Quaternary of Northern England

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Chapter 8

The Holocene (Flandrian) history and record of northern England

INTRODUCTION

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Northern England contains major, contrasting geographical zones that, superimposed upon the legacy of glacial deposition (Johnson, 1985b), have encouraged great diversity in Flandrian sedimentation (Thomas, 1999). Basic to this diversity is the altitudinal gradient from the Pennines and other uplands to the estuarine and littoral areas of the coastal zone, with major river systems adding complexity to intermediate lowlands in The Vale of York, Northumbria, and the Lancashire, Cheshire and Shropshire plains. These major zones form the basis of landscape units within which variability in soils, climate, vegetation and human land use in Flandrian times have all acted to alter depositional regimes.

Flandrian sediments are common throughout northern England, but are increasingly subject to destruction by recent human activities such as ploughing, drainage, forestry, recreation and industry, with deposits of the last few millennia under particular threat. The overview provided here makes reference to key sites in order to describe the range, character and palaeoenvironmental value of the surviving Flandrian sediments. Particular attention is paid to wetland organic deposits because of the range of diagnostic biological data that they preserve and their abundance in the region, but also to associated clastic sediments where appropriate. A series of recent detailed surveys of lowland wetland sediments in north-west England and Humberside (e.g. Cowell and Innes, 1994; Van de Noort and Ellis, 1995a, b, 1999; Leah *et al.*, 1997) have proven the richness and diversity of the Flandrian geological resource. The broad development of Flandrian vegetation in northern England, and the evidence for the human role in changing it, are briefly summarized.

The lake basin to raised bog hydrosere

A key component of northern England's Flandrian geological resource is the sediment record preserved within the very numerous enclosed basins, which occur in all parts of the region, and which contain wetland depositional environments at all stages of hydroseral succession between lake and raised bog (Walker, 1970; Shimwell, 1985). Many depressions originated as a result of glacial erosion or deposition and have accumulated sediment since the early Late-glacial, but they also contain deep Flandrian successions, which are mainly organic but include gyttja, marl, silt, clay and coarser-grained deposits. In many cases Flandrian organic materials form the basal sediment in shallower basins, spreading from deeper centres or becoming waterlogged under wetter climate in the mid- and later Flandrian. Some basins were formed post-glacially as a result of local events, such as salt solution subsidence in the Cheshire Plain (Reynolds, 1979) or behind landslip features (Franks and Johnson, 1964; Simmons and Cundill, 1974b). Lacustrine sediments dominate the early stages of many basins, but only a few of the largest or deepest have persisted as water bodies to the present day, notably the large Lake District lakes, but also those in the upland rock basins of Cumbria (Pennington, 1964) and the meres of the Cheshire–Shropshire plain (Green and Pearson, 1977; Reynolds, 1979). Lake sediments in these areas have preserved sensitive chemical, diatom, mineral magnetic and other proxy records of environmental changes in their catchments (Pennington, 1970; Oldfield *et al.*, 1985), and diatom floras also can be interpreted in terms of recent climate history, as in recent lake sediments from Windermere, Cumbria (Barber *et al.*, 1999). Elsewhere extant lakes are isolated landscape features, such as Gormire Lake (Blackham *et al.*, 1981) or Malham Tarn (Piggott and Piggott, 1963) in Yorkshire. Many more smaller lakes would still exist were it not for their deliberate drainage and reclamation, as in Holderness (Sheppard, 1957; Flenley, 1987) where, compounded by coastal erosion (Gilbertson, 1984a), only Hornsea Mere survives (Beckett, 1981). Many basins have naturally passed through the various stages of hydroseral succession and culminated as acid raised bogs in the later Flandrian, although many remain in an intermediate stage (Tallis, 1973b). Some sites have developed to raised bogs over coastal marine clays (Smith, 1958c; Oldfield, 1960a; Birks, 1982; Tooley, 1978a; Wells *et al.*, 1997). The long pollen records recovered from these lake and mire sequences, particularly those with good dating control (Godwin *et al.*, 1957; Hibbert *et al.*, 1971; Hibbert and Switsur, 1976; Wells *et al.*, 1997), form the basis for study of Flandrian climatic and vegetation change in northern England at focused (Walker, 1966b;

Bartley *et al.*, 1976; Davies and Turner, 1979; Bartley *et al.*, 1990; Dumayne and Barber, 1994; Dumayne-Peaty and Barber, 1998) or more synoptic (Gilbertson, 1984a; Twigger and Haslam, 1991; Cowell and Innes, 1994; Wimble *et al.*, 2000) scales. These data allow regional overviews of environmental history (e.g. Innes, 1999; Innes *et al.*, 1999) and can be enhanced by other evidence such as insects (Buckland, 1979) or Mollusca (Keen *et al.*, 1988). Ombrotrophic raised bogs in northern England are crucial for the study of late Flandrian climate change. Recurrence horizons in bog stratigraphy, most notably the major mid-third millennium BP event, but others also, record wet shifts in climate history (Barber, 1982). Dates of 2447 ± 43 to 2345 ± 45 years BP from Lindow Moss in Cheshire (Leah *et al.*, 1997), 2685 ± 50 years BP from Featherbed Moss in the southern Pennines (Tallis and McGuire, 1972), 2730 ± 110 to 2530 ± 80 years BP from Fenton Cottage in Lancashire (Middleton *et al.*, 1995), 2686 ± 50 years BP from Rusland Moss in south Cumbria (Dickinson, 1975) and 2645 ± 100 years BP from Chat Moss in Greater Manchester (Godwin and Switsur, 1966) are typical for the major Iron Age acceleration in bog growth in northern England. More detailed studies of peat humification, macrofossil stratigraphy and other proxies provide a highly sensitive record of mire palaeohydrology (Dickinson, 1975; Barber, 1993; Barber *et al.*, 1998; Chiverrell and Atherden, 1999; Mauquoy and Barber, 1999b; Hughes *et al.*, 2000). The recognition of tephra layers in such peat sediments at Fenton Cottage (Pilcher and Hall, 1996) is an important recent development. Most raised mire systems in northern England have been extensively damaged by cutting (e.g. Oldfield, 1970), and sites in the whole range of the lake-to-raised-bog hydrosere have been damaged by drainage and other human activity.

Intertidal and coastal wetlands

Post-glacial rise in sea level carried the shoreline of northern England to approximately its present position by the mid-Flandrian (Tooley, 1982, 1978a; Plater, *et al.*, 1993; Zong and Tooley, 1996; Long, A.J. *et al.*, 1998; Shennan *et al.*, 2000a, b), after which a suite of wetland depositional regimes became established in the coastal zone – based upon altitude and salinity, which had either a direct or indirect relationship with marine conditions and tide level. Under direct marine influence were the sand and mudflat environments of the lower intertidal zone, which deposited sands, silts and clays, as well as the upper intertidal vegetated saltmarsh zone, in which increasingly organic peat sediments were laid down. Raised groundwater tables and reduced rates of river flow owing to high sea level caused the creation of a belt of freshwater wetland environments between high tide level and higher ground. This 'perimarine' zone (Tooley, 1985), indirectly stimulated by sea-level change and also accepting seasonal freshwater drainage from landward, was of considerable extent on low-gradient coasts and penetrated well into the lower parts of the region's river valley floodplains. It comprised a mosaic of eutrophic lagoons, swamps, fens and freshwater meres prone to rapid water-level changes and sustaining complex organic sedimentary regimes, but susceptible to penetration by marine conditions. The Downholland Moss and Martin Mere area of south-west Lancashire is a good example of these littoral wetland environments (Tooley, 1978a, 1985). Some peat-forming systems in these coastal fringe locations progressed to raised-bog communities, as in north Lancashire (Smith, 1958c; Oldfield, 1960a; Middleton *et al.*, 1995), the Solway lowlands (Walker, 1966b), south-west Lancashire (Tooley, 1978a) and the Humberhead Levels (Dinnin, 1995). Low-amplitude fluctuations in sea level, local changes in coastal morphology and in sediment flux caused spatial readjustment of coastal depositional environments, so that deep, complex stratigraphical sequences of alternating intertidal and perimarine peats, silts and clays accumulated in low-lying coastal areas. These occur particularly in the major river estuaries, such as the Humber (Metcalfe *et al.*, 2000) and Tees (Plater *et al.*, 2000a), and also in small embayments and adjacent to flat open coasts, such as those of Lancashire and Merseyside (Huddart *et al.*, 1977; Tooley, 1978a; Kenna, 1986; Huddart, 1992; Cowell and Innes, 1994). In these areas mid-Holocene marine sediments are found well inland of the present coast. Clear geological evidence of fluctuating sea level and past coastal change lies in the peat beds of terrestrial origin that are exposed in the present-day intertidal zone, often containing remains of ancient woodland overtaken by rising sea level. These intertidal 'submerged forests' were hitherto much more extensive than today (Reade, 1871), and many recently have undergone severe erosion. Of widely differing ages, they are exposed analogues of the terrestrial elements of the intercalated coastal successions (Tooley, 1985). Good examples exist in the Humber estuary (Long, A.J. *et al.*, 1998), in Merseyside (Tooley, 1978a), in Hartlepool Bay (Tooley, 1978b), and in Cumbria (Oldfield, 1965; Tooley, 1985). The resource-rich environments represented by the intercalated coastal deposits were very attractive to past human settlement and activity, as shown by abundant archaeological sites and palaeoecological evidence in these locations (Fulford *et al.*, 1997). Throughout northern England, from Cumbria (Bonsall *et al.*, 1994) to Wirral (Kenna, 1986), from Hartlepool Bay (Tooley, 1978b) to the Humber estuary (Crowther, 1987), there is evidence of human settlement and activity having been closely affected by shoreline development during several cultural periods. Human exploitation of coastal zone palaeoenvironments around Liverpool Bay is a good illustration (Gonzalez *et al.*, 1996; Huddart *et al.*, 1999a, b), where quantities of animal bone of several species, wild and domestic, have been recovered from intertidal sediments, many with cut marks. Human and animal footprints preserved in the estuarine silts from Mesolithic times onwards are direct evidence of human use of these productive coastal environments.

Eutrophic reedswamp, fen and carr

Sedimentation within eutrophic peat-forming systems became a major element in northern England's mid- and later Flandrian depositional history, as such environments must have been abundant in lowland vales and valley bottoms prior to their large-scale drainage in recent centuries. Fen and carr wetlands fringing floodplain alluvial systems became ubiquitous after mid-Holocene higher sea level, wetter climate and river regime stabilization with low gradients led to increasingly poorly drained soils in low-lying areas. Eutrophic floodplain and valley mires often came to be dominated by *Alnus* associations (Brown, 1988) and these stable fen and fen–carr systems could persist for millennia, replenished by winter flooding, and depositing slowly forming peats and minerotrophic organic silts. The Gowy valley in Cheshire is a good example of long-term persistence of valley fen wetland (Shimwell, 1985), but most such locations would have been similar. Superabundant alder pollen frequencies from valley fen peats, often well over 80% of total pollen, are maintained at some sites for thousands of years, as at Ince Blundell in the Alt valley, Merseyside (Cowell and Innes, 1994). At other sites in the later Flandrian, permanent waterlogging and marsh and fen sedimentation in valley bottoms could be easily stimulated by single events that impeded local drainage, as with possible beaver activity at Briarfield, Lancashire (Wells *et al.*, 2000). The most extensive areas of floodplain marshlands in northern England were probably those of the Humberhead Levels and southern Vale of York (Dinnin, 1997a; Lillie and Gearey, 1999), where between about 5000 years BP and 2500 years BP most of the dry-land plain through which the lower courses of the major Humber tributary rivers flowed was converted to eutrophic mire as a result of gradual waterlogging and peat formation. In some places, as at Thorne Waste and Hatfield Moors (Dinnin, 1997b), the process was so rapid as to drown woodland and preserve tree stumps beneath the peat. Eutrophic reedswamp, marsh and fen–carr wetlands also filled the Vale of Pickering lowland in north-east Yorkshire in the early Holocene, attracting intensive Early Mesolithic occupation and exploitation (Day, 1996; Mellars and Dark, 1998) of these productive environments. Star Carr and adjacent sites are classic locations for study of palaeoecology and environmental archaeology, but such eutrophic wetland areas must have existed in many parts of northern England during most of the Flandrian, although many of these deposits have not survived or have been reduced to thin organic soils.

Drainage channel mires

Deglaciation of northern England has left several areas with distinctive erosional meltwater channel features that comprise a valley depositional environment, and which have accumulated sediments during the Flandrian (Johnson, 1985b; Hemingway, 1993). Often these channels cut through low watersheds and form long but narrow linear hollows, which preserve a spatially continuous record of environmental change on adjacent dry-land areas. Laterally extensive charcoal or mineral inwash stripes, which reflect catchment events, are a characteristic feature of their sediment column (Simmons, 1969a, b; Simmons *et al.*, 1975). Many such features occur on the North York Moors, where key sites such as Ewe Crag Slack (Jones, 1978) and Fen Bogs (Atherden, 1976a) contain multiple inwash horizons within a long organic sequence, and they form a major part of the Flandrian palaeoenvironmental resource in that area. Meltwater channels occur in other parts of northern England also, although their potential has been less developed. They are abundant in the Pennines (Mitchell, 1991a) and the sediments of some in the Cheshire–Staffordshire area have been studied (Yates and Moseley, 1958; Johnson *et al.*, 1970). This category of site holds great potential for future research.

Upland hill peat and blanket bog

A conspicuous feature of the northern England sedimentary environment is the extensive tracts of hill peat and blanket bog that cover much of the uplands of the region, and which have developed and spread during the mid- and late Flandrian (Moore, 1988; Tallis, 1991). Mid-Flandrian climatic deterioration probably prompted the inception of blanket peat in northern England (Moore, 1988), but there is evidence that in most uplands of the region deforestation by humans from Mesolithic times onwards probably also initiated peat formation (Simmons and Cundill, 1974a; Pennington, 1975c; Squires, 1978; Tallis and Switsur, 1983). Blanket peats are best developed on the flat plateaux watershed surfaces of the Pennines, North York Moors and Cheviot Hills, often where they have spread from earlier foci of organic accumulation within shallow basins. Over wide areas, however, thinner blanket peats are unconfined and cover most of the low-gradient, water-shedding upland slopes (Conway, 1954). Although human activity is implicated in peat formation, major blanket peat spread seems to have occurred as a result of the cold, wet climate in Iron Age times after 3000 years BP (Gosden, 1968; Bartley, 1975; Tinsley, 1975). Upland blanket peats in northern England are highly sensitive to climatic change and have great potential for detailed reconstruction of later Holocene climate history through their palaeohydrology. Peat humification studies on moisture-shedding sites (Rowell and Turner, 1985) reflect past temperature and rainfall and such palaeoclimate studies, also using multi-proxy micro- and macrofossil indicators, are now available for the North York Moors and the south Pennines (Blackford and Chambers, 1991; Tallis and Livett, 1994; Tallis, 1995; Chiverrell and Atherden, 1999). Most northern England upland blanket peats are now undergoing erosion and in places this is severe (Tallis, 1985b, 1987), often accelerated by recent moorland fires (Maltby *et al.*, 1990).

Upland basin mires

This category of depositional environment is an important element in Flandrian upland sedimentation in northern England, for although it is much less widespread than the thinner, water-shedding blanket peats, it was a focus for earlier organic accumulation and contains a longer sedimentary record. Upland basin mires often formed the nucleus of peat formation from which blanket peats spread to cover intervening interfluves, and their deeper, basin deposits may be difficult to locate if masked by the later, general blanket peat cover of the uplands. Basin-mire sediments often are exposed at springheads by stream erosion and many preserve detrital stratigraphical evidence of environmental changes in their small catchments in the form of wood, charcoal and inwashed mineral layers. These sites provide much of the available ecological data regarding early and mid-Flandrian environmental conditions in upland northern England. A much researched example is North Gill near the watershed of the North York Moors, summarized by Innes and Simmons (1999), where focused, multi-core, high-resolution temporal and spatial pollen and stratigraphical analyses have recorded multiple episodes of mid-Flandrian forest disturbance during the late Mesolithic and Mesolithic–Neolithic cultural phases, and further human impacts during later periods. Several similar sites have been investigated in the Pennines (Conway, 1954; Tallis, 1975; Tinsley, 1975; Sturludottir and Turner, 1985; Williams, 1985) and the North York Moors (Innes and Simmons, 1988b) and many must exist in other parts of the region.

Alluvial sediments

An important component of the Flandrian sedimentary resource is that laid down within alluvial environments and stored within northern England's fluvial systems. Most of the eastern part of the region is drained by rivers of the Ouse Basin, which feeds into the Humber estuary. Deep sequences of fluvially derived sediments have been shown to be present in upland, piedmont and lower valley floor reaches of this river system and mainly reflect the past effects of climate on river discharge, although the effects of past human land use in releasing sediment to the system by soil erosion also are significant locally (Macklin *et al.*, 2000; Taylor and Macklin, 1998; Merrett and Macklin, 1999; Howard *et al.*, 2000; Taylor *et al.*, 2000). Redeposition of transported fluvial sediment may well have affected upper estuarine depositional regimes, as postulated for the Humberhead Levels by Buckland and Sadler (1985), and demonstrated by Rees *et al.* (2000), after major catchment deforestation (Van de Noort and Ellis, 1997; Long, A.J. *et al.*, 1998) from about 4000 years BP onwards. Similarly, in the Tees estuary, Plater *et al.* (2000b) have considered the role of variations in sediment flux from terrestrial systems through catchment changes during this period. Macklin *et al.* (2000) have identified several periods of river activity in the Yorkshire Ouse in the later Holocene that match climatic changes and are marked by either valley incision or alluviation. The latter usually is fine grained but is coarser during more extreme climatic deterioration, as after 3000 years BP. Flood regimes and the transport of packets of sediment deposited as discrete alluvial units seem to be the norm, with allied organic sedimentation in riparian wetlands usually under long-term *Alnus* carr communities (Brown, 1988). Holocene alluvial histories comparable to that of the Ouse Basin after climatic and human deforestation stimuli have been reported from river valleys of widely different sizes in Northumbria (Macklin *et al.*, 1991; Macklin *et al.*, 1992; Passmore *et al.*, 1992; Tipping, 1992), in Shropshire (Brown, 1990), in the North York Moors (Richards, 1981; Richards *et al.*, 1987) and in the Howgill Fells of north-west England (Harvey, 1985). A good site example is at Seathwaite in Cumbria (Parker *et al.*, 1994), where human deforestation and farming, probably during the Norse settlement of the area, caused the burial of an organic soil by an alluvial debris fan. Alluvial sediments that bury and preserve old land surfaces can contain a rich, long-term data record within larger valleys and form a geological resource of major palaeoenvironmental importance.

Aeolian sand

Cordons of dune sand are characteristic of much of the low-lying coast of northern England, usually comprising a relatively thin dune barrier that fringes the shoreline and with a much wider but shallow blown-sand apron to landward, although blown sand can infill small embayments almost completely. Although narrow, the major dune ridge can be relatively high, approaching 30 m in the Sefton dunes of south-west Lancashire. Dune systems are to be found draped over rock outcrops in the northern part of the region, but they also rest upon till or gravel ridges, or more usually on coastal peat or marine deposits. There is evidence that dune systems may have formed earlier in the Flandrian, but extant coastal dunes mostly lie upon mid-Flandrian sediments and so are one of the region's more recent categories of geological formation (Tooley, 1990). From the mid-Flandrian onwards, dates of phases of dune emplacement have varied considerably both within and between areas. Orford *et al.* (2000) studied phases of dune movement on the Northumberland coast and proposed a model for dune history in north-east England in which landward sand movement occurs under conditions of sea-level rise, often when deceleration in the rate of rise is under way, as in the mid-Flandrian and after. This model generally is supported by other workers' palaeoecological study of dune systems (Tooley, 1978a; Kenna, 1986; Innes and Frank, 1988; Innes and Tooley, 1993; Pye and Neal, 1993a, b; Shennan *et al.*, 2000a). It also is possible, however, that a relative sea-level fall may be required so that sand may be liberated from intertidal areas to support dune building, whereas high sea level may encourage dune-slack formation and system stability. Dates of dune-building phases tend to be site specific and not capable of wider correlation. Some periods, as after 3000 years BP, do seem to have a high incidence of dune establishment, and major dunes of the Cumbrian coast date from this period (Bonsall *et al.*, 1994). The medieval period also seems to have been an important phase of dune building, as in Merseyside (Kenna, 1986) or Northumberland (Innes and Frank, 1988). Dating of dune emplacement is most often indirectly by radiocarbon ages on subjacent peat or wood and on intercalated dune-slack organic layers, although such dates for sand emplacement can vary considerably over quite short distances, for example between Crosby and Formby in Merseyside, with 4510 ± 50 years BP at Sniggery Wood, Little Crosby, 2680 ± 50 years BP at Murat Street, Waterloo and 2335 ± 120 years BP at Lifeboat Road, Formby (Innes and Tooley, 1993). Similar variation continues along the rest of the coast of north-west England (Innes and Tooley, 1993). More direct dating through optical luminescence techniques (Pye *et al.*, 1995) also has been applied, also on the Merseyside coast near Formby. Dunes also have been dated relative to archaeological material beneath or within them. The north of England coastal dune systems probably have experienced many periods of stability, instability and erosion, in both the longer term and more recent times (Pye, 1990; Plater *et al.*, 1993; Pye and Neal, 1994; Orford *et al.*, 2000). Blown sand also occurs in inland locations, particularly where periglacial coversands have been reactivated by climatic or human destabilization in the mid- and late Flandrian. South-west Lancashire and south Yorkshire are the most significant areas for inland-sand reworking (Tooley, 1978a; Innes, 1986; Buckland, 1982), but smaller scale redeposition of sands has occurred locally in many parts of the region.

Buried soils

Buried soil horizons are a category of Flandrian deposit that has provided important environmental data, particularly in areas such as Yorkshire's limestone and chalklands (Evans and Dimbleby, 1976; Smith, 1986) where few other sedimentary records may be available for study. Sealed beneath other sediments, which in many cases accumulated very rapidly, buried soils can preserve old ground surfaces that contain data on local environmental conditions, either natural environmental processes or human land-use activity. Brown Earth soils sealed beneath Bronze Age barrows in the Rossendales (Tallis and McGuire, 1972) and on the North York Moors (Dimbleby, 1962), areas now dominated by acid podsols, give insights into mid-Flandrian soil development and its relationship to vegetation change, forest clearance and agricultural history. Soil profiles in the Pennines buried beneath Roman earthworks at Fortress Dike (Tinsley and Smith, 1973) or a Roman road at High Moor (Brayshay, 1999) illustrate the landscapes in which those monuments were constructed. Northern England contains many instances of mineral or organic soil profiles buried as a result of rapid changes in depositional regime. Such soils are preserved beneath colluvial hillwash at Langdale and alluvial fans at Carlingill and Seathwaite, in Cumbria (Cundill, 1976; Harvey, 1985; Parker *et al.*, 1994), marine sediments at The Starr Hills, Lytham, Lancashire (Tooley, 1978a), blown dune sand at Wallasey in north Wirral (Kenna, 1986), landslides in Longdendale in Derbyshire (Tallis and Johnson, 1980), reactivated periglacial coversands in Merseyside (Innes, 1986) and tree stumps and blanket peat in the Lake District (Pennington, 1965). Analyses of the preserved organic soil surfaces can date emplacement of overlying clastic strata, as with colluvial inwash burying peat at Skipsea Withow Mere in Holderness (Gilbertson, 1984a). Acid soils beneath upland peats can preserve information regarding pre-peat vegetation (Tallis, 1964a) and human occupation (Radley *et al.*, 1974).

The Flandrian landscape: forest development and dominance

The sedimentary record in northern England described above is the prime source for knowledge of the environmental changes that have taken place during the Flandrian. It is a story of the development and spread of forest communities over all but a few parts of the region in the early to mid-Flandrian, with progressive deforestation by natural and, in particular, human agency in the mid- and late Flandrian. Tree remains, including stumps, occur preserved by later sediments in many locations: beneath peat in the uplands (Tallis, 1975; Simmons and Innes, 1981, 1988b; Tallis and Switsur, 1983, 1990), beneath coastal zone deposits (Tooley, 1978b; Kenna, 1986; Pye and Neal, 1993a; Horton *et al.*, 1999b), within lowland basin peats (Lageard *et al.* 1995, 1999), and beneath valley floodplain alluvial sediments (Dinnin, 1997a). Although the highest parts of the Lake District may never have naturally carried forest (Birks, 1988), it seems probable from pollen studies that tree cover extended well up the Cumbrian fells (Pennington, 1965, 1970) and that the summits of the Pennines and all lower uplands were below the mid-Flandrian tree line (Turner and Hodgson, 1983; Tallis and Switsur, 1983, 1990; Turner, 1984) by the early mid-Flandrian forest maximum about 7000 years BP. Coastal and long-term wetland areas would have carried their own specialized climax vegetation, but elsewhere forest dominance would have been the norm. Soils formed on sands or gravel may have carried some heathland throughout the Holocene Epoch, particularly where prone to instability as in the Cheshire or Lancashire plain (Tallis, 1973b; Reynolds, 1979; Kear, 1985) and the southern Vale of York (Lillie and Gearey, 1999). Edaphic conditions on limestone and chalk geology, as in Upper Teesdale, Craven–Lonsdale or the Yorkshire Wolds, also may have limited tree expansion and allowed establishment of specialized grassland or heathland associations (Turner *et al.*, 1973; Smith, 1986; Bush and Flenley, 1987; Bush, 1993). The assembly of the post-glacial forest followed early Flandrian climatic amelioration, and chronologies for the successive establishment of individual tree taxa across northern England during the early and mid-Flandrian have been achieved by radiocarbon dating of major pollen zone boundaries on several standard long-pollen diagrams, supplemented by dates from smaller profiles. Among these key dated profiles (Greig, 1996) are Scaleby Moss in Cumbria (Godwin *et al.*, 1957), Din Moss on the Scottish Border (Hibbert and Switsur, 1976), Red Moss in Greater Manchester (Hibbert *et al.*, 1971), Knowsley Park in Merseyside (Cowell and Innes, 1994), Neasham Fen in Durham (Bartley *et al.*, 1976), Crose Mere in Shropshire (Beales, 1980), Askham Bog in the Vale of York (Gearey and Lillie, 1999) and Robinson's Moss in the Pennine uplands (Tallis and Switsur, 1990).

Transitional *Empetrum*, *Juniperus* and *Salix* shrub communities rapidly supplanted the grassland and tall herb associations with *Rumex* and *Filipendula* that developed under rapid amelioration of climate (Walker *et al.*, 1994; Lowe *et al.*, 1995b; Mayle *et al.*, 1999) at the start of the Flandrian. At Thorpe Bulmer in south-east Durham, for example, *Filipendula* was able to reach 20% of total land pollen before being shaded out by shrub taxa (Bartley *et al.*, 1976). Other than on the areas of specialized geology and soils cited above, only in the highest uplands (Pennington, 1970) did herbaceous vegetation remain important well into the Flandrian. A period of *Juniperus* abundance precedes a transition to wooded environments in much of the northern part of the area, such as Northumberland and Durham (Bartley, 1966; Turner and Kershaw, 1973; Bartley *et al.*, 1976), but in more southerly lowland areas such as the Vale of York and Holderness (Bartley, 1962; Beckett, 1981), the establishment of *Betula* woodland was rapid and restricted major juniper expansion except where sandy or calcareous soils retarded succession. As with almost all of the major Flandrian forest taxa, establishment of *Betula* was delayed in the uplands, as at Robinson's Moss (Tallis and Switsur, 1990) where it was not achieved until after *c.* 8900 years BP. In the north of the region, at all altitudes, *Betula* woodland dominance was not always achieved, and at Langdale Combe (Walker, 1965), Din Moss (Hibbert and Switsur, 1976), Bradford Kames (Bartley, 1966) and Cranberry Bog (Turner and Kershaw, 1973), for example, *Salix* and *Juniperus* remained important until *Corylus* and other forest trees became established. Gradual immigration of *Corylus*, *Pinus*, *Ulmus*, *Quercus*, *Alnus* and *Tilia* took place in turn, although even across small areas there were great variations in forest composition throughout northern England in both early and mid-Flandrian times (Turner and Hodgson, 1979, 1983) because of edaphic and topographical factors (Oldfield, 1965; Pennington, 1965; Turner *et al.*, 1973; Turner and Hodgson, 1981). *Betula*, *Corylus* and *Quercus* were favoured in the lighter upland woods, whereas *Ulmus* and *Tilia* were more common on fertile lowland soils, although the latter was near the northern limit of its range in Cumbria (Piggott and Huntley, 1980). The region-wide differences in forest composition are reflected in the time-transgressive nature of the rise of the main tree types. *Corylus* had become established in many places before 9000 years BP, with some early dates to the south. In the Lancashire and Cheshire lowlands at Hatchmere (Switsur and West, 1975) the date is 9580 ± 140 years BP, at White Moss it is after 9230 ± 85 years BP (Lageard, 1992) and at Knowsley Park it is 9160 ± 80 years BP (Cowell and Innes 1994), whereas it is later at higher altitude in the southern Pennines at Robinson's Moss (Tallis and Switsur, 1990) at 8775 ± 90 years BP. In contrast dates in the north range from 9082 ± 90 years BP at Neasham Fen in the Tees valley (Bartley *et al.*, 1976) and 8940 ± 170 years BP at Din Moss at the Scottish Border (Hibbert and Switsur, 1976), to 8689 ± 50 years BP at Mordon Carr in Durham (Bartley *et al.*, 1976) and to after 8670 ± 70 years BP at Pow Hill in the north Pennines (Turner and Hodgson, 1981). There was a similar variability in the establishment and importance of the other main members of the Flandrian forest. *Ulmus* and *Quercus* immigrated soon after the rise of *Corylus* in most areas with better soils and formed mixed oak woods with a high hazel component, which are characteristic of the Boreal (Flandrian Chronozone I) forest of northern England, with *Quercus* dominant except where conditions particularly favoured elm. The timing of their rise to forest dominance was asynchronous, however, and in some areas with unsuitable soils they never became the major taxa. *Pinus* became common across the region by the early mid-Flandrian, with early establishment on lighter sandy soils as at White Moss, Cheshire at 8625 ± 50 years BP (Lageard, 1992) and at Knowsley Park, Merseyside at 8880 ± 90 years BP (Cowell and Innes, 1994). On heavier clay soils in areas such as east Yorkshire, Lancashire and Cheshire the spread of *Pinus* was later, e.g. 8196 ± 150 years BP at Red Moss (Hibbert *et al.*, 1971). Turner and Hodgson (1979) suggest that where *Corylus* and *Ulmus* formed the lowland woodland, *Pinus* was not able to become established easily. Eventually, however, pine became important almost everywhere. To the east, *Pinus* became abundant in local areas with suitable edaphic conditions, such as the sandstone crests of the North York Moors (Simmons *et al.*, 1993) or the limestone soils of east Durham (Bartley *et al.*, 1976). It remained dominant on poorer soils at high altitude in the Pennines well into the mid-Flandrian, as at Pow Hill and other sites in the north (Turner and Hodgson, 1981) and at Bradwell Sitch in Longdendale in the south (Tallis and Johnson, 1980). Although the decline of *Pinus* was delayed at higher altitude and in some areas of lowland northern England (Oldfield, 1965), it was replaced by *Alnus* across most of its range at some stage during the mid-Flandrian. *Alnus* had been present at many sites throughout the early Flandrian, as at Mordon Carr in the Durham lowlands (Bartley *et al.*, 1976), but its rise to abundance defines the start of post-Boreal Flandrian Chronozone II, the wetter mid-Flandrian Atlantic climate period. At some sites where *Pinus* was never common *Alnus* replaced other taxa, as at Scaleby Moss in Cumbria (Godwin *et al.*, 1957) where it replaced *Betula*. At most sites, however, such as Fellend Moss, Northumberland (Davies and Turner, 1979), *Alnus* directly replaces *Pinus* almost entirely, although in places the process is either delayed, as on the North York Moors with an *Alnus* rise at West House Moss of 6650 ± 290 years BP (Jones, 1977b), or gradual as at Din Moss in the Cheviot Hills (Hibbert and Switsur, 1976) where the *Alnus* rise begins around 7000 years BP but percentages do not rise sharply until 6710 ± 100 years BP. The date of *Alnus* expansion is strongly diachronous and must relate to local environmental factors. In areas of lowland eutrophic fen–carr communities, where conditions would have been very suitable for alder, its expansion occurred very early. Dates of 7640 ± 85 years BP in the Vale of Pickering (Day, 1996), 7759 ± 67 years BP at Mordon Carr in Durham (Bartley *et al.*, 1976) and 7720 ± 50 years BP at Askham Bog in the Vale of York (Gearey and Lillie, 1999) reflect this early spread. Similar early dates occur in the southern Pennines upland, however, with 7675 ± 75 years BP at Robinson's Moss (Tallis and Switsur, 1990) and 7640 ± 40 years BP at Soyland Moor (Williams, 1985). In contrast several very late dates for the *Alnus* rise occur in the northern Pennines, ranging from 6120 ± 50 years BP at Quick Moss (Rowell and Turner, 1985) and 6150 ± 160 years BP at Red Sike Moss (Turner *et al.*, 1973) to 5400±50 years BP at Fox Earth Gill (Harkness, 1981) and 5300 ± 40 years BP at Pow Hill (Turner and Hodgson, 1981). Suitable edaphic conditions rather than altitude seem to have been the major factor in alder expansion. Many dates from the region for this important biostratigraphical feature are close to an average date of around 7000 years BP, such as 7107 ± 120 years BP at the Red Moss, Greater Manchester type site (Hibbert *et al.*, 1971), 6962 ± 90 years BP at Neasham Fen in the Tees valley (Bartley *et al.*, 1976), 7180 ± 120 years BP at Walker's Heath in Cheshire (Leah *et al.*, 1997) and 6948 ± 131 years BP at Scaleby Moss in Cumbria (Godwin *et al.*, 1957). Even within small areas like the Craven district of west Yorkshire, however, the timing of *Alnus* expansion varies markedly between sites only a few miles apart, early at 7590 ± 70 years BP at White Moss on heavier soils and later at Martons Both on limestone soils at 6930 ± 90 years BP (Bartley *et al.*, 1990). *Tilia*, the final major tree of the Flandrian forest, was certainly more common than its underrepresentation in pollen diagrams would suggest (Greig, 1982) and was probably co-dominant with *Quercus* and *Ulmus* in favourable locations. Lowland southern Yorkshire seems to have had substantial *Tilia* within the mid-Flandrian forest (Beckett, 1981; Lillie and Gearey, 1999) and in the lowland limestone area of south-east Durham it had become important as early as 6760 ± 120 years BP

(Bartley *et al.*, 1976). In north Durham and Northumberland *Tilia* frequencies are much lower (Bartley, 1966; Donaldson and Turner, 1977) and the northern limit of *Tilia* as a major forest tree seems to have been the Tees lowlands. This corresponds well with the findings of Piggott and Huntley (1980) that *Tilia* was at its northern limit in Cumbria, although high percentages of lime comparable with those of southern England, have been recorded from sites in the south-east of the Lake District, as at Witherslack Hall (Smith, 1958c). *Tilia* seems never to have been a significant component of the upland forests of northern England (Greig, 1982; Turner and Hodgson, 1983; Tallis and Switsur, 1990).

The most recent pollen stratigraphical marker of significance throughout northern England is the decline in *Ulmus* frequencies around 5000 years BP, which defines the end of the mid-Flandrian forest phase. It is a clear feature of most diagrams of that period and the many dates now available generally conform to around that age. West of the Pennines the type sites of Red Moss, Greater Manchester (Hibbert *et al.*, 1971) and Scaleby Moss, Cumbria (Godwin *et al.*, 1957) provide good average dates of 5010 ± 80 years BP and 4980 ± 119 years BP, as does the date of 5099 ± 50 years BP at Gransmoor to the east (Beckett, 1981) and 5080 ± 110 years BP and 5010 ± 110 years BP at White Moss and Eshton Tarn in west Yorkshire (Bartley *et al.*, 1990), but there is much variation around this mean value and the event is not synchronous. Dates are often a few centuries later than 5000 years BP in the uplands and similarly earlier than 5000 years BP in lowland areas. Earlier lowland dates include 5440 ± 70 years BP at Williamson's Moss (Tipping, 1994b) and 5340 ± 120 years BP at Barfield Tarn (Pennington, 1970, 1975c) in Cumbria, 5468 ± 80 years BP at Neasham Fen and 5305 ± 55 years BP at Mordon Carr in lowland east Durham (Bartley *et al.*, 1976), 5296 ± 150 years BP at Crose Mere, Shropshire (Beales, 1980) and 5290 ± 80 years BP at Knowsley Park, Merseyside (Cowell and Innes, 1994). In the case of these earlier dates a secondary elm decline may occur some centuries later (Oldfield, 1963; Walker, 1966b; Tipping, 1994b). Later upland dates from the Pennines include 4794 ± 55 years BP at Valley Bog (Chambers, 1978), 4875 ± 60 years BP at Robinson's Moss (Bartley *et al.*, 1990), 4770 ± 100 years BP at Hipper Sick (Hicks, 1971), 4865 ± 50 years BP at Soyland Moor (Williams, 1985), 4780 ± 50 years BP at Fox Earth Gill (Harkness, 1981) and 4900 ± 50 years BP at Quick Moss (Rowell and Turner, 1985). In the North York Moors dates of 4767 ± 60 years BP at North Gill and 4720 ± 90 years BP at Fen Bogs (Simmons *et al.*, 1993) and in the Cheviot Hills of 4690 ± 110 years BP at Yetholm Loch (Tipping, 1996) are comparable. Earlier upland dates seem to be related to better soils, often on limestone (Bartley *et al.*, 1990). There is a clear altitudinal dichotomy in age for this feature, which holds true throughout northern England.

The Flandrian landscape: forest decline and clearance

The non-synchroneity of all the major Flandrian pollen stratigraphical changes must be partly the result of natural factors such as regional variations in environmental conditions, competitive relationships amongst taxa and their rates of immigration. Disturbance of ecosystems by exogenic factors also has been very important in the Flandrian, in particular fire and anthropogenic activity, two factors that were often linked. Deposition of both macro- and microscopic charcoal within sediment sequences has been a regular event throughout the Flandrian in northern England, showing that fire, whether of natural or cultural origin, often influenced vegetation patterns. Visible layers of charcoal occur that record local burning and the pollen record usually reflects a major vegetation change, such as deforestation, at those levels. Such layers are most common in the limnic, reedswamp and fen stages of hydroseral succession, when detrital sediment systems are better able to accept inwashed material, and so are often correlated most clearly with Mesolithic and other prehistoric cultures. These productive environments also may have been more attractive to human activity than the later bog phases of mire development, although charcoal does occur in raised-bog contexts, from surface fires or blown in.

Conspicuous charcoal layers of Mesolithic or later age occur at Simonswood Moss in Merseyside (Simmons and Innes, 1987), Hoscar Moss (Cundill, 1981), The Starr Hills, Lytham (Tooley, 1978a) and Little Hawes Water (Taylor *et al.*, 1994) in Lancashire, Malham Tarn Moss (Piggott and Piggott, 1963) and Dunford Bridge (Radley *et al.*, 1974) in the central Pennines, Valley Bog in the northern Pennines (Chambers, 1978), Ewe Crag Slack and Kildale Hall in the North York Moors (Jones, 1977b, 1978), Walker's Heath in Cheshire (Leah *et al.*, 1997) and Willow Garth in the Yorkshire Wolds (Bush, 1988a) as well as many other sites. The most researched examples are probably from north-east Yorkshire, where charcoal layers are extensive at Star Carr and adjacent sites in the eastern Vale of Pickering (Cloutman and Smith, 1988; Day, 1996; Mellars and Dark, 1998) and in the North York Moors upland at North Gill (Simmons and Innes, 1988a, 1996a; Innes and Simmons, 1999). Charcoal also often lies beneath the base of blanket and basin peats at various altitudes and may well have had a role in peat formation as a result of post-fire paludification. Mire Holes in Teesdale (Squires, 1978), May Moss (Atherden, 1979) and Collier Gill (Simmons and Cundill, 1974a) on the North York Moors, Great Close Pasture (Smith, 1986) and Extwistle Moor (Bartley and Chambers, 1992) in the Pennines, White Moss in Cheshire (Lageard, 1992) and Hatfield and Thorne Moors in the Humberhead Levels (Dinnin, 1997b), in the lowland cases with burned tree stumps, are all good examples. Although many sites preserve macroscopic charcoal, microscopic charcoal particles are very common indeed in Flandrian peats. They occur throughout early and mid-Flandrian upland peat in the Pennines at many sites, including Robinson's Moss and Alport Moor (Tallis and Switsur, 1990; Tallis, 1991), Lady Clough Moor (Tallis, 1975; Jacobi *et al.*, 1976) and Pawlaw Mire (Sturludottir and Turner, 1985), and in the North York Moors at Bonfield Gill Head and North Gill (Simmons and Innes, 1981, 1988a, b, 1996a). Microcharcoal is present throughout the mid-Flandrian *Pinus* peak at The King's Pool, Stafford (Bartley and Morgan, 1990) and detailed palaeoecological analyses have shown it to be present throughout eutrophic reedswamp and fen peat stages at Lindow Moss, Danes Moss and Walker's Heath (Leah *et al.*, 1997), Nook Farm (Hall *et al.*, 1995), Top Moss (Leah *et al.*, 1997), Knowsley Park and Simonswood Moss (Cowell and Innes, 1994), Fenton Cottage (Middleton *et al.*, 1995) and indeed virtually all of the surveyed lowland peats west of the Pennines. The same is true of the Mesolithic peats in the Vale of Pickering (Mellars and Dark, 1998). Fire consistently seems to have been part of the regional environment and had the effect of opening the woodland and creating transitional successional vegetation, often promoting particular taxa favoured by fire, such as *Pinus*, *Pteridium* and *Calluna* (Simmons and Innes, 1987). Charcoal is often recorded at the level of major pollen zone changes, as though fire had changed forest composition and given new taxa the opportunity to become established. Charcoal often accompanies the rise of *Alnus* pollen, for example, as at Walker's Heath in Cheshire (Leah *et al.*, 1997), Hoscar Moss, Lancashire (Cundill, 1981), Seamer Carr in the Vale of Pickering (Cloutman and Smith, 1988) or Malham Tarn Moss in the Pennines (Piggott and Piggott, 1963). Fire also has been suggested as encouraging the early Flandrian *Corylus* rise, as at Flixton in the Vale of Pickering (Walker and Godwin, 1954). Natural fire will have occurred, but it may well have been used deliberately by humans to change the vegetation, as has been suggested in particular for Mesolithic people (Jacobi *et al.*, 1976; Simmons and Innes, 1987), as well as for a range of other purposes. The primary association of charcoal layers with Mesolithic flints and activity at Star Carr, Dunford Bridge and many other sites suggests human agency, and at Bonfield Gill Head (Simmons and Innes, 1981, 1988b) and a number of other upland sites, microscopic charcoal is common in the Mesolithic age peat but not in the Neolithic age peat above, suggesting a correlation with the hunter–gatherer mode of land use. Charcoal peaks also often correlate with phases of intensified human activity in all cultural periods, such as at Lindow Moss in Cheshire in the Iron Age (Oldfield *et al.*, 1986) and at several sites during the Bronze Age and Iron Age in the North York Moors (Dimbleby, 1962; Simmons *et al.*, 1993), implicating humans in the use of fire.

In many cases the *Ulmus* Decline in northern England is accompanied by indications of forest opening and although some such as Hoscar Moss in Lancashire (Cundill, 1981) and Barfield Tarn in Cumbria (Pennington, 1975c), are associated with charcoal, many are manifest solely in the pollen stratigraphy as reductions in tree pollen and increases in open ground or agricultural indicator taxa, such as *Plantago lanceolata* and cereal-type pollen. Many disturbance phases without charcoal evidence did occur during the Flandrian prior to the *Ulmus* Decline, as at Quick Moss in the Pennines (Rowell and Turner, 1985) and at sites in the Cheviot Hills (Tipping, 1996), perhaps caused by Mesolithic activity. With or without charcoal, the frequency of disturbance seems to increase in the period leading up to the *Ulmus* Decline, often at the same site. Sturludottir and Turner (1985) have suggested that repetitive forest disturbance, especially using fire, may have caused the decline of elm at Pawlaw Mire and elsewhere in the Pennines as a result of soil degeneration, a process leading to spread of blanket bog and moorland in the uplands (Simmons and Innes, 1987). A few of the forest openings in the millennium before the *Ulmus* Decline in northern England contain cereal-type pollen, as at Little Hawes Water in north Lancashire (Taylor *et al.*, 1994) and Lismore Fields in Derbyshire (Wiltshire and Edwards, 1993), which may be the first evidence of the presence of at least partly agricultural land use in the region. Early dates of 5820 ± 95 years BP (Williams, 1985) at Soyland Moor in the Pennines and 5840 ± 70 years BP (Cowell and Innes, 1994) at Bidston Moss in Merseyside suggest that an incipient Neolithic style economy existed for several centuries before the *Ulmus* Decline, as well as several centuries after it. The *Ulmus* Decline therefore lies firmly within the Neolithic cultural period and itself may be a symptom of human activity, although disease and climate may be implicated. As suggested by Pennington (1970) and Tipping (1994a) for the Cumbrian plain, the *Ulmus* Decline is probably multi-causal and the result of local events.

The five millennia after the elm decline in northern England are characterized by increasing intensity of human agricultural land use, forest clearance and the spread of grassland, heath and bog. Regional variations in the timing, character and intensity of land use are apparent, but a progressively more open and intensively used landscape is a common theme, changing the nature and rate of depositional regimes. The sedimentary signature of this later Flandrian anthropogenic phase is increasingly severe episodic soil erosion and alluviation, from the Cheviot Hills (Mercer and Tipping, 1994) to the Cheshire–Shropshire plain (Twigger and Haslam, 1991), from Humberside (Gilbertson, 1984a; Buckland and Sadler, 1985) and the Yorkshire Wolds (Bush, 1993) to the Lake District (Pennington, 1970). A few areas record heavy woodland clearance in the Neolithic, as at Shibdon Pond in the Tyne valley (Passmore *et al.*, 1992), where it began after 4800 ± 80 years BP and resulted in major alluvial deposition. Major Neolithic clearances also occurred in north Northumberland at this time (Tipping, 1992). At 4543 ± 70 years BP at Mordon Carr in south-east Durham (Bartley *et al.*, 1976) there was major woodland clearance, with cereal pollen found and Tooley (1978b) records similar large-scale clearance at nearby Hartlepool Bay. In contrast in north Durham at Hallowell Moss (Donaldson and Turner, 1977) and at upland Northumbrian sites (Davies and Turner, 1979; Rowell and Turner, 1985) there is no evidence of Neolithic clearance. At Skipsea Withow Mere in Holderness major deforestation with cereal pollen found, occurs at the *Ulmus* Decline and then again until major colluviation after 4500 ± 50 years BP buries the profile (Blackham and Flenley, 1984). Similar major clearance and soil erosion occurred at Prescot Moss, Merseyside between 4650 ± 80 and 4520 ± 140 years BP (Cowell and Innes, 1994), and a marked clearance phase at Hatchmere in Cheshire persists until 4693 ± 90 years BP (Switsur and West, 1975). Later Neolithic people appear to have used fire to clear the upland woods in Cumbria (Walker, 1965; Pennington, 1970) and sites such as Williamson's Moss on the Cumbrian coastal plain record substantial woodland clearance for several centuries after the *Ulmus* Decline (Pennington, 1975c). Sites in the southern Pennines show little evidence for Neolithic activity in the pollen record, with the first clearances at Leash Fen (Hicks, 1971), Rishworth Moor (Bartley, 1975), Deep Clough (Tallis and McGuire, 1972) and Fountains Earth (Tinsley, 1975), for example, delayed until after *c.* 4000 years BP. An exception occurs around Eshton Tarn in the Craven district of West Yorkshire, where major woodland clearance with cereal pollen found, is dated to *c.* 4500 years BP (Bartley *et al.* 1990), although most sites in that area record only sporadic and short-lived low intensity clearance. Honeyman (1985) did, however, report significant clearance in Wensleydale dated 4550 ± 50 years BP. Neolithic age woodland clearance on the North York Moors was also very low scale and sporadic (Simmons *et al.*, 1993). There is good macrofossil and pollen evidence for an expansion of tree cover in most of the uplands during the Neolithic (Tallis, 1975; Simmons *et al.*, 1993) and a rise in the height of the tree line. Despite some exceptions, such as Skipsea Withow Mere, coastal west Cumbria and some locations in the Mersey valley (Birks, 1965b), almost all lowland sites on both sides of the Pennines record only sporadic and limited Neolithic impact on woodland (Twigger and Haslam, 1991; Middleton *et al.*, 1995; Dinnin, 1995; Leah *et al.*, 1997; Gearey and Lillie, 1999). Neolithic impact in the north of England was mainly of a minor nature and significant forest clearance was probably very localized and poorly represented on most pollen diagrams. Some real diversity does exist in the Neolithic environmental record, however, a feature persisting through the late Flandrian.

The great majority of sites in all parts of northern England record major increases in the intensity of forest clearance between about 4000 and 2800 years BP, corresponding to the Bronze Age occupation of the region. In most locations this activity represents the first major reduction in forest cover, and the greatly increased incidence of cereal-type pollen points to the expansion of a mixed farming economy at this time. Individual site histories are too numerous to catalogue, but representative profiles may be cited from various areas. Of particular interest in the mid-fourth millennium BP is a major decline in *Tilia* percentages, in some cases to virtual absence (Beckett, 1981), which is diachronous and often coincides with peak cereal, *Plantago lanceolata* and other agricultural indicator pollen records, supporting Turner's (1962, 1965) recognition of Bronze Age farming as the cause. Turner (1964) dated the *Tilia* decline at Whixall Moss, Shropshire to 3238 ± 115 years BP and a cluster of similar dates are available from that area. Beales (1980) dated the feature to 3714 ± 129 years BP at Crose Mere and mosses and meres in the Baschurch area of Shropshire have provided dates of 3660 ± 50, 3550 ± 50 and 3190 ± 60 years BP (Barber and Twigger, 1987). The last date is similar to that at Top Moss (Leah *et al.*, 1997) of 3220 ± 50 years BP, which was preceded by a substantial phase of clearance at 3800 ± 55 years BP. Local land-use variations caused the event to occur over several centuries in this small area. In south Yorkshire (Dinnin, 1997a), forest clearance episodes, where no cereal pollen has been found, on Hatfield and Thorne Moors have been dated to 3715 ± 70, 3685 ± 65, 3570 ± 70 and 3545 ± 70 years BP, the last two associated with charcoal, and burned and chopped tree stumps. Major clearance occurred in the lowlands of the north-east. *Tilia* declines sharply within a major clearance episode at Bishop Middleham in Durham at 3660 ± 80 years BP, followed by an extensive second clearance at 3360 ± 80 years BP, which left the local fertile east Durham limestone soils virtually deforested (Bartley *et al.*, 1976). Analogous events took place at nearby Hutton Henry at 3544 ± 80 years BP and Neasham at 3242 ± 70 years BP, and *Tilia* almost disappeared at Hallowell Moss near Durham City (Donaldson and Turner, 1977). The same process occurred in Northumberland at all altitudes. Davies and Turner (1979) record major clearance at Fellend Moss at 3688 ± 60 years BP, at Steng Moss at 3594 ± 45 years BP and 3015 ± 45 years BP and at Camp Hill Moss between 3510 ± 70 and 3110 ± 80 years BP. Tipping (1996) reported major woodland clearance, with cereal pollen found, throughout the Bronze Age levels at Swindon Hill in the Cheviot Hills well before 3100 ± 50 years BP. In this area Bronze Age clearance resulted in erosion and alluvial deposition (Tipping, 1992), as at Callaly Moor in mid-Northumberland between 3920 ± 70 and 2540 ± 110 years BP (Macklin *et al.*, 1991). On the North York Moors major upland clearance took place in this period (Simmons *et al.*, 1993) as shown by soil pollen analyses from beneath Bronze Age monuments (Dimbleby, 1962). Cereals were common and extensive deforestation occurred, which converted much of the upland to heather moor and bog permanently. Good dated examples are 3400 ± 90 years BP at Fen Bogs (Atherden, 1976a, b) and 3210 ± 90 years BP at Wheeldale Gill (Simmons and Cundill, 1974a). After 3970 ± 80 years BP at Willow Garth on the Yorkshire Wolds a low amount of *Tilia* falls to zero at a point where cereal pollen first appears and *Plantago lanceolata* peaks (Bush, 1993), suggesting mainly pastoral human activity and spread of grassland in this already poorly wooded area. In the Pennine uplands almost all sites experience their first significant deforestation in the Bronze Age. Major clearances occur at sites in Wensleydale at 3850 ± 50 and 3930 ± 50 years BP (Honeyman, 1985). The first major deforestation at Eshton Tarn (Bartley *et al.*, 1990) occurs at 3600 ± 100 years BP, followed by almost complete woodland removal and high values for cereals at 3160 ± 80 years BP. Bartley (1975) dated the first major clearance in the uplands around Rishworth at 4010 ± 100 years BP, a date similar to those reported by Hicks (1971) for Leash Fen in Derbyshire, who reported a *Tilia* decline and increased clearance between 3740 ± 100 and 3450 ± 110 years BP, Tinsley (1976) who noted major woodland recession at Skell Moor in Nidderdale at 3880 ± 100 years BP, and Tallis and McGuire (1972) at Deep Clough in the

Rossendales, with a date of 3540 ± 120 years BP. The same process occurred throughout the Pennines. In the lowlands to the west at Fenton Cottage in Lancashire records show a large increase in charcoal and *Calluna* pollen during the Bronze Age between 3790 ± 100 and 3180 ± 60 years BP (Middleton *et al*., 1995; Wells *et al*., 1997), the earlier date also marking the deposition of the Hekla 4 tephra at that site, the only tephra layer recovered from northern England to date. In Merseyside (Cowell and Innes, 1994) the first substantial clearances in the central mossland region, although without cereal pollen being found, also occur at this time. Bronze Age clearance is also common in Cumbria (Wimble *et al*., 2000), as at Ennerdale Water before 2996 ± 55 years BP (Pennington 1970, 1975c) and with a series of clearances at Rusland Moss before 2686 ± 50 years BP (Dickinson, 1975). The first significant clearances in the Solway lowlands, with substantial occurrence of cereals indicated by the pollen record, took place in the mid-Bronze Age (Walker, 1966b). In most of northern England the Bronze Age was a time of reduced forest cover and increased human activity. Only in areas with very heavy clay soils and lowland wetland environments, such as the Vale of York, might forest clearance have been very limited. For example, at St George's Field, York (Lillie and Gearey, 1999) environmental evidence indicates mature undisturbed woodland and alluvial wetlands between 3240 ± 70 and 2760 ± 65 years BP in the late Bronze Age.

The impact of Iron Age and Romano-British land use on the woodland in northern England was regionally variable, but in general the earlier part of the period saw moderate-scale episodes of clearance, whereas the later Iron Age and Romano-British period was a time of greatly increased deforestation, in places on a landscape scale (Turner, 1979). Soil quality determined the degree to which arable cultivation figured in local land use. Extensive deforestation took place in the east Durham lowlands, for example, with trees replaced by grassland almost completely at Hallowell Moss between 1956 ± 70 and 1355 ± 50 years BP (Donaldson and Turner, 1977). At Thorpe Bulmer and Hutton Henry (Bartley *et al*., 1976) intensive deforestation occurred at 2064 ± 60 and 1842 ± 70 years BP respectively, the former associated with high values of cereal pollen and a peak of *Cannabis* pollen at 1730 ± 120 years BP, whereas at Bishop Middleham on very fertile soils deforestation had been completed by the start of the Iron Age. In south Northumberland the land was still widely forested through the pre-Roman Iron Age, but became deforested during the Roman period, as at Fellend Moss where clearance took place at 1948 ± 45 years BP and Fozy Moss where massive and rapid clearance occurred from 1820 ± 45 years BP (Turner, 1979; Dumayne and Barber, 1994). Similar major tree removal is recorded in the Northumbrian hills at the same time, at Steng Moss (Davies and Turner, 1979) from 1970 ± 60 years BP, at Bollihope Bog and Steward Shield Meadow in Weardale from 1730 ± 100 and 2060 ± 120 years BP respectively (Roberts *et al*., 1973) and at Quick Moss (Rowell and Turner, 1985) from 2035 ± 50 years BP. The landscape of the Cheviot Hills also was almost cleared of trees during this period and several radiocarbon dates place phases of alluvial aggradation at Powburn (Tipping, 1992) to this time. Major woodland clearance with cereal pollen found preceded major alluviation in the lower Tyne valley after *c*. 2590 years BP (Passmore *et al*., 1992). In all cases the clearance is prolonged as well as massive, lasting through the Roman period and after. A similar record occurs both in north Cumbria (Dumayne-Peaty and Barber, 1998), at Bolton Fell Moss at 1860 ± 60 years BP and Walton Moss with dates of 2000 ± 45 and 1925 ± 40 years BP, and in south Cumbria (Dickinson, 1975) at Rusland Moss, where a cleared landscape with cereals present persisted from 1963 ± 50 to 1361 ± 55 years BP. Supporting data from Helsington Moss (Smith, 1958c) at 1514 ± 100 years BP, Burnmoor Tarn at 1460 ± 130 years BP and Devoke Water after 1750 ± 130 years BP (Pennington, 1970) confirm that widespread clearance and cereal cultivation spread into all parts of Cumbria in the Romano-British period. Massive forest clearance took place on the North York Moors in Iron Age to Roman times, some with high cereal pollen values found, others apparently mainly for pasture (Simmons *et al*., 1993). The best dated examples are Fen Bogs between 2280 ± 120 and 1530 ± 130 years BP (Atherden, 1976a, b) and after 2190 ± 90 years BP at Harwood Dale Bog (Atherden, 1989). Evidence from the Pennines is also clear, and Tallis and Switsur (1973) dated very substantial clearance and the presence of some cereal pollen between 2251 ± 50 and 1400 ± 50 years BP at Featherbed Moss in the southern Pennines, which corresponds well with more limited clearance at Leash Fen (Hicks, 1971) between 2120 ± 100 and 1500 ± 110 years BP. Tinsley (1976) reported a major late Iron Age and Roman forest clearance with high cereal pollen values found dated after 2200 ± 80 years BP in Nidderdale, north Yorkshire Pennines. Bartley and Chambers (1992) recorded similar major woodland clearance between 2260 ± 100 and 1730 ± 75 years BP at Extwistle Moor, Lancashire, as did Mackay and Tallis (1994) between 2025 ± 40 and 1735 ± 45 years BP in the Forest of Bowland, whereas Cundill (1976) reports major clearance at Carlingill in the Howgill Fells as occurring after 2290 ± 80 years BP. The same pattern of low-scale activity in the earlier Iron Age and then major deforestation in Roman times recurs throughout the Pennines, and is also the case in the lowlands of the southern part of the region. Representative examples are Fenton Cottage between 1940 ± 110 and 1590 ± 50 years BP and Winmarleigh Moss at 1680 ± 80 years BP in Lancashire (Middleton *et al.*, 1995), Knowsley Park around 1680 ± 50 years BP and Simonswood Moss after 2380 ± 80 years BP in Merseyside (Cowell and Innes, 1994), after 2090 ± 70 years BP at Rostherne Mere (Leah *et al.*, 1997) and after *c.* 2240 years BP at Lindow Moss (Oldfield *et al.*, 1986) in Cheshire, and after 2086 ± 75 years BP at Crose Mere (Beales, 1980) and around 2195 ± 50 BP at Top Moss (Leah *et al.*, 1997) in Shropshire, all with a consistent pollen curve showing cereal cultivation, intensified agriculture and woodland recession. Oldfield *et al.* (1985) dated major clearance and soil erosion at Peckforton Mere, Cheshire to Romano-British times by mineral magnetic analyses. Regional deforestation for arable agriculture also is a feature of the south Yorkshire lowlands, with a very open landscape indeed, causing major soil erosion and alluviation (Buckland and Sadler, 1985). This deforestation occurred around Thorne and Hatfield Moors after dated levels of 2085 ± 70, 2225 ± 70 and 2145 ± 65 years BP (Smith, 1985; Dinnin, 1997a). Bush (1993) also records a major switch to arable cultivation, indicated by high values of cereal pollen on the Yorkshire Wolds at Willow Garth after 2120 ± 50 years BP, replacing grassland and the little remaining woodland, and Kenward *et al.* (1978) report a sharp decline in oak and hazel and rise in grass and cereal pollen after 2010 ± 90 years BP at Askham Bog near York. Although a few areas, as around Baschurch in Shropshire (Twigger and Haslam, 1991), show some regeneration of woodland in Roman times, there is a similar pattern of major deforestation in late Iron Age and Romano-British times throughout northern England, partly perhaps for the timber itself as much as for farming land (Atherden, 1976a; Dumayne and Barber, 1994).

Fewer sediments of post-Roman age have survived than those of earlier periods and so a systematic view of landscape history in northern England for the past 1500 years is more difficult to attain. Several sites do preserve evidence of medieval and later land use, and these indicate continued forest clearance in all parts of the region with local interludes of regeneration. Woodland regeneration after the Roman period occurs in many places, but persistence of clearance also occurred, as at Baggy Moor in Shropshire (Brown, 1990), dated after 1375 ± 40 years BP but before 1190 ± 50 years BP. Supporting evidence comes from The King's Pool, Stafford (Bartley and Morgan, 1990), where the main arable phase with high values for cereal pollen is after 1370 ± 70 years BP. Beales (1980) and Barber and Twigger (1987) also suggest intensified agriculture in Shropshire from Anglian times onwards, but with an emphasis on pastoralism. At Winmarleigh Moss in west Lancashire (Middleton *et al.*, 1995) the substantial clearance that began in late Roman times at 1680 ± 80 years BP persisted until 900 ± 90 years BP, with increasing cereal pollen values found. At nearby Fenton Cottage a gradual increase in clearance pressure from 1200 ± 70 to 390 ± 50 years BP left the area under mixed agriculture and almost completely without tree cover in the medieval period (Middleton *et al.*, 1995). On Thorne and Hatfield Moors in south Yorkshire, Smith (1985) reported regeneration of woodland in the post-Roman period until after dates of 865 ± 60 years BP and 910 ± 65 years BP, when there was a great expansion of mixed farming and an almost treeless landscape (Dinnin, 1997a). On the Yorkshire Wolds at Willow Garth (Bush, 1993) high cereal values and arable indicators in the pollen record from 1170 ± 50 years BP until modern times shows the intensively farmed nature of the area. At Askham Bog in the Vale of York (Kenward, *et al.*, 1978) peak values for cereal and *Cannabis* pollen occur before 470 ± 80 years BP, and so are medieval in age. In north-east England several dates averaging 1400 years BP are available, from sites such as Steng Moss and Fellend Moss (Davies and Turner, 1979), Quick Moss (Rowell and Turner, 1985), Hallowell Moss (Donaldson and Turner, 1977) and Fen Bogs (Atherden, 1976b; Chiverrell and Atherden, 1999), that indicate a region-wide regeneration of woodland after the Roman period. Exceptions do occur, as at Thorpe Bulmer (Bartley *et al.*, 1976) on the east Durham limestone, where tree recovery did not occur until 852 ± 60 years BP and at Steward Shield Meadow in Weardale (Roberts *et al.*, 1973) until 840 ± 100 years BP. The first major clearance on the heavy soils in the Tees valley did not occur until 1213 ± 60 years BP at Neasham Fen (Bartley *et al.*, 1976). Most sites record massive clearance in the medieval period, at Fellend Moss from 945 ± 40 years BP onwards, at Camp Hill Moss in the Cheviot Hills (Davies and Turner, 1979) from 640 ± 80 years BP onwards and on the North York Moors at Fen Bogs from 1060 ± 160 years BP until 390 ± 100 years BP. Pennington (1970) records major upland woodland clearance in Cumbria in Norse and later times, whereas in the north Cumbrian lowlands and Tyne valley Dumayne and Barber (1994) record sustained clearance for mixed farming at Glasson Moss after 960 ± 40 years BP and at Fozy Moss after 925 ± 45 years BP. In the Lancashire Pennines Mackay and Tallis (1994) record consistently high values of cereal and grassland pollen after 840 ± 45 years BP, with major expansion and sharp decline in *Alnus* and *Corylus* values before a ^{210}Pb date of AD 1847. A similar but less pronounced pattern occurs at Extwistle Moor, Lancashire after 905 ± 70 years BP (Bartley and Chambers, 1992). Tinsley (1975, 1976) in Nidderdale in the North Yorkshire Pennines recorded clearances for grassland pasture but with some cereals in phases after 1050 ± 80 and 480 ± 80 years BP. Clearance on the upland limestone soils of west Yorkshire was earlier, before 1190 ± 40 years BP (Bartley *et al.*, 1990). The evidence from Featherbed Moss in the southern Pennines (Tallis and Switsur, 1973) is again for massive clearance, with high values of cereal and pastoral indicator pollen, and a very open medieval landscape from 1023 ± 50 years BP continuing beyond 491 ± 50 years BP. Many other sites in northern England, although undated, show similar massive deforestation in their near-surface pollen levels, which probably is of medieval and modern age, resulting in the present extremely open landscape.

KEY TO THE STRATIGRAPHICAL SYMBOLS USED IN THE POLLEN DIAGRAMS

The key below (Figure 8.1) is adapted from Troels-Smith (1955) and is used in most of the pollen diagrams in this chapter .

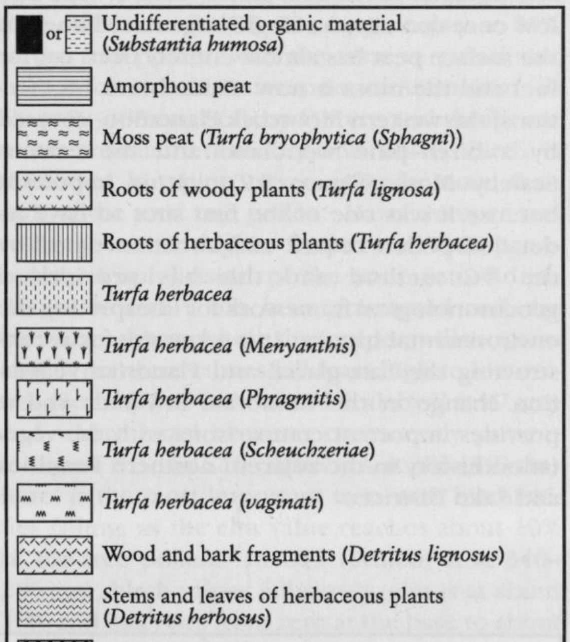


Figure 8.1 Key to the stratigraphical symbols (modified after Troels-Smith, 1955).

SCALEBY MOSS (NY 430 635)

D. Huddart

Introduction

Scaleby Moss, Cumbria, is an ombrotrophic raised bog in an enclosed hollow in the Scottish Readvance till, which developed in a lake that was initiated in the Late-glacial period. The hollow once contained a single large raised bog but the surface peat has almost entirely been cut for fuel and the moss is now divided into two sections, the western Moorcock Plantation, covered by a birch–pine woodland and the eastern Scaleby Moss (Figure 8.2). It is important because it was one of the first sites to have its detailed pollen record and zonation dated by the ^{14}C method and this has provided a geochronological framework for interpreting the environmental history. It is a key site for reconstructing the Late-glacial and Flandrian vegetation change in the Cumbrian lowlands and it provides important comparisons with the vegetation history in the adjacent northern Pennines and Lake District.

Description

The stratigraphy shown from borings and from cut faces records a hydrosere beginning with an open lake and progressing to a *Sphagnum*-dominated, raised bog (Godwin *et al*., 1957). Immediately above the basal till is a thin and intermittent, grey, sandy clay (*c*. 2 cm thick) succeeded by a buff, organic mud, which becomes sandy and contains occasional pebbles near the top. In the middle of the bog this layer is *c*. 60 cm thick. Above, the mud becomes darker and silt-free but contains increasing quantities of *Sphagnum cuspidatum* with grass and sedge stems, fruits of Carex species and seeds of bog bean. The mud changes vertically into a dark-brown, highly humified *Sphagnum* peat, containing locally varying percentages of *Trichophorum caespitosum*, *Eriophorum vaginatum*, *Eriophorum angustifolium* and *Calluna vulgaris*. The maximum depth of *Sphagnum*-rich mud and humified peat is *c*. 5 m. About 1 m below the original bog surface, the humified peat is superseded by a fresher, *Sphagnum* peat and the transition usually is marked by a thin (2–5 cm) of pool peat. The upper peat is not

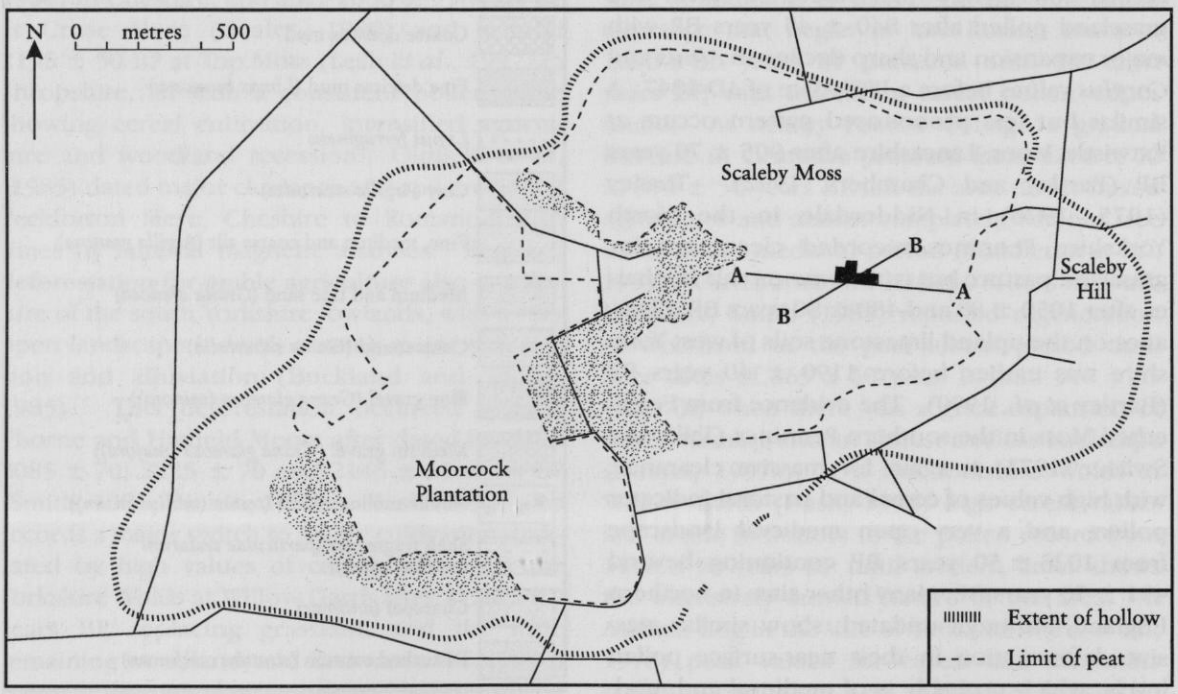


Figure 8.2 Scaleby Moss, showing its position in a hollow, together with the position of Moorcock Plantation, the series of cores (A–A, B–B) and the remnant of uncut peat (after Walker, 1966b).

homogeneous and contains a number of laterally intermittent, recurrence surfaces.

A coring in 1950 near the bog centre resulted in a pollen diagram (Scaleby Moss A) shown in Figure 8.3 from the Late-glacial to the recent past (Godwin *et al*., 1957). In 1955 a pit was dug and a monolith taken close to the core location for laboratory examination and pollen analysis. This resulted in Scaleby Moss B and C pollen diagrams (Figures 8.4 and 8.5), which were zoned (Godwin *et al*., 1957; Walker, 1966b). After each pollen analytical horizon had been identified in the monolith, samples were taken for ^{14}C dating. The dates are shown in Table 8.1 and detailed stratigraphical descriptions are given in Walker (1966b).

Interpretation

The pollen record

It appeared that the results of the dating were trustworthy as the sample ages corresponded to the depositional sequence and the boundary dates were compared with dates in The Netherlands and Denmark (Godwin *et al*., 1957). The earliest pollen zone at Scaleby Moss (C5) corresponds to British Zone 11 and was represented between 127 and 122 cm at Scaleby B, where dry-land herb pollen dominated over trees and shrubs. Birch pollen was almost totally dominant in the trees. *Artemisia* and *Rumex* values fall through the zone and Cyperaceae are much more abundant than Graminae. In C6 (122–112 cm) dry-land herb pollen was dominant over trees and shrubs. Pine was more frequent than previously and *Artemisia* and *Rumex* values were low. *Empetrum* attained a maximum at the top of the zone. Juniper is consistently present though not important and *Betula nana* is present throughout. In C7 (112–95 cm) tree and shrub pollen account for only 10% or under of the total pollen. Willow was more important than earlier. In C8 (95–72 cm) pine rises to 20% of the tree pollen, with the zone very uniform, except for a central maximum of *Artemisia*. In C9 (470–385 cm, Scaleby A) trees and shrubs are more abundant than dry-land herbs, birch is dominant but pine never exceeds more than 10% of the total tree pollen. Hazel, juniper and willow occur in higher frequencies than previously. A subdivision of the zone into British Zones IV and V has been attempted and the boundary drawn where the juniper, willow, grass and Cyperaceae curves fall from high to low values (Scaleby A at 440 cm and Scaleby B at 47 cm) and where hazel becomes consistently present in Figure 8.3. In C10 (Scaleby A at 385–340 cm) hazel is the most important tree, with birch values falling, as the elm value reaches about 10% of the tree pollen. In C11 (Scaleby A at 340–285 cm), birch values fall slowly, elm is at about 10% and oak rises from zero at the base to about 25% at the top.

Zone C12 (Scaleby at A 285–250 cm) coincides with the 'Boreal–Atlantic Transition' and the

Table 8.1 Radiocarbon dated pollen zone horizons at Scaleby Moss (after Godwin *et al*., 1957)

Sample number	Depth related to pollen diagram B or C (cm)	Pollen zonation	Age (years BP)
Q172	67.0–69.0 B	Zone VII*b* base	5030 ± 119
Q171	69.0–71.0 B	VII*a*/VII*b* boundary (Atlantic Sub-boreal/transition)	4975 ± 134
Q173	71.0–73.0 B	Zone VII*a* top	5037 ± 122
Q166	174.5–176.5 B	Zone VII*a* base	6998 ± 131
Q165	176.5–178.5 B	VI/VII*a* boundary (Boreal/Atlantic transition)	7475 ± *c*. 350
Q167	178.5–180.5 B	Zone VI top	7404 ± 146
Q161	–0.5–1.5 C	Zone VI base (V/VI boundary)	9052 ± 194
Q162	3.5–5.5 C	Zone V top	8859 ± 192
Q155	44.5–46.5 C	Zone V base	9790 ± 183
Q154	46.5–48.5 C	IV/V boundary (Pre-boreal/Boreal transition)	9607 ± 209
Q152	69.5–71.5 C	Zone IV base	10 203 ± 193
Q151	71.5–73.5 C	III/IV boundary (Post-glacial/Late-glacial transition)	10 307 ± *c*. 350
Q153	73.5–75.5 C	Zone III top	10 368 ± 215
Q144	109.5–111.5 C	Zone III base	10 878 ± 185
Q147	123.0–125.0 C	Zone II top } Combined	10 748 ± 207
Q148	125.0–127.0 C	Zone II top } Combined	

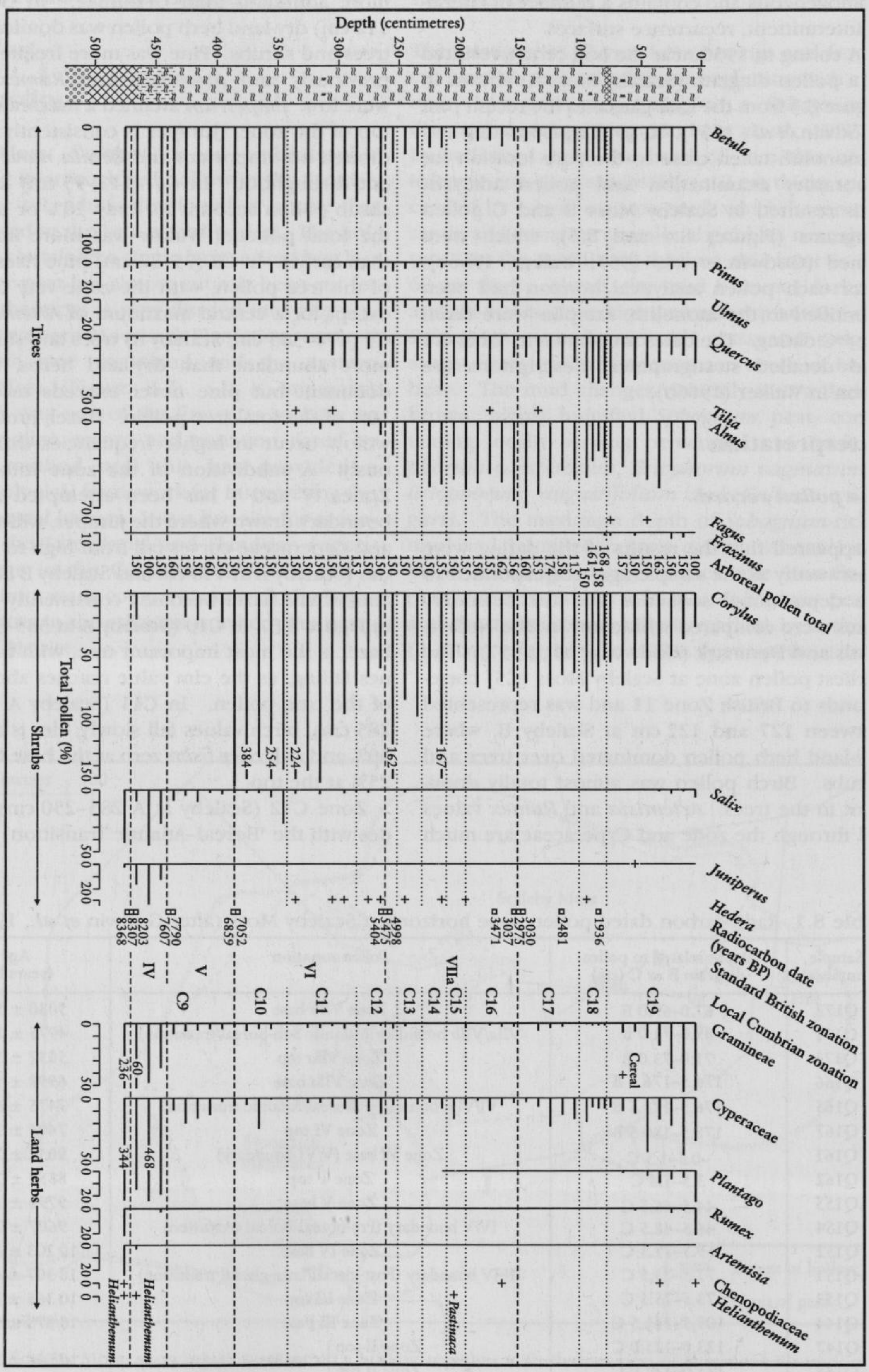


Figure 8.3 Scaleby Moss A pollen diagram through the post-Glacial deposits (after Walker, 1966b). The frequencies of pollen or spores of all taxa are shown as percentages of the arboreal pollen total of the appropriate sample. See Figure 8.1 for key to the stratigraphical log. (Figure continued on opposite page.)

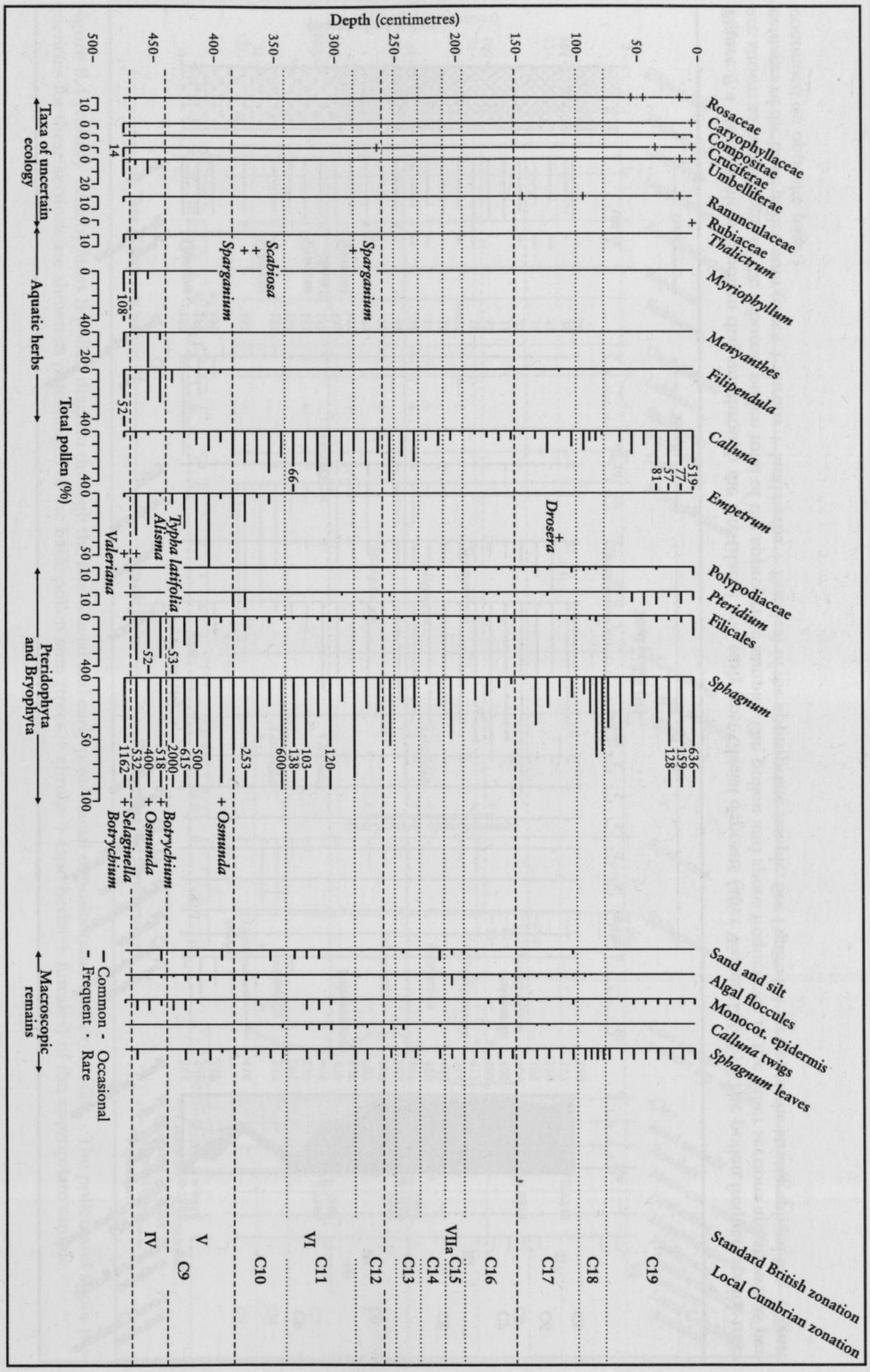


Figure 8.3 *(contd)* Scaleby Moss A pollen diagram through the post-Glacial deposits (after Walker, 1966b).

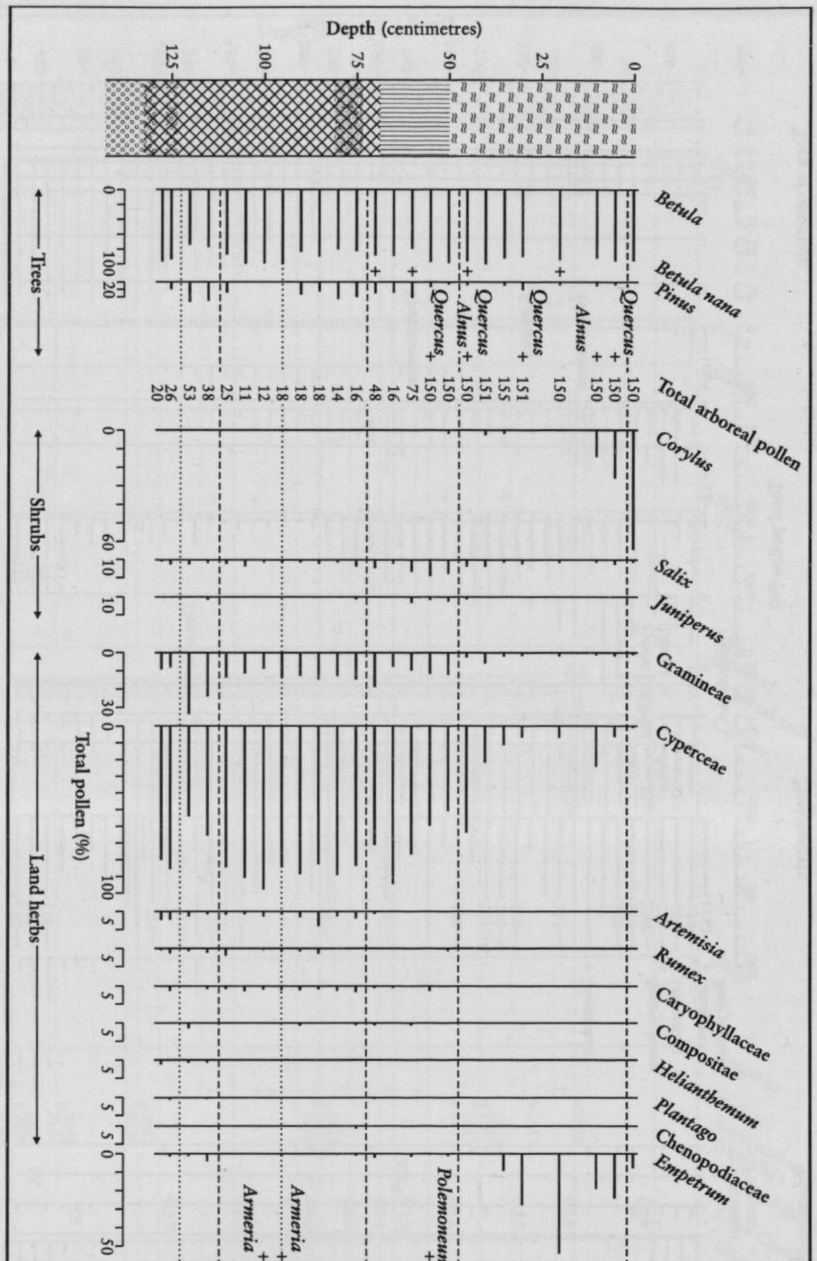


Figure 8.4 Scaleby Moss B pollen diagram through the late-Glacial and early post-Glacial deposits (after Walker, 1966b). The pollen frequencies for trees are shown as percentages of the arboreal pollen total of the appropriate sample. The pollen and spore frequencies in the other sections are shown as percentages of the basic pollen sum (trees + shrubs + land herbs + Ericales) of the appropriate sample. See Figure 8.1 for key to the stratigraphical log. (Figure continued on opposite page.)

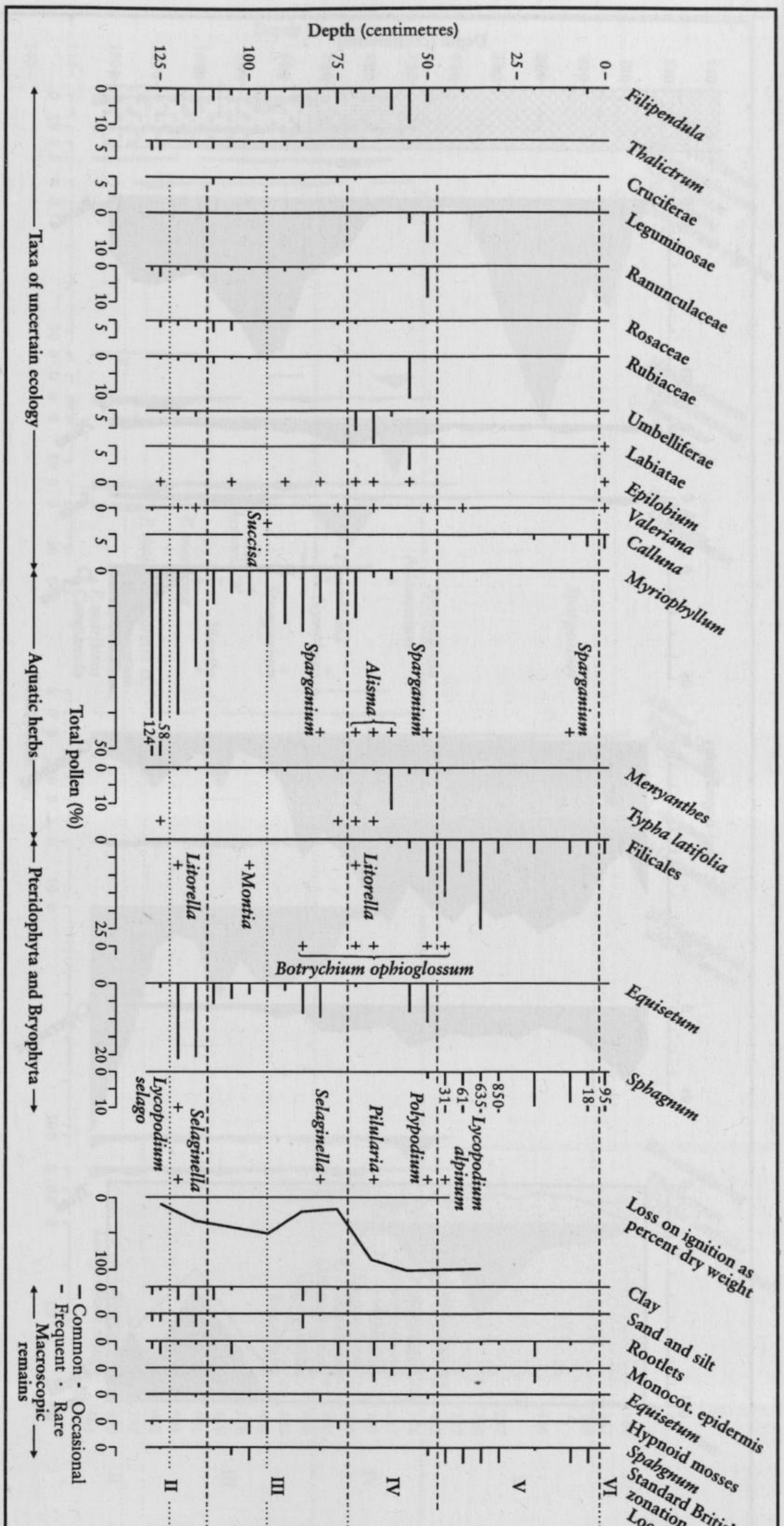


Figure 8.4 (*contd*) Scaleby Moss B pollen diagram through the late-Glacial and early post-Glacial deposits (after Walker, 1966b). The pollen and spore frequencies for these sections are shown as percentages of the basic pollen sum (trees + shrubs + land herbs + Ericales) of the appropriate sample.

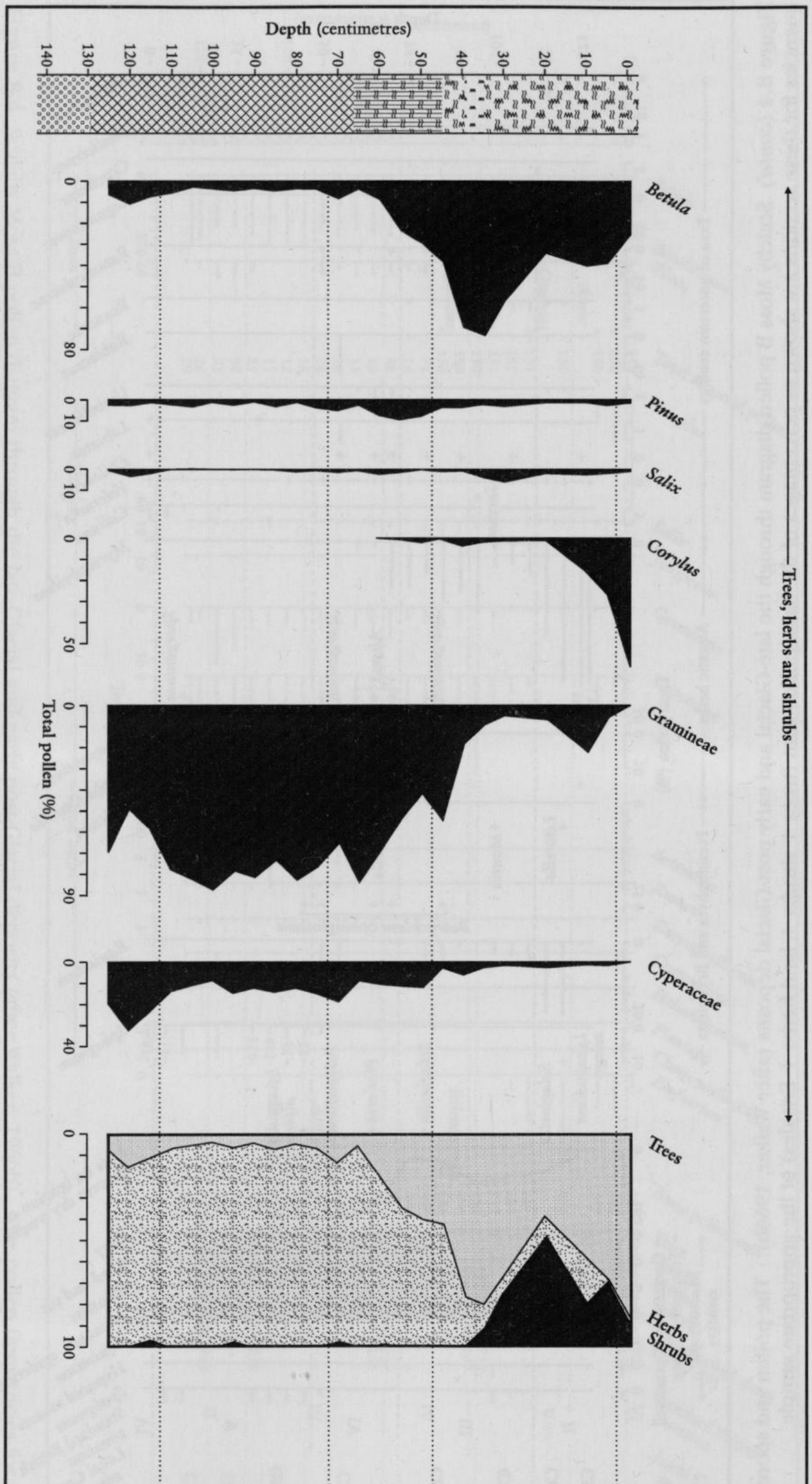


Figure 8.5 Scaleby Moss C pollen diagram from samples at 5 cm intervals through the bottom 1.25 m of the peat monolith (after Godwin *et al.*, 1957; Walker, 1966b). All pollen frequencies plotted as percentages of total pollen of dry-land plants excluding ferns. The main pollen-zones have been identified from the changes in vegetation. See Figure 8.1 for key to the stratigraphical log. (Figure continued on opposite page.)

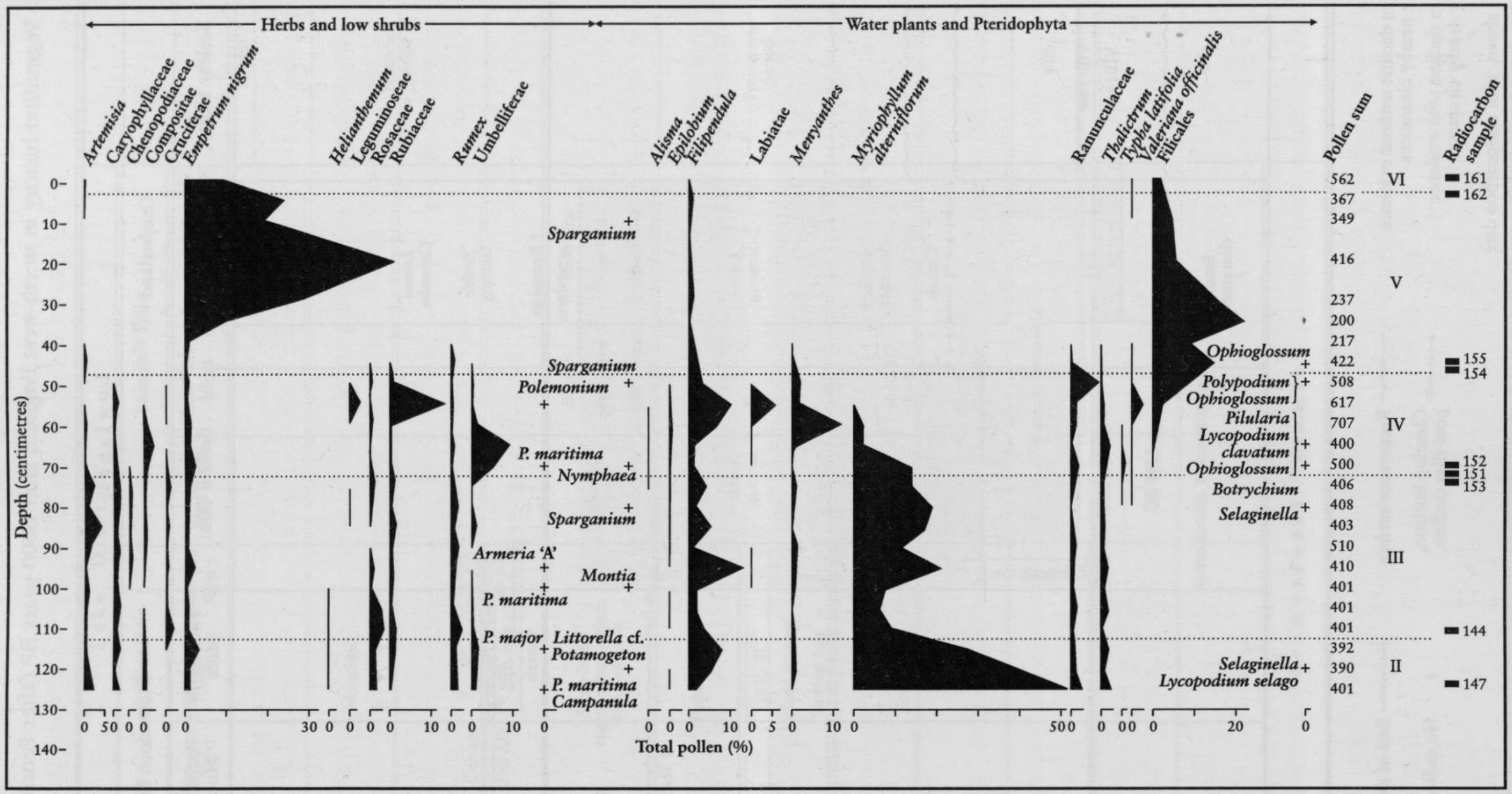
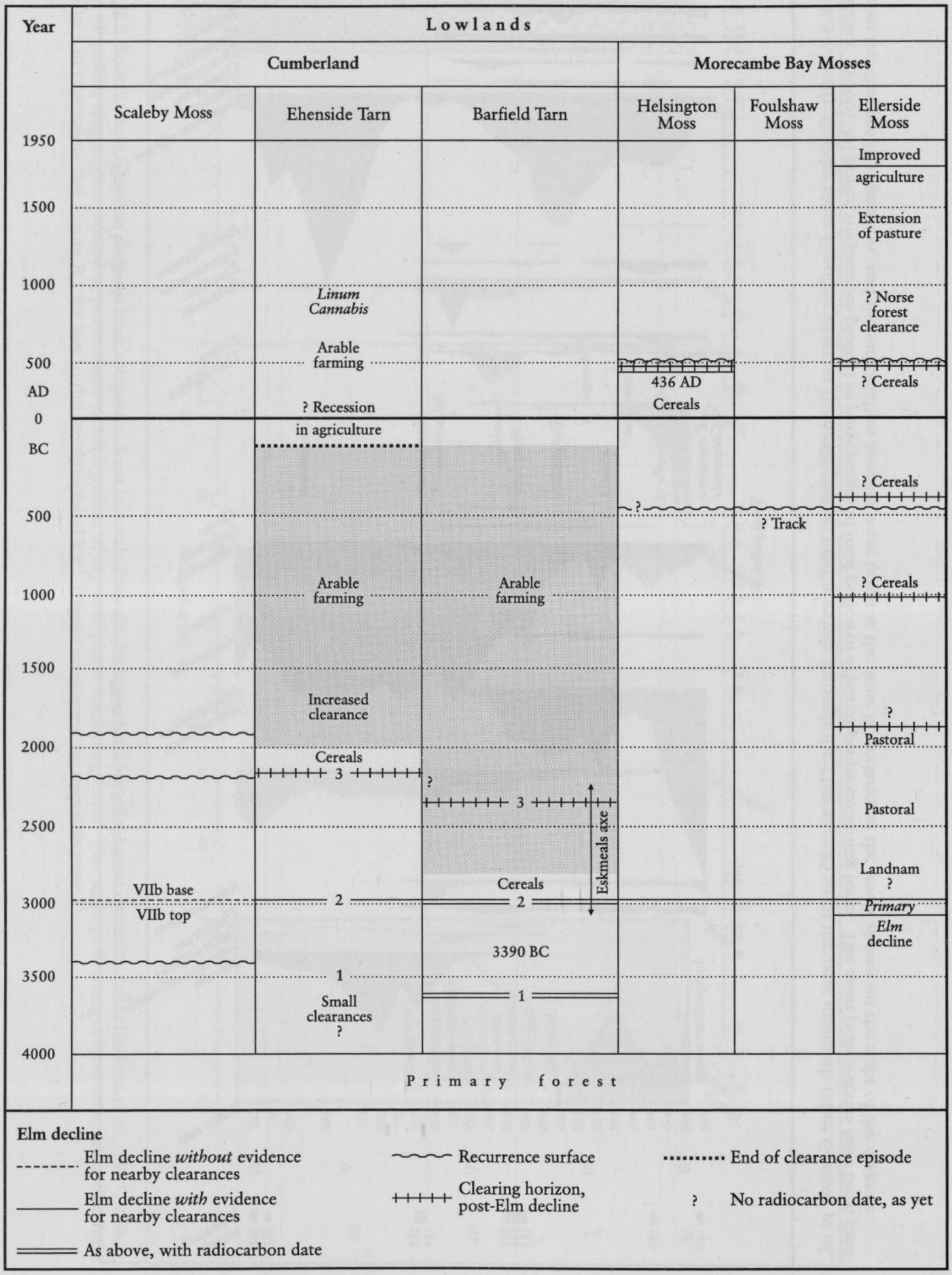
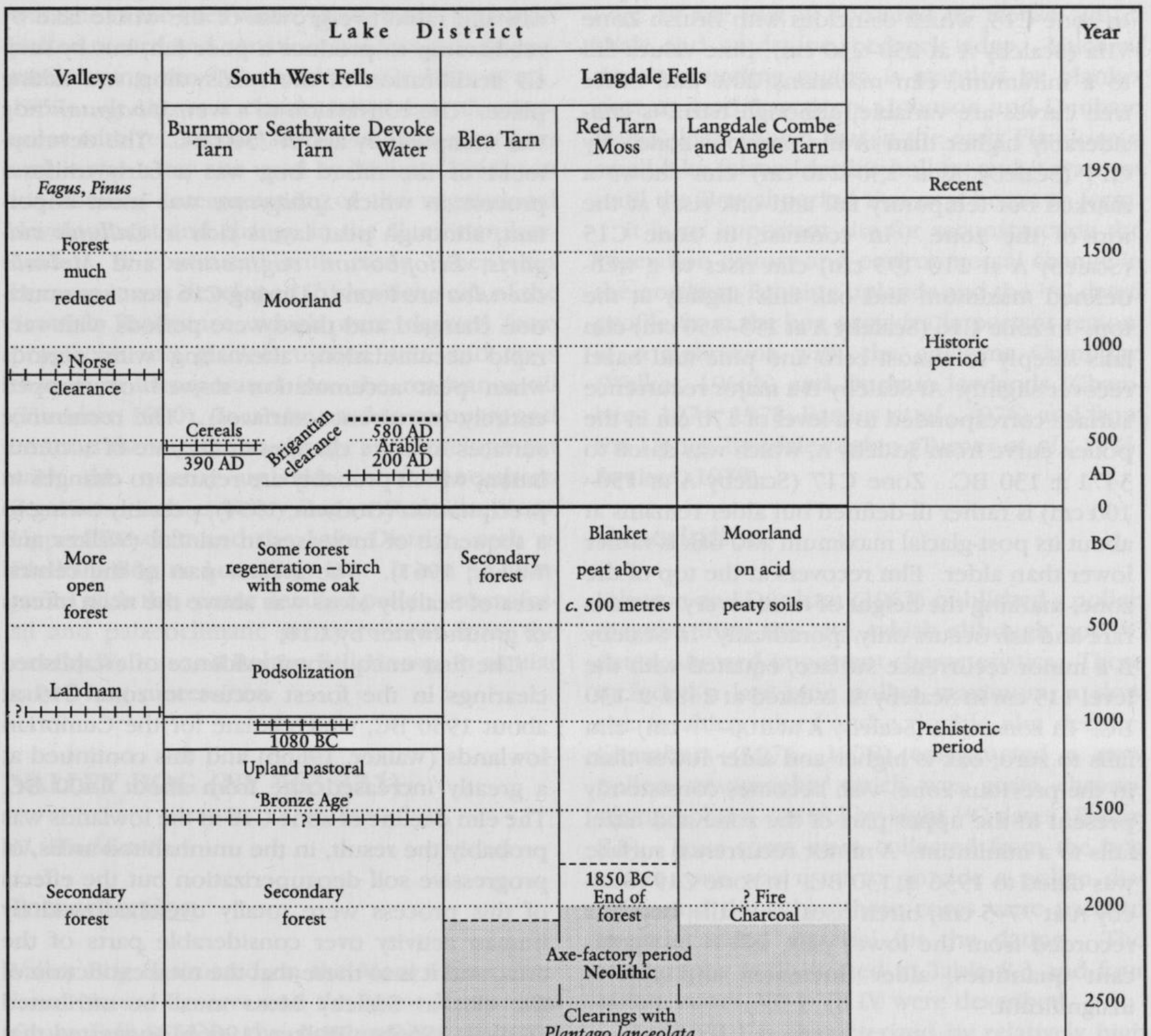


Figure 8.5 (*contd*) Scaleby Moss C pollen diagram from samples at 5 cm intervals through the bottom 1.25 m of the peat monolith (after Godwin *et al*., 1957; Walker, 1966b). All pollen frequencies plotted as percentages of total pollen of dry-land plants excluding ferns. The main pollen-zones have been identified from the changes in vegetation, and the radiocarbon samples were taken at the zone boundaries at the levels shown on the right of the diagram.

Table 8.2 Vegetation history in north-west England after 6000 years BP (after Pennington, 1970)





Lake District
Year
Valleys
South West Fells
Langdale Fells
Burnmoor Tarn
Seathwaite Tarn
Devoke Water
Blea Tarn
Red Tarn Moss
Langdale Combe and Angle Tarn
1950
Fagus, Pinus
Recent
Forest much reduced
1500
Moorland
? Norse clearance
Historic period
1000
Cereals
390 AD
Brigantian clearance
580 AD
Arable
200 AD
500
AD
0
BC
More open forest
Some forest regeneration – birch with some oak
Secondary forest
Blanket peat above c. 500 metres
Moorland on acid peaty soils
500
Podsolization
Landnam
1000
1080 BC
Prehistoric period
Upland pastoral
'Bronze Age'
1500
Secondary forest
Secondary forest
1850 BC
End of forest
? Fire
Charcoal
2000
Axe-factory period
Neolithic
Clearings with Plantago lanceolata
2500
VIIb
3000
3150 BC
3500
VIIa
4000
Primary forest

rise of alder from zero at the base to 45% of the total tree pollen at the top is associated with considerable fluctuations of the other tree types. In zone C13, which coincides with British Zone VIIa (Scaleby A at 250–230 cm), pine values fall to a minimum, elm maintains 20% and other tree curves are variable, although hazel is considerably higher than in the previous zone. By C14 (Scaleby A at 230–210 cm) elm shows a marked but temporary fall and oak rises at the top of the zone. In contrast, in zone C15 (Scaleby A at 210–195 cm) elm rises to a well-defined maximum and oak falls slightly at the top. In zone C16 (Scaleby A at 195–150 cm) elm falls steeply to almost zero and pine and hazel recover slightly. At Scaleby B a major recurrence surface corresponded to a level of 170 cm in the pollen curve from Scaleby A, which was dated to 3471 ± 130 BC. Zone C17 (Scaleby A at 150–100 cm) is rather ill-defined but alder remains at about its post-glacial maximum and oak is rather lower than alder. Elm recovers at the top of the zone, marking the height of its recovery. Lime is rare and ash occurs only sporadically. In Scaleby B a minor recurrence surface, equated with the level 115 cm in Scaleby A, is dated at 2481 ± 130 BC. In zone C18 (Scaleby A at 100–77 cm) elm falls to zero, oak is higher and alder lower than in the previous zone. Ash becomes consistently present in the upper part of the zone and hazel falls to a minimum. A minor recurrence surface was dated to 1936 ± 130 BC. In zone C19 (Scaleby A at 77–5 cm) birch rises, oak falls, beech is recorded from the lower part, ash is in significant quantities, alder fluctuates and elm is insignificant.

The ecological development of the moss

The accumulation of organic mud under open water began before 9000 BC and continued through zones C5–8, probably until about 8200 BC. During zone C5 the basin must have contained a rather barren pool with *Myriophyllum alterniflorum* and *Potamogeton* species and a narrow, fringing reedswamp. In zone C6 the aquatic flora changed little but the reedswamp became more extensive and *Equisetum* species advanced into the shallow water. In zone C8 the inorganic content of the mud increased considerably, with the occurrence of coarse sand and pebbles near the middle of the lake. Walker (1966b) suggests deposition by ice floes to explain this, which increased the base status of the water owing to the influx of inorganic sediments from the banks. The opening of zone C9 saw the rapid overgrowth of the whole lake by reedswamp to produce a poor fen, but by mid-C9 acidification of the reedswamp was taking place. The conversion to a wet, *Sphagnum* bog was complete by about 7500 BC. The development of the raised bog was a fairly uniform process in which *Sphagnum* was most important, although peat layers rich in *Calluna vulgaris*, *Eriophorum vaginatum* and *Molinia caerulea* are found. During C16 peat accumulation changed and there were periods with very rapid accumulation, alternating with periods when peat accumulation slowed or stopped entirely (recurrence surfaces). The recurrence surfaces indicate changes in the rate of accumulation, which probably are related to changes in precipitation (Godwin, 1954), possibly owing to a sequence of increases in rainfall (Walker and Walker, 1961). The greater part of the central area of Scaleby Moss was above the main effects of groundwater by C16.

The first unequivocal evidence of established clearings in the forest occurs in zone C19 at about 1950 BC, which is late for the Cumbrian lowlands (Walker, 1966b) and this continued at a greatly increased rate from about 1400 BC. The elm decline of zone C16 in the lowlands was probably the result, in the uninhabited areas, of progressive soil depauperization but the effects of this process were totally overshadowed by human activity over considerable parts of the area, and it is to these that the total extinction of the elm at Scaleby Moss must be attributed (Walker, 1966b). Walker (1966b) suggested that there were early Neolithic clearances at sites such as Ehenside Tarn and Abbott Moss, and it was in areas of better soils (sandy, gravelly and therefore lighter) that were most early affected, but clearance at sites such as Scaleby Moss, on low-clay areas were delayed until a new economy, more sedentary and utilizing cereals, became established. The vegetational history in north-west England after 6000 years BP is illustrated in Table 8.2, where the comparison can be made between Scaleby Moss, Ehenside Tarn, Barfield Tarn and the valleys of the Lake District. The major differences between Scaleby Moss and the sand and gravelly soils of the coastal plain can be clearly seen and this also contrasts with the inner Lake District, with their early Neolithic clearings.

Conclusions

Scaleby Moss is an important site in Britain as it was one of the sites where the detailed Flandrian pollen record and zonation was first dated accurately using the ^{14}C method (Godwin *et al.*, 1957). It shows a classic raised bog development with recurrence surfaces in its upper part and it was a key site used by Walker (1966b) in his detailed interpretation of the vegetational development and change in the Cumbrian lowlands. It contrasted with many other sites because it was developed in clay-rich soils of the Scottish Readvance, which were derived from proglacial and glaciolacustrine silts and clays that were incorporated into the readvance till (Huddart, 1970). It can be used to compare and contrast the vegetation change and development with the nearby uplands in the northern Pennines, such as Valley Bog or Red Sike Moss, Upper Teesdale and the Lake District, such as Blelham Bog or Low Wray Bay. It can be compared with the recent detailed pollen, macrofossil and palaeoclimatic proxy records from the nearby Walton and Bolton Fell Mosses in similar raised bog successions.

VALLEY BOG (NY 763 331)

D. Huddart

Introduction

Valley Bog is situated on the Moor House National Nature Reserve and the Moor House and Cross Fell SSSI in the northern Pennines, in Cumbria. The bog is the upper and larger of two bogs that are situated in a channel at 549 m OD. To the north, east and west of Valley Bog there are glacial deposits, which form part of a series of moraines (Chambers, 1978). The southern edge of the bog is bordered by House Hill, a shale and sandstone, bedrock ridge. Much of the surrounding region is mantled by blanket peat up to 3.5 m deep. Johnson and Dunham (1963) have shown that in the early Flandrian a small lake formed in this hollow and it was not until the Boreal period that peat began to form.

It is an important site for reconstructing the vegetation history and environmental change in the northern Pennine uplands and the ^{14}C dated profile from the bog provides important regional comparisons with the adjacent Cumbrian (Walker, 1966b) and Durham lowlands (Chambers, 1974, 1978; Bartley *et al.*, 1976) and from the Upper Teesdale region (Turner *et al.*, 1973; Squires, 1978).

Description

Johnson and Dunham (1963) published a pollen diagram from this site, which although not ^{14}C dated, showed important characteristics. These included a late pine pollen maximum, a slow rise in alder values and a double elm decline. Chambers (1974, 1978) constructed a new pollen stratigraphy, which was given chronostratigraphical control by eight ^{14}C dates (Figure 8.6). Four cores were collected from the bog centre, one was used to provide a pollen diagram and the other three cores were used to provide added material for the dating. The stratigraphy is illustrated in Table 8.3 and four pollen zones, VB I–VB IV, were described.

Zone VB I is characterized by relatively high birch and pine pollen values. Elm, oak and hazel pollen is present in abundance but alder in

Table 8.3 Stratigraphy at Valley Bog (after Chambers, 1978)

Depth (cm)	Stratigraphy
0–50	Not sampled
50–75	Sedge peat of low humification (H4) with some *Calluna* remains
75–100	Sedge peat of low humification (H3) with some *Calluna*
100–150	Sedge peat of low humification (H4) with abundant pieces of *Calluna*
150–200	Slightly muddy sedge peat of medium humification (H5–6) with *Calluna*
200–250	Slightly muddy sedge peat of low humification (H3–4) with *Betula* wood
250–290	Slightly muddy sedge peat of low humification (H5–6) with less *Betula*
290–525	Slightly muddy sedge peat of low humification (H5–6) with abundant pieces of *Betula* wood
525–580	Bryophyte peat of low humification (H3) composed mainly of *Paludella squarrosa* together with some *Eriophorum* sedge remains
580–600	Sedge peat of low humification (H3-4) with some *Eriophorum*

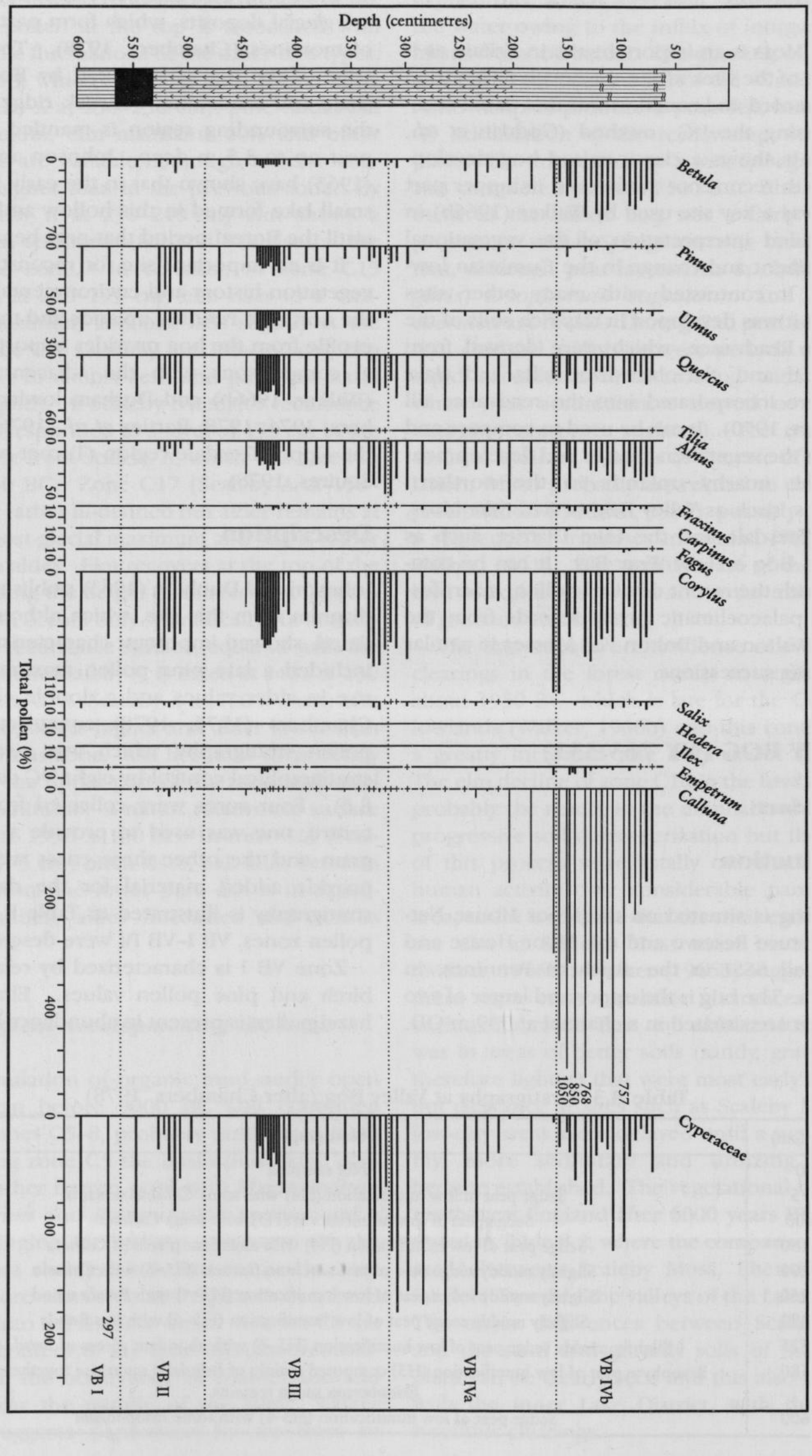


Figure 8.6 Pollen Diagram from Valley Bog (after Chambers, 1978). See Figure 8.1 for key to the stratigraphical log. (Figure continued on opposite page.)

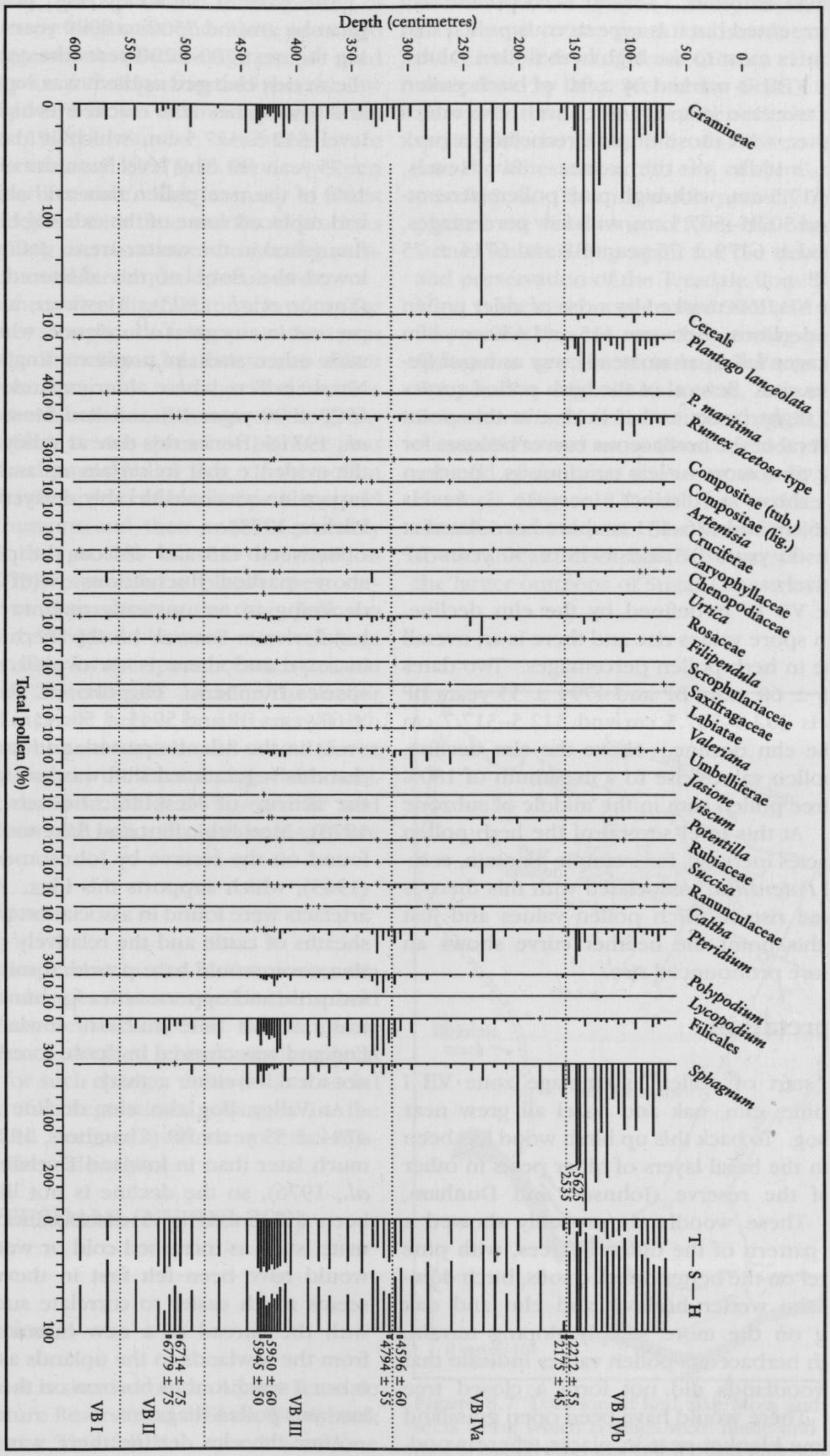


Figure 8.6 (contd) Pollen Diagram from Valley Bog (after Chambers, 1978)

only small amounts. Several herb pollen taxa are represented but it is cyperaceous pollen that contributes most to the high herb pollen values.

Zone VBII is marked by a fall of birch pollen and an increase in oak. Alder and pine values are higher, with those of pine recording a peak in the middle of the zone. Two levels, 512.5–517.5 cm, with high pine pollen percentages, and 502.5–507.5 cm, with low percentages, are dated at 6779 ± 75 years BP and 6714 ± 75 years BP.

Zone VB III is marked by a rise of alder pollen as pine declines. Between 415 and 430 cm, elm percentages fall in an unsteady way as hazel frequencies rise. Several of the herb pollen curves exhibit slight peaks and it is also at this point that several of the herbaceous curves become for the first time more or less continuous. Bracken values show a distinct increase. Levels 421–426 cm and 426–431 cm have been dated at 5950 ± 60 years BP and 5945 ± 50 years BP respectively.

Zone VB IV is defined by the elm decline. Bracken spore values rise and there is an overall increase in herb pollen percentages. Two dates of 4596 ± 60 years BP and 4794 ± 55 years BP for levels 302.5–307.5 cm and 312.5–317.7 cm span the elm decline. Above the elm decline, grass pollen values rise to a maximum of 180% of the tree pollen sum in the middle of subzone VB–IVa. At this level several of the herb pollen frequencies increase, for example plantain, cereals and *Potentilla*. Associated with this there is a marked rise in birch pollen values and just above this point the heather curve shows an even more pronounced rise.

Interpretation

By the start of pollen assemblage zone VB I birch, pine, elm, oak and hazel all grew near Valley Bog. To back this up birch wood has been found in the basal layers of older peats in other parts of the reserve (Johnson and Dunham, 1963). These woodlands probably showed a mosaic pattern of the different trees, with pine and hazel on the better drained soils, birch dominating the wetter hollows and elm and oak growing on the more steeply sloping terrain. The high herbaceous pollen values indicate that these woodlands did not form a closed tree canopy. There would have been open grassland with some blanket peat in places where woodland did not occur.

The base of the Valley Bog pollen diagram must be around 7500 to 8000 years BP and during the next 700–1200 years the composition of the woods changed as birch was replaced by oak and some pine. Pine reached its highest value at level 512.5–517.5 cm, which is dated to 6779 ± 75 years BP. This level has alder values around 10% of the tree pollen sum and alder probably had replaced some of the existing birch that had flourished in the wetter areas. Its migration followed the floor of the sheltered Tees valley (Turner *et al.*, 1973). However, alder was not present in any great abundance, which contrasts with other sites in northern England such as Neasham Fen, where alder expanded soon after 6972 ± 90 years BP, and Red Moss (Hibbert *et al.*, 1971). Hence this date at Valley Bog adds to the evidence that in certain upland areas alder expansion was considerably delayed (Smith and Pilcher, 1973).

Between 415 and 430 cm the pollen curves show marked fluctuations, with elm values declining in an unsteady manner as those of hazel rise. Several of the herb frequencies increase and there is an overall rise in herb species numbers. The two ^{14}C dates of 5950 ± 60 years BP and 5945 ± 50 years BP place this event in the Atlantic period, and it is suggested that this vegetational shift was brought about by the activity of Mesolithic hunters (Chambers, 1978). Mesolithic flint and flake tools have been found on the reserve by Johnson and Dunham (1963), which supports this idea. A few of the artefacts were found in association with the horn sheaths of cattle and the relatively open vegetation cover would have provided suitable grazing for cattle and opportunities for hunting. In contrast, at this time much of lowland northern England was covered in dense forest, which was not ideal for either activity.

At Valley Bog the elm decline is dated to 4794 ± 55 years BP (Chambers, 1978), which is much later than in lowland Durham (Bartley *et al.*, 1976), so the decline is not likely to have been influenced by a deterioration of the climate, such as increased cold or wetness, which would have been felt first in the uplands. It seems much easier to correlate such a decline with the spread of a new human population from the lowlands to the uplands as the decline is not a synchronous horizon on the upland and lowland pollen diagrams.

After the elm decline there was a period of woodland clearance, which is indicated by high

Graminae and plantain values. The opening of the tree canopy began around 3300 years ago and it is suggested that Bronze Age people used the clearings for animal grazing and that the few cereal grains were blown in from lowland areas (Chambers, 1978). Similar clearances were found by Turner *et al*. (1973) in the Cow Green area of Upper Teesdale. The intensity of forest clearance declined with time and the indications from Valley Bog are that once the grazing pressure was removed birch quickly took over.

Towards the pollen diagram top the values for heather rise sharply, but because the peat changes from one without heather to one with it is difficult to determine whether the heather pollen curve reflects a local change on the growing bog surface, or the spread of blanket peat over the areas around Valley Bog. If the pollen curve reflects the spread of blanket peat it would seem that humans and their animals probably accelerated the process of soil deterioration (Moore, 1973), especially as blanket peat certainly grew on the reserve at the time (Johnson and Dunham, 1963) and humans may well have been responsible for the spread of such peat on to the more waterlogged areas. The major deforestation phase begins in subzone VB IVb and was more pronounced and lasted for a longer time than in the Bronze Age clearances. Two levels dated at 2212 ± 55 years BP and 2175 ± 45 years BP indicate a late Iron Age date for this phase.

Conclusions

This site is important in that it documents the vegetational changes that occurred in the upland northern Pennine landscape in the Flandrian and provides a detailed chronostratigraphical framework for such change. It serves as a useful comparison for vegetational changes in the Cumbrian and Durham lowlands and the adjacent Upper Teesdale region.

UPPER TEESDALE (NY 819 290)

D. Huddart

Introduction

Red Sike Moss is situated on the Upper Teesdale National Nature Reserve and SSSI on Widdybank Fell, Durham. This fell is a raised limestone plateau at 523 m OD. To the west is the Cow Green reservoir and the Cautley Snout waterfall; the ground to the south and south-east falls to the River Tees at Falcon Clints. The area is underlain by the Melmerby Scar Limestone (Lower Carboniferous Limestone Series) and the quartz-dolerite Whin Sill, which was intruded into the limestone at the end of the Carboniferous Period. This intrusion partially metamorphosed the limestone to form the Sugar Limestone, which is important for the development and preservation of the Teesdale flora. Red Sike Moss is located between Red Sike and Tinkler's Sike and it has an actively growing surface of *Sphagnum* species, including *Calluna vulgaris*, *Eriophorum vaginatum* and *E. angustifolium*, *Erica tetralix*, *Narthecium ossifragum*, *Drosera rotundifolia* and many *Carex* species. The western end of the moss shows erosion by drainage channels (Figure 8.7) and has a drier surface than the central part. The moss lies just to the south, and at the foot of the slope, where one of the larger outcrops of Sugar Limestone occurs.

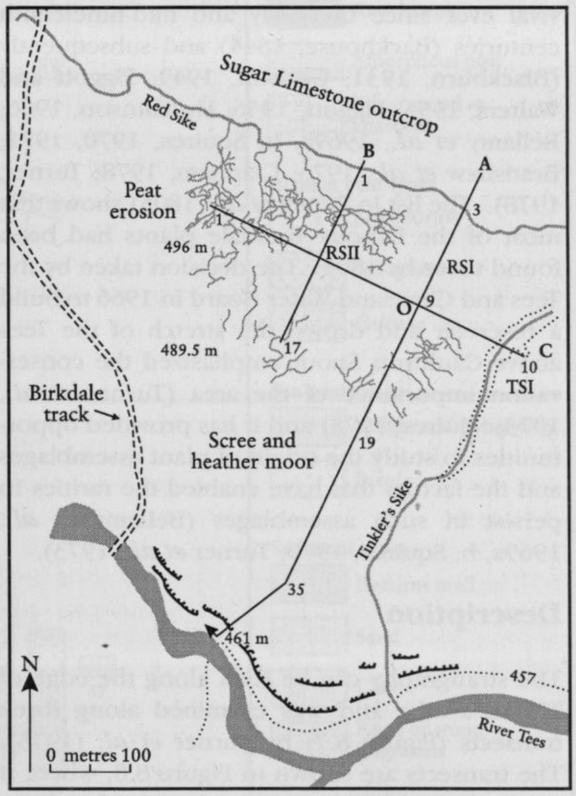


Figure 8.7 Location of Red Sike Moss and the transects along which borings were made and the position of pollen diagrams TSI, RSI and RSII (after Turner *et al*., 1973).

The site is important for reconstructing Flandrian vegetation history in Upper Teesdale, which is an area of international botanical importance, containing a number of species such as *Gentiana verna* and *Kobresia simpliciuscula*, which are found in only a few other locations in Britain and *Minuartia stricta*, which is found nowhere else (Turner *et al*., 1973; Bradshaw and Clark, 1976; Ratcliffe, 1978). These Teesdale rarities represent a wide range of geographical elements in Europe and represent a convergence of phytogeographical elements (Piggott, 1956; Bradshaw and Clark, 1965, Bradshaw, 1970). The rarities occur in communities occupying habitats of diverse character. Many are found in various grassland communities (Shimwell, 1968), existing as windows in the acidophilous moorland cover. Other communities that include rarities have a more limited distribution, and are associated with certain mining spoil heaps with high lead and zinc concentrations, where base-rich drainage water flows over the surface, and those on quartz-dolerite and limestone screes. They have aroused considerable discussion as to their origin and potential survival ever since the early and mid-nineteenth centuries (Backhouse, 1844) and subsequently (Blackburn, 1931; Godwin, 1949; Piggott and Walters, 1954; Piggott, 1956; Hutchinson, 1966; Bellamy *et al*., 1969a, b; Squires, 1970, 1971; Bradshaw *et al*., 1976; Clapham, 1978; Turner, 1978). The list in Winch *et al*. (1805) shows that most of the famous Teesdale plants had been found there by 1805. The decision taken by the Tees and Cleveland Water Board in 1966 to build a reservoir and drown the stretch of the Tees above Cauldron Snout emphasized the conservation importance of the area (Turner *et al*., 1973; Squires, 1978) and it has provided opportunities to study the origin of plant assemblages and the factors that have enabled the rarities to persist in such assemblages (Bellamy *et al*., 1969a, b; Squires, 1970; Turner *et al*., 1973).

Description

The stratigraphy can be seen along the edge of Tinkler's Sike and was examined along three transects (Figure 8.7) by Turner *et al*. (1973). The transects are shown in Figure 8.8, where it can be seen that the peat depth varies from 1 m at the margin to a 4 m maximum in the centre. Samples for pollen analysis were collected from the points marked TSI, RSI and RSII on Figure 8.7. The detailed stratigraphy at TSI is given in Table 8.4 and the pollen diagram in Figure 8.9. This diagram was zoned according to the major assemblages, which are, starting with the youngest:

Zone G a grass–plantain–heather assemblage
Zone A an oak–alder assemblage
Zone O an oak–elm–hazel assemblage
Zone H a hazel–pine assemblage
Zone HJ a hazel–juniper–willow assemblage
Zone J a juniper–willow–herb assemblage

The pollen diagram from RSII is given in Figure 8.10. The results from RSI show only the bottom part of the profile and peat formation did not start until zone H. The bottom of the blanket peat at position BSIII has a zone A pollen assemblage and the upper peat, some 35 cm above in BSIII, has a zone G pollen assemblage. The pollen diagram is shown in Figure 8.11.

The relationships between these local pollen assemblage zones and those of Godwin (1940) and the chronozones of West (1970) are given in Table 8.5. The ^{14}C dates for TSI are shown in Table 8.6. A summary pollen diagram is given in Figure 8.12.

Interpretation

Late-glacial c. 10 000 years BP

The Late-glacial Zone III deposit indicates a period of herbaceous vegetation with abundant grasses and sedges and a wide variety of other species. These include such common Late-glacial species as *Artemisia*, *Filipendula*, *Helianthemum*, *Rumex* and *Thalictrum*. *Betula nana*, juniper and willow species pollen occur in sufficient quantity to indicate that locally these shrubs were flourishing (Figure 8.13a). The area would have provided a wide variety of microhabitats from wet, peaty areas near the sikes to the well-drained, Sugar Limestone outcrops.

At TSI the basal few centimetres are thought to have formed in the Late-glacial pollen assemblage zone J, corresponding to very late in Godwin's Zone III. The evidence comes from the very low tree pollen percentages and over 20% juniper pollen, together with a ^{14}C date of 9900 years BP for the 135 cm level. The peat above this level has a different pollen assemblage, with a high percentage of hazel and both oak and elm recording up to 10%. This

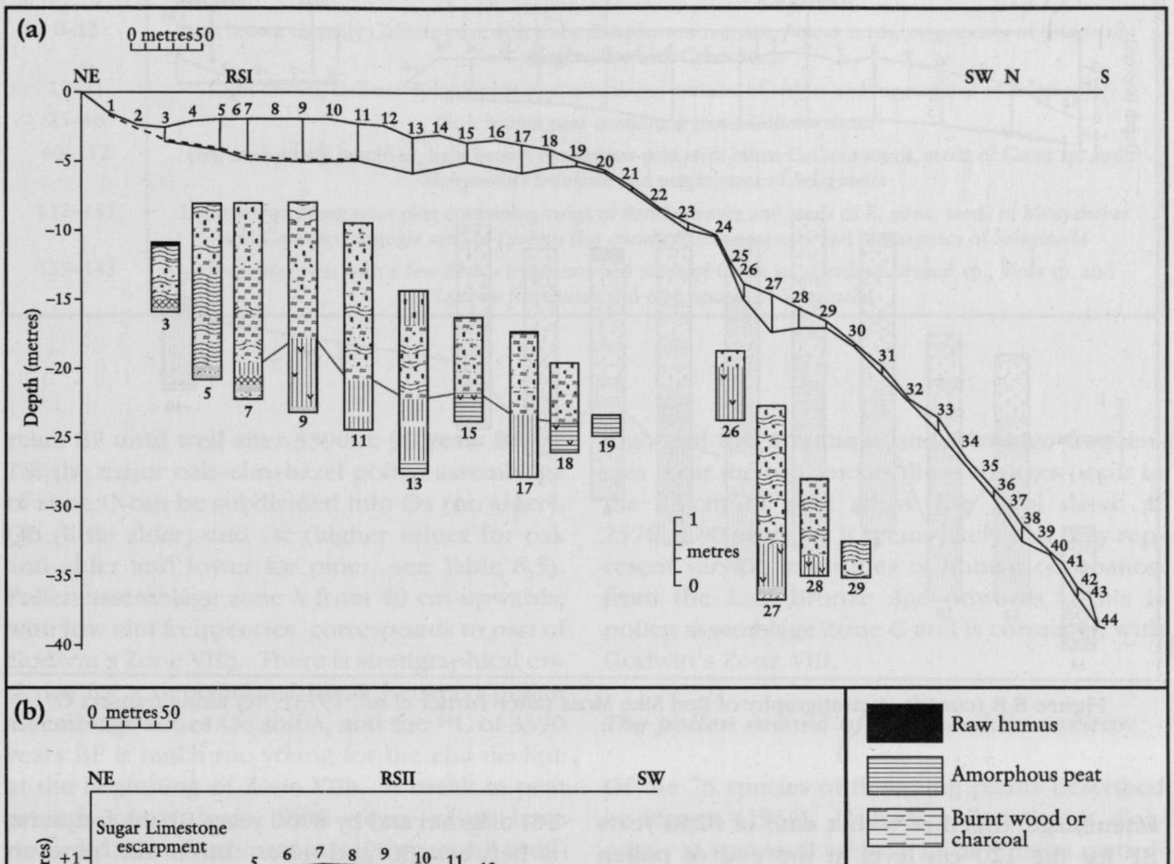


Figure 8.8 A stratigraphy of Red Sike Moss (after Turner *et al.*, 1973): (a) along transect A; (b) along transect B; see overleaf for (c) along transect C.

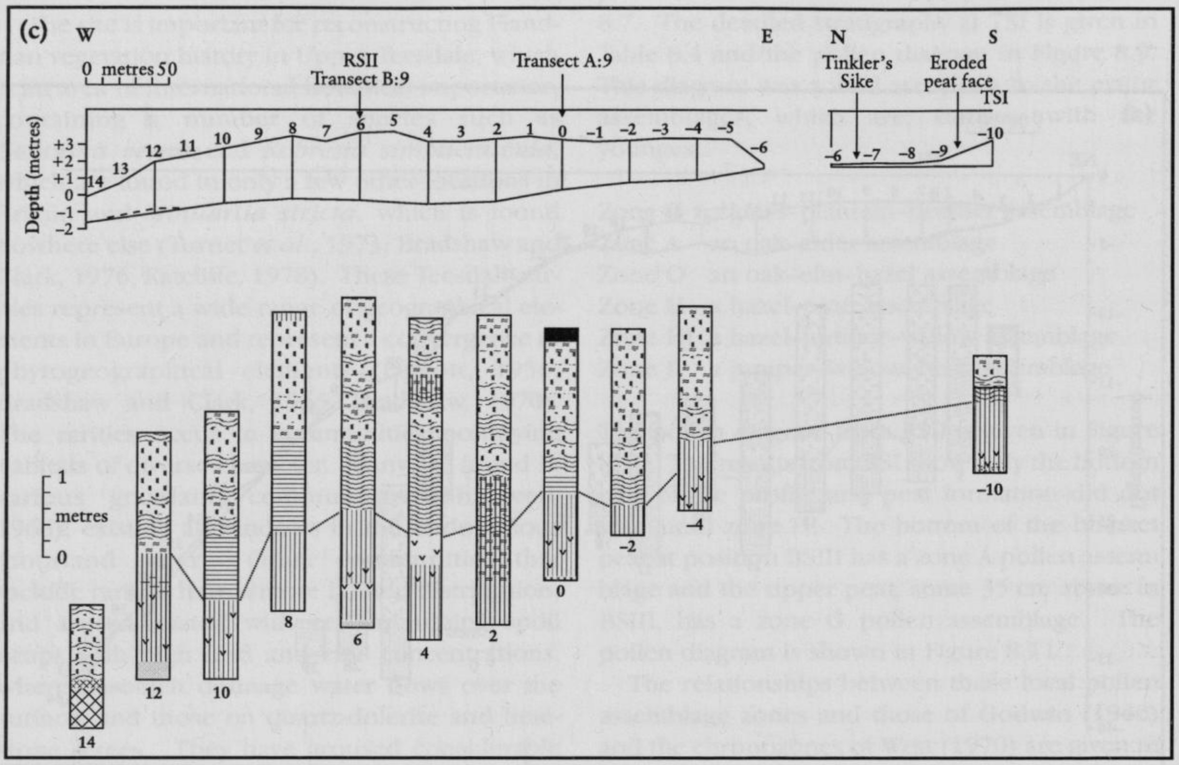


Figure 8.8 (*contd*) A stratigraphy of Red Sike Moss (after Turner *et al*., 1973): (c) along transect C.

assemblage, together with a date of 8250 years BP for the 120 cm level at the end of pollen assemblage zone H, indicates that it is part of Godwin's Zone V1. This means that there is a depositional break during Zones IV and V.

10 000–8800 years BP

In this period the peat stopped growing, but began forming again in pollen assemblage zone H. By the beginning of this zone the total tree pollen appears only marginally higher than at the end of the Late-glacial. Graminae, Cyperaceae and other herbaceous pollen frequencies have decreased correspondingly, but they are still well represented at the end of the period. There is no question of closed woodlands having formed.

8800–8000 years BP

In this period (pollen assemblage zone H), oak and elm pollen appear and the frequency of hazel is high. There is no juniper but willow and herbaceous pollen types are still abundant. The total tree pollen frequencies rise steadily on the TSI diagram and by 8000 years BP the frequency is between 30 and 40%. Birch and pine are increasing but it is unlikely there would be closed woodland.

8000–5770 years BP

At the beginning of this period the pine frequency is high and alder and oak low, or zero. By the end the latter two species have risen to high values and pine has disappeared. It is clear that pine grows best on the limestone and persisted there for longer than it did elsewhere (Turner *et al*., 1973). This conclusion is supported by the fact that *Carex ericetorum* (a Teesdale rarity) is now restricted to the Sugar Limestone. Today it grows in the ground flora of pinewoods in European Russia (Keller, 1927), and where it occurs in the grassland of the East Anglian Breckland it is thought by Watt (1971) to indicate the former existence of pinewoods.

5770–5000 years BP

In this period there is evidence for a long break in peat formation from sometime before 5000

Table 8.4 Stratigraphy at TSI, Red Sike Moss (after Turner *et al.*, 1973)

Depth (cm)	Description
0–12	Dark brown crumbly *Calluna* peat with some *Eriophorum* remains, *Juncus* seeds, megaspores of *Selaginella selaginoides* with *Carex* seeds
12–25	Light brown, *Calluna–Eriophorum* peat containing remains of sedges and megaspores of *Selaginella*
25–40	Dark brown peat containing burnt *Calluna* stems
40–112	Dry, moderately humified, light brown *Phragmites* peat with burnt *Calluna* stems, seeds of *Carex* sp. and *Menyanthes trifoliata* and megaspores of *Selaginella*
112–135	Light brown *Phragmites* peat containing twigs of *Betula*, leaves and seeds of *B. nana*, seeds of *Menyanthes* and *Carex* sp., a single seed of *Lychnis flos-cuculi, Chara* oospores and megaspores of *Selaginella*
135–143	*Phragmites* peat with a few *Betula* fragments and seeds of *Carex* sp., *Carduus cirsium* sp., *Viola* sp. and *Lychnis flos-cuculi* and megaspores of *Selaginella*

years BP until well after 3390 ± 90 years BP. At TSI the major oak–elm–hazel pollen assemblage of zone O can be subdivided into Oa (no alder), Ob (little alder) and Oc (higher values for oak and alder and lower for pine: see Table 8.5). Pollen assemblage zone A from 40 cm upwards, with low elm frequencies, corresponds to part of Godwin's Zone VIIb. There is stratigraphical evidence for a depositional break between pollen assemblage zones Oc and A, and the ^{14}C of 3390 years BP is much too young for the elm decline at the beginning of Zone VIIb. A break in peat growth of at least 1600 years must have occurred. The fresh peat at 3390 years BP is different from that below, being less humified and composed mainly of *Sphagnum* species.

At the time of the forest maximum the tree pollen frequency contributed between 30–50% of the total pollen spectrum on the Teesdale diagrams (Turner *et al.*, 1973), which contrasts, for example, with the Cranberry Bog (County Durham) lowland diagram (Turner, 1970), where the herb frequency averages some 5% as opposed to the 30–40% for Upper Teesdale. Hence the woods were open compared with the lowlands, to allow a varied herbaceous flora to flourish and contribute a large percentage to the total pollen spectrum.

5000 years BP and later

The open woodland covering most of Upper Teesdale by 5000 years BP was replaced where the soil was becoming waterlogged by blanket peat and where it was better drained by grassland. The change from woodland to blanket peat seems to have been an irreversible process. At Tinkler's Sike there are a series of alternating high and low Graminae and *Plantago* frequencies near the top, and as these changes occur in the 15 cm of peat above the level dated at 2570 ± 80 years BP, it seems likely that they represent varying intensities of human occupation from the Late Bronze Age onwards. This is pollen assemblage Zone G and is correlated with Godwin's Zone VIII.

The pollen record of the Teesdale rarities

Of the 75 species of flowering plants described by Piggott (1956), 16 have pollen that is diagnostic at species level and 11 have been recorded in the Upper Teesdale pollen diagrams (Turner *et al.*, 1973). These are indicated in Table 8.7. The species found at Red Sike Moss are discussed below. *Betula nana* (Figure 8.13a) has been present on Widdybank Fell throughout the Flandrian and has been much more abundant than it is today. It was unknown on the fell until Hutchinson (1966) found a few small plants growing on dry, heather-covered peat and demonstrated its presence in the past by finding subfossil leaves in the peat beneath. *Gentiana verna* was recorded from pollen assemblage zone Oa on Red Sike Moss (RSII) and zone H in the blanket peat basal samples II and III (Turner *et al.*, 1973), so confirming its existence in the area during the maximum forest development and later. *Helianthemum* species have been found in pollen assemblage zone H (RSI), zone O (RSII) and zones A and G at many sites in Upper Teesdale. It has been present throughout the Late- and Post-Glacial periods and has expanded with the spread of grassland. *Plantago maritima* has been recorded from pollen assemblage zone G at Tinkler's Sike and

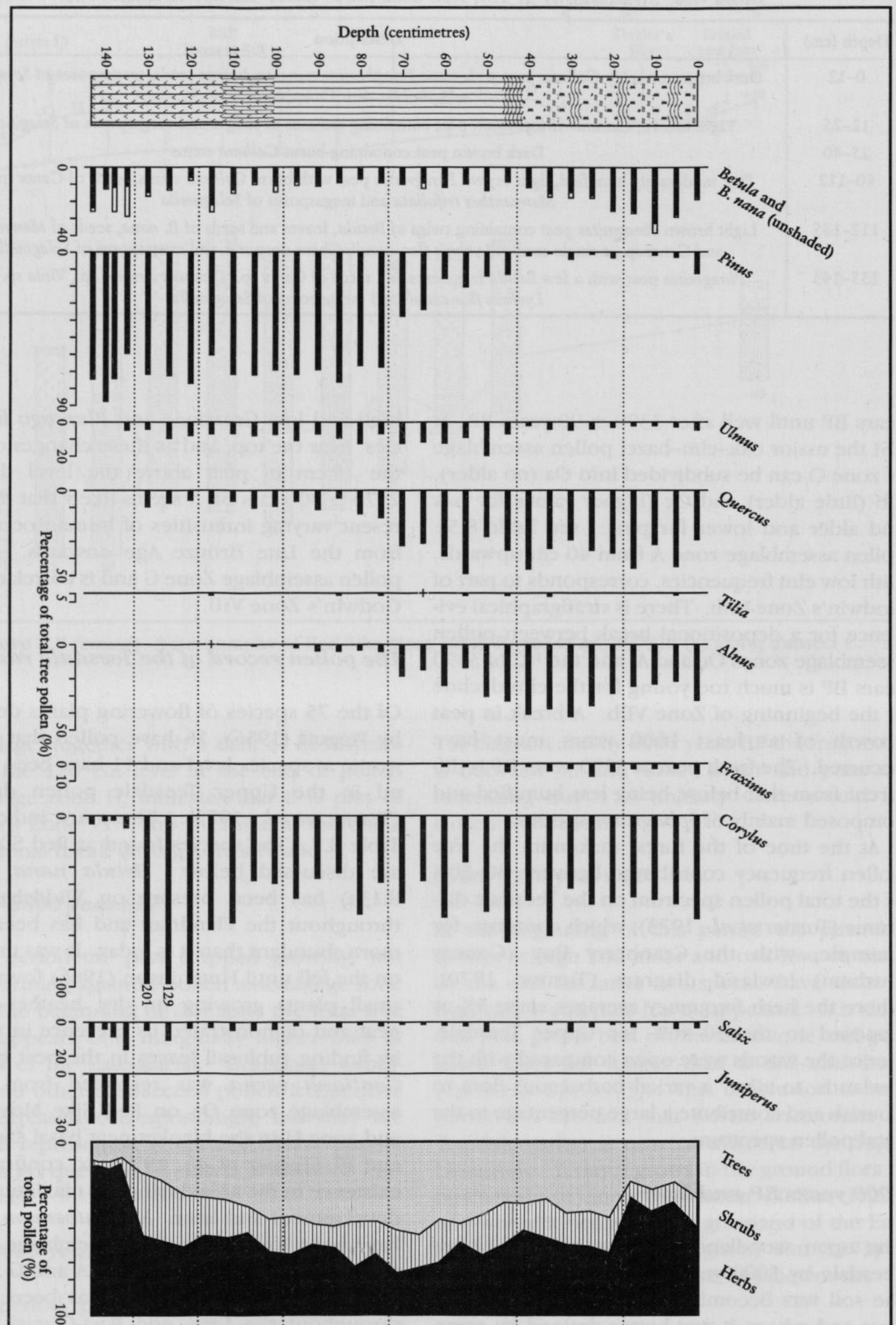


Figure 8.9 Red Sike Moss TSI (after Turner *et al.*, 1973): tree pollen diagram. See Figure 8.8 for key to the stratigraphical log. (Figure continued on opposite page for herb pollen.)

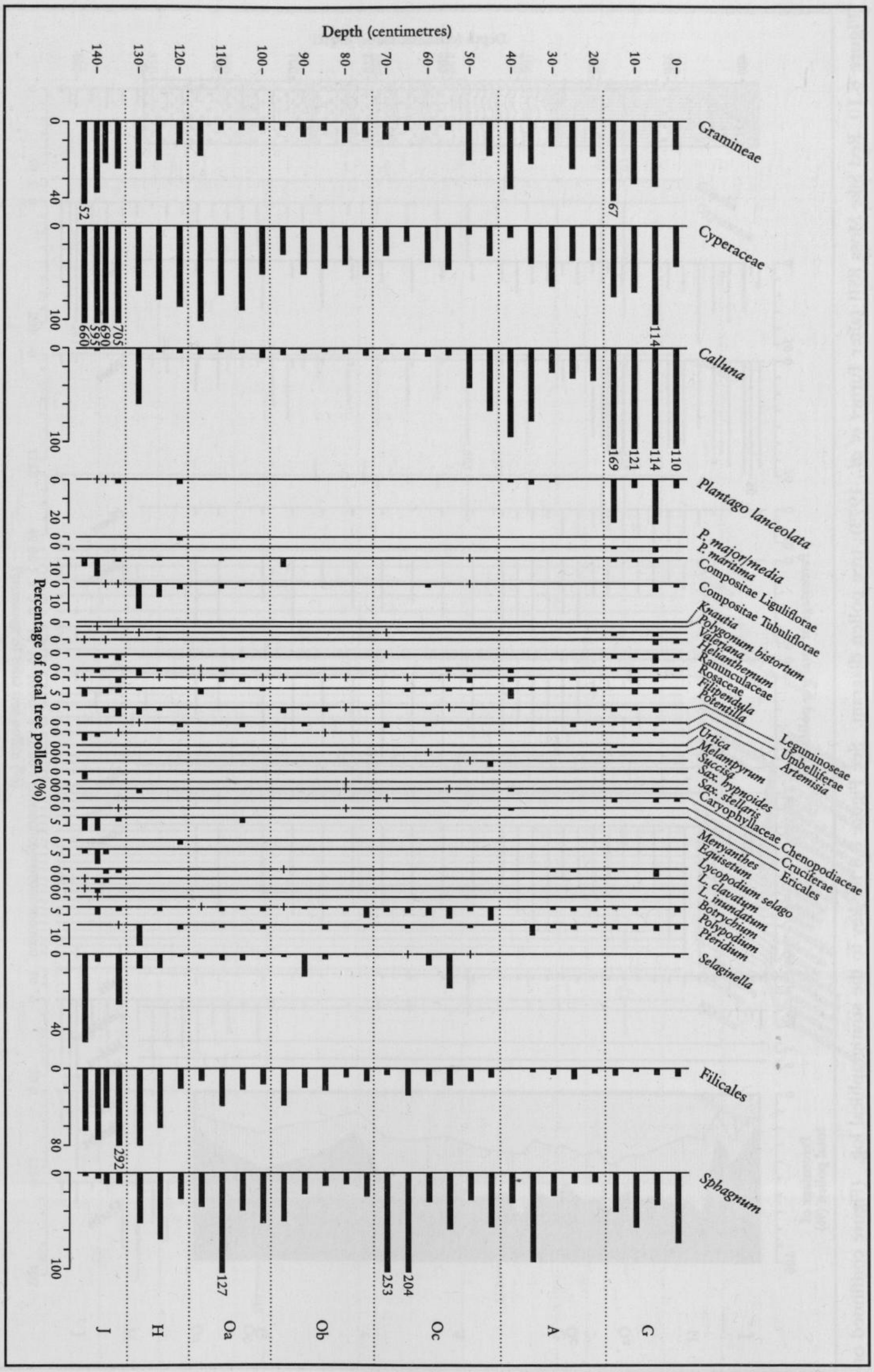


Figure 8.9 *(contd)* Red Sike Moss TSI (after Turner *et al.*, 1973): herb pollen diagram.

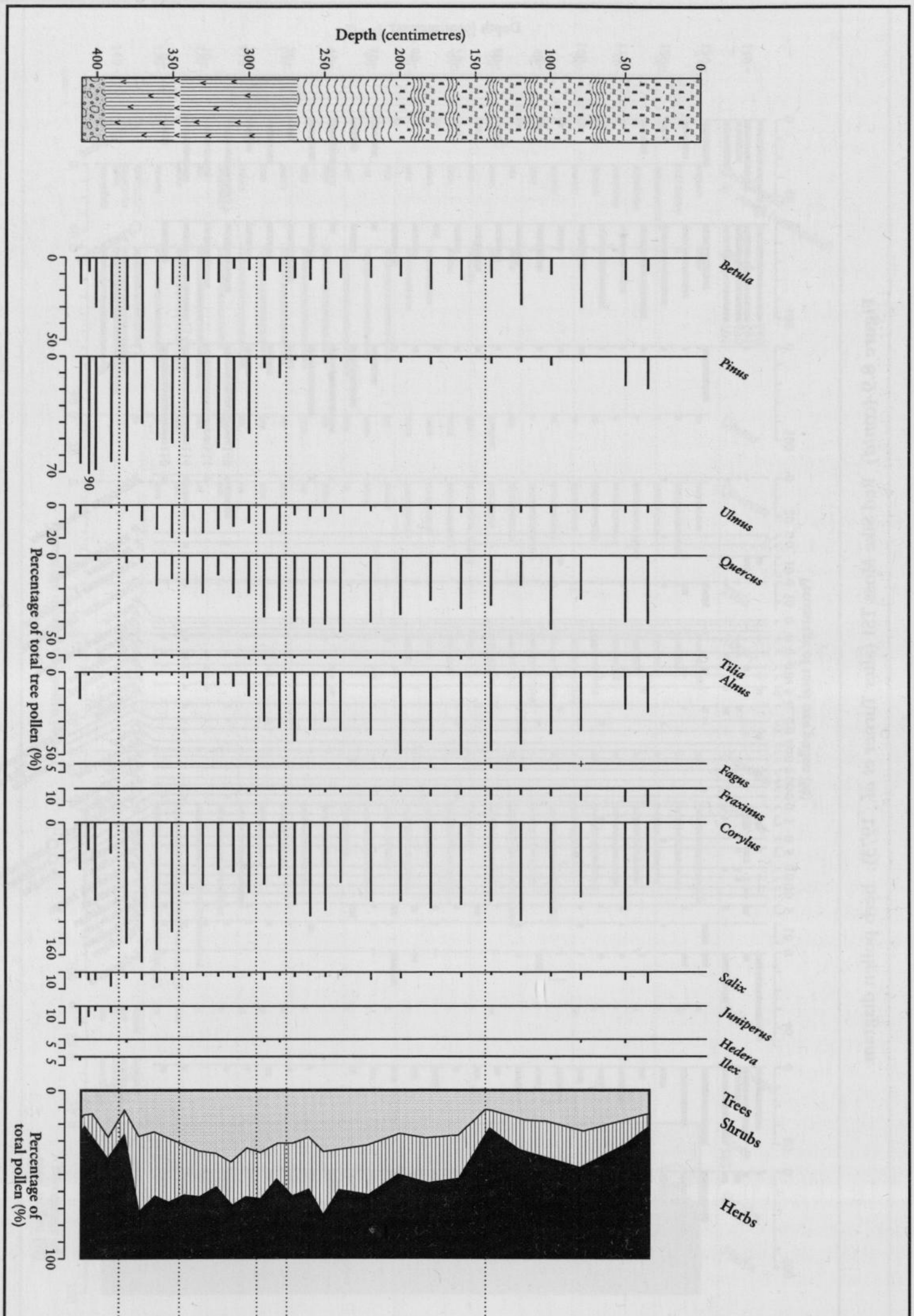


Figure 8.10 Red Sike Moss RSII (after Turner *et al.*, 1973): tree pollen diagram. See Figure 8.8 for key to the stratigraphical log. (Figure continued on opposite page for herb pollen.)

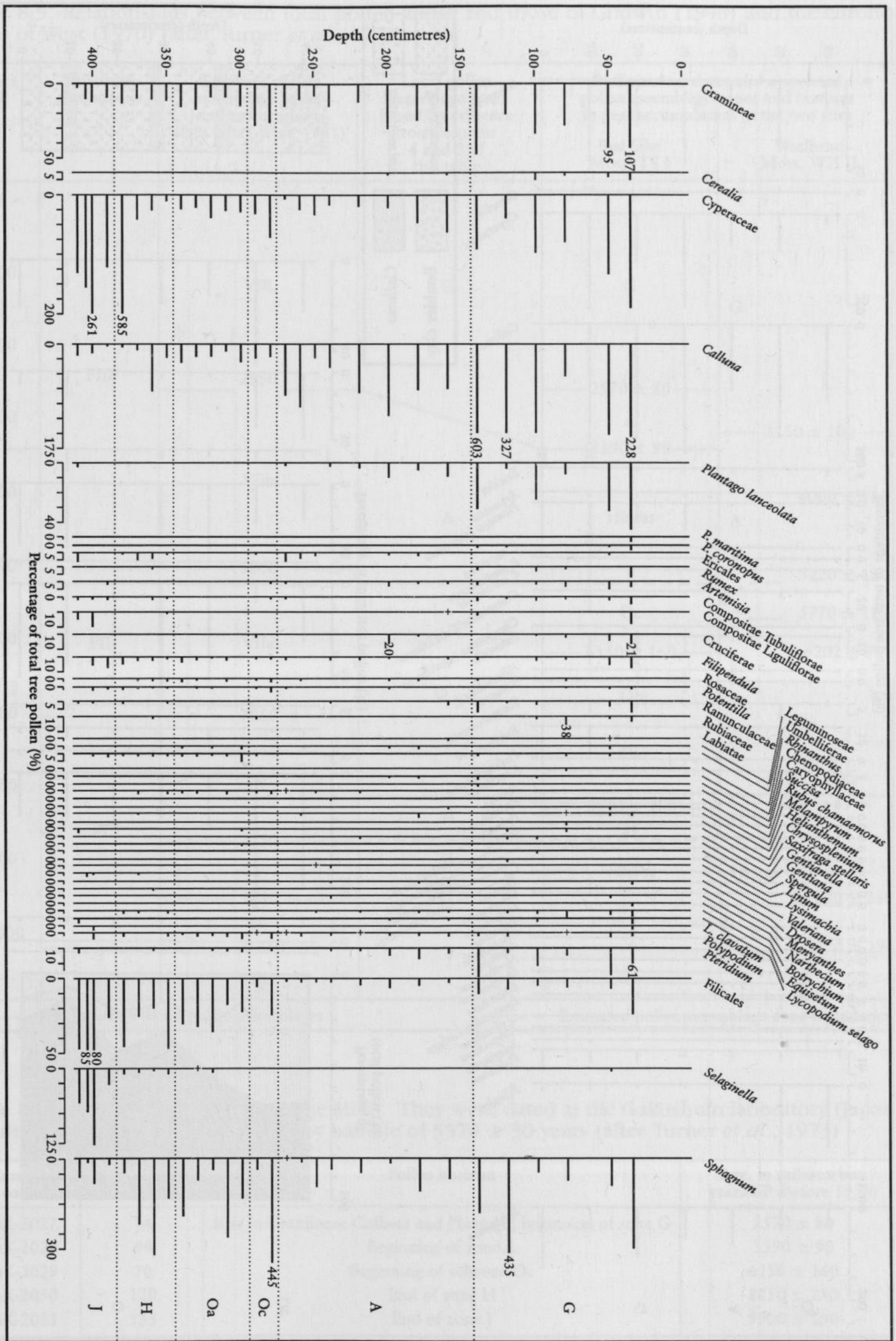


Figure 8.10 (*contd*) Red Sike Moss RSII (after Turner *et al.*, 1973): herb pollen diagram.

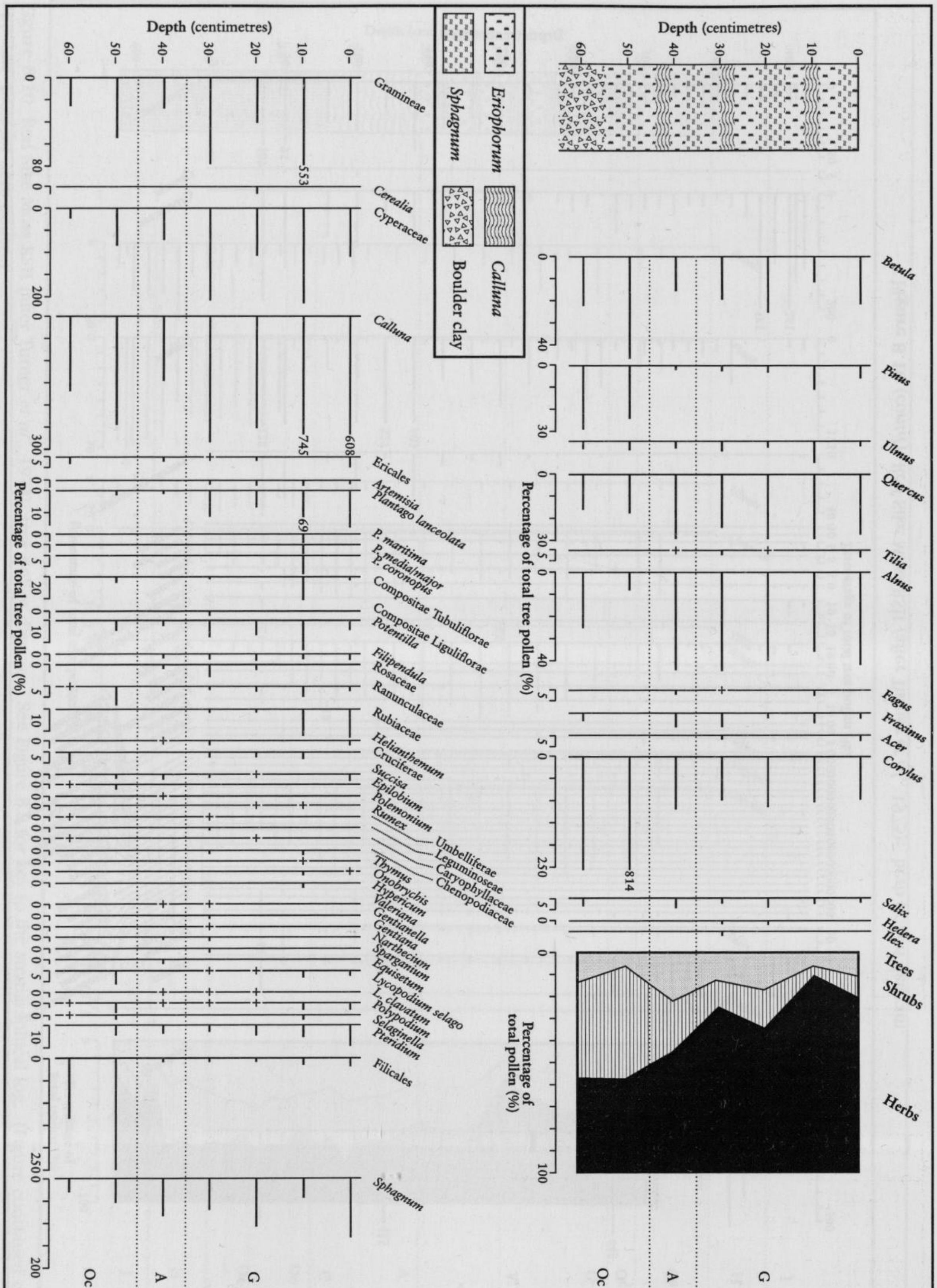


Figure 8.11 Pollen diagram from blanket peat (BSIII). Frequencies plotted as a percentage of the total tree pollen excluding hazel and willow (after Turner *et al*., 1973).

Table 8.5 Relationships between local pollen zones and those of Godwin (1940) and the chronozones of West (1970) (after Turner *et al*., 1973)

Years BP	Flandrian chronozone	Godwin's pollen assemblage zones and radiocarbon dates (after West, 1961)	Local pollen assemblage zones based on evidence from columns 4 and 5 of this table	Radiocarbon dates and associated pollen assemblage zones and hiatuses in peat accumulation at the two sites: Red Sike Moss, TS I	Weelhead Moss, WH II
0					
1000		VIII	G	G	G
2000					
	FIII	2550		2570 ± 80	
3000				A	3150 ± 100
				3390 ± 90	
4000		VIIb			4000 ± 110
			A	Hiatus	A
5000		5000			5220 ± 120
				Oc	5770 ± 110
			c		
6000	FII	VIIa		6150 ± 160	6202 ± 70
			O b	Ob	O
7000		7050			
			a	Oa	
8000		VI			8070 ± 170
				8250 ± 280	8057 ± 85
			H	H	H
	FI	8750			
9000					
		V	Hiatus	Hiatus	Hiatus
		9450			10 020 ± 210
		IV		9900 ± 190	J
10 000		10 150		J	10 070 ± 190
		III	J		

.... Radiocarbon dated pollen assemblages — — — — Estimated pollen assemblage zone boundaries

Table 8.6 ^{14}C dates from TSI, Red Sike Moss. They were dated at the Gakushuin laboratory (Japan) and the dates were based on the Libby half-life of 5570 ± 30 years (after Turner *et al*., 1973)

Laboratory code	Depth (cm)	Pollen horizon	Age, in radiocarbon years BP (before 1950)
GaK-2027	14	Rise in Gramineae *Calluna* and *Plantago*; beginning of zone G	2570 ± 80
GaK-2028	44	Beginning of zone A	3390 ± 90
GaK-2029	70	Beginning of subzone Oc	6150 ± 160
GaK-2030	120	End of zone H	8250 ± 280
GaK-2031	135	End of zone J	9900 ± 190

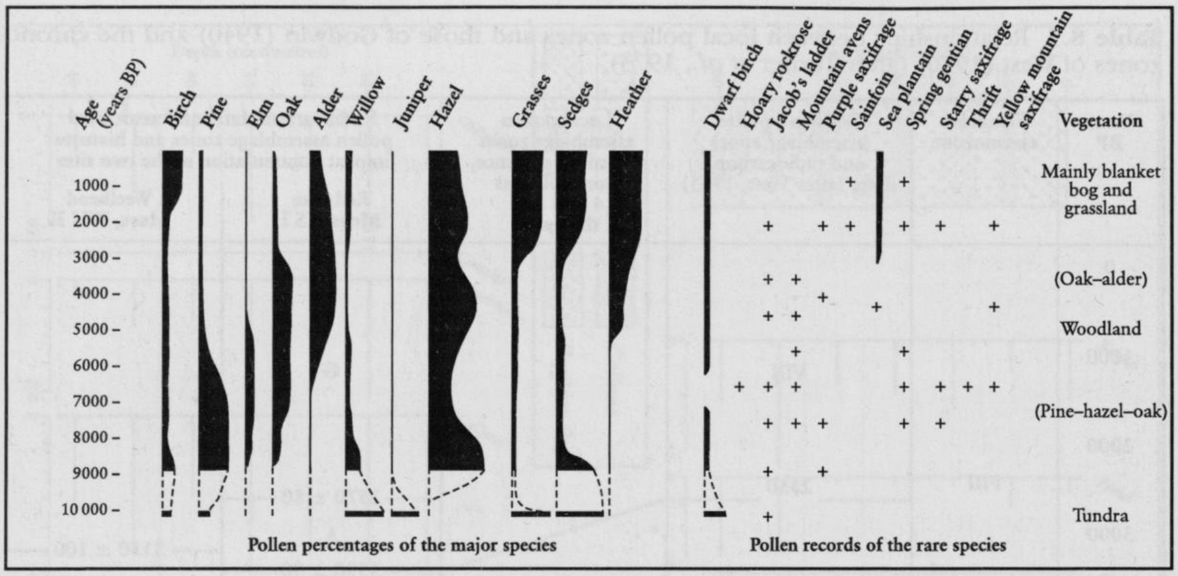


Figure 8.12 Summary pollen diagram from Upper Teesdale (after Turner, 1978).

Red Sike, which indicates that it has increased with the spread of grassland too. *Polygonum viviparum* is recorded in pollen assemblage zone G at Red Sike Moss and zone J at TSI, and seems to have been most abundant when the limestone grassland spread to its maximum extent. *Saxifraga stellaris* has been recorded from RSII in zones J and G, and from RSII and Tinkler's Sike in zone O, so demonstrating its presence in the Late-glacial, at the forest maximum and since the expansion of blanket bog and grassland.

The records for *G. verna*, *Dryas octopetala* (Figure 8.13b), *B. nana*, *Saxifraga azoides* and *S. stellaris*, *H. canum and Thalictrum alpina* from pollen assemblage zone O, the period of forest maximum, confirm that these rare species have been in Upper Teesdale since the Late-glacial to the present day. What is true for them is probably true for the other rare species, even though the diagnostic pollen has not been recorded.

Fossil remains of many of the rarities have been found in many parts of Britain during the Late Devensian and so the Teesdale assemblage is relict. It has been assumed that the present open communities contain those components of the Late Devensian vegetation that persisted in refugia during the Flandrian where positive processes ensured their survival (Piggott and Walker, 1954; Godwin, 1956; Piggott, 1956). However, some of the habitats in which the rarities are found today are the result of ecological change during the Flandrian (Bellamy *et al.*, 1969a, b; Squires, 1970, 1971, 1978; Turner *et al.*, 1973). Minimum competition is a feature of all the communities containing rarities. Their survival during the woodland phase (8000–3000 years BP) depended on the presence of treeless areas, or a low tree density, because many of the rarities are heliophytes and therefore open woodland or woodland with some openings was necessary.

When the blanket peat developed about 5500 years BP there was a critical period for the rarities, all herb values declined because of increasing heather and they were restricted to peat-free areas. After 3000 BP trees were replaced mainly by grassland. At high altitudes in Teesdale, woodland regeneration may have been limited and once trees were eliminated, grassland species were encouraged in the base-rich areas. Tree destruction might have been caused by grazing and browsing of native fauna, such as the aurochs (Johnson and Dunham, 1963), or the red deer (Dimbleby and Simmons, 1974). The grazing intensity may well have increased as the area available decreased with the continued extension of the blanket peat. Increases in the origin and extent of grassland also have been interpreted as a result of prehistoric peoples' activities (Turner *et al.*, 1973). Evidence comes

Figure 8.13 Examples of the Teesdale rarities from modern Arctic habitats: (a) Dwarf birch (*Betula nana*) from Skaftafell National Park, southern Iceland; (b) Mountain avens (*Dryas octopetala*) from Skaftafell National Park, southern Iceland; (c) Scottish Asphodel (*Tofieldia pusilla*) from Morsadalur, southern Iceland. (Photos (a) and (b): Dick Vuijk; Photo (c): D. Huddart.)

from microliths and the remains of domestic cattle (Johnson and Dunham, 1963; Proctor, 1965, 1976) and implicit evidence is present in the pollen diagrams (Squires, 1970, 1971, 1978; Turner *et al.*, 1973), particularly the grass pollen and ruderal pollen, especially ribwort plantain.

During the blanket peat initiation and extension in the mid-Flandrian many refugia disappeared. By contrast removal of woodland promoted the erosion of substrates and one result was the creation of habitats such as those found on the Sugar Limestone. This allowed the migration of rarities from their diminished refugia and resulted in the creation of new communities. Johnson *et al.* (1971) argued that the Sugar Limestone is a subsurface phenomenon formed beneath a thin drift cover or soil, its presence at the surface being the result of subaerial erosion of the overlying sediment. The present Sugar Limestone outcrops are the product of either the past 5000 years (Turner *et al.*, 1973), or the past 3000 years (Squires, 1978). So most of the rarities in the Teesdale assemblage are relict, that is individually they have a history of 10 000–12 000

Table 8.7 Occurrence of Teesdale rarities that possess diagnostic pollen at the species or genus level (after Squires, 1978)

Member of Teesdale assemblage	Geographical element	Taxonomic level of identification	Biostratigraphical position in Teesdale (assemblage zone)
Armeria maritima (Mill.) Willd.	Arctic alpine, maritime	Species	O
Betula nana L.	Arctic alpine	Species	Throughout (J–G)
Dryas octopetala L.	Arctic alpine	Species	O, A
Gentiana verna L.	Arctic alpine, continental	Species	O, G
Plantago maritima L.	Maritime	Species	G
Polemonium caeruleum L.	Continental	Species	Throughout (J–G)
Rubus chamaemorus L.	Arctic alpine, continental	Species	A, G; 3
Saxifraga aizoides L.	Arctic alpine	Species	O, A, G
Polygala amara L.	Continental	Genus	A, G
Polygonum viviparum L.	Arctic alpine	Genus	J, G; 2
Saxifraga stellaris L.	Arctic alpine, continental	Genus	J, O, G; 2
Helianthemum canum (L.) Baumg.	Southern continental	Genus	J, H, O; 2, 3, 4
Thalictrum alpinum L.	Arctic alpine	Genus	Throughout (J–G)
Viola rupestris (Schmidt)	Continental	Genus	O

Schematic Zonation Scheme (Turner *et al.*, 1973)	Possible species other than the one noted
G 3000 BP A 5000 BP O 8000 BP H 10 000 BP J	*P. vulgaris, P. serpyllifolia* *P. bistorta* *S. nivalis, S. tenuis, S. hieraciifolia* *H. chamaecistus* *T. flavum, T. minus* ssp. *minus, T. minus* ssp. *majus* *V. lutea, V. palustris, V. riviniana*

years in the area but the palaeoecological evidence suggests that their present-day distribution is not relict. The rarities are more widespread now than at any other time in their Flandrian history because many of the present-day habitats are recent, or at most 5000 years old. Only those areas such as flushes, springs, quartz-dolerite and limestone screes have been long-standing refugia. These areas must have been crucial in the survival of rare species, particularly during the blanket bog extension. However, at least three species that once belonged to the relict flora have become extinct within comparatively recent times: sainfoin and the purple saxifrage since the spread of grassland and blanket bog, and Jacob's ladder within the past 150 years (Turner *et al.*, 1973). Thus the present widespread distribution of the rarities can be looked at as the latest stage in the following sequence (Squires, 1978).

1. *c.* 10000–8000 years BP: widespread distribution with the predominance of positive processes; i.e. local instability from gelifluction and groundwater fluctuations, minimal competition and base-rich soils and surface water.
2. *c.* 8000–3000 years BP: minimal distribution with the preponderance of negative processes; i.e. soil leaching, tree immigration and blanket peat development restricted rarities to refugia.
3. *c.* 3000 years BP: increased distribution because of increasing positive processes that characterize the modern habitats, which allowed the expansion of refugia; i.e. large-scale and local instability (groundwater fluctuations, erosion especially by wind, trampling and burrowing, minimal competition and human interference).

Conclusions

Red Sike Moss and other sites in Upper Teesdale are very important sites where it is possible to see the changes in the Late-glacial and Flandrian flora and where the complex history of the Teesdale rarities can be established. It is one of the most important locations in Britain for conservation: its rare plant assemblages and its palaeoecological development can be compared with both lowland sites in County Durham, such as Neasham Fen (see site report, this chapter), and higher sites on the Moor House National Nature Reserve, such as Valley Bog (see site report, this chapter).

NEASHAM FEN (NZ 331 115)

D. Huddart

Introduction

Neasham Fen is a small, almost circular, infilled kettlehole, 5.2 km south-east of Darlington. East of the fen a narrow ridge of higher land above 45.7 m OD lies between it and the River Tees meanders. The site provides an important record of Flandrian vegetation history and environmental change, supported by ^{14}C dating. It is a valuable reference location for the north-east England lowlands and allows wider comparisons of vegetation development with the pollen sites from Upper Teesdale and the northern Pennines.

Description

A core for pollen analysis was collected near the centre of the fen by Bartley *et al.* (1976) and the stratigraphy is illustrated in Figure 8.14. The pollen diagram is shown in Figure 8.15 and was divided into five pollen zones by Bartley *et al.* (1976), where values are expressed as percentages of the total tree pollen.

I Characterized by high birch pollen together with some willow, grasses and Cyperaceae. Pine and hazel pollen percentages are low.

II Hazel frequencies rise sharply to reach values of 1020% of the tree pollen sum. Level 590–595 cm, the first with high hazel values, is dated to 9082 ± 90 years BP. Elm pollen values also are high and level 580–585 cm, the first with elm, is dated to 8829 ± 120 years BP. Birch pollen falls as values for oak rise through this zone. The top of the zone is dated to 8202 ± 95 years BP.

III Oak values are higher, whereas those of hazel decline regularly. Elm is less abundant than before and alder pollen is present in small amounts for the first time.

IV Alder values increase as birch frequencies fall, ash and lime are present, although not in

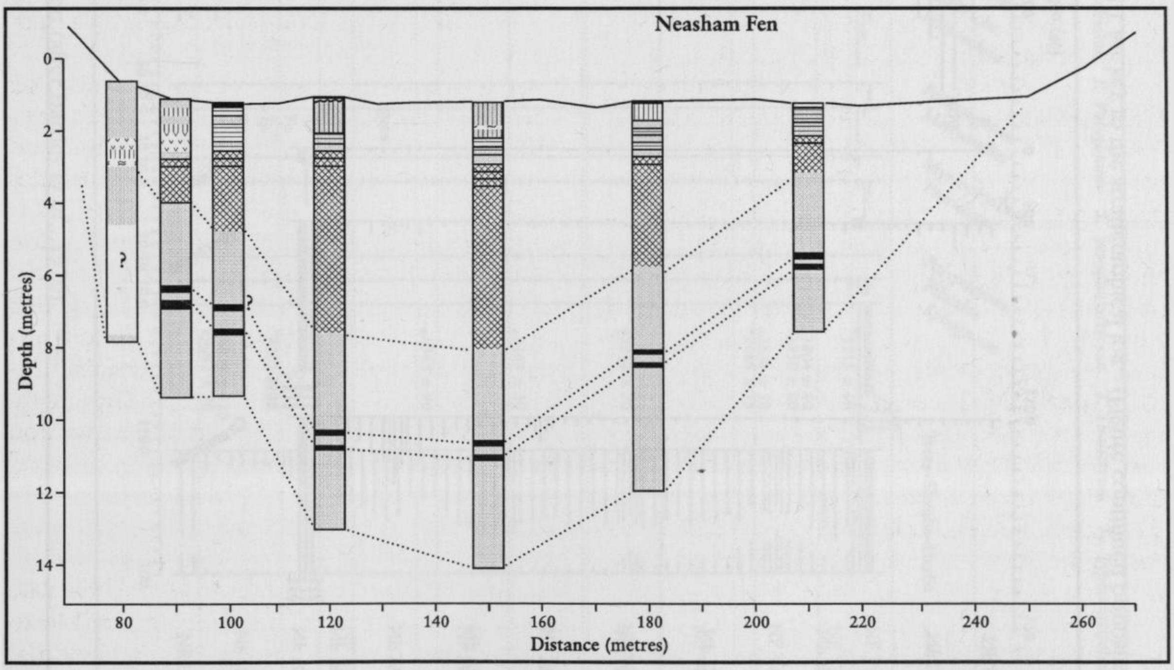


Figure 8.14 Stratigraphy at Neasham Fen (after Bartley *et al.*, 1976). See Figure 8.1 for key to the stratigraphical log.

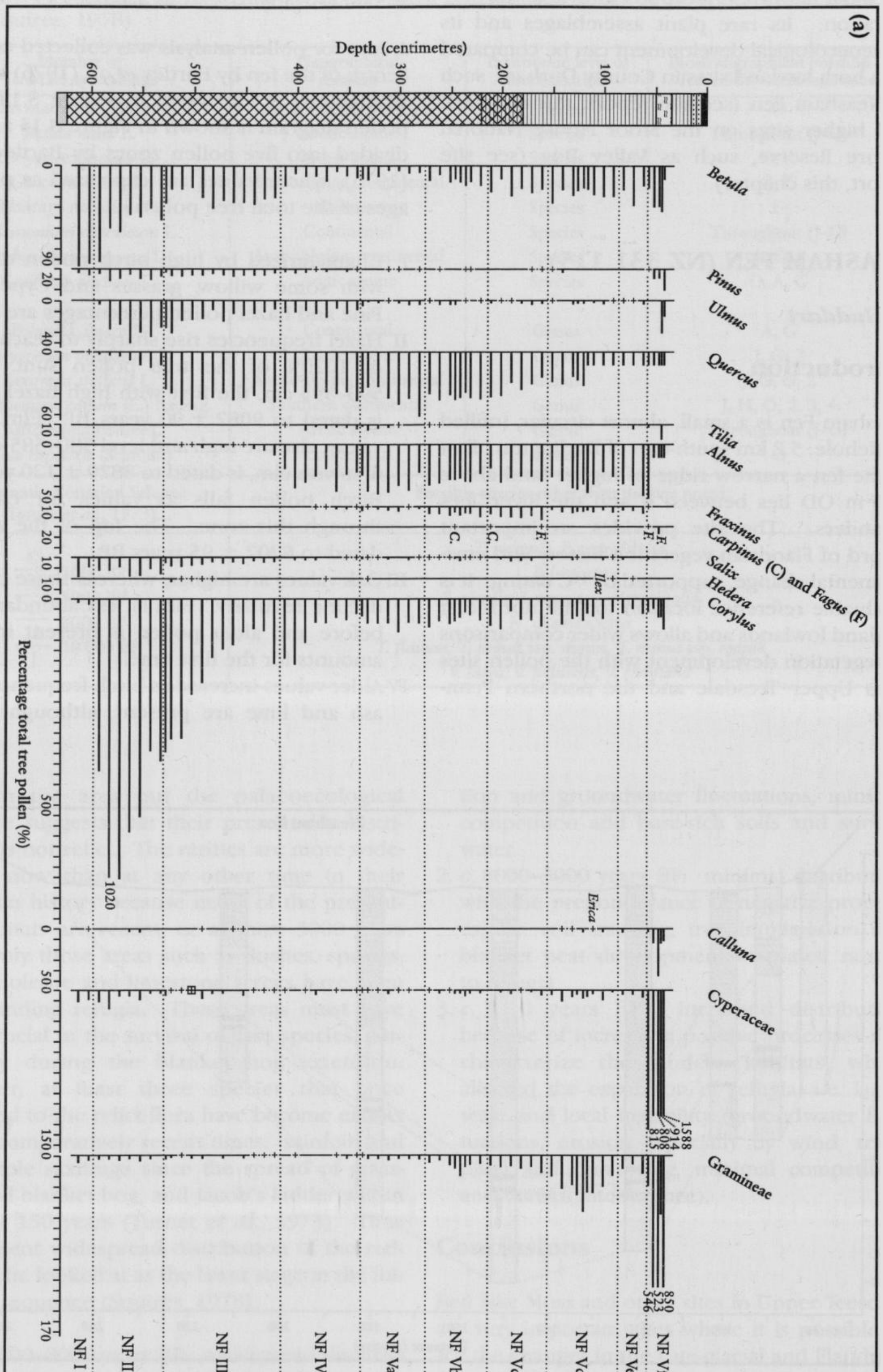


Figure 8.15 Pollen diagram from Neasham Fen (after Bartley *et al.*, 1976). See Figure 8.1 for key to the stratigraphical log. (Figure continued opposite.)

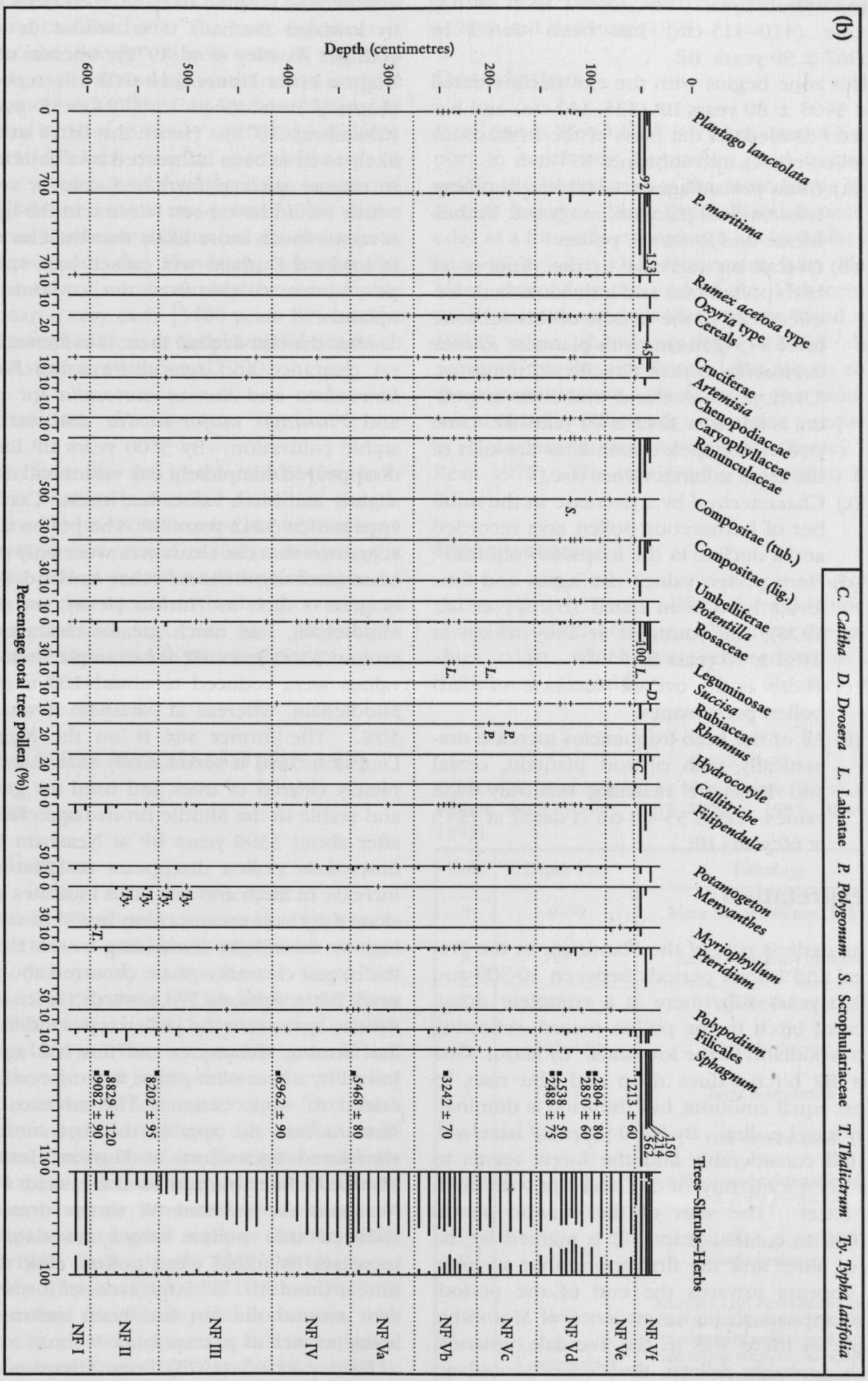


Figure 8.15 (*contd*) Pollen diagram from Neasham Fen (after Bartley *et al.*, 1976).

great abundance. The lowest level of this zone (410–415 cm) has been dated to 6962 ± 90 years BP.

V This zone begins with the elm decline dated at 5468 ± 80 years BP (335–340 cm) and has been divided on the basis of the herbaceous pollen curves into subzones NF Va–f:

V(a) Grass percentages are still low but there is occasional plantain, mugwort, Umbelliferae and Rosaceae pollen.

V(b) Overall an increase in the number of herb pollen taxa and individual herb frequencies. In the middle of this subzone level 245–250 cm, with plantain, *Rumex acetosella*, cereal, Cruciferae, mugwort, Chenopodiaceae and Umbelliferae pollen, is dated at 3242 ± 70 years BP. Lime pollen disappears soon after the start of the zone as birch values rise.

V(c) Characterized by a decrease in the number of herbaceous pollen taxa recorded and a decline in the frequency of each.

V(d) Herb pollen values rise again and four levels have been dated (Bartley *et al*., 1976), the youngest at 140–145 cm at 2488 ± 70 years BP.

V(e) There is an overall decrease of herb pollen percentages.

V(f) All of the herb frequencies increase dramatically, with ribwort plantain, cereal and Rosaceae reaching relatively high values. Level 55–60 cm is dated at 1213 ± 60 years BP.

Interpretation

In the earliest part of the Flandrian, in the Pre-Boreal and Boreal periods between 10 300 and *c*. 7000 years BP), there is a complete dominance of birch in the pollen record, reflecting birch woodland in the lowlands. By about 9000 years BP birch values drop and elm rises to almost equal amounts, but the zone is dominated by hazel pollen. By 8200 years BP hazel values fell considerably and the forest seems to have been a mixture of oak and elm with birch and hazel. The start of the Atlantic period (*c*. 7000 to *c*. 5000 years BP) is marked by the rise of alder and the first appearance of lime. Ash appears towards the end of the period. There appears to be no evidence of Mesolithic people as there was in the Teesdale uplands, which probably reflects the denser woodland and swampy conditions in the Tees valley. The elm decline is dated to 5468 ± 80 years BP and in lowland Durham it is similar or slightly younger (Bartley *et al*.,1976), whereas at Valley Bog on Moor House (see GCR site report, this chapter) it is dated to 4794 ± 55 years BP (Chambers, 1974). Hence the latter site is not likely to have been influenced by a deterioration in climate such as increased cold or wetness, which would have been felt first in the uplands. It seems much more likely that the elm decline in lowland Durham was caused by a spread of new human culture from the lowlands to the uplands.

After the elm decline there is evidence for forest clearance and agriculture using *Plantago lanceolata* and *Rumex acetosella* for pasture and *Plantago major–media* and cereal for arable cultivation. By 3500 years BP lime had disappeared completely, oak values had declined slightly and birch values had risen. Cereal first appeared at 3242 years BP. The pollen diagram suggested that the clearances were only of moderate intensity, whereas other lowland Durham diagrams, such as Hutton Henry and Bishop Middleham, had much greater clearances. At around 3360 years BP, for example, tree pollen values were reduced to about 10% at Bishop Middleham, whereas at Neasham it was about 50%. The former site is on the Magnesian Limestone and it seems likely that it was completely cleared of trees and used for grassland and arable in the Middle Bronze age. However, after about 3300 years BP at Neasham Fen, *P. lanceolata* pollen disappears and there is an increase in birch and ash. This indicates a recession of agriculture and colonization of the clearings by these light-demanding trees. However, the largest clearance phase occurred about 2800 years BP in subzone IVd towards the end of the Bronze age, when the pollen types indicate similar farming techniques and this was again followed by a recession phase when forests regenerated to some extent. The advance of the Romans into the area in the first century AD stimulated agriculture at Hutton Henry and Thorpe Bulmer but not so at Neasham Fen. At Neasham the subzone Vf shows dramatically reduced tree pollen values associated with increases in arable weed pollen. For the first time around AD 737 large areas of forest in the area around the fen had been destroyed for arable as well as pasture land.

Bartley *et al*. (1976) drew attention to the striking difference in the pattern of vegetation change between the sites in the lowlands of south and east Durham. The basic difference is between the poorly drained soils of the Tees lowlands, as illustrated by Neasham Fen, and the well-drained soils of the East Durham Plateau, as illustrated by Hutton Henry and Thorpe Bulmer. Bishop Middleham owes its special features to the close proximity of the Magnesian Limestone and well-drained sands and gravels. This shows that there has been soil control on the vegetation throughout the Flandrian, whereby the best-drained soils and possibly the most fertile were cleared very early (3400 years BP) but the badly drained soils, as at Neasham, did not reach a similar stage until about AD 700.

Conclusions

Evidence from this site in the middle Tees lowlands has enabled a detailed Flandrian vegetation record to be reconstructed, which has been ^{14}C dated. It has proved useful to compare the Neasham pollen record and the influence by humans on this record in both other areas of lowland Durham, as on the Magnesian Limestone Plateau and in the northern Pennines and Upper Teesdale.

MERE SANDS WOOD (SD 448 157)

R.C. Chiverrell

Introduction

Mere Sands Wood provides the best exposures of an extensive (200 km^2) periglacial aeolian deposit (coversand) in south-west Lancashire (Figure 8.16) (Godwin, 1959; Tooley and Kear, 1977; Wilson *et al.*, 1981; Innes *et al.*, 1989; Bateman,1995). The coversands are of Late Devensian age and referred to as the 'Shirdley Hill Formation' (Thomas, 1999). The thick coversands provide evidence for aeolian sedimentation during the Late Devensian, and are overlain by a complex sequence of Holocene sands, and organic sand and muds that have yielded palaeoecological remains (Baxter, 1983; Tooley, 1985; Innes *et al.*, 1989; Middleton *et al.*, 2001). Travis (1909) and Gresswell (1953) first described the Shirdley Hill Formation, with the Mere Sands Wood sequence first investigated by Tooley and Kear (1977) and Tooley (1985). The site subsequently featured in studies by Wilson *et al.* (1981), Baxter (1983), Innes *et al.* (1989) and Bateman (1995).

Description

Mere Sands Wood SSSI is 10 km east of Southport in south-west Lancashire, and is located within the boundaries of the Mere Sands Wood Nature Reserve. Mere Sands Wood is near the edge of a formerly extensive (6 km by 3 km) lowlying (2.7–3.4 m OD) freshwater lake, Martin Mere, drained at the end of the 17th century. Sand was extracted at Mere Sands Wood until 1982, when the Lancashire Wildlife Trust acquired the site and flooded the pits to create the Nature Reserve. Sections in the sand-pits reveal Shirdley Hill Sands underlying a sequence of Holocene organic sediments (Tooley and Kear, 1977). The stratigraphy is listed in Table 8.8.

Interpretation

Gresswell (1957) identified a mid-Holocene shoreline (6000–5000 years BP) near the Lancashire coast, which was referred to as the 'Hillhouse Coastline'. The Shirdley Hill

Table 8.8 Stratigraphy at Mere Sands Wood (after Baxter, 1983; Tooley, 1985; Wilson, 1985; Bateman, 1995).

Unit	Depth (cm)	Lithology
9	0–90	Mere Sands (Wilson, 1985)
8	90–98	Sandy substantia humosa
7	98–105	Fine detrital mud
6	105–139	*Turfa herbaceae*
5	139–140	*Turfa menyanthis*
4	140–141	Fine detrital mud
3	141–157	Fine–sandy detrital mud and *Turfa herbaceae*
2	157–160	Fine detrital mud and *Turfa herbaceae*
1	160– Locally up to 5 metres thick	Shirdley Hill Formation: loose fine to medium moderately to moderately well sorted sands displaying weak cross-bedding and cryoturbation structures

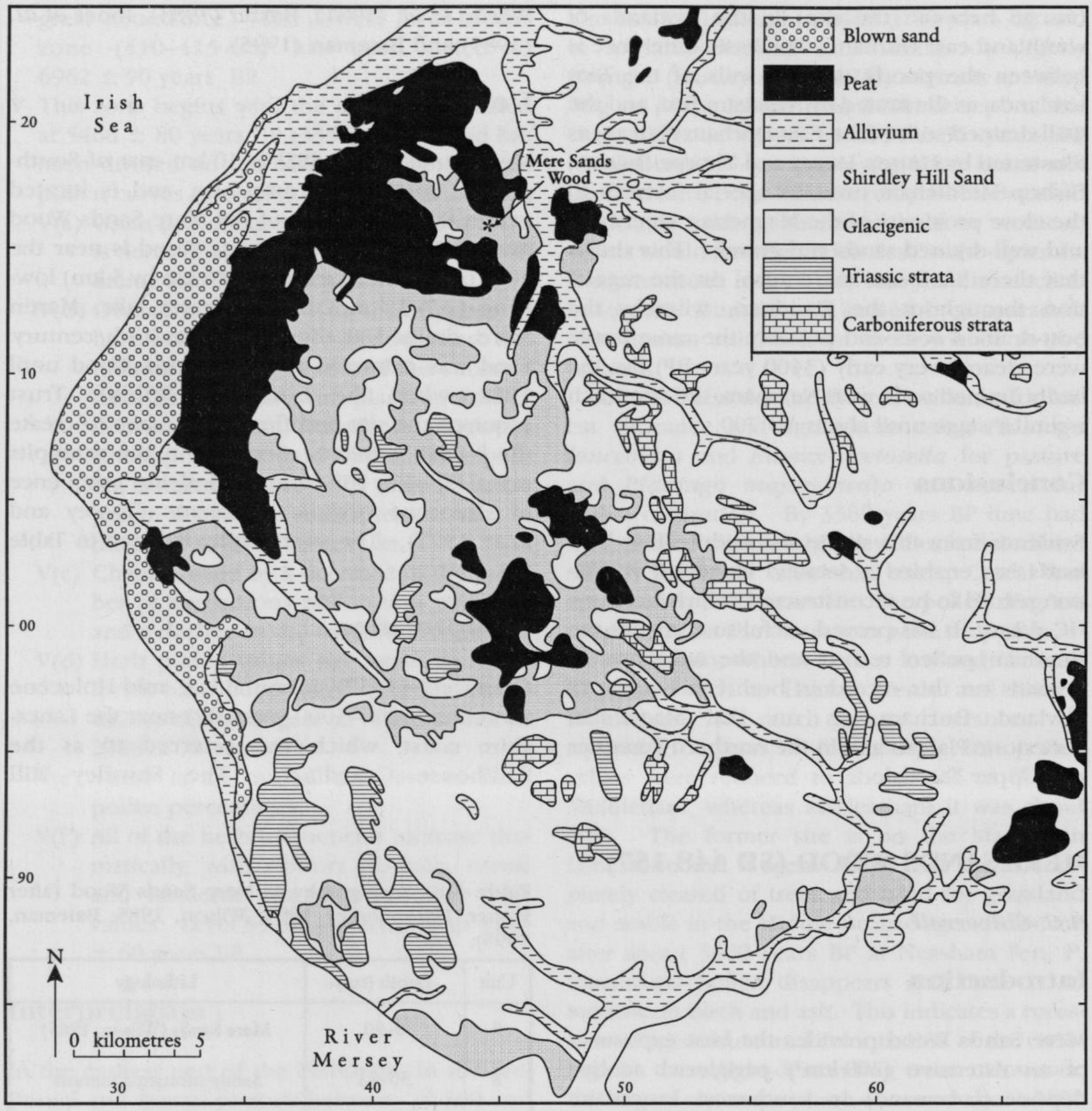


Figure 8.16 Surface geology of south-west Lancashire, showing the distribution of Shirdley Hill Sand (from Wilson *et al.* 1981).

Formation (unit 1, Table 8.8) was identified as relict beach sands associated with this coastline, but re-evaluation of the evidence has radically altered this palaeoenvironmental interpretation. The Shirdley Hill Formation is typically less 1 m but locally can exceed 5 m in thickness, and these sands form a discontinuous cover over Kirkham Formation Devensian glacial sediments (Thomas, 1999). Godwin (1959) identified that the Shirdley Hill Formation overlay Windermere interstadial organic sediments and were covered by earliest Holocene organic muds and peat at Moss Lake in Liverpool. Godwin suggested the sands were deposited by aeolian processes forming coastal dunes during the Late Devensian. Kear (1977) used the uniform particle-size distribution of the Shirdley Hill Formation to indicate an aeolian origin, and suggested that the sediments were derived from local glaciofluvial material. The cryoturbation features that characterize the basal Shirdley Hill Formation also indicate a cold periglacial climate and suggest that the sands accumulated during the Late Devensian (Tooley, 1985).

Wilson *et al.* (1981) used particle-size analysis, scanning electron microscopy of quartz grains and structures in the sediments at Mere Sands Wood to confirm an aeolian origin for the Shirdley Hill Formation. Mineralogical analyses of unweathered Shirdley Hill Formation and several different potential source materials demonstrate that the coversands closely resemble glaciofluvial sediment. Wilson *et al.* (1981) suggested that the Shirdley Hill Formation consists of wind-blown sands derived from outwash sediments left by the Devensian ice sheet. Bateman (1995) correlates the Shirdley Hill Formation with the European coversand chronology presented by Koster (1988) using three thermoluminescence (TL) dates for the sequence at Mere Sands Wood (Figure 8.16). The TL dates on quartz grains confer an age of 11 730 ± 1510 years ago on the upper 20–30 cm of the Shirdley Hill Formation. A ^{14}C date of 10455 ± 100 years BP on organic sediments overlying the Shirdley Hill Formation at Clieves Hill, 10 km south-west of Mere Sands Wood, constrains deposition of the Shirdley Hill Formation to the Late Devensian. Two further TL dates on sand yielded ages of 8740 ± 2060 years BP and 6940 ± 1110 years BP, which are at equivalent depths to units 3 and 7–8 respectively (Bateman, 1995).

The organic sediments at Mere Sands Wood reflect the accumulation of plant-rich detritus and peat in an aquatic to peatland environment on the edge of Martin Mere (Tooley, 1985). Pollen diagrams (Figure 8.17) have been produced for the organic sediments that overlie the Shirdley Hill Formation (Baxter, 1983; Tooley, 1985). The pollen succession in units 2, 6, 7 and 8 overlying the Shirdley Hill Formation is an early Holocene sequence (Figure 8.17). *Betula*, *Pinus* and *Populus* dominate units 2 and 6 before declining as *Quercus*, *Ulmus* and then *Alnus* increase in abundance to dominate units 7 and 8 (Baxter, 1983; Tooley, 1985). The palynology indicates that organic sedimentation began at the beginning of the Holocene Epoch, a view supported by a ^{14}C date of 10 455 ± 100 years BP on equivalent organic sediments at Clieves Hill. Organic sedimentation ceases shortly after the *Alnus* rise (unit 7), which is ^{14}C dated at Red Moss to 7107 ± 120 years BP (Hibbert *et al.*, 1971).

The organic sediments are covered by a further sequence of sands, the Mere Sands (Wilson, 1985), and three explanations exist for the sand's origin and mode of deposition.

1. Lacustrine sands derived from glaciofluvial material and the Shirdley Hill Formation (Crompton, 1966).
2. Wind-blown reworked Shirdley Hill Formation (Tooley and Kear, 1977).
3. Wilson *et al.* (1981) found that the Mere Sands differed substantially from the Shirdley Hill Formation and had a greater affinity with modern beach and dune material on the Lancashire coast.

The pollen spectra in units 4–5 are out of sequence, containing abundant *Quercus*, *Ulmus* and *Alnus* before the early Holocene expansion of these species. Tooley (1985) invokes a scenario with units 3–5 (2.5–2.6 m OD) deposited when unit 6 was floated up as Martin Mere expanded as a freshwater lake. There is evidence for a similar event affecting altitudes of 3.1–3.2 m OD at Tarlscough Moss, a peat sequence on the edge of Martin Mere (McAllister, 2001). The timing for this floating is uncertain, although it has been correlated with higher lake levels in Martin Mere forced by higher local water tables during marine incursions dated to 6890 ± 55 and 6790 ± 95 years BP at Downholland Moss (Tooley, 1974, 1978a, 1985; Huddart, 1992). Organic sedimentation at Mere Sands Wood ceases around 7000 years ago.

Conclusions

Mere Sands Wood is an important reference site for studies of the Shirdley Hill Formation, and for elucidating the geomorphology and environmental history during the Late Devensian and the Holocene in south-west Lancashire. Mere Sands Wood is a former pit from which sand was extracted for glass making up until 1982, when the site was flooded to form the present nature reserve. Currently there is little exposure of the stratigraphy owing to the flooding and natural degradation of the faces. Trenching of drainage ditches reveals the organic sequence and the uppermost metre of the Shirdley Hill Formation. Nevertheless Mere Sands Wood offers some of the best exposures of the Shirdley Hill Formation. The sands were reworked by aeolian processes from glaciofluvial sediments under a periglacial climatic regime during the Late Devensian. This Late Devensian periglacial

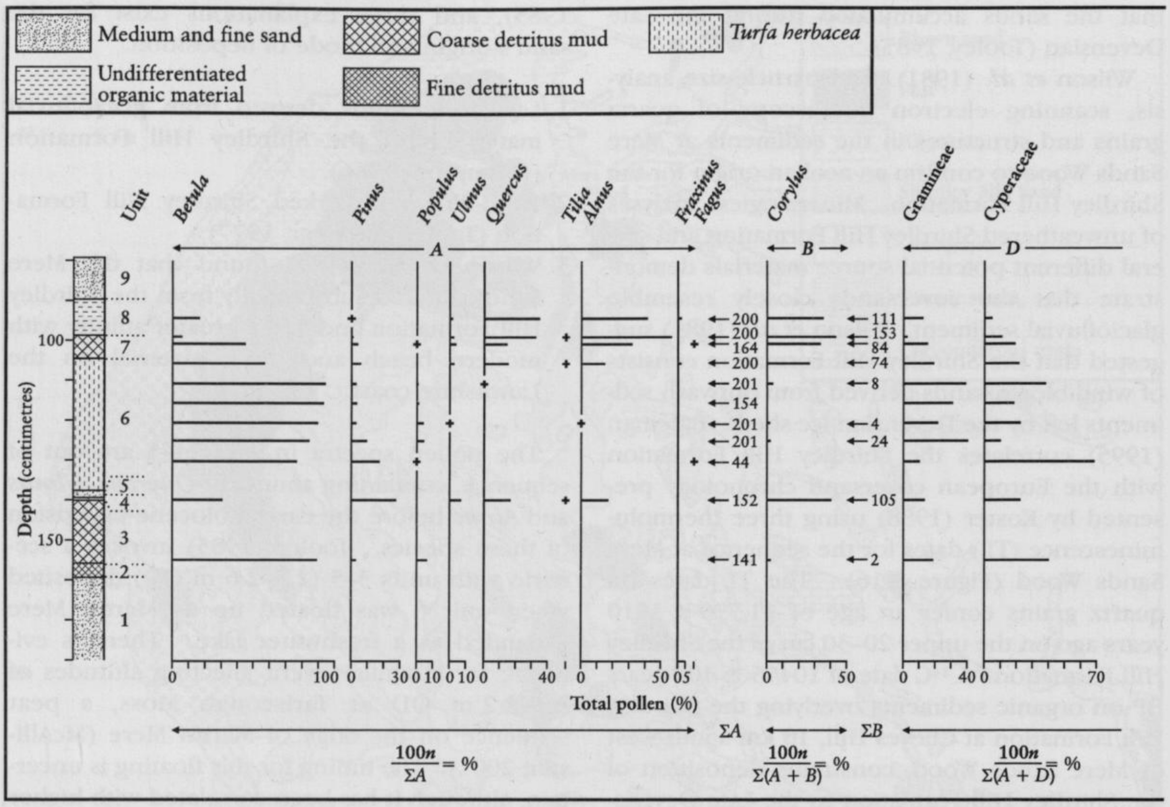


Figure 8.17 Pollen diagram from Mere Sands Wood (after Tooley, 1985).

Figure 8.17 (*contd*) Pollen diagram from Mere Sands Wood (after Tooley, 1985).

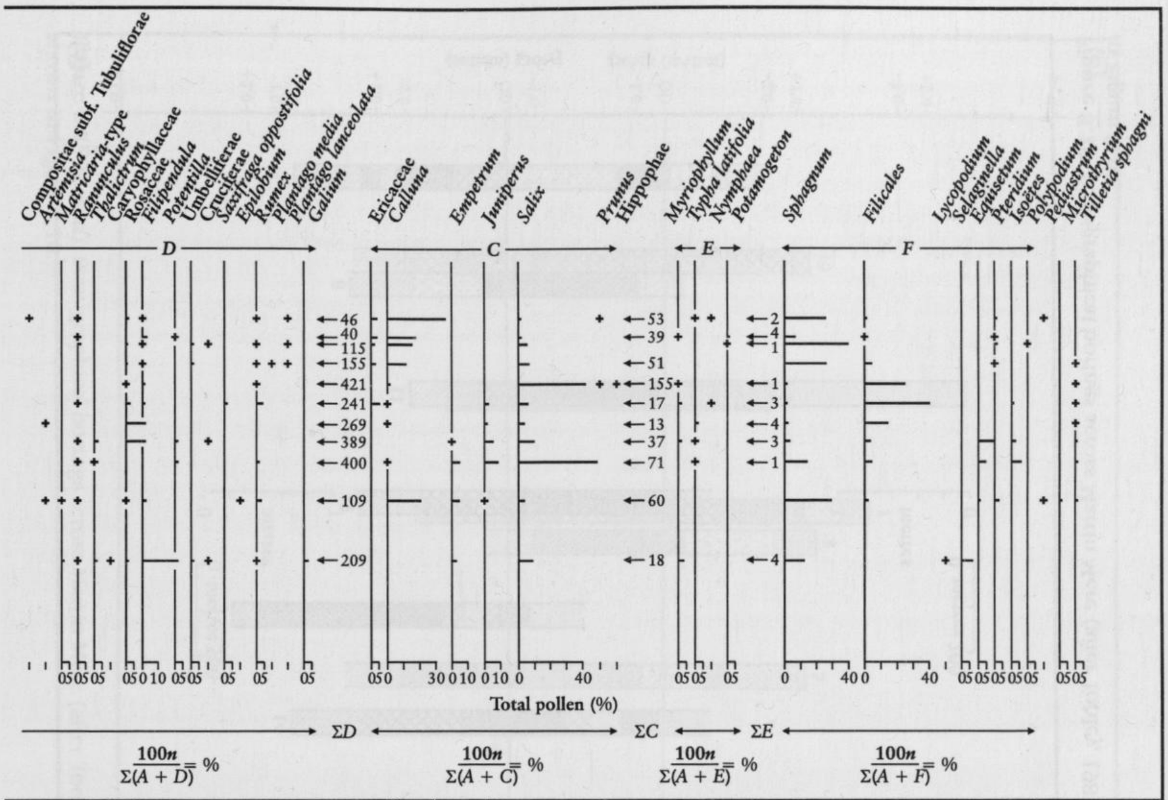
coversand is luminescence dated to 13240–10220 years BP and is analogous to the Younger Coversand II in the European chronology (Koster, 1988; Bateman, 1995). The palaeo-ecology of the organic sediments provides useful information about the early Holocene in south-west Lancashire.

MARTIN MERE (SD 448 157)
POTENTIAL GCR SITE

R.C. Chiverrell

Introduction

Martin Mere is a formerly extensive freshwater lake that is currently a low-lying flat area of reclaimed peatlands and open water that sustain organic sediments ranging from lacustrine mud to terrestrial raised mire peat (Figure 8.18). Sediments from locations across the mere have yielded palaeoecological data that have contributed to an understanding of Holocene vegetation and sea-level changes in south-west Lanca-



shire. Brodrick (1903) first studied the sediments at Martin Mere, identifying a sequence of either estuarine or lacustrine clays and silts. Gresswell (1957) regarded these silts and clays as estuarine sediments deposited in a coastal embayment during the 'Hillhouse Coastline' high sea-level event. Subsequent research has targeted the deposits at several sites across Martin Mere using field stratigraphy, plant macrofossil analysis and pollen analysis to elucidate a history of environmental change during the early Holocene (Tooley, 1977, 1985).

Description

Martin Mere is located 8 km east of Southport in south-west Lancashire. It was an extensive (690 ha) freshwater lake prior to drainage during the 17th century, with the water surface at 2.7–3.4 m OD and maximum depths of *c.* 6 m (Brodrick, 1903). The margins of Martin Mere prior to drainage in AD 1692 reveal the extent of the lake (Figure 8.18). The lake deposited an extensive sequence of unconsolidated sediments reaching a maximum of 7.5 m, but rarely exceeding 3 m of organic sediment (Tooley, 1985). The stratigraphy of Martin Mere has been investigated in some detail with coring across the tidal and lagoonal zone near Churchtown

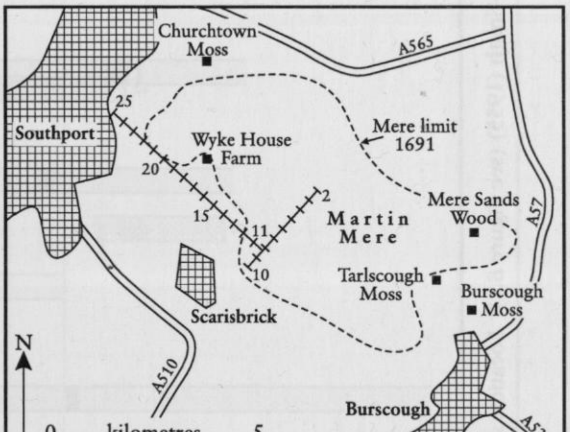


Figure 8.18 Map of Martin Mere, with location of core transects shown in Figure 8.19.

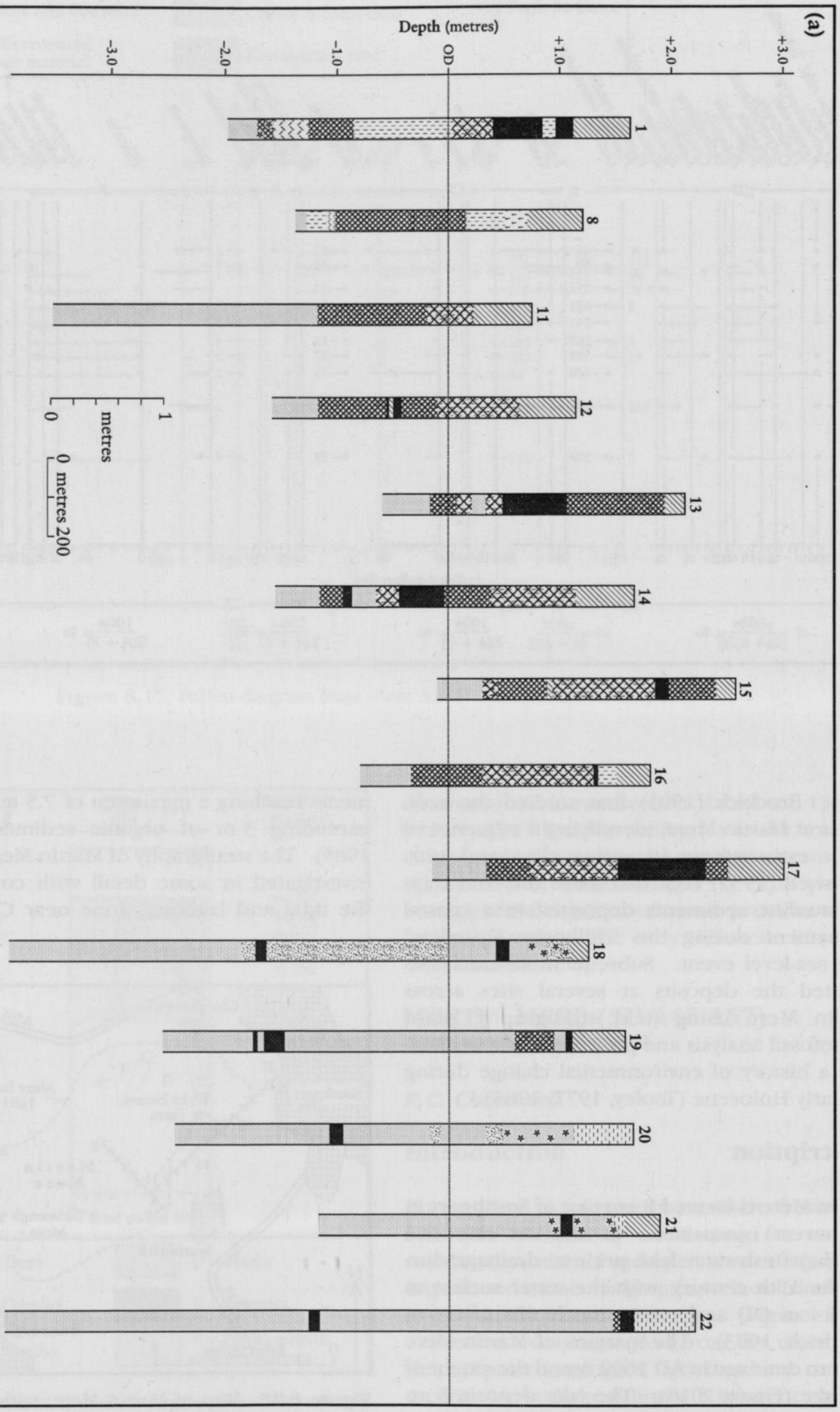


Figure 8.19 Stratigraphical borings across Martin Mere (after Tooley, 1985). Symbols are according to Troels-Smith (1955) (see Figure 8.1). Location shown in Figure 8.18.

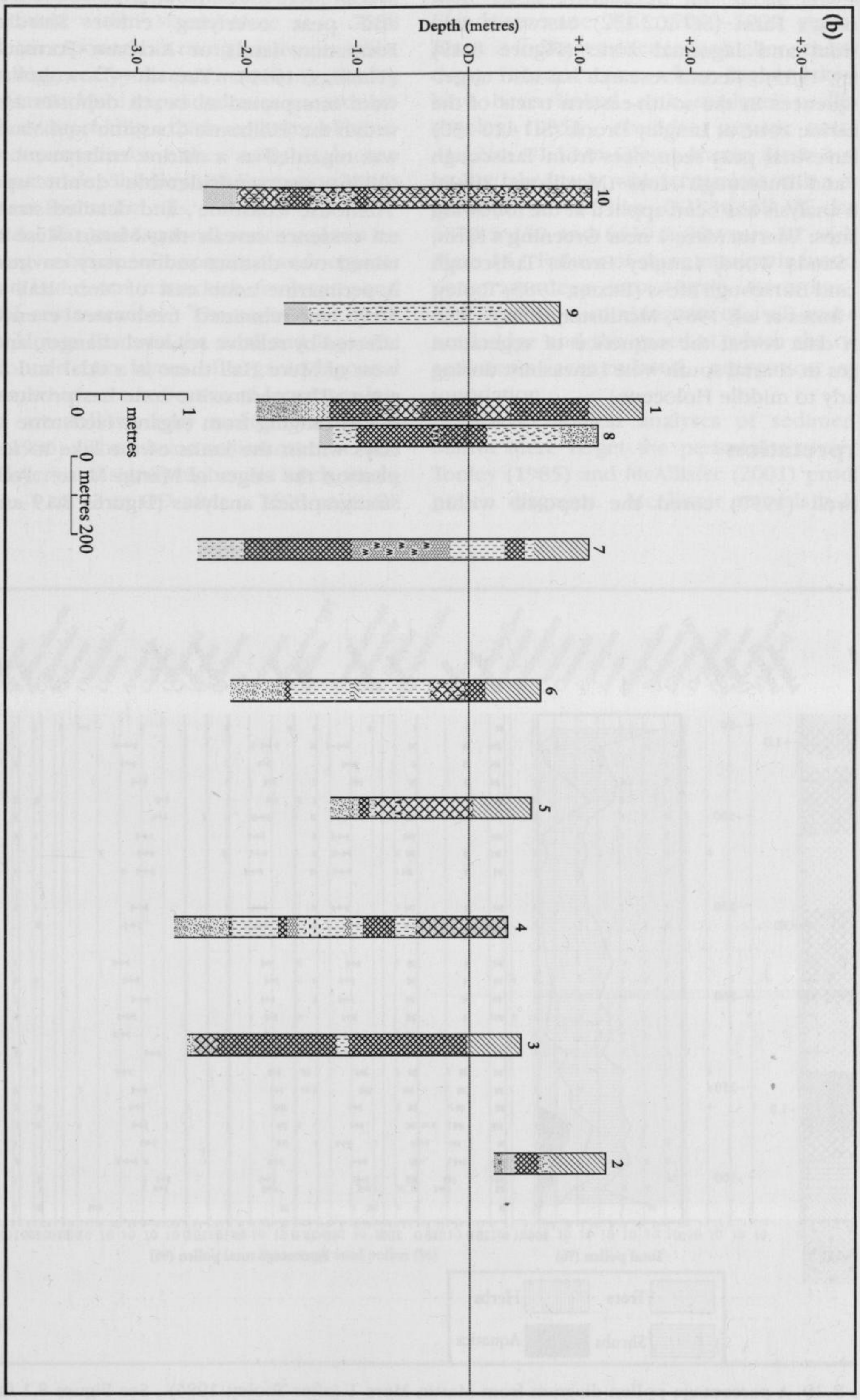


Figure 8.19 *(contd)* Stratigraphical borings across Martin Mere (after Tooley, 1985). Symbols are according to Troels-Smith (1955) (see Figure 8.1). Location shown in Figure 8.18.

Moss and from the perimarine zone near Greening's Farm (SD 402 152) eastwards into the tidal and lagoonal zone (Figure 8.19) (Tooley, 1985). Recent research has also targeted sequences in the south-eastern tracts of the perimarine zone at Langley Brook (SD 410 130) and terrestrial peat sequences from Tarlscough Moss and Burscough Moss (McAllister, 2001). Pollen analysis has been applied at the following core sites: Martin Mere 1 near Greening's Farm, Mere Sands Wood, Langley Brook, Tarlscough Moss and Burscough Moss (Baxter, 1983; Tooley, 1985; Innes *et al.* 1989; McAllister, 2001). The pollen data reveal the sequence of vegetation changes in coastal south-west Lancashire during the early to middle Holocene.

Interpretation

Gresswell (1957) cored the deposits within Martin Mere and encountered sequences of silt and peat overlying either Shirdley Hill Formation sands or Kirkham Formation tills (Thomas, 1999). The silts (Downholland Silt) were interpreted as beach deposits associated within the 'Hillhouse Coastline' and Martin Mere was regarded as a marine embayment. Tooley (1976) cast considerable doubt upon the 'Hillhouse Coastline', and detailed stratigraphical evidence reveals that Martin Mere has sustained two distinct sedimentary environments. A perimarine zone east of Mere Hall (SD 404 165) has remained freshwater even though affected by relative sea-level changes, and to the west of Mere Hall there is a tidal and lagoonal zone. The perimarine zone has produced sediments ranging from organic lacustrine silts and clays within the limits of the lake to terrestrial peat on the edges of Martin Mere. Pollen and stratigraphical analyses (Figures 8.19 and 8.20)

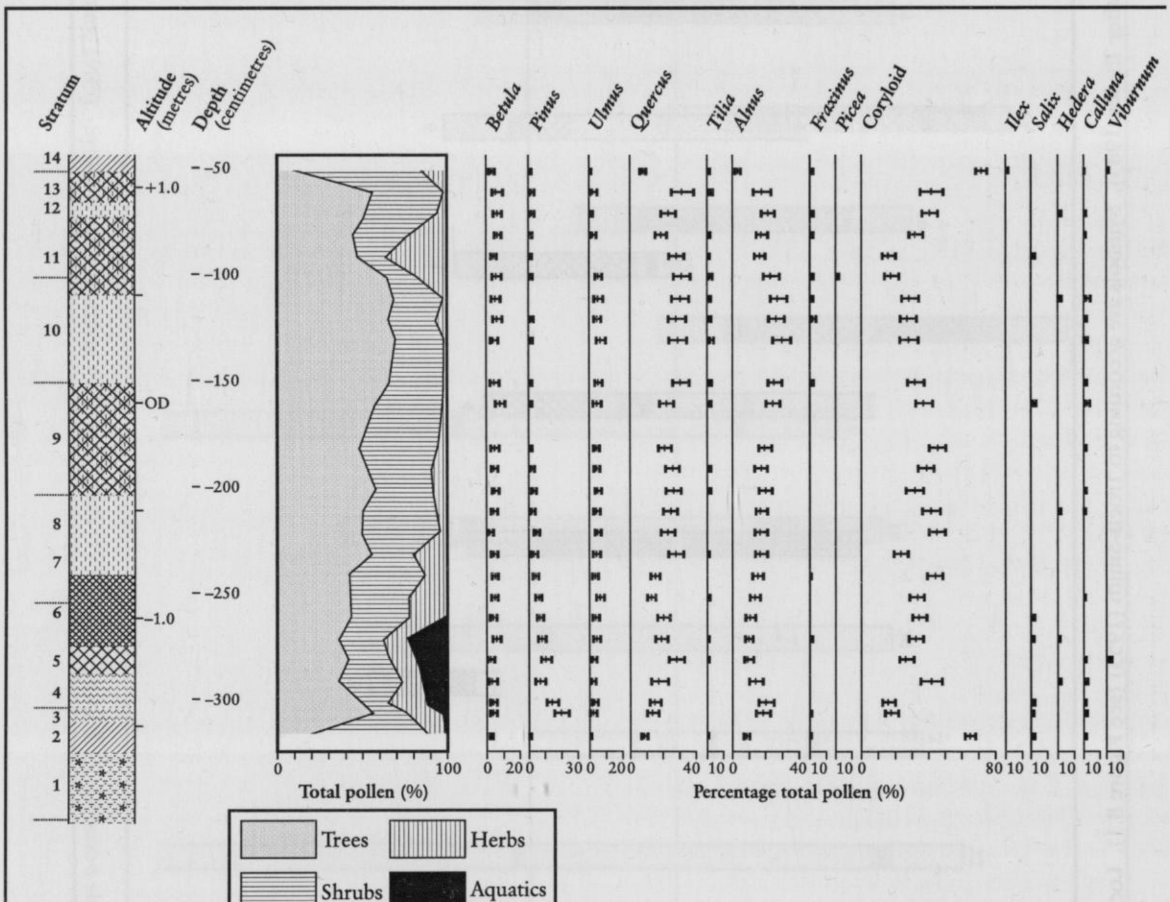
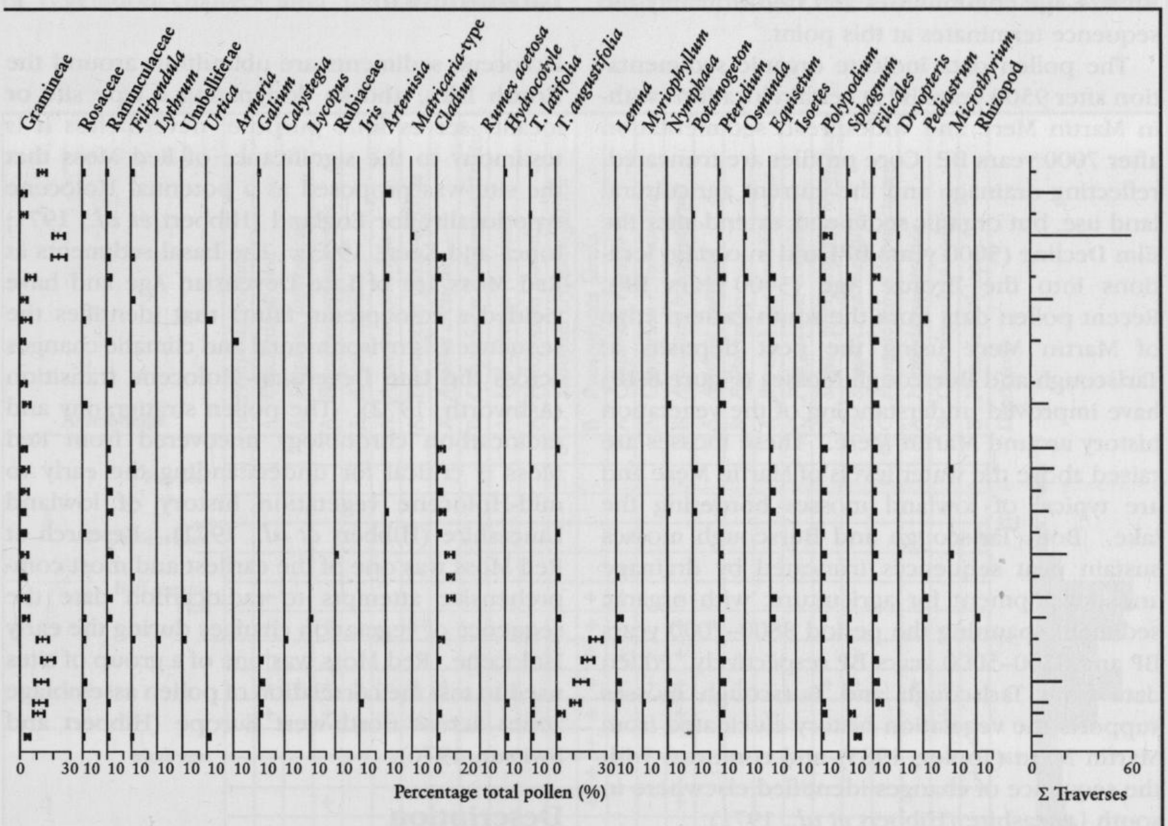


Figure 8.20 A percentage pollen diagram from Martin Mere 1 (after Tooley, 1985). See Figure 8.1 for key to stratigraphical log.

identify the widespread inception of organic sedimentation within Martin Mere at *c.* 7000 years BP, in response to a rising freshwater table, perhaps triggered by marine incursions (Tooley, 1985). Martin Mere expanded over an undulating surface, depositing a thick sequence of lacustrine sediments over either early Holocene organic materials, Shirdley Hill Formation sands or Kirkham Formation tills.

The tidal flat and lagoonal zones contain an alternating mixture of inorganic and organic sediments laid down in brackish, freshwater and terrestrial environments. The sediments north of Wyke House Farm in the tidal and lagoonal zones contain two layers of organic lagoonal sediment, which are indicative of lower sea levels and are followed by marine incursions (Tooley, 1985). The marine sediments cover the organic layers and signify higher sea levels reaching over –1 to –2 m OD after *c.* 7600 years BP and over 0.5 to 1.5 m OD in the middle of Flandrian zone Fl-II (7000–5000 years BP). There is indirect evidence for higher sea levels in the perimarine zone, where higher lake levels have been linked with sea-level oscillations (Tooley, 1985). Abundant aquatic taxa from –1.42 to –1.12 m OD at Martin Mere 1 signify higher lake levels, which correlate with a marine incursion to –1.09 to –0.23 m OD ^{14}C dated to 6870 ± 130 and 6430 ± 90 years BP at Martin Mere 14. Failure of these marine incursions to penetrate deep into Martin Mere has been explained by a combination of a till and coversand ridge and the speed at which the organic sediments accumulated, preventing marine inundation.

Palaeoecological analyses of sediments in Martin Mere target the perimarine zone, with Tooley (1985) and McAllister (2001) producing pollen diagrams. McAllister records a longer



early Holocene sequence at Langley Brook, which pre-dates the widespread expansion of Martin Mere. The basal sediments are dominated by *Betula* and *Pinus* woodland, with the arrival of *Corylus avellana* around 9500–8500 years BP. *Quercus*, *Ulmus* and *Alnus* are the dominant components of the mixed deciduous forest that replaces *Pinus* and *Corylus avellana* woodland at 8080 ± 160 years BP. The sequence from Martin Mere 1 (Figure 8.20) post-dates the expansion of the lake and begins shortly after the *Alnus* rise *c.* 7000–8000 years BP (Tooley, 1985). Towards the top of Martin Mere 1 there is an undated decline in *Ulmus* frequencies and at Langley Brook a decline in *Ulmus* frequencies has yielded a ^{14}C date equivalent to the 5010 ± 80 years BP recorded for the Elm Decline at Red Moss (Hibbert *et al.*, 1971; Tooley, 1985). The record from Martin Mere 1 stops at this point, but at Langley Brook the sequence contains charcoal remains and pollen taxa indicative of arable and pastoral activity, probably by Bronze Age communities and unfortunately the sequence terminates at this point.

The pollen data indicate organic sedimentation after 9500 years BP at certain locations within Martin Mere and widespread sedimentation after 7000 years BP. Core profiles are truncated, reflecting drainage and the current agricultural land use, but organic sediments extend after the Elm Decline (5000 years BP) and in certain locations into the Bronze Age (3400 years BP). Recent pollen data from the south-eastern edge of Martin Mere using the peat deposits at Tarlscough and Burscough Mosses (Figure 8.18) have improved understanding of the vegetation history around Martin Mere. These mosses are raised above the water levels of Martin Mere and are typical of lowland mosses bordering the lake. Both Tarlscough and Burscough mosses sustain peat sequences truncated by drainage and development for agriculture, with organic sediment spanning the period 8800–7000 years BP and 8200–5000 years BP respectively. Pollen data from Tarlscough and Burscough mosses supports the vegetation history elucidated from Martin Mere (Tooley, 1985) and conforms with the sequence of changes identified elsewhere in south Lancashire (Hibbert *et al.*, 1971).

Conclusions

Martin Mere is an important site for elucidating the geomorphology and environmental history during the Holocene Epoch in south-west Lancashire. Martin Mere is a former lake drained in AD 1692, which has yielded important data on the sea-level and vegetation history of the area. The volume of stratigraphical data provides particularly detailed evidence of the evolution of adjacent tidal-flat, lagoonal and perimarine environments, making this site of crucial importance. Detailed pollen records demonstrate the sequence of woodland colonization during the early Holocene. Today Martin Mere is a shadow of it's former glory, drained and utilized for agriculture with little open water remaining and the sediments are buried beneath gley podsols with peaty or humose topsoils.

RED MOSS (SD 634 100)

POTENTIAL GCR SITE

R.C. Chiverrell

Introduction

Holocene sediments are ubiquitous around the British Isles, and so designating a type site or locality serves little purpose; nevertheless it is testimony to the significance of Red Moss that the site was proposed as a potential Holocene type locality for England (Hibbert *et al.*, 1971; Jones and Keen, 1993). The basal sediments at Red Moss are of Late Devensian Age and have yielded a coleopteran fauna that identifies the sequence of environmental and climatic changes across the Late Devensian–Holocene transition (Ashworth, 1972). The pollen stratigraphy and radiocarbon chronology uncovered from Red Moss is critical for understanding the early to mid-Holocene vegetation history of lowland Lancashire (Hibbert *et al.*, 1971). Research at Red Moss was one of the earliest and most comprehensive attempts to radiocarbon date the sequence of vegetation changes during the early Holocene. Red Moss was one of a group of sites used to test the correlation of pollen assemblage zones across north-west Europe (Hibbert and Switsur, 1976).

Description

Red Moss is located on the outskirts of Horwich in south-east Lancashire. Construction of the M61 motorway has damaged the western flank and deepest tracts of the Red Moss, a railway complex impinges on the eastern flanks of the mire and there is a history of peat cutting at the site. The present-day flora does not resemble that of an intact raised mire, but is dominated by grasses and occasional clumps of *Calluna* and *Erica*. Sediment accumulation initiated in a hollow on the watershed of the Croal and Douglas rivers. Hibbert *et al.* (1971) and Ashworth (1972) described and sampled the stratigraphy in the north-western and deepest tract of Red Moss and 18 radiocarbon dates were obtained, providing one of the best-dated sediment successions in the British Isles (Hibbert *et al.* 1971; Ashworth, 1972). Coleopteran fauna within the Late-glacial sediments also have attracted the attention of researchers, identifying the decline of temperatures after the thermal maximum of the Windermere Interstadial into the Loch Lomond Stadial and the subsequent climatic amelioration into the Holocene (Ashworth, 1972; Coope, 1977). Hibbert *et al.* (1971) investigated the pollen stratigraphy, identifying the sequence of vegetation changes after the Loch Lomond Stadial and extending into the Bronze Age, but the upper 1.5 m of sediment were not analysed.

Interpretation

Ashworth (1972) assessed the palaeoecological record of the basal sediments (385–310 cm) at Red Moss, identifying a sequence of climatic and environmental changes during the Late-glacial. He identified over 150 insect species in the Red Moss sediments, of which 26 have sufficiently limited ranges for climatic interpretation. The lowest organic sediments yielded a ^{14}C date of 12 160 ± 140 years BP, which is within the thermal maximum of the Late Windermere Interstadial (Coope, 1977; Lowe *et al.*, 1994a, b). The fossil coleopteran record extends from the basal sediments up to a further ^{14}C date of 9586 ± 200 years BP (Hibbert *et al.*, 1971; Ashworth, 1972). The coleopteran stratigraphy is divided into three assemblages, derived using the abundance of the 26 stenotopic species (Figure 8.21). The basal assemblage (385–360 cm) contains a fauna

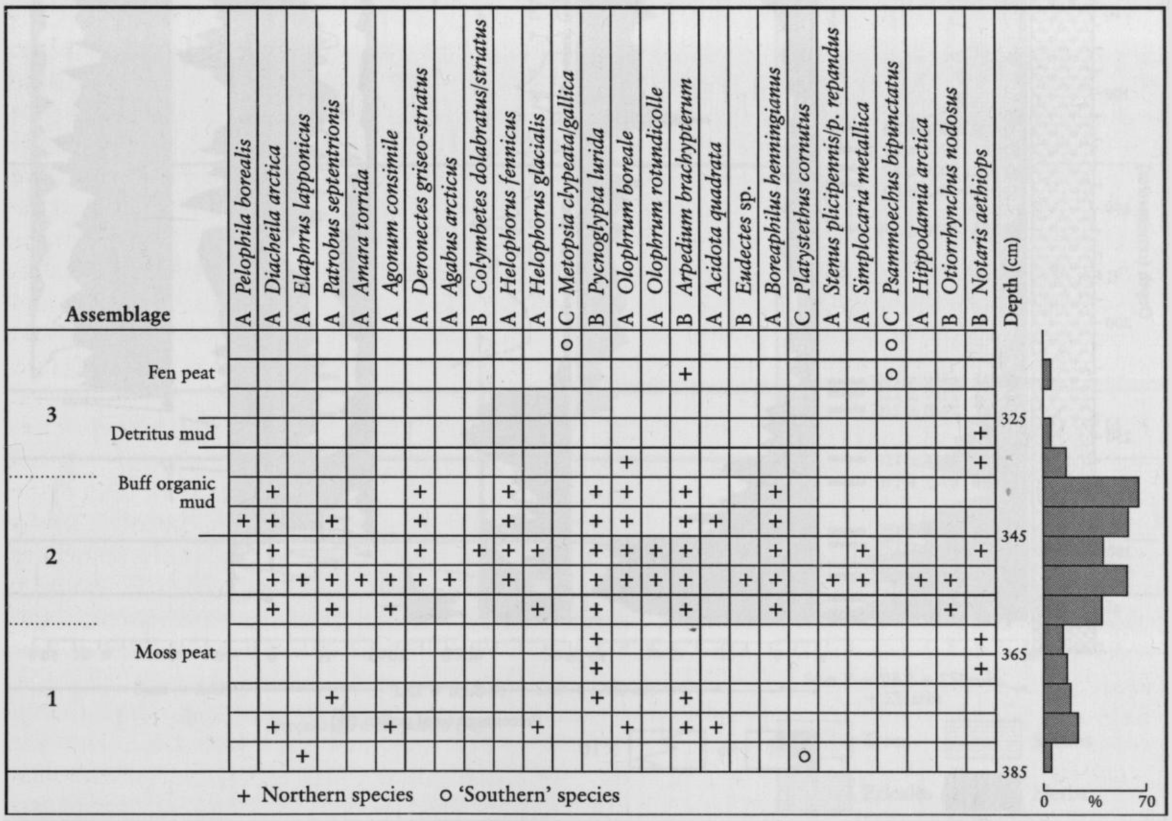


Figure 8.21 Distribution of coleopteran species at Red Moss and percentage frequency of northern species (after Ashworth *et al.*, 1972).

indicative of open exposed wet mossy habitats, with some species identifying a limited cover of *Salix* and *Betula* scrub or woodland, and is ^{14}C dated to between 12 160 ± 140 years BP and 10 850 ± 120 years BP. Eurythermal species and a few species that currently have a clearly northern Scandinavian and montane distribution dominate, and so it is difficult to estimate the climatic regime. Nevertheless, Ashworth (1972) suggested that the period experienced average July temperatures of around 14°C and average January temperatures of around 0°C.

Assemblage 2 (360–335 cm) contains a substantially reduced fauna dominated by northern species that indicate a rapid deterioration of the climate, which is dated to between 10 850 ± 120 years BP and 9798 ± 200 BP. Species inhabited pools of water and areas of patchy vegetation dominated by bryophytes and sedges. Ashworth (1972) suggested average July temperatures fell by 1–3°C and average January temperatures fell by 12°C. Assemblage 3 (335–310 cm) contains a more extensive fauna containing species that either live in wood or require trees and shrubs

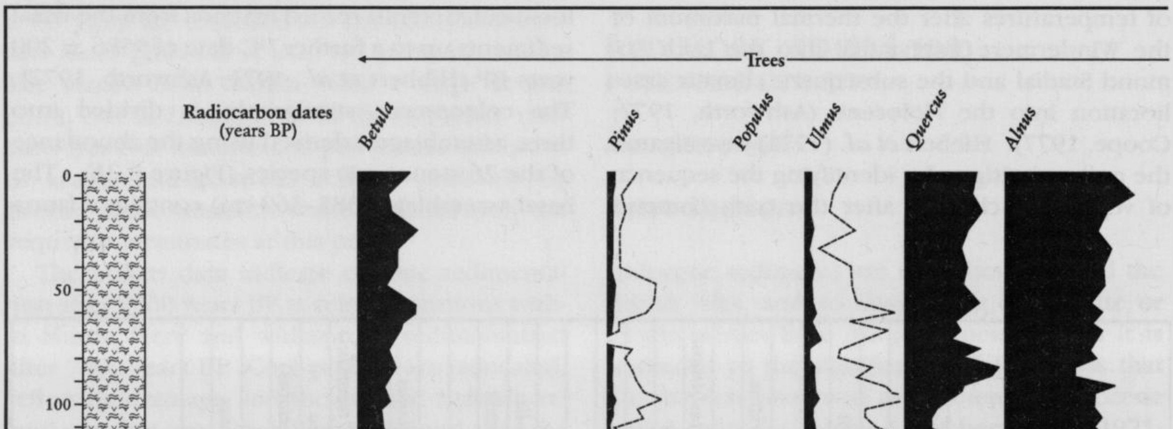
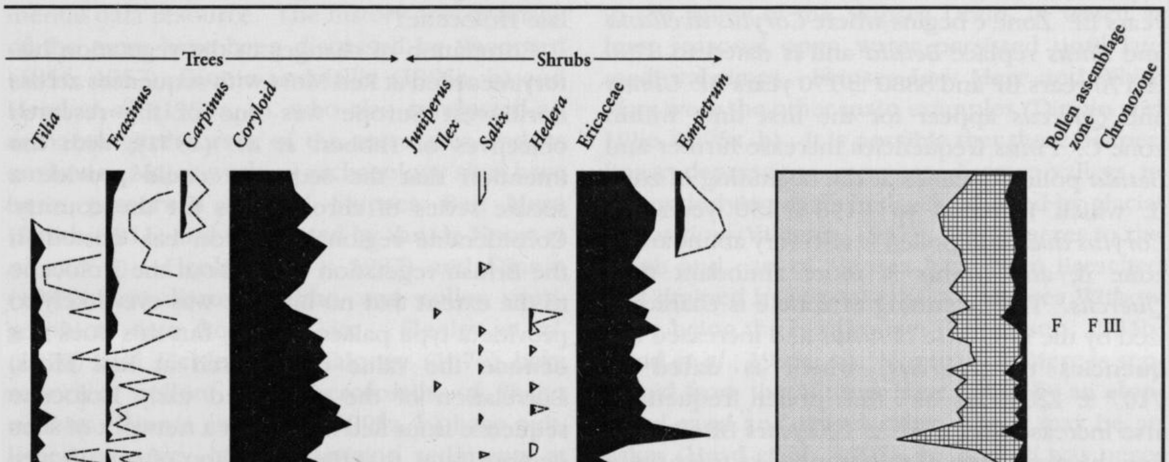


Figure 8.22 Arboreal pollen diagram from Red Moss with radiocarbon dated horizons and pollen zonation scheme (after Hibbert *et al.*, 1971). See Figure 8.1 for key to the stratigraphical log.

for shade, and indicative of a marshy woodland. Between 9798 ± 200 years BP and 9586 ± 200 years BP the Red Moss region experienced average July temperatures of around 16°C and average January temperatures of around 5°C (Ashworth, 1972). Across the boundary between assemblage 2 and 3 sediment accumulation rates at Red Moss indicate that average July temperatures rose from 10°C to 16°C within 350–400 years (Hibbert *et al.*, 1971; Ashworth, 1972).

Hibbert *et al.* (1971) produced a pollen diagram covering the stratigraphy between 330 and 0 cm, which is subdivided into six assemblages zones (Figure 8.22). Basal pollen zones A and B overlap with coleopteran assemblage 3, identifying a landscape dominated by *Betula*, *Pinus*, *Salix* and *Juniperus* woodland or initially scrub woodland. The Late-glacial pollen and coleopteran sequence at Red Moss conforms, with and contributes to, the established sequence of climatic and environmental changes during this period (Lowe *et al.*, 1994a, b). The Red Moss sequence starts within the thermal maximum of the Late-glacial interstadial, and the Late-glacial



stadial is clearly identified by the coleopteran data. The Late-glacial stadial dates conform to the isotope sequence derived from Greenland ice cores (Grootes *et al.*, 1993). Both coleopteran assemblage zone 3 and pollen zone A reflect the amelioration of climate after the Late-glacial stadial and the beginning of the Holocene *c.* 10 000 years BP.

The pollen stratigraphy identifies the sequence of woodland colonization in lowland Lancashire, with 16 radiocarbon dates securing the vegetation sequence for the early to mid-Holocene. *Betula* dominates zone b, with *Pinus*, *Populus*, *Salix*, *Juniperus* and herbaceous pollen declining from the levels in zone a. The expansion of *Betula* is dated to 9798 ± 200 years BP. Zone c begins where *Corylus avellana* and *Pinus* replace *Betula* and is dated to 8790 ± 170 years BP and 8880 ± 170 years BP. *Ulmus* and *Quercus* appear for the first time within zone C. *Pinus* frequencies increase further and *Betula* pollen declines at the beginning of zone d, which is dated to 8196 ± 150 years BP. *Corylus avellana* pollen is also very abundant in zone d, and *Ulmus* is more abundant than *Quercus*. The beginning of zone e is characterized by the rapid rise of *Alnus* and increased frequencies of *Quercus*, which is dated to 7107 ± 120 years BP. *Tilia* pollen frequencies also increase after 6880 ± 100 years BP. Paralleling the expansion of thermophilous trees there are equivalent declines in the shade-intolerant *Betula* and herbaceous pollen taxa, reflecting closure of the forest canopy. *Calluna vulgaris* pollen becomes more abundant around 7450 ± 150 years BP, which supports the stratigraphical evidence for expansion of local raised mire communities dominated by *Calluna vulgaris* and *Eriophorum vaginatum*.

Quercus, *Ulmus* and *Alnus* dominate the pollen stratigraphy until the base of zone f, which is marked by a sharp decline in *Ulmus* pollen frequencies. Four closely spaced ^{14}C dates provide a consistent picture of the timing of the Elm Decline, with a specific date of 5010 ± 80 years BP. Thinning of the forest canopy allows the limited expansion of *Plantago lanceolata*, *Artemisia* and *Filipendula*, but other tree species soon close this opportunity. *Fraxinus* is the last of the deciduous trees to appear in the Red Moss pollen stratigraphy, increasing *c.* 5399 ± 100 years BP. *Quercus*, *Alnus* and *Corylus avellana* are the dominant arboreal trees and shrubs throughout most of zone F, although *Betula*, *Ulmus*, *Tilia* and *Fraxinus* also are significant components of the mid-Holocene mixed deciduous forest of lowland Lancashire. In the upper 30–40 cm of the pollen diagram there is a minor reduction in arboreal pollen frequencies and equivalent increases of Poaceae, *Plantago lanceolata*, *Artemisia*, *Urtica* and Ranunculaceae. These changes probably reflect anthropogenic activity and limited woodland clearances, but unfortunately the upper metre of the pollen diagram is not ^{14}C dated. A further 1.5 m of sediment above the pollen profile analysed by Hibbert *et al.* (1971) was not analysed for pollen content, and so Red Moss has contributed no information about vegetation changes during the middle to late Holocene.

Correlation of changes in the vegetation history recorded at Red Moss with sequences across north-west Europe was one of the research objectives of Hibbert *et al.* (1971), with the intention that the sequence could provide a secure series of chronozones for the country. Considerable regional variation has existed in the British vegetation throughout the Holocene to the extent that no one site was ever likely to provide a type palaeoecology, but this does not denude the value of research at Red Moss. Correlation of the well-dated early Holocene sequence from Red Moss with a network of sites demonstrates that the expansion of certain tree species is clearly diachronous across Great Britain and north-west Europe. Red Moss yields an important pollen record and has contributed to various syntheses of the British vegetation history (Godwin, 1975; Pennington, 1974) and palaeovegetation maps generated for the Holocene Epoch (Huntley and Birks, 1983).

Conclusions

Red Moss is an important site because it has yielded a significant and well-dated record of vegetation change between c. 14 000 and 5000 years ago. Fossil Coleoptera reveal the sequence of climatic and environmental changes during the later stages of the Late-glacial interstadial, and throughout the Late-glacial stadial and transition into the Holocene Epoch. Red Moss has produced one of Lancashire's most complete and best-dated pollen diagrams. Detailed pollen records coupled with radiocarbon dating demonstrate the sequence of woodland colonization during early Holocene times. Today Red Moss is in a poor condition and little resembles a lowland raised mire.

SKIPSEA BAIL MERE (TA 158 558)

J. Innes

Introduction

Skipsea Bail Mere is a drained former freshwater mere to the west of Skipsea village in northern Holderness, East Riding of Yorkshire. Its importance lies in its Late Devensian to mid-Holocene sequence of sediments, and their palaeoenvironmental data resource. The history and drainage of the mere have been discussed by Sheppard (1956, 1957), Dinnin and Lillie (1995a, b) and Head *et al*. (1995a, b), who also conducted an archaeological survey of the mere area and its environs. Major wetland archaeology sites have been preserved around Skipsea Bail Mere (Smith, 1911) and reassessed by Van de Noort *et al*. (1995). Flenley (1984, 1987) and Dinnin (1995) have discussed litho- and pollen stratigraphical data from the site. Flenley *et al*. (1975) and Flenley and Maloney (1976) have reported pollen and macrofossils of *Trapa natans*. Dinnin and Lillie (1995a, b) have published a survey of the surviving sediments at Skipsea Bail Mere. Lillie (1995) has studied the association of colluvial sediments with alluvial mere deposits at the site margins.

Description

The site of Skipsea Bail Mere lies in an area of extensive former wetlands. Classified as the 'Hornsea Member' by Lewis (1999), the undulating hummocky till and outwash plain in this area (Catt and Penny, 1966) contains many depressions, some of which are kettleholes. The confined drainage within these features has led to the formation of meres during the Late Devensian and the early Holocene. Drainage of this landscape was naturally to the west and the River Hull, by low-gradient, flat-bottomed stream valleys, although the streams of the Skipsea area have been artificially diverted to the coast since about 1800 (Head *et al*., 1995a, b). These stream valleys were also inclined to poor drainage, their low altitude enhancing this tendency after the mid- to late Holocene establishment of high sea level. These valleys supported extensive wetlands, with major deposition of organic and alluvial sediments, until their drainage and reclamation for agriculture in historic times. Many of the small meres had naturally silted up by the mid-Holocene, but many larger examples survived until drained during the medieval period (Sheppard, 1957). Dinnin and Lillie (1995a, b) examined several of the possible Holderness mere sites recorded by Sheppard, however, and were unable to find Holocene mere wetland deposits. These were either never true meres, or their sediments have not survived. Skipsea Bail Mere was one of a chain of a few larger meres in deep depressions in the valley of the Skipsea Drain, in which at least seasonal open water persisted until late medieval times. Skipsea Low Mere and White Marr were the other main examples (Dinnin and Lillie, 1995a, b). It is possible that these deeper, linear depressions represent former valleys in the underlying chalk bedrock, masked by glacial deposition (Valentin, 1957). Some meres to the south and east of Skipsea have been breached and drained by coastal erosion, Skipsea Withow Mere being the best known (Gilbertson, 1984b; Head *et al*., 1995a, b). Skipsea Bail Mere is separated from the Withow Mere valley by an elongated sand and gravel ridge, which may be an esker (Head *et al*., 1995a, b), and so was never joined to the Withow wetland system. Bail Mere is almost certainly within the same large depression as the nearby Skipsea Low Mere, however, and the two became separated in antiquity as a result of sedimentation infilling a narrow section of the valley and dividing the water body. Skipsea Bail Mere probably ceased to be open water during increased drainage activity in the period after its last recorded documentary mention in 1367 (Sheppard, 1956; Head *et al*., 1995a, b). The present-day flat land that represents the extent of the former mere is well defined and much prehistoric and later archaeological material has been recovered from survey of the ground around the margins of the old wetland (Head *et al*., 1995a, b). As part of this survey, within a test pit, a wooden stake was found that apparently was driven into the underlying glacial material through the basal organic deposits (Head *et al*., 1995a, b). Pollen analysis showed these to be early Holocene and a radiocarbon date on the wooden stake was 9080 ± 100 years BP. Similar stakes had been reported by Smith (1911) in an earlier study of the land around the mere edge. The former mere surface is under arable land use with a low water table, and is actively being degraded by drainage, ploughing and erosion (Dinnin and Lillie, 1995a, b).

Flenley (1984, 1987) recorded about 5 m of fine-grained lake sediment at Skipsea Bail Mere, which comprised alternating horizons of grey clay–silts and organic limnic muds. The grey clay–silts occupy the lower 1.3 m of the profile and are interrupted by a thin lower band of organic clay up to 5 cm thick, and a higher band of similar organic clay about 50 cm thick. Above the uppermost clay–silt lies about 2.7 m of highly organic limnic mud (gyttja), which is covered by a surficial layer of minerogenic, probably colluvial, material. Flenley (1984, 1987) published outline pollen diagrams for these lake sediments, which showed that the lower clay–silt and organic mud intercalations formed during the Devensian Late-glacial period. The highly organic limnic sequence represents the beginning of the Holocene Epoch to a point in the later Holocene after the *Ulmus* Decline, which marks the end of the mid-Holocene Flandrian II chronozone about 5000 years BP. The main Late-glacial clay–silt is dominated by Gramineae and Cyperaceae. The more organic layers also contain high Gramineae and Cyperaceae pollen values, but *Betula* also is significant. Pollen of more thermophilous trees, such as *Alnus* and *Corylus*, also occur in these very early levels. The Holocene gyttja shows the typical post-glacial succession through *Betula* and *Corylus* stages to a deciduous forest with *Quercus* and *Ulmus* and then high *Alnus*. Post *Ulmus* Decline levels show large-scale forest clearance in the upper sampled metre, with high Gramineae, *Rumex* and *Plantago lanceolata*. A peak of *Plantago lanceolata* also occurs lower in the Holocene record, at the level of the mid-Holocene rise of *Alnus* pollen. *Trapa natans* (water chestnut) pollen and macrofossils were found in the *Alnus–Quercus* pollen zone post-dating the *Ulmus* Decline. They were also found in the equivalent zone at nearby Skipsea Low Mere. Seven pollen assemblage zones were recognized on the pollen diagram (Figure 8.23).

Dinnin and Lillie (1995a, b) completed a transect of cores across the site (Figure 8.24) as part of a recent survey of the Holderness meres and their deepest record contained 6.75 m of wetland sediments, overlying coarse sand. They identified a tripartite Late-glacial stratigraphical succession, but with only a single warm phase organic mud deposit, unlike Flenley's core. Nearly 5 m of highly organic gyttja and woody fen peat overlay the lower clay–silt and organic clay intercalations. Up to 50 cm of fine inorganic clay–silt overlay the organic sequence in all the transect cores, interpreted by the authors as colluvium and as evidence of substantial catchment clearance and erosion. Pollen data were not collected during this lithological study. Work by Lillie (1995) on the colluvial sediments on the slopes around the old mere demonstrates that the onset of colluviation at Skipsea Bail Mere post-dates the mere sediments. Dinnin and Lillie (1995a, b) show that the lake sediments at Skipsea Bail Mere are being affected by dessication, particularly those adjacent to the Skipsea Drain, much of the upper metre being disturbed and oxidized, although the deeper organic sequence appears well preserved.

Interpretation

Although only preliminary palaeoenvironmental studies have been completed at Skipsea Bail Mere (Flenley, 1984, 1987; Dinnin and Lillie, 1995a, b), enough is known to show that the site is clearly one of very high potential for further research. A major significance of the site is the presence of the thin organic mud very low in the Late-glacial stratigraphy. This layer may well be evidence of a warm phase pre-dating the main interstadial (i.e. pre-Allerød) episode of climatic amelioration, equivalent to the dual *Betula* peaks recognized in a few key sites from north-east England, such as Tadcaster (Bartley, 1962), Thorpe Bulmer (Bartley *et al.*, 1976), and close by in Holderness at The Bog, Roos (Beckett, 1981). This early part of the Late-glacial record is becoming better understood through focused research at sites with a high quality lithostratigraphical record, such as Gransmoor (Walker *et al.* 1993; Lowe *et al.*, 1995a, b), which is only a few miles to the west of Skipsea. Although the existence of brief warm oscillations in the early Late-glacial climate record is becoming well established through these studies, the dating of these warm phases remains uncertain owing to difficulties in the radiocarbon dating of samples of low organic content and 'hard water' complications. Correlation of these early warm phases on only lithostratigraphical grounds is very insecure, and existing radiocarbon dates on early

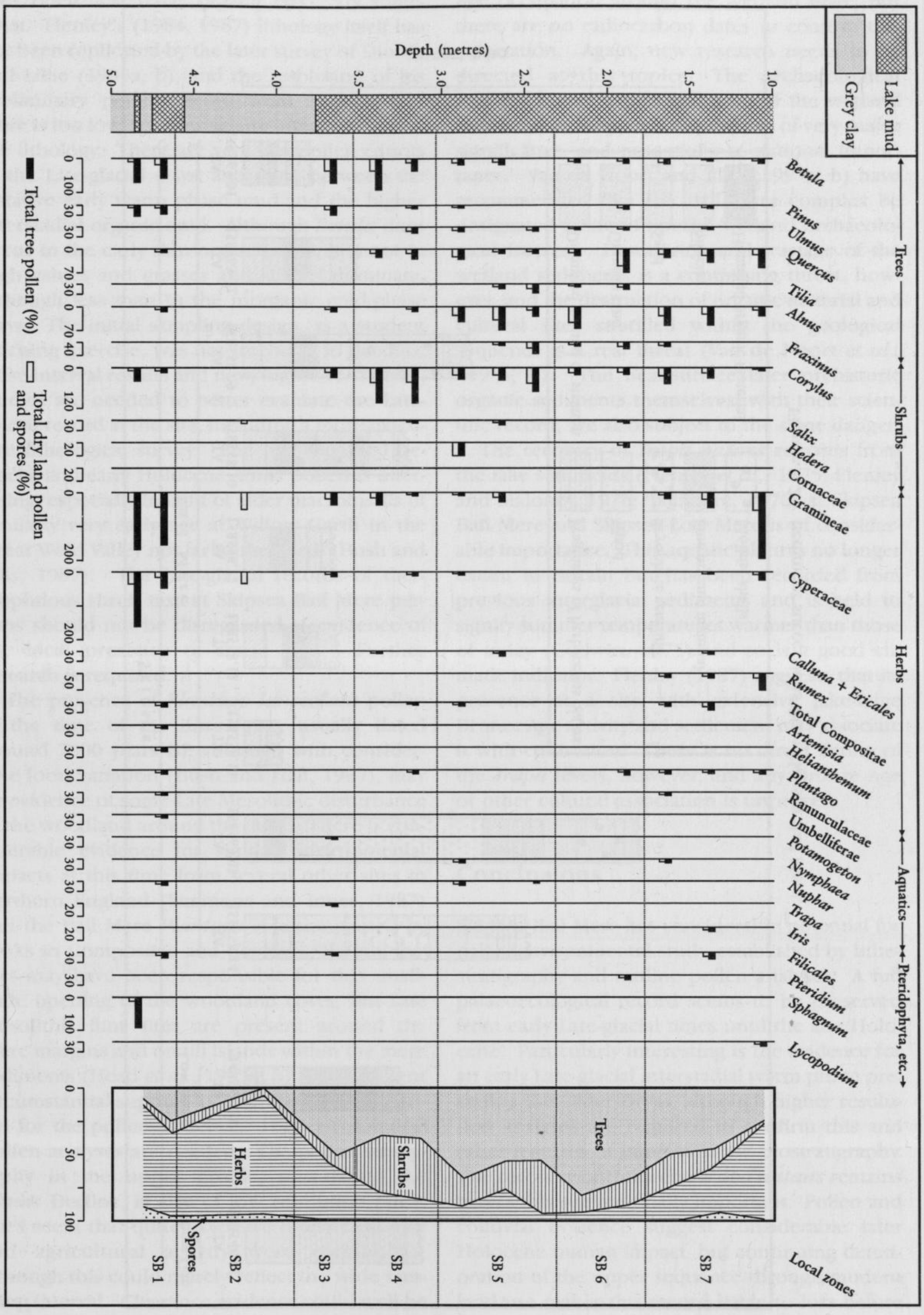


Figure 8.23 Outline pollen diagram from Skipsea Bail Mere.

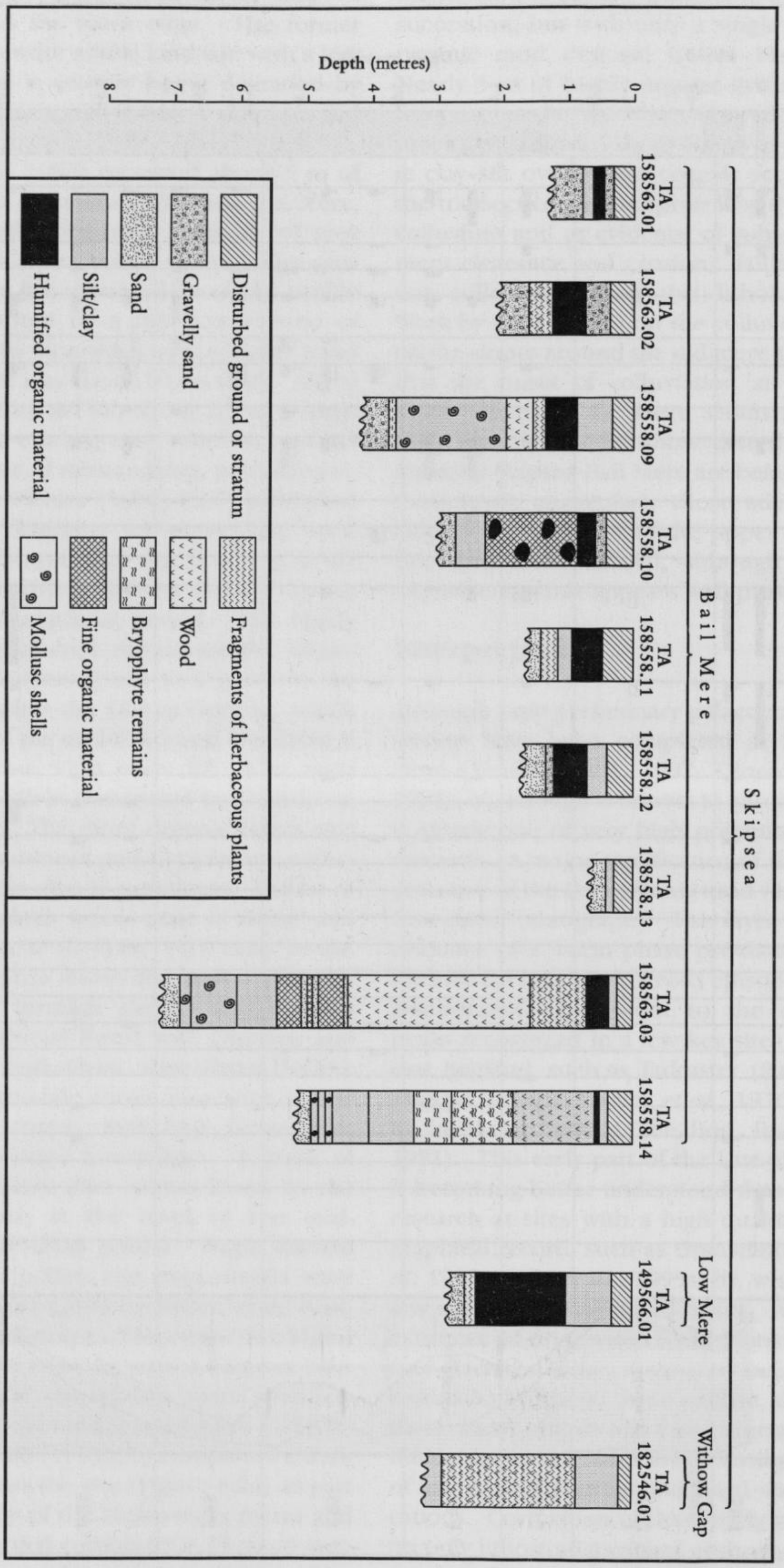


Figure 8.24 Sediment successions from Skipsea Bail Mere.

thin, often near basal, organic clays vary somewhat. Flenley's (1984, 1987) lithology itself has not been replicated by the later survey of Dinnin and Lillie (1995a, b), and the resolution of his preliminary pollen record from Skipsea Bail Mere is too low to allow secure interpretation of the lithology. There are very few pollen counts in the Late-glacial clays, and none between the putative early warm phase mud and the higher interstadial organic mud. Although *Betula* does occur in the early thin organic mud, it is not in high values and grasses and sedges dominate, although less than in the inorganic cold-phase clays. The initial sampling design, as a student teaching exercise, was not intended to produce close interval results and new, higher resolution, studies are needed to better evaluate the Late-glacial record at the site, including a more extensive lithological survey. The presence of Late-glacial and early Holocene *Alnus* pollen is interesting, especially in light of alder macrofossils of similarly very early age at Willow Garth in the Great Wold Valley not far to the north (Bush and Ellis, 1987). The Late-glacial records of thermophilous shrub taxa at Skipsea Bail Mere perhaps should not be disregarded as evidence of the local presence of those taxa. Further research is required.

The presence of *Plantago lanceolata* pollen at the time of the *Alnus* rise, usually dated around 7000 years BP, although with considerable local variation (Bush and Hall, 1987), may be evidence of some Late Mesolithic disturbance of the woodland around the mere. There is considerable evidence for similar environmental impacts at this time from several other sites in northern England (Simmons and Innes, 1987) and the Bail Mere *Plantago* is accompanied by peaks in Compositae and *Corylus*. Natural factors may have been responsible for this small-scale opening of the woodland cover, but Late Mesolithic flint sites are present around the mere margins and on till islands within the mere sediments (Head *et al.*, 1995a, b) and represent circumstantial support for a hunter–gatherer origin for the pollen evidence. Higher resolution pollen analyses are needed. The pollen stratigraphy in the upper profile, post-dating the *Ulmus* Decline, is also of low resolution but it does seem that quite extensive forest clearance and agricultural activity were continuous, although this could merely reflect the wide sampling interval. Clearance evidence could well be linked to the Neolithic and, particularly, Bronze Age occupation around the wetland edge, but there are no radiocarbon dates to confirm this association. Again, new research needs to be directed at this topic. The archaeological remains found within and around the wetland sediments at Skipsea Bail Mere are of very major significance, and potentially of national importance. Van de Noort and Ellis (1995a, b) have recommended that the Bail Mere complex be designated a Site of Special Wetland Archaeological Interest. Dewatering and wastage of the wetland sediments is a continuing threat, however, and the destruction of organic material and cultural sites stratified within the geological sequence is a real threat (Van de Noort *et al.*, 1995a, b). The near-surface later prehistoric organic sediments themselves, with their scientific record, are also subject to the same danger.

The recovery of *Trapa natans* remains from the lake sediments (Flenley *et al.*, 1975; Flenley and Maloney, 1976; Tallantire, 1976) at Skipsea Bail Mere and Skipsea Low Mere is of considerable importance. This aquatic plant is no longer extant in Britain but has been recorded from previous interglacial sediments and is held to signify summer temperatures warmer than those of today (Godwin, 1975) and so is a good climatic indicator. Flenley (1987) suggests that its presence at a site with extensive lake-edge Bronze Age activity and settlement may associate it with cultivation. There is no direct dating of the *Trapa* levels, however, and any Bronze Age or other cultural association is unproven.

Conclusions

Skipsea Bail Mere has considerable potential for palaeoenvironmental study, established by lithostratigraphy and outline pollen analyses. A full palaeoecological record seems to be preserved from early Late-glacial times until the late Holocene. Particularly interesting is the evidence for an early Late-glacial interstadial warm phase preceding the Older Dryas, although higher resolution analyses are required to confirm this and other features of interest in the biostratigraphy. The recovery of Holocene *Trapa natans* remains makes this site nationally important. Pollen and colluvial evidence suggest considerable later Holocene human impact, but continuing deterioration of the upper sequence through modern land use makes this record liable to loss before it is researched properly.

SKIPSEA WITHOW (TA 184 547)

J. Innes

Introduction

Skipsea Withow (or Whitow) is a former freshwater mere to the east of Skipsea village in northern Holderness, East Riding of Yorkshire, which preserves a long sequence of Late Devensian to later Holocene sediments. The mere deposits have been examined in detail (Gilbertson, 1984a; Gilbertson *et al.*, 1987). Blackham and Flenley (1984) have presented a pollen analytical study of the Holocene sediments, which have been investigated for plant macrofossils by Hall (1984). The Late Devensian sediments have undergone molluscan (Thew and Woodall, 1984), palaeomagnetic (Gale, 1984), palaeobotanical (Hunt *et al.*, 1984) and animal bone (Jenkinson, 1984) study. Insect remains from the site have been studied by Kenward (1984). Archaeological material has been studied by Mellars (1984), and Gilbertson (1984b) and McAvoy (1995) have examined the preserved wood evidence. Earlier archaeological, litho- and bio-stratigraphical work was completed by Phillips (1829), Armstrong (1923), Godwin and Godwin (1933), Boylan (1966a) and Robinson (1972). The depositional and vegetation history of Skipsea Withow have been discussed by Beckett (1977a), Flenley (1984, 1987, 1990), Gilbertson (1990), Van de Noort and Davies (1993) and Dinnin (1995). The site has been surveyed archaeologically by Head *et al.* (1995a) and palaeoenvironmentally by Dinnin and Lillie (1995a). Van de Noort *et al.* (1995) have considered the preservation potential of the site. The history and drainage of the mere have been discussed by Sheppard (1912) and Sheppard (1956, 1957).

Description

Skipsea Withow lies on the coast of north-eastern Holderness to the east of Skipsea and north of Atwick villages. The site is a former mere or cluster of meres that occupied a depression in the undulating till and hummocky outwash plain of this area (Catt and Penny, 1966), which has been classified as the 'Hornsea Member' of the Holderness Formation by Lewis (1999). It probably was one of the larger meres of Holderness (Sheppard, 1957; Flenley, 1987; Gilbertson,

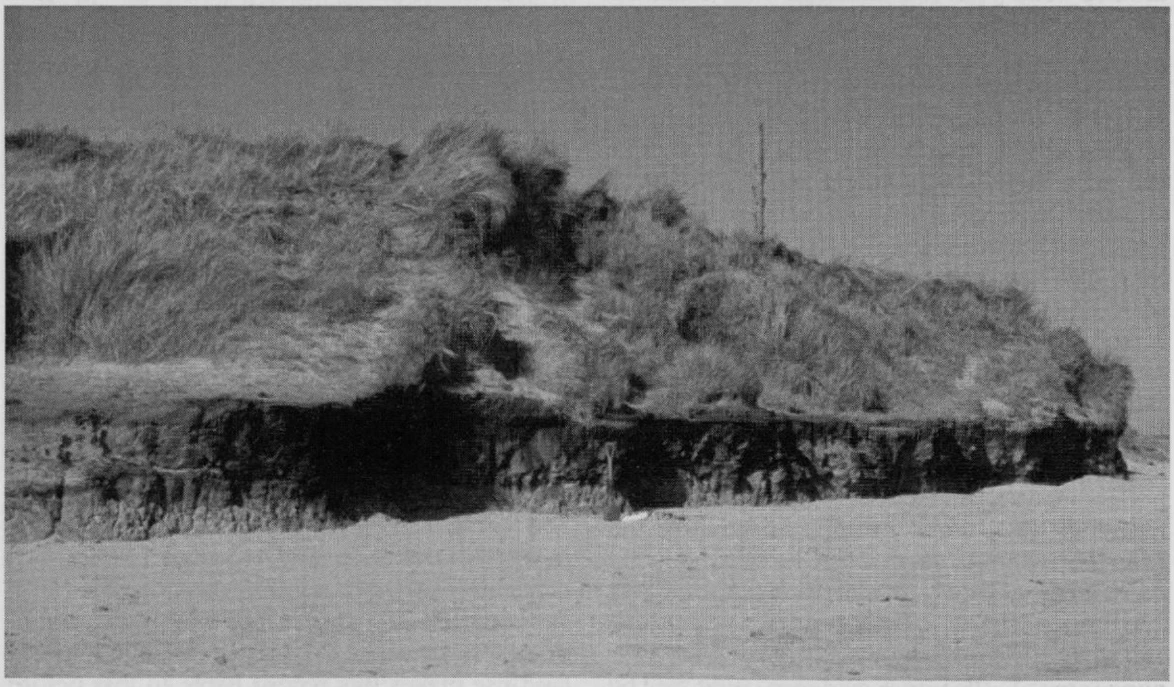


Figure 8.25 Organic sediments exposed at Skipsea Withow. (Photo: J. Innes.)

1990) and in common with some other large mere sites its deeper valley and east–west orientation suggested to Valentin (1957) that it may occupy a pre-glacial valley in the underlying chalk bedrock. Although near to the complex of Skipsea Bail and Low meres, which lies to the west of Skipsea village, Skipsea Withow is separated from them by an elongated sand and gravel ridge composed of glacial material, which may have formed as an esker (Head *et al.*, 1995a). It was never joined to that neighbouring wetland system and has had an independent depositional history. Skipsea Withow ceased to be a freshwater body when it was breached and drained by at least the mid-17th century (Gilbertson, 1984a) by coastal erosion. The surviving mere sediments at Skipsea Withow occupy a hollow about 100 m across (Beckett, 1977a) at the Skipsea Withow Gap (Withow Hole) and have been exposed in section by erosion (Figure 8.25). They represent a highly complicated sequence of lake-margin deposits (Gilbertson *et al.*, 1987).

Early research at Skipsea Withow had established the site as of great potential for palaeoenvironmental research, although the published literature on this early work, reassessed by Gilbertson (1984a) and Dinnin (1995), led to considerable confusion. Phillips (1829) reported a stratigraphical succession in a large hollow up to 400 m wide exposed in a coastal section. It comprised a woody peat of up to 2 m thickness overlying clay with freshwater shells in the southern part of the hollow, with gravel resting upon till completing the lower succession. Although his text did not make this explicit, the drawings of the section by Phillips showed a clear unconformity between the shelly clay and the overlying peat. Also unclear was the stratigraphical position of a skull of giant deer *Cervus megaceros* (cf. *Megaloceros giganteus*) recovered from the deposits, which may have come from the shelly clays or the gravel. Equally uncertain was the context of early Mesolithic barbed 'harpoon' type points recovered in 1903 from within silt and below woody peat in the Skipsea Withow sediments (Armstrong, 1923), which apparently overlay the bones of *Megaloceros giganteus*, as well as of reindeer, red deer and aurochs. Further excavations by Armstrong recovered Mesolithic type flint implements from both the silty clays and the peats. The uncertainty in these early studies regarding the provenance of the sediments, animal bones and human artefacts led to the juxtaposition of Lateglacial and Holocene faunal, floral and cultural elements and the idea of the possible persistence of taxa perceived as diagnostic of the Lateglacial, such as the giant deer, into Holocene times. Such survival of *Megaloceros* into early Holocene times in parts of the British Isles is now supported by recent research (Gonzalez *et al.*, 2000; Kitchener and Bonsall, 1999). The controversy over the Skipsea Withow record, however, prompted Godwin and Godwin (1933) to apply pollen analysis to the Withow Gap sediments to clarify the age of the peat and the underlying silts, which contained the harpoon point. Armstrong's original harpoon site had been destroyed by continuing erosion and at the new site examined for pollen analysis the following stratigraphy was recorded (cf. Beckett, 1977a; Gilbertson, 1984a).

	Depth (m)
Fine brown clay, now cracking into columnar form	0.0–0.76
Solid black or brown amorphous peat with large numbers of horizontal tree branches or trunks, including much oak (*Quercus*) especially in the upper 0.6m, which is almost solid with them. Hazel (*Corylus*) nuts found at 2.01, 2.46 and 2.68 m, and at its base	0.76–2.89
Brown sandy silt with fragments of *Pinus* bark, fins of pike (*Esox lucius*) and flint artefacts. Stone fruit of ?*Prunus*	2.89–3.04
Buttery blue clay	>3.04

Gilbertson (1984a) redrew the pollen diagrams published by Godwin and Godwin (1933) and this redrawn figure is shown as Figure 8.26. The high *Corylus*, *Quercus* and *Ulmus* values, with no *Alnus*, show the brown silt to be predate the *Alnus* rise of early Boreal Holocene age, pollen zone V. It contained Mesolithic-type flint artefacts. The peat above it, with rising and then high *Alnus*, is of late Boreal Flandrian I and then mid-Holocene Flandrian II age, pollen zones VI and VIIa. Study of molluscan assemblages in the brown silt showed it to be a freshwater lake deposit, but were climatically undiagnostic. Although the brown silt analysed by Godwin and Godwin (1933) was clearly of Holocene age, there is no evidence that it was the same silt unit from which skeletal and/or cultural material had

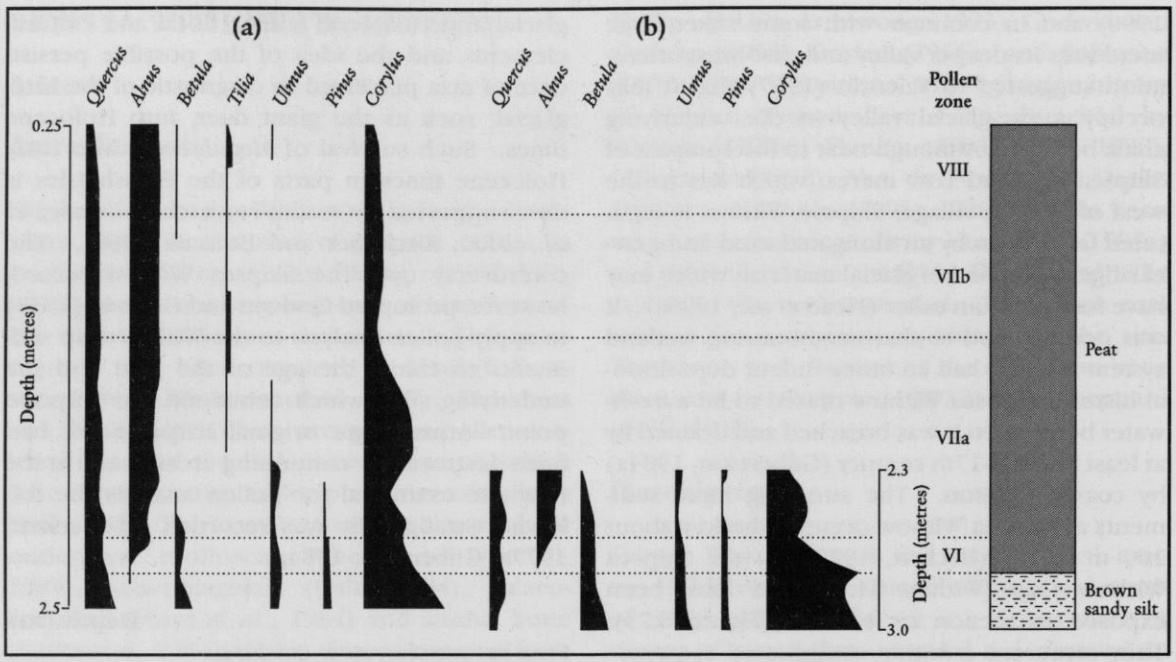


Figure 8.26 Summary of the pollen diagram from Skipsea Withow published by Godwin and Godwin (1933).

previously been derived (Armstrong, 1923), although this correlation was assumed.

Clarification of the environmental history of the site has been achieved by an interdisciplinary analysis of exposures at Skipsea Withow (Gilbertson, 1984a). Detailed sedimentological analyses of several sections allowed the adoption of the following 11 lithological units. Several minor sub-units were recognized within these and are listed elsewhere (Gilbertson, 1984a; Gilbertson *et al.*, 1987).

11. Made ground.
10. Modern, grey clay soil.
9. Grey sandy silt, resting unconformably on unit 8.
8. Dense, well-humified, detrital and carr peat with abundant *Corylus* nuts and wood. Some sand and silt inwash partings. Carved wooden stakes and pegs found within this peat.
7. Brown, laminated, silty peat with abundant detrital brushwood.
6. Dark brown, silty peat with some wood. Unconformable contact with unit 5.
5. Orange-brown, weathered, sandy silt.
4. Poorly sorted coarse gravel with some clay and silt partings. Erosional upper surface with reworked peat and other sediment. A thin *Carex* peat in mid-unit is probably *in situ*, although much affected by reworking.
3. Thinly laminated silts and clays. Blue-grey sticky clays. Locally common fragmented plant debris and shells. Textural varves in the lower unit, with reworked and in-situ thin detrital peats. Rich in molluscan remains, especially *Valvata piscinalis*. Flint blade found above a silty detritus peat.
2. Well-sorted sands resting upon poorly sorted sandy clays.
1. Till.

The top of peat unit 8 was radiocarbon dated to 4500 ± 50 years BP (SRR-1942) and one of the carved wooden stakes in this unit was dated 4770 ± 70 years BP (HAR-3378). The base of the peat of unit 6 was dated to 9880 ± 60 years BP (SRR-1944). A *Betula* log within the *Carex* peat of unit 4 was dated 10440 ± 80 years BP (SRR-1943), and a *Betula* log within the blue clay of unit 3 was dated 10710 ± 70 years BP (Q-3035). These units and dates are shown in outline section in Figure 8.27, as redrawn by Dinnin (1995) after Gilbertson *et al.* (1987).

Blackham and Flenley (1984) and Gilbertson *et al.* (1987) have presented a detailed pollen diagram through most of the above sedimentary sequence (Figure 8.28). The whole of organic

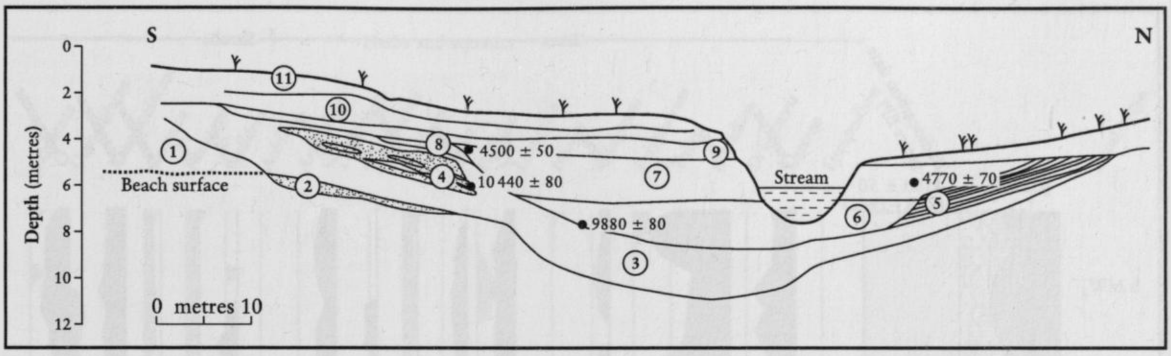


Figure 8.27 Outline of the stratigraphy at Skipsea Withow Gap (for key to numbers, see text).

units 6, 7 and 8 are included, with their bracketing radiocarbon dates of 9880 ± 60 years BP and 4500 ± 50 years BP. A silt sample from unit 3 was added to the base of the diagram at 2.65 m, recovered from lower in the section. The following six local pollen assemblage zones, from the base of the diagram, are recognized.

		Depth (m)
WM1	Dominated by *Betula* pollen with low values for *Pinus*, *Salix* and Gramineae	2.65
WM2	High values for *Betula*, Gramineae and Cyperaceae	1.65
WM3	Dominated by *Betula* and *Corylus* pollen. *Pinus* reaches a peak and *Quercus* and *Alnus* appear at the end of the zone	1.62–1.22
WM4	Dominated by *Alnus*. *Quercus*, and *Tilia* also present. *Fraxinus* appears and *Pinus* disappears during the zone	1.22–0.67
WM5	A reduction in total tree pollen occurs, followed by a recovery. *Alnus*, *Ulmus* and to a lesser extent *Tilia* are the trees most affected. Gramineae reaches a peak. Cereal and *Plantago* are present	0.67–0.42
WM6	Another reduction in tree pollen values with only a partial recovery. *Alnus* and *Ulmus* are most affected. Cereal and *Plantago* are continuously present	0.42–0.02

Plant macrofossils were recorded by Hall (1984) from three points in the section analysed. At 1.4 m near the base, aquatic herbaceous taxa were common, and *Betula*, *Populus* and *Schoenoplectus* occurred. A single *Alnus* fruit was recorded. At 0.5 m in the woody peat *Alnus* was very common with other forest taxa present, including *Corylus*, *Quercus* and *Prunus*. Mosses likely to live on tree bark were present. At 0.2 m in the upper part of the peat *Alnus* was again common, and *Taxus* occurred. Gilbertson (1984a, b) examined the abundant wood remains in the peat of unit 8 and concluded that some showed evidence of woodworking. The main examples were a carved alder rod and peg, which gave the radiocarbon date of 4770 ± 70 years BP and a similar alder rod and peg still *in situ* inserted vertically into the peat, which exhibited a trimmed 'elbow' form typical of coppiced wood. A recent excavation at Skipsea Withow Gap (McAvoy, 1995) has reassessed these wood-rich horizons as a natural log-jam that contains beaver-gnawed wood. The origin and significance of the wood remains of unit 8 are therefore currently a subject of debate.

The Late Devensian succession at the Skipsea Withow Gap has been investigated in detail in several sections at three main exposures (Gilbertson *et al.*, 1984, 1987). It comprises site units 2 to 5 inclusive, between the weathered till of unit 1 and the Holocene sequence of unit 6 onwards. These have been categorized as sandy mudflows and sandy pebbly lag deposits of unit 2, a lacustrine fine-grained suite of silts, clays and rarer organic layers in unit 3, a coarse gravel-dominated suite in unit 4 and weathered sandy silts in unit 5. Erosional breaks in sedimentation were common in these units and faulting also occurred. A flint blade was found within the silts of unit 3. Hunt *et al.* (1984)

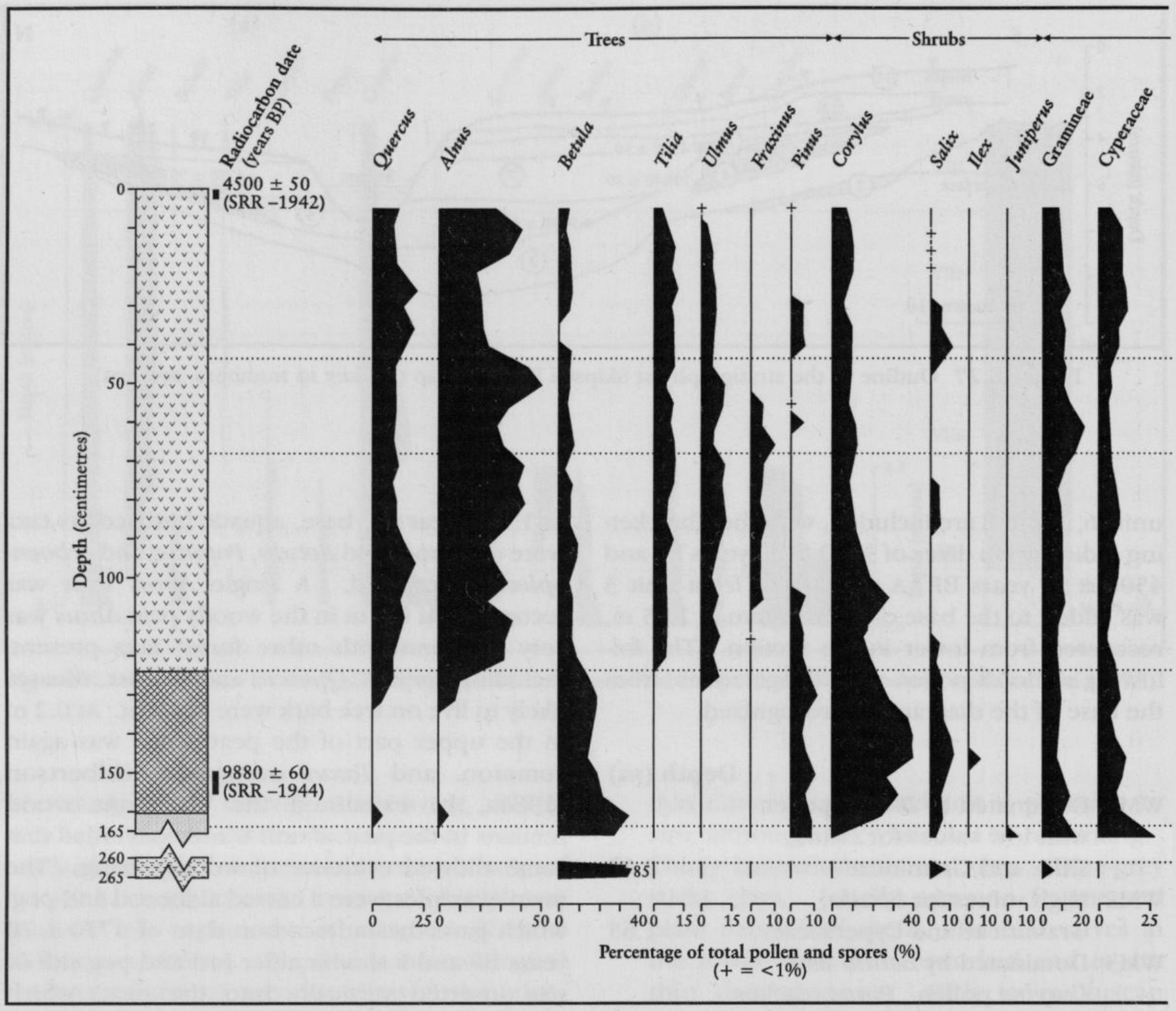
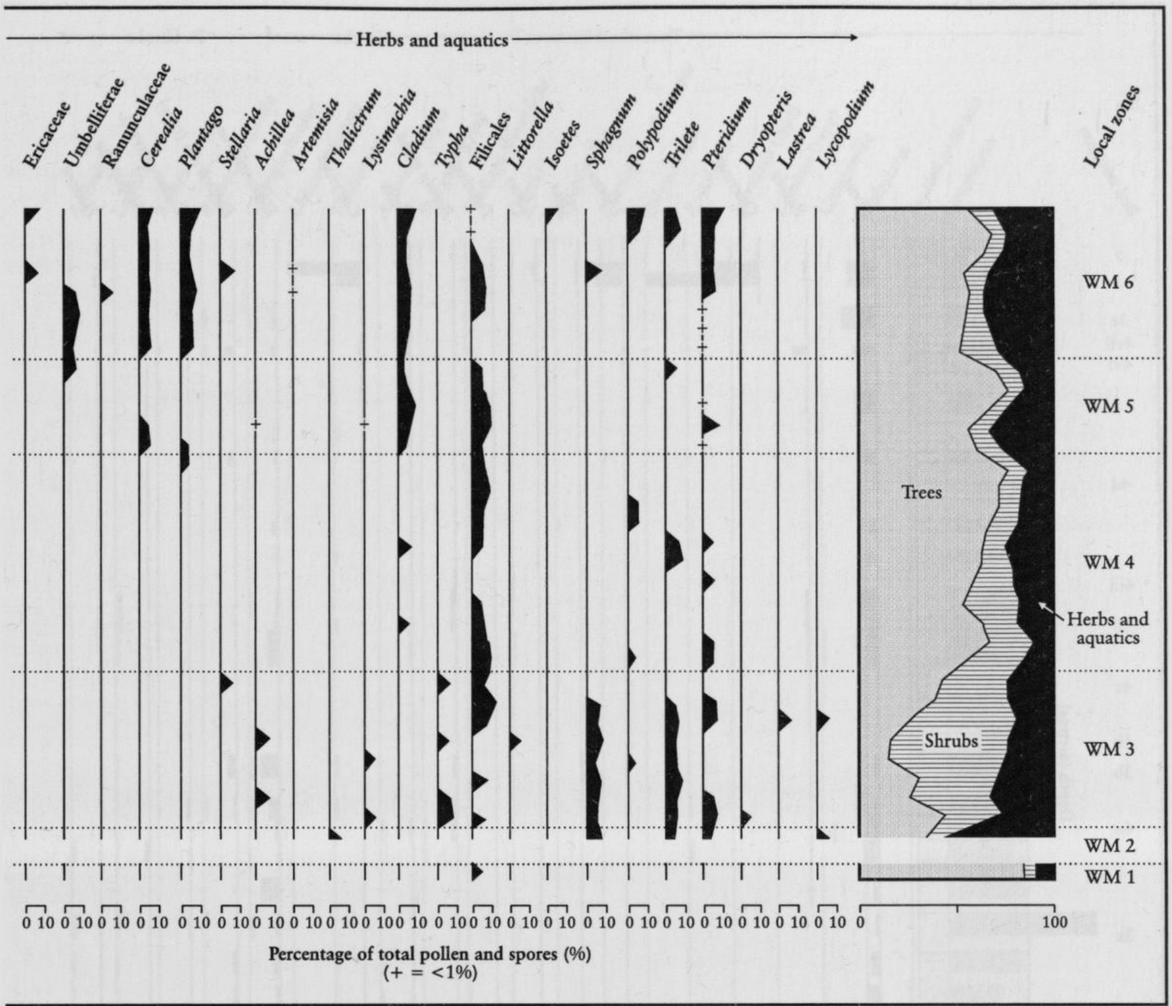


Figure 8.28 Pollen diagram from Skipsea Withow Mere; analysis by A. Blackham. Values are percentages of total pollen and spores. See Figure 8.1 for key to the stratigraphical log.

studied the Late Devensian palaeobotanical evidence at several locations. The lacustrine silt–clay and gravel units all contained cold-stage-type pollen with high frequencies of Gramineae and Cyperaceae. The rich plant macrofossil assemblage also was dominated by Cyperaceae. *Betula* (cf. *B. nana*) pollen was common, as were *B. nana* leaf impressions and macrofossils. *Salix* pollen and macrofossils occurred, but *Juniperus* and *Populus* were represented only by macrofossils. Dwarf shrub pollen was common, with *Helianthemum*, *Hippophäe*, *Rhamnus* and various Ericaceae. A wide range of tundra-type open habitat herbs are present in both the pollen and the macrofossil assemblages. *Artemisia*, *Rumex*, *Thalictrum*, Chenopodiaceae, *Plantago* sp. and various Compositae are characteristic. The Late Devensian pollen diagrams through units 2, 3 and 4 (Hunt *et al.*, 1984) are shown as Figures 8.29 and 8.30. Aquatic plant macrofossils are common, with *Chara*, *Potamogeton*, *Hippuris*, *Eleocharis* and *Sphagnum* notable. Pollen analysis was also completed on the silts of unit 3 in which the flint blade was found. An open Late-glacial type assemblage was recovered, notably Gramineae, Cyperaceae and *Artemisia*, with some *Betula* and *Pinus* and single records of more temperate-type taxa such as *Alnus* and *Quercus*.

Thew and Woodall (1984) performed highly detailed molluscan analyses of several sections at the site, and present full species lists of the forms identified. They were able to distinguish land snail and freshwater snail and bivalve ele-



ments in the faunas, with several species of *Valvata*, *Lymnaea* and *Pisidium* abundant among a very diverse assemblage. Correlation of the molluscan biostratigraphies and their ecological interpretation provided information on changes in water depth and trophic status, local vegetation, erosion, sedimentation and climate, particularly in relation to the Late-glacial lacustrine deposits. Other faunal remains from the site were undiagnostic and consisted of a restricted vertebrate assemblage of frogs, voles and shrews in the Late-glacial sediments (Jenkinson, 1984) and a poorly preserved insect fauna (Kenward, 1984). In the latter case a typical Late Devensian beetle *Arpedium brachypterum* occurred with other cold-adapted types in the Late Devensian levels, whereas swamp and fen woodland species characterized the Holocene peat units.

Mellars (1984) considered the Palaeolithic and Mesolithic archaeological finds from Skipsea Withow. He concluded that the flint blade found in the lacustrine silts of unit 3 had typological affinities with blade industries of Late Devensian age from elsewhere in northern England, and so was consistent with its inferred Late-glacial sedimentary context. Mellars agreed with Clark and Godwin (1956) that the 'harpoon' bone point recovered from sub-peat silt (Armstrong, 1923) typologically resembled other Holderness examples from Brandesburton and Hornsea, rather than the early Holocene examples from Star Carr and the Vale of Pickering to the north. He suggested that a late Late Devensian age for the

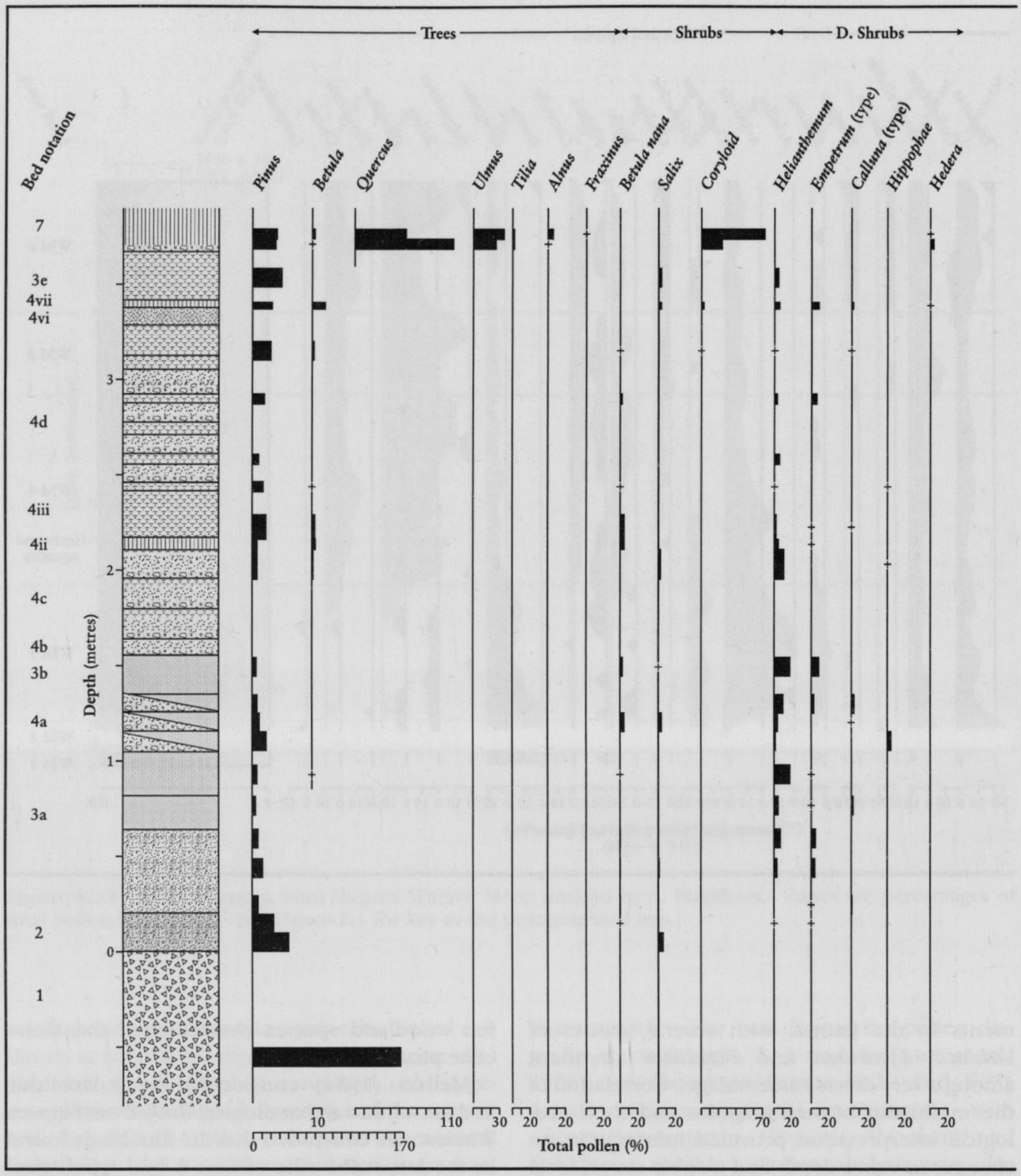
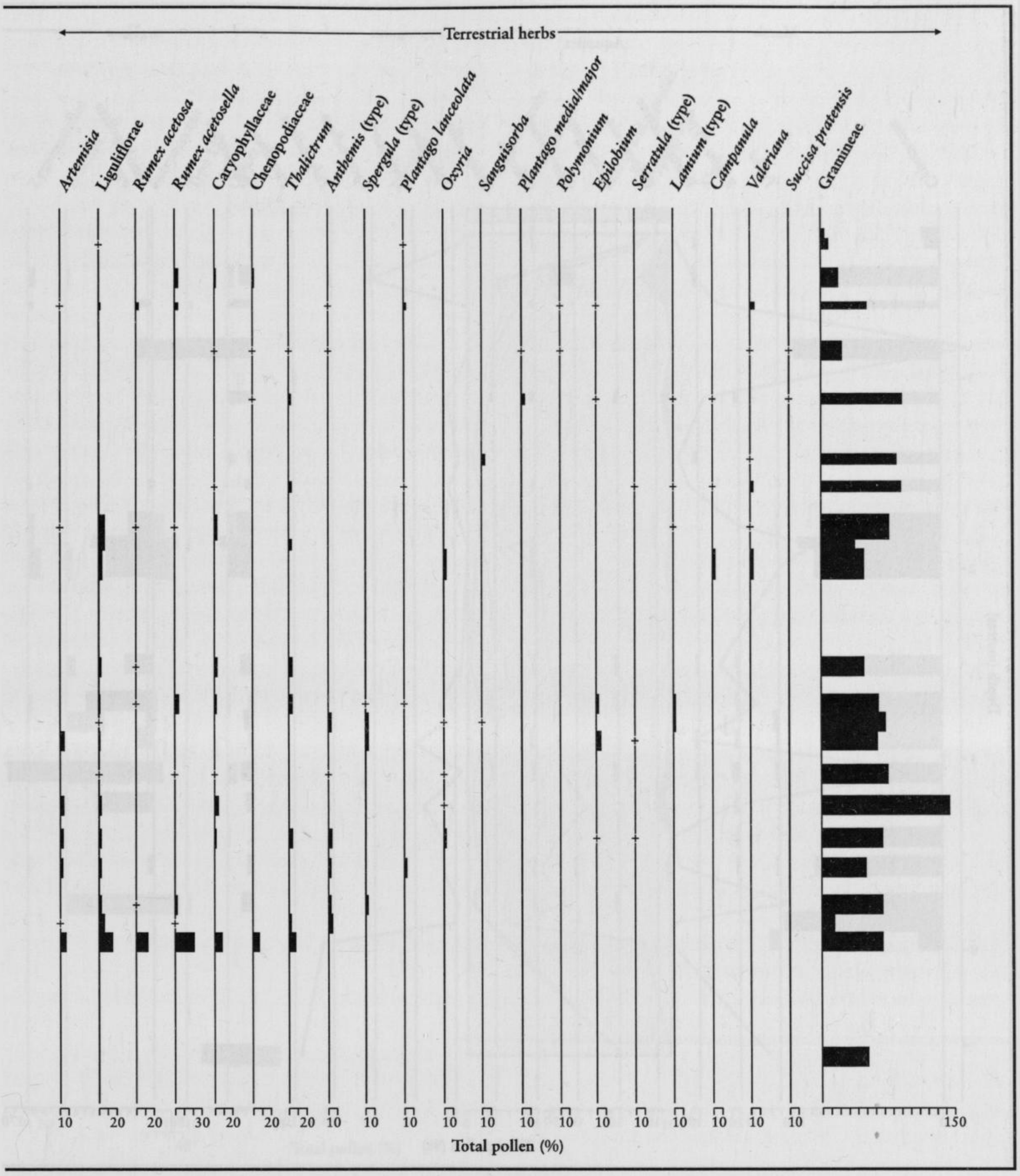


Figure 8.29 Pollen diagram from the Late Devensian deposits at Skipsea Withow showing tree, shrub and terrestrial herb taxa. For stratigraphy, see text.

implement was quite possible and that this would accord with the probable correlation of its original find context with the sandy silt of unit 5. Flint tools recovered by Armstrong (1923) from Skipsea Withow could be assigned to early Mesolithic and Late Upper Palaeolithic typologies (Mellars, 1984).

Interpretation

The exposure by coastal erosion of laterally extensive sections through the sediment succession allows a three-dimensional spatial reconstruction of stratigraphical changes and the evolution of depositional environments at Skipsea



Withow, in contrast to the single core records from most sites. Gilbertson (1984a), Gilbertson *et al.* (1987) and Dinnin (1995) have interpreted the environmental history at the site as follows. The radiocarbon date of 13 045 ± 270 years BP from The Bog, Roos (Beckett, 1981) provides a limiting age for deglaciation of this area. It and similar dates reflect organic inception within basins in the glaciogenic terrain, however, and the true deglaciation date must lie between that age and the localized ice surge (Eyles *et al.*, 1994) that occurred in east Yorkshire after the deposition of the Dimlington Silts about 18 000 years BP (Penny *et al.*, 1969; Catt, 1991b). The

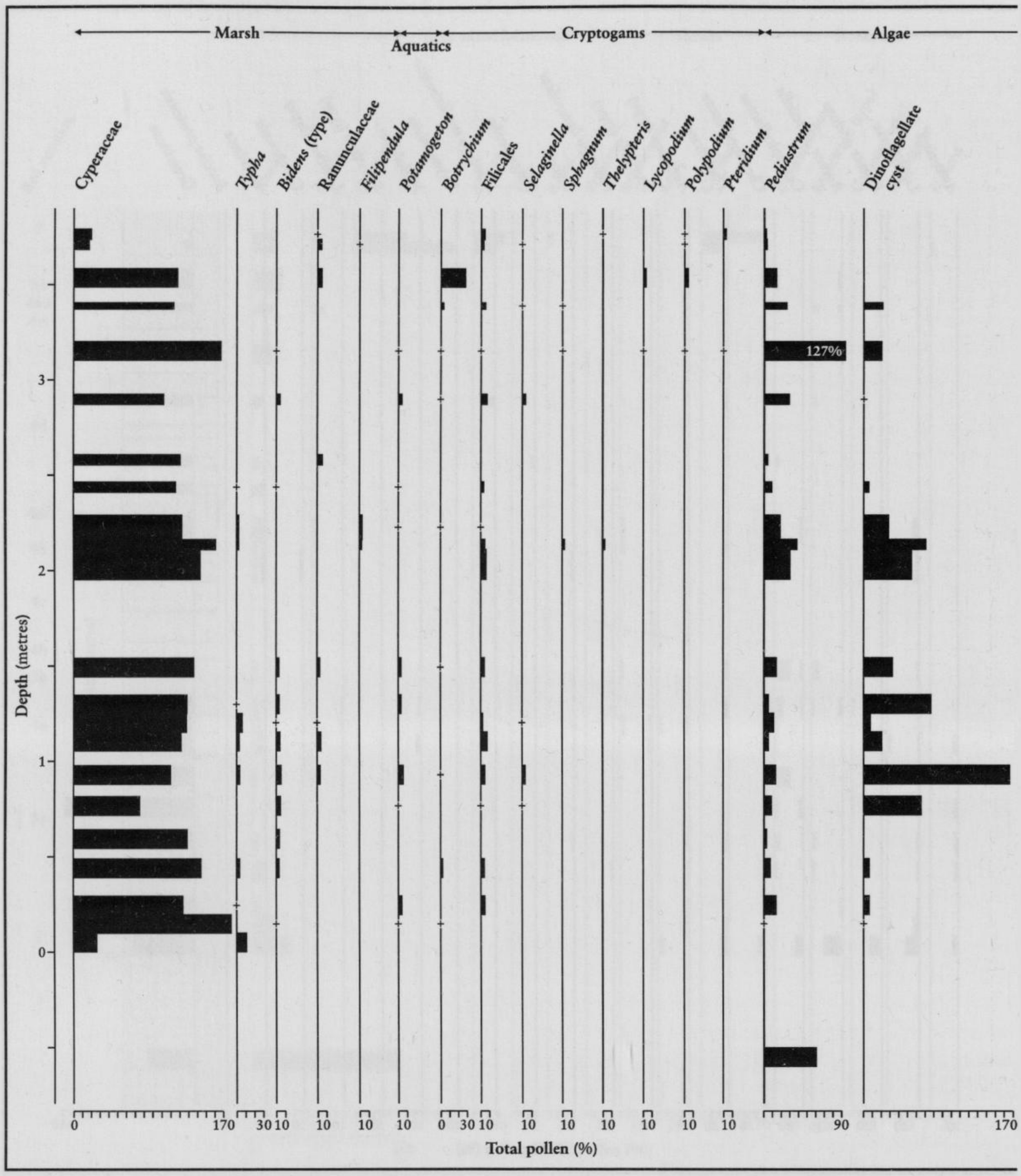
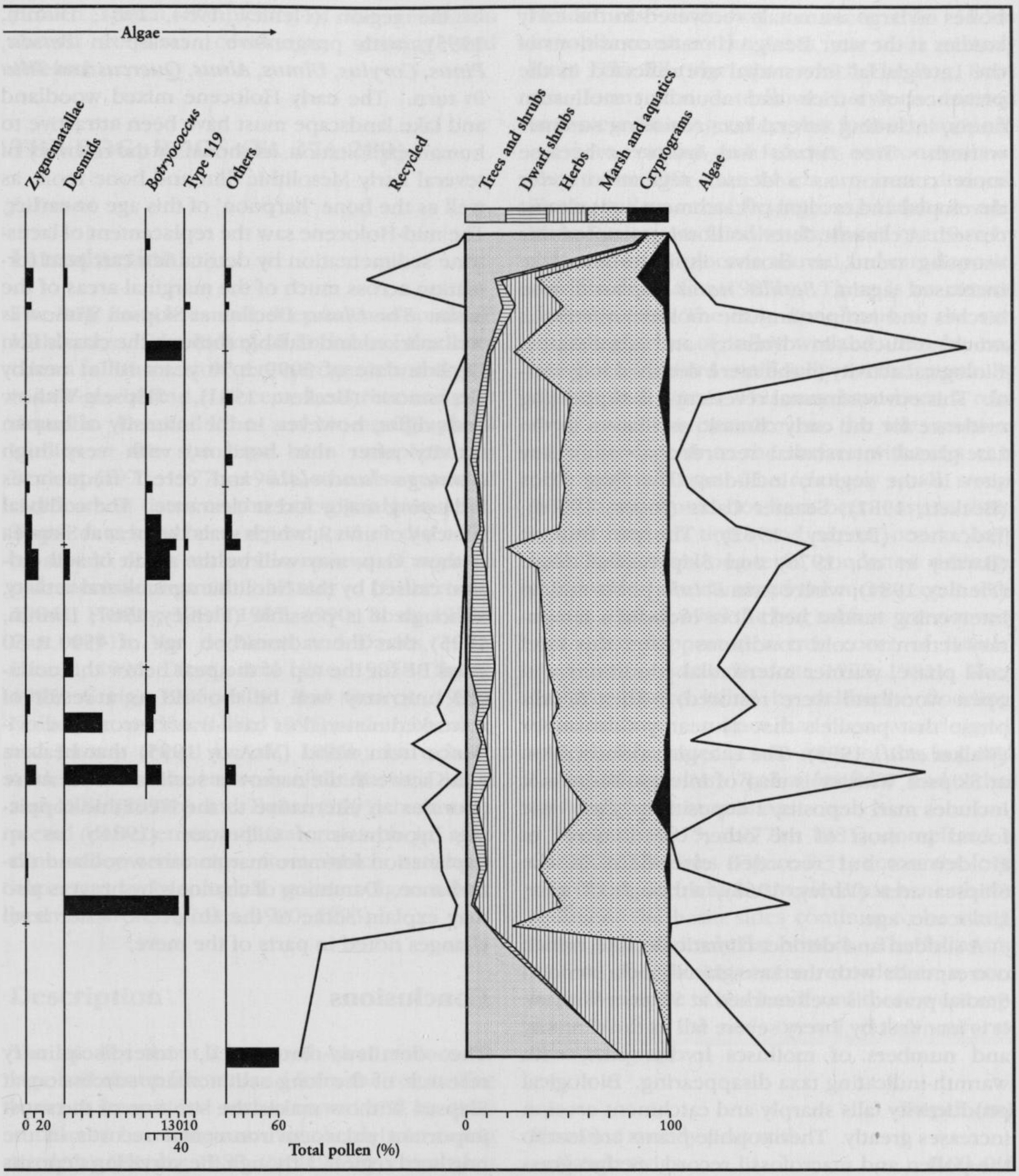


Figure 8.30 Pollen diagram from the Late Devensian deposits at Skipsea Withow showing marsh taxa, cryptograms and algae.

Late-glacial history at Skipsea Withow is likely to be comparable with that investigated in detail at nearby Gransmoor (Walker *et al*., 1993). Prior to the establishment of lake environments in the Skipsea Withow Gap area, desiccation and weathering of the till surface was followed by hillwash mudflows under unstable slope conditions, laying down unit 2. Lake formation in the Withow basin then took place and water levels reached over 8 m higher than present beach level at the Withow Gap, creating pebble lag deposits. Varved lake margin sediments of early unit 3 reflect repeated freezing of the lake. Climatic amelioration allowed the colonization



of the lake and its environs by pioneer plant and animal taxa; freshwater bivalves, *Chara* and pondweeds in the water and tundra type herbs on unstable soils with poor vegetation cover in the catchment. Continued climatic warming in the early Late Devensian encouraged the spread of a rich thermophilous herb flora with increasing *Betula*, *Populus* and *Salix* parkland. Fluctuating lake levels occurred that caused localized erosion and reworking of sediments. The enigmatic bone 'harpoon' find and the flint blade in unit 3 silts suggests human activity around this developing lake ecosystem, probably attracted by the hunting opportunities indicated by the bones of large mammals recovered in the early studies at the site. Benign climatic conditions of the Late-glacial interstadial are reflected in the presence of a rich and abundant molluscan fauna, including several taxa requiring summer warmth. Tree *Betula* and *Juniperus* became more common as a denser vegetation cover developed and erosion of catchment soils almost ceased. A climatic deterioration interrupted this warming trend, as erosive input to the lake increased again, *Betula nana* replaced tree birches and juniper and the molluscan fauna is much reduced in diversity and abundance. Biological activity in the mere declined in general. This environmental revertence is supporting evidence for the early climatic oscillation in the Late-glacial interstadial recorded at other key sites in the region, including The Bog, Roos (Beckett, 1981), Seamer Carrs (Jones, 1976a), Tadcaster (Bartley, 1962), Thorpe Bulmer (Bartley *et al.*, 1976) and Skipsea Bail Mere (Flenley, 1984), where twin *Betula* peaks and an intervening tundra herb flora indicate a temporary return to cold conditions. After this brief cold phase, warmer interstadial conditions and open woodland were restored, with a *Betula* phase that parallels that at nearby Gransmoor (Walker *et al.*, 1993). The Late-glacial succession at Skipsea Withow is also of interest because it includes marl deposits, a depositional facies not found in most of the other cored meres of Holderness but recorded elsewhere in the Skipsea area (Varley, 1968), although of later, Holocene, age.

A sudden and distinct climatic decline, which corresponds with the Late-glacial Loch Lomond Stadial period is well marked at Skipsea Withow. It is manifest by a very severe fall in the diversity and numbers of molluscs in the lake, with warmth-indicating taxa disappearing. Biological productivity falls sharply and catchment erosion increases greatly. Thermophile plants are lost to the pollen and macrofossil record. Sedge–grass tundra replaced the birch parkland. The radiocarbon dates of 10 440 ± 80 years BP and 10 710 ± 70 years BP for wood in sediments of this phase are compatible with Loch Lomond Stadial times. Reworking of earlier sediments and deposition of coarse gravels in parts of the site occurred under these severe climatic conditions.

The Holocene climatic amelioration and spread of forest at Skipsea Withow (Blackham and Flenley, 1984) is similar to that at other sites in the region (Flenley, 1984, 1987; Dinnin, 1995), with progressive increase in *Betula*, *Pinus*, *Corylus*, *Ulmus*, *Alnus*, *Quercus* and *Tilia* in turn. The early Holocene mixed woodland and lake landscape must have been attractive to human exploitation, as shown by the recovery of several Early Mesolithic flint and bone tools, as well as the bone 'harpoon' of this age or earlier. The mid-Holocene saw the replacement of lacustrine sedimentation by detrital fen-carr peat formation across much of the marginal areas of the basin. The *Ulmus* Decline at Skipsea Withow is well marked and datable through the classic Elm Decline date of 5099 ± 50 years BP at nearby Gransmoor (Beckett, 1981). Skipsea Withow does differ, however, in the intensity of human activity after this horizon, with very high *Plantago lanceolata* and cereal frequencies indicating major forest clearance. The colluvial silt–clay of unit 9, which seals the peat at Skipsea Withow Gap, may well be the result of soil erosion caused by this Neolithic agricultural activity, although it is possible (Flenley, 1987; Dinnin, 1995) that the radiocarbon age of 4500 ± 50 years BP for the top of the peat below this colluvial unit may well be too old as a result of reworked material or hard-water error. The evidence from wood (McAvoy 1995) that beavers were active in the narrower sections of the mere provides an alternative to the Neolithic coppicing hypothesis of Gilbertson (1984b) as an explanation for mere-margin carr woodland disturbance. Damming of channels by beavers also may explain some of the Holocene water level changes noted in parts of the mere.

Conclusions

The detailed, integrated, interdisciplinary research of the long sedimentary succession at Skipsea Withow makes the site one of the most important palaeoenvironmental records in the northern region, although the surviving deposits are probably only a small fraction of the original extent of the mere. The site also is important as one of the first in Britain where archaeological remains were relatively dated using pollen analysis (Godwin and Godwin, 1933). The threat to the survival of the geological resource at Skipsea Withow Gap is very great as deposits continue to be lost to coastal erosion, although mere sediments to the west still survive for analysis and may contain the late Holocene record truncated at the Withow Gap site. The proven spatial complexity of the geological record requires continuing scrutiny and research.

THE BOG, ROOS (TA 274 289)

J. Innes

Introduction

The Bog, Roos is a small, deep basin in southern Holderness, East Riding of Yorkshire, that contains a full succession of minerogenic and organic Late-glacial and Holocene lacustrine and peat deposits. Detailed litho- and bio-stratigraphical studies at the site have been completed by Beckett (1975, 1977b, 1981), who also reported radiocarbon dates on the Late-glacial sequence. These data have been reproduced or discussed by several other authors in reviews of regional palaeoenvironmental history (Gilbertson, 1984a; Flenley, 1984, 1987, 1990; Rose, 1989a; Catt, 1991b; Van de Noort and Davies, 1993; Dinnin, 1995; Dinnin and Lillie, 1995b; Taylor, 1995; Greig, 1996; Lillie and Gearey, 1999). Colluvium studies (Lillie, 1995) and archaeological survey (Head *et al*., 1995b) have been undertaken in the environs of the basin. Dinnin and Lillie (1995b) and Taylor (1995) have completed palaeoenvironmental research in adjacent valley sediments, from which nationally important archaeological wooden figures have been recovered (Coles, 1990, 1993).

Description

The site lies about 2 km south-west of Roos village and about 6 km inland from the coast. The basin is steep-sided and is considered by most authors to have originated as a kettlehole during ice decay (Flenley, 1987; Catt 1991b), although it is possibly a collapsed pingo (Berridge and Pattison, 1994), as it is enclosed almost entirely by a till rim of higher ground that may represent the pingo rampart. The bog surface lies at about 5 m OD, and the rim altitude is between 8 and 15 m OD. The oval basin, about 220 m by 120 m in size, lies on a ridge of Withernsea Till (Catt and Penny, 1966; Madgett and Catt, 1978; Catt, 1991b), now reclassified as the Withernsea Member of the Holderness Formation (Lewis, 1999). This till unit is restricted to south-east Holderness, extending no more than 10 km from the coast, so the site is near to its westerly limit. The till ridge upon which The Bog, Roos sits forms the interfluve between incised east–west draining valleys that formerly supported mere, swamp and fen-carr environments at about 3 m OD, but which are now drained. These valleys, the Keyingham, Halsham and Roos Drains, are filled to depths of over 9 m by mid- and late Holocene alluvial and carr deposits (Dinnin and Lillie, 1995b). There is no natural drainage from The Bog, Roos basin to the surrounding areas of valley wetland, but the bog has been drained artificially to the north by ditches and pipes. This may have been a relatively recent occurrence, as the site contains unoxidized peats to the surface without a capping mineral layer. The relatively small, steep-sided basin would not have been easy to drain before modern times and seems not to have been subject to long-term, concerted drainage that would have damaged the more recent sediments. Before 1968 the bog surface was not very wet and supported a tree cover of *Betula*, *Quercus* and *Larix* (Beckett, 1975, 1981). Then the trees were killed by deliberate flooding, since when the surface has been waterlogged, with surface water across most of the bog and *Salix*, *Phragmites*, *Typha latifolia* and other wetland plants increasing.

Beckett (1975, 1977b, 1981) completed a transect of borings (Figure 8.31) across the long axis of the site, which revealed that beneath the sediment the basin sides continue to be steep, down to a relatively flat bottom. The following 11.5 m of sediments were recorded in a core from the centre of the basin and these were representative of the site's overall, rather uniform, lithostratigraphy.

	Depth (m)
Humified woody peat with *Betula* twigs	0.00–0.4
Less humified peat with plentiful *Sphagnum*, *Menyanthes trifoliata* seeds and Ericaceae stems. *Polytrichum* sp. at 1.80 m	0.40–2.00
Humified coarse detritus mud, with *Sphagnum* leaves	2.00–3.90
Fine, dark brown detritus mud with occasional leaves of *Quercus robur*, *Betula pubescens* and *Salix* sp.	3.90–9.30

Pinkish grey clay with occasional moss fragments of *Fontinalis antipyretica*	9.30–10.90
Fine, dark detritus mud	10.90–11.16
Slightly organic grey clay	11.16–11.33
Fine, dark detritus mud with a little clay	11.33–11.36
Grey clay, gravelly at the base, with moss fragments of *Leptodictyum* sp.	11.36–11.50

Pollen analyses were conducted on this central core (Beckett, 1981), and these are shown in Figure 8.32 in the form recalculated and redrawn by Flenley (1984, 1987, 1990), with frequencies as percentages of total dry-land pollen. Eleven pollen assemblage zones are recognized, from the base of the diagram, as follows.

RB1a *Betula* and *B. nana* are the important taxa, with lesser Gramineae and Cyperaceae.

RB1b *Betula* values are greatly reduced. Gramineae and Cyperaceae characterize the pollen assemblage, with high *Helianthemum* and *Hippophäe* and significant *Artemisia* and *Thalictrum*.

RB2 *Betula* dominates the zone, with *Pinus* also increased. *Filipendula* rises in frequency but all other shrub and herb types are greatly reduced.

RB3 Cyperaceae characterizes the assemblage. *Betula* frequencies fall sharply, but *B. nana*, Gramineae, *Juniperus* and *Artemisia* all increase. *Thalictrum* rises late in the zone. *Pinus* is important.

RB4 *Betula* rises sharply to 80% of land pollen and dominates the assemblage, although *Pinus* remains significant. *Filipendula* frequencies show a peak and aquatic herb taxa values rise.

RB5 *Corylus/Myrica* percentages rise sharply to dominate the zone at almost 80%. No other taxa are significant except *Ulmus* and, later in the zone, *Quercus*.

RB6 A rise in *Alnus* pollen to almost 30% characterizes the zone, with *Corylus/Myrica* pollen still at high frequencies. *Ulmus* and *Quercus* are in low but significant values, and *Tilia* rises later in the zone. All other pollen types are very low.

RB7 *Ulmus* frequencies decline to very low values, and *Tilia* also is much reduced before recovering. *Alnus* and *Corylus/*

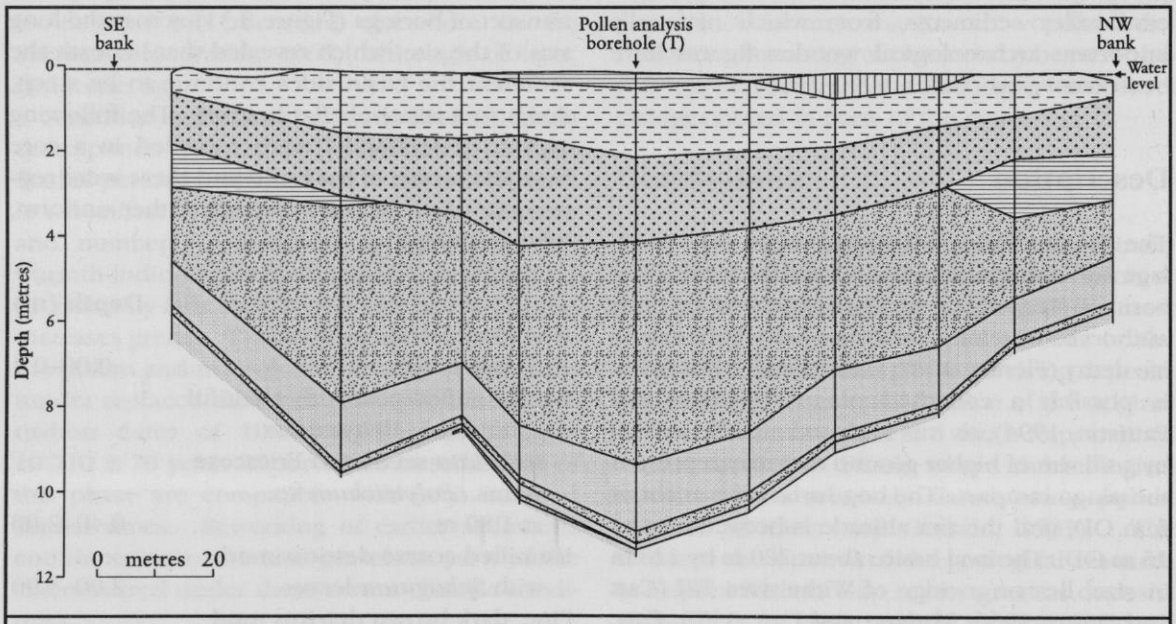


Figure 8.31 Stratigraphy of the deposits along a SE–NW transect.

Myrica continue to dominate the assemblage and *Quercus* and *Betula* are both increased. *Plantago lanceolata* is recorded for the first time, and other dry-land herb types increase. *Sphagnum* spores are common.

RB8 Characterized by a major fall in all tree and shrub taxa. *Calluna* and *Empetrum* show very high peaks in mid-zone. Gramineae and *Plantago lanceolata* frequencies are high. Several ruderal dry-land herb taxa are consistently present. Peaks in *Sphagnum* occur.

RB9 Gramineae dominates the assemblage and *Plantago lanceolata* remains high. A peak of cereal-type pollen occurs, with many ruderal herbs. *Betula* increases slightly and *Sphagnum* spores are important.

RB10 *Betula*, *Alnus* and *Pinus* values rise and all herb pollen frequencies fall, although Gramineae remains in moderate percentages.

Five radiocarbon dates were obtained (Beckett, 1981), all in the Late-glacial sediments, and these are shown on Figure 8.32. The upper 9 m of the core remain undated, other than relative to the major pollen stratigraphy changes. The radiocarbon dating control for the Late-glacial succession allowed Beckett (1981) to calculate sediment-accumulation and pollen-influx rates. The latter are shown on Figure 8.33 for selected, major taxa, calculated as pollen grains per square centimetre per radiocarbon year. In zones RB1a and RB1b influx rates are very low, although higher in zone RB1a. *Betula* and *B. nana* provide most of the pollen influx, in accordance with the pollen percentage data. Rates in zone RB1b are very low indeed.

In zone RB2 tree pollen influx rises to over 300, contributed mainly by *Betula*, and total rates also rise sharply. Fluctuations for other taxa match the percentage changes. In zone RB3 total influx values fall markedly, although they still exceed those for zone RB1b. Values are lowest in the first half of RB3, recovering towards the top of the zone. *Betula* is much reduced. Only *Juniperus* is marginally increased relative to the earlier zones. Sediment accumulation rates are much higher in the clastic sediment of zone RB3 than in the rest of the Late-glacial sequence. In zone RB4 total influx rates increase greatly, contributed mainly by *Betula*, which reaches well over 1000. *Pinus*, *Salix*, Gramineae, Cyperaceae and *Filipendula* all rise sharply also, although none exceed 100. The data on influx rate support the pollen percentage evidence for the whole of the Late-glacial succession.

Interpretation

The major importance of The Bog, Roos lies in the great thickness of the sediments preserved in the basin and the apparently unbroken Late-glacial and Holocene record that they preserve. Resting upon the Withernsea Member near to its arcuate western margin, the basal sediments at the site can provide a limiting age for ice retreat (Catt, 1991b). With the radiocarbon evidence from Dimlington (Penny *et al.*, 1969), the lowermost date from The Bog, Roos therefore can constrain estimates of the age of the latest Devensian Stadial ice-surge advance in the Holderness area (Eyles *et al.*, 1994). The lowest date at the site, 13 045 ± 270 years BP, is closely comparable to the few other dates available from east Yorkshire for earliest organic deposition after ice retreat, for example that of 13 042 ± 140 years BP from Seamer Carrs in the Cleveland lowlands (Jones, 1976a). The interval between deglaciation and revegetation and organic deposition is conjectural, however, as ice removal may well have occurred considerably before *c.* 13 000 years BP. These dates at The Bog, Roos and elsewhere are more a measure of climatic improvement than deglaciation and remain of coarse limiting value. The very large standard deviation on these early Late-glacial dates further increases their uncertainty.

Nevertheless, the Late-glacial date series from The Bog, Roos has been one of the most valuable, at least prior to the major allocation of dates to the Late-glacial sequence at Gransmoor in the Hull valley (Walker *et al.*, 1993; Lowe *et al.*, 1995a, b). The date at The Bog, Roos of 13 045 ± 270 years BP is on a 3 cm thick layer of organic, detritus mud 3 cm thick, which overlies the basal, gravelly grey clay. This thin mud layer lies below the thicker Late-glacial interstadial organic layer, separated from it by an inorganic silt, and so is evidence of a climatic amelioration prior to that of the main (cf. Allerød) interstadial. This early Late-glacial organic mud unit has been observed at other sites in the region, such as Skipsea Bail Mere (Flenley, 1984, 1987). This early warmer climatic phase is also manifest in

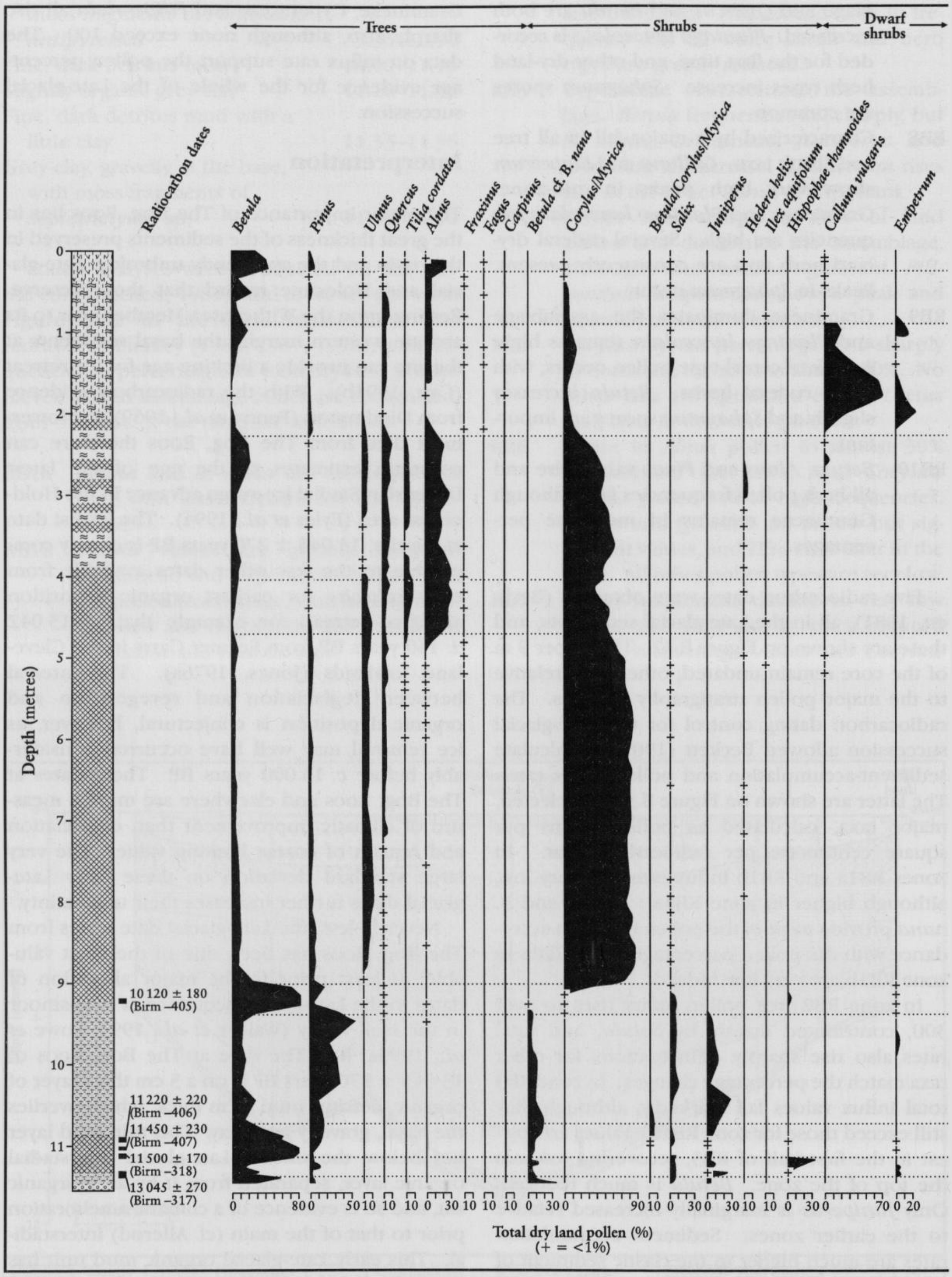
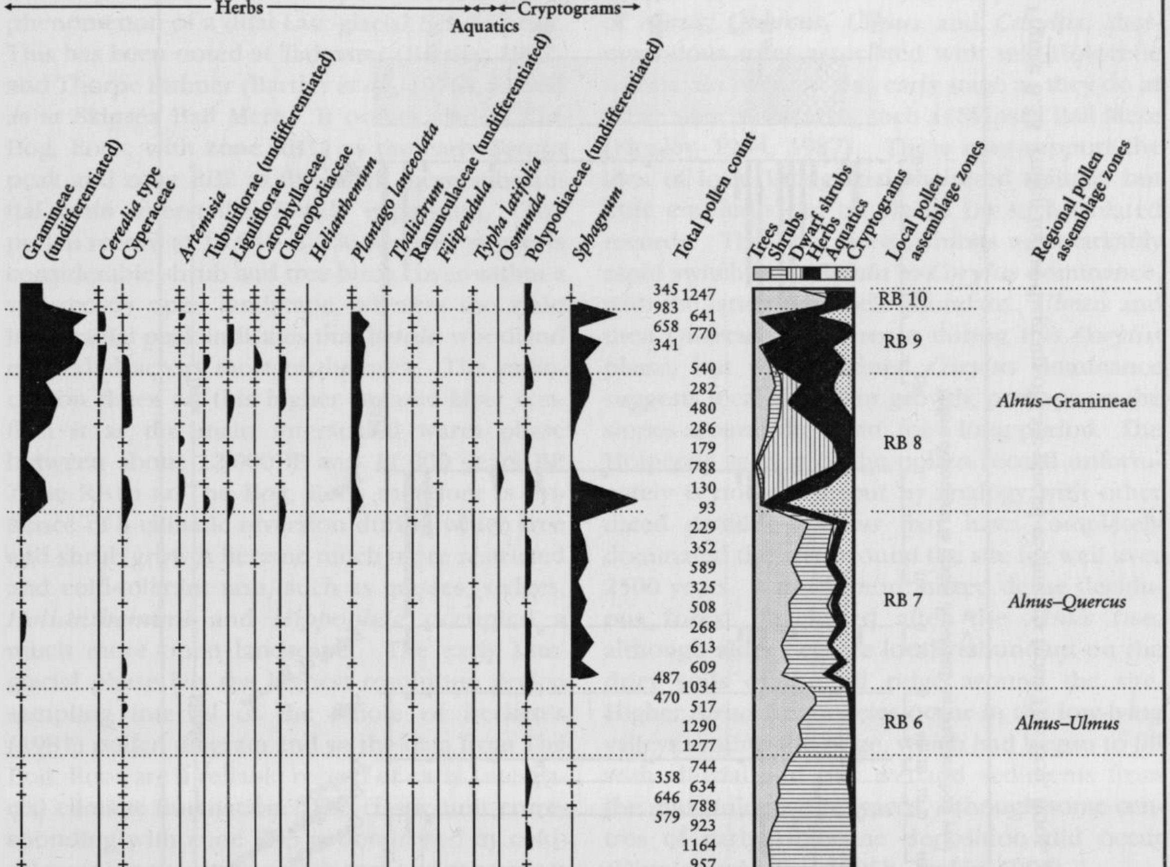


Figure 8.32 Pollen diagram from The Bog, Roos. Only selected pollen taxa are shown. Recalculated and redrawn from Beckett (1981) and Flenley (1984). See Figure 8.1 for key to the stratigraphical log.


Herbs
Aquatics
Cryptograms
Gramineae (undifferentiated)
Cerealia type
Cyperaceae
Artemisia
Tubuliflorae (undifferentiated)
Liguliflorae
Caryophyllaceae
Chenopodiaceae
Helianthemum
Plantago lanceolata
Thalictrum
Ranunculaceae (undifferentiated)
Filipendula
Typha latifolia
Osmunda
Polypodiaceae (undifferentiated)
Sphagnum
Total pollen count
Trees
Shrubs
Dwarf shrubs
Herbs
Aquatics
Cryptograms
Local pollen assemblage zones
Regional pollen assemblage zones
RB 10
RB 9
RB 8
RB 7
RB 6
RB 5
RB 4
RB 3
RB 2
RB 1b
RB 1a
Alnus–Gramineae
Alnus–Quercus
Alnus–Ulmus
Corylus/Myrica–Ulmus
Betula–Pinus
Total dry land pollen (%)
(+ = <1%)

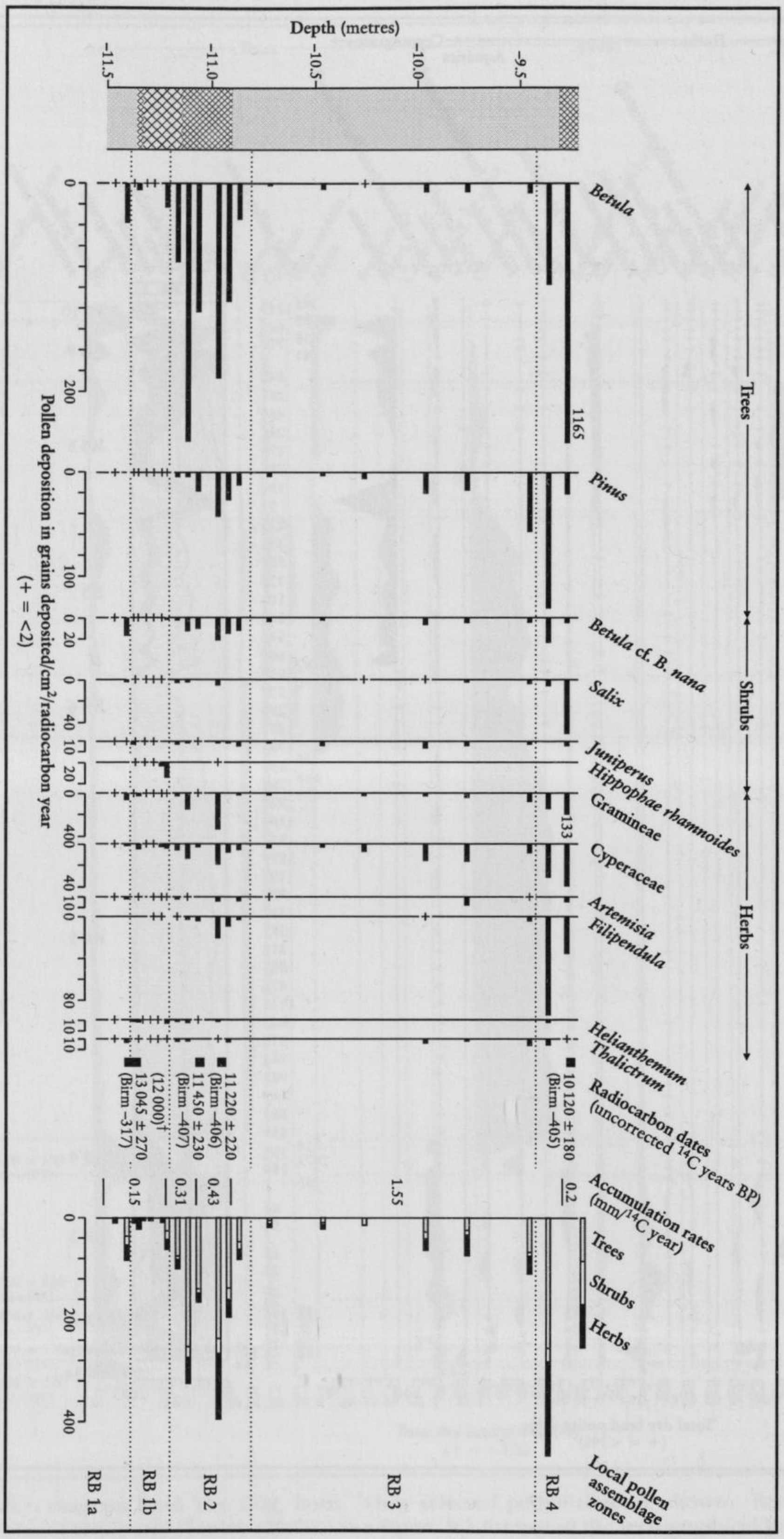


Figure 8.33 Pollen influx diagram of Late-glacial sediments from The Bog, Roos. Selected taxa only. See Figure 8.1 for key to the stratigraphical log.

the pollen record as higher frequencies of tree *Betula* pollen at some sites, so producing the phenomenon of a dual Late-glacial *Betula* peak. This has been noted at Tadcaster (Bartley, 1962) and Thorpe Bulmer (Bartley *et al.*, 1976), as well as at Skipsea Bail Mere. It occurs also at The Bog, Roos, with zone RB1a as the early *Betula* peak and zone RB2 as the much more substantial main interstadial *Betula* expansion. The pollen record at 13 045 ± 270 years BP suggests considerable shrub and tree birch cover within a still mostly open landscape, whereas the main interstadial peak indicates that *Betula* woodland expanded across most of the area. The radiocarbon dates on this higher organic layer confirm it as the main interstadial warm phase between about 12 000BP and 11 000 years BP. Zone RB1b at The Bog, Roos therefore is evidence of a climatic reversion during which tree and shrub growth became much more restricted and cold-tolerant taxa, such as grasses, sedges, *Helianthemum* and *Hippophäe* occupied a much more open landscape. The early Late-glacial phase has the highest resolution pollen sampling interval of the whole of Beckett's (1981) pollen diagram and so the data from The Bog, Roos are a reliable record of early Late-glacial climatic fluctuation. The clastic unit corresponding with zone RB3 is dominated by cold-tolerant pioneer herb pollen taxa and represents a major phase of severe cold climate, probably arid, tundra-type conditions. Its radiocarbon parameters of 11 220 ± 220 and 10 120 ± 180 years BP confirm it as the Younger Dryas (Loch Lomond) Stadial, with poor vegetation cover, soil instability and erosion. Increased *Juniperus* values may represent some growth of that shrub within the rim of the basin on the more sheltered slopes. The return of warm climate conditions at the start of the Holocene Epoch is marked in the lithology by a switch to highly organic detritus lake mud above the Loch Lomond Stadial clay. This limnic sequence persists until the late Holocene and the site was an open water lake for most of its history.

The early Holocene vegetation succession is similar to that of most other sites in the region. A very rapid expansion of *Betula* woodland occurred, supressing shrub taxa such as *Juniperus* before they could become abundant as a transitional community. The classic earliest Holocene *Filipendula* peak does occur at the site, however. Tree *Betula* may well have had local refugia in which it survived the Late-glacial stadial and from which it could spread quickly to establish a dense cover. Sporadic pollen records of *Alnus*, *Quercus*, *Ulmus* and *Corylus*, thermophilous trees associated with mid-Holocene forests, do occur at this early stage as they do at other sites in the area, such as Skipsea Bail Mere (Flenley, 1984, 1987). These may support the idea of local Late-glacial sheltered refugia, but little emphasis can be placed on such isolated records. The Bog, Roos exhibits a remarkably rapid switch from *Betula* to *Corylus* dominance, with the latter extremely abundant. *Ulmus* and then *Quercus* did increase during this *Corylus* phase, but the sustained *Corylus* dominance suggests local abundant growth, perhaps on the slopes around the basin, for a long period. The Holocene section of the pollen record unfortunately is not dated, but by analogy with other dated profiles *Corylus* may have completely dominated the area around the site for well over 2500 years. A much more mixed dense deciduous forest developed after the *Alnus* rise, although alder became locally abundant on the drier soils of the till ridge around the site. Higher *Alnus* frequencies occur in the low-lying valleys around the ridge, which had begun to fill with alluvial and carr wetland sediments from the mid-Holocene onwards, although some centres of early Holocene deposition did occur (Dinnin and Lillie, 1995b; Taylor, 1995).

After a clear *Ulmus* Decline, for which the comparable regional date is 5099 ± 50 years BP at Gransmoor Quarry (Beckett 1975), deposition in the basin changed from limnic to peat-forming conditions as sediment infilling of the lake progressed. There is evidence of some human forest clearance after the *Ulmus* Decline, with *Tilia* also reduced and moderate frequencies of clearance herbs such as *Plantago lanceolata*. Intensive forest clearance occurs in zones RB8 and RB9, with tree pollen reduced to very low values. Cereal agriculture is significant in the later clearance phase. Without radiocarbon support these phases cannot be attributed to any particular culture, although it is probable, from regional comparisons, that this clearance began in the Bronze Age and intensified in the Iron Age (Flenley, 1990; Dinnin, 1995). More dating control is required. The uppermost levels are dominated by *Alnus*, *Betula* and *Salix* pollen derived locally from bog-surface scrub. These local taxa mask any pollen signal from further away from the site at that time. That Bronze Age and Iron Age people were active near The Bog, Roos is demonstrated by the recovery of much archaeological material during survey of the locality (Head *et al.*, 1995b). Of national significance are the early Iron Age Roos Carr wooden figurines, dated 2460 ± 70 years BP, recovered from beneath estuarine alluvium not far to the north of the site (Coles, 1990, 1993; Van de Noort and Davies, 1993).

Conclusions

The Bog, Roos is a classic site for the reconstruction of Late-glacial and Holocene vegetation history and provides a regional standard for the increasing body of palaeoenvironmental data yielded by recent research in Humberside and environs (Dinnin, 1995). For the Late-glacial in particular, it remains one of the key lowland sites for palaeoecological syntheses in eastern north England (Catt, 1977b; Rose, 1989a; Greig, 1996). The lack of radiocarbon dates for the Holocene succession reduces its value, but the potential for research on this apparently hiatus-free, 9 m-long Holocene palaeoenvironmental record is very high indeed.

WILLOW GARTH (TA 126 676)

J. Innes

Introduction

Willow Garth is an area of carr woodland in the Great Wold Valley of the East Riding of Yorkshire that contains a shallow, discontinuous, succession of minerogenic and organic Late-glacial and Early and Late Holocene lacustrine and peat deposits. Detailed litho- and bio-stratigraphical studies and a series of radiocarbon dates from the site have been published (Bush, 1986, 1988a, 1993; Bush and Ellis, 1987; Bush and Flenley, 1987). These data and their interpretation have been debated (Thomas, K.D., 1989; Bush, 1989) and discussed in reviews of regional palaeoenvironmental history (Flenley, 1990; Day, 1996; Greig, 1996). Bush (1988b, 1993) also has completed modern molluscan and pollen studies. Willow Garth sediments have been classified within the Bingley Bog and Ringinglow formations by Thomas (1999). Bush and Hall (1987) have discussed the significance of Late-glacial and early Holocene *Alnus* records from the site, which is notified as an SSSI under the name 'Boynton Willow Garth' and is also a nature reserve sustaining a very rich and important flora and fauna.

Description

Willow Garth is an area of wet fen-carr woodland 200 m in diameter at about 18 m OD and a kilometre to the west of Boynton village in the Great Wold Valley, the main drainage system for the chalk upland of the Yorkshire Wolds. The valley is occupied by a small stream, the Gypsey Race, which is the only major example of surface drainage in the northern Wolds. The site lies just within the east coast Devensian ice limit, and till is present within the valley. Brown silty clay in the valley may be till-derived or reworked Devensian loess. Sands present around the site are probably derived from outwash material from Vale of Pickering ice, which just overtopped the northern Wolds scarp (Catt, 1982, 1987a; Foster, 1987). Willow Garth is a rare example of wetland sediment deposition in the chalk area and has been sustained by several springs in the woodland and by periodic flooding of the stream (Bush and Flenley, 1987). The wooded wetland area is surrounded by arable land and the presence of peaty soils in the cultivated valley bottom suggests that the wetland was once greater than today, and has been much reduced by drainage. Drainage ditches, cut through the modern wetland, and the Gypsey Race, bound the dense carr woodland on the northern side.

An extensive lithological survey of the deposits at Willow Garth was undertaken, comprising over 60 boreholes at less than 10 m intervals and two ditch sections, and the results (Bush and Ellis, 1987; Bush and Flenley, 1987; Bush, 1988a, 1993) have been published as a three-dimensional diagram (Figure 8.34). On this diagram transects A–B and B–K are the two ditch sections where continuous stratigraphical records were recorded. The borehole survey identified some small hollows, about 20 m by 30 m in size, in which the deepest sediments were preserved and these were investigated by cores at 1 m intervals.

A brown silty clay, structureless and without sand or gravel inclusions, is recorded in transect B–K as the lowermost unit at the site. It may be derived from nearby deposits of till in the valley but must have been transported as its characteristics indicate deposition under a low-energy fluvial or lacustrine regime (Bush and Ellis, 1987;

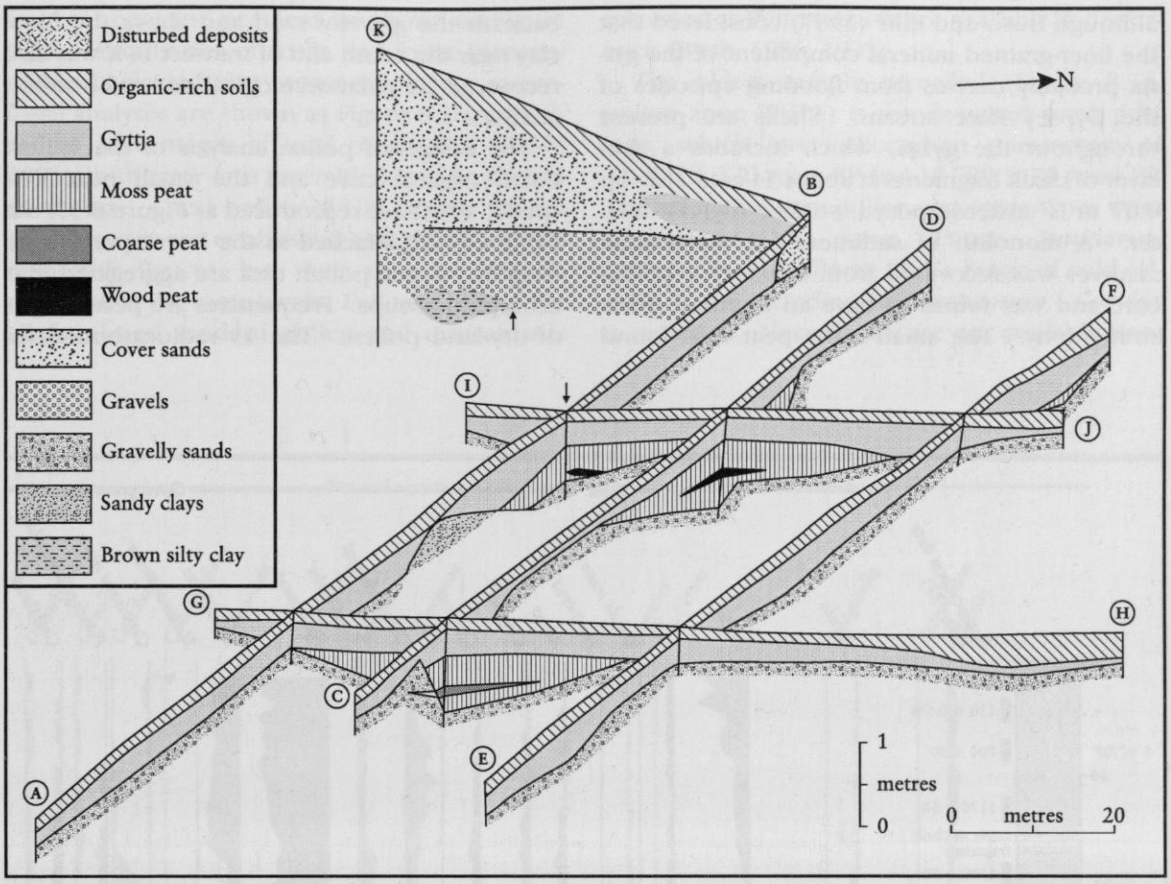


Figure 8.34 Three-dimensional diagram of the sedimentary deposits at Willow Garth (after Bush and Ellis, 1987). The lens of peat on section K–B is too small to show and is marked by an arrow. The main core analysed is arrowed on section A–B.

Bush, 1993). It may well represent reworked loessic material, which covered most of the higher Wolds surface in the later Devensian but was then weathered and eroded, some being redeposited in the Wolds valleys (Catt, 1990b). A small moss peat fragment with attached organic silt was discovered resting upon it in transect B–K. This basal silty clay is overlain by a complex suite of gravels and gravelly sands, which comprise the lowermost unit recorded across most of the site. Bush and Ellis (1987) show these to have had a down-valley fluvial origin and they contain large sand lenses. In places a coarse angular gravel occurs, which is interpreted as a solifluction deposit and it is probable that the main sand and gravel suite is fluvially reworked solifluncted material. A thick coarse sand body containing small chalk and flint clasts overlies the gravels in transect B–K, banked against the valley side slope. Lack of bedding and structure suggested to Bush and Ellis (1987) that this sand was aeolian coversand, its grain size is similar to coversands found elsewhere in the northern Wolds (Foster, 1987) and Vale of York (Matthews, 1970). Overlying these clastic gravelly sand deposits, within the small, wet hollows detected by the borehole survey, were various organic sediments. A core for pollen and radiocarbon analyses was collected from the location with the deepest organic sediment, 1.18 m, with a 7 cm-diameter Livingstone Piston Sampler. Its location on Figure 8.34 is at the intersection of transects C–D and I–J. From 0.48 m to 1.18 m the organic sediment is a moss peat composed mainly of *Amblystegium riparium*, *Calliergon* sp., *Drepanocladus* sp. and *Scorpidium scorpioides*. A wood-rich band occurs within the moss peat between 0.76 m and 0.82 m. From 0.07 m to 0.48 m is a gyttja of low organic content (Bush and Ellis, 1987). The gyttja merges laterally into the subjacent sandy stratum, which accounts for most of its content, although Bush and Ellis (1987) considered that the finer-grained mineral component of the gyttja probably derives from flooding episodes of the Gypsey Race stream. Shells are present throughout the gyttja, which includes a thin layer of chalk fragments at about 34 cm. The top 0.07 m is undecomposed, surface organic matter. A monolith of sediment for macrofossil analyses was recovered from beside the pollen core and was found to have an identical lithostratigraphy. The small moss peat lens found beneath the gravelly sand and above the basal clay near the south end of transect B–K was also recovered for palaeoecological and radiocarbon analysis.

The results of pollen analysis of the Willow Garth organic core and the small peat lens (Bush, 1993) are reproduced as Figure 8.35, the latter shown detached as the lower part of the diagram. Some pollen taxa are aggregated into ecological groups. Frequencies are percentages of dry-land pollen. The 13 radiocarbon dates

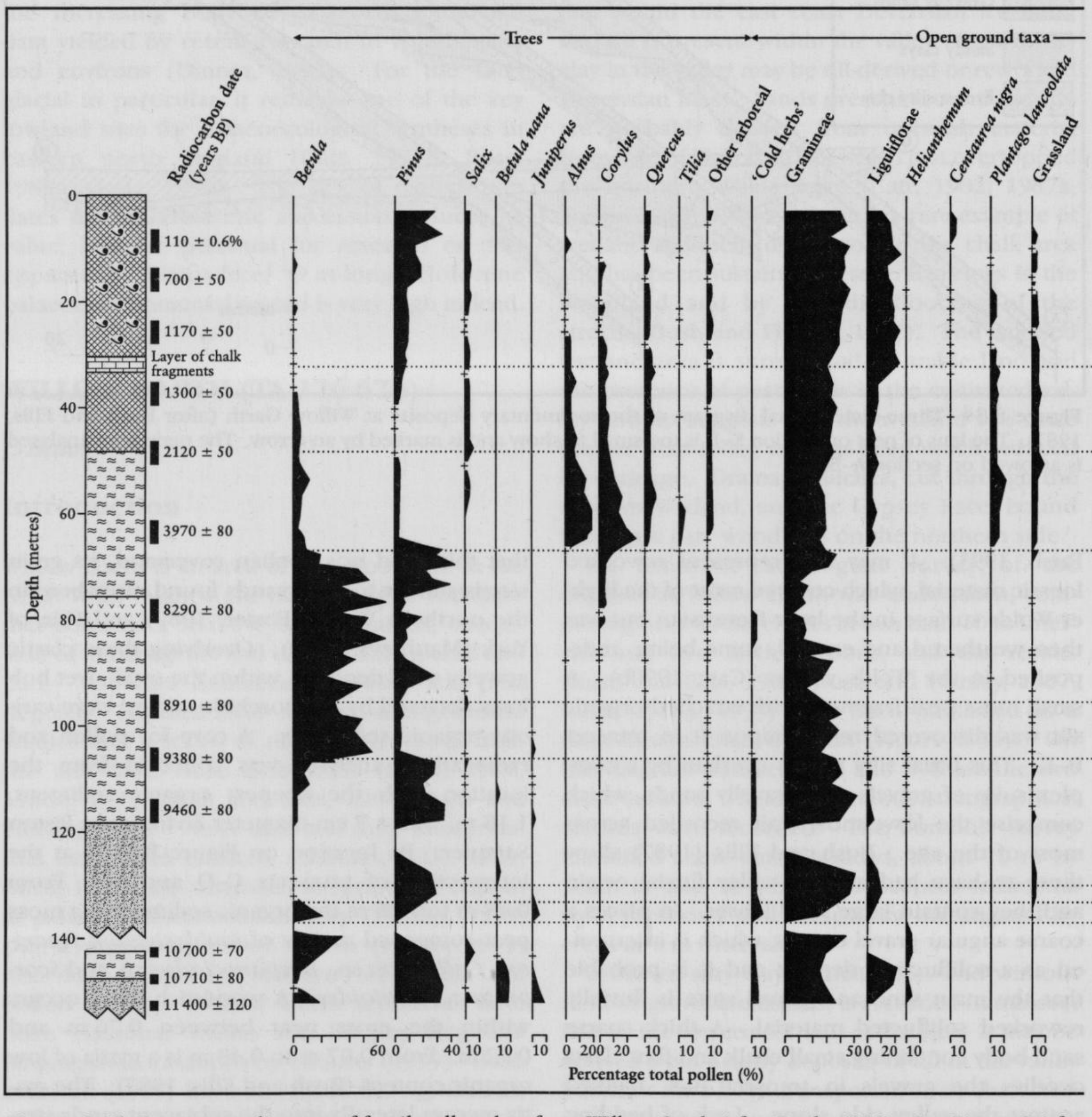
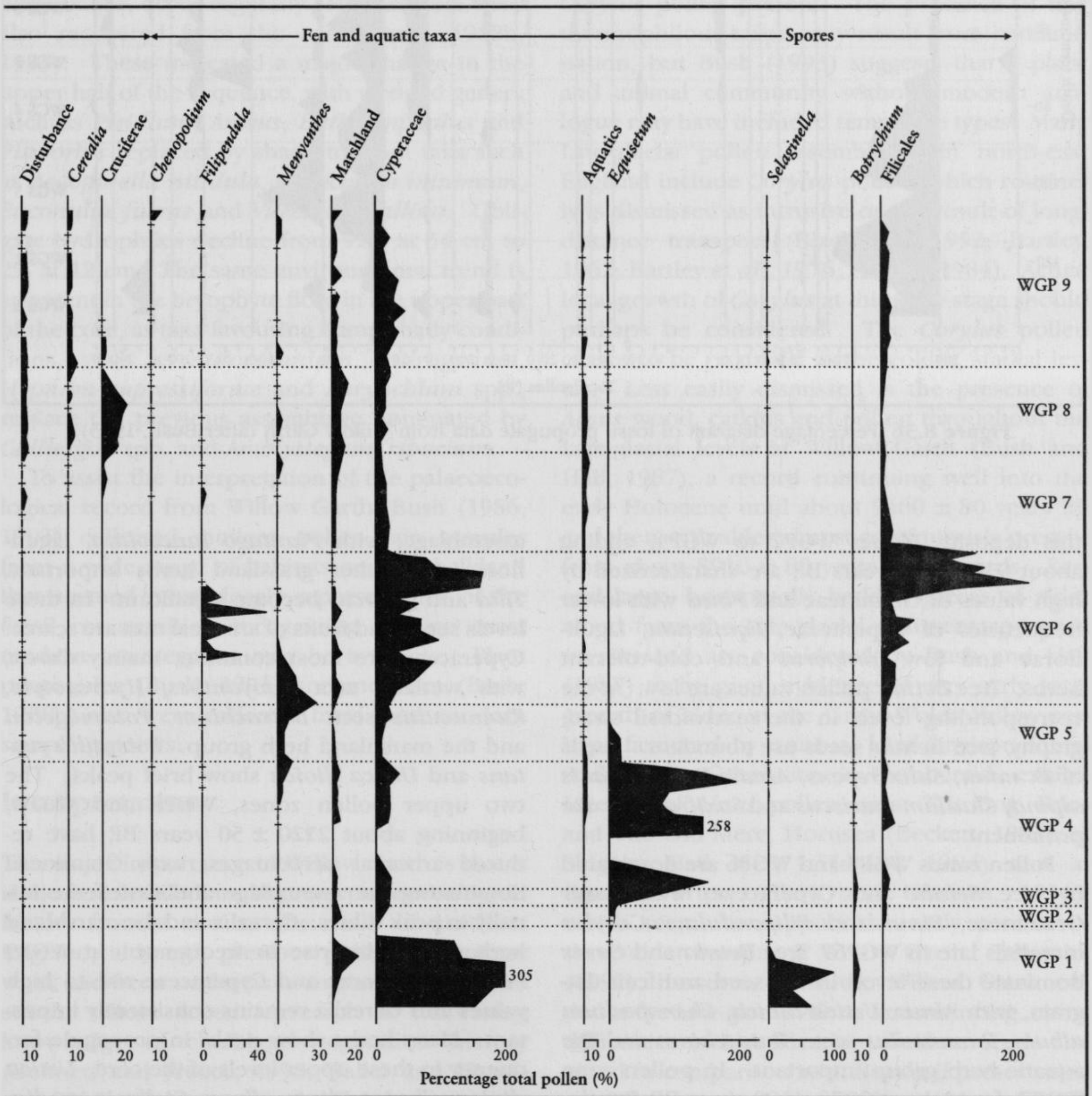


Figure 8.35 Percentage diagram of fossil pollen data from Willow Garth (after Bush, 1993). See Figure 8.1 for key to the stratigraphical log.

from the core and peat lens are shown on Figure 8.35. Frequencies of seeds and fruits (propagules) of selected taxa recovered during macrofossil analyses are shown as Figure 8.36, expressed as percentages of total propagules (Bush, 1993), with rarer taxa assembled into ecological groupings. Local pollen (WGP) or macrofossil (WGS) zones are established and shown on the respective figures, but their zone boundaries rarely fall at the same levels. Full taxa listings for all fossil and modern pollen and macrofossils recorded at Willow Garth are available for reference (Bush, 1986, 1993).

The oldest sample recorded (Figure 8.34, pollen zone WGP1) is the detached moss peat lens, which has radiocarbon parameters of 11 400 ± 120 years BP and 10 700 ± 70 years BP. The most abundant pollen and spore types were Cyperaceae, *B. nana*, *Pinus*, *Juniperus*, *Selaginella*, Liguliflorae and a range of cold-tolerant herbs including *Helianthemum*. *Corylus* is recorded. Macrofossils are not available from



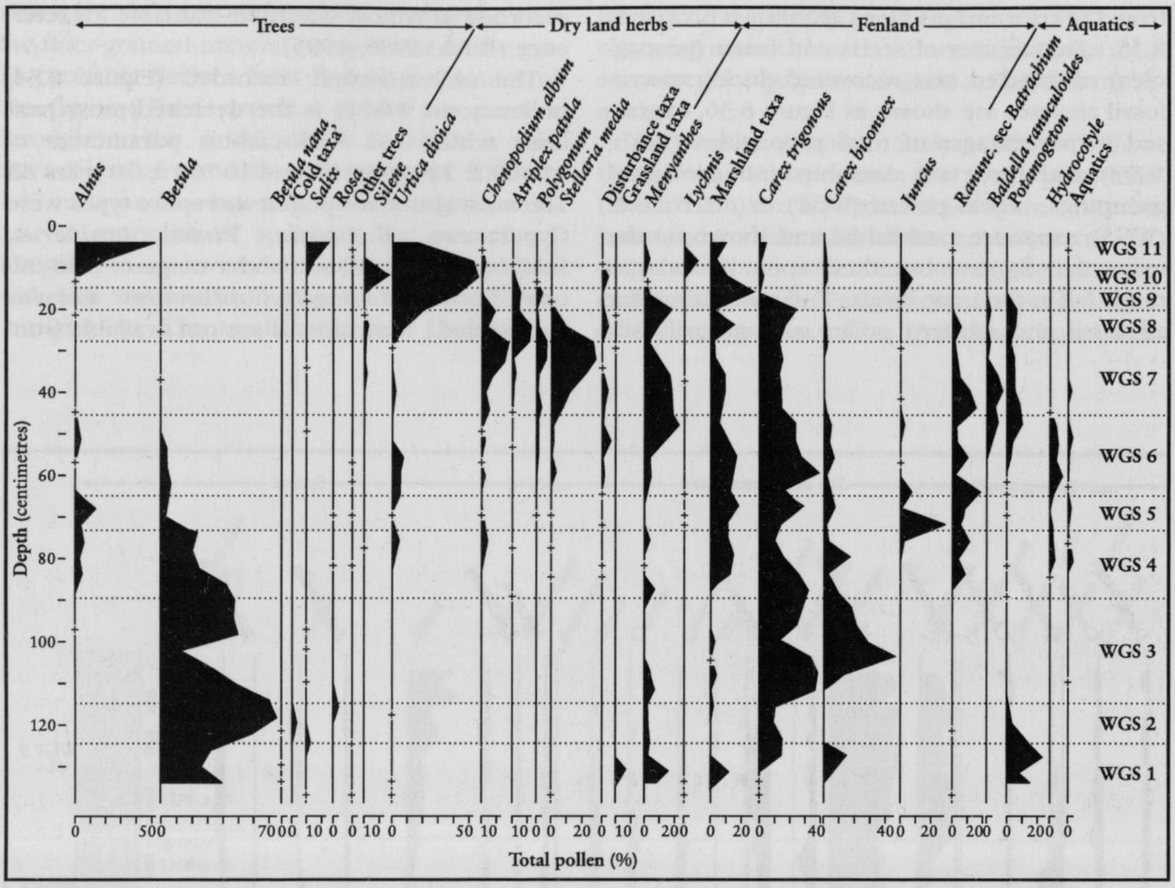


Figure 8.36 Percentage diagram of fossil propugale data from Willow Garth (after Bush, 1993).

this deposit. Zones WGP2 to WGP4, ending about 9380 ± 80 years BP, are characterized by high values of Gramineae and *Pinus* with lower frequencies of Cyperaceae, *Equisetum*, Liguliflorae and low *Juniperus* and cold-tolerant herbs. Tree *Betula* pollen values are low. At the corresponding levels in the macrofossil stratigraphy, tree *Betula* seeds are abundant. Seeds of *B. nana*, *Salix repens*, *Alnus*, *Rosa*, *Lychnis alpina*, *Cladium mariscus* and *Saxifraga* sp. are prominent.

Pollen zones WGP5 and WGP6 are dominated by tree *Betula* and Cyperaceae, with lesser Gramineae, *Pinus* and *Filipendula*. *Corylus* increases late in WGP6. Tree *Betula* and *Carex* dominate these levels in the seed and fruit diagram, with *Alnus*, *Urtica dioica*, *Chenopodium album*, *Ranunculus* sect. *Batrachium* and the aquatic herb group important. In pollen zone WGP7, from about 3970 ± 80 years BP, *Betula*, *Pinus* and Cyperaceae are greatly reduced. *Alnus*, *Corylus* and Gramineae characterize the assemblage, with *Plantago lanceolata*, Liguliflorae and other grassland herbs important. *Tilia* and *Quercus* become significant. In these levels seeds and fruits of arboreal taxa are scarce. Cyperaceae are most common, mainly *Carex*, with wetland taxa *Menyanthes*, *Hydrocotyle*, *Ranunculus* sect. *Batrachium*, *Potamogeton* and the marshland herb group. *Potentilla reptans* and *Urtica dioica* show brief peaks. The two upper pollen zones, WGP8 and WGP9, beginning about 2120 ± 50 years BP, have reduced arboreal percentages, with Gramineae dominating the assemblage and Cruciferae initially in peak values. Cerealia and the marshland herb group also rise in frequency. In WGP9 *Pinus*, Liguliflorae and Cyperaceae rise to high values and Cerealia remains consistently important. Many herbs show peaks in propagule frequency in these upper levels of the core. *Urtica dioica*, *Chenopodium album*, *Stellaria media*, *Atriplex patula*, *Carex* sp. and the marshland herb group all reach peak values. Disturbed ground herbs replace marshland herbs nearer to the surface. In the highest levels peaks of *Alnus*, *Salix* and *Lychnis* occur.

Other palaeoecological data from the Willow Garth core include studies of insect assemblages by Dr H. Kenward (Bush and Ellis, 1987; Bush, 1988a, 1993). Of significance are the 'cold indicator' taxa *Notaris aethiops*, *Otiorhynchus nodosus* and *Olophrum fusca* in the lowest part of the core and the detached moss peat lens. These cold-tolerant taxa give way to *Patrobus* cf. *assimilis*, and above 84 cm *Philopertha horticola*, *Cantharis rustica* and *Serrica brunea*, insects that are indicative of grassland or forest disturbance, occur regularly. Mollusc data were also recovered from the core (Bush, 1988b, 1993). These indicated a major change in the upper half of the sequence, with wetland genera such as *Pisidium*, *Anisus*, *Bathyomphalus* and *Planorbis* replaced by shade tolerant taxa such as *Aegopinella nitidula*, *Carychium minimum*, *Euconulus fulvus* and *Vitrea crystallina*. Obligate hydrophiles decline from 91% at 34 cm to 2% at 12 cm. The same environmental trend is apparent in the bryophyte flora in the upper part of the core, as taxa favouring damp shady conditions, such as *Aulocomnium androgynum*, *Hypnum cupressiforme* and *Eurynchium* spp., replace the previous assemblage dominated by *Calliergon* spp. and *Amblystegium riparium*.

To assist the interpretation of the palaeoecological record from Willow Garth, Bush (1986, 1993) collected modern pollen rain samples from a wide range of habitats on the chalkland that seemed most closely representative of the fossil communities, recognizing that no exact modern counterparts may be available. Plant propagules (Bush, 1993) and mollusca (Bush, 1988b) also were collected from these modern sampling points.

Interpretation

The major importance of Willow Garth is that it is the first detailed, multi-proxy palaeoecological history from a site on the chalk of northern England, although previous pollen or mollusc work had been undertaken in chalk areas of the south, where recent research is increasing knowledge of chalkland vegetation history (Bennett and Preece, 1998; Waller and Hamilton, 2000). The insects and pollen from the lowest sediments at Willow Garth record a dry, open, cold-climate grassland flora with substantial bare ground at a date from 11 400 years BP, covering the latter half of the Late-glacial interstadial and much of the Late-glacial (Loch Lomond) stadial. Both the detached peat fragment and the lower part of the core contain this rich arctic–alpine tundra community. Cold-element components absent from modern chalk grasslands grew around Willow Garth, such as *Armeria maritima*, *Saxifraga oppositifolia* and *Botrychium lunaria*. *Gentiana verna* and *G. pneumonanthe* are unusual and microfossils of *Diphasium alpinum* are the first records for the British Quaternary. There are temperate elements in the earlier part of this phase, with *Corylus* pollen present. The presence of this thermophilous taxon may result from contamination, but Bush (1993) suggests that a plant and animal community without modern analogue may have included temperate types. Many Late-glacial pollen assemblages in north-east England include *Corylus* pollen, which routinely is dismissed as intrusive or the result of long-distance transport (Blackburn, 1952; Bartley, 1962; Bartley *et al.*, 1976; Flenley, 1984). Actual local growth of *Corylus* at this early stage should perhaps be considered. The *Corylus* pollen ceases to be recorded in the colder, stadial levels. Less easily dismissed is the presence of *Alnus* wood, catkins and pollen throughout the Late-glacial levels at Willow Garth (Bush and Hall, 1987), a record continuing well into the early Holocene until about 9460 ± 80 years BP and then with alder almost continuously present from about 8910 ± 80 years BP onwards. Several large, horizontally bedded pieces of alder wood from the Late-glacial sediment appeared *in situ* and are considered by Bush and Hall (1987) to be good evidence of very early local growth of alder in the Great Wolds Valley, perhaps favoured by unusual local environmental conditions. Other sites in Humberside exhibit Late-glacial *Alnus* pollen, as at The Bog, Roos and The Old Mere, Hornsea (Beckett, 1981), at Skipsea Bail Mere (Flenley, 1984) and at Brandesburton (Clark and Godwin, 1956), which seems to support the early context of these alder remains at Willow Garth. Tallantire (1992), however, finds the Willow Garth evidence for Late-glacial *Alnus* unconvincing, citing difficulties in identification of *Alnus* seeds and wood in comparison with *Betula* if preservation conditions are less than perfect. Although Bush and Hall (1987) record the alder wood as horizontally bedded stem wood, Tallantire (1992) points out that alder stem wood is not easy to tell from root wood and the latter may be intrusive, penetrating through sediments from above during times of lowered water tables. Further research is required to resolve the significance of the early *Alnus* records at Willow Garth, including radiocarbon dating of the wood itself.

As with other sites in the region in the early Holocene, at Willow Garth there is a start of the spread of woodland trees and shrubs, but this process is delayed and it is not until after about 9380 ± 90 years BP that tree *Betula* pollen increases substantially and suggests the establishment of woodland around the site. The tree expansion was not maintained, however, and high Gramineae and the persistence of taxa records such as *Helianthemum*, *Campanula*, *Lotus*, *Vicia*, *Teucrium* and *Geranium* show that dry grassland was not displaced by woodland. Temperate forest trees such as *Quercus*, *Corylus* and *Ulmus* never expanded in the early to mid-Holocene as they did at pollen sites elsewhere. *Betula* remained the dominant woodland component at Willow Garth, with *Pinus* pollen possibly contributed mainly via long-distance transport. Whether this apparent persistence of dry grassland was a natural process is conjectural. Bush (1989) suggests that the records of plant and insect taxa favoured by disturbance indicate that from *c.* 8900 years BP onwards Mesolithic hunter–gatherer activity was preventing forest closure and enabling the survival of species-rich grassland. As charcoal records are low, this would have been achieved without the major use of fire. There are many Mesolithic cultural sites within the Great Wold Valley (Bush, 1989). Whether the maintenance of dry grassland was confined to the vicinity of Willow Garth or was true of the wider Wolds is unknown.

The value of Willow Garth for determining the status of the chalk grasslands is much diminished by the presence of sedimentary discontinuities, caused probably by standstill levels in sediment accumulation rather than erosion. Younger and McHugh (1995) have shown that peat accumulation in this area may be linked closely to variations in spring flows and thus water table fluctuations. This may have been the mechanism governing the history of peat formation at Willow Garth, where no evidence is available between *c.* 8000 and 4300 years BP, the entire mid-Holocene period. Grassland was certainly present after *c.* 4300 years BP, with high *Plantago lanceolata* suggesting major pastoral land use, but relatively substantial *Alnus*, *Corylus*, *Quercus* and *Tilia* could indicate that a deciduous forest phase may have occurred during the depositional hiatus. Palynology of buried soils beneath nearby Kilham Long Barrow (Evans and Dimbleby 1976) showed that before *c.* 4900 years BP on the chalk upland there was open grassland and evidence of arable cultivation. Persistence of chalk grassland throughout the mid-Holocene forest phase seems likely.

A second discontinuity between *c.* 3300 and 2200 years BP further reduces the value of the environmental record, but the later Holocene at Willow Garth was a dominantly open grassland landscape. Intensive grazing characterizes land use before the hiatus, but greatly increased arable cultivation occurred after it, with cereals being important. There are almost no other later Holocene pollen records from the Wolds chalk with which to compare Willow Garth, but a similar grassland and cultivated landscape was recorded in pollen data from sediments associated with a later prehistoric field boundary at Foxholes in the northern Wolds (Innes, 1989). The silty gyttja of the upper deposit at Willow Garth probably incorporates much mineral soil eroded after local ploughing, a common feature in the history of the Wolds landscape (Ellis and Newsome, 1991). An important change at Willow Garth occurred after *c.* 1300 years BP when the wetland became much drier, with aquatic pollen being replaced by that of disturbed dry-ground weeds and *Salix* pollen increasing greatly. Bush (1993) suggests that the site became managed as an osier bed, which has saved it from later major drainage and preserved the important environmental record.

Conclusions

The Willow Garth sediments are an important but incomplete archive for the environmental history of the chalk Wolds. If the major discontinuities in the sampled core result from localized erosion, oxidation or failure to form peat because of irregular flooding by the stream or variation in spring flow, it may be that the gaps in the environmental record are specific to that core only. Sediment covering the missing periods may exist at other locations, and other cores could provide a more complete data record. The site is complex and controversial and requires further research.

STAR CARR (TA 028810)

POTENTIAL GCR SITE

S. Gonzalez and D. Huddart

Introduction

Star Carr, North Yorkshire is the most important, exceptionally rich and well-documented, British Mesolithic site. It has been discussed, reviewed and re-interpreted more than any other such site and arguably is the most re-interpreted site in European prehistory (Mellars and Dark, 1998). Discussions of the British Mesolithic usually have relied on the information from the original site monograph (Clark, 1954) and this work influenced excavation techniques and site interpretation throughout the world (Legge and Rowley-Conwy, 1988). Although the investigations by Clark (1954) were a model for their time they left many important questions unanswered . More recent work, summarized by Mellars and Dark (1998), has shown that the site is almost 1000 years older than previously believed. It has demonstrated a complex and repeated pattern of occupation spanning at least three centuries and has documented evidence for the repeated and apparently deliberate burning of the reedswamp vegetation to improve access to the lake and perhaps attract animal populations. There also is evidence for a substantial wooden trackway leading to the lake edge, which is the oldest evidence for systematic carpentry so far documented in Europe. It has been used as a kind of field laboratory for testing different interpretative models of hunter–gatherer sites (Mellars and Dark, 1998).

The site lies at the eastern end of the Vale of Pickering where the 'carr' lands of the vale are formed from calcareous and organic mud and peat accumulated under conditions of a high water level, which overlie till and solifluction sediments from the Late Devensian (Walker and Godwin, 1954). The valley floor is about 3.5 km

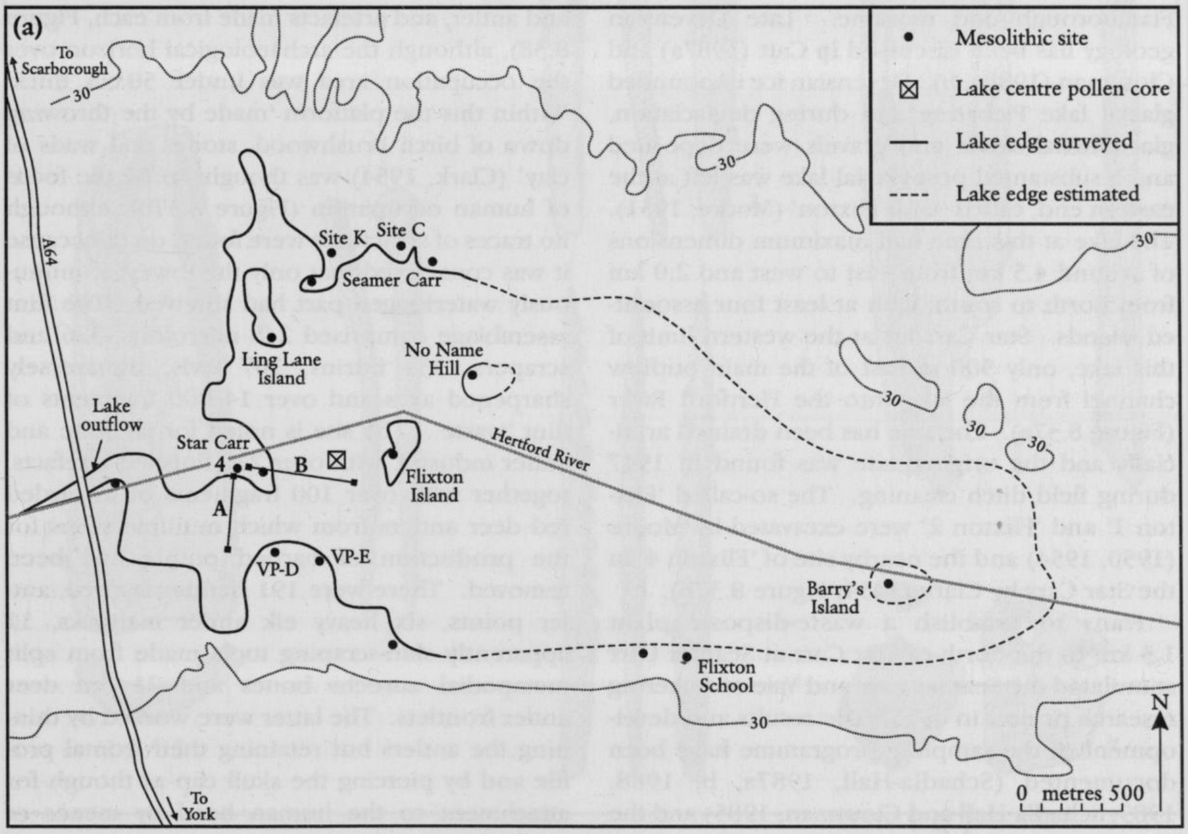


Figure 8.37a Reconstructed plan of the Pre-Boreal Lake Flixton. Note the location of the main early Mesolithic sites, the transects A and B (see Figure 8.40b) from Walker and Godwin (1954) and the pollen core (see Figure 8.40a) from Dark (1996, 1998a, b). (After Clark, 1954; Mellars and Dark, 1998).

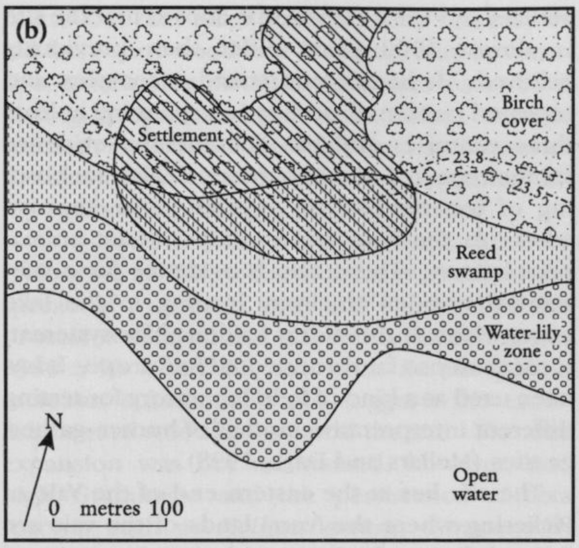


Figure 8.37b Clark's (1954) location of the Star Carr 4 ('Flixton 4') settlement, between the reed swamp and the birch woodland.

wide at Star Carr, with low hills to the north and south. The vale is blocked to the east by the Flamborough end moraine. Late Devensian geology has been discussed in Catt (1987a) and Cloutman (1988a, b). Devensian ice impounded glacial lake Pickering and during deglaciation, glaciofluvial sands and gravels were deposited and a substantial post-glacial lake was left at the eastern end, called 'Lake Flixton' (Moore, 1951). The lake at this time had maximum dimensions of around 4.5 km from east to west and 2.0 km from north to south, with at least four associated islands. Star Carr lay at the western limit of this lake, only 500 m east of the main outflow channel from the lake into the Hertford River (Figure 8.37a). The vale has been drained artificially and the original site was found in 1947 during field ditch cleaning. The so-called 'Flixton 1' and 'Flixton 2' were excavated by Moore (1950, 1954) and the nearby site of 'Flixton 4' in the Star Carr by Clark (1954; Figure 8.37b).

Plans to establish a waste-disposal plant 1.3 km to the north of Star Carr at Seamer Carr stimulated the Seamer Carr and Vale of Pickering research project in 1975. The results and development of the sampling programme have been documented (Schadla-Hall, 1987a, b, 1988, 1989; Schadla-Hall and Cloutman, 1985) and the overall results put into context by Mellars and Dark (1998) and Schadla-Hall and Lane (in press). In 1985 the Vale of Pickering Research Trust was set up to survey the site and systematically test-pit sample the whole of the area around the western and southern shorelines of the former lake and to extend the survey towards the east. This has located ten major new concentrations of early Mesolithic material and the recovery of significant faunal remains from four sites. There also has been a new programme of detailed stratigraphical and palaeoecological research in the lake deposits to investigate the area between Star Carr and Seamer Carr and to re-investigate the limited information published by Walker and Godwin (1954) for Star Carr. This included the excavation of an 18 m long trench, about 20 m to the east of the original excavations and extending from the zone of 'dryland' deposits on the former lake shore into the deeper organic deposits of the lake-edge zone (see Cloutman, 1988a, b; Cloutman and Smith, 1988).

Description

At Star Carr a wide and rich variety of organic materials were found (including wood, bone and antler, and artefacts made from each, Figure 8.38), although the archaeological horizon over the occupation area was under 50 cm thick. Within this the platform 'made by the throwing down of birch brushwood, stones and wads of clay' (Clark, 1954) was thought to be the focus of human occupation (Figure 8.37b), although no traces of structures were found on it, because it was considered that only the lowest, continuously waterlogged part had survived. The flint assemblage comprised 248 microliths, 326 end scrapers, 334 burins, 107 awls, transversely sharpened axes and over 14 000 fragments of flint 'waste'. The site is noted for its bone and antler industry, with over 220 finished artefacts, together with over 100 fragments of discarded red deer antlers from which multiple strips for the production of barbed points had been removed. There were 191 slender, barbed, antler points, six heavy elk antler mattocks, 11 apparently skin-scraping tools made from split metapodial aurochs bones and 21 red deer antler frontlets. The latter were worked by thinning the antlers but retaining their frontal profile and by piercing the skull cap as though for attachment to the human head by means of cords.

Two trunks of birch may have been part of a short trackway leading from the occupation

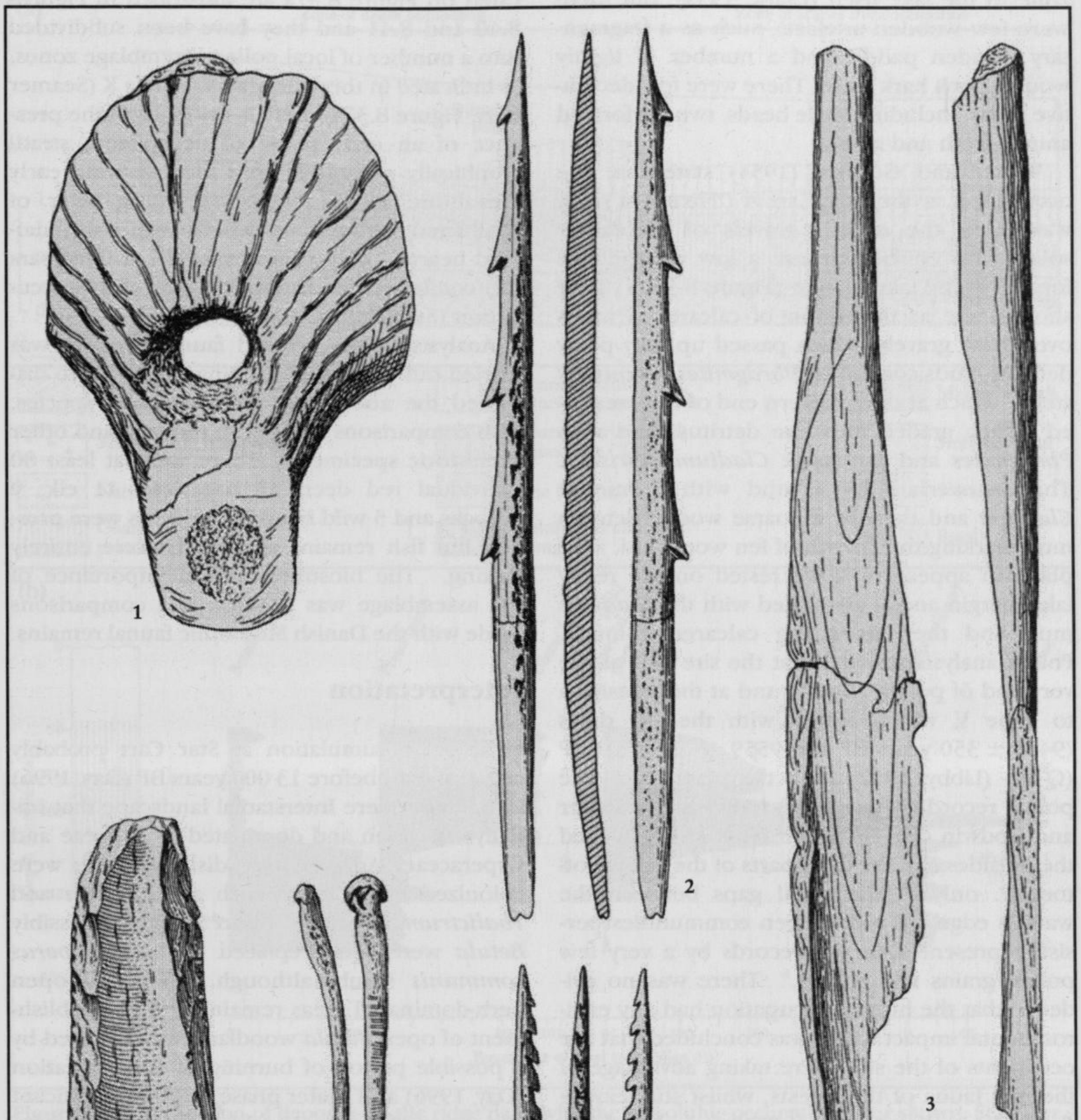


Figure 8.38 Artefacts from the original Star Carr excavations: 1. elk-antler mattock head; 2. barbed antler point (type A); 3. birch-wood 'paddle'; 4. aurochs metapodial scraper; 5. elk metapodial awl; 6. barbed antler point (type E); 7. amber pendant. All half size, except 7, which is actual size (after Clark, 1954).

zone to the lake itself (Clark, 1954), but there were few wooden artefacts, such as a fragmentary wooden paddle and a number of tightly wound birch bark rolls. There were few decorative items, including shale beads, two perforated animal-teeth and amber.

Walker and Godwin (1954) state that the occupation at the Star Carr 4 ('Flixton 4') site was upon the muddy gravels of an earlier solifluction episode, where a low mound was formed at the lake margin (Figure 8.39a). They showed the accumulation of calcareous muds over these gravels, which passed up into peaty detritus muds containing *Phragmites communis* stems, which at the southern end of the excavated trench graded into fine detritus mud with *Phragmites* and the sedge *Cladium mariscus*. This is overlain by a mud with dominant *Cladium* and then by a coarse woody detritus mud marking the growth of fen woodland. The platform appears to have rested on the reedy lake margin and is associated with the *Cladium* mud and the underlying calcareous muds. Pollen analysis showed that the site falls at the very end of pollen zone IV and at the transition to zone V, which agrees with the ^{14}C dates (9488 ± 350 years BP and 9559 ± 210 years BP (Q-14) (Libby, 1952) from the platform). The pollen record (Figure 8.39) indicated to Walker and Godwin (1954) 'closed birch forest clothed the hillsides and the drier parts of the valley bottom ... only in the small gaps between the water's edge did a few open communities persist, represented in our records by a very few pollen grains and spores.' There was no evidence that the human occupation had any environmental impact and it was concluded that the occupants of the site 'were taking advantage of the rich fauna of the forests, whilst still leaving the forest itself untouched' (Walker and Godwin, 1954). Day (1996) and Mellars and Dark (1998) confirm that the earliest lake deposition stages go back to an early stage in the Late-glacial, probably around 13 000–14 000 years BP and lake levels were probably 25 m OD, allowing lake clays to accumulate to at least this height in the channel between the two main excavated areas at Star Carr. Lake-level fluctuations occurred during the Late-glacial and early post-glacial.

Recent palaeoecological research has been detailed in Day (1993, 1996) and Mellars and Dark (1998). The Late-glacial and Flandrian pollen diagrams from the lake-centre core indicated on Figure 8.37a are illustrated in Figures 8.40 and 8.41 and they have been subdivided into a number of local pollen assemblage zones, as indicated in these diagrams. At site K (Seamer Carr, Figure 8.37a) there is evidence of the presence of an early phase of occupation, stratigraphically separated from the overlying early Mesolithic level by an intervening layer of solifluction sand and associated with a well-defined hearth. Dating of this level (11 000 years BP) confirmed the Late-glacial age of this occupation (Schadla-Hall, 1989).

Analysis of the original faunal remains was carried out by Fraser and King (1954), who discussed the animal groups species by species, with comparisons made with modern and other prehistoric specimens. There were at least 80 individual red deer, 33 roe deer, 11 elk, 9 aurochs and 5 wild boar. Some birds were present, but fish remains apparently were entirely lacking. The biostratigraphical importance of the assemblage was stressed and comparisons made with the Danish Mesolithic faunal remains.

Interpretation

Sediment accumulation at Star Carr probably began shortly before 13 000 years BP (Day, 1996) in a Windermere Interstadial landscape that initially was open and dominated by Poaceae and Cyperaceae. Areas of bare, disturbed soils were colonized by pioneers such as *Artemisia* and *Thalictrum*. Scattered dwarf *Salix* and possibly *Betula* were soon replaced by low *Juniperus communis* scrub, although substantial open herb-dominated areas remained. The establishment of open *Betula* woodland was followed by a possible period of burning of the vegetation (Day, 1996) and a later phase, with no significant quantities of charcoal and increases in Poaceae, *Rumex acetosella*, *Artemisia* and mineral inputs, suggests temporary cooling. This also can be found in both the pollen and the coleopteran record at Gransmoor (Walker *et al.*, 1993). This was followed by a recovery in climate, as reflected in the *Betula* curve and a minor increase in *Juniperus communis*. The cessation of marl formation at 507 cm is taken to represent a major climatic deterioration, along with the increase in soil disturbance indicators and a decrease in *Filipendula*. At 456 cm the change to clay with pebbles, followed by increases of *Artemisia* and *Thalictrum* suggest frost disturbance of soils and gelifluction.

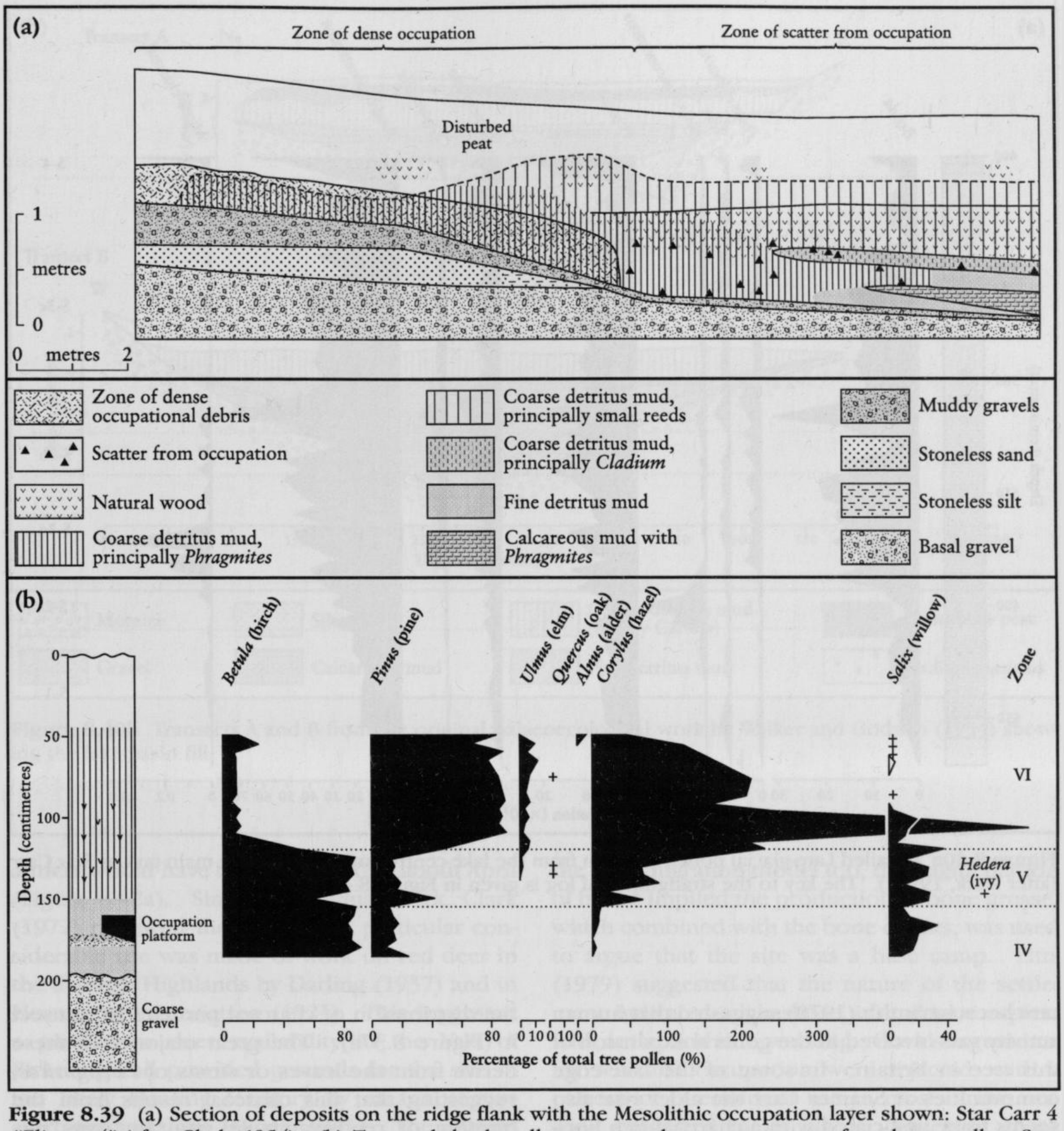


Figure 8.39 (a) Section of deposits on the ridge flank with the Mesolithic occupation layer shown: Star Carr 4 ('Flixton 4')(after Clark, 1954). (b) Tree and shrub pollen expressed as a percentage of total tree pollen, Star Carr (after Walker and Godwin, 1954).

The Flandrian onset is marked by a sharp drop in minerogenic inputs and successive peaks of Poaceae, Rubiaceae, *Plantago media*, *Filipendula* and *Salix* as different plant communities formed in response to the rapid temperature rise. Aquatic plants showed a marked response, beginning with a major phase of vegetative growth of *Chara*, which may have preceded the response of terrestrial vegetation to climatic warming. Birch woodland then spread rapidly, leading to a decrease of herbs. The first arrival of hazel has been dated to around 9400 years BP (Day and Mellars, 1994), which formed a dense woodland about 9000 years BP and was soon joined by elm and then oak, lime and ash by about 7600 years BP. The increase of *Alnus glutinosa* about this time followed the switch from marl accumulation to detritus mud as the lake shallowed. The coincidence of evidence of local burning and the *Alnus* increase is impor-

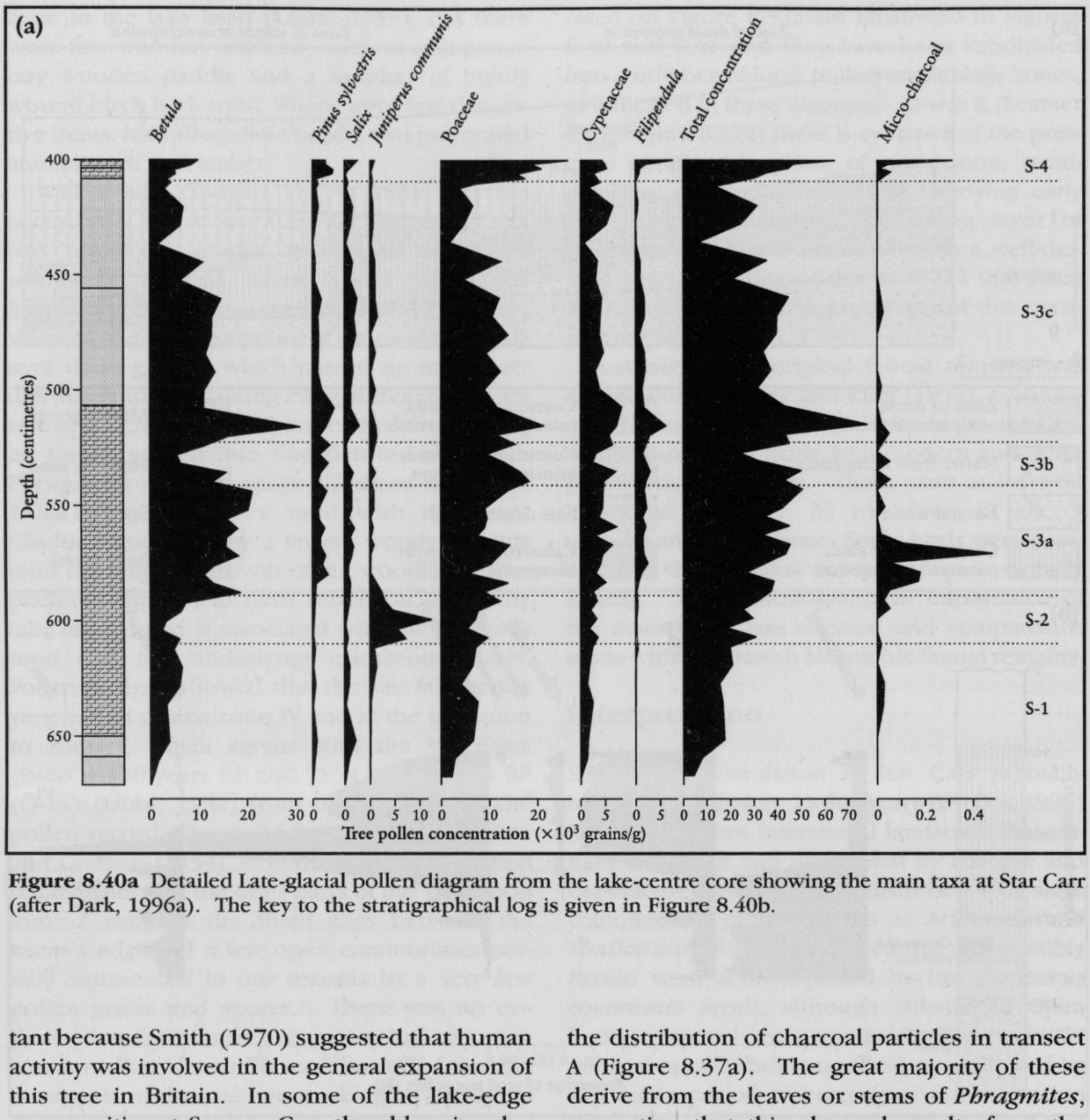


Figure 8.40a Detailed Late-glacial pollen diagram from the lake-centre core showing the main taxa at Star Carr (after Dark, 1996a). The key to the stratigraphical log is given in Figure 8.40b.

tant because Smith (1970) suggested that human activity was involved in the general expansion of this tree in Britain. In some of the lake-edge communities at Seamer Carr the alder rise also seems to be associated with a charcoal-rich horizon (Cloutman, 1988b). Day (1996) suggests that despite the evidence for human activity around the lake in the early Flandrian, including burning of the reed-swamp communities, the overall course of woodland development appears not to have been significantly affected. This contrasts with the Late-glacial burning, which appears to have resulted in the decline of birch and therefore it appears that the Late Palaeolithic peoples may have had a more significant effect on the local environment at Star Carr than the early Mesolithic population.

There is strong evidence for human interference in the local vegetation cover provided by the distribution of charcoal particles in transect A (Figure 8.37a). The great majority of these derive from the leaves or stems of *Phragmites*, suggesting that this charcoal results from the burning of reedswamp *in situ*, but was this deliberate burning? Mellars and Dark (1998) suggest that it was carried out intentionally because of the apparent frequency of burning, the apparently localized nature of the burning and the commonsense rationale that any human group occupying a lakeside location would wish to maintain a clear view and easy access over open water.

The argument for seasonal occupation associated with the deer antlers suggests that as the red deer antlers were attached to pedicels then occupation at some period of the year between October and April was indicated because only then are the deer carrying antlers, and shed

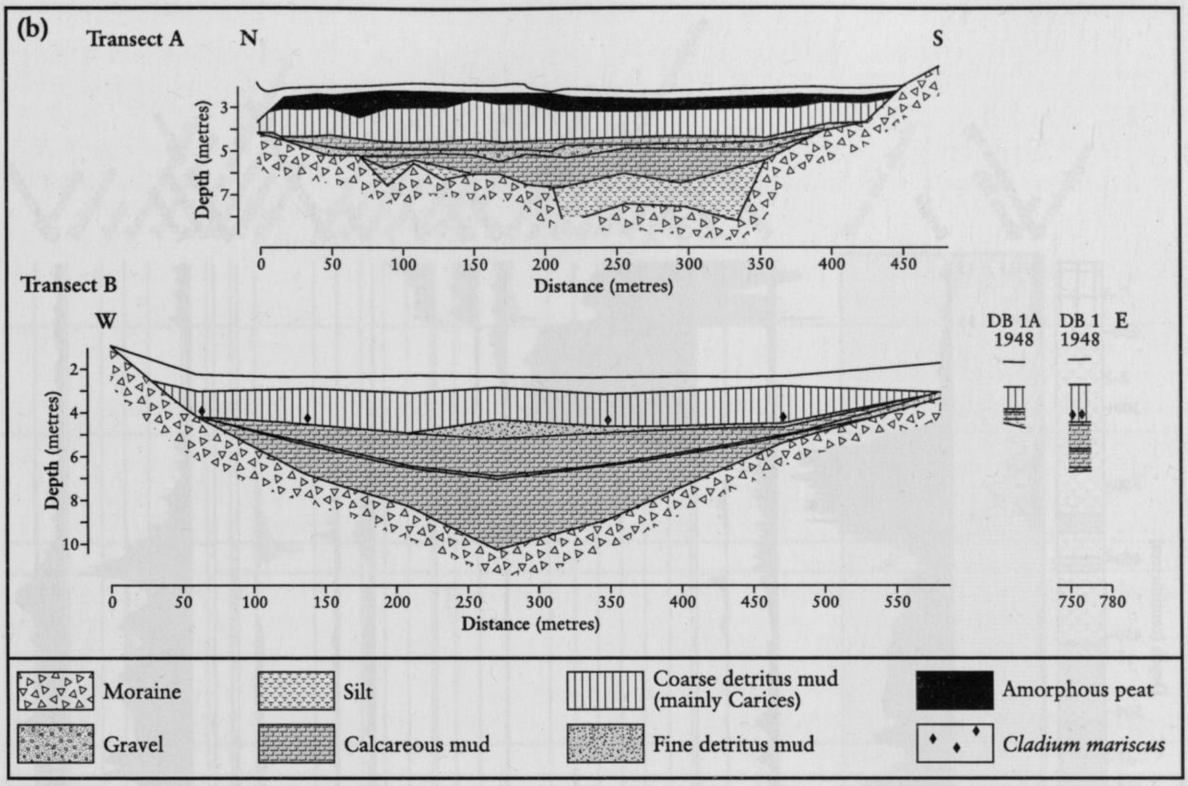


Figure 8.40b Transects A and B from the original palaeoecological work by Walker and Godwin (1954) showing the lake basin fill.

antlers would have been collected in about April (Figure 8.42a). Since the original work, Clark (1972) reviewed the site and in particular considerable use was made of work on red deer in the Scottish Highlands by Darling (1937) and in Norway by Ingebrigtsen (1924). This suggested a parallel to deer availability in the Vale of Pickering and provided grounds for the size of the likely hunting territory by means of size catchment analysis (Higgs, 1975). The pattern of seasonal migration observed in the modern deer also re-inforced Clark's (1954) model of both deer and human migration from the proposed winter base at Star Carr to upland summer territories in the hills to the north and south and even to the Pennines. However, Caulfield (1978) suggested that red deer were less important in the diet of the Star Carr occupants than Clark (1954, 1972) had suggested. Red deer antlers were thought to be imported to the site as raw material and doubt was cast on the red deer migration model. He thought the site was a butchering station and possibly a kill site. Jacobi (1978) expressed some reservation about the deer migration model too, the fragmentation of bones implied the production of bone grease, which combined with the bone counts, was used to argue that the site was a base camp. Pitts (1979) suggested that the nature of the settlement indicated a lakeshore, industrial complex, with antler and skin working. Clark's arguments for seasonal occupation were rejected, with a pattern of intermittent use in most of the seasons suggested in its place. Andersen *et al.* (1981) again rejected the single occupation season and argued that the body part representation suggested a butchering site, with the animals being killed close by. The particular location of the site, on a gravel spit into the lake, suggested that game driving may have been important. Grigson (1981) questioned the usefulness of the red deer antler as a seasonal indicator and the roe deer were used to suggest summer occupation, as was the presence of summer migrant birds. Price (1982) suggested that much of the material recovered had been thrown or dumped into the lake, and rejected the original case for winter occupation. Instead the wide range of

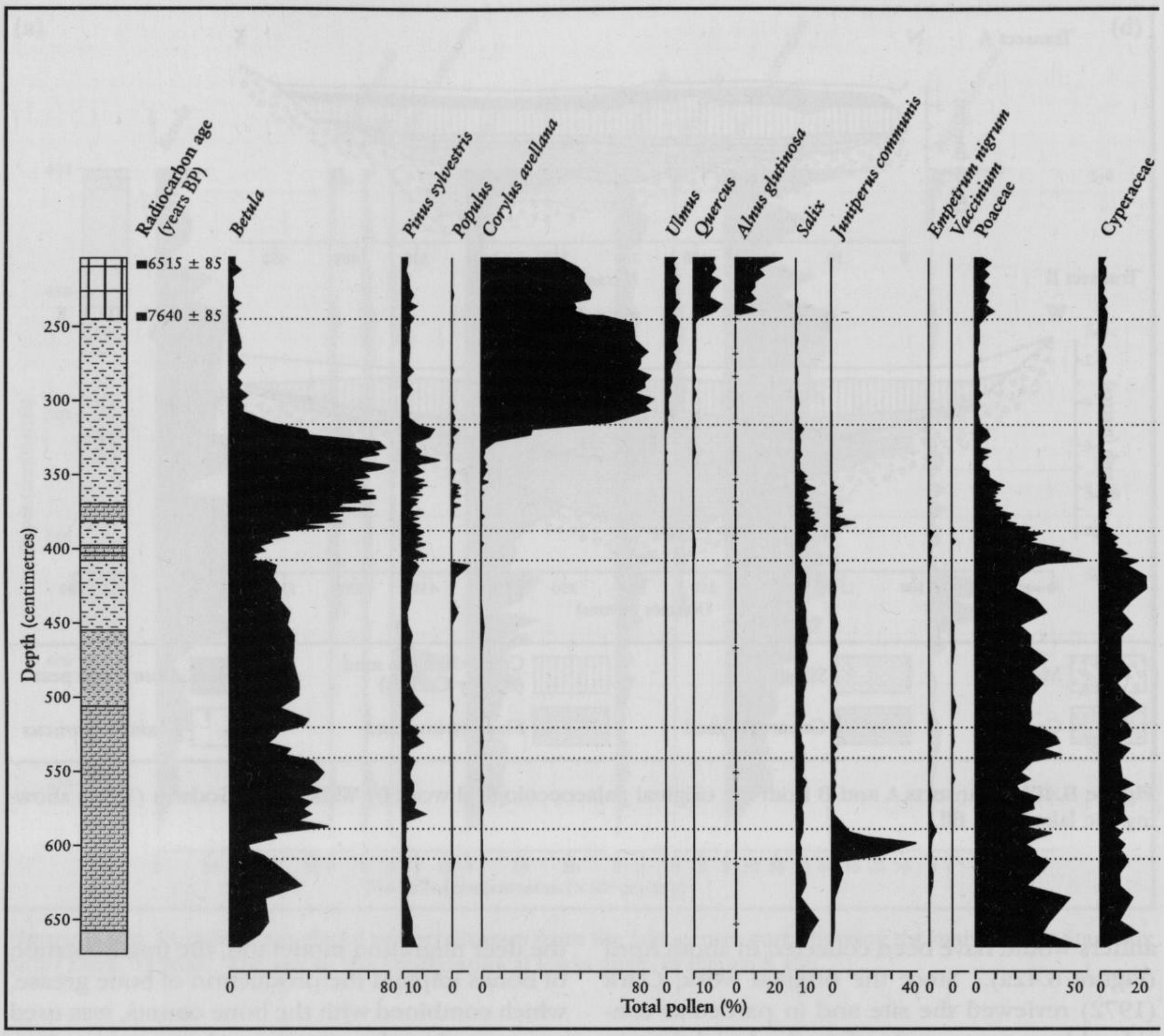


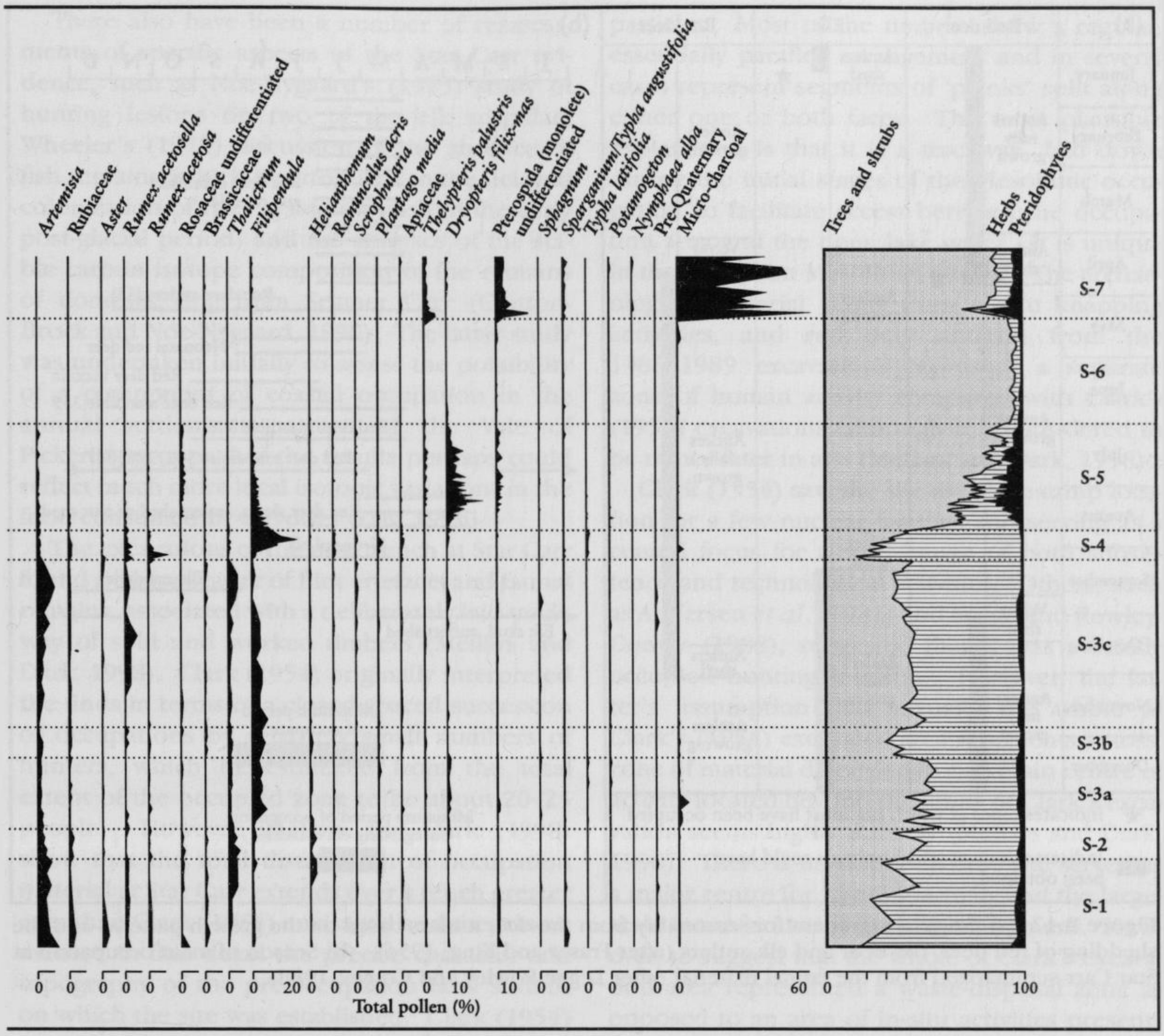
Figure 8.41 Percentage pollen diagram from Star Carr lake-centre core showing all taxa occurring at over 1% of the pollen sum. For *Alnus glutinosa* early records are represented by closed circles (after Dark, 1996a). See Figure 8.39 for key to the stratigraphical log.

mammalian species and artefacts was used to support the base-camp idea. The whole issue of the precise pattern of occupation and seasonality at Star Carr remains controversial.

Legge and Rowley-Conwy (1988) re-analysed the main food animal bones from Star Carr and their main conclusions were:

1. strong support for the use of the site only in late spring and summer;
2. evidence that the cull of red and roe deer was biased towards three- and one-year old animals respectively, for reasons connected with the behaviour of those species;
3. evidence that there is no bias towards the hunting of male red deer, so that the antlers cannot be included in any discussion of seasonality or sex ratios;
4. a revision of the meat available and a downwards reduction of the occupation scale;
5. a more tentative suggestion that the site was a hunting camp from where meat was removed to a base camp elsewhere.

It is clear that the inhabitants obtained their deer by hunting (Legge and Rowley-Conwy, 1988) and it is probable that the animals were hunted singly or in small groups, rather than large-scale planned drives, with no evidence for the herding economy suggested by Jarman



(1972). The seasonal indications from tooth eruption, neonatal specimens and red deer skulls with recently shed antlers combine to indicate early summer occupation. No specimens point to winter visits and only a single young elk mandible may indicate a visit later in the summer or autumn. Following their arguments that the bone representation is not that expected of a base camp, Legge and Rowley-Conwy (1988) interpret Star Carr as a hunting camp to which many short visits were made, mainly in the early summer. The main camps might have been where the flint scatters are on the North Yorkshire Moors, which formed ideal summer sites as long as Star Carr was thought to be a winter base camp. However, the coastal zone, less than 15 km from Star Carr, seems likely to have been important, where alternative resources would have been available, probably at seasons complementary to that in which Star Carr is known to have been occupied. The wide range of manufacturing indicated by the artefacts suggested to Mellars (1976) that the site was a base camp and Dumont (1987) confirmed that a wide range of manufacturing activities was carried out on the site. However, one study of a briefly occupied, Eskimo hunting stand by Binford (1978a, b) has shown that a wide range of activities, particularly connected with hunting equipment, may take place. The red deer

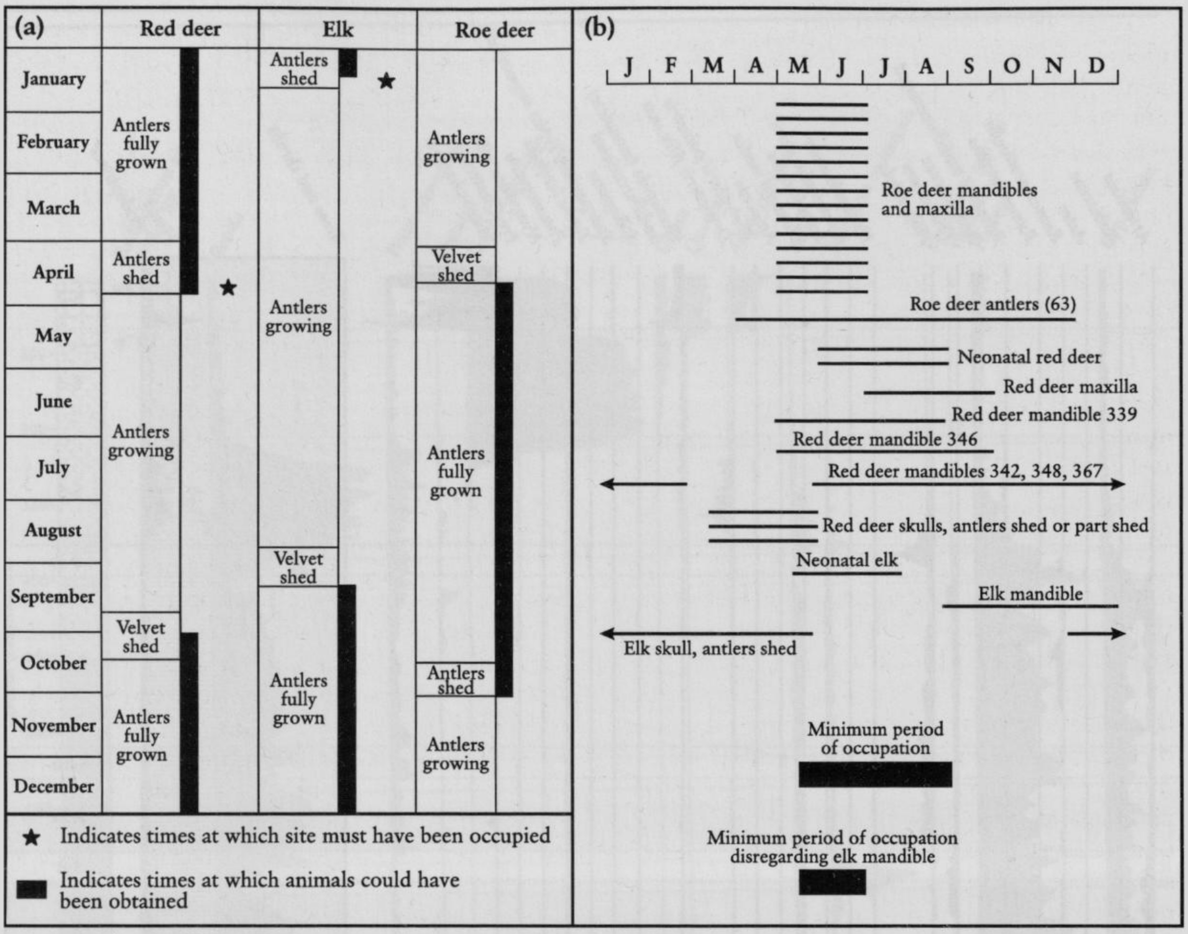


Figure 8.42 (a) Original argument for seasonality from the deer antlers based on the growth patterns and the shedding of red deer, roe deer and elk antlers (after Fraser and King, 1954). (b) Season of main occupation at Star Carr summarized from the faunal material (after Legge-Rowley and Conwy, 1988).

frontlets may have had a practical or a ritual function, and the two explanations usually are presented as equally likely (Clark, 1954). Legge and Rowley-Conwy (1988) suggest that the ritual function seems more likely as they argued that the site was occupied mainly in early summer, a time of year when the red deer stags had just shed and were beginning to regrow their antlers. This is the one time of year when a hunting disguise would not be likely to include antlers as the deer themselves would have partly grown antlers at most, and hence the ritual function seems more feasible.

Legge and Rowley-Conwy's (1988) evidence is summarized in Figure 8.42b. Carter's (1997) study of the lower jaws of roe deer, based on X-ray analysis of the jaws to reveal the patterns of unerupted as well as erupted teeth, and comparative studies of modern samples suggest that the majority of Star Carr jaws derive from animals between 10 and 11 months in age. Therefore they probably were killed in the period from March to April, rather than May–June as estimated in Legge and Rowley-Conwy (1988). Three additional pieces of evidence point to activity in the early spring months. The analysis of the burnt reeds (Hather, 1998) has revealed several pieces from tightly rolled leaf stems, which must derive from plants burnt during the early growing phase around March–April. The identification of clearly charred bud scales of aspen catkins (Dark, 1998a) indicates that they are most likely to have been burnt during, or shortly after, the period of shedding between April and June. Some of the cakes of birch-bark resin recovered from the original Star Carr excavations appear to show high sugar levels, which would point to bark collection during the late April to early May period, when the sugar content in the bark peaks (Mellars and Dark, 1998).

There also have been a number of reassessments of specific aspects of the Star Carr evidence, such as Noe-Nygaard's (1975) study of hunting lesions on two of the elk scapulae, Wheeler's (1978) discussion of the absence of fish remains (which he attributed to the delayed colonization of river systems by fish in the early post-glacial period) and the analyses of the stable carbon isotope composition of the remains of domestic dog from Seamer Carr (Clutton-Brock and Noe-Nygaard, 1990). The latter study was undertaken initially to assess the possibility of a component of coastal occupation in the annual settlement pattern of the Vale of Pickering groups, but the results perhaps could reflect much more local isotopic variations in the food consumed by the dogs (Day, 1996).

The excavations of the new trench at Star Carr found a dense scatter of flint artefacts and faunal remains, associated with a deliberately laid trackway of split and worked timbers (Mellars and Dark, 1998). Clark (1954) originally interpreted the finds in terms of a closed-spaced succession of occupations by relatively small numbers of hunters, which he estimated from the total extent of the occupied zone to be about 20–25 people. However, Mellars and Dark (1998) show that the total distribution of occupation material at Star Carr extends over a much greater area than Clark (1954) envisaged and they documented in fine detail the precise character and topography of the pre-occupation land surface on which the site was established. Clark (1954) nevertheless estimated that the occupation had spanned a substantial period and documented a clear stratigraphical separation of at least two of the major forms of barbed point recovered, but overall the total chronological span remained uncertain. However, close chronological control of the new investigations east of Clark's investigations (Figure 8.37a) has been provided by 12 AMS radiocarbon dates. The overall range of the artefacts and faunal remains extends from *c*. 10 700 to 10 400 calibrated years BP in absolute terms, a period approximately 1000 years earlier than that implied by the original, uncalibrated radiocarbon dates (Mellars and Dark, 1998).

The most remarkable archaeological find in 1985 was the concentration of large timbers of aspen or willow, all lying at the same stratigraphical level close to the base of the organic deposits and closely associated with the earliest occurrences of both flint artefacts and charcoal particles. Most of the timbers show a regular, essentially parallel, arrangement and in several cases represent segments of 'planks' split along either one or both faces. The most plausible explanation is that it is a trackway laid down during the initial stages of the Mesolithic occupation to facilitate access between the occupation area and the open lake water. It is unique in the European Mesolithic period. The archaeological material (flints from in-situ knapping activities, and red deer antlers) from the 1985–1989 excavations represent a separate zone of human activity compared with Clark's (1954) excavations, although it is considered to be rather later in age (Mellars and Dark, 1998).

Clark (1954) saw the site as a base camp location for a few nuclear families and serving as a central focus for a wide range of both subsistence and technological activities. Others, such as Andersen *et al.* (1981) and Legge and Rowley-Conwy (1988), suggested that it was a briefly occupied hunting location. However, the latter's assumption that virtually the whole of Clark's (1954) excavated area represents a 'toss' zone of material discarded from a main centre of activity located beyond the limits of Clark's excavation seems highly unlikely (Mellars and Dark, 1998). There is no doubt that the site served as a major centre for hunting activity and the large-scale manufacture of hunting equipment. However, to suggest that the whole of Clark's excavation area represented a waste-disposal zone as opposed to an area of in-situ activities presents insuperable obstacles (Mellars and Dark, 1998). Firstly, there is a clear pattern in ethnoarchaeological studies of 'toss' zone accumulations in hunter–gatherer sites, in which only the larger refuse fragments are tossed away whereas most of the smaller refuse is allowed to accumulate on the spot in a 'drop' zone (Binford, 1978a, 1983). This presents a marked contrast with Star Carr, where Clark recorded high frequencies of both microliths and flint debitage within the central areas of his excavations. Secondly, the overall distribution and varying densities of the flint artefacts recorded by Clark seems inconsistent with any 'toss' zone hypothesis. He shows four separate areas of high density distribution separated by zones of clearly lower density distribution. Finally, a close analysis of the material represented within each of the artefact concentrations reveals contrasts in the relative frequencies of different tool forms (Mellars and Dark, 1998).

Many different aspects of the evidence suggest that the Star Carr total occupation sequence spans a substantial period, probably over 300 years. These include the stratigraphical spread of the artefacts recorded in both Clark's and the recent excavations, over a depth of 30–40 cm; the clear separation of the two major forms of barbed antler points (types A and E) documented in the original excavations; the overall spread of the radiocarbon dates; and the vertical distribution of macroscopic and microscopic charcoal fragments. The charcoal record points to at least two major episodes of repeated burning at or near the site, the first probably extending from *c.* 10 700 to 10 620 calibrated years BP, and the second from 10 550 to 10 430 calibrated years BP, with the possibility of a third more isolated burning phase at *c.* 10 200–10 300 calibrated years BP. The apparently continuous charcoal input into the deposits during the first major phase would suggest activity either on or close to the site at fairly closely spaced intervals, conceivably on an annual basis (Mellars and Dark, 1998).

In the re-analysis of the faunal remains from Clark's excavations, Legge and Rowley-Conwy (1988) suggested the minimum number of animals at no more than 75 individuals, which translates into a total edible meat-weight of *c.* 12 000 kg. Despite the problems with this kind of analysis related to the food supplies, even if the quantities are doubled this would be sufficient only to support a human group of the size estimated by Clark over a period of continuous occupation for at most two years.

The specific site location of Star Carr seems to have been important for settlement and as Mellars (1998) suggests there are several factors involved. Firstly, the unusual steepness of the early Mesolithic shoreline at this point on the lake edge (Cloutman and Smith, 1988), the narrow reedswamp zone and therefore short distance to open water would have allowed easy access to the lake for various activities and allowed open visibility over the lake. Secondly, the location close to the western end of the lake would have allowed rapid land access to the adjacent western and southern lake shore and therefore exploitation of a long stretch of lake-edge within a relatively short distance. Thirdly, the position close to the end of a major promontory into the lake could have played a role in hunting strategies, such as a means of corralling, or driving game (Andersen *et al.*, 1981), and this type of site location has been found at other sites in Denmark. Fourthly, a feature of Star Carr and sites at Seamer Carr is that the main occupation areas are directly to the south-west of higher ground areas and are sheltered, and they are on the northern lakeshore, where there is maximum sunlight and warmth.

Recent surveys suggest that there are at least ten major Mesolithic artefact concentrations similar in technology and typology to Star Carr in the Vale of Pickering lake basin and more are likely to be found in areas not yet surveyed systematically. The ecological and environmental context of this settlement was within the birch-dominated forests, before the hazel expansion in the Boreal period (Dark, 1998b). The reasons for it being such an attractive location are detailed in Mellars (1998) and probably are related to ease of mobility and visibility across the lake zone; economic productivity and diversity of lake-edge habitats; hunting methods; and the wider food resources available in the immediate surroundings, such as the coast only 12 km to the east, the Wolds and the North Yorkshire Moors.

The shrinkage and impeded lake access is likely to have acted as a deterrent to continued lake-side occupation from the later stages of the Boreal onwards (Walker and Godwin, 1954; Cloutman, 1988a, b; Cloutman and Smith, 1988), but the most critical ecological factor is likely to have been the massive hazel woodland colonization of the dry-land areas. This is reflected in the pollen record by increases of 80–90% (Dark, 1998b) and would have resulted in animal biomass decline (Mellars, 1998).

Mellars (1998) argues that the carefully controlled burning of the reedswamp around the edges of the lake would have substantially increased the quantity and nutritional quality of the new plant growth and hence increase the numbers and predictability of animals feeding at the lake-edge. By selectively burning small areas, hunters could concentrate hunting activities to these areas.

The question of seasonal and annual mobility patterns of the Star Carr population has been reviewed by Mellars (1998). The evidence for the exploitation of upland habitats adjacent to the Vale of Pickering seems strong (Jacobi, 1978; Mellars, 1998), but evidence for the exploitation of coastal locations remains enigmatic. There now seems some evidence for winter occupation, as the excavated site of Barry's Island

(Figure 8.37a) includes remains of two animals killed at some point during the winter or early spring, together with a much higher overall percentage of red deer remains and a much higher frequency of heavily fragmented bones than at Star Carr (Rowley-Conwy, 1995).

Conclusions

Star Carr undoubtedly is the most important early Mesolithic site in Britain and has provided a wealth of archaeological and palaeoenvironmental detail, which has been reviewed and reinterpreted repeatedly. It particularly is important for the range and type of antler artefacts, the evidence for deliberate vegetation modification, for its wooden trackway and for its continued provision of detailed palaeoenvironmental evidence for the Vale of Pickering throughout Lateglacial and Flandrian times. There is no doubt that in the future it will continue to provide further evidence to elucidate a wide range of Quaternary and archaeological problems.

OLD MERE, HORNSEA (TA 208 474)

N.F. Glasser

Introduction

Old Mere, Hornsea, in the East Riding of Yorkshire, is an important site for the study of both local and regional vegetation history and environmental change in the Holderness area during the Holocene Epoch. The importance of the site is twofold. Firstly, palynological analyses of cores obtained from the site have been used in compilation of a regional pollen stratigraphy for Holderness (Beckett, 1981). Secondly, comparisons between Old Mere, Hornsea and the much smaller nearby kettlehole at The Bog, Roos have been used to demonstrate the importance of local factors in pollen deposition and preservation. The main published work on this site is that of Beckett (1975, 1981), who obtained radiocarbon-dated cores from Old Mere, Hornsea and from The Bog, Roos (Beckett, 1977b). The differences in the pollen records from the two sites reflect the much stronger local influences on pollen deposition at Old Mere, Hornsea. This is primarily the result of a greater input from local streams into the lake because of the larger size of the catchment at Old Mere. The site also features in many reviews of the regional palaeoenvironmental history of this area of Holderness (Gilbertson, 1984a; Flenley 1984, 1987, 1990; Catt, 1991b; Van de Noort and Davies, 1993; Dinnin, 1995; Dinnin and Lillie, 1995b; Taylor, 1995; Greig, 1996; Lillie and Geary, 1999).

Description

Old Mere, Hornsea lies on the Holderness coast, immediately south of the town of Hornsea. The entire Holderness Plain is composed of glaciogenic sediments, including a variable thickness of diamicton, sand and gravel. Organic deposits and place names provide evidence that numerous former shallow lakes (meres) existed in Lateglacial and post-glacial times, especially in eastern Holderness. Their disappearance was largely the result of natural silting aided by medieval and later drainage (Sheppard, 1957). The only surviving open-water mere is that of Hornsea Mere. The GCR site at Old Mere, Hornsea is a dry depression of landscaped ground at *c.* 1 m OD, protected from the North Sea by a sea wall. The site occupies the eastern end of a pre-glacial valley, which extends westwards to Brandesburton and includes the surviving Hornsea Mere, which is immediately inland of Old Mere itself (Figure 8.43). Hornsea Mere is approximately 2.5 by 1 km and is drained by a dyke, which runs through the centre of the Old Mere basin. When Old Mere, Hornsea was open water, it had dimensions of at least 300 by 500 m (Beckett, 1981). A terrace approximately 3 m above present water level in the existing Hornsea Mere indicates a previously higher water level, and at some point in the past the two meres probably were joined into a single, much larger lake.

A transect of borings reported by Beckett (1981) revealed up to 14 m depth of lacustrine deposits in the Old Mere basin, comprising mainly clay and fine detritus mud (Table 8.9).

The upper 9.5 m of a core obtained by Beckett (1981) yielded sufficient pollen for palynological analysis and radiocarbon dating (Figures 8.44 and 8.45). The core has been divided into five pollen assemblage zones based upon pollen and spore content, each of which is given the local code HO.

HO-1 (9.45–8.95 m)

The pollen spectra are dominated by *Betula* (40–70%), *Pinus* (10%) and Gramineae (5–10%).

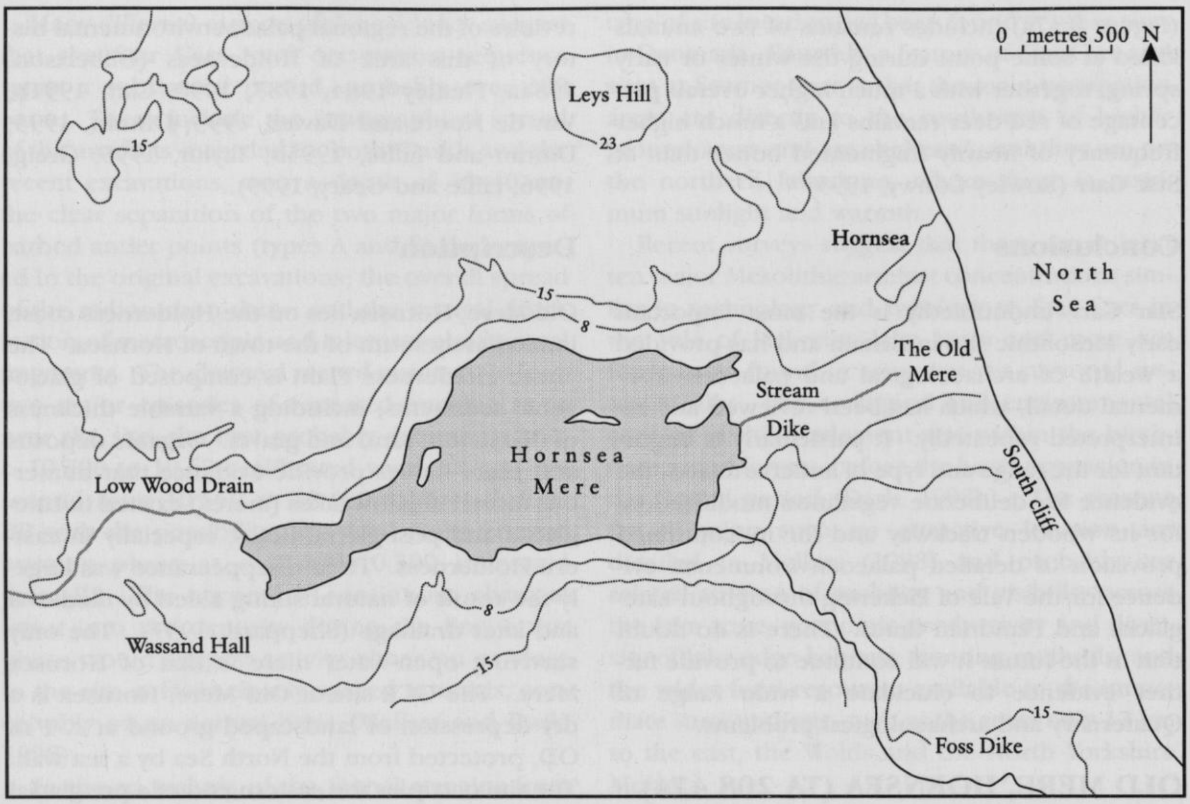


Figure 8.43 Map of the area surrounding Old Mere, Hornsea (modified from Beckett, 1981). (Contours in metres.)

Quercus, *Ulmus* and *Alnus* also occur in small amounts towards the top of this zone. *Salix* pollen is abundant at the base of this zone, but decreases upwards. The pollen of certain herbs, notably *Filipendula*, is present.

HO-2 (8.95–7.55 m)
The lower boundary of this zone is drawn where *Corylus*/*Myrica* pollen shows a sharp increase and *Betula* declines. *Ulmus* and *Quercus* also increase at the base of this zone. *Corylus*/*Myrica* pollen dominates, with values from 70% at the base to 40% at the top of this zone. *Betula*, *Pinus*, *Ulmus* and *Quercus* all contribute around 5–10%. *Tilia cordata* and *Alnus* pollen are present in small amounts. Very few herbs are present, apart from Gramineae and Cyperaceae pollen.

HO-3 (7.55–4.40 m)
The base of this zone is marked by an increase in *Alnus* pollen to exceed 20%. *Alnus* contributes about 30% throughout the zone. Of the other trees, *Betula* and *Pinus* both decline from about 10% at the base to 2 or 3% at the top of the zone. *Ulmus* pollen values (5–10%) are also slightly higher at the base. *Quercus*, at 5–10% for much of this zone, increases to 15–20% near the top. *Tilia cordata* pollen values increase from minimal amounts to around 5% higher in this zone, and *Fraxinus* is sparsely present in the upper horizons. *Corylus*/*Myrica* pollen contributes

Table 8.9 Generalized stratigraphy of Old Mere, Hornsea (source: Beckett, 1981)

Depth in core (cm)	Description
0.00–0.50	Made ground
0.50–1.40	Sandy clay
1.40–1.75	Clayey detritus mud with organic matter
1.75–9.30	Fine detritus mud with no recognizable plant material
9.30–12.30	Silty clay with occasional organic matter
12.30–12.60	Clayey fine detritus mus with some silt
12.60–13.80	Gravelly clay

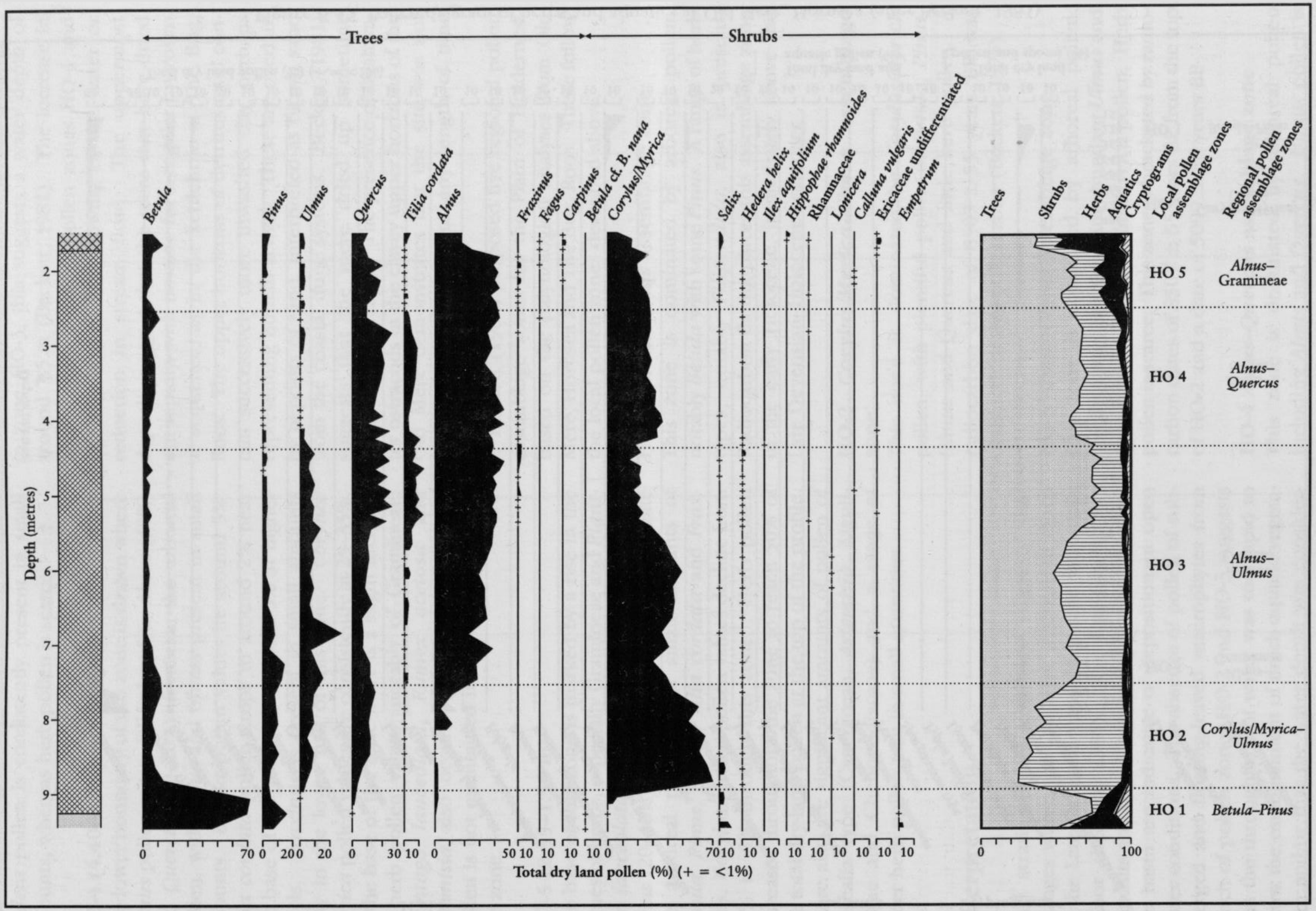


Figure 8.44 Pollen diagram of trees and shrubs at Old Mere, Hornsea (after Beckett, 1981). See Figure 8.1 for key to the stratigraphical log and Table 8.9 for a more detailed description.

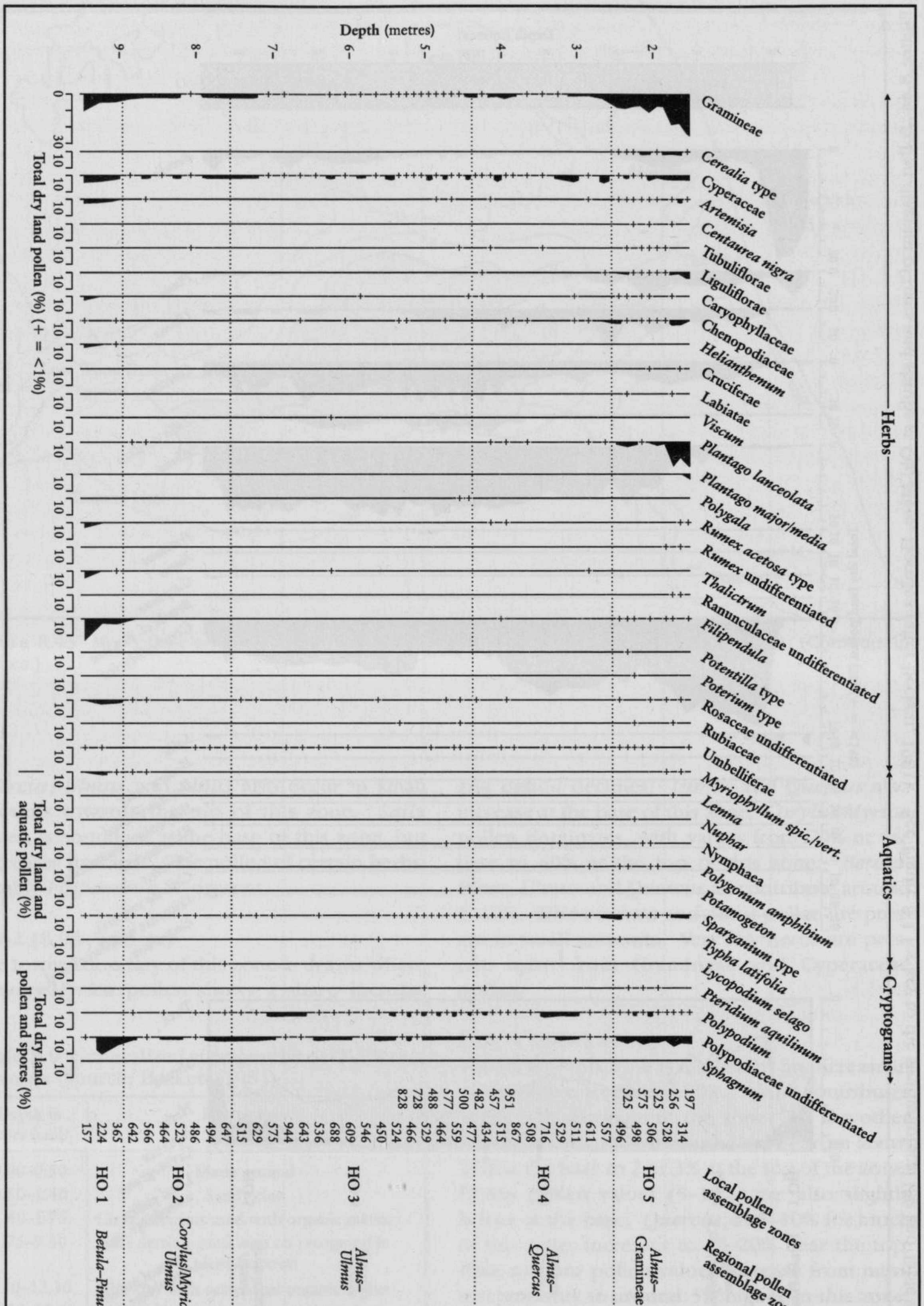


Figure 8.45 Pollen diagram of herbs and aquatics at Old Mere, Hornsea (after Beckett, 1981).

40%, decreasing to 20% at the top of the zone. *Hedera* pollen is consistently present in small amounts, whereas herb pollen is scarce.

HO-4 (4.40–2.55 m)
The lower boundary of this zone is drawn where *Ulmus* pollen values fall below 5%. *Alnus* (40%) and *Quercus* (10–20%) dominate the arboreal pollen, with *Pinus* and *Ulmus* present in small amounts. *Betula* pollen values lie around 5%. *Tilia cordata* pollen drops to around 2% near the base of the zone, but exceeds 5% at higher levels. *Fraxinus* is present in small amounts only in the lower half of this zone. *Corylus/Myrica* pollen values are consistently at 25–35%. At the base of the zone there is a slight increase in herb pollen values, notably of Gramineae; *Plantago lanceolata*, *Rumex acetosa* and *Artemisia* also occur. The increase in herb pollen is not maintained in the higher levels of the zone.

HO-5 (2.55–1.55 m)
The base of this zone is marked by a rise in the values of herbs, especially Gramineae and *Plantago lanceolata*, and a decline in *Tilia cordata*. *Alnus* (20–40%) and *Quercus* (5–20%) dominate the arboreal pollen, with small amounts of *Betula*, *Pinus*, *Ulmus*, *Tilia cordata* and *Fraxinus*. *Corylus/Myrica* pollen values decline from 30% to 5% through this zone. Herb pollen increases throughout the zone to reach 50% of the total dry-land pollen at the top of the profile. There also are significant amounts of pollen of Cerealia type (Cyperaceae, *Artemisia*, Liguliflorae and Chenopodiaceae) and a range of other herb pollen taxa in small quantities.

Interpretation

The stratigraphical evidence suggests that changes within the Old Mere throughout much of the Late Devensian and Holocene were only minor (Beckett, 1981). The Late-Glacial clay deposits and fine detritus mud that fills much of the basin are indicative of deposition in open water conditions. The absence of pollen of submerged and floating-leaved macrophytes from much of pollen zones HO-2 and HO-3 suggests that the throughflow of water was too rapid to allow the establishment of much aquatic vegetation and/or that the water depth was considerable. On the assumption that the raised shorelines of Hornsea Mere represent the water depth in zone HO-3, this suggests a water depth of around 7.5 m (Beckett, 1981). The increase in aquatic pollen during pollen zones HO-4 and HO-5 points to a shallowing of the water or reduction in stream flow. The uppermost deposits suggest that the mere may have dried out rapidly, with no time for a regular hydroseral sequence, as in the kettlehole at The Bog, Roos. The upper horizons of detritus mud contain successively more inwashed clay, perhaps representing human activity. These are overlain by deposits of sand, interpreted as blown sand from the coastal dune systems. Beckett (1981) suggests that the mere dried up sometime around 2000 years BP. The presence of freshwater ostracods in the clayey upper horizons of the Old Mere demonstrates that the site was not inundated by the sea for any length of time (Robinson, 1972).

Beckett (1981) proposed five regional pollen assemblage zones for the Plain of Holderness based on the palynological analyses from Old Mere, Hornsea and The Bog, Roos. These follow the local pollen zones described above.

HO-1 *Betula–Pinus* assemblage zone
This zone is dominated by arboreal pollen, notably *Betula* with some *Pinus*. A range of herb types, notably *Filipendula*, also is present. Radiocarbon dating places this assemblage zone in the early Holocene, immediately above the Late Devensian–Holocene boundary.

HO-2 *Corylus/Myrica–Ulmus* assemblage zone
This zone is dominated by *Corylus/Myrica* pollen, with plentiful pollen of *Betula*, *Pinus*, *Ulmus* and *Quercus* and little herb pollen. A radiocarbon date of 8507 ± 55 years BP was obtained for this pollen zone (Beckett, 1975).

HO-3 *Alnus–Ulmus* assemblage zone
This zone is dominated by arboreal pollen, chiefly *Alnus*, but also with abundant *Ulmus* and *Quercus* and some *Corylus/Myrica* pollen. Herb pollen is scarce. This zone is bracketed by radiocarbon dates of 8507 ± 55 years BP from the top of HO-2 and a date of 5099 ± 50 years BP.

HO-4 *Alnus–Quercus* assemblage zone
This zone is dominated by arboreal pollen including *Alnus* and *Quercus*. *Ulmus* pollen is scarce. Small amounts of herb pollen occur,

notably Gramineae, *Plantago lanceolata* and Chenopodiaceae. The base of this zone is dated to 5099 ± 50 years BP (Beckett, 1975).

HO-5 *Alnus*–Gramineae assemblage zone
This zone contains abundant herb pollen. *Alnus* dominates the arboreal pollen, with smaller amounts of *Betula* and *Quercus*. Many herb pollen types are present, notably Gramineae, *Plantago lanceolata*, and Cerealia type. Dates for the start and end of this zone are unclear, but radiocarbon dates indicate that this pollen zone includes the time spanning 3120 ± 105 years BP to 3433 ± 110 years BP.

Conclusions

Old Mere, Hornsea has yielded a radiocarbon-dated pollen sequence from the Late Devensian–early Holocene boundary to *c.* 2000 years BP. The pollen record from Old Mere, Hornsea is important in defining the regional pollen assemblages for Holderness, of which five regional pollen zones have been recognized. The pollen record from Old Mere is very different to that obtained from nearby at The Bog at Roos, where the pollen spectra are swamped by local pollen produced from the rapidly changing bog surface.

FEN BOGS (SE 853 977)

R.C. Chiverrell

Introduction

Fen Bogs, North Yorkshire is one of the deepest peat sequences in the British Isles, with a maximum depth of 11.2 m. The pollen stratigraphy has contributed evidence critical for understanding the Holocene vegetation history of the North York Moors. Atherden (1972, 1976a, b) first studied the stratigraphy and development of Fen Bogs, focusing upon the impact of people on the Holocene vegetation of the eastern North York Moors. The pollen sequence at Fen Bogs is a cornerstone of wider discussion of vegetation change in north-east Yorkshire (Jones *et al.*, 1979; Simmons *et al.*, 1993; Atherden, 1999). Research at Fen Bogs also has focused upon the finer detail of vegetation changes and climate change during the past 2000 years (Chiverrell, 1998; Chiverrell and Atherden, 1999, 2000).

Description

Fen Bogs developed in the deeply incised glaciofluvial channel of Newton Dale. Newton Dale is a former glacial drainage channel cut in a north–south direction across the Middle Jurassic central plateau of the North York Moors (Gregory, 1962a, b, 1965). The bog has formed at an altitude of 164 m on the watershed between Eller Beck and Pickering Beck, is 1.5 km in length, 0.3 km in width and covers an area of about 300 m². The surface is flat and bordered by steep slopes rapidly rising 90 m on both the east and west sides. The stratigraphical development of Fen Bogs was determined by Atherden (1972, 1976b). Chiverrell (1998) focused upon stratigraphical changes in the top 3 m.

A composite stratigraphy for the centre of the bog is presented in Table 8.10. The basal sequence reflects the early stages of peat inception, with well-humified peat-rich material and inorganic material overlying periglacial clays (beds 6–5). Alder, birch and willow fen woodland deposited peat sediments containing abundant woody remains (beds 4–3). *Phragmites australis* reed-swamp succeeded the fen woodland, and moorland plants (Ericaceae, *Eriophorum* and *Sphagnum*) became more abundant. The upper 1.4 m contains a change to a more ombrotrophic community typical of the present-day mire, with *Sphagnum* and *Eriophorum vaginatum* replacing *Phragmites australis* (Atherden, 1976b). After colonization by an ombrotrophic mire flora the changes in peat stratigraphy appear to identify fluctuations in mire surface wetness (Chiverrell, 1998; Chiverrell and Atherden, 1999). Borings from the bog margins encountered an in-wash stripe of pale grey clay that is attributed to increased erosion after *c.* 4720 ± 90 years BP (Simmons *et al.*, 1975).

Atherden (1972) first studied the pollen biostratigraphy at Fen Bogs and published a complete Holocene pollen diagram from the centre of the mire (Figure 8.46). The basal stratigraphy identifies the sequence of woodland colonization in north-east England during the early Holocene. Five radiocarbon dates secure the age of the vegetation sequence for the past 5000 years, and illustrate the impact that prehistoric and historic communities had upon the landscape. An undated pollen diagram sampled close to Atherden's original site contains an identical sequence of vegetation changes (Chiverrell and Atherden, 1999, 2000).

Table 8.10 Peat stratigraphy at Fen Bogs, North York Moors (after Atherden, 1976a; Chiverrell, 1998).

Bed	Depth (cm)	Environment	Stratigraphy
1	0–140	Ombrogenous mire	Poorly humified *Sphagnum* and Monocotyledonous peat
2	0–140	Ombrogenous mire	Well-humified Monocotyledonous peat
3	140–600	*Phragmites* reed-swamp	Well-humified *Phragmites australis* peat, with occasional other mire plant remains (*Eriophorum* spp., Ericaceae and *Sphagnum*)
	600–820	Partially wooded *Phragmites* reed-swamp	Well-humified *Phragmites* peat, with occasional wood remains (*Betula*, *Alnus* and *Salix*)
4	820–920	Fen/Carr woodland mire	Well-humified wood peat, with *Betula*, *Salix* and occasional *Phragmites* remains
5	920–960	Mire inception	Well-humified peat rich with inorganic material
6	960–	Periglacial valley	Blue-grey clay solifluction deposits

Interpretation

Atherden (1976a, b) synthesized the vegetation development of the eastern North York Moors using the Fen Bogs pollen diagram (Figure 8.46). Peat deposition began during the early Holocene, with pollen zone FB1 dominated by *Betula*, *Pinus* and Cyperaceae. The basal sediments are undated, but clearly it pre-dates the arrival of the thermophilous trees before 8500 years BP during the early Holocene. In pollen zone FB2 the early dominance by *Betula* is relinquished in favour of a mixed coniferous–deciduous forest dominated by *Pinus* and *Corylus*, with *Quercus* and *Ulmus* appearing for the first time. The early Holocene sequence of woodland immigration is not dated at Fen Bogs. Extensive research across the moors has dated key events in the pollen biostratigraphy – including the *Corylus* rise *c*. 9000 years BP and the *Alnus* rise *c*.7500 calibrated years BP at Seamer Carr (Vale of Pickering) and West House Moss in Eskdale (Jones, 1977b; Cloutman, 1988a, b; Day, 1995).

Woodland pollen frequencies peak in FB2 and decline in the subsequent FB3, probably reflecting local expansion of mire communities (Chiverrell and Atherden, 1999). The middle Holocene landscape was covered with *Quercus*, *Corylus* and *Alnus* woodland, with *Ulmus*, *Tilia* and *Fraxinus* significant components of the forest (FB3 and FB4). There is considerable evidence for woodland disturbance by nomadic Mesolithic people elsewhere on the moors – for example North Gill and Bonfield Gill Head (Simmons and Innes, 1988a, b). It appears likely that the combination of early peat inception and Mesolithic woodland disturbance produced a scrubby open-work woodland cover, interspersed with moorland and open-ground communities perhaps producing the lower tree pollen frequencies in FB3 (Simmons and Innes, 1988a; Chiverrell and Atherden, 1999; Atherden, 1999).

The Elm Decline, widely documented at many British sites, is clearly seen at the top of FB3; it has been dated at Fen Bogs to 4730 ± 90 years BP. Pollen zones FB4 and FB5 are characterized by fluctuations in several pollen curves. Cyclic reductions in tree pollen and increases in ruderal and heliophytic taxa identify small temporary clearances. These temporary clearances begin in FB4, which is dated to the Neolithic between 4730 ± 90 years BP and 3400 ± 90 years BP. The clearances increase in frequency during FB5, which is dated to the Bronze Age between 3400 ± 90 years BP and 2280 ± 120 years BP. Cycles of woodland clearance are in keeping with abundant archaeological evidence for human activity (Spratt, 1993). Neolithic settlement was concentrated in the southern North York Moors and there is abundant archaeological evidence for settlement and farming by Bronze Age communities (Spratt, 1993).

Pollen zone FB 6 contains the most substantial woodland clearance event at Fen Bogs, which is dated to 2280 ± 120 years BP. Woodland pollen frequencies decline to values equivalent to the present day, and there are increases in Poaceae and all other non-woodland taxa alongside the

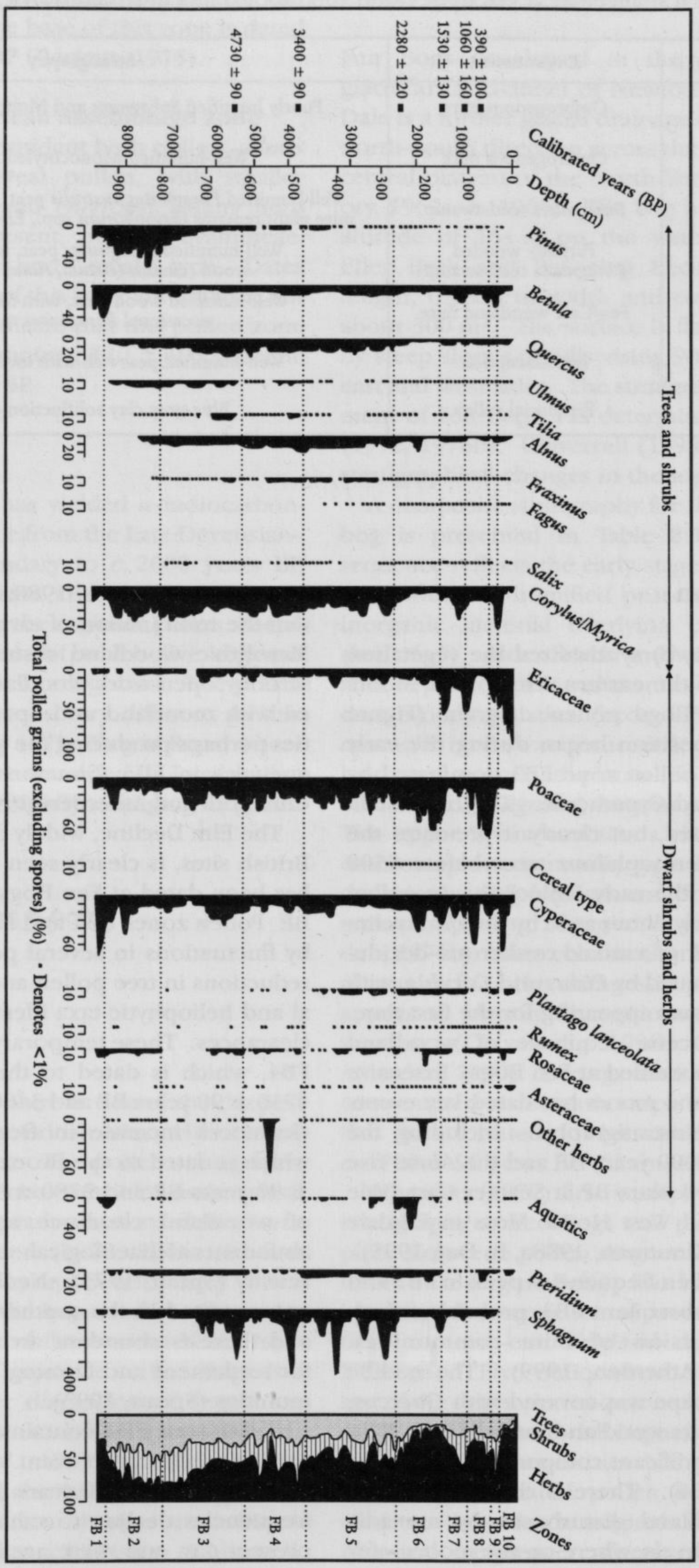


Figure 8.46 Atherden's pollen diagram from Fen Bogs (after Chiverrell and Atherden, 1999).

continuous presence of cereal pollen. The scale of the clearance reflects the establishment of permanent settlements encouraging sustained grazing pressure farther into the moors, preventing the regrowth of trees (Atherden, 1976a; Chiverrell and Atherden, 1999, 2000). The remaining trees may have been coppiced locally to sustain the charcoal needed for iron smelting, which would suppress pollen production. This Iron Age and Romano-British woodland decline was controlled by increased landscape pressure from farming activities arising from population expansion around and into the moors, perhaps combined with a more commercial approach to farming during Roman occupation of Britain (Chiverrell and Atherden, 1999).

Pollen zones FB7 to FB10 contain two phases of woodland regeneration separated by a further decline. All woodland taxa, but particularly *Betula* and *Corylus* recover in FB7, which is dated to between 1530 ± 130 years BP and 1060 ± 160 years BP. This regeneration has been identified and dated at six sites across the North York Moors (Chiverrell, 1998; Blackford and Chambers, 1999; Chiverrell and Atherden, 1999). Woodland recovered perhaps owing to a reduced scale of agriculture linked to economic malaise in the aftermath of the Roman withdrawal from England. Towards the end of the first millennium AD, during FB8, tree species decline and Poaceae and cereal pollen increase, signifying increased agricultural endeavour. Vegetation changes during the Anglo-Scandinavian period reflect economic revival and population expansion, and these trends continue until a sharp increase in woodland taxa below the FB9 boundary.

The woodland recovery in FB9 starts below a radiocarbon date of 390 ± 100 years BP and research at an adjacent mire, May Moss, encounters a similar woodland recovery before *c.* 685 ± 50 years BP, probably during the 12–14th centuries (Chiverrell and Atherden, 1999). Causes of this minor woodland recovery could include the devastation in the wake of the 'harrying of the north' in AD 1069–1070 and demographic collapse during the 14th century owing to incidence of Black Death (Harrison and Roberts, 1989). Woodland decline and expansion of agricultural indicators during the 11–12th centuries and after AD 1500 (FB 8 and FB10) probably reflect concerted attempts to colonize upland Britain and to exploit marginal land for agriculture (Parry, 1976; Harrison, 1993).

The peat stratigraphy of Fen Bogs is dominated by *Phragmites australis*, revealing little environmental information other than the site was a wet upland valley fen. After the mire was colonized by an ombrogenous mire flora the stratigraphy yields useful palaeohydrological information, with the initial expansion of poorly humified *Sphagnum* peat reflecting this colonization. *Sphagnum* species decline between 130 and 95 cm, signifying drier surface conditions. The change to poorly humified *Sphagnum papillosum* at 95 cm indicates a wetter mire surface during FB8. Relatively wet conditions persist until the drier environmental indicator *Sphagnum* section *Acutifolia* replaces *Sphagnum papillosum* during FB9. This dry phase is short-lived, terminating with a return to wetter conditions evidenced by the replacement of *Sphagnum* section *Acutifolia* with the wet environmental indicator, *Sphagnum* section *Cuspidata*.

The sequence of moisture changes in the upper 1.4 m occurs entirely within the past 800–700 years and appears to parallel Little Ice Age climatic fluctuations identified at other peat sites on the moors (Chiverrell, 1998). The uppermost stratigraphy contains evidence of human interference, almost certainly related to the building of the Whitby to Pickering railway across Fen Bogs in 1836 (Statham, 1989). Mires are an easily damaged environment and this unprecedented scale of human activity on the western edge of Fen Bogs could quite conceivably have had a catastrophic impact on the hydrology and flora.

Conclusions

Fen Bogs is an important site because it has yielded a significant record of vegetational and environmental change during the last 10 000 years. Detailed pollen records coupled with radiocarbon dating demonstrate the landscape history of the North York Moors. The sequence is one of few dated pollen profiles focusing upon mid- and late- Holocene times. The sequence identifies the impact that Neolithic, Bronze Age, Iron Age and historic communities had upon the landscape of North Yorkshire. Correlation of vegetation history with archaeological and documentary records has given this site particular importance, because Fen Bogs has produced the region's most complete and thoroughly dated pollen diagrams.

GORMIRE (SE 503 832)

J. Innes and S. Morriss

Introduction

Gormire is a small lake on the south-western fringe of the North York Moors. It is important for preserving a long sequence of Late Devensian and Holocene sediments, which have been investigated by Blackham *et al.* (1981) using pollen and plant macrofossil analyses. Recent research has been undertaken under the Land Ocean Interaction Study (LOIS) research project (NERC, 1999; Oldfield *et al.*, 1999) on magnetic signals of sediment sources to the lake by Barlow (1998), on organic geochemistry by Fisher (1999) and on pollen analyses of recent sediments by Morriss (2001). The origin of the landform containing the lake has been discussed by Kendall and Wroot (1924).

Description

Gormire Lake, North Yorkshire lies at about 150 m OD at the edge of the Hambleton Hills escarpment of the south-western North York Moors at Whitestone Cliff, about 7 km to the east of Thirsk at the eastern edge of the Vale of York. It occupies a depression in probably morainic glacial deposits at the foot of the scarp slope of Sutton Bank and is enclosed to the west by the arcuate ridge of Gormire Rigg (Figure 8.47). The site's location corresponds exactly to the limit of Devensian glacial ice in this part of the Vale of York, as the ice stream, reaching to about 200 m OD, failed to overtop the major upland escarpment (Hemingway, 1993). The origin of the depression is uncertain, but Kendall and Wroot (1924) suggested that it probably represents a lateral drainage channel formed at the ice margin and dammed in the Late-glacial period by a landslip of glaciogenic material from the steep escarpment slopes. If so, the slumped material must today represent the lower sill at the south-east corner of the site. Smaller channels run parallel to the main one, and these probably also originated as meltwater drainage features. Gormire Rigg and Sutton Bank, the slopes that enclose the site, support dense deciduous woodland, with Garbutt Wood on Sutton Bank, a Nature Reserve, perhaps of great age.

Blackham *et al.* (1981) surveyed the lake and found the deepest, central area to have a flat muddy bottom at about 6 m in depth. They recovered three cores for analysis, two (A and B) from the deep central area and one (C) from an area of marginal swamp in the north-east corner of the basin. The short marginal core C was over 4 m in length but was used only to record stratigraphy. It contained alternating coarse detritus mud and brown lake mud with a layer of grey clay in the upper metre and a surface moss peat. Core A was 10 m in length, collected from beneath 6 m of water, and was used for stratigraphical, pollen and plant macrofossil analyses (Figure 8.48). Core B was 2.73 m long (Figure 8.49) and was used for stratigraphical, pollen and magnetic susceptibility analyses. Core A consisted of 6 m of brown lake mud overlying grey clay, whereas core B contained clastic elements within its brown lake mud sequence, in particular a high clay fraction between 1.00 m and 1.70 m depth and a sandy clay unit that occupied most of the upper 0.30 m.

Pollen analyses of the basal grey clay unit in core A showed the lower part to be dominated by Cyperaceae, with lesser frequencies of Gramineae, *Juniperus*, *Betula nana* and *Lycopodium selago*. *Juniperus* increases in frequency in the upper part of this clay and *Salix* becomes important, as do a range of herbs including *Artemisia*, *Filipendula* and Chenopodiaceae. No plant macrofossils were recorded in the clay. In contrast the overlying lake muds are characterized by post-glacial forest tree pollen. *Betula*, *Ulmus* and *Corylus/Myrica* initially are most abundant, and are joined at *c.* 5 m depth by *Quercus* and *Alnus*. High *Potamogeton* values occur. The upper 2 m of the core show declines in tree pollen values, including *Ulmus*, rises in Gramineae and Cyperaceae and the introduction of herbaceous indicators of forest clearance, such as Chenopodiaceae and *Plantago lanceolata* and, above *c.* 1 m, Cerealia type. Plant macrofossils in the lake mud were mostly *Betula*, *Quercus* and *Salix*, with some *Hypnum* and *Pohlia* mosses and many unidentifiable monocot remains.

The pollen record from core B also is dominated by tree and shrub pollen, although *Ulmus* is low from the start, until the clayey lake mud horizon in mid-profile, during which herbaceous pollen values increase considerably. Gramineae, Cyperaceae, *Plantago*, *Rumex*, Leguminosae and *Artemisia* all reach peak values. Tree and shrub pollen return to dominance in the upper profile, although less so than below

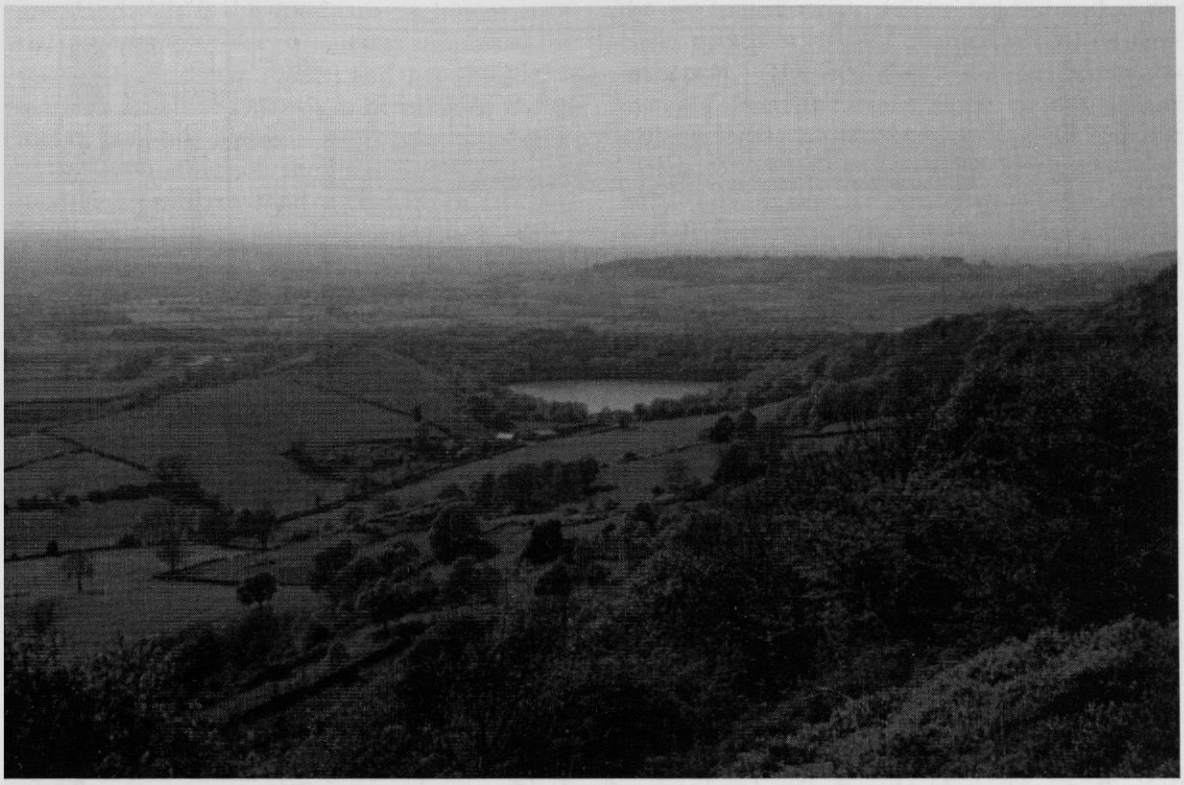


Figure 8.47 Gormire Lake. (Photo: J. Innes).

the clay-rich horizon. Magnetic susceptibility values peak within the clay-rich layer and in the near-surficial sandy clastic unit.

Barlow (1998) has examined the recent sediments of Gormire lake in six cores that could be correlated consistently in detail using magnetic measurements. These, allied to ^{210}Pb and ^{137}Cs profiles, indicated a relatively undisturbed area of sedimentation over much of the bed of the lake, which allowed reliable estimation of mineral sediment accumulation and catchment yields using rates extrapolated from the ^{210}Pb chronology. Mineral sediment influx rates for the period since AD 1630 were measured in this way and showed a doubling of the rate of sediment input from erosion in the small lake catchment since 1949 (Oldfield *et al.*, 1999). During this project (NERC, 1999) the whole of the Holocene sedimentary record at Gormire has been subjected to a range of environmental analyses (Barlow, 1998; Fisher, 1999; Oldfield *et al.*, 1999) of its chemical, physical and magnetic properties and the microfossil record. A radiocarbon chronology has been established. Oldfield *et al.* (1999) have presented geochemical and pollen data plotted against 'Hard' IRM and ARM/SIRM percentage magnetic values. They found that peak ARM/SIRM values coincide with high concentrations of the biophilic elements calcium and bromine and maximum tree pollen percentages, whereas minimum values correspond to the main periods of forest clearance. The forest clearance episodes also correspond to higher values for the lithogenic elements potassium and strontium, as well as peak values for the haematite dominated 'Hard' IRM. The pollen data and radiocarbon dates (Oldfield *et al.*, 1999) show the major period of forest clearance and farming in Romano-British times, followed by forest regeneration and then further clearance continuing to the present. The great increase in grass pollen during the deforestation phases was matched by analyses of the organic matter content of the sediment. Oldfield *et al.* (1999) used qualitative and quantitative analyses of lipid biological markers and found a close relationship between the pollen signal and the organic matter sedimentary record. Depleted total organic carbon levels matched high-grass pollen frequencies and reflected organic-poor

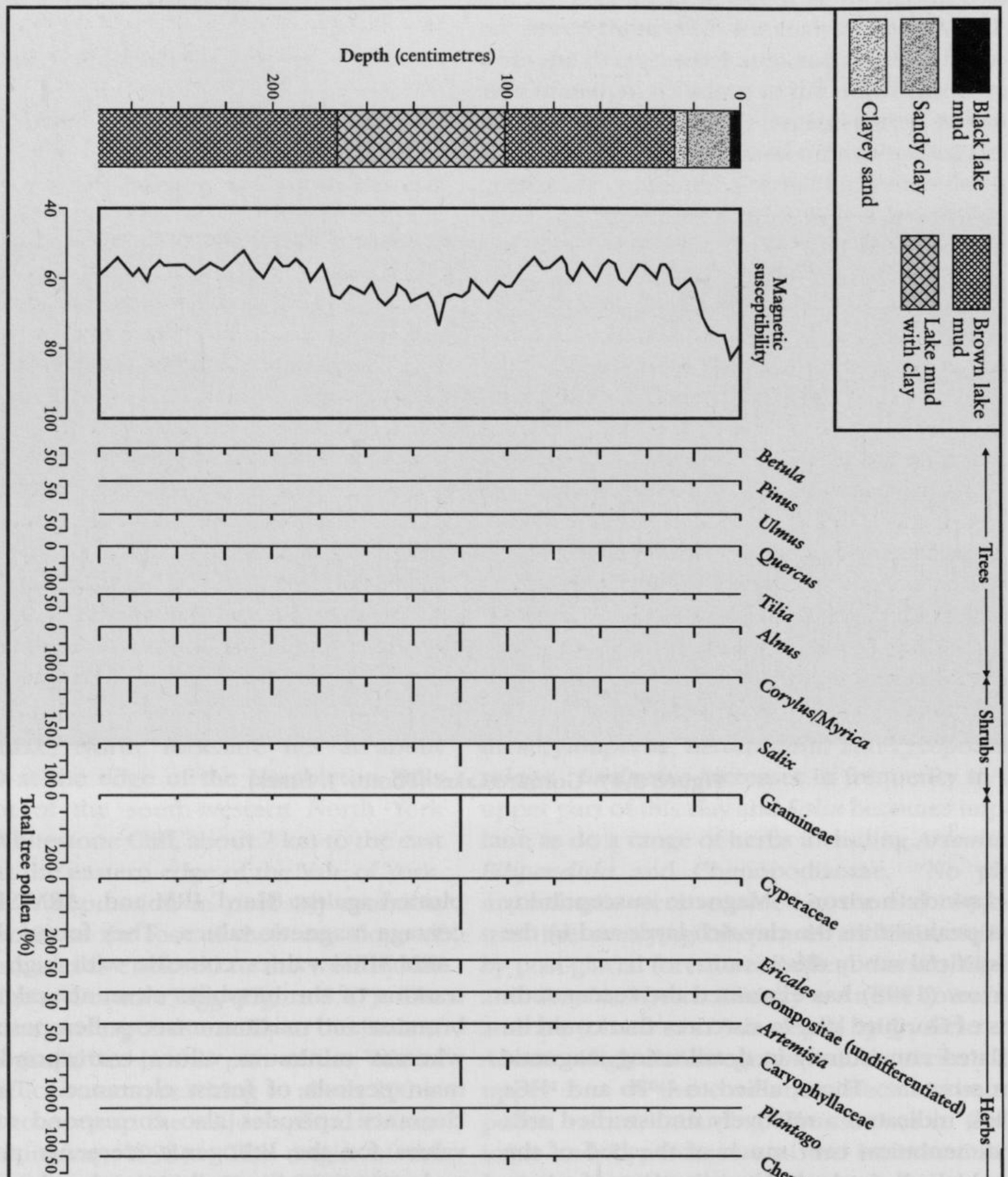


Figure 8.48 Stratigraphy and outline pollen analysis of core A, Gormire Lake (after Oldfield, 1999).

grassland soils as the dominant catchment material. Other molecular parameters supported this result (Oldfield *et al.*, 1999) and these biological markers have been shown to be valuable indicators of land-use change.

Understanding of the medieval and post-medieval environmental record from Gormire has been enhanced greatly by a detailed palynological investigation focusing on the top 70 cm of sediment (Morriss, 2001). The chronology of the profile is yet to be finalized but preliminary magnetic susceptibility data correlated to the ^{210}Pb-dated core from Gormire described above (Oldfield *et al.*, 1999) indicate that the base of the record may be assigned a date of *c.* AD 1350. Assuming that this basal date is correct, the beginning of the pollen record in this study would appear to occur in the middle of the medieval monastic period. There was much Cistercian activity in the vicinity of the lake, with a number of Abbeys established in the 12th century, including Rievaulx, Byland, Fountains and Jervaulx. These Abbeys held significant amounts of land and this may explain the relatively high levels of pollen taxa indicative of arable agriculture such as cereals and *Centaurea cyanus* and to a lesser degree taxa representing pastoral activity. The low arboreal pollen values suggest that cultivation and grazing took place in a largely open landscape.

The pollen profile then records a phase of *Cannabis* cultivation within the mixed farming economy. This is followed by a transitional period in the pollen record. Grazing indicators such as grasses and *Plantago lanceolata* begin to increase, accompanied by a large expansion of Lactuceae (*Taraxacum*-type) pollen and *Pteridium* spores, whereas cereals decline. It would seem that pastoralism was expanding in the surrounding area at the expense of arable agriculture. In the later part of the profile, pastoral frequencies continue to rise, whereas arable indicators remain relatively subdued. There also is a very gradual increase in pine pollen in the top 25 cm of the record as a result of the plantations established in the 19th and 20th centuries. Further documentary research will be undertaken to correlate with the pollen evidence from this profile.

Interpretation

Whereas the most important feature of Gormire is the sedimentary record within the lake basin, the geological origin of the landform that the basin occupies is itself of interest and remains uncertain. The hypothesis that landsliding of material from the scarp slope of the upland plateau was involved in the process (Kendall and Wroot, 1924) remains favoured (Blackham *et al.*, 1981) but unproven. Such events are common and have created depositional basins in the Holocene (Simmons and Cundill, 1974b; Tallis and Johnson, 1980), assuming that sediment accumulation followed soon after stabilization. If a landslip event did create the Gormire lake basin, the pollen data from the grey clay at the base of core A of Blackham *et al.* (1981) show that this occurred at some stage in the Late Devensian Late-glacial period. The sampling interval in this core is very wide, but the clastic nature of the clay deposit and the dominance of a low shrub and herbaceous tundra-type pollen flora within it suggest a Loch Lomond Stadial age. With tree *Betula* virtually absent from the clay there are no indications that the deposits sampled extend to the Late-glacial interstadial. It cannot be assumed, however, that these lowest sediments sampled at Gormire are the oldest present in the basin and the pollen data can provide only a minimum age for the formation of the lake. It is possible that a full Late-glacial succession is present in the lake basin. As the basin at nearby Seamer Carrs in the Cleveland Lowlands was ice-free by at least 13 042 ± 140 years BP (Jones, 1976a, 1999), and possibly rather earlier, the formation of the basin at Gormire could have occurred at any time after this local deglaciation and before the start of the Loch Lomond Stadial about 11 000 years BP.

Blackham *et al.* (1981) recorded an almost continuous Holocene pollen record in their core A at Gormire, although the sampling interval was very wide and sediment was lost from three points in the core during its recovery. One of these sediment gaps occurs at the mud–clay boundary and so the earliest Holocene pollen record may be missing. The potential for a complete Holocene depositional record has been proven, however, and Gormire is an extremely important palaeoenvironmental resource, as sites with such potential are extremely few in this part of the region. Although there are sediment gaps in the upper part of core A, Blackham *et al.* (1981) were able to recognize two clear phases of forest clearance, with the later phase including substantial cereal cultivation. Their core B (Figure 8.49) also showed clear evidence

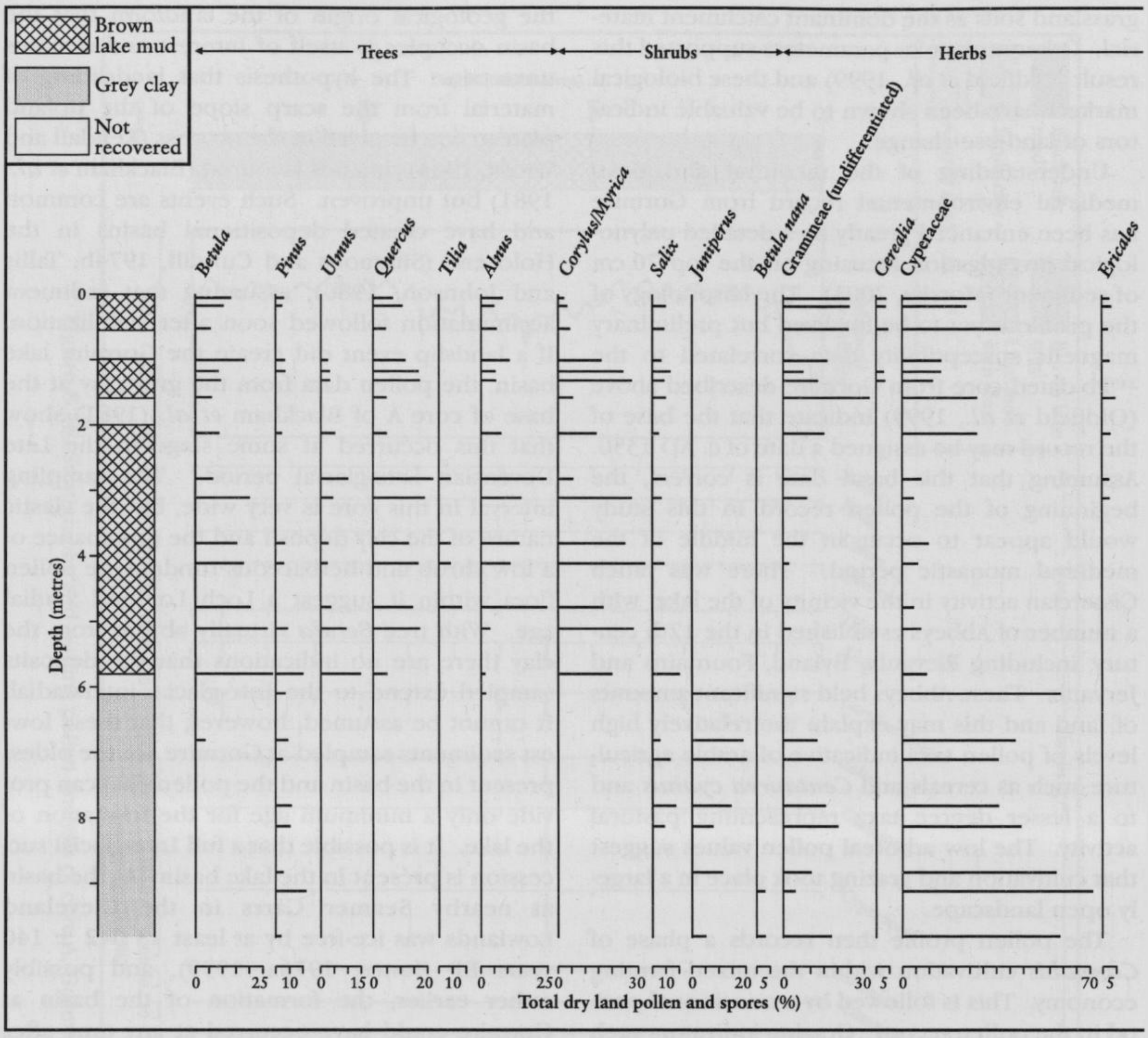


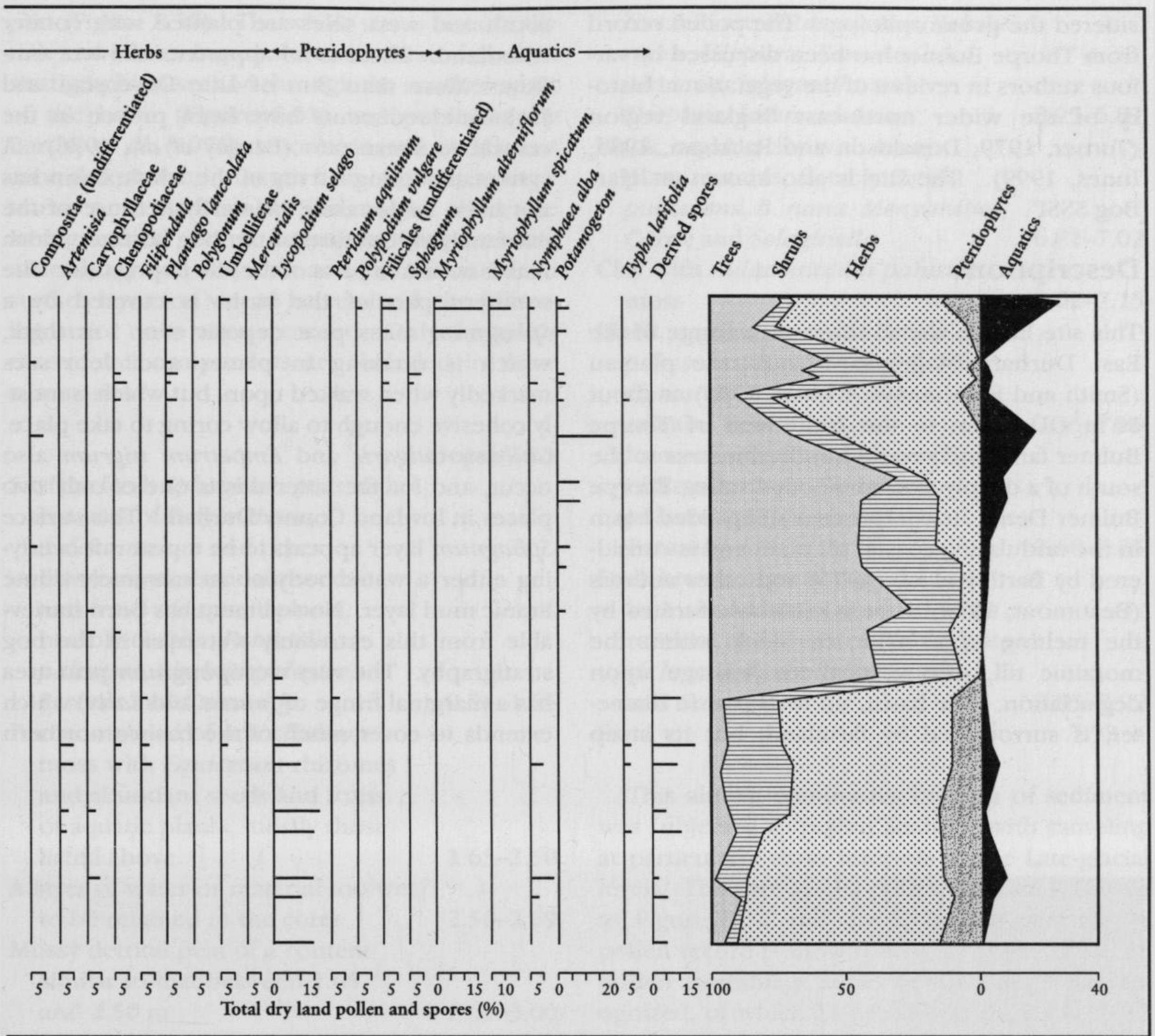
Figure 8.49 Stratigraphy, palaeomagnetic susceptibility and outline pollen analysis of core B, Gormire Lake (after Oldfield, 1999).

of apparently almost total catchment deforestation, with clay mineral inwash and the temporary replacement of trees by open ground.

Blackham *et al*. (1981) were unable to date their two clearance phases, but it is likely that they correspond with the two phases recognized by the recent much more detailed multi-proxy investigations of the lake's sediments (NERC 1999, Oldfield *et al*., 1999). In this case the earlier of the two phases is of Romano-British date and the later corresponds to medieval times and after. These recent investigations are continuing, but already have established the great value of the site and the efficacy of the wide range of integrated techniques used there. Oldfield *et al*. (1999) have proven that at Gormire changes in sediment characteristics and flux can be linked to land-use changes, that sediment source types can be identified on the basis of magnetic properties, and that organic geochemical analyses are highly consistent with the other lines of evidence. These analytical techniques have been less effective at the few other lake sites in the region and their wider application remains to be proven (Oldfield *et al*., 1999).

Conclusions

Gormire has been established as a key site for palaeoenvironmental research by the application of an integrated suite of innovative, sophisticated analytical techniques designed to charac-



terize terrestrially derived deposits in lake sediment sequences. Its probable complete and well-preserved Late-glacial and Holocene depositional record makes it an ideal site for future research.

THORPE BULMER (NZ 453 354)

J. Innes

Introduction

Thorpe Bulmer is a steep-sided depression, probably a kettlehole, in the east Durham coastal till plain. The site is important for preserving full Late Devensian and high-resolution late Holocene sedimentary records. Detailed macrofossil and palynological investigations of the deposits were published by Bartley *et al.* (1976) as part of a wider study of the vegetation history of south-east Durham. They also radiocarbon dated key late Holocene horizons from the upper part of the sequence. The solid and drift geological context of the area has been discussed by several authors (Smith and Francis, 1967; Catt, 1991a; Lunn, 1995; Smith, 1995; Teasdale and Hughes, 1999). Thomas (1999) has classified the Quaternary deposits of the area and assigned the Thorpe Bulmer sediments to the Bamburgh Formation. Stevens and Atkinson (1970) have described the area's soils and Beaumont (1970) and Evans (1999) have considered the geomorphology. The pollen record from Thorpe Bulmer has been discussed by various authors in reviews of the vegetational history of the wider north-east England region (Turner, 1979; Donaldson and Rackham, 1984; Innes, 1999). The site is also known as 'Hart Bog SSSI'.

Description

This site lies on the south-eastern fringe of the East Durham Magnesian Limestone plateau (Smith and Francis, 1967; Lunn, 1995) at about 80 m OD, 1 km to the south-west of Thorpe Bulmer farm and several hundred metres to the south of a deeply incised wooded valley, Thorpe Bulmer Dene. It comprises a steep-sided basin in the undulating coastal till plain and is considered by Bartley *et al.* (1976) and other authors (Beaumont, 1970) to be a kettlehole formed by the melting of a large ice block within the morainic till, during local ice wastage upon deglaciation. The basin, up to 100 m in diameter, is surrounded by farmland, but its steep south and west sides are planted with conifer woodland. There is no apparent inflow or outflow. More than 8 m of Late Devensian and Holocene sediments have been proven in the central southern part (Bartley *et al.*, 1976). A systematic coring survey of the whole basin has not been undertaken, primarily because of the extremely wet nature of the bog surface, which makes access to parts of the site impractical. The southern part of the basin is covered by a *Sphagnum* moss peat deposit over 1 m thick, which is quaking in places and depresses markedly when walked upon, but which is mostly cohesive enough to allow coring to take place. *Calluna vulgaris* and *Empetrum nigrum* also occur, and for the latter this is one of only two places in lowland County Durham. This surface *Sphagnum* layer appears to be a peat raft overlying either a water body or an extremely dilute limnic mud layer. No sediment has been retrievable from this extremely wet layer in the bog stratigraphy. The very wet *Sphagnum* peat area has a marginal fringe of *Juncus* and *Salix*, which extends to cover much of the basin's northern

Figure 8.50 Peat exposed at Thorpe Bulmer. (Photo: Robert van den Nordt.)

part (Figure 8.50). The whole site is botanically rich and has an important invertebrate fauna. A stratigraphical section through the southern part of the basin (Figure 8.51) was published by Bartley *et al.* (1976), who recovered a sediment core almost 8 m in length from a point on their transect not far from the site margin. Here the stratigraphical hiatus caused by the probable water body was very thin and, beneath a 50 cm surface layer too wet to be retained, the following major units were recorded. Only the major macrofossil components are included in these summaries.

	Depth (m)
Surface fresh *Sphagnum* peat with remains of *Polytrichum*, *Aulacomnium* and other mosses and abundant seeds and fruits of aquatic plants, notably *Caltha*, *Potentilla palustris*, *Cirsium palustre*, *Lychnis* and *Carex* spp.	0.51–1.65
Peat composed of *Acrocladium* moss with *Equisetum* rhizomes and abundant seeds and fruits of aquatic plants, mostly those listed above	1.65–2.50
A layer of water or material too wet to be retained in the corer	2.50–2.69
Mossy detrital peat of a content similar to that between 1.65 and 2.50 m	2.69–3.00
Fine wet clay with abundant organic material, including many seeds and fruits of aquatic plants. *Potamogeton*, *Carex*, *Potentilla* and *Acrocladium* important	3.00–3.33
Moss peat dominated by *Hylocomium*, with other mosses and with much *Sphagnum* and *Eriophorum vaginatum* remains nearer the centre of the basin	3.33–3.47
Dark brown detritus with abundant bark, twigs and other remains of *Betula pubescens*, *B. pendula* and *Salix*, with rarer *Quercus* and *Populus*. Abundant remains of a wide range of aquatic plants	3.47–5.72
As above, but with an increasing clay content and remains of *Chara*	5.72–5.95
Grey clay with occasional *Chara* oospores and many plant remains. Notable are *Potamogeton*, *Betula nana*, *Hippuris*, *Ranunculus*, *Armeria*, *Selaginella* and *Drepanocladus aduncus*	5.95–6.81
Black, organic clay with many plant remains, notably *Betula pubescens*, *B. nana*, *Menyanthes*, *Carex* and *Selaginella*	6.81–7.02
Clay with *Aulacomnium palustre* moss	7.02–7.15
Black, organic clay with *Betula pubescens*, *B. nana* and *Carex* cf. *rostrata*. Only *B. nana* remains below 7.40 m	7.15–7.60
Calcareous grey clay with remains of *Betula nana*, *Eleocharis palustris*, *Hippuris* and *Potamogeton*	7.60–7.85
Dark brown organic material with abundant fragmented moss remains and seeds of *Potamogeton* cf. *lucens*, *Lychnis floscuculi* and *Carex* sp.	7.85–7.92
Sand	7.92 +

This almost continuous column of sediment was subjected to pollen analysis, with sampling at particularly close intervals in the Late-glacial levels. The Late-glacial pollen diagram is shown as Figure 8.52 and the Holocene part of the pollen record is shown as Figure 8.53. In all, 16 pollen assemblage zones or subzones were recognized, of which 11 are shown on Figure 8.52 and six on Figure 8.53, with zone TBVI being common to both. The characteristic pollen taxa in these 16 phases of vegetation history are summarized in Table 8.11, with phase TBI at the base of the succession.

Three horizons in subzone TBIXa were radiocarbon dated. The start of the subzone was dated to 2064 ± 60 years BP (SRR-404), the end to 852 ± 60 years BP (SRR-405) and the mid-subzone peak of *Cannabis* pollen to 1730 ± 120 years BP (GaK-3713).

Interpretation

Much of the importance of this site lies in the Late-glacial sediments, which have been subject to very high-resolution pollen analysis and so provide a very sensitive climatic and environmental signal. Their value is diminished, however, without supporting radiocarbon dating so that regional correlation with the established

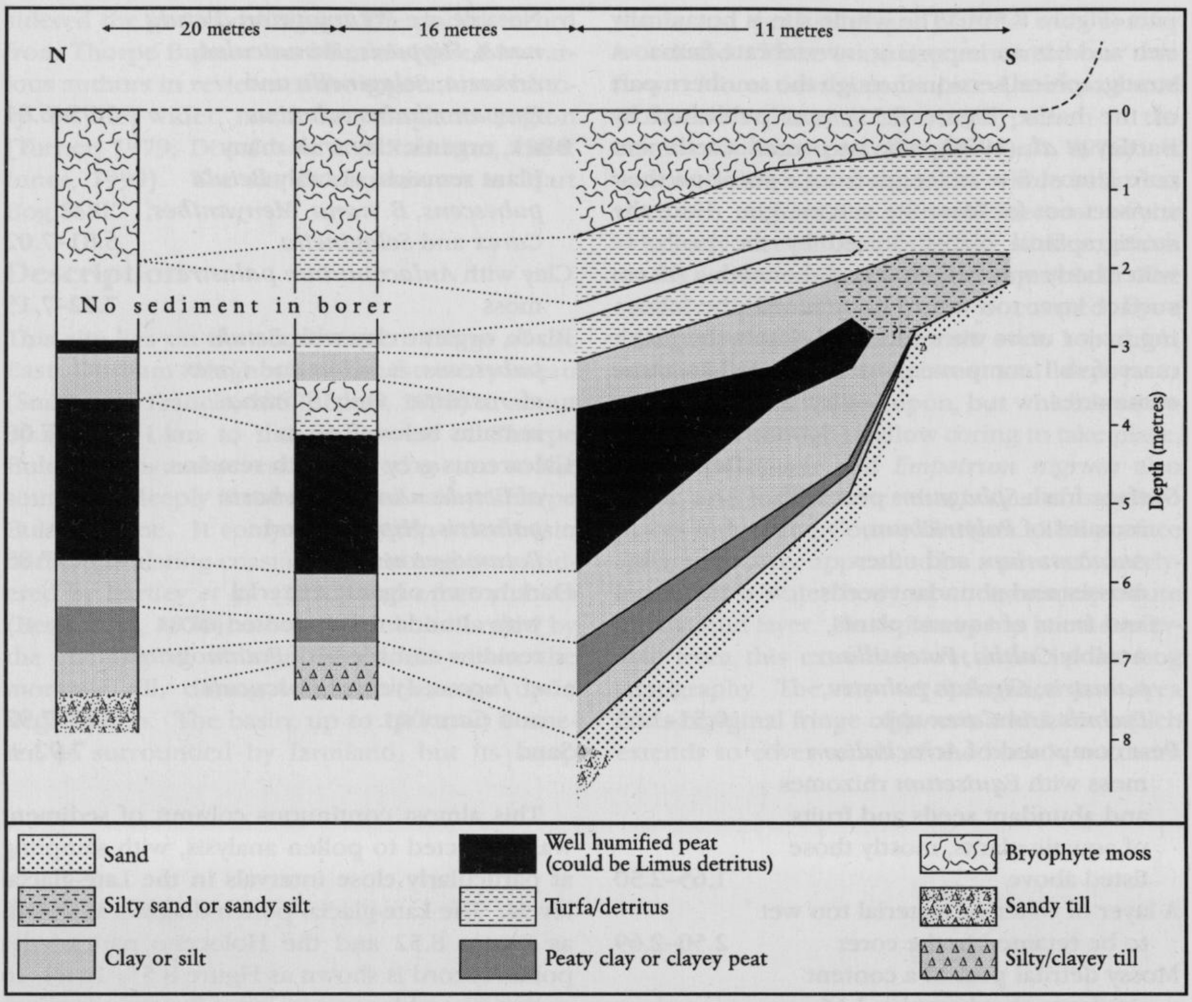


Figure 8.51 Stratigraphical section through Thorpe Bulmer (Bartley *et al.*, 1976).

pattern of Late-glacial events in other pollen diagrams is uncertain. Bartley *et al.* (1976) interpret pollen zones TBII to TBIIIc to be of Late-glacial interstadial age because of greatly increased levels of *Juniperus* and then tree *Betula* pollen and because they correspond with the highly organic layers within the pre-Holocene lithology. The lowermost clay unit and pollen zone TBI therefore were deposited during the cold climatic phase pre-dating the main Late-glacial interstadial. This generally is supported by the zone TBI vegetation of pioneer, cold-tolerant, open habitat taxa, particularly Cyperaceae and Gramineae. Low frequencies of *Betula* pollen and no macrofossil remains suggest no local tree birch growth, although *B. nana* and *Salix* were likely to be present around the site. There are complicating factors in the lower part of this basal zone, including thin, highly organic lenses, a peak of *Juniperus* at the base of the pollen record and a smaller peak of the thermophilous *Corylus* in the same levels. Intrusive reworked material may be a factor and interpretation of the lowest part of the profile requires caution. Early Late-glacial *Corylus* records are not uncommon in the region (e.g. Blackburn, 1952; Bartley, 1962) and may result from long-distance transport of the grains.

The first phase of the main Late-glacial interstadial at Thorpe Bulmer, zone TBII, is characterized by the dominance of *Juniperus* to such a degree that juniper scrub must have been abundant locally. This very high peak of *Juniperus* pollen immediately prior to the first Late-glacial

Table 8.11 Characteristic pollen taxa of the 16 pollen assemblage zones and subzones from Thorpe Bulmer (Bartley *et al.*, 1976).

Phase	Major taxa	Lesser taxa
TBIXc	Gramineae, Cyperaceae	*Plantago lanceolata*, Ericaceae
TBIXb	Gramineae, *Taraxacum*, *Plantago lanceolata*, *P. major-media*	*Alnus*, Cyperaceae
TBIXa	*Alnus*, Gramineae, *Cannabis*	*Plantago lanceolata*
TBVIII	*Corylus*	*Quercus*, *Alnus*
TBVII	*Corylus*	*Betula*
TBVI	*Betula*	*Salix*, *Filipendula*
TBVb	*Betula*, *Filipendula*, *Salix*	*Juniperus*, *Empetrum*
TBVa	Gramineae, *Empetrum*	Cyperaceae, *Betula*, *Galium*
TBIVc	Cyperaceae, *Thalictrum*	Gramineae, *Ranunculus*, *Artemisia*
TBIVb	Cyperaceae, Gramineae	*Thalictrum*, *Artemisia*, Caryophyllaceae
TBIVa	Cyperaceae, Gramineae	*Rumex*, Caryophyllaceae
TBIIIc	*Betula*, *Filipendula*	Gramineae, Cyperaceae, *Empetrum*
TBIIIb	Gramineae, Cyperaceae	*Betula*, *Juniperus*, *Filipendula*
TBIIIa	*Betula*, *Empetrum*	*Juniperus*, *Filipendula*
TBII	*Juniperus*	*Helianthemum*, Cyperaceae, Gramineae
TBI	Cyperaceae, Gramineae	*Betula nana*, *Salix*, *Juniperus*, Ruderals

Betula expansion occurs at other sites in the Tees lowlands area, such as Romaldkirk (Bellamy *et al.*, 1966). A good example is Seamer Carrs (Jones, 1976a) where these pollen changes and the shift to interstadial environments took place substantially before 13 042 ± 140 years BP. Elsewhere, as at Kildale Hall (Jones, 1977b) and Neasham (Blackburn, 1952), this *Juniperus* peak is not apparent. Local factors may be decisive. At Thorpe Bulmer and other sites such as Neasham, the continued substantial presence of *Helianthemum* and other herb pollen during the *Juniperus* phase suggests that juniper cover was never so dense as to suppress growth of lower stature taxa. High *Juniperus* pollen frequencies are a feature of Late-glacial interstadial diagrams to the north of Thorpe Bulmer, although their timing varies. These occur either as percentages continually in excess of that for *Betula*, as at Cranberry Bog in the Durham lowlands (Turner and Kershaw, 1973) or as a very high peak as at Bradford Kames in the Northumberland coastal plain (Bartley, 1966). In the Thorpe Bulmer area these expansions of *Juniperus* are probably successional towards developing *Betula* woodland dominance, but elsewhere may also be linked to climatic fluctuation during the interstadial. There is good evidence for such climatic change at Thorpe Bulmer, where *Betula* pollen abundance indicates birch woodland establishment in zone TBIIIa, with increases in taxa such as *Empetrum* and *Filipendula*. In zone TBIIIb the decline in *Betula* frequencies, which occurs with increases in Gramineae, Cyperaceae and *Juniperus* and peaks for open-ground herbs including *Rumex*, *Ranunculus*, *Helianthemum* and *Plantago major-media*, points to climatic deterioration. This change is reversed again in zone TBIIIc, with *Betula* and *Filipendula* restored to abundance. This very clear interstadial double *Betula* peak has been observed in the north and east Yorkshire area, as at Tadcaster (Bartley, 1962), Kildale Hall (Jones, 1977b), Seamer Carrs (Jones, 1976a) and The Bog, Roos (Beckett, 1981). The vegetation reversion is most likely the result of climatic oscillation. Thorpe Bulmer is unlike the other sites, however, in that the earlier *Betula* peak is much the greater of the two. Usually the earlier *Betula* peak is a weaker signal prior to the main interstadial *Betula* forest phase, as at The Bog, Roos, where it is dated to 13 045 ± 270 years BP (Beckett, 1981). Local factors at Thorpe Bulmer may account for the lower frequencies of the later *Betula* peak, such as high Gramineae and Cyperaceae pollen influx from lake-fringing wetland vegetation. This atypical *Betula* pollen curve and the lack of radiocarbon dating control for the interstadial makes comparison with other regional vegetation histories uncertain, despite the high-resolution pollen record. Consistently lower inter-

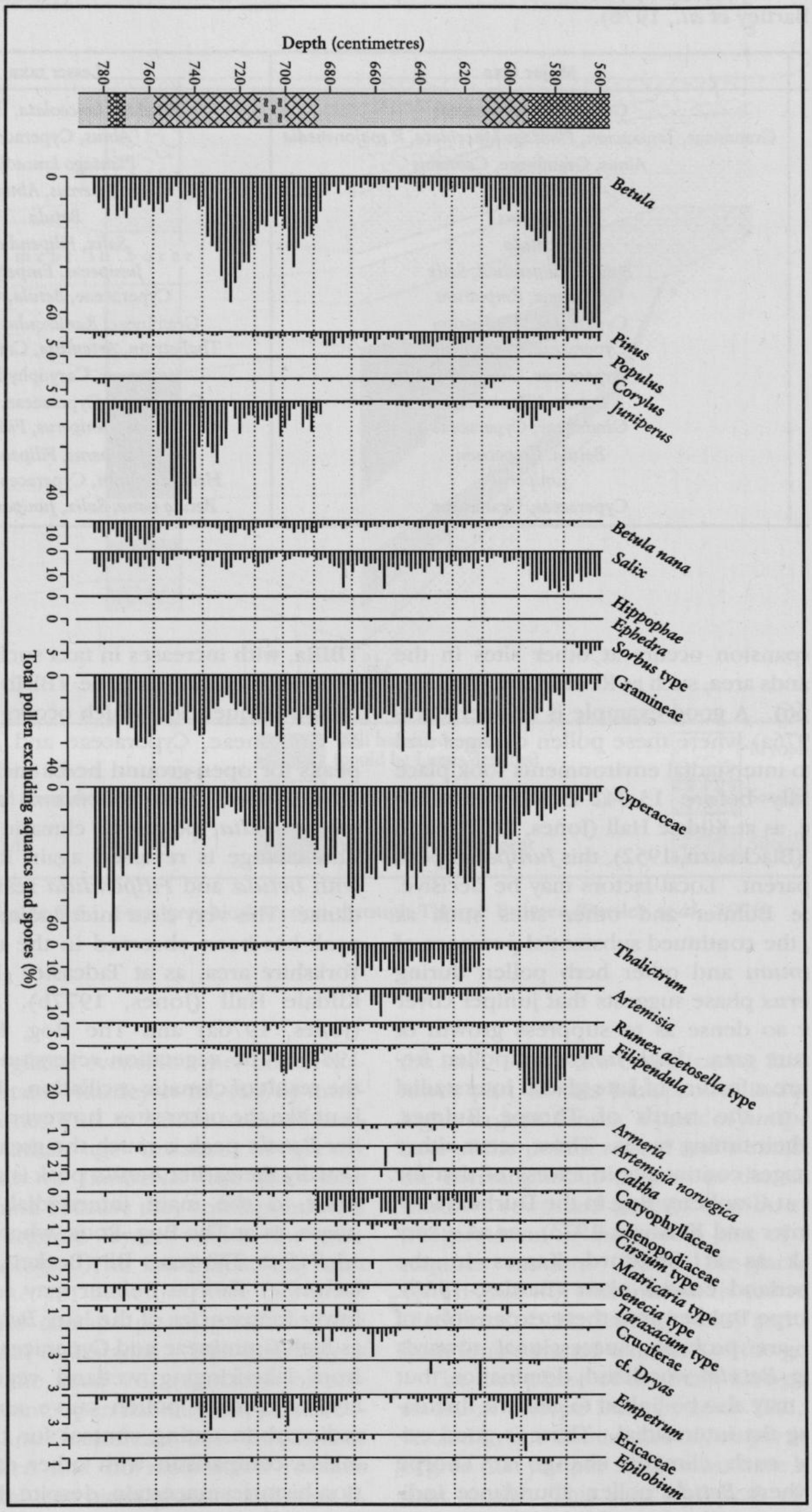


Figure 8.52 Late-glacial pollen diagram from Thorpe Bulmer. Pollen values expressed as percentage of total pollen excluding pollen of aquatics and spores (Bartley *et al.*, 1976) (continued opposite). See Figure 8.1 for key to the stratigraphical log

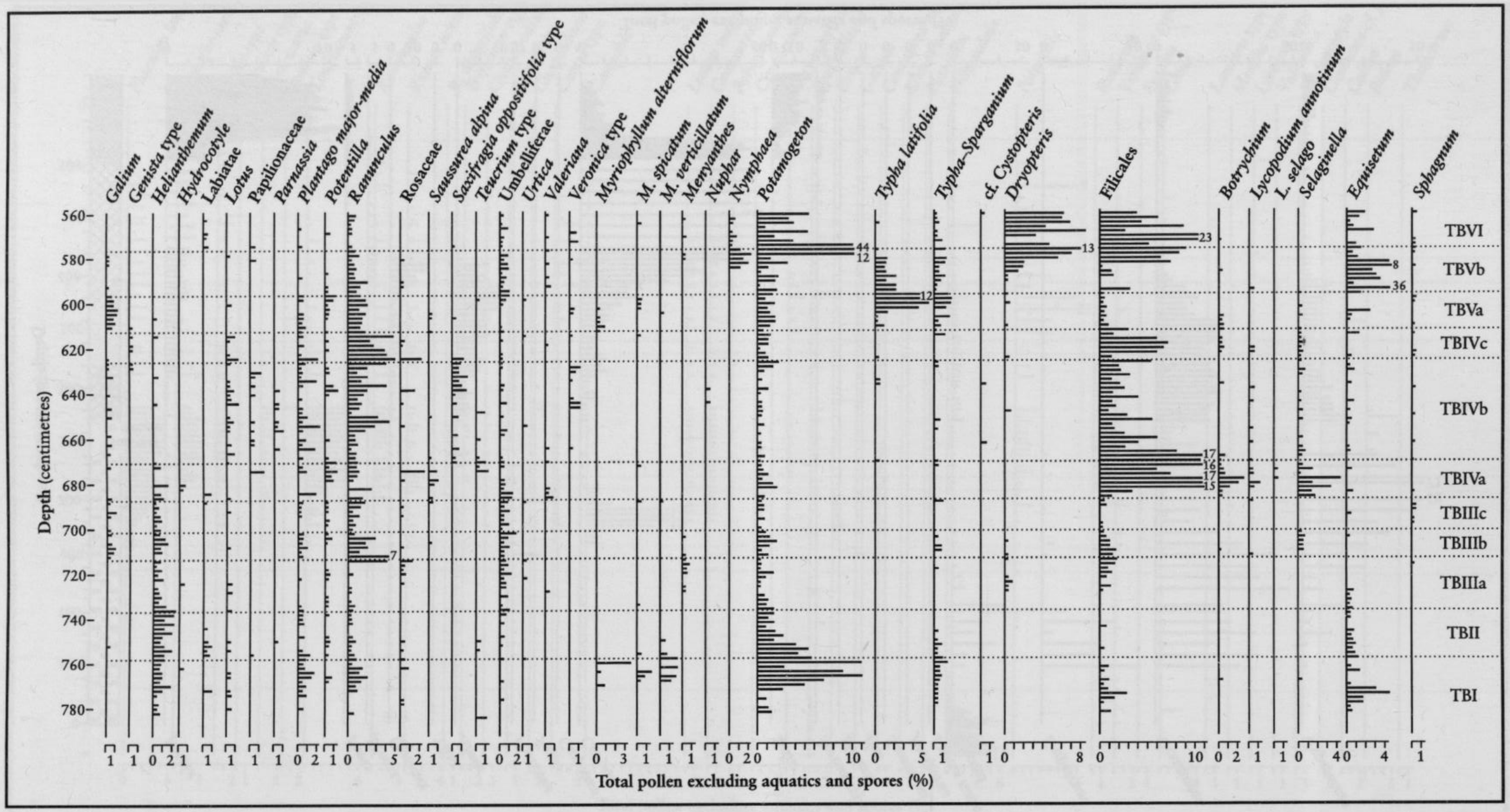


Figure 8.52 (*contd*) Late-glacial pollen diagram from Thorpe Bulmer. Pollen values expressed as percentage of total pollen excluding pollen of aquatics and spores (Bartley *et al.*, 1976).

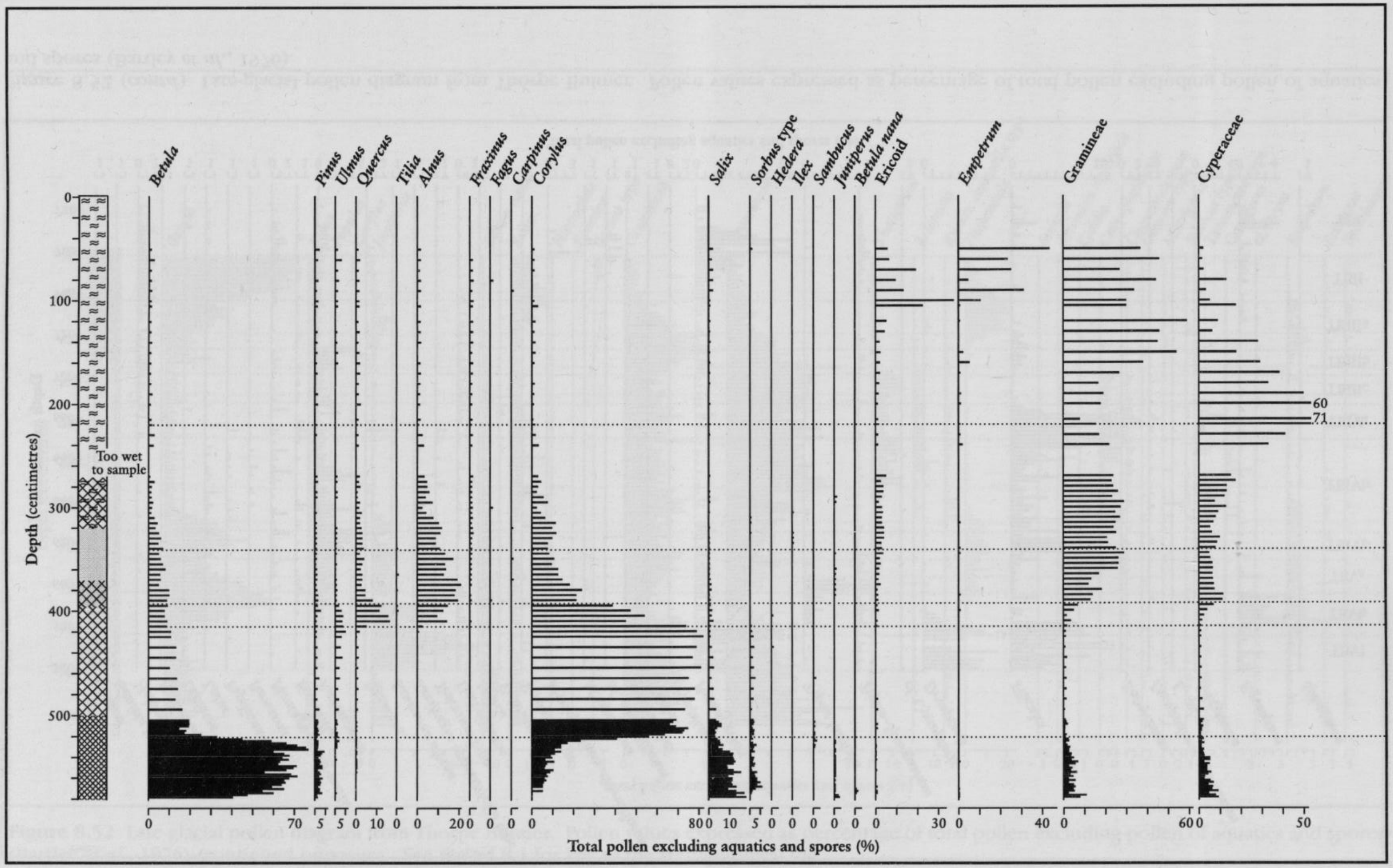


Figure 8.53 Holocene pollen diagram from Thorpe Bulmer. Pollen values expressed as percentages of total pollen excluding pollen of aquatics and spores (Bartley *et al.*, 1976) (continued opposite and overleaf). See Figure 8.1 for key to the stratigraphical log.

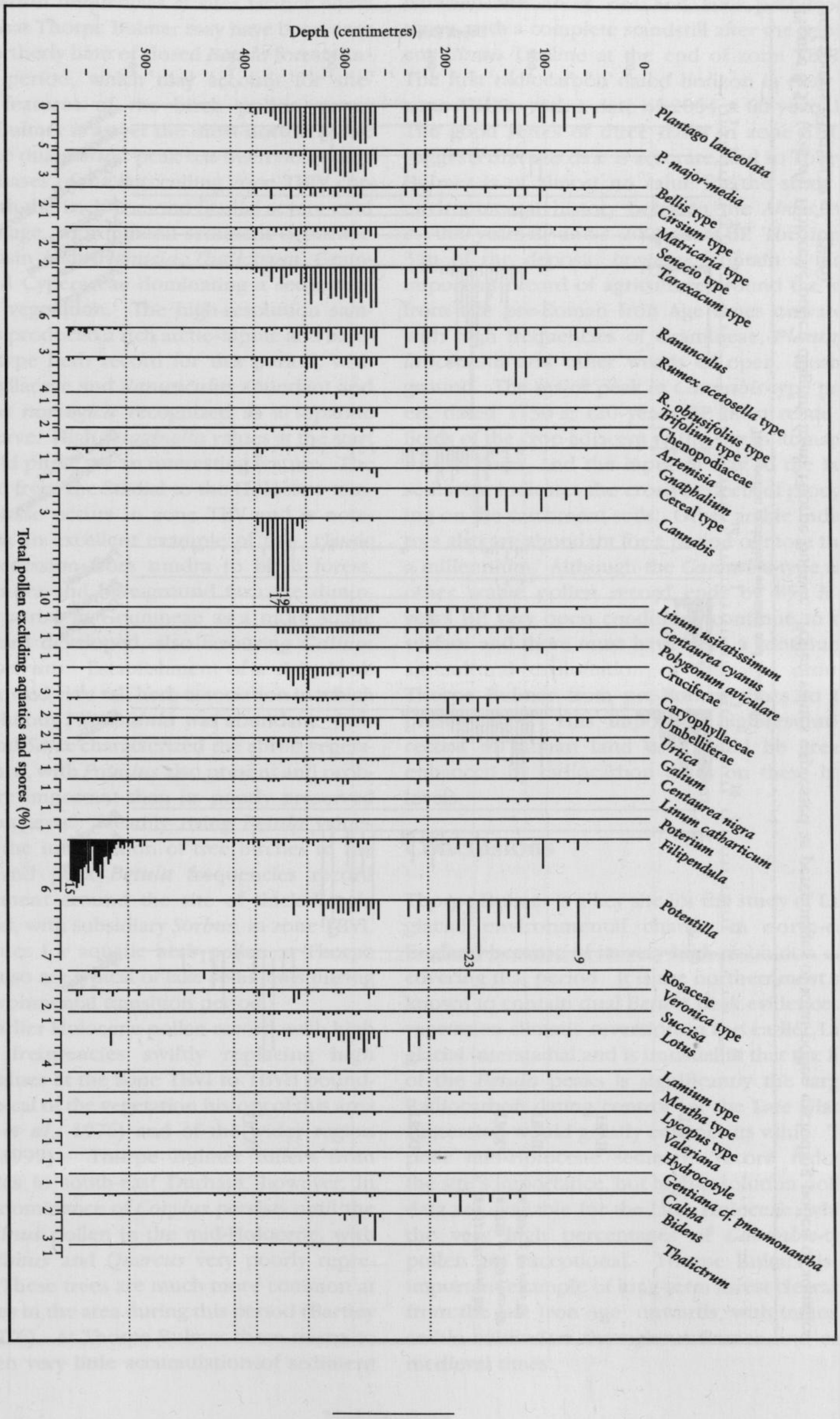


Figure 8.53 *(contd)* Holocene pollen diagram from Thorpe Bulmer. Pollen values expressed as percentages of total pollen excluding pollen of aquatics and spores (Barley *et al.* 1976).

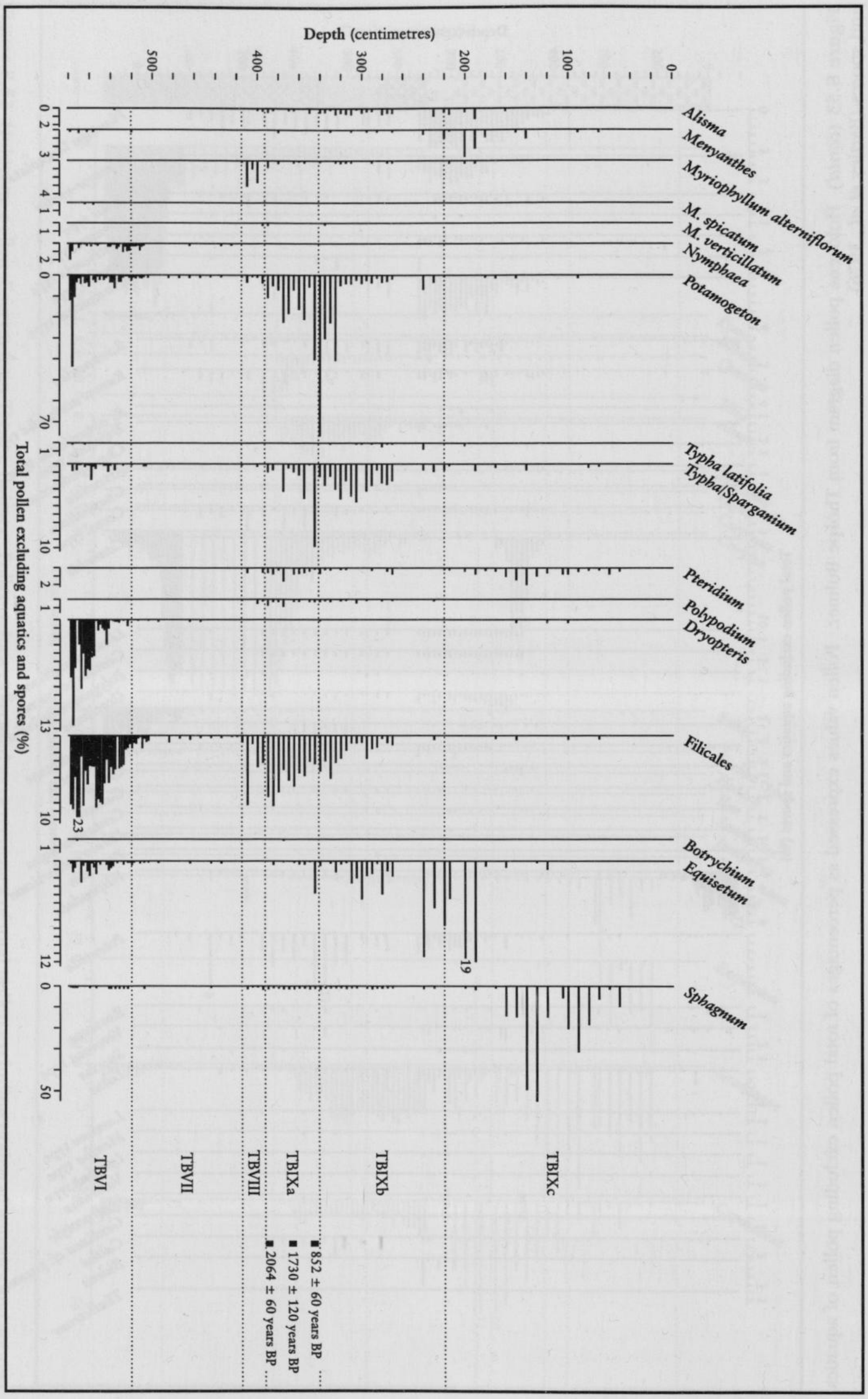


Figure 8.53 (*contd*) Holocene pollen diagram from Thorpe Bulmer. Pollen values expressed as percentages of total pollen excluding pollen of aquatics and spores (Bartley *et al.*, 1976).

stadial *Betula* frequencies at sites farther north suggest that Thorpe Bulmer may have been near to the northerly limit of closed *Betula* forest during this period, which may account for site-specific features of the birch pollen curve. Thorpe Bulmer is as yet the most northerly site where the dual *Betula* peak has been observed.

The phases of the succeeding zone TBIV correlate with the Loch Lomond Stadial severe cold climate stage, with mineral sediment deposited in the basin and *Artemisia*, *Thalictrum*, Gramineae and Cyperaceae dominating a sedge-tundra type vegetation. The high-resolution sampling has produced a rich arctic–alpine and bare-ground type herb record for this period, with Caryophyllaceae and *Ranunculus* abundant and *Artemisia norvegica* recognized as a separate pollen curve. High *Selaginella* values at the start of the cold phase are an interesting feature. The transition from the Stadial to the Holocene temperate phase occurs in zone TBV and is noteworthy as an excellent example of this classic plant succession from tundra to birch forest. Sedge-tundra and bare-ground taxa are diminished by a rise in Gramineae as a more stable grass sward developed, also favouring *Galium* and *Empetrum*. Establishment of a rich shrub flora followed via a tall herb association in which thermophilous *Filipendula* was abundant. *Juniperus* and *Salix* characterized the shrub vegetation in turn, with *Populus* also present and probably more important than its poorly preserved pollen suggests. Steadily rising *Betula* values indicate the immigration of tree birches to the vicinity and peak *Betula* frequencies record establishment around the site of thick *Betula* woodland, with subsidiary *Sorbus*, in zone TBVI. High values for aquatic herb pollen at Thorpe Bulmer also are typical of lake sediments during this environmental transition period.

The earlier Holocene pollen record, with high *Corylus* frequencies swiftly replacing high *Betula* values at the zone TBVI to TBVII boundary, is typical of the vegetation history of this area (Bartley *et al*., 1976) and of the wider region (Innes, 1999). Thorpe Bulmer differs from other sites in south-east Durham, however, in that the dominance of *Corylus* persists until the rise of *Alnus* pollen in the mid-Holocene, with *Pinus*, *Ulmus* and *Quercus* very poorly represented. These trees are much more common at other sites in the area during this period (Bartley *et al*., 1976). At Thorpe Bulmer there seems to have been very little accumulation of sediment between the *Alnus* rise and later prehistoric times, with a complete standstill after the apparent *Ulmus* Decline at the end of zone TBVIII. The first radiocarbon dated horizon is early in zone TBIXa, with a date of 2064 ± 60 years BP. The good series of three dates in zone TBIXa suggests that this date is accurate, and so Thorpe Bulmer is of almost no value for the study of environmental history between the *Alnus* rise *c*. 7000 years BP and *c*. 2000 years BP. The upper 3 m of the deposit, however, contain a most important record of agriculture around the site from late pre-Roman Iron Age times onwards, with high frequencies of Gramineae, *Plantago lanceolata* and other weeds of open, cleared ground. The major peak in *Cannabis*-type pollen, dated 1730 ± 120 years BP, must relate to fields of the crop adjacent to the site in Romano-British times, and the input of clay to the lake sediment indicates the erosive effects of ploughing on the catchment soils. Other arable indicators also are abundant for a period of more than a millennium. Although the *Cannabis*-type and other arable pollen record ends by 852 ± 60 years BP, very open conditions continue to the surface and there must have been a continuous agricultural exploitation of the area around Thorpe Bulmer from pre-Roman times to the present day. This important high-resolution record of human land use would be greatly enhanced by radiocarbon dates on these later levels.

Conclusions

Thorpe Bulmer is a key site for the study of Late-glacial environmental change in north-east England because of its very high-resolution data covering that period. It is the northernmost site known to contain dual *Betula* peak evidence for vegetation-climatic reversion in the earlier Late-glacial interstadial and is unusual in that the first of the *Betula* peaks is significantly the larger. Radiocarbon dating control for the Late Glacial succession would greatly enhance its value. The poor mid-Holocene sediment record reduces the site's importance, but high-resolution pollen data are available for the later Holocene, where the very high percentages of *Cannabis*-type pollen are exceptional. Thorpe Bulmer is an important example of long-term forest clearance from the late Iron Age onwards, with intensive arable cultivation throughout Roman and early medieval times.

LOW HAUXLEY (NU 284 019)

J. Innes

Introduction

The site of Low Hauxley in south Northumberland is important as one of the most extensive coastal exposures of mid-Holocene deposits in north-east England, with complex sequences of fluvial, peat and dune sand sediments, associated with multi-period archaeological material. Lithology, pollen stratigraphy and radiocarbon dating have been published by Frank (1982) and Innes and Frank (1988), and Farrimond and Flanagan (1996) have reported lipid geochemistry results. Data from Low Hauxley have been valuable in reconstructions of sand dune initiation and development (Orford *et al.*, 2000) and sea-level history and coastal change (Plater and Shennan, 1992; Horton *et al.*, 1999a, b; Shennan *et al.*, 2000a, b). Orford *et al.* (2000) have completed a borehole and radiocarbon dating programme in the Hauxley dune system. Archaeological excavations have been carried out by Bonsall (1984), with associated palaeoenvironmental research by Tipping (unpublished) and further radiocarbon dating (Hardiman *et al.*, 1992). Lancaster University Archaeological Unit have undertaken an archaeological excavation and borehole survey of the sand dune cordon, underlying sediments and beach exposures, which included several palaeoenvironmental reports (Lancaster University Archaeological Unit, 1995). The site is within the Low Hauxley Shore SSSI.

Description

Low Hauxley lies at the northern end of Druridge Bay and comprises a variety of mid- to late-Holocene sediment successions preserved in the coastal zone, either exposed in low cliffs at the head of the beach or concealed beneath an overburden of later Holocene sand dunes (Figure 8.54). Cordons of blown dune sand are characteristic of much of the Northumberland coast, usually along shallow embayments between rock headlands; the 9 km stretch of dunes at Druridge Bay is one of the longest (Orford *et al.*, 2000). The sand dunes are draped over the rock outcrops and till deposits of the last glaciation (Taylor *et al.*, 1971; Lunn, 1995). Although the coastal embayments are choked with sand, the dune cordon itself tends to be thin, usually much less than 300 m in width, with only the most seaward ridge of any real height and the rest comprising low ridges, or a thin dune apron. Carboniferous Middle Coal Measures underlie Druridge Bay (Taylor *et al.*, 1971) and opencast mining behind the dunes has further narrowed the sand body and created a lagoon now managed as a nature reserve. Orford *et al.* (2000) note that deep, unvegetated blowouts open to the beach indicate that the dune system at Druridge Bay is undergoing active reworking. The deposits at Low Hauxley are sheltered to the south of the northern rock headland of Druridge Bay, although peats are exposed in the smaller embayment to the north of this headland, both under the dunes and low on the beach. These have been described by Shennan *et al.* (2000a). Intertidal peats analogous to these and the Low Hauxley sequence are exposed at the southern end of Druridge Bay near Cresswell (Horton *et al.*, 1999a, b) and the whole of the Druridge Bay area is a key location for the study of Holocene environmental history.

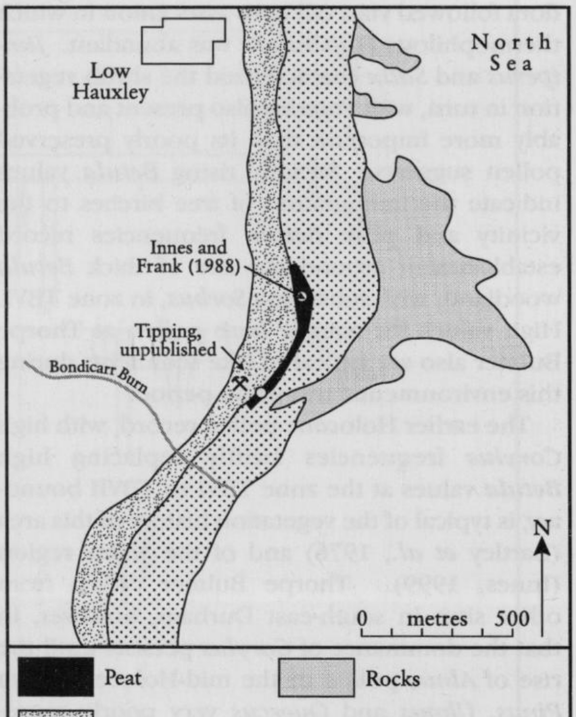


Figure 8.54 Location of the site at Low Hauxley.

The main exposure at Low Hauxley (Figure 8.55) is a thick peat with extensive wood remains, which rests upon till and lies beneath the dune overburden and is exposed in low cliffs at the head of the beach. The peat is about 1 m thick in the northern part of the exposure, where tree stumps protrude from it at the base and in mid-section and branches and other wood remains are common. The peat attenuates rapidly to the south until it becomes increasingly sandy and feathers out against the rising till surface. It has been removed by erosion to the north. The till generally is red, but is heavily weathered to a blue colour in its upper levels where exposed. Sandy, possible fossil-soil horizons can be seen in places at the till surface. The till rests on a probable interglacial wave-cut rock platform, which is covered by modern beach sand (Lunn, 1995). It is this till, peat and dune sand section, represented schematically in Figure 8.55, that has been described and analysed (Frank, 1982; Innes and Frank, 1988; Farrimond and Flanagan, 1996) and used to illustrate the area's coastal evolution (Gehrels and Innes, 1995; Innes *et al.*, 1997) and sea-level history (Plater and Shennan, 1992). The whole exposure is at present undergoing substantial marine erosion.

The base and top of the thick peat bed have been radiocarbon dated to 4720 ± 50 years BP (SRR-1421) and 2810 ± 40 years BP (SRR-1420) respectively (Innes and Frank, 1988), and a tree stump rooted in the till was dated 4890 ± 50 years BP (SRR-1422). A date of 980 ± 50 years BP (SRR-1583) also was obtained on marine shells occurring as a lag deposit in a dune slack horizon almost 1 m above the peat bed. Apart from the wood remains, the peat bed is well humified and contains very few macrofossils. Loss-on-ignition measurements by Frank (1982) showed the peat to contain little inorganic material, except in the upper 20 cm, where an increasing sand fraction occurred prior to the deposition of the blown-sand body. Pollen analysis of the peat (Figure 8.56) revealed a later Holocene assemblage post-dating the Elm Decline, which supported the radiocarbon chronology (Innes and Frank, 1988). Four local pollen-assemblage zones were recognized: LH-1 (*Alnus–Quercus*), LH-2 (*Betula–Alnus*), LH-3 (*Quercus–Alnus*) and LH-4 (*Alnus–Quercus–Calluna*). Pollen concentration was low in zone LH-3, high in zones LH-2 and LH-4, and very high in zone LH-1. Two phases were noted where a range of

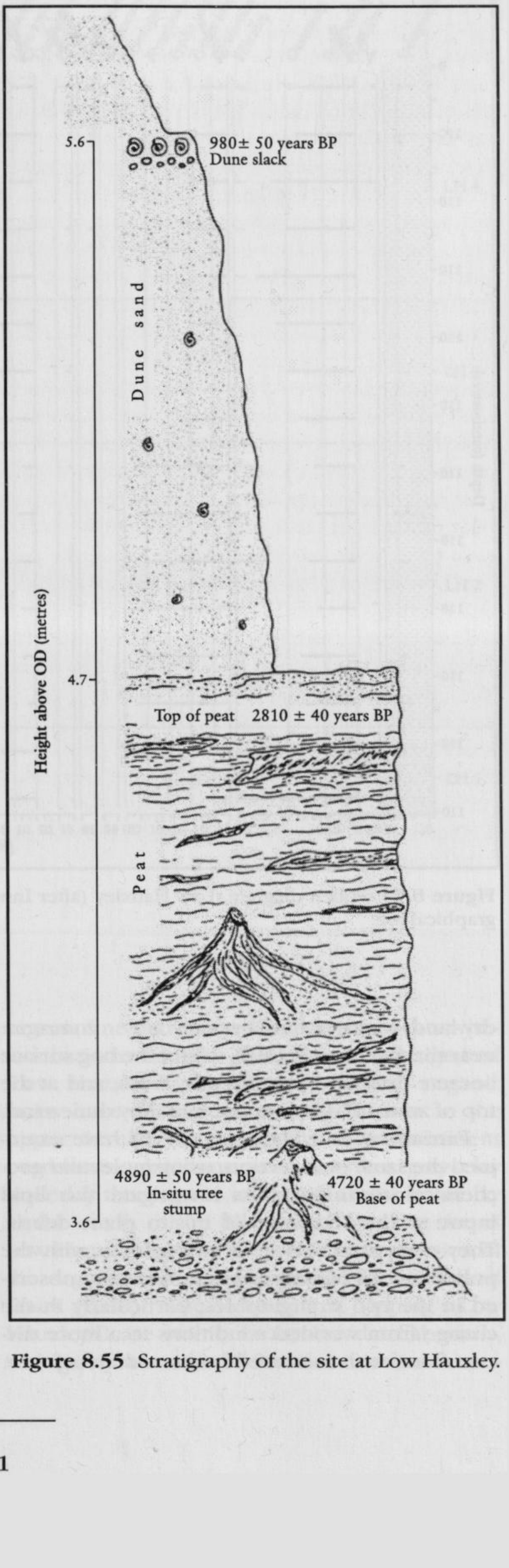


Figure 8.55 Stratigraphy of the site at Low Hauxley.

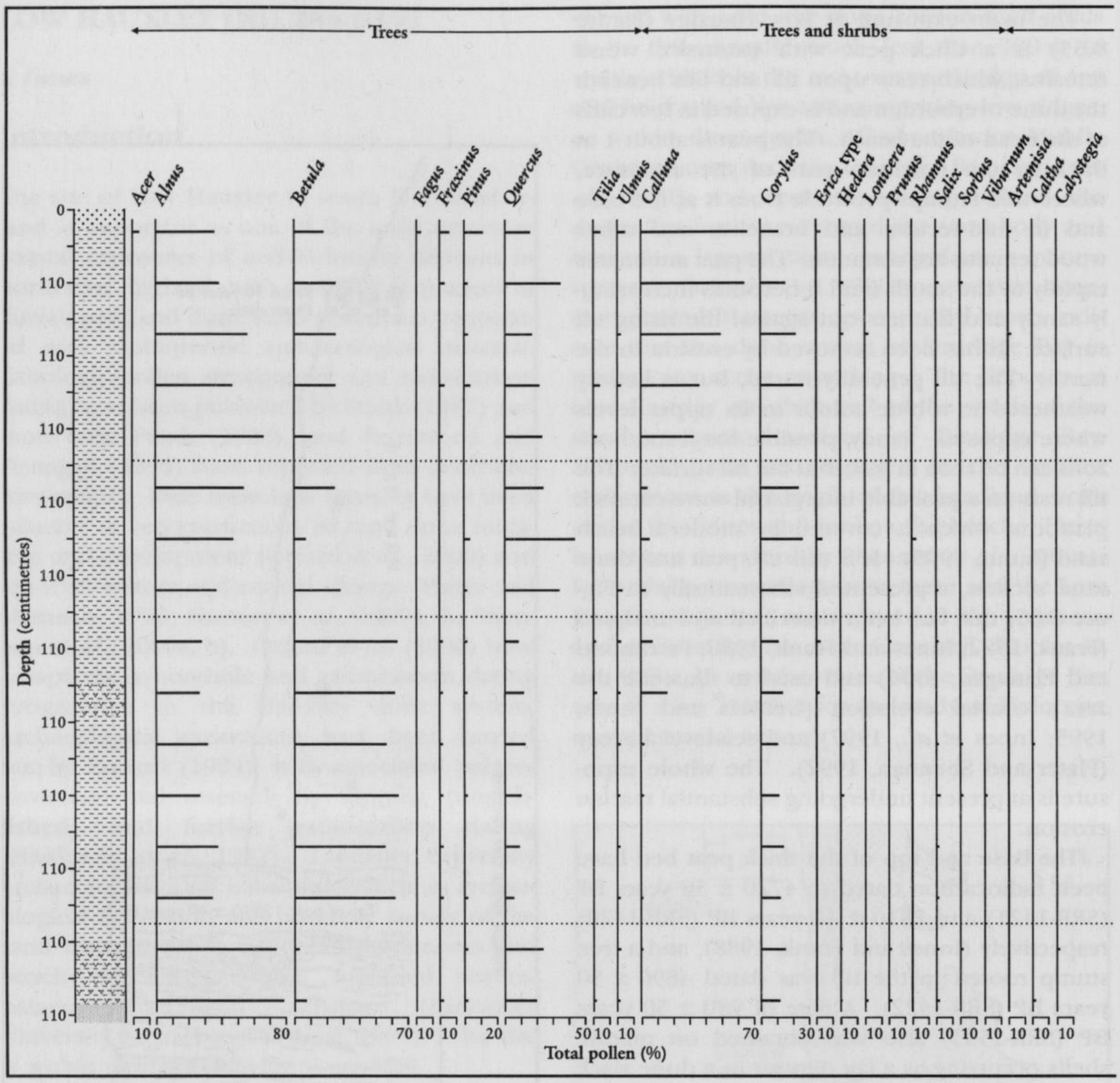
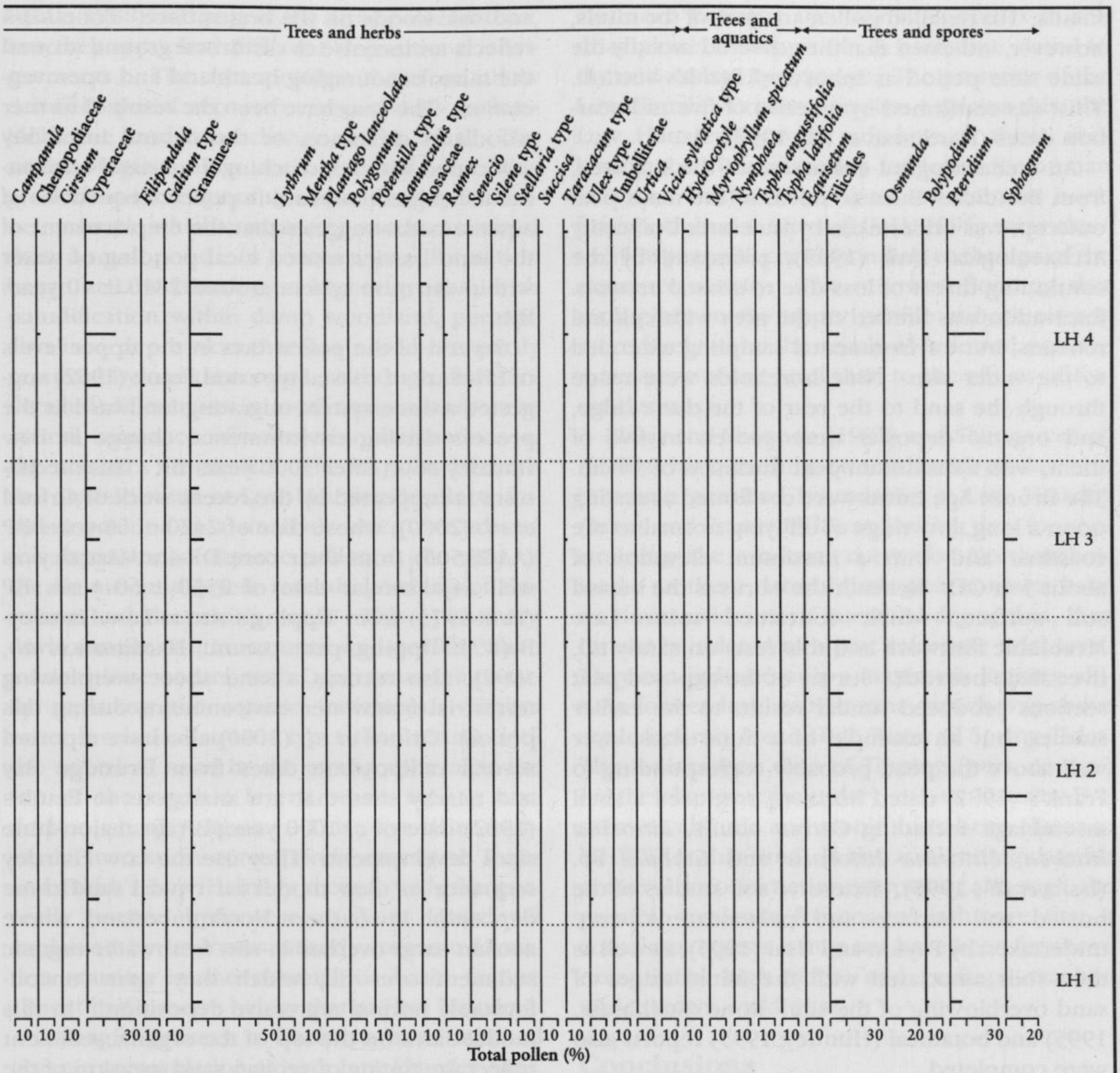


Figure 8.56 Pollen diagram, Low Hauxley (after Innes and Frank, 1988). See Figure 8.1 for key to the stratigraphical log.

dry-land weed pollen became very common: near the top of zone LH-2, where the bog surface became drier and colonized by trees, and at the top of zone LH-4 before the burial by dune sand.

Farrimond and Flanagan (1996) have examined the same peat section, using molecular geochemical techniques, to investigate the lipid input to the sediment of in-situ plant debris. They compared this lipid stratigraphy with the pollen record. Comparable trends were observed in the two stratigraphies, particularly in the change from wooded conditions to a more diverse flora, with increased herbs and *Sphagnum*. With the exception of the *Sphagnum* indicators, the lipid evidence did not correlate well in detail with the pollen evidence of vegetation change. The wider source area of the pollen assemblages compared with the in-situ origin of the plant remains generating the lipid data probably is responsible for the difference. The authors concluded that the two types of data were complementary and could be used together in environmental reconstruction.

Low Hauxley has been the scene of two phases of archaeological excavation, which have produced important geological data. Bonsall



(1984) reported the exposure of two Bronze Age burial cairns resting upon the pre-dune land surface. The buried soil beneath one of the cairns also contained a Late Mesolithic midden deposit, with carbonized plant material, shells, bones and fish remains. This multi-period site and the threat of its loss by erosion prompted detailed study of its associated natural sediments to place it within a secure palaeoenvironmental context (R.M. Tipping, pers. comm.). The archaeological sites were located upon a mound of till and the gleyed palaeosol that underlies the Bronze age cairns was seen to pass northwards into organic fine detritus muds and moss peats within a basin in the till. This basin was designated 'Low Hauxley-B' by Tipping. These organic facies have an increasing sand fraction in the upper levels, which signals their imminent overblowing by dune sand. A fine grey clay underlay the organic muds. A similar basin lay to the south of the till mound, but the site of Low Hauxley-B is the southernmost extension of the organic sediments examined by Innes and Frank (1988). The sediments at Tipping's Low Hauxley-B are mainly lacustrine detrital muds, which contain no wood remains or other macrofossils. His detailed pollen analyses of the muds, however, indicated that they covered broadly the same time period as Innes and Frank's section. This was confirmed by a series of five radiocarbon dates (Hardiman *et al.*, 1992).

An archaeological evaluation of the site area, from Bondicarr Burn to north of the main peat outcrop, was undertaken by Lancaster University Archaeological Unit (1995), prompted by the continuing threat of loss due to coastal erosion. Excavation was limited to the area with cultural remains, but environmental sampling extended to the wider site. Nine boreholes were made through the sand to the rear of the dune ridge, and organic deposits were proven in five of them, with a maximum peat thickness of 50 cm. The Bronze Age cairns were confirmed as resting upon a long, low ridge of till lying normal to the coastline and with a maximum elevation of about 5 m OD. Beneath the cairns is the buried soil surface, which contained some Late Mesolithic flintwork and this rests on sandy till, over shale bedrock. Survey of the exposed peat sections produced similar results to the earlier studies, but an example of a dune slack layer well above the peat, probably corresponding to Frank's (1982) dated horizon, produced a shell assemblage including *Ostrea edulis*, *Littorina littorea*, *Littorina littoralis* and *Gibbula* sp. (Issitt *et al.*, 1995). Extensive soil studies of the buried soil surface and palaeocatena were undertaken by Payton and Usai (1995), as well as the soils associated with the initial stages of sand overblowing of the site. Bone (Stallibrass, 1995) and botanical (Huntley, 1995) reports also were completed.

Interpretation

The pollen stratigraphy of Innes and Frank (1988) can be interpreted as representing the drowning of terrestrial woodland by alder swamp-carr, with a very slow rate of peat accumulation in LH-1 and a natural succession to drier fen-carr environments in LH-2. *Betula* and *Quercus* roots and stumps in the peat in upper zone LH-2 indicate a drier bog surface and the herbaceous weed pollen, including *Plantago lanceolata*, may well indicate local prehistoric woodland clearance. Catchment clearance may have increased runoff to the mire, as zone LH-3 was a period of much wetter conditions and rapid bog growth, causing the demise of birch and oak woods on the bog surface. Zone LH-4 reflects an increase in disturbed ground around the mire, encouraging heathland and open vegetation. This may have been the result of further woodland clearance, or the natural instability associated with approaching dune-sand environments. A great increase in pollen frequencies of aquatic herbs suggests that the emplacement of the sand barrier caused local ponding of water within the mire system around 2810 ± 40 years BP.

Several of the pollen taxa in the upper levels of LH-4 are of coastal type and Frank (1982) suggested a dune system migrating landward as the process driving environmental change at Low Hauxley soon after 3000 years BP. This mechanism is supported by the recent work of Orford *et al.* (2000), whose date of 2420 ± 60 years BP (AA-23505) from their core DR-4 at Hauxley, as well as a similar date of 2330 ± 60 years BP (HAR-8973) from Tipping's site at Low Hauxley-B (R.M. Tipping, pers. comm.; Hardiman *et al.*, 1992), also records a sand sheet overblowing terrestrial freshwater environments during this period. Orford *et al.* (2000) also have reported several radiocarbon dates from Druridge Bay and nearby sites that are analogous to Frank's (1982) date of *c.* 1000 years BP for major dune slack development. They use the Low Hauxley sequence in their model for typical sand dune deposition in southern Northumberland, where aeolian sand overlies in-situ freshwater organic sediment over till, which they term 'unconformable normal regressive deposition'. In this model dates on the top of the organic sediment reflect the timing of aeolian sand invasion of the site rather than initial dune initiation. Hence dates several centuries apart occur within a small spatial area, as around Low Hauxley. Such landward sand movement occurs under conditions of relative sea-level rise, often when deceleration in the rate of rise is occurring. During the past 3000 years, therefore, there has been a transgressive shoreline with onshore sand movement in southern Northumberland. Orford *et al.* (2000) also note that dune building phases tend to be site-specific, so the Low Hauxley dune history cannot be applied directly to other sites in the area, and the history for south Northumberland is quite different to that in more northerly parts of the county.

The organic deposits at Low Hauxley-B (R.M. Tipping, unpublished) are broadly time equivalent to those from Low Hauxley (Innes and

Frank, 1988) and part of the same palaeocatena, but they represent different depositional environments. The two profiles therefore allow detailed spatial reconstruction of mid- to late-Holocene landscape change in the area. The archaeological sites were located on drier ground on the edge of the Low Hauxley-B channel or basin, which contained a watercourse and then a small pond. In contrast, peat formation began at Low Hauxley on higher ground, with paludification within damp woodland, perhaps also in a depression that has now been partly removed by coastal erosion. Although there are detailed differences in local environments, the overall sequence of water-table fluctuations at the two sites are similar. The silty clay beneath the organic strata at Low Hauxley-B reflects fluvial deposition in a small silt-laden stream and the expansion of aquatic pollen at the start of the limnic mud layers records ponding of this stream and deepening water. This open-water environment continues until mid-profile, when a drier phase occurs, with Cyperaceae temporarily replacing the aquatic indicators. Higher water levels and aquatic types then return until the arrival of sand in the stratigraphy near the top of the organic layers, when aquatic taxa decline sharply again. This wet–drier–wet–sandy succession corresponds exactly to Innes and Frank's Low Hauxley record, with the generally wetter hydrosere at Low Hauxley-B explained by its lower altitude, organic formation starting in the channel or basin feature almost half a metre below that at Low Hauxley.

A single process is likely to have initiated organic accumulation at both Low Hauxley sites. Although there is no evidence of coastal conditions at the base of the Low Hauxley sequence, the sub-organic clay at Low Hauxley-B contains dinoflagellate cysts *Spiniferites* and *Operculodinium* and coastal pollen taxa, such as Chenopodiaceae, *Armeria* and *Artemisia*. The sea was clearly not far away when the basal clay and mud deposits were forming at Low Hauxley-B after 5000 years BP. A rising sea level may well have had an influence in raising coastal water levels through ponding at both locations. Although generally there is more evidence of probable woodland clearance activity at Low Hauxley-B and a generally more open vegetation, it is difficult to estimate the impact of human catchment activity on hydrological and depositional changes at either Low Hauxley site. The clearest correlation of pollen evidence for woodland clearance between the two sites occurs in the drier phase in mid-profile in each case, dated to the mid- to late Bronze Age. The presence of the Low Hauxley Bronze Age sites explains this, the Low Hauxley-B profile adjacent to these sites having the clearest pollen evidence for human activity. Radiocarbon dates on human bone from both burial cairns (LUAU, 1995) were close to 3500 years BP, corresponding to the age of the clearance activity. The renewed flooding that took place in the upper profile at both sites is more likely to be the result of renewed positive movement of sea level than to the effects of forest clearance.

Plater and Shennan (1992) and Shennan *et al.* (2000a) have used the two radiocarbon dates on the main peat-bed boundaries at Low Hauxley (Innes and Frank, 1988) as limiting data for constraining the sea-level curve for Northumberland. As these dated lithological contacts have no direct evidence of intertidal deposition, they cannot be related to a past reference tidal level. They can indicate only a maximum limiting altitude for sea level at that time, which was above contemporary Mean High Water of Spring Tides. The radiocarbon date of Tipping (pers. comm.) on the clay–peat boundary at Low Hauxley-B is at lower altitude and would have been closer to MHWST at that time. It also is a limiting date for sea-level reconstruction purposes, although the possible evidence for coastal influence that it contains suggests it may not have been far above exceptional high tides, and perhaps reached under storm conditions.

Conclusions

Low Hauxley is a complex concentration of mid- to later Holocene sediments that is important on several counts. It has been central to the establishment of a sand dune chronology for this part of the Northumberland coast, and has been valuable in the reconstruction of sea-level history. The preservation of a buried soil palaeocatena beneath the dunes allows spatial study of landscape development prior to dune emplacement. Different types of organic deposit originating in contrasting wetland environments permit reconstruction of coastal zone vegetation changes and hydroseral successions. The deposits record the impact of later prehistoric farming and forest clearance on the landscape and are associated with multi-period cultural and environmental remains.

FEATHERBED MOSS (SK 094 924)

D. Huddart

Introduction

Featherbed Moss occupies the centre of a gently rounded ridge which trends NE–SW at 500–540 m OD, connecting the higher ground of Kinder Scout with that of Bleaklow (Figure 8.57). This ridge forms the watershed between the Shelf Brook and the River Ashop and the moss rises gently to the crest of Salvin Ridge and Featherbed Top (Figure 8.58) and forms an unbroken blanket peat cover, now extensively gullied (Conway, 1954; Bower, 1960a, b, 1961; Radley, 1962; Barnes, 1963). The site is important because it has been studied extensively, is the best-documented Flandrian upland-peat site and has provided important data for many peat-related topics in the Pennines. These include the extent and causes of peat erosion (Johnson, 1957; Bower, 1962; Tallis, 1964a–c, 1965, 1973a, 1981a, 1985a, b; Mayfield and Pearson, 1972; Shimwell, 1974, 1981; Evans, 1977; Lee, 1981; Ferguson and Lee, 1983); the dates of the onset of peat erosion (Tallis and Switsur, 1973; Tallis, 1985a); the rates of peat erosion (Tallis, 1973a, 1981a); the hydrology of blanket peat systems (Tallis, 1973a, 1985a, b); the causes and date of the disappearance of *Sphagnum* from the bog surface (Tallis, 1964a, c, 1985a); and the development of the vegetation record through the Flandrian (Conway, 1954; Tallis, 1964a, 1985a, b; Hicks, 1971; Tallis and McGuire, 1972; Tallis and Switsur, 1973). The site is part of the Kinder and Bleaklow SSSI, Derbyshire.

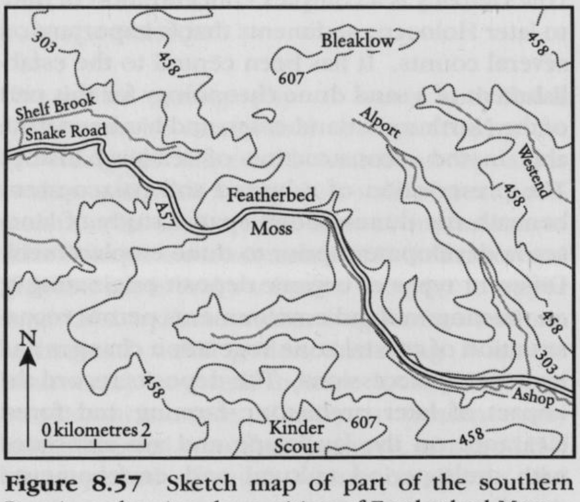


Figure 8.57 Sketch map of part of the southern Pennines showing the position of Featherbed Moss.

Description

Featherbed Moss has been described by Tallis (1973a) and shows the following geomorphological subregions (Figure 8.58):

1. uneroded peat on the dome-shaped, Featherbed Top summit (544 m OD) and on the surrounding, gently convex slopes;
2. closely gullied peat (type 1 dissection of Bower, 1960a, b) on the Salvin Ridge col, forming an erosion complex of haggs, encircling gullies and bare peat flats;
3. bog slope areas with long, sparsely branched and almost parallel gullies (type 2 dissection of Bower, 1960a, b), draining into the River Ashop headwaters;
4. deeply incised headwaters, which occupy channels that were already in existence when peat began to form on the moss (they occur in Thomason's Hollow and the upper reaches of Lady Clough);
5. oversteepened bare peat faces along the peat margin where it abuts on to steeper slopes;
6. slumped peat, which is located along a considerable stretch of the northern moss margin of the moss and is marked by irregular topography, up to 80 m wide.

The drainage gullies can be seen from aerial photographs to fall into four integrated systems (Figure 8.59): (a) two fan-shaped systems on the western part of the moss occurring in the areas of deeper peat overlying the pre-glacial reaches of the River Ashop, (b) a third fan-shaped system draining the deep peat on the Salvin Ridge col, and (c) a radial system on the convex eastern part of the moss. There is considerable variation in peat depth on the moss (Tallis, 1973a) and the extent of peat build-up has been closely dependent on the subsurface contours (Figure 8.60).

Present-day vegetation

The present-day vegetation is dominated over extensive areas by *Eriophorum vaginatum* (cotton grass), with associated *E. angustifolium*. The stream courses, however, are fringed in their upper reaches by *Rubus chamaemorus* and in their lower reaches by *Empetrum nigrum* (crowberry) and *Vaccinium myrtillus* (bilberry).

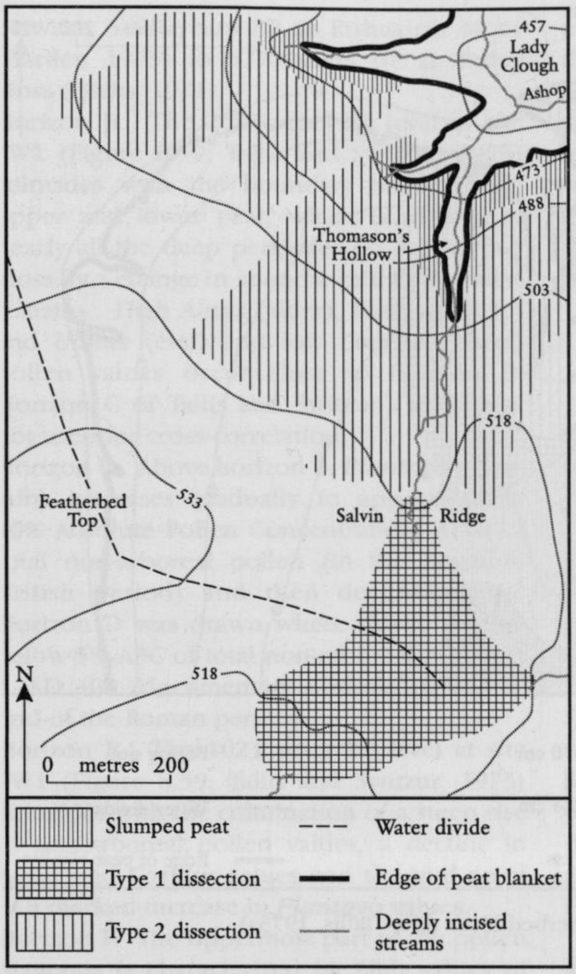


Figure 8.58 Map of Featherbed Moss showing main topographical features (after Tallis, 1973a).

Deschampsia flexuosa (wavy-hair grass) is widespread on the drier peat surfaces. All parts of the moss are regularly shot over for grouse in the autumn and probably have been managed by burning in the past, although regular burning now seems confined to the south side where *Calluna vulgaris* (heather) is widespread (Tallis, 1973a).

Extent of peat erosion

Blanket peat covers some 300 km² of the southern Pennine uplands (Anderson and Yalden, 1981) and erosion, particularly of the deeper peats is widespread, affecting about three-quarters of this area. The extent and severity of this erosion is probably without parallel elsewhere in Britain, and as Conway (1954) described 'there are probably greater expanses of deep and heavily eroded peat than can be found in any other mountain region of the British Isles'. Featherbed Moss has a variety of all the erosional types described by Bower (1960a, b), Radley (1962) and Tallis (1985a). The nature of the peat erosion suggests that dissection is at a relatively early stage and considerable areas of uneroded peat still remain. This erosion appears to be typical of many of the interfluve areas throughout the southern Pennines.

Rates of peat erosion

Peat erosion was directly measured by Tallis (1973a) in a single gully during 1970–1971 as 175 kg, but as he stated this can be no more than a minimum estimate. Accordingly he suggested that a figure of around 300 kg dry weight of peat per year might be more realistic and values of 1000 kg are not impossible. Substantial peat

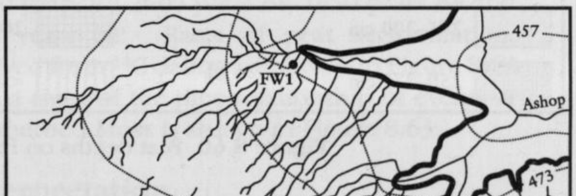


Figure 8.59 Map of Featherbed Moss showing extent and distribution of gullies (after Tallis, 1973a).

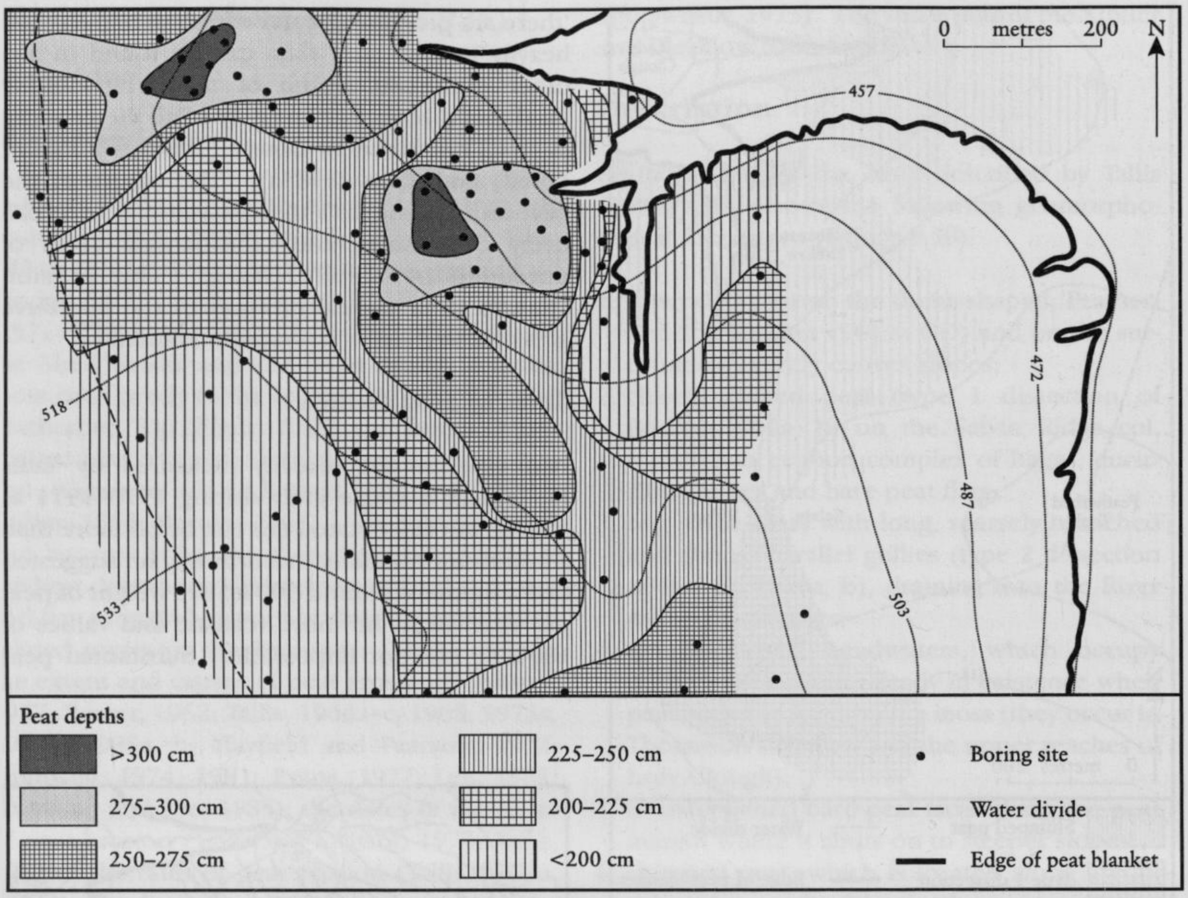


Figure 8.60 Peat depths on Featherbed Moss (after Tallis, 1973a).

erosion was shown to occur during snow melt and during heavy rainfall, when stream flow rates exceeded 40–50 l min^{-1}. An estimate of the former amount of peat in the gully was made at *c.* 50 000 kg and hence, using the estimated rate of peat erosion, the duration of erosion was suggested as between 200–250 years. As the rate of erosion might be expected to increase progressively, as the gullies become wider and deeper, and more bare peat is exposed, this time period could be longer.

Pollen and macrofossil stratigraphical record and its ^{14}C Dating

The pollen record for this section of the Pennines is given in Conway (1954) and Tallis (1964a) from Goyt Moss, Kinder and Wessenden Head Moor and the peat stratigraphy was divided into three horizons, largely using the presence or absence of *Sphagnum* (Conway, 1954; Tallis, 1964a). On Featherbed Moss peat profiles were later obtained by Tallis (1965) and Tallis and Switzur (1973) and also collected from 19 sites by Tallis, (1985a). A summary pollen diagram for three sites is illustrated in Figure 8.61. Earlier work had suggested that distinctive pollen horizons could be recognized in most of the southern Pennine moorlands. Five of these pollen horizons were used to cross-correlate the diagrams, using peat stratigraphy, weed pollen curves and by changes in the pollen spectra from the local bog vegetation, and these were dated by ^{14}C (Tallis and Switzur, 1973), or by documentary evidence where available.

The five horizons were characterized by Tallis (1985a) as follows (Figure 8.61), from the base.

1. An *Ulmus* horizon with the primary decline in the elm pollen about 5500 years BP. No ^{14}C dates are available for Featherbed Moss but dates for other southern Pennine sites vary

between 5490 years BP at Rishworth Moor (Bartley, 1975) to 4770 years BP at Totley Moss (Hicks, 1971).

2. Horizon B. The 2685 years BP level at site FW1 (Figure 8.59; Tallis and Switzur, 1973) coincides with the boundary between the upper and lower peat, which is marked at nearly all the deep peat sites on Featherbed Moss by a change in abundance in *Sphagnum* remains. High *Alnus* (alder), *Fraxinus* (ash) and *Ulmus* (elm) and low *Corylus* (hazel) pollen values occur close to horizon B. Horizon C of Tallis and Switzur (1973) was not used for cross-correlation.
3. Horizon D. Above horizon B, *Plantago* (plantain) increases gradually to approximately 10% Absolute Pollen Concentration (APC) of total non-arboreal pollen (in the Romano-British period) and then declines again. Horizon D was drawn where values decline below 5% APC of total non-arboreal pollen at *c.* AD 400 (documentary evidence from the end of the Roman period).
4. Horizon E. The 1023 years BP level at site FW1 (Figure 8.59; Tallis and Switzur, 1973) coincides with the culmination of a steep rise in non-arboreal pollen values, a decline in *Alnus* and *Corylus* values and the mid-point of a marked increase in *Plantago* values.
5. Horizon F. The uppermost part of the pollen diagrams is characterized by high values of *Plantago* and *Cerealia* pollen, together with increasing values of *Fraxinus*, *Pinus* and *Ulmus*. Values of *Plantago* pollen, typically under 25% APC during the medieval period rise sharply to over 50% APC some distance above the level of the Q-849 ^{14}C date of AD 1515 (Tallis and Switzur, 1973). Horizon F was drawn at the mid-point of this *Plantago* rise and dated partly on documentary evidence to *c.* AD 1580 (Scott *et al.*,1973).

There is a pronounced change in the peat stratigraphy along the gullies on Featherbed Moss, as described in Tallis (1965), and there is no doubt that substantial vegetation differences have existed in the past between different regions, although at present these differences are obliterated by a more or less uniform cover of *Eriophorum vaginatum* away from the drainage channels. In the past conditions appear to have been considerably drier on the crest of the interfluve, as is shown by the rather slower rate of peat accumulation here, by the poor representation of *Sphagnum* in the peat, by the frequent records of *Lycopodium* spores and perhaps by the high Cyperaceae values. On the gentle slopes of the interfluve, however, conditions appear to have been much wetter, with a faster rate of peat accumulation and an abundance of *Sphagnum* and *Drosera*. Conway (1954) had recognized a threefold superimposed peat sequence:

1. a highly humified, basal peat, lacking *Sphagnum*, which began to form around 7000 years BP;
2. a humified lower peat, with abundant Ericaceae and Cyperaceae and some *Sphagnum*, which began to form around 5000 years BP or earlier;
3. a less compacted and less humified upper peat, with abundant *Sphagnum*, which began to form around 3200 years BP.

Periodic horizontal bands of unhumified *Sphagnum* were thought possibly to be true recurrence surfaces by Tallis (1964a), datable to 3200 years BP, 2600 years BP, 1650 years BP and 650 years BP. Rates of peat accumulation, derived from ^{14}C dating are shown in Figure 8.62 and a chart of the time-relationship of events at Featherbed Moss is shown in Figure 8.63.

Interpretation

The three major features of southern Pennine peats i.e. the widespread dominance of *E. vaginatum*, the virtual absence of *Sphagnum* and the extensive erosion of the blanket peats, were thought to be the result of drastic modifying factors (Pearsall, 1950). Some evidence suggested that the changes might be relatively recent:

1. the recollections of gamekeepers and shepherds that *Calluna* had decreased at the expense of *E. vaginatum* during the 20th century;
2. the mention by Farey in 1813 of *Sphagnum* as a prominent member of the upland vegetation;
3. the striking decrease of *Andromeda polifolia* in the past 100 years.

Tallis (1964a) suggested in the light of documentary and palynological evidence that *E. vaginatum* assumed dominance after the 14th century as a result of human interference with the

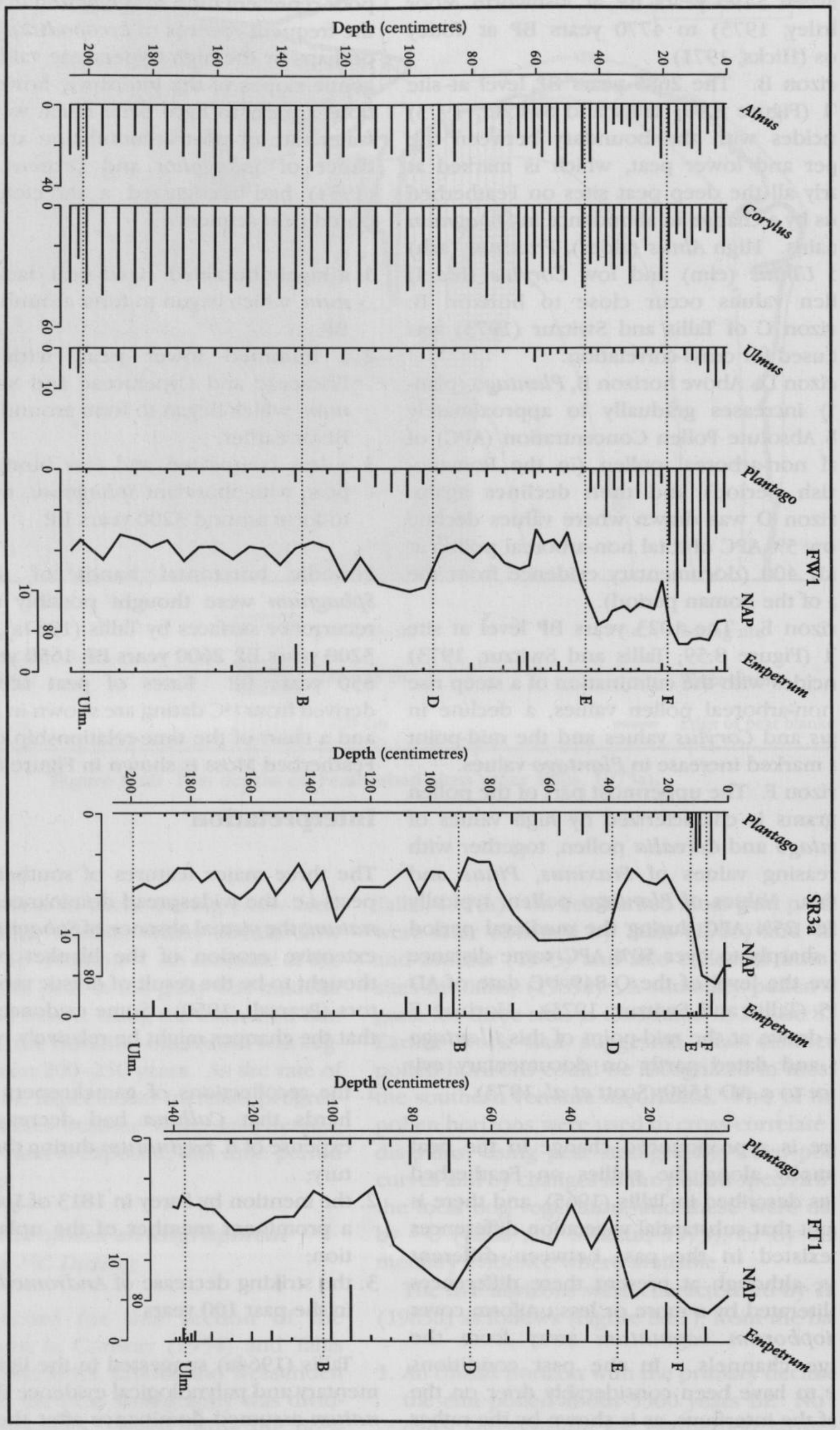


Figure 8.61 Summary pollen diagram for three sites at Featherbed Moss showing percentage values of *Plantago* pollen (left-hand histogram), percentage *Empetrum* pollen (right-hand histogram). Ulm., B, C, E and F refer to pollen horizons described in the text (from Tallis, 1964a).

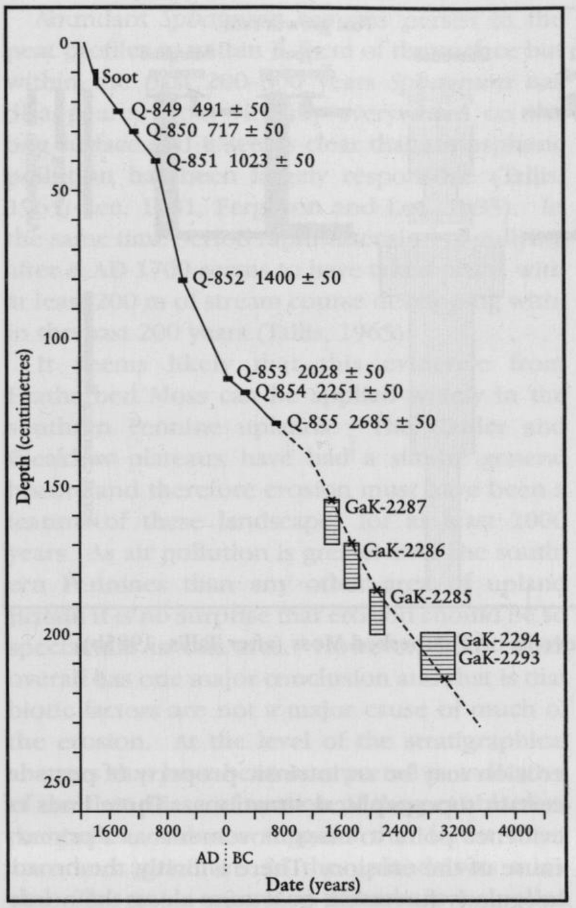


Figure 8.62 The rate of peat accumulation at Featherbed Moss (after Tallis and Switzur 1973).

vegetation. The modifications produced resulted in a *Sphagnum* decline in the vegetation, but he thought that the almost total absence today of *Sphagnum* can be attributed to atmospheric pollution in the past two centuries. The data collected from Featherbed Moss have been used to provide information on the timing of events and the general levels of human activity in the area around the moss from the pollen record; to reconstruct the bog surface vegetation at different times in the past; to trace the development of the peat blanket there; to make inference about climatic change and the causes and mechanisms of peat erosion and the timing of such erosion. Such data have been summarized in Tallis (1985a, b) and can be applied generally to the southern Pennine uplands.

Pollen evidence of forest clearance, settlement and cultivation (cereal and weed pollen) is most marked above horizon E (*c.* AD 1000) and is corroborated by documentary and place-name evidence and must be the result of Viking colonization. The preceding time interval in the 'Dark Ages' appears to have been a time of re-entrenchment of the natural vegetation following disturbance by settlement and grazing in the Iron Age and Roman periods (Hicks, 1971).

Evidence for climatic change

Three main periods of dryness during the bog growth have been recognized, and as the patterns are similar at both eroded and uneroded sites, it is likely that climatic factors rather than local drainage factors were responsible for the *Sphagnum* declines. One coincides with the Little Climatic Optimum of the Early Middle Ages (*c.* AD 1100–1300) (Lamb, 1977), another probably is coincident with the period *c.* AD 800–900 and witnessed by drier conditions at Bolton Fell Moss (see site report, this chapter), and the younger drier phase may coincide with a time period in the past 200–300 years, when there has also been air pollution. Separating these dry phases are two wetter phases, when Sphagnum spread over much of the bog surface at *c.* AD 900–1100 and between AD 1350 and 1750. Earlier there clearly had been wetter climates and *Sphagnum* dominance in all the deeper peat profiles.

The longer term climatic trends during the past 5700 years can be seen from Figure 8.63. There was a period of relatively dry climate (CR-1) between *c.* 2800 and 5700 years ago, when peat growth was slow; a period of wetter climate (CR-2) between 1600–2800 years ago, during which peat growth increased by 50%; a period from AD 400 to 1000 (CR-3) when climatic conditions appear to have been most favourable for rapid peat growth, and a period covering the past 1000 years (CR-4), when peat growth slowed again to a rate similar to that in CR-1. Figure 8.63 forms a framework for considering the possible causes of peat erosion on Featherbed Moss and more generally on the southern Pennine uplands.

The causes of peat erosion

The pollen stratigraphical record for Featherbed Moss shows that the modern, degraded, bog surface is not the product of a single process, or a single time period, but has come into being gradually owing to a variety of factors over many

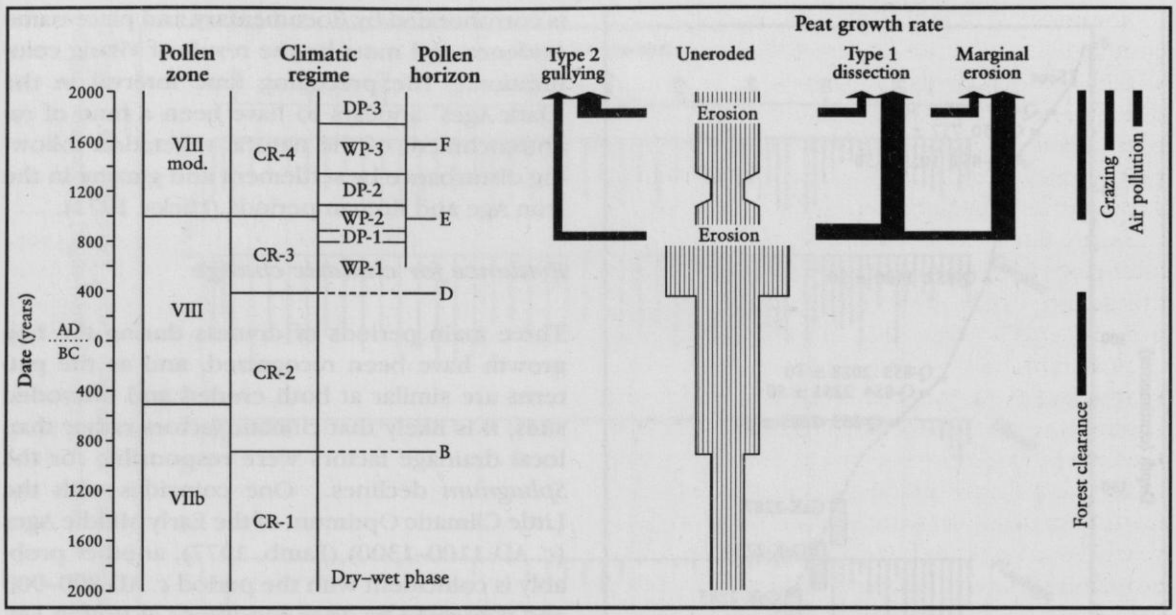


Figure 8.63 Chart showing time-relationship of events on Featherbed Moss (after Tallis, 1985b).

centuries. The erosion that we see today is the result of two widely separated temporal erosion phases, the latest initiated 200–300 years ago and still active, whereas the other was initiated 1000–1200 years ago. The recent erosion phase occurs in the context of intensified grazing and trampling on the moorlands and probably owes little to climatic effects. The earlier phase precedes pollen features in the peat profiles that were the result of major forest clearance of the upland hill-slopes in the tenth and eleventh centuries AD and hence Tallis (1985b) suggests that the erosion was unlikely to be the result of human activity. He suggests that it was generated by naturally induced mass movement of the blanket peat over parts of the moss, as suggested by Colhoun *et al*. (1965). Conway (1954) had believed that instability in southern Pennine blanket peats was accentuated by climatic change around 600 BC, leading to the rapid build-up of uncompacted *Sphagnum* peat above more humified peat, and was then relieved by bog bursts around the margins of the peat. Johnson (1957) too had suggested that bog-bursts were a feature of the 'post-mature' stage of peat accumulation, when erosion of the unstable, water-charged peats became inevitable. Bower's (1961) recognition of climate and landform as the major factors governing the present-day distribution of erosion implies that erosion may be an intrinsic property of peats in certain topographical situations. Three lines of evidence point to mass movements as a primary cause of the erosion. There is firstly, the broad, irregularly furrowed, peat zone along the northern edge of the moss where the topography is similar to some recorded bog slides (Crisp *et al*., 1964) and where more pronounced indentations at intervals along the peat margin could be interpreted as sites of former local bog bursts associated with the slides. Secondly, the peat stratigraphy of sites in this marginal zone differs from sites elsewhere on the moss. The bog surface over this area appears to have suffered a major drying-out some 1100–1200 years ago, from which it never recovered. Although this drying-out was present elsewhere on the moss, in other areas active peat growth recommenced widely after *c*. AD 1000. Thirdly, a case is made by Tallis (1985b) for potential instability, which developed on the margins of the moss as a result of the peat build-up. These factors together caused the various types of erosion between AD 800 and 1000: peat slides at the northern edge of the moss, accompanied by local bog bursting; incision of a deep stream channel at the head of Thomason's Hollow; partial drainage by this stream of pools and hollows in the low relief hummock–hollow complex on Salvin Ridge and the unbranched gullies on the bog slope.

Abundant *Sphagnum* remains persist in the peat profiles to within 3–5 cm of the surface but within the past 200–300 years *Sphagnum* has disappeared from virtually everywhere on the bog surface and it seems clear that atmospheric pollution has been largely responsible (Tallis, 1965; Lee, 1981, Ferguson and Lee, 1983). In the same time period rapid extension of gullying after *c*. AD 1700 seems to have taken place, with at least 200 m of stream course developing within the past 200 years (Tallis, 1965).

It seems likely that this evidence from Featherbed Moss can be applied widely in the southern Pennine uplands. The Kinder and Bleaklow plateaux have had a similar general history and therefore erosion must have been a feature of these landscapes for at least 1000 years. As air pollution is greater over the southern Pennines than any other area of upland Britain it is no surprise that erosion should be so spectacular in this area. However, Tallis' work overall has one major conclusion and that is that biotic factors are not a major cause of much of the erosion. At the level of the stratigraphical changes that have been interpreted as indicative of the first phase of erosion of the peat blanket, the pollen evidence suggests that forest clearance and agriculture in the uplands was minimal. The major expansion of herbaceous pollen associated with Viking penetration of the upland valleys occurred later, its lower limit dated to AD 927 ± 50 years (Tallis and Switzur, 1973). The Viking colonization is indicated from the place-name evidence (Ekwall, 1922; Barnes, 1962; Cameron, 1959), but the Vikings undoubtedly occupied a landscape in which erosion was already active, as indicated by the terms of Norse origin to describe erosion features (hagg, grain and grough).

However, in a sense biotic factors can be held responsible for this early phase of erosion, as the exploitation of the southern Pennine uplands as grazing land for wild game by Mesolithic hunters prior to 5000 years BP led to a gradual degeneration of the upland forest under the combined effects of grazing and regular burning (Jacobi *et al*., 1976) and to a takeover of the flatter ground by peat-forming communities. If this had not happened so early then peat might not even now have built up to a critical instability at the margins.

Nevertheless there are parts of the erosion system that cannot be covered by this explanation. Several large areas of erosion at the present day are known to have been caused by catastrophic accidental fires in the past 50 years and a number of similar areas also are suspected to have been burned at some time (Tallis, 1981b). Active erosion at the peat margin in a number of places also is being accentuated, if not actually caused, by intensive sheep grazing (Tallis and Yalden, 1984). Continual erosion along long-distance footpaths is another recent additional factor (Shimwell, 1981). It appears therefore that the upland peat system is inherently unstable and its break-up can be triggered in a number of ways. Featherbed Moss shows that the peat erosion has been a natural component of the uplands for at least 1000 years and that even a degrading peat blanket can be a long-term component of the landscape.

Conclusions

Featherbed Moss has been shown to be an important southern Pennines site that well documents several themes related to the development of the upland landscape. It shows the development of the vegetation, the chronology of changes, and the timing and the development of the major peat erosion. This erosion is shown to have multiple causes but is thought to be predominantly a natural process, the result of inherent instability within the blanket peat.

LEASH FEN (SK 296 741)

G. Wilson

Introduction

Leash Fen, Derbyshire, a deep (*c*. 6 m) topogenous peat deposit on the Southern Gritstone Moors, is a key site in the context of reconstructing mid- and late Flandrian environmental changes. The pollen record from this site was the first to be radiocarbon dated in the southern Pennines and details an important chronology of successive woodland clearance phases in this area. Hicks (1971) compared Leash Fen with other sites in the south-eastern Pennines and found that it typified environmental changes in this region. Comparison with other work in the southern Pennines as a whole reveals a series of anthropogenic disturbances and climatic perturbations common throughout the area, although data obtained from Leash Fen reveal important differences between sites during and after the Roman occupation.

Description

Leash Fen, previously referred to as 'Leashfield Moss' (e.g. Farey, 1813) is located in the Southern Gritstone Moor area, known also as the 'East Moors' (Coles, 1985) of the south-east Pennines. It lies on the Lower Coal Measures at an altitude of 283 m OD. Prior to peat initiation the Southern Gritstone Moors were covered by a mixed oak forest with occasional breaks in the forest canopy, where areas of damp heath and alder and birch carr occurred (Conway, 1954). The onset of a wetter climate during the Atlantic period (*c.* 8000–5000 BP) coupled with the topography and elevation of Leash Fen probably shifted the local precipitation–evaporation regime to one that was conducive to peat initiation (Moore, 1993). The final stimulus that allowed peat inception, however, has been linked to human activity in the southern Pennines. Tallis and Switsur (1990) suggested that clearance of scrub by fire led to peat initiation at Leash Fen, with the onset placed at 6300 ± 150 BP (4300 ± 150 BC; Hicks, 1971).

Several sites were studied in the south-east Pennines by Hicks (1971). Leash Fen was one of the largest sites in the study and the environmental changes preserved in the pollen record here reflect to a certain extent south-east Pennine landscape changes as a whole. Similar studies at Rishworth Moor, approximately 55 km north-west of Leash Fen, reveal a broadly similar record of environmental change during the mid- and late Flandrian, although there are important local differences, especially during and after the Roman period. Therefore, although the pollen record at Leash Fen could tentatively serve as a record of environmental change for the southern Pennines as a whole, significant local variation does occur.

A detailed pollen record from Leash Fen is shown in Figure 8.64. Based on significant vegetation changes and associated archaeological periods the pollen diagram has been divided into three major zones, A, B and C. Zone A represents woodland clearance in the Neolithic and Bronze Ages, zone B woodland clearance in the Iron Age and in Roman times and zone C vegetation changes after the Roman occupation.

Interpretation

Evidence of human disturbance

Zone A – Neolithic and Bronze Age

Zone A, from 6300 ± 150 BP to 2340 ± 100 BP, records a series of woodland clearances in the area (Figure 8.65). Clearance phases, all of which are radiocarbon dated are identified by successive peaks of *Plantago lanceolata*, an open-habitat indicator (Behre, 1981) with associated rises in one or all of Poaceae, Cyperaceae and Ericaceae and also of cultural indicators such as *Rumex acetosa* type. A small-scale clearance is recorded at 4120 ± 100 BP. In Figure 8.64 this is indicated by the first rise in *P. lanceolata* at 500 cm, together with rising values of Cyperaceae and appreciable values of Ericaceae. This has been related to the activity of Middle and Late Neolithic herdsmen based on the findings of polished stone axes near Hathersage, approximately 10 km north-west of Leash Fen, dated to between *c.* 4500 and 3800 BP (Hicks, 1971).

At 3790 ± 100 BP a further clearance is recorded (Figure 8.64). Here, Poaceae reaches values of 20% TLP (total land pollen). Hicks (1971) has attributed this clearance phase to the 'Food Vessel People' based on the recovery of food vessels dated to 3490 ± 150 BP (Riley, 1966) from Beeley Moor, approximately 5 km south-west of Leash Fen. A more extensive woodland clearance of the Southern Gritstone Moors follows at 3500 ± 110 BP. Together with the presence of *P. lanceolata* and Poaceae, *Rumex acetosa*-type first appears and Ericaceae and Cyperaceae attain values of around 25% and 15% TLP respectively. This clearance episode has been linked to the Early and Middle Bronze Age people, whose presence is evident by the recovery and dating of collared urns in the area (Hicks, 1971).

Woodland still dominates zone A and, together with the small values of Poaceae, indicates the small scale nature of the clearances. Despite evidence of some woodland regeneration after each clearance phase, illustrated in Figure 8.65 as a slight recovery of tree pollen after a period of decline, the trend is one of a move towards a progressively more open landscape, as reflected in the fall of tree pollen from approximately 65% to 40%. The clearances of zone A were probably associated with pastoralism (Hicks, 1971). It is possible that such an agricultural economy was common to the southern Pennines as a whole, as

Bartley (1975) also finds a similar pattern of woodland clearance at Rishworth Moor and he too concludes evidence of a pastoral economy.

Zone B – Iron Age and Roman occupation
The zone A–B boundary at 2340 ± 100 BP marks a significant decrease in tree pollen values (Figure 8.65), together with an increase in clearance indicators such as *P. lanceolata* and Poaceae (Figure 8.64). From 2340 ± 100 BP to 2160 ± 100 BP, Hicks (1971) has attributed the large-scale woodland clearance to Iron Age occupation, although local archaeological evidence for Iron Age settlement is scarce. Hicks (1971) suggested that pastoralism was still very important, with grazing of the cleared areas acting to prevent tree regeneration (Buckland and Edwards, 1984). The date 2160 ± 100 BP should be treated with caution, however, as it is slightly older than the dated horizon nearly 1 m below, 2140 ± 100 BP (Figure 8.64). The change in stratigraphy to a less humified fresher peat between these two dates could indicate a change to a much faster accumulation of peat. A similar change in stratigraphy is seen at Lucas Fen, where a change to a wetter climate is thought to be the possible cause (Long, 1994).

From 1960 ± 100 BP the clearance of woodland continued. Significantly, arable farming became prominent, as suggested in Figure 8.64, with an increase in Cerealia pollen from values close to zero at 140 cm to a peak of 3% TLP at 110 cm. This agricultural practice is seen to occur at other sites on the Southern Gritstone Moors at a similar time, for example, at Stoke Flat (Long, D.J. *et al.*, 1998). Hicks (1971) suggested that the adoption of arable farming could be correlated with the onset of the Roman occupation, with the Romans introducing new crops brought from the continent, such as rye. Rishworth Moor also shows the same general pattern. Here, a pastoral farming economy in the Iron Age was replaced by arable farming, which reached its maximum intensity during the Roman occupation (Bartley, 1975). A mixed farming economy succeeded the dominantly arable economy after the Roman occupation (Hicks, 1971) and continued to the zone B–C boundary.

Throughout zone B it appears that clearance of woodland continued on a larger scale (Figure 8.65). As outlined above, this largely reflects increasing population pressure. Although archaeological evidence reflecting such pressures is limited, fortified hill-top sites are present in the surrounding area (see Butterworth and Lewis, 1968). Varley (1964) stated that these fortified hill-top sites, constructed by Iron Age people between 2100 and 2200 BP, suggest a large organized labour force. Zone B increasingly reflects vegetation changes in the upland as a whole (Figure 8.65). This is a result of the relationship between the presence of woodland and the spatial extent of pollen received. For example, throughout zone A woodland was cleared on a small scale. The pollen diagram at Leash Fen essentially reflects changes in the local vegetation. Zone B, however, increasingly reflects changes in the more regional vegetation. This is a result of the clearance of local woodland, allowing receipt of pollen rain from an extended area (Tauber, 1965).

Zone C – post-Roman occupation
Hicks (1971) has tentatively dated the onset of zone C to between AD 860 and AD 1060. The sharp rise in *P. lanceolata* (Figure 8.64) and continued fall in tree pollen (Figure 8.65) that characterize this boundary reflect the intense woodland clearances attributed to the 10th century Norse invaders (Tallis and Switsur, 1973). As reflected in Figure 8.65, Leash Fen records an upland landscape that is almost treeless. The rising Ericaceae and Poaceae pollen mark the increasing encroachment of grass and heather moorland (Figure 8.64), culminating in a landscape that probably was similar in appearance to that of today. Further small increases in *Pinus*, *Ulmus* and *Fraxinus* respectively indicate secondary woodlands and plantations (Hicks, 1971).

Correlation across the southern Pennines

Tallis (1964a) studied several Southern Pennine blanket peat sites and through pollen analysis picked out several horizons (A to E) marking either climate fluctuations or anthropogenic disturbance that were common to each of these sites. Based on the pollen assemblage and associated radiocarbon date of each of these horizons (Tallis and Switsur, 1973) several of these horizons can be detected at Leash Fen. That such correlation between sites is possible throughout the Southern Pennines suggests periods of extensive environmental change.

Horizons A, B, C and E can all be detected in the pollen diagram at Leash Fen (Figure 8.64).

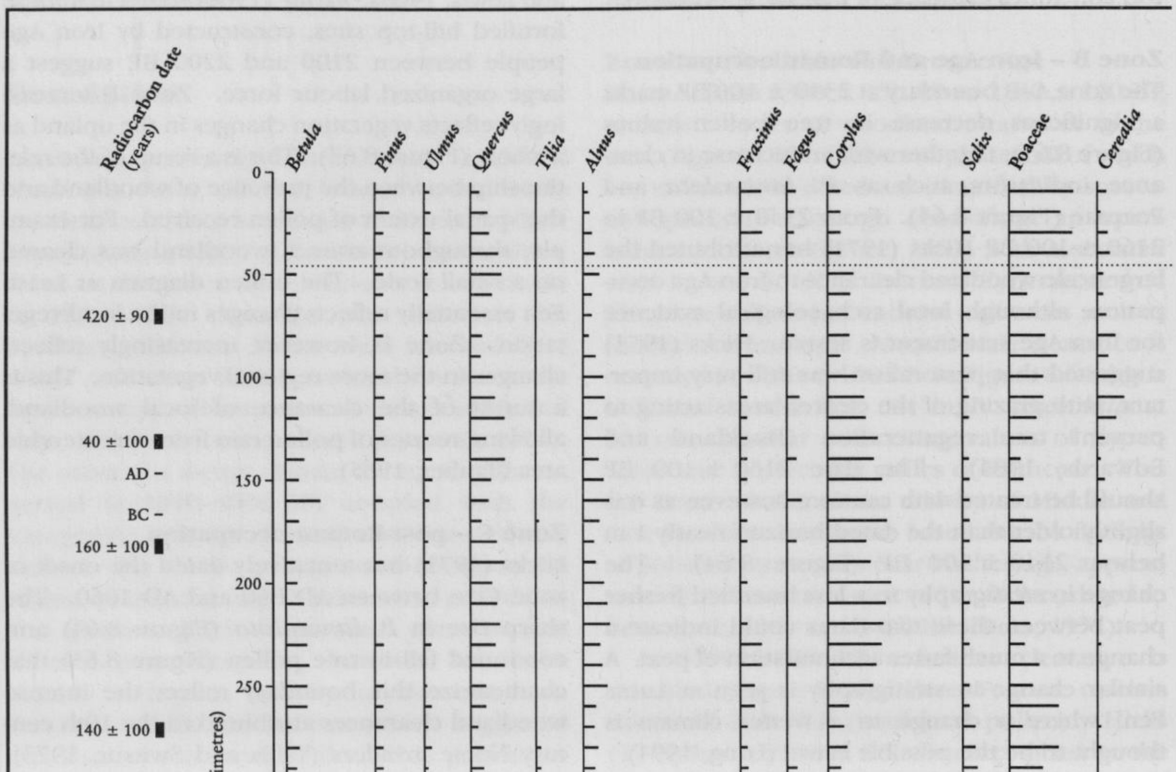
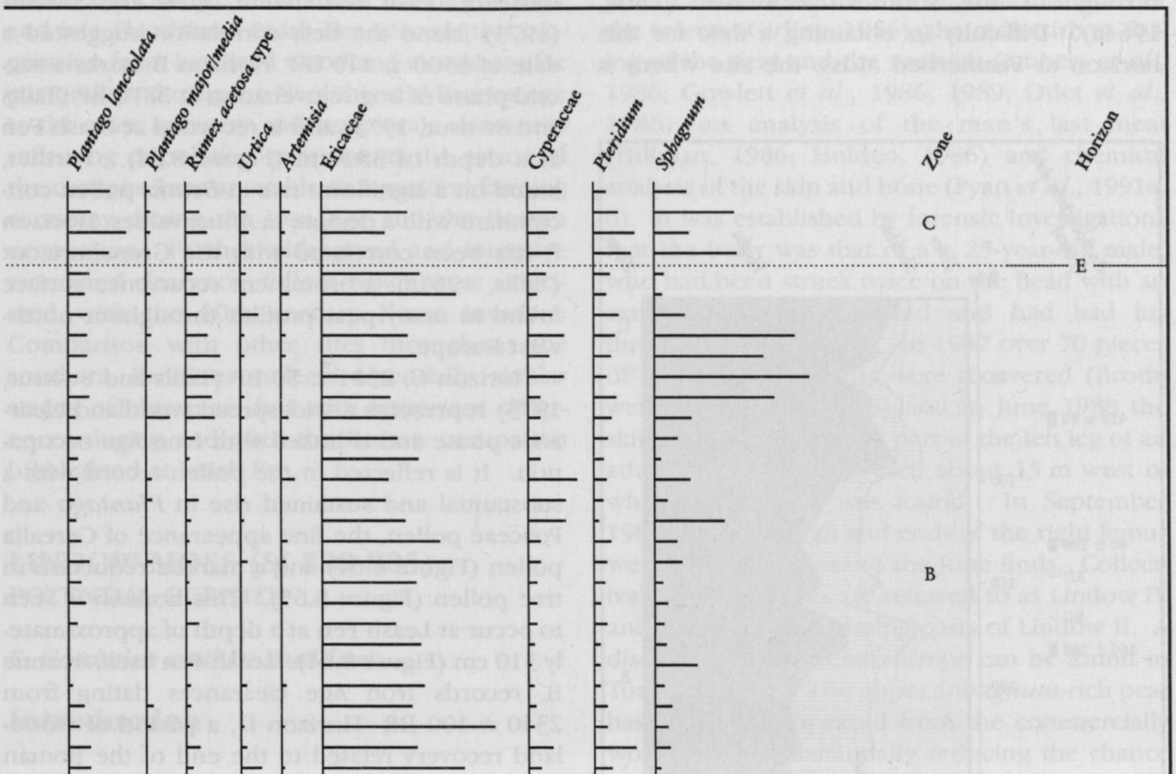


Figure 8.64 Percentage pollen diagram from Leash Fen covering the mid- to late Flandrian. Values are expressed as a percentage of total land pollen excluding aquatic pollen and spores. Several horizons are shown that mark widespread anthropogenic and climate events in the southern Pennines (after Hicks, 1971). Dates are given in years BC and AD.

Leash Fen



Plantago lanceolata
Plantago major/media
Rumex acetosa type
Urtica
Artemisia
Ericaceae
Cyperaceae
Pteridium
Sphagnum
Zone
Horizon
C
E
B
C
B
A
A
0 10 1 0 2 1 1 0 80 0 20 0 10 0 100
Total land pollen (%)

Horizon A denotes a climate deterioration that produced a marked rejuvenation of bog growth throughout the southern Pennines (Tallis, 1964a). Difficulty in obtaining a date for this horizon at Featherbed Moss, the site where it was first recognized, has necessitated tentative correlation of its *Plantago* pollen values with those of Leash Fen. Here, Tallis and Switsur (1973) found the best correlation suggested a date of 3500 ± 110 BP. Horizon B marks a second phase of bog rejuvenation at 2870 BP (Tallis and Switsur, 1973) and is recorded at Leash Fen at a depth of 380 cm (Figure 8.64). Further, based on a significant rise in *Betula* pollen concomitant with a decline in *Alnus* values, Horizon B has been correlated with the Grenzhorizont (Tallis, 1964a), a prominent recurrence surface found in many peat profiles throughout northwest Europe.

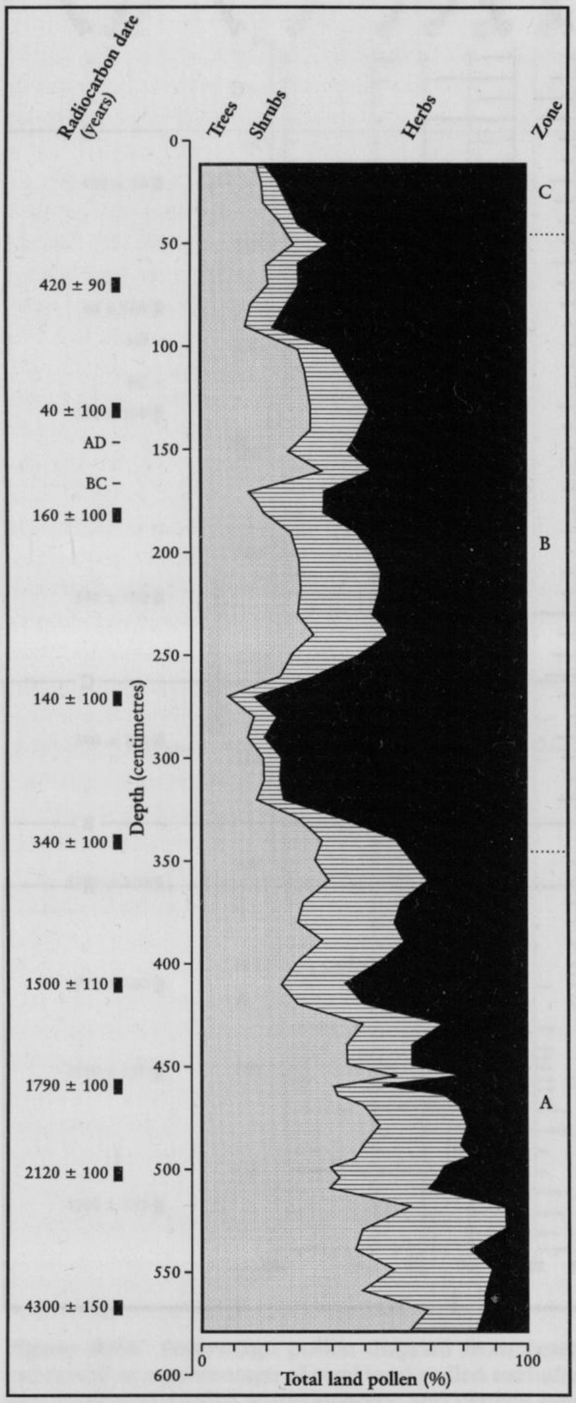


Figure 8.65 Cumulative pollen diagram from Leash Fen summarizing changes in tree, shrub and herb pollen (modified from Hicks, 1971).

Horizon C, 2301 ± 50 BP (Tallis and Switsur, 1973) represents a widespread woodland clearance phase and is linked with Iron Age occupation. It is reflected in the pollen record with a substantial and sustained rise in *Plantago* and Poaceae pollen, the first appearance of Cerealia pollen (Figure 8.64) and a marked reduction in tree pollen (Figure 8.65). This horizon is seen to occur at Leash Fen at a depth of approximately 310 cm (Figure 8.64). Leash Fen itself, in zone B, records Iron Age clearances dating from 2340 ± 100 BP. Horizon D, a period of woodland recovery related to the end of the Roman occupation (Tallis, 1964a), is not apparent in the Leash Fen pollen record. Instead there is evidence of a mixed farming economy continuing throughout the period to the time of the Norman Conquest. This reiterates the point of Bartley (1975), who states that variations between sites in the southern Pennines are particularly noticeable during and after the Roman occupation. Horizon E (AD 927 ± 50), marking the lower limit of intense woodland clearance related to the 10th century Norse invaders (Tallis and Switsur, 1973), is tentatively correlated with the start of zone C in the Leash Fen diagram (Figures 8.64 and 8.65). Although radiocarbon dates are lacking, Hicks (1971) suggests a date of between AD 860 and AD 1060 for the onset of zone C. It is the dramatic increase in *P. lanceolata* pollen values, however, together with the associated decrease in tree pollen values (Figure 8.65), suggestive of a period of intensive clearance, that facilitates correlation between records.

Conclusions

The fossil pollen record of Leash Fen, interpreted within a radiocarbon dated framework, provides detailed information of landscape changes as a consequence of both anthropogenic disturbance and climatic fluctuations during the mid- and late Flandrian. Leash Fen records the progression from localized woodland clearances for pastoralism during the Neolithic and Bronze Age to the more intensive and larger-scale clearances reflecting population pressure in the area and the changes from an arable to a pastoral farming economy during the Iron Age and the Roman occupation. Further widespread and intensive woodland clearances follow, which were associated with the 10th century Norse invaders. Comparison with other sites throughout the southern Pennines reflects essentially similar stages of clearance, but with important differences during and after the Roman occupation highlighted at Leash Fen.

LINDOW MOSS (SJ 820 805)

POTENTIAL GCR SITE

S. Gonzalez and D. Huddart

Introduction

Lindow Moss was originally an extensive (600 ha), lowland peat bog that accumulated in a kettlehole formed during the Late Devensian deglaciation in Cheshire, but now it is about one-tenth of its former size and covered mainly by birch scrub (Brothwell, 1986). The moss margins were gradually reclaimed for agriculture and the moss cut for fuel (Norbury, 1884; Turner, 1995a). Sandy islands remain rising above the moss and are thought to be windblown dunes formed in a periglacial environment after the ice melted. During the 1980s the moss was the location of a remarkable series of four discoveries of well-preserved human remains. The discoveries of Lindow I (Lindow Woman) in 1983 and Lindow II (Lindow Man) in 1984 have been described by Turner (1986). Lindow I consisted of a skull that retained its outer membrane and some hair. The vault contained a decayed brain and part of the left eyeball and optic nerve were also identifiable. Lindow II was the archaeological discovery of the 1980s according to Turner (1995a), and almost a complete body was excavated from a peat block (Figure 8.66). Subsequent discussions established its antiquity and a wide variety of scientific investigations were undertaken (see Connolly, 1985; Stead and Turner, 1985; Brothwell, 1986; Stead *et al*., 1986), including pollen analysis (Oldfield *et al*., 1986), a study of the insects (Girling, 1986), the radiocarbon dating of the peat and the remains (Ambers *et al*., 1986; Gowlett *et al*., 1986, 1989; Otlet *et al*., 1986), an analysis of the man's last meal (Hillman, 1986; Holden, 1986) and chemical analysis of the skin and bone (Pyatt *et al*., 1991a, b). It was established by forensic investigations that the body was that of a *c*. 25-year-old male, who had been struck twice on the head with an axe-like weapon, garroted and had had his throat cut (West, 1986). In 1987 over 70 pieces of a body (Lindow III) were recovered (Brothwell and Bourke, 1995) and in June 1988 the skin of the buttocks and part of the left leg of an adult male were recovered about 15 m west of where Lindow III was found. In September 1988 the right thigh and ends of the right femur were recovered close to the June finds. Collectively these samples are referred to as Lindow IV and represent the missing parts of Lindow II. A discussion of the relationships can be found in Turner (1995b). The upper *Sphagnum*-rich peat has all but disappeared from the commercially worked area, substantially reducing the chance of any more finds. However, considerable amounts of this type of peat still survive at the moss, most notably to the east and to the south, where a fragment of the moss is managed by the

Figure 8.66 Lindow II bog body. © The British Museum

Cheshire Wildlife Trust as a nature reserve. The locations of Lindow Moss, findspot locations and excavation trenches exposed during 1987 are shown in Figure 8.67.

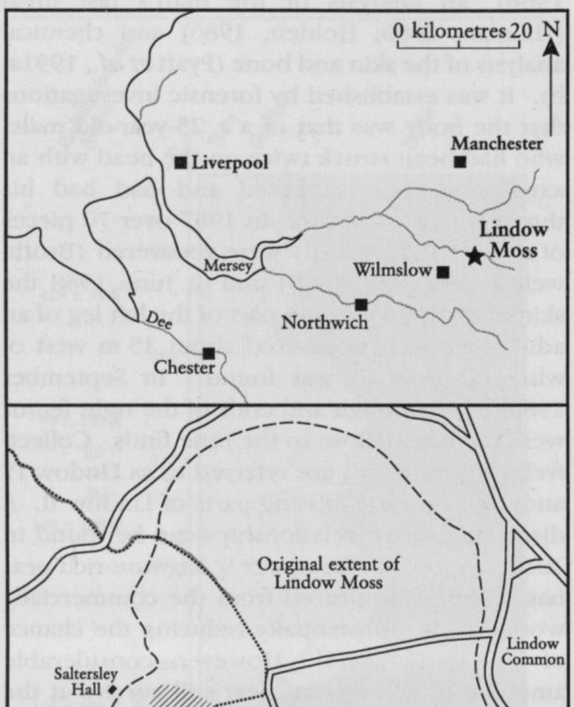


Figure 8.67 Map showing the location of Lindow Moss and significant finds, and the positions of sand islands in the peat and excavation trenches during 1987 (after Turner, 1986, 1995a).

Lindow Moss is important because it has provided the most detailed analysis of bog bodies at any site in Britain (Brothwell, 1986, 1995; Stead *et al.*, 1986; Turner and Scaife, 1995). There is a wealth of associated palaeoenvironmental analysis (Birks, 1965b; Girling, 1986; Oldfield *et al.*, 1986; Branch and Scaife, 1995; Dinnin and Skidmore, 1995; Leah *et al.*, 1997) and the stratigraphy and bodies have been dated. There were initial problems with this dating (Ambers *et al.*, 1986; Gowlett *et al.*, 1986, 1989, Otlet *et al.*, 1986) but the site has been important as the work has shown that ^{14}C dating of human bodies from peat bogs presents special problems because of the biochemical processes involved in the interaction between the peat and human tissue. Nevertheless, there now seems to be agreement on the dating (Housley *et al.*, 1995) and both Buckland (1995) and Barber (1995) have presented a variety of mechanisms by which the ^{14}C dates and stratigraphical positions may be resolved. The archaeological significance of the Lindow bog bodies has also been much discussed (Connolly, 1985; Stead and Turner, 1985; Ross, 1986; Ross and Robins, 1989; Buckland *et al.*, 1994; Briggs, 1995; Magilton, 1995; Turner, 1995c). Specialized techniques have been used to obtain detailed information, for example, related to the food eaten (Hillman, 1986; Holden, 1986; 1995; Scaife, 1995) and the significance of the skin geochemistry (Pyatt *et al.*, 1991a, b, 1995; Cowell and Craddock, 1995). There seems no doubt that the studies on the Lindow bog bodies have helped to advance the investigative standards and techniques used on such remains across the world (Brothwell, 1995).

Description

The stratigraphy and pollen analysis

Lindow Moss was first investigated by Birks (1965b) from the south-western moss (SJ 820 807), where the stratigraphy revealed 3.5 m of gyttjas, reed peats, carr brushwood peats, *Eriophorum* and *Sphagnum* peats. The pollen revealed a segment of vegetational history between *c.* 8000 and 6000 BP and culminating some time in the early 19th century.

Following the discovery of the Lindow II body much interest was focused on the past environment and development of Lindow Moss (Barber, 1986; Oldfield *et al.*, 1986). After the later discoveries, stratigraphical analysis of the peat using plant macrofossils and pollen has been used to give a detailed palaeoenvironmental record for the context of the bog bodies (Branch and Scaife, 1995), although the maximum peat depth proved to be 8 m in trial boreholes. Lateral variation in the peat stratigraphy is marked and a complex of interweaving bands that coalesce to form level horizons is evident, typical of ombrotrophic peat bogs. The pollen diagram (Figure 8.68) has been divided into five recognizable pollen-assemblage zones from the base of the analysed sequence at 190 cm:

		Depth (cm)
LIN:1	*Quercus* (to 35%)–*Corylus*-type (50%)–*Alnus* (27%). There are other tree species present (*Betula* 15%) and lesser percentages of *Pinus*, *Ulmus*, *Tilia* and *Fagus*. Herbs include Gramineae (10%), Cyperaceae, *Plantago* species, *Rumex*, *Chenopodium* and *Urtica*	190–160
LIN:2	*Quercus–Fraxinus–Alnus–Corylus*-type–*Calluna–Sphagnum*. There is some increase in the herb diversity and wetland taxa show some increase	160–120
LIN:3	*Quercus–Corylus*-type–Gramineae–*Plantago lanceolata*–Cyperaceae–*Pteridium–Sphagnum*. There is a marked increase in *Sphagnum* with a dramatic change in peat composition and reduced humification	120–55
LIN:4	*Betula–Fraxinus–Calluna–Sphagnum*. *Pinus*, *Quercus*, *Corylus*-type and herb values are reduced compared with LIN:3	55–35
LIN:5	*Quercus–Alnus–Corylus*-type. Tree pollen becomes dominant	35–20

Leah *et al*. (1997) recorded the peat stratigraphy in the active peat cuttings from 38 cores. The basal organic deposits tend to be composed of reed and sedge peats or fen-carr. In a few places *Scheuchzeria*-dominated assemblages are prominent. Fen-carr peat follows these initial stages. Following this, many locations show wood peats, succeeded by peats characterized by *Eriophorum*/*Calluna*, with frequent *Polytrichum*, *Hylocomium* and *Aulocomnium palustre*. In turn they are succeeded by *S. imbricatum*-dominated peats at the top. The deepest peats recorded reached 7.15 m, although this was at a point situated in the peat workings where perhaps *c*. 3 m had been removed, suggesting original depths approaching 10 m. Away from the central area though peat thicknesses typically were between 2 and 5 m.

Insect assemblages

Two peat samples were taken from around the Lindow III body to give information about the condition of the corpse and the details of the immediate environment at the time of burial, and from a monolith through the peat in the location from which the body is believed to have come. The detailed fauna is given in Dinnin and Skidmore (1995).

Radiometric dating

An extensive ^{14}C dating programme was undertaken to date the bog bodies and the associated sediments and detailed discussions related to the techniques and problems are given in Gowlett *et al*. (1986,1989), Otlet (1986) and Housley *et al*. (1995). The radiocarbon determinations are given in Table 8.12. These indicate differences in age between the bog bodies and the surrounding peat, e.g. Lindow II collagen determinations are about 1800 years BP, whereas the peat surrounding the arm is about 2500 years BP.

Chemical analysis of the skin

As part of the study of the Lindow II and Lindow III bog bodies, geochemical investigations were carried out on skin, bone and associated peat (Pyatt *et al*., 1991a, b; 1995). The skin samples of Lindow III were obtained from the shoulder region, whereas those of Lindow II were from under the right side of the body. Bone samples from Lindow II were from the upper right orbit. The investigations revealed an excess of aluminium, silica and copper, together with traces of titanium and zinc, although the general pattern of inorganic components differed between the two individuals. Cowell and Craddock (1995) analysed further samples from Lindow III from

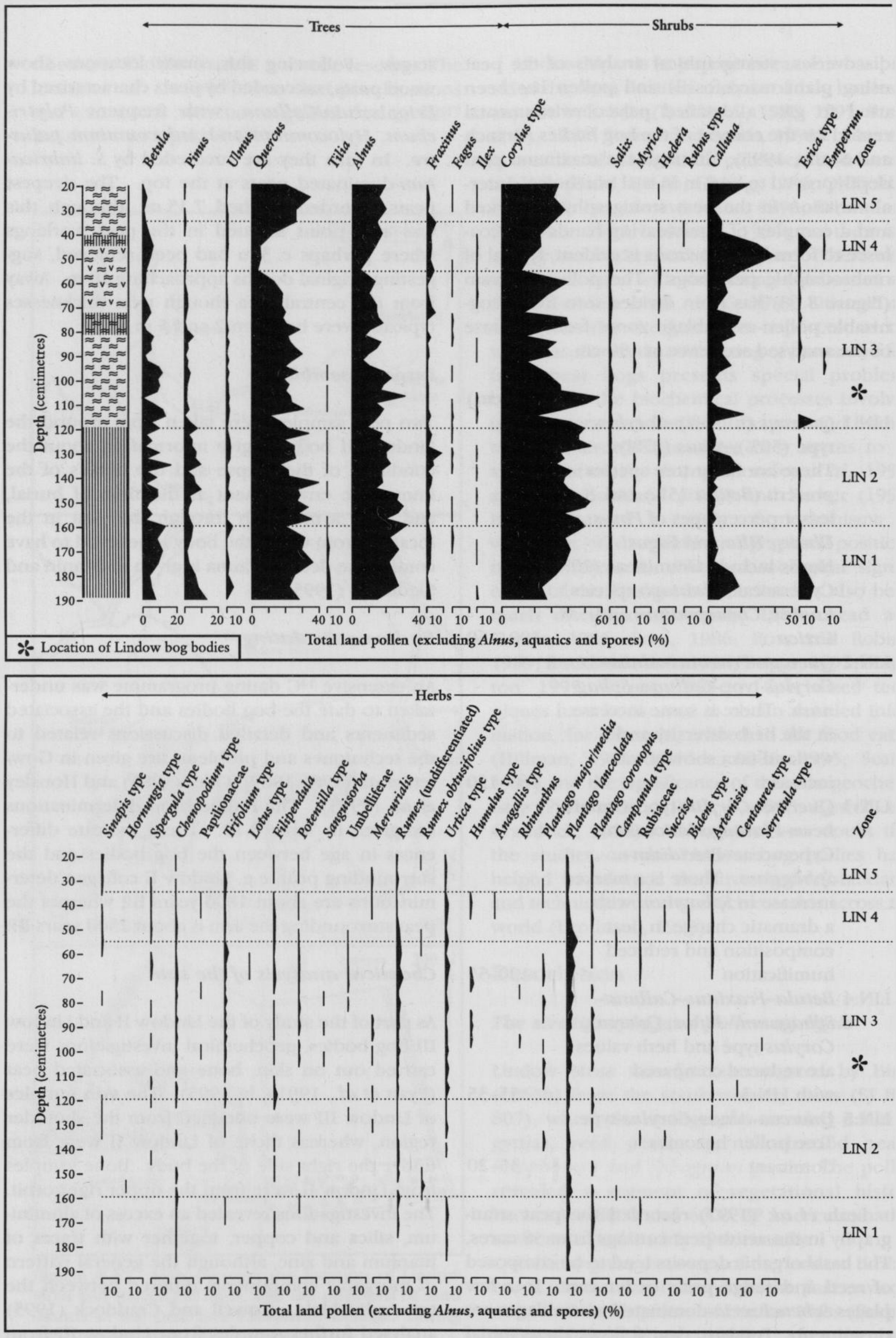
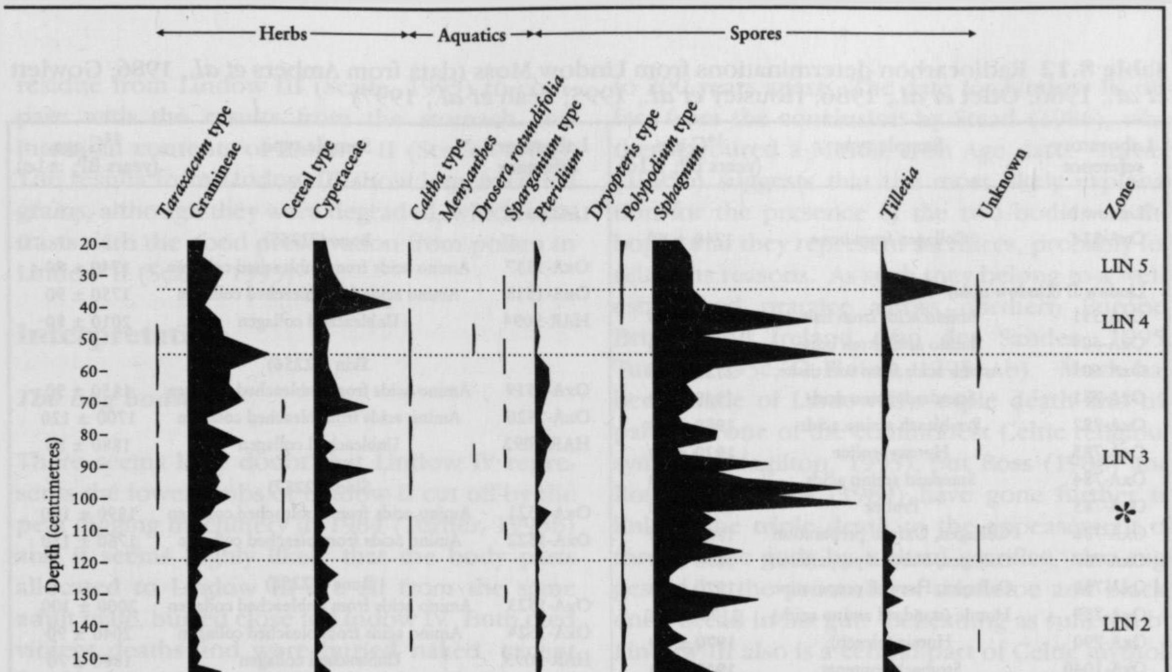


Figure 8.68 Pollen diagram for the in-situ peat section at Lindow Moss (after Branch and Scaife, 1995). See Figure 8.1 for key to the stratigraphical log.


Herbs
Aquatics
Spores
Taraxacum type
Gramineae
Cereal type
Cyperaceae
Caltha type
Menyanthes
Drosera rotundifolia
Sparganium type
Pteridium
Dryopteris type
Polypodium type
Sphagnum
Tilletia
Unknown
Zone
LIN 5
LIN 4
LIN 3
LIN 2
LIN 1
Depth (centimetres)
Total land pollen (excluding Alnus, aquatics and spores) (%)

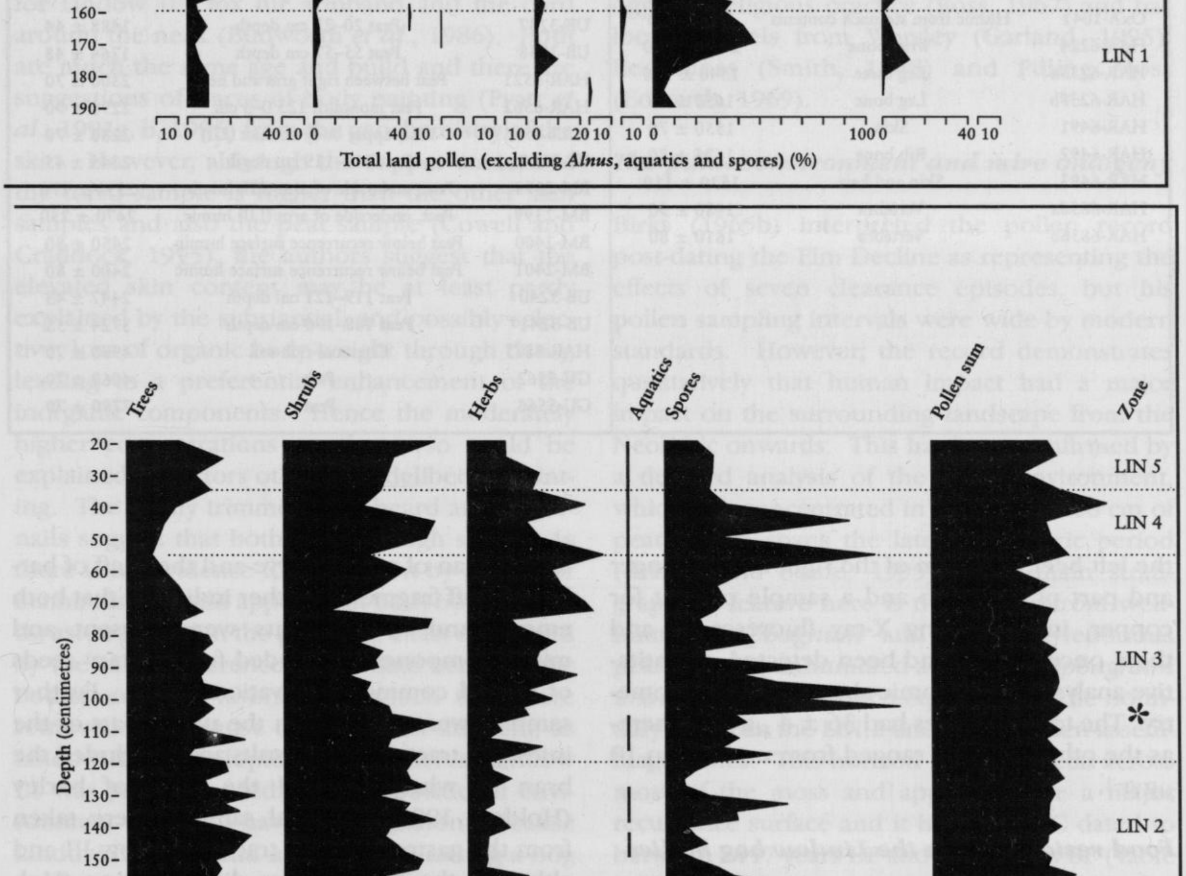

Trees
Shrubs
Herbs
Aquatics
Spores
Pollen sum
Zone
LIN 5
LIN 4
LIN 3
LIN 2
LIN 1
Depth (centimetres)
Total land pollen (excluding Alnus, aquatics and spores) (%)

Table 8.12 Radiocarbon determinations from Lindow Moss (data from Ambers *et al.*, 1986; Gowlett *et al.*, 1986; Otlet *et al.*, 1986; Housley *et al.*, 1995; Leah *et al.*, 1997)

Laboratory reference	Sample type	^{14}C age (years BP; $\pm 1\sigma$)	Laboratory reference	Sample type	^{14}C age (years BP; $\pm 1\sigma$)
Lindow I			**Lindow III**		
OxA-114	Collagen from bone	1740 ± 80		Bone (P2255)	
			OxA-1517	Amino acids from unbleached collagen	1740 ± 90
Lindow II (Lindow man)			OxA-1518	Amino acids from bleached collagen	1750 ± 90
OxA-531	Amino acids from hair	1920 ± 20	HAR-9094	Unbleached collagen	2010 ± 80
OxA-604	Amino acids from bone	1850 ± 80			
OxA-605	Amino acids from soft tissue	2125 ± 80		Skin (P2256)	
OxA-781	Standard amino acids	1940 ± 80	OxA-1519	Amino acids from unbleached collagen	1850 ± 90
OxA-782	Pre-bleach amino acids	1950 ± 80	OxA-1520	Amino acids from bleached collagen	1700 ± 120
OxA-783	Hyroxyproline	1920 ± 80	HAR-9092	Unbleached collagen	1880 ± 80
OxA-784	Standard amino acids	1900 ± 80		Skin (P2257)	
OxA-785	Proline	1900 ± 80	OxA-1521	Amino acids from unbleached collagen	1890 ± 100
OxA-786	Collagen, Oxford preparation	1800 ± 80	OxA-1522	Amino acids from bleached collagen	1760 ± 150
OxA-787	Collagen, Harwell preparation	1870 ± 80			
OxA-788	Collagen, Harwell preparation	1870 ± 80		Bone (P2258)	
OxA-789	Humic (standard amino acids)	2190 ± 100	OxA-1523	Amino acids from unbleached collagen	2000 ± 100
OxA-790	Humic (bleach)	1970 ± 80	OxA-1524	Amino acids from bleached collagen	2040 ± 90
OxA-1040	Stomach contents	1910 ± 60	HAR-9093	Unbleached collagen	1860 ± 70
OxA-1041	Humic from stomach contents	2210 ± 60	UB-3237	Peat 20–22 cm depth	1488 ± 44
HAR-6224	Wrist bone	2420 ± 100	UB-3238	Peat 55–57 cm depth	1764 ± 48
HAR-6235a	Leg bone	1540 ± 100	HAR-6521	Peat between right arm and head	2300 ± 70
HAR-6235b	Leg bone	1650 ± 80	HAR-6562	Peat monolith 125 0–3 cm	2290 ± 90
HAR-6491	Skin	1550 ± 70	HAR-6565	Peat, upper body contact (LII)	2280 ± 70
HAR-6492	Rib bone	1625 ± 80	UB-3239	Peat 117–119 cm depth	2345 ± 45
HAR-6493	Skin and hair	1530 ± 110	BM-2398	Peat, underside of arm (LII) humin	2590 ± 170
HAR-6856a	Vertebra	1480 ± 90	BM-2399	Peat, underside of arm (LII) humic	2470 ± 250
HAR-6856b	Vertebra	1610 ± 80	BM-2400	Peat below recurrence surface humin	2450 ± 80
			BM-2401	Peat below recurrence surface humic	2400 ± 80
			UB-3240	Peat 119–121 cm depth	2447 ± 43
			UB-3241	Peat 188–190 cm depth	3724 ± 55
			HAR-8875	Charcoal-rich soil	4980 ± 70
			GU-5562	Peat	4060 ± 70
			GU-5566	Peat	7780 ± 70

the left heel, the palm of the right hand, a finger and part of the torso and a sample of peat for copper, initially using X-ray fluorescence and then, once copper had been detected, quantitative analysis using atomic absorption spectrometry. The torso samples had 36 ± 4 $\mu g\ g^{-1}$, whereas the other samples ranged from under 2 to 10 $\mu g\ g^{-1}$.

Food residues from the Lindow bog bodies

To investigate the dietary characteristics of the Lindow bog bodies a preliminary analysis of a sample from the fundus and body of the stomach of Lindow II was made by Hillman (1986) and Holden (1986). This indicated that the major part of the last meal was made up of cereal (the bran of wheat or rye and the chaff of barley). Chaff fragments further indicated that both emmer and spelt wheats were present and minor components included fragmentary seeds of several common cultivation weeds. Further samples were taken from the upper part of the intestinal tract and the results again included the bran of wheat/rye and the chaff of barley (Holden, 1995). Several samples were taken from the gastrointestinal tract of Lindow III and although the results were disappointing (Holden, 1995) the food was dominated by fragments of the testa and brown inner layer of the pericarp of the hazelnut and there were smaller quantities of cereal bran. Nineteen samples were examined for pollen analysis from faecal, colonic, gastrointestinal, duodenal and rectal residue from Lindow III (Scaife, 1995) to compare with the results from the stomach and intestinal contents of Lindow II (Scaife, 1986). The results from Lindow III show largely cereal grains, although they were degraded, which contrasts with the good preservation from pollen in Lindow II (Scaife, 1995).

Interpretation

The bog bodies

There seems little doubt that Lindow IV represents the lower limbs of Lindow II cut off by the peat digging machinery in 1984 (Turner, 1995b) and it seems highly likely that the body parts allocated to Lindow III are all from the same adult male, buried close to Lindow IV. Both died violent deaths and were buried naked, except for Lindow II's fox fur armband and the cord around the neck (Budworth *et al.*, 1986). Both are much the same age and build and there are suggestions of traces of body painting (Pyatt *et al.*, 1991a, b, 1995) from the geochemistry of the skin. However, although the copper content of the torso sample is higher than the other skin samples and also the peat sample (Cowell and Craddock, 1995), the authors suggest that the elevated skin content may be at least partly explained by the substantial, and possibly selective, loss of organic body weight through decay, leading to a preferential enhancement of the inorganic components. Hence the moderately higher concentrations on the torso could be explained by factors other than deliberate painting. The neatly trimmed hair, beard and fingernails suggest that both were of high status. As there is no evidence for predation by insects or mammals it would appear that both bodies were thrust or buried in the peat bog, a fact confirmed by the same difference in ^{14}C date between the bodies and the layers from which they were retrieved. The cause of death was different, as Lindow II had multiple injuries whereas Lindow III was just beheaded. The depositional environment seems to have been different because Lindow II was found in a layer indicating a bog pool, whereas Lindow III was on a peat surface. The calibrated date range for Lindow II with a 95% confidence limits based on the Oxford ^{14}C dates is 2 BC to AD 119, whereas the equivalent unified Oxford and Harwell date range for Lindow III is AD 25–230, so the two bodies may be broadly contemporaneous, or deposited up to 200 years apart. The date for Lindow III differs from the conclusion by Stead (1986), who then favoured a Middle Iron Age date. Turner (1995b) suggests that the most likely explanation for the presence of the two bodies in the bog is that they represent sacrifices, probably for religious reasons. As such they belong to a well-established practice across northern Europe, Britain and Ireland (Van der Sanden, 1995; Turner 1995c; Ó Floinn, 1995a, b). Much has been made of Lindow II's triple death and tripalism is one of the commonest Celtic religious symbols (Magilton, 1995), but Ross (1986) and Ross and Robins (1989) have gone further in linking the triple death to the appeasement of three Celtic gods by a ritual sacrifice, also suggested by the presence of mistletoe and blackened bread in his gut. Beheading as suffered by Lindow III also is a central part of Celtic mythology and religious practice (Ross, 1967) and has local parallels from Worsley (Garland, 1995), Red Moss (Smith, 1988) and Pilling Moss (Edwards, 1969).

The palaeoenvironment and mire ontogeny

Birks (1965b) interpreted the pollen record post-dating the Elm Decline as representing the effects of seven clearance episodes, but his pollen sampling intervals were wide by modern standards. However, the record demonstrates qualitatively that human impact had a major impact on the surrounding landscape from the Neolithic onwards. This has been confirmed by a detailed analysis of the palaeoenvironment, which was concentrated in the upper 200 cm of peat, which spans the later prehistoric period (Branch and Scaife, 1995). The main stratigraphical feature here is the change from well-humified *Sphagnum* and monocotyledonous peat to fresh, unhumified and largely *Sphagnum imbricatum* peat, which equates with the boundary between the LIN:2 and LIN:3 pollen-assemblage zones. This horizon has been seen across most of the moss and appears to be a major recurrence surface and it has been ^{14}C dated to between 2447 years BP and 2345 years BP (Table 8.12). It probably represents a period of gradual climatic deterioration after 2900 years BP and a reduction of 1°C for *Sphagnum* pools to commence at the recurrence surface (Barber, 1982).

During the lower two pollen zones prior to the recurrence surface, *Sphagnum* was dominant in wetter areas of pools and damper hummocks, with Gramineae and Cyperaceae. *Calluna* and *Erica* probably were growing on drier hummock areas. *Alnus* and *Salix* probably were growing in carr woodland along the nutrient-enriched marginal areas of the bog. The drier, adjacent land was a dominantly wooded environment, with some areas of open agriculture. *Quercus* and *Corylus* were dominant throughout the pollen record, but with many other woodland species. Openness is shown by *Ilex* and *Corylus*, which require light for flowering. Branch and Scaife (1995) suggested that although woodland was dominant in the period pre-dating the recurrence surface, there had been anthropogenic disturbance and secondary woodland (*Fraxinus* peaks and *Calluna*). Agricultural activity is noted by the presence of *Plantago lanceolata* and other herbs. Archaeologically this disturbance has been substantiated by the presence of Neolithic artefacts from one of the sand islands (Turner, 1995a) and these could be contemporary with the ^{14}C date from the charcoal-rich soil (Table 8.12). Charcoal is not restricted to the immediate environs of the sand islands, however, and it has been noted throughout the mire's history, apart from its very earliest growth stages (Leah *et al.*, 1997). Burning was likely to have affected the local vegetation, as in many raised mires in north-west England (Huckerby and Wells, 1993; Hall *et al.*, 1995; Middleton *et al.*, 1995). The numerous pine stumps exposed on the cut surface of the moss also often appear to show signs of burning (Leah *et al.*, 1997).

Above the recurrence surface there was no cessation of peat growth but a rapid environmental change, although dry-land vegetation continued to dominate the tree pollen, with an open-canopy woodland of mixed deciduous type, but with less lime. From 120 cm in LIN:3 there is a marked and progressive increase in the diversity and percentages of many herbs, especially disturbed and open-ground species, which correlates with the Iron Age to Romano-British period (Branch and Scaife, 1995). There also are sporadic but more abundant cereal-type records, which reflect increased arable and possibly pastoral agriculture, adjacent to the moss. Increases in the pollen of Ericaceae and bracken spores may be indicative of soil deterioration, especially on the sandy soils nearby. This phase of human disturbance also has been documented by Birks (1965b) and Oldfield *et al.* (1986) and it is clear that there was a progressive disturbance from the early Iron Age, which attained a maximum extent during the Romano-British period. Branch and Scaife (1995) suggest that this disturbance was a regional phenomenon and perhaps can be applied to north-west England (Oldfield, 1960a; Mackay and Tallis, 1994; Middleton *et al.*, 1995). The boundary between LIN:3 and LIN:4 pollen-assemblage zones is where there is a reduction in the percentages of herbs and the taxonomic diversity noted in LIN:3, and a late Romano-British date is suggested (Table 8.12, UB-3238). In LIN:5, *Quercus*, *Alnus* and *Corylus* regain their earlier importance, showing a return to woodland dominance and reduced agriculture in the region. From the pollen analysis of the Lindow III body Branch and Scaife (1995) suggest that it was placed on the growing bog surface, some 10–15 cm above the recurrence surface and not in a peat cutting and they suggest an age of *c.* 2300–2200 BP.

From the stratigraphical work of Leah *et al.* (1997) the moss began life as an extensive area of reedswamp, fen and fen-carr occupying hollows within glaciofluvial and aeolian deposits, which gradually became carr woodland. The deepest deposits probably go back to the early Flandrian. The fen-carr phase ended with an increase in wetness, evidenced by the flooding indicator *Scheuchzeria* in places. A reversal of edaphic conditions immediately after this phase led to much drier conditions and the establishment of a slow growing mire and trees, notably pine, with several phases of pine forest establishment, rather than one discrete arboreal phase. Radiocarbon dates from a pine layer suggest that the species formed a significant component of the mire flora at various times between 7780 ± 70 years BP and 4060 ± 70 years BP (Table 8.12). This relatively dry, tree-dominated stage was followed by a return to wetter conditions and the establishment of *Eriophorum–Calluna*-dominated vegetation across almost the whole moss. Conditions became even wetter and *S. imbricatum* became the dominant peat-former.

The beetle fauna (Dinnin and Skidmore, 1995) from around the Lindow III bog body suggests that the corpse was deposited in a wet *Sphagnum* bog with pools of acid water, with this water surrounded by a typical oligotrophic bog flora of cotton grass, mosses and heather. The complete absence of any carrion fauna leads to the conclusion that the body was rapidly submerged. This was the same for the Lindow II body (Girling, 1986). Despite the stratigraphical recurrence surface, the insect assemblages above, below and around this feature revealed no appreciable change in the moisture status of the bog, or any indicators of climatic change.

The dates for the Lindow II and III bodies and the peat, the state of decay of the corpses and the insect evidence require some reconciliation. Buckland (1995) suggests that a human body could have been pushed down into a deep pool and never reappeared, but Barber (1995) refutes the deep-pool hypothesis, especially as at Lindow the evidence from the Cladocera and the Chironomidae point to a shallow pool (Dayton, 1986). He suggests that the body was inserted into the upper layers of the bog, where the felted pool peat could be rolled back, the body laid on the exposed surface and the upper peat rolled back over it. This would create minimal disturbance to the stratigraphy and its included biota and would be impossible to detect.

The results of the analysis of the gut contents of the Lindow II body show that emmer and spelt wheats and barley were the main constituents (Holden, 1995), which probably had been prepared as an unleavened bread. The Lindow III body on the other hand seems to have consumed hazel nuts and a smaller amount of wheat/rye. The palynological evidence (Scaife, 1986, 1995) illustrates a diet composed largely of farinaceous products, although the presence of mistletoe gave rise to speculation as to the possible druidical practices. Scaife (1986) suggests it was more likely that mistletoe was used for medicinal purposes.

Although the Lindow bog bodies are the best known British examples, research has illustrated (Briggs and Turner, 1986; Turner, 1995b, d) that northern England is particularly rich in such remains (Figure 8.69). For example, they have been documented from Austwick Common (Denny, 1871), Scaleby Moss (Turner, 1988), Seascale Moss (Turner, 1989), Whixall Moss (Turner and Penney, 1996), Grewelthorpe Moor (Turner *et al*., 1991), Red Moss (Smith, 1988), Pilling (Edwards, 1969) and Worsley (Garland, 1995).

Figure 8.69 Bog body remains in Britain. Note the predominance in northern England (after Turner, 1995b).

Conclusions

Lindow Moss has proved to be the best-documented site for bog bodies in Britain and has provided a stimulus for new research on these remains elsewhere. The palaeoenvironment of the bog throughout its history has provided much information on mire ontogeny and the influence of humans on the vegetation in the later prehistoric period. The stratigraphy and the bog bodies have been dated extensively and the problems associated with ^{14}C dating of such remains discussed in detail. The archaeological significance of such bog bodies has been debated extensively. Owing to the extensive commercial working of Lindow Moss and landfill development it is unlikely that further bog body remains will be found, but the site, especially the nature reserve, is important as a representative type locality of the large lowland moss in north-west England.

WYBUNBURY MOSS (SJ 697 503)

N.F. Glasser

Introduction

Wybunbury Moss is an outstanding example of a deep basin mire of 'schwingmoor' structure. A schwingmoor is a floating peat raft where the process of infilling occurs by the formation of a *Sphagnum* raft over open water and the gradual settling down of the basal layers of this raft into the water to give a very fluid structure. The site comprises an enclosed basin containing some 4 m of oligotrophic peat overlying up to 16 m depth of water. The history and development of the mire have been studied using a combination of pollen stratigraphy, macrofossil and chemical analyses. The results suggest that subsidence of the basin probably is the result of solution of the subsurface salt deposits in the underlying bedrock. Pollen analysis of the peat raft has provided a detailed record of the vegetation history of the site, which, although undated, represents a Holocene transition from fen woodland to *Sphagnum* bog. Poore and Walker (1959) provided a general introduction to Wybunbury Moss, the stratigraphical record of its peat deposits and its modern vegetation. Green (1965) and Green and Pearson (1968) have provided more recent details of the modern ecology of the site and its relationship to movements of the water table, groundwater and water chemistry. Tallis (1973b) has used the lake basins of north Cheshire, including Wybunbury Moss, to illustrate the processes of terrestrialization of lake basins by peat growth.

Description

Wybunbury Moss is a large enclosed basin (approximately 12 ha in area) to the north of the village of Wybunbury in Cheshire. The basin is one of more than 130 peat-filled or water-filled depressions ('mosses and meres') in Cheshire. Green and Pearson (1968) describe the location of the site as occupying a depression in glacial sands, overlying Triassic strata. The Triassic rocks in this area are primarily salt-rich beds of fluviatile Sherwood Sandstone, consisting of several hundred metres thickness of massive deposits of rock salt interbedded with thin marl bands (Evans *et al.*, 1968). Although glacial deposits mostly obscure the bedrock, it is known that natural ground subsidence occurs throughout this area of Cheshire as a result of the solution of the underlying salt-rich beds. Although Poore and Walker (1959) considered that the hollow in which the site lies is best interpreted as a kettlehole, Green (1965) argued for a salt subsidence origin for the depression at Wybunbury Moss. Evidence cited by Green (1965) for this includes the stepped terracing of the basin sides and the absence of organic deposits in the basin sediments earlier than late Zone VI.

The sides of the depression occupied by Wybunbury Moss slope steeply down to the mire on the west and south sides, with gentler slopes on the north side (Figure 8.70). At the eastern end the hollow is open, with the peat at the same level as the surrounding land, approximately 49 m OD. The peat-covered area is roughly oval in shape, measuring 700 m by 300 m, and totalling 12 ha in area. At the margins of the mire there is a sharp delineation between the peat and the well-drained pastures and arable land of the surrounding agricultural area. The mire is almost completely encircled by a perimeter ditch, which is mostly overgrown. Ditches cut into the mire drain into this outer ditch, which then discharges through the south-east part of the site into Howbeck Brook. There is no canalized inflow into the mire system. The surface of the mire is level, and stratigraphical investigations show that over most of its area the peat at Wybunbury Moss is less that 4 m thick. The peat overlies a maximum water depth of 16 m (Figure 8.71).

Surveys of the modern ecology of the site show that four main vegetational units dominate the surface vegetation of the site: marginal mixed woodland, marginal reedswamp, central *Sphagnum* lawn and central pine woodland (Green, 1965; Green and Pearson, 1968). Instrumentation has been installed at the site to monitor vertical changes in the peat raft and movement of the water table (Green and Pearson, 1968). Results show that rises and falls in the altitude of the peat raft are correlated closely to movement of the water table. Seasonal trends include a gradual fall in the level of the water table during spring and summer, but a rapid winter rise. The greatest annual amplitude of raft movement reported over the

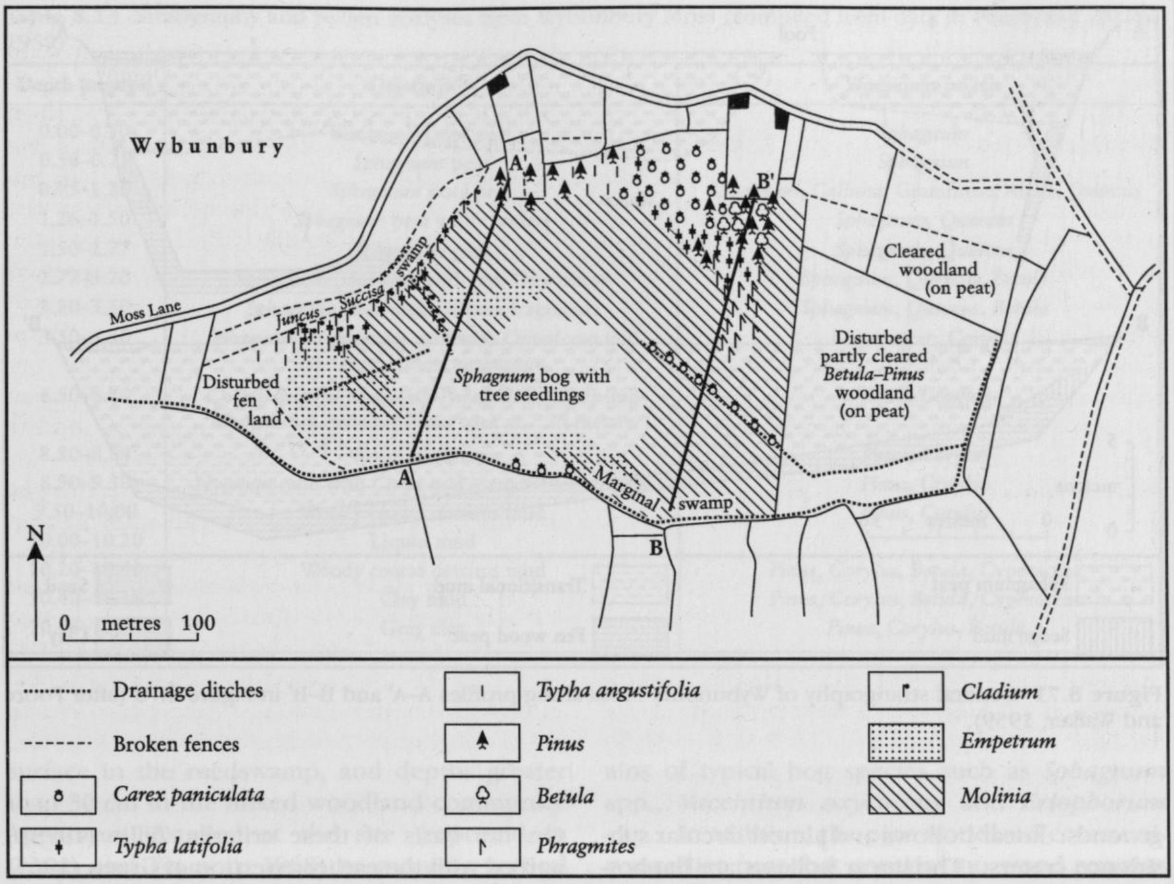


Figure 8.70 Map of Wybunbury Moss, showing the approximate distribution of the main plant communities and the lines of the transects A–A' and B–B' shown in Figure 8.71 (after Poore and Walker, 1959).

four years of observations was 10.6 cm. Chemical analyses of the water from the different plant communities, and from sample sites in and around the peat raft, show that the mire vegetation can be divided into two major units according to the mineral content of the water in which it is growing. The reedswamp and mixed woodland communities are supported by waters far more base-rich than the dilute waters of the *Sphagnum* lawn and pine woodland communities (Green and Pearson, 1968). The reedswamp waters contain at least ten times more calcium than those of the central part of the mire.

Poore and Walker (1959) reported the results of a series of borings made at the site, together with pollen analyses of the sediments in a 10.5 m core (Table 8.13). The lower parts of the core are dominated by muds containing arboreal pollen, with abundant *Pinus*, *Corylus* and *Betula*. Values of *Ulmus* and *Quercus* are relatively low, and *Alnus* is virtually absent. Initially there are high amounts of *Corylus*. Samples collected around the transition from muds to *Sphagnum* peat around 6.50 m in the core contain significant amounts of *Alnus*, *Ulmus* and *Quercus*. The high levels of *Pinus* in the lower parts of the core decline at around 5.00 m and *Sphagnum* dominates above this level. *Pinus* becomes completely absent above 2.05 m.

Interpretation

The origin of the basin

Two types of basins are common in this area of Cheshire: depressions that originated as kettle-holes in glaciogenic deposits during the recession of the last ice-sheet, and those created by more recent subsurface subsidence following solution of salt rich strata. Evans *et al.* (1968) identified two distinct types of subsidence hollow in Cheshire on purely morphological

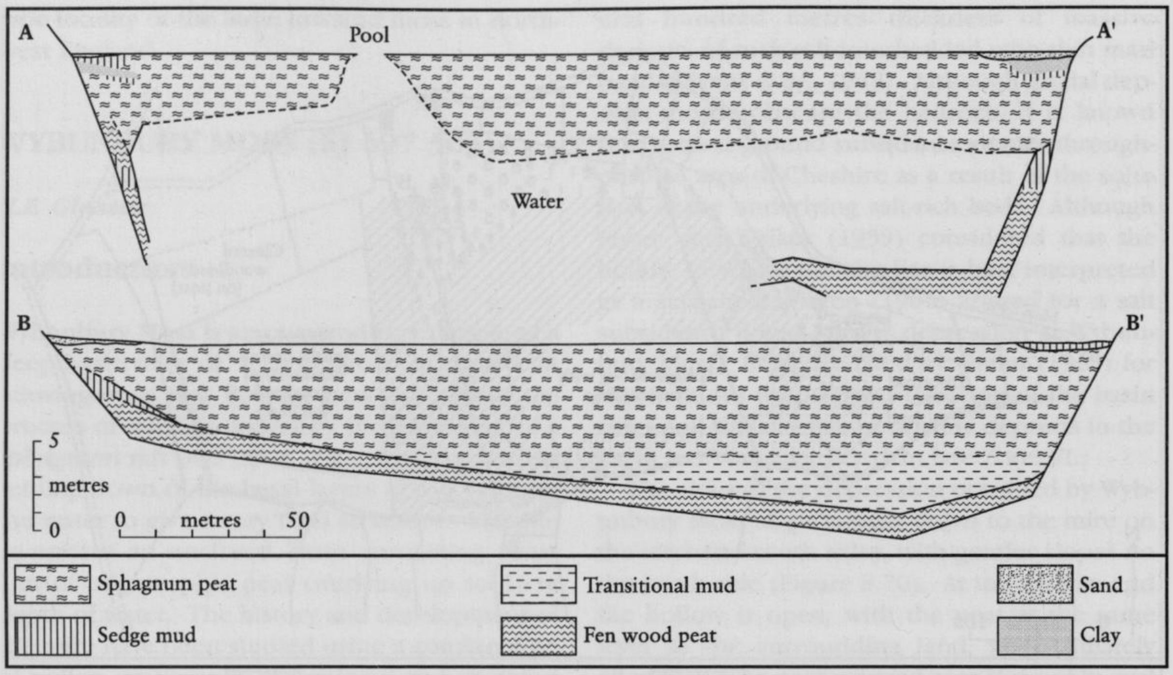


Figure 8.71 General stratigraphy of Wybunbury Moss along profiles A–A' and B–B' in Figure 8.70 (after Poore and Walker, 1959).

grounds: linear hollows and almost circular subsidence craters. The linear hollows are flat-bottomed and steep-sided, with a maximum width of *c.* 200 m and a maximum depth of *c.* 10 m. The near-circular craters are typically 200–300 m in diameter, also flat-bottomed and steep-sided and with the sides sometimes terraced as a result of intermittent subsidence. Based on levelled profiles, Tallis (1973b) produced more diagnostic criteria for discriminating between those depressions representing kettleholes and those created by subsidence.

1. Basins that originated as kettleholes in the glaciogenic deposits contain organic deposits of Zone IV age or earlier. They are small (less that 300 m in diameter), isodiametric or somewhat elongated, steep-sided and with peat depths usually in excess of 5 m.
2. Basins that owe their origin to subsidence resulting from salt solution occupy relatively shallower saucer-shaped depressions. Peat depths in these basins do not generally exceed 3 m and they often contain basal deposits of late Zone VI affinities. These basins are commonly linked to local drainage systems and share the characteristics of these hydrological systems.

On the basis of these criteria, Tallis (1973b) agreed with the earlier assertion of Green (1965) that the origin of Wybunbury Moss is the result of underground subsidence following salt-bed dissolution. Infilling of the basin, termed 'terrestrialization' by Tallis (1973b), occurs by the formation of a *Sphagnum* raft over open water and the gradual settling down of the basal layers of this raft into the water (French and Moore, 1986).

The significance of the modern vegetation communities

In mire ecosystems the geographical distribution of plant communities is related closely to the depth and movement of the water table and the chemical composition of the water. Together, these factors explain most of the features of the vegetation at Wybunbury Moss (Green and Pearson, 1968). For example, water table measurements at the site demonstrate that the reedswamp and *Sphagnum* lawn communities can be differentiated from the two woodland communities by their far higher groundwater levels. Those in the central *Sphagnum* community were less than 10 cm below the peat surface, compared with depths of 10–20 cm below the

Table 8.13 Stratigraphy and pollen analyses from Wybunbury Moss (compiled from data in Poore and Walker, 1959)

Depth (metres)	Description	Dominant pollen
0.00–0.50	Unconsolidated peat	*Sphagnum*
0.50–0.75	*Sphagnum* peat	*Sphagnum*
0.75–1.26	*Sphagnum* pool peat	*Sphagnum*, *Calluna*, Gramineae, *Alnus*, *Quercus*
1.26–1.50	*Sphagnum* peat with rootlets	*Sphagnum*, *Quercus*
1.50–2.77	*Sphagnum* peat	*Sphagnum*, *Quercus*
2.77–3.20	*Sphagnum* pool peat with rare *Oxycoccus*	*Sphagnum*, *Quercus*, *Betula*
3.20–3.50	*Sphagnum* peat with *Calluna* fragments	*Sphagnum*, *Quercus*, *Betula*
3.50–6.50	*Sphagnum cuspidatum* peat with *Oxycoccus* and rare *Eriophorum vaginatum*	Cyperaceae, *Corylus*
6.50–8.80	Coarse detritus mud with *Phragmites*, *Carex* and *Menyanthes* remains; *Pinus* bark at 7.35 metres	*Pinus*, *Corylus*
8.80–8.90	Wood fragments	*Pinus*, *Corylus*
8.90–9.50	*Hypnum* mud with *Carex* and *Menyanthes* remains	*Pinus*, *Corylus*
9.50–10.00	Woody coarse detritus mud	*Pinus*, *Corylus*
10.00–10.20	Liquid mud	
10.20–10.40	Woody coarse detritus mud	*Pinus*, *Corylus*, *Betula*, Cyperaceae
10.40–10.46	Clay mud	*Pinus*, *Corylus*, *Betula*, Cyperaceae
10.46–10.50	Grey clay	*Pinus*, *Corylus*, *Betula*

surface in the reedswamp, and depths greater than 30 cm in the mixed woodland community. Apparatus installed at the site (reported in Green and Pearson, 1968) to monitor vertical changes in the peat raft and movement of the water table showed that the two are closely correlated. This buffering of water table fluctuations favours semi-aquatic plants such as *Sphagnum* species and probably accounts for their dominance of the central community. Chemical analyses show that differences in the mineral content of the waters in which these communities exist also are very marked. According to Green and Pearson (1968) these differences result from fundamental changes in hydrology across the site, for example the marginal reedswamp owes its existence primarily to the ingress of mineral-rich surface waters from outside the mire.

The significance of the pollen stratigraphical record

The pollen record presented by Poore and Walker (1959) shows that the present distribution of the mire vegetation at Wybunbury Moss is a relatively recent development in the history of the mire. The modern reedswamp and mixed woodland communities overlie peat that is oligotrophic in origin and contains abundant remains of typical bog species such as *Sphagnum* spp., *Vaccinium oxycoccus* and *Eriophorum* spp.. In addition, the pine woodland community overlies peat that contains no wood remains and is indistinguishable from that which underlies the *Sphagnum* lawn community. The peat underlying the present vegetation communities is uniform in character, and its palynology indicates that the mire surface in the past was largely covered by open bog vegetation similar to that of the present *Sphagnum* lawn community.

Conclusions

Basin mires with a schwingmoor structure are rare in the UK, and Wybunbury Moss is an outstanding example of this type of feature. The origin of the basin itself is uncertain, but it probably developed as a result of underground subsidence following dissolution of salt-rich beds. Four main vegetational units dominate the modern vegetation at the site: marginal mixed woodland, marginal reedswamp, central *Sphagnum* lawn and central pine woodland. These modern communities are very different to those observed in the palynological record obtained from the site, which is dominated by open bog vegetation similar to that of the present *Sphagnum* lawn community. The pollen stratigraphy of the site shows that the earliest organic deposits are typical fen woodland communities, probably dating from Zone VI. At this time, the water level in Wybunbury Moss was around 12 m lower than present. The transition from fen woodland to *Sphagnum* bog is marked by a change in the deposits from wood peat to mud that is rich in sedges, *Phragmites* and *Menyanthes*, suggesting a rise in the water table. Since this time, *Sphagnum* has spread centripetally across the basin to form the present schwingmoor structure. Unfortunately no radiocarbon dates have been obtained from this site so the precise timing of these events remains uncertain.

MALHAM TARN MOSS (SD 886 669)

D. Huddart

Introduction

Malham Tarn Moss, North Yorkshire, and its associated fens form the western part of the Malham Tarn National Nature Reserve (NNR) and is part of the 5000 ha Malham–Arncliffe SSSI, which is of great overall significance for wildlife and Earth science conservation (Figure 8.72). The NNR is a Ramsar site and falls within the Pennine Dales Environmentally Sensitive Area. Sinker (1960) gave a general account of the topography and character of the vegetation for the area immediately around the tarn, and the North Fen vegetation has been described by Proctor (1974), Adam *et al*. (1975) and Cooper and Proctor (1998). The geomorphology and glacial geology of the area has been discussed by O'Connor (1964), Clayton (1966, 1981) and Clark (1967). Malham Tarn is a shallow marl lake, at 375 m OD, which overlies an impermeable Silurian bedrock, which is mostly covered by calcareous glacial sediments (O'Connor, 1964; Clark, 1967). It is completely surrounded by Carboniferous Limestone and there is a marked escarpment of Great Scar Limestone just north of the North Craven Fault. Malham Tarn Moss is an ombrotrophic, raised bog, covering about 40 ha, some 700–800 m across, built up of acid *Sphagnum* and *Eriophorum vaginatum* peat, which overlie fen peats and lake sediments. Its domed surface rises some 5 m above the level of the adjacent calcareous drainage and tarn level. The surrounding calcareous fens, separating the bog from the surrounding drift-covered limestone, are fed by strongly calcareous water draining from springs that emerge from the limestone at its junction with the Silurian inlier, along the north edge of a broad but shallow drift-filled depression. This intervening belt of fen or lagg varies in width.

Malham Tarn Moss is important because it is one of few sites in the Craven limestone area that provides a continuous pollen and macrofossil sequence from the Late Devensian through to the present day, which allows information about the vegetational history of the limestone to be obtained and gives an understanding of how humans have influenced such vegetation development (Piggott and Piggott, 1959, 1963). Although some bog habitats are equalled or bettered elsewhere, it is considered by Cooper and Proctor (1998) that nowhere can such a range and so many of their characteristic species be seen within such a short distance. The whole Malham Tarn NNR is a remarkable microcosm of the mire habitats and species of upland Britain and it is the close proximity of such fen and raised bog habitats that is so interesting. Nevertheless there are rare bog species present and only in the last 20 years have species such as *Sphagnum riparium* and *Sphagnum warnstorfii* been found. There are other species at Malham Tarn that are rare in Britain as a whole, or that are at, or near to, the southern limit of their English range. Such species are discussed in Cooper and Proctor (1998) and Proctor (1974).

Description

Glacial landforms

Malham Tarn and Tarn Moss have formed in an area of lower ground that has allowed the focusing of mainly subglacial meltwater drainage from the north and west (Clark, 1967; Clayton, 1981). There has been some discussion and dispute as to the origin of some of the associated glacial landforms (Figure 8.73), such as the aligned drift forms of Spiggot Hill and the origin of the landforms in the drift belt to the south of the tarn. Raistrick and Illingworth (1949) called this drift a terminal moraine but Piggott and Piggott (1959) considered them to be drumlins. However, both Clark (1967) and Clayton (1981) consider them to be part of a glaciofluvial kame complex, with associated eskers. There has been similar dispute as to the origin of a 'delta' surface at 380 m behind Tarn Moss, as Clark (1967) suggested that this was subglacial in ori-

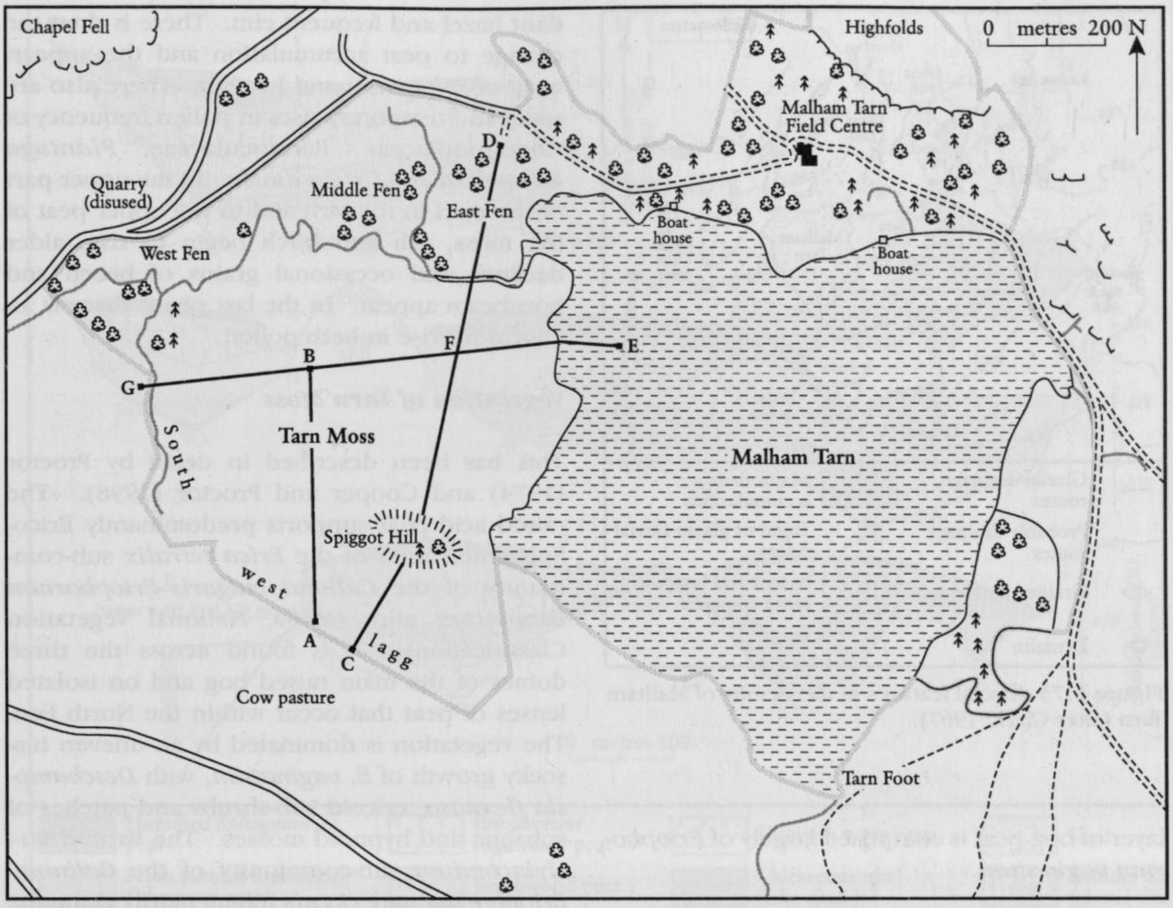


Figure 8.72 Malham Tarn and its surroundings showing the location of Tarn Moss and Fens and transect Lines shown in Figure 8.74 (after Cooper and Proctor, 1998).

gin, whereas Piggott and Piggott (1959) suggested that from its relationship with the basal Lateglacial clays under Tarn Moss it was a delta formed earlier than pollen zone 1 from sediments fed from the West End channel. It seems possible, however, that there was an earlier postglacial lake level at a higher level that overflowed through a channel at 379 m OD, which leads through to Great Close Mire (Clayton, 1981), although Clark (1967) considered this to be subglacial in origin. An even more likely origin is that a pro-glacial lake formed during deglaciation, which drained through this channel.

Stratigraphy in Tarn Moss

The main expanse of Tarn Moss is the result of a coalescence of three raised bog domes (north-east, north-west and south of Spiggot Hill), each initiated on a different part of the undulating drift surface. The stratigraphy was investigated by borings by Piggott and Piggott (1959), and is shown in Figure 8.74. On the drift in the lowest part of the basin there is a grey, silty clay with occasional lines of angular limestone or slate fragments, succeeded by a dark blue, compact clay, with occasional fruits of the aquatic plant *Potamogeton praelongus*. Above this is a layer of laminated silts and clays 20–30 cm thick, with lines of small stones and plant fragments, followed by a thin deposit of blue clay. These sediments grade into a deposit of almost pure calcium carbonate with abundant shell fragments, as well as undamaged molluscs (Stratton, 1956). This sediment is still being deposited in the Tarn, but under the Moss it is succeeded by fen peat. Over the central part of the basin *Sphagnum* appears and increases in abundance until the peat becomes almost pure *S. imbricatum*. Piggott and Piggott (1959, 1963) describe overlapping lens-shaped peat masses that represent compressed hummocks, whereas the uppermost

Figure 8.73 Glacial features in the vicinity of Malham Tarn (after Clark, 1967).

layer of bog peat is composed largely of *Eriophorum vaginatum*.

Pollen analysis in the cores

The pollen diagram is illustrated in Figure 8.75. In the lower grey clay there are only traces of organic matter and occasional damaged pine pollen grains and pre-Quaternary spores. The pollen of various herbs appear in the blue clays, with pine and birch exceeding the total herb pollen. Juniper is plentiful but never reaches over 60%. By contrast the ratio of tree to herb pollen is reversed in the overlying laminated layer, in which grasses, *Artemisia*, *Rumex*, *Plantago* and *Helianthemum* are well represented and juniper and willow become abundant. Above the laminated clay birch remains the only abundant tree pollen, but hazel is now also present, and juniper, willow and herbs decline. Pollen of water plants becomes more frequent and the character of deposition changes to calcium carbonate precipitation. About halfway through the marl deposition the proportion of hazel rises rapidly and becomes about five times more abundant than birch. Pine begins to increase again and then elm and oak. Pine then becomes dominant and is associated with abundant hazel and frequent elm. There is then the change to peat accumulation and the appearance of *Sphagnum* and heather. There also are small and temporary rises in pollen frequency of Chenopodiaceae, Ranunculaceae, *Plantago lanceolata* and *Urtica dioica*. In the upper part of the marl in the tarn and in the upper peat of the moss, ash and birch begin to rise, alder declines and occasional grains of beech and hornbeam appear. In the last phase there is an enormous rise in herb pollen.

Vegetation of Tarn Moss

This has been described in detail by Proctor (1974) and Cooper and Proctor (1998). The raised acid peat supports predominantly Erico-Sphagnion mire of the *Erica tetralix* sub-community of the *Calluna vulgaris–Eriophorium vaginatum* mire (M19a, National Vegetation Classification). It is found across the three domes of the main raised bog and on isolated lenses of peat that occur within the North Fen. The vegetation is dominated by an uneven tussocky growth of *E. vaginatum*, with *Deschampsia flexuosa*, ericoid sub-shrubs and patches of sphagna and hypnoid mosses. The *Vaccinium–Hylacomium* sub-community of the *Calluna–Eriophorum* mire occurs fragmentarily along the steep margin of Tarn Moss at its northern and eastern edge. Here the vegetation is dominated by *Calluna vulgaris* and *Vaccinium myrtillus*, with locally frequent *Vaccinium vitis-idaea* and *Rubus chamaemorus*. Bog pools are poorly developed, but where they do occur, they are of the *Sphagnum cuspidatum/recurvum* type (M2). Around the margins of the bog and where the surface of the peat has been disturbed by peat digging, *Molinia caerulea* often becomes predominant, forming the *Erica tetralix* sub-community of the *Molinia caerulea–Potentilla erecta* mire. Cooper and Proctor (1998) describe poor fens, where wet acidic peat close to the influence of mildly base-rich water is occupied by the *Carex echinata* sub-community of the *Carex echinata–Sphagnum recurvum* mire (M6a), rich in *Carex carta*. The pH of such areas is often little higher than the raised bog communities but cation content is slightly greater and mineralization of nutrients, especially N and P, is probably more rapid. The hydrochemistry is described further in Proctor (1995) and the rich fens and fen-carr are described in detail in Proctor (1974) and Cooper and Proctor (1998).

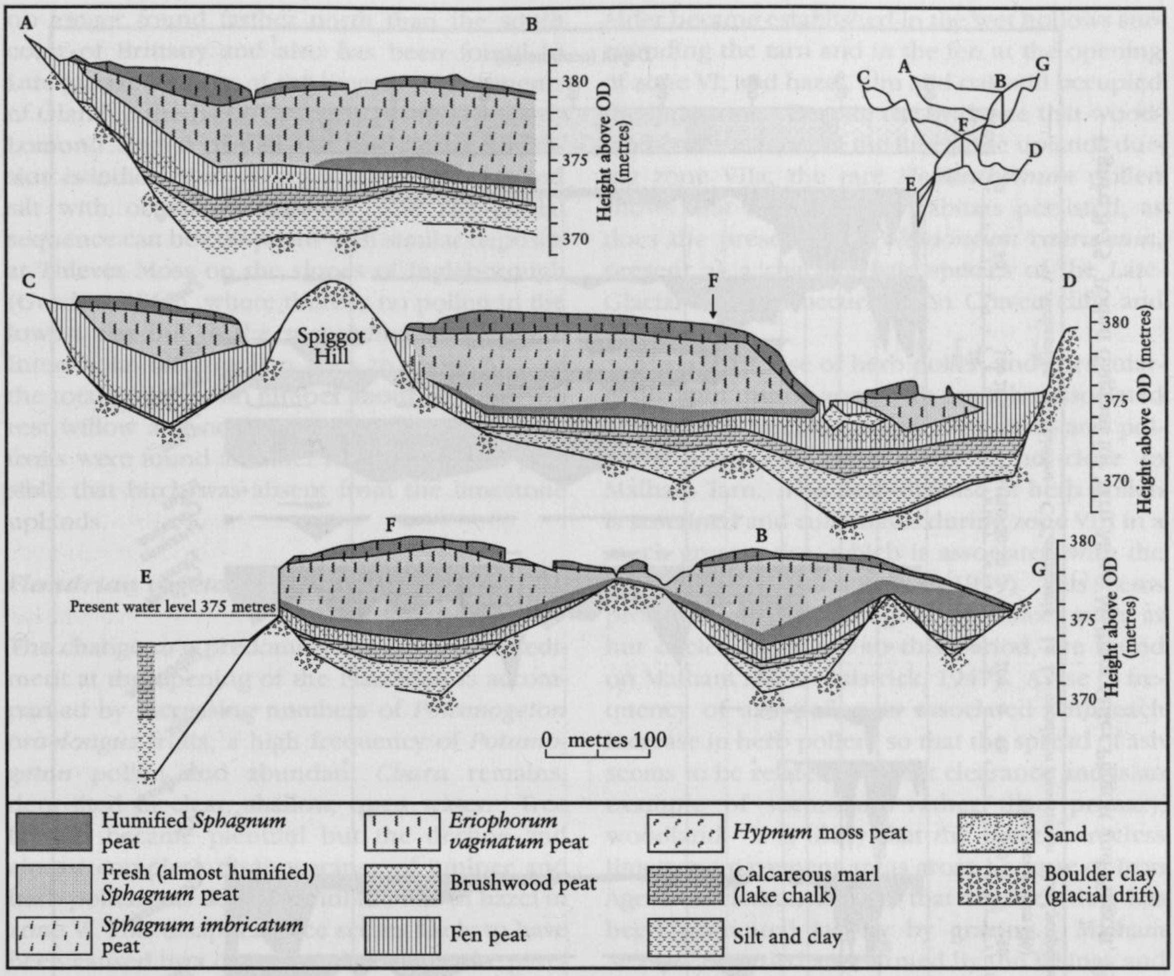


Figure 8.74 Stratigraphy of Tarn Moss, Malham (after Piggott and Piggott, 1959). The location of transects is shown in Figure 8.72.

Around the margins of the Tarn thin (*c*. 5 mm), spongy encrustations of tufa or Krustenstein (Kann, 1941) cover the rocks, which provides a harbour for many protozoa, invertebrates and algae (Lund, 1961; Holmes, 1965). The most conspicuous tufa-forming species, but not the most abundant, is *Rivularia haematites* (Pentecost, 1981).

Interpretation

Late-glacial vegetational history

Piggott and Piggott (1959, 1963) interpreted the lowermost succession in the pollen diagram as showing the Late-Glacial stages. In this period open-water extended over almost the whole area of the present tarn and the Tarn Moss and fens. The absence of organic matter in the zone I clays suggests that vegetation was sparse both in the tarn and in the surrounding uplands. In the Allerød (Windermere Interstadial), fruits of a pondweed in the clay and the proportional increase in birch pollen suggest that the tree may have been present locally, although a high proportion of the pollen grains are of the type characteristic of the dwarf species (*Betula nana*). The abundance of juniper pollen during zones II and III suggests that this was present on the limestone pavements, but the vegetation must have been open because of the high frequencies of grasses, sedges, *Artemisia*, *Rumex* and *Helianthemum* and the occurrence of *Armeria maritima*, *Alchemilla*, *Campanula rotundifolia* and *Poterium sanguisorba*. Rare grains of *Hippophäe rhamnoides* and *Ephedra distachya*, both plants restricted to the coast in western Europe today, are present. *Ephedra* is

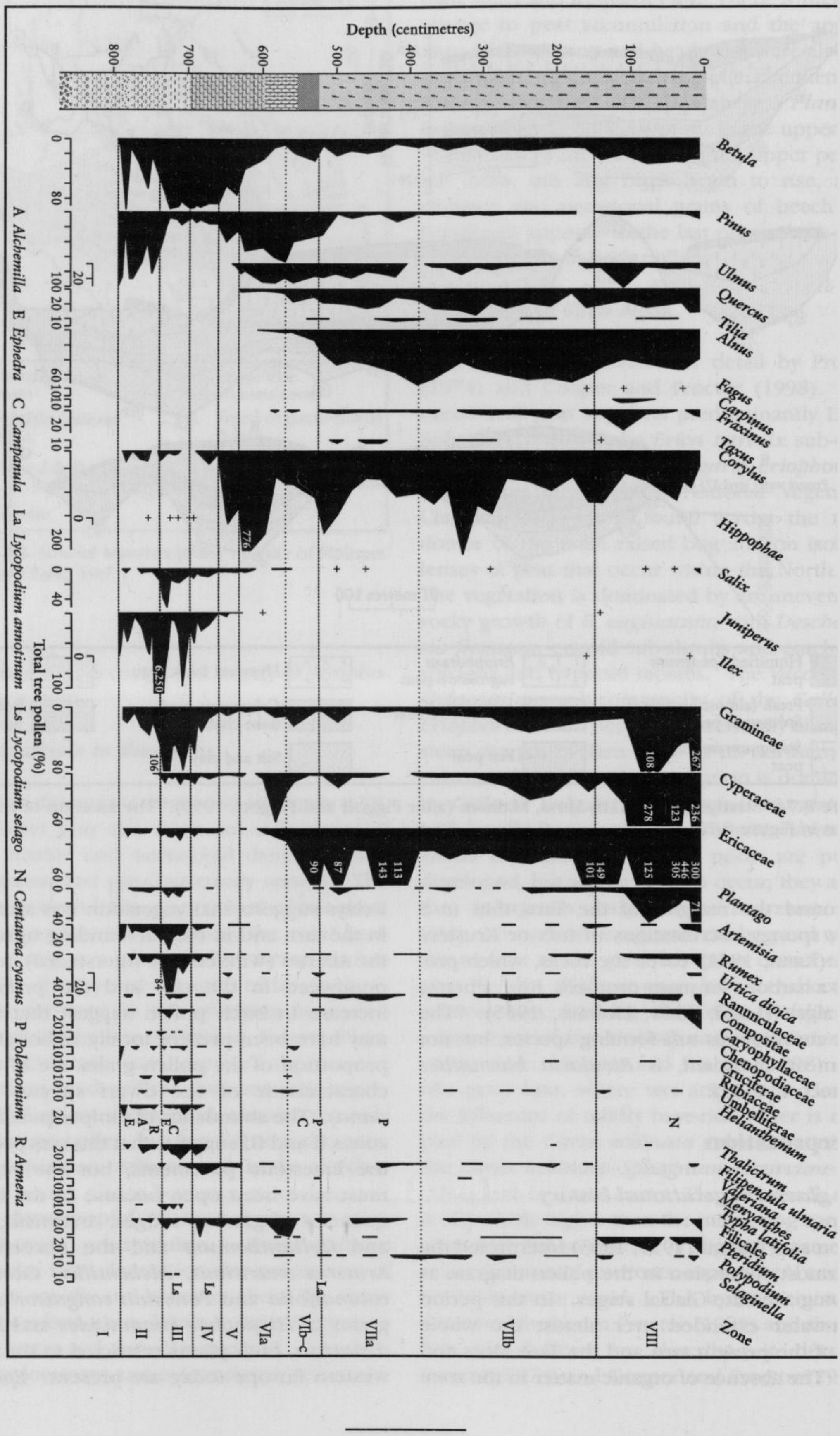


Figure 8.75 Pollen diagram for Tarn Moss, Malham (after Piggott and Piggott, 1959). See Figure 8.74 for key to the stratigraphical log.

no longer found farther north than the south coast of Brittany and also has been found in Late-glacial deposits of the limestone pavements of Öland in the Baltic (Iversen, 1954). The Loch Lomond Stadial part of the Late-glacial succession is indicated by 20 cm of laminated clay and silt with occasional stones. The Late-glacial sequence can be compared with similar deposits at Thieves Moss on the slopes of Ingleborough (Gosden, 1968), where there is no pollen in the lowest clay but in the succeeding Windermere Interstadial birch pollen rises to about 35% of the total pollen, with juniper about 20% and the rest willow and non-arboreal pollen. No birch fruits were found at either locality and it is possible that birch was absent from the limestone uplands.

Flandrian vegetational history

The change to a predominantly calcareous sediment at the opening of the Flandrian is accompanied by increasing numbers of *Potamogeton praelongus* fruits, a high frequency of *Potamogeton* pollen and abundant *Chara* remains, deposited in clear, shallow, open water. Tree birches became plentiful but the decline and almost complete disappearance of juniper and herb pollen was delayed until the rise of hazel in zone V. This disappearance seems likely to have been caused by a dense hazel cover on the pavements. The well-defined pine–hazel phase is unusual in the northern and western uplands, although it was found by Raistrick and Blackburn (1938) from Linton Mires (Wharfedale). Piggott and Piggott (1959) suggest that pine became established on the shallow, well-drained soils of the scars and pavements.

During zone VI by 8000–9000 years ago, sedges were able to spread over the Tarn Moss basin but only the basal part of the fen peat is free of birch and willow, so the spread of carr woodland was rapid. The presence of *Phragmites communis*, at close to its altitudinal limit in the Pennines, and *Lycopus europeus*, well above its limit at Malham, suggest that summer temperatures were warmer than at present. Before the end of zone VI there is a marked increase in *Sphagnum* and *Calluna*. Thin charcoal layers and the small rise in weed pollen (plantain, nettle and Chenopodiaceae) were thought likely to be associated with Mesolithic habitation (Piggott and Piggott, 1959), and microliths have been found around the tarn. Alder became established in the wet hollows surrounding the tarn and in the fen at the opening of zone VI, and hazel, elm and oak still occupied the limestone. Despite the evidence that woodland covered most of the limestone uplands during zone VIIa, the rare *Helianthemum* pollen shows that natural open habitats persisted, as does the presence of *Polemonium caeruleum*, present as a characteristic species of the Late-Glacial and still occurring on Craven cliffs and screes today.

The abrupt rise of herb pollen and particularly plantain during zone VIIb must be associated with a Late Neolithic forest clearance, and polished hand axes have been found close to Malham Tarn. This first increase of herb pollen is sustained and culminates during zone VIII in a much greater rise, which is associated with the Iron Age (Piggott and Piggott, 1959). This seems probable, as archaeological evidence, such as hut circles attributed to this period, are found on Malham Moor (Raistrick, 1947). A rise in frequency of ash pollen is associated with each increase in herb pollen, so that the spread of ash seems to be related to forest clearance and is an example of secondary, rather than primary, woodland. It is likely that the present treeless limestone pavement areas arose because of Iron Age tree clearances and that regeneration has been prevented largely by grazing. Malham Moor was settled and farmed by the Vikings and during the medieval period was part of the monastic estates that maintained large sheep flocks on the limestone uplands (Raistrick, 1947). By this period the limestone uplands were probably as open as today. Grazing by farm stock must have had some influence on Tarn Moss and fens since early times, tending to suppress tree growth and the fen-carr sallows and to increase the area of open fen (Cooper and Proctor, 1998). In 1791 the level of the Tarn was raised by about a metre by a weir construction and sluice (Holmes, 1965) and this inundated the lakeward edge of Tarn Moss, causing the erosion of a peat cliff and flooding of some old peat cuttings, and brought some ombrogenous peat within the influence of calcareous water. Tarn Moss probably has been used for rough grazing and was no doubt sporadically burnt, but the 1785–1786 map from Thomas Lister's estate book (Anon., 1785–1786; Cooper and Proctor, 1998) shows Tarn Moss divided into three compartments, which may have had different management histories. From the latter part of the

19th century onwards grip drains were put in radially across the main raised bog. In the late 1940s the vegetation was dominated by *Eriophorum vaginatum* with abundant *Deschampsia flexuosa*, with little heather, very little *Sphagnum*, except in flooded peat pits, and profuse growth of the mosses *Pohlia nutans* and *Tetrophis pellucida* reflecting recent severe burning (Cooper and Proctor, 1998).

Studies on the important tufa deposits around the shores of Malham Tarn (Pentecost, 1981) indicate that the deposit is continuously disturbed, broken up and redeposited. Its distribution is dependent on the distribution of littoral rocks and stable peat surfaces. Narrow grooves may be seen on the rocks beneath the tufa, giving the surface a gnawed appearance. Similar structures, called galets sculptés, have been ascribed to limestone corrosion caused by respiring algae and bacteria (Le Roux, 1908), however, these features at Malham appear to result from the burrowing activities of the caddis *Tinodes waeneri* (Holmes, 1965).

Conclusions

Tarn Moss is the best site in the Craven limestone uplands illustrating the vegetation development of the area from the Late-glacial to the present day and helps to provide a picture of the development of that landscape throughout this period. It also illustrates the influence of humans in such vegetational change. Unfortunately the pollen work was carried out in the 1950s and the site is important enough to benefit from a modern re-analysis of the succession. The location is important too, because of the contrast between closely associated, modern raised bog vegetation and fens, successional changes in such ecosystems, its complex groundwater development and the development of tufa around the margins of the modern tarn.

BOLTON FELL MOSS (NY 504 667) AND WALTON MOSS (NY 495 695)

POTENTIAL GCR SITE

D. Huddart

Introduction

There are many classic ombrotrophic bogs in Cumbria, known locally as 'mosses', or 'flows' if they are particularly wet. These sites have been an important source for our knowledge of Holocene vegetation, climatic change and the impact of humans on the vegetation record. Barber's (1981) research into the relationship between peat stratigraphy and climatic change as a formal palaeoecological test of the theory of cyclic bog regeneration was centred at Bolton Fell Moss. Here undisturbed stratigraphical sections were analysed for macrofossils to characterize the changing assemblages of bog plants, dated and correlated by means of a master profile that was subject to pollen analysis at close intervals and ^{14}C dated. The close correlation between known climatic changes since the Middle Ages (Lamb, 1977) and the macrofossil and peat stratigraphy changes shown to have occurred across the bog as a whole, falsified the theory of autogenic cyclic regeneration (Barber, 1981). Barber thus demonstrated that the climatic signal was the overriding important factor. Work also has been undertaken on the human impact on the vegetation of the area using the pollen record (Dumayne, 1992) and detailed research has taken place into the sensitive, high-resolution record of Late Holocene climatic changes (Barber *et al.*, 1994b).

Work at Walton Moss (NY 504 667) has allowed a record of natural and cultural change to be made comparable with that at Bolton Fell Moss (NY 490 690) (Barber *et al.*, 1994a; Dumayne-Peaty and Barber, 1998; Hughes *et al.*, 2000), and has shown the conservation problems that beset this rare habitat (Barber, 1993). The work at these two locations forms part of an important overview of the role of climatic change and human impact on the landscape in northern Britain (e.g. Donaldson and Turner, 1977; Davies and Turner, 1979; Turner, 1979; Blackford and Chambers, 1991; Dumayne, 1992; Barber *et al.*, 1993; Stoneman, 1993; Barber, 1994; Tipping, 1995; Chambers *et al.*, 1997; Barber *et al.*, 1999; Mauquoy and Barber, 1999a, b).

Description

Walton Moss (altitude 95 m OD) is possibly the most intact ombrotrophic, raised mire in England and is 283 ha in extent. One kilometre to the south, and separated by a small valley, is Bolton Fell Moss, where peat resources have been worked extensively. It covers 365 ha at an altitude 110 m OD. The plant communities in both are dominated by *Sphagnum magellan-*

icum, which macrofossil analyses show as having only achieved dominance in the past 1000 years, replacing the former community dominated by *S. imbricatum*. The peat stratigraphy shows shallow lake muds succeeded by *Phragmites* reed, with varying amounts of woody material and moss remains. These deposits typically are only *c.* 1 m thick and are succeeded in turn by humified peats rich in cotton-grass and heather remains and then less humified *Sphagnum* peats, which often are very pure and contain thin, algal mud-rich bands. The pollen and macrofossil diagrams for these mosses are given in Figures 8.76 and 8.77 but more detailed description is found in Barber (1981), Barber *et al.* (1994a, b) and Hughes *et al.* (2000).

Interpretation

The interpretation of the peat stratigraphy can be looked at in terms of the influence of climatic change, mire development and how much human impact there has been on the vegetation.

The proxy palaeoclimatic record

Bolton Fell Moss was used as a palaeoecological test site, whose evidence led to the rejection of the cyclic peat-bog regeneration model and established climatic phase theory (Barber, 1981). Barber conducted macrofossil and other analyses on seven peat faces. There are no lenticular structures that would be produced by cyclic regeneration and the dominant stratigraphy is layered, with only moderate relief. Relatively dry bog conditions, dominated by *S. imbricatum*, followed by an intermediate state (*S. cuspidatum–S. papillosum* phase) and then by very wet conditions, dominated by *S. magellanicum*, were found in 16 of the 21 monoliths analysed. These data were used to falsify the cyclic regeneration theory of Sernander (1908) and Osvald (1923) by relating the changes in stratigraphy and macrofossil assemblage to the known climatic variation of the last few centuries, and this was developed into the climatic phase theory, i.e. raised bog growth occurs in climatically forced phases. The curve of bog surface wetness derived from this work is related closely to Lamb's (1977) summer wetness index (see Figure 8.78).

Bolton Fell Moss also was used to characterize and date the main humification change in a transect of bogs across Europe. The quadrat and leaf count method (Haslam, 1987) was used to produce relative hydroclimatic curves. At Bolton Fell Moss the division between the upper and lower peats was characterized in many places by a horizontal layer of green pool muds indicative of widespread surface flooding, and it was one of several changes akin to recurrence surfaces and hence major humification changes. The summary macrofossil diagram and hydroclimatic curve derived from Dupont's method (1986) are shown in Figure 8.79, where wetter and drier phases are indicated (see Haslam, 1987; Barber *et al.*, 1994a, b). A detailed palaeoecological and multivariate analysis of the Bolton Fell Moss profile was undertaken by Barber *et al.* (1994a) and this has shown that the data possess a coherent and robust structure and that the variations in the data are related to the bog water table and hence through that to climate.

Table 8.14 shows the timing of the main wet shifts in the peat stratigraphy at both mosses (Hughes *et al.*, 2000). There seems to be a reasonable degree of correspondence between the two climate archives, but some significant differences are present. Both mires preserve records of four wet shifts between *c.* 4300 and *c.* 2200 calibrated (cal.) years BP that appear to be broadly in phase and this section of the climate reconstruction has been independently radiocarbon dated. The differences are discussed in Hughes *et al.* (2000).

Mire development

In Walton Moss a brief fen phase began shortly before *c.* 10 200 cal. years BP, with the transition to oligotrophic, *Eriophorum*-dominated bog occurring at *c.* 9900 cal. years BP. There also is a significant increase in the frequency of macroscopic charcoal in the pioneer raised-bog phases, which suggests that the dry hummocky surface was burnt periodically. This highly humified peat layer represents a phase when the pioneer raised mire lacked a stable water table. The almost universal occurrence of such a bed of highly degraded *Eriophorum–Calluna* peat lying above the fen levels and below fresher *Sphagnum* peats lends support to the hypothesis that one or more phases of peat surface desiccation and humification favour raised water-mound formation as a consequence of the production of a relatively impermeable, finely comminuted peat layer (Hughes *et al.*, 2000). At Walton Moss the macrofossil remains suggest

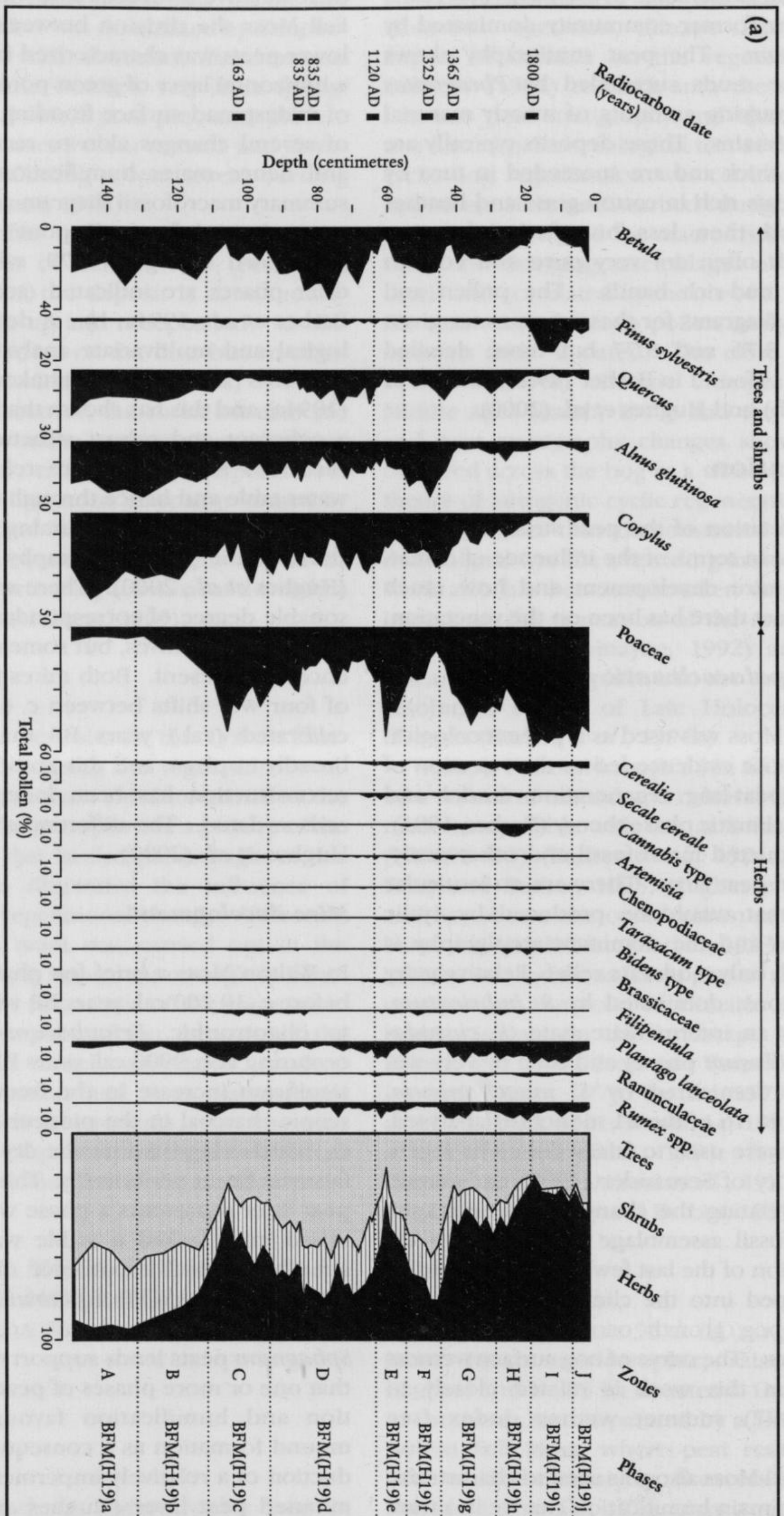


Figure 8.16a Summary pollen diagram from Bolton Fell Moss (after Barber, 1981; Dumayne, 1992)

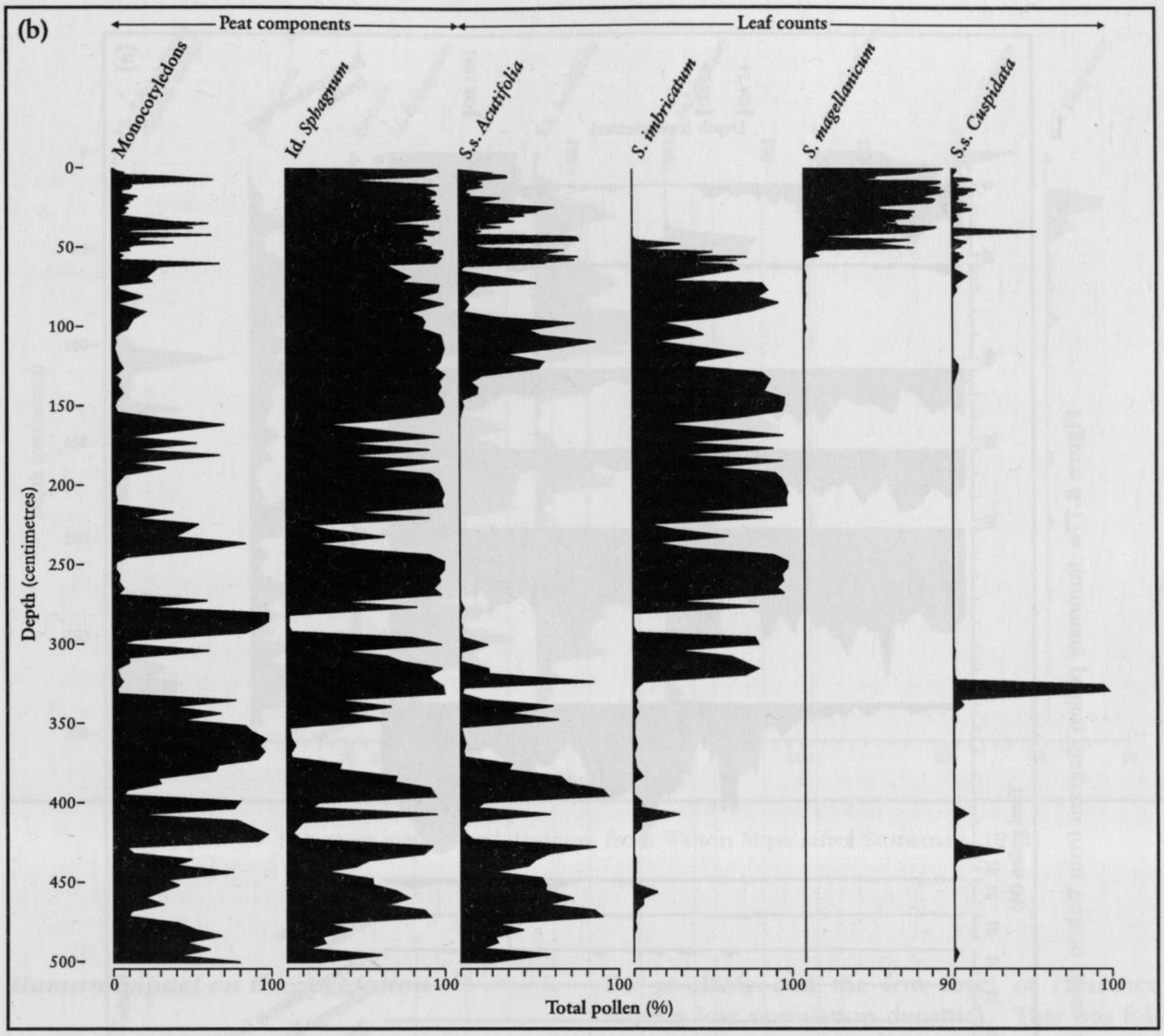


Figure 8.76b Summary macrofossil diagram from Bolton Fell Moss (after Barber, 1981; Dumayne, 1992).

that 2000 years elapsed before the first near-surface water tables were established, as indicated by the arrival of *Sphagnum subnitens* at *c.* 7800 cal. years BP. This 2000 year interval may reflect either a prolonged phase of low effective precipitation in the early Holocene and a major switch to wetter climates at *c.* 7800 cal. years BP, or the time taken for sufficient well-humified ombrotrophic, catotelmic peat to develop to maintain a high water table. A third possibility is that the climatically insensitive phase naturally would be shorter than 2000 years but at the point at which the mire could potentially respond to wet shifts by forming pools, effective precipitation levels were low. It seems, however, that this first wet shift at Walton Moss was a dramatic response to increased effective precipitation. This very wet interval occurred between *c.* 7800 and *c.* 6800 cal. years BP. The marginal development of the mire was different and fen conditions lasted longer than in the main basin and dry, *Eriophorum–Calluna* mire developed here whereas the main mire centre had a *Sphagnum*-rich flora indicative of pool, lawn and hummock microforms. At *c.* 2800 cal. years BP the first appearance of *Sphagnum imbricatum* marked the point at which the water table stabilized at the marginal core site and correlates with a significant increase in surface wetness in the main mire basin.

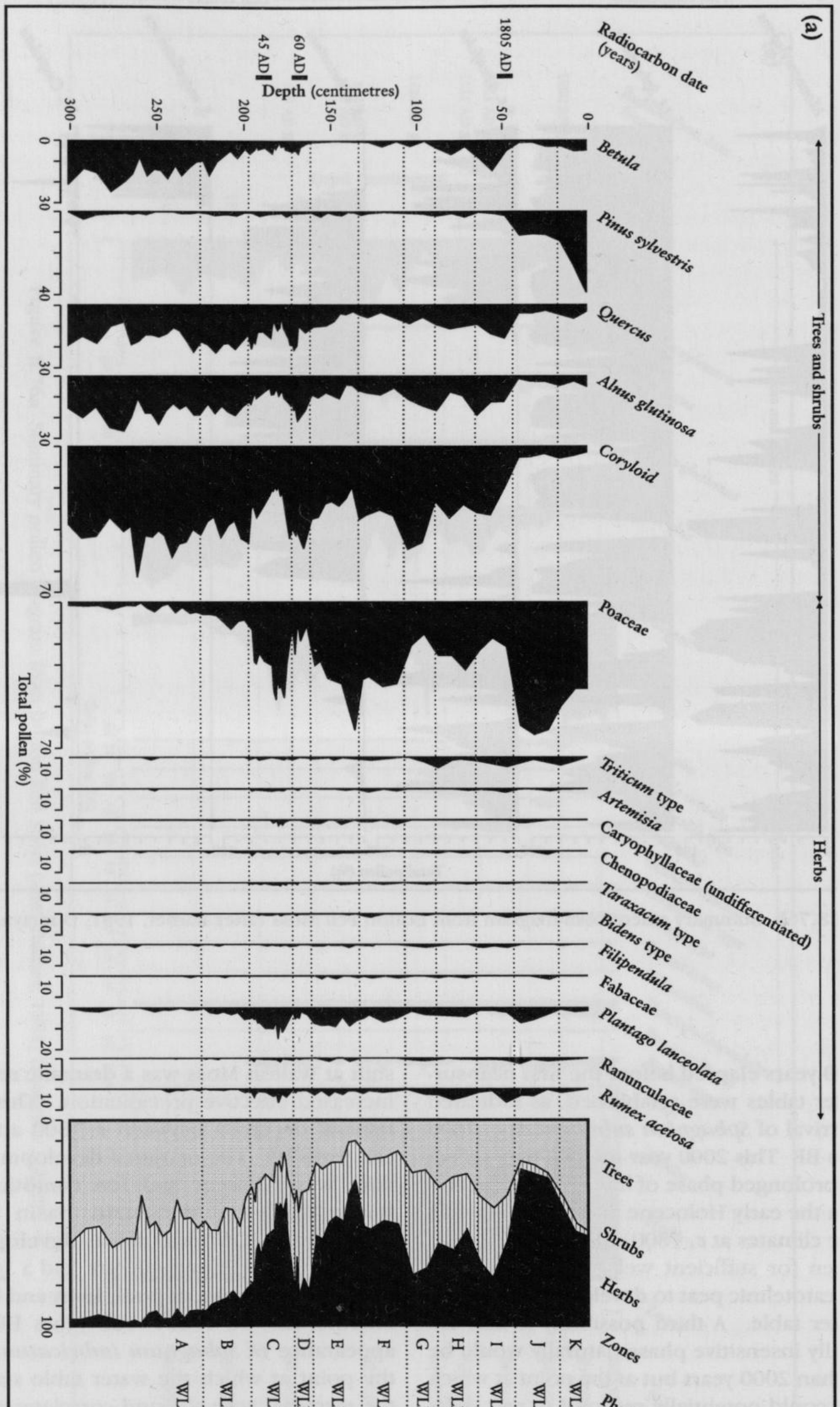


Figure 8.77a Summary pollen diagram from Walton Moss (after Stoneman, 1993)

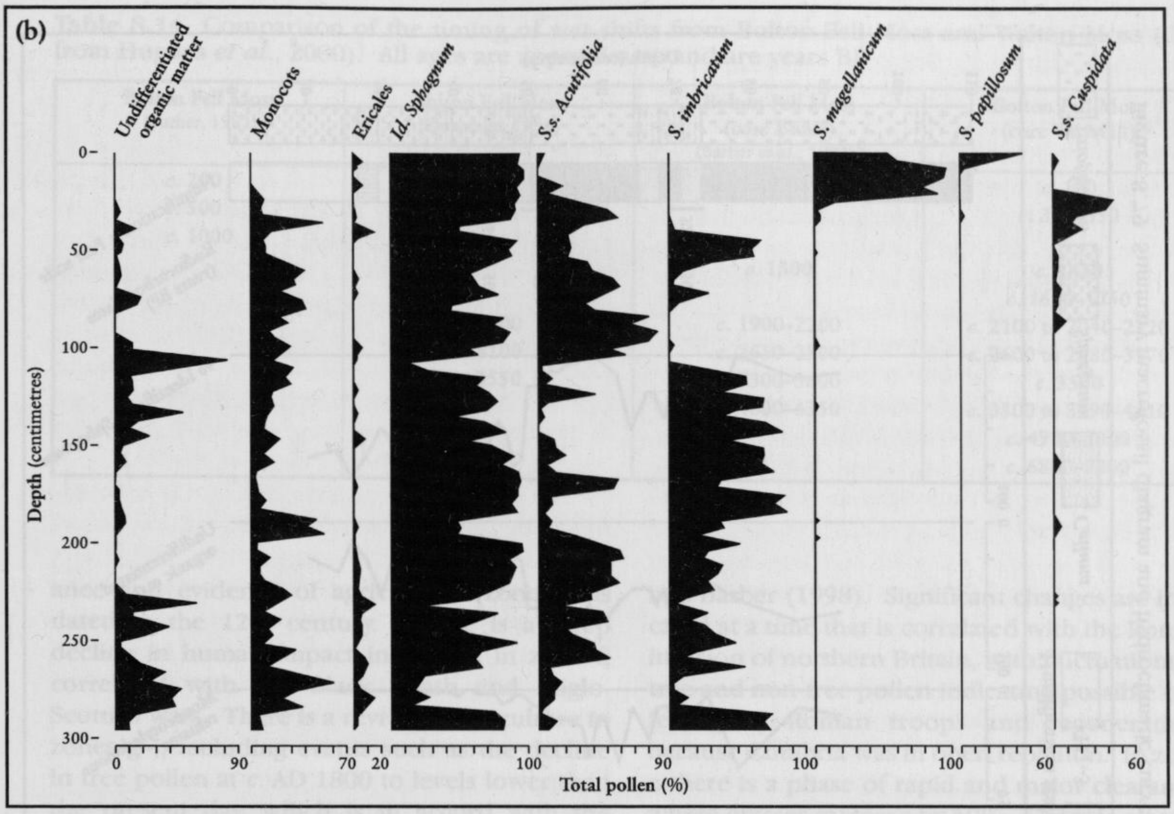


Figure 8.77b Summary macrofossil diagram from Walton Moss (after Stoneman, 1993).

Human impact on the vegetation

Pollen analysis by Barber (1981), Dumayne (1993) and Dumayne-Peaty and Barber (1998) indicate the human impact on the landscape. In zones a and b at Bolton Fell Moss (Figure 8.76) the indications are of a late prehistoric period of subdued agriculture. A date of *c.* 580 cal. years BC at BFMa and the low level of clearance implies low population densities. This was followed by a major expansion of cleared land in Romano-British times (zone c) indicated by grass and plantain pollen reaching levels not seen again until medieval times. After the Roman withdrawal (zone d) there was some increase in woodland and the marked expansion of clear-

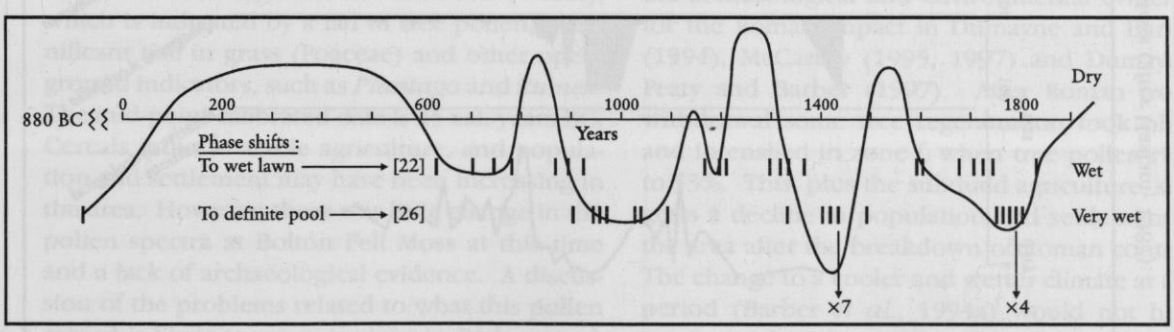


Figure 8.78 Surface wetness curve for Bolton Fell Moss (after Barber, 1981).

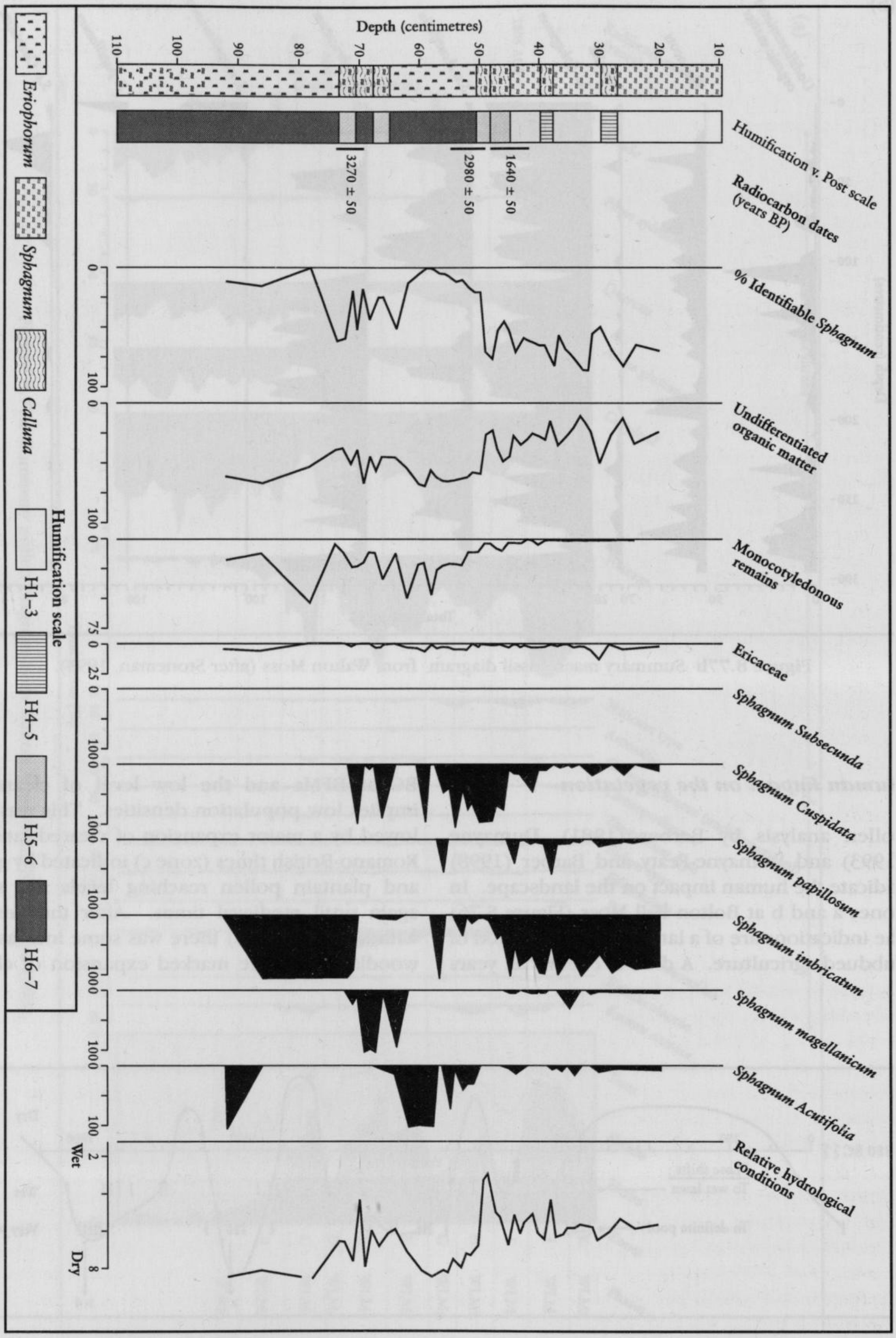


Figure 8.79 Summary macrofossil diagram and hydroclimatic curve for a major humification change at Bolton Fell Moss (after Haslam, 1987).

Table 8.14 Comparison of the timing of wet shifts from Bolton Fell Moss and Walton Moss (data from Hughes *et al*., 2000). All ages are approximate and are years BP.

Bolton Fell Moss (Barber, 1981)	**Bolton Fell Moss** (Stoneman, 1993)	**Bolton Fell Moss (core BFMJ)** (Barber *et al*., 1994b)	**Bolton Fell Moss (core WLM11)**
c. 200			*c*. 100
c. 500	*c*. 350		*c*. 300–350
c. 1000			
		c. 1300	*c*. 1450
			c. 1650–1750
	c. 2400	*c*. 1900–2200	*c*. 2100 to 2040–2320
	c. 3100	*c*. 2650–2900	*c*. 2600 to 2680–3170
	c. 3550	*c*. 3300–3600	*c*. 3500
		c. 4000–4350	*c*. 3800 to 3990–4410
			c. 4900–5300
			c. 6800–7800

ance and evidence of agriculture (zone e) is dated to the 12th century. There is a steep decline in human impact indicators in zone f, correlated with the Black Death and Anglo-Scottish wars. There is a revival of agriculture in zones g–j, including events such as the decline in tree pollen at *c*. AD 1800 to levels lower than the present day, which is in accord with the Napoleonic 'plough-up' campaign and the planting of Scot's pine, dated to about AD 1800, which was accompanied by increased tree species such as beech, elm and ash, which were planted on country estates at that time.

Work at Walton Moss by Dumayne (1992) and Dumayne-Peaty and Barber (1998) is summarized in Figure 8.77. In the earliest phases, the Bronze Age and Early Iron Age landscape was largely forested and typified by small, temporary clearances. Pastoral and/or arable agriculture may have been practiced and affected mainly hazel, which may indicate that secondary woodland and the understorey were being cleared. The Late Iron Age landscape was characterized by a sudden, major forest clearance (WLMc), which is indicated by a fall in tree pollen, a significant rise in grass (Poaceae) and other open-ground indicators, such as *Plantago* and *Rumex*. The mid-point calibrated date is 45 cal. years BC. Cereals indicate arable agriculture, and population and settlement may have been increasing in the area. However, there was little change in the pollen spectra at Bolton Fell Moss at this time and a lack of archaeological evidence. A discussion of the problems related to what this pollen record is likely to mean is given in Barber *et al*. (1994a), Dumayne (1992) and Dumayne-Peaty and Barber (1998). Significant changes are indicated at a time that is correlated with the Roman invasion of northern Britain, with fluctuations in tree and non-tree pollen indicating possible tree felling by Roman troops and regeneration, because Cumbria was in open rebellion. In zone e there is a phase of rapid and major clearance, where grasses increase by 60%. There is a parallel decrease in trees and an increase in taxa indicative of pastoral and arable agriculture. This is correlated with the Romano-British period and the beginning of this phase has been dated to *c*. AD 165 cal. years. The clearance was the result of all or some of the following factors: Roman occupation and native, civilian settlement; the construction of Hadrian's Wall and its related structures needed timber, although originally it was built of turf, indicating an already open landscape; open ground was needed for visibility and intercepting of raiders as well as signalling across the wall and to the forts in front and behind, and for agriculture. There has been a lively debate generated by the reconciliation of the archaeological and environmental evidence for the Roman impact in Dumayne and Barber (1994), McCarthy (1995, 1997) and Dumayne-Peaty and Barber (1997). After Roman troop withdrawal some tree regeneration took place and intensfied in zone f, when tree pollen rises to 53%. This, plus the subdued agriculture, suggests a decline in population and settlement in the area after the breakdown of Roman control. The change to a cooler and wetter climate at this period (Barber *et al*., 1994a) would not have encouraged settlement but woodland regeneration. The later phases of the pollen diagram from Walton Moss are similar to Bolton Fell Moss and are discussed in detail in Dumayne-Peaty and Barber (1998). The decline of *Sphagnum imbricatum* from both mosses, dated to AD 1030–1400 m cal. years may have resulted from interspecific competition between *Sphagnum* species during the 'Early Medieval Warm period' and the 'Little Ice Age' (Mauquoy and Barber, 1999b), as first suggested by Barber (1981). At Walton Moss the decline also may have been the result of climatic change, as its disappearance is associated with large fluctuations of water tables, followed by a phase of increased mire surface wetness (Stoneman, 1993).

Conclusions

Bolton Fell Moss and Walton Moss are important sites in northern England, where a wealth of detailed palaeoecological and palaeoclimatic information has been obtained from a major programme of pollen, macrofossil and dating work. This has allowed the theory of cyclic peat bog regeneration to be disproven and established the climatic phase theory of bog development, controlled by climatic change. The impact of humans on the landscape also has been deduced in detail and the links with the archaeological and documentary records established. The conservation of Walton Moss is important because this raised mire habitat is extremely rare in England.

However, it would be a simplification to suggest that the Holocene proxy climate record is composed solely of a series of coherent, synchronous and far-reaching events. Distinct regional climatic gradients have been identified across Europe during the Holocene Epoch and the relationship between climate and peat stratigraphy may be more complex than was recognized previously. Future goals for peat-based palaeoclimatic research must be to identify the relative importance of the temperature and precipitation elements of the effective precipitation signal and to couple these records to improved tephra-based chronologies (Hughes *et al.*, 2000). Further work also is required to confirm the millennial-scale cycle of wet shifts at Walton Moss and to verify the striking level of agreement with the occurrence of ice-rafted debris events found in ocean cores, which may represent evidence of ocean-driven forcing of the regional climate.

HARTLEPOOL (NZ 520 313)

A. Plater

Introduction

Hartlepool Bay lies to the north of the Tees Estuary and immediately south of Hartlepool Headland (NZ 520 313). The foreshore contains the Hartlepool submerged forest, which is variably exposed along the coast between Seaton Carew and Hartlepool docks. Trechmann (1947) described the ancient forest or peat bed at Hartlepool as 'the most extensive and best known on the north-east coast of England', but marine erosion and engineering works have significantly altered the extent of peat exposure. The submerged forest is easily recognizable as a dark peaty outcrop that contains tree stumps and branches. In addition to its antiquarian interest, the peat bed forms part of a complex intercalated sequence of peats, estuarine and marine sediments, brackish and freshwater organic muds and glacial diamicton. Hence, the sedimentary sequence provides evidence of Holocene coastal change and relative sea-level trends in a crucial region between areas of crustal subsidence to the south and uplift to the north.

Description

The distribution of peat and associated minerogenic sediments at Hartlepool is both discontinuous and complex. Peat beds are exposed in a discontinuous belt from North Sands, Hartlepool, to the foreshore opposite Long Scar at West Hartlepool (Raistrick and Blackburn, 1932; Trechmann, 1936, 1947; Smith and Francis, 1967) (Figure 8.80). These previous studies noted that the peat was generally about 1 m in thickness, but reached a maximum thickness of approximately 5 m in the vicinity of Hartlepool dock. Indeed, excavations near The Slake in 1878 recorded peat up to 2.5 m thick with isolated horns of ruminants as well as tree trunks up to 6 m long and 1 m diameter (Cameron, 1878). Perhaps the best exposure today of the southern part of the submerged forest surveyed by Trechmann (1947) is in the middle part of the beach at Hartlepool Bay, where a depression in the pre-Holocene sediments contains freshwater organic deposits (Horton *et al.*, 1999c). Over 1 m metre of woody peat can be found in the

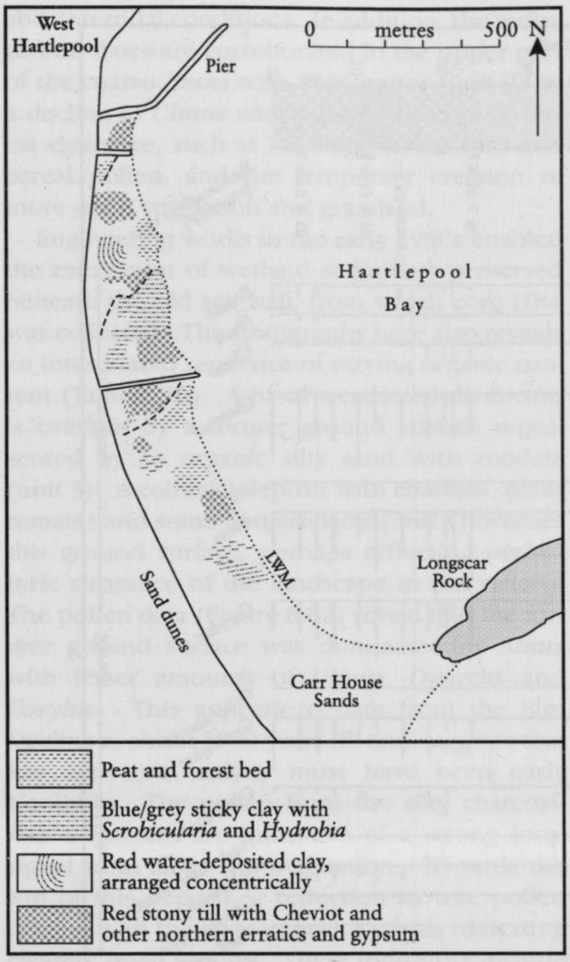


Figure 8.80 Former outcrop of the Hartlepool submerged forest bed between Hartlepool Headland and Long Scar. From Trechmann (1947).

centre of this depression beneath the present-day beach sand. From here, the surface of the underlying diamicton rises seawards, attenuating the peat until the beach sand directly overlies diamicton. Other thin woody peat beds, no greater than approximately 20–30 cm in thickness, are also found within confined basins within Hartlepool Bay, resting either on diamicton or a thin layer of micrite or *Chara* marl.

Although the depressions noted above contain only freshwater sediments, other relict channel features appear to have been open periodically to a marine influence during the Holocene Epoch. The peat beds here are made up of reed material, mainly *Phragmites*, with little wood. Some peaty units are clay-rich organic horizons rather than true peats, and there are several distinct bands of silt and clay. Indeed, Trechmann (1947) noted the presence of an underlying blue-grey clay (of marine origin) that was penetrated by roots from the peat. At these sites, the extent of marine erosion has been considerable but earlier sea defence works afforded some protection; a high-level unit of grey silty clay is preserved as the uppermost sedimentary unit.

The results of micropalaeontological analyses on sediments from two channel fills are representative of the palaeoenvironmental data preserved at Hartlepool Bay (Innes *et al.*, 1993; Horton *et al.*, 1999c). A core from the lower part of the beach, WH19 (NZ 5209 3138) reveals an intercalated stratigraphy of organic muds and minerogenic units (Table 8.15). Selected pollen and summary diatom results, together with five radiocarbon dates, are given in Figure 8.81. The lowest silty *limus* is a former land surface beneath a basal peat (*limus* with herbaceous roots). The sand and overlying silty *limus*, as well as the thick blue-grey silty clay (unit 7), are clearly marine deposits from their predominantly polyhalobous diatom assemblage. These were deposited around 6000 years BP and between approximately 5800 and 5500 years BP respectively. The upper part of the blue-grey silty clay preserves a change towards less saline conditions before the transition to the more brackish water conditions of the overlying fine *limus* with *Phragmites*. The silty *limus* between depths of 4 and 10 cm (unit 11) marks a return to marine clastic sedimentation with dominant poly- and mesohalobous diatom taxa. The pollen data (Figure 8.81) are indicative of deposition in an estuarine and saltmarsh environment under vari-

Table 8.15 Stratigraphy for WH19 (data from Horton *et al.*, 1999c)

Unit	Depth (cm)	Description
12	0–4	*Limus* with herbaceous roots
11	4–10	Silty *limus*
10	10–14	Fine *limus*
9	14–22	Coarse *limus* with *Phragmites*
8	22–24	Fine *limus* with *Phragmites*
7	24–51	Blue-grey silty clay
6	51–55	Coarse *limus*
5	55–58	Silty *limus*
4	58–59	Sand
3	59–63	*Limus* with herbaceous roots
2	63–75	Silty *limus*
1	75+	Stiff clay

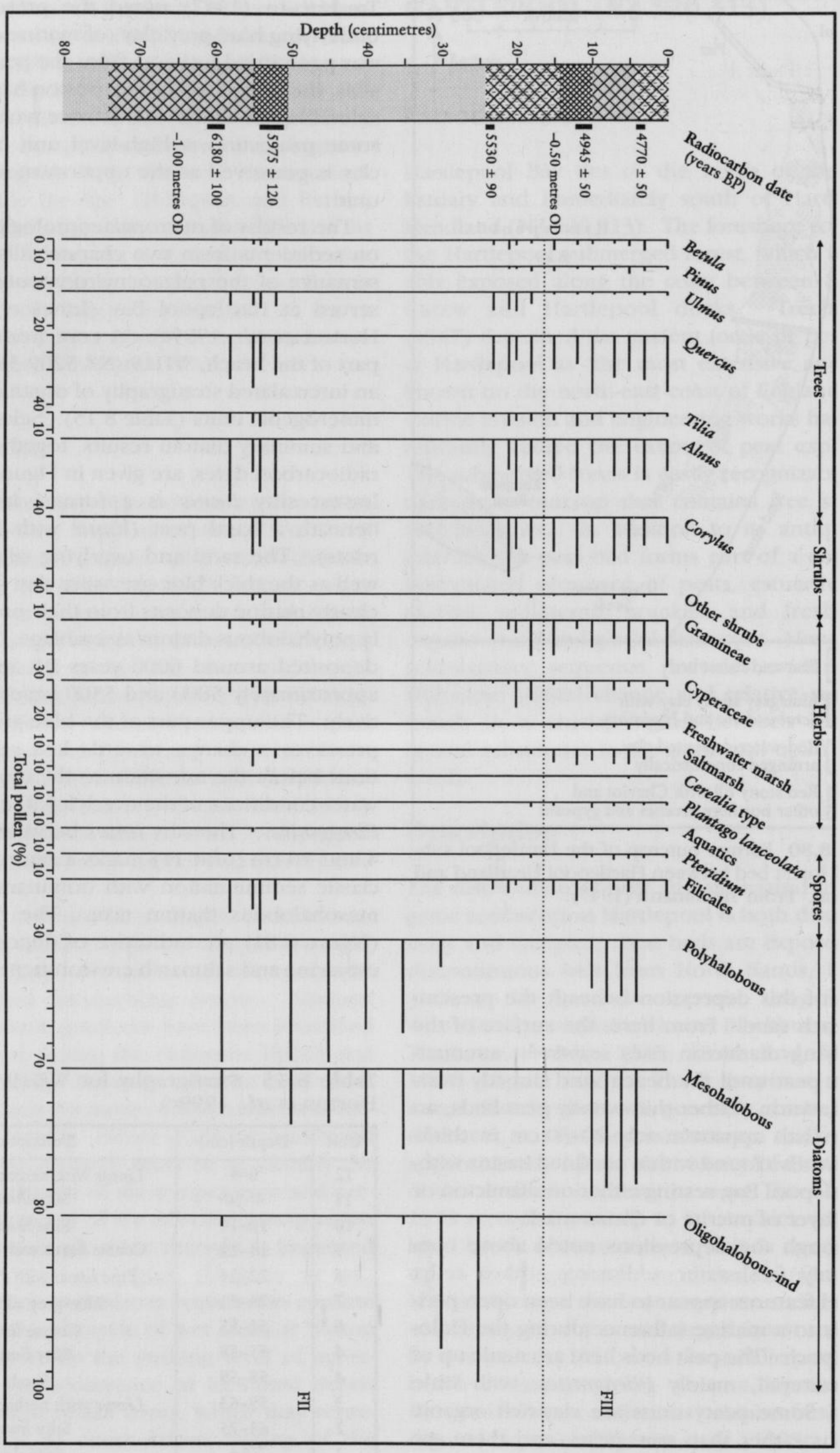


Figure 8.81 Summary stratigraphy, pollen, diatom and radiocarbon data for site WH19. After Horton *et al.* (1999c). See Table 8.15 for a description of the stratigraphy.

able intertidal conditions. In addition, the radiocarbon dates are corroborated in the upper part of the coarse *limus* with *Phragmites* (unit 9) by a decline in *Ulmus* with other indicators of forest clearance, such as *Plantago lanceolata* and cereal pollen, and the temporary creation of more open vegetation and grassland.

Engineering works in the early 1990s enabled the excavation of wetland sediments preserved beneath the old sea wall, from which core HB4 was collected. The stratigraphy here also reveals an intercalated sequence of varying organic content (Table 8.16). A basal weathered diamicton is overlain by a former ground surface represented by an organic silty sand with rootlets (unit 3). A colluvial deposit with charcoal, plant remains and some flint artefacts (unit 4) overlies this ground surface, perhaps reflecting prehistoric clearance of the landscape in the vicinity. The pollen data (Figure 8.82) reveal that the former ground surface was dominated by *Alnus* with lesser amounts of *Ulmus*, *Quercus* and *Corylus*. This appears to date from the Elm Decline at about 5000 years BP and suggests that any clearance activity must have been early Neolithic. The pollen from the silty charcoal-rich colluvium are indicative of a strong local signal from alder carr vegetation. Towards the top of this deposit, a reduction in tree pollen occurs, with the appearance of weeds indicating cleared, open ground. These indicators include *Plantago lanceolata*, *Taraxacum*-type and *Pteridium*. Evidence of local woodland destruction also is found in the lower part of the overlying organic sediments in the form of cereal pollen with *Pteridium*. Similar evidence of major Neolithic clearance activity also is present at other sites around Hartlepool Bay, e.g. charcoal with weed and cereal pollen in peat from The Slake dated to 5240 ± 70 years BP (HV 3459) (Horton *et al.*, 1999c).

The organic sediments in HB4 are mainly organic muds with an increasing content of *Phragmites* reeds. The stratigraphy is indicative of a rise in the groundwater table and the consequent formation of freshwater ponds. These ponds contain a great deal of inwashed cultural debris, mainly from the Bronze Age, in association with butchered bones of domesticated animals, implements and evidence of construction. Hence, a settlement of some kind is inferred. A continued rise in water level led to the site becoming an area of marsh with ponds and reed beds. A peat with a high marine silt content is indicative of periodic flooding by the sea at this time. Indeed, the pollen contains saltmarsh indicators such as several Chenopodiaceae, *Armeria* and *Aster*-types. Saltmarsh and coastal pollen types, together with freshwater aquatics, increase abruptly at a date of 2865 ± 75 years BP before the deposition of the silty grey estuarine clay (unit 10).

In addition to the evidence for Holocene coastal and sea-level change preserved within the peats and associated minerogenic sediments, the submerged forest bed has been the subject of antiquarian interest for nearly 200 years (Sharp, 1816). A wealth of fossil information has been recovered, ranging from freshwater and estuarine Mollusca indicative of warmer climatic conditions (Trechmann, 1947) to vertebrate remains including cod, pochard, roe and red deer, aurochs, domesticated cow, pig and wild boar. A red deer antler from Trechmann's collection yielded ages from 8100 ± 180 ^{14}C years BP (BM-90) to 8700 ± 180

Table 8.16 Stratigraphy for HB4 (data from Horton *et al.*, 1999c)

Unit	Depth (cm)	Description
10	0–17	Slightly organic clayey silt
9	17–40	Silty-clayey *limus* with some *Phragmites*
8	40–45	Laminated light grey-brown silty clay with some *limus* and *Phragmites*
7	45–58	Slightly clayey *limus* with herbaceous *detritus* and *Phragmites*
6	58–66	Woody detrital peat with *limus* and *Phragmites*
5	66–71	Dark brown *limus* with charcoal fragments and herbaceous *detritus*
4	71–76	Light grey, slightly organic silty clay with charcoal and some herbaceous *detritus*
3	76–79	Minero-organic sandy silt with plant rootlets and charcoal
2	79–82	Very sandy clay with some herbaceous rootlets
1	82+	Sandy blue clay with pebbles

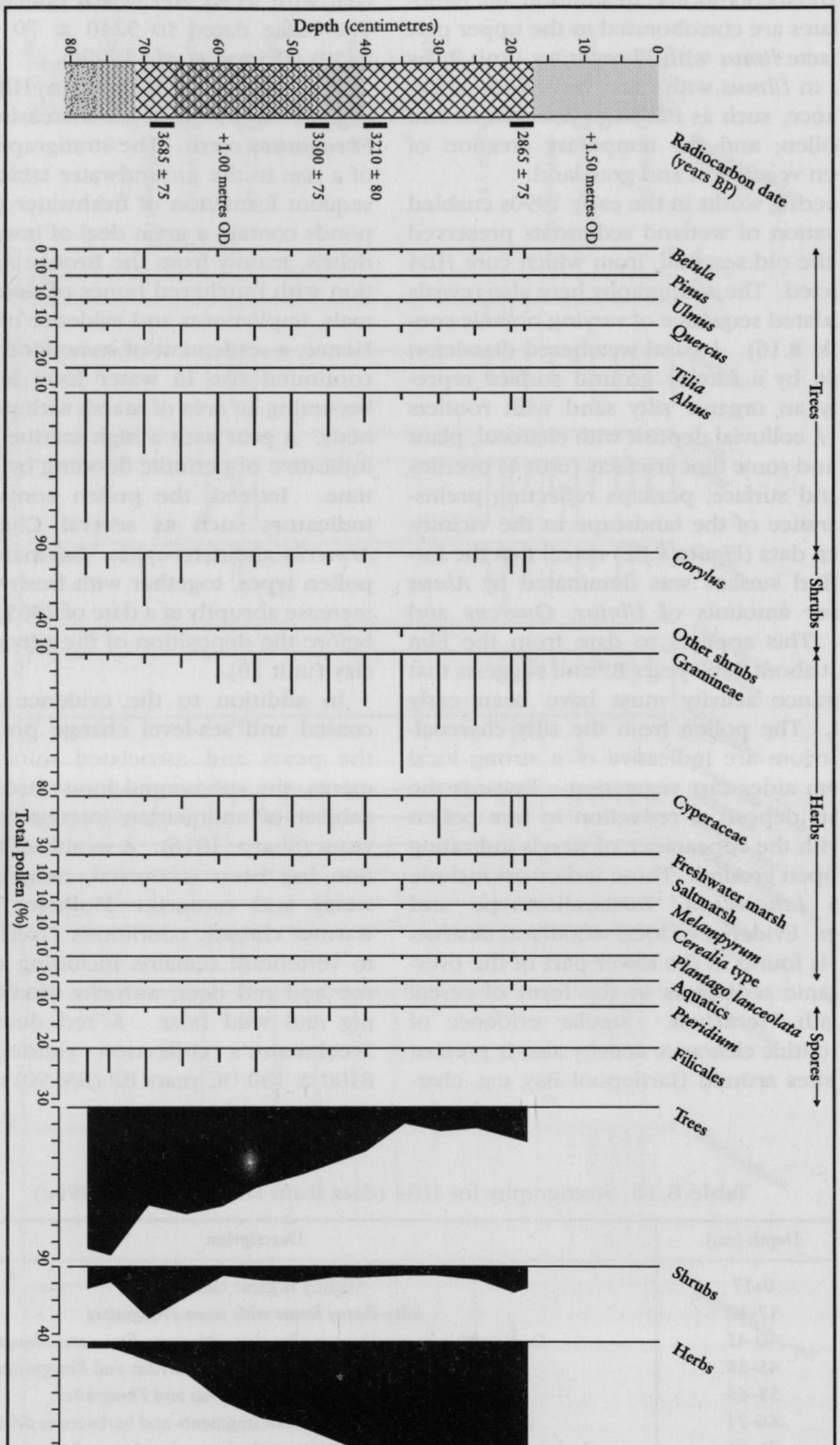


Figure 8.82 Summary stratigraphy, pollen and radiocarbon data for site HR4. After Horton *et al.* (1999c). See Table 8.16 for a description of the stratigraphy.

(BM-80) ^{14}C years BP (Barker and Mackey, 1961), confirming the cultural stage of the associated flint assemblage rather than the age of the peat. Most of the recognizable tree remains are oak wood and leaves, pine wood, birch bark and branches, hazel nuts and leaves, and alder leaves. Cultural remains from the early Mesolithic to the Romano-British periods have been reported as chance finds on the foreshore, but in-situ remains have been found within the peat beds themselves, including Mesolithic flint assemblages (Trechmann, 1936, 1947), Neolithic sherds, stone axes and worked wooden objects, and, more recently, an early Neolithic wattle panel (Annis, 1994). Perhaps the most significant archaeological discovery has been the Neolithic human skeleton from which bones were radiocarbon dated to 4680 ± 60 years BP (HV 5220). The skeleton was disarticulated, with the cranium upside-down (Tooley, 1978b), and was found in association with flints. From the injuries and the pathology this find appears to have been an in-situ wetland burial.

Interpretation

The sediments from WH19 and HB4 are representative of those found across the foreshore at Hartlepool. The intercalated sequence of peats and organic muds with marine and estuarine clays and silts is indicative of a variable marine influence on sedimentation brought about by changing relative sea-level during the mid-Holocene. The blue-grey clays and silts containing marine diatoms and molluscs, such as *Scrobicularia plana* and *Cerastoderma edule*, were deposited during periods of relatively low energy, lower intertidal estuarine conditions (Horton *et al.*, 1999c). The organic deposits range from detrital muds deposited in shallow water to more terrestrial herbaceous and woody peats. These peats and many of the *limus* deposits represent periods when either upper saltmarsh conditions prevailed or freshwater swamps and fens became established on exposed older intertidal units.

The lowermost marine clay on Hartlepool foreshore is dated to shortly before 6000 years BP, and overlies a basal peat that appears from the pollen evidence to have started to accumulate at approximately 7000 years BP. As a result of a slight fall in relative sea-level, peat formation replaced the marine clay at about 5000 years BP. However, an increased marine influence soon after this time deposited a thin layer of marine sediments in restricted areas. In the region protected by the old sea defences, the sediments record two additional phases of sea-level rise, one depositing organic muds in the upper part of a peat bed around 3500 years BP and one that leavens the peat with a grey estuarine silty clay dating from around 2800 years BP.

In terms of the controls on sedimentation, the sequence observed is indicative of both groundwater movements and relative sea-level change. Changes in the groundwater level may, indeed, be linked to sea-level but also are a function of drainage efficiency as a consequence of coastal evolution and landscape changes in the adjacent catchments (Horton *et al.*, 1999c). The history of sea-level change in Hartlepool Bay has been reconstructed from these and other intercalated sequences by Tooley (1978b). The record shows almost no crustal movement in the region since about 5000 ^{14}C years BP, and only a small amount in the earlier Holocene (Shennan, 1989, 1992; Long and Shennan, 1993). Hence, the record from Hartlepool provides evidence of sea-level change for the period of approximately 7000 to 3000 ^{14}C years BP and reflects the combination of Holocene sea-level rise and site-specific factors such as changing catchment land use and sediment supply.

The concentrations of flints and artefacts, human remains and faunal evidence suggest that the wetlands in Hartlepool Bay were attractive to exploitation, especially by early farming communities of the Neolithic and Bronze Age. In addition, these people also may have influenced the nature of wetland sedimentation through clearance activity in the adjacent catchment, potentially altering sediment supply and drainage.

Conclusions

The submerged forest bed at Hartlepool is an important archaeological and geological resource. The wealth of Holocene palaeoenvironmental data obtained from the peats and intercalated tidal sediments includes evidence of past climate and sea-level change, as well as cultural remains from the early Mesolithic to the Romano-British periods.

The sedimentary record from Hartlepool foreshore is the product of considerable changes in land drainage brought about by a variable marine influence between approximately 7000 and 3000 years BP, phases of land clearance, and changes in sediment supply and coastal environment. The observed Holocene relative sea-level trend confirms the unique position of Hartlepool and the Tees estuary on the east coast of the UK as being located on the fulcrum between crustal uplift to the north and subsidence to the south.

HOLY ISLAND (NU 136 418)

A. Plater

Introduction

Holy Island, Northumberland is noted not only for outstanding coastal geomorphology but also for its morphological and sedimentary evidence of late Quaternary sea-level trends. The coast around this mantled outcrop of Whin Sill dykes and adjacent tidal flats is almost devoid of development, allowing virtually unabated development of dunes, blow-outs, barrier beaches and cliff exposures. Relict coastal features also are present in the form of emerged ('raised') beaches that can be traced on Holy Island itself and on the mainland at Beal Point, Whitelee Letch, and west of Ross Links (Gunn, 1900; Hogg, 1972; Plater and Shennan, 1992). The raised shoreline on Holy Island is a cliffline cut into glacial diamicton associated with a series of gravel spits and bars, often truncated, and locally intertidal lagoonal deposits. Recent work has revealed additional sedimentary evidence of Holocene sea-level trends in the region of Elwick immediately to the south of Fenham Flats (Plater and Shennan, 1992), and at Bridge Mill to the north-west of Beal Point (Horton *et al.*, 1999a). At these sites, intercalated sequences of minerogenic and organic sediments reveal the nature and extent of coastal change. Hence, the region of Holy Island preserves morphological evidence of a high post-glacial shoreline and a stratigraphical record of mid-Holocene sea-level trends.

Description

Although raised post-glacial shorelines are well developed in south-eastern Scotland (Sissons, 1983), they are relatively rare in eastern England. Examples such as the Easington raised beach in County Durham now appear to date from Oxygen Isotope Stage 7 (Bowen *et al.*, 1991), although it was first thought to be Late-glacial or post-glacial in origin (Woolacott, 1922). This absence is clearly a spatial function of relative rates of isostatic crustal rebound and sea-level rise (Lambeck, 1991, 1995). The coast of Northumberland is perhaps unique in this context as the low-raised beach features identified in the vicinity of Holy Island by Gunn (1900) appear to be Holocene in age.

Facets of raised beach features are preserved at Beal Point, Whitelee Letch and west of Ross Links. Gunn (1900) records a low-altitude raised beach of considerable extent at Whitelee Letch, approximately 2 m above the high water mark. This was said to be composed of silt and grey clay with traces of marine shells and having a border of shelly sand along the coast. Hogg (1972) carried out a detailed survey of this feature, the base of which was levelled at +7.01 m OD and noted as being best developed in the region of NU 1183 3775 (Figure 8.83). The cliff-line slope was observed at 38° with the beach profile at 2–5°. Shells from a newly cut drainage ditch revealed a molluscan fauna indicative of a sandy or muddy foreshore, with *Mya* sp., *Lutraria* sp., *Scrobicularia plana* and *Lacuna crassior* the most abundant. By far the best example of the raised shoreline is found on Holy Island. Two raised beaches are located in the northern part of the island: a narrow beach partly covered by blown sand is located west of Caves Haven, and a gravel beach located 2.5–3 m above the high water mark on the east side of the island opposite Sheldrake Pool. The examples along the southern shore are broader and larger, and east of the castle there are two distinct levels of gravel (Gunn, 1900). Narrow strips of raised beach features also are found on the west of the island, and patches occur at two levels in The Snook, with the highest being made up of gravel. Gunn (1900) also notes that in 1854, Johnston (1873) recorded shells of *Mytilus edulis* with *Littorina littorea*, *Patella vulgata*, *Cardium edule* and a few broken valves of oyster from a beach on St Cuthbert's Isle not less than *c.* 3.5 m above the level of the sea. According to Tooley (1978a), at the foot of the cliff on Lindisfarne (renamed 'Holy Island' in the 11th century), north-west of the castle in the region of NU 1320 4190 to NU 1350 4175, the mean altitude of the raised beach is +5.89 m OD. From the sandy raised beach between the foot of the cliff and the present shore at an altitude of +4.67 m OD a rich molluscan fauna, dominated by *Littorina littorea*, *Patella vulgata*

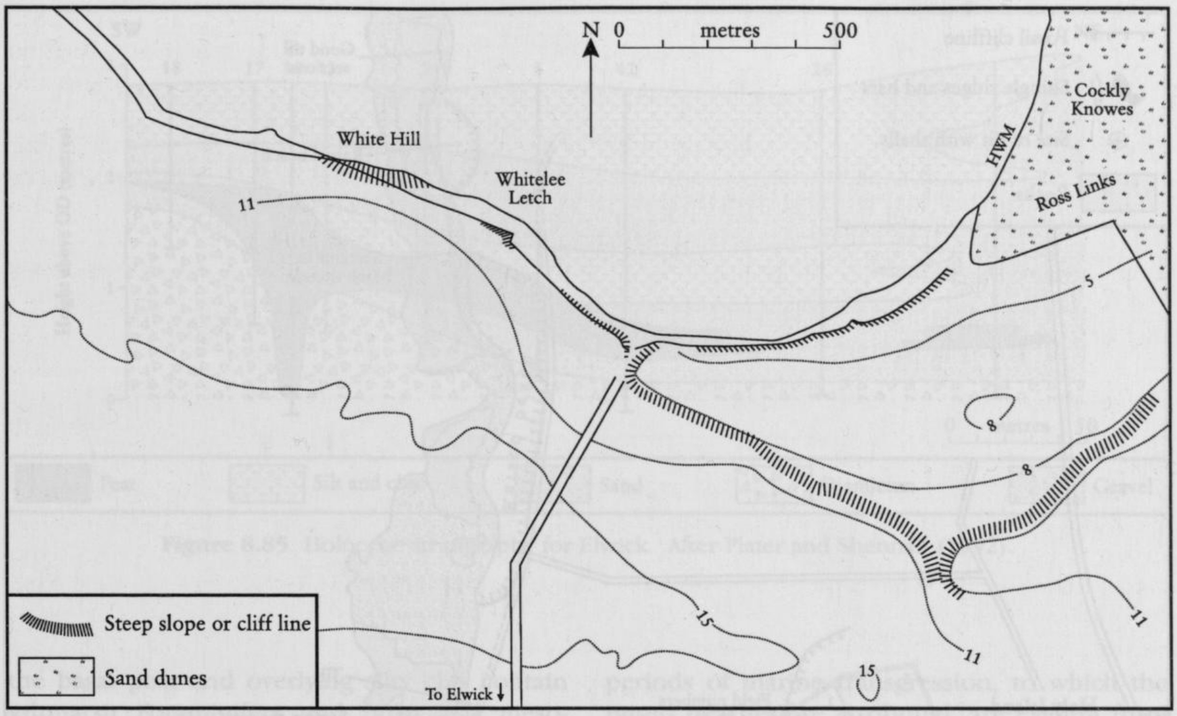


Figure 8.83 Extent of the cliff-line demarcating a raised beach at Whitelee Letch (after Hogg 1972).

and *Litorella littoralis*, proved to be indicative of nearby rocky conditions.

The raised beach in the vicinity of the castle is cut into glacial diamicton and is clearly identified from a series of truncated gravel ridges and bars that run transverse to the present coastline (Figure 8.84). Isolated lagoonal sediments also are noted, but no details appear to have been published on these. Gravel ridges also are present to the east of the castle. These occur at three different levels and, like those to the west, are truncated by the present shoreline.

In addition to the morphological evidence of post-glacial sea-level trends, stratigraphical records are present along the shores adjacent to Holy Island. Gunn (1900) recorded the presence of marine alluvium, mainly silt and sand, in the region of Fenham Flats, Brockmill, and south of Ross Links, although other examples may be covered by blown sand. Plater and Shennan (1992) note that each of these alluvial tracts of estuarine sediment were only a metre or so above the present high water mark of spring tides and were separated from the shoreline by hills of blown sand. At Brockmill, a fining-upward minerogenic sequence overlies a purple-brown diamicton, reflecting progressive sedimentation in an intertidal environment of decreasing energy. Similarly, the Holocene sedimentary record overlying the pre-Holocene surface at Ross Low proved to be almost entirely minerogenic with only a few organic lenses.

The Holocene stratigraphy at Elwick, in the vicinity of Whitelee Letch, also was investigated by Plater and Shennan (1992). Here, a sloping surface of weathered diamicton is overlain by a severely oxidized, fining-upward sequence of grey-brown and orange silty sand and clayey silt. Incised into the diamicton surface is a narrow sinuous channel that preserves a complex Holocene stratigraphy (Figure 8.85). The lowermost unit is a blue-grey and grey-brown clastic unit of variable grain-size with dispersed organic material. The diatoms present within this are predominantly brackish and fresh–brackish (Figure 8.86a). The lower clastic unit is overlain by an equally variable peat with gravel, sand, silt and clay. The altitude of the contact between the peat and the underlying clastic facies rises landward from approximately +0.30 to +2.00 m OD. The pollen in the peat shows a predominance of *Quercus* and *Corylus* types with subordinate *Betula*, *Pinus*, *Ulmus*, *Alnus* and *Salix*, with a significant rise in Gramineae and Cyperaceae

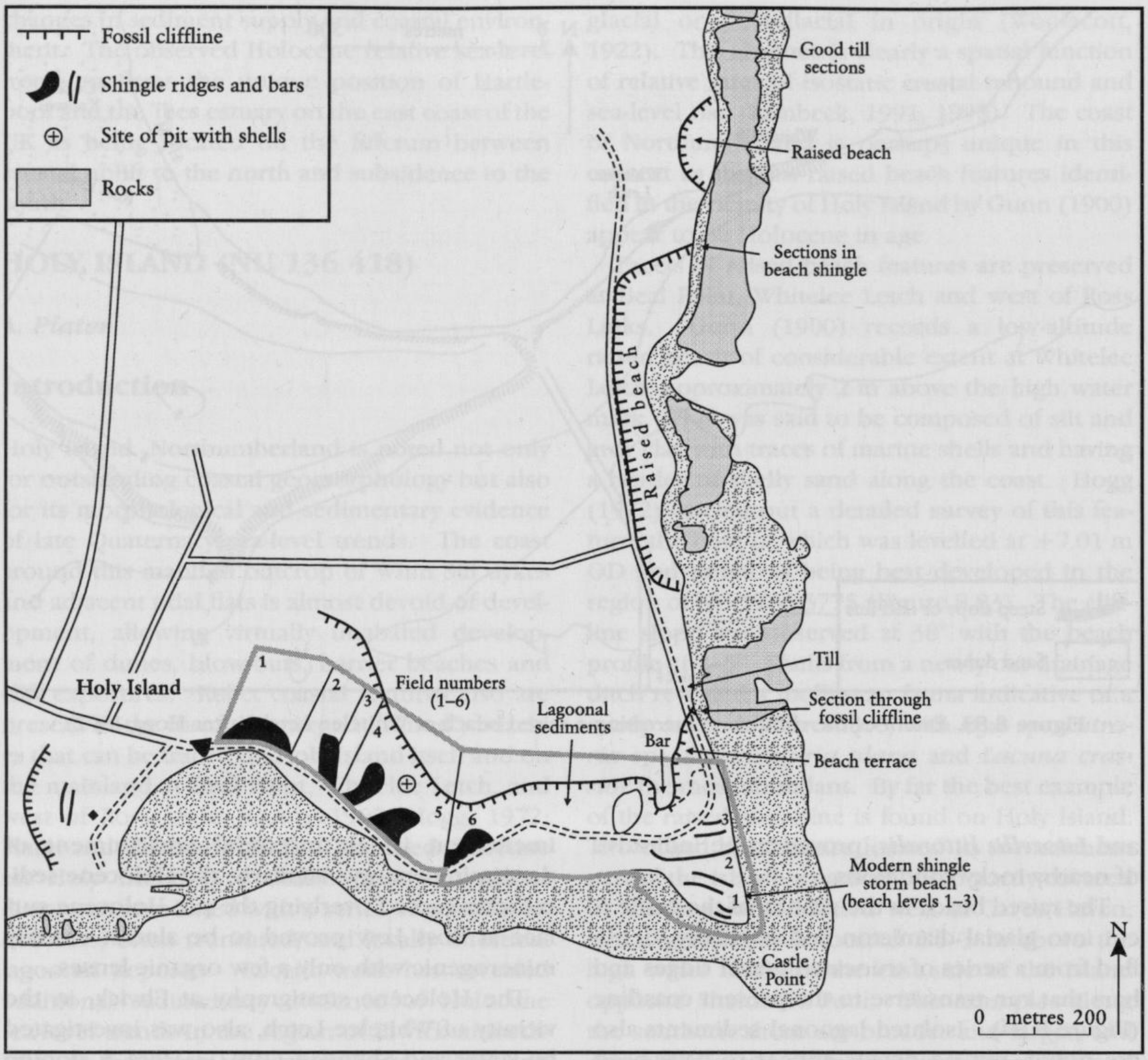


Figure 8.84 The location and morphology of raised beach features on the south-eastern shores of Holy Island. From unpublished English Nature site documentation and management brief (1992).

towards the top accompanied by Compositae tub. and Chenopodiaceae (Figure 8.86b). This indicates a trend towards a more open coastal marsh environment, although brackish diatoms are present throughout the peat (Figure 8.86a). The peat is overlain by an organic blue-grey silty clay between +0.73 and +2.00 m OD, with a landward increase in contact altitude. This overlying minerogenic facies appears to have been deposited in a sand- and silt-dominated open tidal flat, as illustrated by the observed increase in the proportion of the diatom *Paralia sulcata*. From radiocarbon dating, it would appear that the peat started to accumulate approximately 7200 ^{14}C years BP in response to a reduced rate of sea-level rise. Frequent tidal inundation of the peat-forming communities eventually resulted in the return to an intertidal mudflat approximately 6900 ^{14}C years BP.

Horton *et al*. (1999a) also have recorded the presence of an intercalated Holocene stratigraphy to the north-west of Beal Point at Bridge Mill. A basal organic unit that includes saltmarsh indicators is overlain by a clastic marine sequence that includes one main intercalated peat bed and minor organic lenses. The micropalaeontological data obtained for this stratigraphical record (Figure 8.87) reveal that

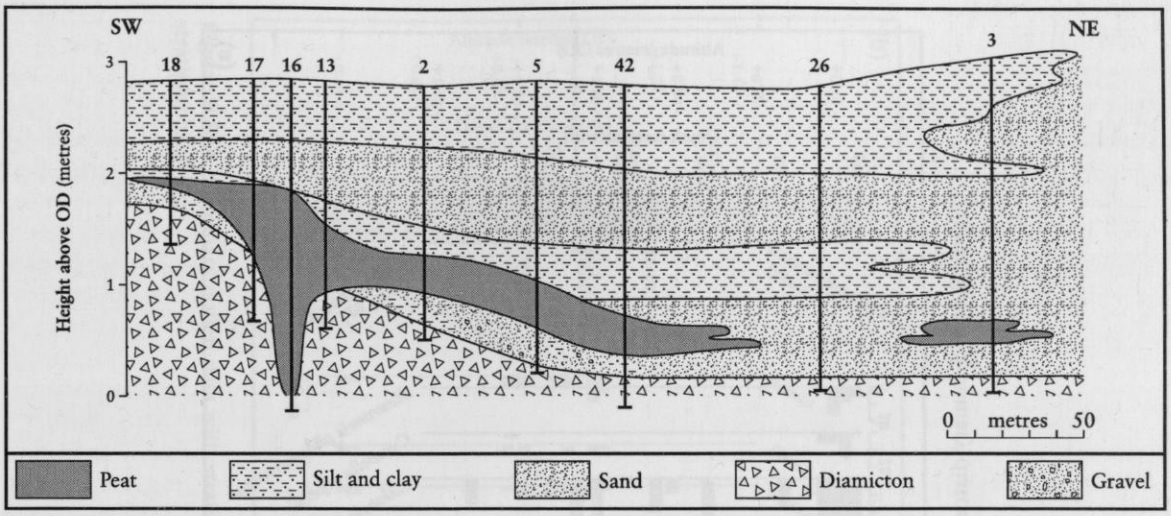


Figure 8.85 Holocene stratigraphy for Elwick. After Plater and Shennan (1992).

the basal peat and overlying silty clay contain saltmarsh Foraminifera and poly- and meso-halobous diatoms. Forams are absent in the upper organic unit but there are halophobous diatoms and saltmarsh pollen indicators, including Chenopodiaceae and *Plantago maritima*. Hence, the sedimentary succession in this location can be used to provide four sea-level index points for the reconstruction of Holocene sea-level trends.

Interpretation

Although the morphological evidence of former relative sea-level is expressed clearly on Holy Island, the chronology is less certain. The age of the raised beaches in the region of Holy Island has been inferred from pollen evidence from the Inner Farne Islands (Hogg, 1972; Tooley, 1978a). South-east of Middle Pond on the Inner Farne, Hogg (1972) described an intercalated peat sequence underlain by a coarse sandy clay. An upper monocotyledenous peat overlies a sandy peat with grey clay lenses, which, in turn, overlies brown peat with grey sand. Detailed pollen analysis of the peat revealed an Atlantic VIIa age for the base, becoming sub-Boreal towards the top. At Long Bog, a similar Holocene sequence of intercalated sands and peats is underlain by a coarse gravelly sand and several metres of peat grading downward into a gravel with clay. Although peat accumulation on the Inner Farne commenced approximately 7000 years BP, two periods of marine transgression, to which the raised beaches are attributed, are evident from the pollen and lithostratigraphy (Tooley, 1978a). The first of these is proposed to have taken place towards the end of Flandrian II (approximately 5200 years BP) and the second early in Flandrian III (approximately 4000 years BP).

The data from Elwick and other sites on the Northumberland coast (Plater and Shennan, 1992) suggest that sea-level has changed by only approximately 2.6 m during the last 8000 years or so. As a result, site-specific factors, such as coastal morphology and sediment supply, have played an important role in shaping the coast. However, a plot of the validated sea-level index points for the coast of Northumberland (Horton *et al.*, 1999a) shows a record of relative sea-level from more than 5 m below present at approximately 9000 years BP to about 2.5 m above at *c.* 2500 years BP. If data points are considered geographically, there is a clear trend of increasing altitude with distance northward for sea-level index points of any given age (Plater and Shennan, 1992; Shennan *et al.*, 2000b; Horton *et al.*, 1999a), which is consistent with geodynamic model predictions of greater uplift towards the north (Lambeck, 1991, 1995; Shennan, 1992). The overall trend is, therefore, one of rising relative sea-level during the Holocene Epoch, with a fall during the past 3000 years or so. However, the expression of this trend is modified by site-specific factors that include changes in the tidal regime, sediment

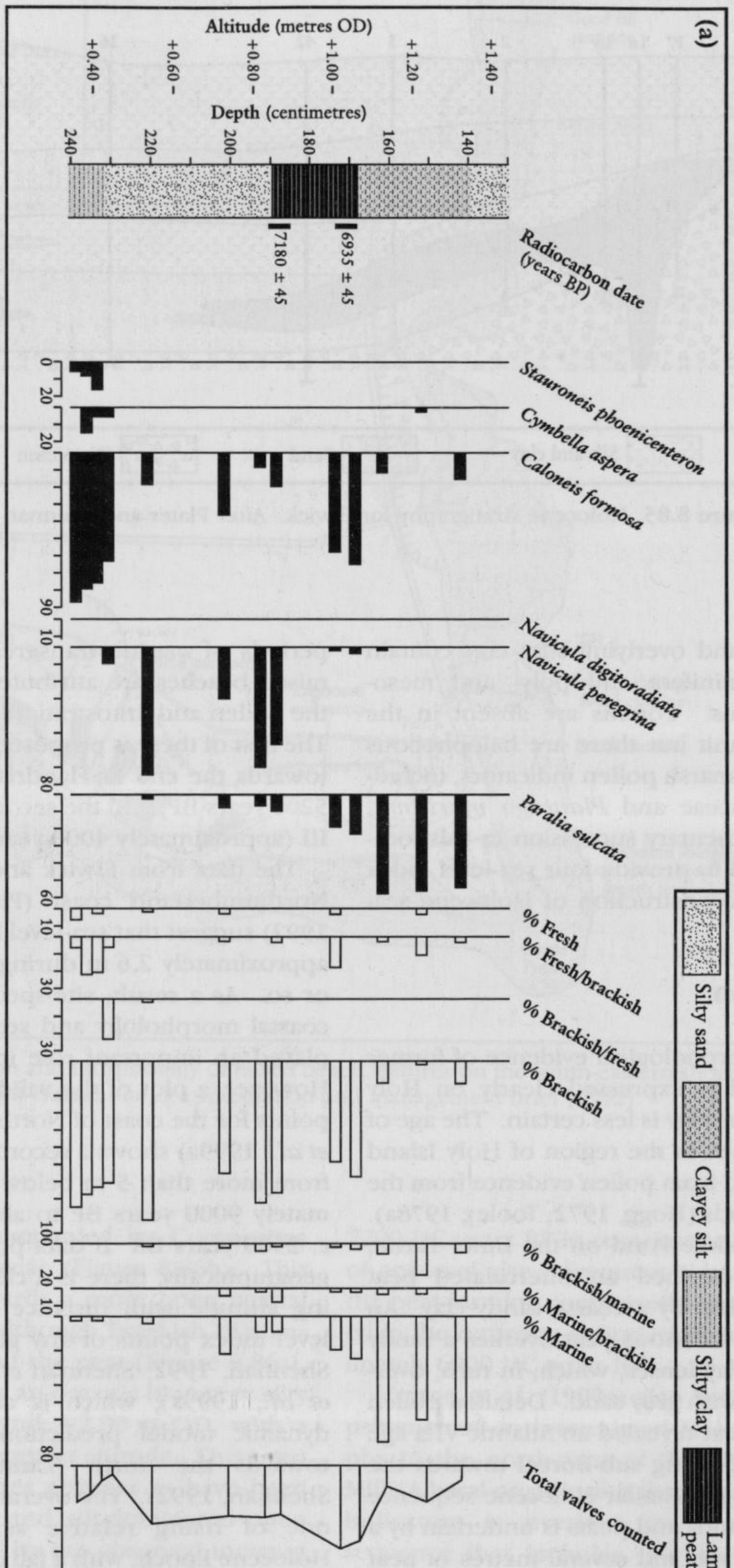


Figure 8.86a Summary diatom data with stratigraphy and radiocarbon dates for Elwick. After Plater and Shennan (1992).

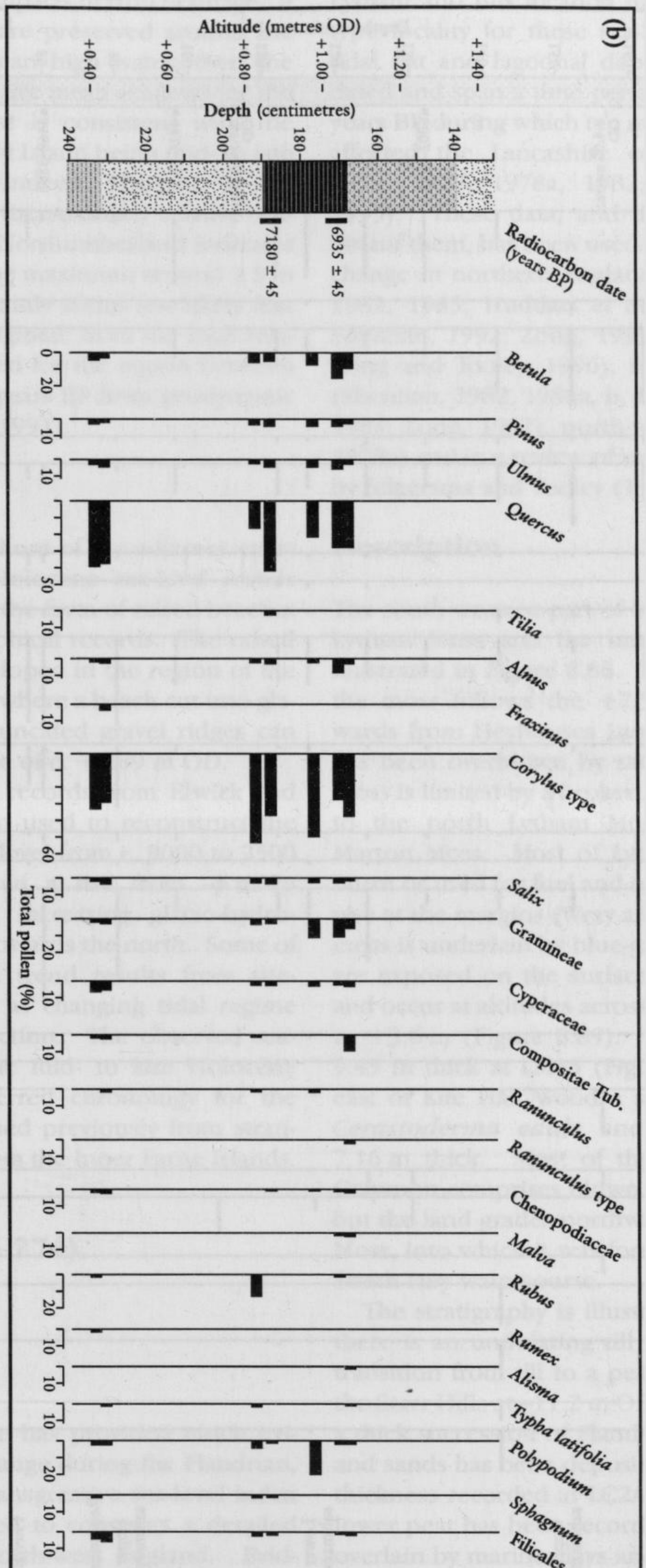


Figure 8.86b Summary pollen data with stratigraphy and radiocarbon dates for Elwick. After Plater and Shennan (1992). See Figure 8.86a for key to the stratigraphical log.

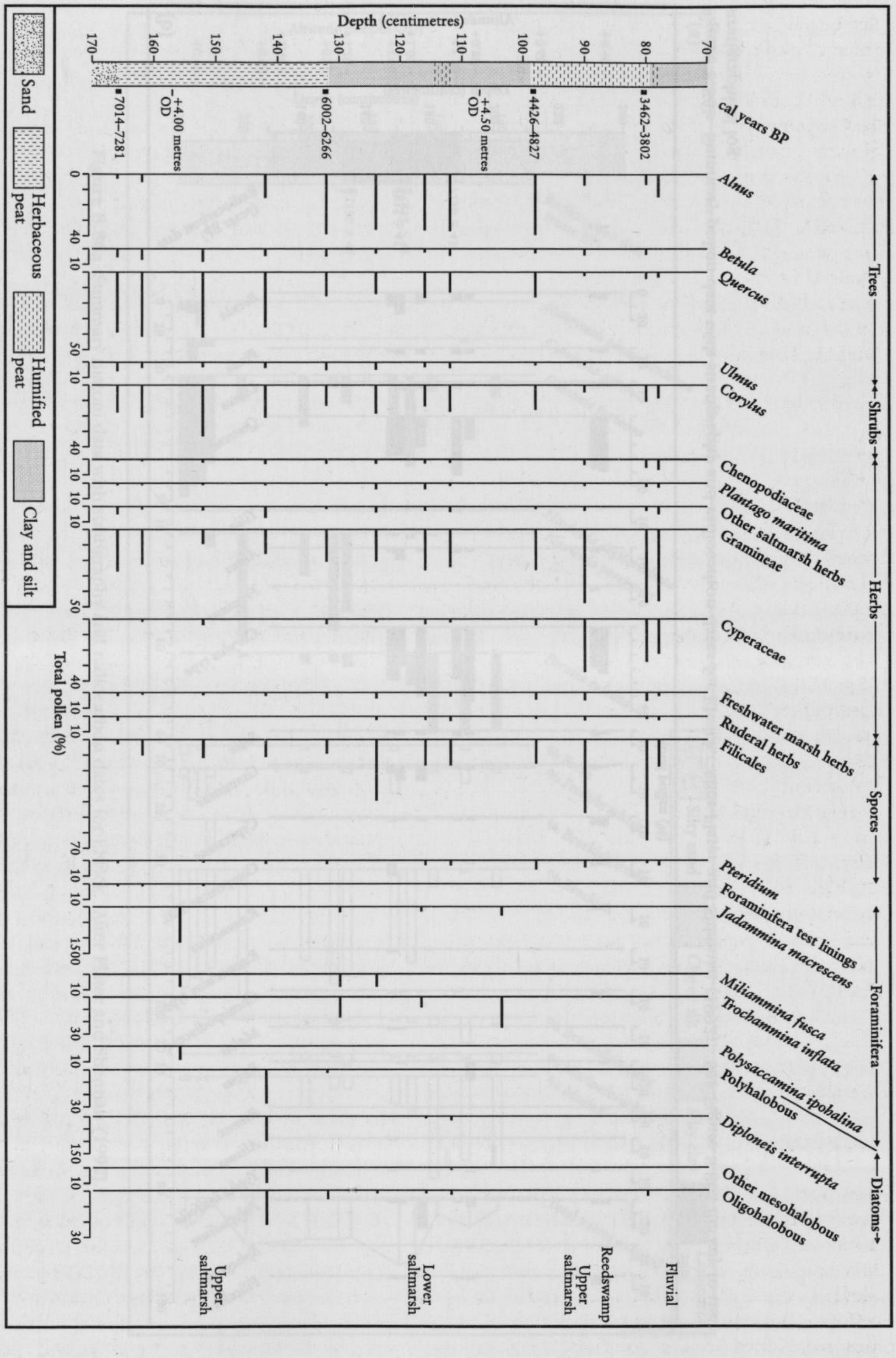


Figure 8.87 Summary stratigraphy, pollen, Foraminifera and radiocarbon data for Bridge Mill. After Horton *et al.* (1999a).

compaction, and the impacts of land drainage. If relict beach features are preserved around the former altitude of mean high water, then the observed trend of relative mean sea-level for the Northumberland coast is consistent with the raised beaches on Holy Island being mid- to late Holocene in age. Indeed, Shennan *et al.* (2000b) observe that the summary relative sea-level curve for north Northumberland indicates a mid- to late-Holocene maximum around 2.5 m above present. It certainly seems less likely that the raised beaches originate from the high relative sea-levels predicted for the region between *c.* 15 000 and 13 000 years BP from geodynamic modelling (Lambeck, 1991).

Conclusions

Holy Island and the shore of the adjacent coast possess records of Holocene sea-level trends and coastal change in the form of raised beaches and lowland stratigraphical records. The raised beaches are best developed in the region of the castle on Holy Island, where a beach cut into glacial diamicton and truncated gravel ridges can be found at an altitude of *c.* +5.89 m OD.

The stratigraphical records from Elwick and Bridge Mill have been used to reconstruct the record of relative sea level from *c.* 9000 to 2500 years BP, which shows a rise from –5 m to +2.5 m OD and an increasing glacio-hydro-isostatic component towards the north. Some of this relative sea-level trend results from site-specific factors, such as changing tidal regime and sediment compaction. The observed sea-level highstand in the mid- to late Holocene corroborates the inferred chronology for the raised beaches obtained previously from stratigraphical evidence from the Inner Farne Islands.

LYTHAM (SD 335 274)

D. Huddart

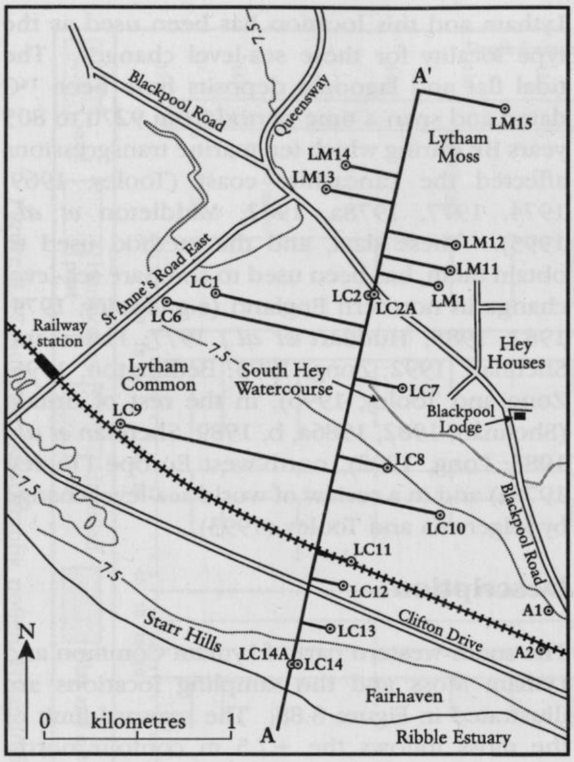
Introduction

South-west Lancashire has provided much evidence for sea-level change during the Flandrian, and regressive and transgressive sea-level index points have been used to construct a detailed sea-level curve for north-west England. Evidence for all but one of the marine transgressions comes from sites in the south Fylde around Lytham and this location has been used as the type locality for these sea-level changes. The tidal flat and lagoonal deposits have been ^{14}C dated and span a time period from 9270 to 805 years BP, during which ten marine transgressions affected the Lancashire coast (Tooley, 1969, 1974, 1977, 1978a, 1982; Middleton *et al.*, 1995). These data, and the method used to obtain them, has been used to compare sea-level change in northern England (e.g. Tooley, 1974, 1982, 1985; Huddart *et al.*, 1977; Plater and Shennan, 1992; Zong, 1993; Bedlington, 1995; Zong and Tooley, 1996), in the rest of Britain (Shennan, 1982, 1986a, b, 1989; Shennan *et al.*, 1983; Long, 1992), north-west Europe (Tooley, 1978a) and in a review of world sea-level change by Jelgersma and Tooley (1993).

Description

The south-western part of Lytham Common and Lytham Moss and the sampling locations are illustrated in Figure 8.88. The seaward limit of the moss follows the +7.5 m contour northwards from Heyhouses Lane; this moss margin has been overblown by sand. To the east the moss is limited by a broken, rising till ridge, and to the north Lytham Moss runs into Great Marton Moss. Most of Lytham Moss has been burnt or used for fuel and only peat veneers survive at the margins (Wray and Cope, 1948). The moss is underlain by blue-grey, silty clays, which are exposed on the surface east of Queensway and occur at altitudes across the moss from +2.6 to +3.6 m (Figure 8.89). These silty clays are 9.45 m thick at LM15 (Figure 8.89) and to the east of Kite Hall Wood a shelly, silty clay, with *Cerastoderma edule* and *Tellina tenuis*, is 7.16 m thick. Most of the surface of Lytham Common comprises blown sand above +7.0 m, but the land grades northwards towards Lytham Moss, into which it was formerly drained by the South Hey watercourse.

The stratigraphy is illustrated in Figure 8.89; there is an undulating till surface, the deepest transition from till to a peaty sand occurring at the Starr Hills at –11.2 m OD. On this till surface a thick succession of Flandrian peats, silts, clays and sands has been deposited, with the greatest thickness recorded at LC2A, with over 17 m. A lower peat has been recorded at LC2 and LC14, overlain by marine clays and silts, locally rich in *Scrobicularia plana*. These marine sediments are replaced either by peat, from LC11 land-



ward, or blown sand from LC12 seaward. In every location peat is overlain by varying thicknesses of blown sand. Where the sand exceeds 1.0 m, it is differentiated into a lower and upper stage, separated by a palaeosol, or an accumulation of sandy peat. At LC6, for example, there is a soil horizon at 55–56 cm and peaty horizons at 75–77 cm and 185–200 cm before woody, detrital peat is reached at 220 cm.

The pollen diagrams from the Starr Hills (LC14A) and Heyhouses Lane, St Annes (LC2A) are illustrated in Figure 8.90 and the pollen diagram from Thomas Gillat's Colley Hey (LC1) is given in Figure 8.91. Two sampling sites occur in the south-eastern corner of Lytham Common

◄**Figure 8.88** Sampling sites across south-west Fylde (Lytham Common and Lytham Moss), projected on to an artificial line A–A'. The area today is more or less built up as part of Lytham St Annes. The sampling codes are: LM, Lytham Moss series; LC, Lytham Common series; A, Ansdell series (after Tooley, 1969, 1978a).

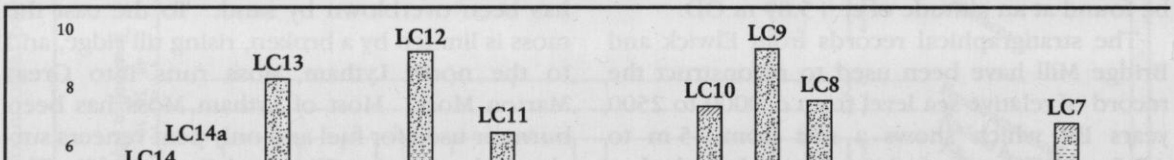


Figure 8.89 Stratigraphical successions at 18 sampling sites in south-west Fylde, projected on to a single artificial line A–A' shown in Figure 8.88 (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log.

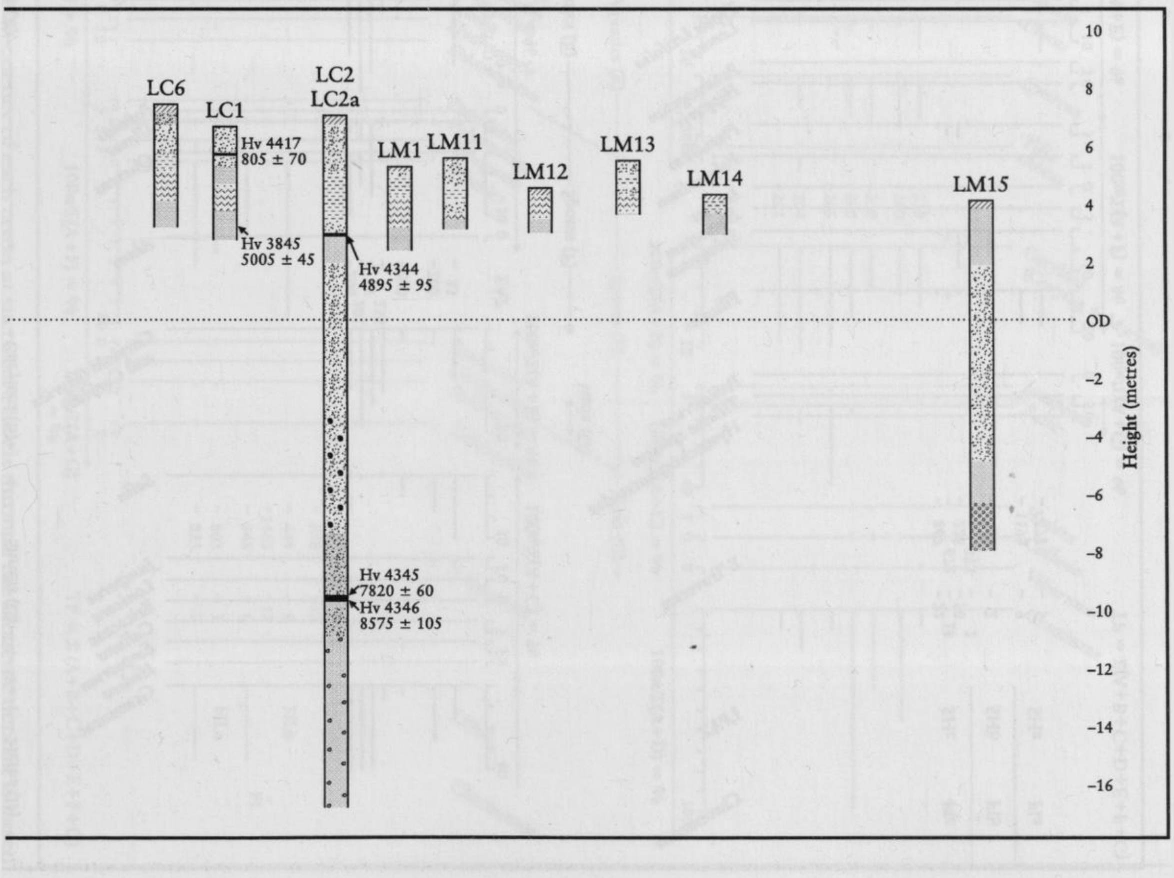
at Ansdell (A1 and A2, Figure 8.88) and the pollen diagram is shown in Figure 8.92.

At Lytham Hall Park the stratigraphy is illustrated in Figure 8.93, which shows a basin in the till, the lowest part of which is at –5.46 m at BH24. This basin is infilled with clays, silts, peats and sands. Again the upper blown sand can be subdivided as at Lytham Common. An example of a pollen diagram is illustrated from LHP5 in Figure 8.94. The detailed stratigraphy and pollen diagrams from Nancy's Bay and the Lytham–Skippool valley can be found in Tooley (1978a).

Interpretation

The south Fylde has provided direct evidence of nine marine transgressions between 8570 and 1370 years BP and indirect evidence of a tenth transgression from the sand-dune area (Figure 8.95). The evidence for these transgressions is from four locations, the three described above and Nancy's Bay (Tooley, 1978a). The marine sequences that bear the Lytham name (Tooley, 1978a) serve as the type succession for Flandrian marine sequences in north-west England and as a basis for inter-regional correlation (Figure 8.96).

The final stages of Lytham I are recorded from Lytham Common, where a grey clay with sandy partings gives way to a gyttja in which the pollen of open habitat, coastal taxa are recorded at an altitude of –9.75 m, although the dominant freshwater environment is indicated by high frequencies of aquatic taxa. This occurred at 8575 years BP. Lytham II is recorded from the Starr Hills as a grey fine clay and transgressed the present coast ending biogenic sedimentation in basins in the till at an altitude –11.13 m OD, shortly after 8390 years BP. The transgressive phase is recorded well landwards at Heyhouses Lane (Figure 8.88), but at a higher altitude (–9.82 m OD). The end of Lytham II is located in Nancy's Bay (Tooley, 1978a) at a mean altitude



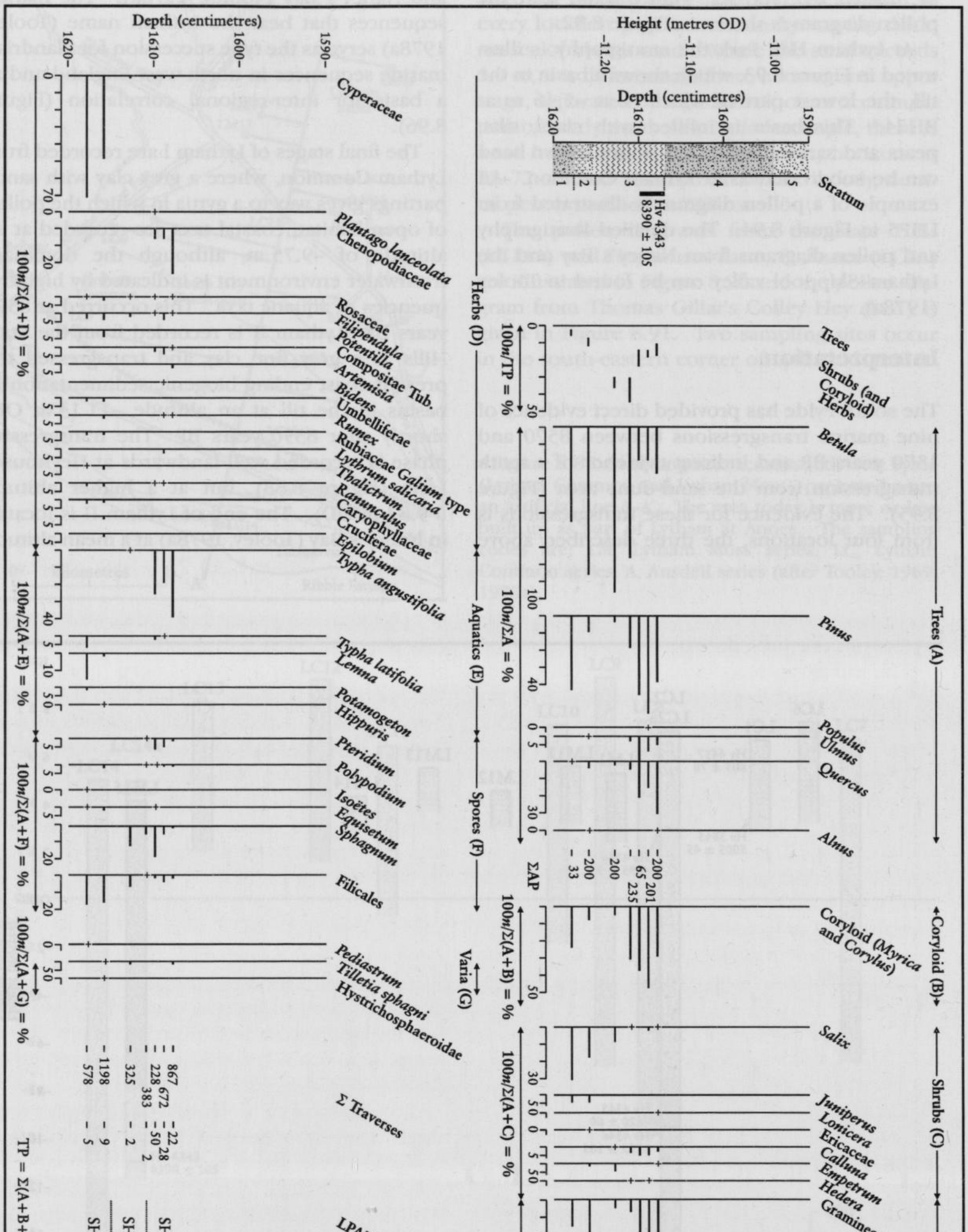


Figure 8.90a Pollen diagram from The Starr Hills, Lytham (LC14A). The frequency of each taxon at successive levels through the biogenic deposit is based on the calculation formula shown for each group (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log.

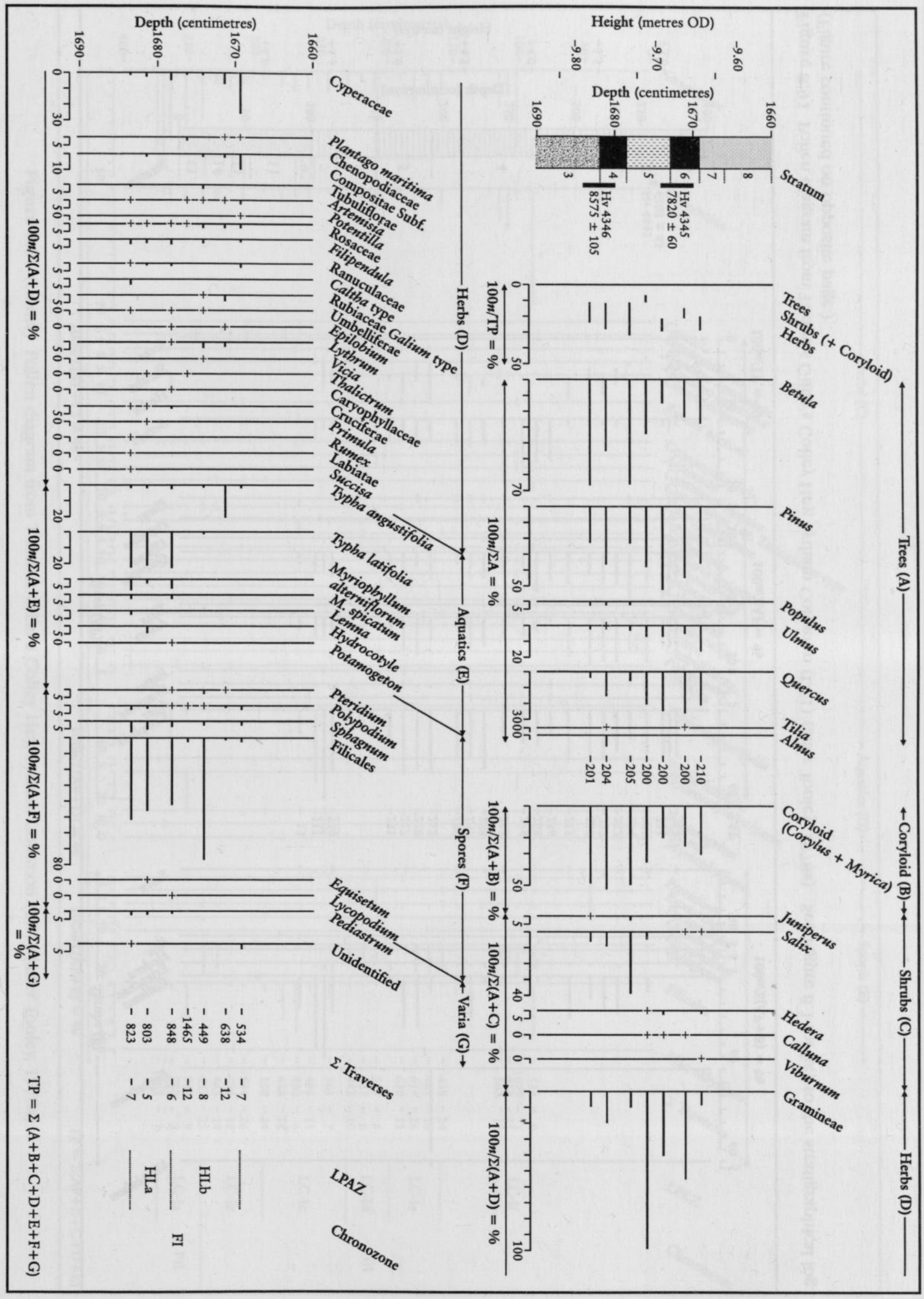


Figure 8.90b Pollen diagram from Heyhouses Lane, St Annes (LC2A). The frequency of each taxon at successive levels through the biogenic deposit is based on the calculation formula shown for each group (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log.

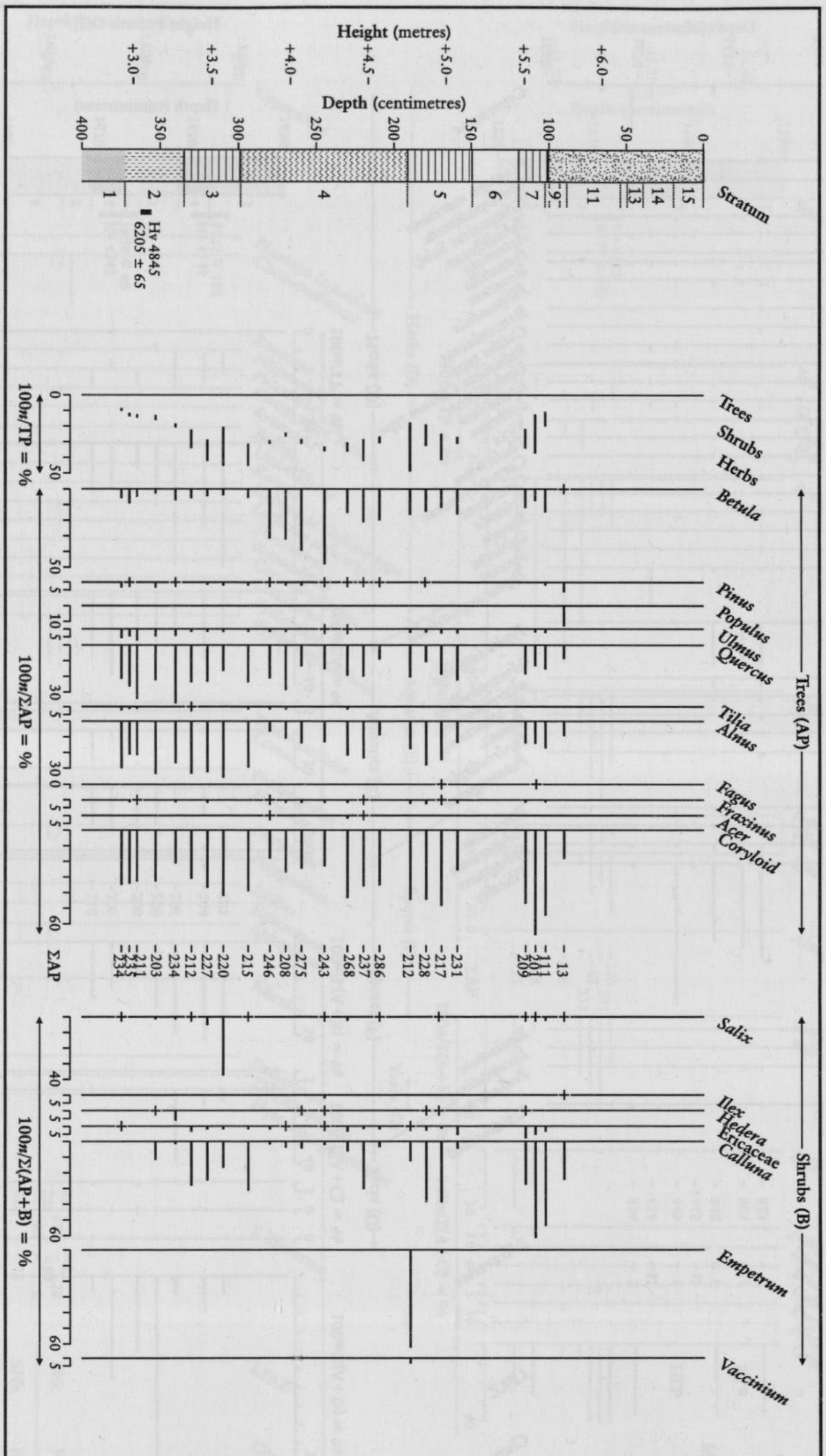


Figure 8.91 Pollen diagram from Thomas Gillat's Colley Hey, Lytham Common (LC1) (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log. (Figure continued on opposite page.)

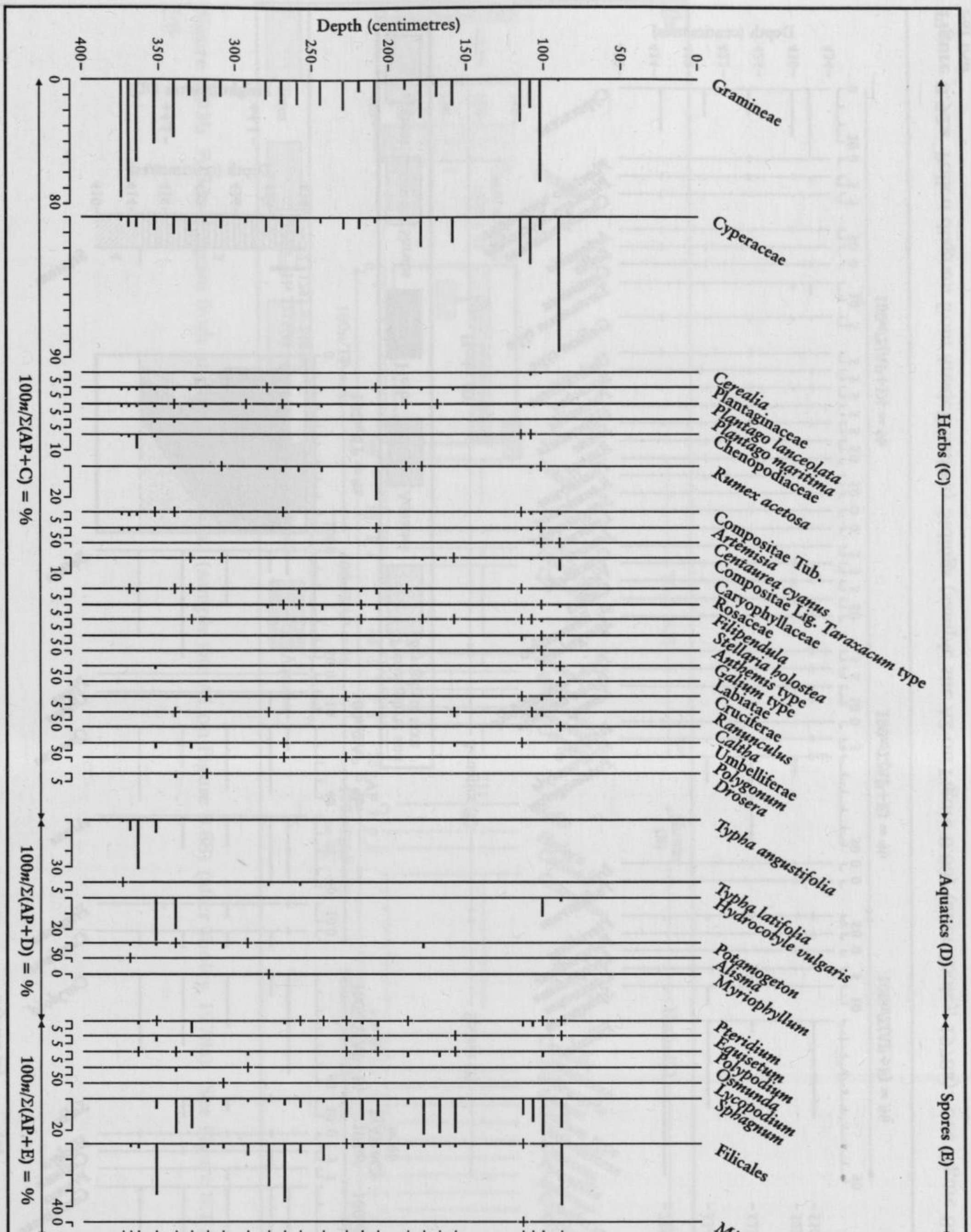


Figure 8.91 (*contd*) Pollen diagram from Thomas Gillat's Colley Hey, Lytham Common (LC1) (after Tooley, 1978a).

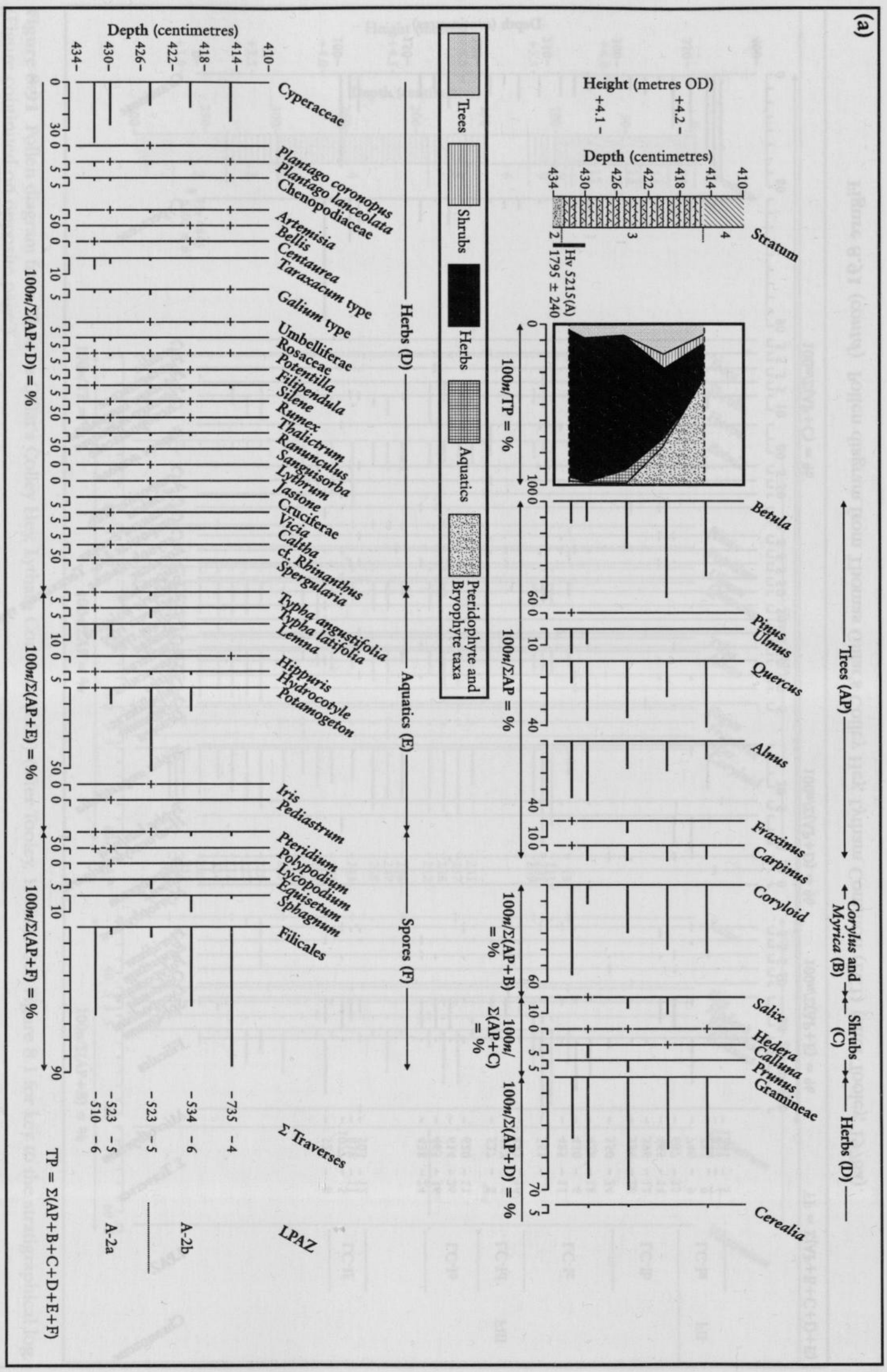


Figure 8.92a Pollen diagram from Ansdell Railway Sidings (sample site A2 on Figure 8.88) (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log.

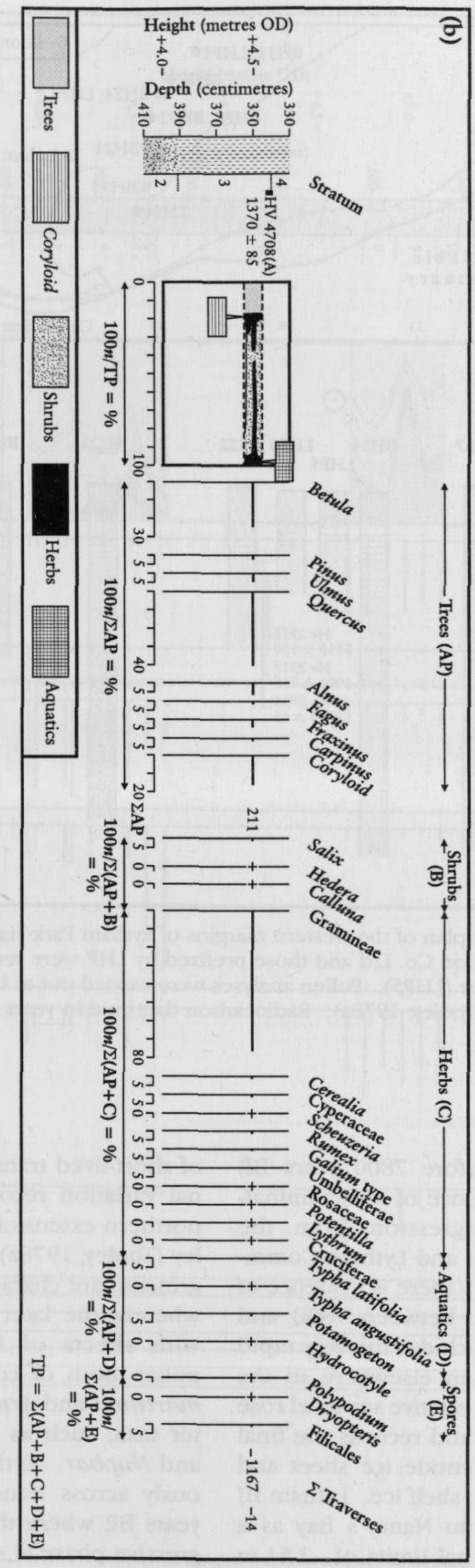


Figure 8.92b Pollen diagram from Ansdell, Rossall Road (sample site A1 on Figure 8.88) (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log.

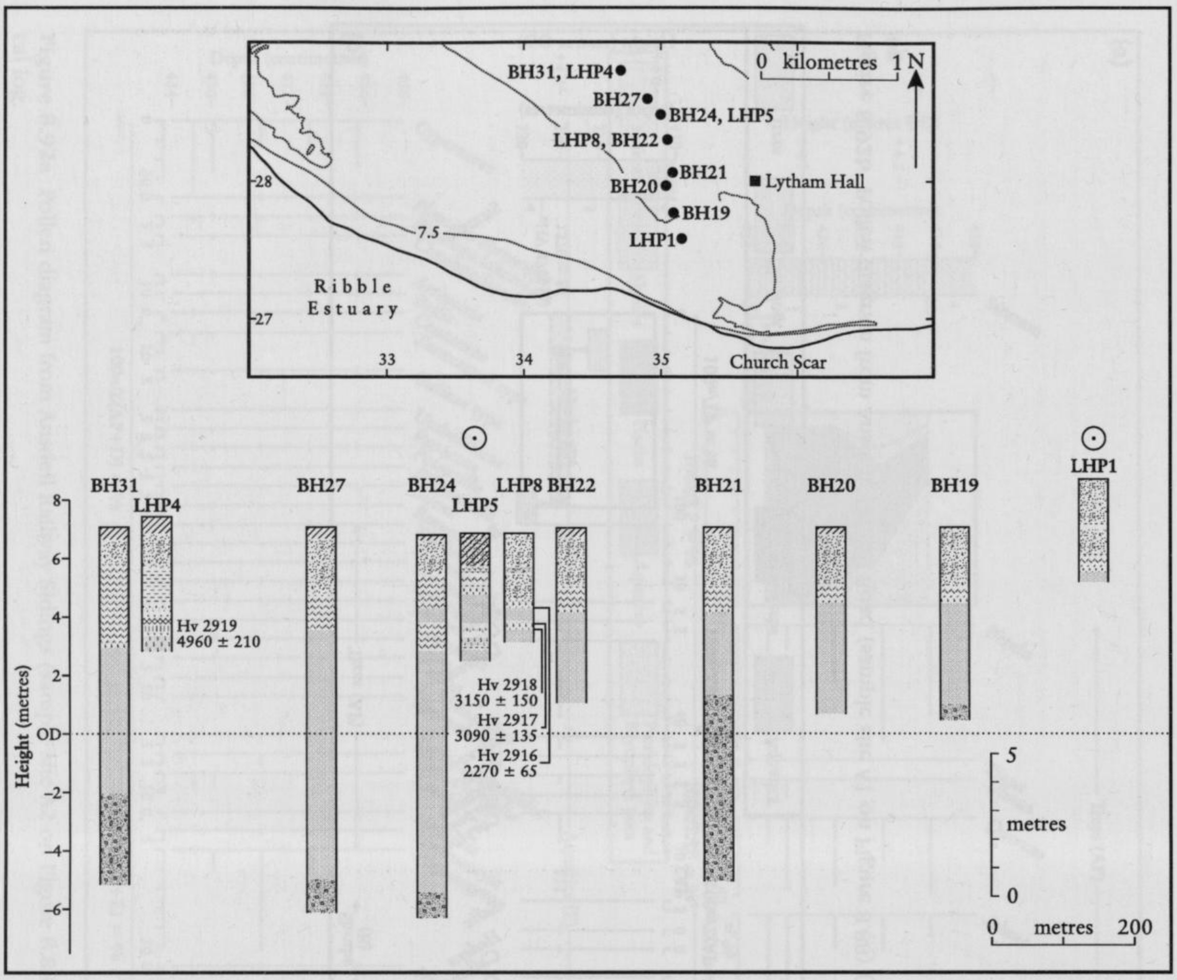


Figure 8.93 Stratigraphy and plan of the western margins of Lytham Park Hall. Sample codes prefixed by BH were carried out by Cementation Co. Ltd and those prefixed by LHP were recorded from open excavations or from a Hiller-type peat sampler (LHP5). Pollen analyses were carried out at LHP1 and LHP5 and are indicated by a circumscribed dot (after Tooley, 1978a). Radiocarbon dates are in years BP. See Figure 8.1 for key to the stratigraphical log.

of −2.58 m OD shortly before 7800 years BP. However, there is no evidence of the culminating stages of this transgression from the sequences on Lytham Moss and Lytham Common, although in Nancy's Bay there is evidence of five marine transgressions between 7800 and 5700 years BP. Lytham II includes the very rapid sea-level rise recorded from elsewhere in the world, and in the Fylde the relative sea-level rose from −9.6 m to 2.5 m OD and records the final disintegration of the Laurentide ice sheet and the attenuation of Antarctic shelf ice. Lytham III is recorded exclusively from Nancy's Bay as a blue-grey silt, with altitudinal limits of −2.51 m to −1.35 m OD. Lytham IV comprises a complex of short-lived transgressions with slight altitudinal variation recorded in Nancy's Bay and its northern extension in the Lytham–Skippool valley (Tooley, 1978a). The early stages of the transgression are characterized by grey sand and silt, whereas the later stages are fine silt and clay with sheets of *Phragmites* peat containing pollen both of coastal taxa, such as *Plantago maritima* and *Armeria maritima*, and freshwater taxa, such as *Cladium*, *Typha angustifolia* and *Nuphar*. Lytham V is recorded simultaneously across Nancy's Bay from 5950 to 5775 years BP, where the mean altitude of the transgressive phase is +1.3 m OD and of the regressive phase +1.59 m OD. Lytham VI is recorded

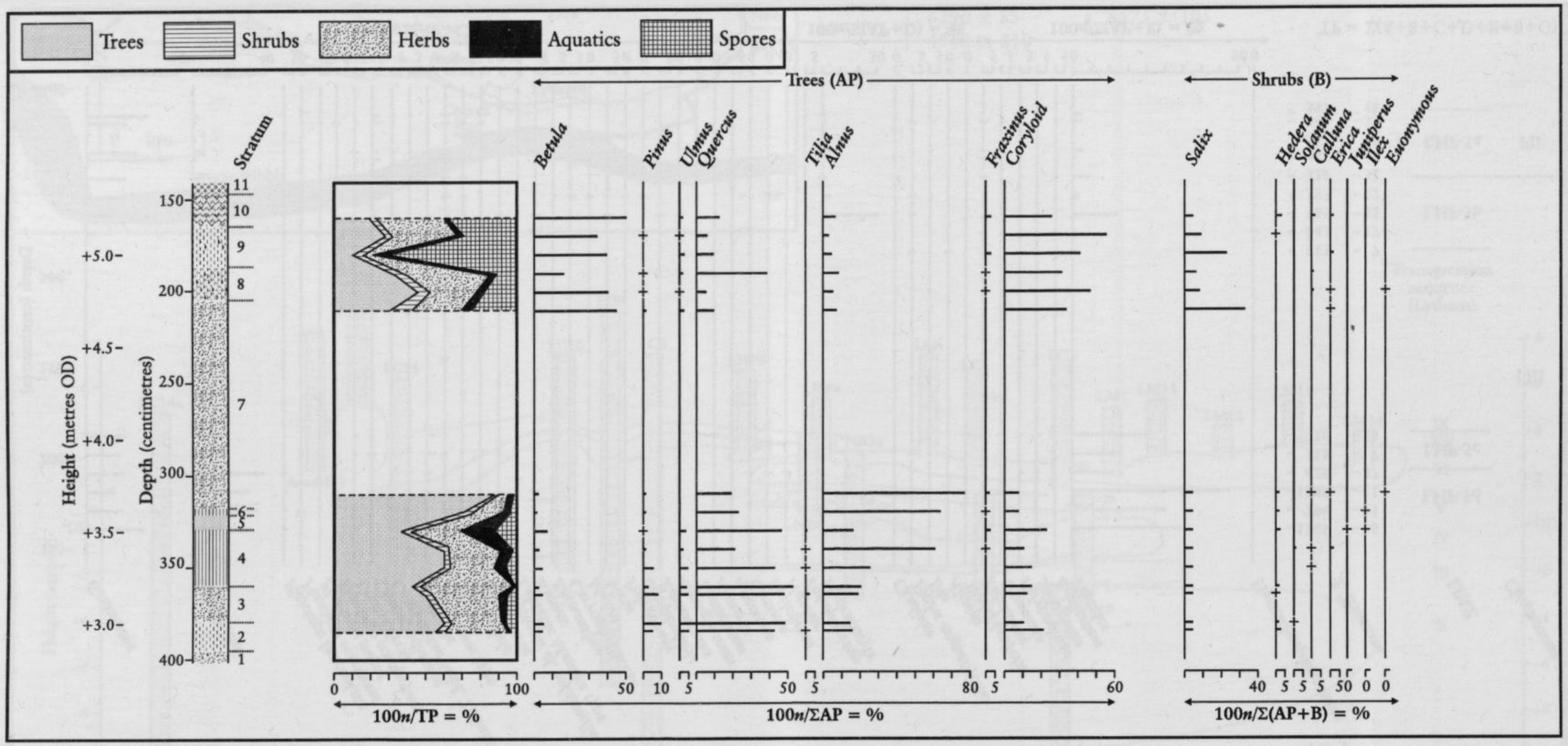


Figure 8.94 Pollen diagram from Francis Fox's South Hey Watercourse Meadow (LHP5) (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log. (Figure continued overleaf.)

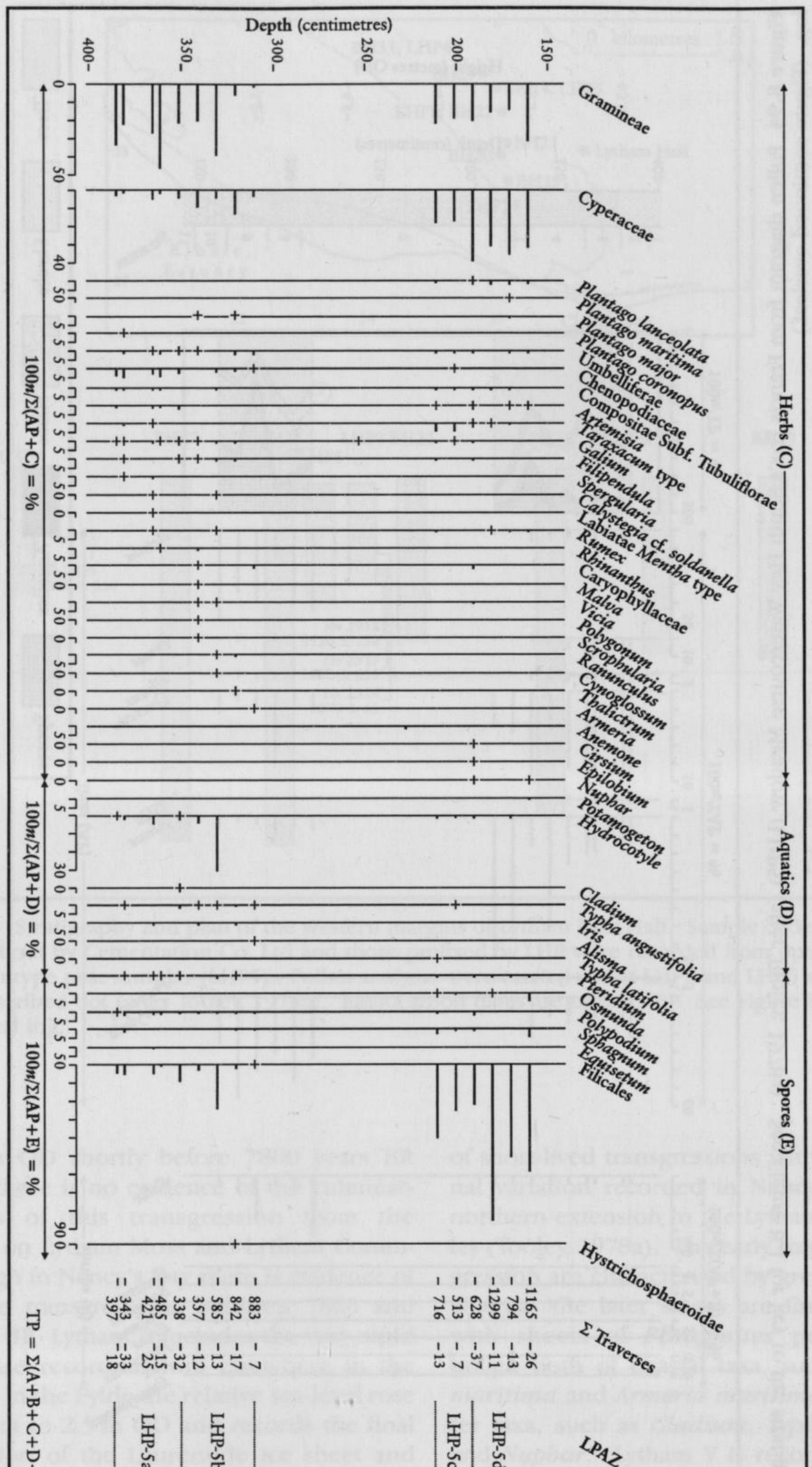


Figure 8.94 *(contd)* Pollen diagram from Francis Fox's South Hey Watercourse Meadow (LHP5) (after Tooley, 1978a).

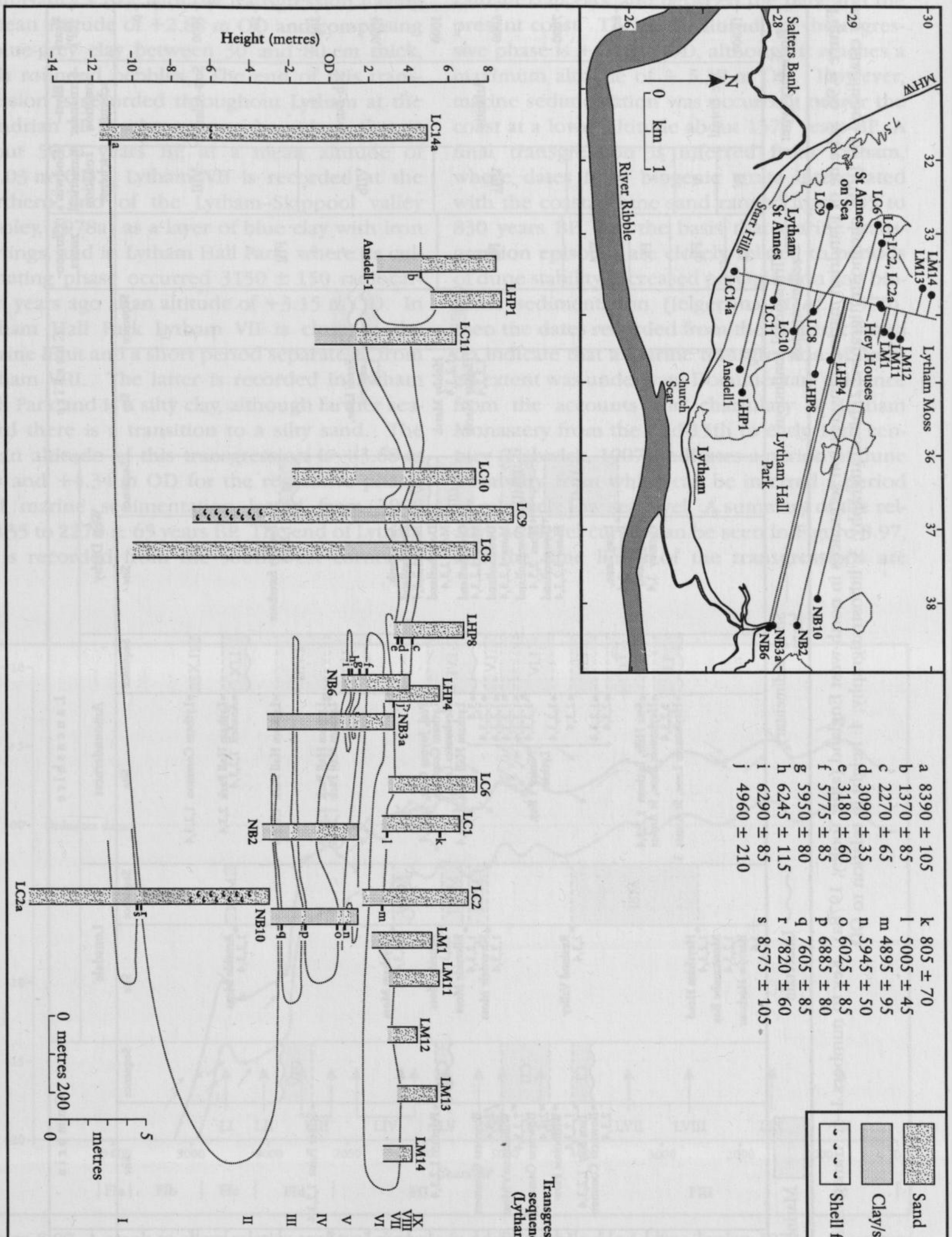


Figure 8.95 Marine transgressions in the Lytham area based on 21 cores (after Tooley, 1978a). The shading in each stratigraphical column is confined to those of a marine or brackish-water origin. Differences in height of the transgression surfaces derive in part from the projection of each sampling site on to a line A–A′ shown on Figure 8.88 and in part on a vertical exaggeration of ×50. Radiocarbon dates are in years BP

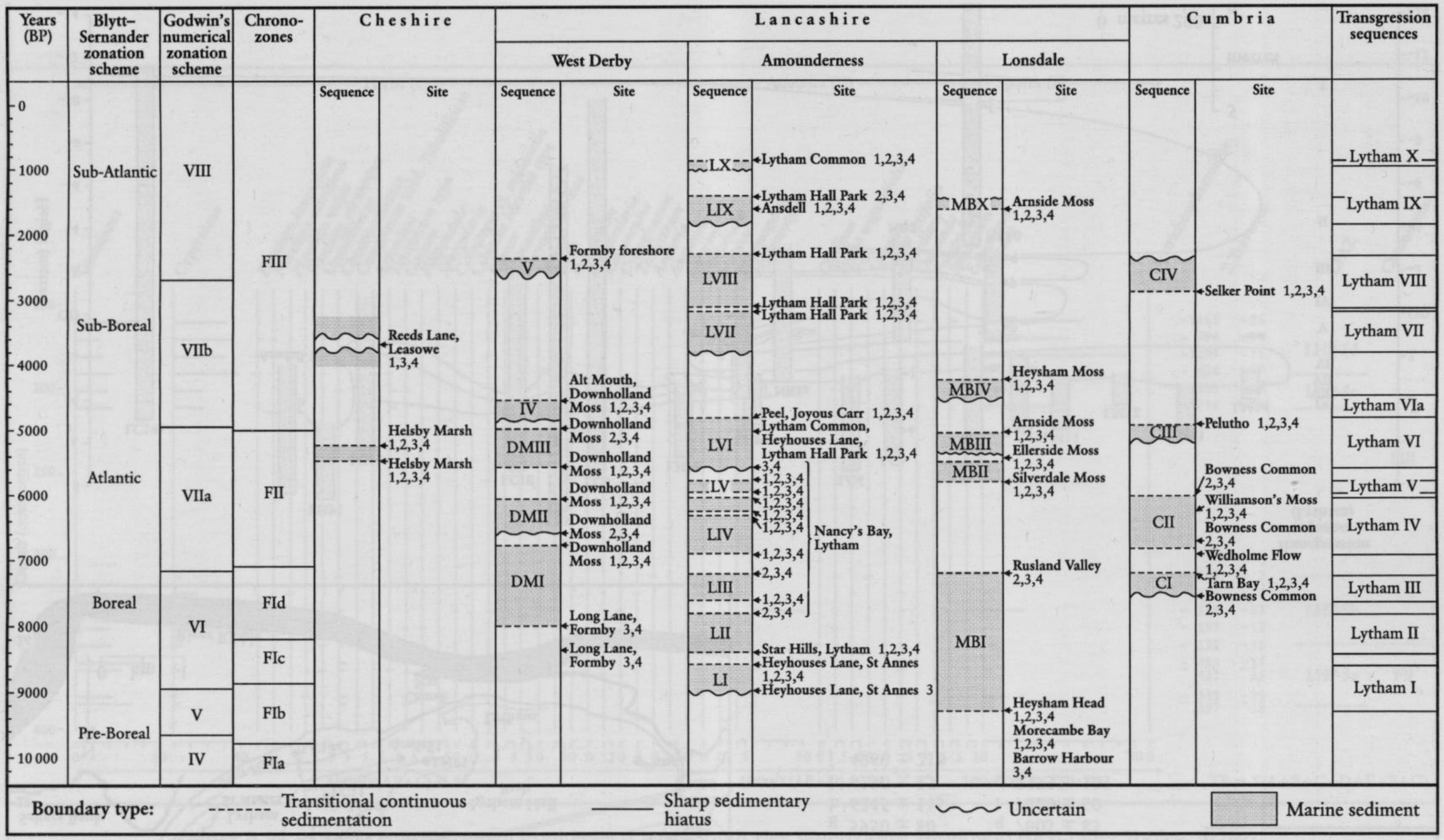


Figure 8.96 Scheme of Flandrian marine transgression sequences in north-west England (after Tooley, 1978a). Key to numbers 1–4: transgression boundary established by 1: ^{14}C chronostratigraphy; 2: biostratigraphy; 3: lithostratigraphy; 4: height in relation to OD.

from Nancy's Bay, with the transgression having a mean altitude of +2.88 m OD and comprising a blue-grey clay between 30 and 80 cm thick, with rounded pebbles. The end of this transgression is recorded throughout Lytham at the Flandrian II–III chronozone boundary, that is about 5000 years BP, at a mean altitude of +3.03 m OD. Lytham VII is recorded at the northern end of the Lytham–Skippool valley (Tooley, 1978a) as a layer of blue clay with iron partings, and in Lytham Hall Park, where its culminating phase occurred 3150 ± 150 radiocarbon years ago at an altitude of +3.15 m OD. In Lytham Hall Park Lytham VII is close to the marine limit and a short period separates it from Lytham VIII. The latter is recorded in Lytham Hall Park and is a silty clay, although farther seaward there is a transition to a silty sand. The mean altitude of this transgression is +3.68 m OD and +4.34 m OD for the regressive phase, and marine sedimentation lasted from 3090 ± 135 to 2270 ± 65 years BP. The end of Lytham IX is recorded from the south-west corner of Lytham Hall Park and between the park and the present coast. The mean altitude of the regressive phase is +4.46 m OD, although it reaches a maximum altitude of + 5.39 m OD. However, marine sedimentation was occurring nearer the coast at a lower altitude about 1370 years BP. A final transgression is inferred from Lytham, where dates from biogenic strata intercalated with the coastal dune sand ranged from 805 to 830 years BP. On the basis that marine transgression episodes are closely related to periods of dune stability, increased precipitation and biogenic sedimentation (Jelgersma *et al.*, 1970), then the dates recorded from the biogenic strata do indicate that a marine transgression of limited extent was underway. Documentary evidence from the accounts and chartulary of Lytham Monastery from the mid-15th to early 16th century (Fishwick, 1907) indicates a period of dune instability, from which can be inferred a period of relatively low sea level. A summary of the relative sea-level curves can be seen in Figure 8.97, and the time limits of the transgressions are

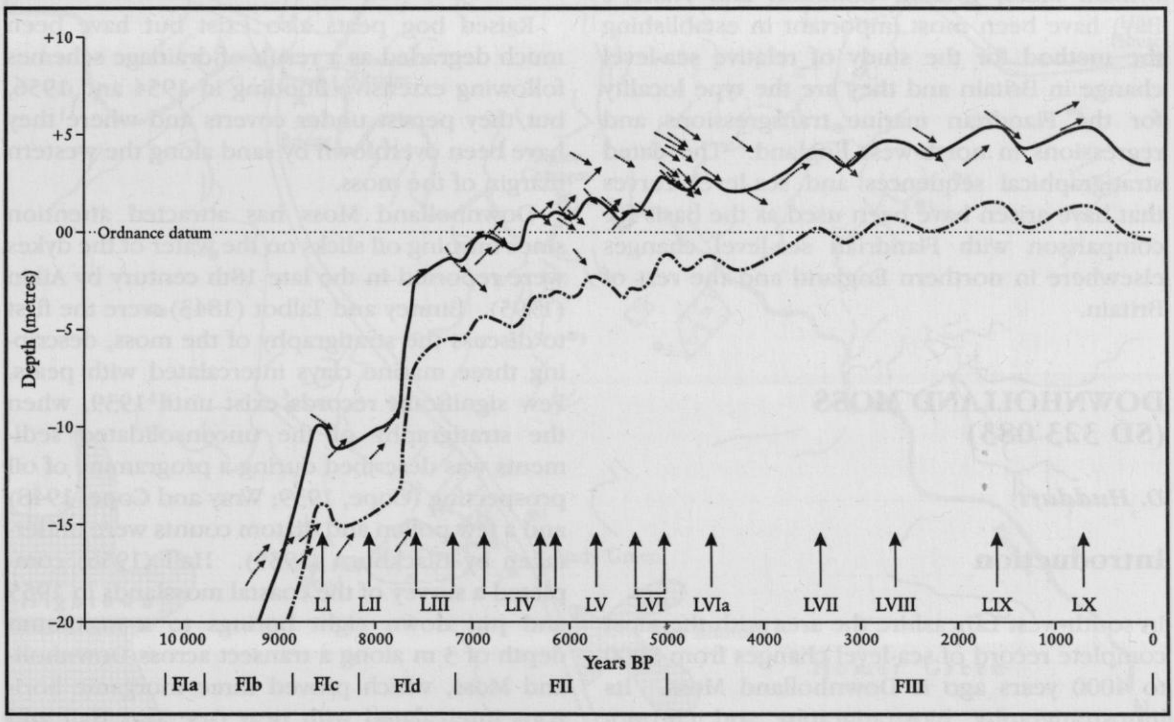


Figure 8.97 A graph to show relative sea-level changes in north-west England (after Tooley, 1978a). An arrow pointing upwards indicates a dated sample immediately below a marine deposit and an arrow pointing downwards a dated sample immediately above a marine deposit. The continuous line curve shows the change in altitude of the spring tide level, whereas the pecked line curve shows the movements of mean tide level. LI to LX are marine transgressions recorded at Lytham. Twenty-six index points establish the amplitude and period of sea-level oscillations in a restricted area of west Lancashire.

Table 8.17 Marine transgressions in the Fylde (after Tooley 1978a).

Transgression	Time limits (radiocarbon years BP)
Lytham I	9270–8575
Lytham II	8390–7800
Lytham III	7605–7200
Lytham IV	6710–6157
Lytham V	5947–5775
Lytham VI	5570–4897
Lytham VII	3700–3150
Lytham VIII	3090–2270
Lytham IX	1795–1370
Lytham X	*c.* 817

summarized in Table 8.17. A much fuller discussion of the importance of the Lytham sequences and their correlation throughout north-west England can be found in Tooley (1978a).

Conclusions

The sites that make up Lytham (Starr Hills, Lytham Moss, Lytham Common and Nancy's Bay) have been most important in establishing the method for the study of relative sea-level change in Britain and they are the type locality for the Flandrian marine transgressions and regressions in north-west England. The dated stratigraphical sequences and sea-level curves that have arisen have been used as the basis for comparison with Flandrian sea-level changes elsewhere in northern England and the rest of Britain.

DOWNHOLLAND MOSS (SD 323 083)

D. Huddart

Introduction

In south-west Lancashire the area with the most complete record of sea-level changes from 8000 to 4000 years ago is Downholland Moss. Its lithostratigraphy, biostratigraphy and chronostratigraphy have been examined in detail (Tooley, 1978a, b, 1980, 1985; Huddart, 1992) and have provided data for the construction of sea-level curves for constraining geophysical models of isostatic uplift (Lambeck, 1991), for the vegetational history of coastal lowlands during the Flandrian and for the impact of prehistoric folk on the coastal environment. It was the key area for the application, development and testing of the concepts of transgressive and regressive overlap tendencies (Shennan *et al.*, 1983).

The following palaeoenvironments are represented in the Downholland Moss sediments.

1. Tidal flat and lagoonal: represented by alternating organic and inorganic sediments of marine, brackish-water, freshwater and terrestrial origin, where the morphology includes old tidal creeks or roddons (as defined by Godwin (1938) and discussed by Huddart (1992); see (Figure 8.98) and sandbanks.
2. Perimarine: represented by alternating organic and inorganic sediments of freshwater and terrestrial origin.
3. Sand dune: represented along the western margin of the moss by surface sheets of sand and by sand layers interfingering the limnic and terrestrial organic deposits and along parts of the eastern margin by a morphology of low dune ridges.

Raised bog peats also exist but have been much degraded as a result of drainage schemes following extensive flooding in 1954 and 1956, but they persist under coverts and where they have been overblown by sand along the western margin of the moss.

Downholland Moss has attracted attention since burning oil slicks on the water of the dykes were reported in the late 18th century by Aikin (1795). Binney and Talbot (1843) were the first to discuss the stratigraphy of the moss, describing three marine clays intercalated with peats. Few significant records exist until 1939, when the stratigraphy of the unconsolidated sediments was described during a programme of oil prospecting (Cope, 1939; Wray and Cope, 1948) and a few pollen and diatom counts were undertaken by Blackburn (1939). Hall (1956) completed a survey of the coastal mosslands in 1955 and put down eight borings to a maximum depth of 3 m along a transect across Downholland Moss, which proved three inorganic horizons intercalated with peat (see also Hall and Folland, 1967). Further stratigraphical, pollen and diatom (Tooley, 1974, 1978a, b, 1985), plant macrofossils (A. GreatRex, pers. comm., 1985) and foraminiferal investigations (Huddart, 1992) have been completed.

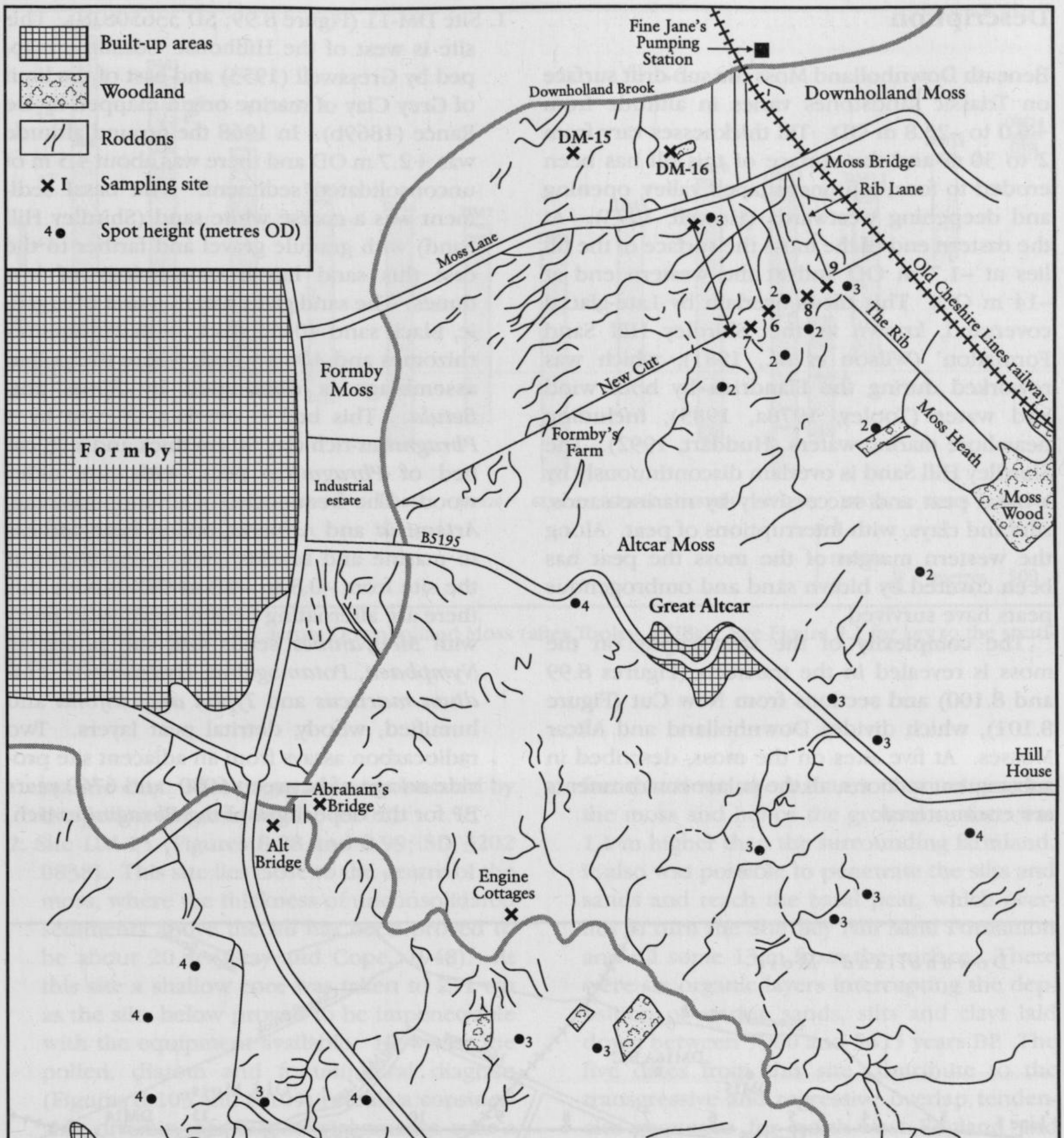


Figure 8.98 Map of the River Alt and Downholland Brook catchments showing location of Downholland Moss and sampling sites (after Huddart, 1992).

Description

Beneath Downholland Moss the sub-drift surface on Triassic sandstones varies in altitude from +6.0 to –26.8 m OD. Till thicknesses vary from 2 to 30 m and the surface of this till has been eroded to form a funnel-shaped valley, opening and deepening westwards (Howell, 1973). At the eastern end of the moss the surface of the till lies at –1.7 m OD and at the western end at –14 m OD. This till is overlain by Late-glacial coversand, known as the 'Shirdley Hill Sand Formation' (Wilson *et al.*, 1981), which was reworked during the Flandrian by both wind and water (Tooley, 1978a, 1985), including nearshore marine waters (Huddart, 1992). The Shirdley Hill Sand is overlain discontinuously by a basal peat and successively by marine sands, silts and clays, with interruptions of peat. Along the western margin of the moss the peat has been covered by blown sand and ombrogenous peats have survived.

The complexity of the stratigraphy on the moss is revealed in the transects (Figures 8.99 and 8.100) and sections from New Cut (Figure 8.101), which divides Downholland and Altcar Mosses. At five sites on the moss, described in subsequent sections, all the palaeoenvironments are encountered.

1. Site DM-11 (Figure 8.99; SD 33650819). This site is west of the Hillhouse Coastline mapped by Gresswell (1953) and east of the limit of Grey Clay of marine origin mapped by De Rance (1869b). In 1968 the ground altitude was +2.7 m OD and there was about 4.5 m of unconsolidated sediment. The basal sediment was a coarse white sand (Shirdley Hill Sand) with granule gravel and farther to the east this sand has a morphology of low dunes. The sand gives way to a unit of organic, black sand 40 cm thick, with *Phragmites* rhizomes and *Alnus* wood, where the pollen assemblage is dominated by *Pinus* and *Betula*. This bed is overlain in turn by a *Phragmites*-rich clay 50 cm thick and a 75 cm bed of *Phragmites* peat, with rare alder wood. The frequencies of Chenopodiaceae, *Artemisia* and *Armeria* pollen bear witness to marine and brackish water conditions at the site from –0.8 to –0.3 m OD. Above this there are alternating beds of freshwater peats with *Menyanthes* seeds and the pollen of *Nymphaea*, *Potamogeton*, *Hydrocotyle*, *Cladium mariscus* and *Typha angustifolia* and humified, woody, detrital peat layers. Two radiocarbon assays from an adjacent site provide an age of between 6980 and 6760 years BP for the deposition of the *Phragmites*-rich,

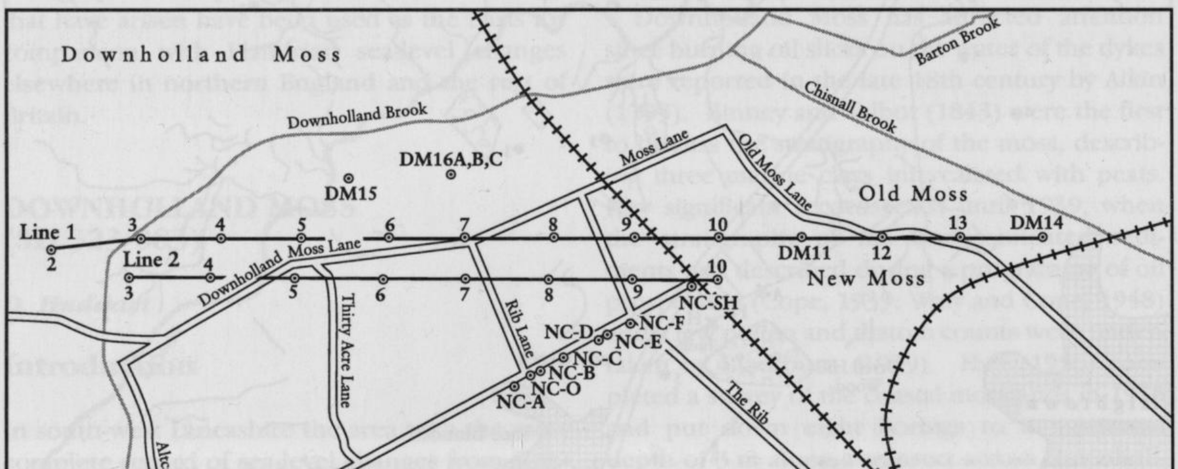


Figure 8.99 Map of Downholland Moss showing the sampling sites (after Tooley, 1985).

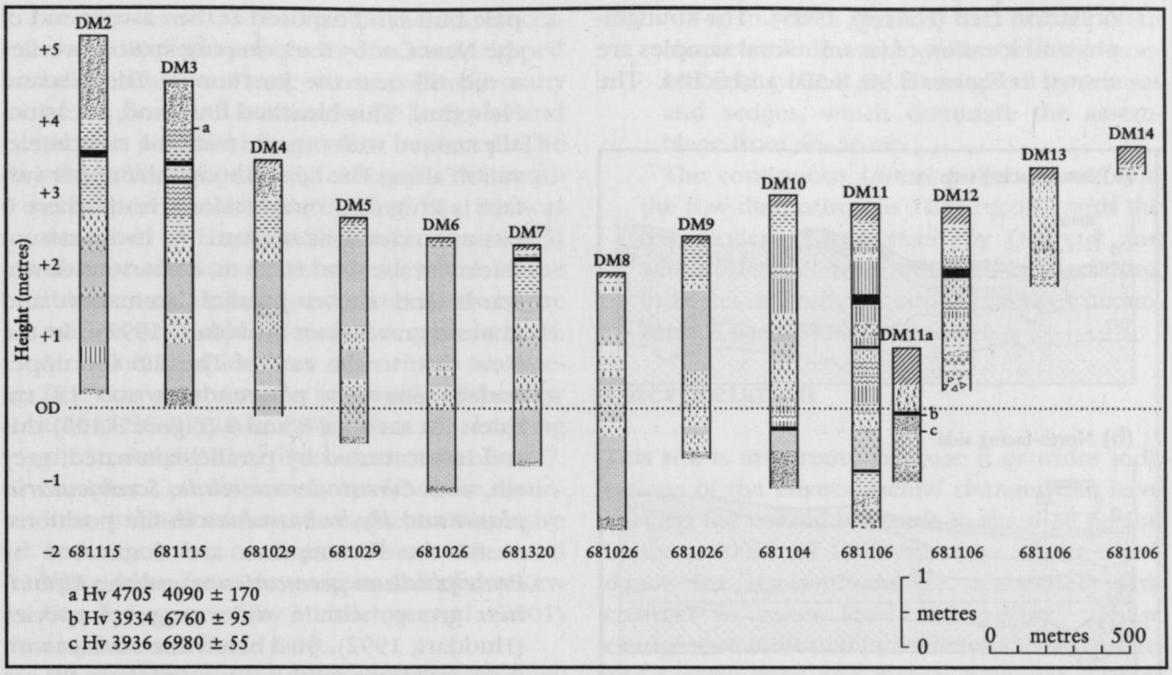


Figure 8.100 Stratigraphy across Downholland Moss (after Tooley, 1978a). See Figure 8.1 for key to the stratigraphical log.

marine clay. These ages are corroborated by the pollen diagram (Figure 8.102).

2. Site DM-15 (Figures 8.98 and 8.99; SD 3202 0838). This site lies close to the centre of the moss, where the thickness of unconsolidated sediments above the till has been proved to be about 20 m (Wray and Cope, 1948). At this site a shallow core was taken to 271 cm as the silts below proved to be impenetrable with the equipment available. However, the pollen, diatom and foraminiferal diagram (Figures 8.103 and 8.104) reveals a considerable diversity of palaeoenvironments over a short time period, established by a series of radiocarbon dates from 6750 to 6050 years BP . In 1968 the ground altitude was +2.4 m OD, which is 1.7 m below the present Mean High Water Mark at Formby. The basal sediments are silts, with sandy laminations, passing up to a *limus*, with *Phragmites* rhizomes and overlain by a laminated silt and a second *limus* layer dated from 5565 to 6050 years BP. This in turn is overlain by a laminated clay and a *limus* with *Phragmites* rhizomes.
3. Site DM-16 (Figures 8.98 and 8.99; SD 3238 0839). The importance of this site lies in the fact that it was the last area of uncut peat on the moss and hence the ground surface was 1.1 m higher than the surrounding farmland. It also was possible to penetrate the silts and sands and reach the basal peat, which overlies in turn the Shirdley Hill Sand Formation and till some 13 m from the surface. There were six organic layers interrupting the deposition of marine sands, silts and clays laid down between 7960 and 5615 years BP. The five dates from this site contribute to the transgressive and regressive overlap tendencies sequence for north-west England and have been used as sea-level index points on the sea-level age–altitude graph for north-west England (Tooley, 1978a, 1982).

   The detailed stratigraphy is given and discussed in Middleton *et al*. (2001).
4. New Cut and The Rib between Moss Bridge and Moss Heath. Stratigraphical and micropalaeontological details from drainage, excavation and construction work in this area on the boundary with Altcar Moss (Figure 8.98) have added to Tooley's earlier work. This area is known from 19th century tithe award apportionments as 'Segar's Moss' and had been reclaimed in the 1750s by Edward Segar

of Barton Hall (Harrop, 1985). The stratigraphy and location of foraminiferal samples are shown in Figures 8.98, 8.101 and 8.104. The

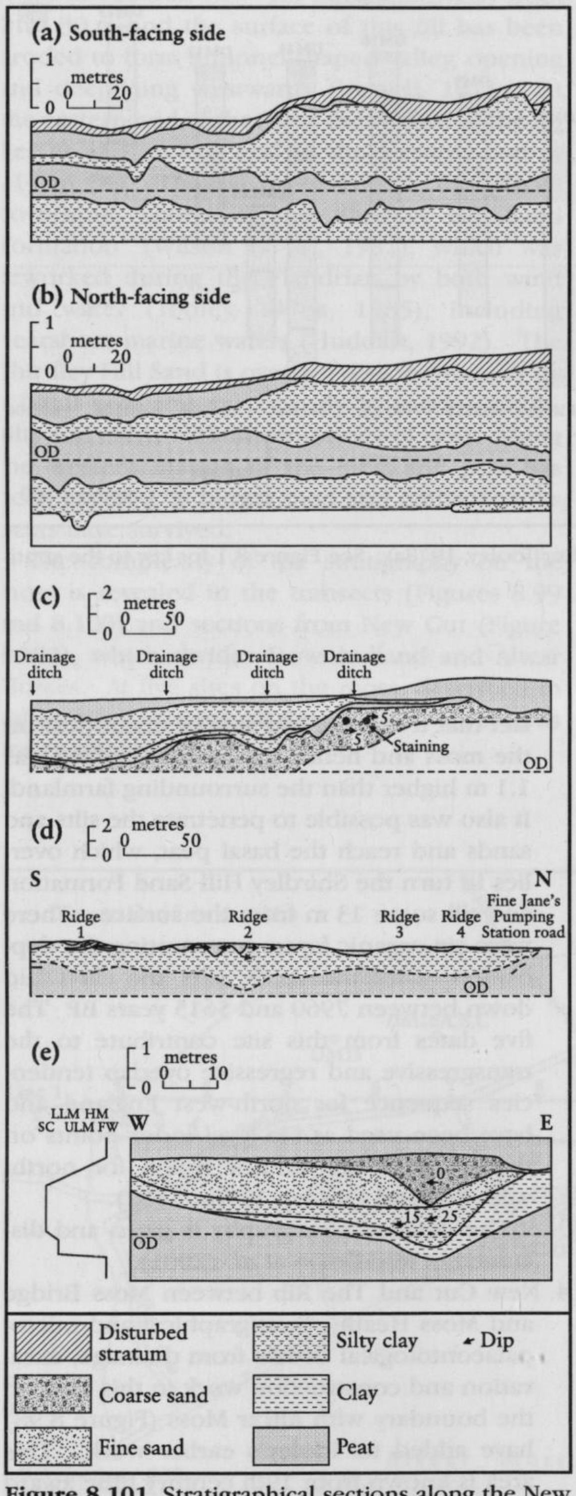


Figure 8.101 Stratigraphical sections along the New Cut (after Huddart, 1992). See Figure 8.103 for key to labels in (e).

pale buff sand exposed at the eastern end of the New Cut by the pumping station overlies a red till near the junction of The Rib and New Cut. This bleached fine sand, occasionally stained with organic material, rises in elevation along The Rib, although the upper surface is irregular, towards Moss Heath where it passes under 3 m of peat. At its maximum thickness the sand is 1.3 m, it coarsens downwards and shows parallel lamination and granule gravel lines (Huddart, 1992). In the New Cut, to the east of The Rib the upper surface shows a pararendzina soil 19 cm thick. In sections 8 and 9 (Figure 8.105) this soil is succeeded by parallel-laminated, grey silt, with *Cerastoderma edule*, *Scrobicularia plana* and *Hydrobia ulvae* in life positions. At 8A the Foraminifera are dominated by *Protelphidium germanicum* and the *Elphidium* group, with a wide range of species (Huddart, 1992). Just below the contact with the overlying peat the Foraminifera in 8B are dominated by *Trochammina inflata*, *Jadammina macrescens* and *Protelephidium germanicum*. Above is 75 cm of peat, which infills hollows in the top of the grey silt. Its upper surface is very irregular because 'flames' of peat rise into the overlying grey silty clay, which slightly coarsens upwards. *Scrobicularia plana* is located in growth position towards the base of this unit. The Foraminifera are dominated by *P. germanicum*. Above are grey silts with occasional fine sand laminae and convolute lamination where these fine sands occur. The Foraminifera are dominated by *P. germanicum* in 8F but in 8G there is a significant percentage of *J. macrescens*. The silts pass to grey silty clay and are capped by 65 cm of peat. There are similar foraminiferal assemblages in section 9 but in 9F, although *P. germanicum* dominates, there are 58 species present and the faunal diversity is much higher than in the rest of this section (Huddart, 1992). Farther south-west along the New Cut (Figure 8.101c) a similar stratigraphy is seen, with grey silts and peat at the base of the cut, grey clay coarsening upwards into horizontal, iron-stained, parallel laminated fine sand and silt, with root channels and peat clasts, capped by peat. Foraminiferal numbers are not high, except in 2D and *P. germanicum* tends to dominate, except in 2I and 2K, where *J. macrescens*, *T. inflata* and *P. ger-*

manicum are co-dominant. There are occasional rippled surfaces in fine sand, capped by silt laminae. Flame structures cause many of the fine sand laminae to be irregular and there is convolute lamination in places. The silts infill a pre-existing channel. The stratigraphy in the New Cut east of The Rib was described by Heptinstall (pers. comm., 1983) and also shows considerable variations in the thicknesses, altitudes and composition of the marine beds and intercalated organic beds. Monolith samples were taken from the free-face excavations on the south side of the New Cut at NC-F (Figure 8.99), close to sampling site 9, and at NC-A, 65 m south-west of site 7. Samples 2 cm thick were cut from the monoliths to radiocarbon date the transgressive and regressive overlaps. These dates and three taken from the eastern end of the New Cut and reported by Middleton *et al.* (2001) are shown in Table 8.18.

A drainage cut, at right angles to the New Cut and parallel to Rib Lane (Figure 8.98), revealed the stratigraphy in the upper part of the succession, as shown in Figure 8.101d. In the field can be seen a series of low-amplitude ridges and the drain bisects four of them. This morphology can be picked out on aerial photographs where they focus on Downholland Brook and the River Alt. These ridges are further described and discussed in Huddart (1992).

5. Altcar 1 (SD 3304 0738). This site is 500 m south of the New Cut and The Rib intersection. From the pollen diagram (Figure 8.106) three local pollen-assemblage zones were identified (Middleton *et al.* 2001).

(i) ALT-a, in which the tree pollen is dominated by *Quercus* and *Pinus* is co-dominant. A feature (Figure 8.106) is the rise and fall of freshwater taxa, particularly *Typha angustifolia*.

(ii) ALT-b where *Quercus* remains dominant with *Alnus* as a subdominant. *Pinus* values decline and *Ulmus* values are low but persistent. *Tilia* values are low and discontinuous. The pollen of Chenopodiaceae is present throughout the zone and rises immediately before the lithological change from a well-humified monocot peat with *Phragmites* rhizomes to a silt clay with monocots, including *Phragmites* and *Cladium* rhizomes.

(iii) ALT-c, in which *Quercus* and *Alnus* pollen characterize the tree pollen and the non-arboreal pollen are dominated by grasses and sedges, which dominate the assemblage from 46–56 cm.

The continuous *Ulmus* pollen record and the low discontinuous *Tilia* record, with the tree pollen characterized by *Quercus* and with *Alnus* and *Pinus* pollen values declining, indicates an early to mid-Flandrian Chronozone II age for the succession.

Interpretation

This site is important because it provides indications of the environmental changes that have affected the coastal lowlands in the time period between 8000 and 4000 years ago. There is evidence for sea-level change, associated with changes in water level and quality. These changes are indicated by variations in sediments and foraminiferal and diatom biozones. There are associated vegetation changes indicated by the pollen and macrofossil record and some evidence of human activity influencing this vegetation. The changes in landscape have been dated and the evidence from Downholland Moss is important in our overall understanding of the coastal ecosystem development in north-west England.

Downholland Moss is the classic site from which coastal palaeoenvironmental change and associated sea-level change has been documented in south-west Lancashire. From this area there has been a long history of theories advanced to explain the distribution of coastal landforms and lithostratigraphy in the tidal flat, perimarine, dune and saltmarsh zone. These have been summarized as follows (Tooley, 1980):

1. the breaching of sand barriers and the shoaling of inlets during a period of static sea-level (Binney and Talbot, 1843);
2. land subsidence and uplift during a period of static sea-level but with relative rises and falls of sea-level (Reade, 1871);
3. an oscillating sea-level caused by the interplay of land uplift and the eustatic rise of sea-level (Gresswell, 1953, 1957).

However, the bio- and litho-stratigraphy from this moss indicate three periods of positive and negative sea-level tendencies between 6980 and 5615 years BP, with age indicated absolutely by

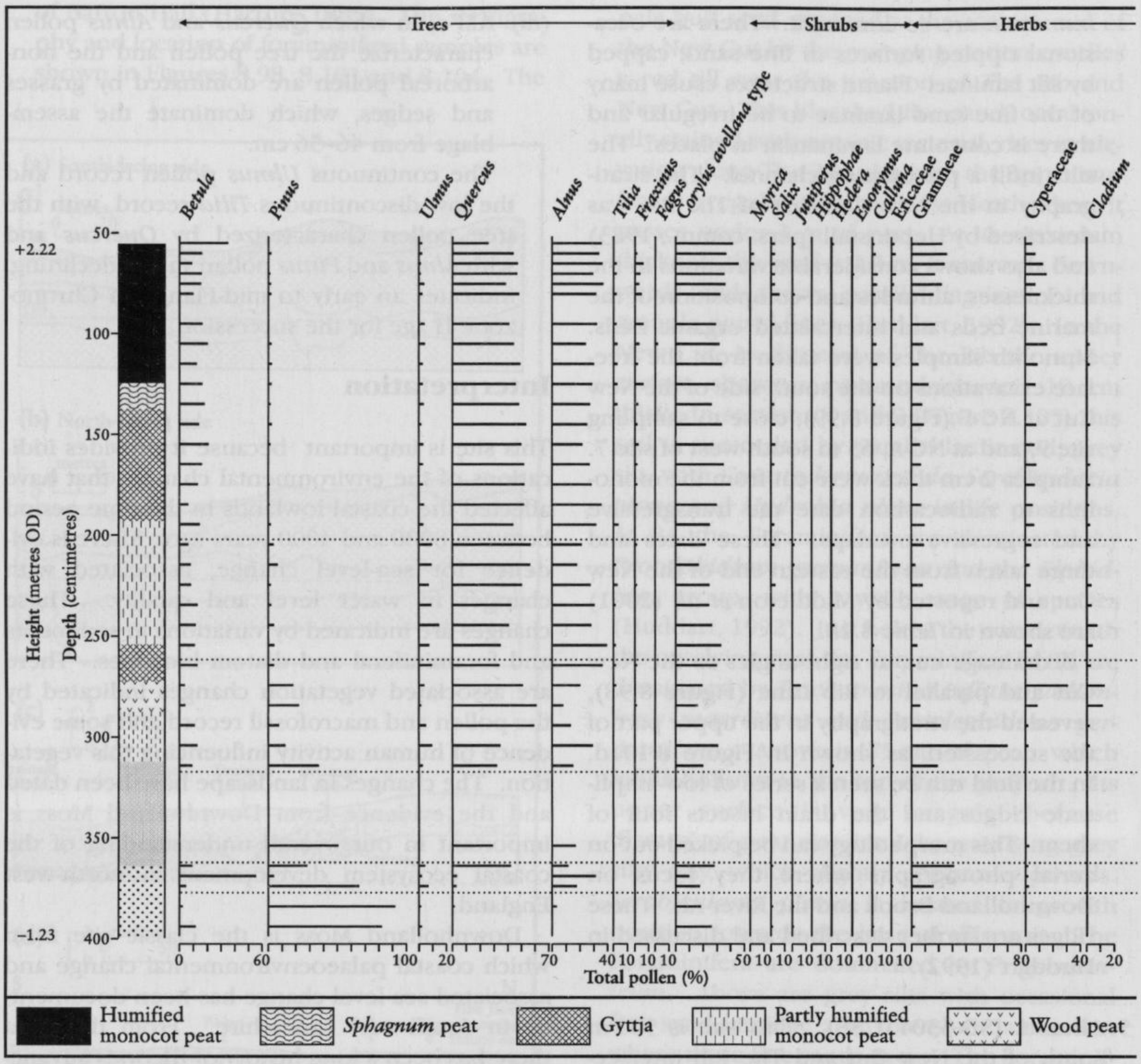
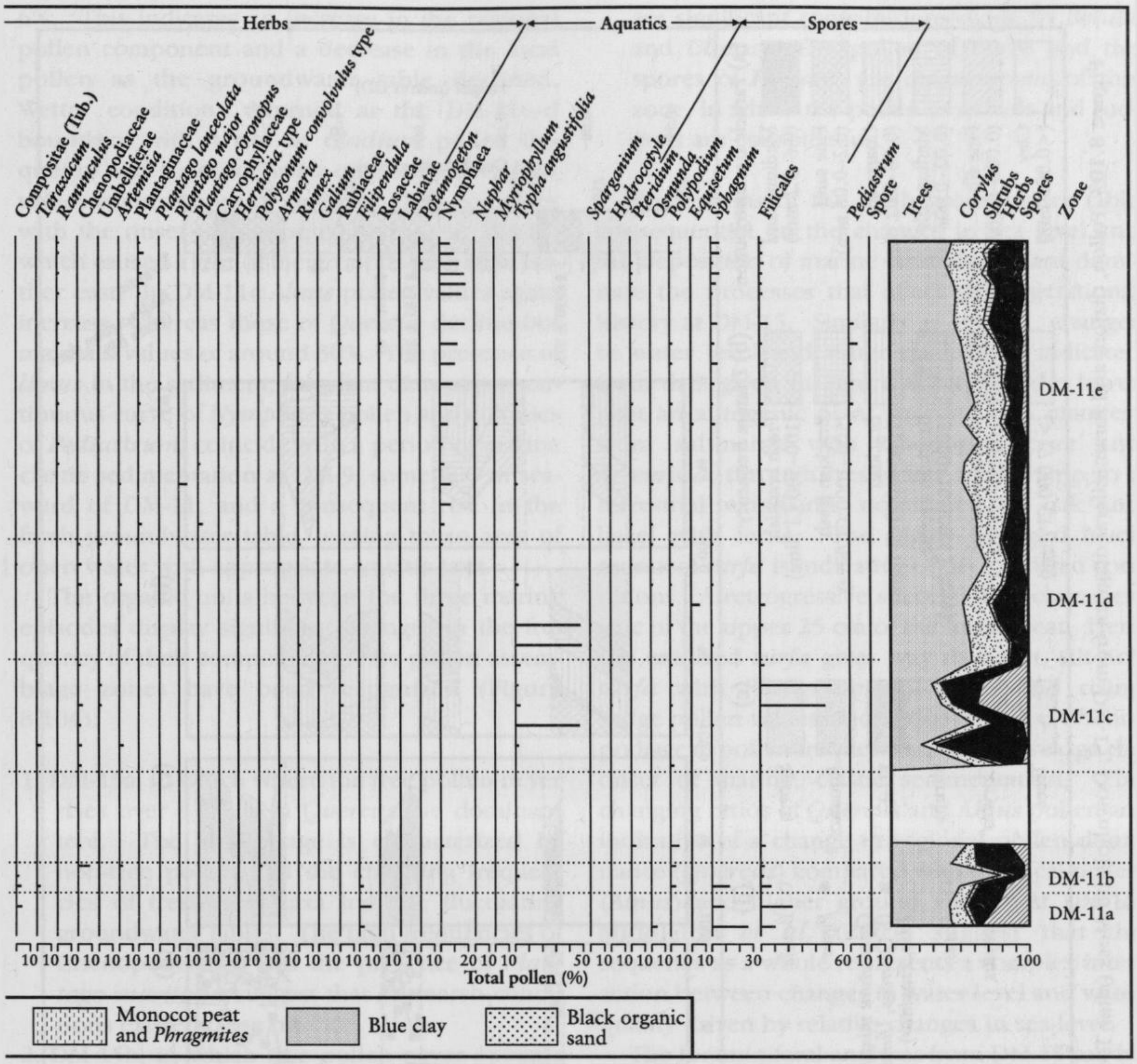


Figure 8.102 Pollen diagram from DM-11, Downholland Moss (after Tooley, 1978a).

^{14}C dating and relatively by the pollen-assemblage zones. Micropalaeontological and sedimentological analyses indicate fundamental changes in water depth and water quality during accumulation of both the minerogenic and the biogenic sediments (Tooley, 1978b; Middleton *et al.*, 2001). In the New Cut section there is evidence for three units of silt and clay with quiet brackish to marine water indicators in the macrofossils and in all the sedimentary indicators. These clastic units are subdivided by terrestrial, telematic or limnic peats. This is similar to the interpretation for DM-15 (Tooley, 1974, 1978a, b, 1985; Huddart, 1992; Middleton *et al.*, 2001), where three transgressive and regressive overlaps occurred progressively higher in the stratigraphical column and with progressively younger dates. Each transgression shows facies changes that permit an interpretation of a rising sea-level followed by a falling sea-level. This is indicated by a saltmarsh regime succeeded by a higher to lower mud-flat sedimentary environment and later by a saltmarsh regime, before brackish water, limnic biogenic deposits accumulated.

In DM-11 the major environmental event is the marine transgression that occurred at the beginning of Flandrian Chronozone II, at alti-



tudes from –0.87 to –0.36 m OD. The onset of marine conditions at 6980 years BP occurred when the regional forest was dominated by pine, in an arrested succession on the sandy soils that had developed on the Shirdley Hill Sand Formation. Nevertheless, the mixed-oak forest taxa were present and marine conditions were anticipated by the presence of wetland pollen taxa, such as *Typha angustifolia* and open freshwater taxa and then saltmarsh conditions, with Chenopodiaceae. The marine event lasted only 200 radiocarbon years and DM-11 and DM-11A are close to the eastern, landward marine limit, which is about 500 m farther inland than the marine limit mapped by De Rance (1869b), but *c.* 1.4 km seaward of the Hillhouse Coastline of Gresswell (1953, 1957). In the overlying 3 m of organic sediments there is abundant evidence of changing freshwater levels, associated with the repetitive marine transgressions and regressions documented at other sites farther west from DM-10 (Figure 8.99).

During pollen-assemblage zone DM-11c (Figure 8.102) there is an autogenic plant succession from saltmarshes to freshwater reedswamps and thence to dry woodland. The pollen frequency of *Quercus* increases to 50% of the tree pollen, whereas that of *Alnus* declines to

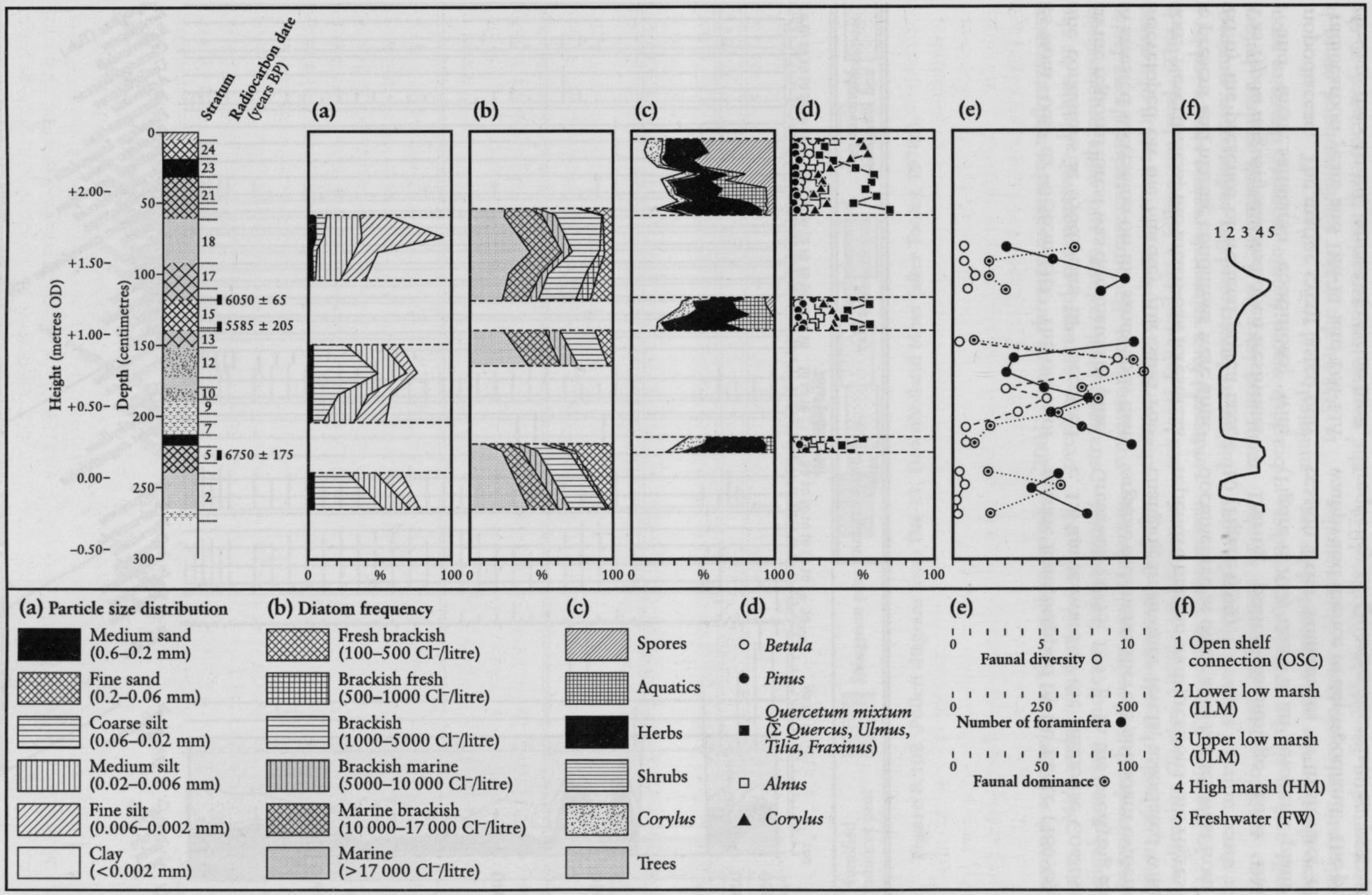


Figure 8.103 Pollen, diatom and foraminiferal diagram from DM-15, Downholland Moss (after Tooley, 1978a). See Figure 8.1 for key to stratigraphy.

6%. This indicates an increase in the regional pollen component and a decrease in the local pollen as the groundwater table declined. Wetter conditions returned at the DM-11c–d boundary, with a rise in *Cladium* pollen frequency and a rise in *Typha angustifolia*. *Alnus* values exceed those of *Quercus*. This coincides with the onset of marine conditions at DM-10, which caused a rise of the groundwater table farther east. In DM-11e *Alnus* pollen values again increase, whereas those of *Quercus* decline but maintain values of around 30%. The presence of *limus* in the sediment, frequent diatoms, a continuous curve of *Nymphaea* pollen and colonies of *Pediastrum*, coincide with a period of marine clastic sedimentation at DM-9, some 600 m seaward of DM-11, and a consequent rise in the fresh groundwater table, leading to an area of open water with appropriate aquatic taxa.

The organic units between the three marine episodes display significant changes in the frequency of their components. Five pollen-assemblage zones have been recognized (Figure 8.104):

1. DM-15a, in which where the tree pollen never rises over 45%, with *Quercus* the dominant tree. The assemblage is characterized by non-tree pollen and the changing frequencies of freshwater taxa indicate fluctuating groundwater tables. The high frequencies of Chenopodiaceae and the presence of *Plantago maritima* suggest that saltmarsh conditions existed close by.
2. DM-15b, in which tree pollen never exceeds 25% and is dominated by *Quercus*. Of the herbaceous pollen, grasses are dominant and the characteristic of this zone is the rise and fall in aquatic taxa frequency.
3. DM-15c shows an increase in tree pollen, dominated by *Quercus*. Gramineae values decline and those of the aquatics rise and fall once again. As the frequency of the aquatic pollen declines so that of saltmarsh taxa increases. In addition there is an increase in the number of dinoflagellate cysts as the boundary with the overlying marine strata is approached.
4. DM-15d is still dominated by *Quercus* tree pollen, although the zone is characterized by non-tree pollen. Grasses are accompanied by saltmarsh taxa.
5. DM-15e, in which the tree pollen is characterized by *Quercus* and *Alnus*, although there are significant contributions made by *Betula* and *Ulmus*. The pollen of herbs and the spores of *Filicales* are characteristic of this zone, in which the pollen of cereals and ruderals are conspicuous.

Fluctuations of the fresh groundwater table consequential on the changes in sea level and the deposition of marine clastic sediment dominate the processes that affect the vegetational history at DM-15. Similarly at Altcar 1, changes in water level and water quality are indicated from data given in Figure 8.106. In the lower peat an autogenic plant succession is manifest from saltmarsh with Chenopodiaceae and *Artemisia*, through freshwater reedswamp to a terrestrial woodland, dominated by oak and hazel with ferns. The highly oxidized black monocot *turfa* is indicative of dry, oxidized conditions. A retrogressive succession is characteristic of the upper 25 cm of the lower peat. Here the oxidized *turfa* gives way to a wet, silt-rich *turfa* with *Phragmites*, *Cladium* and rising sedge pollen values, succeeded by rising Chenopodiaceae pollen frequencies, which presage the onset of marine, clastic sedimentation. The changing ratios of *Quercus* and *Alnus* pollen are indicative of a change in regional pollen dominance (*Quercus*) compared with the local pollen (*Alnus*) and higher ground water. At DM-16, Middleton *et al*. (2001) suggest that the sequence as a whole represents a complex interaction between changes in water level and water quality driven by relative changes in sea-level.

The foraminiferal analyses from DM-15 can be referred to the three marine episodes; they have confirmed the sedimentary environments and have added detail about these marine phases (Huddart, 1992). From the DM-15 core five foraminiferal biozones can be identified. These biozones, with their approximate environmental locations, are as follows:

1. *Jadammina macrescens* biozone on the high marsh;
2. *Jadammina macrescens–Trochammina inflata* biozone on the high marsh;
3. *Jadammina macrescens–Protelphidium germanicum* biozone at the high marsh to low marsh transition;
4. *Ammonia batavus–Jadammina macrescens–Elphidium excavatum–Protelphidium germanicum* biozone on the upper to middle part of the low marsh;

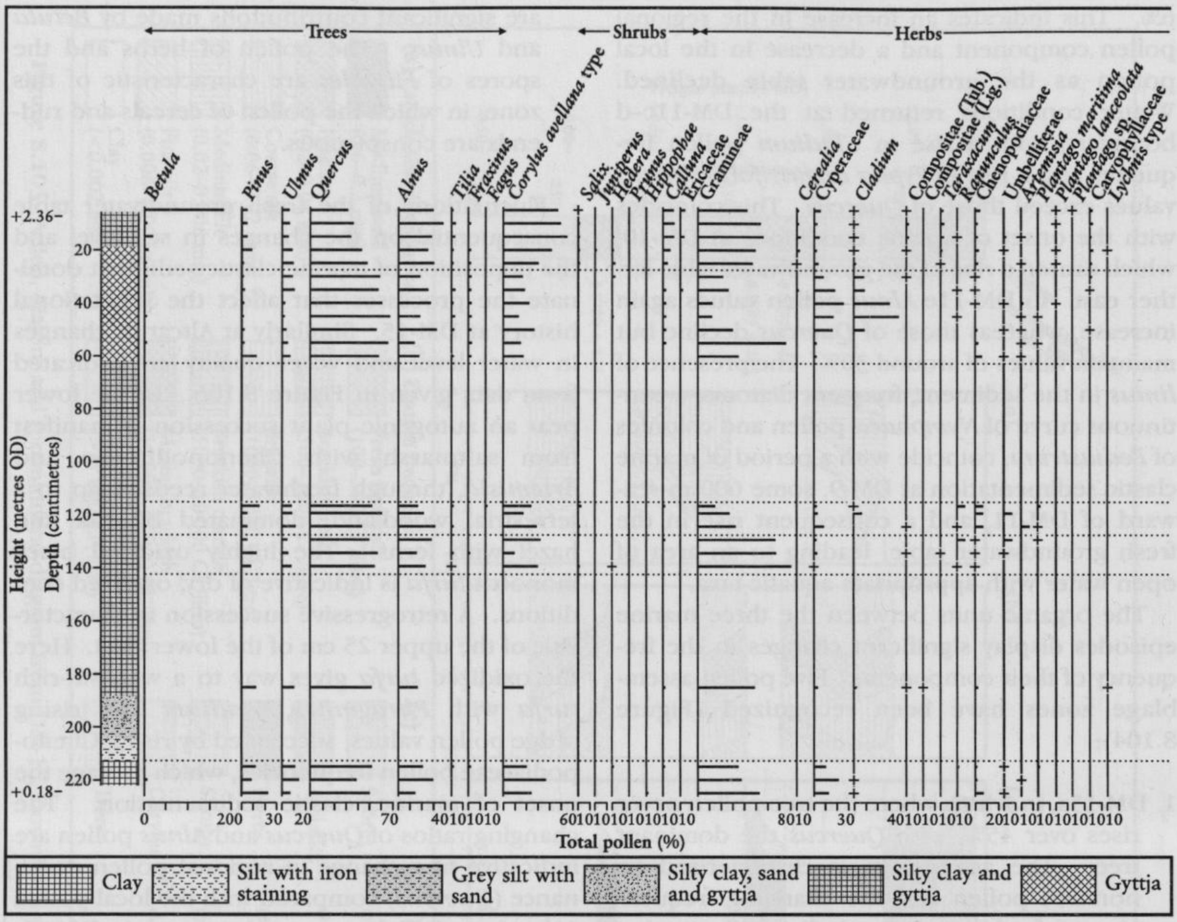
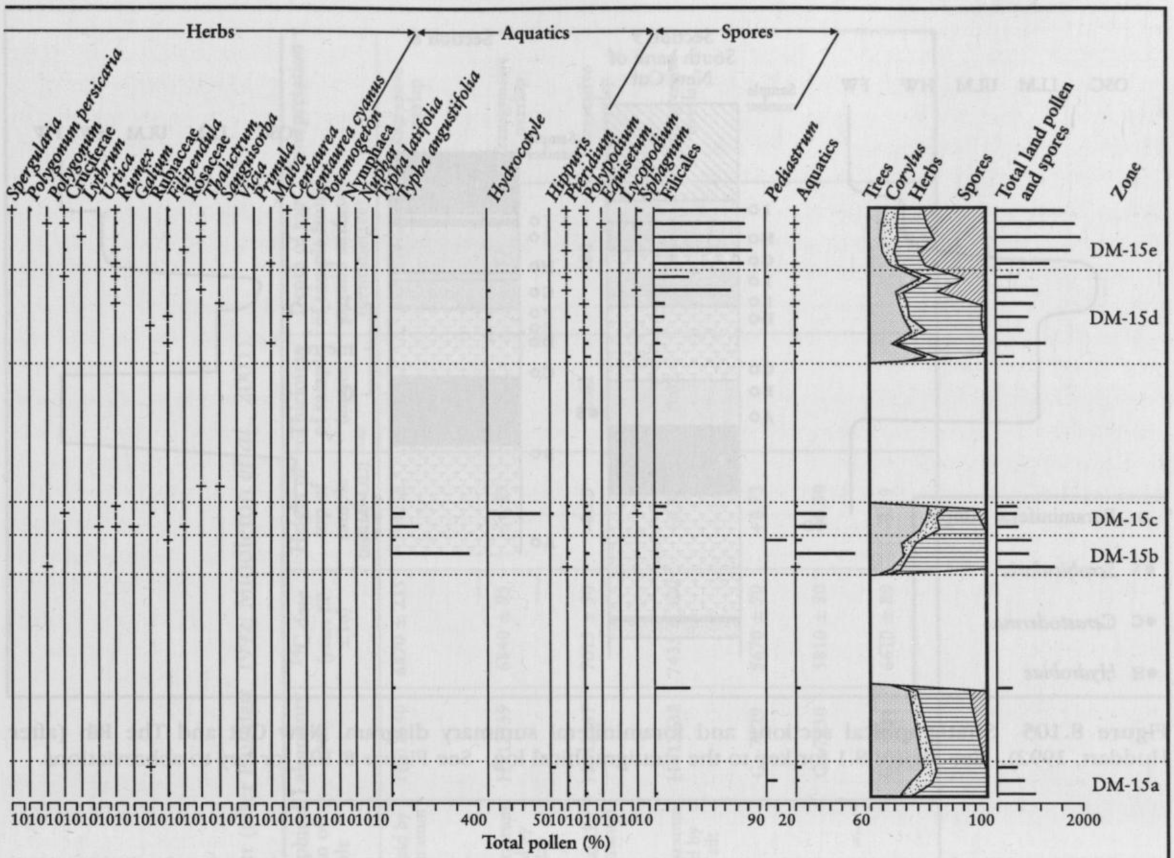


Figure 8.104 Pollen assemblage zones from DM-15, Downholland Moss (after Tooley, 1978a).

5. ***Protelphidium germanicum–Elphidium excavatum–Elphidium articulatum*** biozone from the lower low marsh.

The more detailed work from the New Cut and The Rib adds to this interpretation. The lowest silt passes from a relatively species-rich marine lagoon with an assemblage dominated by *Protelphidium germanicum* and *Elphidium* species and a large number of small, current-transported inner shelf species (8A, Figure 8.105) to the upper and middle section of a salt-marsh indicated by the low-diversity assemblage. The lower assemblage must indicate an active sediment source linking with the open shelf. In fact there are three foraminiferal samples at different stratigraphical positions that are marked by similar characteristics: a high number of species (49–58) and a high number of Foraminifera present in the samples; *Protelphidium germanicum* is dominant (45–85%), with *Elphidium articulatum* (2–8%); other *Elphidium* inner shelf species present; a large number of smaller, inner shelf species (Huddart, 1992). Their origin must be from the main sediment and water input linking the inner marine shelf to the quiet water lagoon environment and this seems likely to have been to the south-west via the Alt estuary (Huddart, 1992). The cause of these exceptional foraminiferal inputs to Downholland Moss, although not exceptional in the Alt estuarine sediments, is probably storm-induced flood conditions, which flushed Foraminifera from the inner shelf far into the quiet water basin via channels linking the salt-marsh to the estuary. Hence a sixth foraminiferal biozone can be added: *Protelphidium germanicum–Elphidium*–small inner shelf species biozone from the lowest low marsh and estuarine channel.



The morphology of low amplitude, sinuous ridges, west of Rib Lane, and obvious elsewhere on Downholland and Altcar Mosses, was described by Tooley (1985). They were interpreted by Huddart (1992) as a series of roddons, which were described first by Skertchly (1877) from the Fenland, and defined by Fowler (1932, 1934) as banks of laminated silt meandering through the peat fens. The banked form of the roddons was, according to Godwin (1938), the result of the natural levee of a tidal creek, and MacFadyen's (1933) work on Foraminifera supported this view, as he suggested that the roddon silts appear to have been deposited from tidal estuarine water flowing up the ancient waterway. He also suggested that the roddon silts were rich in species of estuarine origin brought in up the channel from the sea. This is similar to the explanation for the high-diversity samples lower in the stratigraphical sequence and the roddon characteristics are seen in cross section and vertical succession in Figure 8.101. The pattern of roddon channels feeds towards the present-day location of the Downholland Brook and the River Alt and this last phase of saltmarsh and tidal-channel sedimentation therefore was controlled by higher land to the west, likely to be the dune system and the influence of the palaeoestuary, which may have been in a similar location to the present-day River Alt. Other evidence to suggest this interpretation of the latest sedimentation phase comes from the change in orientation of Downholland Brook from an east–west to north–south flow and the distribution of Downholland Silt, which forms an embayment in the lower Alt valley extending almost to Homer Green (Wray and Cope, 1948). The estuary and connection to the sea must have been to the south-west of Downholland Moss and not directly to the west, which had been the

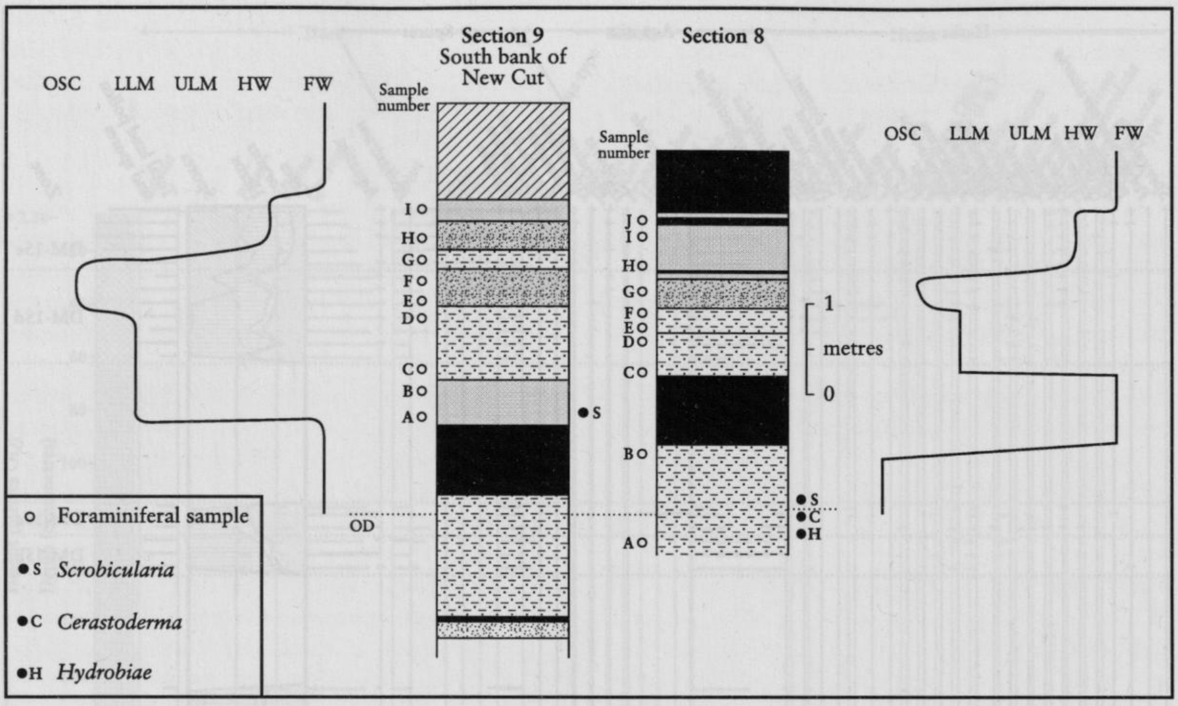


Figure 8.105 Stratigraphical sections and foraminiferal summary diagram, New Cut and The Rib (after Huddart, 1992). See Figure 8.1 for key to the stratigraphical log. See Figure 8.103 for key to abbreviations.

conclusion of Tooley (1978a). There is likely to have been a dune barrier to the west for the whole of the evolution of the sedimentary succession on Downholland Moss. In core 2 (Figure 8.99), sand up to 1 m thick has overblown the peat and forms the landward limit of the modern dunes. A ^{14}C date on peat immediately subjacent to the sand yielded a date of 4090 years BP and blown sand is incorporated into the surface peat farther east. In core 3, 25% of the upper stratum is sand, which levens the limnic peat and indicates that sand was blowing into the lagoon at that time. Tooley (1990) and Innes and Tooley (1993) suggest that the dune formation was younger than the Elm Decline (younger than 5000 years BP) but Pye and Neal (1993a) provide a date of 5110 ± 70 years BP from dune-slack peat in a core on the eastern margin of the Ainsdale National Nature Reserve (Figure 8.107). Blown sand also was recorded in the sieve fractions as occasional fine sand grains in DM-16, which indicates blowing sand on the coast from dunes to the west of the site prior to 6000 radiocarbon years ago (Middleton *et al*., 2001). It appears probable, however, that there has been a sand barrier in this area back to 8400 years BP, in order to provide the quiet-water environment for deposition of the Downholland Silt, associated organic sediments and transgressive wind-blown sand sheets in back-barrier marine lagoons that had a connection to the sea to the south-west from time to time. The stratigraphy west of Downholland Moss (Tooley, 1978a) clearly shows a sand barrier beach with incorporated shell fragments up to +3 m OD, and it is conceivable that windblown sand also may have been incorporated within or on top of this barrier from an early date.

There are some indications of prehistoric human activity on the moss. At Altcar 1, however, these are minor, although the peak frequencies of Gramineae (which may be a manifestation of groundwater fluctuations), the low, discontinuous frequencies of the spores of bracken and of *Taraxacum*, all of which occur prior to the Elm Decline, may indicate human influence on the vegetation. However, the paucity of evidence for human activity both pre-dating and post-dating the Elm Decline may be because of remoteness attributable to the sedimentary environments present around this site. For example, there are two marine episodes separated by episodes in which reedswamps were characteris-

Table 8.18 Radiocarbon dates from the New Cut (after Huddart, 1992; Middleton *et al.*, 2001).

Site name	Coordinates	Grid reference	Material dated (after Troels-Smith, 1955)	Palaeoenvironment represented	Stratigraphical position of sample	Laboratory code	^{14}C date (years BP $\pm 1\sigma$)	Height of top of sample (metres OD)	Thickness of sample (metres)	Depth of top of sample from ground surface (centimetres)	Interpretation
New Cut-A	55°33'39"N 03°01'05"W	SD 3260 0762	Sh4, Th(*Phra*)^2 + Th(*Cladii*)^2 + Humous substance with *Cladium* and *Phragmites*	Saltmarsh to reedswamps	Silt overlaid by organic stratum	Hv.12540	6870 ± 235	+0.52	0.02	134	Regressive overlap
New Cut-A	55°33'39"N 03°01'05"W	SD 3260 0762	Sh4, Th(*Phra*)^2 + Humous substance with *Phragmites*	Reedswamps to saltmarsh	Organic stratum overlaid by silty clay	Hv.12539	6840 ± 95	+0.99	0.02	87	Transgressive overlap
New Cut-F	53°33'47.5"N 03°00'42"W	SD 3304 0787	Ld^{3}3, Th(*Phra*)^{2}1 Laminated *limus* with *Phragmites*	Saltmarsh to reedswamps	Silt overlaid by organic stratum	Hv.12537	7015 ± 90	–0.20	0.02	180	Regressive overlap
New Cut-F	53°33'47.5"N 03°00'42"W	SD 3304 0787	Ld^{3}4, Th^3 + Laminated *limus*	Reedswamps to saltmarsh	Organic stratum overlaid by clayey silt	Hv.12538	7435 ± 300	+0.16	0.02	144	Transgressive overlap
New Cut			Th^2(*Phra*)3, Sh1, Dl$^+$ Dh^{++}	*Phragmites* turfa		Gu-7229	5670 ± 70	+0.73			
New Cut			Dh^3, Sh1, Ag$^+$ Dl$^+$ Th(*Phra*)1$^+$	Woody detritus		Gu-7230	5810 ± 80	+0.60			
New Cut			Th^2(*Phra*)3, Sh1, Ag$^+$ Dh^{++}	*Phragmites* turfa		Gu-7231	6610 ± 80	–0.19			

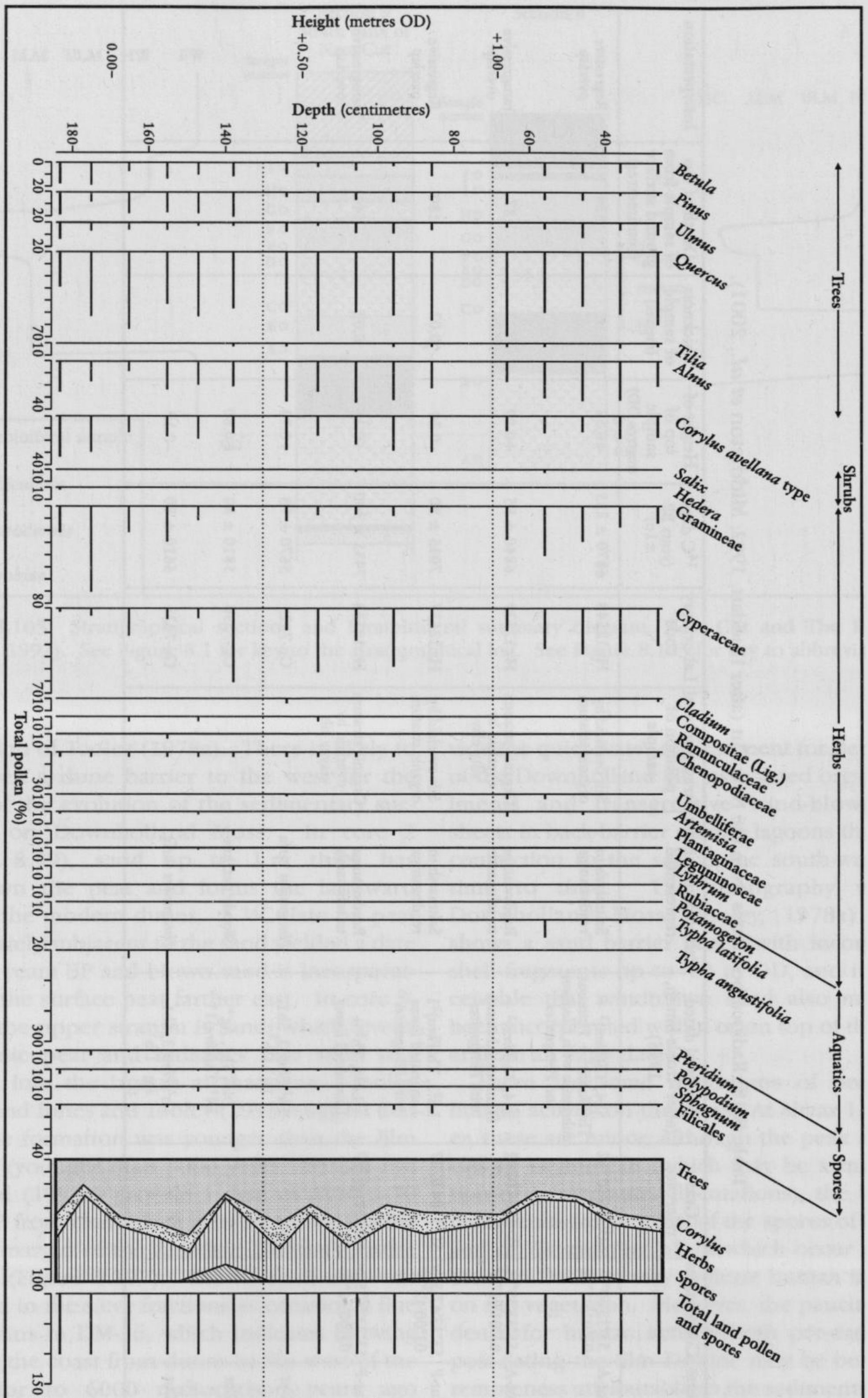


Figure 8.106 Pollen diagram at Altcar 1 (after Heptinstall, unpublished data 1983).

tic, and these environments clearly are not suitable for pastoral or agricultural farming, although they would be attractive as a resource. At several levels in the upper organic stratum at DM-16, between 190 and 5 cm, there is some evidence of human activity. There are low frequencies of ruderals or single grains of *Rumex*, *Urtica*, *Plantago lanceolata* and *Taraxacum* and the spores of *Pteridium*. Charcoal was recorded at several levels pre-dating the Elm Decline. This indicates human activity on dry land nearby, either on islands of till and coversand in the wetland, or the rising subcrops of till farther east. There are some tantalizing correlations, such as the presence of low frequencies of *Pteridium aquilinum* spores associated with charcoal at several levels. As bracken is associated with forest clearance it may be that this remote wetland site clearance on the drier, sandy ground to landward is being recorded. However it is necessary to move closer to the blown sand farther west at DM-15 (Figure 8.98) before indicators of agriculture are present. Here there are unequivocal indicators of human activity, particularly in the upper peat, which accumulated largely as a freshwater deposit. Prehistoric farming must though have been taking place close by as there are records of arable weeds, such as *Polygonum persicaria*, *Centaurea cyanus*, cereal pollen and pastoral weeds, such as *Rumex* and *Plantago lanceolata*. The location of this farming is speculative but as the *limus* has been enriched with blown sand from 54 cm upwards it is probable that the landward margin of the better-drained, warmer sandy soils of the sand dunes to seaward had been settled. Middleton *et al.* (2001) speculate that the early farmers who must have occupied the sand dunes that overran the western margin of Downholland Moss some 4090 years ago were the ancestors of the hunters who left their footprints, along with the hoof prints of both aurochs and domesticated ox, in the silts at Formby Point, and pursued red and roe deer (Huddart *et al.* 1999b).

Conclusions

Downholland Moss has been studied for over 150 years and there has been a wealth of information obtained related to its development over the past 8400 years. It has provided important data related to sea-level change in north-west England and there is evidence for three marine phases in the deeper parts of the moss succession. The detail of the associated vegetation changes and detailed diatom and foraminiferal biostratigraphy have provided much local evidence as to the environmental changes related to transgressions and regressions of the sea. These sea-level changes have been dated across the moss and in stratigraphical succession by ^{14}C dating of the contacts. It has been difficult to correlate marine phases across the moss but major changes from the Alt estuary area to the New Cut and farther to the north at DM-15 suggest that the transgressions observed had their source from the proto-Alt estuary to the southwest rather than from the immediate west. This too suggests a barrier to the west throughout most of the Flandrian, with the connection to the sea via the Alt estuary.

The moss too has given detailed information for the first time in north-west England for the stratigraphy and morphology of roddon landforms, which drain to the Downholland Brook and the River Alt. Throughout Flandrian times Downholland Moss appears to have been a wetland habitat, with a mosaic of freshwater, saltmarsh and brackish-water lagoon environments, behind a sand barrier to the west. Occasionally islands of the Shirdley Hills Sand Formation or till provided a drier habitat for woodland. There are isolated glimpses of prehistoric human land use of the moss, or at least its immediate adjacent dune habitats or drier islands within the wetland.

FORMBY POINT (SD 26 06)

POTENTIAL GCR SITE

S. Gonzalez and D. Huddart

Introduction

Formby Point is part of the Sefton coast dunes, which occupy 1285 ha (Doody, 1989), forming a natural coastal defence some 15 km long and 4 km wide at its maximum extent and locally rising to 30 m above sea level. This coastal dune barrier has been influenced by processes both in the Ribble and Mersey estuaries and in the eastern Irish Sea (Smith, 1982; Pye, 1990) and present-day sediment transport and deposition are dominated by strong tidal currents and moderate wave-energy conditions. The result is a predominantly sandy, multi-barred foreshore

(Parker, 1975) backed by coastal dunes that are of major conservation importance in a European context for animals and plants (Atkinson and Houston, 1993).

The Holocene coastal sediments in this locality preserve a record of changing geomorphology since the sea-level rise that started at the end of the last glacial, and they provide much evidence for changing palaeoenvironments along this coastline, in contrast to the extensive areas of peat mossland inland from the dune barrier. The latter area is underlain by Holocene marine and estuarine silts, peats and other deposits of Late Devensian age where the stratigraphy, age, palaeoecology and evidence for sea-level change has been studied intensively (De Rance, 1869b, 1877; Reade, 1871; Gresswell, 1937, 1953; Wray *et al.*, 1948; Tooley, 1974, 1976, 1978a, 1985a, b; Huddart, 1992; Plater *et al.*, 1993 and Plater *et al.*, 2000c). Until recently few investigations had considered the morphostratigraphy and age of the dune barrier and its underlying deposits, but this has been rectified by the work of Pye (1990), Innes and Tooley (1993), Neal (1993), Pye and Neal (1993a, b), Pye *et al.* (1995), Gonzalez *et al.* (1997) and Huddart *et al.* (1999a, b). The nature of the marine transgressions, their recognition and understanding are crucial to the understanding of the prehistoric archaeology (Cowell and Innes, 1994) and it is in some of these foreshore, marine sediments that finds of contemporaneous human, animal and bird footprints (Cowell *et al.*, 1993; Roberts *et al.*, 1996; Gonzalez *et al.*, 1997; Huddart *et al.*, 1999b) and animal bones (Huddart *et al.*, 1999b) have been made. A multidisciplinary approach to understanding the sediments in which this archaeological material has been found, which includes palaeoenvironmental analysis using Foraminifera, ostracods, diatoms, shells, grain size and the dating of the stratigraphical succession has been presented by Gonzalez *et al.* (1997). The exposure of these sediments is the result of the rapid erosion of the beach and dunes at Formby Point, where the sand dunes have an average rate of erosion of up to 3 m per year (Pye and Neal, 1994).

Description

Sediments below the dunes

A drilling programme to investigate the morphostratigraphy and age structure of the dune belt and its underlying deposits to the west of Downholland Moss was undertaken by Neal (1993) and Pye and Neal (1993a, b, 1994). Shell and auger holes were drilled to a maximum depth of 14.6 m along three transects between Ainsdale and the south side of Formby (Figure 8.107) and samples were collected for sedimentological and micropalaeontological analysis and for ^{14}C and optically stimulated luminescence dating. The Holocene deposits are underlain by Triassic bedrock, glacial till and Shirdley Hill Sand and the stratigraphy recorded is illustrated in Figure 8.108. This stratigraphy provides no evidence that the sequences of brackish water and estuarine silts and peats beneath Downholland Moss and in the banks of the River Alt (Tooley, 1974, 1978a, b, 1980, 1985; Huddart, 1992) passed beneath the central and western part of the dune belt (Neal, 1993; Pye and Neal, 1993a, b, 1994) but instead the area was underlain by shelly marine sands that locally are muddy. A ^{14}C date of 5960 years BP was obtained from *Scrobicularia plana* shells from within these sands at –5.5 m OD (Figure 8.108). The oldest date from within the dune system came from borehole F (Figure 8.108) where an organic unit, interpreted as an intra-dune slack peat, gave a date of 5100 ± 70 years BP. Outcrops of muddier units within the marine sands beneath the western dune margin can be seen along the eroded coast but these generally wedge out landwards. They are important because they contain contemporaneous imprints of human, animal and bird footprints.

The foraminiferal analyses from the boreholes shows the following general characteristics:

1 a generally high faunal diversity (between 38 and 66 species), except in borehole J where it is medium (26–33) and faunal dominance ranges from 13 to 29;
2. the *Elphidium* group is dominant ranging from 46–69%;
3. *Haynesina germanica* (equivalent to *Protelphidium germanicum* (cf. Downholland Moss) is ubiquitous from 3.2 to 32%;
4. there are a number of *Ammonia* species, with *Nonion pauperatum* and *Trifarina angulosa* in most samples;
5. small, often current winnowed species, such as *Buccella frigida*, *Lagena*, *Oolina*, *Bolivina* and *Fissurina* species generally are present;
6. *Planorbulina distoma* and *Cibicides lobatulus* often are present;

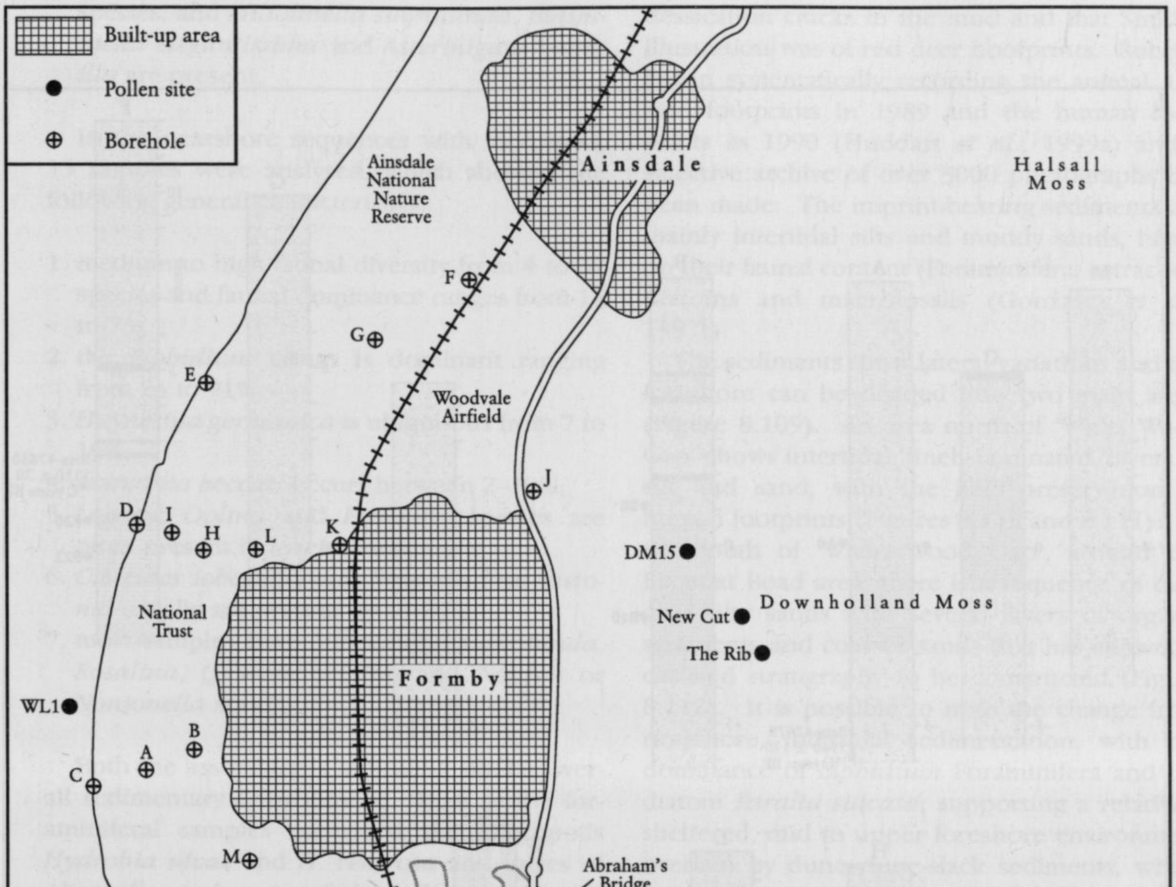


Figure 8.107 Location of boreholes, Formby Point.

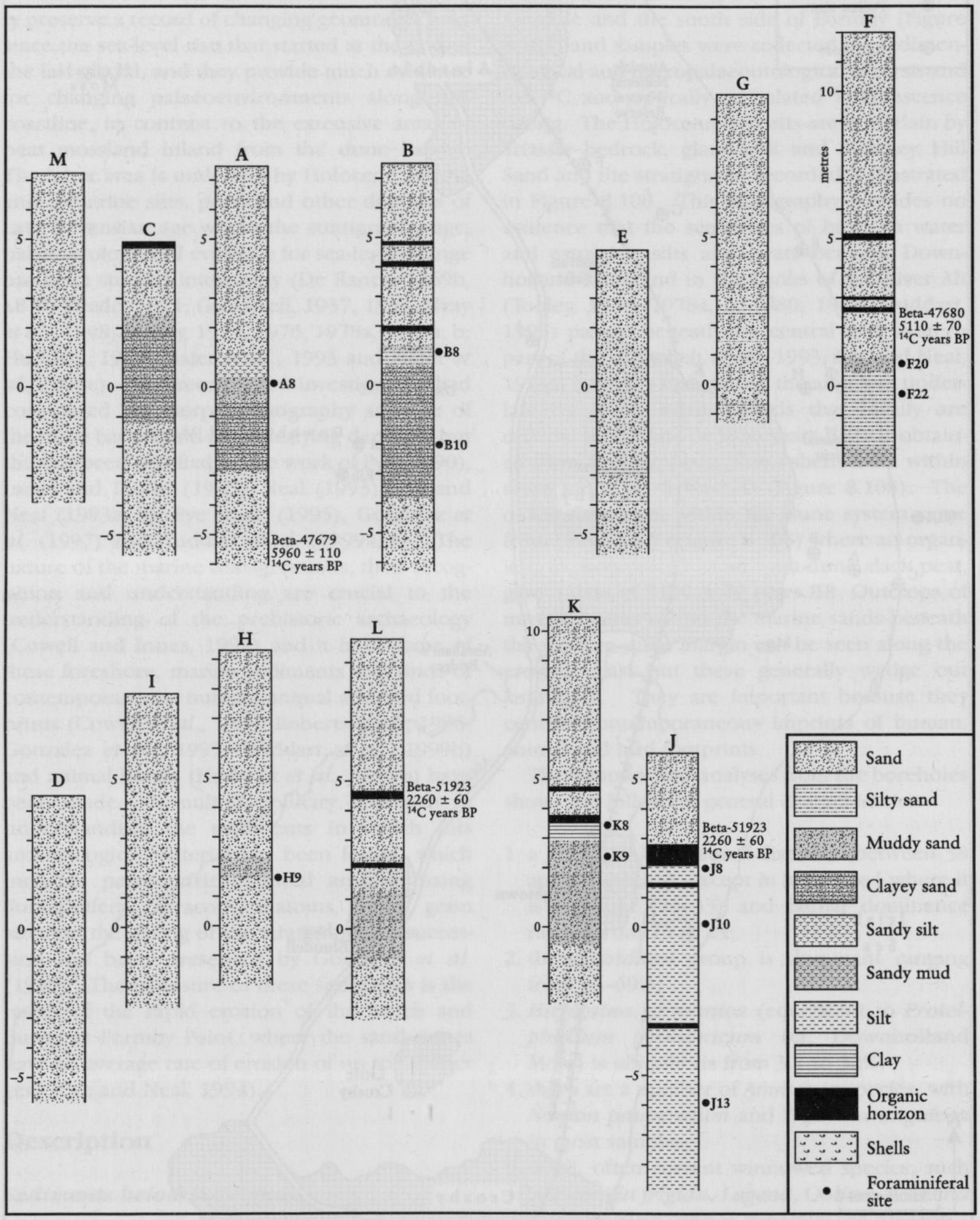


Figure 8.108 Stratigraphy in boreholes, Formby Point (after Neal, 1993).

7. there are occasional forms of *Quinqueloculina*, *Virgulina*, *Bulimina* and *Nonionella* species, and *Miliolinella subrotunda*, *Buliminella elegantissima* and *Asterinigata mamilla* are present.

In the nearshore sequences with footprints, 15 samples were analysed, which showed the following general characteristics:

1. medium to high faunal diversity from 4 to 32 species and faunal dominance ranges from 14 to 73;
2. the *Elphidium* group is dominant ranging from 23 to 91%;
3. *Haynesina germanica* is ubiquitous from 7 to 19%;
4. *Ammonia beccari* occurs between 2–19%;
5. *Lagena*, *Oolina* and *Fissurina* species are often present in low quantities;
6. *Cibicides lobatulus* and *Planorbulina distoma* usually are present up to 10%;
7. most samples have *Miliolinella subrotunda*, *Rosalina*, *Quinqueloculina* and *Nonion* or *Nonionella* species.

Both the assemblages indicate a similar overall sedimentary environment. Most of the foraminiferal samples contained the gastropods *Hydrobia ulvae* and *H. ventrosa* and valves of *Abra alba* and many of the sediment samples contained shell debris, including *Macoma balthica*, *Cerastoderma edule*, *Ostrea edulis*, *Scrobicularia plana*, *Tellina tenuis* and *Littorina littoralis* (Gonzalez *et al.*, 1997).

The footprints and their stratigraphical position

The human, animal and bird footprints have been described by Cowell *et al.* (1993), Roberts *et al.* (1996), Gonzalez *et al.* (1997) and Huddart *et al.* (1999a, b) and appear discontinuously in a 4 km belt within the Holocene sediments (centred on SD 26 06), within a 100 m-wide belt in the intertidal zone, usually about 60 m west of the dune edge. It probably was Tooley (1970) who first observed (in 1968) and then photographed (in 1974), a number of ungulate hoofprints, seemingly associated with small stakes. The first published photograph of 'cattle hoofprints from 4000 year old peat' was taken by P.H. Smith (in Hale, 1985). It was suggested by Gonzalez *et al.* (1997) that Tooley's photograph may have been an aurochs hoofprint, with the stakes being alder roots growing down into dessication cracks in the mud and that Smith's illustration was of red deer hoofprints. Roberts began systematically recording the animal and bird footprints in 1989 and the human footprints in 1990 (Huddart *et al.*, 1999a) and a selective archive of over 3000 photographs has been made. The imprint-bearing sediments are mainly intertidal silts and muddy sands, based on their faunal content (Foraminifera, ostracods, diatoms and macrofossils (Gonzalez *et al.*, 1997)).

The sediments show lateral variations and the foreshore can be divided into two main areas (Figure 8.109). An area north of 'Wicks Wood Gap' shows intertidal, finely laminated, layers of silt and sand, with the best preservation of human footprints (Figures 8.110 and 8.111). To the south of 'Wicks Wood Gap', around the Lifeboat Road area, there is a sequence of dark grey silty sands with several layers of organic sediments and coarser sands that has allowed a detailed stratigraphy to be constructed (Figure 8.112). It is possible to note the change from nearshore, intertidal sedimentation, with the dominance of *Elphidium* Foraminifera and the diatom *Paralia sulcata*, supporting a relatively sheltered, mid to upper foreshore environment overlain by dune–dune-slack sediments, which have only a few reworked or no Foraminifera (Gonzalez *et al.*, 1997). The 'Wicks Wood Gap' where no imprint-bearing sediments have been observed over the past eight years, was investigated using a JCB to excavate 27 trenches up to 2 m deep, but only orange sand and no silts were found. The human and animal footprints have been found in four main areas, forming different assemblages. Sector 1 (Figure 8.109) shows human, red and roe deer, crane and unshod horses; sector 2 shows red and roe deer; sector 3 shows human, red and roe deer, crane and aurochs; sector 4 shows red and roe deer, aurochs and domestic ox. Almost all the human footprints have been found within sectors 1–3 and over 145 trails have been recorded (Roberts *et al.*, 1996). It has been possible to calculate a mean adult male height of 1.66 m. and mean female height of 1.45 m. At Formby Point though the footprints of children are predominant and where adult males are present, they are often associated with red deer tracks and they indicate an above-average speed and cadence (Roberts *et al.*, 1996). The much slower movement of women and children would suggest a different economic activity. In some of the human footprints it is possible to see abnormalities, such as missing toes, the fusion of two toes, congenital bunions and arthritis and the footprints of possibly pregnant women have been observed. In sector 4 there are at least two different footprint sets where the younger set of prints, consisting of cattle, red and roe deer, are preserved in an organic sand, with alder roots and desiccation cracks within it. The alder roots have been dated to 3230 years BP (Pye and Neal, 1993a, b), 3333 years BP and 3649 years BP (Gonzalez *et al*., 1997). The older set of prints is preserved in black, silty sands, interbedded with thin silt layers. Here large aurochs hoofprints are associated with red and roe deer tracks.

At Formby Point the following animal bones have been recovered over the past six years: aurochs and red deer jawbones, both showing cut marks, a complete set of red deer antlers (Figure 8.110) and a dog jawbone. Most of the finds are concentrated at the Lifeboat Road entrance to the beach and are associated with layers of blue silt and organic-rich black sands formed in a dune-slack environment (Gonzalez *et al*., 1997). Although the stratigraphical position of the red deer antlers is not precisely located, it came from intertidal silts and has been

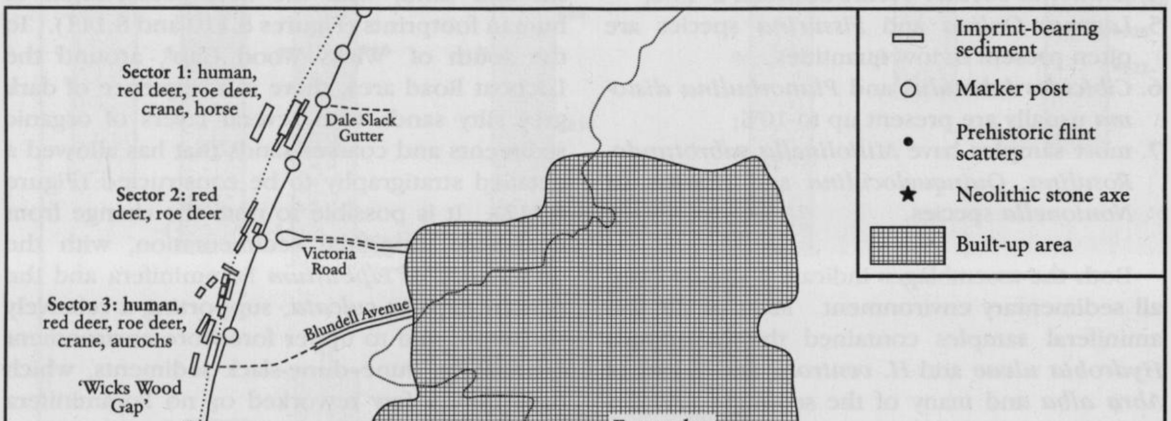


Figure 8.109 Location of imprint-bearing sediments and archaeological finds, Formby Point.

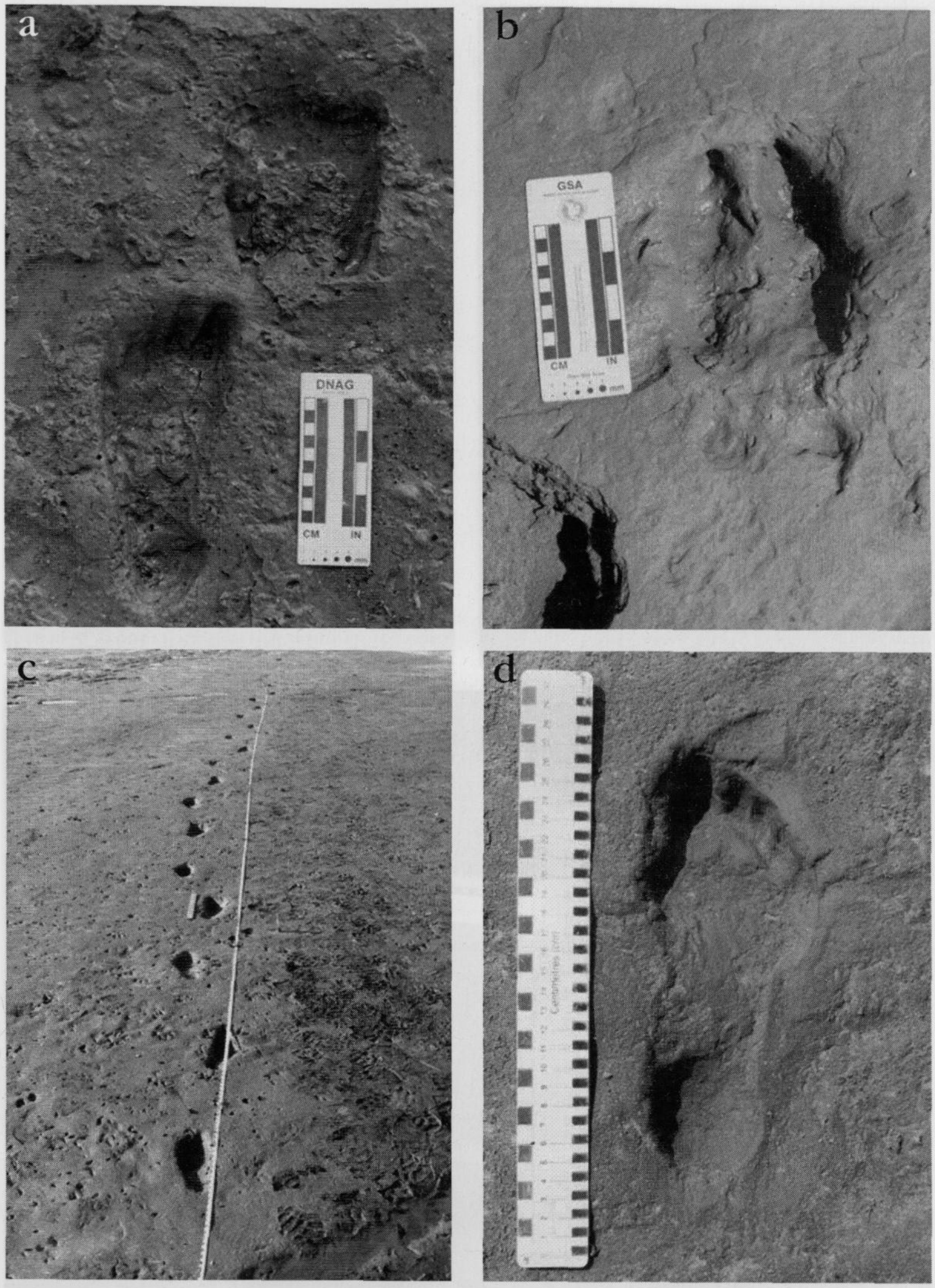
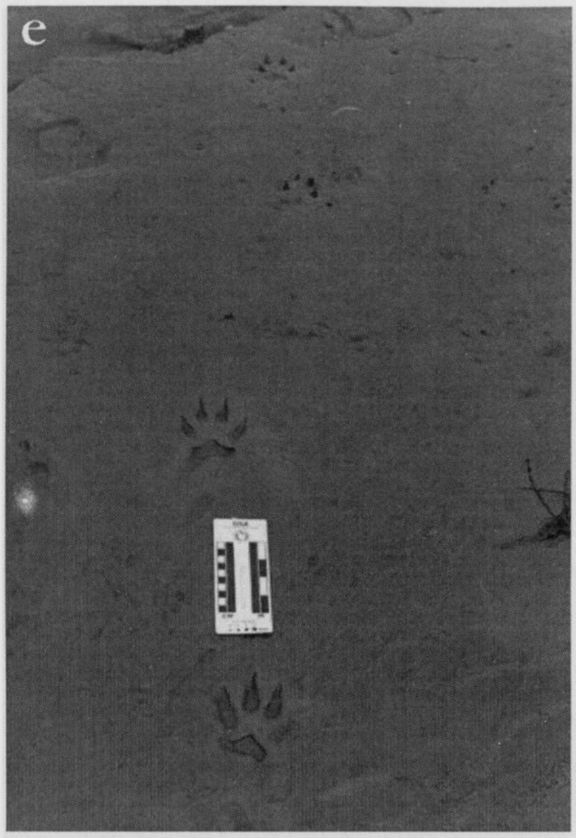


Figure 8.110 Animal and human footprints, Formby Point. (a) Human footprint. (Photo: S. Gonzalez). (b) Red deer hoofprint. (Photo: S. Gonzalez). (c) Sector 1, human footprint trail (from Roberts *et al.*, 1996). (Photo: G. Roberts). (d) Detail from human footprint trail (from Roberts *et al.*, 1996). (Photo: G. Roberts).



dated to 4425 years BP (S. Stallibrass, pers. comm., 2001). The dog jawbone was found in a horizon dated to 3649 years BP (Huddart *et al.*, 1999a, b).

Interpretation

Regional Significance of the Footprints

Over the past 9000 years the coastal areas of south-west Lancashire experienced a total rise in sea level of 20 m (Tooley, 1978a) and it is under these rapid, but oscillating, changes in the coastal environment in which there has been the preservation of human and animal footprints in the intertidal zone at Formby Point. There were three major transgressions that reached as far inland as Downholland Moss. A fourth transgressive event occurred that deposited silts in small embayments at Formby and the Altmouth and a final transgression is inferred from the evidence at Lifeboat Road where an organic sand with alder roots overlies a thin layer of blue silt. Detailed stratigraphical and sedimentological analyses in relation to the dune system (Pye, 1990; Neal, 1993; Pye and Neal, 1993a, b, 1994;

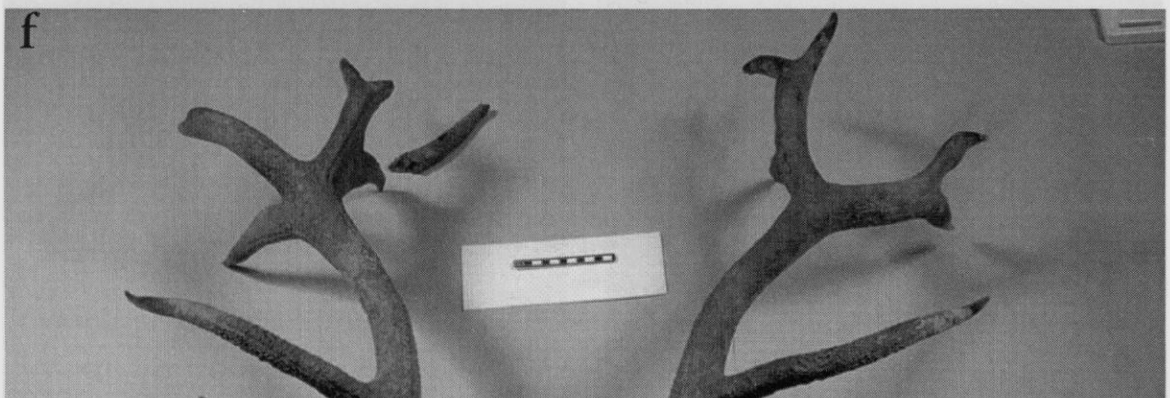


Figure 8.110 (*contd*) (e) Wolf trail. (Photo: S. Gonzalez). (f) Complete pair of red deer antlers, dated at 4425 BP (Stallibrass, pers. comm.). (Photo: Conservation unit of Liverpool Museum.)

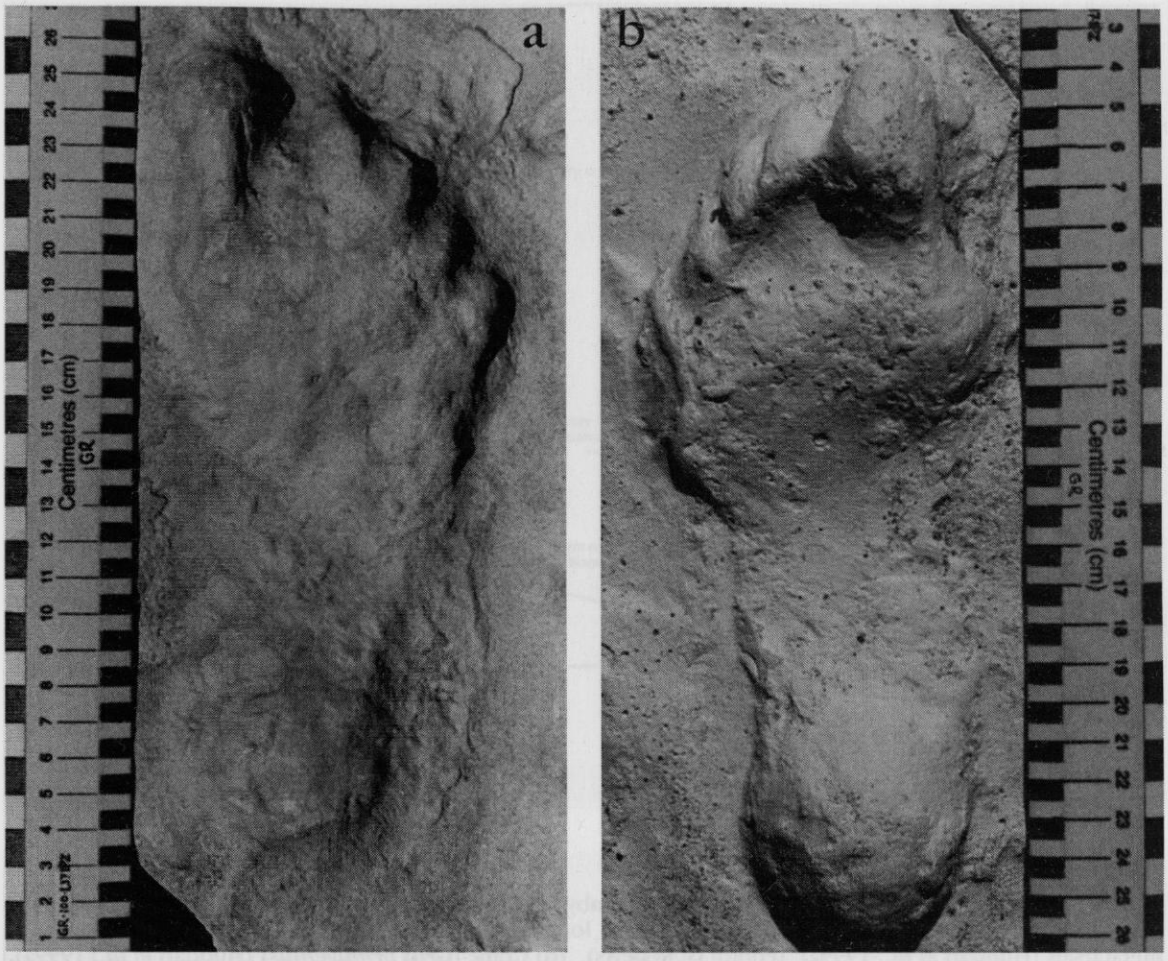


Figure 8.111 Plaster casts of footprints, Formby Point. (a) Habitually unshod healthy foot, stature 1.4 m, showing big toe abduction (Huddart *et al.*, 1999a; photo: G. Roberts). (b) Pregnant female, with congenital bursitis, stature 1.37 m, speed 2.8 km per hour, cadence 89 footfalls per minute; weight on heel to compensate for change in centre of gravity, foot arched and toes curled downwards to obtain grip in the then soft, slippery mud (Huddart *et al.*, 1999a). (Photo: G. Roberts). See overleaf for (c) and (d).

Pye *et al.*, 1995) show that the present central and western dune system, which formed after *c.* 2500 years BP, rests on marine sands to a depth of –7.5 m OD, which were present before 5910 years BP.

The human prints are at least 3649 ^{14}C years old around Formby Point and they indicate that during the Neolithic–Bronze Age there was human activity and although there is no proven age for the lower stratigraphical animal footprint set, it is conceivable that they could be as old as the Later Mesolithic (*c.* 8800–6000 years BP). There is evidence of settlement along the coastal mosslands south of the Alt estuary at this period, where the reduction of flint pebbles and the manufacture of implements took place, although there is an absence of microliths. Although evidence for residential sites is missing, it is possible that coastal–estuarine locations could provide a range of resources to sustain larger groups for longer time periods, as has been suggested by Bonsall (1981) for the Mesolithic sites on the Cumbrian coast. On the Sefton coastal lowland there were large areas of swamp and subsequent fen, in hollows within the deciduous woodland. Easy movement would be difficult across this terrain and therefore a more open coastal foreshore would provide easier access for both animals and humans, in an area rich in food sources (plants, shellfish, fish, wildfowl and animals). The woodland growing on the drier Shirdley Hill Sands soils

Figure 8.111 *(contd)* Plaster casts of footprints, Formby Point. (c) Deformed male foot, stature 1.4 m; loss of toe; compensatory over-development of peroneus longus and long plantar ligament, with collapse of metatarsals (Huddart *et al.*, 1999a). (Photo: G. Roberts). (d) Crane (*Grus primigenius*) (Huddart *et al.*, 1999a). (Photo: G. Roberts).

that surrounded the wetlands would have provided habitats for ungulates, such as aurochs, wild pig and deer, and edible plants. Cowell and Innes (1994) suggest that the sandier soils had less dense woodland and again could have provided easier passage through the woodlands than clay soil areas, and would have been easier to clear if necessary. An even clearer route would have been along the coast and the footprints are some evidence for this. It is possible then that the lower prints are Mesolithic in age, but it seems just as likely that all the prints are Neolithic and the younger set certainly are. Archaeological evidence (Cowell and Innes, 1994) shows Mesolithic–Neolithic sites in the Alt valley, and Neolithic axes have been found at Hightown and west of Little Crosby. In the Early Bronze Age an economic pattern based on mobility and cattle grazing has been suggested (Fleming, 1971), although it is probable that hunting and gathering still played an important role in this area. Evidence from Woodham Knoll, Little Crosby (Cowell and Innes, 1994) suggested that the fens were being used for summer pasture for cattle and that the associated settlement was on the sandy soils above the wetland. During the winter months the fens and the nearby estuary could have been used for wildfowling and fishing, whereas the cattle were moved to higher land. The possibility therefore is that the upper footprints are Late Neolithic in age and that the cattle and human prints were caused by summer grazing on the dunes, utilizing the dune slacks for freshwater. Overall, Huddart *et al.* (1999b) concluded that the Sefton coast was used both as an easy access route for humans and animals, because farther inland it was either extensively wooded or

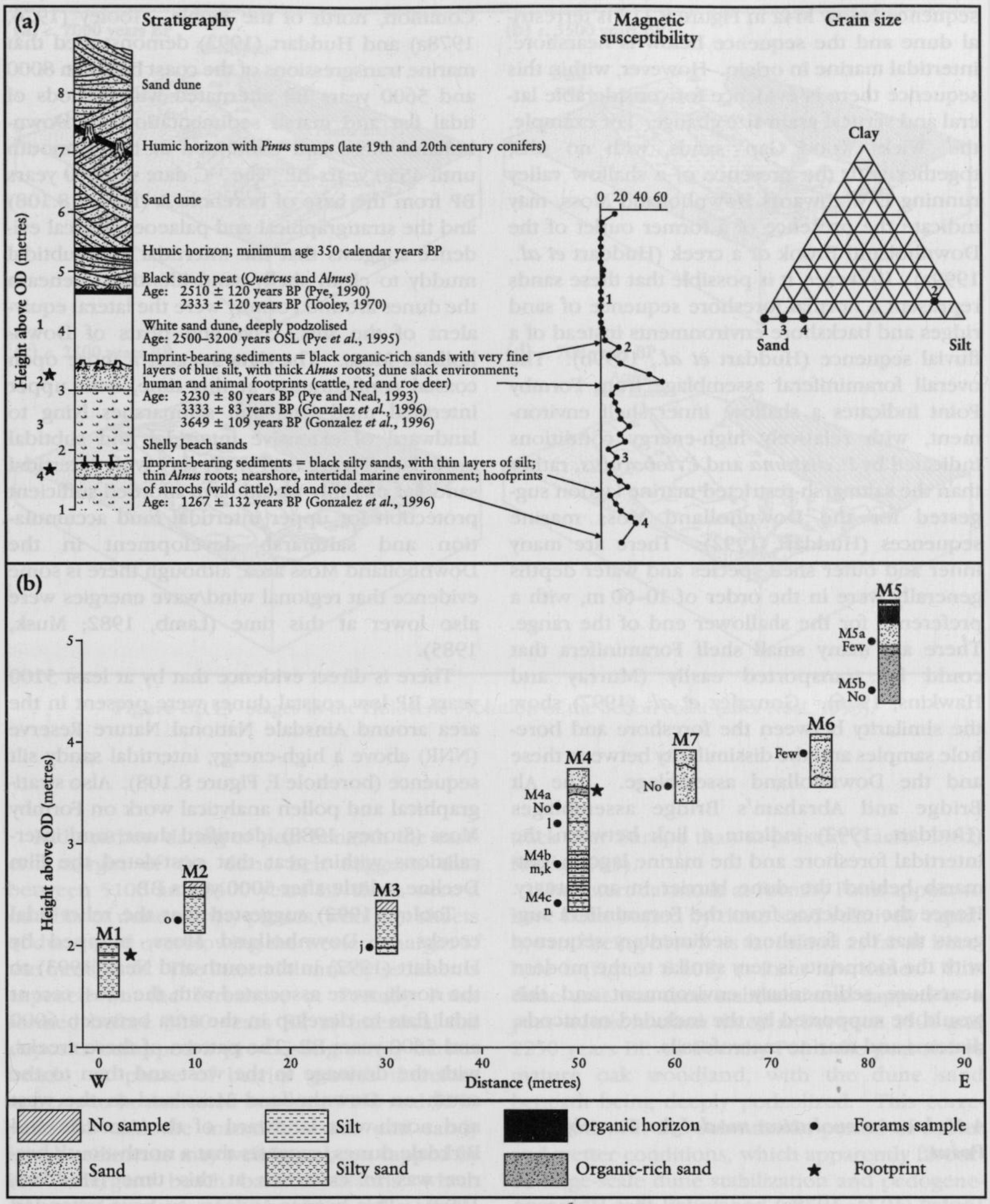


Figure 8.112 (a) Stratigraphy at Formby Point. (b) Stratigraphy at Lifeboat Road area, Formby Point.

extensively waterlogged, and also that it was used for a combination of hunting and gathering and cattle grazing at various prehistoric time periods between the Late Mesolithic and the Early Bronze Age.

Sedimentary environments associated with the footprints

The sedimentological and fossil evidence (Gonzalez *et al.*, 1997) shows that the foreshore sequence below M4a in Figure 8.112 is terrestrial dune and the sequence below is nearshore, intertidal marine in origin. However, within this sequence there is evidence for considerable lateral and vertical grain-size change. For example, the 'Wicks Wood Gap' sands, with no silts, together with the presence of a shallow valley running E–W towards Downholland Moss, may indicate the presence of a former outlet of the Downholland Brook or a creek (Huddart *et al.*, 1999a). However, it is possible that these sands represent a normal foreshore sequence of sand ridges and backshore environments instead of a fluvial sequence (Huddart *et al.*, 1999b). The overall foraminiferal assemblage from Formby Point indicates a shallow, inner shelf environment, with relatively high-energy conditions indicated by *P. distoma* and *C. lobatulus*, rather than the saltmarsh-restricted marine lagoon suggested for the Downholland Moss marine sequences (Huddart, 1992). There are many inner and outer shelf species and water depths generally were in the order of 10–60 m, with a preference for the shallower end of the range. There are many small shelf Foraminifera that could be transported easily (Murray and Hawkins, 1976). Gonzalez *et al.* (1997) show the similarity between the foreshore and borehole samples and the dissimilarity between these and the Downholland assemblage. The Alt Bridge and Abraham's Bridge assemblages (Huddart, 1992) indicate a link between the intertidal foreshore and the marine lagoon–saltmarsh behind the dune barrier in an estuary. Hence the evidence from the Foraminifera suggests that the foreshore sedimentary sequence with the footprints is very similar to the modern nearshore sedimentary environment and this would be supported by the included ostracods, diatoms and marine macrofossils.

A Holocene evolution model for Formby Point

Tooley (1974, 1978a) concluded that Holocene sedimentation started on the Sefton coast around 8000 years BP, when a peat at the base of a borehole at Long Lane, Formby accumulated. A similar peat is found at the base of the Holocene sequences east of Southport (Neal, 1993), but the age should be treated with some caution because it is based on altitudinal correlation with a similar, dated peat beneath Lytham Common, north of the Ribble. Tooley (1976, 1978a) and Huddart (1992) demonstrated that marine transgressions of the coast between 8000 and 5600 years BP, alternated with periods of tidal flat and marsh sedimentation on Downholland Moss and continued at the Altmouth until 4550 years BP. The ^{14}C date of 6000 years BP from the base of borehole A (Figure 8.108) and the stratigraphical and palaeoecological evidence suggests that the intertidal and subtidal muddy to clean shelly sand identified beneath the dunes around Formby were the lateral equivalent of the silts, clays and peats of Downholland Moss. The data suggest a more open coastal configuration at this time, with upper intertidal mud flats and saltmarshes lying to landward of extensive intertidal and subtidal sand flats (Figure 8.113). A very wide intertidal sand flat possibly would have afforded sufficient protection for upper intertidal mud accumulation and saltmarsh development in the Downholland Moss area, although there is some evidence that regional wind/wave energies were also lower at this time (Lamb, 1982; Musk, 1985).

There is direct evidence that by at least 5100 years BP low coastal dunes were present in the area around Ainsdale National Nature Reserve (NNR) above a high-energy, intertidal sandy silt sequence (borehole F, Figure 8.108). Also stratigraphical and pollen analytical work on Formby Moss (Stoney, 1988) identified dune sand intercalations within peat that post-dated the Elm Decline, a little after 5000 years BP.

Tooley (1992) suggested that the relict tidal creeks on Downholland Moss, mapped by Huddart (1992) in the south and Neal (1993) to the north, were associated with the most recent tidal flats to develop in the area between 6000 and 5600 years BP. The pattern of these creeks, with the drainage to the west and then to the south on Downholland Moss and to the west and north-west landward of the Ainsdale and Birkdale dunes, suggests that a north–south barrier was in existence at this time (Huddart, 1992). This barrier appears to have extended from just north of the mouth of the River Alt to at least as far as the present-day Ainsdale NNR. The presence of this barrier, combined with the relatively static sea levels and increased wetness of the Atlantic period, may have assisted the widespread replacement of saltmarsh by freshwater marsh behind the barrier after 5100 years BP.

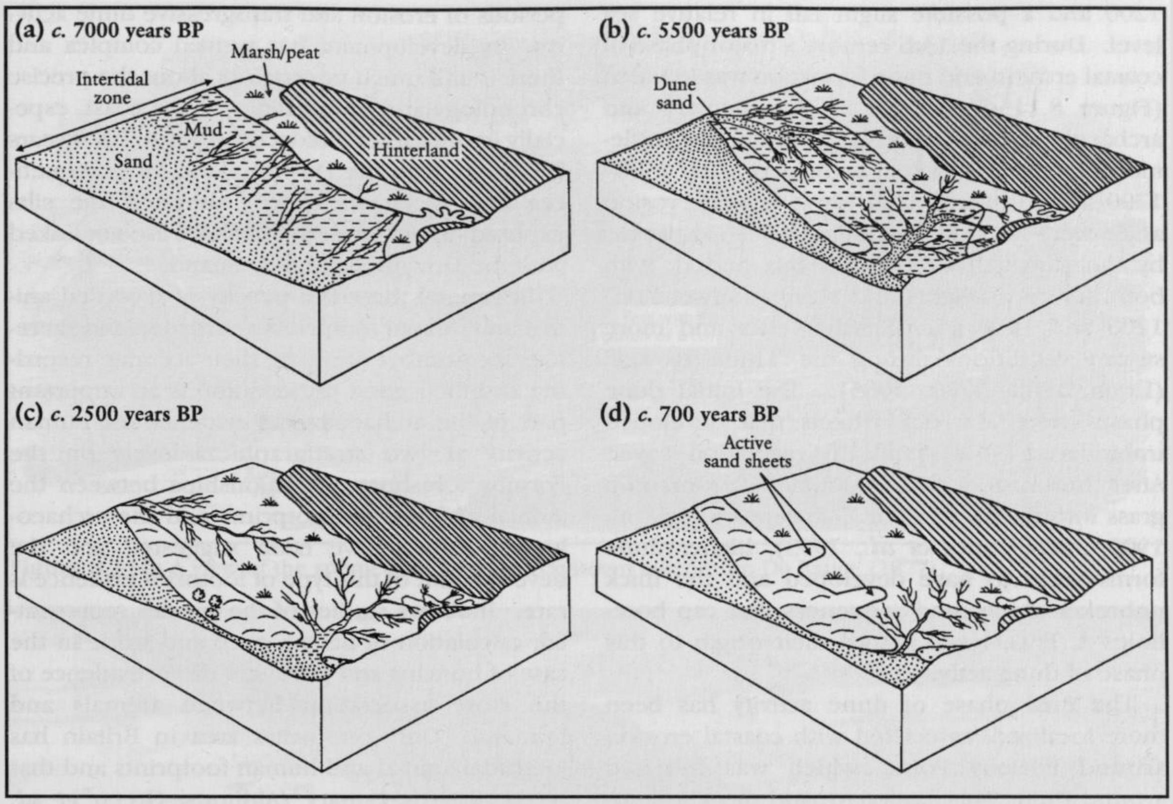


Figure 8.113 Stages in the Holocene evolution of the Sefton coast (after Neal, 1993).

Radiocarbon dating of peat beneath the eastern margin of the dune belt suggests that between 5100 and 3400 years BP there were a number of dune activity phases, with thin sheets of dune sand overblowing the western margin of the mosslands. The intertidal muddy sediments exposed on the foreshore at Formby Point formed before 3250 years BP, as intertidal flat deposits onlapped a pre-existing sandy barrier beach. A phase of barrier growth therefore appears to be indicated between 5100 and 3250 years BP and the intertidal muds and sandy beach sediments may well have been capped by low emergent beach bars or islands, which allowed deposition of the muddier sediments in their lee and also supplied sand-grade sediment for subsequent dune formation. Deposition of the intertidal mud units also may have been favoured by a possible regional reduction in wave energy during the Sub-Boreal period, as atmospheric circulation over Britain had reached its weakest post-glacial phase, with westerlies continuing to take a more northerly track over Europe than at present (Lamb, 1982; Musk, 1985).

The intertidal muds at Formby Point appear to have been buried by dune sand relatively rapidly, sand being blown as far inland as far as borehole L (Figure 8.107) to form low dunes. This dune unit was then stabilized and capped by a peat–humic horizon dated at between 2500 and 2250 years BP. At Formby Point the peat shows mature oak woodland, with the dune sand beneath being deeply podzolized. This corresponded with the Sub-Atlantic period of cooler and wetter conditions, which apparently favoured large-scale dune stabilization and pedogenesis on Formby Point and elsewhere in Britain and Europe (Tooley, 1990). Figure 8.113c indicates a schematic palaeoenvironmental reconstruction for *c.* 2300 years BP.

Documentary and archaeological evidence suggests that much of the sand belt remained largely stable until the early Middle Ages (Jones *et al.*, 1993), probably as a result of relatively storm free conditions in the period AD 400–

1200 and a possible slight fall in relative sea level. During the 13th century a major phase of coastal erosion and dune formation was initiated (Figure 8.113d). There is documentary and archaeological evidence for a number of settlements lost or abandoned in the period AD 1300–1700 as a result of sand blow, coastal erosion and severe flooding (Jones *et al.*, 1993) caused by the climatic instability of this period, with both severe droughts and storms between AD 1200 and 1400 and slightly wetter and more stormy conditions during the 'Little Ice Age' (Lamb, 1982; Musk, 1985). The initial dune phases were low sand sheets that developed unhindered by a significant vegetation cover. After the large-scale introduction of marram grass into the area in the 17th century (Ashton, 1909, 1920; Jones *et al.*, 1993), higher dune forms seem to have developed and the thick unbroken dune sand sequences that cap boreholes A, F, G, I and L owe their origin to this phase of dune activity.

The final phase of dune activity has been more localized, associated with coastal erosion around Formby Point, which was initiated around 1900. This has led to sand sheet and retrogradational foredune ridge development immediately landward of the erosional shoreline. The causes and mechanisms of this erosion have been examined by Pye and Neal (1994), who concluded that the onset of this erosion was mainly the result of an increase in the occurrence of storm-force westerlies and destructive waves around the turn of the century. Erosion subsequently was augmented by bathymetric change and increased wave focusing on to Formby Point, a decrease in the return period of extreme tidal levels, a continued upward movement of relative sea level and the abandonment of dune and foreshore management practices between 1900 and the late 1970s.

Conclusions

Formby Point has provided a chronology and model for the development of the important dune system on the Sefton coast during the Holocene since the rise of sea level after the last glacial. It has been dominated by both regional and local changes in climate, sea level, sediment supply and human activity. The barrier is of the static type as defined by Thom (1974, 1984) and Thom *et al.* (1978), but there have been a number of progradational phases punctuated by periods of erosion and transgressive dune activity. Its development has proved complex and there is still much uncertainty about the precise chronology and morphology of the coast, especially in the earlier Holocene, prior to 7000 years BP. Links with the Downholland Moss sequences have been suggested, although the silts exposed on the modern foreshore are not linked with the Downholland Silts inland.

In general there is a paucity of recorded animal and human footprints worldwide and therefore the number of prints, their accurate recording and their good preservation is an important part of the archaeological evidence for human activity at two stratigraphical levels on the Formby foreshore. Relationships between the animal and human footprints and the archaeological record have been suggested and the development of this type of footprint evidence is rare. It allows studies of the species represented, calculation of height, pace and stride in the case of humans and produces direct evidence of the close association between animals and humans. Only one other area in Britain has intertidal animal and human footprints and that is the Severn estuary (Aldhouse-Green *et al*, 1992; Bell, 1995). The intertidal exposures provide a rich and varied source of palaeoenvironmental and archaeological data but the prehistoric landscapes exposed are ephemeral and are immediately eroded and will continue to do so during rising sea levels.

HIGHTOWN (SD 2950 0290)

POTENTIAL GCR SITE

S. Gonzalez and D. Huddart

Introduction

Hightown, Sefton is famous for an important submerged forest of Flandrian age (Figure 8.114), which currently is being eroded extensively, along with the overlying dunes. Most of the early descriptions of submerged forests are of little scientific value, for example, Camden's description of the Mersey Flats in his Magna Britannia. However, there was correspondence in the July 1786 issue of the *Gentleman's Magazine* where a reader (Holt) reported a submerged forest at Crosby that extended upward of a mile towards Formby and was the earliest account that tried to interpret and understand

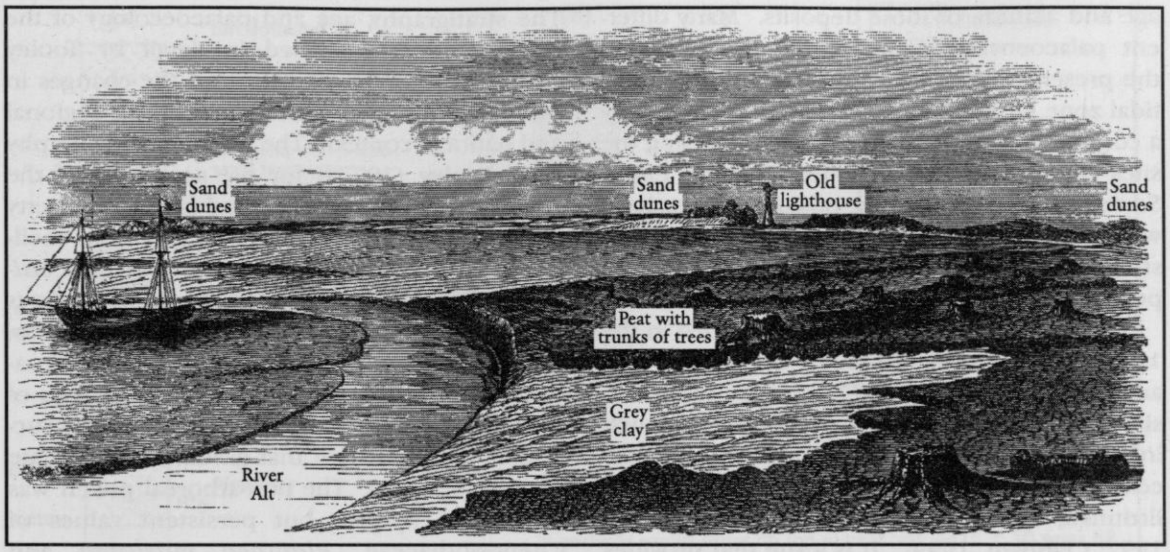


Figure 8.114a A view of the submerged forest at Hightown taken from De Rance (1877).

Figure 8.114b Submerged forest at Hightown. Note surface log and root system through the peat. (Photo: S. Gonzalez.)

the depositional history of the submerged forests. The distribution and characteristics of such submerged forests in England and Wales were described by Reid (1913) and in the British Isles by Wright (1914) and Heyworth (1978, 1986). In the intertidal zone of the eastern Irish Sea there are many records of such submerged forests, reviewed by Huddart and Tooley (1972), Huddart *et al.* (1977) and Tooley (1978a), but there is no consistent relationship between the age and altitude of these deposits. Many different palaeoenvironments are represented and the presence of submerged forests in the intertidal zone, or exposed in cliff sections, is simply a consequence of recent processes exposing the sites that have been recorded (Tooley, 1985). Some are related to former positions of sea level whereas others are not, and only a study of the stratigraphy and fossil content can indicate the proximity of the sea.

The Hightown deposits were described in the 19th century by Reade (1878a, b, 1883a, 1908) and De Rance (1869b, 1877), and the macrofossils and pollen from the peat provide much information about the development of a Holocene forest in a coastal location (Travis, 1926; Erdtman, 1928; Tooley, 1977a; Clapham *et al*, 1997; Clapham, 1999). It is a site that provides one of the regressive overlap points for the sea-level curve for north-west England (Tooley, 1977a, 1978a, 1982) and it has produced several chance archaeological finds that have not been *in situ*. However, in September 1996 the first example of a prehistoric trackway on the south-west Lancashire coast was noted and subsequently excavated in detail.

Description

Intertidal Stratigraphy at Hightown

De Rance (1869b) described the stratigraphy and showed that an upper peat with woody remains overlaid a laminated grey clay that contained the shell *Cyclas cornea* (synonymous with *Sphaerium corneum*), after which the clay was named. Reade (1871) also described the stratigraphy at Altmouth and assigned the laminated clays to the Formby and Leasowe Marine beds on the basis of the included shells, diatoms and Foraminifera. The submerged forest was part of his Superior Peat and Forest Bed. Travis (1926) and Erdtman (1928) were probably the first to examine the pollen content of submerged forests and although Travis provides a list of pollen and spores from the intertidal peat beds at Hightown, Erdtman published the first pollen diagram for such a site and concluded that the peat bed could be assigned to the Atlantic–Boreal period (*c*. 5000 years BP). Travis (1926) described the peat between 20 cm and 1.2 m thick, averaging between 45 and 53 cm, underlying a dark peaty sand, 45 cm to 1.5 m thick, which formed the base of the sand dunes. The stratigraphy, age and palaeoecology of the site has been described in detail by Tooley (1970, 1974, 1978a, 1982) and the changes in palaeoenvironment placed in a local, regional and national context. The intertidal stratigraphy from Tooley (1977a) for the area west of the Blundellsands Sailing Club and north of the jetty described, at the base, at least 1.25 m of shell-rich, grey sand, with fragments *of Macoma balthica*, *Barnea candida*, *Chlamys opercularis* and *Cerastoderma edule* and plant remains, including *Zostera* species. Above were alternations of quiet brackish-water clays with fruits of tasselweed (*Ruppia*) and sands. This was succeeded by a peat unit that gave a basal ^{14}C date of 4545 years BP. The non-arboreal pollen was dominated by low but persistent values of Chenopodiaceae, *Plantago maritima* and *Armeria maritima*. The fen peat in the lower part showed a tree pollen dominated by alder and oak and this passed up into a woody, detrital peat with branches and trunks of alder, oak, birch and much royal fern (*Osmunda regalis*, see Figure 8.116). A more recent stratigraphy is given in Figure 8.115a, where below the peat was located 1.45 m of silt and fine sand that generally fines upward. In samples H7 and H9 there are characteristic saltmarsh Foraminifera, whereas in samples H10, H15 and H16 there are foraminiferal indicators of nearshore environments. Above 21.5 cm of peat at the back of the beach there is a transitional change into 13.5 cm of organic, black, sandy mud with wood fragments and 48 cm of organic sands with woods fragments and tree stumps in some layers. This is overlain by 15 cm grey, leached sands and 3 m of brown and orange sands, with humic horizons. The stratigraphy associated with the trackway approximately 200 m to the south of Figure 8.115a is illustrated in Figure 8.115b. At the base is 18 cm of blue clayey silt overlain by a 20 cm complex interdigitation of wood associated with the trackway (Figures 8.116a and 8.116b) and silts, which is overlain by organic-rich, dark brown, clayey silt, in which are located red deer hoofprints (Figure 8.116c). The ^{14}C dates associated with the Hightown succession are given in Table 8.19.

Palaeovegetation at the Hightown site

Travis (1926), Clapham *et al*. (1997) and Clapham (1999) described in detail the plant macrofossil assemblages from the submerged

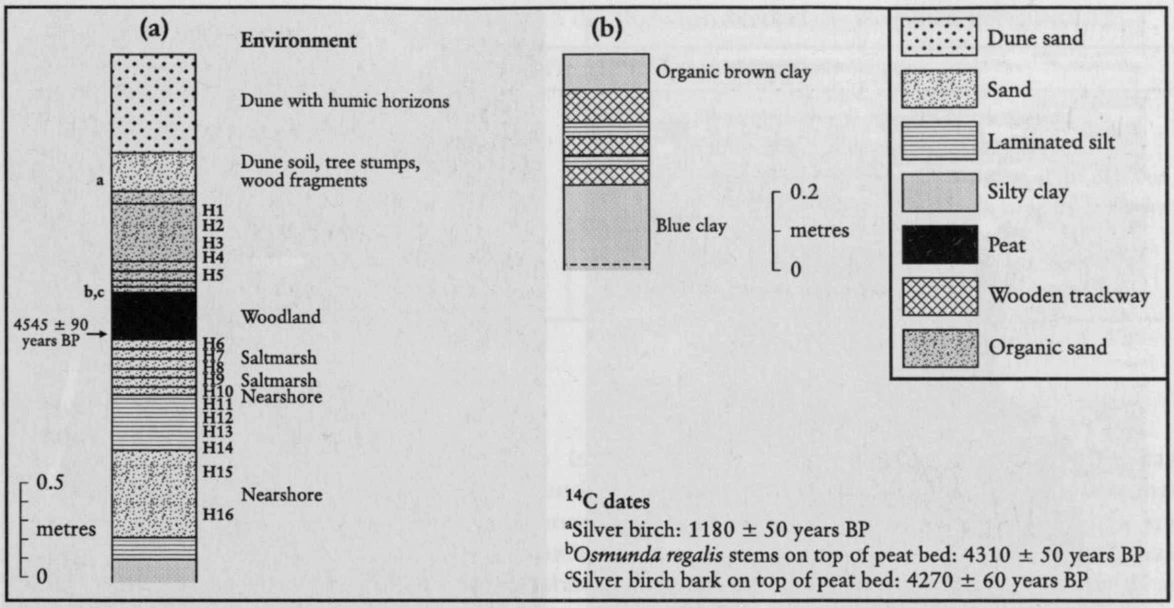


Figure 8.115 Stratigraphy at Hightown: (a) based on recent work by the authors; (b) associated with the prehistoric trackway.

forest and these assemblages represent spatial variation within a fen-carr woodland. The upright stools of the larger trees varied in size from 30 cm to 91 cm and in one case 1.8 m in diameter (Travis, 1926). The prostrate trunks were between 1.8 m and 3.6 m in length (Travis, 1926). The tree and shrub species with the type of fossil remains identified by Travis (1926) are shown in Table 8.20. Pollen analysis from Tooley (1970) in Figure 8.117 shows that the peat is dominated by arboreal pollen, with a maximum of 60% at 27 cm depth. As with Travis' (1926) analyses, Tooley's pollen spectra do not show much variation in the profile. The proportion of hazel appears to be constant throughout the profile within a smaller shrub component.

Interpretation

The first description by Holt (1786) explained the submerged forest and the exposed sediments as an encroachment of the sea on the land at the mouth of the Mersey estuary. Hume (1865), concluded that the submerged forests formed as a result of in-situ woods being engulfed by rising sea levels and the submergence of the land and maintained that the seaward exposure of the forests was an extension of the landward subterranean forests. As De Rance (1869b) interpreted the Lower Cyclas Clay as being of freshwater, lacustrine origin and Reade (1871) suggested the clays were marine there was a lively exchange (e.g. in Reade, 1883a) that was not resolved until further stratigraphical details established the context of the Hightown site. De Rance (1877) explained the stratigraphy of this area as a consequence in part of frequent alternations of estuarine and lacustrine conditions, whereas Reade (1871) argued that three periods of subsidence and two periods of uplift were necessary to explain the lithostratigraphy of south-west Lancashire. Potter (1876) added to the controversy by suggesting the drifted nature of the submerged forest deposit. However, Reade (1878a, b) proved that the roots and stumps of the trees penetrated into the blue clay or silt below the peat and he concluded that the trees of the submerged forest grew and fell in the same location. The arguments for the in-situ formation of the peat and forest bed deposits are now accepted rather than the drifted origin, especially as Birks (1964) for Chat Moss showed no evidence for mass eruptions of peat from this moss.

Travis (1926, 1929) interpreted the plant assemblages from both the Forest Beds at Leasowe and Hightown as representing an area that initially was woodland but which developed

Figure 8.116a Prehistoric trackway, Hightown: as it appeared in September 1996 before excavation. (Photo: S. Gonzalez.)

Figure 8.116b Prehistoric trackway, Hightown: during excavation. (Photo: S. Gonzalez.)

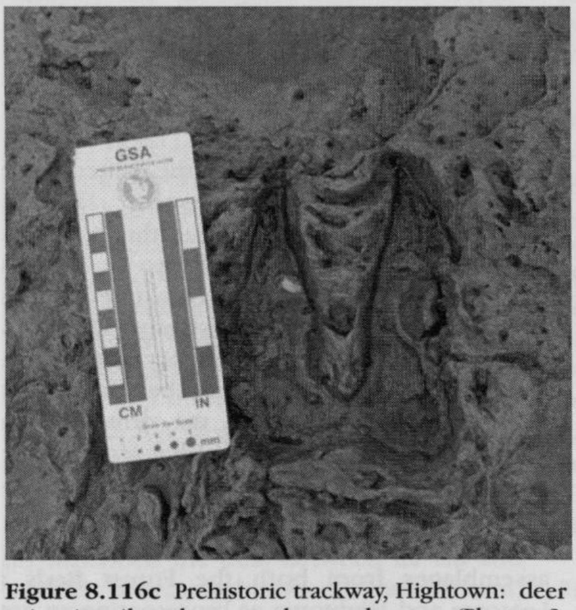


Figure 8.116c Prehistoric trackway, Hightown: deer print in silts close to the trackway. (Photo: S. Gonzalez.)

subsequently, with wetter conditions, into fen- or swamp-carr, brought about by one, or a combination of, the following: impeded inland drainage by the coastal sand dunes, by the depression of the land or by rising sea level. The rhythmic series of clays and sands beneath the submerged forest indicated to Tooley (1977a, 1978a) an unstable coastal environment in which intertidal sedimentation was interrupted by lagoonal and estuarine sedimentation. The organic sedimentation provides evidence of progressive removal of the marine and freshwater influence and their replacement by terrestrial conditions. A complex pattern is indicated by the changing ratio of oak pollen to alder pollen, which shows three oak peaks, interpreted as dry phases, the most recent coinciding with a peak frequency in the spores of *Osmunda*. At the base of the peat, the saltmarsh communities are replaced by reedswamp communities dominated by *Phragmites australis*, with *Typha angusti-*

Table 8.19 Radiocarbon dates associated with the Hightown stratigraphy illustrated in Figure 8.115.

Sample number	Laboratory number	Date (years BP)	Description
56.01	Beta-119011	1180 ± 50	Silver birch tree growing in organic sand
56.02	Beta-119012	4270 ± 60	Silver birch bark from the top of the peat bed
56.03	Beta-119013	4310 ± 50	*Osmunda regalis* (Royal fern) stems from the top of the peat bed
49.01	Beta-119007	4750 ± 80	Intermittent thin band of *Phragmites* peat covering the trackway
49.13	Beta-119009	4430 ± 80	Wooden peg into the trackway
49.16	Beta-119010	4910 ± 60	Part of lowest trackway resting on blue clay
49.11	Beta-119008	5080 ± 60	Part of wooden trackway

folia and *T. latifolia*. The reedswamp is replaced by fen woodland with alder, oak and some birch. This woodland had an understorey of *Myrica* and willow, with woody climbers such as *Solanum dulcamara*. The conditions probably varied considerably, with areas of standing water colonized by *Potamogeton* species and *Hydrocotyle vulgaris*. The pine, elm and lime pollen represent regional components of the pollen assemblage.

The ^{14}C dates in Table 8.19 and Tooley's (1977a) date confirm Erdtman's (1928) estimate based on the pollen assemblage for the end of minerogenic sedimentation in the intertidal zone and the beginning of high saltmarsh sedimentation. The age and altitude (+3.11 m OD) of the stratigraphical contact identifies this position, as it is the highest and youngest, unequivocal sea-level index point in south-west Lancashire and provides evidence for regressive overlap 8 for northern England (Tooley, 1982).

Table 8.20 Tree and shrub species and the type of fossil remains at Hightown (from Travis, 1926).

Species	Type of remains
Pinus sylvestris	Bark, wood
Pinus sp.	Pollen
Myrica gale	Cones, seeds and leaves
Quercus sp.	Bark, wood, acorns, pollen
Betula sp.	Bark, wood, pollen
Alnus glutinosa	Cones, seeds
Corylus avellana	Wood, nuts, pollen
Tilia europaea	Pollen
Salix cinerea	Leaves
Salix aurita	Leaves
Salix sp.	Pollen, wood
Ilex aquifolium	Leaves

Clapham (1999) showed that the fen-carr woodland of the submerged forest was dominated by birch species, but with alder, willow and oak present in smaller quantities. *Cornus sanguinea* and *Frangula alnus* were present as an understorey in some parts, with small pools present where the water table reached the surface as a result of the undulating nature of the underlying sediments. In the pools *Potamogeton* species were present, along with *Alisma* species growing on the margins, and *Menyanthes trifoliata*. The herb layer was dominated by plant species that tolerate high water tables, whereas in other parts large tracts were dominated by a dense growth of royal fern. Other areas, especially those on the edge of the exposure, may have been more open or may represent small clearings within the woodlands where the herb layer was dominated by a mixture of *Phragmites australis*, *Carex* species and Poaceae. Thus both the work of Travis (1926) and Clapham (1999) on the macrofossil species show plants that need, or can tolerate, a high water table. However, it is obvious that this water table was not high enough to stop the tree growth, as indicated by the presence of tree and shrub seeds and the ubiquitous presence of twigs, wood and bark fragments throughout the peat. This indicates a continuous growth of trees since woodland was initiated on the underlying sediment.

Hume (1863) quotes that some of the smaller trees seemed to be in rows at Dove Point on the Wirral and that Dr Aikin had described the same at the mouth of the River Alt, where part of the trunks 'being in a line at equal distances, were undoubtedly planted'. However, there is no evidence for this view today. The quantitative evidence provided by Clapham (1999) suggests that a complex multi-cohort stand was present, as indicated by the large number of small diameter

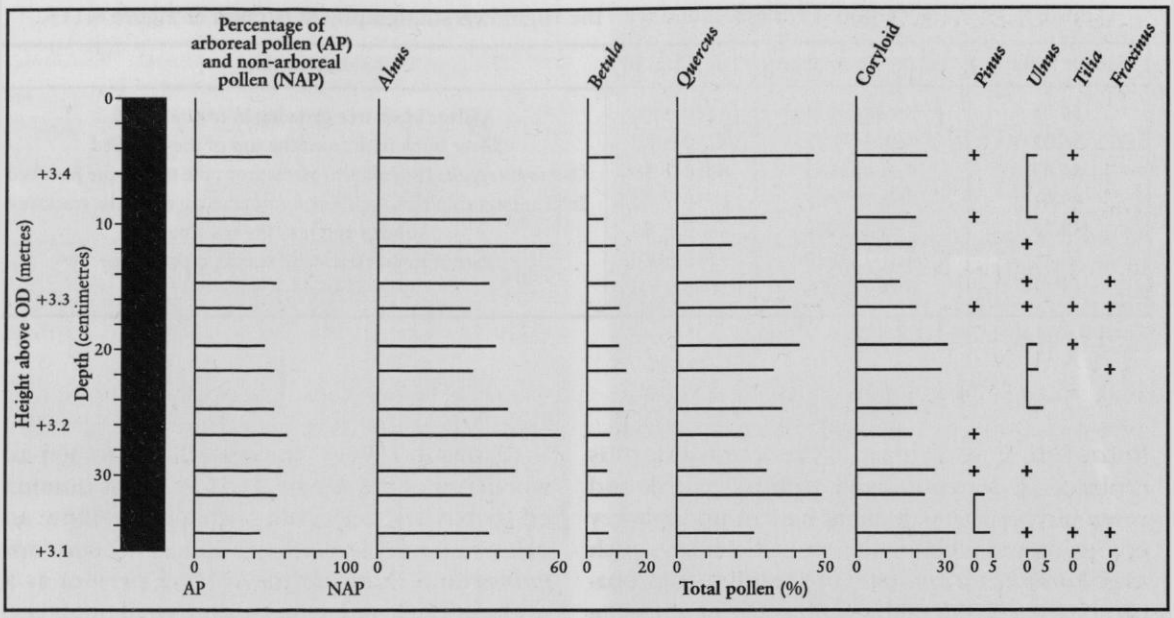


Figure 8.117 Pollen diagram from the Hightown intertidal sediments (after Tooley, 1970).

stumps created by frequent, although not severe, disturbances and the more or less clumped, or random, spatial distribution of these stumps. The angle of fall measured by Clapham (1999) shows that the majority of the trunks lie between ESE and south, unlike the result of Travis (1926), who recorded that most of the stumps, although not showing a main direction of fall, all lay between north-east and north-west.

Clapham (1999) used the community indicator method (Rodwell 1991a, b, 1992, 1995) to interpret the woodland ecology, and found that the dominant woodland types were wetland ones (W1, *Salix cinerea–Galium palustre*; W2, *Salix cinerea–Betula pubescens–Phragmites australis*; W5, *Alnus glutinosa–Carex paniculata*). Aquatic communities were present but not to the same extent as the tall herb fen and swamp communities, which were dominated by S24 (*Phragmites australis–Peucedarma palustre*) tall herb fen. Clapham (1999) also deduced changes in the water table from the samples, with a rise in the water table followed by a drier phase, which he suggested was caused by local changes in the microhabitats as a result, perhaps, of a rise in sea level.

Charcoal fragments were recorded at the junction of the base of the peat and top surface of the underlying sediment, but Clapham (1999) found it difficult to attribute this, to an anthropogenic or natural origin. However, in light of the use of the coast by humans documented both at Formby Point and at Hightown, it is more likely that the charcoal is of anthropogenic origin. The trackway at Hightown has been interpreted as a structure built to facilitate human movement across a saltmarsh towards low water for hunting or fishing activities. There is archaeological evidence too from the recovery of two Early Neolithic axes on the surface of the intertidal zone south-west of Hightown and the presence of Mesolithic–Neolithic sites in the Alt valley to the east (Cowell and Innes, 1994).

Conclusions

Hightown is an important site because it documents important stratigraphical and vegetational changes and woodland development associated with a fall in sea level during the Neolithic period. The site has been documented in detail using many techniques and the environmental changes have been dated accurately. An associated Neolithic intertidal prehistoric trackway has given much new archaeological information and is the first example of its kind in northern England.

CASTLETHORPE (SE 978 077)

D.J.A. Evans

Introduction

The Flandrian deposits of Humberside are dominated by floodplain alluvium and estuarine sediments, but several outcrops of calcareous tufa and shell marl in the Ancholme Valley provide valuable information on the palaeoenvironments of the past 10 000 years (Wright and Wright, 1933; Preece and Robinson, 1984; Gaunt *et al.*, 1992). Tufa is redeposited calcium carbonate or lime. It usually is produced in and around springs or areas of groundwater movement where the water possesses a high calcium carbonate content. The calcium carbonate solidifies into tufa when the water is subject to an increase in temperature or a decrease in pressure. Evaporation of calcium carbonate-rich water also may produce tufa. Shell marl is a calcareous clay containing abundant freshwater molluscs. The most extensive deposits crop out near Castlethorpe Hall (SE 978 077), North Lincolnshire, immediately west of Brigg (Figure 8.118), where the enclosed remains of freshwater and terrestrial snails document the change from marshland to woodland during the early to mid-Flandrian (Musham, 1933; Kennard and Musham, 1937; Preece and Robinson, 1984). Indications of woodland clearance for agriculture during the Bronze Age also are represented by charcoal deposits within the shell marl.

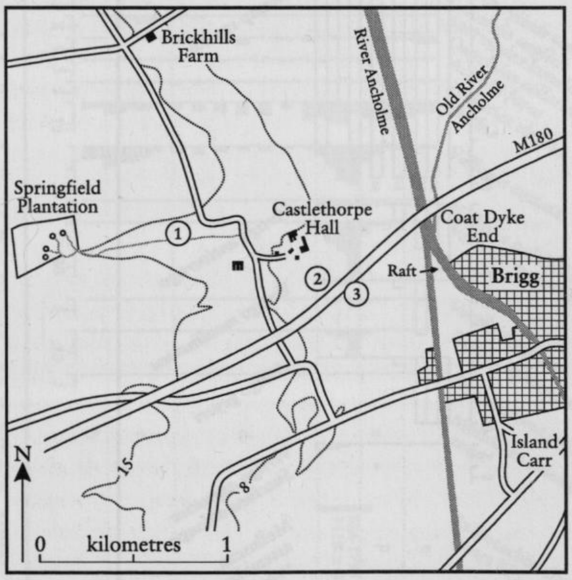


Figure 8.118 Location map of the three Castlethorpe tufa and shell marl sites.

Description

Thin deposits of tufa and shell marl, usually less than 2 m in thickness, are widespread in the small tributaries of the Ancholme Valley, incised into the dip slope of the Jurassic escarpment (Fletcher, 1981). The tufas and marls crop out in the embankments of field drains and streams to the west and south of Castlethorpe (Figure 8.118). Shell marl also underlies the peat and alluvium in the area (Smith, 1958b; Fletcher, 1981). The shell marls, which are creamy coloured, calcareous clays, contain abundant freshwater molluscs associated with the streams that cross the outcrops of the Lincolnshire and Snitterby limestones (Preece and Robinson, 1984).

Exhaustive investigations of the palaeoecological information contained within the tufa and shell marls was undertaken by Preece and Robinson (1984) at three major sites at Castlethorpe (numbered 1–3; Figure 8.118). The details of the lithostratigraphy and Mollusca and Ostracoda from the three sites are reproduced in Figures 8.119 and 8.120. The abundance of the Mollusca allowed Preece and Robinson to identify molluscan assemblage zones (MAZ), defined by the predominance of particular species and based upon a similar sequence of deposits in Kent (Kerney *et al.*, 1980). These zones provide clear indications of changing environmental conditions in the Ancholme Valley.

At Castlethorpe site 1, the Mollusca between 110 and 90 cm (MAZ a) are dominated by swamp species but include some aquatic indicators. The terrestrial Mollusca represent a catholic group, or an assemblage that will tolerate a wide range of habitats, in this case indicative of a wet, open environment. Between 90 and 70 cm (MAZ b) the marshland Mollusca continue to be represented but indicators of more stagnant conditions appear and the terrestrial fauna are enriched. Between 70 and 50 cm (MAZ c) the Mollusca indicate the continuation of swampy conditions but also an environment that is heavily shaded. The Ostracoda at site 1 include 15 species typical of calcareous spring sites. They also indicate that the water at the site was cold and rich in plant debris. Pollen from the under-

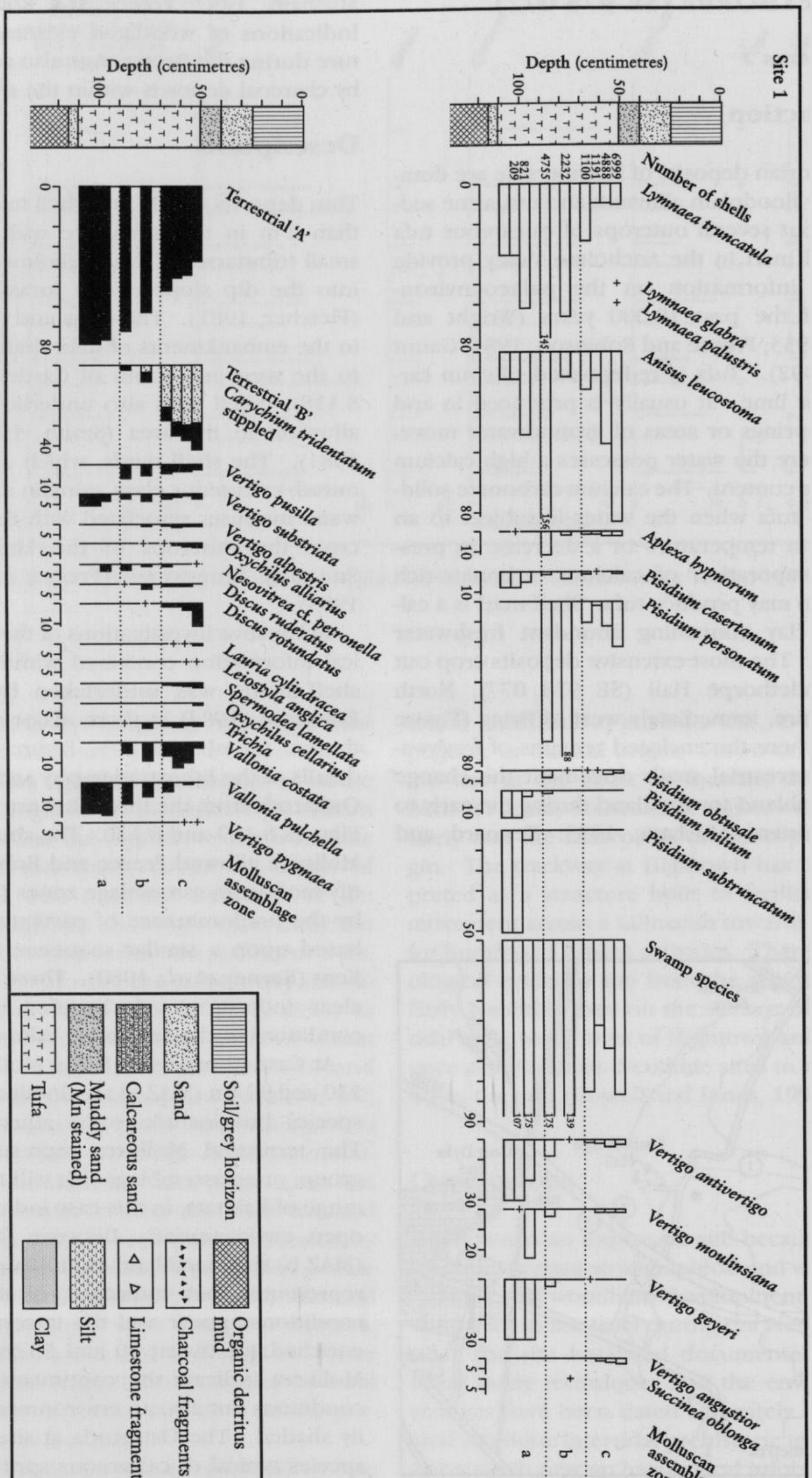


Figure 8.119 Mollusc diagrams with lithostratigraphy of the three Castlethorpe tufa and shell marl sites (after Preece and Robinson, 1984). See opposite for sites 2 and 3.

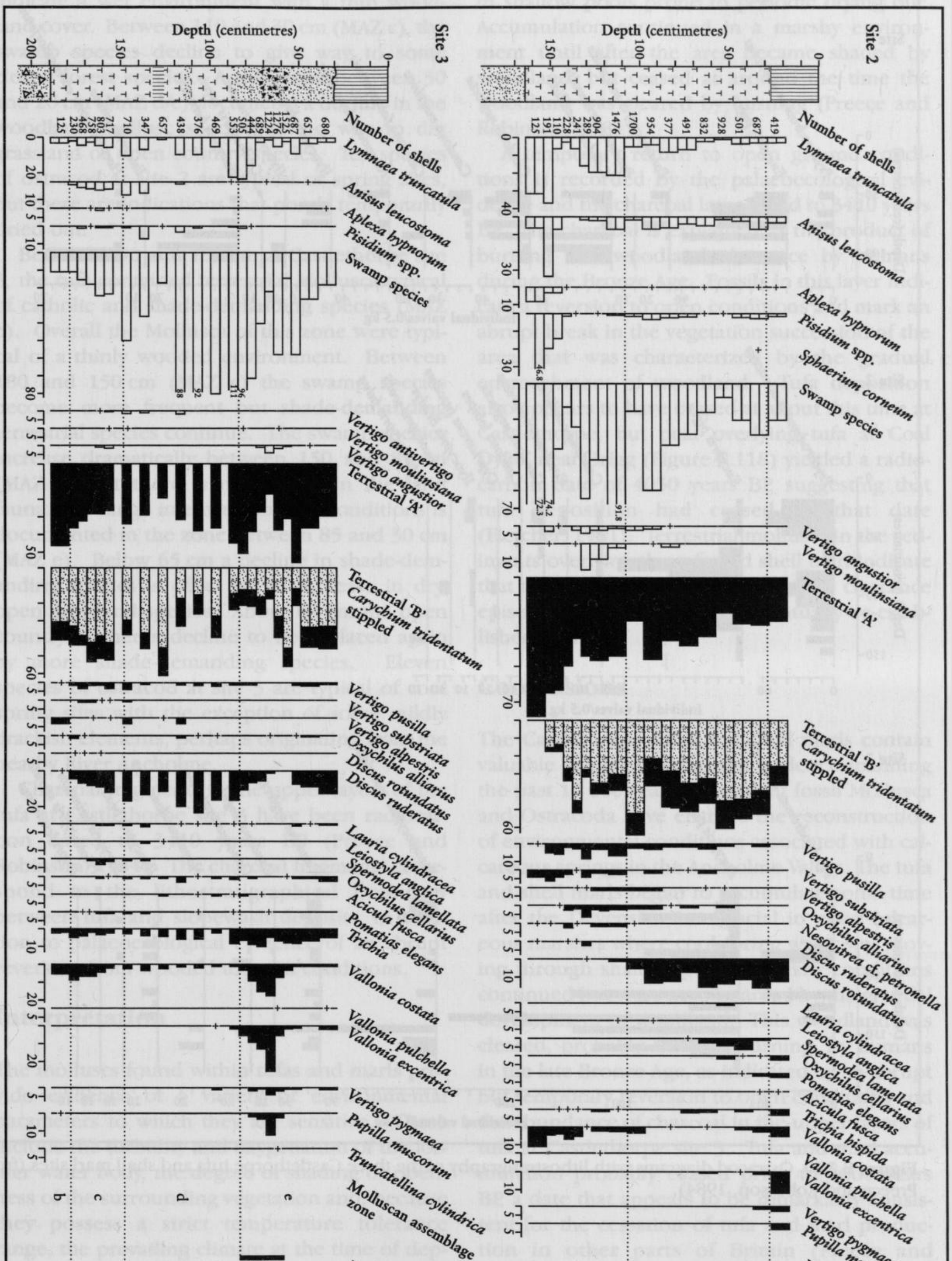


Figure 8.119 *(contd)* Mollusc diagrams with lithostratigraphy of the three Castlethorpe tufa and shell marl sites (after Preece and Robinson, 1984).

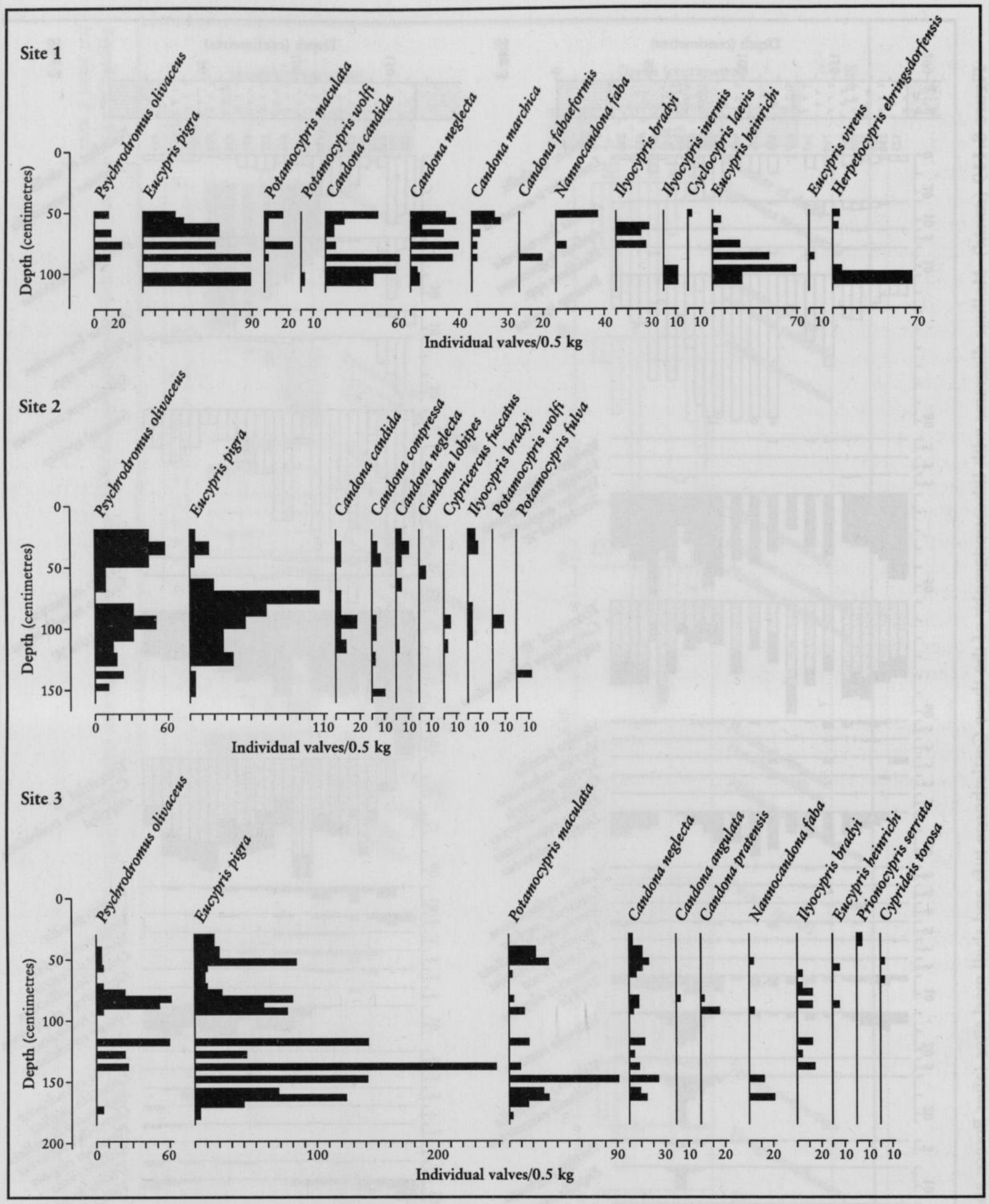


Figure 8.120 Ostracod diagrams with lithostratigraphy of the three Castlethorpe tufa and shell marl sites (from Preece and Robinson, 1984).

lying peat at site 1 are typical of Late Devensian or Late-glacial assemblages, indicating that the accumulation of the overlying tufa may have spanned the whole of the Flandrian.

At Castlethorpe site 2, the Mollusca in tufa between 165 and 110 cm (MAZ b) is dominated by swamp species with a good proportion of shade-tolerant species. The terrestrial molluscs indicate a wet environment with a thin woodland cover. Between 110 and 30 cm (MAZ c), the swamp species decline to give way to some closed forest species. A thin zone between 30 and 20 cm (MAZ d/e) documents a decline in the woodland mollusc species, giving way to dry grassland or open country species. Ten species of ostracod at site 2 are typical of spring sites, but there are indications that ponds temporarily dried out.

Between 190 and 180 cm at Castlethorpe site 3, the tufa contained terrestrial Mollusca typical of catholic and shade-demanding species (MAZ b). Overall the Mollusca of this zone were typical of a thinly wooded environment. Between 180 and 150 cm (MAZ c) the swamp species become more frequent but shade-demanding terrestrial species continue. The swamp species increase dramatically between 150 and 85 cm (MAZ d) but there is no change in the land fauna. A change in environmental conditions is documented in the zone between 85 and 30 cm (MAZ e). Below 65 cm a decline in shade-demanding species is offset by an increase in dry, open grassland species. Above 65 cm the open country elements decline to be replaced again by more shade-demanding species. Eleven species of ostracod at site 3 are typical of cool spring sites with the exception of some mildly brackish elements, perhaps originating from the nearby River Ancholme.

Charcoal fragments in the upper layers of the tufa at Castlethorpe site 3 have been radiocarbon dated to 3410 years BP (Preece and Robinson, 1984). The charcoal fragments correspond to the lithostratigraphical boundary between tufa and slopewash deposits, in addition to palaeoecological evidence of an abrupt reversion from wooded to open conditions.

Interpretation

The molluscs found within tufas and marls provide evidence of a variety of environmental parameters to which they are sensitive. These include the turbidity and oxygenation of the former water body, the degree of shading or openness of the surrounding vegetation and, because they possess a strict temperature tolerance range, the prevailing climate at the time of deposition.

The tufas and shell marls started to accumulate in the early Flandrian in open calcareous marshes that were fed by lime-rich springs and in shallow pools prone to periodic drying out. Accumulation continued in a marshy environment until after the area became shaded by woodland, but ceased at around the time the woodland was cleared by burning (Preece and Robinson, 1984).

A temporary return to open ground conditions is recorded by the palaeoecological evidence and the charcoal layer dated to 3410 years BP. The charcoal is explained as the product of burning and woodland clearance by humans during the Bronze Age. Fossils in this layer indicate a reversion to open conditions and mark an abrupt break in the vegetation succession of the area that was characterized by the gradual encroachment of woodland. Tufa deposition also appears to have ceased at about this time at Castlethorpe, but peat overlying tufa at Coal Dyke, near Brigg (Figure 8.118) yielded a radiocarbon date of 4050 years BP, suggesting that tufa deposition had ceased by that date (Fletcher, 1981). Terrestrial molluscs in the sediments overlying the tufa and shell marl indicate that the area was not farmed after the clearance episode and the woodland eventually re-established itself.

Conclusions

The Castlethorpe tufas and shell marls contain valuable paleoenvironmental evidence spanning the past 10 000 years. Abundant fossil Mollusca and Ostracoda have enabled the reconstruction of environmental conditions associated with calcareous springs in the Ancholme Valley. The tufa and shell marls began to accumulate some time after the Devensian Late-glacial in open calcareous marshes where cool spring water was moving through shallow pools. Marshy conditions continued but were accompanied by the gradual development of woodland. This woodland was cleared, probably through burning, by humans in the late Bronze Age, as indicated by an abrupt but temporary reversion to open conditions and the abundance of charcoal in the upper layers of tufa at Castlethorpe site 3. Tufa and marl accumulation probably ceased prior to 4000 years BP, a date that appears to be remarkably consistent for the cessation of tufa and marl production in other parts of Britain (Preece and Robinson, 1984). Castlethorpe is a valuable geological site with respect to reconstructions of Holocene palaeoclimate and tufa and shell marl deposition in northern England.