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 (Bjork *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 BR 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 Bjork *et al.* (1998) based on the GRIP ice-core oxygen isotope record is also used for intersite 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 Bolling–Alle.rod Interstadial complex of northwest 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 wereclearly 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 (Bjork *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; Bjork *et al.,* 1998). Thus two brief colder phases can be recognized within the Late-glacial interstadial phase (Bjork *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 (Bjork *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, (GI1b), 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 deglaci-ation 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 *Hippophae,* 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 (Bjork *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 Bolling 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 (Bjork *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 Karnes (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 Grose 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 Gill) 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 lithos-tratigraphy. Many unpenetrated clastic units that underlie early Flandrian organic sequences, as in the meres and mires of the Cheshire–Shropshire plain (Talus, 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 levees 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 levee 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 belowShirdley Hill Sand in the Mersey lowlands. In the Vale of York at Cawood a peat above levee 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 (Bucldand, 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 BR 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).

References