure 6.8, which illustrates selected taxa as a percentage of the total pollen. The chemical profile is shown in Figure 6.9. Diatom sequences in the Blelham basin have been reported by Evans (1970) and differences between the flora of the tarn and the kettlehole have been recognized.

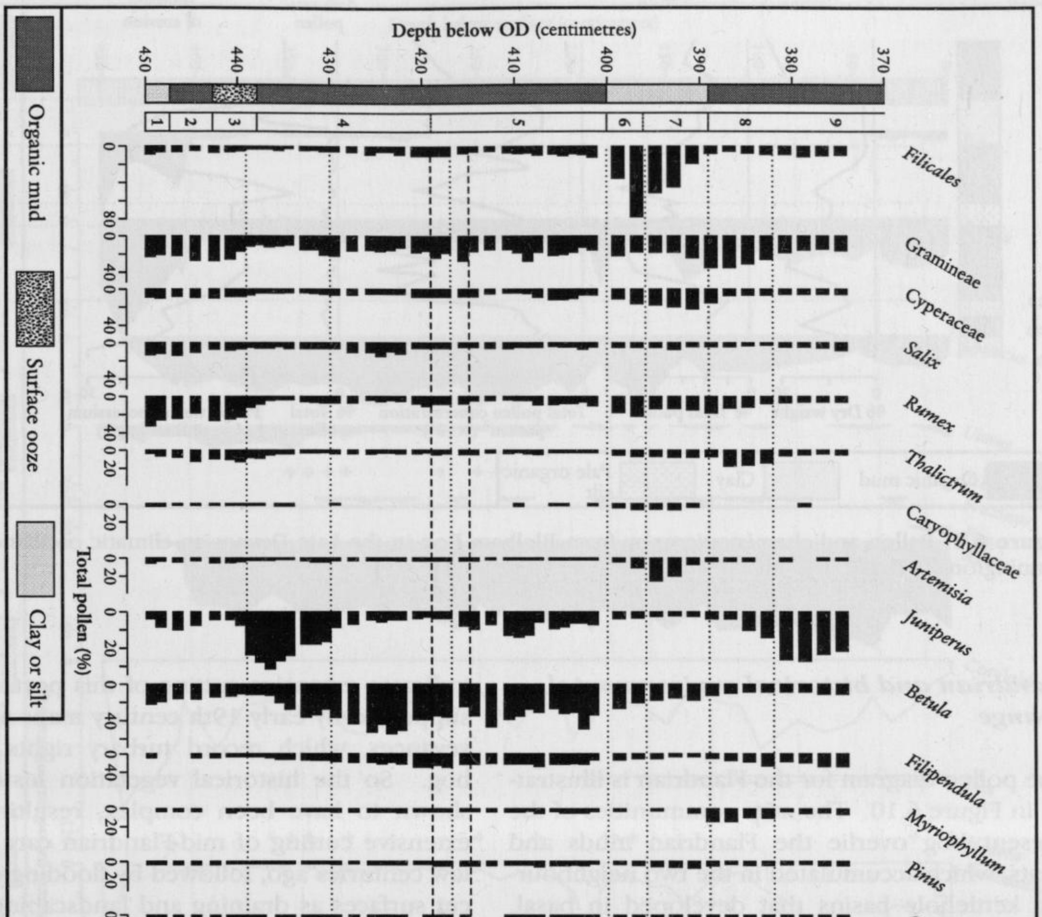



Figure 6.8 Late-glacial pollen diagram from Blelham Bog showing selected taxa as a percentage of total pollen (after Pennington, 1975a).

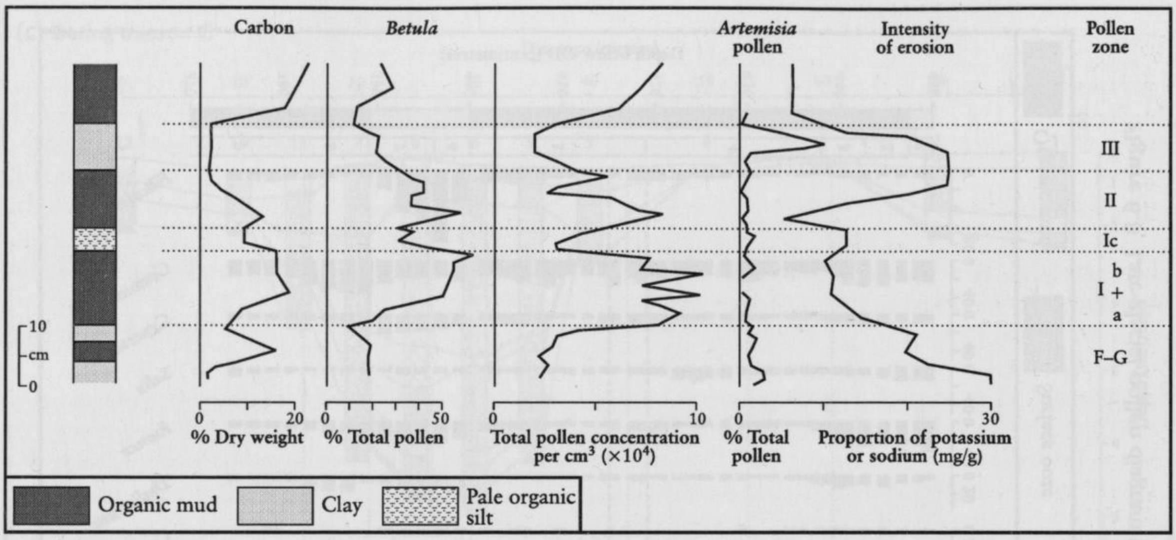



**Figure 6.9** Pollen and chemistratigraphy from Blelham Bog in the Late Devensian climatic oscillation (after Pennington, 1975a).

### *Flandrian and historical environmental change*

The pollen diagram for the Flandrian is illustrated in Figure 6.10. The mire communities of the present bog overlie the Flandrian muds and peats, which accumulated in the two neighbouring kettlehole basins that developed in basal, Devensian gravels. A detailed description of the site, its present vegetation and the stratigraphy and pollen analysis of the subsurface mire was published by Oldfield (1970), who showed that the vegetation history inferred from the pollen and macrofossils in the deposits differed from that originally postulated when the National Nature Reserve was set up. When the site was acquired by the former Nature Conservancy in 1954 the vegetation of the western kettlehole was a wet *Betula–Alnus–Salix* carr and that of the eastern one an acid *Sphagnum–Molinia–Myrica* bog. It seemed possible at this time that these two ecosystems represented two stages in the hydroseral development from carr to ombrogenous bog. A lacustrine bog developed over the lake sediments of the two small kettleholes, which changed from open water in the early post-glacial period to fen, carr and then bog (Oldfield, 1970). However, instead of the postulated hydroseral succession leading to a raised bog, Oldfield (1970) showed that raised bog peat formed on top of mid-post-glacial carr peat and that there is then a hiatus in the profiles that indicates extensive cutting of this peat. This is supported by early 19th century maps and conveyances, which record turbary rights on the bog. So the historical vegetation history was shown to have been complex, resulting from extensive cutting of mid-Flandrian carr peats a few centuries ago, followed by flooding of these cut surfaces as draining and landscaping of the slopes around the bog diverted drainage water over its surface. Other changes in the vegetation dated by Oldfield (1970) to the period 1848–1888 can be attributed to the construction of the present causeway and the diversion of Fish Pond Beck, together with the effects of lowering of the level of Blelham Tarn by deepening and straightening of the outflow stream. Grazing also has influenced the vegetation pattern, and enclosure since 1956 shows recent changes consequent on the cessation of grazing. This history of local land use suggested by Oldfield from stratigraphical and documentary evidence has been confirmed by biological and geochemical analyses of a long core from the neighbouring Blelham Tarn (Pennington and Lishman, 1971).

## Interpretation

### *Late-glacial sequence*

The significance of the western kettlehole stems from the Late-glacial sedimentary sequence. In

# Blelham Bog

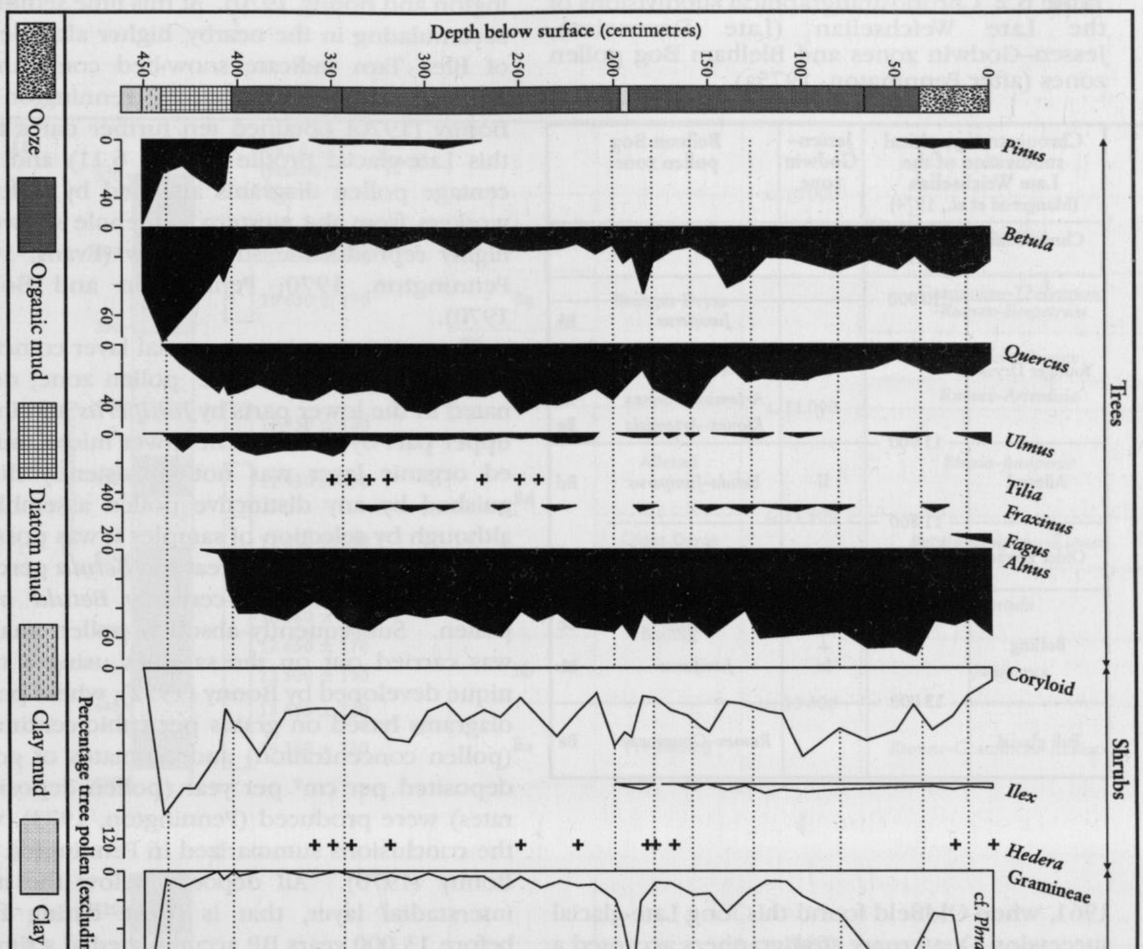



Figure 6.10 Flandrian pollen percentage diagram of Blelham Bog, typical of the vegetational history of the valleys of the southern Lake District (after Pennington, 1975c).

**Table 6.2** Chronostratigraphical subdivisions of the Late Weichselian (Late Devensian), Jessen–Godwin zones and Blelham Bog pollen zones (after Pennington, 1975a).

| Chronostratigraphical subdivision of the Late Weichselian (Mangerud *et al.*, 1974) | | Jessen–Godwin zone | Belham Bog pollen zone | |
|---|---|---|---|---|
| Chronozone | $^{14}C$ date (years BP) | | | |
| | 10 000 | | | |
| Younger Dryas | | III | *Juniperus* | Bh |
| | | | Gramineae–herbs–*Empetrum* | Bg |
| | | | *Artemisia–Rumex* | Bf |
| | | | *Rumex–Artemisia* | Be |
| | 11 000 | | | |
| Allerød | | II | *Betula–Juniperus* | Bd |
| | 11 800 | | | |
| Older Dryas | | Ic | *Betula–Rumex* | |
| | 12 000 | | | |
| Bølling | | Ib + Ia | *Betula* | Bc |
| | | | *Juniperus* | Bb |
| | 13 000 | | | |
| Full glacial | | | *Rumex*–Gramineae | Ba |

1961, when Oldfield found this long Late-glacial succession Quaternary stratigraphers accepted a Late-glacial pollen zonation of I, II and III (e.g. Godwin, 1956; Mitchell *et al.*, 1973) and equated the organic interstadial deposits of Windermere with zone II (see Table 6.2). They were aware of the possible existence of an earlier interstadial (e.g. Walker, 1966b) that would subdivide zone 1 (e.g. Bartley, 1962). Oldfield's demonstration of the microlaminated organic deposit in the silty clays below the main organic interstadial at Blelham Bog suggested that here was a horizon that might correlate with the Bølling Interstadial of mainland Europe (Iversen, 1954). However, Evans (1970), working on the diatom sequence, obtained the first $^{14}C$ date on the microlaminated mud (14 330 years BP), which was much older than the Bølling dates of continental Europe. This indicated that southern Cumbria had become ice-free at what was then considered to be a surprisingly early date, but subsequently this date was confirmed by $^{14}C$ dates from North Wales (Coope and Brophy, 1972), by duplicate dates and by consistent dates above the lowest dated horizon (Pennington and Bonny, 1970). At this time sediments accumulating in the nearby, higher altitude site of Blea Tarn indicate snow-bed communities dominated by *Salix herbacea*. Pennington and Bonny (1970) obtained ten further dates from this Late-glacial profile (Figure 6.11) and percentage pollen diagrams analysed by different workers from the western kettlehole showed a highly reproducible stratigraphy (Evans, 1970; Pennington, 1970; Pennington and Bonny, 1970).

The main organic interstadial layer coincided with a 'woodland biozone' pollen zone, dominated in the lower parts by *Juniperus* and in the upper part by *Betula*. The lower microlaminated organic layer was not consistently distinguished by any distinctive pollen assemblage, although by selection of samples it was possible to suggest a possible increase in *Betula* percentages, and some were certainly *Betula nana* pollen. Subsequently absolute pollen analysis was carried out on the samples using a technique developed by Bonny (1972), where pollen diagrams based on grains per cubic centimetre (pollen concentration) and estimates of grains deposited per cm² per year (pollen deposition rates) were produced (Pennington, 1973), with the conclusions summarized in Pennington and Bonny (1970). All deposits below the main interstadial layer, that is those dating from before 13 000 years BP, accumulated at a time of very low pollen deposition. This profile was interpreted as indicative of a single Late-glacial interstadial between *c.* 13 000 and 11 000 years years BP. Comparison with a $^{14}C$ dated profile from Cam Loch in Sutherland (Pennington, 1975a) was used to suggest that vegetation changes at corresponding dates (13 000 and 11 000 years BP) in these two regions of western Britain having different vegetation patterns, should be interpreted as indicative of climatic changes at these dates (Figure 6.11).

The lower (microlaminated) mud is approximately 40–50% organic content and is of a type not found elsewhere in Cumbria. It contains well-preserved remains of green algae, insects and material resembling decomposed lichen thalli. Although no arctic or subarctic insects were found in it by Coope (unpublished), Harmsworth (1968) found arctic species of Cladocera. In all the sediments older than 13 000 years BP, rates of total annual pollen deposition estimated from the $^{14}C$ dates never exceeded 200 grains $cm^{-2}$ $year^{-1}$, which is

## Blelham Bog

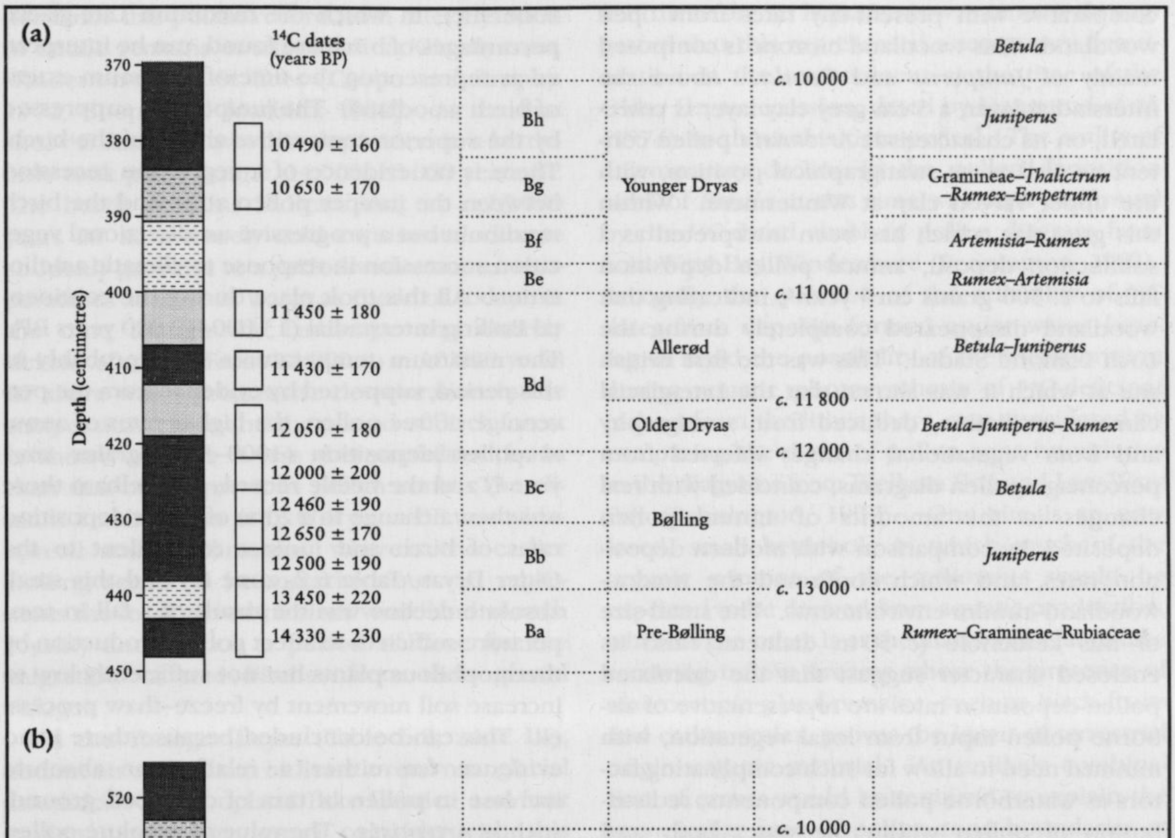



**Figure 6.11** Pollen zones and chronology in the Late-glacial (after Pennington, 1975a): (a) Blelham Bog; (b) Cam Loch, Sutherland.

comparable with rates from the present-day tundra (Davis *et al.*, 1973). In the overlying deposit of the main interstadial layer, total annual deposition increases to 2000–3000 grains $cm^{-2}$ $year^{-1}$, comparable with present-day rates from open woodland. This woodland biozone is composed mainly of *Juniperus* and *Betula*. Above the interstadial layer, a 5 cm grey clay layer is correlated, on its characteristic *Artemisia* pollen content as well as its stratigraphical position, with the upper varved clay at Windermere. Within this grey clay, which has been interpreted as a solifluction deposit, annual pollen deposition falls to *c.* 500 grains $cm^{-2}$ $year^{-1}$, indicating that woodland disappeared completely during the Loch Lomond Stadial. This was the first British site at which it was shown that the Late-glacial climatic oscillation, deduced from stratigraphy and from vegetational changes inferred from percentage pollen diagrams, coincided with real changes in the amounts of annual pollen deposited, by comparison with modern deposition rates, and which confirmed the tundra–woodland–tundra environments. The small size of this kettlehole (*c.* 50 m diameter) and its enclosed character suggest that the calculated pollen-deposition rates are representative of airborne pollen input from local vegetation, with minimal need to allow for such complicating factors as waterborne pollen components, redistribution of pollen within the water body and localization of sediment accumulation (see Likens and Davis, 1975).

### *Late-glacial pollen zone interpretation*

In the Late-glacial at Blelham the sequence of pollen zones described is illustrated in Table 6.2 and Figure 6.11. Prior to 13 000 years BP in pollen zone 'Ba' there was a grass sedge tundra with much *Rumex acetosa* and low percentages of juniper, interpreted as the product of prostrate shrubs dependent on a winter snow cover. The pollen concentration was low at tundra deposition rates. At Blelham in the Windermere Interstadial (*c.* 13 000–11 000 years BP) the lowest boundary of the main organic layer falls at *c.* 13 000 years BP and corresponds with a general climatic amelioration in north-west Europe. Above this horizon there is a rapid rise in the annual deposition rates of all the major pollen taxa and a particularly large increase in juniper, which was thought to be an immediate response to the improved climate by plants already present in the area. Through pollen zone 'Bb' there is a progressive increase in birch, which indicates a steady dispersal of tree birches towards the site. The overlying birch-pollen assemblage zone (Bc), in which the maximum Late-glacial percentages of birch are found, can be interpreted as representing the time of maximum extent of birch woodland. The juniper was suppressed by the superior competitive ability of the birch. There is no evidence of a vegetation recession between the juniper pollen zone and the birch maximum but a progressive unidirectional vegetation succession in response to climatic amelioration. All this took place during the continental Bølling Interstadial (13 000–12 000 years BP). The maximum temperatures were probably in this period, supported by evidence from the percentage of tree pollen, the higher rates of annual pollen deposition (1000–3000 grains $cm^{-2}$ $year^{-1}$) and the beetle record. At Blelham there was then a change to a zone of lower deposition rates of birch and juniper (equivalent to the Older Dryas, Table 6.2, zone Ic) and this small absolute decline was the result of a fall in temperature sufficient to affect pollen production by thermophilous plants but not sufficiently low to increase soil movement by freeze–thaw processes. This can be concluded because there is no evidence for either a relative or absolute increase in pollen of taxa of disturbed ground, such as *Artemisia*. The value of absolute pollen diagrams can be appreciated with regard to this phase. In Figure 6.8 there are only small changes at 419 cm and no separate pollen assemblage zone was originally distinguished. Absolute pollen analysis, however, shows up the real difference with respect to pollen content between the samples from 419 to 415 cm and those below 419 cm. From 419 to 415 cm there are pronounced minima in concentration and annual deposition rates of *Betula*, *Juniperus*, *Filipendula* and *Myriophyllum alterniflorum*, all comparatively warmth-demanding plants. The *Rumex* pollen is raised in the absolute pollen diagram in this narrow zone, which is barely perceptible in the percentage diagram.

This minor recession ended the warmest period in this interstadial and it led from then on to a fluctuating climatic environment, but on the whole with declining temperatures in pollen zone 'Bd' (dated at Blelham to 11 800–11 000 years BP). This zone shows an interplay between birch and juniper, which responded to the fluctuating climate, and at the same time there was a rise in deposition of open-environment herb pollen. The ecological equilibrium of the established birchwoods was broken and there was increased soil erosion. There is back-up evidence for this pollen record from interpretation of beetle remains at sites such as Glanllynnau, Windermere and St Bees (Coope and Brophy, 1972; Coope and Joachim, 1980). At the first site, $^{14}C$ dating shows that deposition in a kettle-hole took place from *c.* 14 000–10 000 years BP. The beetle remains indicate that at *c.* 13 000 years BP an intensely cold continental climate suddenly gave rise to a period with summer temperatures at least as warm as those of today. At the time though the landscape was entirely devoid of trees. From 12 000 to *c.* 10 000 years BP there was a progressive deterioration in the temperature curve and the period of birch forest is shown to have a less thermophilous fauna than that of the previous pollen zone. The climatic amelioration at 13 000 years BP was widespread and synchronous, whereas the earlier Bølling seems to be metachronous. It represents the local arrival of birch woodland at each locality and it took a period of time for the migration of tree birches from their glacial refuges.

In the Younger Dryas (Table 6.2, zone III), sediments at Blelham indicate a return to glacial conditions in the upland Lake District and four pollen assemblage zones coincide with this period. The two *Artemisia* zones (Be and Bf) coincide with a solifluction clay. There is a fall in birch and juniper pollen rates to under 100 grains $cm^{-2}$ $year^{-1}$, which indicates the disappearance of local woodland and pollen is often scarce. There also are occasional leaf macrofossils of *Salix herbacea*, a plant that is found only on mountain summits in the Lake District today. The later two zones (Bg and Bh) coincide with the second half of the period, when there was a rapid temperature rise, a stabilization of the land surface by a vegetation cover, expansion of sedges, grasses and herbs at the expense of the more open community plants, and mud became increasingly organic, and this becomes transitional to the Flandrian when birch percentages expand.

Comparison of the Late-glacial profile from Blelham Bog with other profiles made it possible to develop hypotheses on two problems in the Late-glacial period. Firstly, the comparison with Cam Loch (Pennington, 1975a) included consideration of whether the Late-glacial interstadial in western Britain was a single episode or divided. The minimum in pollen deposition rates of *Betula* at *c.* 420 cm at Blelham Bog is dated to *c.* 12 000–11 800 years BP (Pennington, 1975a) and therefore an hypothesis was tentatively suggested that this represented a temporary deterioration in the environment at the time of the Older Dryas recession, dated by Mangerud *et al.* (1974) to about this time period. The reality of a temporary decline in the pollen deposition rates of *Betula* at this time was then confirmed by the $^{14}C$-dated absolute pollen diagram from Low Wray Bay, Windermere (Pennington, 1981). Secondly, the question of the reliability of $^{14}C$ dates from deposits formed under water, bearing in mind the possibility of the incorporation during aquatic photosynthesis of $^{14}C$-deficient carbon from the lithosphere, was considered by comparing the dates of pollen zone boundaries in the diagrams from Blelham Bog and Low Wray Bay (Pennington, 1977). One site is an enclosed, small kettlehole in which much of the organic content of the sediments would be expected to be derived from aquatic production, whereas the other is at the margin of a large lake receiving inflow streams where the presence of macroscopic plant remains, such as birch fruits and catkin scales, proves the input of terrestrial origin organic material. An unlikely combination of errors would be required to explain the very similar ages of pollen zone boundaries at these neighbouring locations in terms of errors resulting from the incorporation of ancient carbon. Accordingly the view is that the $^{14}C$ dates from Blelham Bog show no evidence of error and are acceptable as a radiocarbon chronology for the Late-glacial stage of the Late Devensian.

### *Interpretation of the chemical analysis of the Late-glacial succession*

A summary of the chemical analysis of the sediments is shown in Figure 6.9. The maximum accumulation of biogenic sediments should take place under the ecologically stable interglacial climates and the minimum concentration of organic matter under glacial conditions, when all the soils have been destroyed by glacial erosion and deposition consists of rock flour and minerogenic sediments. Also, in periods of high erosion rate, soil material rich in potassium, sodium and magnesium is rapidly transferred to lake basins, whereas in periods of low erosion rate these elements are rapidly removed from the surface of the catchment by leaching. What material is transferred to the lakes is relatively poor in the above minerals. For Blelham the Late-glacial profile analyses show relatively high concentrations of the erosion indicators in both pre-interstadial and post-interstadial deposits. The minimum concentration is in the interstadial deposits, where the maximum values of total carbon shows the maximum humus accumulation in the maturing soil profiles.

***Interpretation of the Flandrian succession***

There is a contrast between the Flandrian vegetational history at Blelham and that recorded for upland areas. This can be seen in Figure 6.10 where the changes in the forest composition at the elm decline are similar to other sites, such as Blea Tarn, and the presence of a silt band in the deposits, accompanied by a lack of fern spores, suggests the same type of soil erosion. However, at Blelham a long phase of undisturbed secondary forest with *Fraxinus*, *Mercurialis* and very little Gramineae and *Plantago lanceolata* followed. There is then a distinct clearance phase, accompanied by an influx of silt that shows all the characteristics of a Landnam clearance as described by Iversen (1941). This is a temporary forest clearance, followed by more or less complete regeneration, with short-lived maxima of first, grass and herbs and then, the pioneer tree species, notably birch, followed by other components such as oak, which regenerates completely. This indicates that in the valley woods of the Lake District, the Neolithic to Bronze Age clearances essentially were temporary and were followed by complete forest regeneration, except that in each episode the elm reduced one further stage and it did not regain any ground. Above the Landnam phase there is evidence for a second clearance with an increase in grass and pasture herbs and the first appearance of bracken. The main forest clearance is indicated by a clearly defined band within the top metre of sediment, with cereal pollen and a peak of *Plantago lanceolata* and weeds. Pennington (1965) suggests that this late clearance is the result of the Viking land-takes, recorded by the place names of the two farms that occupy the immediate drainage basin (Tock How and Low Wray). The curve for *Ilex* in Figure 6.10 shows a contrast with the continental curve for this plant, and its expansion after the Landnam phase is interpreted as reflecting the more open character of the secondary forest and the resistance of *Ilex* to grazing. In the oceanic Lake District climate there seems to have been no climatic control (Pennington, 1965). The diatom flora from the kettlehole shows no increase in the ratio of planktonic to non-planktonic forms during the early Flandrian, whereas in the tarn, non-planktonic taxa also dominated at this time. The open water of the tarn was calcareous and included *Cocconeis diminuta* and *Achnanthes suchlandtii* during this period. A water depth of at least 12 m has been estimated for the later Boreal period. An increase in *Eunotia* in the kettlehole has been linked to the development of an increasingly acidic environment by the end of the Boreal period. By the middle of the Atlantic period, the pelagic zone of the tarn included over 70% of planktonic forms and their increase has been thought to be the result of increased amounts of organic compounds and of nitrogen, probably as a result of woodland growth, with *Alnus* being responsible for the enhanced levels of nitrogen. Alkalinity was reduced in the tarn during the Atlantic period, when *Eunotia* and *Frustulia* species either appeared or expanded in occurrence. Although such a trend also characterized the late Flandrian, the tarn water remained alkaline. However, a phase of eutrophication can be identified close to the present day and species such as *Asterionella formosa* and *Cyclotella glomerata* are diagnostic of this, with inputs to the lake from fertilizers and sewage the most likely cause. The formation of Sphagnum peat in the kettlehole in the late Flandrian is indicated by a suite of acidophilic diatoms, such as *Eunotia*, *Frustulia* and *Pinnularia*.

## Conclusions

The importance of this site lies mainly in the relatively high organic character of its Late-glacial layers, which accumulated in an enclosed environment, partly under anaerobic conditions. Its chronology has been established by extensive $^{14}C$ dating and it has been shown that an early climatic amelioration indicated by organic muds with tundra-type plant communities was caused by the onset of a maritime climate, which suggests that 'the regional differentiation of climate now apparent in the warmer winters and higher precipitation of western Britain came into being as soon as the Weichselian/Devensian ice retreated' (Pennington, 1975a). There were warmer conditions here than during the same interval in Holland and Scandinavia. The sediments at this site have recorded important palaeoenvironmental changes throughout pollen zones I–III in the Older Dryas, the Allerød and the Younger Dryas. Regional differentiation of palaeoclimate is possible during these periods, with important differences demonstrated between Blelham Bog and continental north-west Europe and between Blelham Bog and the nearby high-altitude site of Blea Tarn. It is one of the most important sites in Britain, where crucial evidence from a wide range of chemical, lithological, palynological, chronostratigraphical and palaeoclimatological techniques have been used to gain a better understanding of Late-glacial environmental changes, not just in northern England but also in north-west Europe. The site was used to reinforce the view that chronostratigraphy was the best method for correlating Late-glacial sequences. There also are important Flandrian and historical changes documented in the bog stratigraphy.

## BLEA TARN, LANGDALE (NY 293 044)

***D. Huddart***

### Introduction

This small, shallow, upland lake in Cumbria (Figure 6.12) is situated at an altitude of 190 m and is surrounded by rocky, mountainous Borrowdale Volcanic terrain, reaching up to 470 m (Figure 6.13). It lies in a through valley that hangs above Little Langdale, into which it drains. The basin was eroded by ice moving from Great Langdale to the north via a col at 220 m, and the tarn is not dammed by a moraine as the drainage overflows a bedrock ridge west of Tarnclose Crag. The shore of the tarn consists mainly of glacial drift and solifluction deposits covered by a thin peat. The site is important because lake cores have revealed detailed palaeoenvironmental reconstructions of Late-glacial and Holocene vegetational changes in the Lake District, including the Elm Decline (Pennington, 1970, 1973, 1975a, c). Data on changes in environmental geochemistry over these periods also are available (Pennington and Lishman, 1971) and the diatom stratigraphy has been described by Haworth (1969).

### Description

Long Late-glacial cores, including interstadial deposits, were analysed in detail for pollen, diatoms and geochemistry. The Late-glacial pollen and environmental geochemistry diagrams are illustrated in Figures 6.14 and 6.15, and a summary of the diatom record for the Late-glacial and Flandrian (Haworth, 1969) is shown in Figure 6.16. Percentage and absolute pollen diagrams were prepared from replicate cores and the diatoms were analysed from a parallel core. A summary of the stratigraphy and pollen zonation for the Late-glacial period is given in Figure 6.17 and the Flandrian pollen diagram is given in Figure 6.18.

### Interpretation

The results from the detailed core analyses show that there was no Loch Lomond ice in this upland basin (Walker, 1966a). The glacial drift ridges from around the tarn therefore must date from the closing stages of the Devensian deglaciation. Varved clay above a sandy silt indicate annual couplets of sedimentation into a lake fed by glacial meltwater sources. The long Late-glacial profile is important because it provides a record of the Windermere Interstadial in the uplands that can be compared with the lowland Lake District sites of Blelham Bog and Low Wray Bay, Windermere. A detailed comparison of the pollen assemblage zones is given in Pennington (1977).

#### *Late-glacial*

At Blea Tarn the pollen assemblage dominated by *Salix herbacea* has been interpreted as a flora of snow-bed tolerant communities by comparison with Norwegian mountain lakes (Pennington, 1973). It has been linked to the lowermost organic layer at Blelham Bog and both indicate warmer conditions than during the same interval in Holland and Scandinavia. This zone persisted until the opening of the woodland biozone, with juniper. There is evidence of a climatic recession that followed the first juniper phase, with a return to a more chianophilous vegetation with *Betula nana* and *Lycopodium selago*, but no evidence in the stratigraphy for any renewed frost-disturbance of the soils. It suggested to Pennington (1970) that it was quite possible in the oceanic conditions of western

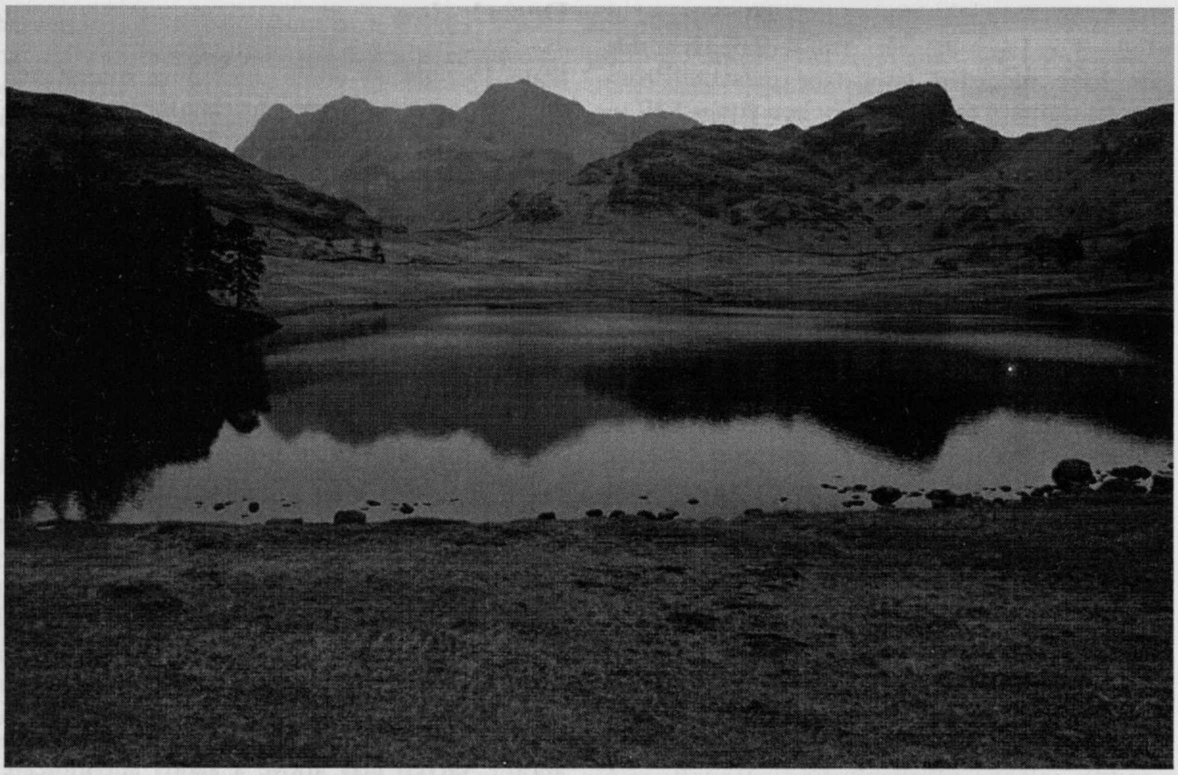


**Figure 6.12** Blea Tarn looking towards Langdale Pikes. (Photo: D. Huddart.)

Britain that temperatures may not have fallen very low during this period but that snowfall may have increased in the mountains, with an adverse effect on the flowering of juniper. The expansion of tree birches was delayed compared with the lowland Lake District sites and the maximum annual deposition of *Betula* pollen coincides with the beginning of minerogenic sedimentation indicative of late-interstadial soil erosion (Pennington, 1973). This shows that soil erosion began at this altitude before the end of the Windermere Interstadial. This is supported by evidence from Coleoptera in the Windermere sediments for falling temperatures before the end of this interstadial. The pollen diagrams from all the higher altitude sites, such as Blea Tarn, Devoke Water and Burnmoor Tarn, show lower percentages of total *Betula* pollen than in the lowlands, with for example 20–30% of the total pollen of land plants contributed by *Betula* at these sites and 40% in Low Wray Bay, Windermere and 60% at Blelham Bog (Pennington, 1970). During the Younger Dryas the sediment accumulation rates and pollen influx rates are high and this is taken to correspond to secondary deposition from reworked slope deposits as periglacial processes operated in the catchment.

The diatom history indicates more stable conditions than in lowland lakes, but a number of nutrient-demanding species were found only in Late-glacial and early post-glacial deposits, whereas in the lowest Late-glacial layers, species that live near glaciers were found. The inception of organic matter deposition in the earliest Flandrian was accompanied by an abrupt expansion in the diatom flora, which was dominated by planktonic species. This has been interpreted as indicative of a rapid increase in temperature at the time. Several alkaliphilous taxa, including *Asterionella formosa*, *Gyrosigma acuminatum* and *Navicula dicephala*, were present throughout the Boreal period, although the pH was probably more or less neutral. These species underwent decline or disappeared about the Boreal–Atlantic transition, when *Achnanthes minutissima* and *Synedra nana* came to dominance. At this time, *Pinnularia* species appeared, and the frequency of *Eunotia* species rose, the overall assemblage being more

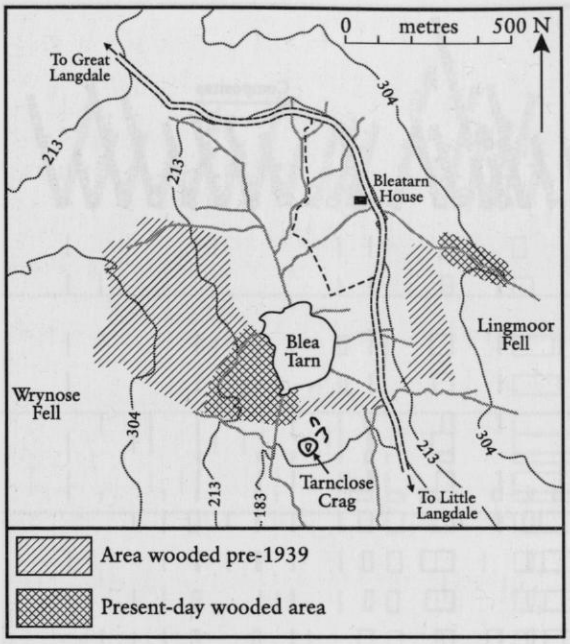



**Figure 6.13** Map of Blea Tarn and its drainage area.

tolerant of acid conditions. It has been suggested that the gradual disappearance of some taxa may have been caused by the inwash of organic material that had been building up constantly in the lake catchment, rather than a sudden climatic change. The Atlantic period shows minor shifts in the flora so that basiphiles disappeared. An increase in epiphytic species probably reflected the onset of *Sphagnum* bog formation adjacent to the lake. Sub-Atlantic acidification as a result of forest removal and soil erosion, as at Devoke Water, has not been detected at Blea Tarn, nor has the recent eutrophication shown by Pennington (1943) from Windermere and in Esthwaite Water by Round (1961).

Geochemical analyses on the first core were correlated by pollen stratigraphy with the core on which absolute pollen analysis was conducted (Pennington, 1973) and the two data sets were compared using principal components analysis, to illustrate the usefulness of this method in zoning hitherto unfamiliar profiles (Pennington and Sackin, 1975). A significant feature of the Blea Tarn absolute pollen diagram is that the concentration of pollen is high in the minerogenic deposits of the *Artemisia* pollen zone (post-interstadial) in the period of the Loch Lomond Advance. Severe periglacial soil disturbance is indicated by this. Using the curves for carbon and sodium (Figure 6.15) as indicating, respectively, the accumulation of soil humus and the relative erosion rate of mineral soils (Mackereth, 1966), the later part of zone I and the whole of zone II are seen as periods of soil stability, during which a continuous vegetation cover became established. The onset of renewed instability of soils, which reached its maximum development in the Younger Dryas, can be seen to coincide in the later part of the Windermere Interstadial with vegetation changes. A fall in the iron and manganese curves is interpreted as showing the presence of humic acids and reducing conditions in the soils, leading to podsolization and the spread of *Empetrum* heaths during the transition from the Windermere Interstadial to the Younger Dryas cold period. At the transition between these periods at Blea Tarn there is an interesting silty mud deposit that has about 50 narrow laminations, which are not graded varves but could represent a series of solifluction silts formed in a period of severe winters (Pennington, 1970). The pollen in this laminated deposit indicates a vegetation of *Empetrum* heath locally, with *Betula* pollen either from distant transport or as a secondary component in the solifluted soils.

Manley (1962) had suggested that between 50 and 100 years of cold summers and heavy winter snowfall could have been sufficient to re-establish glaciers in the Lake District high corries during the Younger Dryas and suggested that persistent flooding, associated with the disturbed chilly climate, prevailed. However, the chemical analyses from Blea Tarn show that the input of material into the tarn soon included mineral material so high in calcium and sodium that it must be regarded as unweathered drift from soil layers deeper than those that had been leached during the Windermere Interstadial. This is evidence that there was more severe periglacial erosion in this basin than had occurred since the Devensian glacial.

Contrasts in sediment composition between, on the one hand, the clays and silts below the interstadial deposits and, on the other hand, the post-interstadial minerogenic sediments, can be interpreted as indicative of a different origin. The high iron content and bright haematite colour of the lowest pre-interstadial sediments are consistent with a drift source from the highly haematized rocks of part of Great Langdale, which was pushed over the col at 220 m in early stages of deglaciation, when ice was still thick in

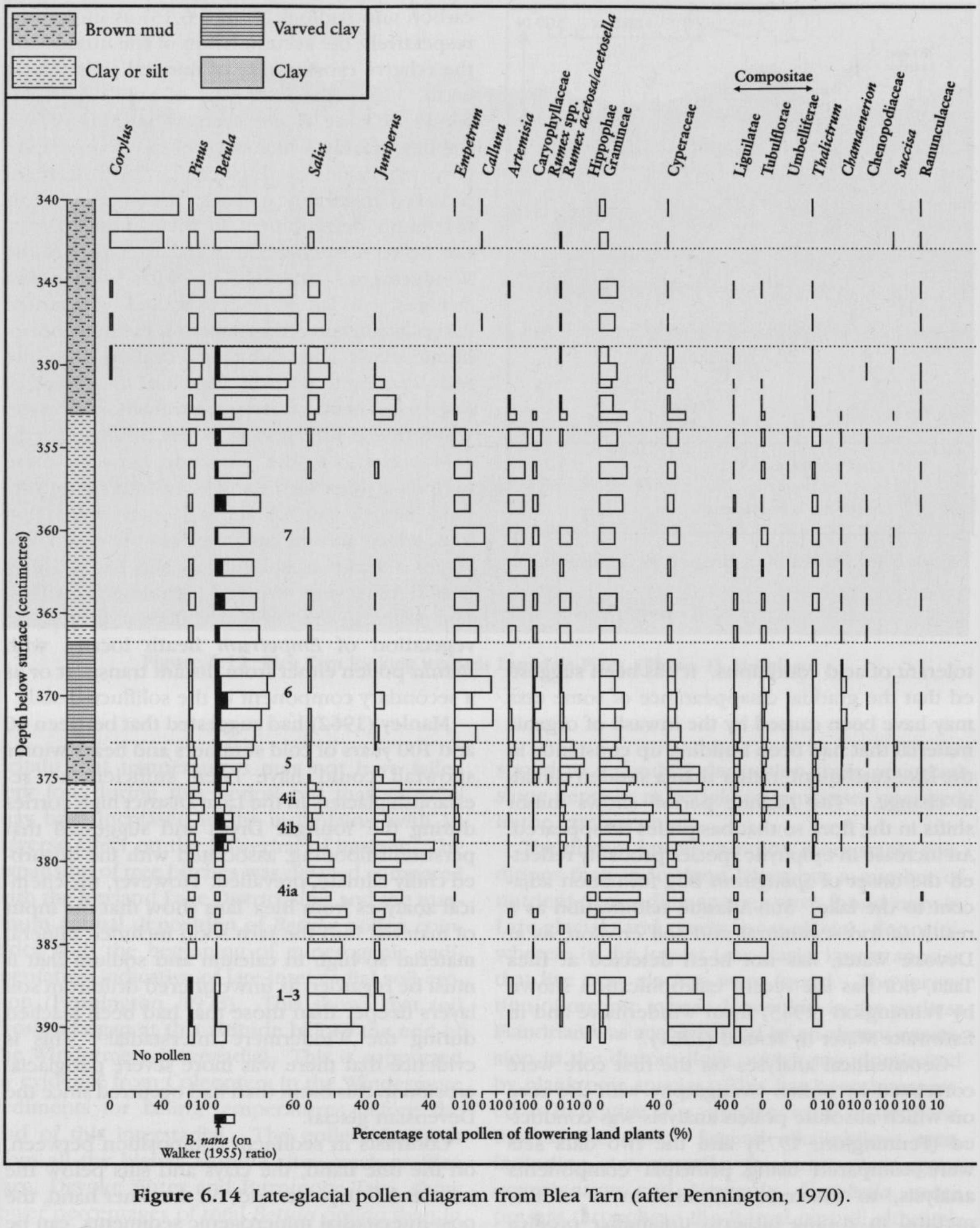
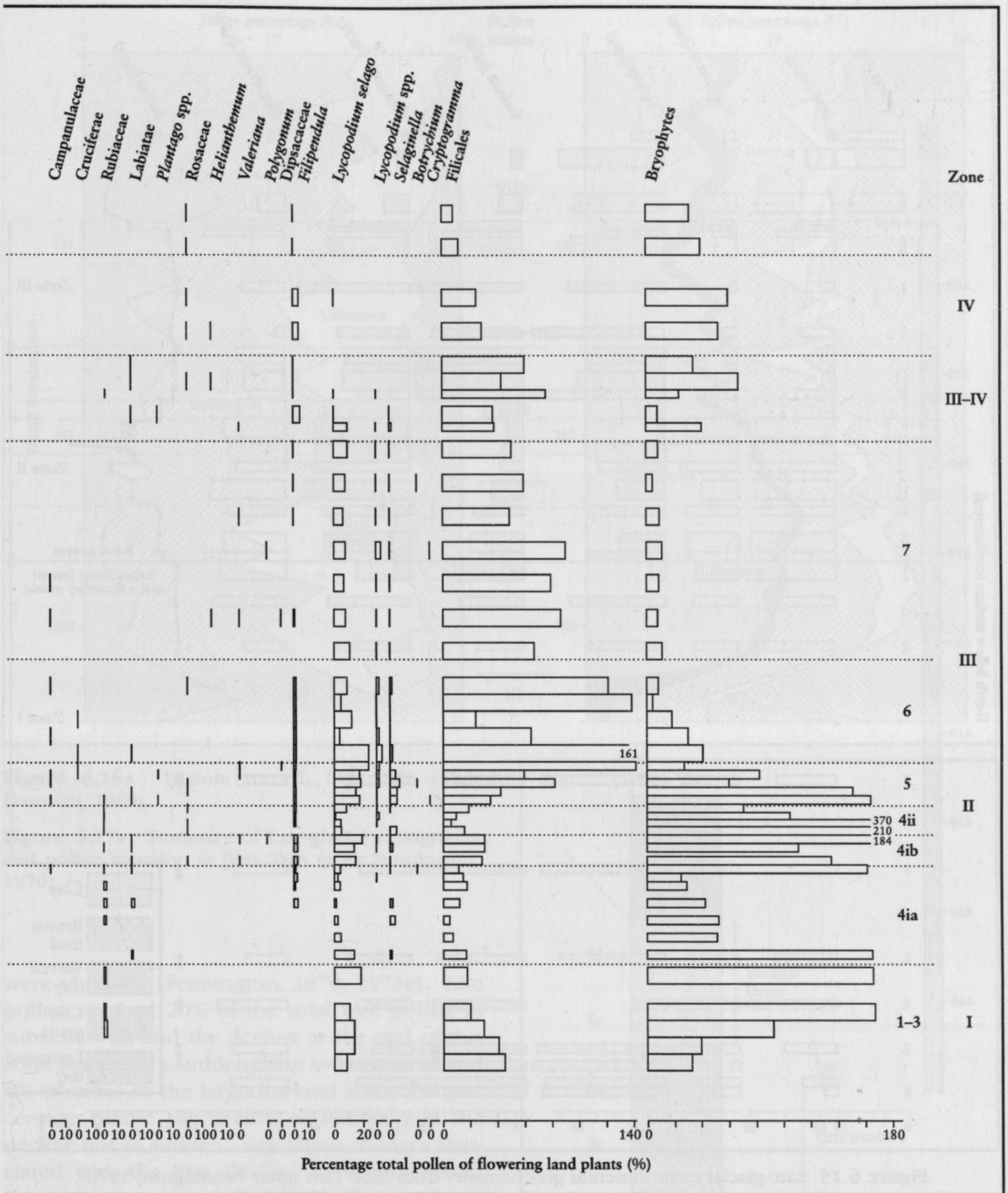



**Figure 6.14** Late-glacial pollen diagram from Blea Tarn (after Pennington, 1970).

Great Langdale. As this ice shrank below the level of the col, the composition of sediment deposited in Blea Tarn changed to that of the local rocks, which have less iron. The low drift ridges probably represent the final deposition from the Great Langdale ice lobe.

### *Holocene*

The post-glacial sediments of this small basin accumulated relatively rapidly in early Flandrian times and this site is a useful location for critical $^{14}C$ dating of the changes in forest composition




as the tree genera reached this site in their early Flandrian migration, and for a comparison with the more lowland valleys. The Boreal hazel maximum is lower than in the valley lakes, for example; furthermore, elm, which appears at the zone V–VI boundary, expands faster and further here than in the lower valleys. This suggested to Pennington (1964) that soils on the Borrowdale Volcanic rocks had a higher base-status in the early Flandrian than now. Late in subzone VI a pine maximum occurs at Blea Tarn, accompanied by small peaks of heather and *Sphagnum*

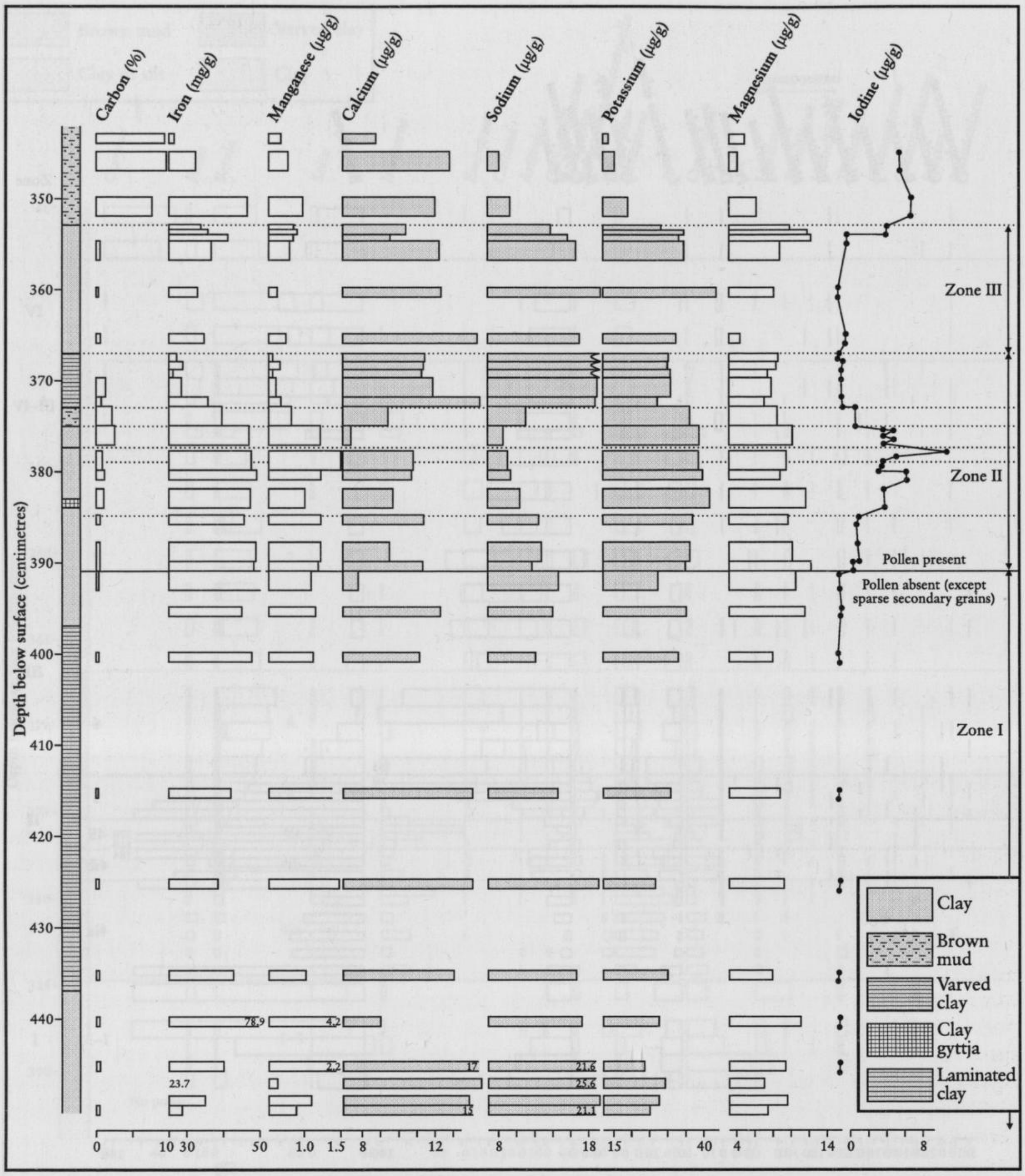



**Figure 6.15** Late-glacial environmental geochemistry from Blea Tarn (after Pennington, 1970).

and it has been suggested that this represents a local succession of well-drained marginal areas during the phase of lowered water levels late in subzone VI. Nevertheless, the presence of *Sphagnum* is surprising in this dry period. The expansion of alder at the VI–VIIa boundary is sudden and coincides with a stratigraphical change and it has been regarded as the tree of basin swamps up to high altitudes. On dry ground the forest consisted of oak, elm and birch.

The mid-Flandrian sediments contain a very precise pollen-stratigraphical record of the Elm Decline, for which entirely consistent $^{14}C$ dates

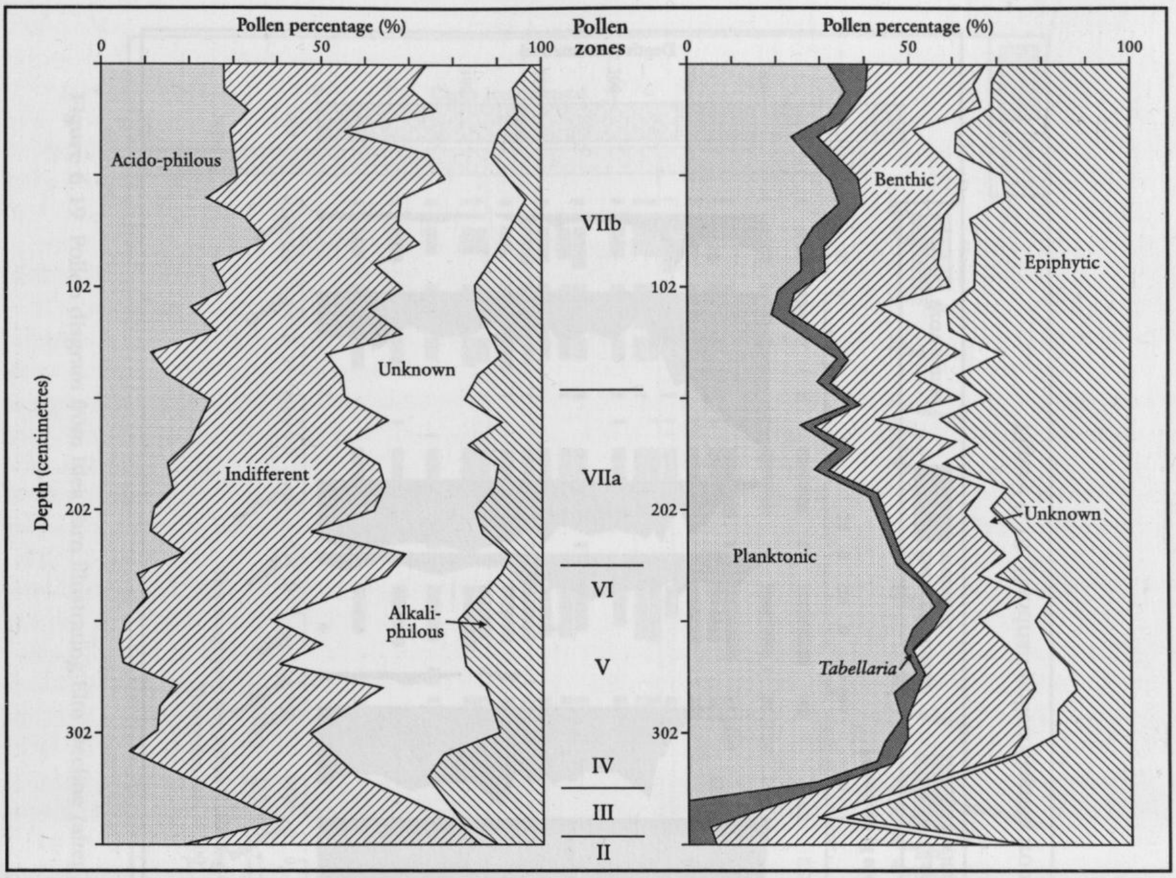
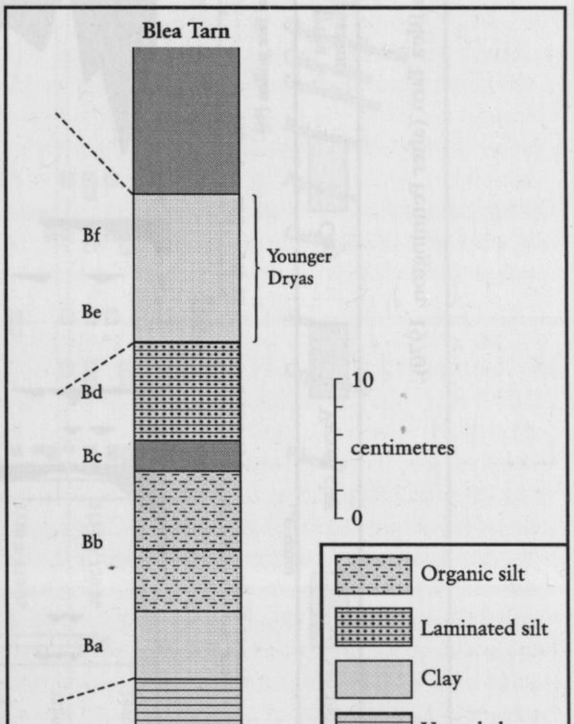



**Figure 6.16▲** Diatom record, Blea Tarn (after Haworth, 1969).

**Figure 6.17▸** Summary of Late-glacial stratigraphy and pollen zonation at Blea Tarn (after Pennington, 1970).




were obtained (Pennington, 1970, 1975c). Elm pollen reached 20% of the total tree pollen in subzone VIIa and the decline at the end of this zone is seen as a sudden drop to between 2 and 4% whereas in the large lowland lakes the percentage before the decline is smaller and the decline not so sudden. Vegetation changes associated with the Elm Decline, as reconstructed from percentage and absolute pollen analyses, have been integrated with the record of soil changes contained in sediment geochemistry, to reconstruct the palaeoenvironment of the Langdale valleys at the time of the activity of the Neolithic axe-factories on the north-eastern crags of Great Langdale (Pennington, 1973, 1975c). A consistent picture emerged of temporary forest clearance in the woods around Blea Tarn, but permanent destruction of forest that

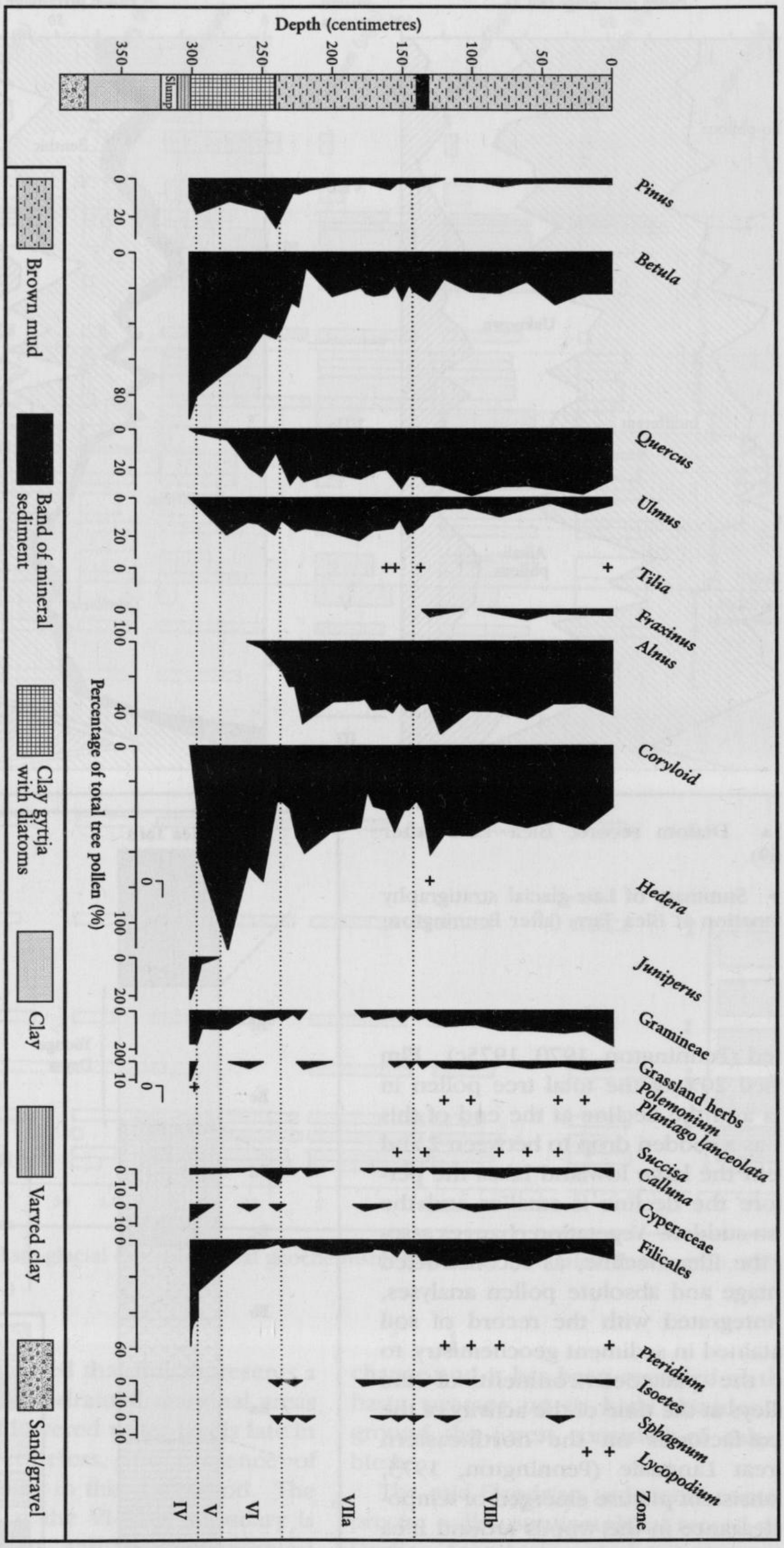



**Figure 6.18** Flandrian pollen diagram from Blea Tarn (after Pennington, 1970).

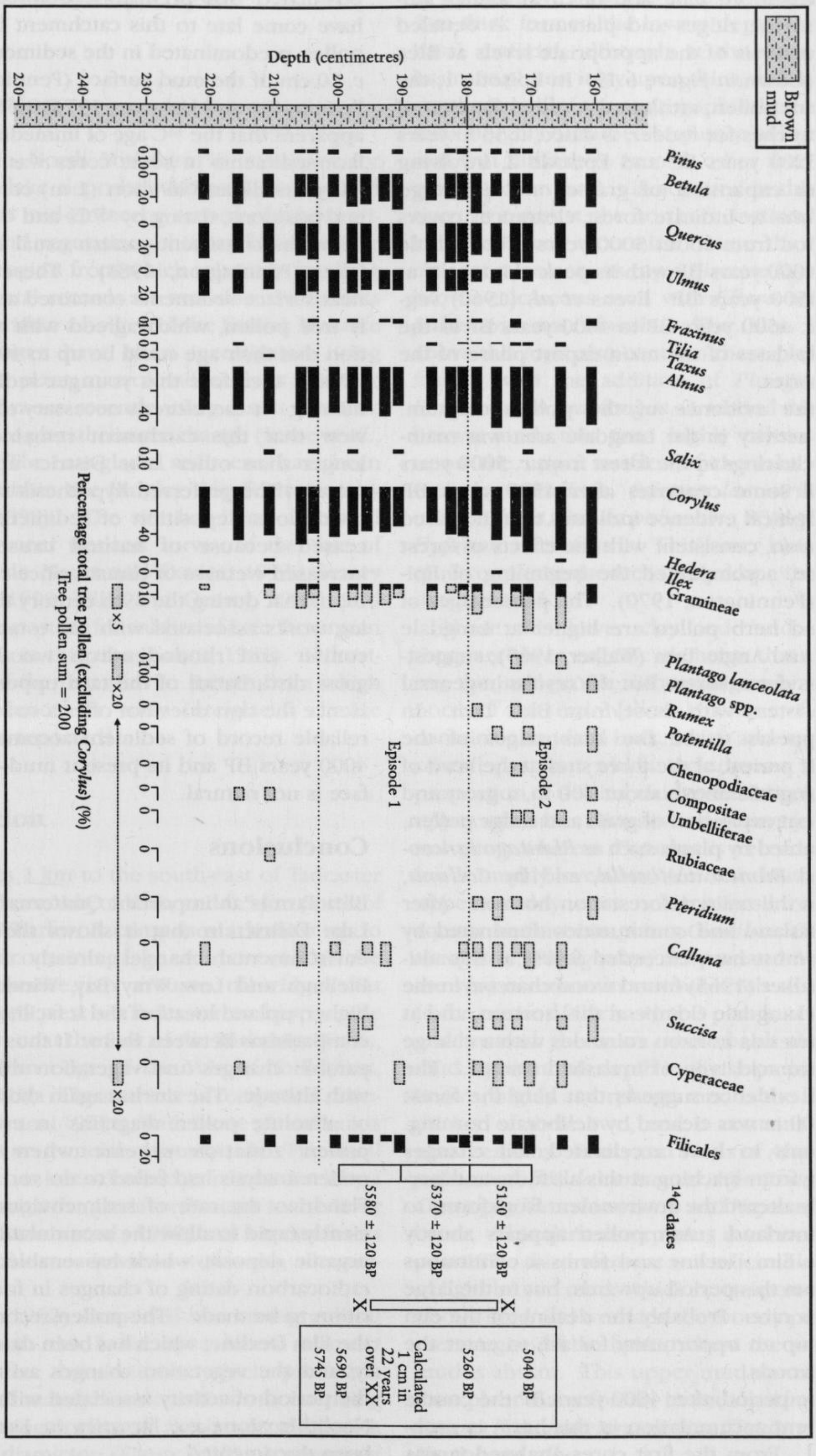



Figure 6.19 Pollen diagram from Blea Tarn illustrating Elm Decline (after Pennington, 1970). X–X marks the positions of two 14C date samples

contained more pine and birch at higher altitudes on the ridges and plateaux. A detailed pollen analysis of the appropriate levels at Blea Tarn is shown in Figure 6.19. In Episode 1, the fall in elm pollen, attributed to the collection of leafy branches for fodder, is dated at 5300 years BP to 5200 years BP, and Episode 2, involving sufficient expansion of grasses and *Plantago lanceolata* to indicate forest clearance, covers the period from about 5000 years BP to a little before 4000 years BP, with its peak of intensity at about 4500 years BP. Evens *et al.* (1962) suggested *c.* 4500 years BP to 4000 years BP as the probable dates of the main export phase of the axe factories.

On the evidence of the pollen diagram, human activity in the Langdale area was maintaining clearings in the forest from *c.* 5000 years BP until some centuries after 4500 years BP. Stratigraphical evidence indicates that increased soil erosion, consistent with the effects of forest clearance, accompanied the beginning of Episode 2 (Pennington, 1970). The percentages of grass and herb pollen are higher at Langdale Combe and Angle Tarn (Walker, 1965), suggesting a less dense forest, but the results in general are consistent with those from Blea Tarn. In what appears to be the later stages of the Neolithic period, at the three sites at the head of Great Langdale above about 400 m, a great and permanent expansion of grass and sedge pollen, accompanied by plants such as *Plantago lanceolata* and *Rumex acetosella*, and by *Calluna*, indicates the main deforestation horizon. After this, grassland and communities dominated by heather must have exceeded forest at this altitude. Walker (1965) found wood charcoal in the muds at Langdale Combe at this horizon, and at Angle Tarn this horizon coincides with a change to a more acid type of inwashed humus. The charcoal evidence suggests that here the forest above 400 m was cleared by deliberate burning. This seems to have accelerated soil changes resulting from leaching at this altitude, and permanently altered the environment from forest to open moorland. Ash pollen appears shortly after the Elm Decline and forms a continuous curve from this period upwards, but in the large lakes it is rare. Probably the decline of the elm opened up an opportunity for ash to enter the upland woods.

For the period after 4000 years BP the course of sediment accumulation in this basin is problematical. From the first cores analysed it was postulated that prehistoric deforestation must have come late to this catchment because tree pollen predominated in the sediments to within *c.* 10 cm of the mud surface (Pennington, 1964; Pennington and Lishman, 1971). It then became apparent that the $^{14}C$ age of immediately subsurface sediments in some cores was nearly 4000 $^{14}C$ years BP and in short (1 m) cores from several positions, dating by $^{137}Cs$ and $^{210}Pb$ failed to show the consistent pattern usual in Cumbrian lakes (Pennington, 1981). These anomalous near-surface sediments contained almost entirely tree pollen, which agreed with the supposition that their age could be up to 4000 $^{14}C$ years BP and therefore that younger sediments were missing. It therefore is necessary to modify the view that this catchment remained forested longer than other Lake District upland catchments. The preferred hypothesis is that either continuous deposition of sediment must have ceased because of natural causes, such as increased wetness of climate since *c.* 4000 years BP, or that during the 19th century the landscaping works associated with the establishment of conifer and rhododendron woods included gross disturbance of the tarn upper sediments. Hence the tarn does not offer a stratigraphically reliable record of sediment accumulation after 4000 years BP and its present mud–water interface is not natural.

## Conclusions

Blea Tarn is an important Quaternary site in the Lake District in that it shows the Late-glacial environmental changes already described at Blelham and Low Wray Bay, Windermere at a higher, upland location and it facilitates valuable comparisons between them. It thus shows comparable changes and vegetation differentiation with altitude. The site has again shown the value of absolute pollen diagrams in establishing a pollen zonation scheme when percentage pollen analysis had failed to do so. In the early Flandrian the rate of sedimentation was sufficiently rapid to allow the accumulation of a thick organic deposit, which has enabled important radiocarbon dating of changes in forest composition to be made. The pollen record picks out the Elm Decline, which has been dated accurately, and the vegetation changes associated with the period of activity associated with the nearby Neolithic stone-axe factories in Langdale have been documented.

## TADCASTER (SE 499 430)

*J. Innes*

### Introduction

This site in North Yorkshire is important for preserving lake deposits with a full Late-glacial and early to mid-Holocene record of lithostratigraphical and vegetational history from the Vale of York, a region from which relatively few long environmental records are available. Sediments occurring within a hollow in the Escrick Moraine near Tadcaster have been investigated using pollen and plant macrofossil analysis (Bartley, 1962), and more recently using insect remains, to study Late-glacial climate change (Lowe *et al.*, 1994a, b). The Late-glacial sediments have been classified within the Bingley Bog Formation by Thomas (1999). The site has been discussed during a recent wetland survey of the Vale of York (Van de Noort and Davies, 1993; Gearey and Lillie, 1999; Lillie and Gearey, 1999), and has been important for studies of Late Devensian deglaciation in this area (Gaunt, 1981). Tadcaster is one of the first sites in Britain from which pollen evidence for early Late-glacial climate change was reported.

### Description

The site lies 1 km to the south-east of Tadcaster in the southern Vale of York, and 0.5 km north of the River Wharfe. It comprises a hollow up to 6.5 m deep containing lake sediments within the Escrick Moraine, an arcuate till ridge that extends from west of Tadcaster to beyond the River Derwent at the foot of the Yorkshire Wolds, at the eastern edge of the Vale of York (Catt, 1977d, 1991b). The moraine is composed mainly of till (Gaunt, 1970a), with some glacial sands and gravels on its north-westerly side. The small lake basin has been drained in recent times by a tunnel through the glacial sediments to the River Wharfe (Edwards *et al.*, 1950) and it today forms an area of flat meadowland with a dry, silty surface soil.

Bartley (1962) examined the stratigraphical succession in the basin by a lateral transect of twenty-five hand cores, which proved a sequence of mainly limnic sediments, overlying the undulating surface of the basal pebble-rich sandy clay diamicton (Figure 6.20). Across most of the transect a calcareous blue lake clay up to 2 m thick formed the lowest deposit above the basal material, although at the edges of the basin, in places, it lay upon thin sand layers. The southern half of the stratigraphical transect contains a highly disturbed sequence of layers which makes it difficult to record, other than a generalized section, and it is not used in this description. In contrast, the northern section exhibits a clear sediment succession. Here the blue lake clay contains near its top a discrete layer of light brown calcareous mud with *Chara* and *Betula pubescens* macrofossils. The blue lake clay above this mud also contains these plant macrofossils, with the addition of *Phragmites* and *Potamogeton praelongus*. A second thin layer of brown mud, this time highly calcareous, lies upon the upper surface of the lake clay, and includes *Cladium*, *Scirpus lacustris*, *Phragmites*, *Chara*, *Betula pubescens*, *Nymphaea* and *Hypnum* macrofossils. This lower blue clay and mud couplet sequence is sealed by a thin layer of detritus mud.

Above this lower clay and mud sequence there are three organic elements that characterize the mid-profile lithostratigraphy. To the north of the section there is a layer of peat that contains abundant bark, twigs and wood in its central part but which includes an increasing proportion of detrital mud towards its lower and upper contacts. *Pinus* wood occurs in the central peat, whereas *Populus* is recorded in the lower muddy levels and *Alnus* and *Betula* in the upper muddy zone. Macrofossil remains of aquatic herbaceous plants *Cladium*, *Carex*, *Potamogeton*, *Phragmites*, *Scirpus* and *Nymphaea* are very common in various levels of this peat. In the centre of the basin this peat is not present, and instead there are two contrasting limnic mud layers at this depth in the sequence. A dark brown, highly amorphous mud is the lower and thicker of these two, and it contains very few plant remains. Fruits and seeds of *Alnus*, *Betula pubescens*, *Betula verrucosa*, *Nymphaea*, *Carex*, *Ceratophyllum demersum* and *Najas marina* do occur. This dark amorphous mud overlaps the peat. The third element is a much thinner layer of light brown mud, which rests upon the amorphous mud and overlaps the woody peat nearer the margins of the basin, where the amorphous mud is absent. This upper mud contains abundant plant macrofossils, including *Alnus*, *Carex*, various *Potamogeton* species, *Cicuta virosa* and *Nymphaea alba*.

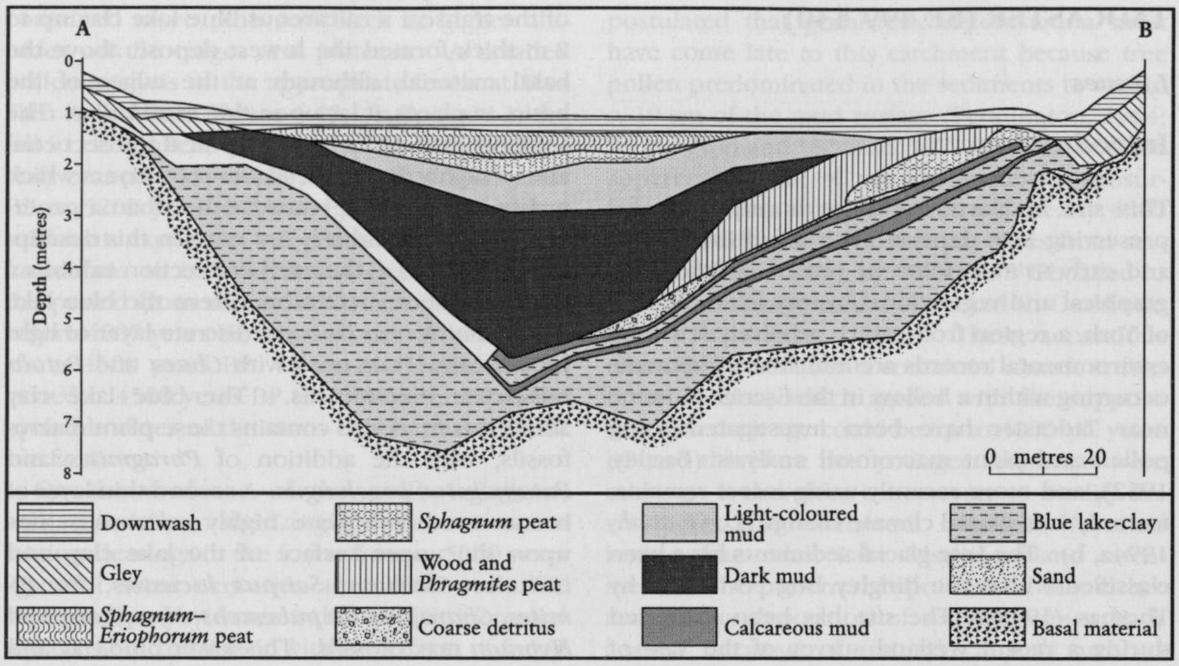



**Figure 6.20** Stratigraphical section through the basin at Tadcaster (after Bartley 1962).

The uppermost sequence of deposits is present in the centre of the basin only, where the surface of the limnic mud layers is concave in shape and upon which firstly fresher brown *Sphagnum*, and then more humified *Sphagnum* and *Eriophorum* peats have accumulated, both containing wood remains of *Alnus*, *Betula* and *Calluna*. The fresher *Sphagnum* peat is composed mainly of *S. cuspidatum* and contains fruits and seeds of *Nymphaea* and *Potamogeton*. The upper, more humified, peats contain abundant *Eriophorum vaginatum* cuticles, and *Sphagnum plumulosum* remains. A peaty, silty mineral soil seals the sequence across the whole of the basin.

### Pollen Analysis

Bartley (1962) constructed pollen diagrams through the northern part of the transect at core 14, where a fully representative sequence of sediments was preserved. These diagrams are shown as Figures 6.21 and 6.22. Figure 6.21 is calculated as percentages of total land pollen and is restricted to the Late-glacial succession, showing a more complete range of herb pollen taxa. Figure 6.22 covers the Holocene succession and is calculated as percentages of total tree pollen to reflect the dominant post-glacial vegetation type, only the major non-tree pollen types being shown. The following pollen zones are recognized:

**Late-glacial**

Ia Low *Betula* values, with *B. nana* consistent at around 10% of total pollen. Cyperaceae and Gramineae contribute more than 50%. Some thermophilous pollen, such as *Corylus*, is present in low numbers. Many open-ground herbaceous taxa occur, most commonly *Artemisia*, *Rumex acetosa*-type, Rubiaceae and a range of aquatic types. *Lycopodium selago* is prominent.

Ib *Betula* pollen rises to about 25%, and Gramineae pollen falls in proportion. *Juniperus* and *Thalictrum*, more thermophilous types, become consistently recorded.

Ic *Betula* frequencies fall again to a minimum of 6%, whereas Gramineae values are almost restored to their previous abundance. Other herbs such as *Helianthemum* increase, but percentages of most taxas remain steady. *Juniperus* frequencies increase in this subzone, and rise to a sharp peak of 44% near the top.

II This zone, which corresponds stratigraphically to the lower calcareous mud layer, includes the disappearance of *Juniperus* pollen and a

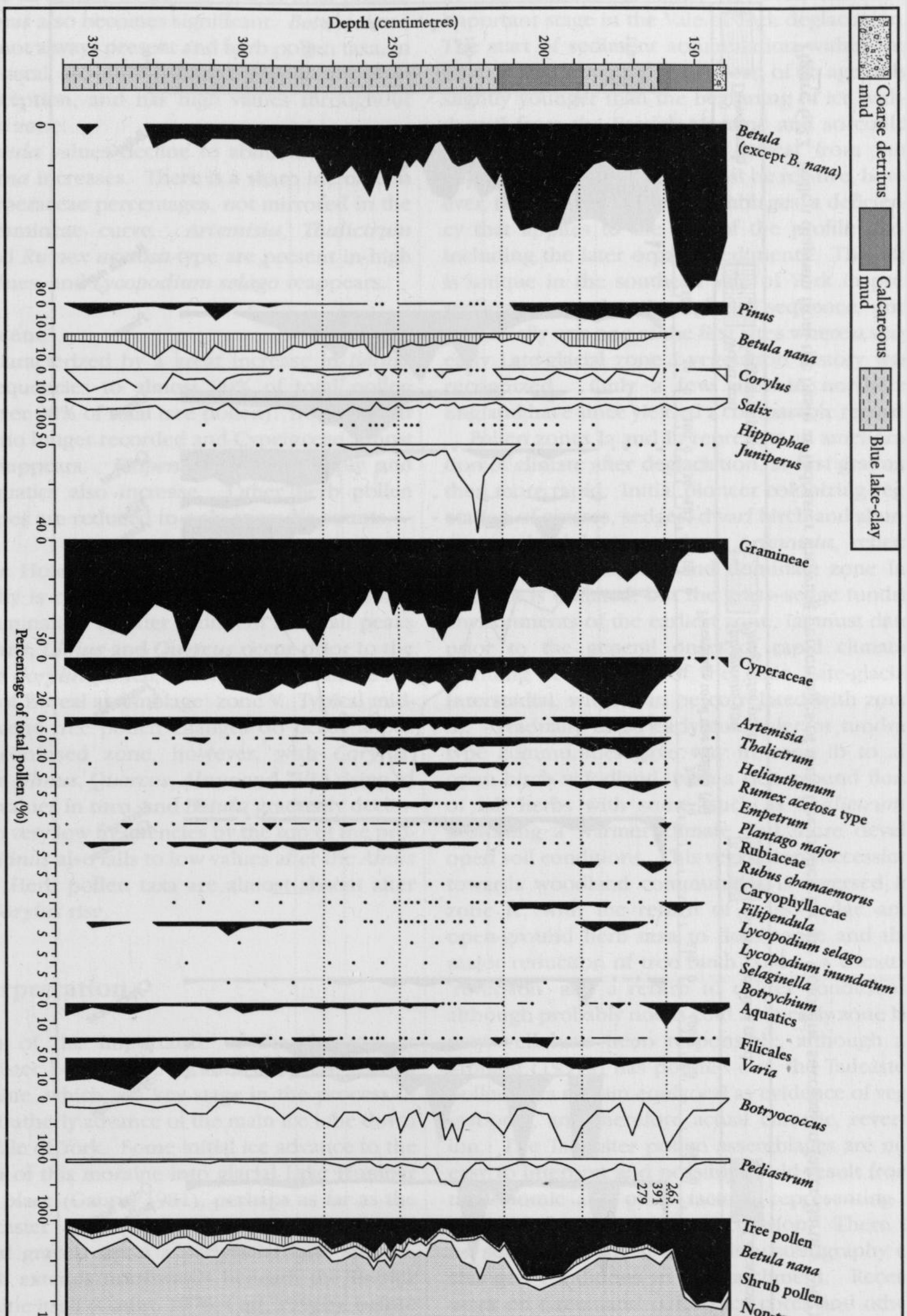



Figure 6.21 Pollen diagram from the basin at Tadcaster showing percentages of total land pollen for the Late-glacial succession (after Bartley 1962)

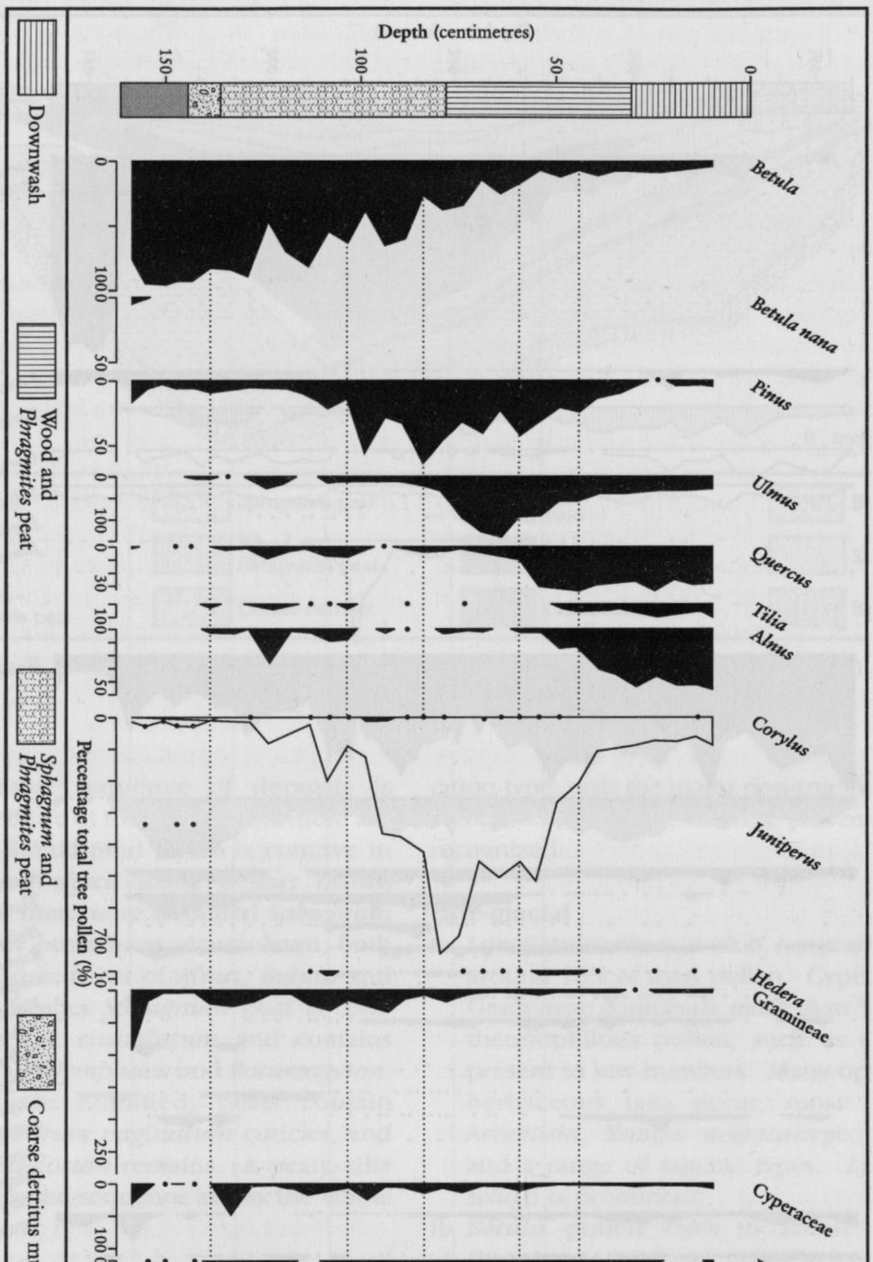



**Figure 6.22** Pollen diagram from the basin at Tadcaster showing percentages of total tree pollen for the Holocene succession (after Bartley 1962).

rise in *Betula* to almost 50% of total pollen. *Pinus* also becomes significant. *Betula nana* is not always present and herb pollen taxa, in general, are much reduced. *Filipendula* is an exception, and has high values throughout the zone.

III *Betula* values decline to about 20%, but *B. nana* increases. There is a sharp increase in Cyperaceae percentages, not mirrored in the Gramineae curve. *Artemisia*, *Thalictrum* and *Rumex acetosa*-type are present in high values, and *Lycopodium selago* reappears.

**Holocene**

IV Characterized by a great increase in *Betula* frequencies to almost 80% of total pollen (over 90% of total tree pollen). *Betula nana* is no longer recorded and Cyperaceae almost disappears. *Filipendula* rises sharply and aquatics also increase. Other herb pollen types are reduced to only sporadic counts.

The Holocene pollen record above zone IV initially is confused and perhaps reflects some contamination by later sediment as small peaks of *Alnus*, *Ulmus* and *Quercus* occur prior to the rise in *Corylus* pollen, which usually delimits the start of Boreal assemblage zone V. Typical mid-Holocene tree pollen changes do occur above this confused zone, however, with *Corylus*, *Pinus*, *Ulmus*, *Quercus*, *Alnus* and *Tilia* rising to high values in turn, and *Betula* gradually declining to very low frequencies by the top of the profile. *Pinus* also falls to low values after the *Alnus* rise. Herb pollen taxa are almost absent after the *Corylus* rise.

## Interpretation

Much of the importance of the deposits at Tadcaster lies in their location atop the Escrick Moraine, which is a key stage in the process of the southerly advance of the main ice lobe down the Vale of York. Some initial ice advance to the south of this moraine into glacial Lake Humber took place (Gaunt, 1981), perhaps as far as the Doncaster area (Catt, 1991b). It deposited glacial gravels upon a periglacial land surface, which extends northwards beneath the Escrick Moraine itself (Gaunt, 1976; Catt, 1991b), before ice-sheet retreat to the Escrick line. It therefore is unlikely that the Escrick Moraine represents a true terminal moraine and static ice front (Catt, 1977d; 1991b). It does, however, mark an important stage in the Vale of York deglaciation. The start of sediment accumulation within the Tadcaster hollow must have been of an age only slightly younger than the beginning of ice withdrawal from the Escrick Moraine and so could provide a limiting date for retreat from this depositional limit. Dating must be relative, however, based upon pollen assemblages, a deficiency that applies to the rest of the profile also, including the later organic sediments. The site is unique in the southern Vale of York in preserving a complete Late-glacial sequence, and historically was one of the first sites where a very early Late-glacial zone I vegetation history was recognized. Only a few sites in northern England have since yielded a comparable record.

Pollen zones Ia and Ib represent an amelioration of climate after deglaciation, at first gradual then more rapid. Initial pioneer colonizing vegetation of grasses, sedges, dwarf birch and abundant ruderal herbs, such as *Artemisia*, reflect tundra-type conditions and dominate zone Ia. The site is undated, but the grass–sedge tundra environments of the earliest zone, Ia, must date prior to the general onset of rapid climatic warming at the start of the main Late-glacial Interstadial, which can be correlated with zone Ib. Gradually these early cold-tolerant tundra-type communities gave way in zone Ib to an open birch woodland, with a rich ground flora of tall herbs with some, such as *Thalictrum*, indicating a warmer climate and more developed soil conditions. This vegetation succession towards woodland communities is reversed in zone Ic, with the return of grass, sedge and open-ground herb taxa to dominance and the major reduction of tree birch cover. A climatic 'reversion' and a return to colder conditions, although probably not as cold as in early zone Ia, may well have been responsible, although as Tipping (1991a) has pointed out, the Tadcaster pollen data remain equivocal as evidence of vegetational, and therefore actual climatic, reversion. The Tadcaster pollen assemblages are not easy to interpret and possibly could result from taphonomic and other factors, representing a unidirectional vegetation succession. There is no supporting signal in the lithostratigraphy of changed conditions in the catchment. Recent work on Greenland (GRIP) ice cores and other proxy data, however, has recorded a short-lived episode of colder temperatures in the Late-glacial Interstadial about 12 000 years BP, which supports the climate 'reversion' theory (Walker *et al.*, 1994; Björk *et al.*, 1998). Although birch trees probably did not disappear from the area during this zone Ic colder phase at Tadcaster, conditions for their existence must have been marginal at best. The great expansion of *Juniperus* towards the end of this phase, however, shows that although this short climatic deterioration was significant enough to suppress tree growth, it was not severe enough to prevent colonization by juniper-dominated shrub communities and the rapid re-establishment of tree birch woodland when amelioration ensued.

Tadcaster zone II is a typical main Late-glacial Interstadial (cf. Allerød) warm climate environmental record, with the rapid spread of birch trees to cover the landscape and shade out *Juniperus* virtually entirely. *Filipendula* is the only herb to be favoured during this phase, reflecting its thermophilous character, and the major decline of all other herb taxa demonstrates the closed nature of the woodland on the drier lands around the site. Most surviving herb types probably had wetland associations. Dates of between 13 000 and 11 000 years BP are inferred for this interstadial, which incorporates the traditional Bølling and Allerød continental warm periods, from many dated sequences elsewhere (e.g. Walker *et al.*, 1993, 1994). The sharp fall in *Betula* tree pollen in zone III and its replacement by a cold-tolerant, sedge tundra-type, herbaceous flora in which grasses, sedges and *Artemisia* are prominent, correlates with the Younger Dryas (Loch Lomond Stadial) arid period of severe cold between 11 000 and 10 000 years BP (Walker *et al.*, 1993). Sparse vegetation cover encouraged soil instability and erosion and in most sites sedimentation at this time is clastic in nature. In this respect the lithostratigraphy at Tadcaster supports the pollen stratigraphy in that the periods of conspicuously warmer climate and more developed vegetation of zones II and IV correlate with more organic mud layers, whereas the colder phases are times of blue lake clay deposition. The end of the zone III cold phase and the start of the Holocene (zone IV) is defined by the very rapid expansion of full *Betula* woodland, with very few open-ground indicators. This transition was achieved without any intervening transitional *Juniperus* phase, unlike several sites farther to the north, such as those in east Durham (Bartley *et al.*, 1976), but similar to nearby Bingley Bog (Keen *et al.*, 1988).

Tadcaster is highly significant in that it contains the dual *Betula* maximum in the Late-glacial succession, with the earlier *Betula* peak allowing the subdivision of Late Devensian zone I at the site. Dating of this earlier of the Late Devensian *Betula* peaks at Tadcaster is still uncertain, but correlation with the start of the main Late-glacial Interstadial would mean that it is almost certainly in the order of 13 000 years BP, indicated by radiocarbon evidence from sites in the region, such as Gransmoor (Walker *et al.*, 1993). At Tadcaster, as in northern England in general, a birch woodland became established in this earlier interstadial phase, which was lighter than that of the later, zone II, woodland biozone. A dual Late-glacial *Betula* peak subsequently has been recognized at other sites in the area, such as The Bog, Roos in Holderness (Beckett, 1981), Seamer Carrs and Kildale Hall on the western edge of the North York Moors (Jones, 1976a, 1977b) and Thorpe Bulmer in east Durham (Bartley *et al.*, 1976). Surprisingly, thermophilous pollen grains occur throughout the Late-glacial record at Tadcaster, explained by Bartley (1962) as the result of reworking. This probably is the case, but Keen *et al.* (1988) have recorded similar thermophilous pollen in Late-glacial contexts, which they ascribe to long-distance transport from refugia to the south. Growth of these taxa, for example *Corylus*, near to Tadcaster in the early Late Devensian seems very unlikely, although Late-glacial *Alnus* macrofossils from Willow Garth in east Yorkshire (Bush and Hall, 1987) suggest such records cannot be dismissed.

The Holocene environmental record at Tadcaster is unexceptional and also truncated, finishing in mid-Holocene post-*Alnus* rise, pre-*Ulmus* decline (Flandrian II) times (Lillie and Gearey, 1999). The successive immigration of thermophilous trees *Corylus*, *Pinus*, *Ulmus*, *Quercus*, *Alnus* and *Tilia* follows a pattern repeated in the other sites in the Vale of York, varying according to local environmental factors. Nearby analogous records are Burton Salmon (Norris *et al.*, 1971) and Askham Bog (Gearey and Lillie, 1999). The demise of *Pinus* at Tadcaster after the *Alnus* rise suggests that these two factors are linked, and this supports the view of Bartley (1962) that the macrofossil content and lithostratigraphy of the upper sediments record a major change from a zone VI dry depositional surface to its replacement in zone VIIa by flooding and deeper water. The peats of zone VI age carried pine trees in the upper layers and the contemporaneous dark oxidized limnic mud in mid-basin also points to low water tables. After the *Alnus* rise, however, the light brown mud with abundant aquatic plant-remains reflects deep water, followed by a succession to wet *Sphagnum* swamp peat. This appears to be a very clear record of mid-Holocene climate change.

## Conclusions

Tadcaster is a classic site in the early study of Late-glacial vegetation and climate change in Britain, providing the earliest analysed, full, pollen record for the tripartite stadial–interstadial–stadial succession. Its evidence of two warm phases encouraged further subdivision of this succession and the recognition that more than one, or at least one complex, Late-glacial Interstadial existed in Britain. This allowed correlation with the data from continental north-west Europe and with its system of classification. Much more detailed research is now available for the history of climate and vegetation change in the British Late-glacial, and the Tadcaster site requires renewed and detailed examination to fulfil its clear potential.

# GRANSMOOR (TA 113 597)

**POTENTIAL GCR SITE**

***D.J.A. Evans***

## Introduction

Valuable information on the floral and faunal changes of the period post-dating the recession of the last (Dimlington Stadial) ice sheet in east Yorkshire occurs in a sand and gravel pit west of the village of Gransmoor. The exposure cuts one of a series of sand and gravel ridges and intervening depressions, or kettleholes (Figure 6.23). The rich flora and fauna have allowed reconstructions of palaeoenvironmental change during the Late-glacial Interstadial and Loch Lomond/Younger Dryas Stadial in Britain, and

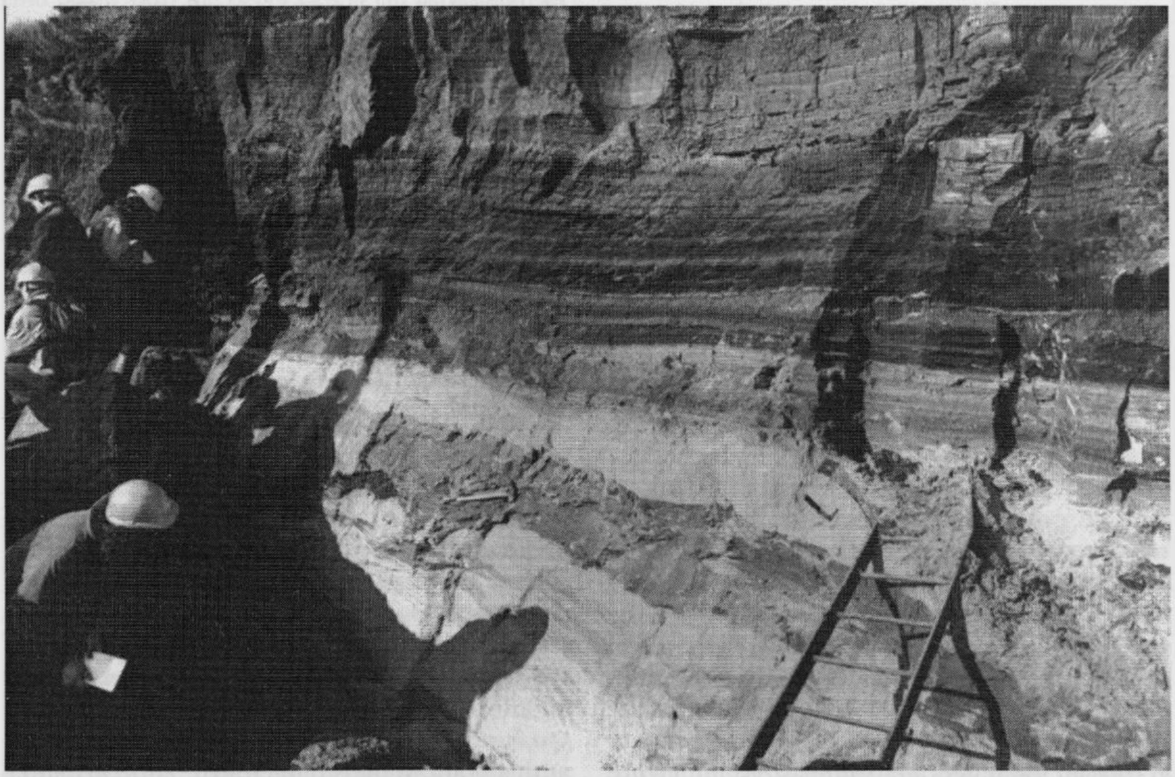


**Figure 6.23** The stratigraphical succession at Gransmoor showing the vertical grading from minerogenic to organic-rich to minerogenic sediments. (Photo: D.J.A. Evans).

comparisons with the high-resolution ice-core records from Greenland (Walker *et al.*, 1993; Lowe *et al.*, 1995b, 1999; Mayle *et al.*, 1999).

## Description

The sand and gravel pit lies at an altitude of 8 m OD and at a distance of 6 km from the North Sea coast. The sand and gravel ridges form part of a network of N–S-aligned features that have been interpreted as kames, eskers and marine storm beaches (e.g. Lamplugh, 1925; Catt and Penny, 1966; Eyles *et al.*, 1994), although a systematic investigation of their sedimentology and regional context is still awaited. The organic remains and associated sands, silts and clays are 235 cm thick and occur in a depression between sand and gravel ridges, where they overlie sands and silts with intermittent horizons of fine-grained gravel and coal fragments (Figure 6.24). The stratigraphical sequence has been subdivided into 16 lithostratigraphical units by Walker *et al.* (1993), but is summarized here using the fourfold scheme of Lowe *et al.* (1995b). First, laminated silts and sands occur from 240–200 cm; second, organic-rich beds occur between 200 and 150 cm; third, laminated silts and sands with occasional moss layers occur from 150 to 20 cm; and fourth, Holocene peats cap the section.

Chronological control on the depositional sequence is provided by 19 radiocarbon dates on plant macrofossils from specific stratigraphical horizons (Figure 6.24). The four oldest radiocarbon age determinations have been rejected by Lowe *et al.* (1995b) owing to the probable contamination by hard-water error. The remaining dates indicate an overall increase in age with depth, and statistical analysis by Lowe *et al.* (1995b) highlights a distinct change in sedimentation rate at 130 cm or at approximately 12 300 calendar years BP (Figure 6.25).

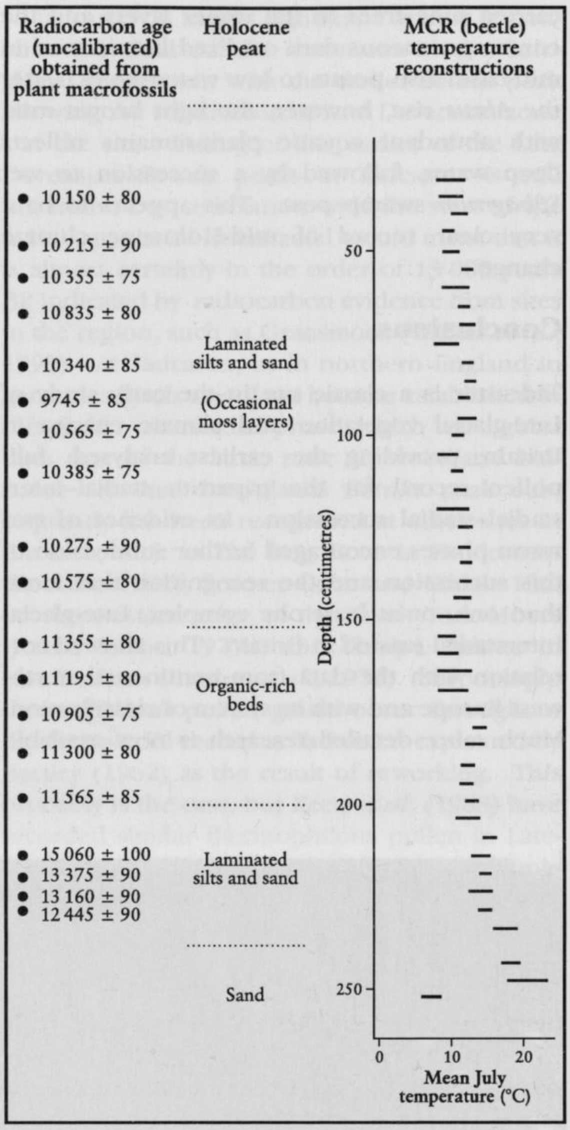



**Figure 6.24** The stratigraphical succession relating to the Late-glacial and Younger Dryas at Gransmoor. Uncalibrated radiocarbon dates and MCR (Mutual Climatic Range) beetle temperature reconstructions (from Lowe *et al.*, 1995b).

The remains of pollen and Coleoptera provide proxy data for reconstructing climatic trends over the period of deposition. A total of 304 different taxa of Coleoptera were identified in the sediments (Walker *et al.*, 1993). The species group has been determined for 214 of the Coleoptera, and 29 species possess present-day geographical distributions outside the British Isles. The pollen data for the stratigraphical sequence are summarized in Figure 6.26 and Table 6.3. The biostratigraphical boundary between the Late-glacial Interstadial and the Loch Lomond/Younger Dryas Stadial is placed at the point where: (a) *Betula* falls to very low levels; (b) Cyperaceae is expanding; (c) *Rumex* and *Thalictrum* are increasing; and (d) there is a marked reduction in *Equisetum*. In addition, the continuous occurrence of pre-Quaternary palynomorphs (pollen) also begins at this point in the stratigraphy.

Of particular interest to the archaeology of the area is a barbed point that was recovered from

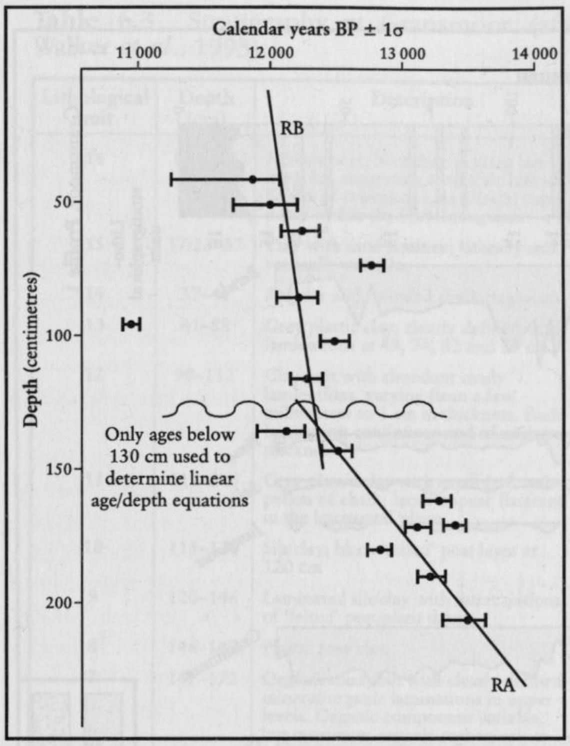



**Figure 6.25** Age–depth plot of radiocarbon dates in calendar years BP for Gransmoor, including the two separate linear regression lines for dates above (RA) and below (RB) the depth of 130 cm (from Lowe *et al.*, 1995b).

between lithostratigraphical units 6 and 7 at 170 cm (Sheldrick *et al.*, 1997). Worked from part of an antler, the point was found embedded in a log that was deposited contemporaneously with its enclosing sediments.

## Interpretation

The radiocarbon dates (Figure 6.24) indicate that the lower organic horizons span most of the Late-glacial Interstadial. The upper minerogenic sediments date from the Loch Lomond/Younger Dryas Stadial (11 000–10 000 years BP). Therefore, the sedimentary succession records deposition of laminated sands, silts and clays in a depression lying between glaciofluvial ridges after the recession of the Dimlington Stadial ice sheet from east Yorkshire.

Palaeoclimatic and palaeoenvironmental reconstructions have been produced for the sequence based upon a combination of pollen and coleopteran evidence. Specifically, the July temperature curve (Figure 6.24) is based on the Mutual Climatic Ranges (MCR) of individual Coleoptera. This method first establishes the range of climates occupied at the present day by the species in the fossil assemblage. The area of overlap of the climatic ranges of all the species present is then taken to reflect the climate of the whole (indicator) assemblage.

The main features of the pollen record represented in Figure 6.26 have been summarized by Walker *et al.* (1993) as follows.

1. A pre-*Juniperus* phase is marked by the dominance of Cyperaceae and Gramineae, with significant amounts of *Artemisia*, *Rumex*, *Thalictrum* and *Helianthemum*, and relatively abundant *Salix*. This phase therefore represents a period of open steppe or grassland vegetation, which developed on the freshly deposited glaciogenic sediments of the area.
2. A *Juniperus* phase (222–208 cm), wherein the *Juniperus* pollen exceeds 15% of the total land pollen (TLP), Gramineae and Cyperaceae remain high, there are peaks in *Artemisia* and *Thalictrum*, and *Betula* accounts for 10–15% of TLP. This is characteristic of an open landscape with few trees and abundant shrub juniper. The expansion of *Juniperus* is thought to reflect significant climatic amelioration, with warmer winters and summers.
3. A *Betula* phase, marked by a decline in *Juniperus* and an increase in *Betula* to values exceeding 50% of TLP. In addition, high counts occur for *Salix* and *Filipendula*, and *Artemisia*, *Thalictrum* and *Helianthemum* levels drop. A distinctive feature of this phase is an oscillation in the *Betula* curve characterized by a fall and rise in the genus following its initial abrupt rise. The fall is accompanied by rises in Gramineae and Cyperaceae and the open-ground taxa *Rumex*, *Thalictrum*, *Helianthemum* and *Artemisia*, leading Walker *et al.* (1993) to suggest a reduction in the woodland cover and an increase in open grassland habitats. A similar pattern in vegetational changes has been documented at other sites in the region, for example, at Tadcaster (Bartley, 1962), Thorpe Bulmer and Kildale Hall (Bartley *et al.*, 1976; Jones, 1977a).
4. The Loch Lomond/Younger Dryas Stadial, reflected in the dominance of Cyperaceae, Gramineae and *Betula* and high counts of *Pinus*. In addition, *Salix* and *Juniperus* are less abundant compared to the preceding

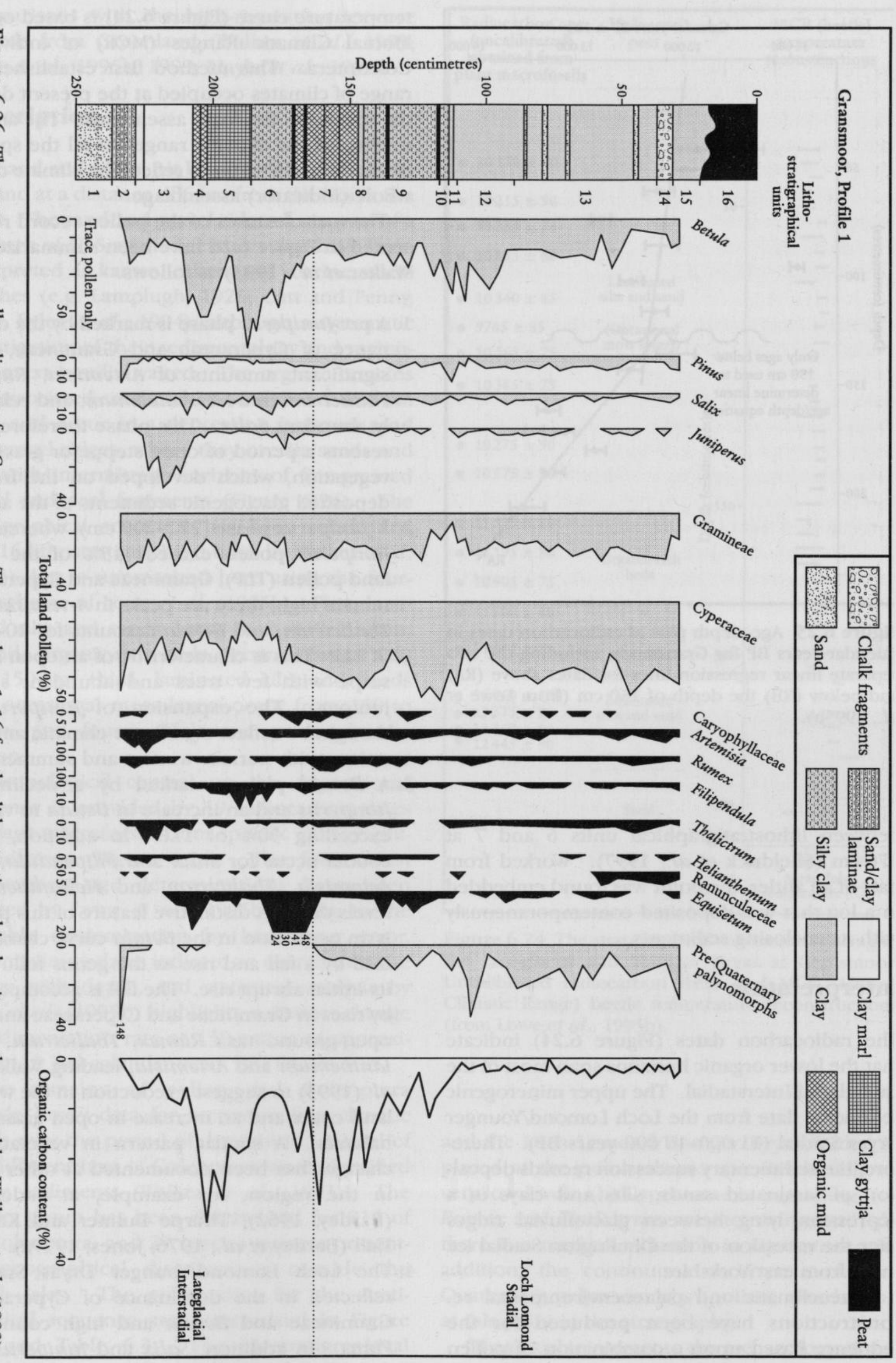



**Figure 6.26** The percentage pollen diagram for Gransmoor. The numbers 1–16 for the lithostratigraphical units are those used by Walker *et al.* (1993) wherein more detailed sedimentological and stratigraphical details are available. After Walker *et al.* (1993).

**Table 6.3** Stratigraphy at Gransmoor (after Walker *et al*., 1993).

| Lithological unit | Depth (cm) | Description |
|---|---|---|
| 16 | 0–17/23 | Fibrous peat; boundary is sharp but irregular, suggesting a possible hiatus. Blocks of reworked Late-Glacial clay occur within the Holocene peats. |
| 15 | 17/23–37 | Clay with sand laminae; laterally and vertically variable. |
| 14 | 37–41 | Angular and rounded chalk fragments. |
| 13 | 41–88 | Grey plastic clay; clearly defined sand laminations at 49, 74, 82 and 89 cm. |
| 12 | 90–112 | Clay unit with abundant sandy laminations, varying from a few millimetres to 1 cm in thickness. Each lamination continuous and of uniform thickness. |
| 11 | 112–115 | Grey plastic clay with small (<1 cm) pellets of chalk; latter appear flattened in the horizontal plane. |
| 10 | 115–120 | Silt/clay; black 'felted' peat layer at 120 cm |
| 9 | 120–146 | Laminated silt/clay with intercalations of 'felted' peat/plant debris. |
| 8 | 146–147 | Plastic grey clay. |
| 7 | 147–172 | Organic mud, but with clearly defined mineral/organic laminations in upper levels. Organic component variable, but maximum organic carbon values (~30%) towards the base of the unit. Bands of compressed plant debris occur in these lower levels. |
| 6 | 172–187 | Grey/brown silt/clay. Slightly organic (10% or less) throughout, but clearly defined clay-rich sub-unit from 174–178 cm; fibrous root material abundant. |
| 5 | 187–203 | Clay gyttja; organic content exceeds 20%, with maximum values (33%) near base of unit. |
| 4 | 203–207 | Transitional unit with intercalations of organic mud and grey silt/clay. |
| 3 | 207–223 | Clay marl with intermittent small (<2 cm) pellets of chalk; slightly organic (<10%) throughout. |
| 2 | 223–235 | Sand and clay laminae (up to 1 cm in thickness); some fine rootlet casts in the upper part. |
| 1 | Below 235 | Sands and silts with intermittent horizons rich in gravel-sized particles of coal and occasional discrete lenses of slightly organic silt. |

interstadial (phases 1–3). The herbaceous taxa (Caryophyllaceae, *Artemisia*, *Rumex*, *Thalictrum*, *Helianthemum* and *Ranunculus*) are typical of poorly developed and/or disturbed soils in arctic and alpine environments, leading Walker *et al*. (1993) to suggest an open steppe or steppe–tundra environment with discontinuous permafrost.

The large variety of coleopteran species allows for detailed reconstructions of changing local environmental conditions in the infilling and subsiding hollow at Gransmoor during the Late-glacial Interstadial and Loch Lomond/Younger Dryas Stadial. Details are available in Walker *et al*. (1993) but only general trends are reviewed here. Because no pollen is available in the basal sands (below 240 cm) and the 'pre-polleniferous' sequence (225–240 cm), the Coleoptera provide invaluable palaeoenvironmental proxy data for these early Late-glacial Interstadial sediments. Specifically, the Carabidae (predatory or scavenging ground beetles) suggest a local environment dominated by sparse vegetation developing on sandy, sterile soils. This reconstruction appears to be supported by the occurrence of certain other species of Coleoptera.

With the exception of some parts of the basal sands and the sediments between 235 and 240 cm, aquatic Coleoptera are represented throughout the stratigraphical succession. However, the climatic deterioration that marks the beginning of the Loch Lomond Stadial is characterized by the complete absence of aquatic beetles between 160 and 170 cm. This suggests that little or no water occupied the hollow during that period of deposition, although the occurrence of abundant Helodidae indicates wet marshy conditions throughout the stadial. Moreover, the occurrence of dytiscids and hydrophilids in the organic, polleniferous part of the sequence record the existence of well-vegetated ponds throughout much of the interstadial and the succeeding stadial.

When viewed as a complete assemblage, the Coleoptera provide information on the changes in palaeotemperature that occurred during the accumulation of the sediments and organics in the Gransmoor depression. Walker *et al*. (1993) have identified 'climatically significant species' and grouped them as either cold-adapted or warmth-adapted species (Table 6.4), the former now living north of the boreal coniferous forest boundary and the latter now living south of the northern limit of oak forests. The reconstruction of the Mutual Climatic Range (MCR) for the stratigraphical succession is compared with the *Betula* and *Juniperus* pollen curves in Figure 6.27. The MCR highlights the abrupt thermal improvement at the beginning of the Late-glacial Interstadial when temperatures peaked at 18–23°C (summer) and –4–9°C (winter). These can be compared with present-day values of 16°C and 4°C respectively for east Yorkshire.

**Table 6.4** List of climatically significant Coleoptera species from the Gransmoor stratigraphy (from Walker *et al.*, 1993).

| Cold-adapted species | Warmth-adapted species |
|---|---|
| *Nebria nivalis* | **Bembidion grisvardi* |
| **Diacheila arctica* | *Bembidion humerale* |
| **Diacheila polita* | *Bembidion quadripustulatus* |
| *Elaphrus lapponicus* | *Bembidion octomaculatum* |
| **Bembidion fellmanni* | *Pterostichus macer* |
| **Bembidion mckinleyi* | **Cymindis angularis* |
| **Agonum consimile* | *Ochthebius pedicularis* |
| *Amara alpina* | **Entomoscelis adonidis* |
| **Pycnoglypta lurida* | |
| **Olophrum boreale* | |
| **Acidota guadrata* | |
| **Boreaphilus henningianus* | |
| **Boreaphilus nordenskioeldi* | |
| *Oreodytes alpinus* | |
| **Colymbetes dolabratus* | |
| *Dysticus lapponicus* | |
| *Gyrinus opacus* | |
| **Helophorus sibiricus* | |
| **Helophorus glacialis* | |
| **Helophorus obscurellus* | |
| **Simplocaria metallica* | |
| **Hippodamia arctica* | |

The Late-glacial Interstadial–Loch Lomond Stadial boundary is marked by a 6°C decline in mean July temperatures and a more pronounced drop in winter temperatures, falling as low as –20°C during the stadial.

Comparisons between the palynological and coleopteran data sets have led to the identification of discordant biostratigraphical signals during the early part of the Late-glacial Interstadial (Walker *et al.*, 1993). Specifically, the biological–climatic disequilibrium previously recognized for the start of the Late-glacial Interstadial (Coope *et al.*, 1971) is emphasized in the Gransmoor data. It is manifest, first, in the occurrence of the thermal maximum, as indicated by the Coleoptera and dated at 13 000 years BP (Coope and Brophy, 1972; Coope, 1977), which occurs prior to the accumulation of polleniferous sediments and, second, by a subsequent fall in July temperatures by as much as 4°C before the first shrubs had firmly established themselves in the area. Moreover, the *Juniperus* curve shows that juniper began expanding in the area only after the climatic optimum of the interstadial had passed, a similar trend to other sites, where the episode of *Juniperus* development is dated at 12 500–12 400 years BP (Walker and Harkness, 1990) or later (Cwynar and Watts, 1989). Because juniper expansion often is regarded as the earliest botanical response to climatic amelioration, it can be used to demonstrate that a several hundred year time-lag in botanical response to climate change occurred at the beginning of the Late-glacial Interstadial.

The high quality of the palynological and coleopteran data at Gransmoor have allowed Lowe *et al.* (1999) to propose an event stratigraphy for the last glacial–interglacial transition based partly upon the site. An event stratigraphy involves the identification and correlation of short-lived geological events that have left some trace in the rock or sediment record. The palaeoecological record at Gransmoor has been compared with the palaeoclimatic data recovered from the Greenland ice cores by Lowe *et al.* (1995b) for GISP 2 (Figure 6.28) and Mayle *et al.* (1999) for GRIP. Comparisons with the snow accumulation record in the GISP 2 core provide a very close match, six points of the palaeotemperature curve being highlighted by Lowe *et al.* (1995b; A–F on Figure 6.28). These points demarcate: (A) a thermal maximum (at 14 700 calendar years BP) occurring within 100 years of the period of maximum snow accumulation in Greenland; (A–B) a rapid decline in both temperatures at Gransmoor (4–5°C) and snow accumulation in Greenland from 14 700 to 14 000 calendar years BP; (C) a slight rise in Gransmoor temperature and snow accumulation in Greenland at 13 700–13 600 calendar years BP; (D, E) a fall and subsequent recovery, respectively, of Gransmoor temperatures towards the end of the Late-glacial Interstadial (from 13 500 to 13 000 calendar years BP), which was paralleled by a decrease and subsequent increase in snow accumulation in Greenland; (F) a rapid decline in Gransmoor temperature values (5°C) and Greenland snow accumulation at the beginning of the Loch Lomond/Younger Dryas Stadial at 12 900 calendar years BP. The apparent synchroneity between climatic changes over Greenland and Britain supports suggestions, for example by Kapsner *et al.* (1995), that major warming trends in the North Atlantic region were characterized by northward storm-track displacement over Greenland.

The Gransmoor palaeotemperature record has been used along with similar records from three other key sites in Britain in a comparison with the $^{18}O$ signal from the GRIP ice core (e.g. Johnsen *et al.*, 1992) for the period covering the last glacial to Holocene transition (Mayle *et al.*,

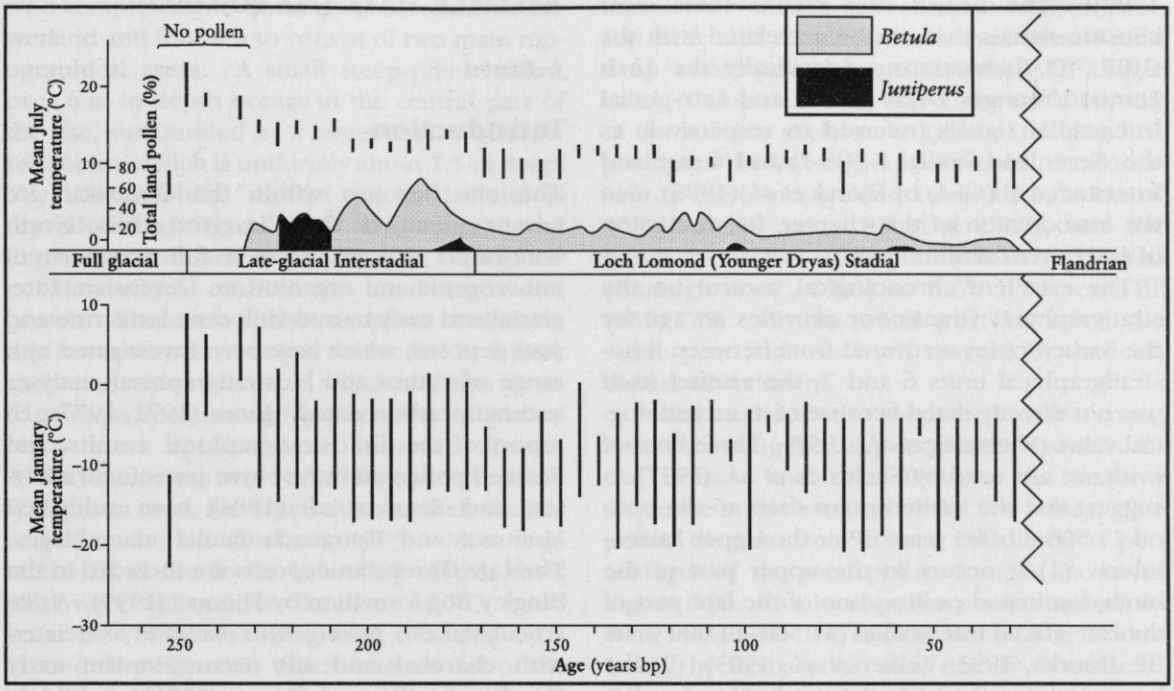



**Figure 6.27** The Mutual Climatic Range (MCR) reconstruction based upon Coleoptera from the Gransmoor stratigraphical succession. Pollen curves for *Betula* and *Juniperus* also are reproduced (from Walker *et al.*, 1993).

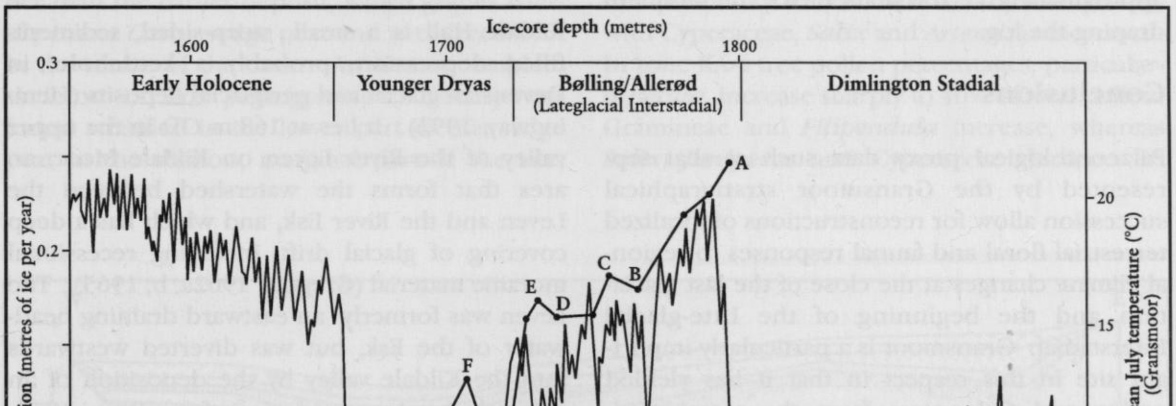



**Figure 6.28** Comparisons of ice accumulation data from the GISP 2 ice core and the palaeotemperature data from Gransmoor (from Lowe *et al.*, 1995b).

1999). The British sites record some clear climatic signals that can be correlated with the GRIP $^{18}O$ fluctuations. Specifically the Loch Lomond/Younger Dryas Stadial and Late-glacial Interstadial signals, referred to respectively as the 'Greenland Stadial 1' (GS-1) and 'Greenland Interstadial 1' (GI-1) by Bjorck *et al.* (1998). See the Introduction of this chapter, for discussion of GRIP event terminology.

The excellent chronological control on the stratigraphy at Gransmoor provides an age for the barbed point recovered from between lithostratigraphical units 6 and 7; the artefact itself was not directly dated because of its archaeological value (Sheldrick *et al.*, 1997). Three lines of evidence are used by Sheldrick *et al.* (1997) to suggest that the barbed point dates to the period 11 500–11 000 years BP or the Upper Palaeolithic: (1) it occurs in the upper part of the birch-dominated pollen zone of the late part of the Late-glacial Interstadial (11 500–11 000 years BP; Bourke, 1993; Walker *et al.*, 1993); (2) the log containing the barbed point has been radiocarbon dated to 11 475 ± 50 years BP (SRR 4920; Lowe *et al.*, 1995b), providing a maximum age for the artefact; (3) the age–depth plot on the 15 radiocarbon dates on terrestrial plant remains from the stratigraphical succession (Lowe *et al.*, 1995b) provide an age estimate of approximately 11 100 years BP for the sediment draping the log.

## Conclusions

Palaeoecological proxy data such as that represented by the Gransmoor stratigraphical succession allow for reconstructions of localized terrestrial floral and faunal responses to regional climate changes at the close of the last glaciation and the beginning of the Late-glacial Interstadial. Gransmoor is a particularly important site in this respect in that it has yielded pollen and Coleoptera from the same stratigraphical context. This allows for a close comparison of the two proxy data sources, focusing particularly on the discordant biostratigraphical signals associated with the use of multiple proxy data sources. Together with other critical palaeoenvironmental records at sites such as Glanllynnau and Llanilid in Wales, and Borrobol and Whitrig Bog in Scotland, the Gransmoor biostratigraphy is a significant component in the recently reconstructed last glacial to Holocene event stratigraphy for the British Isles.

# KILDALE HALL (NZ 609 097)

***J. Innes***

## Introduction

This site lies just within the Devensian ice advance limit in the Cleveland Hills, North Yorkshire. It preserves a full sequence of minerogenic and organic Late Devensian, Late-glacial and early to mid-Holocene lacustrine and peat deposits, which have been investigated by a range of litho- and biostratigraphical analyses and radiocarbon dated. Jones (1971, 1977a, b) reported the lithostratigraphical results and detailed pollen and bryophyte macrofossil analyses, and Keen *et al.* (1984) have published Mollusca and Ostracoda faunal assemblages. The Late Devensian deposits are included in the Bingley Bog formation by Thomas (1999). A disarticulated *Bos primigenius* skeleton associated with charcoal and silt occurs in the early Flandrian sediments (Jones, 1976b). Kildale Hall has been discussed in reviews of vegetation history and early Holocene human environmental impact (Innes and Simmons, 1988; Simmons *et al.*, 1993; Innes, 1999).

## Description

Kildale Hall is a small, steep-sided, sediment-filled depression, probably a kettlehole, in Devensian glacial and periglacial deposits (Hemingway, 1993). It lies at 168 m OD in the upper valley of the River Leven on Kildale Moor, an area that forms the watershed between the Leven and the River Esk, and which has a deep covering of glacial drift, including recessional moraine material (Gregory, 1962a, b, 1965). The Leven was formerly an eastward draining headwater of the Esk, but was diverted westwards into the Kildale valley by the deposition of an outwash barrier on deglaciation (Best, 1956; Gregory, 1962a, b, 1965; Jones, 1999). Dead-ice wastage was a feature of deglaciation in the western Cleveland Hills, with in-situ melting of ice blocks and kettlehole formation. The Cleveland Hills were glaciated in the Devensian but the higher main watershed and the southern valleys of the North York Moors were not (Wilson, 1948; Jones, 1999), and so the site lies close to the maximum Devensian ice margin.

The major investigation at Kildale Hall was conducted by Jones (1971, 1976b, 1977a, b), who completed two transects of cores across the wetland and found it to consist of two main topographical areas. A small steep-sided hollow over 6 m in depth occurs in the central part of the site, surrounded by a larger area of shallow sediments, which is uniformly about 1.5 m deep. The stratigraphical section across the site is reproduced as Figure 6.29. The deep central depression contains a silty clay with subangular pebbles, which overlies the basal sands and gravels. There are almost no plant remains in this lower silty clay but it is succeeded by a thick shell marl, up to 3 m deep, which contains abundant moss remains and is very silty in places. Many moss species were recorded (Jones, 1977a) including two, *Paludella squarrosa* and *Sphagnum teres*, for which this was their earliest record in the UK (Dickson, 1973). Moss fragments from the lower boundary of this thick marl were radiocarbon dated to 16 713 ± 340 years BP (SRR-145). The marl is sealed by a second silty clay unit, which extends out of the central deep hollow and forms the basal unit above the glacial sands and gravels in the shallow wetland area of the site. It is very poor in plant macrofossils, with a few ericoid fragments and some *Carex* stems, probably intrusive from above. Across the whole site this upper silty clay is succeeded by a second shell marl, much thinner than the earlier deposit, which grades laterally into a *Carex* sedge peat and is covered by it in the centre of the basin. An almost complete skeleton of *Bos primigenius* (aurochs) was found stratified in the lower part of this sedge peat in the shallow, marginal part of the site, associated with silt and charcoal. Historically, red deer and reindeer bones had been found deep within peat deposits elsewhere in Kildale valley (Cameron, 1878) and it was the discovery of the aurochs skeleton at Kildale Hall that first prompted the investigation of the site. Peat adjacent to the bones yielded a radiocarbon date of 10 350 ± 200 years BP (Gak-2707), but the bones themselves subsequently have been dated (Burleigh *et al.*, 1983) to 8270 ± 80 years BP (BM-1725). A thin wood peat lay upon the sedge peat and was itself covered by surficial, disturbed sediment.

Pollen analyses were conducted on two cores, core A covering the deep sequence from the central hollow and core B from the shallow sequence at the location of the *Bos* bones. The pollen diagram from core A, calculated as percentages of the total pollen sum, is shown as Figure 6.30. Five local pollen assemblage zones were identified. The earliest, zone KA-1, is dominated by herbaceous taxa, mainly Gramineae, Cyperaceae, *Artemisia*, *Rumex acetosa* and *Epilobium*. Zone KA-2 is characterized by increased tree pollen, reaching 25% of total pollen, with Cyperaceae most reduced and *Juniperus* rising late in the zone. The radiocarbon date of 16 713 ± 340 years BP spanned the lower boundary of this zone. In zone KA-3 tree pollen fall to less than 10% of total pollen, with Cyperaceae, *Salix* and *Artemisia* increased. In zone KA-4 tree pollen percentages, particularly *Betula*, increase sharply to 40% of total pollen. Gramineae and *Filipendula* increase, whereas *Salix*, *Juniperus* and Cyperaceae decline. In

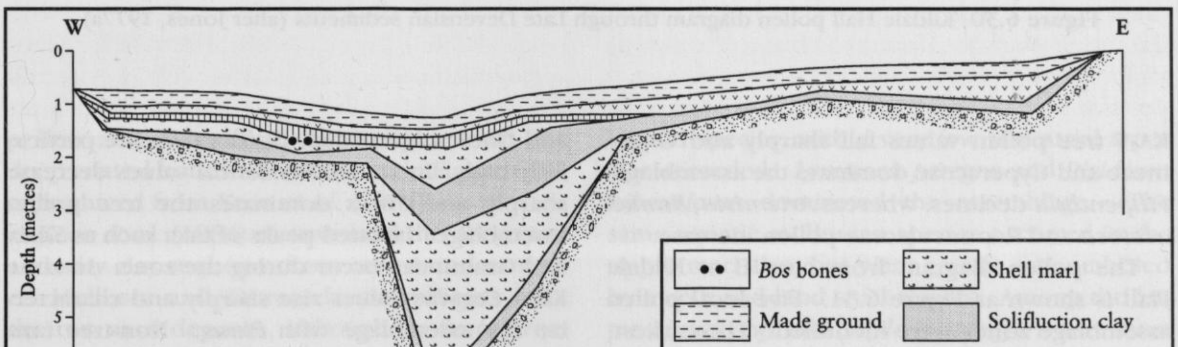



**Figure 6.29** Kildale Hall stratigraphy (after Jones, 1977a).

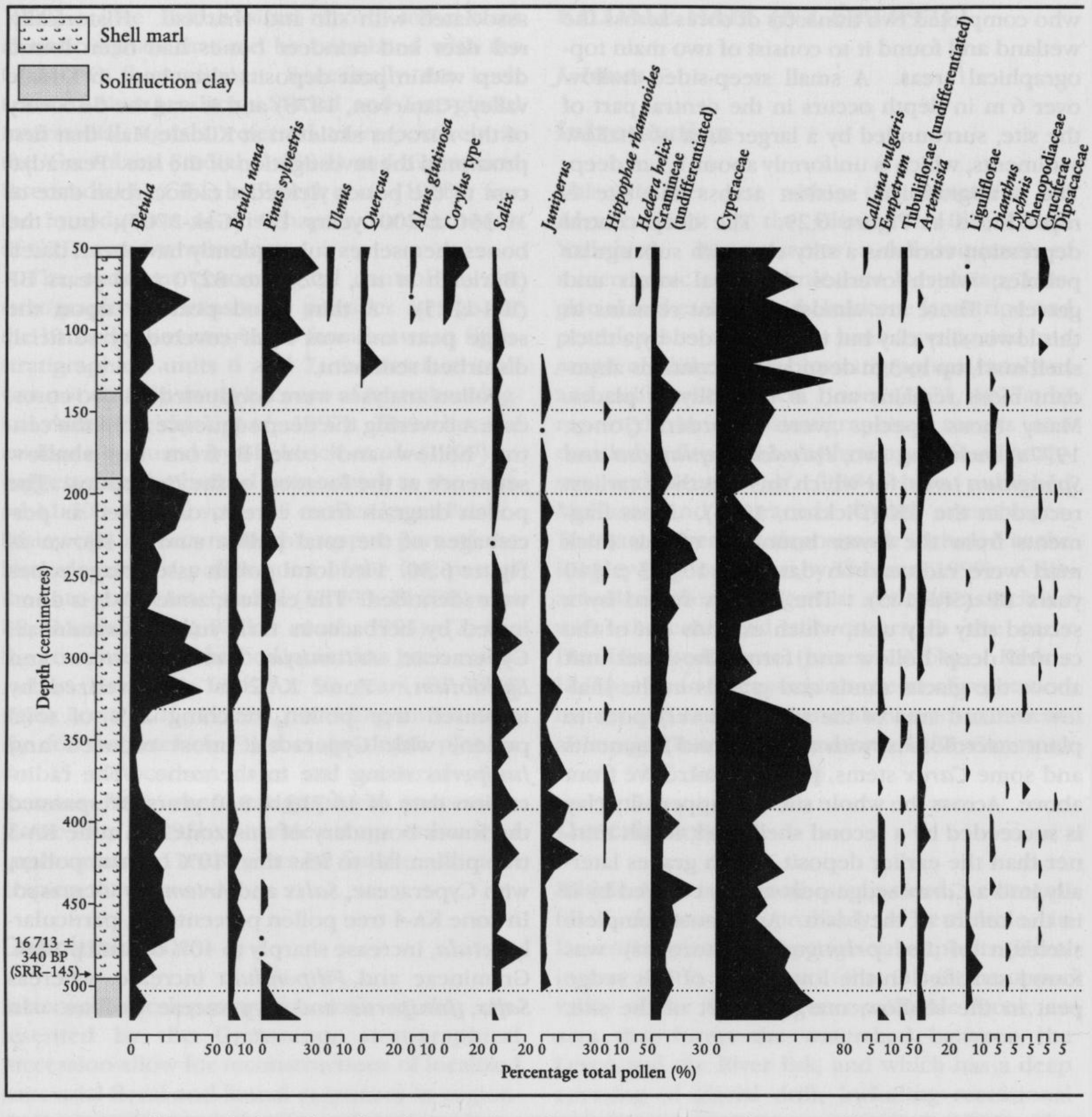
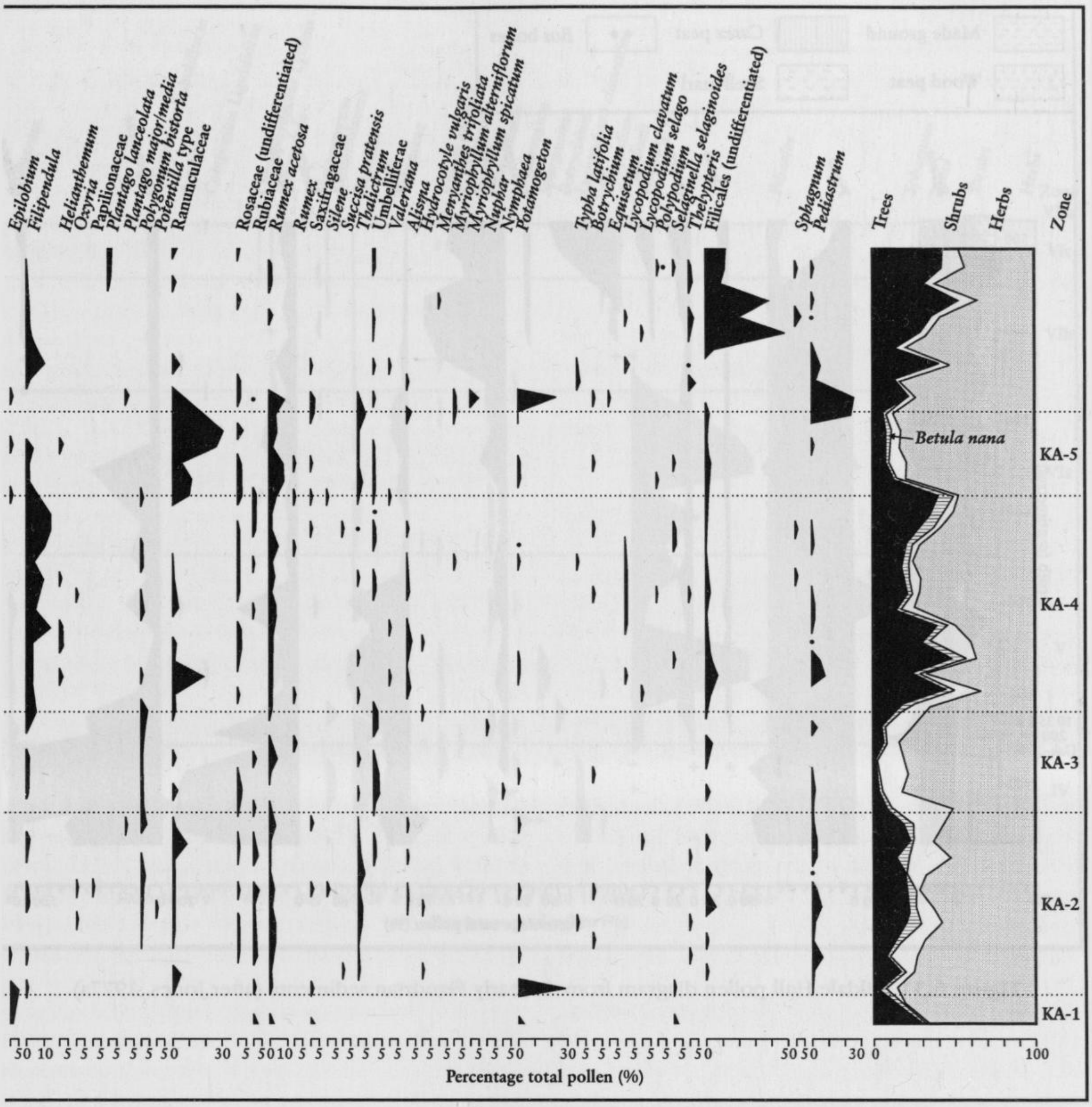



**Figure 6.30** Kildale Hall pollen diagram through Late Devensian sediments (after Jones, 1977a).

KA-5 tree pollen values fall sharply and Gramineae and Cyperaceae dominate the assemblage. *Filipendula* declines, whereas *Artemisia*, *Rumex acetosa* and Ranunculaceae pollen increase.

The pollen diagram from core B at Kildale Hall is shown as Figure 6.31. Five local pollen assemblage zones were identified. The earliest, zone KB-1, contains low tree pollen percentages, mainly of *Betula* and *Pinus*. *Salix*, *Empetrum*, Gramineae and Cyperaceae are the most abundant taxa. Zone KB-2 in contrast is dominated by *Betula* and *Pinus* pollen, with *Ulmus* and *Quercus* recorded. *Salix* values are particularly high. In zone KB-3 *Betula* values decrease sharply and *Pinus* dominates the tree pollen assemblage. Isolated peaks of taxa such as *Salix* and Gramineae occur during the zone. In zone KB-4 *Corylus* values rise sharply and characterize the assemblage with *Pinus*. Non-tree taxa percentages decline sharply to very low values. In zone KB-5 *Alnus* pollen frequencies rise to over 50% of total tree pollen. *Corylus* falls sharply, whereas *Quercus*, *Ulmus* and *Tilia* gradually increase. The main elements of the litho-,




pollen and macrofossil stratigraphy from the deep central hollow are tabulated in Figure 6.32, reproduced from Keen *et al*. (1984).

Keen *et al*. (1984) re-examined the deep marl and silty clay sequence from core A and extracted Mollusca and Ostracoda assemblages from the two marl deposits, although no fauna were recovered from the two silty clay units and Ostracoda were less common in the upper marl. They identified two molluscan suites. Although Mollusca were abundant in the lower marl they represented only a few aquatic bivalve and gastropod species, with a very few terrestrial types also recorded. *Lymnaea peregra* and *Pisidium hibernicum* dominated the assemblage. The same aquatic molluscan species occurred in the upper marl also, but were greatly outnumbered by marsh and land snail counts. Among the latter, *Limax* spp. and *Oxyloma pfeifferi* were most abundant, whereas *Pisidium* spp. were no longer present. In the top of the upper marl and in the *Carex* peat, land snails are the only taxa recorded, with *Nesovitrea hammonis* and *Punctum pygmaeum* most common. Ostracoda were

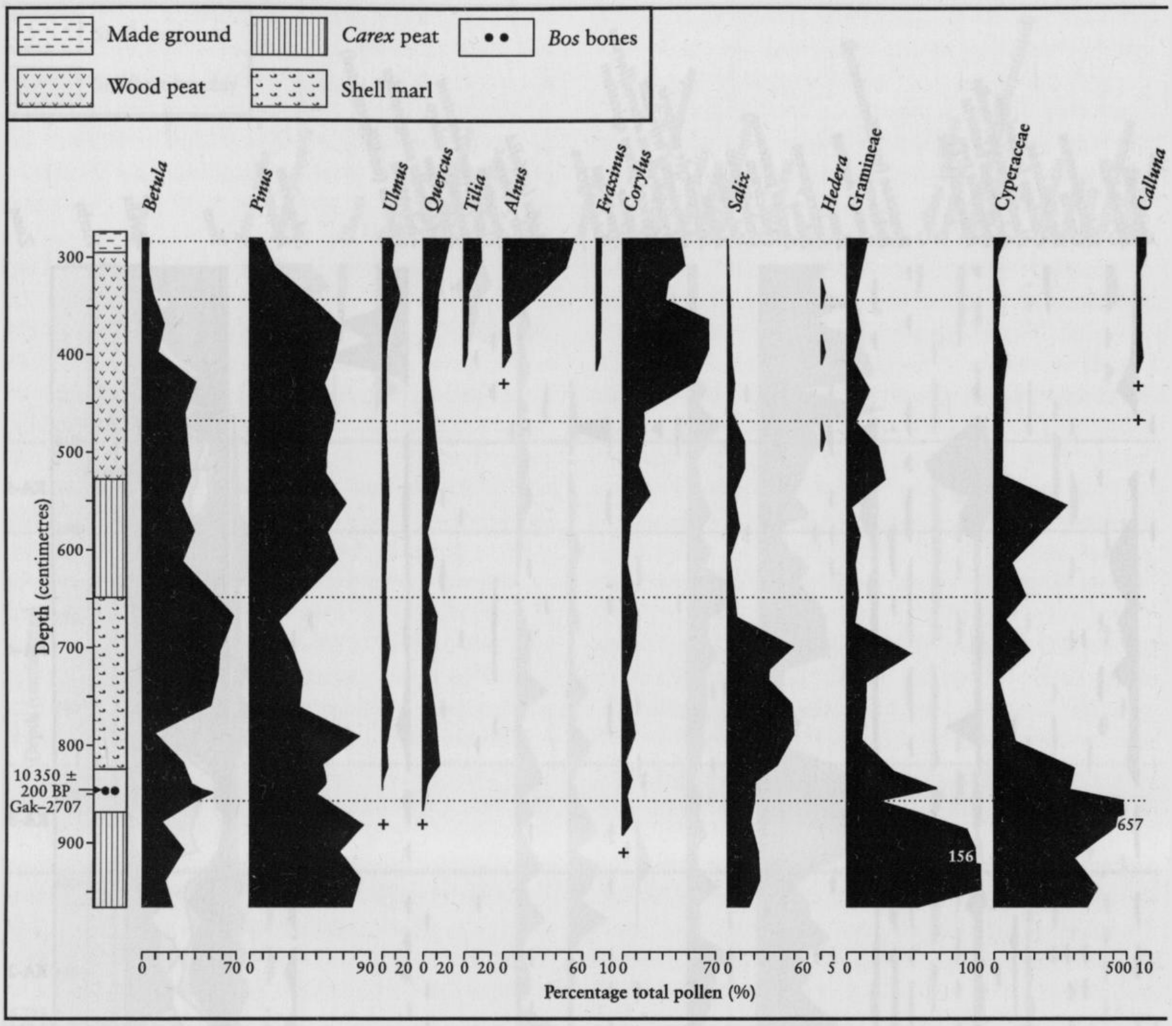
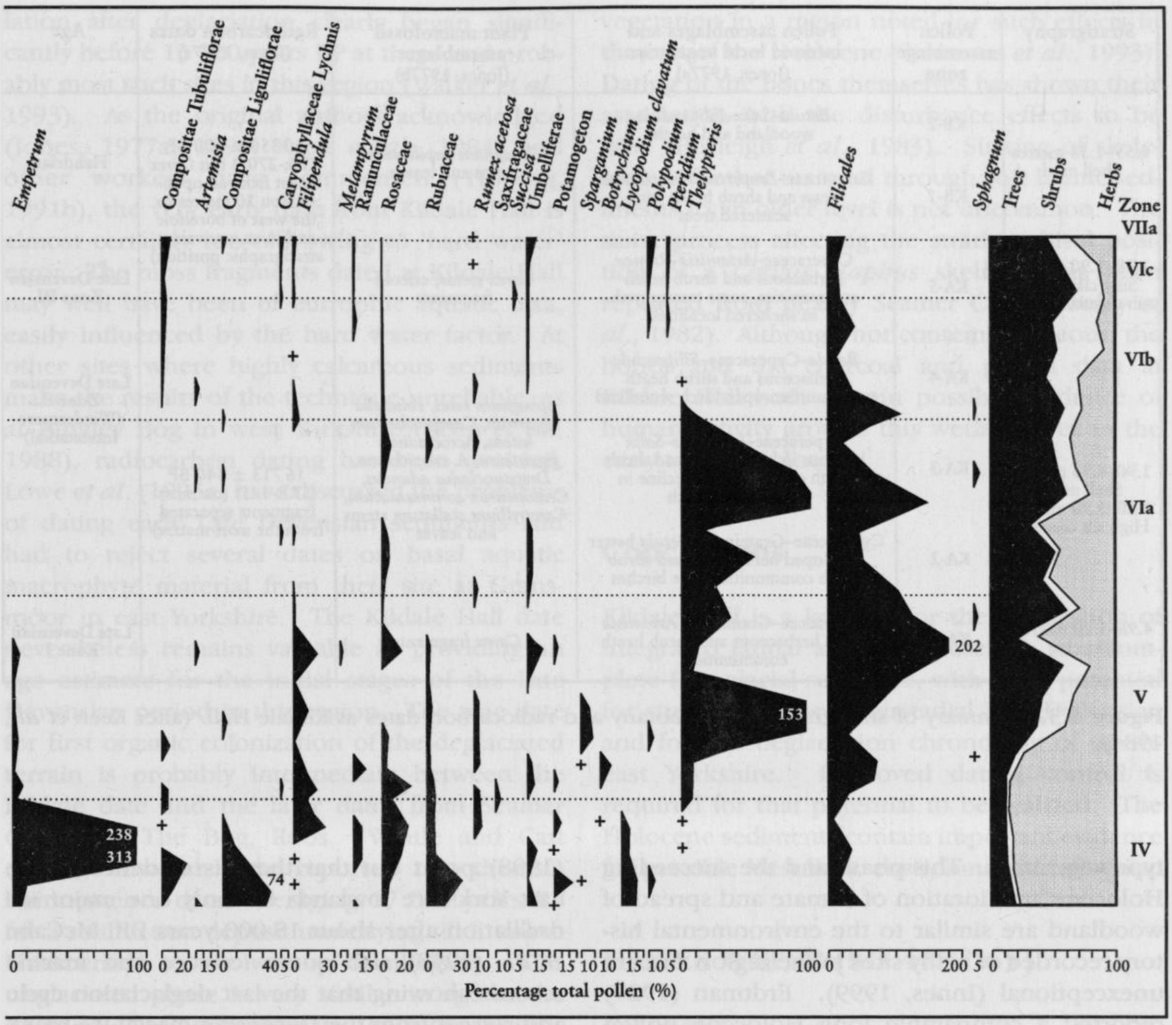



**Figure 6.31** Kildale Hall pollen diagram from the early Flandrian sediments (after Jones, 1977a).

present in the lower marl and the lower part of the upper marl, but did not persist in the stratigraphy as long as the molluscan fauna. Two taxa, *Candona candida* and *C. marchica*, overwhelmingly dominate the assemblage.

## Interpretation

The deep Late-glacial sequence from Kildale Hall is the most important feature of the site, with a multi-proxy data set allowing detailed environmental reconstruction from the early stage of the Late Devensian. The pollen stratigraphy in core A records an initial vegetation of pioneer, open habitat communities dominated by grasses and sedges and tundra-type herbs at the time of deposition of the lower silty clay, in an unproductive lacustrine environment, under still severe cold climate conditions. The change to a shelly marl, rich in moss remains, corresponds with pollen evidence for the spread of juniper and willow shrub-heath communities and some establishment of tree birches. A rich tall herb flora developed and the pollen and moss macrofossils recovered from the shell marl indicate a now highly productive aquatic system, with rapid silting of the lake basin, fringed by eutrophic reedswamp and rich fen vegetation (Jones, 1977a). The restricted ostracod fauna from the lower part of this shelly marl, mainly *Candona candida*, also points to rapid accumulation of




organic-rich, pond-floor litter (Keen *et al*., 1984). The molluscan fauna support this and the water body at Kildale Hall in this period appears to have been poorly oxygenated and rapidly shrinking (Keen *et al*., 1984). Both the pollen and ostracod data record a climatic reversion in phase KA-3 and the marl body contains more minerogenic sediment. Keen *et al*. (1984) interpret the rise of *Candona marchica* in this zone and the record of *Paralimnocythere* cf. *diebeli* as indicating low temperatures. The sharp fall in *Betula* pollen and other woody taxa and the rise of sedge tundra herbs agrees with this climatic interpretation. Zone KA-4 can be interpreted as a return to warm climate, stable soil, interstadial conditions with spread of *Betula* woodland and greatly reduced open ground. Zones KA2, 3 and 4 correlate with the Late-glacial Interstadial. The double *Betula* peak of zones 2 and 4, and the short vegetation reversion of zone 3, mirror the evidence reported from sites in north-east England, such as Seamer Carrs (Jones, 1976a), Tadcaster (Bartley, 1962), Thorpe Bulmer (Bartley *et al*., 1976) and The Bog, Roos (Beckett, 1981), which cover this period. Like these other sites, Kildale Hall preserves clear evidence of the division of the interstadial into two warm phases separated by a period of colder conditions. Zone KA-5 clearly correlates with the Loch Lomond Stadial severe cold stage, with soliflucted clay sediment in the basin and dominantly *Artemisia*, grass and sedge tundra-

| Stratigraphy | Pollen assemblage zone | Pollen assemblages and inferred local vegetation (Jones, 1977a) | Plant macrofossil assemblages (Jones, 1977b) | Radiocarbon dates (Jones, 1977a) | Age |
|---|---|---|---|---|---|
| 0.55–1.38 metres Shell marl | KB-2 | *Betula–Salix–Filipendula* woodland and heath | *Sphagnum papillosum* stems and leaves | 10 350 ± 200 BP (Gak-2707) (on *Carex* peat from an open section 20 metres to the west of borehole and in an equivalent stratigraphic position) | Flandrian |
| | KB-1 | Gramineae–*Empetrum–Filipendula* grass and shrub heath with scattered trees | | | |
| 1.38–1.90 metres Silty clay with sub-angular stones | KA-5 | Cyperaceae–*Artemisia–Rumex* herbaceous and shrub heath communities, some tree growth in sheltered localities | *Carex* stems; ericoid fragments | | Late Devensian Zone III |
| 1.90–4.96 metres Shell marl 3.20–3.90 metres High silt content | KA-4 | *Betula*–Cyperaceae–*Filipendula* herbaceous and shrub heath communities: open birch woodland | *Sphagnum teres, Paludella squarrosa, Camptothecium nitens, Acrocladium giganteum, A. cuspidatum, Drepanocladus aduncus, Cratoneuron communtatum, Campyllium stellatum* stems and leaves | 16 713 ± 340 BP (SRR-145) (on moss fragments separated from the marl matrix) | Late Devensian Zone II (Windermere Interstadial) |
| | KA-3 | Cyperaceae–*Juniperus–Salix* increase in herbaceous and shrub heath communities; decline in tree birch growth | | | |
| | KA-2 | Cyperaceae–Gramineae–*Betula* better developed herbaceous and shrub heath communities; tree birches | | | |
| 4.96–5.20 metres Silty clay | KA-1 | Cyperaceae–Gramineae–*Artemisia* sparse herbaceous and shrub heath communities | *Carex* fragments | | Late Devensian Zone I |

**Figure 6.32** Summary of stratigraphy, palaeobotany and radiocarbon dates at Kildale Hall. (after Keen *et al.*, 1984).

type vegetation. This phase and the succeeding Holocene amelioration of climate and spread of woodland are similar to the environmental history recorded at many sites in the region and are unexceptional (Innes, 1999). Erdtman (1927) reported a comparable long Holocene pollen record from a nearby site at Kildale Moss. The evidence from the few sites that combine pollen and faunal records, such as Bingley Bog (Keen *et al.*, 1988) and Skipsea Withow (Gilbertson, 1984b), is analogous to that from Kildale Hall.

The importance of the Kildale Hall early Late Devensian record is given added significance by the very early radiocarbon date of 16 713 ± 340 years BP for the onset of biogenic deposition after the withdrawal of the local ice cover. The relationship of this date to dates of little more than 18 000 years BP on moss fragments in sub-till silt (Penny *et al.*, 1969) from Dimlington in Holderness would suggest either that the maximum extension of ice cover in east Yorkshire was short lived, or that there were very local differences in the timing of ice advance and retreat in this region. The latter is most likely, as Eyles *et al.* (1994) have shown that deglaciation in the east Yorkshire area was locally complex, with ice-lobe surging and retreat, although McCabe *et al.* (1998) point out that there is evidence in the east Yorkshire lowlands for only one major ice oscillation after about 18 000 years BP. McCabe *et al.* (1998) also quote ice-core and marine records showing that the last deglaciation cycle and retreat from the Devensian glacial ice maximum began about 21 000 years BP, but was a complex process interrupted by major climate shifts that caused ice readvance events. Significantly they conclude that a major deglaciation phase at the southern extremity of the British ice sheet occurred at about 17 000 years BP. This could lend credence to the Kildale Hall date of *c.* 16 700 years BP for the start of organic deposition, which otherwise seems very early, despite its location adjacent to unglaciated, nunatak terrain in the North York Moors. Other dates on basal organic material in kettleholes elsewhere in north and east Yorkshire are in the order of little more than 13 000 years BP, for example, 13 045 ± 270 years BP at The Bog, Roos in Holderness (Beckett, 1981), and 13 042 ± 140 years BP at Seamer Carrs, in the Cleveland lowlands (Jones, 1976a). At both these sites, however, there is a considerable depth of polleniferous sediment, insufficiently organic to date, beneath the dated horizon. Sediment accumulation after deglaciation clearly began significantly before 13 000 years BP at these and probably most such sites in this region (Walker *et al.*, 1993). As the original authors acknowledged (Jones, 1977a, 1999; Keen *et al.*, 1984) and other workers have commented (Tipping, 1991b), the very early date from Kildale Hall is almost certainly too old owing to 'hard water' error. The moss fragments dated at Kildale Hall may well have been of eutrophic aquatic taxa, easily influenced by the hard water factor. At other sites where highly calcareous sediments make the results of the technique unreliable, as at Bingley Bog in west Yorkshire (Keen *et al.*, 1988), radiocarbon dating has been avoided. Lowe *et al.* (1995a) have discussed the problems of dating early Late Devensian sediments and had to reject several dates on basal aquatic macrophyte material from their site at Gransmoor in east Yorkshire. The Kildale Hall date nevertheless remains valuable in providing an age estimate for the initial stages of the Late Devensian period in this region. The true date for first organic colonization of the deglaciated terrain is probably intermediate between the Kildale date and the later dates from Seamer Carrs and The Bog, Roos. Wintle and Catt (1985), however, have reported two thermoluminescence dates averaging 17 000 years BP for a solifluction deposit underlying till in north Humberside to the west of Dimlington. This unglaciated phase is very similar in age to the Kildale Hall radiocarbon date. A further factor is that the interpretive value of all of the conventional radiocarbon dates mentioned above is greatly compromised by their very large standard deviations. Radiocarbon AMS dating of terrestrial plant macrofossils, as advocated by Lowe *et al.* (1995a), would greatly improve the diagnostic value of the early Late Devensian record from Kildale Hall.

The other date of 10 350 ± 200 years BP, on marly peat adjacent to the *Bos* bones, is almost certainly also made too old by hard water influence, although not to such a great degree because the peat contains an early Holocene pollen assemblage. *Betula* and ericaceous charcoal at this level in the mire is accompanied by silt inwash and fluctuations in the pollen spectra. The evidence suggests burning of *Empetrum* heath and birch scrub around a small swamp-fringed lake, encouraging several weed taxa and causing some erosion of catchment soils. It is an early example of fire disturbance of vegetation in a region noted for such effects in the rest of the Holocene (Simmons *et al.*, 1993). Dating of the bones themselves has shown their association with the disturbance effects to be false (Burleigh *et al.*, 1983). Sinking of skeletons and other material through soft limnic sediments to an older level is not uncommon. The same process affecting the stratigraphical position of a *Cervus elaphus* skeleton has been reported from nearby Seamer Carrs (Tooley *et al.*, 1982). Although not contemporaneous, the bones and the charcoal and pollen data at Kildale Hall both remain possible evidence of human activity around this wetland area in the early to mid-Holocene.

## Conclusions

Kildale Hall is a key site for the application of integrated faunal and floral research to a complete Late-glacial sequence, with great potential for study of the pre-interstadial Late Devensian and for the deglaciation chronology of north-east Yorkshire. Improved dating control is required for that potential to be realized. The Holocene sediments contain important evidence for possible Mesolithic environmental influence.

# HAWES WATER (SD 478 765)

***R. Jones***

## Introduction

Hawes Water is an important site because it is one of only four carbonate lakes in northern England and the only lake of natural origin remaining in Lancashire. The site contains a marginal sedimentary record up to 12 m in depth, dominated by authigenic carbonate precipitates, which extends from the end-Devensian through to the Late Holocene (15.0 to *c.* 5.0 ka). The comparative dearth of carbonate sedimentary archives in Britain, which are known to be particularly sensitive to both climatic and anthropogenic impacts on the landscape, makes the site particularly valuable. Hawes Water has been studied in great detail by Oldfield (1960a, b), who demonstrated its significance for the development of regional vegetation and studies of lake-level change during the Late-glacial and Holocene. The importance of the Late-glacial sediments has been highlighted by the recent work of Nash (1995), Nolan *et al.* (1999) and Jones (1999), identifying the existence of four sub-millennial climatic oscillations prior to the onset of the Younger Dryas. The existence of such events reflects a degree of climatic instability during the Late-glacial Stadial not recorded previously at sites in north-western England.

## Description

Hawes Water is a freshwater marl lake situated within the Arnside–Silverdale Area of Outstanding Natural Beauty adjacent to the Gait Barrows National Nature Reserve and 2 km south-west of Silverdale village. Along with Malham, Cunswick and Sunbiggin Tarns, it represents one of only four carbonate lakes in northern England. Hawes Water and its smaller sister basin Little Hawes Water, which lies to the north-east, rest approximately 8 m above sea level, within a catchment of approximately 1.7 km$^2$. The main lake is *c.* 400 m in length and 225 m wide, reaching a maximum depth of 12 m towards the centre. It occupies a shallow depression within the local Carboniferous Limestone. The depression is thought to be pre-Devensian in age, possibly a polje (Vincent, 1985). The limestone crops out to the north-east of the lake, forming an extensive limestone pavement. In places it is overlain by thin glacial diamicton, comprising a fine-grained, clast-rich blue-grey clay, deposited under ice during the end-Devensian.

The lake is fed by ephemeral springs, the majority of which flow into the north-east part of the basin. It also receives drainage from Little Hawes Water via a stream that runs directly into the lake. The main outflow from Hawes Water is situated at the southern end of the lake, where it drains by seepage and stream flow on to Hawes Water Moss (Figure 6.33). The principal characteristics of the lake are presented in Table 6.5.

A terrestrialized marl bench, currently supporting reed beds, fen carr and more importantly calcareous meadows, rings the modern lake. These marginal sediments were first studied by Oldfield (1960a, b), whose combined stratigraphical and palynological study revealed a record of uninterrupted lacustrine deposition from the end-Devensian through to the mid-Holocene (*c.* 15.0–5.0 ka). The marginal sediment stratigraphy is illustrated in Figure 6.34. Despite its close proximity to the coast, low altitude and close proximity to Silverdale Moss (where marine sediments can be found), there is no evidence in the sedimentary record, for a marine incursion into the catchment (F. Oldfield, pers. comm., 2001).

## Interpretation

### *The Late-glacial (c. 15–10.0 ka)*

The Late-glacial sequence is represented by a sequence of micritic carbonates deposited during the Late-glacial (Windermere) Interstadial, bounded above and below by clay-rich sediments corresponding to the end-Devensian and Younger Dryas cold periods (Figure 6.34). Identification of the upper clay unit as equivalent to the Younger Dryas has been confirmed recently by two $^{14}$C AMS dates that bracket the unit (10 980 ± 60 and 9 600 ± 100 years BP) (Jones, 1999; cf. Pennington, 1975a).

The Late-glacial sequence was first studied by Oldfield (1960a, b), who undertook a detailed stratigraphical and palynological study. The results of the pollen analysis subdivided the Late-glacial into Jessen's classic tripartite pollen zonation scheme (e.g. Jessen, 1938), which was at that time being used as a chronostratigraphical framework to define the period. This work, along with that from Windermere (Pennington, 1947), Skelsmerg Tarn (Walker, 1955b), Helton Tarn and Witherslack (Smith, 1958c), represent the first published Late-glacial profiles for northern Britain.

Oldfield's initial pollen study has been updated recently by a higher resolution study (Jones, 1999; core HW1/1 – see Figure 6.33), which identified a total of seven LPZs (HW1–7, Figure 6.35). The results chart the development of the catchment vegetation from tundra (end Devensian, HW1) through to juniper scrubland (Early Interstadial, HW2) and then closed *Betula* woodland (Late Interstadial, HW3–4). The Younger Dryas (HW5) is characterized initially by the development of open grassland communities, which colonized the area after the destruction of the interstadial birch woodland. As the climate continued to deteriorate the grasslands were replaced by tundra communities, thus reflecting the severity of the climate.

Pollen analysis at Hawes Water has been complemented by a high-resolution mineral magnet-

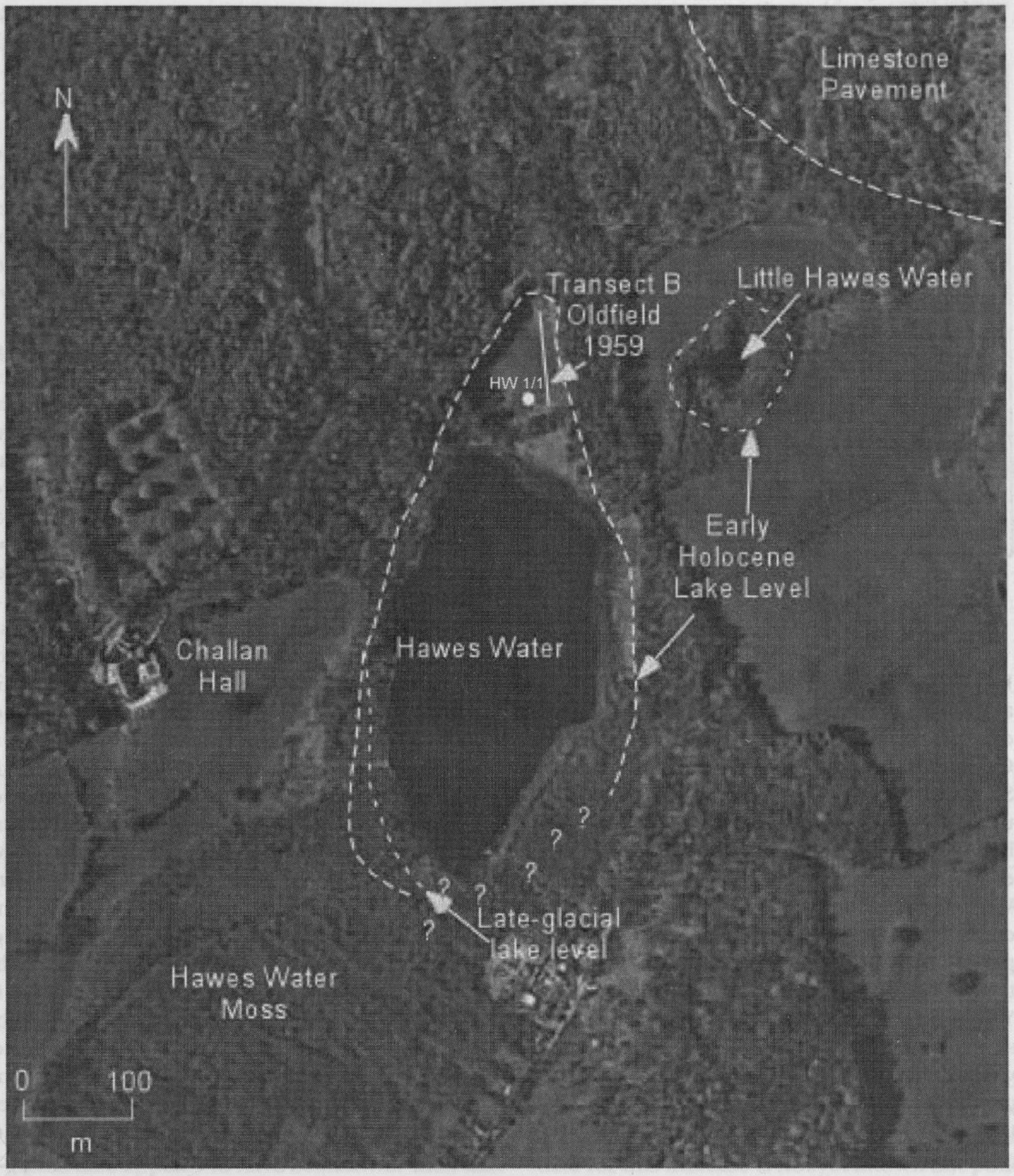



**Figure 6.33** Aerial photograph of Hawes Water and its catchment. The position of Transect B (Oldfield, 1960a, b) and core HW1/1 (Jones, 1999) are recorded. The provisional location of the Late-glacial and Holocene lake levels also are noted.

ic and geochemical study (Nolan *et al.*, 1999), the results of which revealed the existence of two short-term, low-amplitude climatic oscillations during the Windermere Interstadial. These events were thought to be contemporary with the Aegelsee and Gerzensee oscillations identified at sites across the Swiss Plateau (Lotter *et al.*, 1992) and at several sites in the UK (Mayle *et al.*, 1999). Subsequent oxygen isotopic analysis of the bulk carbonate sediments (Nash, 1995;

**Table 6.5** Limnological characteristics of Hawes Water

| Hawes Water | |
|---|---|
| Dimensions | ~400 × 225 metres |
| Elevation | 8 metres above sea level; distance to sea ~ 5 kilometres |
| Water depth | Marl shelf ~1.2 metres; maximum ~12 metres |
| Lake volume | ~390 000 cubic metres |
| Water supply | Direct precipitation (~1350 mm/a), spring, groundwater |
| Water temperature | Surface water 5–18°C; deep water –5–8°C |
| Catchment area | 1.77 $km^2$ |
| Residence time | 2–6 months? |
| Carbonate precipitation | Seasonal<br>Biogenic (skeletal): gastropods, bivalves, ostracods, Chara<br>Biologically mediated (plants/microbial) |
| Sediment record | 'Marl': (bio)micrite, clay, peat, gyttja |

Jones, 1999) has extended the number of low-amplitude events to four, producing a record directly comparable to the high resolution Chironomid record from Whitrig Bog, Southern Scotland (Brooks and Birks, 2000). The impact upon the lake-catchment system appears to have been muted in most cases, with none of the events producing a discernable inwash horizon. Within the pollen record (Figure 6.35) each event is characterized by a decline in arboreal pollen (AP) and an expansion of herbaceous taxa. The fourth event, prior to the onset of the Younger Dryas (LPZ HW4) appears to have been severe enough to prompt the partial loss of woodland cover, leading to the re-expansion of grassland communities previously dominant in the catchment during the latter stages of the end-Devensian.

### *Early to mid-Holocene (10.0–c. 5.0 ka)*

The Early to mid-Holocene sedimentary record is dominated by the formation of the carbonate bench system. The bench system reaches a maximum thickness of 10–12 m at the southern end of the lake. It is composed of almost pure carbonate (95% $CaCO_3$) with very little detrital or organic components, reflecting the predominantly autochthonous sedimentary regime in the lake during this time. The bench is capped by a sequence of marginal fen peats, with the contact between the two units effectively marking the previous margins of the lake.

Carbonate precipitation rates during this period are estimated to have been as high as 2 mm $year^{-1}$. The sediments are rich in shell remains, particularly molluscs and gastropods, primarily *Lymnaea peregra* and *Pisidium nitidium*. Sparks (1962), who studied the stratigraphical distribution of this shell material, found them to be 'the most monotonous' he had ever encountered.

A detailed pollen diagram for the Holocene marginal sediments was published by Oldfield (1960a) (Figure 6.36), which recorded the classic Holocene sequence from tundra to mixed deciduous woodland. Using the pollen data from Hawes Water and several other sites around southern Lakeland, Oldfield (1965) highlighted for the first time the potential problems associated with the use of local pollen zones (LPZs). He demonstrated that, on a countrywide scale, LPZs could not be used as a chronostratigraphical tool for dating the Holocene, as advocated by Godwin (1956). Concentrating on the boundary between zones VI and VIIa, the 'Boreal–Atlantic transition', (pinpointed by the spread of *Pinus*), Oldfield highlighted a marked discrepancy in the timing of events between southern and northern Britain. In south and eastern England, the spread of *Pinus* (usually at the expense of *Betula*), occurred before the decrease in *Ulmus* and *Corylus* frequencies. However, at Hawes Water and other sites around north-west England, the *Pinus* rise was seen to occur after the rise in *Ulmus* and *Corylus* and was never entirely at the expense of *Betula*. This research was of considerable importance at the time of publication because palaeoenvironmental studies relied heavily on LPZs for inter-site correlation, radiocarbon dating being a technique still very much in its infancy.

The existence of marginal sediments provides clear evidence of significant changes in the lake level at Hawes Water during the Holocene. Pollen and stratigraphical data from the northern perimeter (Transect B, Oldfield, 1960a, b) indicate that the lake-level was at its highest during the early Holocene, with the lake covering the field to the north of the present-day lake, and extending south towards Hawes Water Moss (Figure 6.33). More recent stratigraphical investigations of the southern end of the basin

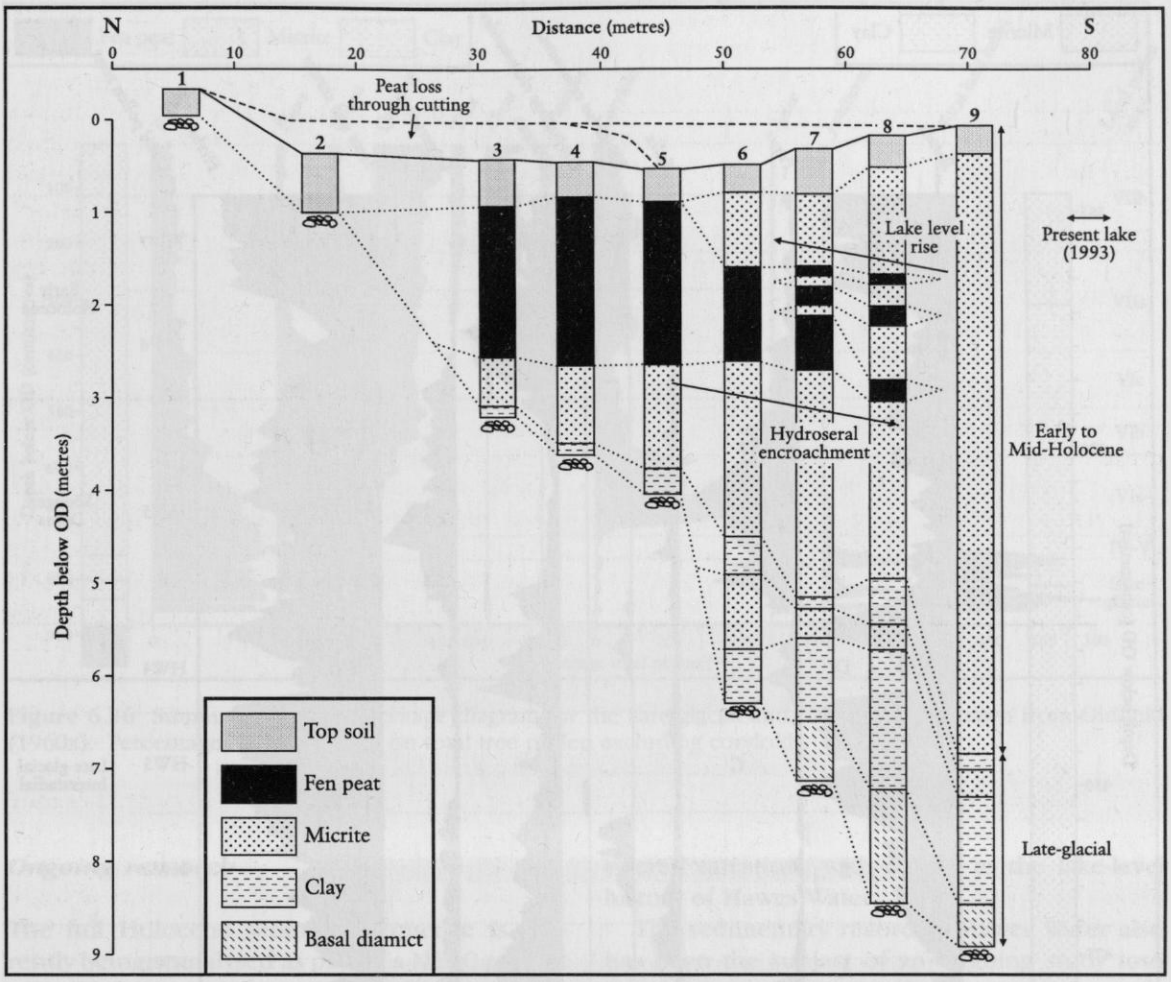



**Figure 6.34** Marginal lacustrine sediment stratigraphy underlying the area to the north of the present lake (after Oldfield, 1960a).

confirm Oldfield's initial conclusions and have identified the former Late-glacial lake level (Jones and Gedye, unpublished data). The Late-glacial lake appears to have been lower then that recorded during the early Holocene (Figure 6.33) but higher than the present lake.

It is clear from the stratigraphical evidence that the Early Holocene was characterized by the gradual infilling of the basin by the progradation of the carbonate bench system and hydroseral encroachment of the marginal peat deposits. Pollen analysis of the peat–marl contact shows a distinct younging trend lakewards for both the peat and carbonate sediments (Figure 6.37). This declining trend was reversed during the mid-Holocene by a substantial rise in lake-level, as reflected in the stratigraphical record from Transect B by the deposition of carbonate sediments atop the fen peat (Figure 6.34). A precise date for this increase is not yet known but pollen analysis places it at some point during or just after LPZ VIc, during the alder rise.

The next major phase of lake-level change appears to have been the decline to that of the present day. The absence of marginal late-Holocene sediments suggests that no further increases in lake-level occurred. The timing of the fall is unknown at present but must post-date the date for the top of the marginal sequence, which Oldfield (1960a, b) dated to LPZ VIIb, sometime after the elm decline, *c.* 5500–5700 years BP (Skog and Regnel, 1995).

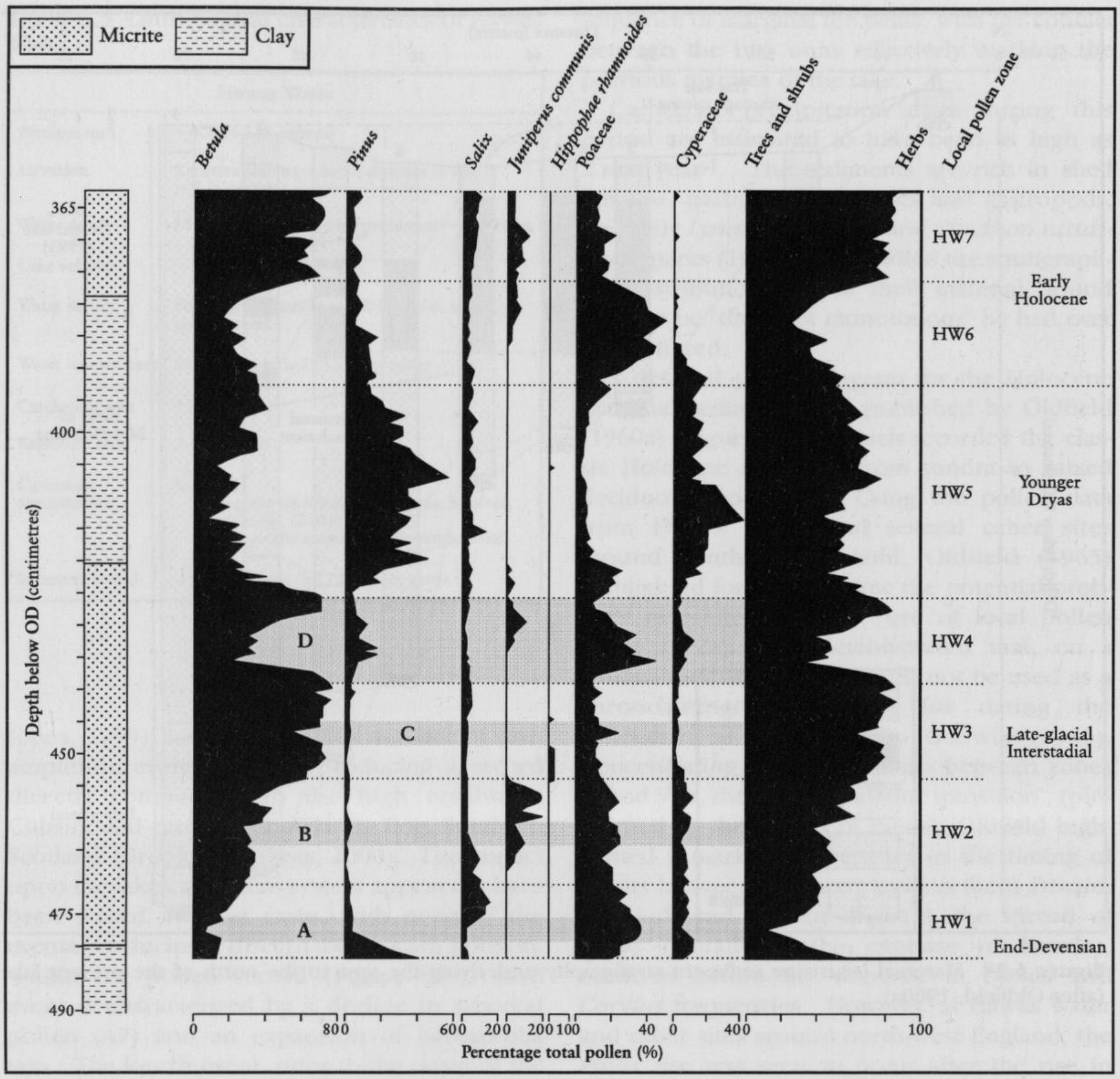



**Figure 6.35** Simplified Late-glacial pollen diagram from core HW1/1 (Jones, 1999). The four short-term oscillations (A–D) identified within the oxygen isotope record are highlighted in grey.

The situation is complicated owing to the loss of marginal sediment through peat cutting, producing an 'artificial' date in places for the top of the marginal sediments.

The mid-Holocene fall in lake-level prompted a dramatic shift in sedimentation patterns. Analysis of a 3 m Mackereth core from the modern lake reveals sedimentation during the latter half of the Holocene being dominated by organic-rich gyttja (J.D. Marshall, Liverpool University, pers. comm. 2001) with a carbonate content of <10%. There is no evidence for renewed bench formation during the latter stages of the Holocene. The marked decrease in carbonate precipitation in the lake is likely to be linked to the drop in lake level, increasing the extent of the marginal peat deposits and the rate of organic inwash to the lake. Increased levels of organic decomposition would lead to increased $CO_2$ production, which effectively would 'acidify' the lake, causing the carbonate precipitates to re-dissolve back into the water column. A similar situation has been observed within the modern lake system.

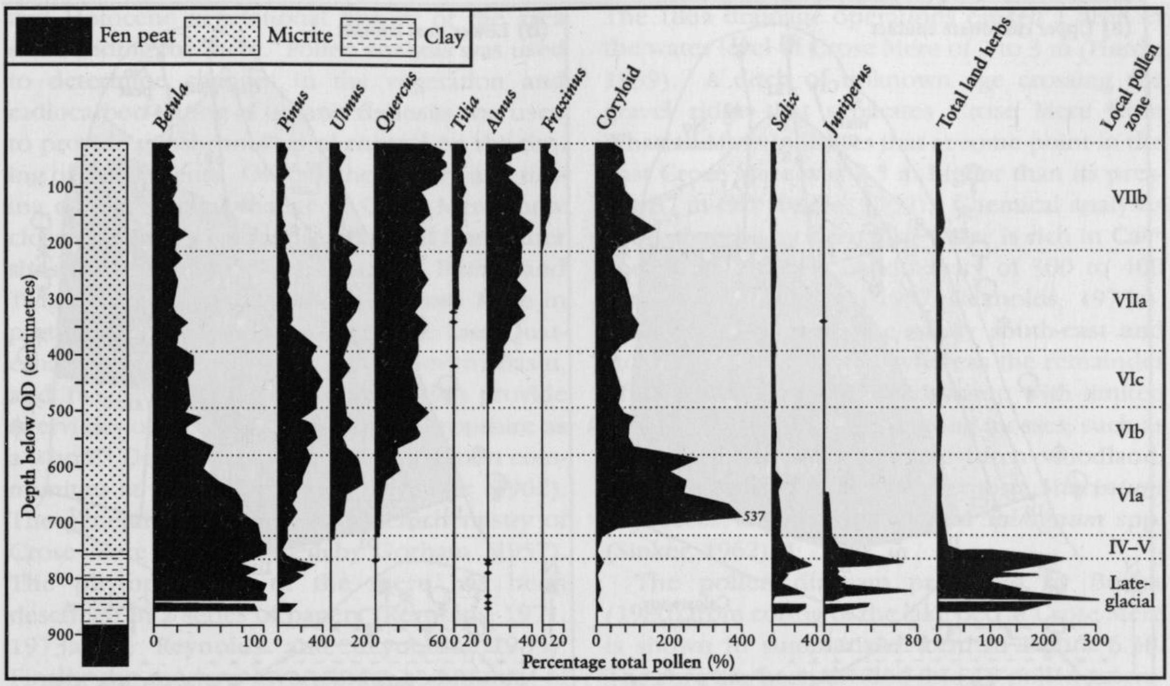



**Figure 6.36** Summary pollen percentage diagram for the Late-glacial and Holocene. Redrawn from Oldfield (1960a). Percentages are calculated on total tree pollen excluding coryloids.

## *Ongoing research*

The full Holocene sediment sequence is currently being reanalysed as part of a NERC project 'A high-resolution lacustrine record of Holocene climate change' (J.D. Marshall (co-ordinator), Liverpool University, Earth Sciences). The project aims to produce the first high-resolution Holocene stable isotope record ($\delta^{18}O$, $\delta^{13}C$, 1 cm contiguous sampling resolution) for northern Britain, one of only a handful in Europe. The isotopic work on micritic carbonate and ostrocods is supported by a parallel study incorporating pollen (F. Mann, Edge Hill College), and Chironomids (B. Lang, Edge Hill College). A daily monitoring programme of the lake system also has been undertaken since 1999, collecting data on 4-hourly temperature records, and monthly measurements of the isotopic composition of lake waters and precipitates. Initial results have identified a number of climatic oscillations, particularly during the early Holocene, that are evident within both the isotopic and Chironomid records. Two such oscillations have been linked tentatively to the Pre-Boreal and 8.2 ka events. It is hoped that these investigations will shed light on some of the unanswered questions with regard to the lake-level history of Hawes Water.

The sedimentary record at Hawes Water also has been the subject of an ongoing study into the radiometric dating of carbonate lake sediments. This followed pilot work on the uranium-series signatures of lake waters and sediments to investigate the controls on radionuclide ($^{238}U$, $^{234}U$, $^{232}Th$ and $^{230}Th$) flux to Hawes Water and their transfer to the sediment column. The lake was seen to be buffered from the seasonality observed in radionuclide input as a result of changes in the prevailing hydrological regime (Blake *et al.*, 1998). In addition, a seasonal change in the mechanism of lake sedimentation was observed, from allochthonous organic input during winter to autochthonous carbonate formation during summer. Subsequently, Holocene sediments from the margin of Hawes Water and a shallow-water island populated by charophytes were studied in order to develop a novel technique for dating lake sediments using $^{226}Ra$. The results suggest that the application of this dating method is valid for up to 5000 years under the initial conditions of radioactive disequilibrium displayed at the present sediment–water interface. The sedimenta-

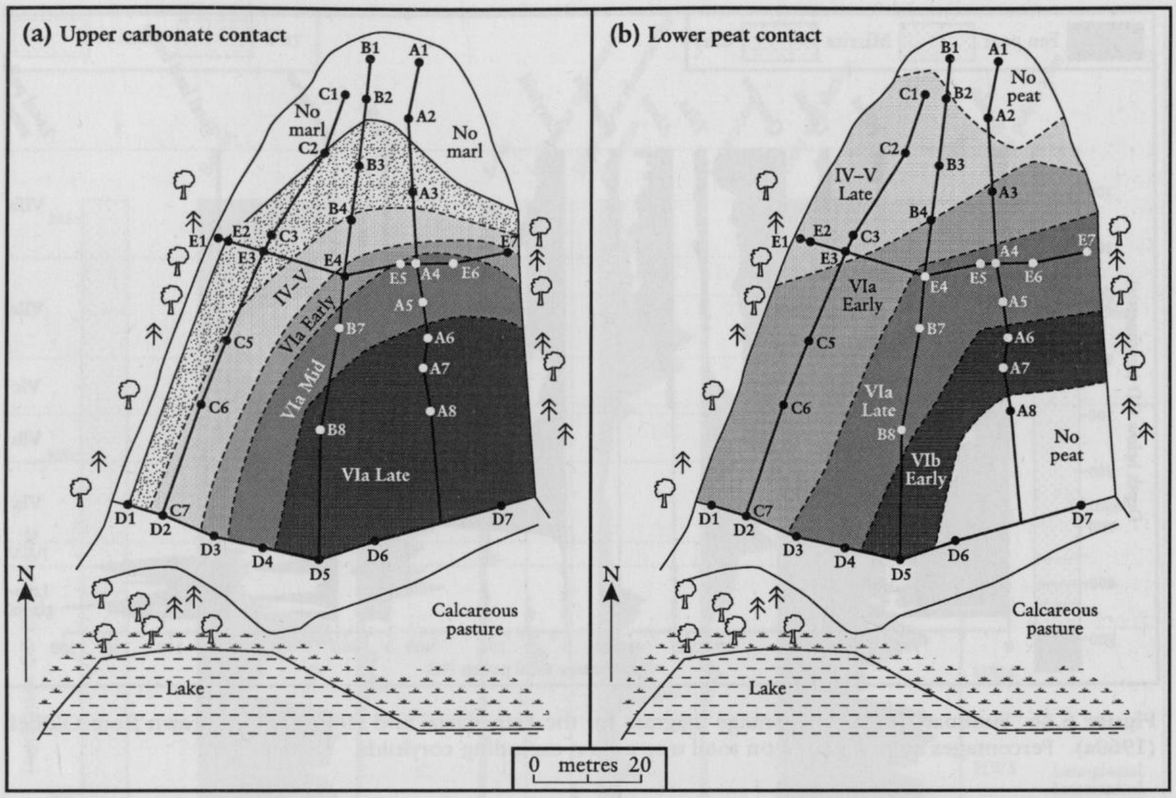



**Figure 6.37** Dating of the peat–marl contact from the north field. A1–E7: location of core samples. (A) Lakeward younging charting the progradation of the carbonate bench system. (B) Subsequent hydroseral encroachment of the marginal peat deposits.

tion rate for the mid- to late Holocene determined using the decay of excess $^{226}$Ra was less than half that obtained from $^{210}$Pb dating of sediments deposited during the past 150 years or so (Fairclough, 1999).

## Conclusions

Hawes Water is an important site because it is one of only four carbonate lakes in northern England and the only lake of natural origin remaining in Lancashire. Its marginal sedimentary record presents a record of uninterrupted lacustrine deposition dating from the end-Devensian through to the mid-Holocene (*c.* 15.0–5.0 ka). Analysis of the Hawes Water sediments has provided information on the development of regional vegetation and lake-level change for the Late-glacial and Holocene. The importance of this sedimentary archive is further re-inforced by the carbonate nature of the sediments, which are known to be particularly sensitive to climatic and anthropogenic impacts. The sensitivity of the Hawes Water record is best exemplified by the identification of four short-term climatic fluctuations during the Late-glacial, recording a degree of climatic instability previously unrecorded at sites in England.

# CROSE MERE (SJ 430 306)

*N.F. Glasser*

## Introduction

Crose Mere is an important site for reconstructing the Devensian Late-glacial and Holocene vegetation and environmental history of the north Midlands. The mere is typical of the many shallow depressions (the 'mosses and meres') of this part of Shropshire and its pollen record provides important comparisons both regionally and nationally. Beales (1980) obtained a sediment core from beneath the deepest part of the lake and used this to reconstruct the Late Devensian and Holocene vegetational history of the area surrounding the mere. Pollen analysis was used to determine changes in the vegetation and radiocarbon dating of organic deposits was used to provide geochronological control on the timing of these events. Overall, the pattern and timing of vegetational change at Crose Mere show close similarities with those obtained from other sites in the north-west of England. Barber and Twigger (1987) have considered Crose Mere in part of their wider discussion of the Late Quaternary vegetational history of the Severn Basin, and Twigger (1988) and Hobby (1990) provide overviews of the palaeoecology of Shropshire as a whole. Details of the modern vegetation communities at the site are given by Sinker (1962). The general limnology and water chemistry of Crose Mere are described by Gorham (1957). The phytoplankton of the mere has been described in a series of papers (Reynolds, 1971, 1973a, b; Reynolds and Reynolds, 1985). Finally, the modern invertebrate community of the mere has been documented by Harrison and Hildrew (1998).

## Description

Crose Mere is located 6 km to the south of Ellesmere in Shropshire. In this area are numerous shallow depressions, many of which are developed on an irregular topography composed of glaciogenic deposits. This irregular terrain traditionally has been interpreted as part of an end moraine complex extending from Ellesmere to Bar Hill, probably associated with the recession or a minor readvance of the Late Devensian ice-sheet in this area (Boulton and Worsley, 1965; Worsley, 1969; Shaw, 1972b). Crose Mere is typical of the lakes associated with the Ellesmere-Bar Hill moraine. It is a shallow dimictic lake, with an overall area of 15.2 ha and a maximum water depth of 9.2 m (Beales, 1980). Underlying the lake deposits are (from top down): glaciogenic sands and gravels, till, and Triassic bedrock.

Together with Whattall Moss and Sweat Mere, Crose Mere is part of a complex of three lakes. There is some evidence of a recent lowering of the levels of these meres. In 1864, drainage operations at the site yielded an iron-fashioned oak dug-out canoe in Whattall Moss, similar to one from Breconshire attributed to Roman times (Fox, 1926). This has been taken as evidence for open water conditions in the area at the time. The 1864 drainage operations caused a drop in the water level of Crose Mere of 2 to 3 m (Hardy, 1939). A ditch of unknown age crossing the gravel ridge that separates Crose Mere from Whattall Moss indicates that at some point in the past Crose Mere was 2.5 m higher than its present 87 m OD (Peake, 1909). Chemical analyses show that the modern lake water is rich in $Ca^{2+}$ and $HCO_3^-$, with a conductivity of 300 to 400 $\mu$mho cm$^{-1}$ (Gorham, 1957; Reynolds, 1973a). Bare ground covers the sandy south-east and stony north-east shores, whereas the remainder of the lake is typically reedswamp with limited areas of alder. The surrounding mosses, such as Whattall Moss, support pine–birch woodland, *Myrica gale*, *Deschampsia flexuosa*, *Vaccinium oxycoccos*, *Dryopteris* spp. and *Sphagnum* spp. (Sinker, 1962).

The pollen diagram produced by Beales (1980) from coring of the lake bed at Crose Mere is shown in summarized form in Figure 6.38. The core has been divided into 10 pollen assemblage zones (LPAZ) based upon pollen and spore content, each of which is given the local code CMCP.

**CMCP-1 (612–598 cm) Gramineae–Cyperaceae LPAZ**
Herb pollen is over 60% with Gramineae pollen the dominant type, followed by Cyperaceae, *Artemisia* and *Thalictrum*. Arboreal pollen is less than 15%, with some *Juniperus communis* pollen values over 10%.

**CMCP-2 (598–563 cm) *Betula*–Gramineae LPAZ**
*Betula* pollen values are 20 to 40%, with Gramineae over 30%. Cyperaceae pollen is less that 10%. Shrub pollen consistently exceeds 10%, with *Salix* and *Juniperus* the major contributors.

**CMCP-3 (563–532 cm) Gramineae–*Betula*–Cyperaceae LPAZ**
Herb pollen values are 60 to 80%, with Gramineae pollen 20 to 50% and Cyperaceae pollen 10 to 25%. Some *Artemisia* and *Rumex acetosella* pollen values exceed 10%. Shrub pollen is less than 10%, with *Salix* the major type.

**CMCP-4 (532–516 cm) *Betula* LPAZ**
Arboreal pollen (predominantly *Betula*) is 50 to 80%. *Salix* pollen is the dominant shrub type, with values of 5 to 15%. *Corylus* pollen is less

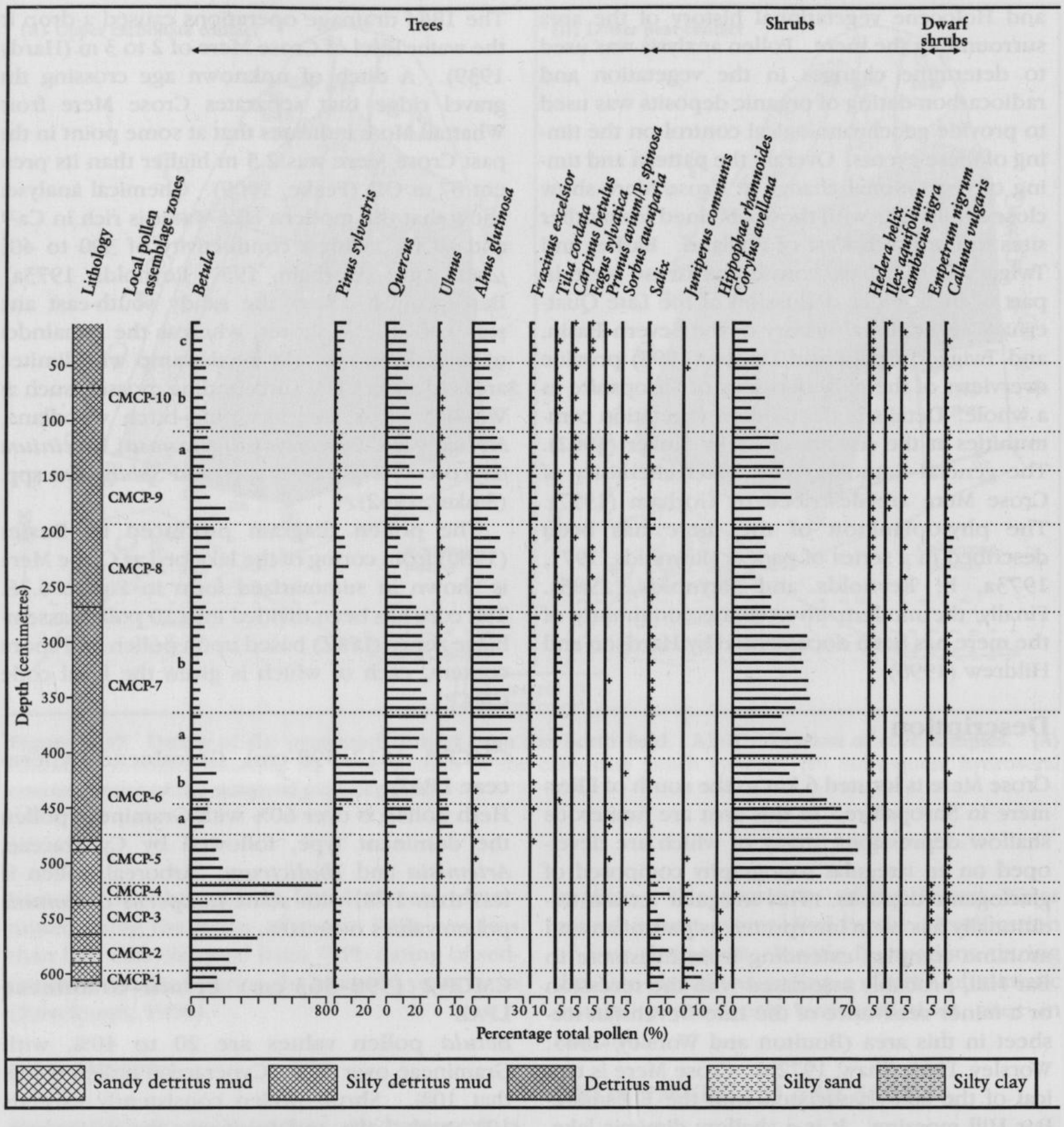
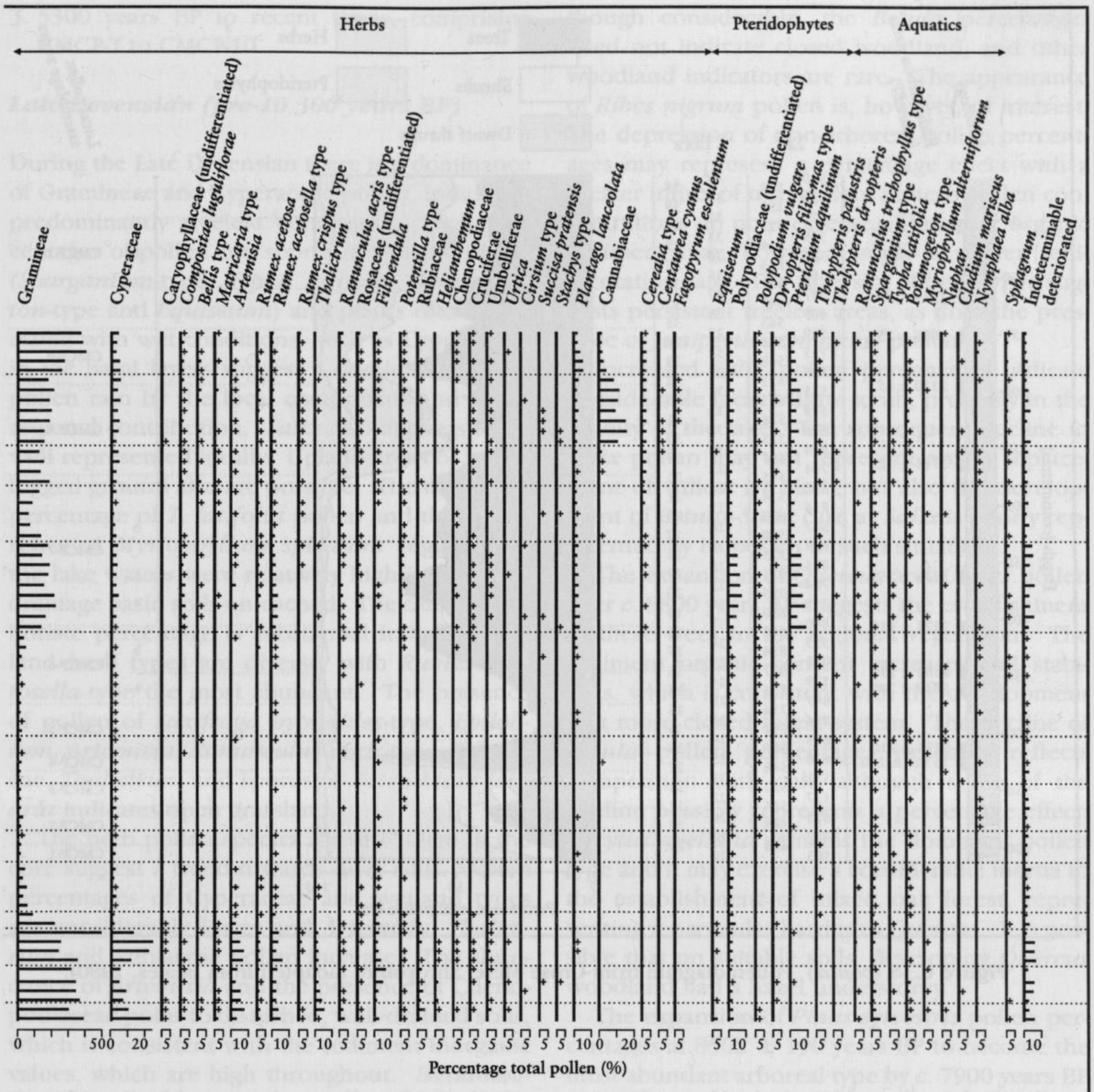


**Figure 6.38** Pollen diagram from Crose Mere, Shropshire (adapted from Beales, 1980). (Continued on p. 282.)

than 1%. *Calluna vulgaris* and *Empetrum nigrum* show continuous curves. Herb pollen values are between 10 and 30%, with Gramineae dominant. *Filipendula* pollen values are more than 2%.

**CMCP-5 (516–447 cm) *Corylus avellana* LPAZ**
*Corylus avellana* pollen percentages are 50 to 75%, with diverse arboreal pollen present, including *Quercus* and *Ulmus*. Herb pollen is less than 10%.

**CMCP-6 (447–407 cm) *Pinus sylvestris–Quercus* LPAZ**
*Pinus sylvestris* pollen is 10 to 25% and *Quercus* pollen is consistently over 15%. *Alnus glutinosa* pollen also is present but less than 10%. *Ulmus* pollen exceeds 5%. *Tilia cordata* is present. *Corylus avellana* is the major pollen contribu-




tor (30 to 60%) and herb pollen is consistently below 2%.

**CMCP-7 (407–274 cm) *Quercus–Alnus glutinosa* LPAZ**

Arboreal pollen is consistently 50% or more, with *Quercus* pollen values consistently 20% or greater and *Alnus glutinosa* pollen between 15 and 25%. *Pinus sylvestris* pollen is consistently under 5% and *Betula* is less than 10%. *Ulmus* pollen values vary, and *Tilia cordata* and *Fraxinus excelsior* pollen are continuously present. *Corylus avellana* pollen is over 30%. Herb pollen is less than 10%.

**CMCP-8 (274–193 cm) Gramineae–*Pteridium aquilinum* LPAZ**

Herb pollen is 20 to 35% with Gramineae over 10%. Pteridophyte spores are 5 to 15%, with *Pteridium aquilinum* values over 5%. Arboreal

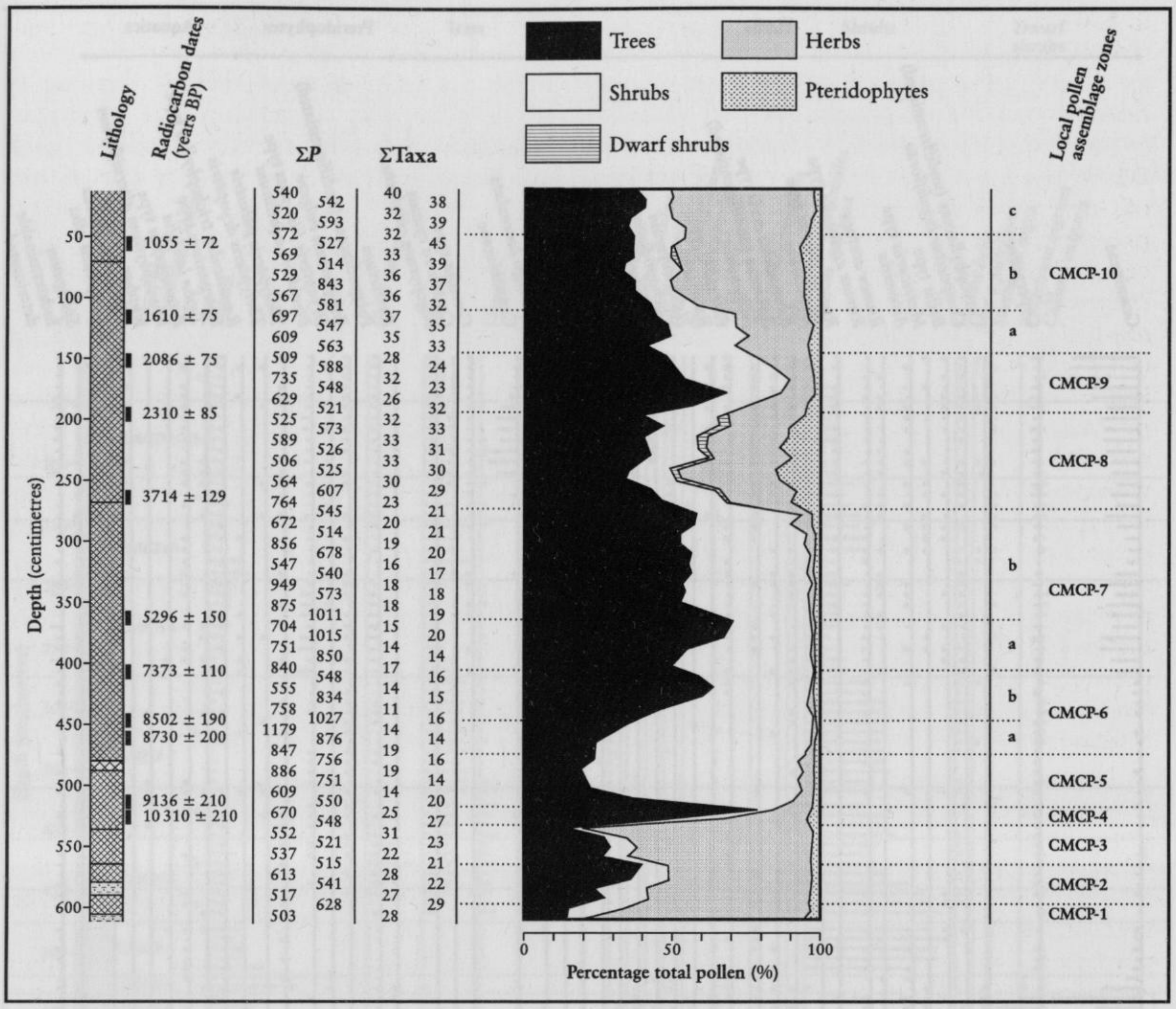



**Figure 6.38** (*contd*) Pollen diagram from Crose Mere, Shropshire (adapted from Beales, 1980).

pollen is less than 50%, with *Alnus glutinosa*, *Quercus*, *Betula* and *Pinus* the major contributors. *Calluna vulgaris* pollen is consistently over 1%.

**CMCP-9 (193–146 cm) *Quercus–Betula* LPAZ**
*Quercus* pollen levels are consistently over 20% and *Betula* values 5 to 25%. *Alnus glutinosa* pollen generally exceeds 15% and *Fraxinus excelsior* pollen ranges from 2 to 5%. Other arboreal pollen values are low. Herb pollen is diverse, but values are less than 20%, with Gramineae dominant.

**CMCP-10 (146–11 cm) Gramineae–Cannabiaceae LPAZ**
Herb pollen is 20 to 50%, with Gramineae 10 to 40%. Herb pollen is diverse with, *Cerealia*-type pollen consistently over 2% and *Plantago lanceolata* pollen 1 to 6%. Cannabiaceae pollen shows values over 15%. Arboreal pollen is less than 50%, with diverse types. *Picea* and *Acer* pollen are restricted to this zone.

## Interpretation

The pollen diagram shows evidence for three temporal divisions in the environmental history of Crose Mere (Beales, 1980):

1. The Late Devensian (pre-10 300 years BP), comprising CMCP-1 to CMCP-3;
2. 10 300 to 5300 years BP, comprising CMCP-4 to CMCP-6;

3. 5300 years BP to recent times, comprising CMCP-7 to CMCP-10.

### *Late Devensian (pre-10 300 years BP)*

During the Late Devensian there is a dominance of Gramineae and Cyperaceae pollen, indicating predominantly treeless conditions. High percentages of pollen and spores of wetland plants (*Sparganium*-type, *Typha latifolia*, *Potamogeton*-type and *Equisetum*) and plants often associated with wet conditions, such as Cyperaceae, in the basal layers suggest a dominance of the pollen rain by the local component and a low regional contribution. *Salix* also appears to be well represented locally. Upland areas of waterlogged ground also are possible. The maximum percentage of *T. latifolia* pollen and the occurrence of *Myriophyllum spicatum* suggest that the lake waters were relatively high in ions and drainage basin soils unleached. The calcium carbonate percentage is substantial initially. Dryland herb types are diverse, with *Rumex acetosella*-type the most abundant. The presence of pollen of *Saxifraga hypnoides*-type, *Epilobium*, *Artemisia*, *Ranunculus acris*-type, Rubiaceae, *Thalictrum* and spores of *Botrychium lunaria* indicates open grassland.

The herb pollen spectra above 610 cm in the core suggest a trend towards drier soils. Pollen percentages of Cyperaceae and wetland types are considerably lower and *Artemisia*, *Thalictrum* and Rubiaceae pollen increase. The abundance of *Artemisia* and the presence of Chenopodiaceae point to disturbed, well-drained soils, which is consistent with the sediment inorganic values, which are high throughout. *Helianthemum* (probably *chamaecistus*) pollen appears and is suggestive of short turf communities on well-drained soils. Throughout the Late Devensian there appears to be a steady increase in concentrations of Gramineae and *Betula*, representing climatic amelioration during and after the deglaciation of Shropshire.

### *10 300 to 5300 years BP*

During the period 10 300 to 5300 years BP, there is a substantial increase in birch woodland around 10 310 ± 210 years BP. The fourfold increase in sediment organic content points to stabilizing drainage-basin soils, and early peaks of *Filipendula* and *Juniperus communis* are consistent with a climatic improvement. Although considerable, the *Betula* percentages need not indicate closed woodland, and other woodland indicators are rare. The appearance of *Ribes nigrum* pollen is, however, of interest. The depression of non-arboreal pollen percentages may represent a percentage effect with a greater influx of tree pollen, as herb pollen concentrations do not decrease. The more organic sediment may, however, imply a slower sedimentation rate. The diversity of herb types suggests persistent treeless areas, as does the presence of *Juniperus communis* pollen.

Increased *Salix* pollen percentages indicate considerable local willow scrub, probably in the vicinity of the lake. The subsequent decline in *Salix* pollen may not represent only a replacement of willow by birch, but also the development of *Betula–Salix* carr, as *Salix* is poorly represented by its pollen in such situations.

The expansion of *Quercus* and *Ulmus* pollen after *c.* 8800 years BP suggests the establishment of these trees in the regional vegetation. The sediment organic content increases and stabilizes, which is consistent with the development of a more closed forest system. The decline of *Betula* pollen percentages probably reflects competition with oak, although some of the decline possibly represents a percentage effect. *Corylus avellana* remains the dominant pollen type and it may exercise a considerable inertia to the establishment of mixed oak forest, representing a particular barrier to *Quercus*. It is possible that on suitable soils, developing *Quercus* woodland had a hazel understorey.

The expansion of *Pinus sylvestris* pollen percentages at 8502 ± 190 years BP to become the most abundant arboreal type by *c.* 7900 years BP suggests a substantial spread of pine into the area. On the continent and in south and eastern England it clearly expands before *Corylus*. Its failure to do so in Shropshire may result from its slow dispersal and the relatively strong preferences of its seedlings, resulting in a potentially slow migration rate that possibly allowed it to be overtaken by competitors on the way north. *Pinus sylvestris* seedlings cannot regenerate under hazel, and even birch woodland provides a barrier.

The environment into which pine expanded is difficult to determine. Development in established mixed oak forest is unlikely. The slight increase in *Betula* percentages may indicate colonization of openings, caused either by fire or windthrow, but these openings are probably insufficient to explain the high pine pollen for over 1000 years duration. Drying out of waterlogged areas with an initial colonization by birch succeeded by pine provides a plausible explanation. An increase in sediment minerogenic content contemporaneously with the pine maximum would be consistent with a lowering of water levels and some marginal erosion. It is probable that pine occupied areas previously devoid of extensive woodland. Permanent soil deterioration is unlikely owing to the later increase of plants of richer soils.

The decrease of *Corylus avellana* pollen percentages probably reflects its restriction by expanding mixed oak forest, pollen production being considerably reduced by shading. The continued presence of the light-demanding trees indicates the persistence of openings in the canopy.

*Alnus glutinosa* pollen first appears *c.* 9000 years BP, but from the commencement of a continuous curve at *c.* 8150 years BP to the first value over 5% takes about 500 years and to the maximum plateau, where percentages reach 20%, takes a further 1000 years, although maximum concentration and deposition rates are reached slightly earlier. *Alnus glutinosa* can be restricted by late spring frosts or high winds, but the presence of suitable damp soils would seem the most important determinant of distribution in the region. Its slow expansion therefore can be related to lack of such soils, which is consistent with the expansion of pine during a phase of lower water-tables.

### *5300 years BP to recent times*

A modification in forest composition is indicated at 5296 ± 150 years BP, with a decline of *Ulmus* pollen percentages and accompanying declines in *Alnus glutinosa* and *Tilia cordata*, and a rise in *Fraxinus excelsior*. *Ulmus* values fall from 9% to 1% in 30 cm (400 to 500 years). The main percentage increase is that of *Corylus avellana*, possibly representing a response to the removal of forest canopy or some replacement of *Ulmus*. Herb pollen percentages shown no significant increase and remain low. The *Pteridium aquilinum* spore curve becomes continuous and points to a more open woodland, as does the reappearance of *Calluna vulgaris* pollen.

More conclusive evidence of openings in the forest canopy are provided by the small peaks of herb pollen after 303 cm. Percentages reach 7% at 287 cm in the core, coinciding with a further decline in *Ulmus* and *Tilia cordata* pollen percentages. Herb pollen types are not diverse, but include *Plantago lanceolata*, *P. major*/*P. media*, Chenopodiaceae, *Rumex acetosella*-type and *Ranunculus acris*-type. *Ilex aquifolium*, a tree favoured by high light intensities (Godwin, 1960b) whose pollen production is severely limited in forests, first appears in the pollen record. The appearance of *Lonicera periclymenum* and *Viscum album* and the start of a continuous curve for *Hedera helix* pollen also may be interpreted as a reflection of openings in the forest.

Beales (1980) considered this pattern of change and the lack of progressive deforestation to be consistent with a more settled form of agriculture of a homestead nature. Herb communities appear varied and pollen of plants of open and sandy soils and woodland continue to be represented.

## Conclusions

The pollen diagram from Crose Mere represents the only pollen sequence from north Shropshire that includes the Late Devensian, the first from a site characterized by open water conditions throughout its history, and the only pollen record from this area with a long sequence of radiocarbon dates. The pollen record indicates that during the Late-glacial Pollen Zone I the area was treeless and dominated by grasses and sedges. During Zone II, tree birches expanded in the area but substantial frequencies of herbaceous taxa suggest the absence of closed woodland. In Zone III, frequencies of herbaceous pollen increased and grassland vegetation dominated. The early Holocene is dominated by arboreal pollen, before a sharp decline in elm at 5296 ± 150 years BP. The first significant human impact on vegetational composition appears around 3900 years BP, where the pollen record indicates a mainly pastoral economy with some cereal crops. Environmental reconstructions from Crose Mere therefore provide a framework within which shorter records from other sites in north Shropshire and Cheshire can be compared. Overall, the site provides valuable information regarding the nature and rate of vegetation change in the north Midlands from the Late Devensian through to the present day.