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Holocene floodplain vegetation dynamics and sea-level change in the lower Aire valley, Yorkshire

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Holocene floodplain vegetation dynamics and sea-level change in the lower Aire valley, Yorkshire

Perimarine river valleys are a rich resource for studying palaeoecological change which may provide information concerning catchment and sea-level change. This thesis reconstructs the environmental history of the lower Aire valley during the Holocene, and investigates the factors influencing floodplain and vegetational development.

Reconstruction of environmental change in the lower Aire valley in the mid- to late-Holocene is based on lithological and palaeoecological records from three sites in the upper, middle, and lower parts of the study reach. Techniques used include pollen, diatom, wood macrofossil, loss on ignition, and radiocarbon analysis.

Paludification of the valley floor was time transgressive, apparently responding to gradually rising sea level, beginning at *c*.7000 BP (*c*.8000 to *c*.7600 cal. yrs BP) at the lowermost site, whereas conditions were not wet enough for preservation of organic sediment in the upper reach until *c*.4200 BP (*c*.5000 to *c*.8500 cal. yrs BP). Accumulation of floodplain peat was interrupted by the deposition of finely laminated humic clays some time after *c*.7000 BP (*c*.8000 to *c*.7600 cal. yrs BP) in the lower tract of the Aire valley, near Goole, suggesting a change to lagoonal conditions. This was apparently caused by the ponding of freshwater against the rising estuary. It is also possible that drainage was impeded, associated with widespread deposition of organic sediment in the lower valley areas, which may have contributed to the creation of a lagoonal environment. The lagoon had silted up by *c*.6000 BP (c.7200 to *c*.6600 cal. yrs BP), probably due to an increase in tidal asymmetry, and range, which resulted in a net surplus of sediment into the floodbasins and enabled the re-invasion of fen carr onto the site.

The main period of organic sedimentation lasted for several millennia at each of the study sites, during which time *Alnus glutinosa* fen carr communities dominated the wet floodplain backswamp areas. During mid-Holocene times, the vegetation of the surrounding dryland, was colonised by a mixed woodland, with *Tilia*, *Ulmus*, *Quercus*, and probably, *Corylus avellana* and *Fraxinus excelsior*. *Pinus sylvestris* was also prevalent in the region.

Alnus carr was progressively replaced by fen meadow communities, and then saltmarsh or freshwater reedswamp communities, due to a phase of positive sea-level tendency, which was recorded throughout the lower Aire valley between *c*.4600 and *c*.2700 BP (*c*.5500 to *c*.2700 cal. yrs BP). Remnants of a possible upper peat unit and diatom evidence from the upper clastic sediment is tentatively interpreted as indicating the contraction of estuarine conditions and a phase of negative sea-level tendency some time during the late-Iron Age.

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List of Abbreviations and Acronyms

AMS	Accelerator Mass Spectrometry
BP	Before Present
cal. yrs BP	Calibrated years before present
CONISS	Constrained Incremental Sum of Squares
HAT	Highest Astronomical Tide
HWP	Humber Wetlands Project
LAT	Lowest Astronomical Tide
LDAZ	Local Diatom Assemblage Zone
LOEPS	Land-Ocean Evolution Perspective Study
LOI	Loss on Ignition
LOIS	Land-Ocean Interaction Study
LPAZ	Local Pollen Assemblage Zone
MHW	Mean High Water
MHWST	Mean High Water of Spring Tides
MSL	Mean sea level
MTL	Mean Tide Level
NERC	Natural and Environmental Research Council
NNR	National Nature Reserve
OD	Ordnance Datum
OS	Ordnance Survey
PQS	Pre-Quaternary Spores
RPAZ	Regional Pollen Assemblage Zone
RSL	Relative sea level
SBAB	Stable-Bed, Aggrading-Bank
SLIP	Sea-Level Index Point
SSSI	Site of Special Scientific Interest
TDV	Total Diatom Valves
TLP	Total Land Pollen

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Chapter 1 Introduction

1.1 Preamble

Palaeoenvironmental investigations of floodplain and backswamp deposits contained within coastal river valleys provide both ecological and chronological information regarding the Holocene evolution of such systems. Long uninterrupted organic sequences are characteristic of these quiet water environments, which can yield a variety of palaeobotanical data.

The sequence of biogenic and clastic sediments occurring within the upper part of a tidal system has been laid down as a direct consequence of the interplay between estuarine and fluvial processes, which control the level of the water table. This area is called the ***perimarine zone***, and can be defined as 'the area where the sedimentation or seditation (*sic*) took place under the direct influence of the relative sea-level movements but where marine or brackish sediments themselves are absent' (Hageman, 1969; 377).

Odum (1988) and Orson *et al.* (1992) have recognised the lack of research focusing on long term biological and physical aspects of tidal freshwater and oligohaline marshes situated within the upper tracts of estuarine systems. The physical conditions operating in this environment are often transient in nature and complex. The lower reaches of river systems comprise an interesting environmental gradient, between the upstream headwaters and lowland reaches, that changes in character both temporally and spatially. The tidal freshwater zone occurring within this interface is characterised by severe environmental stresses such as fluctuating salinity, reduced dissolved oxygen, and increased turbidity (McLusky, 1994). Fluvial geomorphologists, marine ecologists, and palaeoenvironmental scientists have therefore avoided research into this important environment because they tend to either focus on the freshwater fluvial dynamics where complications due to maritime tidal influence are absent, or on the estuarine and marine processes that dominate seawards (Odum, 1988; Adam, 1990).

The main focus of research into Holocene relative sea-level change has traditionally concentrated on transitory muddy shores such as those found typically in estuaries, where the relatively quiet water environments and fine sediment regime enable the deposition of intercalated layers of marine clastic sediments and semi-terrestrial peats that directly record vertical changes in sea-level and proximity of marine conditions (Shennan, 1992). Stratigraphic and micropalaeontological investigations of such intercalated sequences allow the collection of sea-level index points which identify transgressive (marine overlying terrestrial sediments) and regressive (terrestrial overlying marine sediments) sediment contacts (Shennan, 1982; Tooley, 1982). Sea-level index points can be based on lithology (i.e. sedimentary contacts, Heyworth & Kidson, 1982), but for higher precision they are more commonly identified by microfossil analysis (biostratigraphic boundaries) which enables greater precision (Tooley, 1978a; 1978b, 1982; Devoy, 1979; 1982; Shennan, 1982; 1986a; 1986b; Long, 1992; Plater & Shennan, 1992; Long & Innes, 1993; Long *et al.*, 1998a).

The analysis of such sequences enables a detailed chronology for sea-level movements to be obtained from coastal contexts, but gathering further palaeoenvironmental information regarding the nature of the vegetation communities during different phases of coastal evolution is limited due to the discontinuous nature of the biogenic horizons. This means that attention is focused on periods of changing sedimentary environment over the lithologic contacts, which is restrictive when it is the intervening period which is of palaeoecological interest. Furthermore, precise palaeobotanical information is lost from the record during phases of marine deposition.

Tooley (1985a; 1986) highlights the importance of palaeoenvironmental studies in the perimarine area and notes the largely untapped potential this environment has for study in the British Isles. His assertion is based on the amount of palaeoecological and sedimentological data obtained from studies in the perimarine zone in England by Godwin & Clifford (1938), Godwin (1941), and Godwin & Vishnu-Mittre (1975) as well as in European countries such as Holland (Hageman, 1969; Jelgersma *et al.*, 1979; van de Plassche, 1982) and Germany (Behre *et al.*, 1979). In the period since Tooleys' pertinent observation and the pioneering work of Sir Harry Godwin in the Fenland and Somerset Levels (see references above), there have been notable studies of perimarine environments in Britain (Tooley, 1985a; Burrin & Jones, 1991; Passmore, *et al.*, 1992; Waller, 1994a; 1994c; Hewlett & Birnie, 1996), although this still lags behind the excellent and ongoing research into the Dutch perimarine area (e.g. Verhoeven, 1992; Törnqvist, 1993a).

Research in perimarine fluvial areas has many advantages owing to the diversity of information that can be gleaned from the sediments regarding their palaeoenvironmental development. Not only are there long organic sequences preserved on river floodplains (e.g. Waller, 1993; 1994a) but palaeochannel features, river levees and backswamp lagoons also provide a plethora of information regarding the evolution of coastal fluvial systems and their sedimentary and palaeoecological history (Van der Woude, 1983; Tooley, 1985a; Jennings & Smyth, 1987; Hofstede *et al.*, 1989; Passmore *et al.*, 1992; Törnqvist, 1993a; Weerts and Bierkens, 1993; Waller, 1994c; Janssen *et al.*, 1995; de Klerk *et al.*, 1997a; 1997b; Vos & van Heeringen, 1997; Vos & de Wolf, 1997).

The onset of biogenic sedimentation in the perimarine hinterland often provides the first indirect evidence for rising Holocene sea-levels (Hageman, 1969). However, the relationship between sea-level change, sedimentation, and fluvial activity in the perimarine zone is far from simple (Berendsen, 1984a). For example, in The Netherlands, Hageman (1969) posited that transgressions and regressions impacting on the coastal zone had a direct response registered in the perimarine zone by the synchronous deposition of alternating phases of clastic and biogenic sediments. These lithostratigraphic units relating to transgressions and regressions were then formally classified into members of the Westland Formation and correlated chronostratigraphically with the associated facies in the perimarine zone (Zagwijn & van Staalduinen, 1975). The validity of this model has subsequently been challenged and tested with a large number of radiocarbon dates, (Van der Woude, 1979; De Mulder & Bosch, 1982; Berendsen, 1984a; 1984b; 1993) which has lead to this conceptual model being rejected. It has been shown that differing rates of sea-level rise or fall result in complex response patterns within the fluvial realm, such as shifts from anastomosing to meandering river regimes and periods of increased avulsion, and not simply phases of peat or clay deposition (Törnqvist, 1993a; 1993b; Berendsen, 1995).

Despite the complexity of the relationship between the marine and freshwater system, sediments from the perimarine zone provide a range of environmental data on water-level changes, nutrient status and vegetation history, which are intimately related to sea-level and tidal changes, thereby enlarging the potential area for sea-level studies (Tooley, 1986). The presence of peat deposits within a fluvial catchment also allows insights into the vegetational and sedimentary history of the surrounding dryland environment. Therefore, the perimarine zone of a fluvial area represents a unique resource for palaeoenvironmental study into both water-level changes and vegetation dynamics.

Chappell (1990) notes the highly vulnerable nature of riverine and coastal floodplains to future changes in sea-level. He states that in order to estimate or model the effects of rising sea-level on coastal and riverine lowland basins, it is necessary to establish and understand their behaviour during the Holocene. By collecting data about past fluvial system responses to water-level changes during the Holocene, (which were of similar magnitude to future predictions) engineers, biologists and geologists can plan and begin to mitigate possible changes in the future.

This thesis examines the evolution of the lower tract of one of the main river valleys that drains the eastern extent of the inner Humber estuary. The inter-relationship between long term floodplain development, vegetation dynamics, and changing sea-level is investigated in detail.

1.2 Background to the Study

The work presented in this thesis is part of a research plan set out in 1993 (Van de Noort, 1993), which was intended to supplement the English Heritage funded Humber Wetlands Project (HWP). From the outset of the wetlands survey, it was clear that specific palaeoecological questions could not be fully addressed within the remit of the project. This was mainly due to the time constraints associated with the size of the geographical area to be studied. As a result, the Department of Geography planned to fund several PhD students to study certain aspects of the Humber wetlands in more detail, to compliment and expand upon the palaeoenvironmental data already collected by the HWP. This research is the result of the first PhD produced as part of this initiative, which has enabled a high spatial and temporal resolution model of wetland development in relation to Holocene sea-level change to be obtained.

In addition to the HWP, this research also relates to the Land-Ocean Interaction Study (LOIS) funded by NERC. In part, the LOIS project aimed to model Holocene depositional regimes in the western North Sea. Another element of LOIS is the Land-Ocean Evolution Perspective Study (LOEPS). One project, involving collaboration between the Universities of Durham, Edinburgh, Hull, Newcastle and St. Andrews, is entitled 'Holocene Evolution of the Humber Estuary'. This aimed to reconstruct the changing land-ocean boundaries during the Holocene in the Humber and aid in the identification of the critical geomorphological and ecological processes that determine that boundary. This type of holistic study is critical in the understanding of broad-scale patterns of sea-level change and coastline evolution in the Humber, but is lacking in high resolution data from the perimarine river valleys that drain into the inner Humber Estuary.

This study will build on the framework established by the HWP and LOIS, with the focus on providing new information concerning the timing and nature of peat development within marginal river environments, and to establish the relationship of these deposits to Holocene sea-level change. Therefore, this thesis is a contribution to the considerable amount of ongoing palaeoenvironmental and archaeological research currently being undertaken in the Humber region.

1.3 Aims and Objectives

1.3.1 Aims

The aims of this thesis are to:

1. *Identify sites that characterise the range of palaeoenvironments in the perimarine zone of the lower River Aire valley, within the Inner Humber estuary.*

There is a general paucity of data regarding the Holocene evolution of inner estuarine environments. Data concerning the coastal evolution of middle and outer estuaries are abundant as are those from upper reaches of catchment systems. However, the interface of these two important systems has received scant attention. It is hoped that exploration into this enigmatic palaeoenvironment using lithological and biostratigraphical techniques will shed light on the processes that operate in such environments and the relative extent to which sedimentation and vegetation are controlled by fluvial or estuarine processes.

2. *Identify water-level changes within inner estuarine environments and assess the extent to which they coincide with known periods of marine transgression and regression within the middle and outer estuary during the Holocene.*

The response of marginal fluvial riverine systems to Holocene sea-level changes is the subject of much debate. Therefore, the registration of changes in water-level close to the tidal limit and the timing, duration, and elevation of such changes are of particular interest. Godwin & Godwin (1933a) have shown how subtle changes in the water table are manifested by distinct vegetative changes, which can be observed within the biostratigraphic record. Research of this nature may then provide analogues for the possible effects of predicted future sea-level changes which may be of similar magnitudes.

3. Examine the composition of local riverine vegetation communities and character of the surrounding vegetation.

To date, detailed palaeobotanical studies in the Humber basin have been based largely on lake (mere) deposits (e.g. Beckett, 1981), raised mires (e.g. Smith, 1985a), the chalk uplands (Bush, 1986), or fragmentary records from the shore of the Humber estuary (e.g. Gaunt & Tooley, 1974; Long *et al.*, 1998a). The only data available from perimarine river valleys is Smith's (1958a; 1958b) work within the Ancholme which, although an important contribution to our knowledge of environmental change in the Humber river valleys, is now somewhat outdated and of limited use owing to the lack of a radiocarbon chronology. This study aims to address this gap in information, by concentrating on the nature of the vegetation colonising floodplain and backswamp areas within a river valley draining into the inner Humber estuary, and identifying the forcing factors that influence vegetation change during the Holocene. The recognition of the off-site vegetation signal from this context may also provide a different perspective on the character of the dryland vegetation communities within the catchment area.

4. Use empirical data to evaluate existing theories of floodplain development in perimarine environments.

On a broad scale, patterns of deposition within coastal valley systems during the Holocene can be considered paradoxical. Generally, the very early Holocene can be characterised by minerogenic sedimentation during a time of rapidly rising sea-levels, a response to climatic amelioration, whilst the mid-Holocene is dominated by the deposition of organic sediments as the rate of sea-level rise slowed. Coastal environments during mid- to late-Holocene times usually register a switch to clastic sedimentation during a period when the rate of sea-level rise is lower, when a continuation of peat forming conditions may be expected. It has been suggested that the reasons for this apparent contradiction are related to the increasing impact of human activity (Burrin, 1985; Brown, 1987a; 1987b; Jennings & Smyth, 1987). The integration of litho- and biostratigraphic techniques is necessary to provide a link between changes in vegetation and sedimentary environment which may shed light on the factors influencing sedimentation in littoral environments.

On a more local scale, previous hypotheses regarding the cause of late Holocene sedimentation within catchment systems in the Humber basin has been based upon largely archaeological evidence from a limited number of sites. It is hoped that the quantity of stratigraphic and palaeoecological data gathered in this study, in conjunction with a radiocarbon chronology, can contribute to this debate by critically examining the validity of these informal conceptual models of floodplain alluviation to other valley systems in the inner Humber area (Samuels & Buckland, 1978; Buckland & Sadler, 1985; Riley *et al.*, 1995).

1.3.2 Objectives

The specific objectives of this thesis are to:

1. Establish the pattern of Holocene sedimentation in the lower Aire valley through lithostratigraphic exploration and schematic data displays.

2. Identify the range of sedimentary palaeoenvironments within the perimarine zone and sample sites suitable for palaeoenvironmental analysis from the upper, middle and lower reaches of the study area.

3. Reconstruct the palaeoecological history of the study area using biostratigraphic techniques.

4. Identify regional trends in vegetational development.

5. Elucidate the palaeobotanical evidence for changes in vegetation communities within organic and inorganic deposits as proxy data for changes in the altitude of the water table.

6. Employ ^{14}C dating to establish an absolute chronology for vegetational history and the pattern of Holocene water-level movements in the study area.

7. Study the evidence for tendencies of sea-level movements in the Humber estuary.

8. Analyse the evidence for synchroneity between the chronology of sea-level tendencies in the middle and outer estuary against those identified from the study site within the inner Humber estuary.

9. Critically examine previous models of floodplain development and sedimentation patterns in the light of these results

1.4 Structure of Thesis

This thesis presents the results of a palaeoecological study into environmental change within perimarine environments from a riverine context. Data have been collected from three sites along the floodplain of the lower Aire in the inner Humber estuary and are used to elucidate the relationship between sedimentary and vegetational environments at the junction of the estuarine and fluvial catchment system. The thesis has been organised into four main parts, consisting of a total of 11 chapters, as follows:

Part I Background

Following the present chapter, Chapter 2 provides the context for the work by introducing floodplain environments and reviewing previous concepts and ideas regarding the causal mechanisms controlling the evolution of low-lying sedimentary river environments. Following this, an introduction to the terminology and methodology of sea-level data is given, along with a consideration of how Holocene sea-level data is interpreted and the possible problems with these interpretation methods in the perimarine zone. Chapter 3 is an introduction to the study area. The site locations are defined and the specific site conditions (geology, land-use, climate etc.) associated with the area under investigation are outlined. In addition, the Quaternary development of the landscape in the Humber region is reviewed and the physical characteristics of the Humber estuary are also set out. The literature (archaeological, palaeoenvironmental and geomorphological) concerning research undertaken within the Humber basin in general is reviewed in Chapter 4.

Part II Methods and Techniques

Chapter 5 addresses the methodology employed in all aspects of data collection, (the lithostratigraphic approach is justified and the biostratigraphic investigations are introduced). The methodology used in the pollen data collection is outlined and an account of the problems associated with interpreting palynological data from perimarine riverine areas is given. Other methods such as diatom, loss-on-ignition and wood analysis are presented. Finally the principles of radiocarbon dating are considered and the advantages and limitations associated with this technique in floodplain and coastal environments is assessed.

Part III Results

Chapter 6 is a summary of the coring program, with brief written descriptions of the sediments recorded within the river valley, and the presentation of simplified schematic data displays. Chapters 7, 8, and 9 are site reports from the three principal study areas, within which the litho- bio- and chronostratigraphic results are described and interpreted.

Part IV Discussion and Conclusions

Chapter 10 provides a synthesis of the findings from the three sites in terms of the chronology of wetland vegetation development and patterns of sedimentation. Broader scale inferences are also made as to the character of the dryland vegetation communities within the study region. A chronology of sea-level changes is presented and comparisons made between the study sites and other data from the outer estuary areas. The dominant controls of floodplain development and local vegetation change are outlined and inferences made as to the applicability of previous models of floodplain alluviation within the Humber basin. The conclusions and recommendations for future research are given in Chapter 11.

1.5 Conventions and Definitions

1.5.1 Chronological Framework

The timescale of this investigation spans the latter part of the **Quaternary** period. The Quaternary covers the last 2.5 million years and is the most recent subdivision of the geological record (Lowe & Walker, 1997). Hedberg (1976) has formally designated two epochs within the Quaternary; the **Pleistocene** (up until *c.*10000 radiocarbon years before present) and the **Holocene**, which covers the current warm period in which we live (from *c.*10000 years ago until and including the present). The brief climatic oscillation that occurred towards the end of Pleistocene in northwestern Europe is referred to as the **Lateglacial**. The Lateglacial effectively marks the end of the last cold stage, and the term **Postglacial** is informally used to designate the following warm stage. The term **Flandrian** is the formal name for the present interglacial stage in Britain and some parts of northwest Europe (Lowe & Walker, 1997). This term is derived from the European practice of naming temperate stages after characteristic marine transgressions; in this instance on the Flemish coastal plain (Hyvärinen, 1978; West, 1979). However, because this term has not been widely adopted internationally, the term Holocene is used hereafter to refer to the present interglacial stage (Lowe & Walker, 1997).

The use of the term 'before present' (BP) has been adopted, whereby 0 BP ('present') is conventionally defined as before 1950 AD. All ages are also expressed on a calendrical timescale (see section 5.10) and have been calibrated using the CALIB 3.0 program of Stuiver & Reimer (1993). The beginning of the Holocene is therefore referred to as *c.*11500 calibrated years BP (cal. yrs BP).

1.5.2 Nomenclature

Vascular plant nomenclature follows Stace (1997) and pollen nomenclature is based on the suggestions of Bennett (1994a) and Bennett *et al.* (1994). Where other authors have used a different naming system, it has been converted as far is possible to this method. Diatom nomenclature follows Hartley (1986). Place names follow Ordnance Survey nomenclature.

Chapter 2 Floodplain Development and Sea Levels

2.1 Introduction

This chapter outlines the physical and biotic characteristics of river floodplain and backswamp environments. The influence of autogenic and allogenic factors are discussed in relation to the development of Holocene floodplain vegetation. A review of the low-lying fluvial sequences investigated by previous authors, and the different models of floodplain formation they advance are considered. In particular, the debate concerning the relative contribution of climatic and anthropogenic factors in determining the development of Holocene fluvial systems is discussed.

Information concerning the nature of both fluvial deposition and coastal processes is essential when studying the evolution of sedimentary sequences in the perimarine zone where these two realms interdigitate. Pethick (1988) notes that the majority of sediments in the Humber today are derived from North Sea sources (particularly erosion of the Holderness coast), but what proportion was derived from river or sea sources during the Holocene is not known. This is of particular significance, as changing catchment processes will affect the supply of terrestrial and fluvial sediment to the coastal zone (Shennan, 1994), which plays an important role in the balance between terrestrial and marine sedimentation.

An understanding of sea-level change is particularly relevant to this study as this is a major control on groundwater levels, and therefore floodplain development in coastal river valleys. The methodology and terminology used in the reconstruction of Holocene sea-level change is dealt with in the latter part of this chapter. The factors that influence the relative position of land-sea boundary, the methods of reconstructing and interpreting former sea-levels, and the problems associated with these approaches in the perimarine zone are then discussed.

2.2 Holocene Floodplain Development

2.2.1 Definitions and Background

Floodplains can be defined as the flat areas adjacent to rivers liable to flooding (Brown, 1997). A study of the alluvial facies of floodplains can provide valuable insights into factors influencing the natural evolution of the fluvial system, provide evidence of climate change, ecological dynamics, and nutrient status, and yield information regarding prehistoric human utilisation of such environments .

The term 'floodplain' is rather misleading. Fenneman recognised as early as 1906 that floodplains were not solely the result of floods but also comprise deposits formed by river migration (Fenneman, 1906); they are a product of both erosional and depositional processes. The floodplain is made up of a complex array of features including levees, point bars, oxbows, terraces, pools, and backswamps; formed by a variety of different lateral and vertical processes (Brown, 1997). In a coastal valley setting, the floodplain evolves under the combined influence of physical (catchment, fluvial, and estuarine) and human processes. Detailed investigation of environmental change from these transitional perimarine contexts can provide information relating to a variety of allogenic and autogenic biological processes (see section 2.2.3). It is important for the relative influence of fluvial and estuarine parameters to be considered together, as they are fundamental to the understanding of the evolution of tidal floodplain environments.

2.2.2 Floodplain Sediments

The alluvial architecture of floodplain sedimentary sequences consists of a variety of sediment types, derived from different depositional processes. In temperate regions, coarse grained clastic material such as sand and gravel may derive from the inwash of colluvium (Bell, 1982; Brown, 1982; Burrin, 1988; Waller *et al.*, 1988; Brown, 1992), windblown dunes or river levees (Törnqvist, 1993a; van de Plassche, 1995), or reworked loess deposits (Burrin & Scaife, 1984). Channel fill deposits and peripheral bank/valley side sediments are usually made up of such coarse grained material. Fine grained sediments usually dominate the valley fills of low lying coastal river valleys.

In the perimarine zone of coastal river valleys, clays and silts are deposited by tides in a low energy environment, forming wide mud flats flanking the channel, with peat forming in the vegetated backswamp wetland due to high water tables on the upper parts of the floodplain. Such sediments make up the majority of Holocene floodplain sediments within coastal river valleys, and can be over 10 metres thick in large estuarine floodplains (see Gaunt & Tooley, 1974; Devoy, 1979). Channel switching and meander migration result in palaeochannel features which are also usually infilled with fine sediments such as clay, silt, and peat.

Modern fen systems, which appear to closely resemble the Holocene wetlands within the perimarine river valleys in low-lying areas, such as the Humber Wetlands, with similar morphological and biological characteristics, are known as 'floodplain fens' (Wheeler, 1984). These are 'topogeneous fens (i.e. with a high watertable maintained primarily by the topography of the site), which developed on waterlogged, and often periodically inundated floodplains, alongside rivers and streams' (Wheeler, 1984; 239).

2.2.3 Holocene Floodplain Vegetation

The most common community present on eutrophic floodplain wetlands is *Alnus* carr (close modern analogues are the *Osmundo-Alnetum* of Wheeler, 1980b and *Alnus-Carex* woodland (W5) of Rodwell, 1991). *Alnus glutinosa* is the only species of alder native during the Postglacial period in Britain (Godwin, 1975a). Other tree species such as *Corylus avellana* and *Salix* are often important components of *Alnus* carr (Rodwell, 1991), although such communities are often species poor due to their impenetrable nature and high degree of canopy shading (Brown, 1997). Favourable conditions during much of the mid-Holocene (e.g. waterlogging and substrate stability, Brown, 1988) made the *Alnus glutinosa* woodland community very dense and almost non-invadable resulting in a low species diversity (Brown, 1988; 1997). Whilst *Alnus* carr may dominate the central areas of floodplain, Wheeler (1984) suggests taxa such as *Glyceria*, *Filipendula*, *Urtica*, *Epilobium*, *Phragmites* and Cyperaceae are associated with silty substrates along the banks of lowland rivers, with *Glyceria* particularly prominent in areas of strong tidal influence.

The development of vegetation patterns in wetlands are the product of both autogenic and allogenic factors (Walker, 1970; Jackson *et al.*, 1988; Singer *et al.*, 1996; Bunting & Warner, 1998). Autogenic factors are usually biotic processes which operate naturally within an ecosystem to dictate the direction of change in plant succession until the vegetation reaches 'equilibrium' with the environment. Tansley (1939) identified sequential stages, or **seres**, from open water through to reedswamp, fen, fen 'carr' (a term used to describe woody scrub vegetation) and ultimately a 'climax' vegetation consisting of mixed deciduous woodland or raised bog. Godwin (1978) applied this theory to vegetation change in the Fenland, and described a progressive series of vegetation succession whereby open water passes to reedswamp, sedge fen, fen woodland, poor fen and then bog purely as a result of autogenic processes (see Figure 2.1a).

Early models of hydroseral succession, whereby waterbodies and wetlands gradually become terrestrialised by autogenic mechanisms (i.e. the accumulation of autochthonous organic and inorganic sediments), only apply to closed systems and do not explain most patterns of floodplain vegetation communities or other pathways of vegetation change. More recent authors have stressed the importance of allogenic factors (external parameters) in determining the stages of hydroseral succession (e.g. Walker, 1970; Wheeler, 1984; Jackson *et al.*, 1988; Burrows, 1990; Delcourt & Delcourt, 1991; Singer *et al.*, 1996; Bunting & Warner, 1998; Waller *et al.*, 1999). Stratigraphic investigations provide evidence that successional sequences are variable and often show reversals in direction and skipped stages in open systems, controlled by allogenic processes (e.g. Walker, 1970; Delcourt & Delcourt, 1991; Bunting & Warner, 1998), which may be physical or biotic, such as the dominant species first reaching a site (Walker, 1970). In fact, all alluvial wetland successions in perimarine areas are influenced by a range of allogenic processes, such as the sedimentation of allochthonous material (organic and inorganic), changes in groundwater level and quality, flooding, climate change, subsidence, and vegetation disturbance due to human impact, which reflect the interaction of coastal and terrestrial processes (e.g. Clark & Patterson, 1985; Clark; 1986).

One of the features of plant communities in coastal and floodplain wetlands is that they exhibit a spatial zonation, which is related to elevation (i.e. distance from the river channel or the sea). This zonation can be regarded as a spatial manifestation of vegetation changes occurring through time, that are recorded in the biostratigraphic column (see Godwin, 1940; Shennan, 1986b; Waller, 1994c).

The most important allogenic factor controlling groundwater level, and thus sedimentation and vegetation development in coastal river valleys, is sea-level change (Behre, 1986). Although sea level may be the dominant regional control on watertable height, other factors influence the local groundwater level, such as the stratigraphy and hydraulic conductivity of sediments, the slope of the floodplain watertable, the degree of rainfall and flooding, and the rate of evapotranspiration from plants and the ground surface (Brown & Bradley, 1996). However, it is clear that the principal factor initiating peat formation over formerly dry basal surfaces was rising water levels during the early stages of the Holocene (e.g. Jelgersma, 1961; Hageman, 1969; Tooley, 1978a; van de Plassche, 1982; Pons, 1992; Shennan, 1994; Waller, 1994c; Vos & van Heeringen, 1997). Indeed, basal peats were thought to be so intimately related to sea-level rise that early curves charting the Holocene rise in sea-level in The Netherlands were derived entirely from radiocarbon dates from the base of basal peats on the gently inclined Pleistocene surface (Jelgersma, 1961; van de Plassche, 1982), where the effects of compaction would be minimal. Although the validity of sea-level curves derived in such a manner has recently been questioned (e.g. van Dijk *et al.*, 1991; Pons, 1992; Kiden, 1995; Vos & van Heeringen, 1997, see section 2.4.6), there can be little doubt that change in vegetation from dryland to fen woodland (see Figure 2.1b) was driven by an allogenic process (Waller, 1994c).

Behre (1986) has proposed a progressive and regressive series of vegetation changes for coastal areas (see Figure 2.1). The direction of these vegetation changes have been taken to indicate rising or falling sea-level. The progressive series, from intertidal environments, through reedswamp and finally into fen carr (Figure 2.1c), may be explained as a consequence of allogenic factors such as falling sea levels, or increasing allochthonous sedimentation rates, although the gradual build up of the ground surface may be brought about by autogenic processes. The retrogressive series describes the sequence of vegetation changes from fen carr, through reedswamp, into intertidal flats (Figure 2.1c). An allogenic process, such as rising sea level, can again be invoked to explain the shift from terrestrial to mudflat environment. However, this explanation may not apply in all cases because the same vegetation change could result from a decrease in rate of local sedimentation (either allogenic or allochthonous) during a period when the rate of relative sea-level change remains constant. Therefore, the nature and direction of changes in vegetation 'are ultimately determined by the rate at which both allogenic and autogenic processes are operating' (Waller, 1994c; 42).

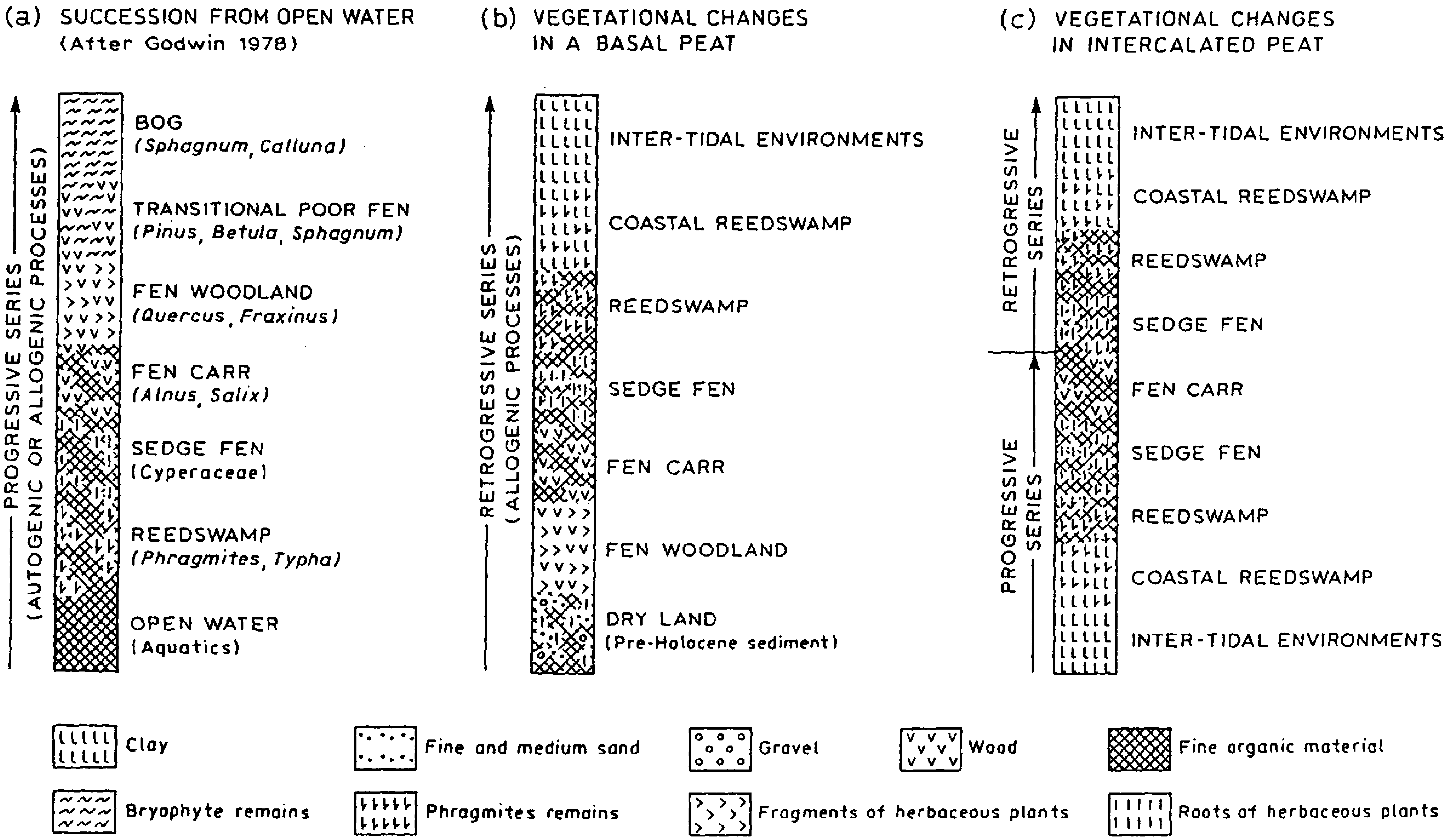

Figure 2.1 Patterns and pathways of vegetation change, as shown in sediment cores. The diagram shows Godwin's model of succession (a) and the retro- and progressive series envisaged in basal (b) and intercalated peats (c). Based on Godwin (1978) and Waller (1994c). Terminology follows Behre (1986).

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2.3 Theories of Valley Sedimentation and Floodplain Genesis

2.3.1 Definition of terms

Floodplain sequences have been subdivided into two broad categories by Macklin & Lewin (1986) who classify British drainage basins into 'upland' and 'lowland' types.

- **Upland valley systems** are characteristic of rivers in northern and western Britain where the landscape has been glaciated several times during the Pleistocene. Such catchments generally have steeper gradients and higher unit discharges than their lowland counterparts.

- **Lowland valley systems** are characteristic of rivers in southern and eastern Britain which are generally beyond the limits of Quaternary glaciation (or experienced glaciation only in their headwaters). Such catchments have low relief, minimal gradients, and low unit discharges.

Burrin & Scaife (1984) divide lowland Holocene alluvial fills into two further categories:

- **The coastal zone sequence**, which occupies the lowest portion of the valley tract. This sedimentary sequence generally consists of deep interbedded clastic and biogenic deposits reflecting both freshwater and estuarine depositional environments (e.g. Devoy, 1979; Heyworth & Kidson, 1982; Jennings & Smyth, 1987; Smyth, 1986; Burrin, 1988; Burrin & Jones, 1991; Waller *et al.*, 1988; Hewlett & Birnie, 1996). These sequences occur within the perimarine zone and coastal parts of the valley, where alluvium is deposited under the influence of sea-level movements. The processes through which such coastal sequences form, and the debate concerning the interpretation of such sediments in relation to sea-level change, is considered later in this chapter (Section 2.3.3 and 2.4.6).

- **The floodplain association**, which consists of predominantly fine grained sediment overlying bedrock or glacial drift deposits. Such sequences occur at the interface between the upper reaches of the catchment and the coastal zone sequence (e.g. Beckinsdale & Richardson, 1964; Dury, 1964; Shotton, 1978; Hazelden & Jarvis, 1979; Burrin & Scaife, 1984; Buckland & Sadler, 1985; Burrin, 1988;

Burrin & Jones, 1991; Waller, 1993; 1994a; Waller *et al.*, 1988; Hewlett & Birnie; 1996). Beckinsdale & Richardson (1964) note that valleys narrow, and the sediments in such floodplains typically shallow and become younger in an upstream direction. This feature is shown in the long profiles presented in Burrin & Jones (1991) and Scaife & Burrin (1992).

The complex causal factors and mechanisms of valley fill alluviation have long been the subject of considerable debate. In southern and eastern England, over the last 10000 years, there has been a general vertical tendency of progressive alluviation of fine grained sediment, largely as a result of glacio-eustatic relative sea-level rise (Straw & Clayton, 1979; Waller *et al.*, 1988), itself a consequence of climatic amelioration. However, later alluvial phases have often been attributed to increased catchment soil erosion as a consequence of human activity (e.g. Shotton, 1978; Bell, 1982; Brown, 1983). Therefore it is appropriate to consider the principal arguments put forward to explain valley alluviation in Britain during the Holocene.

First it is necessary to consider each of the factors that combine to influence floodplain development. Burrin & Scaife (1984) and Burrin (1985) define exogenic and endogenic factors, which influence valley systems by initiating phases of alluviation and floodplain construction, and terrace development, or periods of erosion.

Exogenic factors are external, environmental parameters, that are independent of the fluvial system. These include:

- Sea-level and base level change
- Tectonic activity
- Climate change (e.g. affects flood frequency and magnitude)
- High magnitude, low frequency events (e.g. catastrophic events such as storm surges)
- Anthropogenic activity (direct and indirect – land-use change)

Endogenic factors are internal responses by the fluvial system to exogenic environmental change, which often involve complex response mechanisms, including thresholds (Schumm, 1977; 1979) and autogenic processes.

2.3.2 Introduction to Geomorphological Change

Early models of geomorphic change in the late 19th century had climate as the primary control, until Davis (1899; 1902) proposed that cycles of erosion and base-level changes were the major driving factors in landscape change and floodplain evolution. Davis' model applied over a geological timescale, and floodplain formation was seen as the 'old-age' stage of landscape evolution after orogenic uplift. Baulig (1935) also recognised the importance of base-level change in floodplain evolution, stating that glacio-eustasy was a major control in defining the long profiles of river systems. These theories were important as they suggested that factors other than climate were influential in driving landscape change. It is clear that climate is the central factor concerning floodplain evolution over geological timescales, spanning different glacial periods. However, the importance of smaller, catchment scale processes could not be readily incorporated into, or explained by such models, that were essentially concerned with long time periods.

Schumm & Lichty (1965) developed a model to explain geomorphic change on a historical scale. It assumed that river environments operated as a process-response system, and variables such as sediment load and water flow vary in response to a range of external environmental controls. The model includes the fact that the status of the variables differs with the timescale of the investigation. Over shorter timescales, channel forms develop in response to both the amount of water and sediment supplied to the system as well as valley characteristics inherited from geological time, such as gradient and valley fill deposits. Therefore, major climatic episodes may have influenced the antecedent conditions of the catchment, but other physical and/or anthropogenic processes must be invoked to explain smaller scale, shorter term changes in river regime.

2.3.3 Causal Mechanisms and River Characteristics of Holocene alluvial systems in Britain

During the late-Devensian, deglaciation occurred following climatic amelioration (Manley, 1964; Godwin, 1975a; Goudie, 1992) which resulted in rapidly rising Postglacial sea levels (Gaunt & Tooley, 1974; Tooley, 1978a; Devoy, 1982; Heyworth & Kidson, 1982) and subsequently in valley aggradation and floodplain formation (Straw & Clayton, 1979). There are many examples in Britain where floodplain aggradation reflects the elevation of local base levels in the perimarine area associated with rapid early to mid-Holocene sea-level rise (e.g. Martin Mere, Tooley 1985a; the Brede valley, Waller *et al.*, 1988; the Sussex Ouse, Burrin & Jones, 1991; the lower Tyne, Passmore *et al.*, 1992; the lower Severn, Hewlett & Birnie, 1996). The similar pattern of fluvial activity, and widespread deposition of organic rich fine sediments on coastal valley floodplains, are a corollary of their low gradients and perimarine location (Passmore *et al.* 1992). Straw & Clayton (1979) suggest that aggradation of river floodplains upstream of estuarine zones in eastern England was largely complete by *c.*5000 BP (*c.*6000 to 5500 cal. yrs BP) or earlier, in response to rising sea level, and prior to the establishment of significant vegetation cover on the valley sides. However, apart from rather general remarks concerning valley evolution in coastal lowlands, reflecting broad environmental changes (e.g. Beckinsdale & Richardson, 1964; Kidson & Heyworth, 1976; Straw & Clayton, 1979), detailed examination of how and when such activity occurred remained elusive until the past few decades. Since this time, there has been considerable debate concerning the precise causal mechanisms of floodplain activity.

Wolman & Leopold (1957) developed a model of floodplain formation from observations of a small meandering stream in the USA which was generally accepted for many years. They postulate that floodplains are comprised largely of coarse point bar sediments, formed by lateral migration of river meanders across the floodplain floor. They reason that vertical accretion from overbank sediments contributes only a minor component of the overall valley fill. In their model, fine overbank sediments become progressively limited due to successive deposition of flood deposits raising the floodplain relative to the channel, making later floods more and more infrequent. There are two important assumptions inherent in this model that have subsequently restricted its applicability to many fluvial situations. First, it requires there to be no within channel sedimentation, or channel bed aggradation, as this would facilitate a constant rate of overbank deposition as the floodplain/river bed level would remain relatively the same. Secondly, it assumes that the channel migrates evenly over the entire floodplain which most channels do not do. The relative contributions of lateral and vertical overbank sediments to the floodplain fill depends very much on the sediment type and supply (Schumm & Lichty, 1963), and the lack of knowledge regarding the balance between vertical and lateral accretion has led to much uncertainty (Burrin & Scaife, 1984).

Nanson & Young (1981) recognise that the processes of floodplain construction are variable, and influenced by many exogenic and endogenic factors specific to certain geographical areas, and regard the Wolman & Leopold (1957) model as inappropriate for an idealised model of general floodplain evolution. Burrin (1985) advocates a move away from simplistic, generalised models of floodplain alluviation to ones which reflect the complexity of the range of fluvial environments. It is now realised that there can be no universal floodplain model (Brown, 1997) due to the different amounts of sediment, and types of river system. Although the Wolman & Leopold (1957) model may still accurately describe floodplains created by a small meandering channel, a floodplain constructed by a low sinuosity, sandy, braided (unstable multiple channel) system, is expected to exhibit a totally different fluvial architecture (e.g. Allen, 1970; Brown, 1997).

This is an important consideration when evaluating the evolution of floodplain environments in the British Isles, as there is evidence suggesting many low lying fluvial systems were characterised by fine grained sediments during the earlier Holocene, formed primarily by overbank processes within silty braided channels (Burrin & Scaife, 1984; Brown, 1987b), and not necessarily meandering systems. Many divided channel systems (defined as **anastomosing** river channels, *cf.* Smith, 1983) are evident in the Holocene alluvial record but are rare or absent from Britain today (Burrin, 1985). Anastomosing systems consist of multiple interconnected channels bordered by wetlands and floodbasins (Macklin & Needham, 1992) and were common in many low gradient river valleys in south and east England, particularly during the mid-Holocene period (e.g. Seale, 1979; Burrin & Scaife, 1984; Brown, 1987b; Limbrey & Robinson, 1988; French *et al.*, 1992). Such systems have also been identified in The Netherlands (Törnqvist, 1993a; 1993b) and Belgium (Kiden, 1991). Various authors (e.g. Smith, 1983; Miall, 1992) suggest that low river gradients favoured the development of anastomosing fluvial systems, although clearly other factors are equally important, since not all low gradient valleys are characterised by such multi-channel systems. French *et al.* (1992) studied the Holocene anastomosing fluvial systems of the lower Welland and Nene rivers in Cambridgeshire, and suggest that such systems are a reflection of the interaction between two extrinsic factors. Firstly, the increase in local base levels associated with sea-level rise helped maintain the dominance of vertical aggradation over lateral accretion, and secondly, the influx of clays and silts provided cohesive stable banks, which inhibited lateral movement of the channel, and promoted the development of anastomosing systems (see Nanson & Young, 1981; Brown & Keough, 1992a; 1992b; Törnqvist, 1993b).

Vegetation cover is also an important factor in determining channel style (Miall, 1992). Dense fen carr vegetation on the floodplain (in particular *Alnus* carr) has been suggested as being a contributory factor in promoting bank stability, thus maintaining anastomosing channel systems during the mid-Holocene, in the Nene and Soar river valleys in the East Midlands valleys (Brown & Keough, 1992a, 1992b), the middle to lower Trent valley (Knight & Howard, 1984; Howard *et al.*, 1999), the middle reaches of the Thames at Runnymede Bridge (Needham, 1992), the lower Severn (Brown, 1987b), and the river Gipping in Suffolk (Rose *et al.*, 1980). However, it may be very difficult to distinguish an anastomosed pattern in the sedimentary record from successive single channel positions caused by periodic avulsion (Brown, 1987b; Kiden, 1991).

A recent model of floodplain construction, the stable-bed, aggrading-banks hypothesis (SBAB), applicable to the lower reaches of river systems, has been proposed by Brown & Keough (1992a). This model assumes that mid-Holocene vertical aggradation of the floodplain and channel margin, is related to changing hydrological regime and an increase in fine sediment delivery. These conditions result in flat floodplains with low gradients, and the development of cohesive clayey river banks inhibits lateral channel migration. The theory is supported by stratigraphic evidence for limited reworking of floodplain sediments within an anastomising fluvial regime, suggesting relatively stable channel networks. Accretionary processes continue through overbank sediment diffusion (*sensu* Pizzuto, 1987), as long as there is adequate supply of fine sediment and continued floods or rises in water level. This results in relative channel incision due to increasing disequilibrium between the aggrading floodplain surface and stable river bed, and an enlarged cross sectional area (*cf.* Brown, 1987b). Brown & Keough (1992a) note that the processes that initiate such floodplain metamorphosis (i.e. rise in water tables, and increase in sediment supply) can be explained in terms of both human activity and climate change (which may have been an important secondary factor). The arguments in support of the relative importance of these causal factors are discussed in section 2.3.4 below.

2.3.4 The Climate Versus Human Activity Debate

There are two major schools of thought concerning the causes of Holocene floodplain development in Britain and elsewhere (e.g. the Mediterranean, Vita-Finzi, 1965; 1975; Bintliff, 1975). One proposes that climate is the main control and the other suggests human activity, although some authors have also stressed the importance of local geomorphic conditions in explaining catchment history (e.g. Taylor & Lewin, 1996).

Zuener (1959) formulated a simple model whereby cold or dry periods are the principal cause of aggradation (rivers removed exposed sediments in barren landscapes), and wet and warm phases are associated with phases of erosion (vegetation colonisation hinders denudation, which restricts the sediment supply, causing rivers to incise and erode deposits from the previous dry phase). However, this model could only be applied to arid areas, away from the influence of changing sea level. Dury (1964; 1965; 1970; 1977; Dury *et al.*, 1972) argued that valley sedimentation took place in response to stream shrinkage caused by palaeohydrologic variations, induced by climate change. However, subsequent advances in geomorphology demonstrated the complex (often episodic) manner in which fluvial systems can respond to environmental stimuli, often involving thresholds (Schumm, 1979).

There can be little doubt that climate is the major forcing factor underlying fluvial processes (Knox, 1995) over Quaternary timescales. However, during the Holocene, although the role of climate change may be important, the signal is less clear cut, and it is generally believed that human activity was the main agent for change. Authors advocating anthropogenic activity as a hypothesis for explaining valley alluviation have dominated the literature over the past two decades. Macklin & Needham (1992) attribute this bias to the apparent small magnitude of climate change during the Holocene compared to earlier Quaternary periods. Holocene climate change may be viewed as a contributory factor (Brown, 1988) but is often not considered to have been of a significant magnitude to have affected the fluvial system (see reviews in Bell, 1982; Limbrey, 1983; Burrin & Scaife, 1988). The main physical features of many extant British valley systems are 'antecedent' landforms, 'inherited' from the Pleistocene, which often have considerable influence on later processes of floodplain formation (Brown, 1990; Croke & Nanson, 1991). These morphological characteristics have remained relatively stable since the Pleistocene, giving the impression that climate has been largely ineffective in reshaping the valley during the Holocene. Macklin & Needham (1992) state that this has led to the assumption that fluvial systems have played a rather passive role in Postglacial alluviation in response to climate, and that significant changes in the nature and rate of sedimentation coincided with, and were therefore to be attributed to, the expansion of human population during the Holocene.

The impact of prehistoric culture is seen as an important agent in both the cause and exacerbation of soil erosion from the valley sides, and subsequent inwash onto the valley bottom (Brown, 1982). Hence, the timing and magnitude of river sedimentation is often related to the scale, type and history of catchment disturbance (Burrin & Scaife, 1984; Brown & Barber, 1985; Robinson & Lambrick, 1984). Brown & Barber (1985) have stressed the importance of catchment characteristics such as slope angle and soil type in influencing sediment yield and the timing of floodplain deposition.

The general principle of anthropogenically driven floodplain alluviation is simple. The clearance of dryland woodland is likely to have caused a reduction in evapotranspiration rates, interception, and infiltration, and an increase in surface run-off, which in turn gave rise to higher groundwater tables and increased waterlogging of valley bottoms (Moore & Willmot, 1976). The consequence of this may be enhanced valley floor peat formation, which has implications for the correct interpretation of peat accumulation within valley fills, (stability or instability?, Scaife & Burrin, 1992). The increase in run-off and stream flow is also likely to result in erratic flow rates and the increased likelihood of flooding (Limbrey, 1978). The process of soil erosion is also facilitated by the loss of soil stability associated with woodland destruction (Moore, 1985), which is also likely to lead to phases of minerogenic alluviation on floodplains.

Floodplain studies by Scaife & Burrin (1983) and Burrin & Scaife (1984) stress the importance of woodland clearance by Mesolithic populations in initiating soil erosion. This was cited as the reason for increased colluvial inwash and sediment supply to valley bottoms, inducing alluviation at Sharpsbridge, in the Sussex Ouse. They suggested that floodplain sedimentation occurred periodically, in relation to intense phases of anthropogenic activity, interspersed with erosion phases. Similarly, Scaife & Burrin (1985) consider that anthropogenic activity during the Neolithic period played an important role in the development of the alluvial fill at Chiddingly in the adjacent Cuckmere valley. They suggest that the bulk of the valley aggradation of the Sussex Ouse had taken place by the Bronze Age, and that there has been little floodplain development since (Burrin & Scaife, 1984). Scaife & Burrin (1987) and Burrin (1988) investigated additional sites from the eastern Rother valley and were able to establish that prehistoric human activity was widespread elsewhere in the High Weald, with erosion and floodplain alluviation prominent from the Neolithic period (Burrin, 1985). Needham & Macklin (1992) also state that alluviation had begun in most river valleys by *c*.4800 BP (*c*.5700 to 5300 cal. yrs BP).

Other authors have identified similar phases of alluviation, but suggest that this sedimentation occurred later, during the late-Bronze Age, Iron Age and Roman periods. These phases of floodplain construction, dominated by minerogenic sediments, were also attributed to anthropogenic valley side clearance (e.g. Limbrey, 1978; Shotton, 1978; Brown, 1983), and inferences were made as to the nature of the human activities, such as expanding and changing agricultural practices (Burrin & Scaife, 1984; Buckland & Sadler, 1985). The main reasons cited in most cases for the acceleration in alluviation during the mid- to late-Holocene period are changing land-use due to an increase in agricultural utilisation of catchments, which led to the increasing occurrence of floods, and erosive storms (Brown, 1997).

Brown (1987) and Barber & Twigger (1987) note that many sites in the Severn basin experience increased alluviation during the late-Bronze Age/early-Iron Age period. In the Severn-Avon valley, Shotton (1978) also suggests increased sedimentation between *c*.3100 and *c*.2600 BP (*c*.3500 to *c*.2400 cal. yrs BP). This is attributed to a change in agricultural practice (e.g. ploughing), which resulted in soil erosion, following widespread deforestation, and was further substantiated by Brown (1983). Brown & Barber (1985) cite a shift from pastoral to arable agriculture as the reason for the apparent acceleration in valley alluviation between *c*.2900 and *c*.2300 BP (*c*.3300 to *c*.2100 cal. yrs BP) in the Ripple Brook in the Upper Severn. In a study of the alluvial history of the Severn around Welshpool, Taylor & Lewin (1996) attributed the deposition of a silty alluvial unit (dated to 2850±60 BP, between 3148 and 2791 cal. yrs BP) as, in part, a response to the human occupancy of the surrounding landscape. Brown *et al.* (1994) state that in the East Midlands, all sites studied in the Soar and Nene valleys were undergoing alluviation by *c*.2500 BP (*c*.2800 to *c*.2300 cal. yrs BP).

By the Iron Age there are further reports of valley sedimentation. In the Thames tributary valleys (e.g. Windrush), Hazleden & Jarvis (1979) interpret fine grained floodplain deposits as a result of a rapid increase in local forest clearance and ploughed land. A date of 2660±85 BP (2938 to 22496 cal. yrs BP) provides a *terminus post quem* for this phase of alluviation. Robinson (1992) and Lambrick (1992) confirmed the earlier work by Robinson & Lambrick (1984) that alluviation, due to a rise in water table and increased flooding in the Upper Thames valley, was particularly severe during the late Iron Age (*c*.2400 BP, between *c*.2700 and *c*.2200 cal. yrs BP), and continued into the Roman period. Lambrick (1992) establishes links between the palaeoecological record of vegetation disturbance from the Oxford region (Day, 1991), and the alluviation phase, which is attributed to human interference with the hydrological cycle through deforestation, rather than a climatic cause. A combined hypothesis of anthropogenic activity and estuarine activity has been suggested as the causal mechanism in floodplain evolution in the Combe Haven valley in East Sussex (Smyth, 1986; Jennings & Smyth, 1987; Smyth & Jennings, 1988; 1990). Iron Age forest clearance, which released colluvial material enabling floodplain development, has been held responsible for increased river discharge (reduced evapotranspiration and increased run-off) and a subsequent marine incursion in to the lower course of the Combe Haven valley (Jennings & Smyth, 1987).

Roman occupation has also been invoked as a cause of increased alluviation. Samuels & Buckland (1978) and Buckland & Sadler (1985) suggest changing agricultural practices, during the late-Roman period are a major factor in the enhanced erosion of soil and deposition of oxidised alluvium which buried the Roman site at Sandtoft on the river Idle floodplain in the Humberhead Levels. Increased minerogenic sedimentation has also been noted at this time from elsewhere in the Humber region; at Brough-on-Humber (Wacher, 1969), North Ferriby (Buckland *et al.*, 1990), the river Trent (Salisbury *et al.*, 1984; Riley *et al.*, 1995), and from other areas such as the Sussex Ouse (Burrin & Scaife, 1984), and Fengate in East Anglia (Pryor, 1984); (although a radiocarbon chronology is lacking in most cases). Robinson & Lambrick (1984) associate increased alluviation with the development of Roman villas and extensive arable field systems in the Cotswolds.

Reviews by Bell (1982), Limbrey (1983), and Burrin & Scaife (1988) all consider that the diachronous nature of floodplain alluviation excludes any climatic control. They concluded that climate change during the Holocene was not of a sufficient magnitude to have significantly influenced fluvial sedimentation. However, recent work has begun to suggest otherwise. Macklin *et al.* (1992a) explain the difference between the nature of early-, mid- and late-Holocene valley alluviation in the Tyne basin as a reflection of the variation in sediment supply (i.e. increased erosion and delivery of sediment to valley floors following agricultural intensification from the Neolithic onwards). However, they suggest that the redistribution of this material occurred during abrupt climate shifts characterised by major changes in flood frequency and magnitude. Similarly, Passmore *et al.* (1992) suggest that variations in the rate and nature of sedimentation in the lower reaches of the Tyne coincide with rising sea level and periods of catchment change during the early- to mid-Holocene, but that changes to the hydrological regime during the late-Holocene are attributable to climatic deterioration (i.e. the Little Ice Age). Macklin & Lewin (1993) acknowledge that anthropogenic ecosystem disturbances (e.g. agricultural practice and forest clearance) are important in initiating soil erosion, which is the obvious precursor to accelerated floodplain alluviation, but believe that the rates of sediment delivery to rivers, and the timing of valley floor sedimentation, are related to climatic controls characterised by major changes in the magnitude and frequency of storms and floods. Thus the combination of human and climatic factors may be important, whereby human activities create the environmental instability, and climatic events are responsible for the transportation and redistribution of the sediment (Macklin & Needham, 1992; Macklin & Lewin, 1993).

Macklin & Needham (1992) and Macklin & Lewin (1993) review all relevant literature concerning Holocene river activity (archaeological, geomorphological, and palaeoecological) and note significant regional and national patterns in sedimentation that apparently occur synchronously, which may suggest climatic control. Macklin & Lewin (1993) present a frequency histogram showing the timing and duration of significant phases of Holocene river activity, which they use to evaluate the links between factors such as climate, vegetation and land-use, and periods of alluviation. Three phases of river activity are identified, which they consider to be relatively synchronous (Macklin & Lewin, 1993). The period between *c*.8000 and *c*.5000 BP (*c*.9200 and *c*.5500 cal. yrs BP) was characterised by channel stability or incision and a lack of alluviation. A significant change in fluvial activity is marked at *c*.5000 BP (*c*.6000 to *c*.5500 cal. yrs BP) with alluviation recorded in most valley floors by *c*.4800 BP (*c*.5700 to *c*.5300 cal. yrs BP), but this phase slowed or had ceased by *c*.4200 BP (*c*.5000 to *c*.4400 cal. yrs BP). Between *c*.3800 and *c*.3300 BP (*c*.4400 and *c*.3300 cal. yrs BP) deposition of gravel and later fine grained alluvium was widespread. Major alluviation phases also occurred between *c*.2800 and *c*.2400 BP (*c*.3200 and *c*.2200 cal. yrs BP) in southern and parts of northern Britain, but the most important phases of river activity took place between *c*.2000 and *c*.1600 BP (*c*.2300 and *c*.1300 cal. yrs BP) and *c*.1200 to *c*.800 BP (*c*.1300 to *c*.600 cal. yrs BP). Work by Brown (1988; 1990) and Macklin *et al.* (1991; 1992a) on the pattern of alluviation and river channel change also provides evidence in support of these phases of sedimentation.

The frequency of dated British Holocene alluvial units and phases of Postglacial river activity identified by Macklin & Lewin (1993) correspond well with the timing of climatic discontinuities identified by Wendland & Bryson (1974) in the USA, and with marked periods of hydroclimatological change in Britain (Macklin *et al.*, 1992a). Furthermore, there appear to be international correlations, with simultaneous climate changes occurring in Scandinavia (Karlen, 1991), and similar alluvial discontinuities in central and northern Europe (Becker & Schirmer, 1977; Starkel, 1991a; 1991b) and in the USA (Knox, 1983). Clear parallels in river history between countries and continents led Macklin & Lewin (1983) to suggest that the widespread synchrony of fluvial events is a reflection of climatic control. Certainly, the short history of land disturbance in the USA is more likely to imply that correlations there are driven by climate (Knox, 1995).

Macklin *et al.* (1992a) suggest that anthropogenic and climate causes are not to be viewed as competing hypotheses, but a continuum between two factors, each of which can exhibit a control on alluviation. Furthermore, the broad synchrony in Holocene fluvial episodes leads Macklin & Lewin (1993) to suggest that climate was the principal driving mechanism, with the degree and nature of the fluvial response varying spatially because of catchment physiography and anthropogenic manipulations of vegetation. They conclude that the British fluvial record is one that has been "climatically driven but culturally blurred" (Macklin & Lewin, 1993, p 119). Tipping *et al.* (1999) consider this compromise an adequate generalisation but question whether human activity is always of secondary importance to climate.

A closer and more critical inspection of the evidence by Brown (1997) reveals that the apparent synchroneity of these fluvial events is not as clear cut as Macklin & Lewin (1993) suggest. For example, major climatological and hydrological change occurred in northern Europe during the Bronze Age/Iron Age transition (van Geel *et al.*, 1996), between *c*.2600 and *c*.2300 BP (*c*.2900 and *c*.2100 cal. yrs BP) which is not recorded in North America. In North America the nearest climatically induced period of fluvial activity identified by Knox (1983) occurs between *c.*2000 and *c.*1800 BP (*c.*2300 and *c.*1500 cal. yrs BP). Such disparities are perhaps inevitable when attempts are made to relate regionally continuous phases of alluviation to global climatic discontinuities. However, there are also methodological problems, since some of the synchronous phases of fluvial activity may be a product of the clustering of radiocarbon dates during phases of ^{14}C overproduction (*c.f.* Bartlein *et al.*, 1995; Van Geel *et al.*, 1996), which may give rise to false patterns and correlations. This is more pertinent to some Holocene periods than others (e.g. Bronze Age/Iron Age transition, *cf.* Pilcher, 1991). Coincidental and spurious correlations may also be attributable to the 'suck in' and 'smear' effects described by Baillie (1991).

Brown (1997) notes that considering the number and duration of the many climatic and alluvial discontinuities, it may be statistically improbable for such events not to overlap in some cases. Significant lags may also exist in the response of the fluvial system which further complicate any attempt at chronological correlation. The chronological model Macklin & Lewin (1993) propose is also open to criticism because, as they admit, it only marks accretionary events and not periods of incision or erosion, which appear as depositional hiatuses in the stratigraphic record.

Synchroneity remains the major test for climatic causation, but phases of alluviation in the British fluvial record are seldom dated by more than one control (Tipping *et al.*, 1999). Brown (1997) suggests that the dating control used by Wendland & Bryson (1974) and adopted by Macklin & Lewin (1993) for comparison is not truly independent of climate change. The basis of their climatic discontinuities come from both radiocarbon dates from peats within alluvial units and 'cultural' dates, neither of which are independent or direct evidence of climate change (Brown, 1997). A good source of data for testing the anthropogenic/climate hypotheses (see Macklin *et al.*, 1992a; Tipping *et al.*, 1999) is changing peat humification recorded in raised peat bogs, which are dependent upon rainfall and sensitive to direct changes in climate (Barber, 1981; 1982). Barber *et al.* (1994) develop a bog wetness curve (based on the humification record) for the last 2000 years from a bog in Cumbria which has been transformed into a proxy climate record using general linear modelling, which is highly correlated with known periods of climate change.

Despite these dating problems, parallels between the Holocene alluvial record of Britain and North America do seem to exist, which Macklin & Lewin (1993) suggest reflect large scale (hemispheric) atmospheric circulation patterns exerting a climatic control. Further support for the influence of continental teleconnections has since been provided by Van Geel *et al.* (1996). There does seem to be widespread agreement of a relative lack of alluviation and channel stability common to both Europe and North America between *c.*8000 and *c.*5000 BP (*c.*9200 and *c.*5500 cal. yrs BP), and to a certain extent between *c.*5000 and *c.*4500 BP (*c.*5500 and *c.*4900 cal. yrs BP, Knox, 1983; Starkel, 1991a; 1991b; Macklin & Lewin, 1993). However, there are always dangers inherent in making such correlations, given that temporal coincidence does not necessarily imply causality. Correlation of fluvial events with climatic shifts on this basis alone must be viewed with caution (Brown, 1997).

Bell (1992) notes that a clear dichotomy has emerged between those who favour climate or human activity as the responsible agency for fluvial change. This bias is partly due to personal preference but is also perhaps a product of the fact that many studies take place in an archaeological context (Bell, 1992), in which case any environmental disturbances are often simply attributed to the effects of land-use and anthropogenic disturbance, rightly or wrongly. This debate has not stagnated progress in research, but rather stimulated further analysis. Knox (1995) favours the climate change hypothesis but acknowledges that climatic variables contribute strongly to a plethora of other multiple explanatory factors that influence fluvial change, such as vegetation, tectonic and eustatic base-level, anthropogenic factors, intrinsic thresholds, and geomorphic stability (*cf.* Patten & Schumm, 1981). However Gregory (1995) and Tipping *et al.* (1999) still regard human activity as fundamental in interpreting the cause of change in fluvial systems.

2.3.5 Conclusions

It is obvious that there is not a simple relationship between climate, anthropogenic activity, hydrological regime, and processes of floodplain construction (Burrin (1985). In reality climate and human activity are variables which are not truly independent of one another, because climatic events (such as non-random decadal scale fluctuations in rainfall, Macklin *et al.*, 1992b) are responsible for the transport and redistribution of sediment (Macklin & Lewin, 1993), which may or may not have been mobilised by human activity (Brown, 1997). Therefore it is difficult to draw firm conclusions concerning the relation between climatic and cultural controls on Holocene fluvial activity. Brown & Quine (1999) suggest that the key to which cause dominates is controlled by their relative magnitude and state of the system.

Floodplain deposits are a product of spatially variable interactions between people and physical factors, such as climate change, ecological dynamics, and soil development (Brown, 1997), which operate over different spatial and temporal scales. Climate is seen as being an unequivocal forcing factor influencing floodplain development over long timescales (that span different glacial periods), but over shorter timescales many other factors are important. Scale is an equally important consideration in this respect . As Schumm & Lichty (1965) note, the status of the variables in process-response systems (i.e. the river system) varies with the timescale of the operation.

Shennan (1992) explains how the sentiments associated with Schumm & Lichty's model are also applicable to studies of coastal zone evolution. The coastal system is dependent on the interaction of global scale factors such as glacial/interglacial climate change, regional scale parameters such as glacio-isostasy, and local scale effects including coastal morphology, sediment type and supply, tidal regime, meteorological conditions, river discharge, and vegetation dynamics. The inter-relationship of sedimentation between river and coastal systems is central to the theories of floodplain evolution put forward by Smyth (1986), Jennings & Smyth (1987), and Smyth & Jennings (1988; 1990), which are testimony to the complex nature of the perimarine system, and the need for an holistic, multidisciplinary approach to reconstructing such environments.

2.4 Holocene Sea-Level Change

2.4.1 Introduction

The river and floodplain ecosystem can be regarded as a continuum from headwaters within upland reaches of the catchment downstream to the sea (Brown, 1997). The perimarine zone therefore constitutes an important link between the fluvial and coastal environment. As highlighted by Ashley & Renwick (1983), the influence of sea level extends from the marine realm to the head of the tide, which can be situated many kilometres inland within the fluvial system (Pizzuto & Rogers, 1992). In the Humber Estuary, the tidal length is approximately 140 km on the River Ouse (Pethick, 1990), but this is an artificial limit set by locks and weirs, so the natural extent of the tide would be even longer, and the effects of sea-level change will be felt at least this far inland. Therefore, a record of Holocene sea-level change is preserved not only in coastal sedimentary environments, but also in freshwater and brackish tidal rivers within the inner estuary, where increases and decreases in marine influence are reflected by changes in vegetation community (Godwin & Godwin, 1933a).

The importance of sea-level change as an allogenic control on vegetation and sedimentation patterns in the perimarine zone is clear (Behre, 1986). The following sections address the methodological aspects of Holocene sea-level investigations, and highlight some of the complications associated with reconstructing sea-level change from sediments in the perimarine zone.

2.4.2 Operational Definitions in Sea-Level Studies

The vertical extent of a tidal oscillation (**tidal range**) ranges between the lowest astronomical tide (LAT) which occurs within the subtidal zone, to the highest astronomical tide (HAT) which is situated within the supratidal zone, which is inundated only occasionally (see Table 2.1).

Mean tide level (MTL) is calculated from the mid-point of the tidal range, and may vary from mean sea-level (MSL) if the tide is asymmetric. Because most tides in the world are asymmetrical, due to shallow water distortion of the tidal wave, MSL actually occurs slightly above or below MTL. Jardine (1986) has pointed out that sea-level curves that are constructed using MSL are actually based on MTL. MSL for the UK is referenced to the tide gauge at Newlyn, Cornwall, and is based on the average level of the sea recorded between 1915 and 1921 (Jardine, 1986). For any given location, a tide level can be calculated, ranging from the LAT, to the HAT, using Newlyn as the national reference.

		Supratidal Zone
Highest Astronomical Tide	HAT	
Mean High Water Spring Tide	MHWST	
Mean High Water Neap Tide	MHWNT	
Mean Sea Level	MSL	Intertidal Zone
Mean Tide Level	MTL	
Mean Low Water Neap Tide	MLWNT	
Mean High Water Spring Tide	MLWST	
Lowest Astronomical Tide	LAT	
		Subtidal Zone

Table 2.1 Tide levels within the intertidal zone (adapted from Jardine, 1986)

The transition between fresh-terrestrial and brackish-marine (saltmarsh) environments occurs at approximately the level of MHWST. Therefore, in terms of analysis of past sea-level change, MHWST is ecologically the most important tide level because, unlike MSL indicators, microfossil assemblages occur in and around MHWST that leave a long term record in sediments that allows this level to be identified in the stratigraphic record. This is the most frequently adopted sea-level indicator used to reconstruct past tide levels (Tooley, 1978a; Kidson & Heyworth, 1979; Kidson, 1986; Long, 1992; Shennan, 1992; Zong, 1992; 1997; 1998; Zong & Tooley, 1996; Zong & Horton, 1998; 1999).

2.4.3 Terminology and Methodology of Holocene Sea-Level Studies

In the last few decades, advances have been made to avoid the ambiguous usage of the various terms used to describe sea-level changes, to enable effective international comparison of data between researchers. For instance, the terms 'transgression' and 'regression' have long been synonymous with sea-level studies, but their inconsistent use, in different contexts, has led to confusion (Shennan, 1980; 1982; 1983a; Tooley, 1982; 1985c). Shennan (1980; 1982) and Tooley (1982) develop strict operational definitions that are now been widely adopted and applied in the recommended manner (e.g. Shennan, 1986a; 1986b; Long, 1992; Plater & Shennan, 1992; Long & Innes, 1993; 1995). These are outlined below:

Transgression and **regression** are descriptive terms that are used to refer to a landward or seaward migration of the coastline, due to any combination of factors. They should not be used as formal litho- or biostratigraphic units.

Transgressive and **regressive contacts** are lithostratigraphic descriptive phrases that are used to define the boundaries between marine-brackish and freshwater-terrestrial sediments at a single location. A transgressive contact describes a change in lithology within a core from semi-terrestrial to a marine deposit, and a regressive contact refers to the replacement of marine conditions by semi-terrestrial sediments in a core. It is not appropriate to use these terms when sedimentary boundaries are disturbed or eroded. Although such terms are most often used in conjunction with lithological expressions, a biostratigraphic transgressive or regressive contact may be located from microfossil analyses. When such a change is observed between a series of cores or along a section, the above terms are referred to as **transgressive** and **regressive overlaps**.

The terms transgressive contact and overlap only describe a process, which may be associated with a variety of factors. Their use does not imply any vertical change in sea level (which it may indeed reflect, but this must be supported with additional evidence). For example, a rising sea level will not cause a retreat in coastline if the rate of sedimentation is greater than the rate of sea-level rise (Wilks, 1979). Similarly, shoreline advance may occur during periods when the rate of sea-level rise is accelerating, if the rate of terrigenous sediment supply (i.e. peat accumulation) exceeds the rate of sea-level rise (see examples in Shennan *et al.*, 1983; Gerrard *et al.*, 1984).

The **tendency of sea-level movement** refers to the nature of the change in sea-level which may be inferred from a single sea-level indicator. It describes whether the point records an increase or decrease in water level or salinity (Shennan, 1983b). A tendency is interpreted as a reflection of the movement of marine water towards or away from a site, but such movements do not necessarily equate with rises or falls in relative sea level (Nelson *et al.*, 1996). A **positive sea-level tendency** reflects an apparent increase in marine influence and a **negative tendency** is the apparent decrease in proximity of marine conditions. Tendencies of sea level are usually manifest as lithological or vegetational stratigraphic changes, therefore transgressive contacts are commonly interpreted as indicating a positive tendency and regressive contacts as periods of negative tendency. However, a study of biostratigraphic fossil assemblages is required to confirm such lithostratigraphic interpretations.

Changes in microfossil assemblages within a homogeneous lithostratigraphic unit may also mark a tendency. For example, changes in diatom zones indicative of a shift from MHWST to HAT within saltmarsh sediments provides evidence for a negative sea-level tendency (Zong, 1997; 1998). This approach, in addition to the traditional use of transgressive and regressive contacts, can provide additional information concerning sea-level changes (Zong, 1997; 1998). A positive tendency may be first recorded several centimetres below a transgressive contact, possibly in biostratigraphic evidence, or a gradual increase in salinity and/or water depth (Long, 1992; Nelson *et al.*, 1996). A freshening of the depositional environment within a minerogenic unit (Jordan, 1987) or within a clastic unit below a regressive contact may reveal the onset of a negative tendency which may then continue across the contact into the overlying unit (Smith *et al.*, 1981; Shennan *et al.*, 1995a). The biostratigraphic record of positive or negative tendencies within and/or across boundaries of dated units allows rates and the direction of changes in water depth and salinity to be estimated (Nelson *et al.*, 1996).

The correlation of tendencies from a number of sites in a region through radiocarbon dating of index points can help establish the **dominant tendency** during particular time periods (Shennan *et al.*, 1983; Shennan, 1986a; 1986b; Long, 1992). The consistent application of this methodology can then be used to establish the importance of local and regional factors in determining the relative sea level chronology of an area (Shennan, 1983b; 1992; Shennan *et al.*, 1983).

2.4.4 Determination of Former Sea Levels

Sea-level investigations rely on the collection and analysis of **Sea-Level Index Points** (SLIPs). A sea-level indicator is a sample which may be used to indicate a past sea level. However, a sea-level indicator only becomes a SLIP when certain criteria are met, and only then can the sample be used to establish a fix on a past sea level. A SLIP, must comprise four components:

- A known geographical location
- A known age
- A known altitude that can be related to a former sea level (i.e. indicative meaning and range)
- A known sea-level tendency

Various lithostratigraphical and biostratigraphical techniques can be employed as sea-level indicators. Detailed stratigraphic descriptions of transgressive and regressive overlaps can enable basic mapping of coastline palaeogeography (Shennan, 1994; Waller, 1994d), and the analysis of microfossils (such as diatoms, foraminifera, and pollen) provide the precise environmental information necessary for the accurate identification of SLIPs. For example, the proximity of marine conditions is often indicated by the occurrence of saltmarsh associated pollen types within an assemblage, such as Chenopodiaceae, Poaceae, *Plantago maritima*, *Glaux maritima*, *Armeria maritima*, *Limonium*, *Artemisia*-type and certain members of the Asteraceae family. The presence of a combination of these pollen types is characteristic of transgressive and regressive contacts.

To enable the construction of a sea-level graph showing age/altitude information, the SLIP must have an indicative meaning. **The indicative meaning** of an index point describes the altitudinal relationship between the local depositional environment in which the sample accumulated and a contemporaneous **reference water level** (Shennan, 1982; 1986b; Shennan *et al.*, 1994; van de Plassche, 1986; Horton, 1997). The reference tide level assigned to transgressive and regressive contacts is usually taken to approximate MHWST (Tooley, 1978a), although see Godwin (1940), Heyworth & Kidson (1982), and

Shennan (1986b) for alternatives. For example, Godwin (1940) suggests that the transition from reedswamp to fen wood peat will occur at MHWST in a coastal fen, but in a backwater area, local groundwater may be the controlling factor which may be around MTL (Shennan, 1986b).

In addition to this uncertainty, the reference tide level may not be constant, and therefore an **indicative range** must be incorporated into any interpretation. This accounts for the uncertainty regarding the accuracy of the tide level to which the index point is referenced. Errors are most likely on coasts with large tidal ranges, or where there are significant spatial variations in tidal level, for example in estuaries (Kidson, 1982; 1986; Shennan, 1986b). Shennan (1986b) suggests that the indicative range can be reduced by dating the level at which microfossil and stratigraphic evidence exhibits a change in sedimentary environment. Therefore, the indicative meaning can vary according to the type of stratigraphic evidence, and is expressed in terms of an indicative range (typically, vertical ranges for transgressive/regressive contacts are *c.*0.20m (Shennan, 1982), and a reference water level to which the assemblage is assigned (e.g. MHWST, MTL, HAT, etc., van de Plassche, 1986).

Shennan (1982; 1986b) determines the reference water levels and indicative ranges for a variety depositional contexts and suggests some correction factors. For example, where a *Phragmites* peat occurs directly above a fen wood deposit, the reference water level is assigned to MHWST-10 cm with a vertical indicative range of 20 cm. The accurate determination of the indicative meaning of coastal samples is hindered by the lack of information on the contemporary relationships between tide levels and the succession of coastal plant communities which lead to the formation of transgressive or regressive overlaps of peat and clay beds (Tooley, 1978a). The indicative meaning of indicators within a saltmarsh sequence are also complicated by ecological and habitat processes (Allen, 1990a; 1990b). However, recent work on contemporary coastal zonation of diatoms, foraminifera, testate amoebae, and pollen will enable a more precise quantification of the indicative meaning of such samples (Gehrels, 1994; Shennan *et al.*, 1995b; 1996; 1998; Haslett *et al.*, 1997; Charman *et al.*, 1998; Horton, 1997; Horton *et al.*, 1999a; 1999b; in press; Zong and Horton, 1998, 1999).

SLIPs can potentially be produced from a wide variety of palaeoenvironmental data including archaeological evidence (Louwe Kooijmans, 1980; Flemming & Webb, 1986; Devoy, 1990; Fulford *et* *al.*, 1997) and morphological features such as raised beaches (Sissons & Dawson, 1981; Smith & Dawson, 1983; Firth, 1989) and marine notches (Pirazzoli, 1986). By far the most common type of sea-level index point used in the analysis of sea-level change in North-west Europe (and are used in this study) are transgressive and regressive contacts, identified from bio- and lithostratigraphic changes in boreholes, marking distinct environmental and lithological changes between fresh terrestrial and marine littoral deposits (Tooley, 1978a; Devoy, 1979; 1982; Shennan, 1986a; 1986b; Long, 1992; Long & Innes, 1993; 1995; Zong & Tooley, 1996).

2.4.5 Controls on Sea-Level

Regional sea-level changes at a site are a product of the interplay between various long- and short-term controls on coastal evolution (Long & Roberts, 1997). The long-term controls identified by Fairbridge (1961) can include oceanic (or 'eustatic') variables (i,ii,iii below) which control the global volume of water in the oceans and crustal variables (iv,v below) and are defined below following Mörner (1980; 1987):

i) **Glacio-eustasy**. This includes changes in ocean volume caused by the build-up and decay of land-based ice masses.

ii) **Geoidal-eustasy**. This is the change in distribution of ocean water due to differential gravitational attraction over the surface of the earth associated with planetary forces, and the uneven distribution of materials of different density within the crust and mantle of the earth.

iii) **Tectono-eustasy**. This is sea-level change as a result of long-term changes in the size of ocean basin size associated with ocean ridge spreading and tectonic uplift.

iv) **Glacio-isostasy**. This is associated with crustal movements due to ice loading and unloading.

v) **Hydro-isostasy**. This is sea-level change associated with crustal movements due to water loading and unloading effects.

Shorter term local to regional changes in sea-level are caused by a combination of oceanographic, meteorological, hydrological, and sedimentological factors such as changes in tidal regime, ocean temperature and currents, air pressure and wind, river discharge, sediment supply, and vegetation patterns (Tooley, 1985b; Long & Roberts, 1997). Changes in weather such as air pressure can affect the geodetic sea level by up to 2 m (Mörner, 1981) and exceptional storm conditions can raise sea water levels by over 3 m (Rossiter, 1962; Kidson, 1982).

Recognition of these controlling factors, which are not only spatially and temporally variable but difficult to quantify accurately, lead to objections to a single globally valid sea-level curve (Tooley, 1985c). The concept of 'global eustasy' has been questioned and discredited by Mörner (1976), who drew attention to geoidal deformation. He subsequently redefined eustasy as ocean-level changes, irrespective of cause, which are determined by climate, earth movements, and gravity (Mörner, 1980). The **geoid** can be defined as the surface of the constant geopotential (combined potential of the gravitational attraction and rotational acceleration of the earth) which, over the seas, coincides with MSL (Mörner, 1976). The geodetic sea-level or equipotential surface of the geoid has been shown to be unstable in both space and time, and possesses an irregular configuration, with swells and depressions with a maximum calculated range of 180 m at any one time (Mörner, 1976).

2.4.6 Interpretation of Former Sea-Levels

The Coastal Zone

Two established methods of studying former sea-levels are time/altitude analysis and tendency graphs. The first method provides information on the former altitude of sea-level (e.g. Godwin, 1940; Fairbridge, 1961; Tooley, 1978a; 1978b; Devoy, 1979; 1982) and the SLIPs used must possess all four attributes listed in section 2.4.4. Tendency analysis provides information on the timing of increases or decreases in marine influence, and is based solely on the age and tendency of the index points (Morrison, 1976; Shennan *et al.*, 1983; Long 1992).

The data used for the construction of sea-level curves in north-western Europe are similar but their interpretations differ (Tooley, 1978b). Many stratigraphic sequences record alternating marine inorganic and semi-terrestrial organic deposits but the extent to which the changes between these types of deposit represent distinct oscillations in sea-level is debatable. The construction of time/altitude diagrams, showing a single line sea-level curve, resulted in polarisation and two opposing views as to the nature of sea-level recovery during the Postglacial period. Some researchers accepted the ideas of Fairbridge (1961), who proposed that Holocene sea-level rise was characterised by a series of transgressions separated by regressive phases which could be linked with fluctuations in climate (Mörner, 1969, Ters, 1973; Tooley, 1974; 1976; 1978a; 1978b; 1979; Devoy, 1979). While they accept that some of the alternations between peat and marine clays may be explained by local site morphological characteristics, such as the breaching of coastal barriers, they interpret the intercalating layers of sediments as reflecting actual changes in sea level.

Other researchers suggest sea-level recovery took the form of a smooth exponentially rising curve (Jelgersma, 1961; 1966; Shepard, 1963; Hawkins, 1971; 1972; Kidson & Heyworth, 1973; 1976; 1978; 1979; Kidson, 1977). They interpret alternations in sediment stratigraphy in terms of local geomorphic changes and/or consequences of differential rates of sea-level rise, sedimentation and isostatic rebound (see Kidson, 1977; Heyworth, 1978). For instance, Godwin (1956) interpreted the thicker marine clays found on the coastal fringes of the Somerset Levels in terms of a renewed transgression in the late-Holocene (Romano-British transgression). However, Kidson & Heyworth (1978) suggest these deposits were explained by the cumulative effects of occasional high sea levels caused by increased frequency of large storm surges and long periods of high tides. Beckett & Hibbert (1979) suggest, that such explanations are too generalised.

Jelgersma (1966) rejects oscillating sea-level curves on the basis of inadequacy and insufficiency of the data. Tooley (1978a) acknowledges the presence of minor oscillations in his North-west England sea-level curve, but proposes that these low amplitude fluctuations reflected local factors such as sediment compaction, and that high amplitude oscillations were primarily eustatic in origin. There is now a growing body of evidence from subsequent coastal investigations in the Mersey Basin of the existence of former sand dunes, forming a discontinuous coastal barrier during the mid-Holocene, behind which many of the deposits dated by Tooley (1974, 1976, 1978a, 1978b, 1980) were laid down (Plater *et al.*, 1999). The original findings of Tooley, whilst remaining the definitive work for the area, have been revised in the light of this evidence, and the evidence for an oscillating sea-level is equivocal (Huddart, 1992; Plater *et al.*, 1999).

Recent theoretical and methodological advances in sea-level studies have largely dispelled the earlier debate on smooth or oscillating sea-levels (Devoy, 1982; Heyworth & Kidson, 1982; Shennan, 1982; Tooley, 1982). Little progress was made as this debate ensued, and the controversy stagnated sea-level research. Shennan (1982) believed the way forward was to develop a standard methodology for litho- and biostratigraphical analysis of sea-level data and produced a flow diagram of research methods based upon continual assessment of errors (Shennan, 1983b). It was realised that significant errors were incorporated in the calculation of vertical and horizontal components of SLIPs, which meant that sea-level curves based on a single line gave a misleading impression of precision (Kidson, 1982, Shennan, 1982, Heyworth & Kidson, 1982). Shennan (1982) has estimated the magnitude of errors affecting the measured altitude of stratigraphic boundaries based on data from the Fenland. Altitudinal errors included the identification and measurement of the height of the stratigraphic boundary; levelling to the benchmark and its accuracy to OD, and non-vertical sampling. Long (1992) has assumed vertical errors of ± 1 m for each data point, designed to approximate altitudinal errors caused by post-depositional compaction, as well as errors involved in sampling and levelling. In addition, the age estimates of SLIPs have errors associated with them, such as the calibrated ^{14}C age standard deviations error, counting error set by the laboratory, and non-counting errors such as contamination.

Other errors concern the indicative meaning of the sample due to a lack of information concerning the uncertainty surrounding the relationship between sea level and contemporary plant communities (Tooley, 1978a). Modern data are lacking due to the effects of drainage, construction of sea-defences, and industrial development of saltmarsh areas. Extensive alteration of coastal areas has largely destroyed the saltmarsh to fen transitional environment (although see Ranwell, 1974). The assumption that tidal range has not changed through time is a further simplification.

An appreciation of these uncertainties resulted in the plotting of error boxes (Kidson & Heyworth, 1979; Streif, 1979; Devoy, 1982; Shennan, 1982; Tooley, 1982) or ellipses (Heyworth & Kidson, 1982). This permits the identification of only a broad sea-level band and it is therefore not possible to elucidate minor or individual fluctuations in sea level. Shennan *et al.* (1983) note it is difficult to evaluate the accuracy of chronological correlation schemes based on a comparison of sea-level curves from a broad sea-level band. For example, two SLIPs, one transgressive and the other regressive from the same borehole, separated by

20 cm of marine clay, will have error boxes that greatly overlap. When such SLIPs are plotted on a graph in a band, they will be hidden within the general trend and will not indicate their differing tendencies.

An alternative approach is tendency analysis which is based on direct investigation of the timing of changes in marine influence (Morrison, 1976; Shennan, 1983b). Shennan (1980) suggests the construction of local sea-level chronologies based on the tendency concept to overcome the problems associated with time/altitude analysis. This method utilises only the age and tendency of index points, ignoring altitude altogether (Roeleveld, 1974; Griede, 1978). Tendencies provide an objective framework for the correlation of regional sea-level chronologies between rising and subsiding areas (Shennan, 1982; 1986b; Tooley, 1982; Long, 1992) and also allow the application of statistical techniques to sea-level studies (Shennan *et al.*, 1983). The method also relies on the consistent definition of the terms transgression and regression. The relationship between transgressive and regressive overlaps and former sea-levels has been questioned by Godwin (1975a), Tooley (1978a), and Long (1991). For example, a change in tendency of sea-level movement from negative to positive will take place within the organic sediments and the actual lithological change from semi-terrestrial to a marine or brackish deposit will be a point within the continuous process of a positive tendency (Long, 1991; Long & Shennan, 1994). This highlights the importance of microfossil analyses in identifying the onset of sea-level tendencies, which do not always coincide with lithological boundaries. Extensive collection of corroborating data is needed to confirm the meaning of a particular sea-level indicator. Also, unless the method of definition is clearly stated, then the correlation between sites and regions is meaningless. This is particularly important as the great advantage of the tendency approach is that it can incorporate a range of SLIPs from a variety of palaeoenvironments, thus increasing the scope of sea-level investigations.

The Perimarine Zone

Although sea-level change is likely to be the major allogenic factor influencing vegetation change on perimarine floodplains, the relationship between the water table in the perimarine zone and sea level is not at all clear (see Tooley, 1978a; van de Plassche, 1982; Berendsen, 1984a; 1984b; van Dijk *et al.*, 1991; Pons, 1992; Törnqvist, 1993a; Kiden, 1995; Vos & van Heeringen, 1997). Tooley's (1978) reservations have been confirmed by research into the Rhine-Meuse system in the Dutch perimarine area. The variation in height of basal peats used for the construction of early sea-level curves in this area (Jelgersma, 1961; 1966) were initially thought to be related to differences in tidal range or tectonic subsidence. However, van de Plassche (1980; 1982) has shown that basal peat formation there is mainly governed by local hydrological conditions and challenges the validity of sea-level curves based on such criteria (e.g. Jelgersma, 1961). The use of basal peat samples for the analysis of age-altitude trends in sea level is problematic where groundwater induced peat growth occurs above contemporaneous sea level (Pons, 1992; Kiden, 1995; Vos & van Heeringen, 1997). The altitudinal position of stratigraphic contacts in the perimarine area is also affected by tidal irregularities associated with their inland location (see section 2.4.7).

The simple relationship between trangressive and regressive phases, and synchronous deposition of alternating layers of clastic and organic sediment, in the perimarine zone of The Netherlands (Hageman, 1969) has been disproved (Berendsen, 1984a; 1984b; 1993; 1995; Törnqvist, 1993a, see section 1.1). The relationship between sea-level change, sedimentation, vegetation systems, and fluvial activity in the perimarine zone can be complex (Berendsen, 1984a; Törnqvist, 1993a) and are, in part at least, related to the groundwater conditions. Groundwater levels are a function of a large number of factors, such as tidal range, river gradient, permeability of subsurface sediments (seepage), coastal morphology, size of sediment units, precipitation, and drainage, which can lead to irregularities over short distances, and can also be variable over time (Törnqvist, 1993a).

Further complications arise when attempting to assign a reference water level to the dated sample which can vary from MTL, through MHW to MHWST (Godwin, 1940; Kidson & Heyworth, 1979; van de Plassche, 1980; 1982; Shennan, 1982; 1986b; 1987). Shennan (1994) also points out that the inception of peat growth itself can influence the local groundwater conditions.

2.4.7 The Floodbasin and River Gradient-Effect

Van de Plassche (1980; 1982) has observed that some sea-level index points collected from tidally influenced river valleys within the Dutch perimarine area do not plot within the expected band depicting coastal MHW (Jelgersma, 1961; 1966; 1979; Louwe Kooijmans, 1974; van de Plassche, 1980). In fact, the data collected from peats formed on 'donken' (buried river dunes and levees) converge with the MSL curve in an irregular manner. The fact that peat growth on the donken has occurred under strong influence of river water was thought to be indicated by the peat composition (fen wood) which suggests formation in eutrophic freshwater environments. However, it is clear that local MHW level curves do not always plot at the same level as coastal MHW curves (Kiden, 1989).

The data points from the donken in the Brandwijk and Brandrecht area of the Rhine-Meuse delta plot above the curve for MSL due to the presence of a **river gradient-effect**. Due to the distinct seaward slope characteristic of rivers in the early to mid-Holocene, MHW and other tidal levels tend to be raised in an upstream direction. Louwe Kooijmans (1974) called this the river gradient-effect. Hageman (1969) suggests that this effect is less marked before *c.*5000 BP (*c.*6000 to *c.*5500 cal. yrs BP), because the rivers were forced to deposit much of their load upstream, which caused a tendency for the gradient to maintain itself. Therefore, the greater the distance from the coast, the greater the river gradient-effect. Van de Plassche (1980) depicts this effect schematically (see Figure 2.2). Figure 2.2a illustrates the progressive decrease in river gradient with time as a result of the rising sea-level and simultaneous raising of the river bed due to fluvial deposition. As river gradients level off towards the mid to late-Holocene, the data collected from river valleys in the perimarine area gradually converge with the MSL curve (Figure 2.2b). Van de Plassche (1980) calls this a **gradient-effect reduction curve**. A similar pattern of convergence of water levels in the Holocene between the outer and inner Thames estuary is also evident (Devoy, 1979). In a funnel-shaped estuary, tidal amplitude often rises landward anyway, due to confinement of the tidal wave as it progresses up estuary. This has been termed the **estuary effect** by Fairbridge (1961).

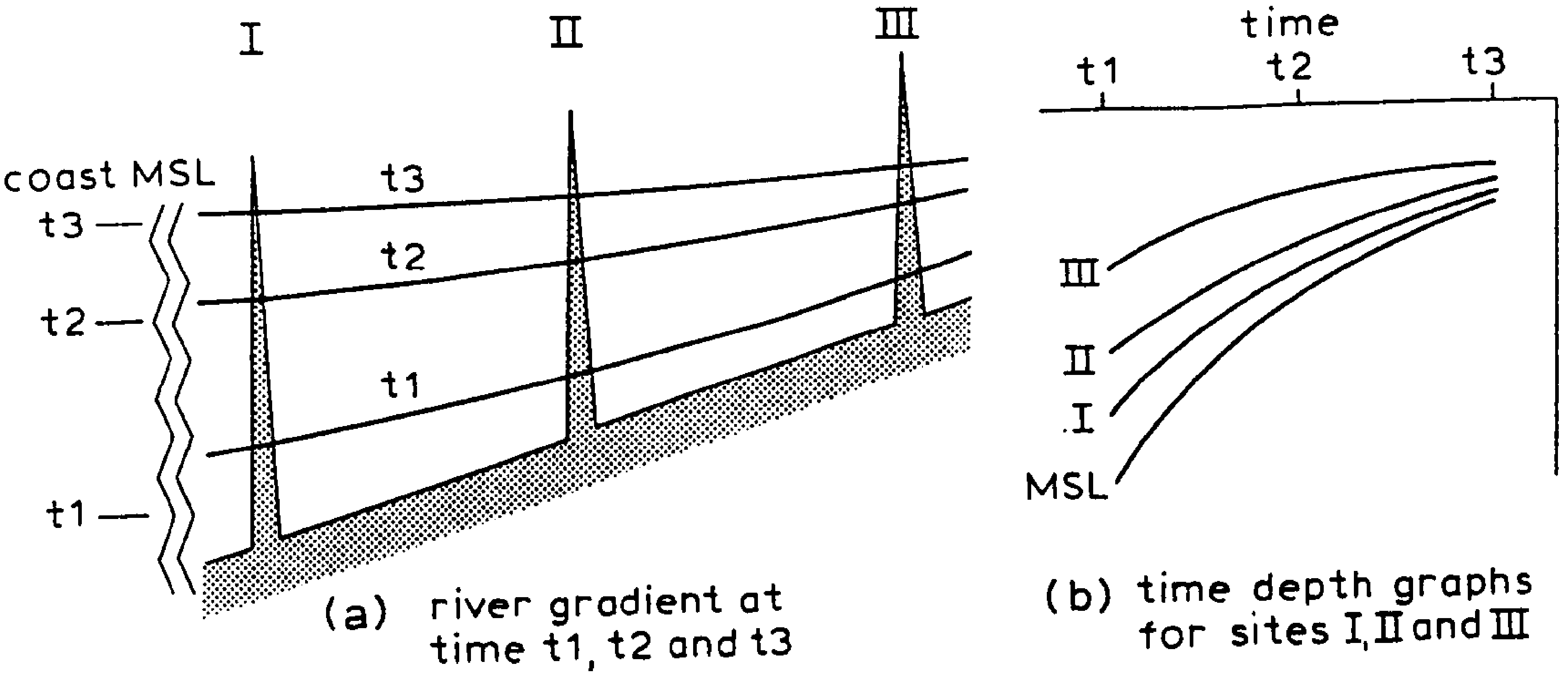


Figure 2.2; a) Illustrates the river gradient effect associated with the lower reaches of river valleys in the perimarine area; b) shows the decrease in river gradient effect over time as a result of estuarine infilling. (Adapted from van de Plassche, 1980). See text for full explanation.

Conversely, some samples from Brandrecht and Brandwijk plot below the contemporaneous coastal MHW curve.· Jelgersma (1980) accepted that, since *c*.4800 BP (*c*.5700 to *c*.5300 cal. yrs BP), the mean tidal range along the mid-western Netherlands coast has not changed significantly (although see Roep & Beets, 1988 for more recent evidence to the contrary), so this effect could only be ascribed to the reduction in tidal amplitude as the tidal wave moves inland (van de Plassche, 1984). Van de Plassche (1980) attributes this to frictional dissipation of energy caused when the tidal wave moves into a lagoon or wide estuary. This phenomenon has been termed the **floodbasin effect** (Van Veen, 1950; Zonneveld, 1959). This refers to the reduction in the tidal amplitude and lowering of local MHW caused by strong frictional dissipation of tidal energy at the water-bed interface (Allen *et al*, 1980) and/or the increase in storage capacity of storage basins and lagoons within estuaries, which can absorb part of the flood volume entering (that part of) the estuary (van de Plassche, 1980; Kiden 1989). This effect is shown schematically in Figure 2.3. Van de Plassche (1984) has even observed the coexistence of a steep river gradient-effect in a reach and a floodbasin effect downstream of that reach.

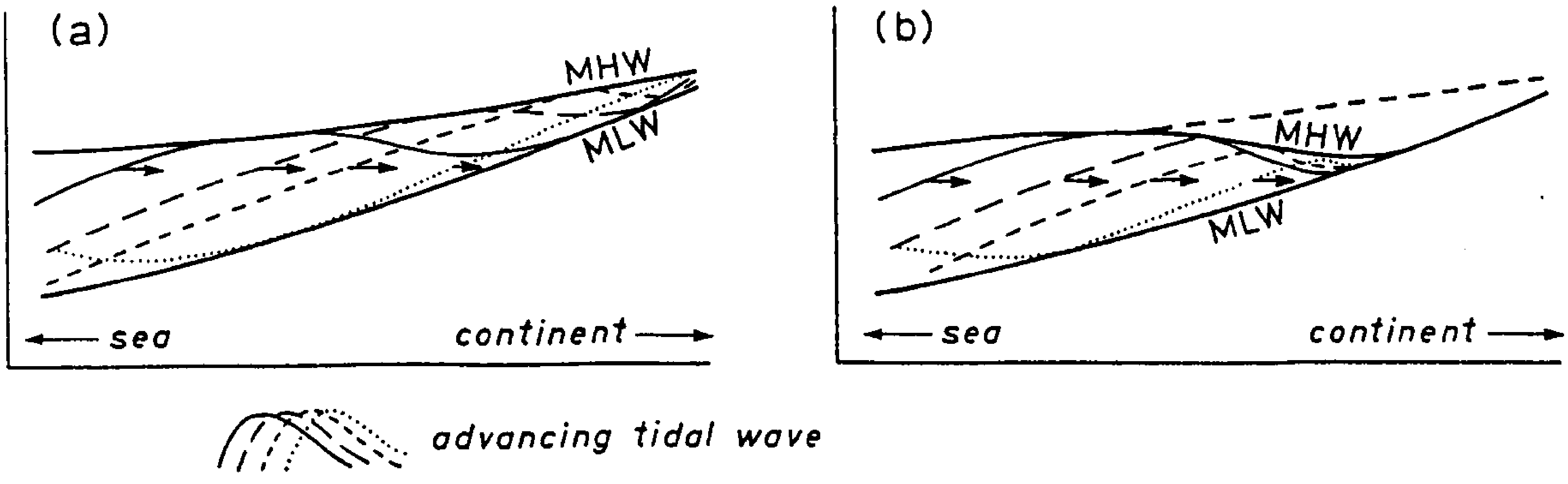


Figure 2.3; a) Illustration of the rise of MHW up an estuary with no floodbasin effect, and b) a marked floodbasin effect. Adapted from van de Plassche (1980). See text for full explanation.

Van Dijk *et al.* (1991) have evaluated the general trend of Holocene water-level changes in the Rhine-Meuse delta and attempted further explanations for the irregularities in evolution of this trend using a relatively large number (45) of ^{14}C dates. The plot of isochrones of peat formation on river dune sand shows a distinct knick-point with a sea-level dominated downstream part and river gradient dominated upstream part (van Dijk *et al.*, 1991). All curves converge upon MSL which is consistent with a gradually decreasing river gradient-effect (van de Plassche, 1980; 1982). Compared to an adjusted MSL curve, the water level curves from the downstream area show slight gradients that are more closely related to MSL than MHW which confirms the reduction of tidal amplitude inland caused by the floodbasin effect.

The presence of these effects has been noted elsewhere along the south-western North Sea coast. In Belgium, Kiden (1989; 1991) notes a significant floodbasin effect in the perimarine area of the River Schelde near the Dutch-Belgian border from *c.*4500BP (*c.*5500 to *c.*4900 cal. yrs BP). He ascribes a later rise in MHW level, and upstream penetration of tidal conditions, as the result of a decreasing floodbasin effect in the late-Holocene. Elsewhere on the Belgian coastal plain, no pronounced river gradient-effect is noticeable, but rather a lowering of tide levels with increasing distance from the sea (Denys & Baeteman, 1995) is inferred.

Studies of the inner Severn estuary by Hewlett & Birnie (1996) note that the contact between the extensive floodplain peat and overlying clay decreases in altitude inland and not seawards as would be expected with a river gradient-effect. Detailed palaeoecological analyses of the sediments suggest that a rise in water table began after *c*.3100 BP (*c*.3500 to *c*.3000 cal. yrs BP) as a result of rising relative sea level, which allowed sediment-charged tidal waters to penetrate inland and inundate the former wetlands forming extensive pools of standing water. The lowering of the tide height due to the expansion of the tide into a series of floodbasins is suggested as the reason for the decrease in contact height inland (Hewlett & Birnie, 1996). This floodbasin hypothesis may also account for the fairly sharp sedimentary contact over this horizon due to sudden flooding upon breach of a threshold.

2.5 Summary

The background to the processes affecting floodplain development has been presented. Sea level is a major control on groundwater level and paludification within coastal river valleys. Therefore analysis of the litho- and biostratigraphic record of sediments within perimarine environments can potentially provide sea-level indicators which extend the effective area for sea-level studies (Tooley, 1986). Similarly, the collection and evaluation of data from such transitional environments can provide information on the importance of catchment processes in the evolution of the fluvial system. Former periods of fluvial activity or sea-level change may be examined with reference to transitions between floodplain fen peats and brackish estuarine sedimentation, although relationships are not always precise. For example, variations in past tidal regimes, wave climates, and sediment supply mean that alternations between terrigenous organic deposits and estuarine sediments are not always indicative of sea-level change. Within coastal valley systems there are further complications for determining former altitudes of river or sea-level due to the river gradient and floodbasin effects (van de Plassche, 1980, 1982).

Chapter 3 The Study Area

3.1 Introduction

This chapter introduces the geographical setting of the three principal study areas. The geological background of the region is summarised, followed by an account of the late-Quaternary evolution of the Humberhead Levels and Vale of York area. This is based on the conceptual model of Postglacial landscape development proposed by Gaunt *et al.* (1971). This is important because processes acting on the landscape during this period were influential in shaping the landforms and creating the physical valley characteristics evident in the present topography. A brief summary of the available information regarding the Holocene palaeoenvironmental history of the area before this investigation is presented. The physical aspects of the Humber estuary are given, and the sediment supply discussed. The history of attempts at land drainage are briefly considered, in addition to modern factors such as land-use, soils, and climate.

3.2 Site Location and Description

The sites under investigation lie within the lower tract of the rivers Aire and Ouse in Yorkshire (Figure 3.1). The Ouse begins north of York, beyond the confluence of the rivers Swale, Ure and Nidd, which drain the western Pennines. It flows down the Vale of York, where it usurps the rivers Wharfe, Derwent and Aire, before entering the Humber at Goole. The headwaters of the river Aire occur in the central Pennines. It originates near Malham in the southern Yorkshire Dales, passes through Leeds and Castleford, and continues eastwards towards the lowland catchment, where it combines with the Ouse to form the eastern extent of the inner Humber estuary.

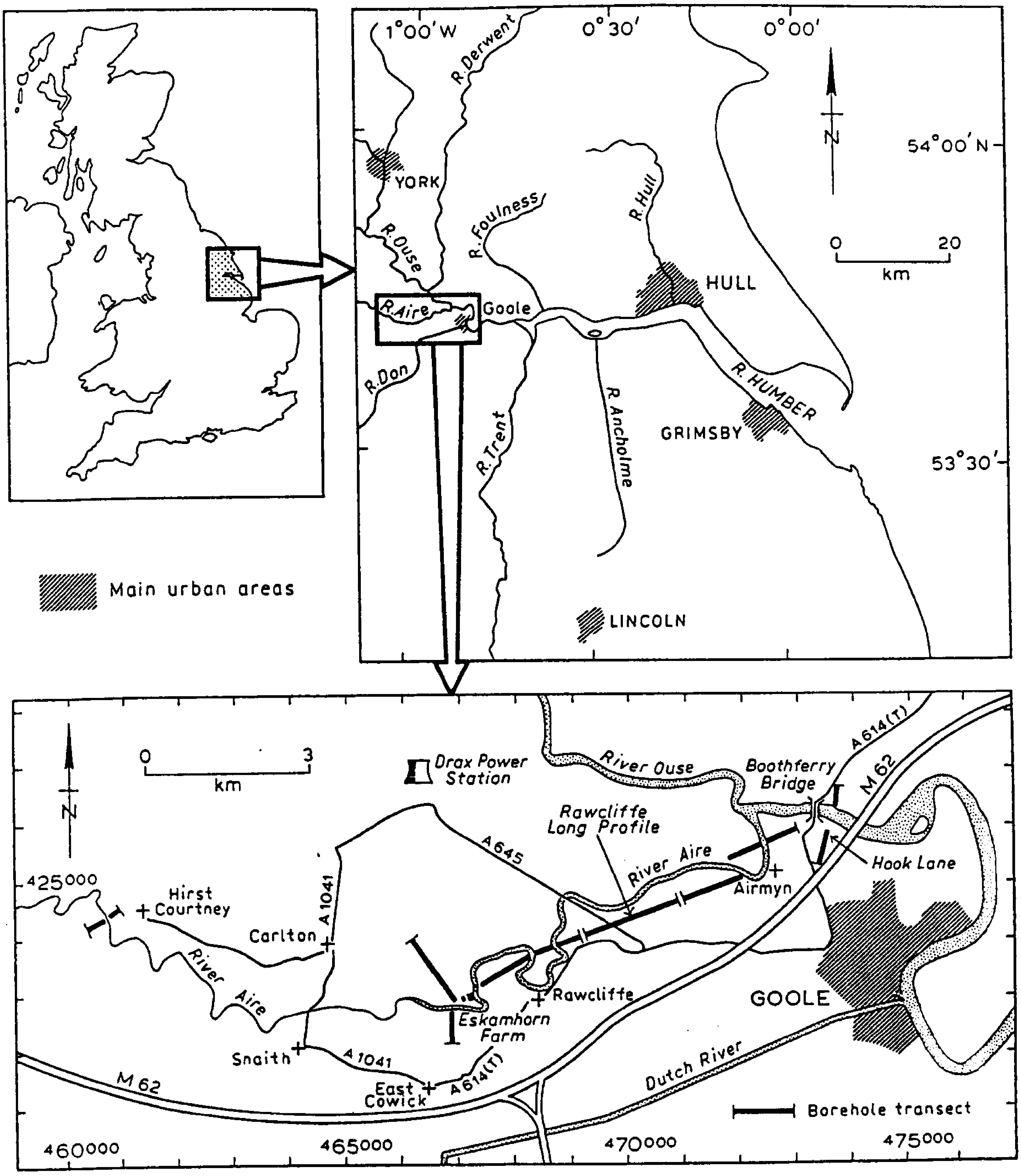
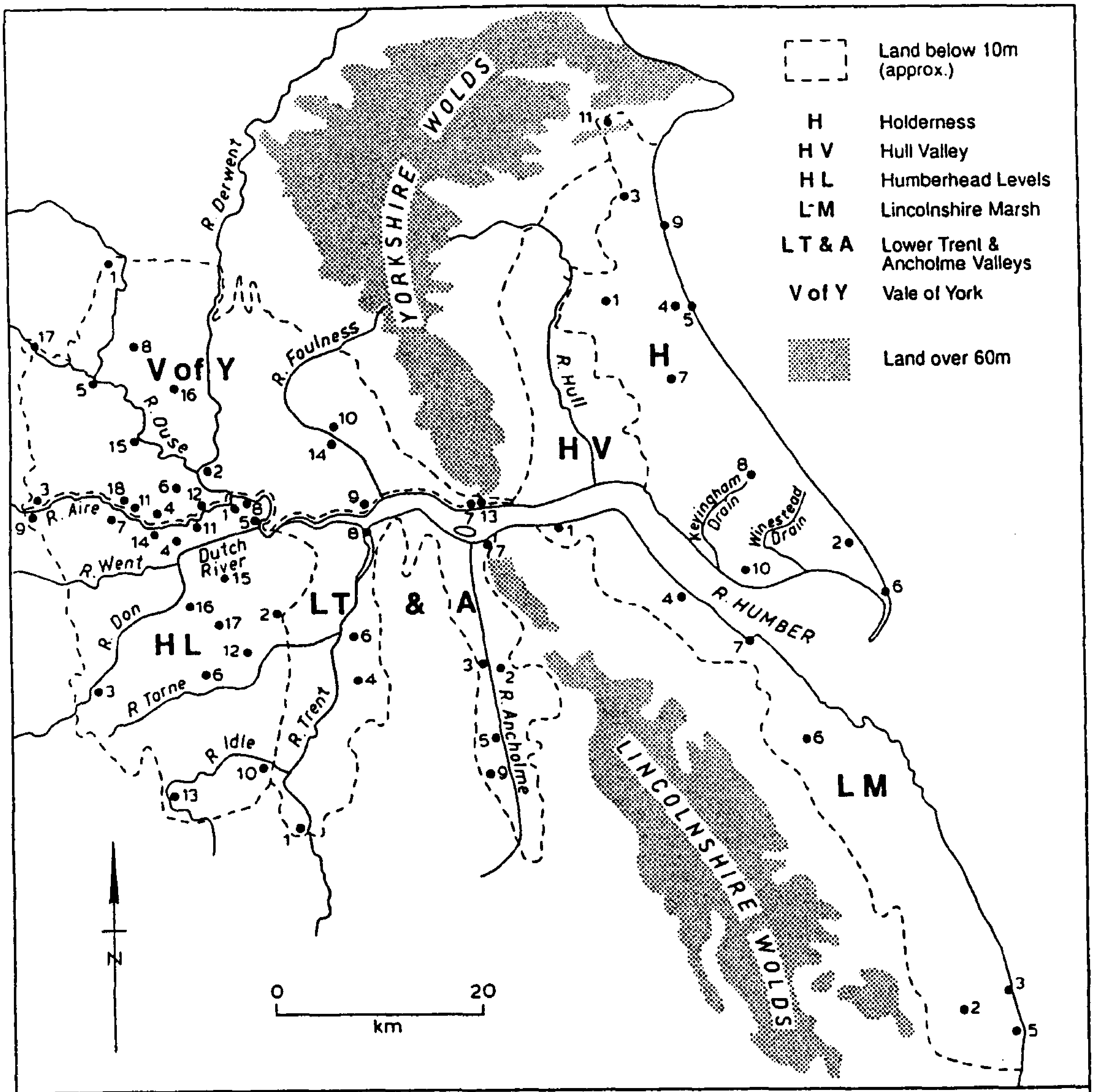


Figure 3.1 Location map of the study area within Britain and the Humber region, showing the position of the borehole transects.

The current study investigates a reach from close to the tidal limit of the Aire at Hirst Courtney, to the Ouse at Airmyn (Figure 3.1). Detailed analysis has concentrated on three sites, named after villages, farms, or other landmarks in the immediate vicinity of the study area. The sites at Hirst Courtney village (OS Grid Reference, SE605245), and Eskamhorn (SE670230), east of Rawcliffe, lie within the floodplain of the river Aire, whilst the site at Hook Lane (SE7352550) is situated within the floodplain of the river Ouse north of Goole, downstream of the confluence with the Aire at Airmyn (Figure 3.1). A long transect of cores has been undertaken down the length of the Aire floodplain, past Rawcliffe and Airmyn, linking Eskamhorn with Hook Lane. Although the study reach impinges on the Ouse floodplain, the study area is referred to as the lower Aire valley hereafter.

For the administrative purposes of the HWP, the Humber area has been divided into wetland regions (Figure 3.2). The sites investigated in this study broadly lie within the southern-most tract of the Vale of York, and the northern limits of the Humberhead Levels (Figure 3.2), and form the arbitrary boundary between these two geographical areas (Van de Noort & Davies, 1993). Detail is given to the sequence of environmental changes within these two generic areas during the late-Quaternary period in section 3.4 and 3.5 below.



HOLDERNESS

1. Brandesburton
2. Dimlington
3. Gransmoor Quarry
4. Hornsea Mere
5. Hornsea Old Mere
6. Kilnsea Warren
7. Lambwath Mere
8. Roos Bog
9. Skipsea Withow Mere
10. Sunk Island
11. Willow Garth

HUMBERHEAD LEVELS

1. Airmyn
2. Crowle
3. Doncaster
4. East Cowick
5. Goole
6. Hatfield Moors
7. Hensall
8. Hook Lane
9. Knottingley
10. Misterton
11. Rawcliffe
12. Sandtoft
13. Scaftworth
14. Snaith
15. Thorne Moors
16. Thorne Waterside
17. West Moor

LINCOLNSHIRE MARSH

1. Barrow Haven
2. Butterbump
3. Chapel Point
4. Immingham
5. Ingoldmells
6. Newton Marsh, Telney
7. Union Dock, Grimsby

LOWER TRENT AND ANCHOLME

1. Bole Ings
2. Brigg
3. Castlethorpe
4. Messingham
5. Redbourne Hayes
6. Scunthorpe
7. South Ferriby
8. Trent Falls
9. Waddingham Holmes

VALE OF YORK

1. Askham Bog
2. Barmby on the Marsh
3. Burton Salmon
4. Carlton
5. Cawood
6. Drax
7. East Clough, Melton
8. Escrick
9. Faxfleet
10. Hasholme
11. Hirst Courtney
12. Newland
13. North Ferriby
14. Sandholme Lodge
15. Selby
16. Skipwith Common
17. Tadcaster
18. Temple Hirst

Figure 3.2 Location of the wetland regions in the Humber Basin (according to Van de Noort & Davies, 1993) and the principal palaeoenvironmental sites mentioned in the text.

3.3 Geology

Much of the solid geology that underlies the study area (Figure 3.3) comprises north-south trending bands of Triassic Sherwood Sandstone (formerly known as Bunter Sandstone), which overlie Upper Magnesian Limestones and Upper Permian Marls in the western margin of the area (Kent & Gaunt, 1980; Gaunt, 1994). Along the lower Ouse east of Goole, and in the eastern Humberhead levels, Triassic Mercia Mudstones (formerly known as Keuper Marls) overlie the Sherwood Sandstone (Gaunt, 1994). Between the lower Trent and Ancholme valleys, the Triassic strata are overlain by Jurassic marls, limestones, sandstones and clays, and form the Lincoln edge (Ellis, 1998). In the east of the Humber estuary the Jurassic strata are overlain by, Liassic Oolitic Limestones and Cretaceous Chalk.

The Humber is divided by the narrows of the Humber Gap, where the estuary cuts through the chalk escarpment of the Yorkshire and Lincolnshire Wolds (Figure 3.2). Throughout much of the area however, this bedrock is coated with a veneer of much younger Quaternary drift deposits (see Figure 3.5 and 3.6), relating mainly to the Devensian glaciation and subsequent Holocene epoch (Ellis, 1997). There are some outcrops of pre-Devensian deposits within the study area which are probably remnants of the Ipswichian interglacial or Anglian glacial period (Ellis, 1997). The origins of these sands and gravels are considered in the next section.

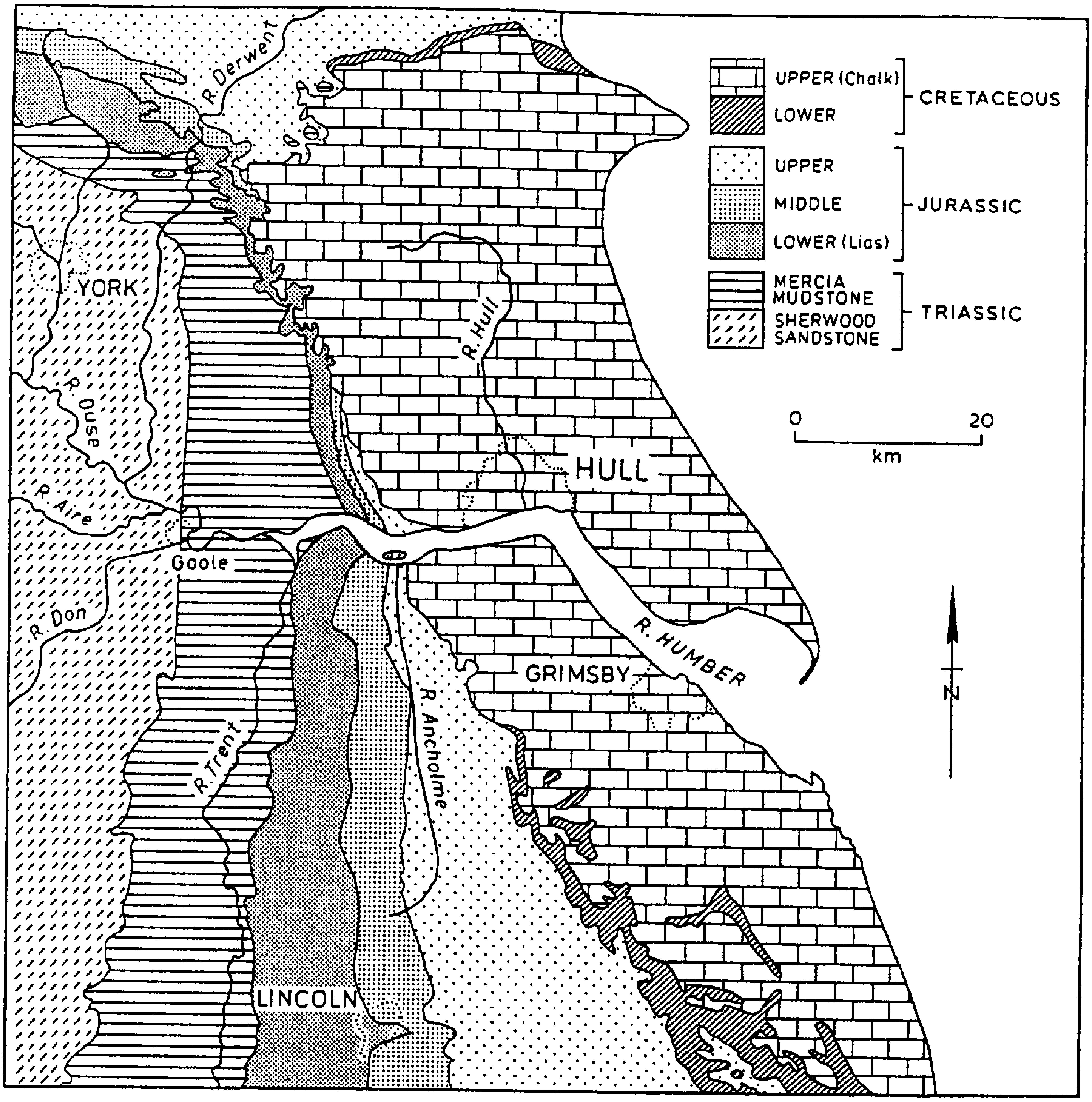


Figure 3.3 The solid geology of the Humber basin (based on Kent & Gaunt, 1980 and Gaunt, 1994).

3.4 Quaternary Landscape Development

There are no early-Quaternary deposits in existence in the Humber basin. Gaunt (1994) states that there are deposits originating from mainly the last three stages of the Quaternary (the Ipswichian, Devensian, and Holocene), and also an older pre-Ipswichian glacial period (probably the Anglian). The evolution of the Humberhead Levels and southern Vale of York during this time period is summarised here.

The occurrence of clay till containing erratics (and other denuded fluvioglacial, and sub-glacial channel sand and gravel deposits, deeply incised into bedrock) within the Vale of York, Doncaster area, and elsewhere, is thought to be remnant of the deglacial phase of a pre-Ipswichian cold stage (Gaunt, 1994). Such deposits occur locally, overlain by river terrace deposits containing fossils of known Ipswichian age (*c.*130 to 10 ka yr BP), which suggests they belong to an earlier glacial period (Gaunt, 1994). This glacial period has been provisionally correlated with deposits of Anglian age (*c.*480 to 430 ka yr BP) east of the Lincolnshire Wolds by Gaunt *et al.* (1992).

The margins of the river Aire floodplain west of the study area are delineated by the Snaith ridge which occasionally outcrops at the surface of the floodplain between Hensall and Snaith (see Figure 3.5). This ridge is comprised of Sherwood Sandstone running in a south-easterly direction, but is covered by *c.*5 m of sand (often clasts of weathered Carboniferous Sandstone) and gravel with some silt, clay and abundant coal particles, eroded from coal measures to the west. These deposits are thought to be of fluvioglacial origin, and associated with the pre-Ipswichian glacial phase mentioned above (Gaunt, 1994). Processes operating upon deglaciation reworked these sediments into terrace-like features which are of Ipswichian or early-Devensian age, and buried in most localities by thick late-Devensian deposits (Gaunt, 1994). A buried extension of the Snaith ridge runs under Rawcliffe station towards Goole (Gaunt, 1994) at the southern margin of the study site.

Fluvial deposits of Ipswichian age occur throughout the Humberhead Levels and Vale of York region, consisting of sand, clay and gravel, and are fossiliferous. Plant and wood remains have been recorded, with high frequencies of *Carpinus* among the pollen types preserved, indicating a temperate Ipswichian climate (Jones & Keen, 1993; Gaunt, 1994). These lithostratigraphical units have been formally named as 'older river gravel' (Gaunt, 1994) and form extensive spreads and terrace-like features, for instance, where the river Aire enters the Vale of York near Knottingley.

The vast majority of sediments extant within the Humber basin have a Devensian origin. Evidence from marine records suggest that this last cold stage started *c.*120000 years ago, and lasted until *c.*10000 BP (*c.*12300 to *c.*10600 cal. yrs BP, Imbrie *et al.*, 1984). Major ice sheets formed in northern Britain during this time but the main phase of glacier expansion took place during the late-Devensian, after *c.*25000 BP (Bowen *et al.*, 1986). Gaunt (1994) states that there are few, if any, Devensian deposits older than *c.*18000 BP (*c.*22300 to *c.*20600 cal. yrs BP), due to fluvial incision (sea level had fallen as low as 120 m below its present level, Fairbanks, 1989), denudation, and severe periglacial conditions, so most of the sediments of Devensian age in the Humber region relate to this final phase of glacial activity.

During the late-Devensian period, the Vale of York and Humberhead Levels underwent considerable landscape change. The timing of this ice-advance is transgressive across the region but most dates indicate glacial activity soon after *c.*18000 BP (*c.*22300 to *c.*20600 cal. yrs BP, Penny *et al.*, 1969). Radiocarbon dates from organic silts beneath late-Devensian boulder clay at Dimlington Farm near Easington, Holderness, the mean of which is 18730 BP (Madgett & Catt, 1978), indicate that ice reached this point after this time. This late-Devensian stage of the Quaternary is termed the Dimlington Stadial after this type site (Rose, 1985). During this stadial, ice approached the Humber from the north down the Vale of York (see Figure 3.4) almost as far as Doncaster (Gaunt, 1976), and from the western part of the North Sea over Holderness and the Lincolnshire Marsh (Catt, 1990). A minimum age for ice retreat from this area is provided by a radiocarbon age of 13045±270 BP (16299 to 14596 cal. yrs BP) from basal sediments infilling a kettle hole at Roos Bog, Holderness (Beckett, 1981). The main stages in the late-Quaternary evolution of the region are outlined below and are based on the conceptual model proposed by Gaunt *et al.* (1971).

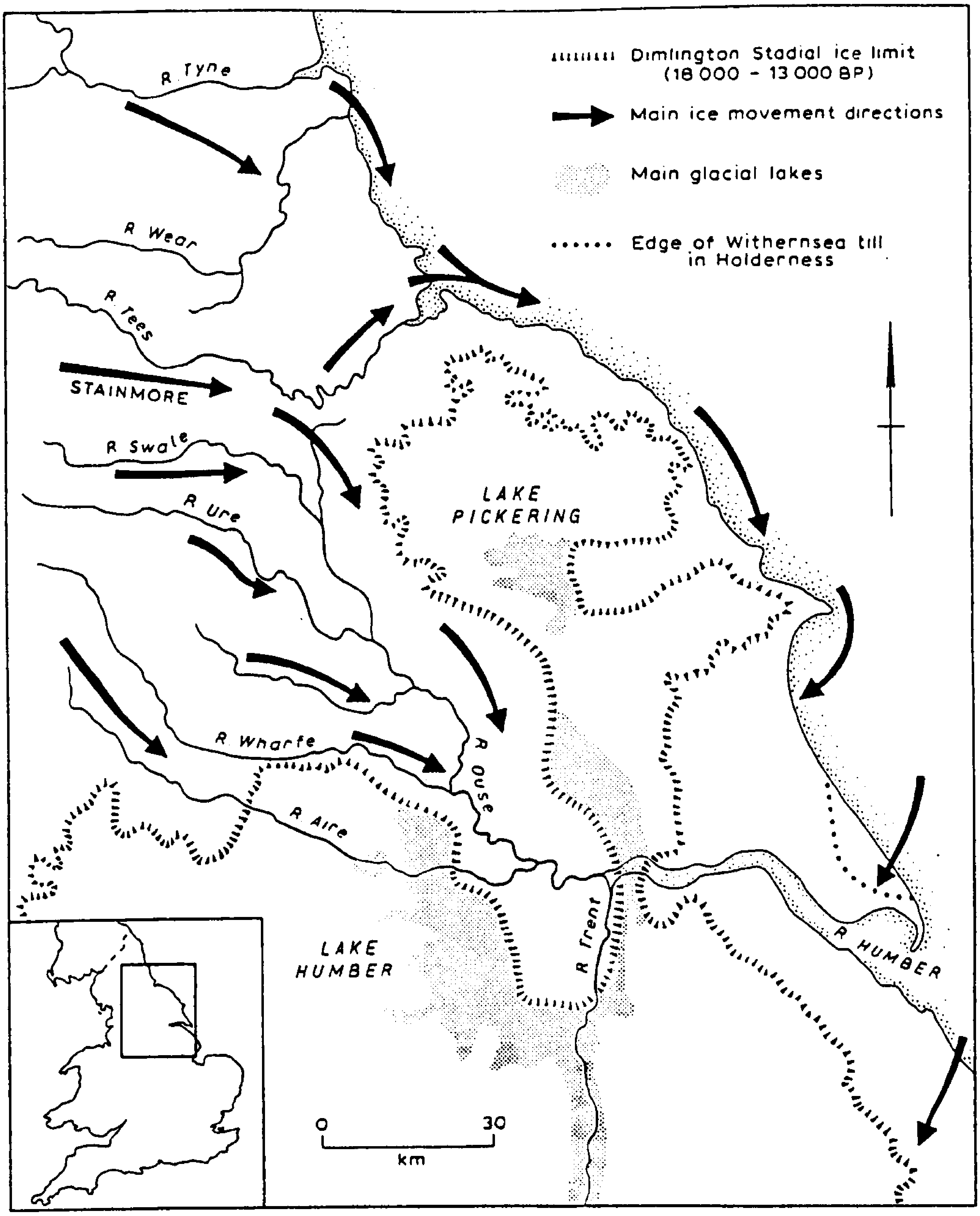


Figure 3.4 The ice limits during the Dimlington Stadial of the Devensian glaciation in East Yorkshire (based on Madgett & Catt, 1978 and Catt, 1990).

The Lake Humber Phase *c*.21000 to *c*.12000 BP (>*c*.23000 to *c*.13300 cal. yrs BP). The expansion of the North Sea ice sheet on the eastern side of the Yorkshire and Lincolnshire Wolds, which reached a height of *c*.60 m OD (Catt, 1990) during the Dimlington Stadial, caused a drainage blockage across the Humber gap. Since ice was also occupying the Vale of York, meltwater was impounded and a large lake developed (Lake Humber) in the southern Vale of York and Humberhead Levels (Figure 3.4). Initially, the lake level rose to over 30 m OD (Catt, 1991), with sand and gravel deposited at the margins (Gaunt, 1994). During this high lake level stage, dated to between *c*.21000 and 18000 BP (>*c*.23000 to *c*.20600 cal. yrs BP, Gaunt, 1974; 1994), the ice tongue extended down the Vale of York (the Stainmore-Vale of York ice advance), partly floating on the lake, and deposited sand and gravel into the water (Gaunt, 1976). After this initial surge, the ice retreated northwards and stabilised at Escrick, forming an arcuate morainic ridge of till (Catt, 1990; 1991) and the lake level stabilised around 9 m (Gaunt, 1994).

As the ice front slowly wasted, drainage of Lake Humber was inhibited by the blocking of the Humber gap by glacial deposits, allowing the lake to survive after the melting of the ice *c*.13000 BP (16300 to 14500 cal. yrs BP). Lake Humber eventually disappeared *c*.11100 BP (*c*.13400 to *c*.12600 cal. yrs BP). This is the age of a radiocarbon date obtained from a buried soil located between Lake Humber deposits and blown sand at West Moor (Gaunt *et al.*, 1971). The demise of Lake Humber was deemed to be a result of sediment infilling, rather than draining by breaching the moraine (Gaunt, 1981). The exact timing of this event may be earlier, since insect assemblages from a peat sealing the lake deposits at Sandtoft, indicate the lake was wetland *c*.12500 BP (*c*.15600 to *c*.13800 cal. yrs BP, see Buckland in Dinnin, 1997a).

The sediments that infilled Lake Humber as it silted-up are termed '25 foot drift' by Gaunt (1994) and form most of the flat plain of the southern part of the Vale of York, and Humberhead Levels area, at a height not exceeding 25 feet or *c*.8 m OD (Gaunt, 1994). Sand is present towards the base of this unit, and reaches a variable thickness, between 1 and 7 m, in the area of Drax Power Station (Gaunt, 1994). Overlying the sand are up to 20 m of blue-grey to red-brown laminated silty clays, with sand and coal particles locally abundant. The thickest sequences are found within deeply incised river valleys. These late-Devensian drift deposits attain thicknesses of *c*.5 to 10 m in the region of the river Aire between Snaith and Goole (Gaunt, 1994). An upper sand is often found resting on the silt and clay of the '25 foot drift', which is generally not more than 2.5 m thick (Gaunt, 1994). In some areas it is clayey and contains coal particles.

The Braiding Phase *c*.12000 to *c*.10500 BP (*c*.14900 to *c*.11000 cal. yrs BP). After the disappearance of Lake Humber, rivers and streams began to flow across the emergent plain towards the Humber gap. These rivers did not incise into the drift and lake deposits but actually aggraded, reworking the upper lacustrine deposits, forming sandy levees either side of their braided channels (Gaunt *et al.*, 1971; Gaunt, 1981). The lack of incision suggests that rivers were responding to a regional drainage base-level that was at about the same height as the emergent plain. This provides additional evidence that the lake did not drain suddenly by breaching a barrier of glacial deposits but rather silted up with sediment (Gaunt, 1981). This also suggests that the knick point of the estuary was well out into the present North Sea basin, and had not yet transgressed westwards through the Humber gap (Van de Noort & Davies, 1993; Dinnin, 1997a).

These levees form ridges and mounds with a linear distribution visible along the flanks of present rivers (see Figure 3.5). The fact that old river courses can be recognised from these sand levees led Gaunt (1987) to suggest that rivers such as the Aire and Ouse followed essentially the same course during the early-Holocene as today. A levee is evident along the river Aire from Hirst Courtney to Carlton (Figure 3.5), and further remnants are visible above the floodplain surface of the river Aire between Rawcliffe and Airmyn (see section 6.3). The duration of this phase is bracketed by dates of 11100±200 BP (13444 to 12615 cal. yrs BP) from soil on top of lacustrine clays at West Moor (Gaunt *et al.*, 1971) and 10469±60 BP (12569 to 12142 cal. yrs BP) from peat between levee deposits and blown sand at Cawood in the Vale of York (Jones & Gaunt, 1976).

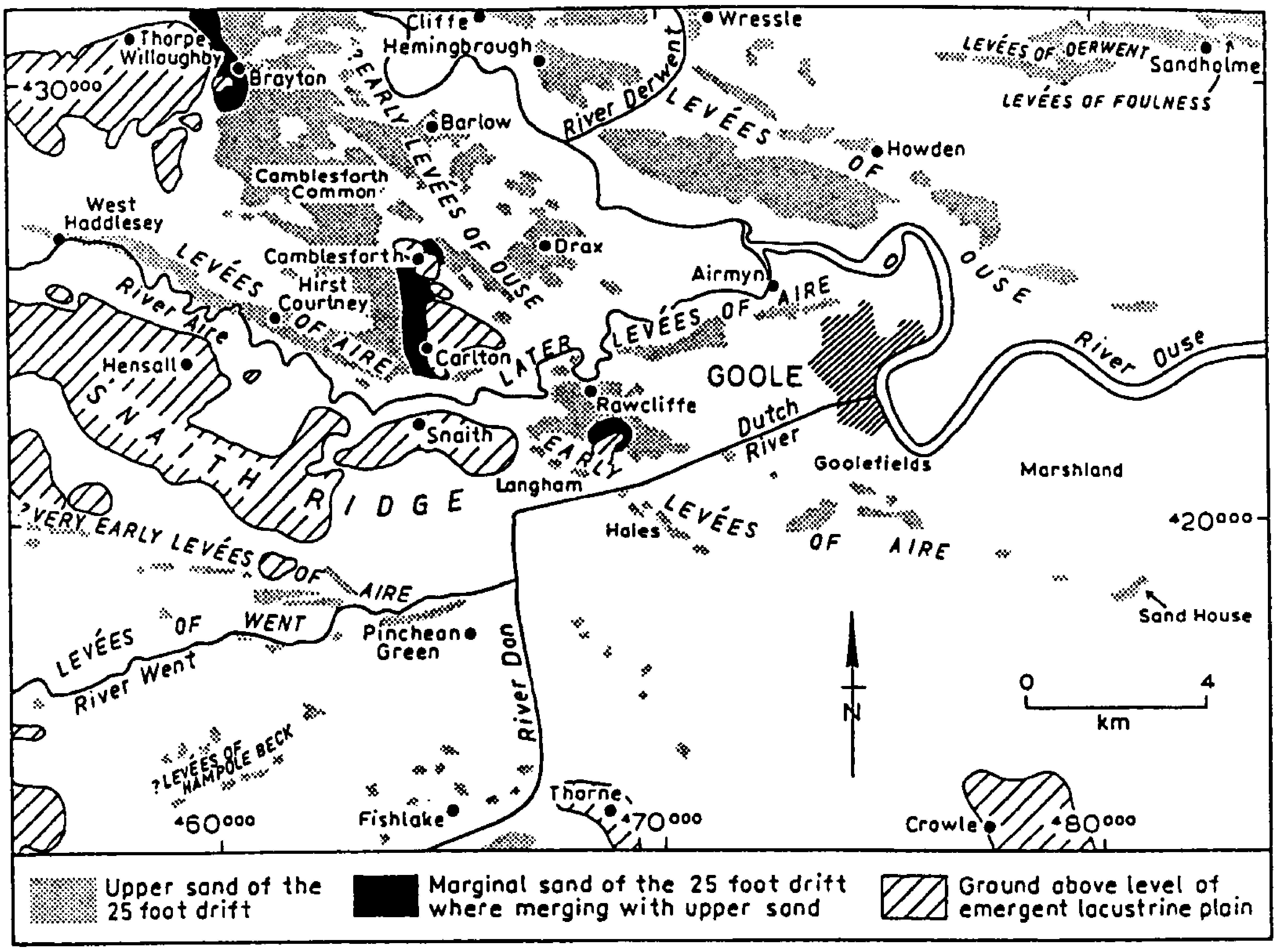


Figure 3.5 The Snaith ridge and the distribution of Lateglacial river levee deposits in the proximity of the study area (after Gaunt, 1994).

Aeolian Sand Phase *c*.10500 to *c*.10000 BP (*c*.13000 to *c*.10600 cal. yrs BP). The exposed sandy ridges within the Vale of York and Humberhead Levels were partly reworked by aeolian action during the Lateglacial and early-Holocene times, which gave rise to the formation of blown sand (coversand) deposits and extensive dune systems (Gaunt, 1987; Halkon, 1990). There are problems associated with the chronology of this phase of reworking, which relate, in part, to contamination, and the uncertainty regarding radiocarbon dating this period, which has resulted in a large spread of ages.

Radiocarbon dates from the base of the aeolian deposits in the lower Trent valley at Messingham near Scunthorpe, and West Moor and Cawood in the Vale of York, gave dates of 10280±120 BP (12475 to 11138 cal. yrs BP), 11100±200 BP (13444 to 12615 cal. yrs BP) and 10469±60 BP (12569 to 12142 cal. yrs BP) respectively (Buckland & Dolby, 1973; Gaunt *et al.*, 1971; Jones & Gaunt, 1976). Thermoluminescence dates indicate substantial amounts of blown sand had begun to accumulate near Scunthorpe potentially as early as *c*.12500 cal. yrs BP, and continued until *c*.11400 cal. yrs BP (Bateman, 1995; 1998). Dates of 10700±190 BP (13006 to 12152 cal. yrs BP) and 9950±180 BP (12203 to 10587 cal. yrs BP) from within the blown sands in the Vale of York (Matthews, 1970), and 10550±250 BP (12974 to 11224 cal. yrs BP) in the Humberhead Levels (Buckland, 1982), place this aeolian phase within the last millennium of the Devensian, a time of known climatic deterioration, and indicate sand continued to accumulate up to the early-Holocene (Bateman, 1995; 1998).

Analysis of pollen and insect faunas from such dated organic sediments indicate a cold, arctic climate, and a tundra environment with limited vegetation cover. Such conditions are typical of the terminal cold stage of the Devensian, *i.e.* the Loch Lomond Stadial/Younger Dryas (Matthews, 1970; Gaunt *et al.*, 1971; Gaunt 1981; Buckland, 1982; Bateman, 1998). Bateman (1995; 1998) provides thermoluminescence dates for the cessation of this aeolian phase of 10780±840 yrs (Q) and 10040±790 yrs (Q) from North Lincolnshire.

Incision and aggradation <*c*.10000 BP (<*c*.12300 cal. yrs BP). The migration of the knick point of the proto-Humber past the Humber gap towards the unconsolidated lacustrine deposits of the Humberhead Levels and Vale of York caused a drop in regional base-level, and deep fluvial incision ensued, down to contemporary sea-level of lower than -20 m OD (Gaunt, 1981; 1994; Gaunt & Tooley, 1974). The timing of the onset of this incision phase is not known precisely, although it certainly post-dates formation of the sand levees (see above). Gaunt (1981) suggests that it may have started during the late-Devensian, when a sudden lowering of groundwater level would have facilitated sand mobilisation. Incision continued until equilibrium was reached with local base level (associated with the regional sea-level position), from which point channels and floodplains would have begun to aggrade following the trend of sea-level rise (Gaunt & Tooley, 1974).

Gaunt (1994) suggests incision had largely ceased by *c*.8500 BP (*c*.9800 to *c*.7600 cal. yrs BP) when lacustrine shelly marl had begun to form in an incised channel at Burton Salmon in the Vale of York (Norris *et al.*, 1971) although Van de Noort & Davies (1993) and Dinnin (1997a) caution that such deposits can form due to localised channel blockage rather than a response to rising sea level. Aggradation had certainly replaced incision in the outer Humber estuary between *c*.8000 BP and *c*.7000 BP (*c*.9800 and *c*.7600 cal. yrs BP), as deep basal peats formed at Union Dock, Grimsby and Market Place, Hull, by which time sea levels were in the region of -9 m OD (Gaunt & Tooley, 1974; Dinnin & Lillie, 1995; Long *et al.*, 1998a).

Evidence for earlier peat formation is suggested by palynological analysis of basal peats from rivers in the southern Humberhead Levels and south-eastern Vale of York (Dinnin, 1997b; Lillie, 1997a and Lillie & Gearey, 1999). However, these deposits may represent peat formation due to waterlogging in isolated depressions in the pre-Holocene subsurface, and therefore, may not be directly related to regional base-level change. Subsequent channel and floodplain aggradation correspond to Gaunt *et al.*'s (1971) 'younger fluvial deposition phase' and resulted in the deposition of a range of alluvial deposits (see Figure 3.6). The factors influencing the spatial and temporal distribution of these deposits are discussed in section 3.5 below.

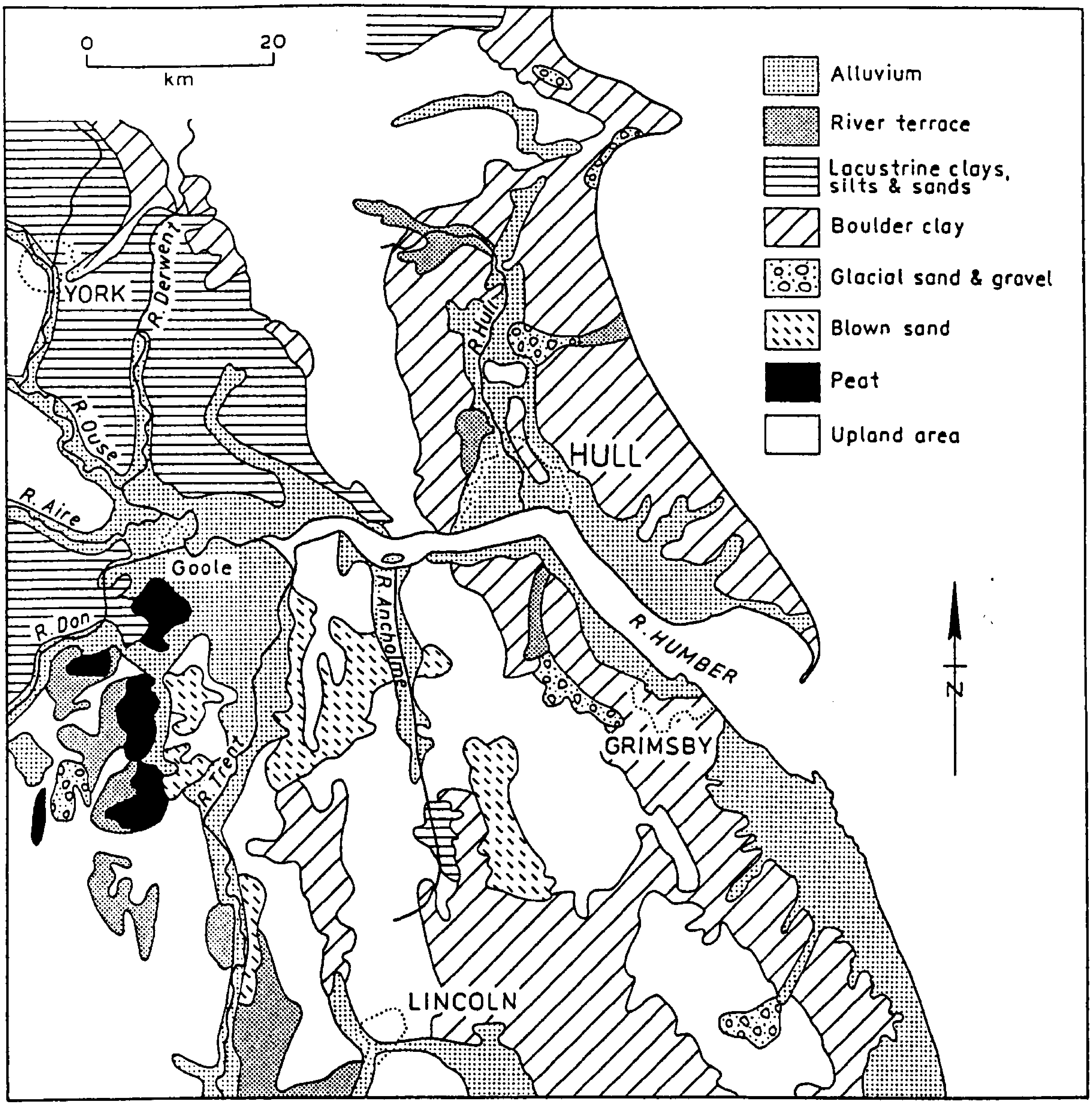


Figure 3.6 The drift geology of the Humber basin (after Van de Noort & Davies, 1993).

3.5 Holocene Landscape Development

Continued rising sea levels during the Holocene resulted in both channel and floodplain aggradation, through a combination of estuarine penetration and impeded freshwater drainage (Dinnin, 1997a). Gaunt (1994) states that river alluvium is up to 20 m thick locally where it fills incised river valleys and that peat mostly occurs flanking the deeply incised river courses at elevations between -4 and +3 m OD, although as pointed out above, deeper peat deposits occur at the base of deeper floodplains (see Figure 3.6). Despite this observation he later states that "adjacent to the Aire there is little evidence of concealed peat" (Gaunt, 1994; 127). Gaunt (1994) and Lillie (1997a) suggest that most of the alluvium within the Humberhead Levels and Vale of York is of freshwater origin, except the most eastern parts of the Ouse floodplain and northern parts of the Trent. The findings of this research offer further light on these observations.

Gaunt & Tooley (1974), Dinnin & Lillie (1995) and Long *et al.* (1998a) suggest that by *c.*5000 BP (*c.*6000 to *c.*5500 cal. yrs BP), mean sea-level (MSL) had risen to within a few metres of OD, by which point the river channels must have been largely infilled with accreted deposits (Gaunt, 1994). The date at which paludification (peat formation) was initiated on floodplains along the lower river courses depends on the depth of fluvial incision. This depth determines the time at which incipient sea level would begin to influence groundwater levels, thus, older ages may be anticipated for deposits lying at lower altitudes. Floodplain aggradation and peat formation are therefore likely to be time transgressive across the region, occurring initially in the bottom of deeply incised channels, and spreading outwards and upwards onto progressively higher ground (Dinnin, 1997a).

In the Ancholme valley at Brigg, Neumann (1998) has obtained a minimum age for basal peat formation at Brigg of 6170±90 BP (7225 to 6802 cal. yrs BP) between -4.59 and -4.62 m OD. Subsequent paludification began soon after on the adjacent valley floor with the development of fen carr on the floodplain. In the lower Aire valley at Airmyn, Lillie & Gearey (1999) have obtained a *terminus ante quem* for paludification of 6200±80 BP (7230 to 6883 cal. yrs BP) by dating wood macrofossils from basal organics between -5.87 and -5.97 m OD. Temporal differences in timing of peat inception on floodplains will also occur according to local variations in subsurface topography.

At Hasholme in the Foulness valley, peat was forming at -4.31 m OD by 5710±100 BP (6739 to 6298 cal. yrs BP) in response to raised water levels (Millett & McGrail, 1987). Evidence from the Humberhead Levels indicates that base levels had risen sufficiently to cause waterlogging on the river floodplains which initiated peat formation soon after this time. Dates from the base of the floodplain of the river Don at Thorne Waterside and Crowle of 4230±100 BP (4865 to 4575 cal. yrs BP) and 4230±70 BP (4874 to 4560 cal. yrs BP) and from the base of the river Idle at Misterton of 4330±120 BP (5296 to 4560 cal. yrs BP), indicate floodplain aggradation during the mid- to late-Holocene (Buckland & Dolby, 1973; Buckland & Sadler, 1985; Smith, 1985a).

Elsewhere in the Humberhead Levels, impeded run off led to the development of extensive fen and bog peatlands at Thorne and Hatfield Moors (Smith, 1958a; Buckland, 1979; Smith, 1985a). Two dates obtained from the base of these deposits at Hatfield (Smith, 1985a) suggest peat formation at 4180±70 BP (4863 to 4453 cal. yrs BP) and 4335±75 BP (5214 to 4654 cal. yrs BP), a similar time to elsewhere on the Humberhead levels. Peat development on the northern part of Thorne Moors gave dates of 4545±75 BP (5452 to 4878 cal. yrs BP) and 4515±70 BP (5442 to 4874 cal. yrs BP) but was somewhat later in the southern and western areas (Smith, 1985a), with radiocarbon dates of 3060±65 BP (3387 to 3069 cal. yrs BP) and 3475±65 BP (3891 to 3571 cal. yrs BP). Subsequent growth of these peatlands led to segregation from the influence of the local groundwater table and acidification. The details of palaeoecological research undertaken on these ombrotrophic raised mire deposits are dealt with in Chapter 4.

By *c.*3500 BP (*c.*4100 to *c.*3500 cal. yrs BP), it is suggested that MSL had risen to OD, and thereafter, fluctuated around this level (Gaunt & Tooley, 1974). It has been postulated that by this time, river channels had largely been infilled, which caused floodwaters to spread beyond the channel confines, resulting in overbank alluviation. Much of the late-Holocene floodplain aggradation has been attributed to accelerated soil erosion during the late-Roman period (Buckland & Dolby, 1973; Samuels & Buckland, 1978; Buckland & Sadler, 1985; Riley *et al.*, 1995). However, Long *et al.* (1998a) identify a significant change in estuarine dynamics prior to the Roman period, with an expansion of marine conditions occurring throughout the estuary during the late Bronze Age/early Iron Age, followed by a contraction of the estuary in the late Iron Age (between *c*.3200 and *c*.1900 cal. yrs BP). This sequence of events is corroborated by archaeological information from the Humber region (see Chapter 4 and Chapter 10), and provides an alternative hypothesis for floodplain alluviation during the late Holocene.

3.6 The Humber Estuary

3.6.1 Physical Characteristics

The Humber estuary originates on the south-eastern side of the Vale of York from the confluence of the Ouse and Trent, and flows eastwards between the Humber gap (a broad gap in the chalk escarpment of the Yorkshire and Lincolnshire Wolds between North and South Ferriby). The estuary widens considerably as it passes between the Holderness area and low-lying Lincolnshire Marsh before debauching into the North Sea between Spurn Point and Grimsby (Figure 3.2).

The Humber can be classified as a macro-tidal estuary (Davies, 1964), has a maximum tidal range of 7.2 m and a present tidal length of 140 km along the river Ouse (Pethick, 1990). The Humber estuary is well mixed and has a considerable freshwater drainage capacity that covers 24240 km, approximately one-fifth of the area of England (Pethick, 1990). Its tributaries extend to Birmingham in the south, the Pennine watershed in the west and Swaledale in the north, and pass over a variety of different geology (see section 3.3 and Figure 3.3). In high tidal range estuaries such as the Humber, the volume of water between high and low tide, **the tidal prism**, is large compared with the freshwater river contribution and low tide volume (Dyer, 1997). Major rivers such as the Ouse, Aire, Derwent, Trent, Ancholme and Hull contribute over 1.3×10^7 m^3 of water to the estuary each day (Pethick, 1990), which is small in comparison to the 16×10^7 m^3 of tidal flow (the tidal prism) which moves up and down the estuary during each tide.

An analysis of records of relative sea-level (RSL) change in east England (Shennan, 1992) notes possible differential uplift in north-east England of *c*.0.2 mm yr^{-1} since *c*.5000 BP (*c*.6000 to *c*.5500 cal. yrs BP), which is attributed to variations in the magnitude of isostatic rebound since the end of the Devensian. The

Humber estuary has experienced slow subsidence at a rate of <1 mm yr^{-1} since the mid-Holocene (Shennan, 1989).

3.6.2 Sediment Supply

The balance in contribution of sediment from terrestrial and/or offshore sources during the Holocene is an important consideration when addressing research questions relating to the extent to which prehistoric human activity influenced floodplain and estuary development. At present, there is a net surplus of sediment to the estuary, with suspended sediment concentrations as high as 2000 p.p.m during winter months, equating *c*.3 x 10^6 tonnes of total suspended sediment load at any one time (Pethick, 1990). This is due to the asymmetry of the tidal wave. For example, the flood tide rises at Goole in 2 or 3 hours, whereas the ebb tide takes 9 to 10 hours to fully recede (Heathcote, 1951; Pethick, 1988; 1990). Pethick (1988; 1990) suggests that the implications of this phenomenon are strong inland currents, which carry with them high suspended sediment concentrations, and a weaker ebb flow which carries much less sediment seawards, hence, a net input of sediment in the estuary from the North Sea.

The low cliffs of glacial diamict along the Holderness coast form an important source of coastal sediment and are currently eroding at a rate between 1 and 2.5 m yr^{-1} (Furness, 1985), which is among the highest rates of coastal erosion in the world (Valentin, 1971). Based on an average retreat rate of 2 m yr^{-1}, Sheppard (1912) suggests retreat of the Holderness coast of 4 km since Roman times (*c*.2000 BP, *c*.2300 to *c*.1700 cal. yrs BP). Littoral drift along the outer coast is north-south, which results in longshore drift, delivering sediment to the Lincolnshire and north Norfolk coast (Sheppard, 1912; Steers, 1964; Evans, 1965). Associated with this littoral drift is the development of Spurn Point, an ephemeral feature of the outer estuary built out across the mouth of the Humber (de Boer, 1964; 1978; 1988). From an analysis of historical data, de Boer (1978) suggests that the evolution of this feature is of a cyclical nature. Every 250 years or so, the Holderness coastline retreats to a critical position where it no longer offers the necessary shelter or protection for the spit, which then is breached by waves and gets washed away, only to reform inside the position of its predecessor.

The extent to which the eroded coastal sediment from Holderness, which feeds the east coast of England, also provides the majority of suspended solids in the Humber estuary is a matter of debate. The assumption that the high sediment concentrations of Humber tidal waters is due to the inwash of material from Holderness was initially challenged by Wheeler (1901). He postulated that at best, this assumption could only apply to the Humber itself (the 60 km stretch between the confluence of the river Ouse and Trent at Trent Falls and the sea), and that within the tidal rivers such as the Aire, Ouse and Trent, the sediment is probably derived from the river catchment and bank erosion. His argument was based on the nature of the tidal asymmetry described above, which implies that sediment laden water at Spurn could only reach a maximum of *c*.50 km up river, before the ebb tide, reinforced by freshwater discharge, carried it back out further into the North Sea. This model was over-simplistic in many ways, not least because it does not recognise the possibility of sediment deposition and remobilisation at consecutive tides, enabling steady successive accumulation up-estuary. Indeed, Wheeler himself toned down his argument somewhat as he later withdrew his (1901; 568) statement that "it was physically impossible" for the mud in the Humber to be supplied from the sediment of the Yorkshire coast and conceded that it was possible that some of the alluvium washed from the Holderness coast goes into the Humber and has been deposited on Sunk Island (Wheeler, 1906 in Sheppard, 1912). Thus, what was "physically impossible" in 1901, became 'possible' in 1906!

Sheppard (1912) opposed Wheelers' (1901) views, as he observed clear river water during ebb flows and 'coffee-coloured' muddy water associated with the rising tide, and suggested most of the sediment was derived from sea rather than river sources. However, this scenario is more likely to be a result of the stronger velocity of the flood tidal wave, which rises rapidly, and stirs up sediment in the estuary which, according to Pethick (1988), enables the estuary to remove and carry more sediment for deposition up-estuary during the flood tide than the ebb. Sheppard (1912; 248) provides an amusing paradox in the context of this early debate, when remarking upon the failed attempts to help a stranded whale out of the estuary with the tide, which kept returning, even after being towed out to the North Sea: "If a dead whale could not be kept out of the estuary, how is it possible to make us believe (as some endeavour to) that no fine particles of mud can reach the Humber from outside?"

Recently, Al-Bakri (1986) has stated that sediment supply to the Humber estuary is dominated by material supplied from the eroding cliffs of the Holderness coast, with only limited inputs from the catchment. Chang & Evans (1992), however, identify similarities between offshore sands in the North Sea and littoral sediments on the east coast, suggesting important onshore sediment transport from this source. This is partly supported by Pethick (1988) who notes that most of the sediment in the estuary comes from the North Sea rather than from rivers, but that just how much of this material is derived from river or sea sources is not known. Pethick (1988) presents an estimated value for sediment deposition in the Humber in one year of *c.*63400 tonnes. This sediment is made up of varying proportions from inland and offshore sinks. Pethick (1988) also estimates that *c.*200000 tonnes of sediment are derived from river sources (but much of this is simply on its way out to the North Sea) and that 1.4 x 10^6 tonnes of material is contributed to the North Sea by the Holderness cliffs each year, most of which enters the 'giant mixing bowl' of the North Sea and does not enter the Humber in the short term. Less sediment from river sources is likely to be deposited in the estuary because fresh water floats over the surface of the more dense saline flow, so terrigenous material tends to be carried out to sea (Pethick, 1988).

However, the extent to which present sediment loads in the Humber compare with those of earlier Holocene periods is questionable, considering the greater vegetation cover and limited prehistoric human activity at the coast, and the ability of estuaries to expand and accrete freely in the past compared to their restricted and constricted nature today (Plater *et al.*, 1999). This has caused tidal prism reduction over the past 100 years due to a gradual decrease in estuary width (of 14%) and volume associated with sediment accretion in the channel (Wilkinson *et al.*, 1973; Pethick, 1990). Similarly, the relative contribution of each sediment source is likely to have changed over the Holocene, with significant amounts likely to be derived from the (now submerged) continental shelf as the rising Postglacial sea transgressed its surface during early to mid-Holocene times, compared with an acceleration in terrigenous input to the system due to agricultural intensification from the Bronze Age onwards (see Long *et al.*, 1998a).

3.7 Drainage

At present, much of the low-lying land around river courses and flat plains of the southern Vale of York and Humberhead Levels lies well below the contemporary level of Mean High Water of Spring Tides (MHWST), which is +4.30 m OD at Goole. This area is only kept from flooding today by considerable reinforcement of the river banks with artificial levees. Early attempts at wetland drainage and reclamation probably date back to between the Roman and early-Medieval periods (Palmer, 1966; Gaunt, 1975; Buckland & Sadler, 1985; Jones, 1995), as in many other British lowland areas (Fulford *et al.*, 1997) such as the Severn Estuary (Rippon, 1996). However, the earliest documentary and cartographic evidence for drainage improvements in the Vale of York and Humberhead Levels region date from the 11th century, when areas of saltmarsh which were reclaimed along the lower river Ouse *c.*1086 AD (Sheppard , 1966).

Palmer (1966) suggested that along the inland clayey river margins where flooding is prolonged, damp *Quercus* fenwoods and *Alnus* carrs may have persisted until Romano-British times. Indeed, such areas were not extensively settled until after 1150 AD (Palmer, 1966). From then on, until 1250 AD, dykes were built on top of natural levees, and marsh areas behind the river banks were dewatered by a series of drains, enabling settlement and agricultural activities along the tidal river margins (Saltmarshe, 1920). However, such early attempts at drainage were often ineffective. Excluding the tidal river water and draining precipitation and run-off from the land are just two problems to overcome, and the seasonally flooding rivers that traversed across flat ground emptying into tidal estuaries also had to be considered. Medieval settlers found it much easier to keep the sea out than to control swollen rivers (Palmer, 1966; de Boer, 1988), and floods were still a regular occurrence.

The first serious attempts at large scale and effective drainage alterations in the Humberhead Levels area were begun by the Dutch engineer Cornelius Vermuyden in the 17th century (Dinnin, 1997c). Prior to 1626 AD, it has been estimated that the low lying land in this area consisted of huge areas of fen carr and raised bog, and that up to 60000 acres (24000 hectares) were regularly or permanently under water (Van de Noort & Davies, 1993). Most of the natural river courses that drain the Humberhead Levels have been greatly altered by dredging schemes to deepen the channels, in addition to the construction of sluices, artificial drains, and ditches during the mid- to late-Medieval period. Rivers were canalised into straight courses confined by embankments and often drastically re-routed to lead the rivers away from flat areas or shorten their courses. For example, the north branch of the river Don (the Turnbrigg Dike) was blocked near East Cowick (see Figure 3.1), where it flowed into the river Aire at the study site at Eskamhorn as a result of an earlier unsuccessful diversion by Vermuyden (Gaunt, 1975; de Boer, 1988). A right angled turn was engineered forming a canal (the Dutch River), which flows into the Ouse at Goole, and receives drainage from Thorne and Hatfield Moors (Figure 3.1). This new channel was completed in 1635 AD (Gaunt, 1975). Pumping stations were also widely constructed so water could outflow at all stages of the tide.

By the end of the 18th century, the majority of land around the Humber was drained in the sense that tidal river water was effectively excluded, and winter floods largely controlled. Nevertheless, the ground was so low, and water table so close to the surface, that the soils were waterlogged practically all year round in the river valleys, and useless for arable agriculture (Palmer, 1966). The solution to this was to raise the surface level of the land, using a technique called **'warping'** (Heathcote, 1951). The earliest record of flood-warping is from the 1730s AD, near Rawcliffe (Gaunt, 1994), and it was practised extensively throughout the lower tract of the Ouse up to Barmby on the Marsh, and along the Aire as far as Temple Hirst and Hirst Courtney (Palmer, 1966). Between 1800 and 1860 AD, huge tracts of land were warped along the river Aire, Ouse and Trent, and around Thorne Waste (see Figure 3.7).

Warping utilises the high suspended sediment load characteristic of the Humber tributary rivers and involves the controlled and repeated flooding of land. The area to be warped is embanked and, at high tide, a sluice is opened allowing sediment charged tidal waters to flow into the area along a warping drain. The sluice is then closed as the tide subsides allowing the fine sand, silt and clay to settle out and at low water the water is allowed to drain gently back into the river. Evidence for this practice is often masked by subsequent ploughing but typically the flood-warped ground is silty with traces of laminations. Each flood deposited up to 2 mm of silt or clay and almost 1 m could accumulate in a year (Heathcote, 1951). In some areas, the depth of the warp exceeds 3 m in thickness (Van de Noort & Davies, 1993). Flood-warped ground is also identifiable by distinct differences in field level either side old embankments and drains (see Allen, 1993; 1996 for analogous examples). Reviews on warping practice, and other technical details are provided by Heathcote (1951), Ellis (1990), and Lillie (1997b).

In addition to raising the level of the land, and making it less susceptible to inundation, warp soils were very fertile and resistant to wind erosion, which led to dramatic increases in the amount of agricultural land (Metcalfe, 1960). Further drainage improvements, which enhanced agricultural propensity, were made during the early 19th century by the introduction of tile drains (Palmer, 1966). These helped to prevent the accumulation of surface water, and subsequent development of surface water gleying (Heathcote, 1951).

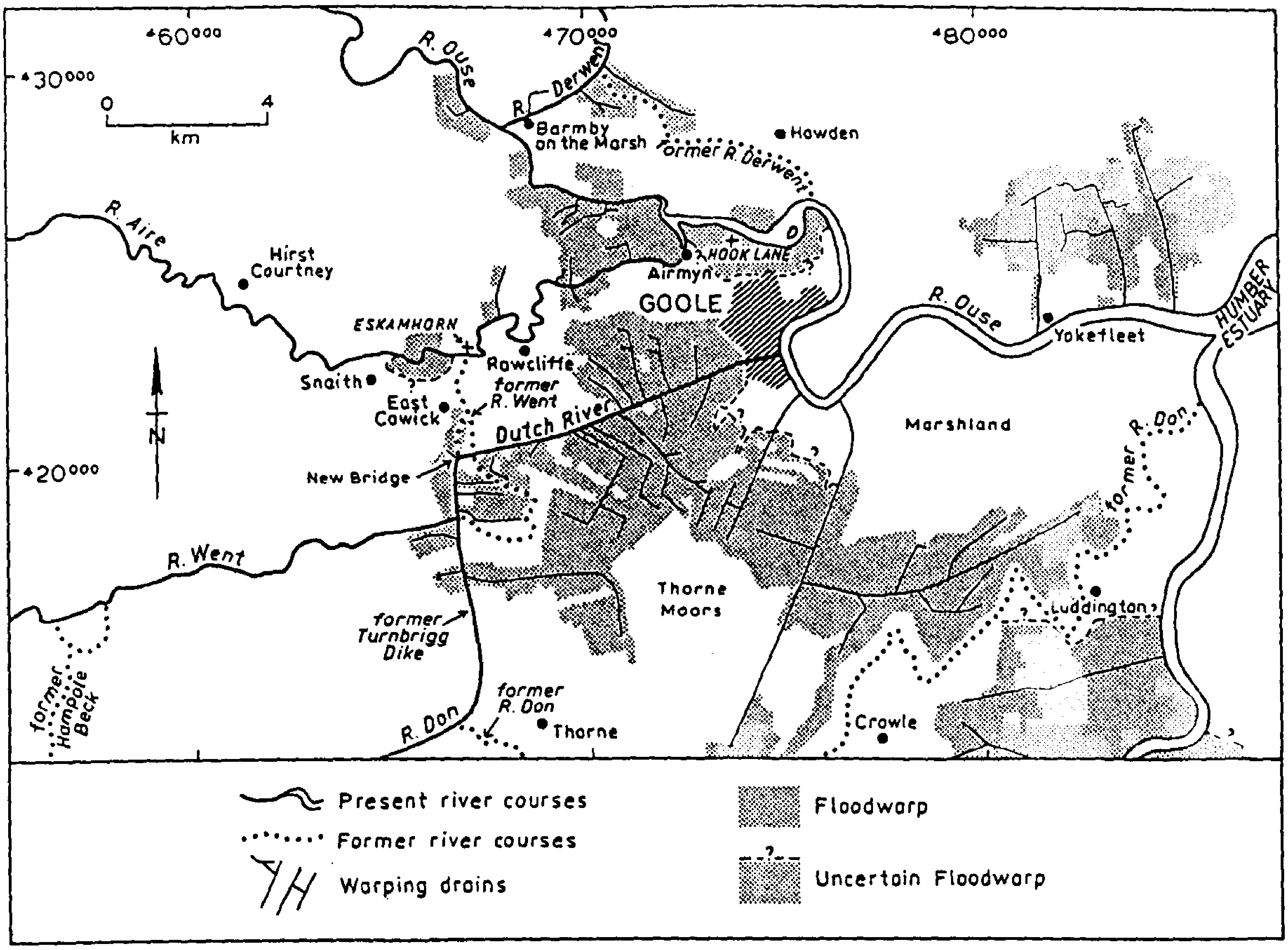


Figure 3.7 The distribution of warped land in the Humberhead Levels and southern Vale of York (after Gaunt, 1994).

3.8 Land-Use

Apart from the main urban centres around Doncaster, Selby and Goole, the dominant landuse is agriculture (Middleton, 1997; 1999). Arable crops account for 83% of the agricultural land-use in Humberside (Symes, 1987). In the Selby District, 80 to 90% of the land area of the parishes which encompass the study area (Hirst Courtney, Carlton, Drax and Newland), are taken up by arable land (Warburton, 1989; Middleton, 1999). Although this is dominated (*c.*75%) by cereal crop cultivation such as wheat and barley, there is some grassland pasture which reaches coverage of up to 5% in the immediate study area (Middleton, 1999). Other more intensive farming methods occur in the Selby District, such as poultry and pig rearing, a mushroom farm and even a maggot hatchery (Warburton, 1989). Cultivation of horticultural crops such as oil seed rape, sugar beet, beans, potatoes and peas are subordinate, and whilst there are variations with some areas reaching almost 20% coverage, this usually makes less than 5% of the total (Middleton, 1997; 1999).

The exploitation of extensive coal-bearing strata in the Yorkshire region had enabled large scale industrial and urban development, concentrated in areas such as Doncaster and Selby. Several large coal fired power stations are prominent in the Humberhead Levels and Vale of York region, which are supplied by coal from the Selby coalfield. The largest coal-fired power station in western Europe is situated at Drax, just 3 km to the north of the Eskamhorn study site (Figure 3.1).

Several notable areas of natural landscape remain in the area. The peat bogs of Thorne and Hatfield Moors cover an area of *c.*3300 ha (Middleton, 1997) and retain some original wetland characteristics despite a long history of peat extraction (Smart *et al.*, 1986; Eversham, 1997). Certain areas of this landscape are now protected, with 1400 ha of Hatfield Moors designated as SSSI, and parts of Thorne Moors are leased as a National Nature Reserve (Middleton, 1997). Elsewhere, in the Vale of York, areas of lowland heath and fen survive, such as Skipwith Common (SSSI) and the NNR, Askham Bog (Van de Noort & Ellis, 1999). The existence of woodland is extremely fragmentary, with values in the order of 1%, but as low as 0.1% near Thorne (Middleton, 1997).

3.9 Soils

Soils develop in response to natural factors such as geology, relief, climate and vegetation (Jenny, 1941), as well as a range of anthropogenic factors (Ellis, 1990). The soils within the Humberhead Levels and southern Vale of York closely reflect the local geological and drainage conditions (Ellis, 1997). The soils that have developed from the alluvium contained within the low-lying river valleys studied, are either warp soils, or ground- and surface-water gleys, reflecting the influence of low, flat relief, high water table, and poor to moderate drainage of the periodically saturated clay substrate (Ellis, 1990; 1997). The blue-grey-brown, and often orange mottled nature of these soils, reflects the iron content and redox conditions within the ground, and are typical of alluvial deposits (Dury, 1964). Totally blue-grey clay soil reflects permanent saturation and a total lack of oxygen, so that all the iron in the soil is in reduced form. More characteristic however, is a brown-grey mottled colour, reflecting a lower and seasonally variable water table, causing periodic saturation, with iron occurring in both reduced and oxidised forms (Ellis, 1990; 1997). Such soils are potentially very fertile but require artificial drainage for successful arable use, and are also used for grazing pasture (Ellis, 1990; 1997).

Occasionally, brown earth soils develop on more freely drained land, beyond the margins of the floodplain and on slightly higher ground where there are gentle to moderate slopes (Ellis, 1990; 1997). Brown earth soils occur around Goole, are highly fertile, and therefore used almost exclusively for arable farming (Ellis, 1997)

3.10 Climate

Low lying areas in the extreme east of England generally receive less rainfall than elsewhere in northern Britain with mean annual precipitation ranging from 600 to 700 mm in the areas close to sea level, rising to 800 mm on the Yorkshire Wolds (Ellis, 1990). Mean annual temperatures vary with altitude in a similar manner, ranging between 8 and 10°C, with median accumulated temperature above 0°C between January and June, ranging from 1250 day-degrees on higher ground, to over 1400 day-degrees in low-lying areas (Jarvis *et al.*, 1984). This means that the Humberhead Levels and southern Vale of York, are 2 to 4°C warmer, and up to 1000 mm drier than higher areas to the west, making it one of the driest regions in England (Ellis, 1997).

Chapter 4 The Humber Wetlands

4.1 Introduction

This chapter is a review of the literature relating to Holocene palaeoecological research undertaken within the Humber basin. Research projects employing palaeoecological techniques have been carried out from a variety of different depositional contexts, within a diverse range of research fields. Examples are given from work where the main aim was not to define vegetational history, but concerned environmental archaeology, or sea-level change and coastal geomorphology.

The results of such investigations often reveal the nature of the local vegetation in the immediate vicinity of the study area. A review of all the palaeobotanical research is necessary to enable regional trends to be identified. Mere sites from the Holderness region, and a site situated on the eastern side of the Wolds are also considered as they may reflect a different pattern of upland vegetation distribution, associated with the Chalk geology.

The majority of sites are from lowland regions, a natural bias attributable to the topography of glacial drift deposits. The development of wetlands, and their vegetation associations, are dealt with in sections, according to geographical distribution throughout the region. Most detail is given to sites in close proximity to the study area. An analysis of the evidence for sea-level change in the Humber estuary, and the influence of sea-level change on landscape evolution is also summarised. Comprehensive reviews of much of the information summarised here can be found in Van de Noort & Davies (1993) and Van de Noort & Ellis, (1995; 1997; 1998; 1999).

4.2 Vale of York

Two sites occurring within till depressions in the Vale of York moraine (at Tadcaster and Askham Bog) provide most of the data concerning vegetation history of the Lateglacial and Holocene periods. These small lake basins infilled and progressed through a successional sequence of fen peats into *Sphagnum* and *Sphagnum-Eriophorum* raised mire. The deposits studied by Bartley (1962) from a site near Tadcaster provide an undated early pollen sequence, which is complemented now by further analysis of similar deposits from Askham Bog which have been radiocarbon dated (Gearey & Lillie, 1999, see Figure 3.2).

Bartley (1962) provides details for the Lateglacial vegetation, which reflects climatic amelioration from *c.*13000 BP (*c.*16300 to *c.*14500 cal. yrs BP), with an open landscape dominated by Poaceae and Cyperaceae, giving way to scrub *Betula* woodland. A short climatic oscillation is recorded which is consistent with the Bölling-Allerod interstadial complex of northern Europe (Mangerud *et al.*, 1974), and is also a feature of Beckett's (1981) pollen diagram from Roos Bog (see section 4.4).

The Lateglacial interstadial (Windermere) is represented at Tadcaster by rising *Betula* pollen frequencies, and decreasing values for *Juniperus*, and many of the herb pollen types, due to the increasingly shaded conditions (Bartley, 1962). Similar conditions are recorded from the earliest sediments at Askham Bog (Fitter & Smith, 1979). The arctic conditions of the Loch Lomond/Younger Dryas stadial are characterised by rapidly declining *Betula* tree pollen, with a corresponding rise in *Betula nana*, Cyperaceae, *Artemisia*-type, and *Thalictrum* (Bartley, 1962). Similarly aged deposits have also been reported from pollen spot samples from the valleys of the river Went and Idle in the Humberhead Levels (Dinnin, 1997b; Lillie, 1997a, see section 4.3) and from pollen diagrams in the Holderness area (Beckett, 1981; Walker *et al.*, 1993, see section 4.5).

The opening of the Holocene period is registered at Tadcaster by a very sharp rise in *Betula* pollen, followed by increases in *Corylus avellana*, *Pinus sylvestris*, and *Ulmus* (Bartley (1962), at the expense of light demanding shrubs and herbs such as *Betula nana* and *Thalictrum*. This latter expansion of thermophilous trees is also recorded at Askham Bog, where Gearey & Lillie (1999) date a *Corylus* *avellana* dominated landscape to 9150±55 BP (10289 to 9985 cal. yrs BP). High frequencies of *Corylus avellana* also characterise similar pollen zones at Tadcaster (Bartley (1962) and in the southwestern Vale of York at Burton Salmon (Norris *et al.*, 1971, see Figure 3.2).

Considerable inter-regional differences occur in the representation of woodland habitats from the early-Holocene, perhaps reflecting site specific conditions, although comparison is hampered by the lack of radiocarbon dates in the early work by Norris *et al.* (1971) and Bartley (1962). Gearey & Lillie (1999) date an increase in *Quercus* and *Ulmus* to 8605±50 BP (9797 to 9454 cal. yrs BP), reflecting the continued spread of these woodland taxa into the area surrounding Askham Bog, with *Corylus avellana* still remaining an important taxon. However, Norris *et al.* (1971) and Bartley (1962) show *Betula* and *Pinus sylvestris* to be important components of the vegetation at this time, taxa which only reach subordinate levels at Askham. In general, the main woodland species characteristic of the early-Holocene include *Quercus*, *Ulmus*, *Corylus avellana*, and *Pinus sylvestris*, confirming that these were the typical arboreal components of the vegetation in this region (Norris *et al.*, 1971; Bartley, 1962; Beckett, 1981; Gilbertson, 1984a; Gearey & Lillie, 1999).

The *Alnus glutinosa* rise is a prominent feature of British vegetation history, and is recorded at Tadcaster (Bartley, 1962) and Askham, where it is dated to 7720±50 BP (8557 to 8372 cal. yrs BP, Gearey & Lillie, 1999). Although this is a feature of many pollen diagrams in the Humber region, the only other date for this event prior to this study is at Bole Ings on the Trent (see Figure 3.2 and section 4.4), where a date of 6290±70 BP (7361 to 7012 cal. yrs BP) has been obtained (Dinnin & Brayshay, 1994; Brayshay & Dinnin, 1999). Although the *Alnus glutinosa* rise is known to have been a metasynchronous phenomenon (Birks, 1989), and the date from Askham Bog is within the expected age range (Bennett & Birks, 1990), Bayliss *et al.* (1999) conclude that this Askham date may to be too old by an unknown amount due to hard water error.

A *Tilia* rise is also evident at Askham Bog and has been dated to 6310±45 BP (7271 to 7092 cal. yrs BP), which is almost identical to the date for this event at Bole Ings (Dinnin & Brayshay, 1994; Brayshay & Dinnin, 1999) where it coincides with the *Alnus glutinosa* rise. This suggests that the *Alnus glutinosa* rise at Askham may in fact be an early expansion of this taxa and not a reflection of an erroneous date. Bush

& Hall (1987) have present evidence which lends support to the Askham date, as they suggest that *Alnus glutinosa* was present from the Lateglacial and early-Holocene times in East Yorkshire. The rise in *Tilia* at Askham seems to replace *Pinus sylvestris* and *Betula*. The dry land vegetation is typical of lowland woodland during the mid-Holocene with *Corylus avellana*, *Quercus*, *Tilia* and *Ulmus*, and scattered *Fraxinus excelsior* (Gearey & Lillie, 1999).

Subsequent events at Askham Bog, such as the *Ulmus* decline and some perturbations in tree pollen curves, occur in conjunction with herbs indicative of open ground. This may be interpreted as palynological evidence for human activities. Later, further clearance of *Corylus avellana*, *Tilia*, and *Pinus sylvestris* occurs, along with herbs such as Chenopodiaceae, *Plantago lanceolata*, *Urtica*, and Cereals (Gearey & Lillie, 1999). Unfortunately, all of the dates constraining such events are extremely unreliable, with some samples giving modern ages. Whilst no obvious disturbance to the pollen sequence is noticeable, the biostratigraphic record must be treated as suspect in light of this (Bayliss *et al.*, 1999).

Lillie (1997a) and Lillie & Gearey (1999) recently investigated several sites within the Aire and Ouse valley as part of the HWP survey. Considerable thicknesses of floodplain peat were uncovered within these river valleys, which palynological evidence suggest formed throughout the region during the early- to mid-Holocene. Floodplain peat initiation is tentatively ascribed to the late-Mesolithic period (*c.*6000 BP, *c.*7200 to *c.*6600 cal. yrs BP) due to the presence of *Ulmus* values of 3% TLP (Lillie & Gearey, 1999), at Birkin, east of Tadcaster, along the river Aire (see Figure 3.2). Further down-river at Carlton, floodplain peat formation is indicated by the late-Mesolithic, although these ages are also based on spot pollen samples, and must be treated with caution in the absence of an independent radiocarbon chronology. A pollen sample from the south side of the river Aire floodplain at East Cowick (Lillie, 1997a) indicates a post-*Ulmus* decline age, although because the sample was not taken from the deepest part of the floodplain, this represents a minimum age for peat development at this site (see section 8.2). The lowermost site investigated by Lillie & Gearey (1999) is situated at Airmyn in the Aire valley. A date from *Alnus glutinosa* macrofossils of 6200±80 BP (7230 to 6883 cal. yrs BP) is obtained from the central area of the floodplain, at *c.*-5.90m OD, which provides a minimum age for peat development at this site. A pollen spot sample from *c.*+0.30 m OD from a borehole situated on the north edge of the floodplain indicates a local environment consisting of *Alnus-Quercus* fenwood around 3650±80 BP (4223 to 3721 cal. yrs BP). Although this precision of this age is poor (the sample consisted of a combination of *Alnus glutinosa*, *Quercus*, *Fraxinus* and *Rosa* wood macrofossils obtained from the base of this core which are unlikely to be contemporaneous), it suggests that the gradual spread of peat formation within the Aire valley at Airmyn occurred over three millennia (Lillie & Gearey, 1999).

Close to the study site at Hook Lane (see Chapter 9), Smith (1958a) has constructed a short pollen diagram from an undated peat profile situated within the northern limits of Goole itself (see Figure 3.2). The pollen spectra obtained indicate *Alnus* fen wood with *Quercus*, but towards the top of the sequence, these taxa are replaced by *Betula* and *Pinus sylvestris*, with Ericaceae and *Sphagnum* equally dominant. Smith (1958a) correlates these deposits with similar deposits from South Ferriby and Redbourne Hayes in the Ancholme (see section 4.4) that exhibit similarly high percentages of *Betula* (50% of the total tree pollen), and are thought to be Romano-British in age. No radiocarbon dates are available to verify this correlation.

Elsewhere in the southern Vale of York, work has been undertaken to provide a palaeoenvironmental setting for the Hasholme and North Ferriby logboats (Millett & McGrail, 1987; Wright, 1990), and with reference to sea-level change (Long *et al.*, 1998a). Turner (1987) studied the palynological properties of the peat deposit below the Hasholme logboat in the Foulness valley (see Figure 3.2), and has shown that *Alnus* carr dominated the wetland environment in the Foulness valley from 5710±199 BP (6739 to 6298 cal. yrs BP), with *Filipendula*, Cyperaceae, and Poaceae also prominent. Lillie (1999) suggests poor fen carr, dominated by *Betula* and *Alnus glutinosa* was prominent within Walling Fen at this time. It is suggested that the dryland vegetation along the periphery of Walling Fen supported a mixture of *Quercus*, *Corylus avellana*, *Betula*, and *Tilia* during this time (Turner, 1987). A reduction in *Tilia* and perturbations in the *Corylus avellana* curve are interpreted as possible evidence for woodland clearance and management (i.e. coppicing) respectively. Evidence for a rising water table is evident towards the top of the peat, as frequencies of Cyperaceae, *Typha angustifolia*, and Poaceae pollen rise, and *Alnus glutinosa* pollen values decline. *Frangula alnus* and *Fraxinus excelsior* also increase at this point, which Turner (1987) interprets as an opening up of the fen carr associated with the increasing water levels. The transgressive contact with the overlying estuarine clay at -1.85m OD, within which the logboat was sealed, is dated to 2530±70 BP (2764 to 2335 cal. yrs BP), although the truncated nature of the sedimentary boundary renders it inadequate as a reliable sea-level index point (Jordan, 1987).

A more complex suite of deposits have been uncovered from Sandholme Lodge in the Foulness valley (see Figure 3.2) by Long *et al.* (1998a). Up to three peat layers are identified, separated by clastic sediments, which indicate both freshwater and estuarine environments. Peat underlying the middle minerogenic horizon is shown by pollen analysis to have formed under freshwater conditions. *Alnus glutinosa* pollen dominates, with *Quercus*, *Tilia*, and *Corylus avellana* prominent dryland arboreal taxa (Long *et al.*, 1998a). Saltmarsh conditions are indicated at the top of the peat which is dated to 5615±45 BP (6484 to 6301 cal. yrs BP). Overlying the clay is the uppermost peat, which began forming in a saltmarsh environment at 4170±45 BP (4835 to 4534 cal. yrs BP). After this time, it is apparent that *Pinus sylvestris* became an important contributor to the dry land pollen rain, along with *Quercus* and *Corylus avellana*, with each taxa recording values of *c.*15% TLP (Long *et al.*, 1998a).

A number of palaeoenvironmental studies have been undertaken from intertidal peat outcrops on the shore of the Humber estuary, in the south-east corner of the Vale of York. Hulme & Beckett (1973) analyse the pollen from two peat horizons separated by clay, which they believe to be of estuarine origin, at Faxfleet (see Figure 3.2). The lower peat unit suggests formation initially in an *Alnus* carr, followed by a reedswamp environment containing Chenopodiaceae pollen, which the authors (rightly) attribute to saltmarsh rather than agricultural origin. The dryland vegetation is dominated by *Quercus* woodland, with *Tilia, Corylus avellana*, and *Pinus sylvestris* of secondary importance. The upper peat formed in a reedswamp, with *Alnus glutinosa* in close proximity, and *Quercus* within the surrounding woodland. Although no radiocarbon dates were obtained, the authors attribute the lower and upper peats to early- and late-Bronze Age respectively, due to stratigraphic similarities with radiocarbon dated deposits at Chapel Point in Lincolnshire (Smith, 1958a) and North Ferriby (Wright & Churchill, 1965).

The sediments at North Ferriby have been studied in detail due their association with the three Bronze Age boats found within estuarine clays at Ferriby (Wright, 1990). Buckland *et al.* (1990) have analyse pollen and insect remains from peat deposits directly below the boat finds, which reveal a wooded landscape consisting of *Quercus, Corylus avellana, Tilia, Ulmus*, and *Pinus sylvestris*. The peat itself formed in a saltmarsh context due to rising sea level. High representation of *Alnus glutinosa* and *Quercus* reflect fen carr environments fringing the estuary. Macrofossil remains from the peat suggest that both *Alnus glutinosa* and *Tilia* were growing in close proximity to the site (Wright & Churchill, 1965).

Good evidence for sea-level change is preserved in the deposits at North Ferriby, but these analyses merely establish the intertidal origin of the sediments, and provide minimum ages for the onset of positive sea-level tendency. Relative rise in sea level pre-dates *c*.3100 BP (*c*.3500 to *c*.3000 cal. yrs BP), the age obtained from *Alnus glutinosa* wood embedded in estuarine clay beneath a prehistoric boat (Wright & Churchill, 1965). Similar ages are obtained from wooden rails found in the foreshore at Melton (Crowther, 1987), which date to 2910±90 BP (3337 to 2792 cal. yrs BP) and 2990±70 BP (3356 to 2948 cal. yrs BP), and from a wooden paddle recovered from an analogous clay at North Ferriby, dating to 3062±31 BP (3353 to 3166 cal. yrs BP, Bayliss & Dinnin, 1995).

More recently, Long *et al.* (1998a) have studied stratigraphically similar deposits at East Clough, Melton and provide a more precise chronology for an analysis of sea-level change (see Figure 3.2). Pollen data reveal similar results to Buckland *et al.* (1990), with *Quercus* and *Alnus glutinosa* dominating the lower part of the peat indicating fringing fen carr, and Poaceae occurring in high frequencies towards the middle of the sequence, reflecting local saltmarsh and reedswamp environments. The combined litho- and biostratigraphic data presented by Long *et al.* (1998a), indicate a continuous positive sea-level tendency, which initiated peat accumulation at 3770±45 BP (4268 to 3983 cal. yrs BP), resulting in its eventual inundation at 3640±45 BP (4084 to 3831 cal. yrs BP).

4.3 Humberhead Levels

The vast expanse of former wetland known as Thorne and Hatfield Moors (see Figure 3.2) have been the source of detailed information regarding climatic change and the vegetational history of the region (Buckland & Dinnin, 1997; Dinnin, 1997a). Smith (1958a) studied a pollen profile from Hatfield Moors where bog peat growth was initiated directly above the nutrient poor basal sand deposits. Arboreal pollen is initially dominated by *Betula*, *Pinus sylvestris*, and *Quercus*, with *Alnus glutinosa* and *Corylus avellana*-type pollen abundant throughout. Pollen profiles from the later work by Smith (1985a) suggests a similar wooded environment, with *Calluna* and *Sphagnum* well represented in the basal layers. This confirms that the basal amorphous, highly humified black peat, is associated with raised mire conditions. This type of peat is rare at the base of raised mires where, under typical situations, organic lake muds or fen peats are more usual (Godwin, 1975a). Smith (1958a) suggests that peat formation took place in a manner analogous to blanket peat (*cf.* Moore, 1993).

Smith (1958a) identifies four flooding horizons at Hatfield on the basis of large variations in humification of the *Sphagnum* and *Scheuchzeria* peat in the stratigraphy. These are correlated with changes in the non-arboreal pollen biostratigraphy; drier phases are characterised by ericaceous heath and wetter phases dominated by *Sphagnum* spores. Smith (1958a) demonstrates that these wet shifts can be correlated on a lithostratigraphic basis across his bog transect. However, he refrains from naming these horizons, or giving them any widescale climatic significance, in view of the then recent work of van Zeist (1954), who has shown that recurrence surfaces need not be contemporaneous, and that dates differing by over 1000 years can be obtained for a single wet shift, even in the same bog.

Evidence for agricultural activity is prominent towards the top of the pollen diagram, and reaches maxima during the dry periods, when human activity is presumed to have been most extensive. Smith (1958a) postulates that two major forest clearance episodes were associated with Iron Age and Romano-British human activity. This latter period was followed by a decline in agricultural activity and woodland regeneration. The coincidence of this with one of the four 'flooding horizons' led Smith (1958a) to postulate a loss of agricultural land during the post-Roman period due to flooding. However, in the absence of radiocarbon dates, fitting a chronology to this pollen sequence to support such interpretations is problematic. Smith's (1958a) study agrees with the earlier findings of Erdtman (1928), that peat growth is pre-*Ulmus* decline in age on Hatfield Moor (Zone VIIa of Godwin, 1940b), and initiated during 'early to late-Atlantic times' (his terminology). However, subsequent radiocarbon dates by Smith (1985a) contradict these findings, as the basal deposits reveal later dates of 4180±70 BP (4863 to 4453 cal. yrs BP) and 4335±75 BP (5214 to 4654 cal. yrs BP).

Thorne Moors were also the subject of early research. Turner (1962) uses a pollen diagram from Thorne to illustrate her argument for an anthropogenic cause for the *Tilia* decline, as opposed to climate change (Turner, 1970). She concludes that two *Tilia* declines are evident, which are attributed to the effects of Bronze Age and early-Iron Age activities. She dates a 'primary decline' to 3179±115 BP (3680 to 3078 cal. yrs BP) and 2942±115 BP (3375 to 2782 cal. yrs BP) and a later 'secondary decline' to 2329±110 BP (2728 to 2061 cal. yrs BP). Her work also draws attention to the structure of the woodlands surrounding the bog prior to this time, which at Thorne were dominated by *Tilia* and *Corylus avellana*, as well as woods with significant quantities of *Ulmus*, *Fraxinus excelsior* and *Quercus*, and locally, *Betula* and *Alnus glutinosa*.

The lowest level of peat on Thorne Waste consists of fen deposits which are of archaeological interest, owing to the presence of a wooden trackway buried within this unit (Buckland & Kenward, 1973). Buckland (1979) studied part of this trackway, comprised mainly of *Betula* and *Quercus* wood, and analysed the surrounding peat for beetle remains. A late-Bronze Age date of 2980±110 BP (3393 to 2851 cal. yrs BP) is obtained from a *Pinus sylvestris* trackway timber (Buckland (1979), which agrees closely with a later date for a *Quercus* timber from the trackway horizon, obtained by Smith (1985a), of 2900±65 BP (3215 to 2855 cal. yrs BP). Both these dates are similar to the date for the primary *Tilia* decline of Turner (1962). The construction of this trackway was probably a response to expanding mire conditions, and increased wetness of the bog (Buckland & Kenward, 1973; Buckland 1979; Smith, 1985a), which buried and preserved the prehistoric forest and trackway *in situ*. No other trackways or bridging structures have been reported from this area, although a more sophisticated structure, also belonging to the late-Bronze Age, has been found at Brigg in the Ancholme valley (Smith, 1958b) along with hurdle type trackways uncovered recently on the Humber foreshore at Melton (Crowther, 1987; Fletcher *et al.*, 1999).

Smith (1985a) made a large contribution to the hitherto piecemeal palaeobotanical studies of the Humberhead Levels area. Analysis of pollen, and variations in peat humification and composition, are used to reconstruct a detailed record of changes in mire vegetation, hydrology, and regional landscape development. He obtained 77 dates from sites on both Hatfield and Thorne Moors, which proves peat initiation is generally later at Thorne than at neighbouring Hatfield Moors, and is characterised by an initial reedswamp and fen carr stage due to a base rich groundwater influence. However, it is also apparent that wetland development had been time transgressive at Thorne Waste, with peat growth at Rawcliffe and Goole Moors, central mire sites in the north, dated to 4545±75 BP (5452 to 4878 cal. yrs BP) and 4515±70 BP (5442 to 4874 cal. yrs BP), and at Crowle Moors in the south western part of the region, to 3060±65 BP (3386 to 3069 cal. yrs BP) and 3475±65 BP (3891 to 3571 cal. yrs BP). Smith (1985a) concludes that peat development at Thorne Waste ensued largely as a result of waterlogging ahead of rising sea level (Gaunt & Tooley, 1974) and climatic deterioration.

Smith (1985a) notes that there are obvious similarities between local pollen assemblage zones (lpazs) from individual sites. He uses this as justification for constructing regional pollen assemblage zones (rpazs), reflecting the vegetation of the higher and drier land of the surrounding Humberhead Levels area, and human impact on the landscape from Bronze Age times through to the post-Norman era. The main characteristics of these rpazs are outlined below:

Zone HHL/A: *Quercus - Corylus - Pinus* zone c.4300 to c.3600 BP (c.5200 to c.3600 cal. yrs BP)

Tree and shrub pollen contribute *c*.95% of the total pollen in this rpaz. Fossil beetle faunas from the basal woody peat at Thorne also attest to the relatively undisturbed nature of the woodland at this time (Buckland, 1979). The vegetation mosaic is similar to the base of the pollen diagram by Smith (1958a). Smith (1985a) suggests that *Alnus glutinosa, Salix* and *Betula* scrub occupy wet areas locally, with relatively high (up to 30%) values for *Pinus sylvestris* pollen indicating growth of this taxa on drier areas of the peat surface, and forming stands on the sandy exposures surrounding the bog. Drier soils on higher ground were colonised by mixed deciduous forest dominated by *Quercus*, *Tilia*, *Corylus avellana* with subordinate *Fraxinus excelsior*, *Ulmus* and *Betula* (Smith, 1985a). Although percentages of *Tilia* pollen are fairly low, this species is entomophilous, and often under-represented in the pollen rain (Godwin, 1975a; Huntley & Birks, 1983). Smith (1985a) considers the 5%, or less, *Tilia* pollen to represent a substantial population of lime trees. This assumption is supported by unpublished insect evidence for the *Tilia* feeding beetle *Ernoporus caucasicus* from Misterton Carr (Buckland & Dolby, 1973), which suggests a high proportion of *Tilia* in the forest of the Isle of Axholme (see Dinnin, 1997a), and from the mid-Holocene deposits along the river Trent floodplain at Bole Ings (Dinnin & Brayshay, 1994; Brayshay & Dinnin, 1999).

High *Tilia* pollen frequencies have been found from sites with small source pollen areas. Greig (1982b) reports high *Tilia* percentages from Butterbump in Lincolnshire, which he integrates with evidence from other sites, and concludes (as had others before him) that *Tilia* was a major constituent of mid-Holocene forests (Godwin, 1975b; Birks *et al.*, 1975; Girling & Greig, 1977; Bradshaw, 1981b; Greig, 1982a). Smith (1985a) envisages a 'modified mixed oak forest', upon which Mesolithic and Neolithic peoples had already exerted an influence. Smith (1985a) suggests palynological evidence in support of this is provided by the low *Ulmus* values (which may have been used for cattle fodder, Troels-Smith, 1960), and low values of the light demanding species *Fraxinus excelsior* (Tinsley, 1981) which may indicate opening areas within the woodland. There is also archaeological evidence for human disturbance as Buckland & Dolby (1973) and Buckland (1979) report findings of stone tools dating from Meso- and Neolithic times from the Isle of Axholme area.

Zone HHL/B: *Quercus - Corylus - Alnus* zone c.3600 to c.2300 BP (c.4200 to c.2100 cal. yrs BP)

The opening of this zone is characterised by a decline in *Pinus sylvestris* and *Tilia*, and a spatially discontinuous reduction in *Quercus* and *Ulmus*, which is accompanied by an increase in certain herb taxa (e.g. Poaceae, *Plantago lanceolata*, *Pteridium*, *Urtica*). Forest openings are also indicated by high *Corylus avellana* frequencies, and fluctuating *Fraxinus excelsior* values, but because levels of arboreal pollen remained generally high (*c.*80% TLP), this is only thought to represent relatively small scale agricultural activity (Smith, 1985a). The presence of dated charcoal layers in the peat at both Thorne (3715±70 BP, 4246 to 3847 cal. yrs BP) and Hatfield Moors (3570±70 BP, 4080 to 3647 cal. yrs BP) and a burnt and chopped pine trunk dated to 3545±70 BP (4048 to 3634 cal. yrs BP) at Crowle, together with palynological evidence for disturbance, leads Smith (1985a) to conclude that this zone represents a major phase of deliberate tree clearance and burning activity by Bronze Age people. The *Pinus sylvestris* decline is dated to 3715±70 BP (4262 to 3847 cal. yrs BP) on Thorne Moors and 3685±65 BP (4225 to 3835 cal. yrs BP) on Hatfield, which Smith (1985a) considers to be a result of possible tree clearance and human activity on the Lindholme outcrop, a raised island in the middle of the Hatfield Moor. The importance of *Alnus glutinosa* is thought to relate to the variable development of fen woodland locally at sites marginal to the raised bog (Smith, 1985a).

The extent and character of human activity is rather difficult to discern during this zone. Whilst sites peripheral to Thorne Moors indicate phases of clearance and regeneration, (such as the 'Trackway' site, Buckland, 1979; Smith, 1985a) sites central to the mire studied by Turner (1965) and Smith (1985a) suggest little evidence for temporary clearance but, that imply one extended period of Bronze Age clearance. This probably reflects the smearing effect of different taphonomic processes and variable pollen source areas (Tauber, 1965). Also certain aspects of the pollen evidence for farming activities is not secure. For example, Buckland & Kenward (1973) and Buckland (1979) report significant quantities of Poaceae (30% TLP) and Cereal pollen (14%) from the trackway horizon, which they use as evidence for mid-Bronze Age clearance and cultivation. However, Smith (in Buckland, 1979) cautions against such interpretations, because the Cereal pollen is most probably derived from the natural wetland grass *Glyceria*, which is recorded from the same context by Smith (1985a). Similarly, Smith (1985a) indicates problems with interpreting the Poaceae pollen as a clearance indicator, as it could have originated from *Phragmites* communities or other wet ground grass species surrounding the mire.

In spite of difficulties in recognising the extent of woodland clearance and human activities during this time, it is evident from Smith (1985a) that varying degrees of vegetation disturbance, largely for pastoral activity, were taking place during the Bronze Age. This is concordant with Turner (1962) who dates a decline in *Tilia* with an associated increase in Poaceae and *Plantago lanceolata c.*3000 BP (*c.*3400 to *c.*2900 cal. yrs BP) at Thorne Moors. However, this 'primary decline' in *Tilia* is not a pronounced feature of the pollen diagrams in Smith's (1985a) research.

Smith (1985a) also provides evidence for a re-expansion of landscape clearance and subsequent woodland regeneration during the late-Bronze Age and Iron Age, although this is more marked in the vicinity of Thorne than Hatfield Moors. Smith (1985a) tentatively suggests that this may relate to a period of positive sea-level tendency, which has been widely reported from the Humber region during this time (e.g. Smith, 1958a; Gaunt & Tooley, 1974 and subsequently, Long *et al.*, 1998a). He postulates that the backing up of drainage water from the rivers could have caused flooding in low lying areas, which would have resulted in a reduction in agricultural activity, and a period of woodland regeneration during the Iron Age (see section 10.3 for alternative explanation). The coincidence of flooding horizon R.Y.HHLIV (dated to *c*.2500 to *c*.2300 BP, *c*.2800 to *c*.2100 cal. yrs BP, Smith, 1985a), broadly contemporaneous with this sea-level rise, is therefore attributed to sea-level change rather than climatic deterioration (Smith, 1985a). However, previously, Smith (1958a), using evidence from Godwin (1954), doubted whether raised mire growth could be affected by sea-level change, and Smith (1985a) acknowledges the comments of Turner (1981), that it is extremely difficult to distinguish between the effects of climate and sea-level change. More recently, it has been suggested that factors such as sea-level change can affect the seepage and hydrological stability of bogs (Barber, 1994).

Zone HHL/C: *Gramineae - Plantago - Pteridium* zone *c*.2300 to *c*.1400 BP (*c*.2700 to *c*.1100 cal. yrs BP)

This zone spans the middle-Iron Age to the end of the Roman period. It is characterised by high frequencies of herb pollen (up to 80% TLP), of which, Poaceae, Cyperaceae, and *Calluna* are most abundant, which reflects the development of raised mire vegetation on Thorne and Hatfield Moors. The opening of the zone is marked by a decline in total tree and shrub pollen percentages, and a parallel increase in the pollen of clearance herbs such as *Plantago lanceolata*, *Rumex*, *Artemisia*-type and *Urtica*, and spores of *Pteridium aquilinum* (Smith, 1985a). This provides substantial evidence for the intensification of agricultural activity during this period, which is supported by many other sites in the British Isles (Turner, 1981). In particular, the frequencies of *Tilia* pollen decrease markedly, and thereafter, this species is almost totally absent (Smith, 1985a). This is also a feature of pollen diagrams from Hatfield by Smith (1958a) and Thorne by Turner (1962). The boundary dates Smith (1985a) obtains are very similar to the 'secondary *Tilia* decline' of Turner (1962), which is dated to 2329±110 BP (2728 to 2061 cal. yrs BP).

Smith (1985a) dates the *Tilia* decline to 2145±65 BP (2325 to 1947 cal. yrs BP) and 2085±70 BP (2305 to 1876 cal. yrs BP) on Hatfield Moors, and 2335±40 BP (2360 to 2214 cal. yrs BP), 2350±60 BP (2703 to 2191 cal. yrs BP), and 2225±70 BP (2350 to 2012 cal. yrs BP) on Thorne Moors. The metachronous nature of this feature at lowland sites across the British Isles is due to the influence of Iron Age people (Turner, 1962), perhaps utilising the bast and nutritious leaves for cattle fodder, or removing the tree preferentially due to its tendency to grow on better soils, although see Waller (1994b) an alternative explanation.

Smith (1958a) and Turner (1962) both contend that the major period of clearance was during Romano-British times. Smith (1985a) shows that relative percentages of dryland trees decrease to *c*.30% TLP at the height of the agricultural activity during this period, indicating a very open environment. However, the continued presence of tree pollen frequencies of over 50% from Hatfield Moor sites, indicates some areas remained wooded during this time. Smith (1985a) indicates that high frequencies of Poaceae, and cereal pollen, represented by both *Triticum* and *Secale cereale*, point to a mixed farming economy after the Roman conquest (studies have shown that rye pollen first appears in the Roman period, e.g. Birks, 1965; Godwin, 1967).

Buckland (1979) summarises the considerable amount of archaeological evidence for human occupation and agriculture at this time, such as field systems, enclosures and artefacts. This was a time when there was extensive modification of the natural environment, by an increasing population, whose numbers were bolstered by European invaders. Buckland (1979; 1973) considers that by the end of the Roman period, primary forest on all but the most marginal land had been cleared for agriculture. Buckland & Sadler's (1985) investigation of Roman insect faunas from Sandtoft is also suggestive of extensive landscape clearance, which supports the hypothesis that deforestation and agricultural intensification at this time caused extensive alluviation in the region (Samuels & Buckland, 1978; Buckland & Sadler, 1985; Riley *et al.*, 1995).

Pollen investigations from the Roman site at Scaftworth, along the river Idle floodplain in the southern Humberhead Levels (see Figure 3.2), indicate that away from the *Alnus*-dominated carr woodland, the landscape was essentially open and farmed (Gilbertson & Blackham, 1985; Van de Noort *et al.*, 1997). However, the extent to which the extensive upper oxidised facies present throughout the region can be demonstrably linked to increased soil erosion during this time, has not been rigorously tested, and the theory has been generally accepted due to the archaeological context in which such mottled clays are situated at sites in the Humberhead levels (Samuels & Buckland, 1978; Buckland & Sadler, 1985).

Subsequent studies that have alluded to similar conclusions (e.g. Dinnin, 1997b; Lillie & Grattan, 1995) have done so without substantial additional evidence in support of the argument, or precise chronological control. This can be attributed to the 'suck-in' effect by Baillie (1991) whereby the coincidence of archaeological evidence is conveniently used to 'suck-in' a date for an event. Although the application of mineralogical sourcing and dating (e.g. mineral magnetics etc.) is the preferred method of elucidating this hypothesis, it is hoped that by extensively surveying similar sediments from valleys in the Humberhead Levels in this study, and analysing their microfossil content, this information may help resolve this uncertainty.

Zone HHL/D: *Quercus - Corylus - Betula* zone *c*.1400 to *c*.900 BP (*c*.1500 to *c*.700 cal. yrs BP)

The time period spanned by this rpaz is basically the post-Roman period through to the Norman conquest. The zone commences with a decline in herb pollen and a recovery in percentages of tree and shrub pollen, which contributes *c*.65% TLP, suggesting a reduction in agricultural activity at the end of the 4th century (Smith, 1985a). At most sites, dryland tree frequencies reach *c*.80%, similar values to HHL/B during the Iron Age, although the woodland composition differs. Taxa such as *Quercus*, *Betula*, *Fraxinus excelsior*, and *Corylus avellana* are moderately high, whilst percentages of *Ulmus*, *Pinus sylvestris*, and *Tilia* remain low. Other arboreal species such as *Acer campstre*, *Fagus sylvatica*, and *Carpinus betulus* also occur. Although such tree taxa were present before this time, Godwin (1975a) suggested that they may only have been able to compete with other tree species when the environment opened up and they were allowed to share the recolonisation of cleared areas. Smith (1985a) suggests that scrub woodland prevailed on the abandoned agricultural land, dominated by *Quercus*, *Fraxinus excelsior*, and *Corylus avellana*. (Rackham, 1980 notes how *Quercus* can act as a pioneer species on abandoned agricultural land). However, this agricultural decline did not affect the frequency of clearance herbs, which are present throughout.

Smith (1985a) proposes four reasons for the pollen changes which characterise zone HHl/D. First, as in zone HHL/B, he suggests a rise in sea level, which considerably reduced the land available for agricultural activity. Evidence in support of this hypothesis is forthcoming from other sites in the Humber and elsewhere in Britain (Smith, 1958a; Gaunt & Tooley, 1974; Godwin, 1978; Tooley, 1978a). The second theory Smith (1985a) suggests, is climatic deterioration, which was registered by a recurrence surface in the mire stratigraphy, and may have resulted in conditions not conducive to crop cultivation and ripening. Third, Smith (1985a) introduces another flooding hypothesis, whereby the widespread deforestation during Romano-British times, caused increased run off into the rivers with a negative impact on agricultural productivity. The final, hypothesis is that the change may be a cultural one. Smith (1985a) suggests that there was less pressure on the land after the decay of the Roman society, which caused woodland regeneration (*cf.* Arnold, 1984). The collapse of population and settlement nucleation during the Dark Ages is also seen as attributable to woodland regeneration (*cf.* Cunliffe, 1974). Buckland (1973) also attributes agricultural decline during the post-Roman period as responsible for regeneration of woodland.

A slight resurgence in agricultural activity towards the end of this zone is suggested by the appearance of *Cannabis sativa* pollen (Smith, 1985a), which is evidence for the migration of Anglo-Saxon people into the region after the 3rd century (Godwin, 1967, Bradshaw *et al.*, 1981).

Zone HHL/E: *Gramineae - Cannabis - Secale* zone <*c*.900 (<*c*.1000 cal. yrs BP)

Peat cutting activities and drainage have disturbed the upper 1000 years of peat on Thorne and Hatfield Moors (Smith, 1985a; Buckland & Dinnin, 1997), so that the end of this zone remains undated and Smith's (1985a) conclusions are limited to evidence from only a few sites on Thorne Waste. The opening of the zone is marked by a decline in tree and shrub pollen suggesting a resurgence in woodland clearance, with herb pollen increasing to *c*.80% TLP. Poaceae, Cyperaceae, and *Calluna* are important components of this, which partly reflects the presence of heath environments locally. Pollen of crop species, such as *Cannabis sativa*, *Secale cereale*, and *Triticum* are high, along with weed species characteristically associated with arable subsistence (e.g. *Centaurea cyanus*, *Spergula arvensis*, *Polygonum arvensis* etc.). Species such as *Plantago lanceolata*, *Rumex*, *Artemisia*, and *Urtica*, in association with high values of Poaceae and Cyperaceae suggest considerable pastoral activities.

4.4 Trent and Ancholme valley

The most detailed palaeoecological study from the Trent is based on the analysis of floodplain peat at Bole Ings by Dinnin & Brayshay (1994) and Brayshay & Dinnin (1999), and provides the only chronology for the vegetation history of this area. The palynological record dates from 8240±70 BP (9426 to 8986 cal. yrs BP), with *Salix* and a range of wetland herbs dominating the floodplain community. *Corylus avellana* and possibly *Pinus sylvestris* may have colonised drier areas locally. The *Alnus glutinosa* rise is dated to 6290±70 BP (7361 to 7012 cal. yrs BP), and persists for a further *c*.3500 radiocarbon years. During this time, inferences as to the nature of the dryland community are hampered by the overwhelming amounts of *Alnus glutinosa* pollen (up to *c*.90%) which is included in the pollen sum (Dinnin & Brayshay, 1994). Brayshay & Dinnin (1999) present a pollen diagram excluding wetland types, which shows *Quercus*, *Ulmus*, *Pinus sylvestris*, and *Tilia* to be major components of the woodland community in the surrounding landscape. *Tilia* is a particularly well represented taxa and considering the poor nature of its pollen dispersal, and the identification of *Tilia* dependent beetles, it is suggested to be dominant in the dryland community (Brayshay & Dinnin, 1999).

The *Ulmus* decline is weakly expressed and tentatively dated by interpolation to *c*.5200 BP (c.6300 to 5700 cal. yrs BP, Brayshay & Dinnin, 1999) although there are also unexplained pre-*Ulmus* decline pollen fluctuations. A decline in both *Tilia* and *Pinus sylvestris* is dated to 3570±70 BP (4080 to 3647 cal. yrs BP), which (Brayshay & Dinnin, 1999) interpret as the first evidence for major clearance of woodland due to anthropogenic activity. This event is correlated with similarly dated declines in *Pinus sylvestris* and *Tilia* on Thorne and Hatfield Moors by Smith (1985a) during rpaz HHL/B. Turner (1962) and Greig (1982b) also note clearance of *Tilia* during the Bronze Age period within the Humber region.

A decline in *Alnus glutinosa* is dated to 2690±100 BP (2985 to 2496 cal. yrs BP), and this taxon is replaced by Poaceae and *Typha angustifolia*, which suggests a significant increase in wetness at the site. Brayshay & Dinnin (1999) suggest the cause of this flooding, and increase in inorganic alluviation, may have been induced by deliberate catchment tree clearance and resulting increased soil erosion. However, this environmental change could equally be attributed to the ponding of freshwater, associated with the expansion of estuarine conditions during this time (Long *et al.*, 1998a, and sections 7.5 and 11.1).

Preece & Robinson (1984) have investigated aspects of the Lateglacial stratigraphy in the Ancholme valley, gathering evidence from pollen, ostracods, and mollusca. Above the Lake Humber deposits at Castlethorpe (see Figure 3.2), Preece & Robinson (1984) identify a detrital mud. They obtain pollen assemblage typical of the Lateglacial period from this context, which indicates a wetland environment, surrounded by open ground. Mollusc remains contained within aeolian sands overlying the detrital mud unit indicate an open, marshy environment, with numerous shallow water bodies (Preece & Robinson, 1984). By *c.*10000 BP (12300 to 19600 cal. yrs BP), tufa had begun to form in the vicinity of valley side-tributary gullies (Fletcher, 1981). Preece & Robinson (1984) map these tufa deposits, and show them to be widespread across the valley bottom. Molluscan analysis of these deposits reveal that tufa initially formed within an open environment, with shallow pools prone to drying (Preece & Robinson, 1984). Throughout the sequence the mollusca indicate progressive shading until eventually closed woodland conditions prevail (Preece & Robinson, 1984).

Smith (1958a) presents pollen data from three sites in the Ancholme valley (Island Carr, Brigg, and Redbourne Hayes, see Figure 3.2), which he uses in his early analysis of sea-level change in the Humber. At Island Carr, Smith (1958a) describes a clay wedge that fills the Ancholme valley, and overlies a wood peat via a transitional reed clay horizon. Below the peat, small pockets of tufa deposits and shelly peats are also recorded (*cf.* Preece & Robinson, 1984) and are found to be the lowest stratigraphic horizon, within a deeply incised channel of the Old River Ancholme (Smith, 1958a).

High *Tilia* pollen values are noted from the base of the peat, which are accompanied by increasing values of *Alnus glutinosa* and *Quercus* pollen, indicating fen wood, with *Corylus avellana*, *Betula*, *Pinus sylvestris*, and *Ulmus* also represented on drier ground. The *Tilia* pollen is considered by Smith (1958a) to be over-represented, due to the robust nature of the pollen grains, and their possible differential preservation in soils (e.g. Havinga, 1984; Dimbleby, 1985). However, later work by Greig (1982a; 1982b) has shown that *Tilia* was an important component of the woodland during this time. The fact that *Tilia* pollen values rapidly decline as *Alnus glutinosa* rises in the basal samples (see Smith, 1958a), implies the decrease in *Tilia* may simply be the product of waterlogging and expansion of wetland area, which progressively pushes the *Tilia* trees away from the coring site (Waller, 1994b). Smith (1958a) ascribes this peat to the latter part of Godwin's (1940b) pollen zone VIIb on the basis of this pollen spectrum. A radiocarbon date of 4046±50 BP (4808 to 4410 cal. yrs BP), from a stratigraphically undefined piece of wood within the basal peat, provides a *terminus ante quem* for peat formation (Fletcher, 1981; Smith *et al.*, 1981), and lends support to Smith's (1958a) assertion.

An increase in Poaceae and Chenopodiaceae, in conjunction with a large decline in *Alnus glutinosa* below the clay, is indicative of increasingly wet local conditions prior to marine transgression. The recognition of an array of archaeological finds dating from Bronze Age/Iron Age times encased within the clay, in conjunction with the pollen assemblage, led Smith (1958a) to conclude that the sedimentary change from peat to estuarine clay occurred at the Bronze Age/Iron Age transition. A similar date of *c.*2800 BP (*c.*3200 to *c.*2800 cal. yrs BP) is postulated for the deposition of marine clay opposite the Ancholme valley, at North Ferriby (Wright & Churchill, 1965).

On the surface horizon of the lower peat bed at Brigg, a trackway of *Quercus* planks was discovered, upon a foundation of small trees and branches, secured by stakes. Associated with this trackway are various finds of late-Bronze Age/early-Iron Age date, including a bronze pin, pottery and spear head. Pollen analysis from sediment contained within the canal of a large animal vertebrae and the spear head have been carried out by Smith (1958a; 1958b), and are shown to be broadly contemporaneous with the earliest stage of clay deposition. Wood from the level of the trackway is dated to 2552±120 BP (2867 to 2339 cal. yrs BP, Godwin & Willis, 1960) which corresponds to the Bronze Age/Iron Age transition.

A dugout canoe has been uncovered from deposits at Brigg, which Smith (1958b) contends was of Bronze Age, but had probably been used into the Iron Age. A 'raft' was also discovered sealed within the lower levels of this clay, but has subsequently been identified as more like a part of a Bronze Age sewn boat (Wright & Wright, 1947; McGrail, 1981). Various parts of the 'raft' have been dated, the mean of which is 2597±117 BP (2938 to 2349 cal. yrs BP, Switsur, 1981), and has subsequently been recalculated to 2603±21 BP (2756 to 2729 cal. yrs BP, Van de Noort *et al.*, 1999). A statistically indistinguishable date of 2625±65 (2850 to 2499 cal. yrs BP) from the reed clay immediately overlying the 'raft' (Smith *et al.*, 1981), supports the previous suggestions by Smith (1958a), Wright & Churchill (1965), and Gaunt & Tooley (1974) that there was a large marine incursion in the Humber region at this time. Smith (1958a) concludes that the Ancholme valley around Brigg would have been an important site during the Bronze Age and later, because the narrowing of the valley at this point makes it an ideal place for a river crossing, during a time when the environment was becoming wetter.

Fletcher (1981) contends that the maximum extent of the marine transgression had occurred at Brigg by *c.*2700 BP (2995 to 2508 cal. yrs BP), when estuarine sediment accumulated 22 km inland at Waddingham Holmes. Similar findings have been reported from elsewhere in the Humber estuary, and have enabled a better spatial and temporal picture of sea-level changes at the Bronze Age/Iron Age transition to be established (Long *et al.*, 1998a).

Soon after the deposition of the 'raft', marine influence began to wane, probably due to both sediment accretion elevating the local land surface, and marine regression (Van de Noort & Davies, 1993). Smith *et al.* (1981) suggests the environment around the 'raft' was marginal saltmarsh, with *Salix* and *Betula* scrub. The regional vegetation appears to have remained densely wooded, although pollen and insect evidence suggests there were open areas of grassland (Buckland, 1981; Smith *et al.*, 1981). Whilst, these may be of natural saltmarsh origin, Preece & Robinson (1984) note that some areas of the valley sides had experienced anthropogenic clearance, which induced slopewash by 3410±80 BP (3846 to 3465 cal. yrs BP).

The upper clay layers at Brigg are sparse in pollen, and the upper peat, which had been recorded in the late 19th century, has disappeared (Smith (1958a). However, the site at Redbourne Hayes, 8 km south of Brigg, is stratigraphically similar to Island Carr, and the upper peat was then, still in existence. Here, a lower reedswamp peat is overlain by a clay with reeds, and then the upper peat. Smith (1958a) notes ploughing disturbance within this surface organic layer, and deemed it safe to only take one sample from its base. The tree pollen curves from the base of the reedswamp peat are similar to those at Island Carr, with high values of *Tilia*. High *Betula* values within the upper peat lead Smith (1958a) to correlate this with similar Romano-British horizons of peaty clay at South Ferriby, which occur with adjacent Roman pottery within the underlying clay.

Although high *Betula* values may only be of local significance, archaeological data is also available at Redbourne which supports this date. Within the upper peat horizon in the vicinity of Redbourne, a timber trackway and associated finds of Roman archaeology has been reported (Smith, 1958a). Smith *et al.*, (1981) provide a maximum age for the upper peat with a date on the regressive contact at Brigg of 2625±65 BP (2851 to 2499 cal. yrs BP). Fletcher (1981) suggests that the upper peat was not a single depositional unit, but was attributed to separate wetlands that accumulated as sea-level continued to fall during the period *c.*2600 to *c.*2200 BP (*c.*2900 to *c.*1900 cal. yrs BP).

Lower sea levels during the Roman period, followed by a transgression, have been documented in other coastal areas of Britain (e.g. Godwin, 1978; Fulford *et al.*, 1997). Smith (1958a) reports an upper clastic facies at South Ferriby, that he suggests is related to a renewed estuarine phase at the end of the Romano-British period. Gaunt & Tooley (1974) also put forward evidence for a sea-level rise at this time, which they suggest corresponds with the Lytham IX transgression, between *c.*1800 and *c.*1400 BP (*c.*1900 to *c.*1100 cal. yrs BP, Tooley, 1978a). This provides an approximate minimum age for upper peat deposits at South Ferriby. This organic unit is now absent from most of the valley (Van de Noort & Davies, 1993).

These deposits at Redbourne and Brigg have been extensively re-evaluated by Neumann (1998) and her synthesis of wetland development generally confirms the sequence of events outlined by Smith (1958a). A total of 21 radiocarbon dates have been taken from the base of the lower peat from the sites investigated by Smith (1958a), which provides data on the onset of peat development during a period of generally continued sea-level rise. An age-altitude plot shows the distribution of dated samples up the valley in relation to rising ground water levels. The upward tendency of the height of peat inception in relation to radiocarbon age reflects the progressive rise in base level in the Ancholme, which is similar to established sea-level trends in the Humber estuary (Gaunt & Tooley, 1974; Long *et al.*, 1998a, see Chapter 10).

The oldest sample at Brigg records rising water levels by 6170±90 BP (7225 to 6802 cal. yrs BP) at *c.*-4.60m OD (Neumann, 1998), which is considerably earlier than the late-Neolithic date suggested by Smith (1958a) whose sample core attained a very similar depth below OD. However, a deeper peat *c.*-6.35m OD from an adjacent core reveals a much younger date of 4990±75 BP (5912 to 5591 cal. yrs BP).

Neumann (1998) suggests the sand underlying this peat, which occurs down to *c.*-8m OD, contains the marine epipsammic diatom *Psammodiscus* sp. Comparison with the sea-level graph in Long *et al.* (1998a), and chapter 10, suggests that if the sand does indeed represent an early marine phase, the age obtained from the peat must be several thousand years too young, although Neumann (1998) provides no explanation for this discrepancy. The downward intrusion of plant rhizomes is a likely explanation, considering reeds are indicated as a component of the overlying peat. Neumann (1998) maintains peat formation was confined within a deep palaeochannel feature *c.*6000 BP (*c.*7200 to *c.*6800 cal. yrs BP), and the spread of wetland over the higher parts of the floodplain did not commence until *c.*5000 BP (*c.*6000 to *c.*5500 cal. yrs BP) which is similar to the crude date hypothesised by Smith (1958a). Therefore, if the identification of marine diatoms is correct, perhaps this represents the landward migration of tidal channels during a positive tendency, which eroded the central valley floor sediments. Peat formation may then ensue at a later date, as rates of organic sediment accumulation overtake the rate of sea-level rise.

At Redbourne, the radiocarbon dates by Neumann (1998) suggest basal peat formation from 4494±31 BP (5291 to 4988 cal. yrs BP, based on an average of two replicate measurements), which is more in agreement with Smith (1958a). Peat had spread onto the higher margins of the floodplain by *c.*3500 BP (*c.*4100 to *c.*3500 cal. yrs BP, Neumann, 1998).

At South Ferriby, at similar stratigraphy to Redbourne and Brigg is described by Neumann (1998), with peat inception dating to 6000±50 BP (6983 to 6731 cal. yrs BP) at c.-3.30m OD. However, again there are inconsistencies, as a younger date is obtained from a stratigraphically lower position in an adjacent core. Peat initiation was time transgressive up the valley sides with the youngest dates of c.2600 BP (*c.*2900 to *c.*2400 cal. yrs BP) occurring at +2m OD, along the floodplain margin. An upper peat is recorded between +1 and 2m OD, from which microfossil interpretation was not possible (Neumann, 1998). This is overlain by grey clay, as reported by Smith (1958a), which he suggested was deposited during a post-Roman transgression. Brushwood from a Roman road exposed on the foreshore at South Ferriby is buried by marine silty clay, and provides a radiocarbon date of 1960±50 BP (1995 to 1808 cal. yrs BP, Neumann, 1998). This provides a *terminus ante quem* for this transgression in the Ancholme region.

4.5 Holderness and the Yorkshire Wolds

The vast majority of palaeobotanical research in the Holderness region has focused on the limnic sediments infilling the numerous meres, which occur in depressions in the undulating till surface (Sheppard, 1957). These mere basins vary in size, and are described by Flenley (1987) and Dinnin (1995). The large, elongate meres (e.g. Hornsea Mere, Hornsea Old Mere, Skipsea Bail Mere and Lambwath Mere, see Figure 3.2) are probably related to pre-till valleys. The smaller ovoid meres reflect natural hollows in the till surface due to uneven deposition of till (e.g. Sproatley Mere), or are ice-melt features (e.g. Bog at Roos and Gransmoor), such as kettle holes (Flenley, 1987) or collapsed pingos (Berridge & Patterson, 1994).

Early studies by Godwin & Godwin (1933b), and Clark & Godwin (1956), applied pollen analysis in an attempt to provide a relative chronology for the vegetation changes at Skipsea Withow Gap, and organic deposits at Brandesburton as part of an investigation into Maglemosian harpoon finds and fossil bones of the giant elk (*Megacerus giganteus*). However, the stratigraphic origins of these artefacts are open to question (*cf.* Gilbertson, 1984a; 1990; Dinnin, 1995), not least because they suggest an unlikely survival date for this beast into the Holocene (Simmons *et al.*, 1981; Dinnin, 1995).

A detailed re-investigation of the exposed cliff section of the Skipsea Withow Gap deposits by Gilbertson (1984a; 1984b) and Gilbertson *et al.*, (1987) using a combination of pollen, plant macrofossils, molluscs and sedimentary analyses, have firmly established the environmental history of the mere from late-Devensian times. Climatic oscillations and vegetation changes are identified, which correspond with the typical Lateglacial sequence recorded from elsewhere in the region (Beckett, 1981; Walker *et al.*, 1993). A pollen diagram shows the development of the vegetation between 9880±80 BP and 4500±50 BP (11594 to 4879 cal. yrs BP) and is divided into 6 lpazs (Gilbertson 1984a; Gilbertson *et al.*, 1987). *Betula*, *Pinus sylvestris* and *Corylus avellana* are characteristic taxa of the early-Holocene period, which become supplanted by *Quercus*, *Alnus glutinosa* and then *Tilia* and *Ulmus*. Inorganic lake muds are replaced by organic gyttjas as a response to this increase in biological productivity around the mere.

Alnus carr dominates the catchment from *c.*7000 BP (*c.*8000 to 7600 cal. yrs BP), from which time fen woodland began to extend across the site, although some open water remained (Gilbertson, 1984a; Gilbertson *et al.*, 1987). By 4500±50 BP (5307 to 4879 cal. yrs BP), the mere had silted up completely, and the fen and fen carr conditions were replaced by colluvial silts. However, Gilbertson (1984a; 1984b; 1990) and Gilbertson *et al* (1987) suggest this date may be too old, due to the input of older carbon (hard water error) from water draining the adjacent chalk rich tills.

This site is of particular interest from an archaeological perspective due to the evidence for the human management of the *Alnus* carr (Gilbertson, 1984b). A carved wooden rod, stake and peg have been uncovered, of early-Neolithic age, suggesting coppicing and perhaps trackway construction from 4770±70 BP (5645 to 5316 cal. yrs BP). The pollen record corresponding to this time (lpaz WM5 and WM6) indicates a two phase of reduction in *Alnus glutinosa* pollen, and an associated increase in *Fraxinus excelsior*, Cereal-type, Poaceae, and *Plantago lanceolata* pollen, which has been interpreted as human disturbance associated with pastoral agriculture (Gilbertson, 1984a 1984b; Flenley, 1987).

Recent reports of significant finds of beaver gnawed wood from these deposits (McAvoy, 1995), has cast doubt on the original pollen and wood macrofossil evidence for human induced Neolithic woodland clearance (Gilbertson, 1984a; 1984b; 1990; Gilbertson *et al.*, 1987; Flenley, 1987). It is possible that the decline in woodland taxa is the result of beaver activity, which may be exacerbated by the rise in water level associated with beaver damming. This is supported by an increase in *Cladium* in WM5, and Apiaceae in WM6 (which includes wetland taxa such as *Peucedanum*, *Apium*, *Oenanthe* and *Cicuta*), which may indicate wetter conditions. The impact of beaver activity in creating open areas (which would traditionally be interpreted as evidence for human activity in pollen diagrams) has been stressed recently by Coles & Orme (1983) and Coles (1992). It is likely that prehistoric people would have taken advantage of these clearings and may have maintained them for there own purposes (*cf.* Coles, 1992).

The studies of Beckett (1975; 1981) provide a virtually continuous record of vegetation changes from the Lateglacial, through the Holocene, to the present. Rpazs are based on the pollen record from three sites; a valley fill sequence from Gransmoor Quarry, and mere deposits from Roos Bog, and Hornsea Old Mere (see Figure 3.2). The pollen records from Gransmoor and Hornsea Old Mere are temporally limited, as they only provide information up until the mid-Holocene. Therefore, the 11.5 m sequence at the Bog at Roos forms the basis for interpreting the later Holocene vegetational history for Holderness (Gilbertson, 1984a; Flenley, 1984; 1987; 1990).

Despite the high quality of the palynological information in Beckett (1981), Taylor (1995) has drawn attention to some limitations with using these rpazs as a basis for interpreting palaeobotanical changes during the Holocene. The absence of a radiocarbon chronology for the Holocene is not ideal, and there is uncertainty regarding 3 of the 5 dates that cover the Lateglacial period, as they yield statistically similar ages (11500±170, 11450±230 and 11220±220; 13900 to 12700 cal. yrs BP). In addition, the samples for pollen analysis and radiocarbon analysis were obtained from separate core locations. Many of the characteristic Holocene vegetation changes are constrained by a chronology determined from elsewhere (e.g. Gransmoor) or even outside the region, which does not enable the identification of intra or inter regional disparities in the timing of vegetation change.

Nevertheless, it is worthwhile summarising the rpazs of Beckett (1981), as they provide a useful comparison of the vegetation growing on a different lithology (Quaternary deposits of till and boulder clay and perhaps the regional signal from the Chalk uplands of the Yorkshire Wolds), to the study sites in this thesis in the lower Aire valley. The Lateglacial sequence is only clearly represented at Roos Bog, and is subdivided into 3 lpazs. The record of landscape, climatic and vegetation changes closely matches the early work by Bartley (1962) and the recent work of Lowe *et al.* (1995) and Walker *et al.* (1993; 1994). Consideration of the development of the Lateglacial vegetation is beyond the scope of this study, so is not included here, but this information is comprehensively reviewed by Dinnin (1995). The Holocene is subdivided into 5 rpazs and are summarised below.

Betula-Pinus assemblage zone _c._10200 to _c._9000 BP (_c._12500 to _c._9700 cal. yrs BP)

A radiocarbon date of 10120±180 BP (12421 to 10971 cal. yrs BP) secures this rpaz within the early Holocene (Beckett, 1981). *Betula* woodland dominates with *Pinus sylvestris,* and some *Salix*. Conditions are relatively open, indicated by the presence of light demanding herbs such as *Filipendula*. The type site is Hornsea Old Mere (lpaz HO1), although similar conditions are also evident at surrounding sites in

Holderness (Godwin & Godwin, 1933b; Clark & Godwin, 1956; Beckett, 1975; 1981, Gilbertson, 1984a; Flenley, 1987).

***Corylus/Myrica-Ulmus* assemblage zone *c*.9000 to *c*.7000 BP (*c*.10300 to *c*.7600 cal. yrs BP)**

This rpaz is dominated by *Corylus/Myrica* pollen (most likely to be *Corylus avellana* pollen during this time in Holderness, Flenley, 1987) with *Quercus* and *Ulmus* also abundant. Pollen of *Betula* and *Pinus sylvestris* persist, and dense wooded conditions result in a reduction in the frequency of herb pollen (Beckett, 1981). This rpaz is poorly dated, but a radiocarbon age of 8507±55 BP (9531 to 9386 cal. yrs BP) has been obtained from close to the top of this zone at Gransmoor (Beckett, 1975). The type site is Hornsea Old Mere (lpaz HO2) although vegetation of similar composition is recorded from other sites in the region (Godwin & Godwin, 1933b; Clark & Godwin, 1956; Beckett, 1975; 1981, Gilbertson, 1984a; Flenley, 1987).

***Alnus/Ulmus* assemblage zone *c*.7000 to *c*.5000 BP (*c*.8000 to *c*.5500 cal. yrs BP)**

Non-arboreal pollen is at a minimum during this zone, and arboreal pollen reaches a maximum. *Betula* and *Pinus sylvestris* pollen frequencies decline as *Alnus glutinosa* and *Tilia* increase in frequency. *Ulmus*, *Quercus* and *Corylus avellana* pollen is also abundant (Beckett, 1981). The woodland composition probably varied considerably at this time, according to local edaphic conditions, with *Alnus glutinosa* frequent around the large wetland areas and meres, and *Quercus*, *Ulmus* and *Corylus avellana* dominating areas of drier ground. Poorly drained till and boulder clay in western Holderness probably supported *Quercus*, *Alnus glutinosa* and *Ulmus*, with *Tilia* more abundant in some sites in eastern Holderness (Clark & Godwin, 1956; Beckett, 1975), where the soils are better drained glaciofluvial gravels and sands (Dinnin, 1995).

Fraxinus excelsior appears at some sites, taking advantage of openings in the woodland canopy. Beckett (1981) and Flenley (1987) tentatively suggest that fluctuations in the abundance of arboreal pollen may relate to Mesolithic woodland disturbance, although there is little evidence to relate this to anthropogenic activity (Dinnin, 1995). This rpaz is undated in Holderness, but Beckett (1981) and Flenley (1987) use a date of *c*.7000 BP (*c*.8000 to *c*.7600 cal. yrs BP), as an approximate date for the *Alnus glutinosa* rise from sites elsewhere in Britain. The type site is Hornsea Old Mere (lpaz HO3), but this assemblage is also recorded from other sites in Holderness (Godwin & Godwin, 1933b; Clark & Godwin, 1956; Beckett, 1975; 1981, Gilbertson, 1984a).

Alnus/Quercus assemblage zone *c.*5000 to 2500 BP (*c.*6000 to *c.*2300 cal. yrs BP)

The zone opens with the *Ulmus* decline, dated to 5099±50 BP (5936 to 5730 cal. yrs BP) at Gransmoor Quarry by Beckett (1975). Undated declines in *Ulmus* pollen are evident at Roos Bog, Hornsea Old Mere (Beckett, 1975; 1981), and the Skipsea meres (Gilbertson, 1984a; Flenley, 1987). Arboreal pollen remains high, with *Quercus* and *Alnus glutinosa* particularly abundant. Low frequencies of ruderal herb pollen such as Poaceae, Chenopodiaceae, *Plantago lanceolata*, and occasional cereal pollen is interpreted as evidence for human influence on the vegetation (Flenley, 1987). *Fraxinus excelsior* is the main beneficiary of the falls in *Ulmus* and *Tilia* within this zone, although the detailed pollen diagrams by Beckett (1975; 1981), and Gilbertson (1984a), suggest subsequent regeneration of both *Ulmus* and *Tilia*.

Pollen grains and fruits of the water chestnut *Trapa natans* are recorded from the mere deposits at Skipsea (Flenley *et al.*, 1975; Flenley, 1987) during this rpaz. The climatic implications of the occurrence and disappearance of this species in Britain is the subject of subsequent debate (Tallantire, 1976; Flenley & Maloney, 1976). Flenley & Maloney (1976) and Flenley (1987) even suggest that the occurrence of *Trapa natans* may be related to its deliberate cultivation by Bronze Age people, although archaeological evidence in support of this is merely coincidental. It is probably just as likely that the disappearance of this taxon was related to the silting-up of many of the mere habitats, that was under way during this time, as opposed to any climatic determinant. The type site for this rpaz is Hornsea Old Mere (lpaz HO4), although comparable pollen assemblages are evident in the sites studies by (Godwin & Godwin, 1933b; Clark & Godwin, 1956; Beckett, 1975; 1981, Gilbertson, 1984a; Flenley, 1987).

Alnus/Gramineae assemblage zone *c.*2500 to present (<*c.*2800 cal. yrs BP)

This rpaz is poorly defined due to the variations in the timing and character of vegetation changes recorded in the pollen diagrams at this time, and the lack of a radiocarbon chronology. However, the beginning of this rpaz is characterised by the first major indication of woodland clearance, as herb pollen predominates, and arboreal pollen declines. *Alnus glutinosa* (reflecting *Alnus* carr at the infilling mere sites) is the dominant tree taxa, with *Quercus* and *Betula* present in small amounts. A wide range of herb taxa, such as Poaceae, *Plantago lanceolata*, Chenopodiaceae, and cereal pollen are represented, and are interpreted as reflecting both pastoral and arable cultivation (Beckett, 1981; Flenley, 1987). Several phases of clearance and regeneration (with *Fraxinus* as secondary woodland) are indicated at Roos Bog, which Beckett (1981) attributes to Bronze Age clearance. Due to the absence of radiocarbon dating control, this assumption is based on correlation with sites outside Holderness, such as a similar pollen spectrum from the context containing the Ferriby boats (Wright & Churchill, 1965). This correlation is extremely tenuous, not least because it assumes that the nature and timing of the impact of prehistoric people was synchronous throughout the region. The contrasting character of the uppermost levels of the pollen diagrams from this area, shows that this is clearly not the case.

Assuming that the date of 4500±50 BP (5307 to 4879 cal. yrs BP) from Skipsea Withow Mere is erroneous, Flenley (1987) propose a date of *c*.2500 (*c*.2800 to *c*.2300 cal. yrs BP) for the opening of this zone. However, the imprecise nature of this rpaz, and the inevitable temporal and spatial variations in the pattern of vegetation change and disturbance, either natural or anthropogenic, means that this rpaz is too poorly defined to be instructive (Dinnin, 1995). The type site for this rpaz is Hornsea Old Mere (lpaz HO5), although it also occurs at many other sites within and beyond the Holderness region (Wright & Churchill, 1965; Clark & Godwin, 1956; Hulme & Beckett, 1973; Beckett, 1975; 1981, Gilbertson, 1984a; Flenley, 1987).

Information on the Holocene vegetation of the Chalk uplands is available from the research by Bush at Willow Garth in the Great Wold Valley (Bush, 1986; Bush & Hall, 1987; Bush & Ellis, 1987; Bush & Flenley, 1987; Bush, 1989; Bush, 1993). Much of this work concentrates on the Lateglacial period, and unfortunately there is a hiatus in the record spanning the period *c*.8000 to *c*.4000 (*c*.9200 to *c*.4200 cal. yrs BP). The Holocene vegetational record begins at 9460±80 BP (10901 to 10220 cal. yrs BP), and points to a cold-climate grassland flora (Bush & Ellis, 1987; Bush & Flenley, 1987; Bush, 1993). *Alnus glutinosa* wood macrofossils are recorded from this moss peat from *c*.9300 BP (*c*.10800 to *c*.10000 cal. yrs BP), which is evidence for the early/continued occurrence of this taxa throughout the Lateglacial and early-Holocene (Bush & Hall, 1987; Bush, 1993). By 9380±80 BP (10796 to 10139 cal. yrs BP) *Betula* woodland expanded in the area, although not to the exclusion of grassland species such as *Helianthemum* and *Campanula* (Bush & Ellis, 1987; Bush, 1993). By *c*.8800 BP (*c*.10000 to *c*.9500 cal. yrs BP) *Betula* had declined and become replaced by *Pinus sylvestris*. This phase is short lived however, and *Betula* populations expand once more in the Wolds, at a time when *Corylus avellana* and *Pinus sylvestris* were dominant elsewhere in lowland Britain (Bush, 1993). Thermophilous taxa, such as *Quercus*, *Corylus avellana*, and *Sambucus* appear for the first time at 8290±80 BP (9445 to 8992 cal. yrs BP, Bush & Ellis, 1987; Bush & Flenley, 1987; Bush, 1993). During this period, there is insect and pollen evidence of the survival of chalk grassland flora, as Poaceae frequencies remain high (20 to 33% dry land pollen) and pollen types such as *Plantago media/major*, *Campanula*, *Vicia*, *Lotus*, *Teucrium botrys*, and *Centaurea nigra* all persist (Bush, 1986; Bush & Ellis, 1987; Bush & Flenley, 1987; Bush, 1993). Cereal-type pollen is also recorded along with herb taxa indicative of disturbance such as Chenopodiaceae.

This provides evidence to suggest that Chalk grassland communities are of considerable antiquity, and were maintained by Mesolithic people, using the Wolds as a hunting ground (Bush & Ellis, 1987; Bush & Flenley, 1987, Bush, 1989; Bush, 1993). Bush & Flenley (1987) suggest that the resurgence of *Betula* *c.*8900 BP (*c.*10000 to *c.*9600 cal. yrs BP), is related to secondary woodland regeneration, which was perhaps managed, and enabled grasslands to prosper. Between *c.*8200 and *c.*8000 BP (*c.*9400 to *c.*8500 cal. yrs BP), *Betula* pollen declines, and *Pinus sylvestris* pollen increases to 43%. Bush & Ellis (1987) and Bush (1993) suggest this may relate to the abandonment of a Mesolithic camp situated on the sandy soils, previously supporting stands of *Pinus sylvestris*, which were able to recolonise the area. The decline in *Betula* may relate to a successional change as *Quercus*, *Alnus glutinosa* and *Corylus avellana* populations expand (Bush & Ellis, 1987). Bush & Flenley (1987), Bush (1989), and Bush (1993) contend that the existence chalk grassland is a continuous feature during the Holocene period, and not just a product of Neolithic woodland disturbance (*cf.* Smith, 1980; Waton, 1982).

The Willow Garth pollen record resumes at *c.*4000 BP (*c.*4800 to *c.*4200 cal. yrs BP), and indicates a largely cleared, open landscape (Bush & Ellis, 1987; Bush, 1993), in contrast to the lowland areas of Holderness (Flenley, 1990). Remnant woodland of *Tilia*, *Quercus* and *Corylus avellana* is indicated, although pollen of Cereals, Poaceae, and *Plantago lanceolata* are abundant, suggesting Chalk grassland, and pastoral and arable agricultural activity was widespread (Bush & Ellis, 1987; Flenley, 1990; Bush, 1993). *Tilia* had all but disappeared by *c.*3400 BP (*c.*3900 to *c.*3400 cal. yrs BP, Bush & Ellis, 1987) and the pollen record from the Wolds shows no sign of subsequent woodland recovery (Flenley, 1987).

A further hiatus interrupts the pollen record between *c*.3300 and *c*.2200 BP (*c*.3800 to *c*.1900 cal. yrs BP), and mineral rich gyttja sedimentation after this time may be related to ploughing of the valley sides leading to soil erosion (Bush & Ellis, 1987). From *c*.1300 BP (*c*.1400 to *c*.1000 cal. yrs BP) the site begins to dry out and a *Salix* fen woodland community dominates from *c*.1100 BP (*c*.1300 to *c*.800 cal. yrs BP), suggesting the site was managed as an osier bed (Bush & Ellis, 1987; Bush, 1993).

Chapter 5 Methods and Techniques

5.1 Introduction

The first part of this chapter introduces the theoretical aspects of the research methodology. The range of techniques available for reconstructing different aspects of environmental change are discussed and the methods adopted in this study are justified. The field techniques used in the collection of borehole data are also presented. The sampling design reflects both the need to locate a suitable borehole site to allow the collection of material for microfossil analysis and to collect sufficient data to enable accurate palaeogeographic reconstructions of the valley sediments. Good spatial coverage is required when collecting borehole data so that the range of sedimentary palaeoenvironments buried within the valley fill can be determined. Specific sites that represent the floodplain palaeoenvironments at 3 stages up-valley were selected for detailed examination using criteria set out below. The techniques used in borehole data collection and description are discussed first, followed by the laboratory procedures necessary for sample preparation, data handling and interpretation. The theoretical background to pollen data analysis is given particular attention. The supplementary palaeoenvironmental techniques that were employed (e.g. loss-on-ignition and wood macrofossil analysis) are discussed and lastly, the radiocarbon dating method is introduced.

5.2 Conceptual Framework of Study

The methodological approach adopted in this research requires a consideration of the theoretical, philosophical and practical underpinnings of palaeoecology. The status of palaeoecology as a science has been the subject of increasing discussion over recent years (see Edwards, 1983; Birks, 1985, 1992; Oldfield, 1993). Traditionally, palaeoecologists have adopted an inductive approach, where an empirical dataset is collected and patterns of vegetation change are identified, usually in the context of reconstructing past environments. This is the normal route followed by a science during the data collection phase (Harvey, 1969). That the theoretical development of palaeoecology is limited can, to some extent, be attributed to the relatively young age of the discipline; it is still only 80 years since Von Post pioneered the subject of Quaternary palynology. Developing more deductive research strategies has proved difficult for palynologists as palaeoecology is essentially a descriptive, historical science, with results that are frequently ambiguous and open to interpretation in a variety of ways. Edwards (1983) recommended that a more critical deductive approach must be adopted in palaeoecology, through the route of hypothesis testing, if the discipline is to progress. However, it can be very difficult to postulate testable hypotheses from palaeoecological records (Birks, 1985; 1992; MacDonald, 1993) as many patterns in palynology are localised and unique and may not be reproducible in time or space.

Shennan (1994; 1995) has recently advocated a move towards deductive research strategies in the study of Holocene sea-level change by adopting the approach of multiple working hypotheses (*cf.* Birks & Birks, 1980). This approach has been successful in developing models of coastal evolution (Shennan, 1994) and also to establish theories concerning a hypothesised coastal response to a given sea-level change (Shennan, 1995).

Whilst it is recognised that a more analytical deductive approach based on methodological falsification of hypotheses is important if science is to progress from narrative to an analytical phase (Birks, 1985; 1992), it must also be appreciated that such an approach is not always realistic. For example, it is feasible to examine the processes controlling the timing and nature of coastal development in an area such as the Fenland, where a large dataset is available (Shennan, 1994), by testing a series of hypotheses, but this approach is difficult to apply to areas such as the inner Humber where there is little data available. As Long *et al.* (1998a; 229) point out, 'For much of the last 20 years, methodological and technical advances in Holocene relative sea-level research have bypassed the Humber estuary…'. Considering the limited sea-level chronology for the Humber, and the lack of litho- or biostratigraphic information available from the lower Aire valley, an inductive approach was adopted, with an aim to describe, categorise and interpret the data to form a series of plausible interpretations and generalisations. At this stage of research, such a dataset can be used to compare with broad patterns in the region and with estuarine events. The development of a deductive methodology using testing and faslification of hypotheses can be implemented further to this study to critically examine the trends observed.

5.3 Evaluation of Techniques for Palaeoenvironmental Reconstruction

The principle techniques that can be adopted for the purpose of investigating palaeoenvironmental change are shown in Table 5.1. The decision as to which of these lines of evidence was to be used for this research depended not only on the nature of the research questions being asked but also on a combination of resource and time constraints which are briefly discussed below.

Palaeoenvironmental Reconstruction	Technique	Major Applications
Local vegetation	Pollen	Palaeoecology
	Plant macrofossils	Palaeoecology
	Fauna (e.g. beetles)	Palaeoecology
Regional vegetation	Pollen	Biostratigraphy (dating)
Palaeohydrology (salinity)	Diatoms	Type of depositional environment and salinity
	Foraminifera	Type of depositional environment and salinity
	Ostracods	Type of depositional environment and salinity
Sediment sourcing	Geochemistry	Identifying sediment provenance

Table 5.1 Principle palaeoenvironmental techniques for reconstructing floodplain environments.

The techniques selected for reconstructing the local and regional vegetation and sedimentary environment were those which offered the most information feasible given the time and resource constraints associated with the study. Pollen analysis is the main technique used (see section 5.6) as it provides an insight into the nature of both the local floodplain vegetation and vegetation communities colonising the wider landscape. However, the taphonomic restrictions of using pollen, particularly in fen and fen carr situations means that sometimes, it is difficult to distinguish between the local (wetland) and regional (dryland) vegetation. One way to address this problem is to analyse the plant macrofossil remains (seeds, fruits, wood, leaves, moss etc.) in the sediments (see Behre, 1986; Grosse-Brauckmann, 1986; Mannion, 1986; Wasylikowa, 1986). Although such analyses provide little or no information about the wider environment, they can provide precise information regarding the nature of the on-site vegetation.

Macrofossil remains may also be particularly useful in identifying plants whose pollen is rarely preserved in fossil sediments (e.g. *Populus*, *Juncus*) or in detecting plants with a low pollen productivity (e.g. *Rubus*).

However, the technique of plant macrofossil analysis does have weaknesses which can limit its usefulness. Although plant macrofossil analysis is frequently used in studies of raised mires (e.g. Barber, 1981), its application to fen carr deposits in Britain is minimal except as a supplement to pollen analysis (e.g. Walker *et al.*, 1998; Waller *et al.*, 1999). In such situations, it is rarely possible to obtain results that can be used in a quantitative manner. This is because macrofossils are produced in small quantities, and consequently a large sample size is required (i.e. *c.*100cm^3) compared to pollen (Mannion, 1986) where a much smaller sample can be used (i.e. *c.*1cm^3). Ideally, an open section is required to sample for plant macrofossils which is rarely possible in studies of floodplain sediments. The analysis of macro-remains in a core sample usually only gleans enough information for a qualitative assessment of the presence of certain species. To add to this problem, floodplain sediments are often comprised of peats with a high clay content (see Chapter 6), which further dilutes the concentration of macro-remains. Also, fragile macrofossil remains are often not preserved due to increasing dessication associated with drainage. For example, the macrofossil study undertaken by Godwin & Clifford (1938) in the Fenland could not be replicated as the remains are highly humified and no longer preserved in the peats (Waller, pers comm.).

Plant macrofossil analysis can be a useful supplement to pollen analysis as identification can be made to species level, which is seldom possible in palynology. In theory, plant macrofossil data may then be used to infer which part of the pollen assemblage is derived from local plant taxa (Rybnickova & Rybnicek, 1971). However, this is often of limited use, as even if, for example, macrofossil finds of *Cicuta* are made, this does not necessarily mean that all the Apiaceae pollen identified in the pollen diagram is from this species. Furthermore, Grosse-Brauckmann (1986) points out that the above ground part of plants are seldom preserved (with the exception of fruits, seeds and mosses), and that it is generally the remains below the wetland surface that form identifiable macrofossils. Since the subterranean parts of plants are intrusive (i.e. they are growing within a peat matrix which is older than the above ground vegetation), they cannot really be considered as being contemporary with the pollen sample obtained from the same stratigraphic position.

Lowe & Walker (1997) warn that detailed investigation of plant macrofossils is more time consuming than pollen analysis, and only recommend they are studied when there are abundant, well preserved remains that can be easily extracted. Borehole investigations (see Chapter 6) revealed very few noticeable plant macrofossil remains apart from woody detritus. Grosse-Brauckmann (1986) also states that a comprehensive reference collection is essential for the correct identification of macrofossils. However, no resources or expertise was available in Hull to assist with identifications. In the light of these limitations it was decided not to undertake plant macrofossil analysis in this study. However, wood macrofossil analysis has been undertaken to provide some indication of the local carr taxa.

Palaeoenvironmental investigations into the minerogenic sediments has been undertaken using analysis of diatoms (see section 5.7). This was preferred to forams or ostracods as the technique was already familiar to the author. As the focus of this research was palaeoecological, geochemical analysis of the sediments, which may establish possible sediment sources, has not been attempted as it was considered beyond the scope of the study. An appraisal of the techniques chosen and the recommendations for future work is given in section 11.2.

5.4 Site Selection

The location of sites for palaeoenvironmental study was primarily predetermined by the objectives (see section 1.3). If the range of depositional environments within the perimarine zone is to be analysed comprehensively, the sites chosen must characterise this spatial variability. Therefore, the collection of a core representative of the freshwater catchment system close to the present tidal limit and two other sites characterising the more estuarine environment seaward were necessary to give good geographical coverage. The lower river Aire provided a good opportunity to study these changes as it forms the innermost extent of the Humber estuary and contains a suite of relatively undisturbed sediments that have hitherto not been studied. Spot samples for microfossil analysis were taken during the initial investigative phase to examine the type of depositional environment encountered and the range of salinity regimes within the reach of the study area. Sites for detailed study were located through a series of cross valley coring transects linked at their downstream extent by a 6 km long profile. These transverse profiles are situated within the perimarine area, between the tidal flat and lagoonal zone, extending close to the tidal limit. Following an extensive coring program, specific sample core locations were determined partially by accessibility but mainly for their suitability for addressing important specific palaeoecological questions.

5.5 Lithostratigraphic Investigations

5.5.1 Introduction

Preliminary stratigraphic explorations revealed sediments that varied greatly both spatially and in terms of thickness. Therefore, it was clear that a high resolution, multiple core sampling strategy would be required to enable lateral correlation of sedimentary units with a high degree of certainty. The strategy also needed to be practical and manageable within the time and resource constraints of the study. Previous lithostratigraphic surveys of coastal river valleys and lowlands in the UK where laterally persistent stratigraphic units occur have used borehole long transects with individual cores initially spaced 100 m apart (Waller, 1987; Waller *et al.*, 1988; Long & Innes, 1995; and Long *et al.*, 1998b) with proven success. For cross profile sections and areas where the stratigraphy is more complex, a closer coring interval is adopted (Waller, 1987; Waller *et al.*, 1988; Burrin, 1988; Plater, 1992; Spencer *et al.*, 1998a; 1998b). Occasionally, the interval may be as little as 1 m where stratigraphic variability is high (Van Dijk *et al.*, 1991; van de Plassche, 1995). In a very detailed study of the fluvial architecture of the complex anastomising and meandering channel systems and overbank deposits of the Rhine-Meuse delta in The Netherlands, Weerts and Bierkens (1993) use a quantitative method to devise a suitable sampling strategy. They use semivariograms to describe the spatial variation of the thickness of overbank deposits which also reveals maximum sampling intervals required for accurate mapping of the sediments. A semivariogram is basically a plot of semivariance (defined as the degree of spatial dependence between attribute values measured at a certain distance apart) against separation distance (Weerts and Bierkens, 1993) which allows for the description and interpolation of geological properties whose values behave irregularly in space. They recommend a sampling distance of 100 m with resolution increasing to 25-30 m for boreholes perpendicular to the main flow where there is more lithological variation (i.e. cross valley profiles), confirming the approach of Waller (1987), Waller *et al.* (1988), Burrin (1988), Long & Innes (1995), Long *et al.* (1998b), and Spencer *et al.* (1998a; 1998b).

5.5.2 Sampling Strategy

In perimarine areas, water regime and sediment supply may be dominated either by estuarine or fluvial processes, with water derived from the sea, the immediate hinterland, or distant sources further upriver. Therefore, the relative distance of a site from the estuary, river or upland is an important consideration, which necessitates comprehensive lithostratigraphic survey of sediments. Transects are a good basis for the study of lateral variability of sedimentary facies changes within Holocene fluvial systems (Törnqvist, 1993a). Conditions are not uniform down a river valley and so at any one time, the sedimentary regime may vary from reach to reach, making it important to investigate the alluvial long profile (Brown, 1997). A long profile transect was carried out between the lowest two sites downstream along a bearing of 60° roughly parallel to the river, taking boreholes every 100 m where possible (see Figure 3.1). This enabled the longitudinal downstream variation in sediment fill of the valley to be established, and was also necessary to locate stratigraphic changes that determined the location of the cross valley transects. The nature of the latitudinal changes in sediment type were determined by three cross valley transects which completed the spatial dimension. These were carried out in the lower Aire and Ouse valleys, with core spaces ranging from 25 to 100 m depending on the width of the floodplain and the stratigraphic complexity.

5.5.3 Coring Methods

All transect cores were carried out by hand using an Edelman type clay auger to remove the stiff upper disturbed sediments. A 1 m long, 25 mm diameter Eijkelkamp gouge corer was used to penetrate the softer sediments below the water table down to depths occasionally in excess of 10 m. Sediment recovery was usually 100% except where channel sands or wood were encountered. The thixotropic nature of saturated sands means that their penetration and subsequent recovery are not possible using open gouge coring techniques. Similarly, the dense nature of large wood remains caused boreholes to be abandoned on occasion and re-bored close by to avoid the obstruction.

5.5.4 Sediment Classification and Description

Accurate and comprehensive borehole recording is a critical first stage in palaeoenvironmental research. These records must be understandable and relied upon months or years after the notes were taken.

Reproducibility and consistency are of utmost importance in the recording of sediments from a multi-core study of inherently variable sediments. Therefore an easy to use, descriptive and qualitative method of recording sediment composition is essential. The Troels-Smith (1955) classification scheme was adopted which enabled the stratigraphy to be recorded in a semi-objective manner.

This stratigraphic recording system has been widely used in palaeoenvironmental studies of coastal stratigraphy and is now almost an international convention among Quaternary scientists. A simple guide to its use is provided by Long *et al.* (1999a) who demonstrate its versatility and applicability to soft sediments from a plethora of palaeoenvironments. A major advantage of the Troels-Smith (1955) scheme is that it is independent of any knowledge of depositional processes and therefore does not presuppose any origin of environment of any sediments. For a full account of the method and Latin terminology used, the reader is referred to the original publication (Troels-Smith, 1955).

The stratigraphic transects have been displayed graphically using the TSPPlus plotting program (Waller *et al.*, 1995) which produces an output based on the Troels-Smith (1955) symbols.

5.6 Levelling

All cores were surveyed to Ordnance Survey (OS) benchmarks by hand using a 'dumpy' level and telescopic staff (Clancy, 1991). Benchmarks give a height in metres above MSL, which is the national Ordnance Datum (OD).

The location and altitude of several benchmarks were obtained from the OS, and temporary benchmark heights were established from these due to the large area over which the cores were situated. Acceptable closing errors of <0.03m (Shennan, 1982) were achieved by carrying out traverses in opposite directions between the benchmark and cores (a survey loop).

5.7 Pollen Analysis

5.7.1 Introduction

This section outlines the criteria used in the selection of sites for palaeoecological study. Pollen analysis is the standard technique used for the reconstruction of past vegetation communities. However, the pollen grains that reach a site are a product of the local, extra local and the regional component (Jacobson & Bradshaw, 1981). Different plants produce different amounts of pollen and have differing dispersal mechanisms which acts to complicate vegetation reconstruction. The distinction between local and regional pollen is a difficulty faced by all palaeoecologists but this may be further obscured when part of the pollen is brought in by rivers. This particular problem is inherent when interpreting pollen derived from clayey peats (Van der Woude, 1983). The chronological significance of fluctuations in regional pollen types may then be restricted (Janssen & Törnqvist, 1991). Therefore, some of the modern theoretical concepts used in the collection, evaluation and interpretation of pollen data need to be considered.

5.7.2 Site Selection

The main aim was to obtain a good geographical coverage of floodplain environments representative of both the lower and higher reaches of the river system. Therefore sites were selected which encompassed the different sedimentary regimes characteristic of the lower and upper reaches of the perimarine zone. This strategy was designed to enable the dominant controls on sedimentation and vegetation to be evaluated. Following the completion of lithostratigraphic survey, a strato-type core was chosen which was lithologically representative of the area (Hedberg, 1976) and contained all sedimentary units with no apparent depositional hiatuses.

The choice of cores suitable for micropalaeontological analysis was partly based on the nature and thickness of the organic sequence. Aiming for longest organic sequences maximises the age range giving better temporal resolution. Pollen cores were sampled away from the floodplain edge to try and minimise the occurrence of wood but marginal enough to reflect regional components of the pollen rain (see Jacobson & Bradshaw, 1981; Waller, 1998).

5.7.3 Sample Collection and Storage

Sample cores were taken using a Stitz piston corer with a sample tube diameter of 50 mm. Cores were extruded in the field and cased between two halves of gutter piping, wrapped in plastic packaging and stored in the dark at *c*.4°C in the Hull University Geography Department cold store. The detailed sediment stratigraphy was recorded in the laboratory using the Troels-Smith (1955) system.

5.7.4 Pollen Sample Preparation

Subsamples of *c*.1 cm^3 of sediment were taken from the core for pollen analysis, initially at 10 cm intervals, with a finer sampling resolution of 5 and 2 cm around critical horizons. Irregularities in this sampling interval are due to the occurrences of wood in the core. Samples were subjected to the standard preparation procedure of Moore *et al.* (1991) for the concentration of pollen and spores. First, the samples were heated in 10% KOH for *c*.10 minutes to allow for the disaggregation of the material and the removal of soluble organic compounds (humic acids). After treatment with 10% HCl, samples were boiled in HF (often twice) to effectively remove all silica. Then the samples underwent acetolysis (to remove unaltered lignin and cellulose), nitric acid treatment if excess lignin or iron sulphide was present, and stained with 0.2% aqueous Safranin solution. Finally the sample was washed in alcohol with TBA used as a dehydrant (Davis, 1966) and mounted in silicone oil ready for counting (Andersen, 1960). For full details of this procedure, see Appendix 1.

5.7.5 Palynological Data Collection

Pollen Counting

Pollen slides were counted using a Nikon Optiphot light microscope under x400 magnification and a x1000 oil immersed objective for the identification of difficult grains. Counting along regularly spaced horizontal traverses to avoid the problems associated with non-random distribution of pollen grains on slides (Brookes & Thomas, 1967) continued until at least 300 land pollen grains (not including *Alnus glutinosa*) were identified. Birks & Birks (1980) suggest that a count between 300 and 500 grains is sufficient to produce a statistically reproducible assemblage. A pollen count must be adequate to enable reproducibility of results and therefore count size must balance the need to reduce the statistical significance of fluctuations or 'spikes' that may be apparent in low counts against time available.

Therefore, to a certain extent, percentages depend upon the size of the pollen count. This issue has been addressed by Maher (1972) who suggests the use of nomograms to calculate the 0.95 confidence limits of counts of various amounts. Nomograms show that as the count increases, the confidence limits rapidly decrease until a point is reached where the analyst must balance the extra work of counting more pollen against the relatively small improvement that a larger count will achieve. A minimum of 300 dry land pollen grains was deemed a sufficient proportion of the total pollen sum to be counted to fulfil these requirements.

Pollen Identification

Pollen was identified to the lowest taxonomic level wherever possible using the keys within Moore *et al.* (1991), Punt *et al.* (1974a; 1974b), Andrew (1984) and the reference slide collection housed at the Geography Department, University of Hull. Often identifications are not possible down to the lowest taxonomic (species) level. In such cases, pollen grains are classified as a 'type' which will consist of a variety of grains of similar morphology. A comprehensive guide as to the taxonomic precision of British pollen types is provided by Bennett (1994a). Pollen grains of *Corylus avellana* and *Myrica gale* have been separated using the pore characteristics described in Moore *et al.* (1991) and consultation with type slide material. However, Edwards (1981) doubts that consistent separation of these two pollen types is possible using a light microscope. This uncertainty is expressed by the use of two categories; *Corylus avellana*-type is used for grains of uncertain taxonomy (resembling *Corylus* and possibly *Myrica*); *Myrica gale* is used for grains that meet the criteria in Moore *et al.* (1991). Therefore, the values for *Myrica gale* can be considered as a minimum of the true abundance of that taxa.

Due to inadequate reference material and poor pollen preservation, grains of *Rumex acetosella* and *Rumex acetosa* are included under the aggregate heading *R. acetosella/acetosa*. Similarly, *Plantago media* and *Plantago major* are labelled as *P. media/major*.

The identification of Cereal pollen was based on the criteria of Andersen (1979). Grains with an annulus diameter of >8 μm were classified as Cereal-type (which not only includes Andersen's *Hordeum* group but also several wild grass species such as *Glyceria*). No attempt has been made to make identifications based on surface sculpturing of cereal pollen or Poaceae grains (e.g. *Phragmites*) due to the lack of well preserved grains. Diameter distribution curves for three grass species show that distinction may be possible using grain size (Clark & Patterson, 1985). However, due to the number of flattened and folded grains present, the sphericity of Poaceae grains could not be readily determined. These problems prevent the separation of Poaceae pollen grains (see Hall, 1991).

The frequent absence of an outer perine made the identification of monoaperturate fern spores (except *Polypodium*) using the criteria of Moore *et al.* (1991) an impossibilty. All such spores are categorised into Pteropsida (monolete) indeterminable following the recommendations of Bennett *et al.* (1994).

5.7.6 Palynological Data Handling

Pollen Concentration/Influx

The addition of a known amount of exotic pollen species to a known subsample volume enables the calculation of the density of pollen grains in the sediment or pollen concentration (Benninghoff, 1962). This enables the determination of absolute pollen fluctuations through time, or pollen influx rates, which are usually expressed as the number of pollen grains settling on a given area in a given time year (e.g. grains cm^2 yr^{-1}). This has an advantage over the usual expression of pollen data in relative terms (percentages) because the abundance of one taxon is not dependent on fluctuations in others (Moore *et al.* 1991). However, the calculation of pollen influx rates does not just reflect the amount of pollen reaching a site but also requires an appreciation of sedimentation rates because variations in pollen concentration are mostly related to changes in peat accumulation (Middeldorp, 1982). Pollen influx values can therefore, potentially provide important information on both the pollen rain and the sedimentary environment. This is not a problem where annual laminations occur in lakes or where a high resolution radiocarbon dating chronology has been established, but its use is much more problematic in fen peats because of their irregular rates of accretion (see Beckett & Hibbert, 1979). The relationship between sediment accumulation rates and pollen recruitment is not simple as they are not dependant on each other. Peat generally consists of autochthonous material, but the pollen is largely incorporated from elsewhere and as a result an inverse relationship between rates of sedimentation and pollen deposition is likely (Bennett, 1983a). The determination of the pollen influx rate is therefore particularly dependent on the rate of sediment accumulation, which has unfortunately been shown to be highly variable in fen floodplain environments (Waller, 1987; Brown, 1988). Pollen influx values have also shown great variation from ombrotrophic peats (Donner *et al.*, 1978; Middeldorp, 1982) where retardations or standstills in peat growth cause exceptionally high influx figures. Problems of sediment compression due to dessication and sediment overburden means that factors other than sediment accumulation rate, such as post depositional changes, can equally be invoked to explain any increases in pollen concentration (see Old Place results in Waller 1987).

Often, absolute pollen frequencies exhibit a high degree of variability between adjacent samples (Brown, 1988; Day, 1991; Waller, 1993; 1998; Long *et al.*, 1998c). If data on pollen influx are to be used for comparison of results within or between sequences, sedimentation must be constant on the scale of the individual sample. However, this too has been shown to be highly irregular, with large temporal variation between adjacent samples, and has been attributed to the lack of sediment homogeneity (Waller, 1993; 1998; Long *et al.*, 1998c). The difficulties involved in volumetric sampling of coarse fen peats (Beckett, 1979) exacerbate this problem and are a serious limitation on the use of absolute pollen frequencies in these contexts regardless of the number of radiocarbon dates obtained. Moreover, the deposition time at the sampling point is not known, but only the average deposition time for the section of sediment between the radiocarbon dates (Bennett, 1994b) which can result in spiky pollen influx curves.

As a general rule, usually only percentage data are available for comparing fossil and modern pollen data, as only percentage data are available for modern samples. Percentage data have proved useful in regional studies and for revealing information about vegetation and vegetation dynamics because they are much less variable than absolute values of pollen deposition among sites within a region (Davis *et al.*, 1973; Webb *et al.*, 1981; Grimm 1988). In recognition of these drawbacks, it was decided not to use absolute pollen concentrations for this study.

The Pollen Sum

A major feature of pollen analysis which has implications for subsequent interpretation is the use of percentages in data analysis. This results in interdependence; fluctuations in one pollen taxon have a knock-on effect on all other types regardless of whether a change in the parent vegetation of these types has occurred (Faegri & Iverson, 1989; Moore *et al.*, 1991). Another problem is caused by the choice of which taxa to include within the pollen sum. An arboreal pollen sum was used by early workers (Erdtman, 1928; Godwin, 1934; 1940) because they were primarily concerned with the reconstruction of forest history and because many herbaceous pollen types were not recognised at that time (Moore *et al.*, 1991). Even some important arboreal types such as *Fraxinus excelsior* and *Taxus baccata* could not be distinguished. However, in many situations, it is clear that fully forested landscapes did not exist all the time during the period of interest. The use of a sum based solely on arboreal pollen in situations where fully forested conditions did not persist results in the under-representation of the non-arboreal component in pollen diagrams.

Specific problems arise when plants growing locally on the sediment surface produce large amounts of pollen. If this component is included within the pollen sum then it will dilute or mask the signal of changes in vegetation occurring elsewhere which may be of interest to the investigator. This problem can be overcome by selectively removing over-represented local taxa from the pollen sum (Rybnickova & Rybricek, 1971). For example, *Alnus glutinosa* is a high pollen producer (Andersen, 1970, 1973) and as a result, it is over-represented in many pollen assemblages (Janssen, 1966, 1984). Dominance of *Alnus glutinosa* is a striking feature in the pollen spectra in this study due to its anemophily and autochthony so, following the recommendations of Janssen (1959), it has been excluded from the pollen sum.

Other attempts to overcome these difficulties are presented by Tooley (1978a) and Shennan (1982) who group taxa into life-form classes (*cf.* Wright & Patten, 1963), so that percentages can be calculated from a basic sum (% arboreal pollen) to which one of 4 ecological groups are added (shrubs, herbs, aquatics and spores). The exclusion, separation or grouping of taxa so that they are expressed as percentages of several sums prevents some types from adversely influencing others which are in the basic sum, but to a certain extent, it also presupposes that pollen types can be unequivocally separated into their environment of origin. Waller (1993) emphasised that such a precision is not easy to make mainly because of the lack of taxonomic precision in pollen identification which potentially leads to considerable overlap in establishing the environment of origin. For instance, many of the herbs used as traditional indicators of anthropogenically cleared environments (*cf.* Behre, 1981) also occur naturally in fen systems or coastal environments (Wheeler, 1980a; 1980b; Jones, 1988; Rodwell, 1991; 1995; Wiegers, 1992).

More recent workers have stressed the importance of including all members of an assemblage within one pollen sum (Birks & Birks, 1980; Moore *et al.*, 1991). Following these recommendations, it was felt that for this study the pollen sum should be kept as complete as possible and for judgements as to the significance of changes in particular taxa to be made only at the interpretation stage. The basic sum used is Total Land Pollen (TLP) excluding *Alnus*; this includes all tree, shrub and herb types, and involves no assumptions regarding local/non-local sources. For non-terrestrial pollen types the sum used is TLP excluding *Alnus* plus a group.

The pollen sums used are as follows:

TLP-*Alnus* - All terrestrial land types expressed as % of sum minus *Alnus*

TLP+*Alnus* - Expressed as a % of the TLP sum (of which it forms a part).

TLP-*Alnus*+aquatics - Expressed as a total of land pollen types minus *Alnus* but including obligate aquatics.

TLP-*Alnus*+spores - Expressed as a total of land pollen minus *Alnus* but including pteridophytes.

TLP-*Alnus*+pre-Quaternary Spores (PQS) - Expressed as a total of land pollen minus *Alnus* but including pre-Quaternary Spores.

All types are included within the sum in which they are expressed (Mosimann, 1965; Faegri & Iverson, 1989) to avoid percentage frequencies in excess of 100. Obligate aquatics have been excluded from the basic sum as there is little doubt as to their separate environment of origin. Spores are excluded due to their incompatibility with pollen. Spores are usually omitted from the pollen sum as they have a different function to pollen, therefore there is no basis for assuming changes in their abundance reflect actual changes in vegetation (Faegri & Iverson, 1989; Moore *et al.*, 1991). All data are displayed graphically using the TILIA program by Grimm (1993).

Pollen Zonation

Biostratigraphic data, particularly those from long sequences, are too cumbersome to be described in their entirety. Therefore, it is conventional to subdivide pollen diagrams into smaller units or zones to aid in interpretation of the salient features. In addition to ease of description, the diagram is subdivided into relatively consistent and homogeneous pollen groups (pollen assemblage zones) that can be discussed together and enable inter-site comparison with other such zones. This is in accordance with Hedbergs' (1976) biostratigraphic assemblage zone. Birks (1973) advocates the identification of local pollen assemblage zones (lpazs) specific to single diagrams which then enable the description, discussion and comparison of regional pollen data (Birks & Birks 1980).

The delimitation of pollen zones has in the past been made by visual inspection by the investigator. However, more recently quantitative techniques have been introduced which reduce the subjectivity and bias of the visual inspection process (Birks & Berglund, 1979) and allow consistent application of the criteria (Gordon & Birks, 1972). For this reason, numerical techniques are preferable and the stratigraphically constrained statistical zonation method CONISS (Constrained Incremental Sum of Squares) contained within the TILIA package (Grimm, 1993) has been used for zone calculation allowing equal treatment of all types (Grimm, 1987). CONISS is based on cluster analysis which are formed by hierarchical agglomeration of stratigraphically adjacent samples. The analysis examines the amount of variance associated with each cluster and indicates significant differences between samples that form potential zones. The method of cluster analysis for the pollen and diatom data is based on a matrix of Euclidean distances giving equal weight to all pollen taxa regardless of their respective variances. Stratigraphically constrained data analysis has been used, with no data transformation (Grimm, 1987).

Multivariate methods can objectively define and compare pollen assemblages in time and space, thus eliminating the need for detailed zone descriptions (Birks & Gordon 1985). Bennett (1996) has shown that after rigorous simulation of various techniques of numerical analysis, that the number of zones identified is affected by the number of samples in the dataset as well as the method of zonation used and introduces a method whereby the significance of zones can be assessed. This is particularly important when the aim of the research is the identification and correlation of chronostratigraphic units.

It is recognised however that pollen zonation is a simplification of events. Similarly, the act of sectioning a diagram into zones implies that one can characterise a diagram by periods of uniformity interspersed by repeated phases of rapid change (Walker, 1982). Although palaeoecologists tend to focus attention to periods of change, zones also highlight the constancy and stability of vegetation systems (Bennett, 1988) which is a prominent feature of many pollen diagrams from perimarine floodplain areas where *Alnus* carr dominates for thousands of years (e.g. Brown, 1988; Waller, 1993; 1994a; Waller *et al.*, in press). However, care must be taken when dealing with events that are assumed to be of chronostratigraphic significance which are subsequently used as biostratigraphic boundaries (e.g. the *Ulmus* decline) even though the assumed change does not indicate a major change in the pollen assemblage (Birks, 1974).

Bennett (1996; 169) defines a biostratigraphic assemblage zone as "the smallest significant unit found in a sequence following numerical analysis". He also advocates the usage of 'superzones' (*sensu* Tzedakis, 1994) where changes in pollen assemblage are of high significance, and suggests the use of 'subzones' for lower order groupings where the unit may not be of particular statistical significance but is worth remarking upon. The equal treatment of all types means that often significant biostratigraphic zone boundaries are indicated by CONISS but are obviously related to insignificant fluctuations in a pollen type (usually of local origin). In such cases, where the integrity of the zone is in doubt or the boundary reflects a consistent shift in several local taxa, sub-zones have been used. Therefore some subjectivity has been used in these cases on ecological grounds (*cf.* Faegri & Iverson, 1989) although the recommendations of Birks (1992) have also been considered, and zonation procedure was primarily determined by CONISS.

5.7.7 Interpretation of Pollen Data

The problem with pollen analysis is that the pollen deposited at a site does not necessarily directly correspond to the surrounding vegetation. As well as problems associated with the variability of random depositional processes, there is bias in both pollen production and dispersal, factors which have long been recognised (Tauber, 1965; Erdtman, 1969). This is also complicated by local conditions. For example, *Corylus avellana* flowering is suppressed as an understorey shrub, but is prolific when growing in open areas or as a canopy tree (Godwin, 1975a; Rackham, 1986). Before an accurate interpretation of palynological data can be made, which usually translates as the reconstruction of a past vegetation community, a detailed understanding of the nature of palynological data is required.

The fact that anemophilous plants (wind pollinated), in general, produce more pollen grains per anther than entomophilous plants which are insect pollinated (Erdtman, 1969; Whitehead, 1983) has led to the development of methods which calibrate pollen representation against forest composition. This has usually been undertaken to improve the recognition of regional vegetation changes through studies of this relationship in modern conditions. Davis (1963) uses a correction factor (R-values) based upon the ratio of pollen in a surface assemblage to the abundance in the surrounding vegetation. A major limitation of this method is that it depends upon the radius from which pollen is assumed to have come (source area). The source area contributing pollen to any one site is usually unknown and depends on many factors from vegetation structure to local topography (Moore *et al.*, 1991). The application of this technique is therefore restricted to forest environments where the pollen catchment is very small and well defined (e.g. Bradshaw, 1988). Parsons & Prentice (1981), aware that R-values depended on the correct sampling radius being chosen for estimating forest composition, modified the R-value using statistical methods.

Andersen (1967; 1970) attempted to overcome the problem of dispersal bias and unknown source area by investigating pollen fallout in a Danish forested area where the pollen is likely to have been derived from a very small area around the sampling site. He sampled moss polsters which he was confident contained the pollen from a catchment of no larger than a 30 m radius under the assumption that all pollen originating from outside the area consisted of a background component common to all sites. Andersen's (1970) correction factors have been successfully applied to other sites within forests such as small hollows and soil profiles, (similar situations to those from which they were derived) to adjust pollen percentages to account for differential production (Andersen, 1978; Aaby, 1983). Work by Bradshaw (1981a) in southern England broadly agrees with Andersen's data and confirms the general rank order of tree pollen production. However, there are considerable restrictions on the use of Andersen's correction factors since they cannot be directly applied to any situation other than closed sites located directly beneath a forest canopy (Moore *et al.*, 1991). Other constraints are imposed by the confinement of these methods to tree taxa and by the extrapolation of relative pollen productivity of tree species, in a limited number of species mixtures, back into the past. Andersen (1980) contends that because pollen filtration in the canopy is largely impartial (Andersen 1974a; 1974b) correction factors are applicable to sites with large source areas. However, Prentice (1985) maintains that in such situations, there must be a dispersal bias with heavier grains being removed more rapidly from airstreams than lighter grains (*cf.* Gregory, 1973). Similarly Gearey & Gilbertson (1997) find that concentrations of woodland pollen falls from >20% within a plantation, to background levels within <40m from the woodland edge. Clearly there are differences in pollen taphonomy between different sites and from different geographical environments which are highly variable.

Later studies on the relationship between pollen source area and basin size have been carried out by Prentice (1985). He presents a theoretical model to explain different means of pollen recruitment that can predict pollen source areas and representation characteristics for any basin size. Field data presented by Bradshaw & Webb (1985) supports the theoretical predictions of Prentice's model. Jackson (1990; 1994) further evaluated the theoretical model of pollen dispersal, representation and source area by sampling set distances away from the shores of small ponds in North America. His results are consistent with the predictive model of Prentice (1985; 1988), that pollen source area is positively related to basin size, although he concludes that more comprehensive tests were needed to extract vegetational information from pollen data with accuracy and precision.

Similar work by Janssen (1966; 1973; 1984) has been undertaken along spatial gradients from closed forest to open wetland or across vegetated islands in a peatland. He studied the relationship between modern pollen assemblages and vegetation and demonstrates it is possible to spatially delineate the source areas of pollen production (e.g. local and regional) that contribute to the pollen assemblages encountered, which reveals the fine-scale pattern of peatland vegetation. He also shows the significant effects of local plant populations on pollen assemblages, as pollen from some species are transported laterally by only a few metres. However, the nature and magnitude of local effects are poorly known (Jacobson & Bradshaw, 1981; Birks & Gordon, 1985).

More recent modifications of Prentices' (1985) model (which only enables estimates of pollen deposition to be made at lake centre locations) have been conducted by Sugita (1993) who developed a model to estimate pollen deposition over an entire lake surface. This overcame the problem of sediment focusing and pollen redistribution and highlights the importance of vegetation growing in nearby sources around a lake (Bradshaw & Webb, 1985; Jackson 1990). Sugita (1994) then conducted a computer simulation to define vegetation heterogeneity which provides a theoretical insight into pollen representation of patchy vegetation that empirical studies and simplified models cannot provide (Prentice, 1985; Sugita, 1993). Calcote (1995) carried out a spatially defined vegetation survey which tested Sugita's 1994 model for lakes and applied it to predict 'relevant pollen source area' (an estimate of the scale at which fossil pollen in sediment reflects the surrounding vegetation, Sugita, 1993) for forest hollows. This empirical test confirms Sugita's (1994) model results of vegetation heterogeneity and the estimates for relevant pollen source area, beyond which, little additional information is gained by sampling a larger area (Sugita, 1994).

Despite a number of comparative pollen-vegetation studies which attempt to calibrate pollen assemblages from lake and forest environments, they lack sufficiently detailed information to address this issue (Bradshaw & Webb 1985). Similarly, in palaeovegetation studies, we still cannot distinguish between a few species growing near the pollen site and many individuals at a much greater distance away, a problem initially recognised by Oldfield (1970). Despite the vast amount of theoretical and empirical work carried out to address the issue of pollen source areas and representation (see reviews in Bradshaw, 1994; Jackson, 1994), they only apply to restricted circumstances (i.e. closed canopy environments or lake basins of certain sizes). Correction factors have therefore not been applied to this work due to the lack of a clear empirical understanding of models of pollen populations from comparable river and tidally influenced floodplain environments.

Pollen dispersal bias becomes particularly important when pollen grains are transported over some distance to the site. Factors such as weight of grains, time of year of pollen production and mode of transport result in differential dispersion (Gregory, 1973; Greig, 1982a; Huntley & Birks, 1983). Knowledge of the size of the pollen source area is fundamental to the interpretation of pollen data, as it will also determine the scale of reconstruction which can be attempted from a given site. This is especially pertinent to this study where pollen is potentially incorporated into the sediment from conventional pathways, e.g. atmospheric fallout, and also from river deposition (Van der Woude, 1983; Scaife & Burrin, 1992). Similarly, the relatively open nature of marginal riverine and coastal environments invites pollen transport from fairly large areas which are likely to change with time. The complex relationship between the pollen source area, route of pollen transportation and size of sampling site is therefore worthy of further consideration.

The pollen source area for a site can be divided into 4 components - local, extra-local and regional (Janssen, 1966; 1973). Jacobson & Bradshaw (1981) have defined these terms more quantitatively: local pollen being derived up to 20 m from the edge of the sampling basin; extra-local pollen derived from between 20 and several hundred metres from the edge of the sampling area, and regional pollen from even greater distances. Unfortunately these terms have not been used consistently (see Prentice, 1985 for a different set of definitions). An additional problem is that this simple model is based entirely on aerial transport. Water transport of pollen from both fluvial and tidal sources may alter the distribution of pollen under such circumstances (Traverse, 1990; Scaife & Burrin, 1992; Chmura, 1994). Therefore extra-regional pollen deposition, the fourth term devised by Janssen (1973) is applicable in such cases to describe allochthonous pollen grains transported from beyond the area of vegetation formation being studied (beyond 200 km, Prentice, 1985). Here, a more ecological division can be envisaged, with local pollen originating from the floodplain environment, extra local from the floodplain edge and the first few hundred metres of the dry land source area and regional pollen from further afield is preferred. This is justified because distinct communities are likely to occur in the local environment.

Tauber (1965; 1977) first attempted to define the modes of transport of pollen to a site, suggesting that within forests, pollen is transported above the canopy, through the trunk space, by rainfall, surface run-off, and streams. Jacobson & Bradshaw (1981) also define a gravity component, to describe vertical deposition. Pollen is unlikely to travel any distance by the gravity component or trunk space, where wind velocities are low and filtration rapid. Long distance transport will occur above the canopy and particularly the rainfall component. Jacobson & Bradshaw (1981) develop the ideas of Tauber (1965; 1977), and suggest that the size of the basin will influence the proportion of pollen transported to the site via each of these components. For example, in a small basin, a forest soil will receive more pollen from the gravity and water components, with some from the trunk space, and hence most will be of local origin. Conversely, large lakes should receive more pollen from the canopy and rainfall components and the pollen assemblages should reflect the regional vegetation. Other studies have suggested that most pollen reaches lakes via surface run-off and stream transportation (Peck, 1973; Pennington, 1979; Brown, 1985) and hence is largely derived from the catchment (Bennett, 1986). Jacobson & Bradshaws' (1981) model is restricted in its applicability, as it applies to only a limited number of palaeoenvironmental settings.

Recent palynological studies from floodplain backswamps and oxbows (e.g. Brown & Keough, 1992b; Moores *et al.*, 1999) indicate that the pollen source is spatially restricted to predominantly local and extra-local sources, and therefore primarily reflects that valley floor vegetation (Brown, 1996). However, the high influx of minerogenic sediments in many floodplain deposits suggests that much of the pollen may be derived from further afield, and is deposited with the washing-in of sediment. Although flood water has been shown to contain high amounts of suspended sediment and pollen (Brown, 1985), the associated flood deposits contain low proportions of pollen (Brown, 1985; 1996). Despite the reservations concerning pollen taphonomy in alluvial deposits and intertidal sediments, palynological analysis of such sediments has provided ecologically coherent information, which suggests reliable results can be obtained from such contexts (e.g. Scaife & Burrin, 1984; Hall, 1989; Schirmer, 1991; Scaife & Burrin, 1992; Long *et al.*, 1999b). Scaife & Burrin (1992), Brown (1996), and Caseldine & Barrow (1997) present modified versions of the Tauber (1965) and Jacobson & Bradshaw (1981) models which help in the understanding of pollen taphonomy in fluvial, perimarine and coastal settings.

In theory, the models by Jacobson & Bradshaw (1981) and Bradshaw (1988) suggest that it could be possible to choose sites which relate to the spatial scale of the vegetation reconstruction desired. However, within this study, it is the type of sites which is limited, and the main problem is to resolve the source area for the available sites. Despite the literature on pollen dispersal, fen environments are rarely considered (in contrast to lakes, bogs and saltmarshes) and the modern studies which have been undertaken (e.g. Turner, 1964; Tinsley & Smith, 1974; Caseldine, 1981; Innes *et al.*, 1993; Shennan *et al.*, 1995b; 1996; 1998) are not from comparable situations. The use of theoretical models such as that provided by Jacobson & Bradshaw (1981) presupposes a clear understanding of the environments involved, both the source area and the depositional environment. In order to test these hypotheses, data from modern analogous undisturbed floodplain environments must be obtained, which are no longer in existence in the UK (Brown, 1997). Because these requirements cannot be met, the investigations of Waller (1998) into pollen dispersal in a floodplain wetland with distance from the dryland edge are used as a model to help with interpretations. In this study, Waller (1998) compares 4 pollen diagrams from the Brede valley, south-eastern England, constructed from peat cores at distances of 40, 80, 150 and 500 m away from the dryland edge. The results of detrended correspondence analysis (DCA) show that higher proportions of poorly dispersed types such as *Tilia* and cereals occur at sites close to the floodplain edge reflecting the increased input of extra local pollen, the values of which fall away with increasing distance into the floodplain wetland. Also curves for local taxa are often not reproducible between sites indicating that there is considerable variation in the distribution of fen vegetation between sampling sites (Waller, 1998).

5.8 Diatom Analysis

5.8.1 Introduction

Diatoms are microscopic unicellular algae belonging to the class Bacillariophyceae, which live in a range of habitats wherever there is water (Barber & Haworth, 1981). The exoskeleton or *frustule* is composed of highly resistant silica (Lewin, 1962) which enables their preservation in fossil sediments. Diatoms are a valuable aid to palaeoenvironmental reconstruction because of their sensitivity to a range of environmental parameters (e.g. Mannion, 1982; Battarbee, 1986; Dixit *et al.*, 1992). Some diatom species have specific ecological requirements and as such respond to subtle changes in environmental conditions such as salinity, pH, light, temperature, dryness, water velocity and turbidity (Admiraal, 1977a; 1977b; Mannion, 1982; Brugam, 1982; Brown & Barber, 1985; Battarbee, 1986; Dixit *et al.*, 1992). As a consequence they have been used to address a wide range of topics: e.g. the effects of human activities upon lakes and coastal systems such as eutrophication (Battarbee, 1978; 1984; Anderson & Vos, 1992), reconstructing sea-level change (Devoy, 1979; Palmer & Abbott, 1986; Long, 1992; Zong & Tooley, 1996; Zong, 1997; Denys & de Wolf, 1999), identifying tsunami deposits, (Hemphill-Haley, 1996) and as an aid to archaeological investigation (Juggins & Cameron, 1999).

In this study diatoms are used as an aid to interpreting the environment of the inorganic deposits where the pollen record cannot be relied upon due to poor preservation and the possibility of high proportions of tidally reworked derived pollen. In particular, they are used to reconstruct past hydrological conditions, such as salinity, so that the relative importance and extent of estuarine water penetration up the river valley can be assessed during the Holocene. For a review of the methods and applications of the study of diatoms, see Mannion (1982), Battarbee (1986), and Stoermer & Smol (1999).

5.8.2 Diatom Sample Preparation

Diatom samples were taken in the same way as those intended for pollen analysis. Subsamples of 1 cm^3 of sediment were taken and subjected to digestion in hydrogen peroxide to remove all organic material (Barber & Haworth, 1981; Palmer & Abbott, 1986) and glued to slides using Naphrax which has a suitably high refractive index (≅1.7, Cooper, 1999).

5.8.3 Diatom Data Collection and Handling

Diatoms were routinely counted under a magnification of x400 and x1000 with oil immersion for difficult or fragmented valves. This continued until at least 200 but usually 300 total valves were counted. Where a valve was broken but recognisable, it was counted as 'whole' if more than half remained. Diatoms can usually be identified to species level. Identifications were made by reference to the photographs and sketch drawings in Van der Werff & Huls (1958-1974), Hendey (1964), Germain (1981), Carter & Bailey-Watts (1981), and Hartley *et al.* (1996).

5.8.4 Interpretation of Diatom Data

Interpretations are based on the salinity group classifications (based on Hustedt, 1957) of Vos & de Wolf (1988; 1993a) and the ecological codings of de Wolf (1982; 1993), Denys (1991; 1994) and Van Dam *et al.* (1994). Data are expressed as a percentage of total diatom valves (%TDV) and are displayed graphically using the TILIA program of Grimm (1993). Diatom diagrams have been zoned using CONISS; the stratigraphically constrained statistical cluster package contained within TILIA (Grimm, 1987, see section 5.5.6 for discussion of zonation). Definitions of the terminology used diatom habitat preferences and ecological requirements are given in Appendix B.

5.9 Wood Thin Sectioning

The cores sampled were found to contain significant amounts of fossil wood. The presence of large wood remains is a hindrance to coring and pollen sampling. However, establishing the provenance of wood that may be non-local can reveal important information regarding the processes operating in the depositional environment. The identification of the species of wood contained in the cores may also be an important aid to pollen interpretation because it may help establish the nature of the local vegetation since macrofossils, wherever found, are unlikely to have been moved far from their place of origin (Birks & Birks (1980). AMS dates on wood macrofossils have been used in this study so the identification of the species provides a check that it is not anomalous for the general time period, or an allochthonous component of the organic assemblage of the site. Following the advice of Maisie Taylor (pers. comm.) samples were frozen and sections cut by hand using a razor blade. Identifications were made using the keys and photographs in Wilson & White (1986), Schweingruber (1990), Godwin (1956).

5.10 Loss on Ignition

The determination of organic content was of primary importance in this study due to the clay rich nature of the organic floodplain deposits. This was necessary to establish the amount of carbon in the sediments to select large enough samples for radiocarbon dating. It is also useful in interpreting lithostratigraphy because it enables more precise definition of diffuse sedimentary boundaries. Although it can be used in palaeolimnology as a guide to biological productivity (Lowe & Walker, 1997), it is most often used in conjunction with pollen analytical and mineralogical techniques to identify periods of soil erosion and inwashing of sediment in to lakes in relation to human activities and vegetation disturbance (Pennington, 1965; Oldfield, 1977; Jones *et al.*, 1985 and Lageras & Sandgren, 1993).

Recently, loss on ignition has been somewhat superseded by more modern, precise sedimentological and geochemical techniques such as mineral magnetics (see Oldfield & Clark, 1990), and its value in comparison with these new methods has been put into question (Lageras & Sandgren, 1993). In this study, it has simply been used as a way of supplementing the Troels-Smith (1955) lithological descriptions (which often under-estimate clay content in sediments, see Long *et al.*, 1999a) and to help with sampling for radiocarbon dates. It has not been used as primary data for interpretation but has been used to support inferences made from the pollen results where necessary. The method used is described by Bengtsson & Enell (1986) in which the amount of carbon in a sample is indicated by the percentage weight loss-on-ignition following combustion in a furnace at 550°C for 3 hours.

5.11 Radiocarbon Dating

5.11.1 Introduction

Radiocarbon dating (^{14}C) has revolutionised Quaternary science by providing an absolute chronology for the late-Devensian and Holocene periods. It works on the principle that the ^{14}C isotope is formed continually from nitrogen in the atmosphere by the action of cosmic rays and was first devised by Libby (1952). Carbon dioxide containing the unstable isotope is continually exchanged in the biosphere through photosynthesis and indirectly by other life forms through the food chain. Thus all things living contain the same amount of ^{14}C but upon death the equilibrium is lost, replacement of ^{14}C ceases, and the radiocarbon will begin to decay back to nitrogen. The rate of decay follows a negative exponential curve and the point at which the amount of remaining ^{14}C falls below the limit of measurement, provides an upper limit of *c.*50000 years for the technique (Pilcher, 1991). For samples younger than this, the constant decay rate can be measured and is expressed as the half-life; the time taken for half the isotope to decay. The half-life for ^{14}C used for radiocarbon date calculation is 5570±30 years (Mook, 1986) and by measuring the ^{14}C in a sample, an age can be calculated from the time elapsed since death. For details on the measurement of ^{14}C activity see Bowen (1978) and Mook and van de Plassche (1986). A simple and comprehensive guide to its principles, uses and applications can be found in Bowman (1990).

A recent advance in the technique has been with the development of Accelerator Mass Spectrometry (AMS) dating. This enables the amount of ^{14}C atoms in the sample to be determined, rather than relying on the detection of the decay products as in conventional ^{14}C dating (Linick *et al.*, 1989). AMS dating allows the dating of very small samples which has increased the range of datable sedimentary deposits (Hedges, 1991). For example, it has enabled the dating of annually laminated sediments (Hajdas *et al.*, 1995; Oldfield *et al.*, 1997), small plant macrofossils (Zbinden *et al.*, 1989), calcareous microfossil assemblages (Austin *et al.*, 1995; Heier-Nielsen *et al.*, 1995), insect assemblages (Hedges *et al.*, 1989), and even pollen grains (Brown *et al.*, 1989; Regnell, 1992).

5.11.2 Sample Selection: Problems and Contamination

When dating sediments from fluvial systems there is often a possibility of bulk samples being contaminated with old carbon derived from catchment erosion and reworking of earlier deposits and subsequent redeposition in floodplain sediments during times of high river flow (Blong & Gillespie, 1978; Törnqvist *et al*, 1992). Schoute *et al.* (1981) found that when dating vegetation horizons the organic content of the deposit contained resistant older organic debris transported by rivers. Similarly, Dickson *et al.* (1978) encountered problems associated with inwash of material containing old carbon during the onset of a marine transgression of the Clyde estuary in to the Loch Lomond area. However, since the proportion of allochthonous carbon is usually low, only sediments low in organic matter are liable to suffer this contamination (Olsson, 1972; 1979). This is also a problem associated with the radiocarbon dating of deposits of late glacial age (Donner & Jungner, 1974; Sutherland, 1980; Walker & Harkness, 1990). Mook & Waterbolk (1985) have shown how a sample containing a 1% old carbon contamination results in an age 80 years too old.

The incorporation of reworked particles is a problem that potentially affects all peat samples although the overall effect on the final age will be much diluted if the sample is dominated by *in situ* organic material (Shore *et al.*, 1995). Various experiments have been undertaken to try and address this problem either by dating different fractions (humic, humin and fulvic acid) of the peat (Dresser 1971; Walker & Harkness 1990; Bartley & Chambers 1992; Tipping, 1994; Shore *et al.*, 1995), or by comparing bulk sample dates with AMS dates performed on macrofossils (Lowe *et al.*, 1988; Törnqvist *et al.*, 1992). Studies have suggested that different fractions of the sample yield different dates. For instance, Johnson *et al.* (1990) and Tipping (1994) suggest humic fraction dates are more reliable whereas Bartley & Chambers (1992) recommend a chronology based upon humin dates. Shore *et al.* (1995) also demonstrate that distinct variations in the dating of different chemical fractions can occur. They were unable to demonstrate any consistent trend in the different dates either within or between sites and question the validity of different pre-treatments of the various radiocarbon laboratories. Therefore, it seems that the observation by Fowler *et al.* (1986), that finding an ideal chemical compound for dating for the many different types of context, does not seem likely.

As a result of this uncertainty and the high cost of dating separate peat fractions, mainly conventional bulk samples have been used in this research. This is in accordance with the recommendations of the NERC Steering Committee who test dated some preliminary samples from a site at Eskamhorn (see Chapter 8) against the pollen based biostratigraphic chronology which acted as 'range finders'. The results of these dates are in stratigraphic order, and in accordance with the pollen defined chronohorizons (e.g. *Ulmus* decline), and are accepted as valid. However, some of the dating horizons from the Hook Lane site contained up to 40% (TLP-*Alnus*+PQS) pre-Quaternary spores (see Chapter 9) which brought into question the integrity of bulk dates from these samples (*cf.* Törnqvist *et al*, 1992). Tankersley *et al.* (1987) warn that the recognition of certain fossil spores in palaeobotanical analysis are indicators of the presence of bituminous coal contamination which can result in ages for samples many thousands of years too old and that standard laboratory procedures are usually unsuccessful in removing it. However, Nambudiri *et al.* (1980) suggest that the percentage of pre-Quaternary microfossils in a sample can be used as a guide to the amount of old carbon and may even provide a basis for correcting any anomalous dates.

In recognition of this possible contamination, the bulk dates from Hook Lane have been coupled with AMS dates on wood macrofossils from adjacent stratigraphic levels to enable the reliability of the conventional dates to be tested at this site. The use of dates on wood macrofossils has been queried due to problems associated with provenance (Streif, 1972) as well as the 'old wood' age offset (Bowman, 1990). This occurs when wood that is not *in situ* is dated, and will only give a minimum age for the deposit which may considerably pre-date deposition (Brown, 1997). However, this was the only viable option due to the absence of smaller less transportable non-woody fragile macrofossils of known local origin (*cf.* Törnqvist & van Dijk, 1993). Despite these disadvantages, Shore *et al.* (1995) have stated that the smaller the sample, the greater the effect of any reworked particles on the sample will be. Also Brown (1997) points out that sites where wood is preserved by burial are only occasionally sites of subsequent erosion and that wood that is not buried but exposed on the floodplain decays rapidly. Dating of wood has provided an accurate chronolgy in previous studies (*cf.* Nelson *et al.*, 1988; Moores *et al.*, 1999). Therefore the double dating of bulk samples with AMS dates on wood macrofossils was deemed a satisfactory method to address the problem.

5.11.3 Calibration of Radiocarbon Dates

The fundamental assumption that underpins both AMS and 'conventional' ^{14}C dating is that the production of atmospheric ^{14}C has remained constant in time and space. However, de Vries (1958) showed that this principle is violated, at least in the short term of a few hundred years, as wood samples collected in the 17th century contained greater than expected levels of ^{14}C. Dendrochronological cross-matching with the Bristlecone-pine tree record confirmed that ^{14}C production had fluctuated through time (Suess 1970; 1978; 1980; Suess & Linick, 1990). The major difficulty with radiocarbon dating therefore is that the amount of ^{14}C in the atmosphere has varied over the Holocene, with for instance, levels being significantly higher *c*.6000 and *c*.2600 years ago relative to the present day (Bell & Walker, 1992; Van Geel *et al.*, 1996). At present there is no way of predicting the fluctuations in ^{14}C production in the atmosphere. The only means for creating a correction factor for changes in the past is to create an empirical calibration curve by radiocarbon determinations on already dated dendrochronological sequences to enable identification of calendar age/radiocarbon age differentials to be identified (Bartlein *et al.*, 1995). In this way, radiocarbon years can be converted into equivalent calendar years over the past *c*.18000 ^{14}C years (Stuiver & Reimer, 1993).

Superimposed on the calibration curve are '**Suess wiggles**'. These are non-random short term changes in the production of atmospheric ^{14}C on an approximate 100 year timescale (Suess, 1980; 1986; Stuiver *et al.*, 1991; Stuiver & Braziunas, 1989; 1993) which cause steps and plateaus to occur in the relationship between radiocarbon years and calendar years. These features lead to indeterminacy of ^{14}C ages and existence of multiple calibrated ages for ^{14}C ages from certain time periods (Stuiver *et al.*, 1986). The calibration relationship is not a smooth one and consequently a single radiocarbon date often represents a band of several ages. This effect is more serious in some time periods than others, producing what are referred to as a '**radiocarbon plateaus**'. This causes the spurious clustering of radiocarbon ages around specific time periods. For example, Van Geel *et al.* (1996) show that a rapid rise in atmospheric ^{14}C occurred on a radiocarbon time-scale between 2750 BP and 2450 BP (300 radiocarbon years) but that this equivalent phase lasted for only *c*.90 years on a calendar time-scale (namely, from *c*.2800 to 2710 cal. yrs BP). During this period, radiocarbon years were of very short duration and during the following 300 calendar year period (as a consequence of declining atmospheric ^{14}C), the radiocarbon age remained at the level of about 2450 BP (Van Geel *et al.*, 1996). This effect is called the **Hallstatt plateau** in the calibration curve. Misleading impressions of synchrony or diachrony of events seen in palaeoecological records can result, and duration of events, rates of sedimentation or population change can be biased (Bartlein *et al.*, 1995). Therefore it is essential that calibration of the radiocarbon timescale of palaeoecological records is undertaken to enable conclusions to be expressed correctly.

Accordingly, the radiocarbon dates in this thesis have been calibrated to sidereal years using the intercept method (i.e. where the end-points of the $\pm 2\sigma$ range intersect the calibration curve). However, it must not be assumed that the highest probability is in the centre of the range as the date has a statistically equal chance of falling anywhere in the range with 95% confidence limits. Dates are calibrated using the CALIB 3.0 program of Stuiver & Reimer (1993), using the bidecadal dataset and intercept method. Radiocarbon ages are quoted as years before present (BP), BP being defined as AD 1950, with a one sigma age range (1σ = 68% confidence limits) and as calibrated years before present (cal. yrs. BP) with a two sigma range (2σ = 95% confidence limits). The age range for one or two standard deviations indicate the precision of measurement of the sample's radioactivity and the likelihood that the level of radioactivity represents a particular age. In general there is a tendency for radiocarbon ages to be younger than calendar ages (Stuiver *et al.*, 1991; Bartlein *et al.*, 1995) which is thought to be the result of secular variations in geomagnetic intensity that govern the production rate of ^{14}C.

5.11.4 Age-Depth Models

An age depth model has been constructed for each of the sites in this study purely for the purpose of establishing approximate age estimates for the local pollen assemblage zone boundaries (Figure 7.2, 8.3 and 9.2). Radiocarbon dates are plotted on a graph against depth, and the points connected by a straight line. This is a simple and the most frequently used age-depth model (Webb & Webb, 1988). Assuming a constant rate of deposition, sedimentation rates (SR, after Bennett, 1994b) can be calculated using linear interpolation between the mid-point of the dates, and are expressed in units of cm yr BP^{-1}. The time elapsed during accumulation of a sediment unit or 'deposition time' (DT) can also been estimated, with units of BP yrs cm^{-1} using the inverse of SR (Bennett, 1994b). Although these methods of calculation have been standard practice in palaeoecological research over the past decades, there are severe limitations associated with the precision of such an approach, related to both the age determinations themselves and the type of palaeoenvironment in question.

Stuiver (1971) demonstrates that secular variations in the abundance of atmospheric ^{14}C can influence the calculation of sediment accumulation rates. Using uncalibrated records, sedimentation rates are overestimated during periods in which the radiocarbon timescale is compressed (i.e. during a plateau), and spuriously low rates may be estimated during periods when the ^{14}C timescale is stretched (Bartlein *et al.*, 1995). Webb & Webb (1988) show that accumulation rates calculated from radiocarbon dates in the 5000 to 2000 BP range are most drastically affected, which is the time period in which many of the dates obtained from this study fall. During this time, uncorrected radiocarbon ages underestimate the duration of calendric time, so that processes appear to have occurred faster than they actually did (Webb & Webb, 1988). Therefore, uncalibrated BP ages do not provide a particularly reliable chronology for the calculation rates of change (Pilcher, 1991) which are generally lower during the mid- to late-Holocene on a calibrated timescale (Webb & Webb, 1988).

The obvious solution to this problem is to calibrate the radiocarbon dates and construct the age-depth curve in the same way using these calendar ages. These dates are usually expressed as an age range which provides a problem when constructing the age-depth graph. Following the suggestion of Bennett (1994b) calibrated dates are plotted using the mid-point between the pair of ages as a crude estimate of the sample calendar age, with confidence intervals (1σ) derived by halving the distance between this age and either of the calibrated BP ages (Bennett, 1994b).

In addition to the problems associated with the non-linear nature of the radiocarbon calibration curve, the calculation of rates of change from interpolation between approximate ages are likely to be affected by errors associated with the nature of the depositional context (e.g. Donner *et al.*, 1978). It has already been mentioned that sedimentation rates (and therefore interpolated deposition times) are highly variable in fen carr environments such as those found on river floodplains (see section 4.6.2 on pollen influx/concentrations). This is even more apparent when attempting linear interpolation (or extrapolation) over areas characterised by dramatic changes in sedimentary environment. Resolving this uncertainty would require an impractical number of radiocarbon determinations.

Therefore, the resolution of the data and precision of the radiocarbon dates are such that reliable conclusions as to precise sedimentation rates are not appropriate in this context. However, the calculation of sedimentation rates does enable approximate age estimates to be determined. Therefore, the purpose of the age-depth models in this study is to estimate the duration of pollen zones not constrained by radiocarbon dates. Age estimations based on linear interpolation are tentative due to the problems acknowledged above. The results of the uncorrected BP dates are also included in the age-depth models as they provide the standard age datum from which to make palaeoecological comparisons with other work, although reference to the calibrated ages is made in all cases.

Linear extrapolation between the uppermost radiocarbon date and the surface has not been attempted due to the large changes in depositional environment between the biogenic and clastic sediments and problems associated with upper sediments such as erosion, drainage and the possibility of rapid gains from warping. The assumption that rates of deposition in such circumstances are constant is not valid.

Chapter 6 Summary of Lithostratigraphy

6.1 Introduction

This chapter presents an overview of the lithostratigraphic results from the Holocene valley fill of the lower Aire. The results of detailed stratigraphic investigations from three principal sites within the river floodplain are summarised. The following descriptions concern the sedimentary characteristics of sites at Hirst Courtney, Eskamhorn and Hook Lane, situated in the upper, middle and lower reaches of the study area respectively (Figure 3.1). Lithostratigraphic cross profiles are constructed from interpolation of data between adjacent boreholes across the floodplain transect. More detailed descriptions are given in the site reports of the three sites in Chapters 7, 8 and 9. The stratigraphic complexity is more variable towards the lower portion of the study reach, where the sediments become deeper, which necessitated longitudinal examination. Therefore a long profile is described in more detail, which links the area between Eskamhorn and Hook Lane, and encompasses the confluence with the river Aire and Ouse. The maps depicting borehole locations in this chapter are based on OS 1:10000 maps with OS national Grid References shown. A key to the symbols used in the maps is given in Figure 6.1a below. A key to the units used in the schematic sections is shown in each Figure, and a detailed description of the lithostratigraphy of these units is given in Table 6.1.

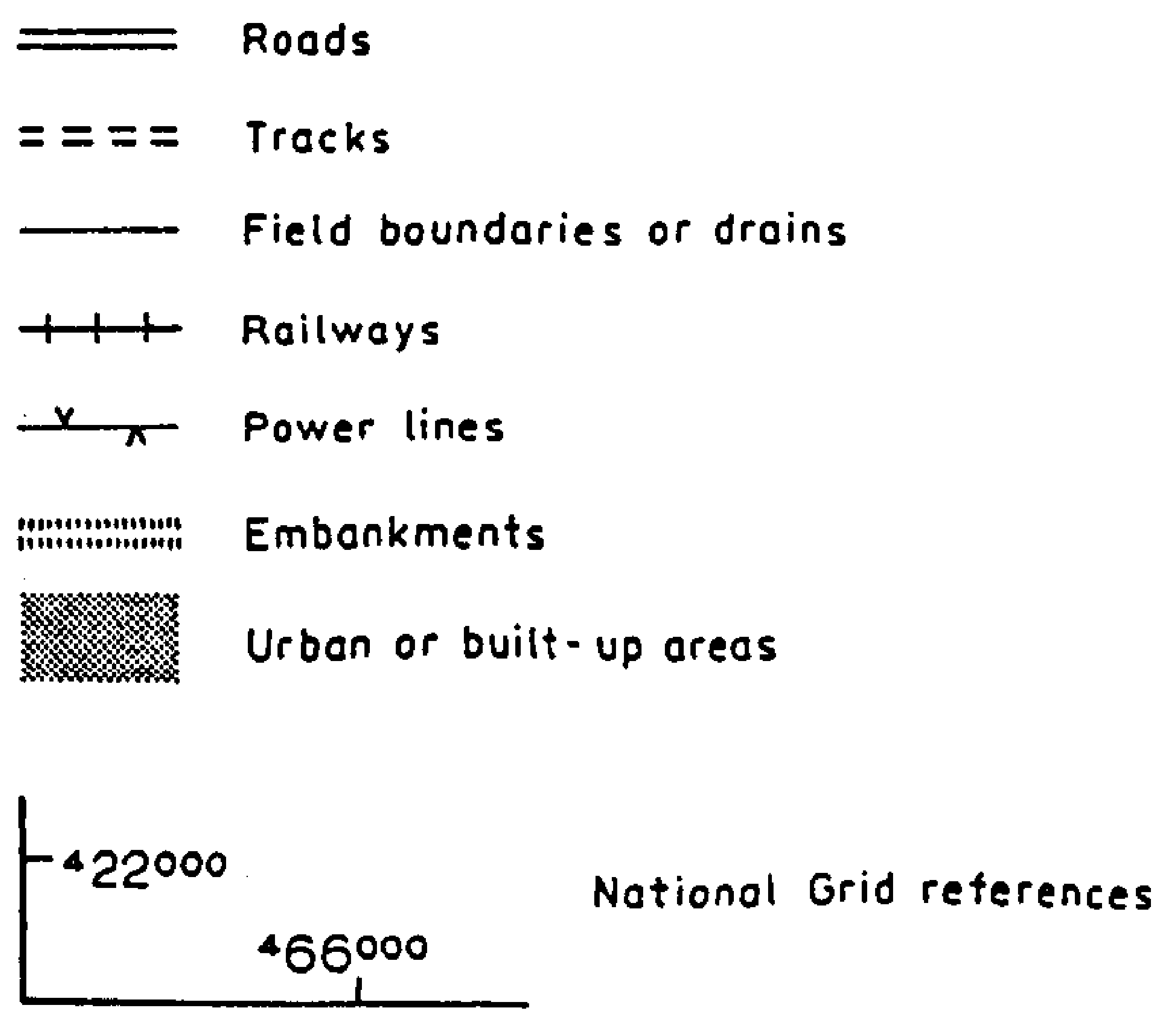


Figure 6.1a Key to symbols shown in OS maps depicting borehole locations

6.2 The Cross Profiles

6.2.1 Hirst Courtney

This transect is situated close to the present tidal limit of the river Aire adjacent to the village of Hirst Courtney (Figure 3.1). A total of 8 boreholes were sunk across the floodplain to elucidate the extent of the wetland area and establish the floodplain width (Figure 6.1). A fairly symmetrical suite of inorganic and organic sediments were uncovered representing a range of different wetland palaeoenvironments (Figure 6.2). The lowermost sediments closest to the floodplain margins consist of mainly consolidated silty, clay-sand (cores HC 3, 4 and 5) with sand with gravel in cores HC 7 and 8 at the southern side of the transect (see Unit 1 in Table 6.1).

Cores HC 1 and 2 penetrate deeper into the ground but in these cases the sediments consist of a sequence of laminated silts, clays and sands representing a more recent channel fill, presumably representing the extent of the old river Aire prior to embankment during modern times (Unit 2 in Table 6.1). Therefore the following summary refers to the boreholes excluding these.

Typically, a woody peat of variable thickness is found overlying the basal sediments, but is usually in the order of 1 to 2 m thick (Unit 4 in Table 6.1). This unit becomes shallower and thinner towards the valley sides. Overlying the peat (and the laminated clays in HC 1) is a purple-grey silty clay (Unit 5 in Table 6.1) which is laterally persistent across the floodplain and recovered in all boreholes. This is overlain by a green-grey silty clay which is generally thicker on the northern side of the floodplain (Unit 6 in Table 6.1). This clastic unit is lithologically similar to the underlying sediment, and is differentiated on the basis of the distinct colour difference.

The penultimate unit, overlying the green-grey clay of Unit 6, is a mottled orange-brown-grey clay facies (Unit 7 in Table 6.1). This deposit was devoid of any sedimentary structures such as laminations, so it is tentatively assumed not to have been deposited as a result of flood warping (although see section 6.3 below). This is overlain by top soil which caps the alluvial valley fill.

Unit	Generalised Lithostratigraphic Description
8	**Top soil/disturbed ground.**
7	**Orange-brown-grey mottled silty and sandy clay.** This formed the uppermost oxidised unit in most boreholes. In some cores, sedimentary structures such as laminations were recorded, and the sediments often became more sandy towards the river channel. In some areas, this unit is likely to be derived through the process of artificial warping although differentiation was not possible on this basis.
6	**Olive green-grey silty clay.** This unit is largely spatially restricted to the site at Hirst Courtney, although small lenses were occasionally observed along the Rawcliffe long profile.
5	**Blue-grey organic clay.** This unit is recorded overlying the peat and is variable in both colour and composition. At Hirst Courtney, it is purple-grey and contains few organic remains whereas at other locations it is blue or medium grey and contains herbaceous turfa (including *cf. Phragmites*) and woody detritus.
4	**Brown-grey clayey wood peat.** This organic unit is generally recorded mantling the valley floor and consists of coarse fragments of detrital and turfaceous (*sensu* Troels-Smith, 1955) herbaceous plant material with large wood branches particularly abundant. The peat matrix is typically highly inorganic and discrete, laterally impersistent clay horizons also occur.
3	**Grey laminated humic clay.** This horizon is restricted geographically to the lower part of the Rawcliffe long profile and the cores at Hook Lane, where it separates the main floodplain peat. It consists of soft clays with fine laminations consisting of leafy detritus (identified as *Salix* sp., M. Waller pers. comm.) and occasional woody branches.
2	**Light to medium grey strongly laminated sands, silts and clays.** This unit is characterised by distinct stratified sedimentary structures and its stratigraphic position within the valley usually close to the present river channel. The laminations consist of fine sand lamina, separated by thicker bands of silt and clay. Reworked woody detritus and subtle colour changes from bands of light to medium grey clays are occasionally observed.
1	**Stiff blue-grey-pink sandy clay or clayey sand.** This basal unit was the deepest recorded deposit defining the valley floor. It is generally dense and ranges in texture from sands to silty clay with sand. Coal dust is occasionally evident. The sediment is often observed to become coarser with depth, grading from stiff silty clays to stiff sandy clays and sand. Humified organic material and roots occur towards the top of the unit.

Table 6.1 Description of lithostratigraphic units used in schematic sections

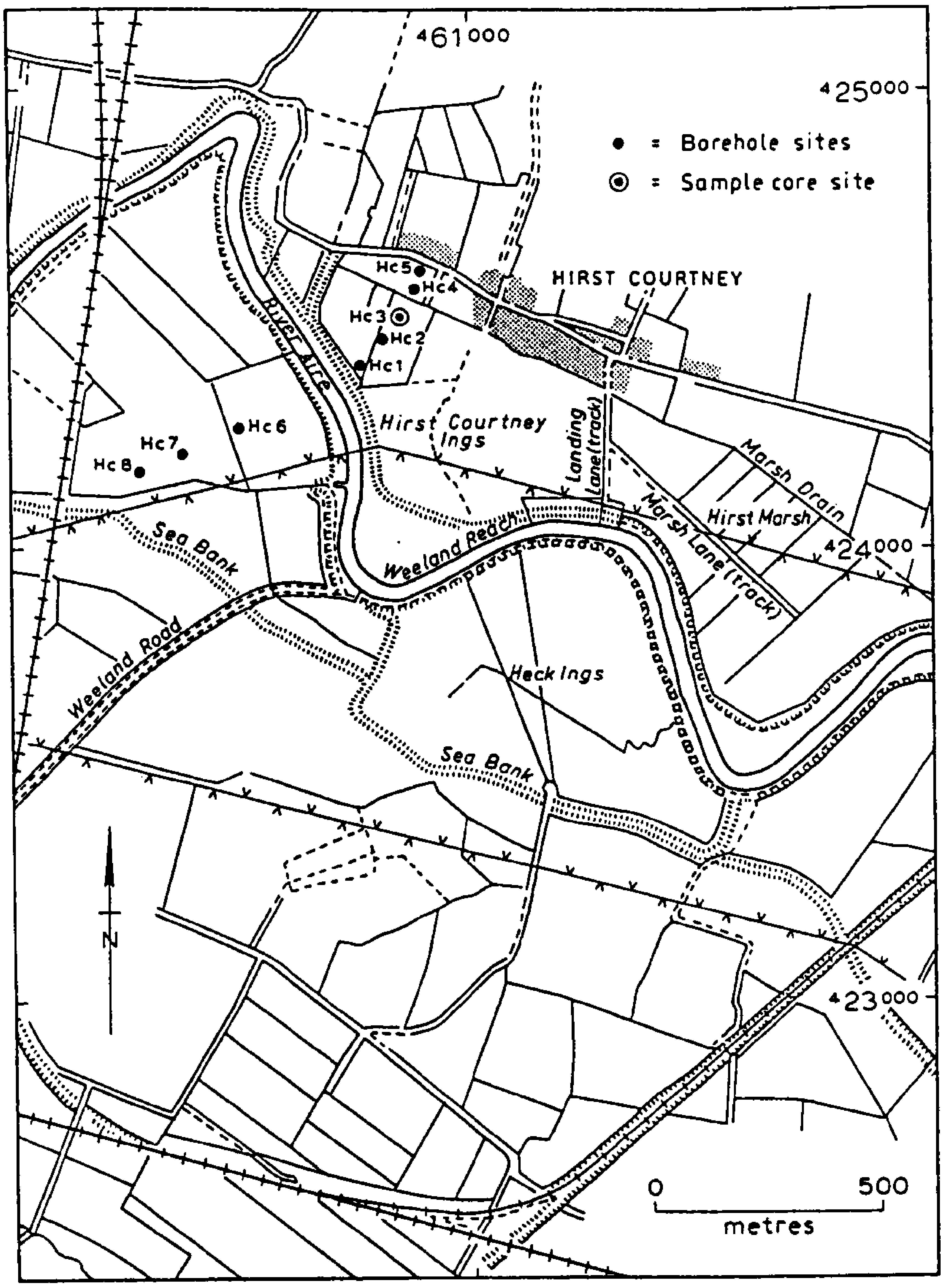


Figure 6.1b Borehole locations at Hirst Courtney in the lower Aire valley

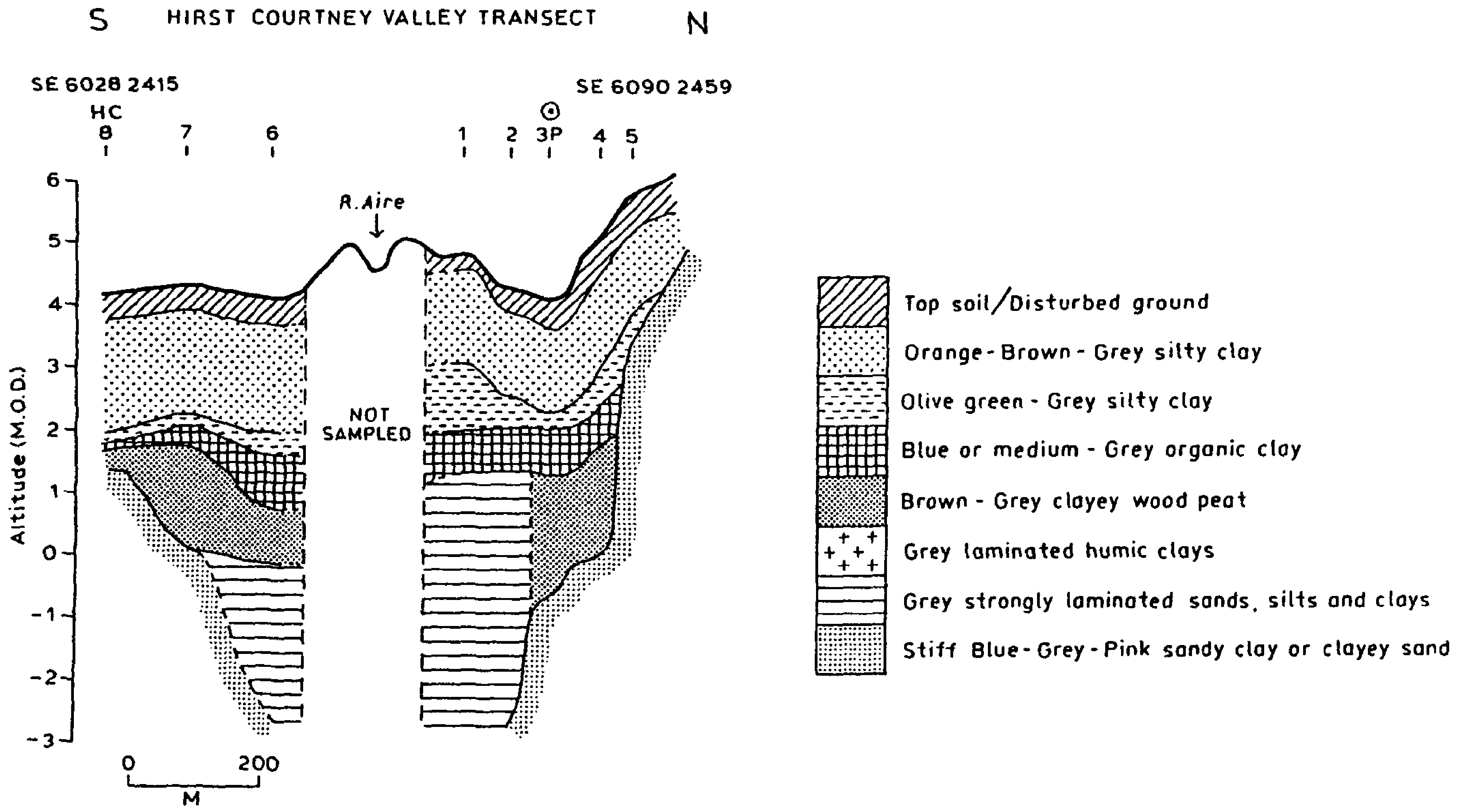


Figure 6.2 Schematic cross section of the lower Aire valley at Hirst Courtney

A more detailed examination of the sediment stratigraphy from Hirst Courtney is presented in Chapter 7 along with the litho- bio- and chronostratigraphic results obtained from the analysis of sample core HC 3P.

6.2.2 Eskamhorn

A transverse line of boreholes were sunk across the floodplain of the river Aire in the vicinity of Eskamhorn Farm, east of Rawcliffe village (Figure 3.1). 26 boreholes define the extent of the floodplain from north to south (Figure 6.3), although the depth of wetland deposits is greater in the north (Figure 6.4). The central area of the valley fill is again comprised of laminated palaeochannel deposits (Unit 2 in Table 6.1) and another palaeochannel feature is evident at the southern extremity of the floodplain near East Cowick. This is most likely to be a previous course of the river Don (the Turnbrigg Dike), which was artificially diverted into the Aire as part of an early drainage scheme (see Gaunt, 1975).

Overlying the pre-Holocene stiff sandy clay (Unit 1 in Table 6.1) is a wood peat (Unit 4 in Table 6.1) which can be traced in all boreholes except in the central floodplain area where it is replaced by minerogenic channel deposits (Unit 2 in Table 6.1). The peat sequence is thicker than upstream at Hirst Courtney and generally consists of detrital and *in situ* organic components within a rather clayey matrix. Wetland deposits rise sharply towards the northern edge of the floodplain where they abut an outcrop of Lateglacial '25-foot' drift deposits (Unit 1 in Table 6.1), consisting of sand and clay (*cf.* Gaunt, 1994).

Overlying the floodplain peat is an alluvial horizon consisting of blue-grey silty clay with vertically aligned monocotyledonous stems and detrital organic debris within it (Unit 5 in Table 6.1). This clastic unit is generally thickest peripheral to the present channel, and thins towards the floodplain edge. The undulating nature of the peat surface may reflect differential compaction or the presence of small channels or tidal creeks cutting down into the organic unit to allow tidal runoff associated with the deposition of unit 5 (*cf.* Devoy, 1979).

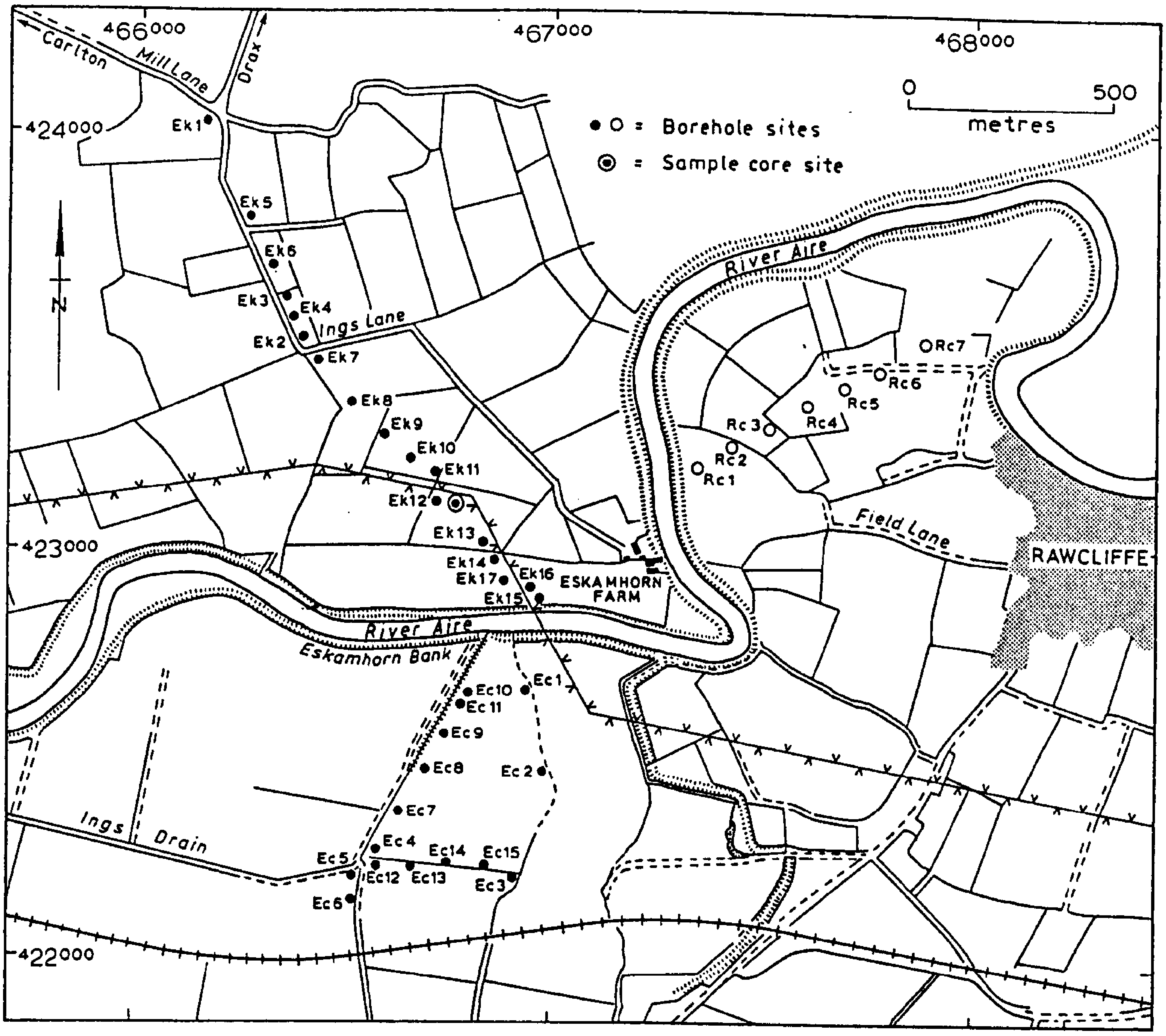


Figure 6.3 Borehole locations at Eskamhorn in the lower Aire valley

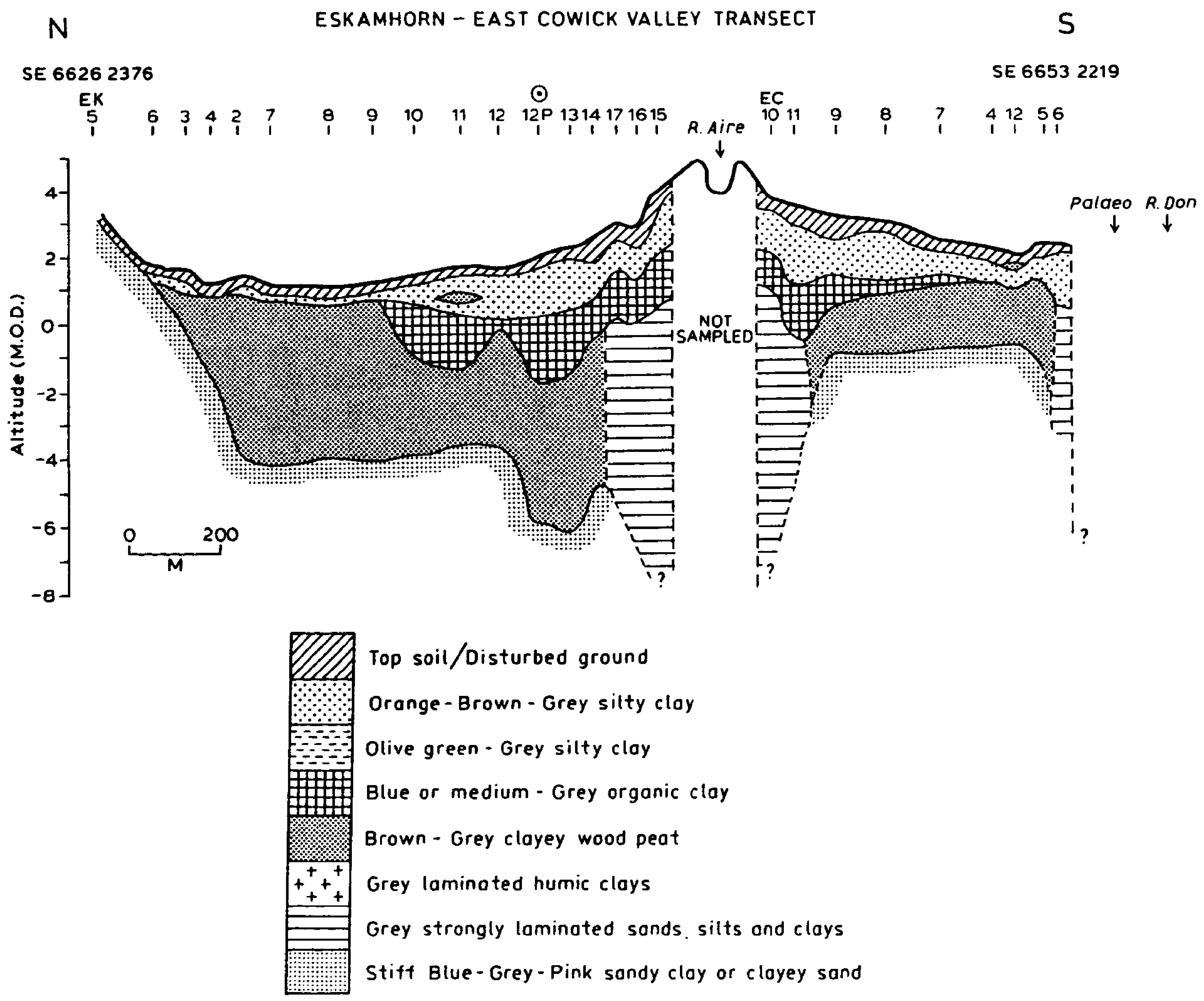


Figure 6.4 Schematic cross section of the lower Aire valley at Eskamhorn

The upper facies is a mottled orange-brown-grey silt-clay (Unit 7 in Table 6.1) and closely resembles the stratigraphic equivalent unit at Hirst Courtney, where it is generally thicker. This unit is recorded in all boreholes that penetrate the floodplain except EK 4 where the peat is capped by a thin soil layer. A small humified pocket of peat is evident within Unit 7 in core EK 11, and between Unit 7 and top soil in core EC 12, but this well oxidised organic layer is poorly preserved and laterally truncated (see section 6.3). A sample core was extracted from EK 12P which contained all the representative stratigraphic units and the deepest sequence of peat. The detailed lithostratigraphy is examined further in Chapter 8, where the results of the microfossil analyses conducted on EK 12P are also presented.

6.3.3 Hook Lane

This site is situated across the river Ouse floodplain (Figure 3.1) and is the most seaward site studied. The stratigraphy is represented by a transect of 11 boreholes either side of Hook Lane (Figure 6.5). Five main sedimentary units have been identified which together exceed 10 m in depth in places. These sediments are displayed schematically in a 2-D plot of the floodplain fill in Figure 6.6.

Mantling the consolidated pre-Holocene clay-sand deposits (Unit 1 in Table 6.1) of presumed Lake Humber origin (Gaunt, 1994) on the southern side of the floodplain is a dark brown well humified wood peat (Unit 4 in Table 6.1). The lowermost deposits uncovered on the north side are extensively laminated channel sediments, which were not bottomed by the corer. In cores HK 1 to 6, the peat is overlain by an unusual clastic facies not present at either of the sites above. This unit (Unit 3 in Table 6.1) consists of finely laminated humic clays and large pieces of detrital wood. The laminations are unevenly spaced throughout the unit and usually consisted of partings of well preserved leaf and other organic detritus. These stratified clays are replaced by a thick floodplain wood peat (Unit 4 in Table 6.1) with locally variable and spatially inconsistent clayey horizons. This peat is recovered at a much higher altitude in cores BB 1 and 2 in the north and HK 6 in the south, where it directly overlies the pre-Holocene sediments at the margin of the valley.

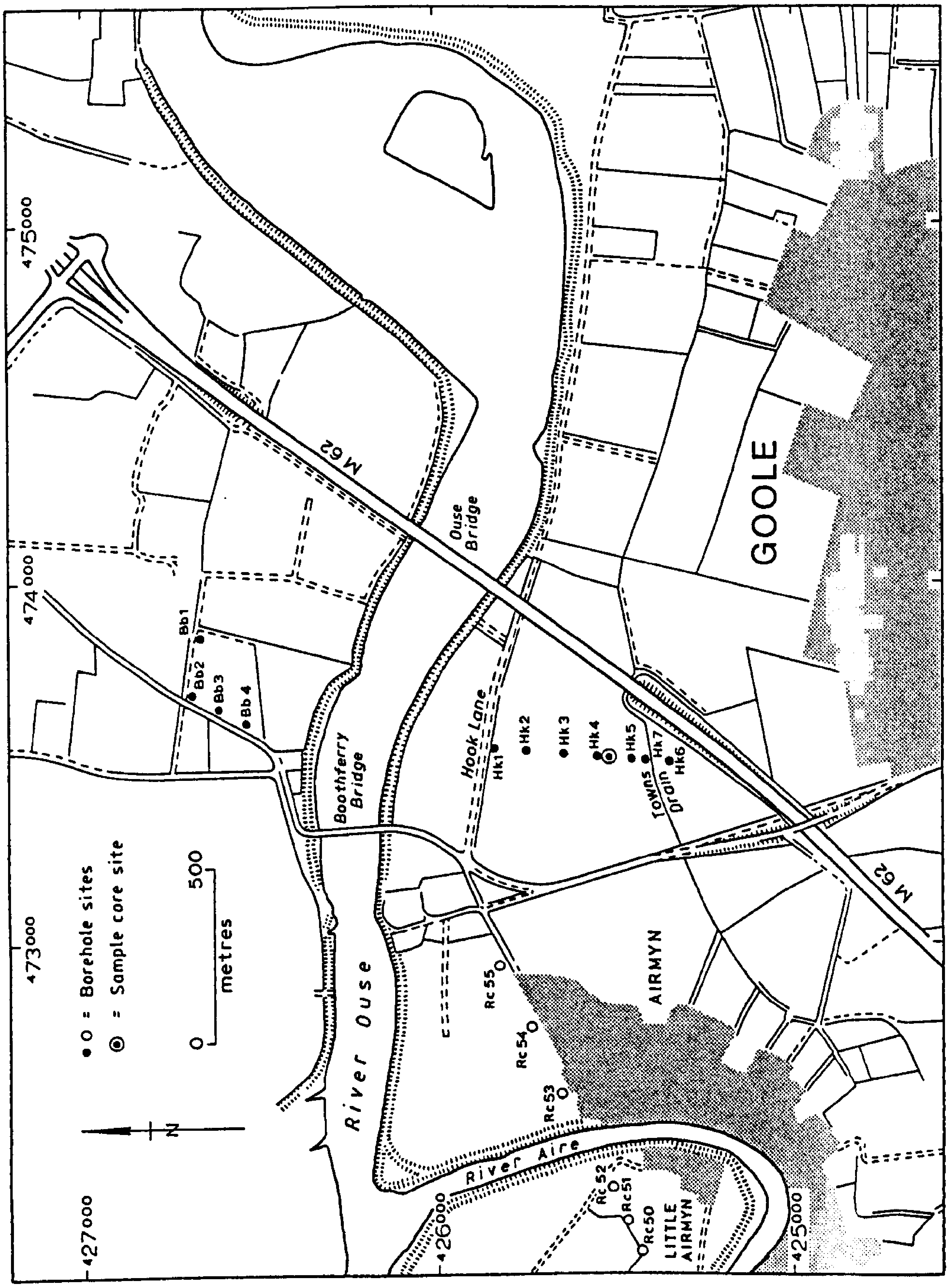


Figure 6.5 Borehole locations at Hook Lane in the lower Aire valley

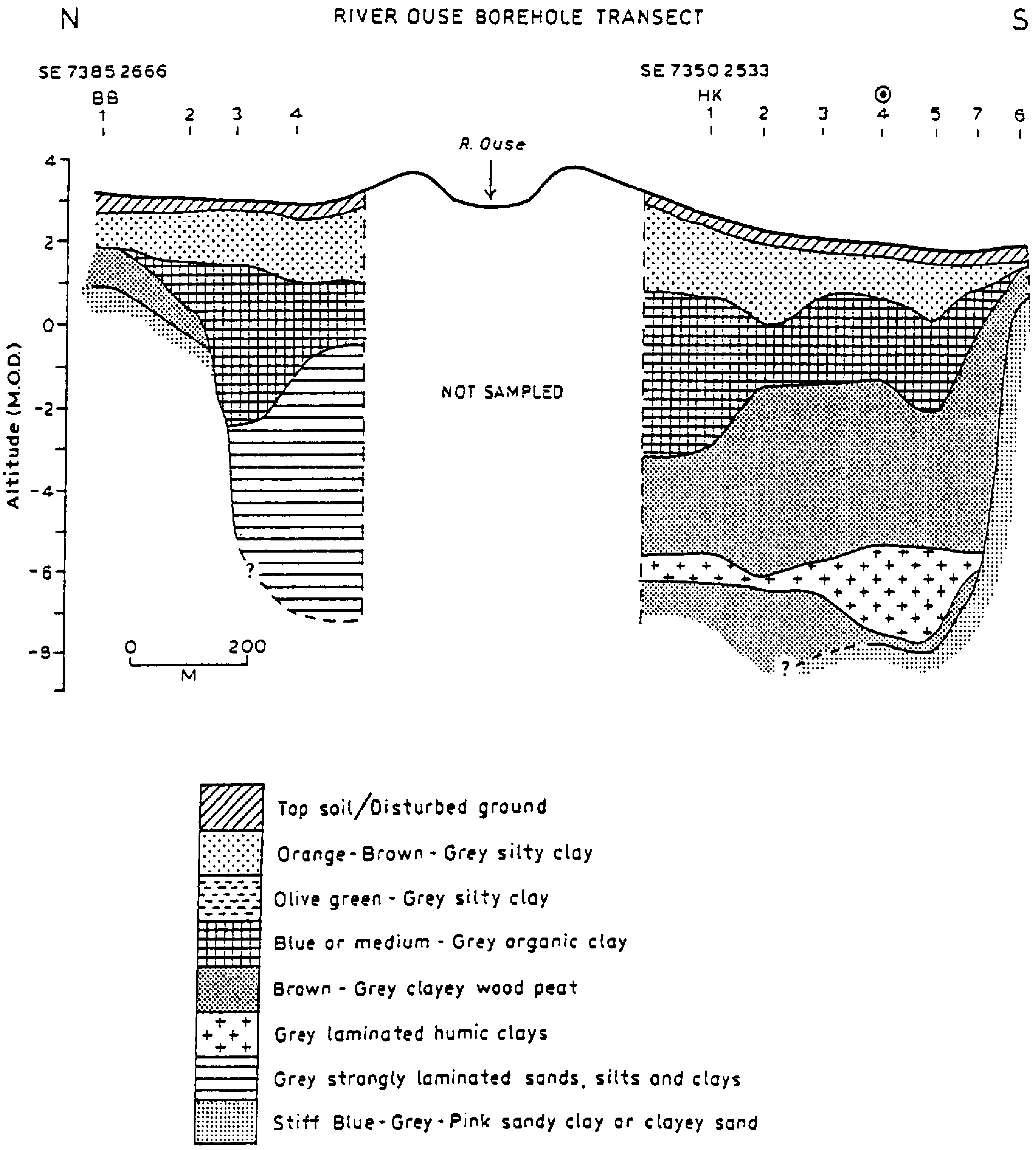


Figure 6.6 Schematic cross section of the lower Aire valley at Hook Lane

The transition to the overlying blue-grey organic silty clay (Unit 5 in Table 6.1) is very gradational and difficult to define in the field. The wood peat gradually becomes more clayey and is eventually replaced by an organic clay with woody and monocotyledonous detritus (probably *Phragmites* stems). This is the stratigraphic equivalent to the blue-grey silty clay unit (Unit 5) that overlies the peat at Eskamhorn. The uppermost unit is the oxidised clay (Unit 7 in Table 6.1) that seals the alluvial floodplain deposits at each of the study sites. However, faint traces of laminations are detectable within the upper metre or so of this unit in some cores, so it may, in part, be artificially derived, through the process known as warping (Heathcote, 1951; Ellis, 1990; Gaunt, 1994, see section 3.7). A representative sample core was taken close to HK 4 (HK 4P) and was subjected to extensive micropalaeontological analyses and radiocarbon dating (see Chapter 10).

6.3 The Long Profile

In order to elucidate the longitudinal variation in floodplain deposits in the lower part of the study reach, the valley long profile was cored, connecting the lowermost study sites at Eskamhorn and Hook Lane (Figure 3.1), to the north-east of Rawcliffe village. A total of 55 boreholes are used to reconstruct the sedimentary environment, largely from the south side of the river Aire floodplain, except where the transect crosses the contemporary river at several large meander bends (see Figures 6.7, 6.9, and 6.11). The detailed core lithologies from the long profile transect are divided into three sections and shown in Figures 6.8, 6.10, and 6.12. Lithostratigraphic symbols follow Troels-Smith (1955). A schematic borehole transect which combines and summarises all the information in Figures 6.8, 6.10 and 6.12 is presented in Figure 6.13. The stratigraphy is described from east to west below.

Boreholes RC 1 to RC 19 (Figures 6.7, 6.8 and 6.13)

Borehole RC 1 represents a marginal channel fill sequence (Unit 2 in Table 6.1) associated with its position close to the Aire channel. Boreholes RC 2 and 3 basically replicate the typical stratigraphy recorded at many sites in the lower Aire valley whereby the pre-Holocene subsurface (Unit 1 in Table 6.1) is overlain by a clayey wood peat (Unit 4 in Table 6.1) which in turn gives way to the deposition of blue-grey organic clay (Unit 5 in Table 6.1) and lastly the inorganic oxidised sediments (Unit 7 in Table 6.1).

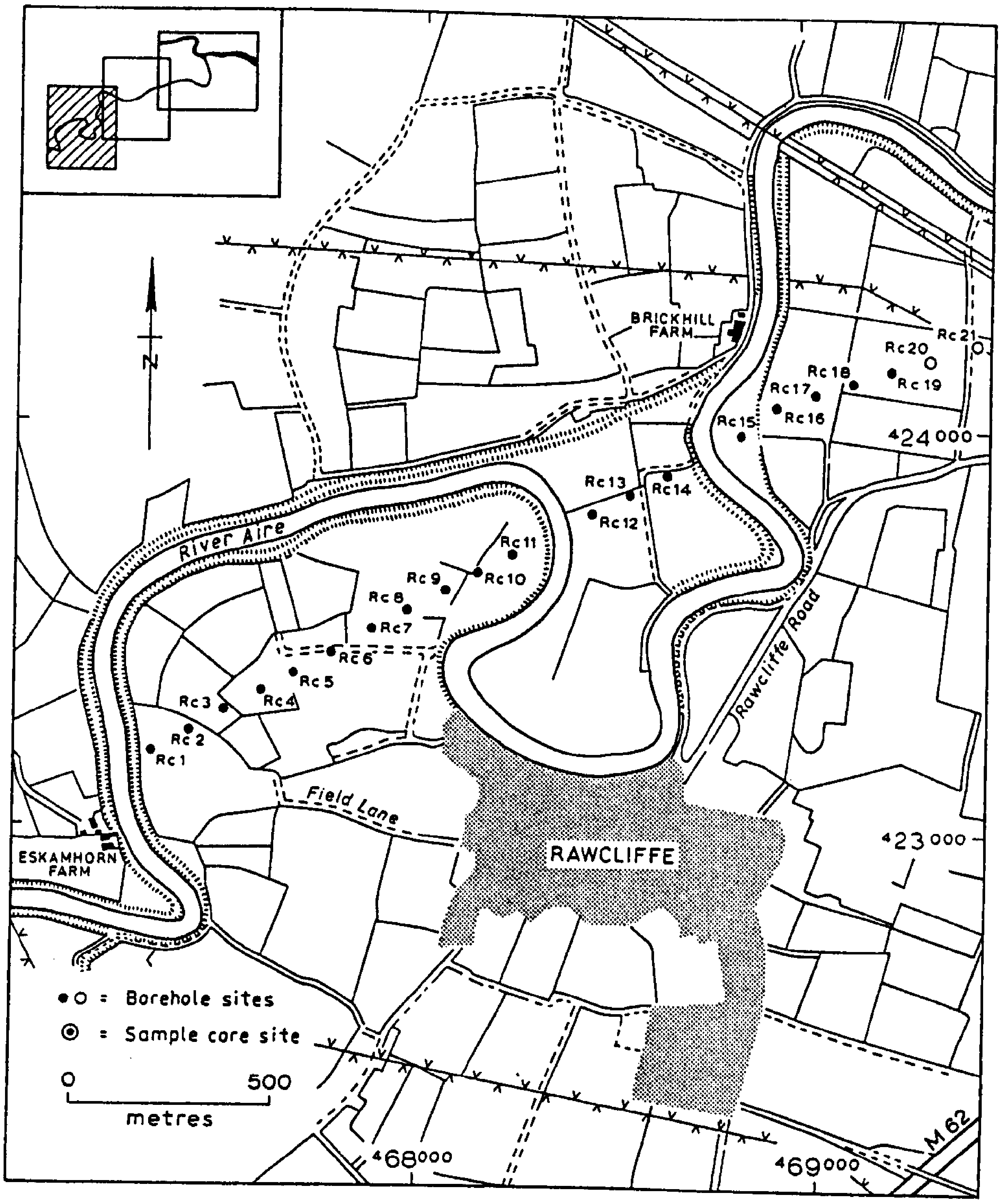


Figure 6.7 Borehole locations at Rawcliffe long profile (RC 1 to RC 19) in the lower Aire valley

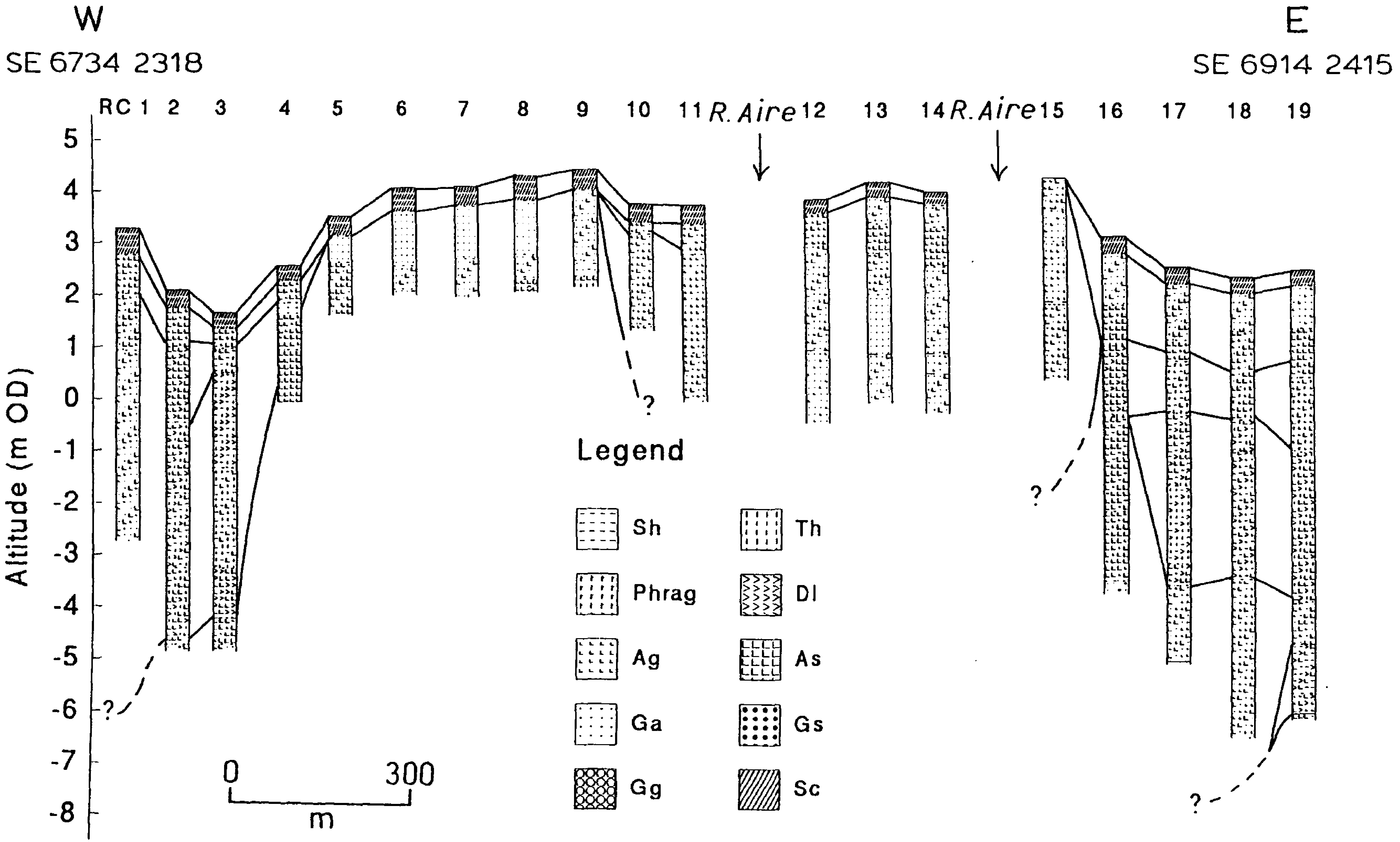


Figure 6.8 Lithostratigraphy of Rawcliffe long profile (RC 1 to RC 19) in the lower Aire valley

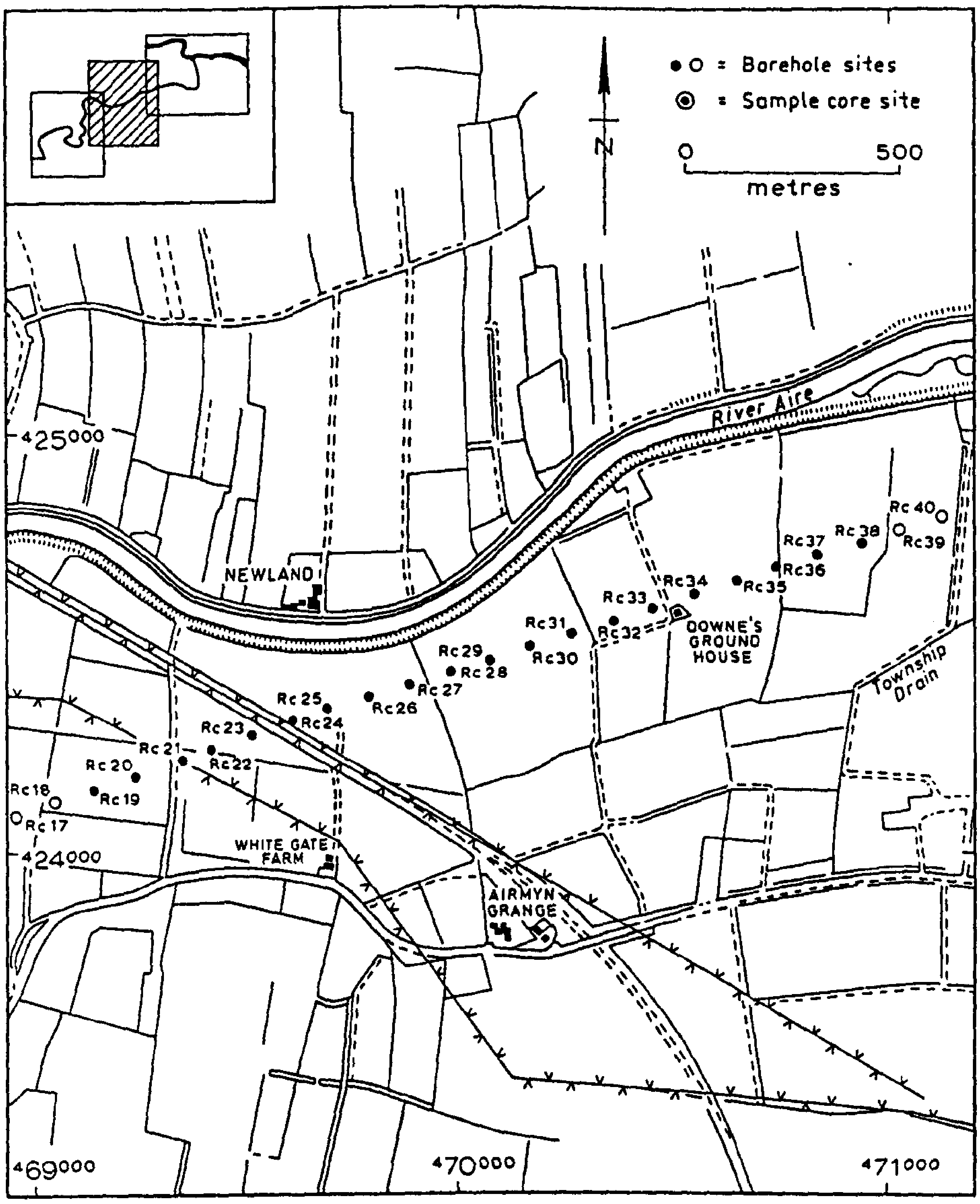


Figure 6.9 Borehole locations at Rawcliffe long profile (RC 19 to RC 38) in the lower Aire valley

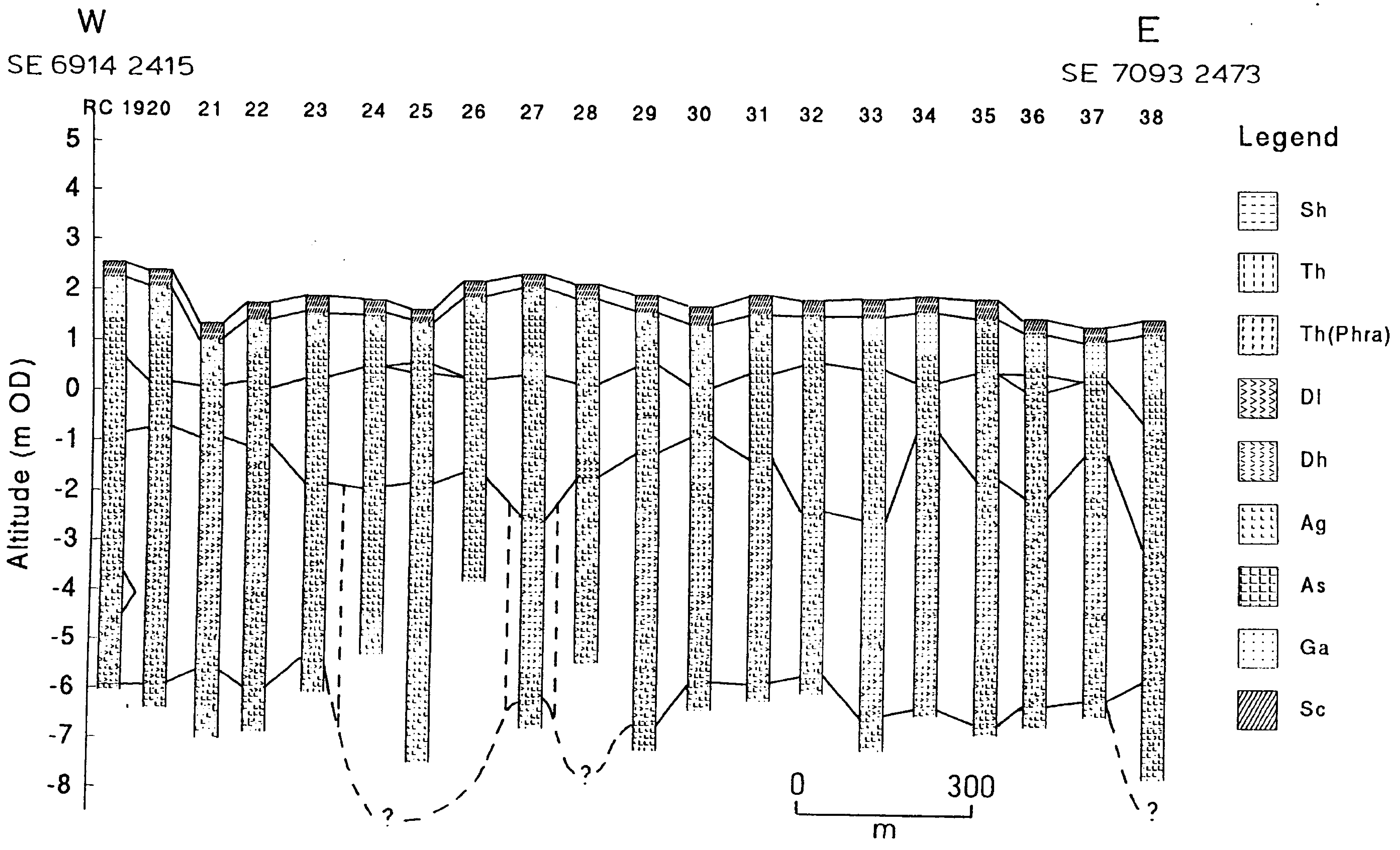


Figure 6.10 Lithostratigraphy of Rawcliffe long profile (RC 19 to RC 38) in the lower Aire valley

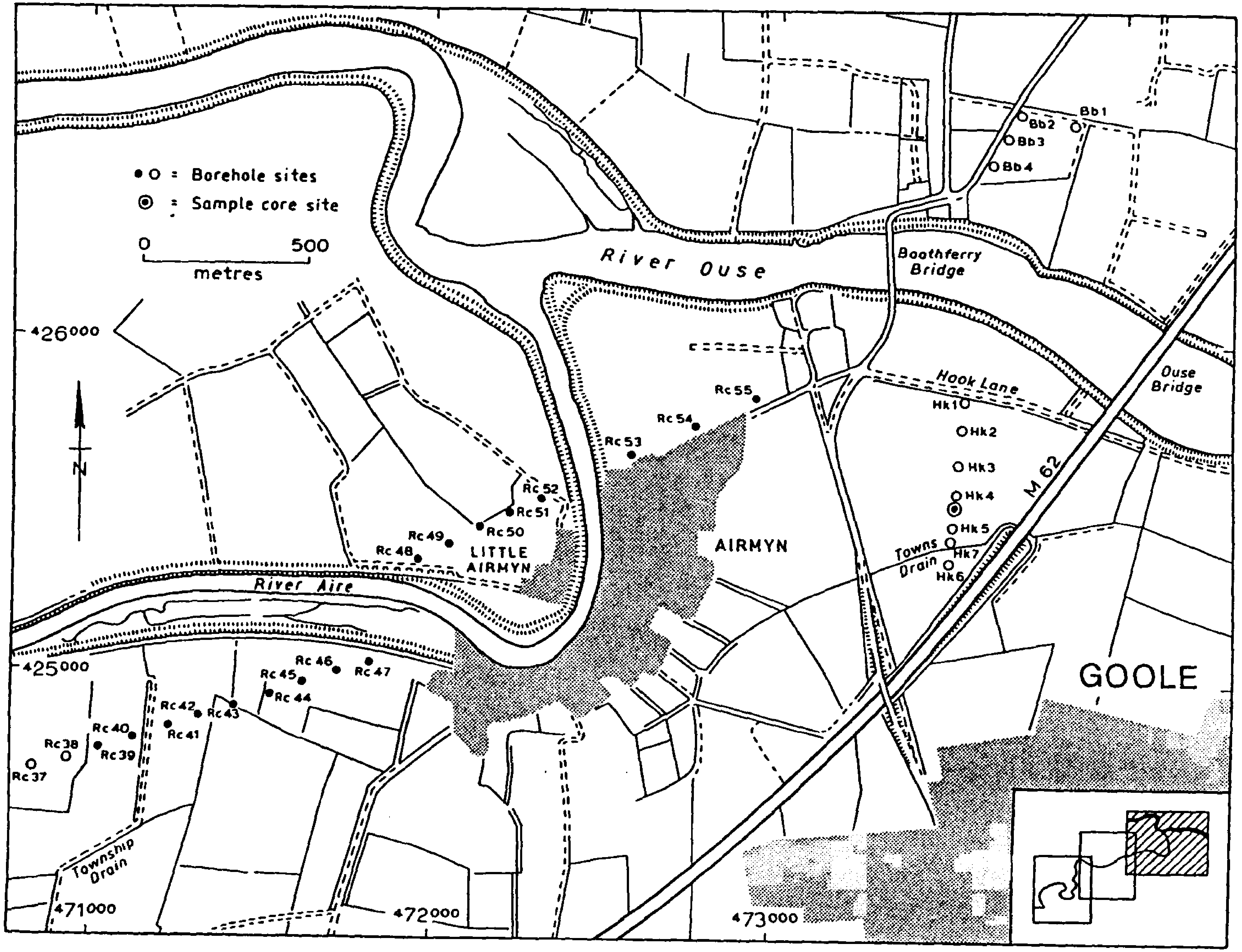


Figure 6.11 Borehole locations at Rawcliffe long profile (RC 38 to RC 55) in the lower Aire valley

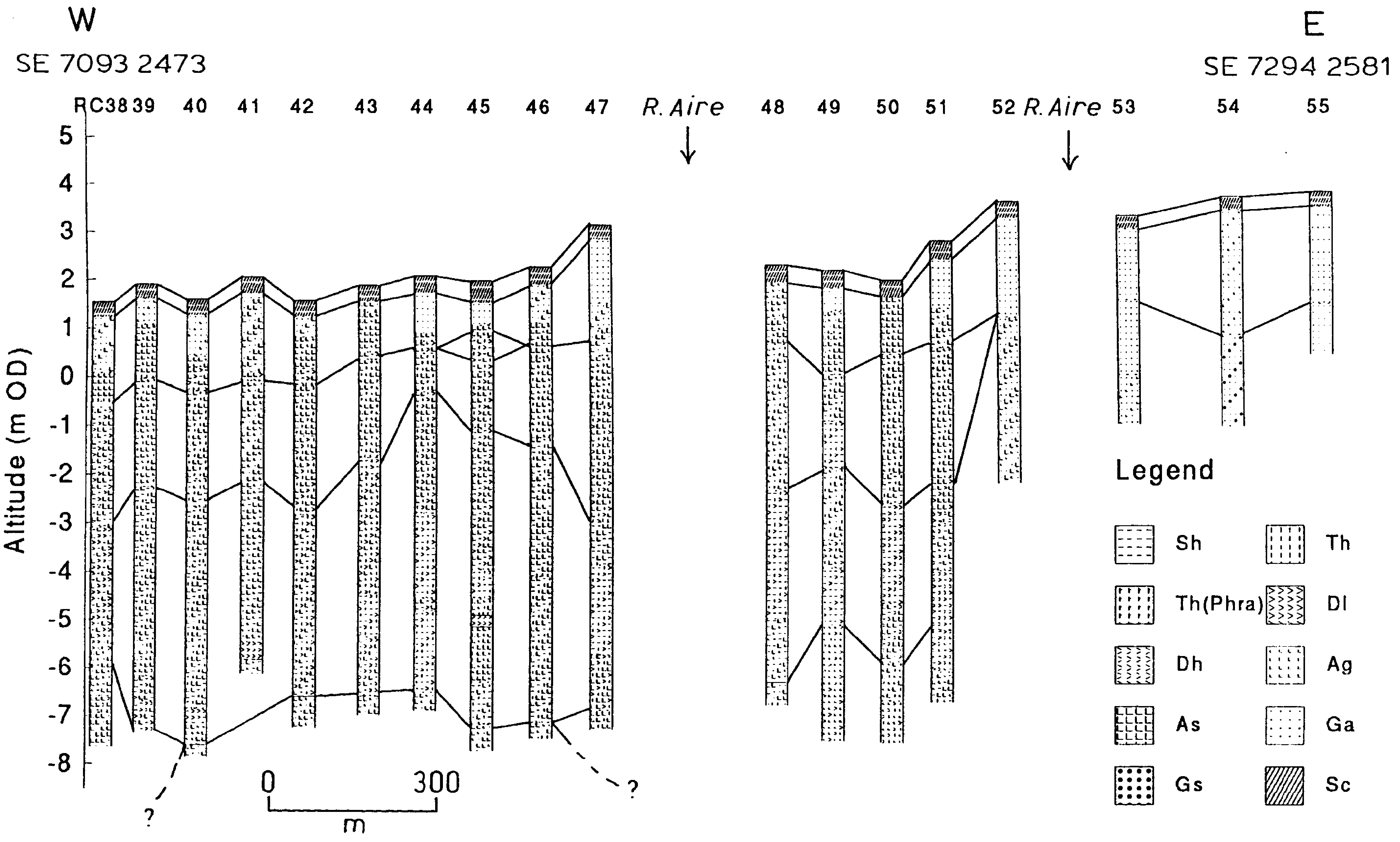


Figure 6.12 Lithostratigraphy of Rawcliffe long profile (RC 38 to RC 55) in the lower Aire valley

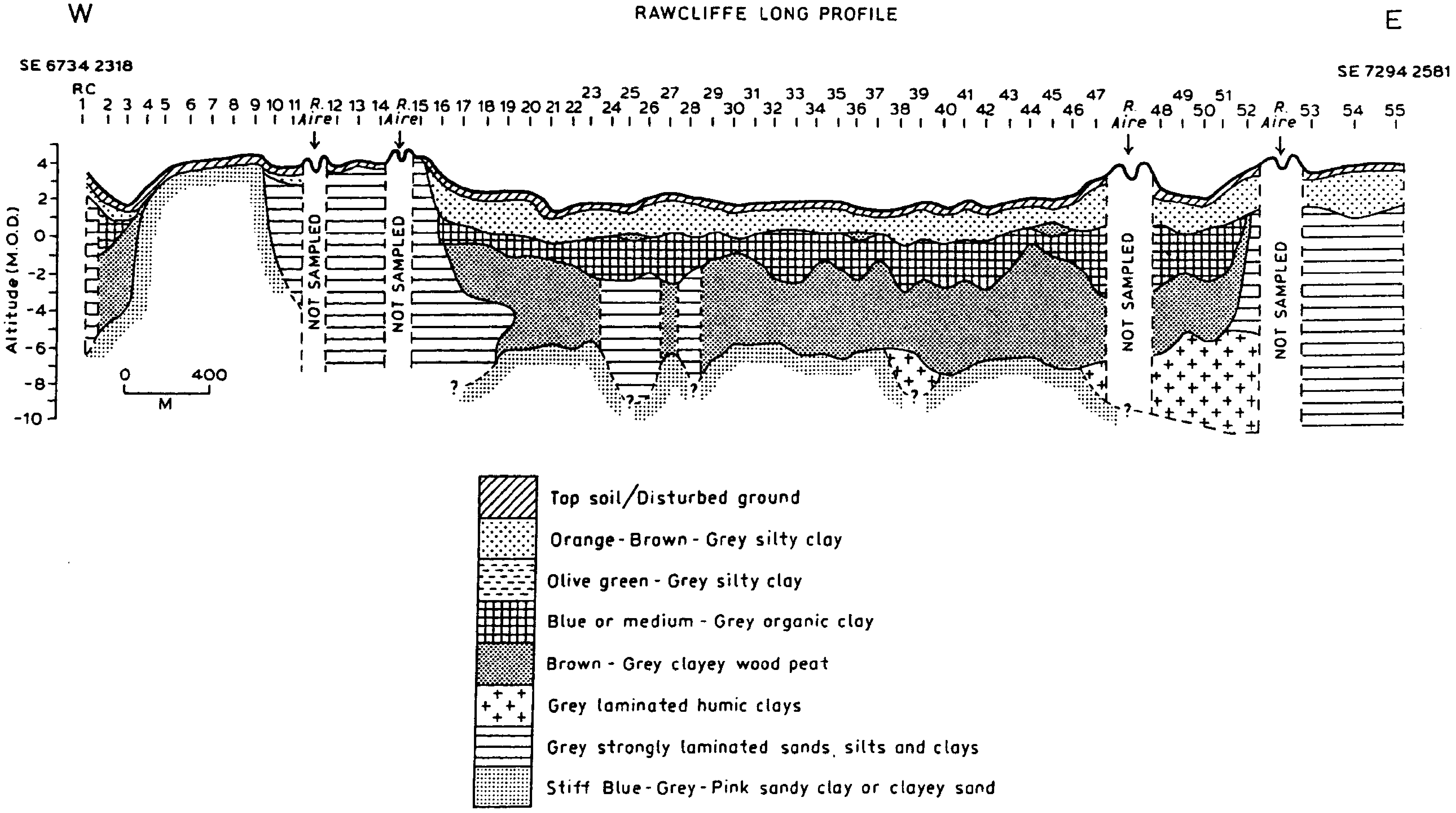


Figure 6.13 Schematic section of Rawcliffe long profile in the lower Aire valley

Between boreholes RC 4 and 9, this typical suite of floodplain wetland deposits is not recorded but instead, the sediments consist of coarse and fine grained sands, clays and silts with abundant black particles of presumably coal dust (*cf.* Unit 1 in Table 6.1). This ridge feature has all the characteristics of a relic levee feature, associated with aeolian and fluvial activity, which reworked the Late-Devensian drift sediments subsequent to the drainage of Lake Humber (see Gaunt, 1994, section 3.4, and Figure 3.5). Boreholes RC 10 to 15 reveal a deep palaeochannel sequence of laminated clay, silt and sand (Unit 2 in Table 6.1) associated with the migration of the large meander feature evident in the modern river Aire at this point (see Figure 6.7). The junction between these channel sediments and the peat unit recorded in RC 16 forms a ragged edge, and the contact was often of an unconformable nature. Between cores RC 16 and 19, some erosion of the main floodplain peat unit (Unit 4 in Table 6.1) is evident, but generally, the characteristic sequence of peat (overlying a basal surface in RC 19) is recorded, which is replaced by the minerogenic alluvial units 5 and 7.

Boreholes RC 19 to RC 38 (Figure 6.9, 6.10 and 6.13)

Between boreholes RC 19 and 38, detrital peat (Unit 4 in Table 6.1) is found overlying the basal deposits (Unit 1 in Table 6.1) in the majority of cores. However, in the area between cores RC 23 and 28, peat is absent, and replaced by channel sediments which have incised deeply into the valley. Large wood macrofossils are evident (identified to *Alnus glutinosa*, *Salix* sp., and *Fraxinus excelsior*) both within the peat and occasionally from the channel fill context. The boreholes where peat is recorded bottom-out in pre-Holocene drift deposits (*cf.* Gaunt, 1994). No indication as to the depth of bedrock has been obtained by hand coring due the deep stratigraphic position of the drift deposits and their consolidated nature. However, Gaunt (1994) has recorded a borehole in the vicinity of Newland (SE69932474) in the northern Aire floodplain (see Figure 6.9), where Sherwood Sandstone is recorded at -14.6 m OD, and is overlain by sand then clay deposits containing wood fragments.

Overlying the peat between boreholes RC 19 and RC 23, and RC 29 and RC 38, is a blue-grey organic clay (Unit 5 in Table 6.1). Along the Rawcliffe long profile, as at Eskamhorn, the upper surface of the peat is characteristically uneven. This may, in part, be a consequence of the difficulty in exactly defining the lithological boundary associated with this transitional sedimentary change in the field, but in other cases is probably attributable to networks of creeks that have eroded into the floodplain during the post-peat depositional period (*cf.* Devoy, 1979). The upper blue-grey organic clay contains grass and sedge reed remains, as well as a residue of woody detritus and *Substantia humosa* (*sensu* Troels-Smith, 1955).

Towards the top of Unit 5 in boreholes RC 25 and 36 (Figure 6.10), and RC 45 (Figure 6.12), a thin humified peat layer is recorded. The presence of isolated peat 'islands' at the upper surface of the blue-grey clay unit at three locations along the Rawcliffe long profile is intriguing. These peats were generally very dessicated and often had abrupt upper contacts (e.g. cores RC 36 and 45). Interestingly, they occur at a comparable stratigraphic altitude to the single peat lens identified in core EK 11 at Eskamhorn, suggesting they perhaps represent a formerly more extensive bed of peat (see Figure 6.13). It is feasible (and likely) that these patches of near surface crumbly peat represent an upper organic layer similar to the reported, and for the most part non-existent, dessicated upper peat within the Ancholme valley (*cf.* Smith, 1958a; Neumann, 1998). Unfortunately, the extent of the drainage in the Aire valley and the poorly preserved nature of this upper deposit at these coring sites renders it unsuitable for further sedimentological and palaeoecological investigation at these sites.

The uppermost sediments recorded in this part of the long profile transect are the oxidised silty and sandy clays (Unit 7 in Table 6.1) which are consistently uncovered from throughout the study reach. The possible origin of this unit is discussed below.

Boreholes RC 38 to RC 55 (Figures 6.11, 6.12, and 6.13)

Between boreholes RC 52 and 55 at the eastern extent of the long profile, high energy laminated channel silts and sands occur (Unit 2 in Table 6.1), which are associated with the junction of the river Aire and Ouse. Floodplain peat (Unit 4 in Table 6.1) overlies the pre-Holocene surface (Unit 1 in Table 6.1) between boreholes RC 40 and 46. However, between boreholes RC 38 and 39, and RC 46 and 51, the lowest sediments recorded are laminated humic clays (Unit 3 in Table 6.1). This clay is lithologically and stratigraphically comparable to a similar unit recorded at Hook Lane, with laminations composed of fine leaf detritus interspersed with coarse woody remains, although the depth of the deposits prevented recovery of the lower peat. Spot diatom samples from this unit were taken from boreholes RC 38, 48, 49, 50, and 51, and revealed similar assemblages to those obtained from core HK 4P at Hook Lane (see Chapter 9). Geographically, this facies is restricted spatially in the lower Aire valley to the area around the present day confluence of the river Aire and Ouse in the vicinity of Airmyn and Little Airmyn (see Figure 6.11). A similar unit is also recorded around -6 m OD from boreholes in the northern floodplain deposits of the Ouse (Lillie & Gearey, 1999).

A floodplain peat unit (Unit 4 in Table 6.1) is found overlying the humic clays (Unit 3 in Table 6.1). This organic unit attains thicknesses of up to *c*.5 m between boreholes RC 40 and 46. The clay content of the wood peat is variable along the long profile transect, in part, relating to the relative distance from the contemporary river channel or creek systems during accumulation which would have changed over time. Spot diatom samples were taken from the clayey horizons within the peat at RC 43 and 45 but preservation of valves was generally poor. The low diatom concentration meant that counts were very low (<10). However, brackish species such as *Cyclotella striata* and *Nitszchia navicularis* were identified, which probably represents the allochthonous component of the assemblage. Autochthonous valves, such as epiphytic or benthic taxa, may include freshwater species such as *Pinnularia* sp., but their abundance was very low and preservation generally poor. In environments unfavourable for diatom preservation, such as high salt marshes and marginal peat areas, robust species such as *Cyclotella striata* are preferentially preserved, so that only a small proportion of the species that actually lived in the habitat remain (Sherrod *et al.*, 1989). The resulting assemblages may have little to do with the actual palaeoenvironment (Denys, 1989; 1994; Vos & de Wolf, 1994; Hemphill-Haley, 1995). Therefore, the diatom assemblages obtained from the peat deposits along the Rawcliffe long profile record evidence for flooding of brackish water, but little information concerning *in situ* environment of deposition (Vos & de Wolf, 1997).

The floodplain peat is overlain by a blue-grey organic silty clay (Unit 5 in Table 6.1) in all boreholes between RC 38 and 55. Unit 5 is replaced by an orange-brown-grey unit (Unit 7 in Table 6.1). This oxidised clay is the uppermost unit of the valley fill and was generally differentiated from the underlying unit on the basis of a lack of organic remains and colour. This was mainly for ease of field recording of sediment stratigraphy and it is recognised that these variables are rather unsuitable for strict lithostratigraphic correlation because colour and organic content is strongly related to the post depositional level of the water table (*cf.* Dury, 1964). The fluctuating level of the water table determines the redox status of the soil and results in reduced grey colours or oxidised orange-browns. However, because this is the basis upon which Samuels & Buckland (1978), Buckland & Sadler (1985), and Dinnin (1997b) determined the phase of supposed Roman and post-Roman alluviation in the region, this practise was adopted in this study.

The extent to which these upper sediments are related to natural or artificial flooding is an equivocal matter. Gaunt (1994) maps large parts of the fields covered by the Rawcliffe long profile and Hook Lane as having been warped extensively during the last few hundred years (see Figure 3.7). However, very few boreholes into the upper alluvial sediments revealed the stated characteristics of artificial flood warp (i.e. finely laminated with silt clay, some fine sand and a sharp contact with the underlying deposits, (*cf.* Heathcote, 1951) and, where these diagnostic features were noted, they were spatially very restricted. Furthermore, a defined warping drain has been mapped, along with its associated warping field, by Gaunt (1994) in the area around cores RC 21 and 25 (see Figure 3.7). However, surveying showed these cores to be lower than the surrounding landscape (see Figures 6.10 and 6.13) which contradicts the geomorphological common sense that such areas should be noticeably raised (*cf.* Gaunt, 1994; Allen, 1993; 1996). Lillie (1997b; 1999) reports similar problems when trying to separate artificial warp deposits from sediments resulting from natural flood alluviation, even when employing laboratory analysis in addition to field description.

Another interesting feature of the upper oxidised sediments along the long profile is that they clearly show a coarsening upwards in grain size in many boreholes (see Figures 6.8, 6.10, and 6.12). This trend would not be expected if Unit 7 was derived from warping because the aim of the technique was a deposit fine grained sediments in a carefully controlled manner (see Heathcote 1951; Ellis, 1990; Lillie, 1997b). Tidal velocities were slowed so that the flood tide did not have the capacity to carry large quantities of sand as it entered the warping area. Also warping would not result in an upwards change in grain size (apart from on the small scale of successive flood laminations). Therefore, it is apparent that the upper, more sandy part of Unit 7 may be derived from natural overbank flood processes. Whilst it is likely that in some cases, perhaps the upper metre or so of this oxidised sediment has been deposited as a result of deliberate warping practices, it is difficult to identify sediments derived from such a practice using the techniques employed in this research.

6.4 Summary of Lithostratigraphic Units

The basic stratigraphy of the valley sediment has been broadly categorised into units (see Table 6.1). Overlying the basal minerogenic subsurface (Unit 1) is a wood peat (Unit 4) which is recorded to some extent at all sites within the study reach (see Chapters 7, 8 and 9 for details). The contact between the basal sediments and the floodplain peat increases in altitude up the valley. Also the sediments that comprise the floodplain fill generally narrow and shallow with distance up-river. Where the peat is not recorded, it is usually because of the presence of a palaeochannel feature (Unit 2). Sites at the seaward extreme of the study reach, such as the lower part of the Rawcliffe long profile and Hook Lane, stand out because of an additional lower laminated humic clay facies (Unit 3), that separates the floodplain peat into two, which is not recorded elsewhere.

Overlying the peat is an alluvial unit at all sites, but the nature of this clay is different at the most landward site at Hirst Courtney. Throughout the lower valley tract, from Eskamhorn downstream, the peat is slowly replaced by a blue-grey silty organic clay with monocotyledonous and wood remains (Unit 5), which grades upwards into a less organic oxidised clay (Unit 7). However, at Hirst Courtney, the peat is overlain by a purple-grey clay (*cf.* Unit 5) which is replaced by a green-grey clastic unit (Unit 6), and finally, the oxidised unit (Unit 7), perhaps reflecting a different sedimentary regime during that time within the upper reaches of the study area.

The presence of an upper peat within the floodplain is hinted at by the variable recovery of a generally well oxidised and dessicated organic unit between the blue-grey clay (Unit 5) and the upper oxidised clastic unit (Unit 7). The depositional history of this discontinuous unit is equivocal and the sediment is poorly preserved which precludes further micropalaeontological investigation.

For further details regarding the precise sedimentological descriptions of the sedimentary fill can be obtained from the author.

Chapter 7 Hirst Courtney Site Report

7.1 Introduction

This chapter presents the results of the palaeoenvironmental investigations from the Holocene floodplain deposits at Hirst Courtney. This site is the most inland area investigated (see Figure 3.1) being situated close the present tidal limit. The site was chosen to represent the fluvially dominated floodplain association. After preliminary lithostratigraphic survey (section 7.2 below), a core was collected from HC 3P on the north side of the river Aire floodplain for micropalaeontological analysis and radiocarbon dating. These results provide the basis for the detailed bio- and chronostratigraphic descriptions and palaeoenvironmental interpretations in sections 7.4, 7.5, and 7.6 below.

7.2 Lithostratigraphy

The sequence of sediments recorded from the river Aire valley are described in detail in this section. A total of 8 boreholes demonstrate the nature of the Holocene valley fill sediments at Hirst Courtney (see Figure 6.1). Detailed core records are drawn using the Troels-Smith (1955) symbols (Figure 7.1). A simplified schematic section has been described in section 6.2.1 and Figure 6.2.

In cores nearest the valley sides, a basal unit of stiff blue-grey silty and sandy clays is present, recorded at altitudes between -0.63 m OD in HC 3P, to +1.36 m OD in HC 8 (see Figure 7.1). This sediment coarsens up the valley sides; in cores HC 7 and 8, it contains a significant gravel component. HC 5 terminates in sand at +4.12 m OD; this is apparently part of the linear sandy levee feature of supposed Lateglacial origin upon which the village of Hirst Courtney is situated (see Gaunt, 1994 and Figure 3.5).

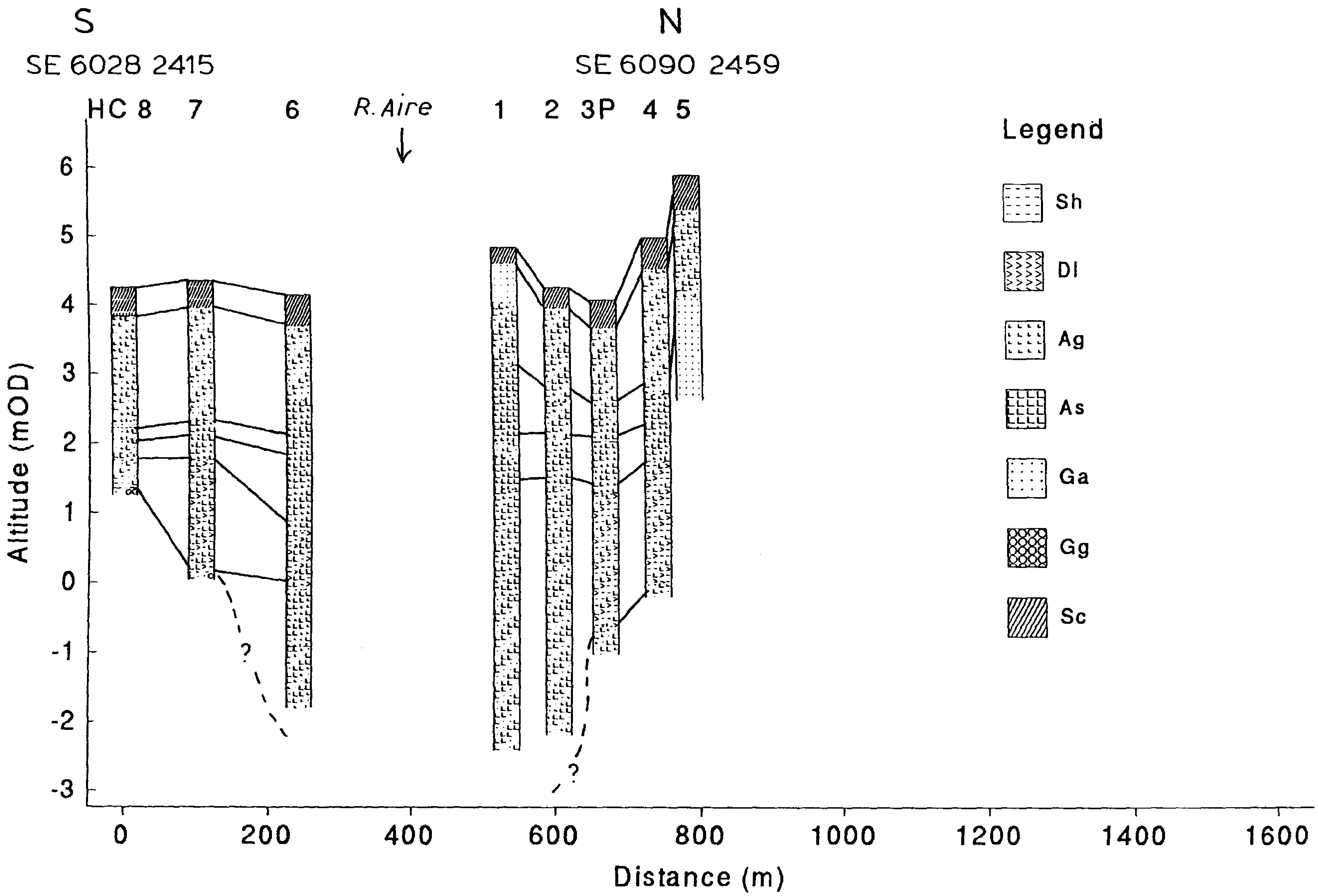


Figure 7.1 Lithostratigraphy of the cross valley transect at Hirst Courtney in the lower Aire valley (symbols follow Troels-Smith (1955))

Overlying the pre-Holocene surface in cores HC 3, 4, 7, and 8 is a detrital wood peat, which was found between -0.63 m OD in HC 3P, and 1.72 m OD in HC 4. The lower contact between the base of peat and the underlying pre-Holocene unit is transgressive up the valley sides. At the base of HC 8, this transitional unit consists of mixed sandy peat at +1.26 m OD. This organic unit typically consisted of large pieces of detrital wood and leaves, with some turfaceous rootlets, in a humified organic matrix (*Substantia humosa, sensu* Troels-Smith, 1955). Although the unit appeared to be predominantly biogenic in the field, there is a high clay content within the peat matrix, which characteristically increased around woody branches in the peat. This peat is 189 cm thick in borehole HC 3P where it overlies the basal unit of stiff silty clay at -0.63 m OD.

The floodplain peat is gradually replaced by a blue-purple-grey silty clay (suggesting a transitional sedimentary contact) with occasional plant detritus. This unit is recorded in all cores, at a range of heights from +0.74 m OD in HC 6 to +1.72 m OD in HC 4, and contains some sand in borehole HC 1. Overlying this deposit is a sedimentologically similar unit which is distinguished separately by its green-grey colour. This silty clay unit is present in all cores, between +1.60 m OD in HC 6 and +2.69 m OD in HC 4, contains occasional organic detritus, and a trace of sand in boreholes HC 2 and 8. This unit is generally thicker on the north side of the floodplain and attains its maximum thickness of 107 cm in HC 1. Although these two alluvial units are laterally persistent across the valley, they generally occur at a lower altitude on the south side of the floodplain, possibly due to incision of a small creek channel into the peat in the vicinity of HC 6, or because of differential compaction of the organic unit.

The uppermost alluvial unit is a mottled orange-grey-brown silty, sandy clay or clayey, silty sand, with sparse organic debris, some rootlets (probably of modern origin), and is dessicated. It overlies the green-grey clays sharply in HC 2, 3P, and 8, but in all other cores the contact appears gradational. It occurs between +2.00 m OD in HC 6 and +5.38 m OD in HC 5, where it is overlain by topsoil.

The boreholes taken from the central valley area are atypical of the stratigraphy recorded from the floodplain deposits flanking the valley sides described above. Boreholes HC 1, 2 and 6 are the deepest cores and consist of laminated light and medium grey silty clays and sands. The lowest sediment is recorded at -2.38 m OD in a 721 cm borehole at HC 1, and forms the infill of an incised channel of unknown depth. This unit is characterised by a fining-up sequence of clays with sandy partings, grading into laminated silty clays.

A strato-type core was extracted from HC 3P, which contained all the representative lithostratigraphic units and the thickest and deepest sequence of peat. The detailed sediment descriptions are provided in Table 7.1. These sediments were sub-sampled in the laboratory for biostratigraphic analysis.

Site: Hirst Courtney, River Aire valley	**Core**: HC 3P

Grid reference: SE 6084 2449	**Date of fieldwork**: November 1997
Latitude: 53° 43' N	**Sampling method**: Stitz Piston Corer
Longitude: 1° 5' W (Greenwich Meridian)	**Bench mark datum**: Hirst Courtney School
Collector: J. R. Kirby/P. Goodwill	**Ground Altitude**: 4.07 m OD

Observed depth (cm)	**Altitude (mOD)**	**Troels-Smith Sediment Description** **nig., strf., elas., sic., lim. sup. (Troels-Smith, 1955)**
120 to 175	+2.87 to +2.32	Orange-brown heavily oxidised silty, sandy clay. As2, Ag1, Ga1, Lf++; 3, 0, 3, 0, 0.
175 to 206	+2.32 to +2.01	Abrupt contact onto blue-grey-green silty clay with trace of fine sand and organic detritus As2+, Ag2, Ga+, Sh+; 3, 0, 3, 0, 4.
206 to 281	+2.01 to +1.26	Transition to purple-grey clay with silt and occasional organic detritus. As2+, Ag2, Sh+; 3, 0, 3, 0, 0.
281 to 470	+1.26 to -0.63	Grades onto dark brown clayey detrital wood peat with turfa. Sh2, Dl1, As1+, Ag+, Dh++, Th^1++; 3+, 0, 3, 0, 0.
470 to 480	-0.63 to -0.73	Blue-grey silty clay with sand, organic detritus and occasional gravel sized sandy concretions and chalky flecks. As2, Ag1, Ga1, Gg min+, Sh+, Dl+; 3, 0, 3, 0, 0.

Table 7.1 Lithostratigraphic details from core HC 3P described using the Troels-Smith (1955) terminology.

7.3 Radiocarbon dating

Four bulk samples were selected for conventional radiocarbon dating and sent to East Kilbride for analysis. These dates form a consistent series (Table 7.2) and provide a chronology for the timing of specific biostratigraphic events (see section 7.4). However, radiocarbon ages are not equivalent to calendar ages due to the non-linear nature of the calibration curve (see section 5.9.4) and as a result, the radiocarbon timescale compresses or stretches calendar time due to under and over-production of ^{14}C respectively (Bartlein *et al.*, 1995). To account for the effects of secular variations in atmospheric ^{14}C on the calculation of processes and rates of change in Holocene environments, dates have been calibrated using the CALIB 3.0 program of Stuiver & Reimer (1993).

Mid cal. age estimates have also been calculated in Table 7.2 for the purposes of the age-depth model using the method explained in section 5.9.4, after Bennett (1994b).

Laboratory Code	Depth (cm)	Altitude (m OD)	Age (^{14}C yr BP$\pm 1\sigma$)	Calibrated Age Range * (cal. yrs BP $\pm$ 2σ)			Mid Cal. Age (cal. yrs BP$\pm 1\sigma$)
				max	(cal ages)	min	
SRR-6268	281 to 286	+1.25 to +1.21	2560±45	2757	(2734)	2480	2619±138
SRR-6269	310 to 315	+0.97 to +0.92	2665±45	2850	(2760)	2740	2795±55
SRR-6270	360 to 365	+0.47 to +0.42	3195±40	3471	(3388)	3346	3409±63
SRR-6271	465 to 470	-0.58 to -0.63	4150±45	4831	(4808, 4764, 4642 4621, 4619)	4525	4678±153

* Calibrated using Stuiver & Reimer (1993)

Table 7.2: Radiocarbon dates and calibrated age estimates from Hirst Courtney.

7.3.1 Age-Depth Model

Due to the variable nature of sediment deposition in fen floodplain environments, it is not appropriate to calculate precise sedimentation rates or deposition times (see section 5.9.4). However, to establish an approximate chronology for the local pollen zone boundaries in the following description, an age-depth curve has been plotted using linear interpolation between dates (Figure 7.2). Average sedimentation rates (cm [cal.] yr BP $^{-1}$) enable tentative age estimations to be made for undated parts of the core (see section 5.9.4 for methods). Error bars are based on the 2σ age range of the date, and sample thickness. In the discussion of pollen assemblage zones (section 7.4.2), dates quoted without the error term are based on linear interpolation of ages from the age depth model. Calibrated ages are also stated in all cases.

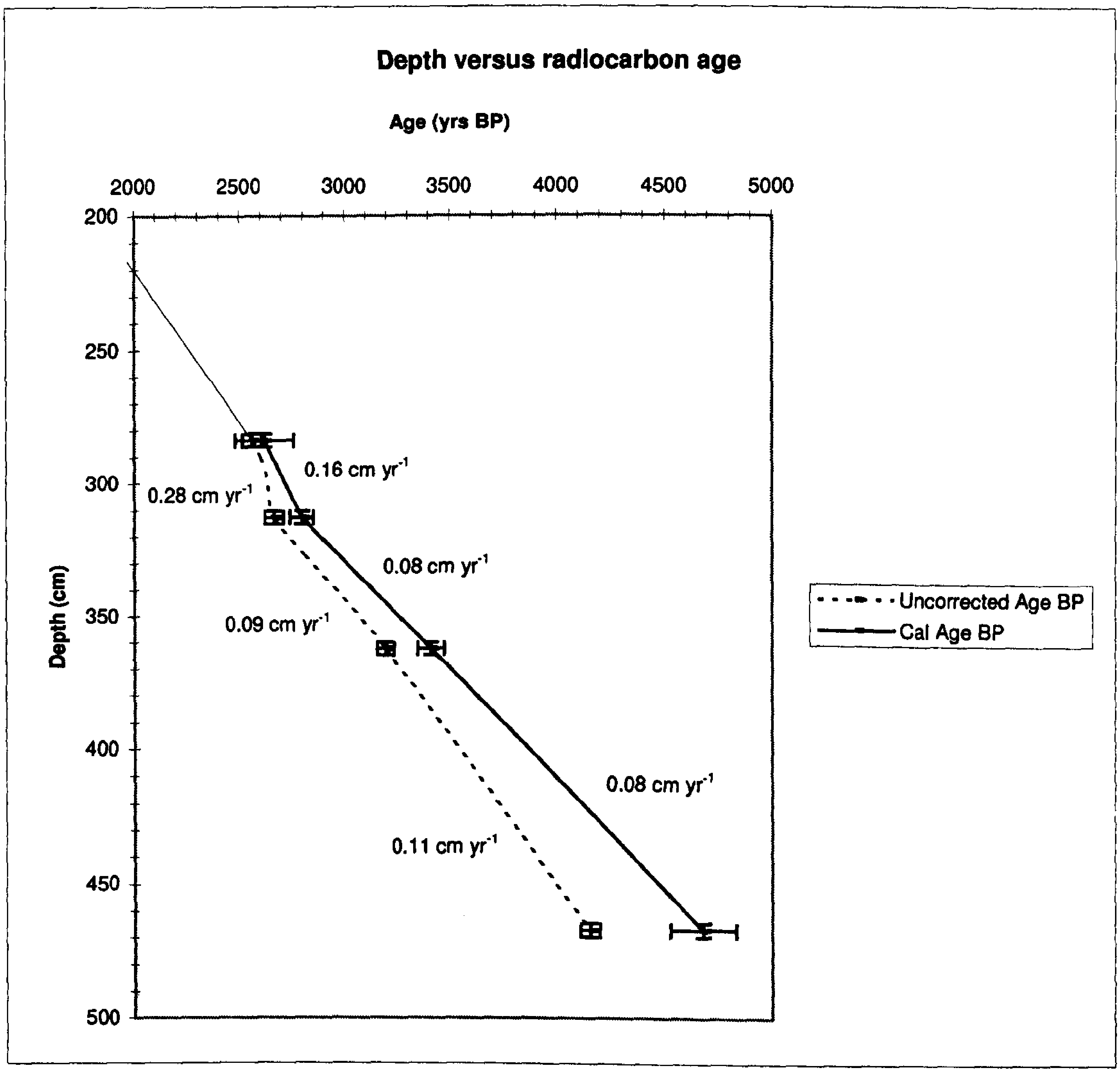


Figure 7.2 Hirst Courtney Age-Depth Model

7.4 Biostratigraphy

7.4.1 Wood Macrofossils

A variety of wood fragments were recorded from the cores at Hirst Courtney. Several large pieces of wood were sampled within HC 3P (Table 7.3) and have been identified using the methods outlined in section 5.7.

Depth Range (cm)	Macrofossils	Depth Range (cm)	Macrofossils
325 to 340	*Alnus glutinosa* wood	465 to 466	*Fraxinus excelsior* wood
373 to 374	*Fraxinus excelsior* wood	490 to 491	*Fraxinus excelsior* wood
405 to 415	*Fraxinus excelsior* wood	515 to 516	*cf. Alnus glutinosa* root wood
452 to 460	*Fraxinus excelsior* wood		

Table 7.3 Wood macrofossils from core HC 3P.

7.4.2 Pollen Stratigraphy

The results of the pollen analysis are presented in the form of a percentage pollen diagram (Figure 7.3) which has been drawn using the TILIA program (Grimm, 1993). In the description of pollen assemblages which follows, the basic sum for percentage calculations is TLP-*Alnus* except where stated. Information regarding the pollen types used, the pollen sum, and the method of calculation is given in section 5.5.5 and 5.5.6. A pollen diagram of rare pollen types that did not exceed 1% of the pollen sum is shown separately (Figure 7.4).

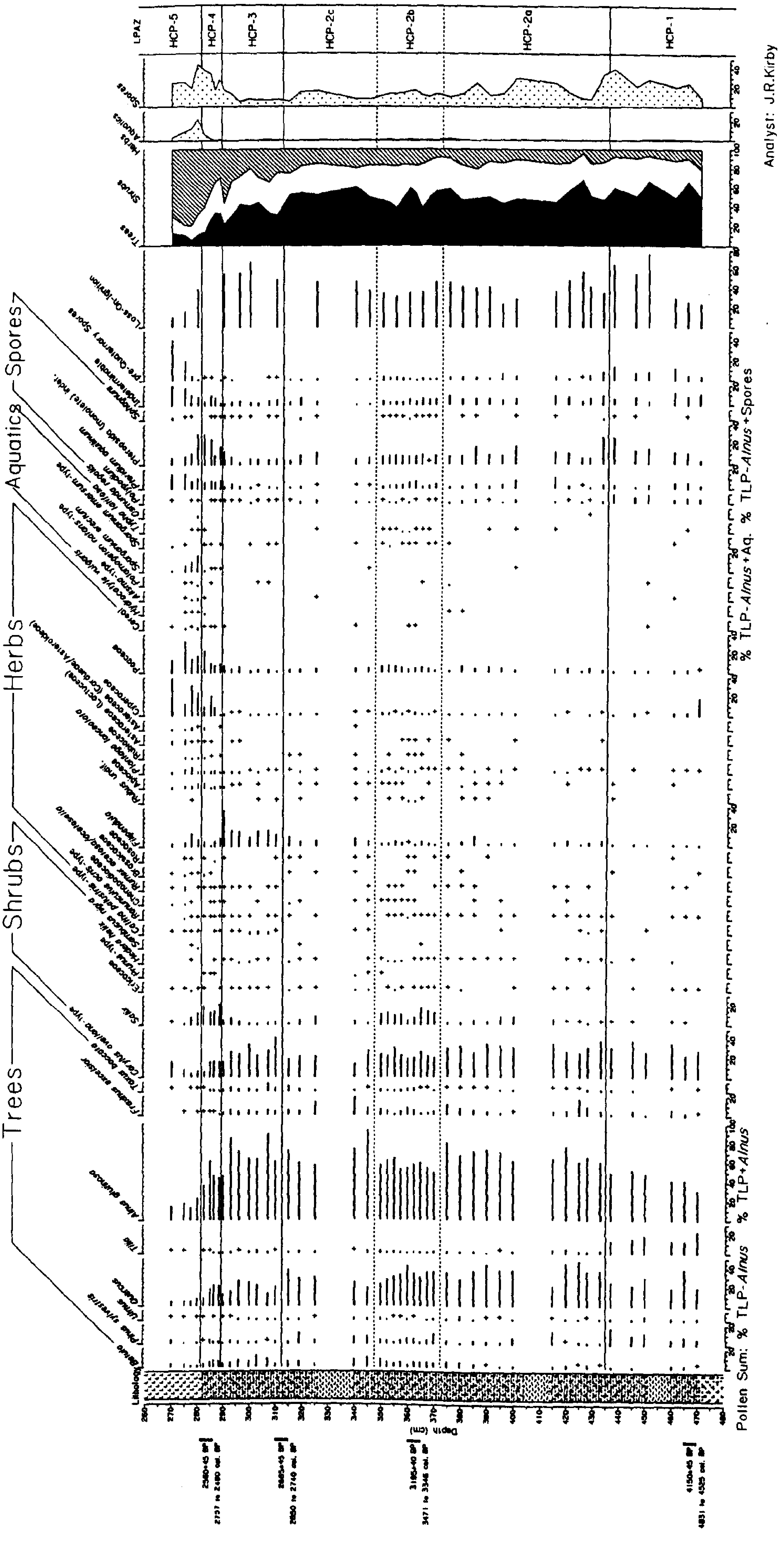


Figure 7.3 Hirst Courtney percentage pollen diagram (+ represents frequencies <1%)

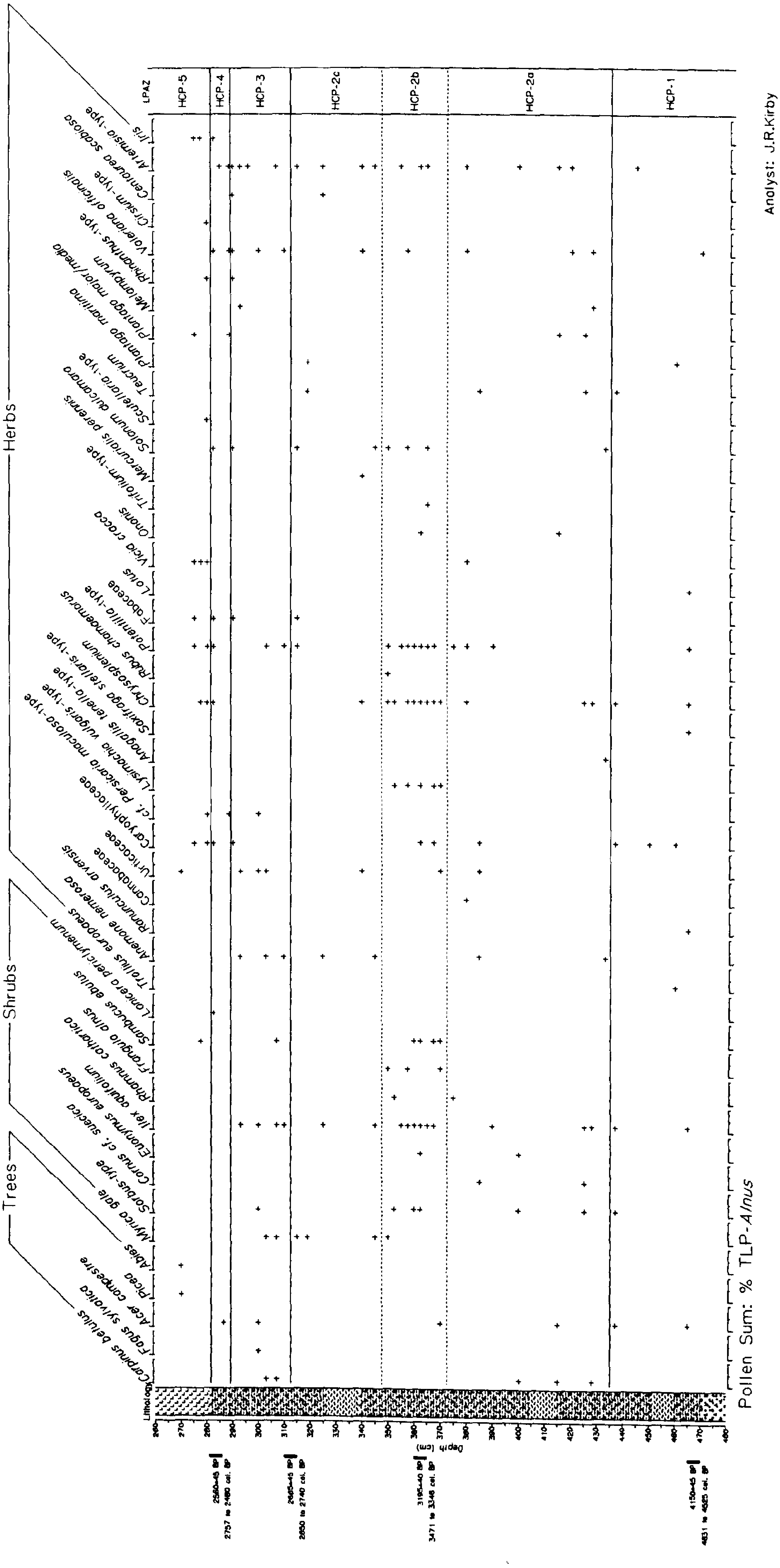


Figure 7.4 Hirst Courtney percentage pollen diagram showing minor types (+ represents frequencies <1%)

HCP-1 470 to 435 cm *c*.4200 to *c*.3900 BP (*c*.4700 to *c*.4300 cal. yrs BP)

Figures 7.3 and 7.4

The opening of this pollen local pollen assemblage zone (lpaz) is dated to 4150±45 BP (4831 to 4525 cal. yrs BP) and is dominated by arboreal pollen (51 to 67%) and shrub pollen (24 to 39%) types. The main characteristics are high frequencies of tree taxa pollen such as *Pinus sylvestris*, *Quercus*, *Tilia*, and *Alnus glutinosa*. Of the arboreal pollen taxa, *Alnus glutinosa* is most abundant and values rise slowly from 31% to *c*.50% (TLP+*Alnus*) towards the close of the lpaz. *Quercus* (20 to 38%) and *Pinus sylvestris* (5 to 17%) pollen is also a prominent constituent of the lpaz. *Tilia* pollen is most frequent in the lowest level at 22% and then drops to an average of 14% throughout the rest of the lpaz. Pollen of *Betula*, *Ulmus*, and *Fraxinus excelsior* is poorly represented.

Shrub pollen percentages are dominated by *Corylus avellana*-type which averages 30%. Of the herb pollen types, Cyperaceae attains values of 17% in the basal sample but then declines to <3% in subsequent levels. Spores are represented mainly by *Polypodium* (3 to 9%), *Pteridium aquilinum* (1 to 6%), and Pteropsida (monolete) indet., which reaches 26% (TLP-*Alnus*+spores) at the close of the zone. Pre-Quaternary spores are consistently present with an average value of 7% (TLP-*Alnus*+PQS). LOI values rise abruptly midway through the zone from 25% to 65% dry weight.

HCP-2 435 to 312.5 cm *c*.3900 to *c*.2700 BP (*c*.4300 to *c*.2800 cal. yrs BP)

This zone is characterised by the same dominant arboreal and shrub pollen taxa as described for HCP-1, namely, *Quercus*, *Alnus glutinosa*, and *Corylus avellana*-type. However, important variations occur in the representation of other pollen taxa. Therefore, the zone is divided into three subzones accordingly.

HCP-2a 435 to 372.5 cm *c*.3900 to *c*.3300 BP (*c*.4300 to *c*.3500 cal. yrs BP)

The opening of this subzone is located where frequencies of *Alnus glutinosa*, *Quercus*, *Fraxinus excelsior* pollen begin to rise. There is a marked fall in *Pinus sylvestris* and *Tilia* pollen frequencies, from *c*.15 to 2%, which causes the overall tree pollen percentages to drop to *c*.50%. Whilst *Tilia* pollen percentages remain below 4% during the zone, *Pinus sylvestris* pollen values increase to *c*.10% either side of a large wood macrofossil, identified as *Fraxinus excelsior*, between 405 and 415 cm (Table 7.3). *Alnus glutinosa* pollen dominates the subzone (60 to 80% TLP+*Alnus*), with *Quercus* pollen values ranging from 21 to

46%. Frequencies of *Fraxinus excelsior* pollen rise to 17% at 425 cm, decline towards the centre of the subzone, and gradually increase to 11% at the upper subzone boundary. Shrub pollen percentages are dominated by *Corylus avellana*-type (25 to 38%), which rise towards the centre of the subzone at the expense of *Quercus* and *Alnus glutinosa* values. *Salix* pollen percentages match this trend, and rise to 8% between 400 and 395 cm.

Percentages of herb morphotypes such as *Filipendula* and Poaceae are better represented throughout this subzone than in HCP-1, although frequencies never rise above 9% for either taxon. Cyperaceae, *Ranunculus acris*-type, and *Plantago lanceolata* pollen occur consistently in each sample but contribute little to the total herb pollen percentages, which range between 4 and 20%. Percentages of *Polypodium*, *Pteridium aquilinum*, and pre-Quaternary spores exhibit minor fluctuations and are generally lower than in the previous zone. Values for LOI decline from a maximum of 57% at 425 cm to a minimum of 24% at 395 cm and then rise to 46% dry weight at the close of the subzone.

HCP-2b 372.5 to 347.5 cm *c*.3300 to *c*.3000 BP (*c*.3500 to *c*.3200 cal. yrs BP)

There is a sharp drop in *Alnus glutinosa* pollen values to a minimum of 48% (TLP+*Alnus*) at the beginning of this subzone. Simultaneously, *Salix* pollen values rise sharply, and attain an average value of 11% throughout the subzone. Several other tree pollen taxa achieve greater percentages accompanying this fall in *Alnus glutinosa* pollen, but to a lesser extent than *Salix*. For example, pollen percentages of *Betula* (3 to 9%), *Pinus sylvestris* (1 to 10%), and *Taxus baccata* (1 to 3%) increase during this subzone. *Quercus* pollen percentages decline from 42% at 360 cm to 24% at 350 cm.

Filipendula (1 to 5%) and Poaceae pollen (2 to 8%) remain the most important herb taxa with lesser amounts of Cyperaceae (<3%) and *Plantago lanceolata* pollen (<2%). A bulk sediment sample from between 360 and 365 cm has been dated to 3195±40 BP (3471 to 3346 cal. yrs BP). This corresponds with a reduction in *Alnus glutinosa*, *Corylus avellana*-type, and *Salix* pollen values, a coincident increase in *Quercus* and *Fraxinus excelsior* pollen percentages, and the first occurrence of cereal pollen grains (*cf. Hordeum*-type). Although aquatic pollen contributes less than 2% of the pollen sum, pollen from *Sparganium emersum*-type and *Typha latifolia* are recorded with greater frequency than in HCP-1 and 2a. *Pteridium aquilinum* (2 to 7% TLP-*Alnus*+spores) and Pteropsida (monolete) indet. (6 to 10% TLP-

Alnus+spores) morphotypes occur as consistent components of the spore category. LOI values range between 32 and 47% dry weight.

HCP-2c 347.5 to 312.5 cm *c.*3000 to *c.*2700 BP(*c.*3200 to *c.*2900 cal. yrs BP)

The opening of HCP-2c is marked by a rise in *Alnus glutinosa* pollen percentages from 58% TLP+*Alnus* during the last level of HCP-2b, to (93% TLP+*Alnus*), and remains above 57% (TLP+*Alnus*) for the subzone duration. *Quercus* pollen percentages rise from 20% at 345 cm to 38% at 315 cm. Arboreal pollen is slightly better represented, with percentages averaging 57%. *Betula* and *Pinus sylvestris* pollen is more abundant than previously and *Fraxinus excelsior* pollen values in particular peak at 18% at 340 cm. *Corylus avellana*-type (14 to 28%) and *Salix* pollen (2 to 8%) are the main constituents of the shrub pollen component.

Herb pollen comprises *c.*16% of the TLP-*Alnus* sum and the main components are pollen of Poaceae (4 to 5%) and *Filipendula* (5 to 10%); percentages of the latter rising towards the end of the subzone. *Pteridium aquilinum* spores are no longer an important constituent of the spore sum which is mainly comprised of Pteropsida (monolete) indet. morphotypes (4 to 12% TLP-*Alnus*+spores). LOI values are generally higher than lpaz HCP-2b and range between 38 and 47% dry weight. The lpaz HCP-2c/3 boundary is dated to 2665±45 BP (2850 to 2740 cal. yrs BP).

HCP-3 312.5 to 289 cm *c.*2700 to <*c.*2600 BP (*c.*2900 to <*c.*2600 cal. yrs BP)

This lpaz is characterised by a fluctuating but generally declining level of arboreal pollen values and an increase in pollen percentages from shrubs and herbs. *Alnus glutinosa* (46 to 89% TLP+*Alnus*), *Quercus* (13 to 27%), and *Corylus avellana*-type (16 to 41%) pollen dominate the assemblage, and *Betula* pollen (3 to 12%) is also well represented. The transition from HCP-2c is marked between 315 and 307 cm by a fall in *Quercus* pollen percentages from 38 to 14%, and a corresponding increase in percentages of *Corylus avellana*-type pollen (18 to 41%), then *Betula* and *Taxus baccata* pollen, which attains percentages of 8% at 307 cm.

Coincident with this change is a rise in *Filipendula* pollen values from 10 to 37%, which peak with a spike at the close of the zone. Frequencies of *Corylus avellana*-type pollen fall to 23% as *Filipendula* pollen values continue to rise, but then recover. Pollen frequencies for herb types such as *Caltha palustris*-type (<3%) are higher than lpaz HCP-2c, and Poaceae pollen values (2 to 8%) rise towards the top of the zone. Spores attain an average value of 7% (TLP-*Alnus*+spores) and are again mainly represented by Pteropsida (monolete) indet. LOI rises to 67% at 300 cm and then falls to 55% dry weight at 290 cm.

HCP-4 289 to 281 cm <*c.*2600 BP (<*c.*2600 cal. yrs BP)

Major shifts occur in the pollen curves of arboreal, shrub and herb types in this zone. Frequencies of the main tree pollen taxa are lower than in previous lpazs and total arboreal pollen values fall from 36 to 15% towards the zone end. *Alnus glutinosa* pollen values decline from 44 to 36% (TLP+*Alnus*) and frequencies of *Quercus* pollen fall from 22 to 9%. Percentages of *Betula*, *Fraxinus excelsior*, and *Taxus baccata* pollen are similarly reduced in importance although the frequency of shrub pollen continues to be well represented (25 to 38%). *Corylus avellana*-type pollen occurs initially at 17% and falls to 5% at the zone close. *Salix* pollen rises sharply across the lower zone boundary, and maintains percentage values between 13 and 21% throughout the zone.

Herb pollen contributes 27 to 60% of the total pollen assemblage, despite the rapid decline in *Filipendula* pollen from a peak in the uppermost level of lpaz HCP-3 of 37%, to much lower frequencies (between 2 and 9%) in lpaz HCP-4. Cyperaceae (1 to 21%) and Poaceae pollen (8 to 23%) values show rising trends, and Apiaceae and *Plantago lanceolata* pollen becomes more abundant. Spore frequencies rise from 17% to 38% (TLP-*Alnus*+spores) at the top of the zone mainly due to *Pteridium aquilinum* (7 to 9% TLP-*Alnus*+spores) and Pteropsida (monolete) indet. (9 to 31% TLP-*Alnus*+spores) morphotypes. The close of the zone is dated to 2560± BP (2757 to 2480 cal. yrs BP).

HCP-5 281 to 270 cm >*c.*2600 BP (> *c.*2600 cal. yrs BP)

This final lpaz is dominated by herb (67 to 79%) and aquatic pollen (3 to 22% TLP+aquatics) types, with arboreal and shrub pollen only contributing, on average, 11 and 15% respectively. *Quercus* (5 to 8%) and *Alnus glutinosa* pollen (12 to 23% TLP+*Alnus*) frequencies continue their decline from the previous zone with none of the other tree pollen types reaching values over 3%. Frequencies of *Corylus avellana*-type pollen rise from *c.*5% to 16% during the zone, whereas *Salix* pollen frequencies show an opposite trends declining from 13% to <1%.

As in HCP-4, Cyperaceae (13 to 39%) and Poaceae pollen (13 to 33%) are the most important of the herbs, with high proportions of Poaceae pollen occurring where frequencies of Cyperaceae are low. Frequencies of other herb pollen types expand in number and diversity. Percentages of *Filipendula*, Chenopodiaceae, Brassicaceae, Apiaceae, *Plantago lanceolata*, Asteraceae (Cardueae/Asteroideae) and Asteraceae (Lactuceae) pollen are all higher than in previous zones. Cereal-type pollen grains (*Hordeum*-type) also occur at low frequencies (<2%). Aquatic pollen increases in variety although the abundance of *Sparganium erectum* pollen declines from 17 to 1% (TLP-*Alnus*+aquatics) during the zone. Other aquatic pollen taxa such as *Hydrocotyle vulgaris*, *Alisma*-type, and *Potamogeton natans*-type are consistently present. Spore frequencies of *Pteridium aquilinum* rise from 10 to 15% (TLP-*Alnus*+spores) whereas those of Pteropsida (monolete) indet. fall from 31 to 7% (TLP-*Alnus*+spores). The abundance of pre-Quaternary spores rise sharply between 280 and 270 cm, from 2 to 42% (TLP-*Alnus*+PQS), and LOI values decline from 39 to 9% dry weight.

7.4.3 Diatom Stratigraphy

The following section describes the salient features of the local diatom assemblage zones (ldazs) from core HC 3P. Diatoms are absent from the organic deposits at Hirst Courtney and first appear in the overlying clastic sediments. The record therefore post-dates 2560±45 (2757 to 2480 cal. yrs BP). TILIA graphs (Grimm, 1993), representing common and rare diatoms types classified according to salinity groups, are presented in Figures 7.5 and 7.6, and a summary of diatom life form classes is shown in Figure 7.7 (see section 5.6.4). Diatom frequencies are expressed as a percentage of total diatom valves (see section 5.6.3). For a definition of the salinity and habitat classifications, see Appendix B.

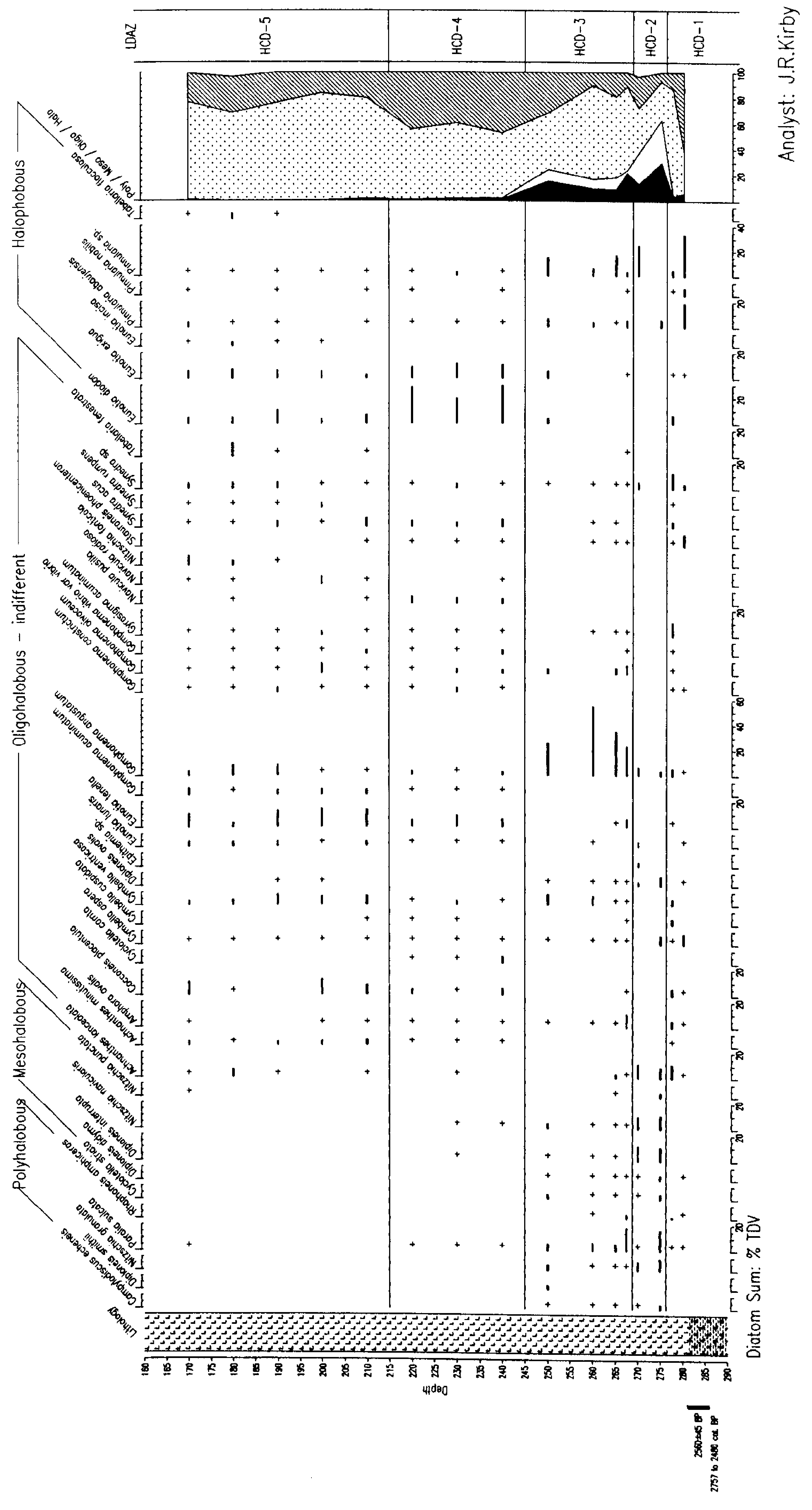
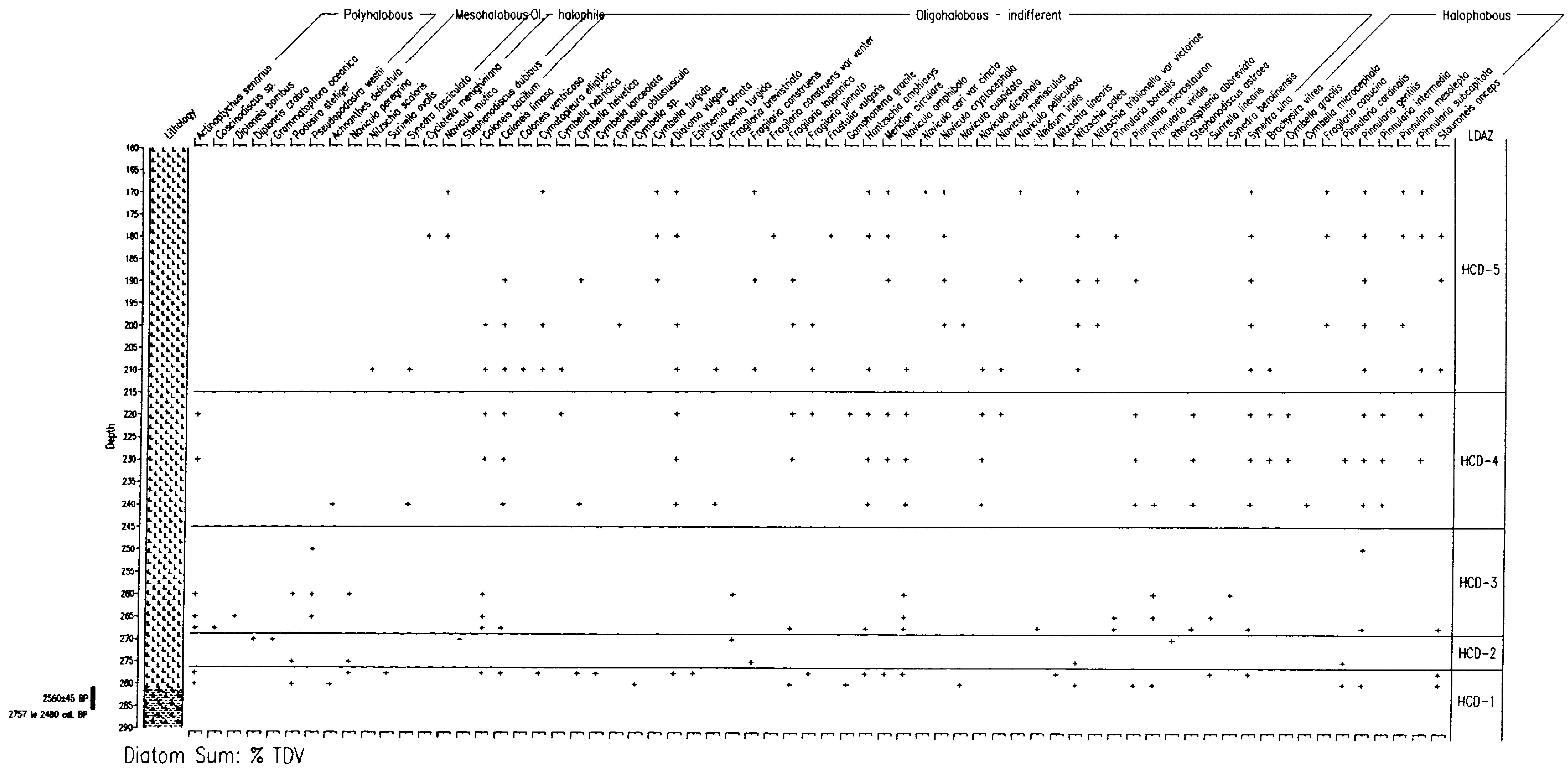


Figure 7.5 Hirst Courtney percentage diatom diagram based on salinity (+ represents frequencies <3%)



Analyst: J.R.Kirby

Figure 7.6 Hirst Courtney percentage diatom diagram (minor taxa) based on salinity (+ represents frequencies <3%)

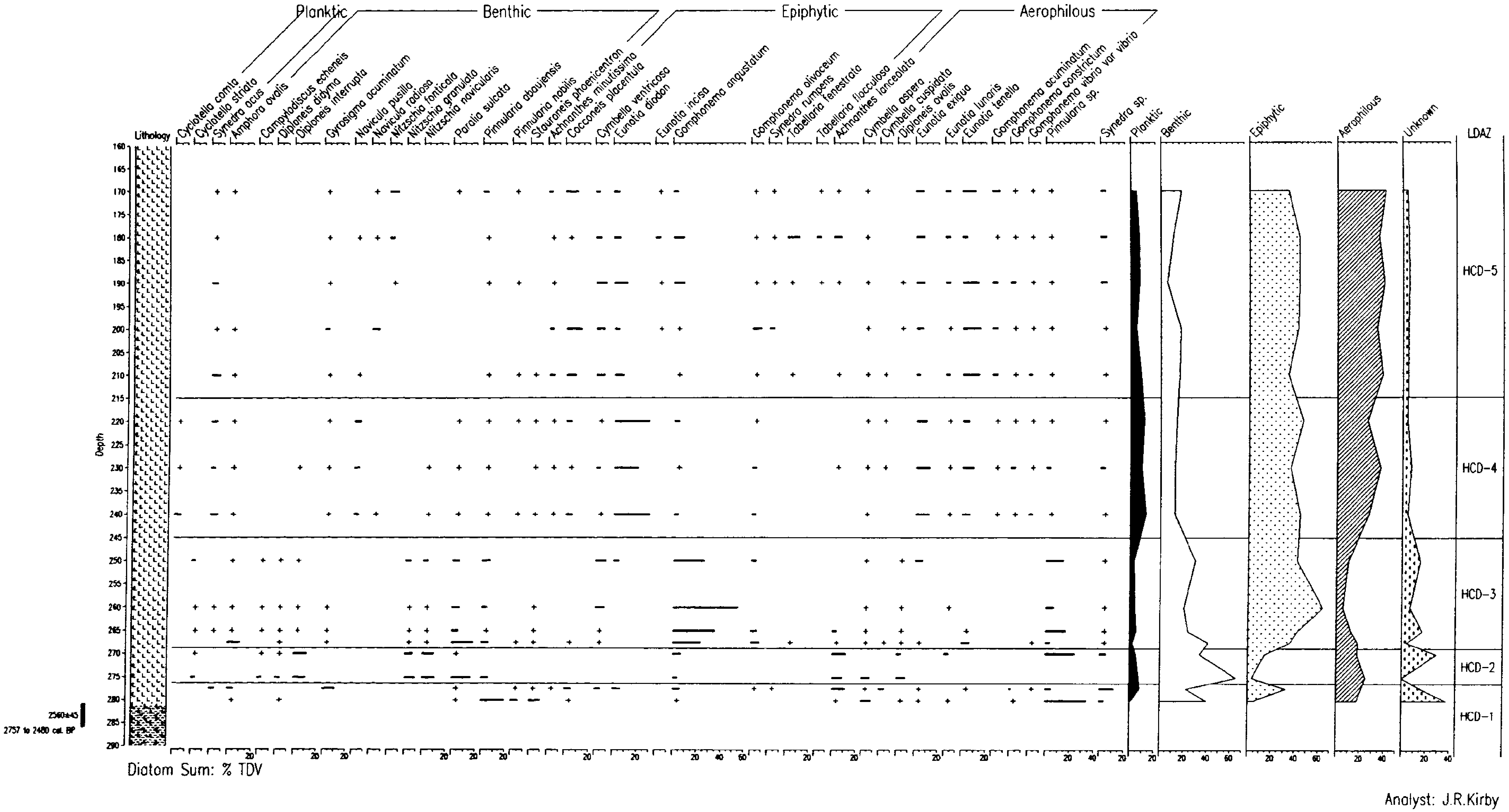


Figure 7.7 Hirst Courtney percentage diatom diagram based on life-form (+ represents frequencies <3%)

HCD-1 280 to 276 cm <2560±45 (<2757 to 2480 cal. yrs BP)

Figures 7.5, 7.6, and 7.7

Only 2 samples are included in this ldaz and therefore, it is not appropriate to put interpretative weight on the trends that occur within ldaz HCD-1. However, the fact that some distinct shifts in composition of the assemblage occurs, enables some circumspect observations to be made. The lowest level in this zone is dominated by halophobous diatoms (63%), which decline to 13% in the second level where the dominant group are oligohalobous varieties (83%). The two most important halophobous diatoms are *Pinnularia abaujensis* (19%) and *Pinnularia nobilis* (6%). The *Pinnularia* sp. category has been included in this halophobian group because it consisted mainly of broken valves of the *Pinnularia* genus that strongly resembled fragments of these two species. *Cymbella aspera* (8%) and *Stauroneis phoenicenteron* (9%) are the main species from the oligohalobous group.

In the next sample, frequencies of diatoms of the genus *Pinnularia* fall away and are replaced by a variety of oligohalobous diatoms such as *Achnanthes lanceolata* (11%), *Cocconeis placentula* (5%), *Gomphonema angustatum* (6%), *Gyrosigma acuminatum* (10%), *Synedra acus* (4%), and broken valves resembling *S. acus* (*Synedra* sp., 12%). Planktonic (3 to 9%), epiphytic (5 to 33%), and aerophilous (17 to 21%) forms increase in importance throughout this zone, and benthic (38 to 21%) species decline.

HCD-2 276 to 269 cm

The main feature of this ldaz is the increase in polyhalobous diatoms to 30% and mesohalobous diatoms to 33% during the opening of the zone. Statements are of a tentative nature due to the few samples within the zone and the low number of diatom valves (<100) counted at these levels. However, despite the low abundance of diatoms, clear changes in assemblage composition are observed. Several polyhalobous species are abundant in this zone such as the tychoplanktonic species *Paralia sulcata* (16 to 3%) and the benthic species *Nitzschia granulata* (7 to 9%), and *Campylodiscus echeneis* (2 to 4%). *Diploneis interrupta* (11%) and *Nitzschia navicularis* (10 to 11%) are the most important mesohalobous species and, along with *Nitzschia granulata*, contribute to the peak of benthic taxa of 49%. Oligohalobous taxa make up 30 to 36% of the total, with *Gomphonema angustatum* (4 to 7%), *Achnanthes lanceolata* (9 to 11%), and *Diplonies ovalis* (3 to 7%) among the main contributors. These latter two taxa are also the main constituents of the aerophilous group which reached 25% in this zone. Broken valves of halophobous *Pinnularia* species are abundant at the close of the zone which puts the total for this category to 24%.

HCD-3 269 to 245 cm

The dominant feature of this ldaz is the rise in frequency of *Gomphonema angustatum* diatoms from *c*.6% in HCD-2 to a peak of 54% at 260 cm in the centre of HCD-3. The abundance of this species, and the presence of diatoms such as *Amphora ovalis* (2 to 10%), *Cymbella ventricosa* (1 to 8%), and *Gomphonema olivaceum* (4 to 7%) cause the proportion of oligohalobous diatoms to rise to between 44 and 73% of the total. These diatoms (excepting *Amphora ovalis*) are members of the epiphytic group which contributes between 35 and 64% of the total. Percentages of benthic diatoms (*c*.20%) are lower than in ldaz HCD-2. This is largely attributable to a decline in abundance of benthic mesohalobous diatoms (such as *Diploneis interrupta* and *Nitzschia navicularis*), which total *c*.7%, but rarely do individual species occur at percentages >3%.

Polyhalobous diatoms exhibit a small peak, making up 22% of the total at the opening of the ldaz, but thereafter they contribute between 9 and 15%. The relatively high frequency of polyhalobous diatoms is mainly attributable to the abundance of *Paralia sulcata* (5 to 18%) which also strongly influences the percentage of planktonic diatoms which are present between 8 and 20%. The frequency of aerophilous diatom taxa declines from 18 to 6% in the middle of the zone and then rises to 10% at the zone close. Halophobous diatom percentages increase from *c*.13 to 36% at the end of the zone which coincides with a rise in *Pinnularia abaujensis* to 6% and *Pinnularia* sp. to 14%, and the appearance of *Eunotia diodon* (4%) and *Eunotia exigua* (6%).

HCD-4 245 to 215 cm

This zone is characterised by a marked increase in abundance of halophobous diatoms to an average of 44%, and a decline in values of meso- and polyhalobous diatoms. The main diatom species present are the halophobes *Eunotia diodon* (20 to 29%) and *Eunotia exigua* (9 to 12%). Oligohalobous diatoms occur consistently at *c*.55% of the total, with no one species dominant. Epiphytic species make up *c*.42% of the total, with *Eunotia diodon*, *Cocconeis placentula* (3 to 5%), *Gomphonema angustatum* (*c*.3%), and *Gomphonema olivaceum* (*c*.3%) the main species. Aerophilous diatoms increase to 36% of the total in the middle of the zone before declining to 26%. This diatom group is mainly represented by *Eunotia exigua* with *Eunotia tenella* (5 to 9%) also important. Benthic and planktonic species are consistently present and each comprise *c.*11% of the total.

HCD-5 215 to 170 cm

This ldaz is characterised by an increase in the frequency of oligohalobous diatoms and a decline in halophobous taxa. Polyhalobous and mesohalobous diatoms are only recorded at 170 cm and attain frequencies of <1%. Above this point, they are absent from the record. Oligohalobous diatoms comprise between 68 and 84% of the total with *Cocconeis placentula* (2 to 12%), *Cymbella ventricosa* (4 to 8%), *Eunotia tenella* (3 to 14%), and *Gomphonema angustatum* (1 to 8%) the main constituents of this group. *Eunotia diodon* (4 to 11%) and *Eunotia exigua* (3 and 8%) are the main species of the halophobous group, which reaches percentages between 16 and 28%. Although the dominant diatoms are different, the prevailing life form characteristics of the diatoms within this zone are similar to ldaz HCD-4. Epiphytic taxa reach values between 33 and 43%, with *Cocconeis placentula*, *Cymbella ventricosa*, *Eunotia diodon* and *Gomphonema angustatum* the main species. Aerophilous diatoms account for between 33 and 41% of the diatom sum. This group is characterised by *Eunotia exigua*, *Eunotia lunaris* (1 to 4%), *Eunotia. tenella* and *Gomphonema acuminatum* (2 to 6%). The planktonic (4 to 8%) and benthic (5 to 17%) groups are minor components of ldaz HCD-5.

7.5 Inferred Local Vegetation History and Depositional Environment

Distinguishing between dry and wetland vegetation is a fundamental problem in palaeoecological research. This uncertainty is partly attributable to the lack of taxonomic precision when identifying pollen but is also blurred by the wide ecological amplitude of many of the tree, shrub, and herb taxa; many plants within certain 'pollen-types' have both wetland and dryland affinities (Waller, 1993). Therefore, the interpretation of the pollen assemblages from Hirst Courtney are based on shifts in local and extra-local vegetation (see section 5.5.6), which may reflect wetland and dryland changes. The following palaeobotanical reconstructions are discussed in stratigraphic order according to the lpazs.

HCP-1 1 470 to 435 cm *c*.4200 to *c*.3900 BP (*c*.4700 to *c*.4300 cal. yrs BP)

Figures 7.3 and 7.4

The pollen evidence suggests the pre-peat environment supported a mid-Holocene woodland typical of this region (see Beckett, 1981; Smith, 1985a; Brayshay & Dinnin, 1999), consisting of *Quercus*, *Tilia* and *Corylus avellana* (the eutrophic nature of the environment suggests that the *Corylus avellana*-type pollen in this study is most likely to be attributable to *Corylus avellana* and not *Myrica gale*), with lesser amounts of *Fraxinus excelsior* and *Ulmus*. Whether these taxa formed a mixed community with *Corylus avellana* forming the woodland understorey, or relate to two or more communities occurring in separate areas on different soil substrates is unknown. The relatively low abundance of *Ulmus* pollen throughout this sequence suggests the deposits are post-*Ulmus* decline in age. This is confirmed by a radiocarbon date from the base of the peat which suggests that the onset of paludification began at 4150±45 BP (4831 to 4525 cal yrs BP).

High pollen percentages for *Tilia*, which is poorly dispersed (Godwin, 1975a; Bradshaw, 1981a; Greig, 1982a; Huntley & Birks, 1983) suggests the dry land surface of the valley floor prior to paludification was populated locally by this taxon. High *Tilia* pollen values occur in buried soils at the base of Holocene sedimentary sequences prior to peat initiation (e.g. Smith 1958a; Waller, 1988; Waller 1994b) and characteristically decline at the transition to the overlying peats due to the increasingly wet conditions. However, given the intolerance of *Tilia* to waterlogging, the lack of macrofossil evidence, and the fact that pollen percentages for *Tilia* do not decrease appreciably subsequent to peat formation, a peripheral environment of origin is suggested for this taxon.

Brown (1988) notes that *Tilia* often occurs on terraces raised above the floodplain surface. It is feasible that the occurrence of significantly high *Tilia* frequencies during this lpaz represents a community growing on such areas that remained dry during the initial stages of peat accumulation. Godwin (1975a) has also suggested that *Tilia* is prevalent at the ecotone of the floodplain margin and valley side where its growth is favoured by seepage. In the Brede valley in East Sussex, Waller (1998) records *Tilia* pollen percentages of *c*.30% TLP 40 m from the floodplain edge. *Tilia* values peak at 22% and average 14% in HCP-1, which suggests an extra-local source for the *Tilia* pollen in this lpaz.

Pinus sylvestris is also a notable component of the pollen assemblage in HCP-1. This taxon is often associated with sandy habitats. The presence of a sand levee features which forms the northern margin of the floodplain at Hirst Courtney (see Gaunt, 1994 and Figure 3.5) suggests it is possible that *Pinus sylvestris* may have been prevalent on this ground, and therefore its pollen is also part of the extra-local component of the pollen rain (*cf.* Jacobson & Bradshaw, 1981). Brown (1988) suggests *Pinus sylvestris* dominated the low sandy terraces of the river Stour in the Stourport area during the early-Holocene, and that it could have extended onto the drier areas of floodplain (Brown, 1984). Indeed, *Pinus sylvestris* is tolerant of waterlogged soils (Carlisle & Brown, 1968) and it is probable that *Alnus glutinosa* replaced *Pinus sylvestris* growing in wet valley bottoms and alluvial habitats (Clark & Godwin, 1962; Bennett, 1984) during the spread of *Alnus glutinosa* across Britain *c.*7500 BP (Smith & Pilcher, 1973; Bennett & Birks, 1990). It is ecologically unlikely *Pinus sylvestris* would occur with *Tilia* at the floodplain edge or on terrace features in the floodplain as *Pinus sylvestris* is easily out-competed when mixed with hardwoods (I.S Matyuk, quoted in Carlisle & Brown, 1968). Therefore it is perhaps more likely that these taxa occurred separately, with *Pinus sylvestris* growing on the sandy levee areas, and *Tilia* occupying the drier floodplain terraces and floodplain edge/valley side ecotone.

However, Huntley & Birks (1983) and Bunting *et al.* (1998b) suggest that frequencies for *Pinus sylvestris* of >25% TLP are required to indicate the presence of pine near a site, whereas Bennett (1984) suggests values of >20% TLP are sufficient. Within lpaz HCP-1, *Pinus sylvestris* pollen values range between 5 and 17%, but average only 9%. Therefore, the *Pinus sylvestris* pollen is perhaps more likely to be a part of the regional pollen rain and not from local or extra-local sources such as the adjacent sandy levee. *Pinus sylvestris* produces well-dispersed pollen in abundance and it is considered to be over-represented in the pollen rain (Andersen, 1970; Bradshaw, 1981a). Unfortunately it is not possible to discern between a small amount of pollen produced by a local stand or a larger amount, travelling from a long distance (*cf.* Oldfield, 1970; Bennett, 1984). In addition to this uncertainty, a fluvial origin for the *Pinus sylvestris* pollen cannot be ruled out. Work from a similar environment in the inner Thames by Devoy (1979) suggests that *Pinus sylvestris* pollen in these circumstances may be related to the inwashing of river water into the fen carr (see below). This may also explain the frequency of pre-Quaternary spores in this lpaz (see below).

The local depositional environment is easier to interpret in HCP-1. Cyperaceae was initially dominant close to the coring site during the onset of peat formation, but this open river marsh community (perhaps growing adjacent to the river margins) becomes replaced by the invasion of wetland trees such as *Alnus glutinosa* in subsequent levels. Huntley & Birks (1983) suggest that values of >25% TLP are common when alder is growing locally at a site. The frequencies of *Alnus glutinosa* pollen within lpaz HCP-1 (31 to 52%) therefore point to the presence of *Alnus glutinosa* locally at the coring site. This fen carr community then spread across the valley floor as peat development ensued.

The environment of deposition during HCP-1 therefore, is one of *Alnus glutinosa* dominated fen carr, with an initial marsh phase of sedge communities dominating the wetland. Open areas of wetland are suggested by low levels of fen herbs such as *Ranunculus acris*-type, *Filipendula,* and Cyperaceae. These could occur as local groundflora within the open fen carr or in wetter areas adjacent to the river Aire channel as suggested above. Herbs such as *Plantago lanceolata* and Poaceae also occur throughout the zone which indicates open ground of either dry or wetland in the vicinity of the site.

An environment of origin for the spores present in HCP-1 is difficult to determine. *Polypodium* often occurs as a epiphyte within fen carr (see Wheeler, 1980b) and many of the spores within the Pteropsida (monolete) indet. category could be from ferns colonising the wetland understorey. For example, some of the fern spores may be from *Thelypteris palustris* which favours shady areas in wet fen carr (Devoy, 1979). Walker *et al.* (1998) also suggest Pteropsida (monolete) indet. may be derived from tidal inwashing. The presence of pre-Quaternary spores is a persistent feature of the pollen diagrams in this area and may be derived from reworking of deposits containing crushed coal fragments (*cf.* Walker *et al.*, 1993). Such spores may be deposited from fluvial sources (e.g. bank erosion) or soil erosion from Devensian deposits (*cf.* Gaunt, 1994) in the dryland environment of the valley sides (see models by Scaife & Burrin, 1992; Caseldine & Barrow, 1997).

HCP-2a 435 to 372.5 cm *c.*3900 to *c.*3300 BP (*c.*4300 to *c.*3500 cal. yrs BP)

The opening of this lpaz is characterised by a marked biostratigraphic change as frequencies of both *Tilia* and *Pinus sylvestris* decline abruptly and *Alnus glutinosa* and *Quercus* pollen frequencies increase. This is of particular interest in this region as a decline in *Tilia* and *Pinus sylvestris* has been used to define the boundary between rpaz HHL/A and HHL/B *c.*3600 BP (*c.*4600 to *c.*4000 cal. yrs BP) on Thorne and Hatfield Moors (Smith, 1985a) in the Humberhead Levels. A similar decline in these taxa is also recorded at this time by Brayshay & Dinnin (1999) from Bole Ings in the Trent valley. This feature is attributed to deliberate human clearance of woodland by these Brayshay & Dinnin (1999), although there are other possible interpretations (see Chapter 10). Whilst the coincidence in timing of these events may imply a similar cause at Hirst Courtney, consideration of the changes occurring in the local environment at this time lends support to a possible, and it is argued, more likely alternative explanation for this event.

It is clear from Figure 7.1 that peat accumulation was spreading out and up the valley sides due to continued waterlogging during this time. A feasible explanation for this vegetation change is that peat development encroached upon the terrace feature or floodplain edge ecotone supporting *Tilia* as the floodplain environment expanded, and conditions became too wet for *Tilia* to tolerate. This is analogous to many of the pollen diagrams from buried soil/basal peat transitions in the Fenland (Waller, 1994b; 1994c), and Devoy (1980) also suggests that *Tilia* declines are coincident with significant rises in water table.

Tilia is more sensitive to high water levels than *Quercus* and *Alnus glutinosa*, which are able to survive at least until the period preceding peat formation. However, the rise in *Quercus* which accompanies this transition may also be a proportional rise in the pollen from surviving trees (with the demise of *Tilia*), rather than an actual increase in *Quercus* abundance (Waller, 1994b), or even a product of interdependence associated with percentage data. The rise in *Alnus glutinosa* and reduction in *Pinus sylvestris* pollen is likely to be attributable to the closing of the *Alnus* fen carr during this time which filters out much of the *Pinus sylvestris* pollen previously derived from airborne regional sources (e.g. Tauber, 1965; 1977; Jacobson & Bradshaw, 1981).

Although the possible influence of human activity is not discounted, the above interpretation is favoured in this instance. A Bronze Age *Tilia* decline is a marked feature of many pollen diagrams from Britain, which is generally accepted as having an anthropogenic cause (Turner, 1962). Turner (1962) has dated a primary *Tilia* decline at Thorne Moors to *c.*3100 BP (*c.*3500 to *c.*3000 cal. yrs BP) which is not replicated elsewhere in the Humberhead Levels. However, the secondary decline she dated has been identified elsewhere in the Humberhead Levels by Smith (1985a) and dates to *c*.2250 BP (*c*.2500 to *c*.2000 cal. yrs BP). The date for the pronounced *Tilia* decline in the surrounding area is therefore much later than the decline at Hirst Courtney, which further suggests a separate explanation is justified.

The increase in *Fraxinus excelsior* during the opening of lpaz HCP-2a may also have been a beneficiary of the reduction in *Tilia* due to differential tolerance to waterlogging. *Fraxinus excelsior* has been recorded as a component of *Alnus* carr fen woodland from elsewhere in the British Isles and the Netherlands (Godwin, 1975a; 1978; Wheeler, 1980b; Wardle, 1981; Van der Wiel, 1982; Wiegers, 1992; Waller, 1994c; Rodwell, 1991) and has been shown to be particularly tolerant of waterlogging (Iremonger & Kelly, 1988). *Fraxinus excelsior* is a light-demanding tree (Tinsley, 1981), so the open areas also encourage increased flowering and greater production of pollen. Huntley & Birks (1983) suggest values of >5% TLP indicate woodland in which *Fraxinus excelsior* was either dominant or co-dominant. Therefore, it is likely that *Fraxinus excelsior* (along with *Quercus* and *Alnus glutinosa*) colonised the areas left by *Tilia*, and formed part of the marginal fen community on the newly waterlogged floodplain terraces or valley sides. The peak in *Fraxinus excelsior* pollen of 17% in HCP-2a, also indicates that it may have been present more locally. This is supported by the presence of *Fraxinus excelsior* wood macrofossils from within the peat at Hirst Courtney. Wood macrofossils also indicate *Fraxinus excelsior* was a local component of fen carr in the perimarine area of The Netherlands during the Holocene (Van der Wiel, 1982).

The identification of a *Fraxinus excelsior* wood macrofossil between 405 and 415 cm coincides with some small changes in the pollen assemblage which may be related to taphonomic and depositional processes. The clay content of the sediment is greater around many of the wood macrofossils uncovered during coring, which is reflected here by a drop in LOI values to *c*.30% in the centre of HCP-2a. The presence of large wood macrofossil remains appears to locally affect the hydrology of the floodplain and may act as a conduit for water containing silts and clays. During times of flood, such large obstructions may also trap river sediments. Waller (1994a) similarly attributes such local, laterally impersistent clay lenses to avulsion of small drainage networks within the fen carr. In addition, Devoy (1979) suggests the large inorganic fraction of fen peats situated in the inner Thames estuary indicates the importance of river inwashing.

In this instance, pollen from *Alnus glutinosa* declines and that of *Pinus sylvestris* and *Pteridium aquilinum* increases. This may be a reflection of the dilution of local pollen by the inwashing sediment (*cf.* Wheeler, 1995) and the input of pollen susceptible to floatation. *Pinus sylvestris* pollen has been shown to be particularly prone to floatation by Hopkins (1950). De Jong (1970-71) has shown *Pinus sylvestris* to be typical of inorganic river marsh sediments and Robinson (1993) interpreted high counts of *Pinus sylvestris* pollen as indicative of the presence of water borne grains in a perimarine environment. Clark & Patterson (1985) report abundant *Pinus sylvestris* pollen in intertidal sediments and Long *et al.* (1999b) use the recent introduction of *Pinus sylvestris* as a chronostratigraphic marked to establish recent sedimentation rates of mudflats in relation to sea-level rise.

Waller (1994c) provides examples of how *Pteridium aquilinum* spores show a similar relationship during the deposition of more clastic Fenland sediments. The presence of allochthonous, fluvially derived sediments on the floodplain at this time inevitably means that the pollen deposited with this sediment potentially has a source area as large as the fluvial catchment. The discontinuous nature of these lenses of peaty clay suggest that they reflect very minor drainage networks upon the floodplain but, nevertheless, the potential expansion of the pollen source area is an important consideration in interpreting the pollen assemblages from such horizons.

In addition to the difficulties in reconstructing the nature of the extra-local vegetation from the pollen signal, a similar problem is encountered when attempting to establish the local fen vegetation. The pollen spectra from the biogenic deposits in HCP-2a (and indeed Hirst Courtney in general) are dominated by *Alnus glutinosa, Quercus* and *Corylus avellana*-type pollen. The relative contribution of pollen from these taxa, from dryland and wetland sources, is difficult to establish. Local vegetation components are often indicated by very abundant and fluctuating values for the particular taxon. This is pertinent to species such as *Alnus glutinosa*, which is a prominent local taxon of mid-Holocene floodplain corridor vegetation.

Cowell & Innes (1994) report 'superabundant' alder within local fen carr communities at Ince Blundell in Merseyside with percentages >80% TLP. Smyth & Jennings (1990) record values of 352% TLP-*Alnus* from the Combe Haven valley in East Sussex, and similarly high values are common from many lowland

British sites (e.g. Devoy, 1979; Thorley, 1981; Scaife, 1982; Brown, 1988; Waller, 1993; 1994a; 1994c; Waller *et al.*, 1999). The abundance of *Alnus glutinosa* is indicative of a high degree of soil moisture, with the water table permanently at, close to, or above the surface (McVean, 1953). *Alnus glutinosa* is particularly suited to these conditions and is specially adapted to hold a competitive advantage over other native British tree species due to the production of adventitious roots (McVean, 1956a; 1956b). Iremonger & Kelly (1988) provide evidence that *Alnus glutinosa* is one of the most tolerant trees to waterlogging. These optimum conditions enable alder to dominate river valley environments for thousands of years (Brown, 1988). However, some degree of caution must be asserted here, given that some of the *Alnus glutinosa* pollen appeared within clumps. This feature has been observed elsewhere and is indicative of macrofossil deposition of inflorescences (Waller, 1993; 1994c). Also, the scalariform perforation plates of *Alnus glutinosa* wood were recognised on the pollen slides. Therefore, in addition to the fact that *Alnus glutinosa* pollen is produced in abundance and well dispersed compared to other tree taxa (Andersen, 1970), the over-representation of *Alnus glutinosa* pollen may be exacerbated further due to the possibility of local deposition of anthers. Although this evidence implies that *Alnus glutinosa* was prominent in the 'on-site' vegetation, it is also possible that some of the *Alnus glutinosa* pollen was derived from trees growing within the dryland woodland.

Assigning an environment of origin for the *Corylus avellana*-type pollen is notoriously difficult, not least because of the fact that *Corylus avellana* pollen is not easily separated from that of *Myrica gale*, which although is generally associated with mires and bogs, can also occur in rich fen systems (Wheeler, 1980b). Early studies by Sir Harry Godwin noted how hard the curves for *Corylus avellana* pollen are to interpret; 'Of all the pollen shown in the (Fenland) diagrams, none is so difficult to interpret as the hazel: it differs very greatly between neighbouring series and the more closely samples are analysed the more intricate do its fluctuations become' (Godwin, 1940; 280). Even if the pollen can be accurately assigned to *Corylus avellana*, it cannot be assumed to be entirely of dryland origin (Long *et al.*, 1998c). If, as often is suggested, *Corylus avellana* formed a substantial part of the dryland forest understorey (e.g. Waller, 1994a), then this may not be represented in the pollen record, because hazel apparently flowers sparsely beneath a canopy layer (Godwin, 1975a; Rackham, 1986). *Corylus avellana* can also form part of the woodland canopy (Rodwell, 1991; Waller, 1994c) despite the fact that it is usually classified as a shrub in pollen diagrams.

Brown (1988) and Waller (1994c) suggest *Corylus avellana* is able to grow on the drier parts of valley fen alder woods (see Wheeler, 1980b; Rodwell, 1991). Scaife & Burrin (1987) interpret *Corylus avellana* as being present locally on damp floodplain carr, and Brown (1997) suggests that hazel is often co-dominant with *Alnus glutinosa*, with *Quercus* and *Fraxinus excelsior* also often derived from floodplain environments. Similarly, Devoy (1979; 1980) suggests *Quercus* and *Corylus avellana* occurred in the drier areas of fen in the inner Thames, which was in part supported by the occurrence of macrofossils. Therefore, it is likely that at least some of the pollen assigned to *Corylus avellana*-type and *Quercus* is derived from local wetland sources. Evidence from the Fenland suggests that *Quercus* is most likely to have occupied the floodplain edge areas (Waller, 1994c).

Shifting patterns of local wetland vegetation communities over time are an important factor in reconstructing fen communities from pollen diagrams (Waller, 1998; Long *et al.*, 1998c). Such a scenario is the probable reason for the increase in *Salix* pollen in the middle of HCP-2a so one should be cautious of over-interpreting palaeoecological records from such environments where changes in the pattern of plant distribution on the valley floor are often not reproducible between sites (Waller, 1998). *Salix* is dioecious (having the two sexes on different plants) and characterised by extensive hybidisation. Many of the hybrids produced are female varieties only (Stace, 1997), so in addition to being under-represented in the pollen record due to low pollen production (Bradshaw, 1981a), low abundance of *Salix* is compounded by the presence of female shrubs that produce no pollen. Therefore, a brief, discontinuous rise in the representation of *Salix* pollen, whilst may be interpreted as increased local abundance due to higher water tables (see below) or better light availability in appropriate cases (Smyth, 1986), in this context is most probably due to the changing composition of the fen community through time, with an increase in male willow trees in the vicinity.

HCP-2b 372.5 to 347.5 cm *c*.3300 to *c*.3000 BP (*c*.3500 to *c*.3200 cal. yrs BP)

A definite change in composition of the fen carr community occurs in HCP-2b as *Alnus glutinosa* pollen frequencies are reduced coincidental with a significant increase in *Salix* pollen. Pollen from aquatic species such as *Sparganium emersum*-type and *Typha latifolia* also occur with increased frequency during this zone, with *Filipendula* and Poaceae important herb species. The high *Salix* pollen values are particularly significant as it is usually considered to be heavily under-represented in the pollen record (see

Bradshaw, 1981a; Huntley & Birks, 1983; Janssen, 1984). Values as low as *c*.3 to 5% TLP have been taken to indicate the local presence of *Salix* (e.g. Long *et al.*, 1998c), while percentages of 10 to 20% probably indicates it was co-dominant with *Alnus glutinosa* (Waller, 1994c). This suggests the increase in *Salix* pollen percentages could be due to an increase in wetness which shifted the ecotone so that *Alnus glutinosa* was growing further away from the coring site along with sub-dominant constituents of the fen carr, such as *Fraxinus excelsior* and *Corylus avellana*-type which also decline at this time. Iremonger & Kelly (1988) provide experimental evidence to support Dupont (1987) who suggests that *Salix* spp. are the most tolerant wetland trees to saturated conditions. However, the extent to which changes in pollen abundance of taxa such as *Fraxinus excelsior* and *Corylus avellana* reflect extra-local changes in dryland populations is not known. Therefore, it is possible that this change was only of local significance, possibly caused by channel position adjustment of small drainage networks in the fen carr, which may shift a wetter ecotone closer to the coring site.

In addition to changes in the pattern of vegetation in the wetland community, apparent shifts in the dryland woodland may be suggested by the increase in pollen of *Quercus*, *Fraxinus excelsior,* and *Taxus baccata* in the middle of the zone, which has been dated to 3195±40 BP (3471 to 3346 cal yrs BP). *Salix* pollen percentages exhibit a sharp fall at this point and other herb species such as *Rubus* undif., *Plantago lanceolata* and *Typha latifolia* are registered in greater quantity. Cereal-type pollen grains are also recorded from these levels. This event is difficult to interpret due to the difficulty in separating relative dryland/wetland affinities for the taxa concerned. Whilst this may be because *Salix* is under-represented and therefore these taxa are proportionately increased, the effect of interdependence does not readily explain the appearance of cereal-type grains. A plausible explanation may be that the local environment was subjected to continued wetter conditions which drowned some of the *Salix* carr and promoted the development of a tall herb aquatic community. The presence of Cereal-type pollen grains (of *Hordeum* class size) in this context are likely to originate from certain genera of wetland grasses such as *Glyceria* or *Elymus* (members of which grow in fen and brackish environments). In this scenario, the increase in arboreal pollen types such as *Quercus*, *Fraxinus excelsior,* and *Taxus baccata* could be a product of reduced filtration of extra-local pollen surrounding the site subsequent to opening up of the willow community. However, this is only tentatively suggested as not only may *Quercus* and *Fraxinus excelsior* grow in damp areas, but *Taxus baccata* may also be found in fen woodland (Waller, 1994c). The slight decline in frequencies of *Alnus glutinosa* coincident with the decrease in *Salix* mid-subzone, suggests that some of the *Alnus glutinosa* tress remained as part of an *Alnus-Salix* carr and so its pollen may also have been reduced by the increased wetness. The average values for *Salix* of 11% within this zone support the suggestion of Waller (1994c) who interprets values between 10 and 20% as indicative of *Salix* being a substantial component of the *Alnus* carr.

There is no substantial evidence to imply the presence of human activities or cultivation in initiating these changes, despite their usual association with the occurrence of *Plantago lanceolata* and Cereal-type pollen. However, the fact the *Plantago lanceolata* increases, may be taken to suggest pastoral activities (*cf.* Iverson, 1941). *Plantago lanceolata* responds positively to trampling and grazing of animals by increased flowering, and may occur in wetland areas (Wheeler, 1980b) so small scale disturbance caused by browsing on the wet river marshes remains a possibility. Pollen percentages of *Salix* and *Alnus glutinosa* increase again towards the end of HCP-2b as *Quercus* percentages decline. This is suggestive of a re-expansion of wooded fen carr conditions locally which acts to filter out some of the extra-local *Quercus* pollen.

HCP-2c 347.5 to 312.5 cm *c.*3000 to *c.*2700 BP(*c.*3200 to *c.*2900 cal. yrs BP)

At the opening of HCP-2c, frequencies of *Alnus glutinosa* rise sharply (maximum of 93% TLP+*Alnus* at 345 cm), and replaces *Salix*. The 'spiky' nature of the *Alnus* curve is indicative of local dominance of this taxa throughout. *Fraxinus excelsior* pollen frequencies also fluctuate, and are particularly high either side of a large *Alnus glutinosa* macrofossil in the centre of this zone, which suggests continued local growth of *Fraxinus excelsior* and also perhaps *Betula*, within an *Alnus* dominated carr. *Pinus sylvestris* appears to be well represented within the dryland vegetation community during this zone. Towards the top of this lpaz, the frequencies of *Filipendula* begin a steady rise. This is interpreted below.

HCP-3 312.5 to 289 cm *c.*2700 to <*c.*2600 BP (*c.*2900 to <*c.*2600 cal. yrs BP)

The opening of lpaz HCP-3 is marked by an increase in *Filipendula* percentages and a fall in *Quercus* pollen, and is dated to 2665±45 BP (2850 to 2740 cal yrs BP). This is the suggests a second significant rise in the local water table, although this time it is likely to be related to more than just a local event. The wettest areas of the floodplain appear to become dominated by fen meadow species such as *Filipendula* (probably *Filipendula ulmaria* in this context) with *Caltha palustris*-type, Cyperaceae, and Poaceae (*cf.* Wheeler, 1980b). Such species are common in the wetter areas of floodplains and on river banks (Grime *et al.*, 1988) and may form a transitional community between the fen carr and river channel environment. The decline in *Quercus* pollen may be an indication of the presence of *Quercus* within the drier areas of the carr community (Wheeler, 1980b), which is one of the first trees to die due to its relative intolerance of waterlogged conditions compared to *Alnus glutinosa*. *Taxus baccata* and *Betula* pollen frequencies increase during this time which may be a result of these species invading the space left by *Quercus* in the drier part of the fen woodland or simply able to flower more profusely due to increased light in open areas created by the decline in *Quercus*.

HCP-4 289 to 281 cm <*c.*2600 BP (<*c.*2600 cal. yrs BP)

The transition to HCP-4 is characterised by marked shifts in vegetation suggesting a further rise in water table. Arboreal pollen declines appreciably, which may reflect the increasing distance between these taxa growing in the dryland community and the coring site as the floodplain expands. It is apparent that local conditions become too wet for *Alnus glutinosa*, which is replaced by *Salix* and dominates the local fen carr community. This suggests that the *Alnus glutinosa* pollen within this zone is increasingly from extra-local sources.

The fen meadow community also experiences changes in composition as values of *Filipendula* pollen drop away rapidly, and are replaced by pollen from Poaceae and Cyperaceae species. Significant increases in pollen of other herbs such as Apiaceae and *Plantago lanceolata* also occur. Members of the Apiaceae family have a wide ecological amplitude (e.g. *Peucedanum, Apium,* and *Oenanthe* commonly occur as part of the field layer in fen systems, Long *et al.*, 1998c), so this increase could either be indicative of open areas within the surrounding woodland, or from wetland types colonising the fen meadow. Spores of *Pteridium aquilinum* increase in abundance within this zone which may reflect opening up of the catchment vegetation. Spores of Pteropsida (monolete) indet. also rise, which could possibly be from ferns growing in the open areas of fen meadow in the transitional floodplain communities or attributable to increased spore production subsequent to the canopy opening of the wet woodland. These spores are very resistant to decay, and may also reflect increasing inwash. However, this is a less likely explanation in the absence of other pollen types associated with such as process (e.g. *Pinus sylvestris, Pteridium aquilinum*, pre-Quaternary spores and other exotic grains, see Chapter 9).

A broadly similar pattern of vegetation changes has been observed from similar floodplain deposits in the Combe Haven valley in East Sussex (Smyth, 1986; Smyth & Jennings, 1988). As the floodplain peat gradually gives way to a silty clay the biostratigraphy records a decrease in arboreal pollen percentages (e.g. *Quercus, Alnus glutinosa,* and *Corylus avellana*) and an increase in *Filipendula* and *Salix*. Smyth (1986) and Smyth & Jennings (1988) suggest several possible explanations in their interpretation of such a sequence of vegetation change.

Firstly, an increase in moisture is suggested due to a rising water table, which explains the nature of the vegetation change and corresponding rise in wetland herbs. Scaife (1987, in Balaam *et al.*, 1987) suggests *Salix* and *Filipendula* are indicative of areas of greater wetness, and locally, standing pools of water. Second, and the interpretation the authors favour, is that the changes observed can be explained as a response to increased light availability. However, it is apparent that this argument is circular. Both the litho- and biostratigraphical evidence suggest a significant rise in water levels (increasing inwash of minerogenic sediment and a rise in wet ground shrubs and herbs). The effect of this may be to flood the fen carr and drown many of the trees which would create open area, but for such a scenario to occur, a rise in water table is necessary. Devoy (1979) has obtained similar palynological results from Crossness in the inner Thames estuary where high values for *Salix* and *Filipendula* indicate rising water tables at the top of the peat layer.

Therefore, increasing water level is seen as the driving mechanism underlying the very similar vegetation changes observed at Hirst Courtney. The close of HCP-4 and the contact between the peat and overlying clay is dated to 2560±45 (2757 to 2480 cal yrs BP).

HCP-5 281 to 270 cm >*c.*2600 BP (> *c.*2600 cal. yrs BP)

Further reductions in arboreal pollen are evident during the opening of lpaz HCP-5 (which marks the end of peat deposition and the opening of the clay phase) as conditions locally become too wet for the fen woodland. The low input of tree pollen thereafter, may suggest that the catchment was largely cleared of trees by this time. Poaceae and Cyperaceae are the most important pollen types indicating that *Alnus glutinosa* carr is replaced by herbaceous grass and sedge dominated communities. Pollen of aquatic plants also become frequent in HCP-5 with *Sparganium erectum* the dominant species. This species occurs in the quiet backwaters of river environments in water depths up to 1 m (Cook, 1962; Haslam *et al*., 1975). A very similar sequence of vegetation changes have been recorded by Dinnin & Brayshay (1999) in the lower Trent valley and have been dated to 2690±100 BP (2985 to 2496 cal. yrs BP). This is attributed to flooding due to extensive clearance of catchment trees although this feature may be interpreted in a different manner (see section 10.2.1 and 11.1) in the light of these results and those of Long *et al.* (1998a). This period coincides with the culmination of a well documented phase of estuarine expansion throughout the Humber, which can also be invoked to explain these changes.

Frequencies of *Sparganium erectum* decline as percentages of Cyperaceae and Poaceae increase during HCP-5. As water levels rise, the deposition of clastic sediments continues and it is likely that flooding from tidal inundation becomes more regular (see section 7.6). The fall in *Sparganium erectum* percentages may therefore be a reflection of the intolerance of this species to wave action or the fact that it is incapable of competing effectively with the taller *Phragmites australis* (Cook, 1962; Grime *et al*., 1988), which is a possible source of much of the Poaceae pollen in this case.

Other herb taxa increase in abundance during lpaz HCP-5 such as *Filipendula*, Apiaceae, *Plantago lanceolata*, Rubiaceae, Asteraceae (Lactuceae), Asteraceae (Cardueae/Asteroideae) and cereal-type pollen grains. A range of different habitats support plant species within these families but within this context, the most likely is within fen meadow (Wheeler, 1980b). The increase in cereal-type pollen is probably attributable to grass species such as *Glyceria,* which can occur on waterlogged alluvial soils next to slow flowing water (Lambert, 1947) in association with Cyperaceae and *Phragmites australis* (Wheeler, 1980a; Rodwell, 1995). *Plantago lanceolata* is not exclusively a dryland herb of disturbed ground and occurs locally in permanently saturated ground (Grime *et al*., 1988) and fen meadows environments (Wheeler, 1980b; Wiegers, 1992). The presence of Chenopodiaceae with Asteraceae (Cardueae/Asteroideae) type pollen may indicate slightly saline conditions on the river marshes.

A prominent feature of the upper levels is the out of phase relationship between Poaceae and Cyperaceae. Similar variations between the relative representation of these two families are apparent in studies by Smyth (1986), Orson *et al.* (1990; 1992), and Hewlett & Birnie (1996) which may relate to fluctuating water levels within tidal floodplain environments. Work on the freshwater marshes of the inner Delaware estuary (Orson *et al.*, 1990; 1992) suggests that Poaceae and Cyperaceae pollen are out of phase; Poaceae pollen percentages rise and Cyperaceae values fall during the onset of tidal conditions. However, Wheeler (1980a) mentions that members of the Cyperaceae family such as *Carex* sp. are particularly associated with a strongly fluctuating water table. It is probably not possible to fully explain this matter palynologically because species and even genus specific identification is not possible with these families. Further comments regarding the competitive relationship between these two taxa is not appropriate here, as other factors may be important such as salinity. It equally likely that the spiky nature of the pollen curves of Cyperaceae and Poaceae during HCP-5 reflects the variable local deposition of pollen within a mixed reed community (e.g. a sub-community of a *Phragmites australis* reedswamp with locally abundant Cyperaceae, Rodwell, 1995).

The development of these types of late-Holocene tidal freshwater marshes have been extensively studied in relation to sea-level rise in the inner Delaware estuary (e.g. Pizzuto & Rogers, 1992; Fletcher *et al.*, 1993a; 1993b; John & Pizzuto, 1995). Transitional environments between estuarine saltmarshes and non-tidal inland river marshes where "tidal freshwater marshes develop at the head of a transgressed coastal plain" are termed **palustrine** marshes (Fletcher *et al.*, 1993a; 1810).

7.6 Palaeohydrological Interpretation

HCD-1 280 to 276 cm <2560±45 (<2757 to 2480 cal. yrs BP)

Figures 7.5, 7.6, and 7.7

The diatom record begins with the deposition of the silty clay above the peat at 281 cm so post-dates 2560±45 BP (2757 to 2480 cal. yrs BP). However, the lowest two diatom zones (HCD-1 and 2) overlap with the final zone (HCP-5) of the pollen diagram. The diatoms present within HCD-1 are dominated by freshwater (salt-intolerant) species of the *Pinnularia* genus. *Pinnularia* sp. diatoms are common in contemporary peat environments but are often fragmented by diagenetic processes such as dissolution in fossil sediments (Vos & de Wolf 1988; 1994); the relatively dry conditions prevent good preservation (Denys, 1994). Consequently, the strongly silicified raphe structures of these diatoms are the only parts of the valve preserved in such situations. The poor diatom preservation in this ldaz and presence of frequent raphe structures of *Pinnularia* sp. in ldaz HCD-1, perhaps suggests the occurrence of fen peat environments close by (*cf.* Denys, 1994).

Generally, *Pinnularia* sp. are indicative of eutrophic conditions (Robinson, 1993) although some (e.g. *Pinnularia nobilis*) relate to more oligotrophic conditions. Although the life-form ecology of unidentified *Pinnularia* sp. cannot be reliably determined, *Pinnularia abaujensis* and *Pinnularia nobilis* are commonly occurring benthic species in HCD-1 (Denys, 1991; de Wolf, 1993) which dwell *in situ* within the sediments with other species such as *Stauroneis phoenicenteron* and therefore indicate transitional freshwater fen environments in the vicinity and exposed muddy substrates locally (Denys, 1994). The presence of oligotrophic species such as *Pinnularia nobilis* may be the result of local acidification due to leaching. The epiphytic group is also well represented which is indicative of diatom species growing attached to plants on the river marsh. Aerophilous taxa such as *Achnanthes lanceolata* indicate a certain degree of subaerial exposure which suggests the site was perhaps not permanently submerged. However, other authors suggests this species is epiphytic (Denys, 1991; Vos & de Wolf, 1993a; 1997) which leads to an uncertain interpretation. De Wolf (1993) also classifies *Achnanthes lanceolata* as rheophilous (i.e. requires running water), this may indicate some degree of fluvial input across the area, so if this species is epiphytic, it may be allochthonous.

Although it is often difficult to separate the autochthonous from the allochthonous component in diatom assemblages (*cf.* Beyens & Denys, 1982; Vos & de Wolf, 1988), here the evidence suggests the local environment was a predominantly freshwater muddy river marsh subjected to flooding by tides. The presence of patches of vegetation can be inferred from the occurrence of epiphytic species and the abundance of grass and sedge pollen in lpaz HCP-5, although the occurrence of benthic epipelic species such as *Gyrosigma acuminatum* (Vos & de Wolf, 1993a) suggest exposed clayey substrates were abundant within the marsh. The prevalence of *Pinnularia* sp. indicates the proximity of fen peat environments probably on the higher areas not regularly flooded or peripheral to the site.

HCD-2 276 to 269 cm

The change in diatom assemblage during HCD-2 indicates a greater estuarine influence at the site, although interpreting the diatom taphonomy is complicated by the range of different taxa recorded and perhaps the small number of samples. Marine and brackish planktonic diatoms, such as *Paralia sulcata* and *Cyclotella striata*, increase in abundance which suggests the expansion of saline conditions up the river Aire valley at this time. Whilst these taxa are indicative of the allochthonous component (*cf.* Vos & de Wolf, 1988) carried up the river by the tide, the occurrence of marine and brackish benthic species such as *Nitzschia granulata* and *Nitzschia navicularis,* and a reduction in epiphytic forms, may suggest the development of saline tidal flats locally. Both these species have an epipelic life form and migrate freely within muddy brackish sediments (Vos & de Wolf, 1993a). These species may be part of the autochthonous component colonising bare substrates close to tidal creeks where slightly saline water may flow, although it is also possible for them to have been washed up from more estuarine locations down river. Brackish and freshwater aerophiles are also prevalent (e.g. *Diploneis interrupta* and *Achnanthes lanceolata*). Although Denys (1991) and de Wolf (1993) classify *Diploneis interrupta* as benthic, Vos & de Wolf (1988) state that this species is a benthic aerophile that occurs in the supratidal zone. In this context, it can be interpreted as occurring in the drier muddy marsh areas, with perhaps *Achnanthes lanceolata*, where freshwater input is high and flooding less frequent. It is interesting to note here that the occurrence of marine and brackish diatom species and estuarine tidal flooding coincides with the decline in *Sparganium erectum* in HCP-5 which was tentatively associated with an increase in tidal/wave activity.

HCD-3 269 to 245 cm

HCD-3 is dominated by the freshwater epiphytic species *Gomphonema angustatum* (Vos & de Wolf, 1993) which indicates abundant aquatic and wetland plants on the river marsh. Estuarine influence is diminished in this zone as planktonic marine species (e.g. *Paralia sulcata*) are reduced in abundance and the frequency of brackish species such as *Nitzschia granulata* and *Nitzschia navicularis* decline. However, it is unclear whether this reduction in estuarine conditions reflects an actual drop in water levels or simply a relative shallowing due to the re-colonisation of bare areas of mudflat with herbaceous vegetation and an increase in sedimentation rate (or both).

HCD-4 245 to 215 cm

The boundary between ldaz HCD-3 and 4 marks a shift in salinity status of the local environment as halophobous taxa increase in abundance. Epiphytic diatom taxa are still an important group with *Eunotia diodon* the main species. *Gomphonema angustatum* is classified as oligohalobous-indifferent (Hustedt, 1957). This group consist of freshwater species that can tolerate a certain degree of salinity. Vos & de Wolf (1997) suggest that *Gomphonema angustatum* commonly occurs in eutrophic, perhaps permanently submerged environments with low water velocities. *Eunotia diodon* is an epiphytic halophobe, and cannot withstand any degree of saline conditions, which indicates the environment has become fresher. This is also indicated by the rarity of polyhalobous and mesohalobous diatoms and planktonic taxa within this ldaz. The rise in aerophilous diatoms coincident with these changes may be indicative of falling water levels and a reduction in tidal activity.

Diatom taxa such as *Gomphonema* spp. and *Eunotia* spp. have also been reported from clay deposits in freshwater tidal marsh areas in the inner Severn estuary (Hewlett & Birnie, 1996), and in the Delaware estuary (Fletcher *et al.*, 1993a; John & Pizzuto, 1992).

HCD-5 215 to 170 cm

Drying is indicated in ldaz HCD-5 by an increase in the abundance of aerophilous taxa (e.g. *Eunotia tenella* and *Eunotia exigua*). Generally, the *Eunotia* genus is indicative of fresh oligotrophic water and live in areas that are not permanently submerged (Van Dam *et al.*, 1994). This indicates that large parts of the marsh are only inundated by floods and the extreme highest parts of the tide. Leaching associated with lower water tables may result in some acidification of the local environment. Oligohalobous and halophobous epiphytic taxa such as *Cocconeis placentula*, *Cymbella ventricosa*, and *Eunotia diodon* are also important components of the assemblage. Together, this indicates the development of a freshwater river marsh, with diatoms occurring on exposed damp substrates and attached to plants.

7.7 Hirst Courtney Palaeoenvironmental Summary

The inferred palaeoecology of the local peat forming environment at Hirst Courtney is summarised is Figure 7.8. After an initial spike in Cyperaceae in HCP-1, perhaps indicating herbaceous marsh communities growing along the fringes of the river, the floodplain is dominated by *Alnus* carr vegetation. It has been suggested that the persistence of *Alnus* carr in perimarine areas is indicative of a continually rising water table (Kidson & Heyworth, 1973; Hewlett & Birnie, 1996).

Expanding wetland conditions is the suggested cause for a decline in *Tilia* which may have been the dominant vegetation at the floodplain edge and on raised terrace or levee features close to the site. Closed *Alnus* carr conditions in lpaz HCP-2a acts as a buffer which dilutes pollen from surrounding dryland vegetation. This may have resulted in reduced pollen representation of taxa such as *Pinus sylvestris* which is possibly derived from regional sources and transported through the canopy component (*cf.* Tauber, 1965). Although it is not possible to state precisely the composition of the extra-local dryland vegetation, the likely constituents are *Quercus* and *Corylus avellana*, with lesser amounts of *Ulmus*, *Fraxinus excelsior*, and *Tilia*.

Two phases of increased wetness are evident. During lpaz HCP-2b, *Salix* takes the place of *Alnus glutinosa* as the fen carr opens up, accompanied by a range of wetland herbs. The evidence suggests this phase is temporary, as *Alnus glutinosa* communities increase in HCP-2c and shade out the *Salix* scrub. The implication is that this change was perhaps only of local significance, possibly caused by shifting channel position within the carr drainage network, which temporarily favoured the growth of *Salix*. During, lpaz HCP-3 and 4 there is a more sustained rise in water table. This first affects the composition of the floodplain vegetation communities, promoting open fen meadow and *Salix* carr conditions, and then the sedimentary regime, as the site becomes flooded with clay and silt laden river water.

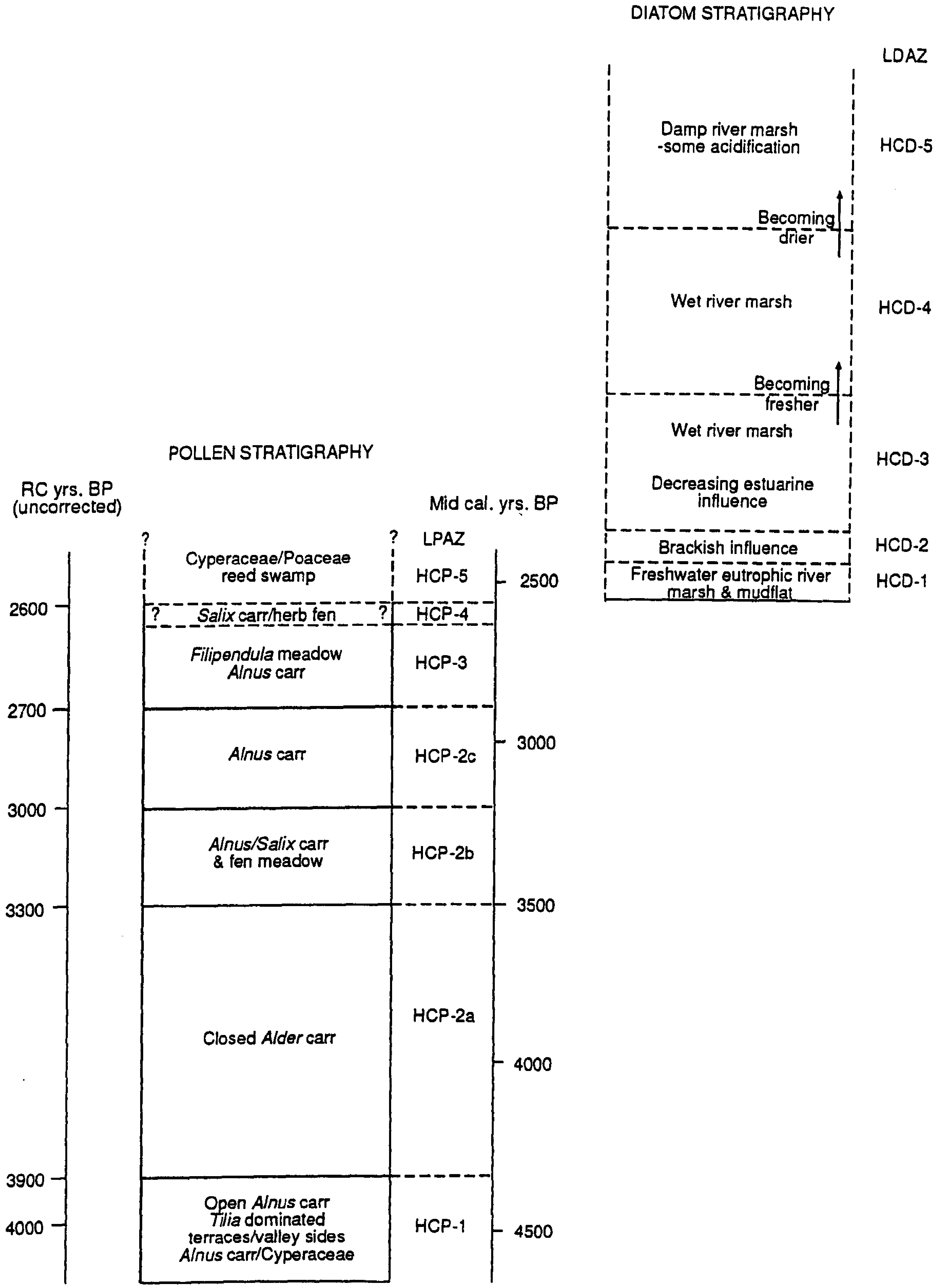


Figure 7.8 Hirst Courtney palaeoenvironmental summary diagram

The sedimentary contact at 281 cm is dated to 2560±BP (2757 to 2480 cal yrs BP) and marks the end of organic sediment deposition and the beginning of the diatom record. The inferred palaeoenvironmental reconstruction from these clastic sediments is given in Figure 7.8. Locally, the river marsh supports a wide range of autochthonous diatom communities including epipelic varieties dwelling the exposed substrate and epiphytic forms attached to vegetation forming a marsh. Initially, freshwater eutrophic conditions are recorded and pollen and diatom evidence suggests a patchy river marsh environment with grasses and sedges and bare areas of mudflat.

An increase in estuarine conditions is apparent during ldazs HCD-2 and 3. This brackish influence is short lived, and possibly represents the expansion of the marine limit of the Humber estuary (*cf.* Long *et al.*, 1998a). Salt-intolerant diatoms replace the brackish species within HCD-3 and 4 and the taxa are indicative of a vegetated river marsh with sub-aerially exposed muddy areas. Progressively more drier conditions are suggested within HCD-4 and 5, which may represent a drop in the frequency of flooding and tidal inundation, possibly related to a reduction in the rate of sea-level rise or a relative increase in sedimentation rates. Drier and perhaps leached conditions are indicated in the latter part of the diagram as both eu- and oligotrophic diatom varieties are abundant.

Chapter 8 Eskamhorn Site Report

8.1 Introduction

Eskamhorn is situated within the central area of the study reach (see Figure 3.1). In this chapter, the results of a transect of boreholes sunk across the river Aire floodplain near Eskamhorn Farm (see Figure 6.3) are presented (Figure 8.1 and 8.2). These deposits are characteristic of the sedimentary association uncovered within the perimarine environment of the lower Aire valley (see section 6.4). A sample core taken from EK 12P on the north side of the floodplain is described in detail (Table 8.1). Palaeoecological investigations, based on pollen, diatom and wood macrofossil analyses from this core, are used to elucidate the sequence of environmental changes at Eskamhorn. A chronology is provided by a series of radiocarbon dates. The results of the microfossil analyses are presented in section 8.4, and a palaeoenvironmental interpretation is given in section 8.5 and 8.6.

8.2 Lithostratigraphy

A summary of the stratigraphic results is provided in section 6.2.2 and Figure 6.4. In this chapter the details of the lithological findings are presented in two parts. A total of 27 boreholes, which straddle the river Aire floodplain at Eskamhorn, are recorded (see Figure 6.3). A transect of 18 cores from the north side are plotted in the Eskamhorn borehole transect (Figure 8.1) and a further 9 boreholes from the south side of the floodplain near East Cowick are shown in Figure 8.2. A simplified schematic summary of the lithostratigraphy is presented in Figure 6.4.

It is evident from Figures 8.1, 8.2, and 6.4, that the valley form is asymmetric, being deeper and hence the sedimentary fill thicker on the north side of the valley (reaching a maximum recorded depth of 914 cm in EK 13) than the south. Rivers commonly display a preferred position within the floodplain (Brown, 1996), which is one possible cause of this floodplain and valley cross sectional asymmetry (Palmquist, 1975).

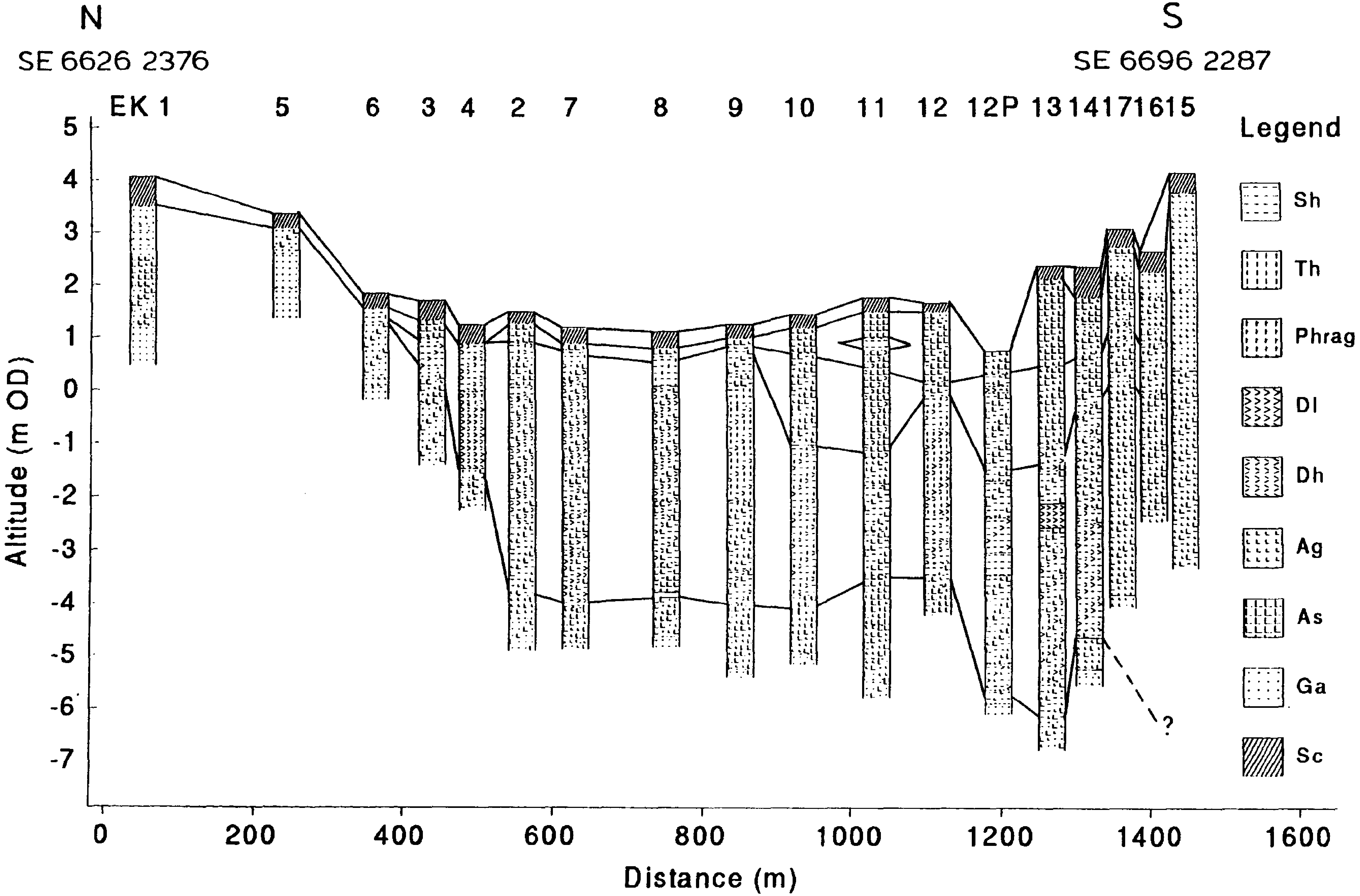


Figure 8.1 Lithostratigraphy of the cross valley transect at Eskamhorn in the lower Aire valley (symbols follow Troels-Smith (1955))

The lowest unit recorded in the sedimentary fill of the north Aire valley at Eskamhorn (Figure 8.1) is a variable minerogenic facies, comprising mainly stiff grey silty, sandy clay, or sand in places. A trace of humified organic detritus is present in the upper portion of this unit along with occasional wood remains and roots. The upper surface of this unit undulates, and is cut into by a deeper channel-like feature between boreholes EK 12 and 14. The basal sediments within this depression are grey consolidated sands with a minor clay component, and in EK 13 are recorded to a depth of -6.68 m OD. In boreholes EK 1 and 5, this basal sediment unit outcrops at the surface and represents the limit of the Holocene floodplain wetland community.

Overlying the basal unit between boreholes EK 6 and 14 is a thick grey-brown clayey wood peat. The contact between the basal sediments and the organic unit rises from an altitude of -6.24 m OD in EK 13 in the central part of the valley, up to +1.29 m OD in EK 6 at the margin of the floodplain. The peat also thins from EK 13, where it is 481 cm thick, to EK 6 where it is a mere 18 cm thick. The organic components include abundant large wood branches, twig detritus, and *turfa* rootlets, with occasional monocotyledonous stems. The peat matrix is a heterogeneous mixture of *Substantia humosa* (*sensu* Troels-Smith, 1955) and silty clay. Spatially discontinuous clay lenses are also common. The inorganic fraction within the peat is often high, but also variable. This is a feature of most boreholes in this study as well as other perimarine fluviatile areas, and is the result of river inwashing (Devoy, 1979; Pons, 1992; Waller, 1993; 1994a).

Between boreholes EK 10 to 14, the peat gives way to a blue-grey-brown organic silty clay. This unit also contains woody detritus, herbaceous rootlets and monocotyledonous stems (*cf. Phragmites*), which are particularly common. Although the transition from the underlying wood peat is recorded as gradational, the altitude of the change in stratigraphy varies between -1.73 m OD in EK 12P and -0.12 m OD in the adjacent core EK 12. The lack of convincing evidence for abrupt sedimentary contacts suggests this is probably the result of differing compaction associated with the weight of the sedimentary overburden and de-watering, as opposed to erosion of the surface of the peat deposit by channels during the deposition of the clays. A lithologically similar horizon is also recorded between *c*.0 m OD and *c*.+1 m OD in boreholes EK 16 and 17, situated close to the present river channel, which overlies a different suite of sediments (see below).

The blue-grey-brown organic silty clay unit is replaced by a grey-orange-brown silty clay, which occurs between +0.15 m OD in EK 12 and +3.74 m OD in EK 15. This clastic layer directly overlies the main peat unit in boreholes EK 2, 3, 6, 7, 8, and 9, but is absent from EK 4 where the peat is present just below the top soil. The sediments are heavily stained by oxidation mottles which are concentrated around cracks, root channels, and woody organic detritus. This unit contains few organic remains except rootlets (many of which are modern) and more resistant wood timbers, which may reflect oxidation; the sediments are above the recorded water table height and become increasingly dessicated towards the surface. This horizon also contains an increasing sand component in boreholes situated closer to the river from borehole EK 12 towards the EK 15. This may represent fairly recent overbank deposits, which characteristically fine away from their source (Nanson & Young, 1981; Pizzuto, 1987; Reid & Frostick, 1994).

A thin, well humified and dessicated clayey peat layer is recorded within the grey-orange-brown silty clay unit in borehole EK 11. This oxidised organic unit is atypical of the surrounding stratigraphy at Eskamhorn but similar peaty horizons are recorded within the upper mineral sediments along the long profile transect (see section 6.3) and also towards the top of core EC 12 on the East Cowick transect (Figure 8.2).

The stratigraphy recorded in boreholes 15, 16, and 17 is dissimilar to the description of the majority of the floodplain fill outlined above. The lowest sediments in these boreholes (recorded down to -4.03 m OD in EK 17) are characterised by alternating bands of light and dark grey finely laminated silty clays, separated with sandy partings. Patches of eroded peat and detrital woody remains are occasionally recorded and reed stems (*cf. Phragmites*) increase in abundance towards the top of the unit. Such sequences are typical of depositional conditions within a tidal channel. In EK 16 and 17 these deposits are overlain by blue-grey-brown silty clays containing frequent reed stem remains. The uppermost unit in cores 15, 16 and 17 is an orange-brown-grey silty clay (becoming more sandy towards the surface), similar to other near surface deposits throughout the study area.

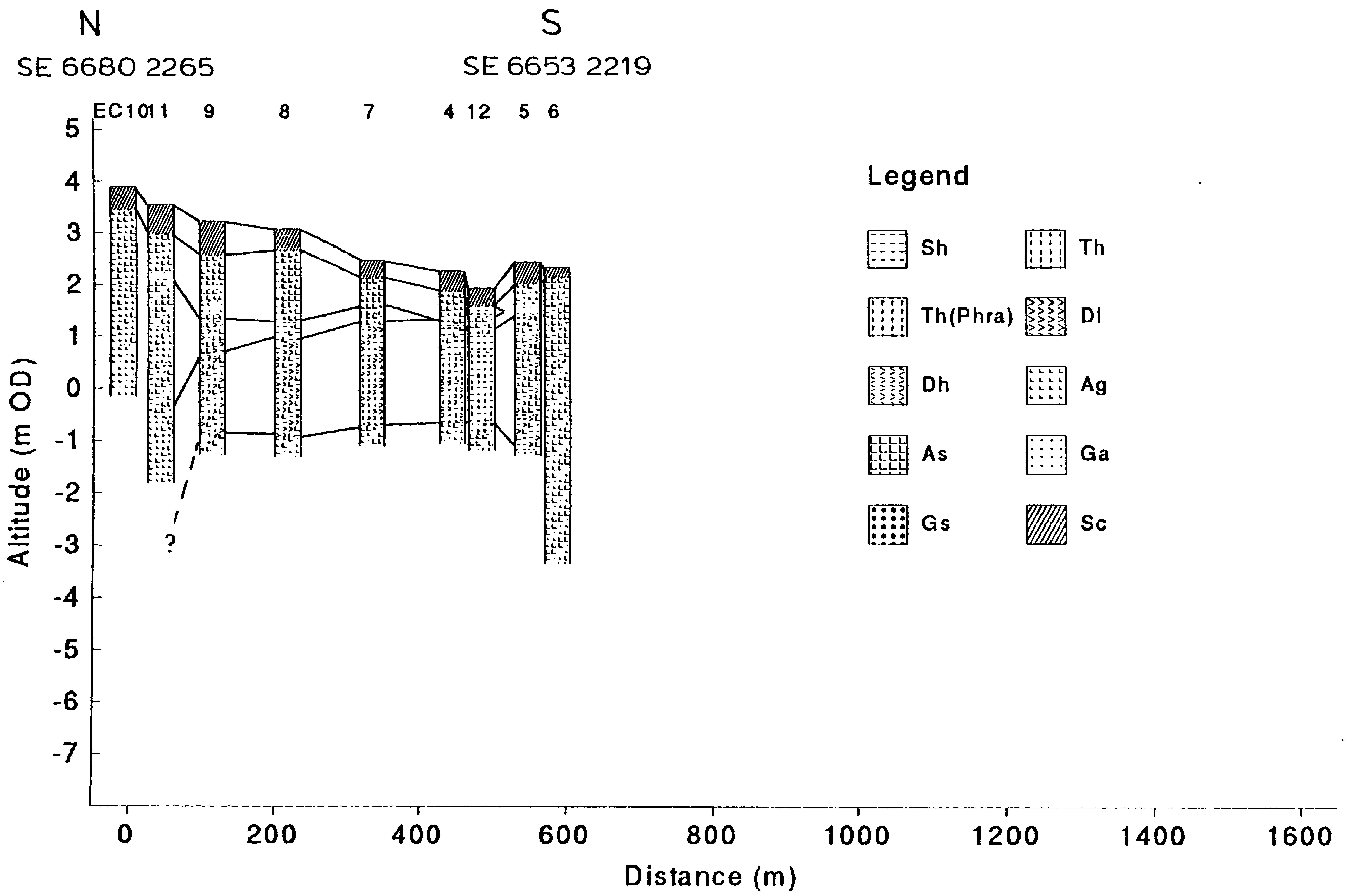


Figure 8.2 Lithostratigraphy of the cross valley transect at East Cowick in the lower Aire valley (symbols follow Troels-Smith (1955))

The sedimentary sequence on the south side of the floodplain near East Cowick (Figure 8.2) is similar to Eskamhorn Farm. However, the boreholes are shallower because the basal unit is altitudinally higher due to valley asymmetry. The basal sediments are recorded in boreholes EC 9, 8, 7, 4, 12, and 5, and consist of stiff pink-blue-grey sandy clay or clayey sand, although some coarse sand is recorded in EC 9 and 12. This unit is overlain by a brown wood peat with a clayey matrix. This organic unit is mainly detrital in nature although *turfa* rootlets and herbaceous stems are recorded with frequent clayey horizons. The peat is recorded between cores EC 9 and 5 but reaches the greatest thickness of 255 cm (between +1.32 and -1.23 m OD) in EC 5. The upper contact of the peat at East Cowick shows signs of burning in all cores, with charred wood remains and black charcoal fragments abundant in the upper layers.

The peat is overlain by a blue-grey brown organic clay between +0.58 m OD in EC 11, and +1.17 m OD in EC 7. This organic clay is replaced by an orange-brown-grey silty clay with oxidation mottles, occasional sandy horizons, and few organic remains apart from rootlets. This unit directly overlies the peat in boreholes EC 4, 12 and 5. These clastic units have the same characteristics as the upper clastic horizons along the Eskamhorn transect. Separating the top soil and the mottled silty clay in EC 12 (between +1.38 to +1.50 m OD) is a dark brown well humified dessicated peat, comparable to that recorded from a similar altitude in EK 11.

Boreholes adjacent to the present river Aire channel (EC 10 and 11) record a similar stratigraphy to cores EK 15, 16, and 17, with a sequence of laminated clay, silt, and sand overlain by the uppermost oxidised clays. At the other end of the transect, a similar sequence of channel deposits is evident which are recorded down to -3.35 m OD in EC 6. These deposits probably relate to an artificial channel of the river Don that was diverted into the River Aire in the vicinity as part of the modern drainage scheme in the Humberhead Levels (Gaunt, 1975).

A piston core was taken from a location representative of the valley stratigraphy at Eskamhorn. Core EK 12P was selected, from between EK 12 and 13, because it contained a thick peat sequence and gradational boundaries between the adjacent sediment units. The detailed lithological description of this core is presented in Table 8.1.

Site: Eskamhorn, River Aire valley	**Core**: EK 12P

Grid reference: SE 6680 2310	**Date of fieldwork**: August 1996
Latitude: 53° 42' N	**Sampling method**: Stitz Piston Corer
Longitude: 0° 59' W (Greenwich Meridian)	**Bench mark datum**: Rawcliffe Church
Collector: J. R. Kirby / D. M. Taylor	**Ground Altitude**: +1.72 m OD

Observed depth (cm)	**Altitude (m OD)**	**Troels-Smith Sediment Description** **nig., strf., elas., sic., lim. sup., (Troels-Smith, 1955)**
100 to 143	+0.72 to +0.29	Orange-brown-grey heavily oxidised silty clay. Trace of woody detritus. As3, Ag1, Lf++, Sh+, Dh+, Dl+; 2, 0, 2, 0, -.
143 to 166	+0.29 to +0.06	Transitional layer of organic silty clay with wood and monocotyledonous detritus. As1+, Ag1, Sh1+, Dh+, Dl+; 2, 0, 2, 0, 0.
166 to 345	+0.06 to -1.73	Brown-grey organic clay with abundant woody detritus and occasional rootlets and reed (*cf. Phragmites*) stems. Faint clay bands apparent towards base of unit. As2+, Sh2, Ag+, Dl+, Dh+, Th^{2}+, Th$^{Phrag.}$+; 2++, 0, 2, 0, 0.
345 to 520	-1.73 to -3.48	Brown clayey detrital wood peat with turfa rootlets and occasional monocotyledonous (*cf. Phragmites*) stems. Large wood branches common, and variable clay content. Sh2, Dl1, As1, Ag+, Dh++, Th^{2}+; 3, 0, 2, 0, 0.
520 to 715	-3.48 to -5.43	Variable grey-brown organic silty clay and clayey peat. Abundant woody detritus and sand trace in places. Sh2, As1, Ag1, Dl++, Ga+, Dh+; 3, 0, 2, 0, 0.
715 to 762	-5.43 to -5.90	Grey-brown silty peat with abundant woody detritus becoming sandy towards the base. Sh2, Ag1, Ga1, As++, Dl++; 2, 0, 2, 0, 0.
762 to 780	-5.90 to -6.08	Grey clayey sand with woody detritus and turfa. Occasional chalky blue flecks of authigenic vivianite (e.g. Bennett, 1983b; Denys, 1985). Ga4, As++, Sh+, Dl+, Tl^{2}+; 3, 0, 2, 0, 0.

Table 8.1 Lithostratigraphic details from EK 12P described using the Troels-Smith (1955) terminology.

8.3 Radiocarbon Dating

Five bulk samples of organic sediment were sent to East Kilbride for conventional radiocarbon dating. Eskamhorn was the first site dated in this study and initial samples were selected as 'range finders' to address the possible effect of old carbon in the samples. Although no macroscopic coal fragments were recorded (*contra* Bayliss *et al.*, 1999), microscopic pre-Quaternary spores were recorded on the pollen slides (see Figure 8.4). Pre-Quaternary carboniferous material is a potential contaminant which may cause a possible ageing effect (see section 5.9.2). Therefore, these samples were chosen to assess the reliability of bulk sediment dates from such contexts. Specific chronohorizons have been dated, such as the mid-Holocene decline in *Ulmus pollen* which is a well documented, broadly synchronous time-marker throughout the British Isles (Smith & Pilcher, 1973; Huntley & Birks, 1983). This can be used as a dating control, against which any significant age offset, which may be attributable to old carbon, may be recognised. An additional test is provided by a date from a part of the core where evidence for old carbon was not recorded on the pollen slides.

The results of the dating program are presented in Table 8.2. The ages are all in stratigraphic order (which may not be expected if some of the samples were adversely affected by old carbon) and consistent with the anticipated age associated with the chronohorizons This suggests that the proportion of old carbon in the form of pre-Quaternary microfossils is negligible, and the use of conventional bulk sample dating is reliable from these floodplain contexts. Törnqvist *et al.* (1992) also found that dates from samples of clayey *Alnus* peat were comparable to AMS dates on coexisting macrofossils, suggesting that such dates, although not recommended when samples contain a low *in situ* carbon content (e.g. Olsson, 1979), are still useful.

Laboratory Code	Depth (cm)	Altitude (m OD)	Age (^{14}C yr BP±1σ)	Calibrated Age Range * (cal. yrs BP±2σ)			Mid Cal. Age (cal. yrs BP±1σ)
				max	(cal. ages)	min	
SRR-6218	355 to 360	-1.83 to -1.88	3440±45	3828	(3689)	3572	3700±128
SRR-6219	380 to 385	-2.08 to -2.13	3670±45	4130	(3980, 3938, 3940)	3852	3991±139
SRR-6220	450 to 455	-2.78 to -2.83	4220±45	4860	(4826, 4740, 4738)	4573	4717±143
SRR-6221	621 to 626	-4.49 to -4.54	5440±45	6304	(6276, 6224, 6216)	6216	6260±44
SRR-6222	745 to 750	-5.73 to -5.78	6385±45	7380	(7259)	7202	7291±89

* Calibrated using Stuiver & Reimer (1993)

Table 8.2 Radiocarbon dates and calibrated age estimates from Eskamhorn

8.3.1 Age-Depth Model

The two age series (uncorrected and calibrated dates) are plotted on an age-depth model in Figure 8.3 which is used as a means of estimating the age ranges of the biostratigraphic zones using linear interpolation. Calibrated ages are plotted using the mid point of the calibrated age (after Bennett, 1994b) using the method outlined in section 5.9.4. Calculation of the average sedimentation rate between the ages (cm yr^{-1}), enables the deposition time for the lpazs to be inferred. Error bars for the uncalibrated BP ages show the stated laboratory counting error. The depth error bars on both sets of data account for the width of sediment sampled for dating from the core.

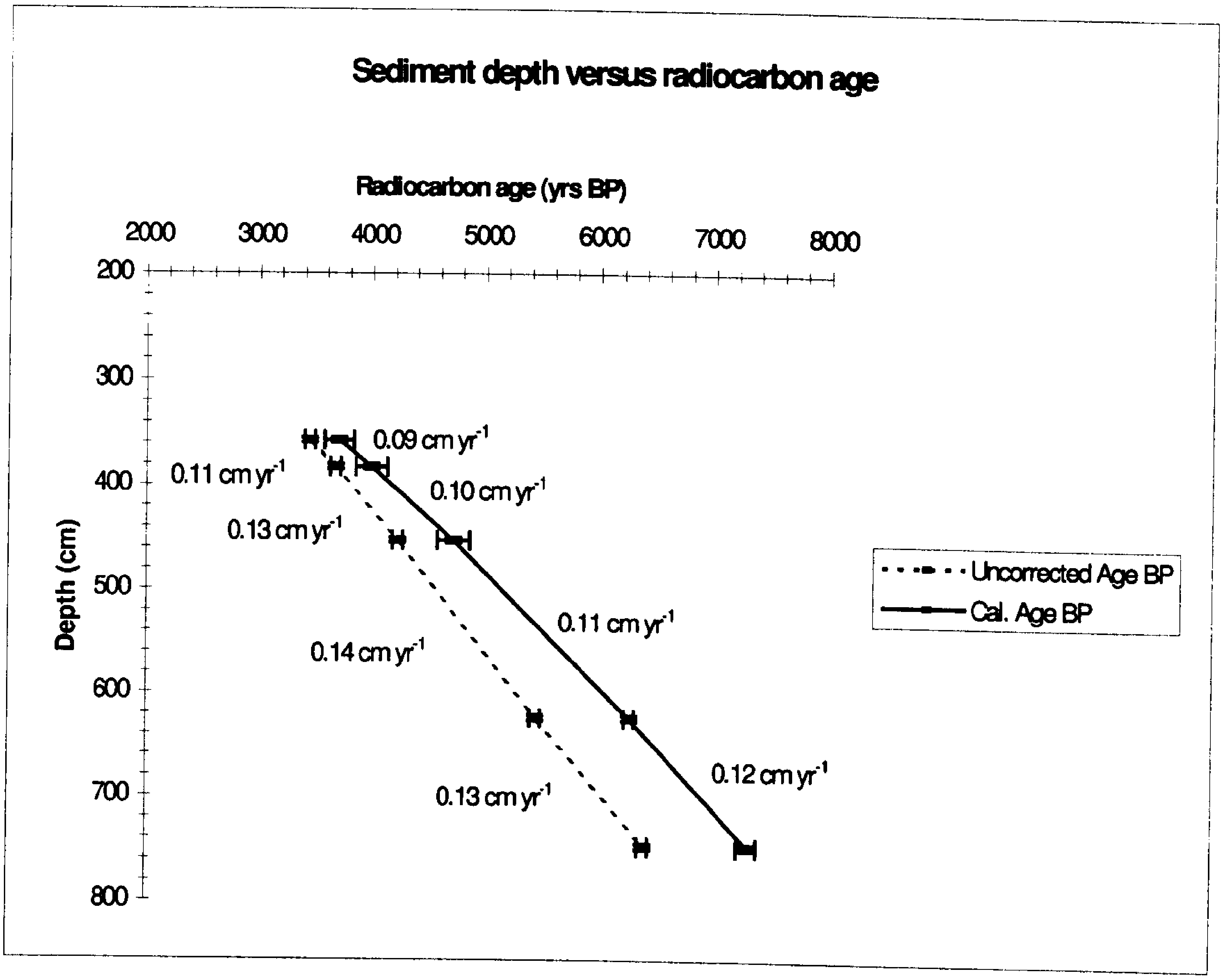


Figure 8.3 Eskamhorn Age-Depth Model (see section 5.9.4 and 8.3.1 for details)

8.4 Biostratigraphy

7.4.1 Macrofossils

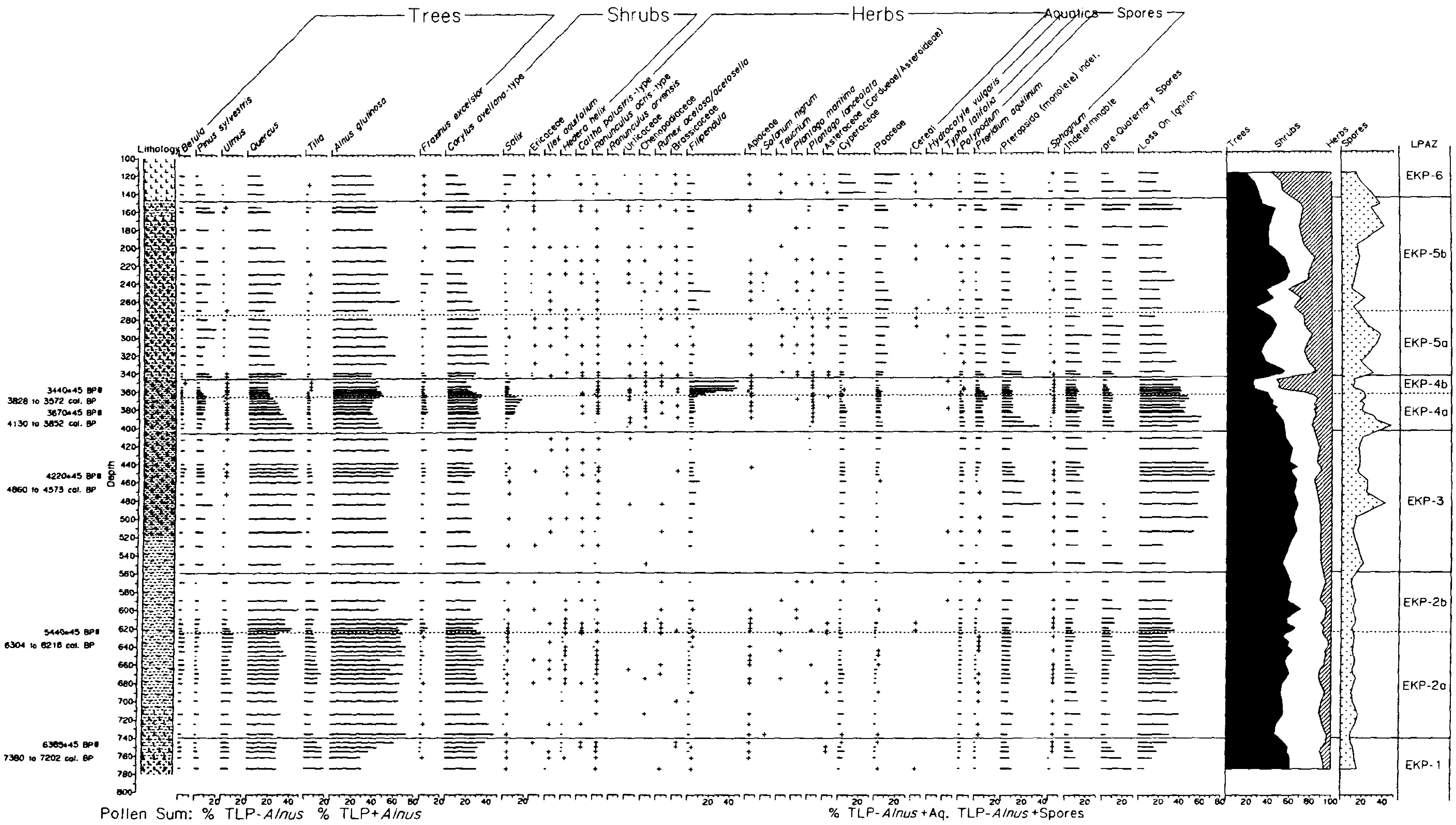
Wood macrofossils have been identified from core EK 12P using the thin sectioning technique outlined in section 5.7. *Corylus avellana* nuts were also recorded from within the sediments and are listed with the wood macrofossils in Table 8.3.

Depth Range (cm)	Macrofossils	Depth Range (cm)	Macrofossils
171 to 175	*Alnus glutinosa*	502 to 505	*Alnus glutinosa*
180 to 185	*Alnus glutinosa*	510 to 512	*Alnus glutinosa*
349 to 350	*Alnus glutinosa*	550 to 551	*Ulmus*
359 to 360	*Alnus glutinosa*	572 to 575	*Alnus glutinosa*
360 to 364	*Salix* sp.	672 to 674	*Ulmus*
400	*Corylus avellana* nut	683 to 685	*Fraxinus excelsior*
415 to 420	*Alnus glutinosa*	709 to 710	*Corylus avellana*
425 to 428	*Alnus glutinosa*	715	*Corylus avellana* nut
455 to 457	*Alnus glutinosa*	715 to 726	*Quercus*
466 to 478	*Fraxinus excelsior*	727 to 734	*Ulmus*
489 to 490	*Alnus glutinosa*	748 to 749	*Ulmus*
495 to 502	*Alnus glutinosa*	754 to 755	*Alnus glutinosa*

Table 8.3 Wood macrofossils from core EK 12P

8.4.2 Pollen Stratigraphy

The results of the pollen analyses are presented in the form of a percentage pollen diagram (Figure 8.4), drawn using the TILIA program (Grimm, 1993). The basic sum used is TLP-*Alnus* except where stated. The diagram has been divided into zones using CONISS as a guide (Grimm, 1987, see section 5.5.6). The lpazs are prefixed EKP, and form the basis for the following description. A separate diagram showing rare pollen types (those not exceeding 1%) is also presented (Figure 8.5).



Analyst: J.R.Kirby

Figure 8.4 Eskamhorn percentage pollen diagram (+ represents frequencies <1%)

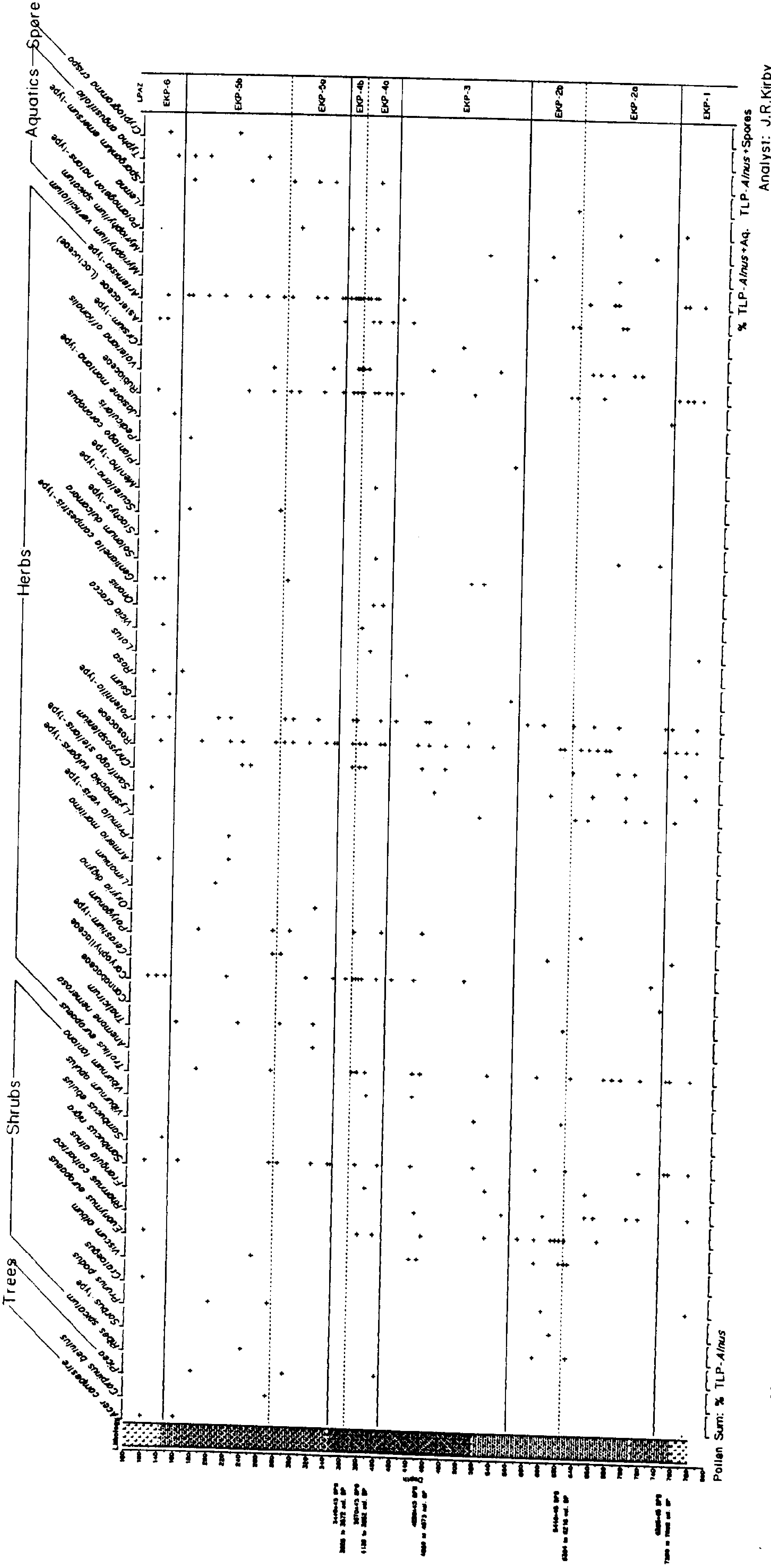


Figure 8.5 Eskamhorn percentage pollen diagram showing minor types (+ represents frequencies <1%)

EKP-1 774 to 740 cm >*c*.6400 BP (>*c*.7300 cal. yrs BP)

Figure 8.4 and 8.5

This lpaz is dominated by pollen from tree (54 to 61%) and shrub (31 to 40%) taxa with herb pollen making up only 5 to 11% of the basic sum. The main constituents of the arboreal pollen group are *Quercus* (26 to 28%), *Tilia* (12 to 19%), *Ulmus* (7 to 8%), and *Alnus* glutinosa. Frequencies of *Alnus glutinosa* pollen rise from 30% in the first sample, to 60% (TLP+*Alnus*) at the close of the zone. Pollen from *Betula*, *Pinus sylvestris*, and *Fraxinus excelsior* is less well represented with percentages <5% throughout. Shrub pollen is dominated by *Corylus avellana*-type (29 to 38%), with *Salix*, *Ilex aquifolium*, and *Hedera helix* consistently recorded but at percentages <2%. A range of herb pollen taxa occur sporadically but only Cyperaceae pollen is notable, with percentages rising from 1 to 7% mid-zone, falling to <1% at the zone close.

Values of spore morphotypes range between 9 and 16% (TLP-*Alnus*+spores), with Pteropsida (monolete) indet. the most abundant (5 to 10% TLP-*Alnus*+spores). Percentages of pre-Quaternary spores are high in the basal sample (29% TLP-*Alnus*+PQS) and fall throughout the zone to values of 6% (TLP-*Alnus*+PQS) at the zone end. LOI values rise steadily during the zone from to 6% at the base to 28% dry weight at the zone close.

The pattern of pollen distribution is marked by the increase in frequency of pollen from *Alnus glutinosa* and *Corylus avellana*-type towards the top of the zone, and the Cyperaceae pollen curve, which increases in the middle of the zone but declines at the zone close. The culmination of the rise in *Alnus glutinosa* coincides with a change in stratigraphy from a sand deposit to a sandy peat. This transition has been dated to 6382±45 BP (7380 to 7202 cal. yrs BP) and marks the boundary of lpaz EKP-1/2a.

EKP-2 740 to 560 cm *c*.6400 to *c*.5000 BP (*c*.7300 to *c*.5700 cal. yrs BP)

This zone is characterised by similar dominant arboreal and shrub taxa to lpaz EK-1. However, significant fluctuations are evident in some of the less well represented taxa, and the *Ulmus* decline has been located within this zone. Therefore, this lpaz has been separated into two subzones, which are described below.

EKP-2a 740 to 625 cm *c*.6400 to *c*.5400 BP (*c*.7300 to *c*.6300 cal. yrs BP)

This subzone is marked by high percentages of arboreal (46 to 59%) and shrub (29 to 49%) pollen, with herb pollen (4 to 9%) less well represented. The transition from the previous lpaz is characterised by a decline in frequency of *Tilia* pollen from an average of 16% in EKP-1, to 6% at 725 cm in EKP-2a. *Tilia* pollen values rise again towards middle and end of the subzone with percentage values between 7 and 15%. Pollen of *Alnus glutinosa* is particularly high, ranging between 47 and 73% (TLP+*Alnus*). Other important arboreal pollen taxa are *Quercus* (25 to 36%), *Tilia* (6 to 15%), and *Ulmus* (5 to 11%). Frequencies of *Betula*, *Pinus sylvestris*, and *Fraxinus excelsior* pollen remain at consistently low values (<5%) throughout the subzone. Pollen of *Corylus avellana*-type (28 to 47%) dominates the shrub group and frequencies of *Salix* pollen (*c*.1 to 2%) are better represented than in the previous lpaz.

Although herb pollen is not well represented, *Filipendula* and Poaceae morphotypes are recorded in most levels (*c*.1 to 2%), and values for Cyperaceae range between *c*.1 and 6%. Spore frequencies range between 8 and 16% (TLP-*Alnus*+spores). *Polypodium* (1 to 3%), *Pteridium aquilinum* (*c*.1 to 2%), and Pteropsida (monolete) indet. (6 to 12% TLP-*Alnus*+spores) are the main morphotypes represented. Pre-Quaternary spores are present in lower frequencies than in EKP-1, with values ranging from 8 to 16% (TLP-*Alnus*+PQS). LOI percentages (27 to 40% dry weight) are higher than the previous zone.

EKP-2b 625 to 560 cm *c*.5400 to *c*.5000 BP (*c*.6300 to *c*.5700 cal. yrs BP)

Subzone EKP-2b is marked by similar proportions of arboreal (57 to 70%), shrub (25 to 37%), and herb (5 to 12%) pollen percentages to EKP-2a but there are marked differences in individual taxa. In particular, there is a gradual decline in *Ulmus* pollen from an average of 8% in previous zones to 5% in EKP-2b. The boundary between EKP-2a and 2b has been drawn at the point where *Ulmus* pollen begins to fall and has been dated to 5440±45 BP (6304 to 6216 cal. yrs BP).

Other changes in arboreal taxa occur at the opening of EKP-2b. Frequencies of *Fraxinus excelsior* pollen increase to 7% and *Quercus* pollen is better represented than in lower lpazs, with values ranging between 35 and 48%. Pollen of *Alnus glutinosa* has a spiky distribution within this zone with percentages varying from 52 to 78% (TLP+*Alnus*). *Tilia* pollen rises to 12% in the centre of the subzone from 5% at the opening of the subzone.

The shrub pollen group is dominated by *Corylus-avellana*-type pollen with values ranging from 23 to 34%, and *Salix* pollen is consistently recorded at low values (<1%). There is no appreciable increase in pollen from ruderal herbs during the decline in *Ulmus* pollen, although pollen of plants such as Chenopodiaceae, *Rumex acetosa/acetosella*, Brassicaeae, Apiaceae, *Plantago lanceolata* and Poaceae occur at frequencies <2%. Two cereal-type pollen grains of the *Hordeum* size class (Andersen, 1979) were also recorded.

Spores make up between 9 and 14 % (TLP-*Alnus*+spores) and are mainly represented by Pteropsida (monolete) indet. (6 to 10%), with *Polypodium* and *Pteridium aquilinum* morphotypes both averaging <3% (TLP-*Alnus*+spores). LOI values range between 25 and 34% dry weight.

EKP-3 560 to 406 cm *c*.5000 to *c*.3800 BP (*c*.5700 to *c*.4200 cal. yrs BP)

Arboreal pollen is better represented in this lpaz with percentages reaching an average of 62%. This is partly due to increased frequencies of *Pinus sylvestris* pollen (5 to 15%) which peak mid-zone, and a general decrease in percentages of *Corylus avellana*-type pollen (16 to 37%), which average 24%. *Alnus glutinosa* pollen frequencies remain high (40 to 59% TLP+*Alnus*) but values are reduced in the middle of the zone. *Quercus* pollen percentages remain high (31 to 50%) throughout the lpaz whereas values of *Ulmus* pollen continue to decline and remain low (<2%). Frequencies of *Tilia* pollen (2 to 9%) are also generally lower than previous zones and average only 4%. Pollen of *Betula* and *Fraxinus excelsior* attains frequencies of <5% throughout.

Although herb pollen makes up a relatively small proportion of the sum (11 to 17%), frequencies of *Filipendula* pollen rise to 9% at 440 cm in the centre of the zone, with Cyperaceae (4 to 6%) and Poaceae (1 to 5%) also important herb taxa. A radiocarbon date had been obtained from between 450 and 455 cm which yielded an age of 4220±45 BP (4860 to 4573 cal. yrs BP).

The dominant spore morphotype is Pteropsida (monolete) indet., which averages 13% (TLP-*Alnus*+spores) but attains a peak of 37% (TLP-*Alnus*+spores) in the middle of the zone. Pre-Quaternary spores decline from 19% (TLP-*Alnus*+PQS) at the opening of the zone and disappear towards the end of the zone. LOI values reach their maximum within this zone, attaining percentages between 23 and 72% and averaging 57% dry weight.

EKP-4 406 to 346 cm *c.*3800 to *c.* 3400 BP (*c.*4200 to *c.* 3700 cal. yrs BP)

Significant changes occur in the representation of pollen from arboreal, shrub and herb pollen within this zone. However, it is argued that these changes relate to the same allogenic event and are therefore split into two subzones (see section 8.5 below).

EKP-4a 406 to 366 cm *c.*3800 to >*c.*3400 BP (*c.*4200 to >*c.*3700 cal. yrs BP)

Declining arboreal pollen percentages (54 to 39%) and increasing proportions of shrub (27 to 45%) pollen are the main characteristics of this subzone. This is due to the reduction in percentages of *Quercus* (42 to 22%) and *Alnus glutinosa* (47 to *c.*38% TLP+*Alnus*) pollen, and an expansion in the values of *Salix* pollen (1 to 16%) towards the top of the zone. The horizon over which *Salix* pollen increases and arboreal taxa decline has been dated to 3670±45 BP (4130 to 3852 cal. yrs BP). Frequencies of *Fraxinus excelsior* pollen rise from 1% during the initial phase of the subzone to an average of 5% during the latter part of the subzone. Arboreal and shrub pollen types such as *Pinus sylvestris* (4 to 8%), *Tilia* (2 to 4%), and *Corylus avellana*-type (26 to 32%) remain at relatively steady values throughout the zone.

Percentages of herb pollen (15 to 19%) also increase in importance during EKP-4a with *Filipendula* (3 to 7%), Cyperaceae (2 to 7%), and Poaceae (4 to 6%) the main taxa. Pollen of Chenopodiaceae and *Plantago lanceolata* begins to be consistently present with values of <*c.*1% in each level. Frequencies of spore morphotypes decline (47 to *c.*20% TLP-*Alnus*+spores), as percentages of Pteropsida (monolete) indet. fall from 35% to *c.*10% (TLP-*Alnus*+spores). Values of *Polypodium* (2 to 5%) and *Pteridium aquilinum* (5 to 11% TLP-*Alnus*+spores) remain fairly stable. Pre-Quaternary spores are present (7 to 11%) and LOI values are slightly lower than EKP-3, varying between 36 and 58% dry weight.

EKP-4b 366 to 346 cm *c.*3400 BP (*c.*3700 cal. yrs BP)

Percentages of herb pollen types increase dramatically during this subzone from 25 to 54%, whilst pollen of arboreal (24 to 38%) and shrub (22 to 38%) taxa are relatively low. This is related to the sharp rise in percentages of *Filipendula* pollen (which attains an average of 36%) from 7% at the close of lpaz EKP-4a to 47% at the close of EKP-4b. This increase in *Filipendula* has been radiocarbon dated to 3440±45 BP (3828 to 3572 cal. yrs BP).

Of the arboreal pollen types, *Alnus glutinosa* (41 to 47% TLP+*Alnus*) is dominant, *Quercus* (17 to 19%) and *Fraxinus excelsior* (2 to 4%) occur with moderate frequencies, with *Betula* and *Ulmus* minor types (<2%). Pollen of *Pinus sylvestris* decline from 9% at the opening of the subzone to 2% at the close.

Shrub pollen is less well represented in this subzone due to the decline in frequency of *Salix* pollen which attains values between 3 and 6%. Percentages of *Corylus avellana*-type pollen also decrease in this subzone from 32 to 18%. Spores are more infrequent than the previous lpaz (11 to 21% TLP-*Alnus*+spores). *Pteridium aquilinum* morphotypes decline from 9 to 3% (TLP-*Alnus*+spores), as Pteropsida (monolete) indet. (6 to 9% TLP-*Alnus*+spores) remains relatively stable. Pre-Quaternary spores decline from 9 to 4% TLP-*Alnus*+PQS and LOI values range between 32 and 40% dry weight.

EKP-5 346 to 148 cm *c*.3400 BP (<*c*.3700 cal. yrs BP)

This zone is characterised by markedly fluctuating arboreal and shrub pollen curves. Significant changes in some of the less well represented pollen taxa enables the delineation of two subzones.

EKP-5a 346 to 275 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

This zone is characterised by higher proportions of arboreal and shrub taxa than the previous lpaz. After a peak of 55% at the start of the subzone, arboreal pollen frequencies decline to 32% before increasing again towards the close of the subzone to *c*.43%. Shrub pollen frequencies (27 to 41%) are dominated by the percentages of *Corylus avellana*-type morphotypes which increase from 24% to *c*.38% in the centre of the zone and then decrease slightly to *c*.27% at the close of the zone. The main arboreal pollen taxa are *Alnus glutinosa* which increases from 36% initially to 60% in the centre of the subzone, before falling to 37% (TLP+*Alnus*) at the zone close, and *Quercus* which declines from an initial level of *c*.33% to *c*.19% for the remainder of the subzone. Pollen percentage values of *Betula* (1 to 4%) and, in particular, *Pinus sylvestris* (6 to 18%) are also more important than lpaz EKP-4b.

Although percentages of herb pollen are lower in this subzone (16 to 27%) (due to the decline in *Filipendula* pollen), Cyperaceae (3 to 7%) and Poaceae (6 to 13%) pollen percentages both increase. Pollen of Chenopodiaceae, *Plantago maritima*, and *Plantago lanceolata* is better represented and minor frequencies (<1%) of cereal-type (*Hordeum*-type) pollen are recorded at the close of the zone.

Frequencies of spores (18 to 36% TLP-*Alnus*+spores) increase towards the centre of this subzone, and decline towards the close. *Polypodium* (1 to 6%), *Pteridium aquilinum* (5 to 10%) and Pteropsida (monolete) indet. (5 to 22% TLP-*Alnus*+spores) are the most important morphotypes. Pre-Quaternary spores are moderately high with percentages between 8 and 20% TLP-*Alnus*+PQS. LOI values are generally lower than below, ranging between 15 and 35% dry weight.

EKP-5b 275 to 148 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

Arboreal pollen percentages are generally higher than EKP-2a, and increase to 59% towards the centre of this subzone before declining to 33% at the close. Frequencies of *Quercus* and *Fraxinus excelsior* pollen increase during the first part of the subzone, to 35% and 11% at 230 cm respectively, but both decline thereafter. *Alnus glutinosa* pollen (33 to 64% TLP+*Alnus*) is generally better represented in this lpaz, with frequencies of *Betula* (2 to 5%) and *Pinus sylvestris* (4 to 17%) pollen remaining relatively well represented. Shrub pollen (18 to 37%) is dominated by *Corylus avellana*-type, which declines from 31 to 14% in the middle of the subzone and increases to 36% at the close. Pollen of other shrub taxa such as *Salix* and Ericaceae are recorded at frequencies <4%.

Herb pollen (17 to 41%) is recorded at slightly higher percentages than lpaz EKP-5a. The most important taxa are Cyperaceae (3 to 9%) and Poaceae (5 to 15%) although *Filipendula* pollen reaches 20% during the lower part of the zone. Chenopodiaceae, *Plantago maritima*, *Plantago lanceolata* and cereal-type (*Hordeum*-type size class) pollen grains continue to occur at minor frequencies.

Frequencies of spores increase towards the end of the zone due to a rise in the Pteropsida (monolete) indet. percentages (6 to *c*.21% TLP-*Alnus*+spores) towards the close of the subzone. Percentages of pre-Quaternary spores also increase from *c*.11% to *c*.39% (TLP-*Alnus*+PQS) at the end of the subzone. LOI values vary between 19 and 53% dry weight.

EK-6 148 to 120 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

This zone is characterised by declining arboreal (29 to 18%) pollen percentages and a rise in the frequency of herb morphotypes (47 to 58%). *Alnus glutinosa* pollen percentages average 36% (TLP+*Alnus*) and frequencies of *Quercus* pollen are low, with an average value of only 13%. Shrub pollen averages 25% in this zone. The main taxa are *Corylus avellana*-type pollen which declines in frequency from 21 to 11% and *Salix* which increases from 2 to 12% during the zone.

Herb pollen is dominated by Poaceae, which increases from 10 to 24%, and Cyperaceae pollen, which declines from 25 to 11% throughout the zone. *Filipendula* pollen frequencies attain 6% and a range of herb pollen types are represented at low frequencies such as *Caltha palustris*-type, *Ranunculus acris*-type, Chenopodiaceae, *Plantago maritima*, *Plantago lanceolata*, Asteraceae (Cardueae/Asteroideae), and cereals.

Representation of spore morphotypes declines from 27 to 14 % (TLP-*Alnus*+spores) due to lower frequencies of Pteropsida (monolete) indet. (9 to 3%) and *Polypodium* (3 to 1% TLP-*Alnus*+spores). Spores of *Pteridium aquilinum* (*c.*11% TLP-*Alnus*+spores) are better represented than EKP-5b. LOI percentages are generally low, and decrease from 30 to *c.*18% dry weight.

8.4.3 Diatom Stratigraphy

The important aspects of the local diatom assemblage zones (ldazs) within core EK 12P are described in the following section (Figure 8.6, 8.7, 8.8 and 8.9). Diatoms are either rare and broken or absent from the clayey peats in the middle of the sequence. However, two samples were counted from the base of the sequence and are shown in Figure 8.6. A consistent diatom record is preserved in the organic clays overlying the peat which therefore post-dates 3440±45 BP (3928 to 3572 cal. yrs BP). Diatom diagrams are presented showing the common and rare species classified according to salinity tolerance (Figures 8.7 and 8.8) and also according to life-form groupings (Figure 8.9, see section 5.6.4 for references used). Diatom abundances are expressed as a percentage of total diatom valves (TDV). Definitions of the terms used regarding salinity and habitat preferences are given in Appendix B.

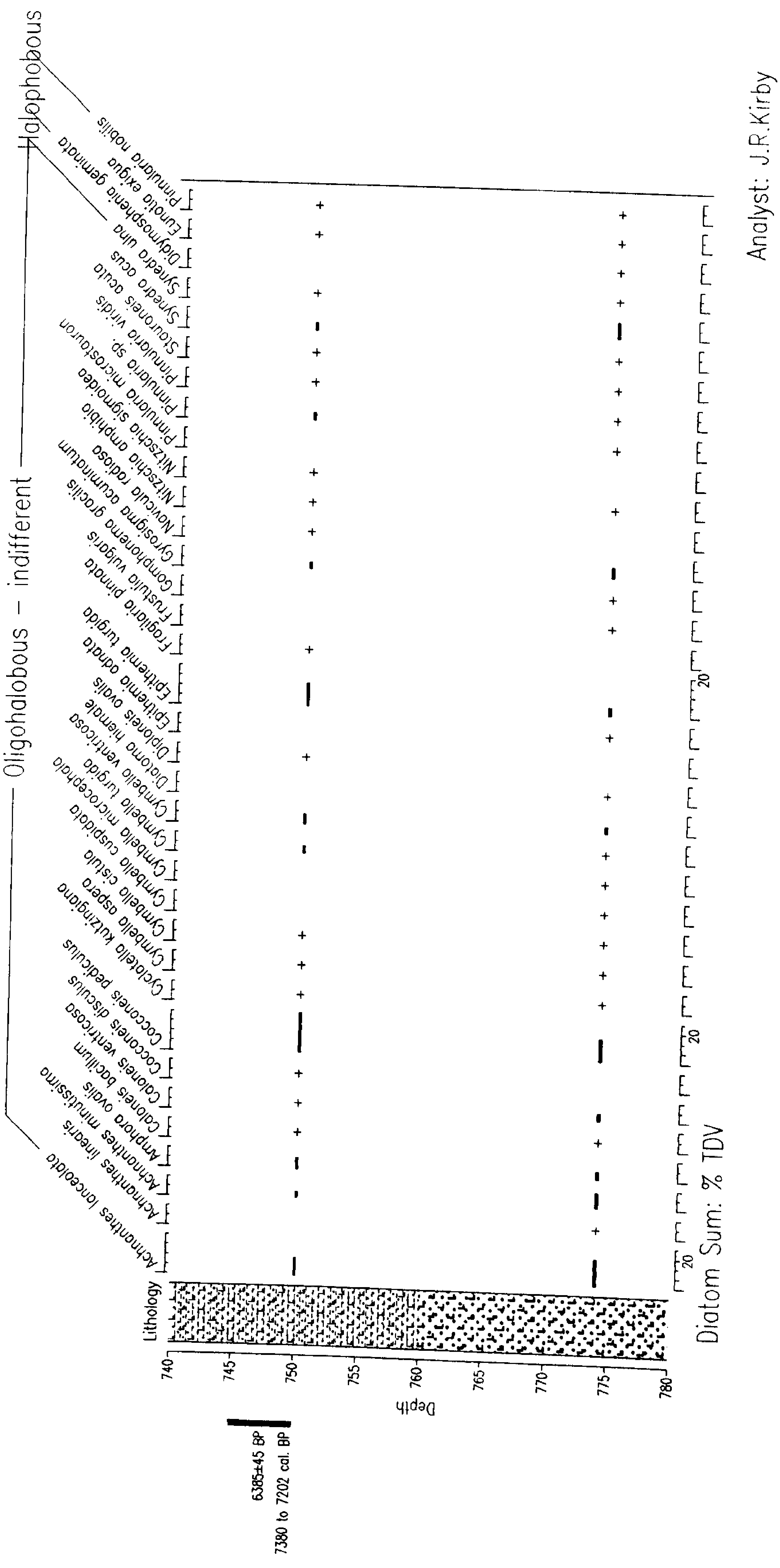


Figure 8.6 Eskamhorn percentage diatom diagram based on salinity (base of core EK 12P, + represents frequencies <3%)

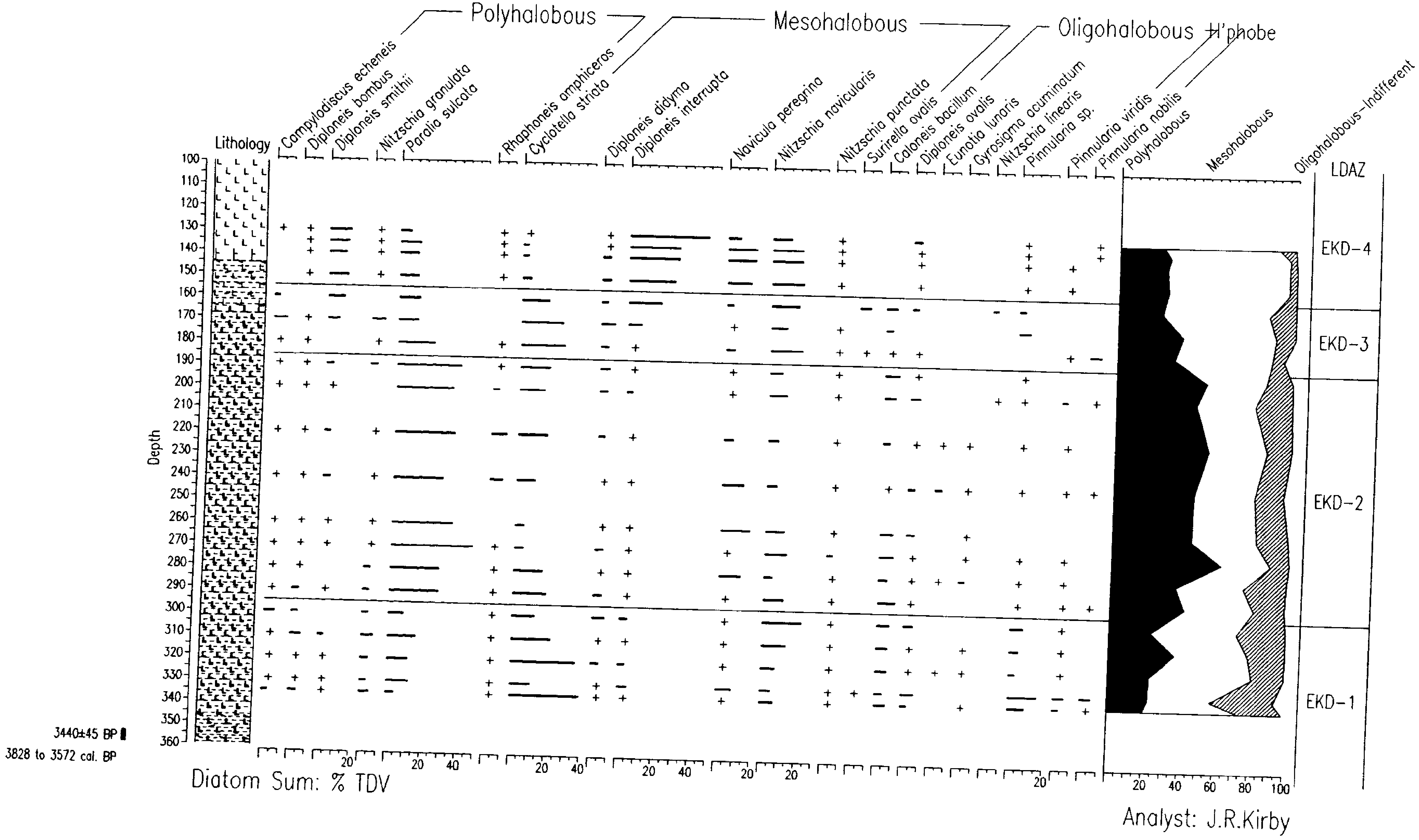


Figure 8.7 Eskamhorn percentage diatom diagram based on salinity (+ represents frequencies <3%)

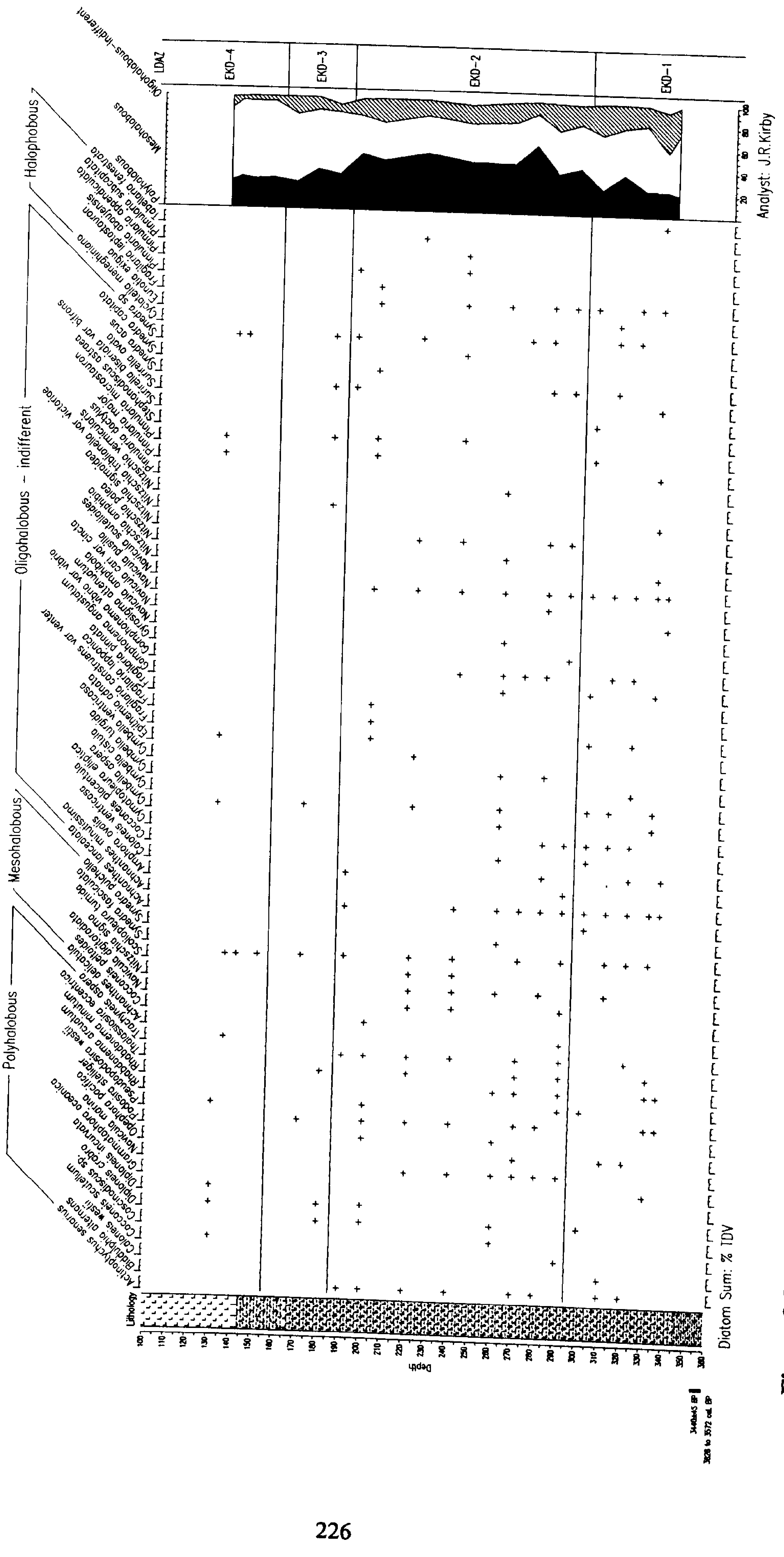


Figure 8.8 Eskamhorn percentage diatom diagram (minor taxa) based on salinity (+ represents frequencies <3%)

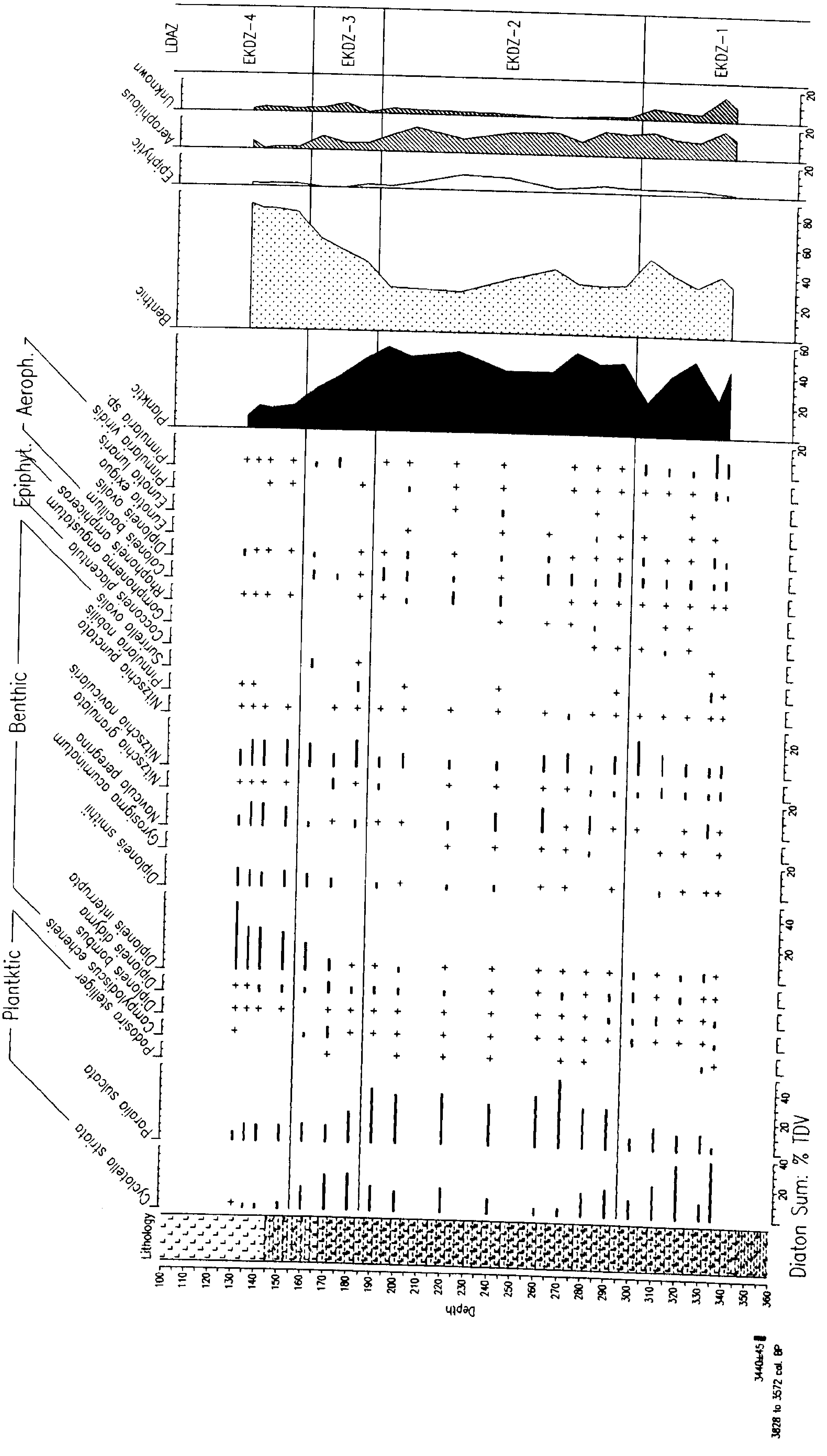


Figure 8.9 Eskamhorn percentage diatom diagram based on life-form (+ represents frequencies <3%)

Base of core 750 to 775 cm >*c*.6400 BP (>*c*.7300 cal. yrs BP)

Figure 8.6

The two samples analysed from the basal clayey sand and base of peat contain assemblages dominated by oligohalobous-indifferent taxa (>96%). The most important species present are benthic varieties. In particular, epiphytic, rheophilous taxa such as *Cocconeis pediculus* (17 to 25%), *Epithemia turgida* (4 to 12%), and *Achnanthes minutissima* (3 to 8%) are common. Aerophilous diatoms are also well represented and include *Achnanthes lanceolata* (15 to 19%) and *Gyrosigma acuminatum* (8 to 10%). Planktonic species such as *Synedra acus* (4 to 8%) are moderately well represented.

EKD-1 335 to 295 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

Figure 8.7, 8.8 and 8.9

Mesohalobous diatoms (35 to 58%) are generally the most abundant taxa within this zone although polyhalobous (20 to 37%) and oligohalobous-indifferent (18 to 36%) types are also prominent. The most important life-form groupings are planktonic (22 to 49%) and benthic (33 to 51%) diatoms. The most frequent diatom species present at the opening of the zone is the mesohalobous planktonic taxon *Cyclotella striata*, which declines from a maximum of 39% to 12% at the zone close. *Nitszchia navicularis*, a mesohalobous benthic species rises, from *c*.6% to 22% at the end of the zone. Of the polyhalobous taxa, the planktonic species *Paralia sulcata* (4 to 16%) and benthic *Nitszchia granulata* (3 to 7%) are the most important.

Oligohalobous-indifferent taxa are relatively abundant at the opening of the zone with aerophilous species such as *Caloneis bacillum* (4 to 7%), *Diploneis ovalis* (1 to 7%), and *Pinnularia viridis* (c.1 to 6%) the main species. Frequencies of *Pinnularia* spp. also decline from *c*.12% at the start of the zone to 5% at the close.

EKD-2 295 to 185 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

This ldaz is marked by high frequencies of polyhalobous species (36 to 61%) such as *Paralia sulcata* (27 to 46%) and *Rhaphoneis amphiceros* (*c*.2% to 8%). Mesohalobous diatoms (28 to 40%) are generally less well represented than in EKP-1 due to the lower abundance of *Cyclotella striata* (5 to 19%). Despite the decline in *Cyclotella striata*, planktonic taxa dominate (40 to 53%) due to the high frequencies of *Paralia sulcata*.

Benthic taxa (27 to 43%) such as the mesohalobous epipelic species *Navicula peregrina* (*c*.1 to 16%) and *Nitzschia navicularis* (5 to 12%) are also relatively abundant. Epiphytic taxa reach 10% at 220 cm due to an increase in *Rhaphoneis amphiceros*. Frequencies of oligohalobous-indifferent species (11 to 24%) are slightly lower than EKD-1 due to the decrease in percentage of *Pinnularia* spp. from *c*.5% at the close of EKD-1 to *c*.1% in EKD-2. Aerophilous taxa are moderately frequent (9 to 16%), with *Caloneis bacillum* (1 to 10%) and *Diploneis ovalis* (*c*.2 to 5%) remaining the main species.

EKD-3 185 to 155 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

This zone is characterised by an increase in the representation of mesohalobous taxa (*c*.54 to 61%) and a decline in the percentage of polyhalobous (*c*.34 to 24%) species. *Paralia sulcata* percentages (11 to 21%) are lower than in EKD-2 which causes the overall percentage of planktonic taxa to decline (46 to 27%) despite the increase in *Cyclotella striata* (15 to 25%). Percentages of benthic species are higher (46 to 61%), mainly due to an increase in mesohalobous species such as *Diploneis interrupta* (1 to 18%), *Diploneis didyma* (3 to 7%), and *Nitzschia navicularis* (9 to 18%), and polyhalobous species such as *Campylodiscus echeneis* (2 to 7%) and *Diploneis smithii* (0 to 9%).

Oligohalobous-indifferent species are poorly represented (6 to 15%), in part due to the decline in frequency of aerophilous taxa such as *Caloneis bacillum* (1 to 6%) and *Diploneis ovalis* (0 to 3%).

EKD-4 155 to 130 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

Mesohalobous diatoms dominate this zone (66 to 69%). In particular, there is a marked increase in mesohalobous benthic taxa such as *Diploneis interrupta* (25 to 44%) and *Navicula peregrina* (7 to 16%), with *Nitszchia navicularis* (10 to 17%) also well represented. Polyhalobous diatoms (25 to 28%) are moderately well represented with planktonic species such as *Paralia sulcata* (6 to 12%) and benthic species such as *Diploneis smithii* (9 to 12%) the most frequent types.

Benthic diatoms (78 to 83%) are the main life form group, with planktonic taxa declining from 16 to 8% throughout the zone. Epiphytic and aerophilous groups are minor at <5%.

8.5 Inferred Local Vegetation History and Depositional Environment

EKP-1 774 to 740 cm >*c*.6400 BP (>*c*.7300 cal. yrs BP)

Figure 8.4 and 8.5

The local environment prior to paludification at Eskamhorn is likely to have been colonised by *Alnus glutinosa* and Cyperaceae and a wet, river edge environment is suggested. Relatively high pollen percentages for *Tilia*, *Ulmus* and *Quercus* pollen imply that the valley floor adjacent to the coring site (and the valley side) was dry and perhaps dominated by these taxa. Relatively low frequencies of *Alnus glutinosa* suggests that the carr vegetation was relatively open. *Ulmus* wood macrofossils were identified between 748 and 749 cm which may suggest this taxon in close proximity to the coring site. The substantial rise in *Alnus glutinosa* pollen percentages mid-zone onwards, is indicative of increasingly wet conditions, probably a response to a significant and continued rise in water table. Assigning an environment of origin for the *Corylus avellana*-type pollen is not easy (see section 7.5). It is possible that *Corylus avellana* formed part of the understorey of the dry land woodland, or formed a transitional community between the stream side wetland and the dryland woodland community.

Paludification ensues at -5.80 m OD, coincident with a decline in the influx of pre-Quaternary spores. As peat accumulation occurs LOI values increase accordingly. Two different causes can be invoked to explain the decline in pre-Quaternary spores. It is possible that the source of the pre-Quaternary spores is fluvial, and as water levels rise, the rate of organic sedimentation increases, raising the peat forming surface above the average flood level, so the frequency of inwashed spores is reduced. Alternatively, the pre-Quaternary spores could be derived from the pre-Holocene basal sediments, and as the input of autochthonous organic material continues, this seals in these sediments and the proportion of secondary pre-Quaternary spores declines. It is also possible for both of these explanations to be valid.

The pollen and spore changes that occur within EKP-1 are typical of a buried soil/basal peat transition. The high percentages for *Tilia* pollen begin to decline towards the close of the zone as the water level rises and peat development encroaches on the area supporting *Tilia* trees on the valley floor. *Quercus* percentages are maintained which is a reflection of oak being more tolerant to waterlogging than *Tilia* (but less so than *Alnus glutinosa* which increases in abundance), and is able to survive until at least the early stages of peat formation. Waller (1988; 1994b; 1994c) provides examples of analogous biostratigraphic changes from the Fenland. The decline in *Tilia* and simultaneous rise in *Alnus glutinosa* as water levels rise coincides with the onset of peat accumulation at Eskamhorn, and is dated to 6385±45 BP (7380 to 7202 cal. yrs BP).

EKP-2a 740 to 625 cm *c.*6400 to *c.*5400 BP (*c.*7300 to *c.*6300 cal. yrs BP)

Pollen of *Alnus glutinosa* dominates this lpaz. The spiky nature of the pollen curve and the presence of abundant scalariform perforation plates on the pollen slides is indicative of on site growth of alder, and *in situ* deposition of woody remains (*cf.* Van der Wiel, 1982; Smyth, 1986). Percentage values of up to 73% (TLP+*Alnus*) suggest the closure of the carr canopy. The wetland ground flora possibly consisted of wetland herbs such as *Ranunculus acris*-type, *Filipendula*, Cyperaceae, and Poaceae.

Although *Alnus glutinosa* is likely to have been the dominant carr taxa, the continuous presence of *Salix*, which is severely underrepresented in the pollen spectrum (Bradshaw, 1981a; Huntley & Birks, 1983; Janssen, 1984), indicates that it may too have formed part of the wetland community which is difficult to identify palynologically (Bunting *et al.*, 1998a; 1998b). *Quercus* and *Corylus avellana* are also likely components of the drier areas of fen carr (see section 7.5) although it is not possible to distinguish precisely between the proportion of pollen derived from local wetland or extra-local dryland sources. A *Corylus avellana* nut is recorded at 715 cm in the core (Table 8.3) which suggests at least some of the *Corylus avellana*-type pollen originates from trees growing locally on the floodplain. However, Godwin (1975a; 1978) and Brown (1988) suggest that *Corylus avellana* nuts are often of allochthonous origin due to their easily transportable nature. Waller (1993) contends that such nuts are equally as likely to be derived from vegetation colonising the valley slopes. *Corylus avellana* wood macrofossils also occur between 709 and 710 cm at Eskamhorn (Table 8.3) which tends to favour the local explanation for the source of at least some of the *Corylus avellana* pollen. Wood macrofossil remains of *Fraxinus excelsior* and *Ulmus* within this zone may suggest that the floodplain wetland is limited in size, and the dry valley side community is still close enough to the coring site for senescent trees to fall into the wetland.

The high percentages for *Tilia* (up to 15%) suggests woodland dominated by lime was colonising the proximal valley sides or terraces in the floodplain (Brown, 1988) and contributed to the extra-local pollen rain.

EKP-2b 625 to 560 cm *c*.5400 to *c*.5000 BP (*c*.6300 to *c*.5700 cal. yrs BP)

The transition between EK-2a and 2b has been drawn at the point where *Ulmus* pollen begins to decline. Other pollen changes accompany this fall in *Ulmus* pollen frequencies, most notably an increase in *Quercus* and *Fraxinus excelsior* pollen percentages and an expansion in the abundance of herb pollen. Taxa such *Rumex acetosa/acetosella*, Chenopodiaceae, *Plantago lanceolata*, Asteraceae (Cardueae/Asteroideae), and Poaceae are recorded (but not for the first time) along with a few cereal pollen grains of *Hordeum*-type. The most likely explanation for this sequence of changes is that the surrounding woodland environment opens up, with the gaps left by dead elm trees enabling better pollen transport through the trunk space (*cf.* Tauber, 1965). This would encourage flowering of light demanding herbs. Trees such as *Quercus* and *Fraxinus excelsior* would benefit from these more open conditions, which would enhance pollen production and enable new saplings to become established in place of *Ulmus*. *Corylus avellana* is usually the main beneficiary of declining *Ulmus* values (Waller, 1994c), with significant increases in its pollen immediately after the classical decline (see Clark & Godwin, 1962; Godwin & Vishnu-Mittre, 1975; Peglar, 1993). Therefore it is strange that *Corylus avellana*-type pollen does not begin to rise until after *Quercus* and *Fraxinus excelsior*, as one would expect this taxon to respond quickly to such a perturbation. This may suggest that *Corylus avellana* was not a significant component of the dry woodland understorey but was abundant in the drier areas of *Alnus* carr wetland.

Towards the centre of the zone, there is a slight rise in indeterminable deteriorated pollen grains (*sensu* Cushing, 1964) and pre-Quaternary spores, which may be evidence for the remobilisation of penecontemporaneous pollen in surrounding soils, and increased erosion from valley sides owing to the death of *Ulmus* stands (Pennington, 1979; Hirons & Edwards, 1986).

A date of 5440±45 BP (6304 to 6216 cal. yrs BP) has been obtained for the onset of declining *Ulmus* frequencies at Eskamhorn. This date is slightly early for the traditionally quoted approximation of *c*.5000 BP (Smith & Pilcher, 1973; Huntley & Birks, 1983), although an older age for the beginning of the decline, as well as some regional diachroneity, may perhaps be expected. Dates which apply to the first indications of falling *Ulmus* pollen have their means between *c*.5300 and 5100 BP (Smith and Pilcher 1973).

Unfortunately, no radiocarbon dates are available for the *Ulmus* decline in the Humberhead Levels or Vale of York area. However, Beckett (1975) has obtained a date from just above the *Ulmus* decline horizon of 5099±50 (5936 to 5730 cal. yrs BP) at Gransmoor Quarry in northern Holderness which provides a *terminus post quem* for the onset of declining *Ulmus* values. Although this date is somewhat later than at Eskamhorn, early ages for this event are not unusual. In the Fenland, early dates for the onset of the *Ulmus* decline are *c*.5500 BP (*c*.6500 to *c*.6000 cal. yrs BP, Waller, 1994c) and even earlier ages have been reported from other parts of the east coast lowlands. Bennett (1983b) obtained a date for the *Ulmus* decline at Hockham Mere of *c*.6000 BP (*c*.7200 to *c*.6600 cal. yrs BP). Waller (1994c) shows that there is considerable variability between *Ulmus* decline dates (*c*.900 years) within the Fenland itself, and suggests it may not be such a synchronous chronostratigraphic marker for this region. Other early dates for the *Ulmus* decline are also forthcoming from floodplain sites elsewhere in England such as the Severn basin where Brown (1988) has obtained an *Ulmus* decline similar to that from Eskamhorn (*c*.5400 BP, *c*.6400 to *c*.5900 cal. yrs BP) from lowland floodplain deposits. Research in the Dutch coastal river valleys has revealed dates for the *Ulmus* decline as early as *c*.6000 BP (*c*.7200 to *c*.6600 cal. yrs BP, Hofstede *et al*., 1989).

In upland regions of Northeast Yorkshire, the *Ulmus* decline generally appears later than in the lowlands (Simmons & Innes, 1987; 1988; Simmons *et al*., 1993). This provides evidence for diachroneity of up to a millennium between lowland and upland sites in this region which perhaps has implications with regard to possible causal mechanisms for this phenomena. It is possible that the *Ulmus*-decline in this region was attributable to the spread of a pathogen (Rackham, 1980) through already disturbed woodlands (*cf.* Peglar & Birks, 1993), from oceanic lowland sites such as the Fenland and Humber, into the uplands. Direct evidence in support of such an argument is not yet forthcoming although Perry & Moore (1987) provide a modern example of how a recent outbreak of Dutch elm disease shows similar changes in pollen assemblage to the fossil counterpart. Also, the identification of the beetle responsible for carrying the fungal pathogen that causes the death of *Ulmus* trees (*Scolytus scolytus*) within fossil deposits lends further support to the possibility of disease as a principal cause (Girling & Greig, 1985; Girling, 1988). However, a slow spread of disease such as this is not in agreement with recent experience for a rapid decline in *Ulmus* populations following the outbreak of Dutch elm disease although the dense woodland cover during the mid-Holocene period may to some extent act as a buffer to the spread of the disease carrying vector.

The lack of convincing evidence for human activity around the time of the *Ulmus* decline, such as increases in the pollen of *Plantago lanceolata* etc., (Iverson, 1960) does not discount anthropogenic clearance as a hypothesis and absence of ruderal herbs is a feature of many *Ulmus* decline events (e.g. Clark & Godwin, 1962). However, the fluctuating nature of the pollen curves for *Quercus*, *Alnus glutinosa*, and *Corylus avellana*-type, in conjunction with light demanding herbs and some cereal pollen, may be a weak (or distant) manifestation of deliberate woodland disturbance in association with the *Ulmus* decline.

EKP-3 560 to 406 cm *c*.5000 to *c*.3800 BP (*c*.5700 to *c*.4200 cal. yrs BP)

Subtle changes in the pollen derived from various different source areas occurs within this zone which may be indicative of slight opening up of the local carr vegetation, although other explanations may be put forward. The pollen curve of *Alnus glutinosa* declines towards the centre of the zone, during which time pollen of *Pinus sylvestris*, *Filipendula*, and Cyperaceae is better represented than the previous zone. It could be argued that the increase in pollen percentages of *Pinus sylvestris* is attributable to inwashing of fluvially transported pollen, as it has already been established that this particular taxa has a high propensity for floatation (section 7.5). If this was the case, it could be expected that input of minerogenic sediment would increase, the *Pinus sylvestris* pollen may be poorly preserved, and other secondary morphotypes, such as pre-Quaternary spores would be recorded with greater frequency. However, the evidence does not support this explanation. This event occurs mid-zone where the core is dominated by organic matter and LOI values are at their highest, the *Pinus sylvestris* grains are normally well preserved and other possible indicators of increased fluvial activity such as pre-Quaternary spores and *Pteridium aquilinum* are low or absent, although Pteropsida (monolete) indet. percentages do peak. Therefore, it is more likely that the increase in *Pinus sylvestris* pollen is related to better representation of the regional pollen rain due to canopy openings in the *Alnus* carr. EKP-3 occurs between *c*.5000 and *c*.3800 BP (*c*.5700 to *c*.4200 cal. yrs BP) which approximately coincides with the date obtained from basal peat deposits at Hirst Courtney (see Chapter 7) where palynological evidence also suggests high regional input of *Pinus sylvestris* pollen during HCP-1, *c*.4200 to *c*.3900 BP (*c*.4700 to *c*.4300 cal. yrs BP).

The better representation of *Filipendula* and Cyperaceae may also be a product of this change in the local community, which either reduces the filtration effect, whereby dense local vegetation acts as a buffer the extra-local and regional pollen rain (*cf. Pinus sylvestris* above), or is due to increased flowering of the fen ground flora due to better light availability. The coincident peak in Pteropsida (monolete) indet. may also be a response associated with these local wetland community changes but any shifts in spore production are difficult to link with the pollen record due to the different function of spores and uncertainty regarding patterns of sporulation (*cf.* Tinsley & Smith, 1974).

Quercus pollen is also relatively better represented in this zone, coincident with reduced *Alnus glutinosa* frequencies. The *Alnus/Quercus* ratio (see Shennan, 1980) has been used to identify wet and dry phases related to water level changes in the Fenland (Godwin & Clifford, 1938; Godwin, 1940; Shennan, 1980). In this model, *Quercus* frequencies may increase in wetter periods (in which local *Alnus glutinosa* frequencies are reduced due to excessive wetness) due to increased extra-local input of pollen. The absence of *Quercus* wood macrofossils from this part of the core means there is no evidence to suggest that *Quercus* was colonising the fen woodland (although it was possibly surviving within it) so it is feasible that the site became locally wetter. This scenario would favour wetland herbs such as *Filipendula*, which are better represented, and the decline in locally produced *Alnus glutinosa* pollen frequencies may lead to increased extra-local input of *Quercus* pollen (Waller, 1994c). However, if *Quercus* is an undetected significant component of the fen carr, the slight expansion of *Quercus* frequencies over *Alnus glutinosa* may also be explained as a result of competition between alder scrub and fen oak woods, with the latter expanding during dry periods (Godwin & Clifford, 1938; Waller, 1994c). Therefore, due to the lack of additional evidence, this inverse relationship between *Alnus glutinosa* and *Quercus* is of little value in determining wet or dry shifts in this case.

EKP-4a 406 to 366 cm *c*.3800 to >*c*.3400 BP (*c*.4200 to >*c*.3700 cal. yrs BP)

Dramatic shifts in the pollen percentages of tree and shrub taxa occur within this zone that are suggestive of a rise in water table. The main features are the declining frequencies of *Alnus glutinosa* and *Quercus* pollen, both of which decrease simultaneously and consistently throughout the zone. Towards the top of the zone, frequencies of *Salix* increase to significant percentages for an underrepresented taxa (up to 16%), which is suggestive of co-dominance of this taxon with *Alnus glutinosa* (Waller, 1994c). It is unclear whether this increase in importance of *Salix* is the result of a shift in ecotone or due to increased flowering of previously underrepresented *Salix* shrubs within the carr vegetation prior to the decline in *Alnus glutinosa*. The recovery of a *Salix* wood macrofossil from sediment at the close of this zone attests to the local presence of this taxa (see Figure 8.3). Dupont (1987) and Iremonger & Kelly (1988) suggest *Salix* is more tolerant of wet conditions than *Alnus glutinosa* which infers a rise in water levels during this time. The onset of this increase in wetness is dated to 3670±45 BP (4130 to 3852 cal yrs. BP).

A rising water table would be one explanation to explain this scenario, whereby some of the *Alnus glutinosa* and *Quercus* trees growing in the fen carr are ousted due to the increasingly wet conditions which become more suited to *Salix*. It is noteworthy that *Corylus avellana*-type percentages are unaffected which indicates that *Corylus avellana* was not a significant component of the carr wetland during this time but instead may have been predominant on the damp floodplain edge or dry valley sides. The presence of a *Corylus avellana* nut from 400 cm (Table 8.3) suggests that this ecotone was not a significant distance from the coring site. *Fraxinus excelsior* pollen is better represented towards the close of the zone which may also be evidence of increased extra-local input of pollen from proximal damp floodplain edge or dryland sources.

The decline in *Quercus* pollen percentages during this lpaz could be interpreted as a reduction in the amount of *Quercus* growing within the fen carr as a result of waterlogging. However, given the explanation for better *Quercus* representation in EKP-3, and the lack of evidence for significant local *Quercus*, it is most likely that *Quercus* was dominant at the floodplain margins where conditions would have been drier until the rise in water table. The rising water levels in this scenario would drown the *Quercus* trees colonising this area and *Alnus glutinosa* would become established at what was previously the floodplain edge. This represents ecotone movement on the floodplain with the expansion of wet conditions.

Opening up of the fen carr in the vicinity of the site is suggested by the consistent recording of herbs such as Chenopodiaceae and *Plantago lanceolata*. There is also evidence for an increased inwashing of river water, as pre-Quaternary spores reach moderately high proportions along with other morphotypes associated with fluvial deposition such as *Pteridium aquilinum* and *Pinus sylvestris* (*cf.* De Jong, 1070-71; Devoy, 1979; Clark & Patterson, 1985; Robinson, 1993; Waller, 1994c; Walker *et al.*, 1998). Evidence for encroaching brackish conditions or the advancement of tidal conditions is suggested by the appearance of a range of saltmarsh taxa towards the end of EKP-4a such as *Plantago maritima*. Other herbs present such as Chenopodiaceae, *Artemisia*-type, and Poaceae may also be indicative of saline marshes near to the coring site.

EKP-4b 366 to 346 cm *c.*3400 BP (*c.*3700 cal. yrs BP)

A continuation of rising water levels is indicated by the vegetation changes occurring within this lpaz. The most striking feature is the increase in *Filipendula* values to a maximum of 47% (TLP-*Alnus*). This rise is dated to 3440±45 BP (3828 to 3572 cal. yrs BP). *Salix* percentages decline which is perhaps an indication of the local removal of the shrub carr community and the establishment of fen meadow conditions due to the high water table. Frequencies of other herb pollen taxa such as Cyperaceae and Poaceae decline slightly during this zone although this is probably the result of interdependence within the dataset and not a reflection of a reduction in their abundance as these taxa would be expected to occur within a wet fen herb community (Wheeler, 1980b). The apparent decrease in percentages of *Corylus avellana*-type may also be a product of bias associated with percentage calculation. Re-examination of these samples using pollen concentration data would be the only method of resolving this uncertainty. Nevertheless, the general trend of rising water levels has been established.

Alnus glutinosa pollen frequencies are greater in this lpaz which, along with the persistence of *Salix*, is suggestive of the persistence of this fen carr community elsewhere on the floodplain. The identification of an *Alnus glutinosa* wood macrofossil from *c.*360 and *c.*350 cm (Table 8.3) suggests this community was near to the coring site. A range of open ground indicators occur, such as *Ranunculus acris*-type, Chenopodiaceae, Apiaceae, *Plantago lanceolata*, and *Artemisia*-type. These herbs are likely to be of local wetland origin (some members of the Apiaceae family occur in fen systems; *Apium*, *Cicuta*, *Peucedanum* and *Oenanthe*, Wheeler, 1980b; Rodwell, 1995), although they may also be an indication of cleared areas within the surrounding catchment given the poor representation of arboreal taxa within this zone. However, as mentioned above, this may be due to the overrepresentation of local vegetation.

The biostratigraphic changes during EKP-4a and 4b are similar to those occurring within the upper levels of the peat at Hirst Courtney (see Chapter 7), albeit *c.*1000 years later, and in the research by Smyth (1986) and Smyth & Jennings (1988) in the Combe Haven valley in East Sussex. There are some differences in the vegetation response to rising water levels between the sites, however, as at Hirst Courtney, *Filipendula* pollen percentages rise first, followed by *Salix*, whereas the opposite trend is evident at Eskamhorn. This is likely to be an expression of different local factors influencing vegetation dynamics on the floodplain.

EKP-5a 346 to 275 cm <*c.*3400 BP (<*c.*3700 cal. yrs BP)

This lower lpaz boundary coincides with a change in sedimentary regime from a brown wood peat to a brown-grey organic clay. Percentage values of *Filipendula* pollen decline sharply and pollen percentages of arboreal and shrub pollen recover. Although no depositional hiatus is evident in the stratigraphy, some erosion of this contact is possible due to the abrupt nature of the decline in *Filipendula.* However, there is evidence to suggest tidal flooding increased and the inwash of saline water occurred which may also explain the sudden decline in *Filipendula.* Percentages of *Pinus sylvestris* and pre-Quaternary spores are significantly higher than the previous zone and their frequencies are erratic, which is suggestive of tidal deposition. LOI values are relatively low (<35%) which also suggests significant inwash of minerogenic sediment due to increased river flooding (Devoy, 1979; 1980).

Values of Poaceae, Cyperaceae, Chenopodiaceae, and *Plantago maritima* pollen are greater, which suggests the development of riverine saltmarshes and reedswamp with a brackish water influence at the coring site. Poaceae pollen percentages are only moderately high (*c.*10%) and the inorganic nature of the sediments suggests that the local vegetation may have been patchy, consisting of a mosaic of exposed muddy substrates. The presence of extensive saltmarsh environments is indicated at the coring site, with carr vegetation occupying a separate ecotone on the higher areas of floodplain bordering the saltmarshes and river mudflats. The occurrence of frequent tidal inundation and creeks distributing estuarine water is also suggested by the diatom record which coincides with this lpaz (see section 8.6).

Percentages of *Alnus glutinosa* pollen recover further which suggests *Alnus* dominated carr was established at the margins of the river/salt marsh environment and contributed to the extra-local pollen rain. Frequencies of extra-local pollen possibly of dry land origin are low considering the apparently relatively open nature of the environment in the vicinity of the coring site suggesting a more distant pollen source. *Quercus* and *Corylus avellana*-type pollen are the likely major constituents of such peripheral woodland, which may also suggest some degree of woodland clearance has been undertaken. The occurrence of Ericaceae and *Plantago lanceolata* may be indicative of significant open areas within this woodland. The presence of a few cereal-type pollen grains towards the top of the zone may be related to human activities within these open areas although they were of *Hordeum*-type size category which includes wild grasses that could grow on river margins and saltmarshes (e.g. *Glyceria* and *Elymus*). The suggestion that the river saltmarsh is fringed by a belt of carr woodland may filter much of this dryland pollen. Also, the increase in water level and expansion of wetland communities extending the floodplain over perhaps previously dry ground may simply have pushed the dryland community further from the coring site so that it is not easily detectable. Examination of the stratigraphy of the valley (Figure 8.1) suggests that the floodplain wetland would have extended at least 800 m to the north meaning that much of the dry land signal is from the regional pollen rain some distance away.

EKP-5b 275 to 148 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

Minor changes occur within this lpaz which suggests small scale vegetation changes of a localised nature. *Filipendula* pollen is more important at the opening of the zone which suggests that such fen herbs were colonising the higher, fresher parts of the saltmarsh away from tidal creeks where inundation by brackish water is less common. The abundance of minerogenic sediment in the core, with occasional monocotyledonous stems suggests the immediate environment of deposition was a saltmarsh, with exposed areas of mudflat. Cyperaceae, Poaceae, and *Filipendula* were perhaps abundant in the fen vegetation colonising the areas of floodplain further away from frequent tidal inundation at the floodplain edges. Pollen of *Alnus glutinosa* dominates which indicates that *Alnus* carr was at least prevalent close to the coring site. The presence of several *Alnus glutinosa* wood macrofossils within this zone (see Figure 8.3) suggests it may have encroached onto the high saltmarsh where dead trees could fall onto the exposed muddy areas. *Alnus glutinosa* can withstand a fluctuating water table through the development of adventitious roots (McVean, 1956a) and has been found in tidal woodland (Walker *et al.*, 1998) where it is tolerant of occasional flooding by saline water (Ranwell, 1974).

Frequencies of *Fraxinus excelsior* and *Quercus* pollen are better represented towards the centre of this lpaz which may reflect these tree taxa invading the fringing carr community, or increased frequencies of regional pollen input. Fluctuating frequencies of *Pinus sylvestris* pollen and high percentages of pre-Quaternary spores suggests continued tidal input of estuarine or fluvial water. Occasional grains of saltmarsh taxa such as *Armeria maritima* and *Limonium* attest to the presence of estuarine conditions and saline river marshes at the coring site. *Alnus glutinosa* dominated fen carr is likely to be fringing the wettest areas and invading drier part of the marsh.

EK-6 148 to 120 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

The pollen assemblages within this lpaz indicate that the environment has become fresher and perhaps drier. Pollen frequencies of the dominant herb taxa (Poaceae and Cyperaceae) increase and other wetland herbs such as *Caltha palustris*-type, *Ranunculus acris*-type, and *Filipendula* are well represented. Such taxa are not tolerant of significant inwashing of brackish water. This suggests an expansion of freshwater wetland communities (e.g. reedswamp or fen meadow) in the vicinity. There is further evidence for a reduction in tidal inundation from declining frequencies of *Pinus sylvestris* pollen and pre-Quaternary spores. Values of *Salix* pollen also increase significantly, which suggests that the local river marsh environment became drier and more suitable for the colonisation of such shrubs.

Frequencies of *Alnus glutinosa* pollen remain high enough to indicate continued local growth within the floodplain. Evidence for human activities is equivocal due to the ecological range of the herb species recorded, although percentages of possible non-wetland tree taxa such as *Quercus*, *Tilia*, *Fraxinus excelsior*, and *Corylus avellana*-type are low indicating perhaps a rather open landscape. The occurrence of Urticaceae, Chenopodiaceae, *Rumex acetosa/acetosella*, Brassicaceae, Apiaceae, *Plantago lanceolata*, Asteraceae (Cardueae/Asteroideae), *Artemisia*-type, and cereal-type pollen (*Hordeum*-type), cannot be considered as indicating the presence of possible human activities such as agriculture, due to the possible wetland or saltmarsh affinities, for these taxa.

8.6 Palaeohydrological Interpretation

The diatom record covers the lowest and uppermost part of the core, with diatoms either extremely rare or absent from the intervening peat, probably due to dissolution of silica skeletons as a result of silica deficiency during or after deposition (*cf.* Vos & de Wolf, 1993b; 1994; 1997). The diatom assemblage obtained from the sand at the base of the core is discussed first although conclusions are tentative due to the few samples counted, then the diatom results from the upper part of the core are summarised.

Base of core 750 to 775 cm >*c*.6400 BP (>*c*.7300 cal. yrs BP)

Figure 8.6

The diatoms recorded from the base of the core are all oligohalobous-indifferent or halophobous which indicates that prior to paludification and as peat formation ensued, the site was flooded by fresh water. The stratigraphic transect at Eskamhorn (Figure 8.1) suggests that the sample core EK 12P was taken from a depression in the pre-Holocene surface which, on morphological evidence, has the characteristics of a small channel. This suggestion is supported by the diatom assemblage which is dominated by epiphytic taxa such as *Cocconeis pediculus*, *Epithemia turgida*, and *Achnanthes minutissima*, which have a preference for rheophilous conditions (de Wolf, 1993). Diatom taxa such as these live attached to aquatic plants and require the presence of running water. Species such as *Cocconeis pediculus*, which rarely ever occur outside water bodies (Van Dam *et al.*, 1994), together with planktonic taxa such as *Synedra acus* give additional evidence for the presence of a flowing stream in the vicinity of the coring site. This is likely to be a small subsidiary channel because the main Aire watercourse would have down-cut to much deeper depths in the central valley areas (Gaunt & Tooley, 1974).

Other taxa present such as *Achnanthes lanceolata* and *Gyrosigma acuminata* are aquatic and often live attached to plants and silt respectively (Vos & de Wolf, 1993a). De Wolf (1993) also classifies these species as aerophilous which means they can withstand occasional exposure to air and may be colonising shallow water around the stream edge or wet stream bank environment.

This diatom assemblage coincides with lpaz EKP-1. The Cyperaceae and *Alnus glutinosa* pollen recorded within these basal levels are likely to be the very local component of the pollen rain, and from vegetation colonising the wet bank sides of this drainage channel. Subsequent to widespread water-level rise, due to high base levels *c*.6400 BP (*c*.7300 cal. yrs BP) associated with rising sea-level, peat accumulation and *Alnus* carr development was of a sufficient pace that the water course became blocked off and the palaeochannel feature filled in with organic material. Brown (1996) suggests that such cut-offs easily become choked with vegetation as succession proceeds.

EKD-1 335 to 295 cm <*c*.3400 BP (<*c*.3700 cal. yrs BP)

Figure 8.7, 8.8 and 8.9

The diatoms within EKD-1 coincide with lpaz EKP-5a in the pollen record and therefore post-date 3440±45 BP (3828 to 3572 cal. yrs BP). Mesohalobous diatoms are most common, particularly the estuarine planktonic species *Cyclotella striata*. This diatom is associated with tidal channels (Vos & de Wolf, 1988; 1993a) and therefore is probably part of the allochthonous component which is washed on to the site with polyhalobous planktonic species such as *Paralia sulcata* (Denys & Verbruggen, 1989. High frequencies of *Cyclotella striata* have been reported previously from inorganic deposits within inner estuarine environments such as those in the Thames (Devoy, 1982; Milne *et al.*, 1983; Milne, 1985; Juggins, 1992), the Hawkesbury valley in New South Wales, Australia (Devoy *et al.*, 1994), and Chesapeake Bay estuary in the USA (Cooper, 1995). Similarly, Verbruggen & Denys (1989), Denys & Verbruggen (1991), and Vos & de Wolf (1997) describe brackish river deposits ('lower Schelde deposits') from the Province of Zeeland in the Dutch coastal lowlands, and in the Belgian perimarine area with a preponderance of *Cyclotella striata*, indicating estuarine influence in the lower tracts of tidal rivers.

A transition from riverine to estuarine conditions is apparent in EKD-1 as oligohalobous-indifferent, aerophilous species such as *Caloneis bacillum* and *Pinnularia viridis* are present at the base of the zone which probably reflects the autochthonous diatom flora colonising the supratidal freshwater peaty river marsh (*cf.* Denys & Verbruggen, 1989). As estuarine conditions expand and marine water floods the local environment and deposits clay and silt, benthic species such as *Nitzschia navicularis* and *Nitzschia granulata* occupy exposed muddy substrates. The abundance of planktonic and benthic species and the lack of epiphytic taxa may suggest the local environment was not densely vegetated, but was characterised by river mudflats and tidal creeks within a saltmarsh environment which may have been submerged for long periods of time.

EKD-2 <*c*.3400 BP (<*c*.3700 cal. yrs BP)

Frequencies of *Cyclotella striata* are reduced in this zone but are replaced by high values of *Paralia sulcata*. This suggests a further expansion of marine conditions up the Aire valley associated with a positive tendency in sea-level. A brackish influence in the river Aire is evident by the presence of *Cyclotella striata* but the dominance of *Paralia sulcata* suggests that the environment has changed from a sheltered tidal river valley to an open tidal flat and marsh environment with subtidal channels. This situation is similar to the change registered in the lower Holocene deposits of the Province of Zeeland due to rising sea-level (Vos & de Wolf, 1997).

Care must be taken when interpreting assemblages dominated by marine tychoplanktonic species such as *Paralia sulcata* and this is given consideration below. *Paralia sulcata* is particularly problematical due to its cosmopolitan distribution and the uncertainty regarding its habitat palaeoecology (Cooper, 1995; Zong, 1997b). For instance, Denys (1991) classifies it as marine tychoplanktonic whereas de Wolf (1993) groups it within the benthic community. Cooper (1995; p. 55) states that '*Paralia sulcata* is a classic example of our ignorance of the microhabitats in coastal marine regions - it is widespread and common, yet there is no precise information on its actual niche...' She goes on to suggest that it may be part of the unattached bottom community of diatoms that drift along the bottom of coastal waters.

Although *Paralia sulcata* may be loosely associated with benthic habitats (hence its classification as tychoplanktonic, Vos & de Wolf, 1993a), it is easily lifted into the water column (especially during storms, e.g. Cullingford *et al*., 1989; Haggart, 1988) and is found throughout the year in the phytoplankton community (Roelofs, 1984). Therefore, considering the nature of the palaeoenvironment under study, *Paralia sulcata* can be regarded as primarily planktonic and probably represents the transport of marine diatoms by extreme spring tides into predominantly freshwater/brackish environments (Cameron, 1997). *Paralia sulcata* should be interpreted as allochthonous but since they are capable of living in quiet coastal water habitats with fluctuating salinity (Zong, 1997b), they may not have been transported far from their source.

In the USA, Holland, and in the Holocene and present day sediments in the Thames estuary, marine planktonic diatom taxa such as *Paralia sulcata* are the most common allochthonous species which are transported by tidal currents into the internal part of tidal inlets and estuaries in large numbers (e.g. Devoy, 1977; 1979; Cooper, 1995; Long, 1992; Nelson & Kashima, 1993; Juggins, 1992; Plater & Shennan, 1992; Robinson, 1982; 1993; Vos & de Wolf, 1993b; 1994; 1997; Zong, 1992). The behaviour of marine tychoplanktonic species may therefore be largely determined by sediment influx (Vos & de Wolf, 1994; Denys & de Wolf, 1999), and it is possible that much of the *Paralia sulcata* count recorded at the site was washed in with the clastic sediments recorded in the stratigraphy. Denys & Baeteman (1995) suggest the occurrence of *Paralia sulcata* in such situations (whilst probably allochthonous), indicates regular tidal inundation.

A predominance of tide transported diatoms is often found in deposits in or close to tidal channels and inlets, since in such environments, a combination of poor light conditions and high current velocities creates turbid conditions unfavourable for the development of an autochthonous assemblage (Vos & de Wolf, 1994; 1997; Spencer *et al.*, 1998b). However, the assemblage within EKD-2 consists of a wide range of autochthonous benthic and epiphytic taxa, which indicates that the incidence of *Paralia sulcata* is related to inwashing of brackish estuarine water and not enhancement of diatom proportions due to turbid conditions or differential preservation.

Mesohalobous benthic taxa occurring within EKD-2 include *Navicula peregrina* and *Nitzschia navicularis*. These are epipelic species that live within the muddy sediments in the intertidal zone (Vos & de Wolf, 1988; 1993a). Marine-brackish epiphytic species such as *Rhaphoneis amphiceros* (de Wolf, 1993) are more important towards the close of the zone, perhaps indicating an expansion of vegetated saltmarsh environments close to the site. However, this evidence is equivocal, as Vos & de Wolf (1993a; 1997) define *Rhaphoneis amphiceros* as a tychoplanktonic species which suggests this taxa may have a similar origin to *Paralia sulcata*. Aerophilous species such as *Calonies bacillum* occur throughout and indicate the presence of aerial freshwater peat communities in the vicinity which are not regularly inundated by brackish water (Denys & Verbruggen, 1989).

EKD-3 <*c*.3400 BP (<*c*.3700 cal. yrs BP)

An apparent reduction in marine influence is indicated in this ldaz as frequencies of *Paralia sulcata* decline and those of *Cyclotella striata* increase slightly, as does the proportion of benthic taxa. Polyhalobous and mesohalobous species such as *Diploneis didyma, Diploneis smithii*, and *Nitzschia navicularis* are probably components of the autochthonous assemblage and represent intertidal mudflat conditions within low saltmarsh environments at the site (Denys & Baeteman, 1995). The lack of epiphytic diatom species may suggest the coring site was sparsely vegetated at this time. However, the use of foraminifera may be required to clarify this interpretation as they are often a better means of indicating saltmarsh conditions locally at a site because the diatom data can be blurred by taphonomic processes (Evans *et al.*, submitted).

EKD-4 <*c*.3400 BP (<*c*.3700 cal. yrs BP)

Further reduction in tidal activity is indicated in this ldaz by the lack of marine or brackish planktonic species and the dominance of benthic taxa. Supratidal conditions are indicated by the high frequencies of *Diploneis interrupta* (Vos & de Wolf, 1993a, 1997). Vos & de Wolf (1993a) list this diatom as aerophilous, and common at the higher parts of brackish tidal marshes beyond the reach of the regular tide and flooded only occasionally during spring and storms tides (Vos & de Wolf, 1997). Denys & Baeteman (1995) suggest *Diplonies interrupta* indicates rather dry, subaerially exposed environments with limited tidal influence. This suggests drying out of the local environment associated with the removal of marine conditions. Vos & de Wolf (1997) suggest that diatoms such as *Diploneis interrupta* are often found in wet humic and peaty clays where the natural vegetation is not broken down completely. Such a scenario is consistent with the pollen evidence from EKP-6 where a reduction in tidal inundation and relative drying of the environment appears to enable the expansion of Poaceae, Cyperaceae, and *Salix* communities onto the river mudflats and saltmarshes which were previously perhaps less densely vegetated due to flooding and marine influence.

Relatively saline conditions close by are indicated by the presence of polyhalobous and mesohalobous benthic species such as *Diplonies didyma*, *Diploneis smithii*, *Navicula peregrina* and *Nitzschia navicularis*. High proportions of epipelic species such as this with low frequencies of epiphytic taxa are indicative of exposed muddy substrates within the intertidal zone close to the site.

Diatoms from the oxidised overlying clastic deposits are either entirely absent or fragmented. Diatoms become dissolved or broken as a result of recent/subrecent pedogenic processes within such sediments (Mayer *et al.*, 1991; Vos & de Wolf, 1997; Spencer *et al.*, 1998a; 1998b; Plater *et al.*, 1999b).

8.7 Eskamhorn Palaeoenvironmental Summary

The biostratigraphic data presented above give an insight into the processes influencing wetland development. The changing local pollen assemblages enable estimates to be made as to the vegetation dynamics of the floodplain flora. Figure 8.10 shows a summary of the timing and character of the floodplain vegetation changes at Eskamhorn, and the diatom data also highlight the palaeoenvironmental conditions subsequent to, and during estuarine flooding.

The local environment prior to *c.*6400 BP (*c.*7500 to *c.*7000 cal. yrs BP) is characterised by a streamside wetland community of *Alnus glutinosa* and Cyperaceae. Diatom evidence indicates flowing fresh water at the site, or in close proximity. Rising water tables result in the paludification of the floodplain and an expansion of *Alnus glutinosa* dominated wetland carr communities over the valley floor. Evidence suggests the dryland vegetation colonising the valley slopes was dominated by *Tilia*, *Quercus*, *Ulmus*, with *Corylus avellana* either as an understorey shrub or floodplain edge community.

Declining *Ulmus* pollen percentages are dated to 5440±45 BP (6304 to 6216 cal yrs. BP). An early date for this phenomena is registered from many lowland sites, with upland areas of north-east England recording later ages. Little evidence is forthcoming to support human induced clearance so perhaps the spread of a pathogen is a possible alternative.

A rise in water table is evident during EKP-4a as *Salix* carr rises to dominance at the expense of *Alnus glutinosa*. This event is dated to 3670±45 BP (4130 to 3852 cal yrs. BP). A further increase in wetness is indicated in lpaz EKP-4b and dated to 3440±45 BP (3828 to 3572 cal. yrs BP). Carr vegetation is reduced, and fen meadow conditions prevail, dominated by *Filipendula*. Expansion of the floodplain wetland is suggested due to the decline in extra-local dryland tree taxa.

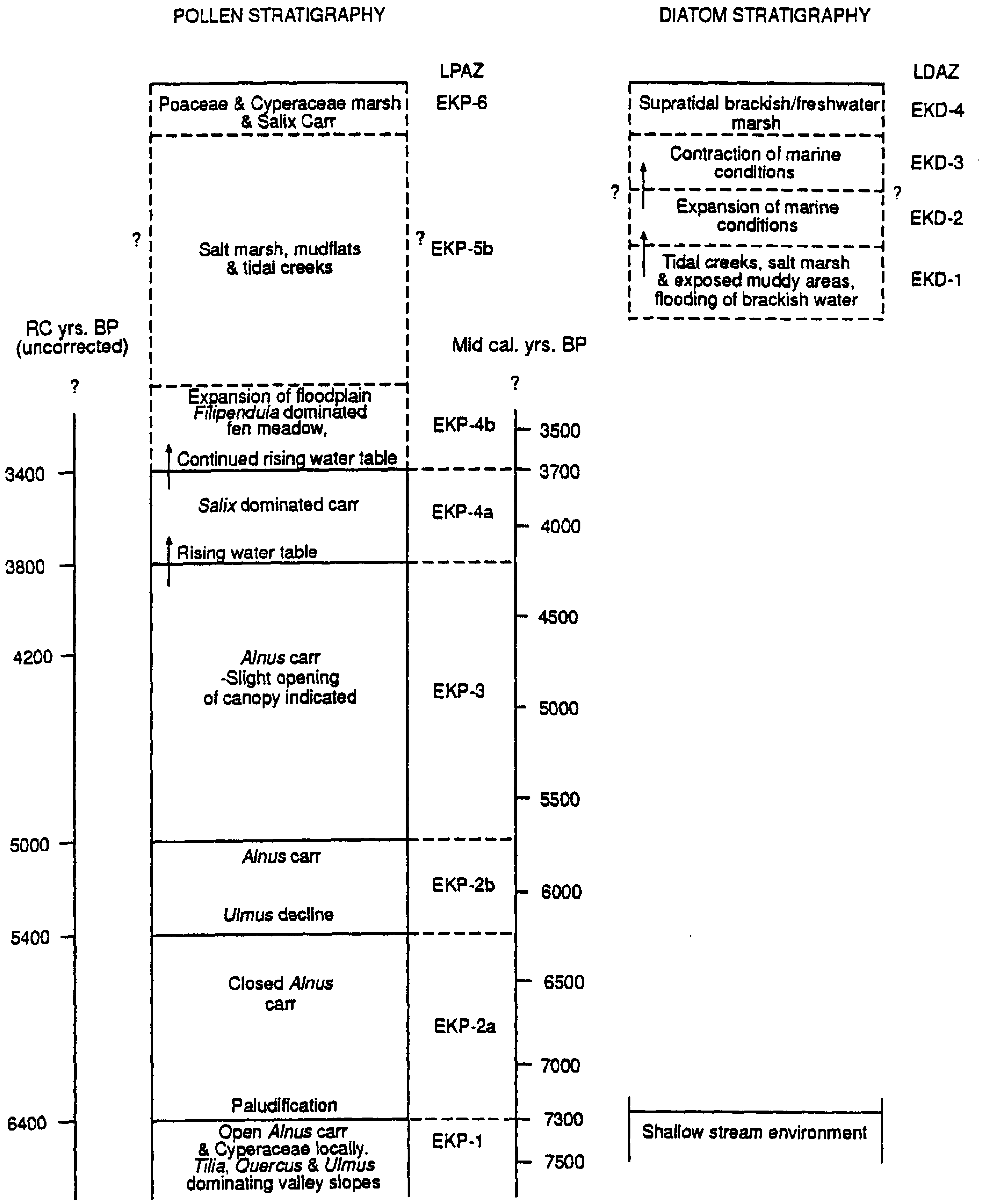


Figure 8.10 Eskamhorn palaeoenvironmental summary diagram

Continued rising water levels initiate a change in stratigraphy as the site is flooded and organic clay dominates the sedimentary sequence. The diatom record from this unit is dominated by marine and brackish planktonic and benthic taxa, which suggests frequent flooding of estuarine water and the development of tidal creeks within a low saltmarsh environment with abundant exposed areas of mudflat. The pollen evidence suggests the presence of Poaceae dominated saltmarshes close to the coring site. The persistence of salt intolerant wetland herbs in the vicinity suggests the input of freshwater is also maintained from terrestrial sources. This fen vegetation probably colonised higher areas at the back of the saltmarsh, adjacent to the valley sides.

Estuarine conditions expand in ldaz EKD-2 as marine planktonic species dominate. Consistent presence of aerophilous diatom taxa suggests the environment was not permanently submerged although the presence of high frequencies of planktonic and benthic taxa indicates the area was inundated for considerable parts of the tidal cycle.

Thereafter, the influence of marine activity begins to wane. Frequencies of aerophilous diatom taxa indicate supratidal conditions and the pollen evidence suggests an expansion of freshwater reedswamp communities and *Salix* carr close to the site.

The sequence of events at Eskamhorn are apparently related to changes in water level associated with changes in estuarine regime. Important information has been gathered as to the nature of the vegetation response during the onset, continuation and withdrawal of estuarine conditions (positive sea-level tendency) recorded from the litho- and biostratigraphy at Eskamhorn. This suggests there is good evidence for sea-level change recorded in the sediments at Eskamhorn. This is the subject of further discussion in Chapter 10.

Chapter 9 Hook Lane Site Report

9.1 Introduction

The results in this chapter are concerned with the palaeoecological development of the most seaward site in this study. Hook Lane is situated just downstream of the confluence of the river Aire with the Ouse north of Goole (Figure 3.1). This site was chosen because the sediments were deposited under more direct estuarine influence, thus, establishing a contrasting environmental gradient from the freshwater fluvial dominated regime upstream at Hirst Courtney (Chapter 7). A lithostratigraphic survey was undertaken to identify a suitable coring site, from which a borehole could be collected and sampled for detailed micropalaeontological analysis. The findings of the palaeoecological analysis (based mainly on pollen and diatom studies) are combined with radiocarbon dates to enable the palaeoenvironmental history of the site to be established. The various controls on sedimentation and vegetation dynamics are discussed.

9.2 Lithostratigraphy

A transect of 11 boreholes is utilised to establish the main sedimentary units characteristic of the site and establish the extent of the wetland area (see Figure 6.5). A summary of the lithological findings at Hook Lane has been given in section 6.2.3 and Figure 6.6. A more detailed description of the lithostratigraphy is presented here, which provides additional information regarding the nature and variability of the sedimentary sequence (see Figure 9.1). The sedimentary fill at Hook Lane is thicker than at the previous sites (up to 1060 cm in HK 1) and is of particular interest owing to the presence of a previously unsampled depositional unit, which separates the main floodplain peat into two units.

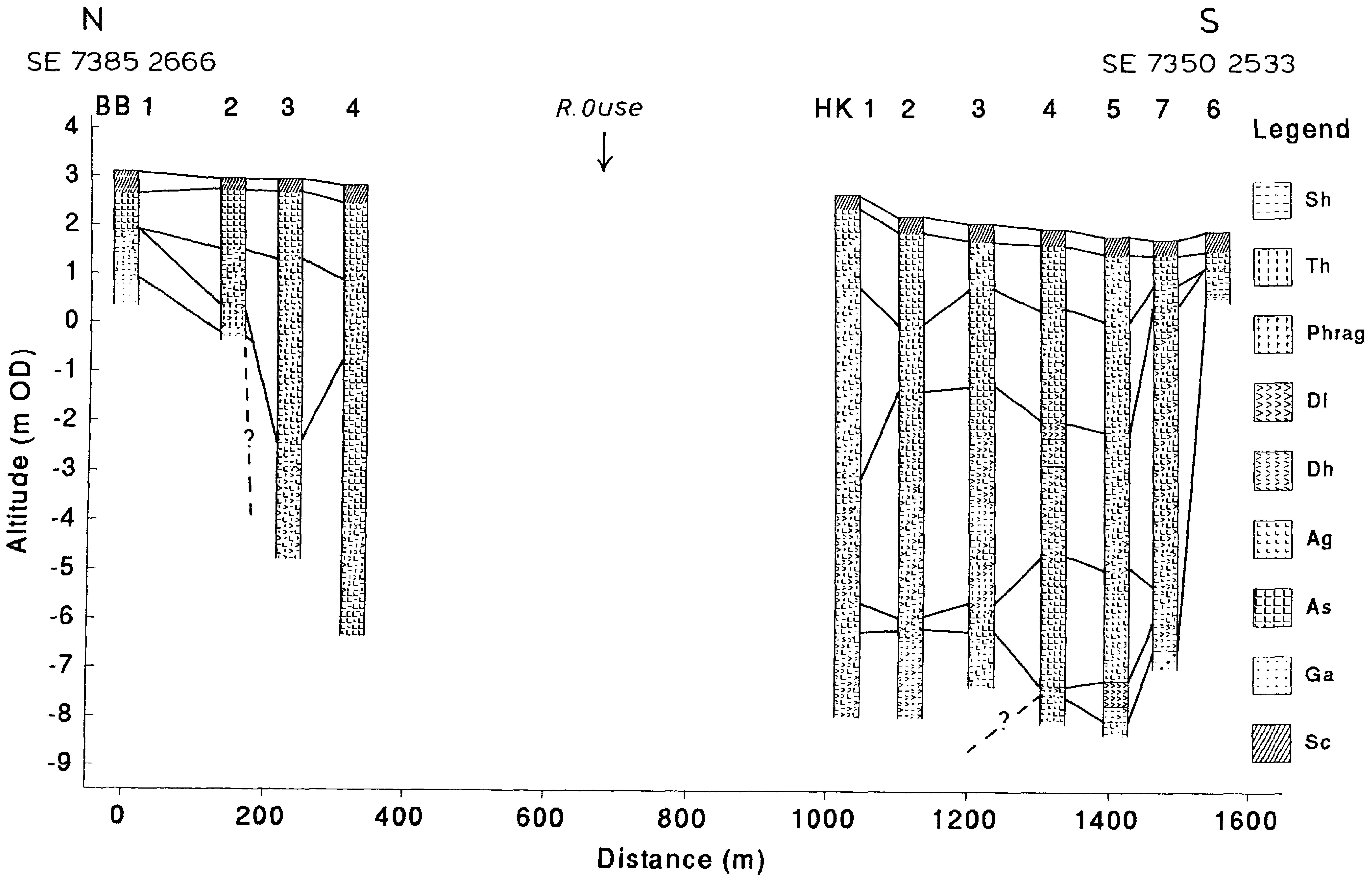


Figure 9.1 Lithostratigraphy of the cross valley transect at Hook Lane in the lower Aire valley (symbols follow Troels-Smith (1955))

The valley fill is asymmetric, with a deeper sedimentary sequence at the south side of the transect, which is undisturbed, and typical of the lower part of the study area. This sequence is not well reproduced by the transect near Boothferry Bridge on the north of the floodplain (see Figure 6.5, 6.6 and 9.1), where the sedimentary fill has been reworked and largely replaced by channel deposits.

A basal inorganic deposit is consistently found in the shallower boreholes flanking the valley sides. Close to the floodplain edge, this unit generally consists of light brown-pink-grey sand (BB 1 and 2; HK 6), and is recorded at relatively high altitudes (between -0.80 and +0.49 m OD) as the buried valley sides rise to the surface. At both sides of the floodplain, the valley sides are incised relatively steeply so that recovery of the basal unit in boreholes away from the valley edges was only possible in HK 4, 5 and 7. The basal sediments in core HK 7 comprise medium and coarse sands, but in boreholes HK 4 and 5, sand is recorded at -8.09 and -8.34 m OD respectively, and the sediments then fine upwards into a stiff blue-grey-pink silty clay with sand.

The deepest sediment in boreholes HK 1, 2, and 3, down to a maximum depth of -8.01 m OD in HK 1, is brown clayey wood peat. This lower peat deposit overlies the basal clastic unit in boreholes HK 4, 5, and 7, with a contact altitude between -8.04 m OD in HK 5 to -6.61 m OD in HK 7. The peat consisted of a matrix of *Substantia humosa* (*sensu* Troels-Smith, 1955) with clay, abundant woody detritus, and occasional herbaceous plant stems, and is considerably thinner in cores HK 4, 5 and 7 than in boreholes HK 1, 2 and 3 which bottom out on the valley floor.

Overlying the lower peat at Hook Lane, between -7.36 in HK 4 and -6.07 m OD in HK 7, is a brown grey humic silty clay with distinctive sedimentary characteristics. Within this facies, lighter, beige coloured bands are evident, *c.*1 to 2 cm thick, particularly at the base, and the sediment appears finely laminated by leafy detritus. These single leaf layers often occur at millimetre intervals, and are particularly apparent when the sediment is examined in the field for description, as the sediment tends to break along the plane of a leaf. Large wood detritus is also recorded in this unit, along with occasional herbaceous stems. This sedimentary unit is confined to the lower reaches of the study area, and can be traced as far as core RC 38 in the long profile transect (see Figure 6.12 and 6.13). This sedimentary unit is similar to the fluviolagoonal facies described by Van der Woude (1983; 1984) from the Rhine-Meuse perimarine system in The Netherlands.

Overlying this humic clay unit between -6.24 m OD in HK 2 and -4.82 in HK 4 is a clayey wood peat, of similar composition to the main floodplain peats from other sites in the study area (see section 6.4). Large wood remains are common with herbaceous *turfa* and detritus, and the peat matrix is characterised by a variable but generally high clay content. Where this peat does not overlie the humic clays, it mantles the valley sides in cores BB 1 and 2 and HK 6 at the margins of the floodplain, at altitudes between -0.24 m OD in BB 2 to +0.54 and +0.87 m OD in HK 6 and BB 1.

The wood peat layer is overlain by a brown-grey organic silty clay in all boreholes except BB 1 and HK 6. This unit is generally thicker towards the present river Ouse channel, where it overlies peat at -3.05 m OD. The contact altitude rises and the deposit thins towards the floodplain edge. In boreholes HK 7 and BB 2, it is recorded at +0.32 m OD and +0.36 m OD respectively, but is absent from the most peripheral cores, BB 1 and HK 6. This silty clay is characterised by a high organic content, with woody detritus and monocotyledonous stems (*cf. Phragmites*) recorded throughout. The organic content gradually decreases towards the top of the unit, which influences a subtle colour change from brown-grey to blue-grey. Orange oxidation stains are also apparent.

This organic clay layer gradually gives way to an orange-brown mottled silty clay which is recorded as the uppermost unit in all the boreholes in the transect between -0.04 m OD in HK 2 and +1.83 m OD in BB 1. Recent and subrecent *turfa* rootlets are recorded throughout this layer, which is heavily iron stained and contains a significant sand component in cores HK 3 and 5. Although these boreholes are collected from an area known to have been warped (*cf.* Gaunt, 1994) no visible sedimentary evidence, such as an abrupt contact with the underlying unit or distinct laminations, is preserved in support of this (see section 6.3).

This sedimentary succession is not replicated in boreholes BB 3 and 4 on the north floodplain. Here, the lowermost unit, recorded down to -6.29 m OD in BB 4, is a brown-grey moderately organic silty clay, with subtle colour changes from dark to light bands of clay with peaty detritus. Although the sediments are not laminated with bands of silt or fine sand (*cf.* channel sediments in RC 53, 54 and 55, section 6.3; Figure 6.12 and 6.13), much of the organic material appeared reworked, which suggests that these sediments represent a part of a low energy channel that eroded away pre-existing peat deposits as it migrated northwards. Overlying these fluvial deposits is a blue-grey silty clay containing monocotyledonous stems, *turfa*, and occasional woody detritus. This is overlain by the uppermost oxidised silty clays. These two units are comparable with the sedimentary horizons described above that are found along the length of the floodplain.

Following the lithostratigraphic investigation at Hook Lane, a sample borehole was chosen from a representative location (HK 4P), and sampled for palaeoecological and radiocarbon analysis using a Stitz piston corer. The detailed sediment description of the major lithological units is shown in Table 9.1, and the results of the biostratigraphic analyses are presented in sections 9.4, 9.5 and 9.6.

Site: Hook Lane, River Ouse valley	**Core**: HK 4P

Grid reference: SE 73522554	**Date of fieldwork**: September 1997
Latitude: 53° 43' N	**Sampling method**: Stitz Piston Corer
Longitude: 1° 53' W (Greenwich Meridian)	**Bench mark datum**: Gatepost SE 73182561
Collector: J. R. Kirby / N. Tunstall	**Ground Altitude**: +1.91 m OD

Observed depth (cm)	**Altitude (m OD)**	**Troels-Smith Sediment Description** **nig., strf., elas., sic., lim. sup. (Troels-Smith, 1955)**
100 to 160	+0.91 to +0.31	Orange-brown-grey mottled silty clay. Trace of peaty detritus, *turfa* rootlets, and vertically aligned monocotyledonous stems. As3, Ag1, Lf++, Sh+, Dg+, Dl+, Th^0+; 3, 0, 3, 0, 0.
160 to 270	+0.31 to -0.79	Light blue-grey organic silty clay with woody detritus and herbaceous *turfa* (rootlets and reed stems *cf. Phragmites*). As3, Ag1, Sh+, Dl++, Dh+, Dg+, Th^2+, $Th^{Phrag.}$+; 3, 0, 3, 0, 0.
270 to 396	-0.72 to -2.05	Grey organic clay with abundant woody detritus and occasional herbaceous *turfa*. As3, Ag1, Dl++, Dh+, Dg+, Th^2+, Sh+; 3, 0, 3, 0, 0.

* **Continued over**

Table 9.1 Lithostratigraphic details from HK 4P described using the Troels-Smith (1955) terminology

Observed depth (cm)	Altitude (m OD)	Troels-Smith Sediment Description nig., strf., elas., sic., lim. sup. (Troels-Smith, 1955)
396 to 481	-2.05 to -2.90	Brown-grey organic clay with abundant detrital wood remains, occasional herbaceous *turfa*, and well humified organic matrix. As3, Sh1, Ag+, Dl++, Dg+, Dh+, Th^2+; 3, 0, 2, 0, 0.
481 to 673	-2.90 to -4.82	Variable grey-brown clayey peat with abundant large woody detritus. 2 light grey clay bands evident towards base (between 649 and 660 cm) Sh2, As2, Dl++, Dh+; 3+, 0, 3, 0, 0.
673 to 773	-4.82 to -5.82	Grey organic clay with abundant large woody detritus. Well laminated with leafy detritus and flecks of chalky blue vivianite abundant (*cf.* Bennett, 1983b; Denys, 1985). As3, Ag1, Dl++, Sh++, Dg+; 3, 3, 3, 0, 0.
773 to 821	-5.82 to -6.30	Brown-grey organic clay with lighter beige coloured bands of silty clay. Fine leaf laminations and some flecks of blue vivianite around occasional woody detritus. As3, Ag1, Sh++, Dl+, Dg+; 3, 4, 4, 0, 0.
821 to 927	-6.30 to -7.36	Brown-grey organic clay with detrital wood remains and slightly laminated leaf detritus. As3, Sh1, Ag+, Dl++, Dg+, Dh+; 3, 0+, 3, 0, 0.
927 to 947	-7.36 to -7.56	Dark brown well humified peat with some clay. Sh3, As1+, Dl+; 4, 0, 3, 0, 0.
947 to 994	-7.56 to -8.03	Blue-grey-pink stiff silty clay with sand and some flecks of chalky light blue vivianite. As2, Ag2, Ga++; 2, 3, 3, 0, 0.
994 to 1000	-8.03 to -8.09	Light brown sand. Ga4; 2+, 0, 3, 0, 0.

Table 9.1 Lithostratigraphic details from HK 4P described using the Troels-Smith (1955) terminology

9.3 Radiocarbon Dating

Selecting horizons for radiocarbon dating from Hook Lane was problematical for several reasons. First, the biostratigraphic analysis revealed that the entire sequence contained moderately high frequencies of pre-Quaternary spores. The results at Eskamhorn (section 8.3) show that as long as there is sufficient *in situ* carbon in the sample, the presence of pre-Quaternary microfossils (which may make up a small proportion of the sample) does not necessarily have a significant effect on the age of the sample. However, ageing effects resulting from fluvial transport of older reworked geological organic material is a cause for concern considering the low amounts of carbon in some of the samples proposed for dating the sequence at Hook Lane (*cf.* Blong & Gillespie, 1978; Olsson, 1972; 1979; Schoute *et al.*, 1981). Second, the sedimentary sequence sampled is representative of very different depositional environments, including fen carr, lagoon and saltmarsh. Therefore, the dating strategy has for this site had to be revised and is discussed below.

Following the recommendations of the NERC Radiocarbon Dating Facility Steering Committee, wood macrofossils were sampled from three of the proposed dating horizons for AMS dating. Samples of this nature are not likely to have been mechanically contaminated in the same manner as bulk samples and could be expected to give accurate radiocarbon ages in the absence of a hard water error (see below). In addition, bulk samples were collected from sediment above and below the wood macrofossil in the core. These samples contained a high clay content and abundant pre-Quaternary spores, and were dated using conventional methods.

The use of large wood macrofossils is not ideal, mainly due to the uncertainty regarding their provenance and the depositional processes involved in their burial (see section 5.9.2). However, in the absence of other terrestrial macrofossil remains (e.g. seeds or fruits), the coupling of dating samples in this way was the only method of assessing the accuracy of the age estimates of bulk sediment samples from this site. The results of a similar study by Törnqvist *et al.* (1992) suggest that there is little risk of using reworked (and, hence, too old) macrofossils for ^{14}C dating, and Brown (1997) suggests wood branches decay rapidly on the floodplain surface which indicates the incorporation of old wood into peat is unlikely (although see

Streif, 1972). This suggests that any age discrepancy may be related to mechanical contamination of the bulk sample. The results of these age determinations are reported in Table 9.2, discussed, and interpreted following the recommendations of Pilcher (1991). Decisions as to the significance of each are made with respect to the depositional relationship between the sample and the peat forming environment, the nature of the sample, and the biostratigraphic record.

The estimates obtained show internal inconsistencies (Table 9.2). The most striking anomaly is the inversion at the base of the sequence. The bulk sample of humic clay (SRR-6394) produced a date that is *c.*1400 years older than the AMS estimate on the wood sample (AA-32268) from the same horizon. The age estimate for SRR-6394 is also over 200 radiocarbon years older than the age obtained from a peat sample at the base of the core (SRR-6395), which came from *c.*180 cm lower.

Clearly, one or more of these age estimates are erroneous, and it is reasoned here that the bulk date from the humic clay (SRR-6394) may be too old. There are two possible reasons for this anomaly. The first, and the favoured explanation, is that the sample is adversely affected by mechanical contamination (NERC Steering Committee, pers. comm.). Considering the depositional context of the sample, is probable that fluvial and estuarine processes not only resulted in the admixture of clastic sediment but also the deposition of reworked fine grained organic particles, probably of Carboniferous age (*cf.* Gaunt, 1994). Pre-Quaternary spores are common in sediments throughout this region, derived from the erosion of Pleistocene sediments containing coal fragments crushed by glacial transport (Walker *et al.*, 1993). This type of contamination is likely to be more significant in this sample due to the low (LOI averages *c.*20% dry weight) proportion of carbon within this unit (Olsson, 1972; 1979). In The Netherlands, it has been demonstrated that humic clays can return radiocarbon age estimates which are too old (for the stratigraphic position) due to this effect (e.g. Shoute *et al.*, 1981; Törnqvist *et al.*, 1992).

A second explanation for the anomalously old estimate obtained may be that the sample has been affected by hard water error. This effect is not considered likely due to the absence of carbonate bedrock in the vicinity. Also, it is suggested that this sediment formed in a tidally influenced, relatively shallow fluviolagoonal environment (see section 9.5 and 9.6), where there would probably be a rapid mixing of

CO_2 between the water and the atmosphere (Törnqvist *et al.*, 1992). The bulk date from the humic clay (SRR-6394) is considered unreliable on these grounds.

In addition, it is apparent that the wood macrofossil AMS date in each sample pair is consistently younger than the bulk sediment age, even where the *in situ* organic content of the sample is relatively high (e.g. SRR-6391 and AA-32269). When the coupled dates are compared with the biostratigraphic record (section 9.5), it is argued that the AMS determinations from the wood samples appear too young (for the stratigraphic level they coincide with) and the bulk dates from the floodplain peat reveal expected ages. Therefore, a justifiable decision has to be made as which age series to use for the age-depth model and biostratigraphic interpretation. Is it more appropriate to use AMS dates from wood, where the mean age of the organic material dated is not in question, but the stratigraphic integrity is, or to use bulk samples which certainly contain *in situ* organic material from the selected horizon, but also contain mechanical contaminants that may age the sample? Another alternative is to simply reduce the error by averaging the combined ages of the dated samples. These options are discussed below.

Given the uncertainty regarding the depositional and taphonomic processes surrounding the sequence of events between the death of a tree and its subsequent burial within a peat, it is possible that the pieces of dated wood are intrusive. The relatively large size of some of the branches sampled for AMS dating (20 cm diameter for AA-32270) suggests that at least some of the shrubs were reasonably large specimens, and it is necessary to consider some of the pathways by which wood macrofossils may be incorporated into a floodplain fill. It is highly unlikely that when a woody shrub or tree dies, it rests horizontally on the peat surface it is contemporaneous with and then quickly becomes enveloped in peat of a similar age. It is probable that as a tree/woody shrub dies, it sinks into the wetland and many branches will become embedded in peat stratigraphically lower, and therefore older than, the age of the tree. For this scenario to be a full explanation of anomalies observed at this site, however, some of the wood branches must have pushed considerable distances into the sediment. If the AMS date on the *Salix* wood in the humic lagoonal clay (AA-32268) is accepted as too young, and the bulk date from the peat at above it is correct (SRR-6393), the wood must have been forced down at least 1 m. This may be possible given the soft nature of the sediment although no visible signs of sedimentary disturbance were evident in the cores, which may perhaps be expected if such intrusion has occurred. Another possibility is that within fen carr environments, as water washes over the floodplain surface, sediment will be scoured from beneath wood branches, causing them to sink to lower depths, and also that peat may grow up around a trunk, incorporating younger wood with older peat. It is also likely that the woody roots from the fen carr vegetation would penetrate into the floodplain peats resulting in younger wood in older deposits.

Unfortunately, there is a lack of information regarding such depositional processes or the taphonomy concerning the incorporation of trees into fen carr peats, to determine whether such a scenario is feasible. The averaging of the paired ages has not been considered as a viable option because this method is as likely to increase as reduce any errors, which would result in a low resolution chronology. Therefore, on the balance of evidence discussed above, at this stage, the 3 AMS dates (AA-32270; AA-32269, AA-32268) have been rejected, and the bulk sediment ages have tentatively been used as a basis for the chronology.

9.3.1 Age-Depth Model

Owing to the inconsistencies regarding the radiocarbon dates (discussed above), only the conventional bulk age determinations (excepting SRR-6394 from the humic clay) are used for the calculation of sedimentation rates (SR, given in cm yr^{-1}) in the age-depth model (Figure 9.2). The determination of the SR between the dated samples for both uncorrected and calibrated ages enables the calculation of deposition times using linear interpolation (see section 5.9.4 for methods). This provides a means of estimating the age range of the biostratigraphic assemblage zones. However, interpolation is not used where a significant change in sedimentary environment is apparent in the stratigraphy (i.e. between the basal peat and the lagoonal clay). This is due to the likelihood of variations in sedimentation rate associated with such a stratigraphic change.

Laboratory Code	Sample Material	Depth (cm)	Altitude (m OD)	Age (^{14}C yr BP±1σ)	Calibrated Age Range * (cal. yrs BP ± 2σ) max	(cal. ages)	min	Mid Cal. Age (cal. yrs BP±1σ)
SRR-6389	Organic clay	300 to 305	-1.09 to -1.14	3950±45	4517	(4410)	4257	4387±130
•SRR-6390	Organic clay	390 to 395	-1.99 to -2.04	4555±45	5439	(5290)	5043	5241±198
•AA-32270	Waterlogged wood (*Salix* sp.)	396 to 400	-2.05 to -2.09	3885±55	4433	(4340, 4330, 4290)	4096	4265±169
@SRR-6391	Clayey wood peat	479 to 488	-2.88 to -2.97	4850±45	5655	(5593)	5478	5567±89
@AA-32269	Waterlogged wood (*Vibernum cf. opulus*)	482 to 486	-2.91 to -2.95	4185±50	4846	(4820, 4750, 4720, 4670, 4650)	4538	4692±154
SRR-6392	Clayey wood peat	585 to 590	-3.94 to -3.99	5655±45	6528	(6416)	6313	6421±108
SRR-6393	Clayey wood peat	665 to 670	-4.74 to -4.79	5965±45	6889	(6842, 6835, 6789)	6719	6804±85
#SRR-6394	Humic clay	763 to 773	-5.72 to -5.82	7225±50	8120	(7970)	7913	8017±104
#AA-32268	Waterlogged wood (*Salix* sp.)	765 to 770	-5.74 to -5.79	5875±55	6850	(6720)	6547	6699±152
SRR-6395	Clayey wood peat	942 to 947	-7.51 to -7.56	7030±45	7914	(7878, 7865, 7812)	7699	7807±108

•, @ and # denotes coupled dates

* Calibration follows Stuiver & Reimer (1993)

Table 9.2 Radiocarbon dates and calibrated age estimates from Hook Lane

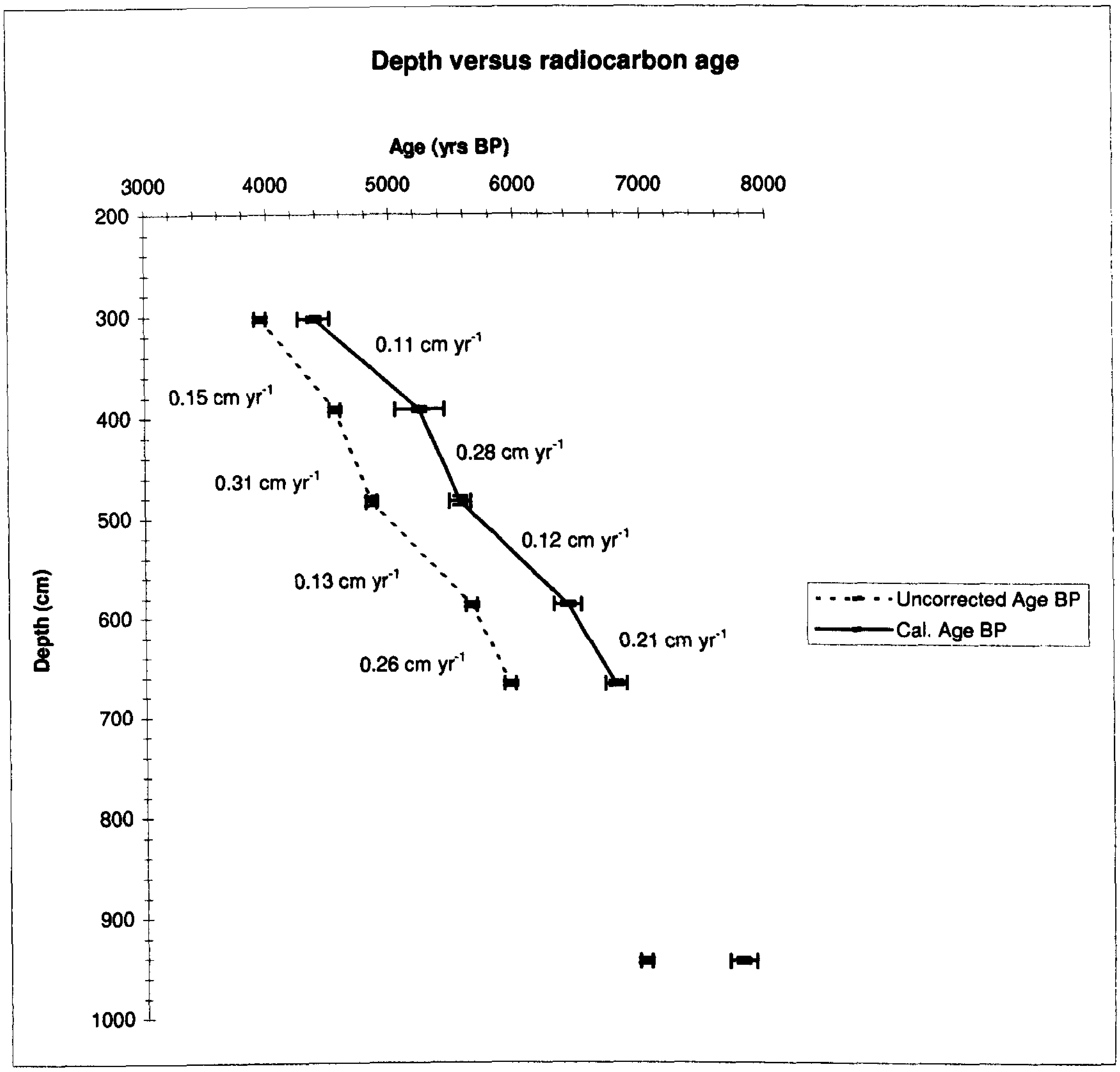


Figure 9.2 Hook Lane age-depth model. Calibrated ages are plotted using the mid cal. age (after Bennett, 1994b) using the method outlined in section 5.9.4 with error bars representing the 2σ range of the age either side of the mid-point. Error bars for the uncalibrated BP ages show the laboratory counting error. Depth error bars on both sets of data of represents the sample thickness.

9.4 Biostratigraphy

9.4.1 Plant and Wood Macrofossils

Depth Range (cm)	Macrofossils	Depth Range (cm)	Macrofossils
396 to 426	*Salix* sp.	615	*Corylus avellana* nut
482 to 486	*Vibernum (cf.) opulus*	725 to 734	*Salix* sp.
520 to 527	*Quercus*	765 to 770	*Salix* sp.
544 to 547	*Fraxinus excelsior*		

Table 9.3 Macrofossils from core HK 4P

In addition to the woody macrofossils listed in Table 9.3, the humic clay horizon between 673 and 927 cm contains distinct laminations of leafy detritus. These leaf layers have been identified as *Salix* spp. (Dr Martyn Waller, pers. comm.).

9.4.2 Pollen Stratigraphy

The pollen analytical results are presented here in a percentage pollen diagram (Figure 9.3) drawn using the TILIA program (Grimm, 1993). The basic sum used is TLP-*Alnus* unless stated otherwise (see section 5.5.6). Local pollen assemblage zones (lpazs) have been determined using CONISS as an objective guide (Grimm, 1987). These lpazs are prefixed HKP and described in the following section. A separate pollen diagram showing rare morphotypes (those not exceeding 1%) is also presented (Figure 9.4).

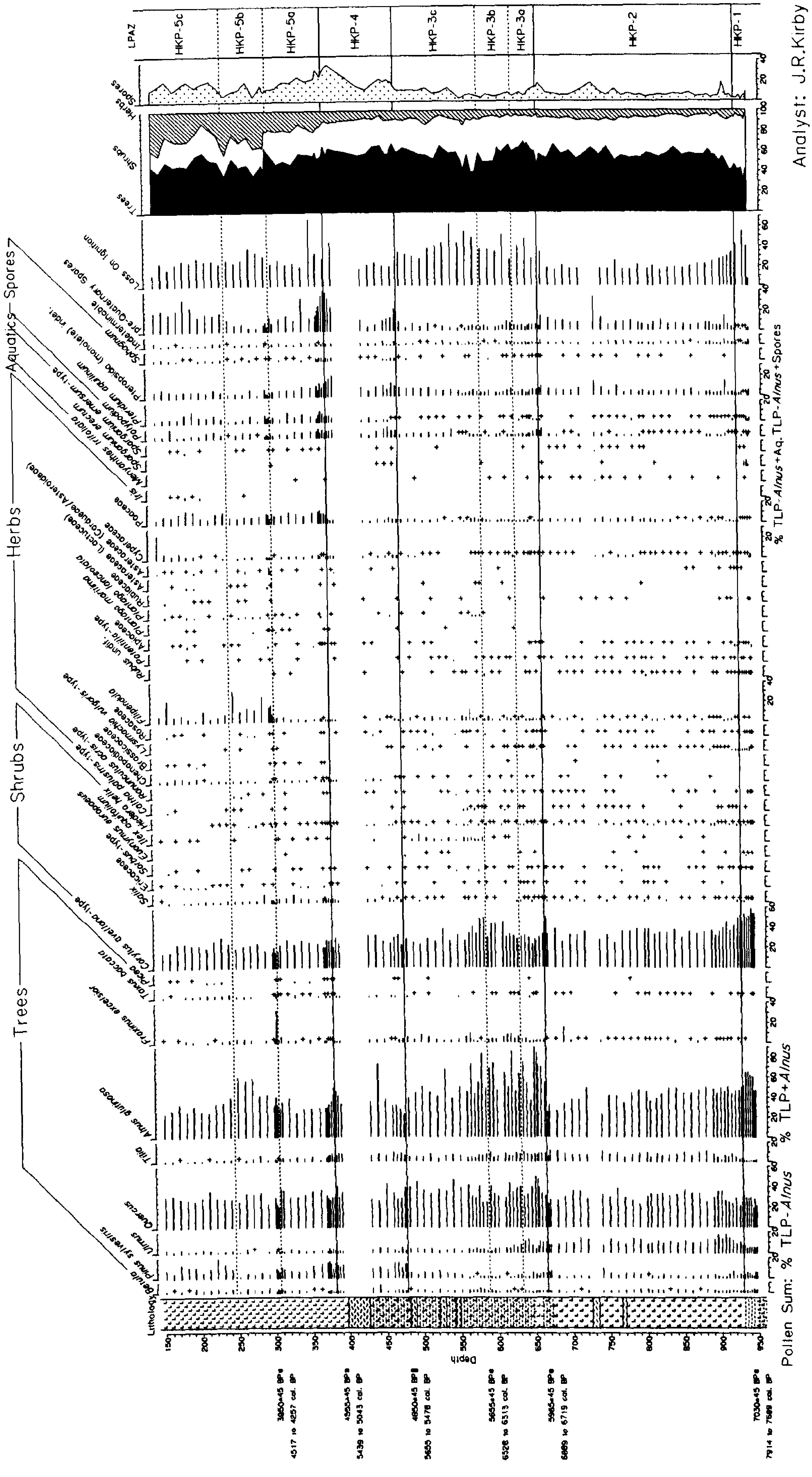


Figure 9.3 Hook Lane percentage pollen diagram (+ represents frequencies <1%)

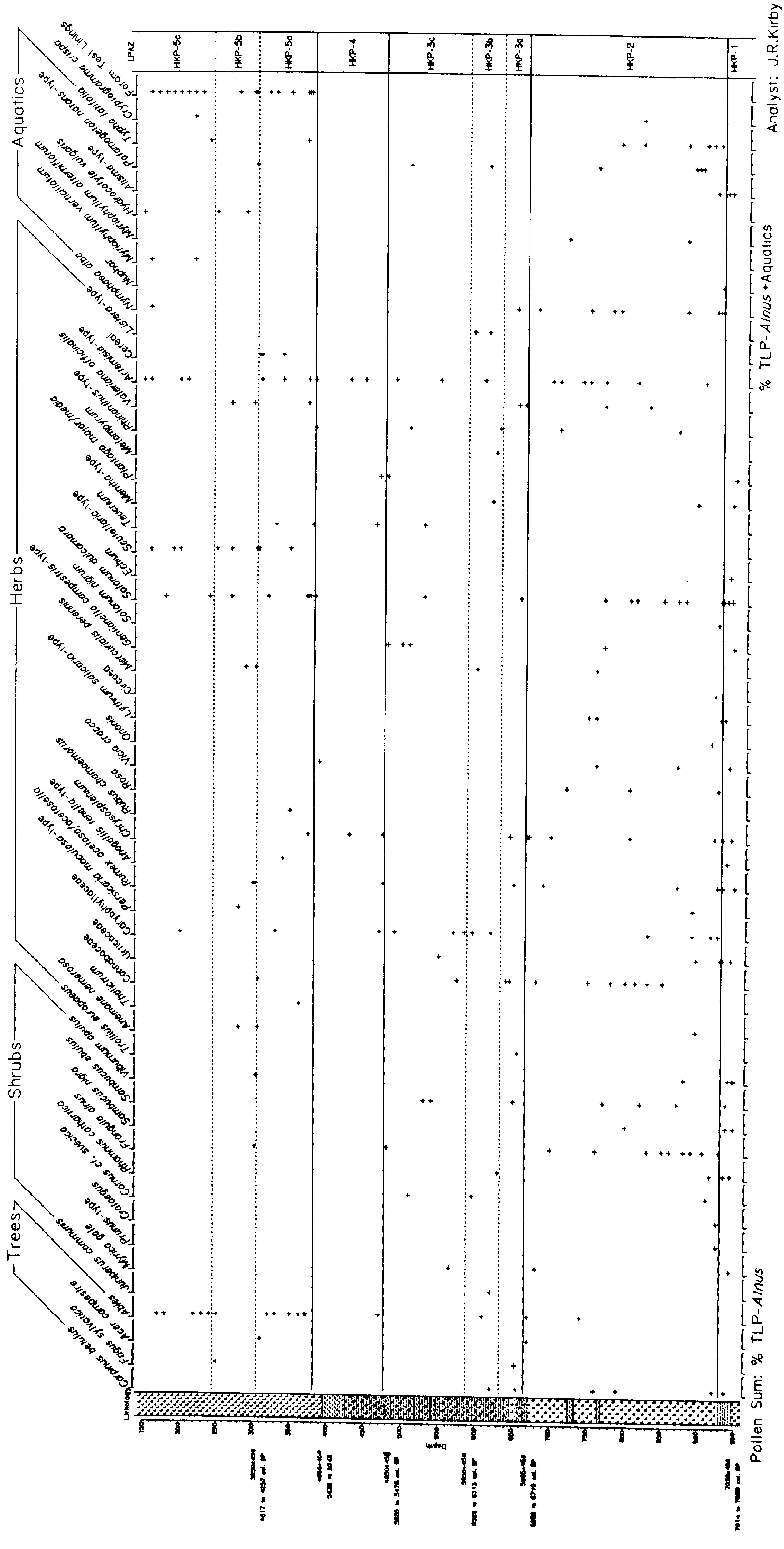


Figure 9.4 Hook Lane percentage pollen diagram showing minor types (+ represents frequencies <1%)

HKP-1 947 to 929.5 cm *c.*7000 BP (*c.*7800 cal. yrs BP)

Figure 9.3 and 9.4

This lpaz is dominated by arboreal (31 to 46%) and shrub (47 to 60%) pollen, with herb pollen comprising between 3 and 9% of the total. *Alnus glutinosa* (46 to 65% TLP+*Alnus*) and *Quercus* (19 to 31%) are the main arboreal pollen types with *Ulmus* (4 to 7%), *Pinus sylvestris* (2 to 6%), *Betula* (1 to 4%), and *Tilia* (1 to 3%) relatively poorly represented morphotypes. A date has been obtained from sediment at the base of the zone between 942 and 947 cm of 7030±45 BP (7914 to 7699 cal. yrs BP).

Shrub pollen is dominated by *Corylus avellana*-type (45 to 58%), with *Salix*, *Sorbus*-type, *Euonymus europaeus*, and *Hedera helix* pollen present at <1%. Herb pollen is represented at the base of the lpaz by Apiaceae and Cyperaceae, which attain frequencies of up to 3%. *Lysimachia vulgaris* pollen occurs between *c.*1 and 3% throughout the lpaz.

Pollen from aquatic plants reaches 6% (TLP-*Alnus*+aquatics) within the lower part of HKP-1 and is mainly comprised of *Sparganium erectum* (up to 6% TLP-*Alnus*+aquatics), although *Menyanthes trifoliata*, *Sparganium emersum*-type, and *Alisma*-type are also recorded at frequencies <1%. Spore morphotypes are not well represented (*c.*1 to 9% TLP-*Alnus*+spores), although *Polypodium* (*c.*1%), *Pteridium aquilinum* (*c.*1 to 4%), and Pteropsida (monolete) indet. (*c.*1 to 6% TLP-*Alnus*+spores) occur in each level. Pre-Quaternary spores are recorded at low frequencies (between *c.*1 and 2% TLP-*Alnus*+PQS). LOI values rise from 7% at the base of the zone to *c.*46% dry weight during the remainder of the lpaz.

HKP-2 929.5 to 666 cm <*c.*7000 to *c.*6000 BP (<*c.*7800 to *c.*6800 cal. yrs BP)

Arboreal pollen taxa (40 to 67%) are better represented in this lpaz despite the decline in percentage frequency of *Alnus glutinosa* (22 to 50% TLP+*Alnus*) from an average of 57% in HKP-1 to 42% TLP+*Alnus* in HKP-2. In particular, *Ulmus* (6 to 18%) and *Tilia* (3 to 12%) pollen percentages rise during the opening of the lpaz, and *Quercus* pollen frequencies (21 to 38%) are also generally higher than in HKP-1. Other pollen types, such as *Betula* (*c.*1 to 4%), *Pinus sylvestris* (*c.*1 to 10%), *Fraxinus excelsior* (<*c.*1 to 15%), and *Taxus baccata* (<*c.*1%) are not important components of the arboreal pollen group (except at 690 cm where *Fraxinus excelsior* pollen frequencies attain 15%).

Frequencies of shrub pollen (26 to 51%) are lower than in the previous zone. Despite a decline in frequency over the HKP-1/2 zone boundary, *Corylus avellana*-type pollen (26 to 50%) remains the most prominent morphotype with other consistently represented pollen taxa such as *Salix*, *Sorbus*-type, and *Hedera helix* generally attaining values less than *c.*1%.

Frequencies of herb pollen types are generally poorly represented, ranging between 4 and 11%. Poaceae (*c.*1 to 5%) pollen is the most frequent with *Lysimachia vulgaris*-type, *Filipendula*, *Rubus* undif., *Potentilla*-type, Apiaceae, and Cyperaceae morphotypes consistently recorded at values generally less than *c.*1%. Pollen from obligate aquatics contributes <*c.*1% (TLP-*Alnus*+spores). However, the diversity of morphotypes present in the lpaz is greater than HKP-1. Pollen from taxa such as *Menyanthes trifoliata*, *Sparganium erectum*, *Nymphaea alba*, and *Typha latifolia* is recorded in numerous levels at frequencies <*c.*1% (TLP-*Alnus*+aquatics), with *Sparganium erectum*, *Myriophyllum alterniflorum*, and *Potamogeton natans*-type also present sporadically.

Pteropsida (monolete) indet. (2 to 15% TLP-*Alnus*+spores) morphotypes are the most important constituent of the spore group with *Polypodium* attaining generally lower frequencies (between <*c.*1 to 8% TLP-*Alnus*+spores). Pre-Quaternary spores typically occur at levels of between 1 and 15% (TLP-*Alnus*+PQS), although a spike of 34% is recorded at 740 cm. LOI declines over the lower zone boundary and rises again at the closing zone boundary, with values ranging between 16 and 34% dry weight. The close of this subzone has been dated to 5965±45 (6889 to 6719 cal. yrs BP).

HKP-3 666 to 477.5 cm *c.*6000 to *c.*4800 BP (*c.*6600 to *c.*5500 cal. yrs BP)

The shifts in arboreal pollen in this zone are not significant enough to justify the separation of the pollen assemblages into different zones. However, there are changes evident in the representation of tree, shrub and herb taxa, which enable the identification of three subzones.

HKP-3a 666 to 632.5 cm *c.*6000 to *c.*5800 BP (*c.*6800 to *c.*6600 cal. yrs BP)

This lpaz is marked by a significant increase in the frequency of arboreal pollen (60 to 70%). This is mainly attributable to the rise in frequency of *Alnus glutinosa* (37 to 90% TLP+*Alnus*) and *Quercus* (31 to 50%) pollen, both of which peak in the centre of the subzone. Values of *Ulmus* (4 to 13) and *Tilia* (7 to

10%) pollen remain moderately high, whereas percentages of *Betula* (1 to 3%), *Pinus sylvestris* (1 to 6%), *Fraxinus excelsior* (1 to 4%), and *Taxus baccata* (>*c*.1%) pollen remain low.

Corylus avellana-type morphotypes (23 to 33%) are the dominant member of the shrub pollen group (25 to 34%) with *Salix* and *Hedera helix* pollen types only occasionally recorded at frequencies greater than 1%. Pollen from herb taxa contributes little to the TLP-*Alnus* sum, with frequencies generally lower than the previous lpaz. The range of morphotypes is similar to those in HKP-2 with Poaceae (*c*.1 to 2%), *Filipendula* (*c*.1 to 4%), Cyperaceae (<*c*.1%), and *Lysimachia vulgaris*-type (<*c*.1%) pollen consistently recorded taxa.

Pollen from aquatic taxa is insignificant and spore frequencies (5 to 14% TLP-*Alnus*+spores) are relatively low, with *Polypodium* (1 to 6% TLP-*Alnus*+spores) and Pteropsida (monolete) indet. (3 to 6% TLP-*Alnus*+spores) the best represented morphotypes. *Pteridium aquilinum* and *Sphagnum* are also consistently recorded at frequencies <c.1% (TLP-*Alnus*+spores). Pre-Quaternary spores are represented at consistently low values (3 to 6% TLP-*Alnus*+PQS) and LOI percentages are higher than the previous lpaz, ranging between 27 and 47% dry weight.

HKP-3b 632.5 to 587.5 cm *c*.5800 to *c*.5700 BP (*c*.6600 to *c*.6400 cal. yrs BP)

Representation of arboreal pollen (49 to 66%) is broadly similar in this subzone to HKP-3a. The most significant changes are the gradual decline in frequency of *Ulmus* pollen from 8% at the opening of the subzone to 1% at the close, and an increase in *Fraxinus excelsior* pollen values (3 to 9%). Pollen from *Alnus glutinosa* (46 to 86% TLP+*Alnus*) remains dominant, with *Quercus* (29 to 43%) and *Tilia* (4 to 9%) pollen also important. Pollen from shrub taxa (30 to 46%) is generally better represented in this subzone than in HKP-3a with frequencies of *Corylus avellana*-type increasing from 28% at the base to *c*.44% towards the subzone close. *Hedera helix* (<*c*.2%) and *Sorbus*-type (<1%) remain consistently recorded pollen types although the representation of other morphotypes such as *Salix* is low and infrequent in this subzone.

Herb pollen (4 to 9%) is not well represented although a range of taxa are sporadically recorded at frequencies less than 1%. *Filipendula* pollen (*c*.1 to 4%) attains the highest frequency, with *Potentilla*-type (<1%), Cyperaceae (<*c*.1%), and Poaceae (1 to 3%) other consistently recorded morphotypes. Pollen from aquatic plants is sparsely represented. Spore frequencies (4 to 8% TLP-*Alnus*+spores) are also low, with *Polypodium* (*c*.1 to 2% TLP-*Alnus*+spores) and Pteropsida (monolete) indet. (2 to 7% TLP-*Alnus*+spores) the main morphotypes. Percentages of pre-Quaternary spores average only *c*.5% TLP-*Alnus*+PQS and LOI values occur between 26 and 51% dry weight.

HKP-3c 587.5 to 477.5 cm *c*.5700 to *c*.4800 BP (*c*.6400 to *c*.5500 cal. yrs BP)

The opening of this subzone has been dated to 5655±45 BP (6528 to 6313 cal. yrs BP). The proportion of pollen from arboreal taxa (39 to 65%) is similar to the other subzones in HKP-3. *Alnus glutinosa* (38 to 82% TLP+*Alnus*) is the dominant tree pollen type and *Quercus* pollen (24 to 47%) is also frequent. Pollen from *Pinus sylvestris* (*c*.1 to 8%) becomes more abundant towards the close of the subzone with *Tilia* (3 to 8%) and *Fraxinus excelsior* (*c*.1 to 9%) morphotypes also moderately well represented. *Betula* (*c*.1 to 5%) and *Ulmus* (2 to 4%) pollen frequencies are low.

Shrub pollen frequencies are higher at the opening of the subzone (*c*.52%) than at the close (*c*.31%), largely due to the decline in abundance of *Corylus avellana*-type pollen from *c*.48% to *c*.25% throughout the lpaz. However, *Ilex aquifolium* pollen (*c*.1 to 7%) is better represented during this lpaz. Herb pollen frequencies (5 to 14%) are also higher within HKP-3c. Pollen of *Catha palustris*-type (<*c*.1 to 3%), *Filipendula* (<*c*.1 to 11%), *Plantago lanceolata* (<*c*.1 to 2%), and Poaceae (<*c*.1 to 3%) increases in abundance, particularly during the opening of the subzone.

Aquatic pollen is not significant in this lpaz. The proportion of spores however, increase from *c*.5% at the opening of the subzone to *c*.12% (TLP-*Alnus*+spores) at the close as frequencies of *Polypodium* (c.1 to 7% TLP-*Alnus*+spores) and Pteropsida (monolete) indet. (2 to 8% TLP-*Alnus*+spores) morphotypes become more abundant. Frequencies of pre-Quaternary spores and LOI values range between c.1 to 9% (TLP-*Alnus*+spores) and 43 and 62% dry weight. The close of this subzone has been dated to 4850±45 BP (5655 to 5478 cal. yrs BP).

HKP-4 477.5 to 381.5 cm *c*.4800 to *c*.4500 BP (*c*.5500 to *c*.5200 cal. yrs BP)

This lpaz is dominated by arboreal pollen (56 to 68%), although the proportion contributed by the various tree taxa is markedly different from the previous subzone. The main changes are the considerably lower

Alnus glutinosa pollen values (23 to 46% TLP+*Alnus*), despite an isolated spike of 73% (TLP+*Alnus*) at 440 cm, and an increase in the abundance of *Pinus sylvestris* pollen (8 to 16%). *Quercus* pollen is recorded at values between 25 and 42%, and *Betula* (2 to 8%), *Ulmus* (1 to 5%), and *Tilia* (6 to 9%) generally occur at slightly higher frequencies than in HKP-3c. Pollen of *Fraxinus excelsior* and *Taxus baccata* occur at frequencies generally <3%.

The shrub pollen sum (24 to 36%) is dominated by *Corylus avellana*-type pollen (19 to 32%), with *Salix* morphotypes (*c*.1 to 4%) better represented than previous lpazs. Herb pollen (4 to 11%) is lower than HKP-3c with Chenopodiaceae, *Filipendula*, Cyperaceae (<*c*.1 to 2%), and Poaceae (1 to 5%) the main types.

Some aquatic taxa (*Sparganium erectum* and *Sparganium emersum*-type) occur at the base of the zone at low frequencies (<1% TLP-*Alnus*+aquatics). Spore morphotypes (12 to 37% TLP-*Alnus*+spores) generally increase in abundance throughout the zone, with *Polypodium* (2 to 10% TLP-*Alnus*+spores), *Pteridium aquilinum* (*c*.1 to 7% TLP-*Alnus*+spores), and Pteropsida (monolete) indet. (6 to 21% TLP-*Alnus*+spores), all showing a rising trend. *Sphagnum* spores (<*c*.1 to 4% TLP-*Alnus*+spores) are also more abundant. Pre-Quaternary spores are abundant (4 to 38% TLP-*Alnus*+PQS), and LOI values range between 17 and 44%, averaging 29% dry weight. A bulk sediment sample from between 390 to 395 cm, close to the zone end has been dated to 4555±45 BP (5439 to 5043 cal. yrs BP).

HKP-5 381.5 to *c*.155 cm *c*.4500 to <*c*.4000 BP (*c*.5200 to <*c*.4400 cal. yrs BP)

Although there are consistent shifts in some of the arboreal pollen taxa in this lpaz, the main changes concern the herb pollen types. Therefore, this lpaz has been subdivided into three subzones.

HKP-5a 381.5 to 305.5 cm *c*.4500 to *c*.4000 BP (*c*.5200 to *c*.4400 cal. yrs BP)

This subzone is characterised by a marked increase in the proportion of herb pollen taxa (12 to 21%) recorded and an overall decrease in the representation of arboreal morphotypes (60 to 47%). Frequencies of *Alnus glutinosa* pollen decline from *c*.40% at the opening of the lpaz to *c*.30% (TLP+*Alnus*) for the remainder of the zone. *Quercus* pollen percentages are generally higher than those of *Alnus glutinosa* and range between 27 and 38%. *Pinus sylvestris* (6 to 16%) and *Tilia* (1 to 8%) pollen frequencies are moderately high, whereas pollen of *Betula* (*c*.1 to 3%), *Ulmus* (1 to 4%), *Fraxinus excelsior* (*c*.1 to 6%), and *Taxus baccata* (<*c*.1 to 2%) occurs at relatively low frequencies. Exotic pollen grains of *Picea* and *Abies* are also regularly recorded during this lpaz at frequencies <*c*.1%.

The proportion of shrub pollen (21 to 33%) is lower than HKP-4 but remains dominated by *Corylus avellana*-type (16 to 28%), although *Salix* morphotypes (*c*.1 to 8%) are more important during this lpaz. The increase in representation of herb pollen is mainly attributable to the higher frequencies of Poaceae (7 to 12%), although pollen from several other taxa are notable, such as Chenopodiaceae (<*c*.1 to 3%), *Filipendula* (*c*.1 to 5%), and Cyperaceae (<*c*.1 to 3%). A few cereal pollen grains (*Hordeum*-type) are also recorded towards the close of the lpaz. Apart from *Sparganium emersum*-type (<1% TLP-*Alnus*+aquatics), pollen from obligate aquatics is not well represented.

Frequency of spore morphotypes declines from *c*.28% to *c*.13% (TLP-*Alnus*+spores) at the close of the subzone. *Polypodium* (3 to 8%), *Pteridium aquilinum* (3 to 10%), Pteropsida (monolete) indet. (4 to 13%), and *Sphagnum* (<*c*.1 to 3% TLP-*Alnus*+spores) are the main morphotypes recorded. Pre-Quaternary spore frequencies decline in the lower part of the subzone (8 to 40% TLP-*Alnus*+PQS) but are generally high throughout. Test linings from foraminifera are recorded in numerous levels. LOI values range between 19 and 33% except a spike of 67% dry weight at 362 cm.

HKP-5b 305.5 to 245 cm <*c*.4000 BP (<*c*.4400 cal. yrs BP)

This zone is characterised by a sharp increase in the frequency of herb pollen, with values ranging between 19 and 42%. Arboreal (36 to 64%) and shrub (17 to 28%) pollen still occur at relatively high proportions. *Alnus glutinosa* is the most abundant tree pollen taxon, with percentage values increasing from *c*.34 at the beginning of the lpaz to *c*.58% (TLP+*Alnus*) towards the close. *Quercus* pollen (20 to 34%) is well represented, but frequencies of all the other arboreal pollen types except *Betula* (2 to 5%), e.g. *Ulmus* (<*c*.1 to 4%), *Tilia* (<*c*.1 to 6%), and *Taxus baccata* (<*c*.1%), are lower than HKP-5a. Frequencies of *Fraxinus excelsior* pollen (<c.1 to 4%) are also relatively low, despite a spike of 32% at 304 cm. The opening of this subzone is dated to 3950±45 BP (4517 to 4257 cal. yrs BP).

Shrub pollen (17 to 28%) is dominated by *Corylus avellana*-type (14 to 24%), although *Salix* (<c.1 to 8%) is also relatively well represented. *Filipendula* (11 to 31%) and Poaceae (5 to 10%) pollen are important herb

taxa, although other morphotypes are moderately abundant, such as Chenopodiaceae (<*c*.1 to 2%), *Plantago lanceolata* (<*c*.1 to 2%), Asteraceae (Cardueae/Asteroideae) (<*c*.1 to 6%), and Cyperaceae (<*c*.1 to 2%).

Frequencies of spores range between 5 and 20% (TLP-*Alnus*+spores), with *Polypodium* (<*c*.1 to 6% TLP-*Alnus*+spores), *Pteridium aquilinum* (1 to 7% TLP-*Alnus*+spores), and Pteropsida (monolete) indet. (2 to 7% TLP-*Alnus*+spores) the main morphotypes. Pre-Quaternary spores (3 to 11% TLP-*Alnus*+PQS) are not as abundant as the previous lpaz but increase in frequency towards the close of the subzone. Foram test linings remain present in this lpaz and LOI values range between 23 and 38% dry weight.

HKP-5c *c*.245 to 150 cm <*c*.4000 BP (<*c*.4400 cal. yrs BP)

The proportion of arboreal taxa (40 to 59%) in this lpaz is similar to HKP-5b although there are differences in the relative frequency of individual morphotypes. *Alnus glutinosa* (23 to 39% TLP+*Alnus*) and *Quercus* (27 to 36%) morphotypes are co-dominant, and the frequency of *Pinus sylvestris* (4 to 19%) and *Taxus baccata* (<*c*.1 to 3%) is higher then the previous lpaz. Pollen of *Betula* (*c*.1 to 4%), *Ulmus* (1 to 3%), *Tilia* (*c*.1 to 2%), and *Fraxinus excelsior* (<*c*.1 to 2%) is also present at low frequencies. There is a consistent occurrence of *Picea* and *Abies* pollen at minor percentages (<*c*.1%).

Shrub pollen (17 to 32%) is dominated by *Corylus avellana*-type (15 to 30%), with *Salix* (<*c*.1 to 2%), and Ericaceae (<*c*.1%) morphotypes consistently represented taxa at low levels. Herb pollen (11 to 43%) is well represented, with frequencies generally higher at the close of the subzone. *Filipendula* and Poaceae (4 to 12%) are generally the most abundant morphotypes throughout the zone. Chenopodiaceae (*c*.1 to 4%) and Cyperaceae (*c*.1 to 24%) pollen is also relatively abundant, and the latter increases in frequency from the centre of the lpaz and peaks at the close. Other herb pollen taxa occur frequently with low percentage values, such as *Plantago lanceolata* (<*c*.1 to 2%), Asteraceae (Cardueae/Asteroideae) (<*c*.1%), and *Iris* (<*c*.1 to 2%).

Minor frequencies of aquatic pollen (<*c*.1% TLP-*Alnus*+aquatics) are recorded throughout this lpaz, with *Sparganium emersum*-type pollen the most frequent. *Polypodium* (2 to 9% TLP-*Alnus*+spores), *Pteridium aquilinum* (2 to 11% TLP-*Alnus*+spores), Pteropsida (monolete) indet. (5 to 8% TLP-*Alnus*+spores), and *Sphagnum* (<*c*.1 to 3%) are the main spore morphotypes, which are better represented in this subzone (12 to 21% TLP-*Alnus*+spores). Frequencies of pre-Quaternary spores (15 to 31% TLP-*Alnus*+PQS) are high and LOI values (12 to 28% dry weight) are generally lower than HKP-5b. Foram test linings are frequently encountered throughout the subzone.

9.4.3 Diatom Stratigraphy

The main characteristics of the local diatom assemblage zone within core HK-4P are described in the following section (Figures 9.5, 9.6, 9.7, 9.8, 9.9, and 9.10). Diatoms were not either not preserved, or too infrequent to count from within the main floodplain peat unit in the central part of the stratigraphy. However, diatom valves were abundant within the humic clay which lies between the basal peat dated to 7030±45 (7914 to 7699 cal. yrs BP), and the main peat dated to 5965±45 BP (6889 to 6719 cal. yrs BP). A percentage diagram showing the common and rare diatom species from this unit, arranged according to salinity tolerance is presented (Figure 9.5 and 9.6), and the life form preferences of the major species (see section 5.7.2 for references used) is shown in Figure 9.7.

In addition, a diatom record has been obtained from the organic clays overlying the main floodplain wood peat and thus dates from 4555±45 BP (5439 to 5043 cal. yrs BP). The salinity classification of the common and rare diatoms characteristic of this part of the core is presented (Figure 9.8 and 9.9) and the habitat requirements of these taxa is shown in the life-form diagram (Figure 9.10). Diatom frequency is expressed in all diagrams as a percentage of total diatom valves (%TDV). Definition of the terms used regarding salinity and habitat preferences are given in Appendix B.

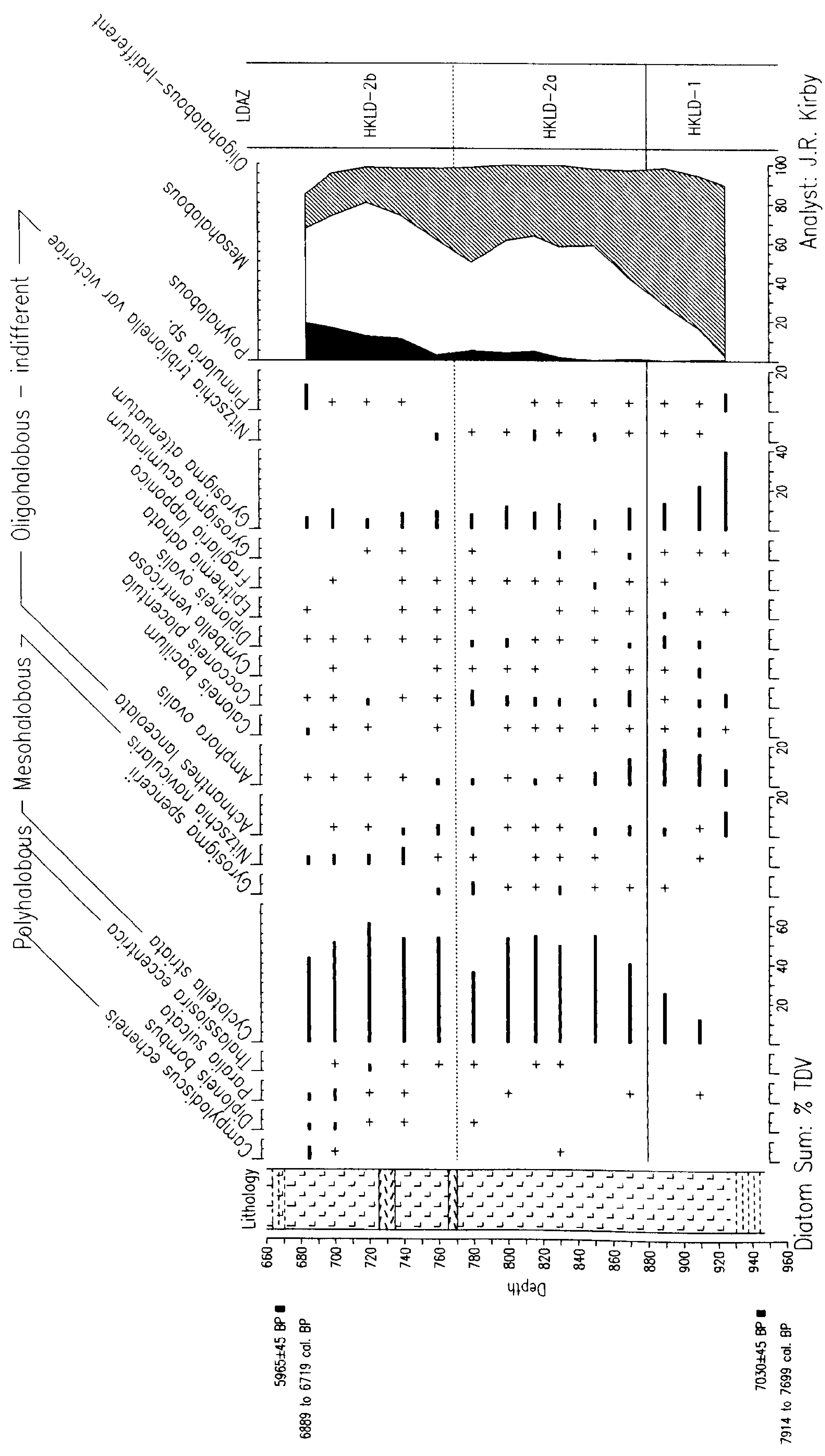


Figure 9.5 Hook Lane lower percentage diatom diagram based on salinity (+ represents frequencies <3%)

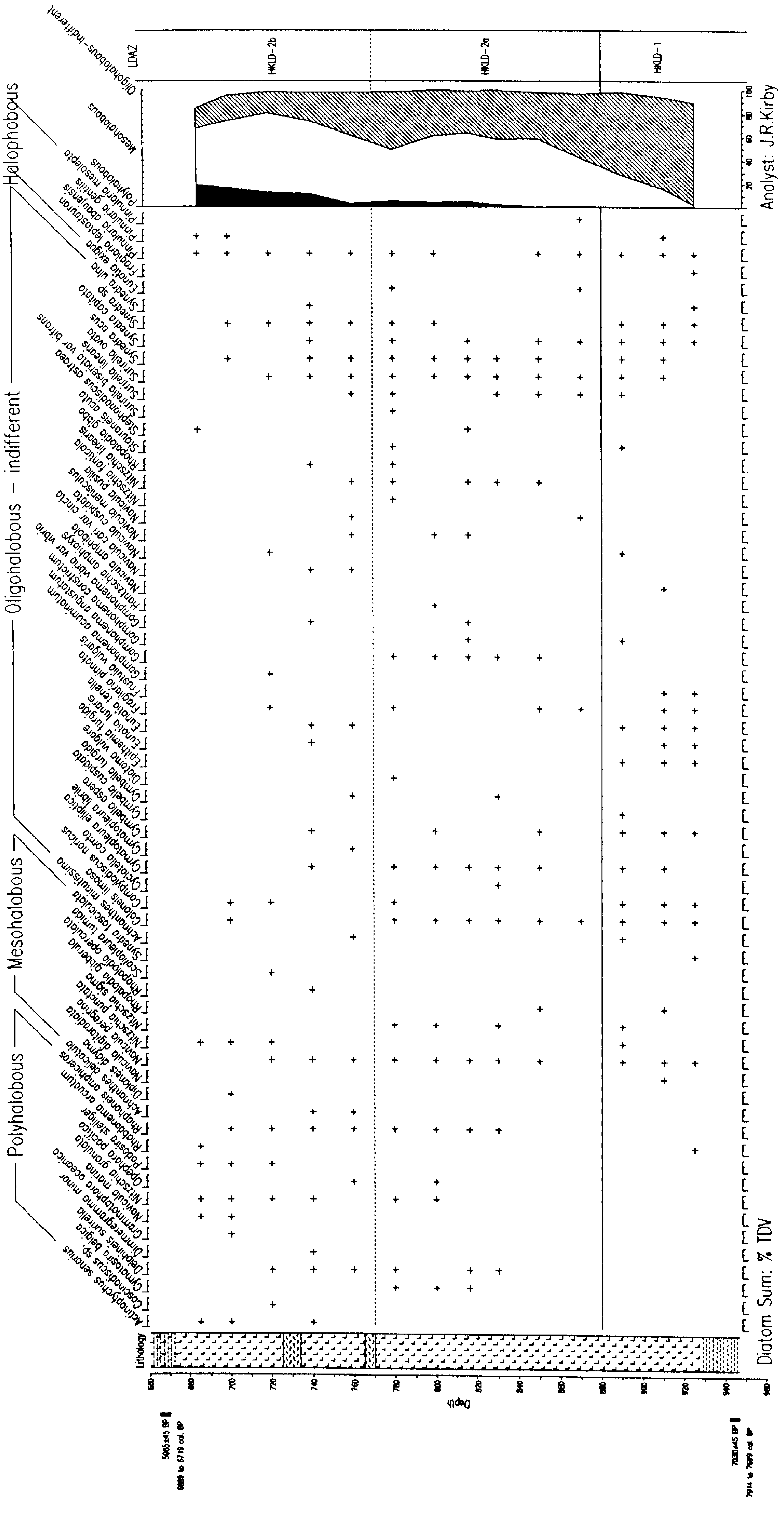


Figure 9.6 Hook Lane lower percentage diatom diagram (minor taxa) based on salinity (+ represents frequencies <3%)

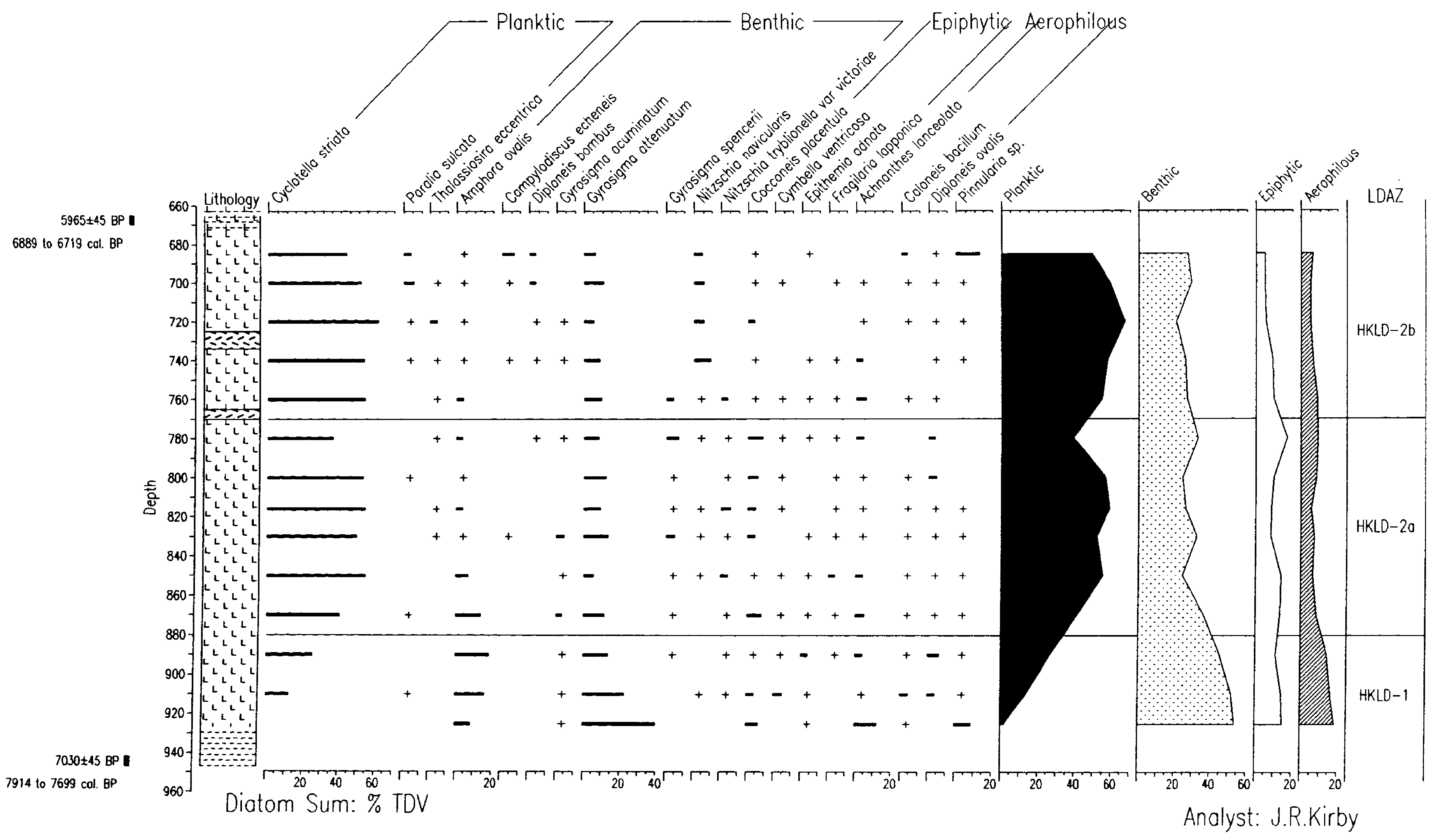


Figure 9.7 Hook Lane lower percentage diatom diagram based on life-form (+ represents frequencies <3%)

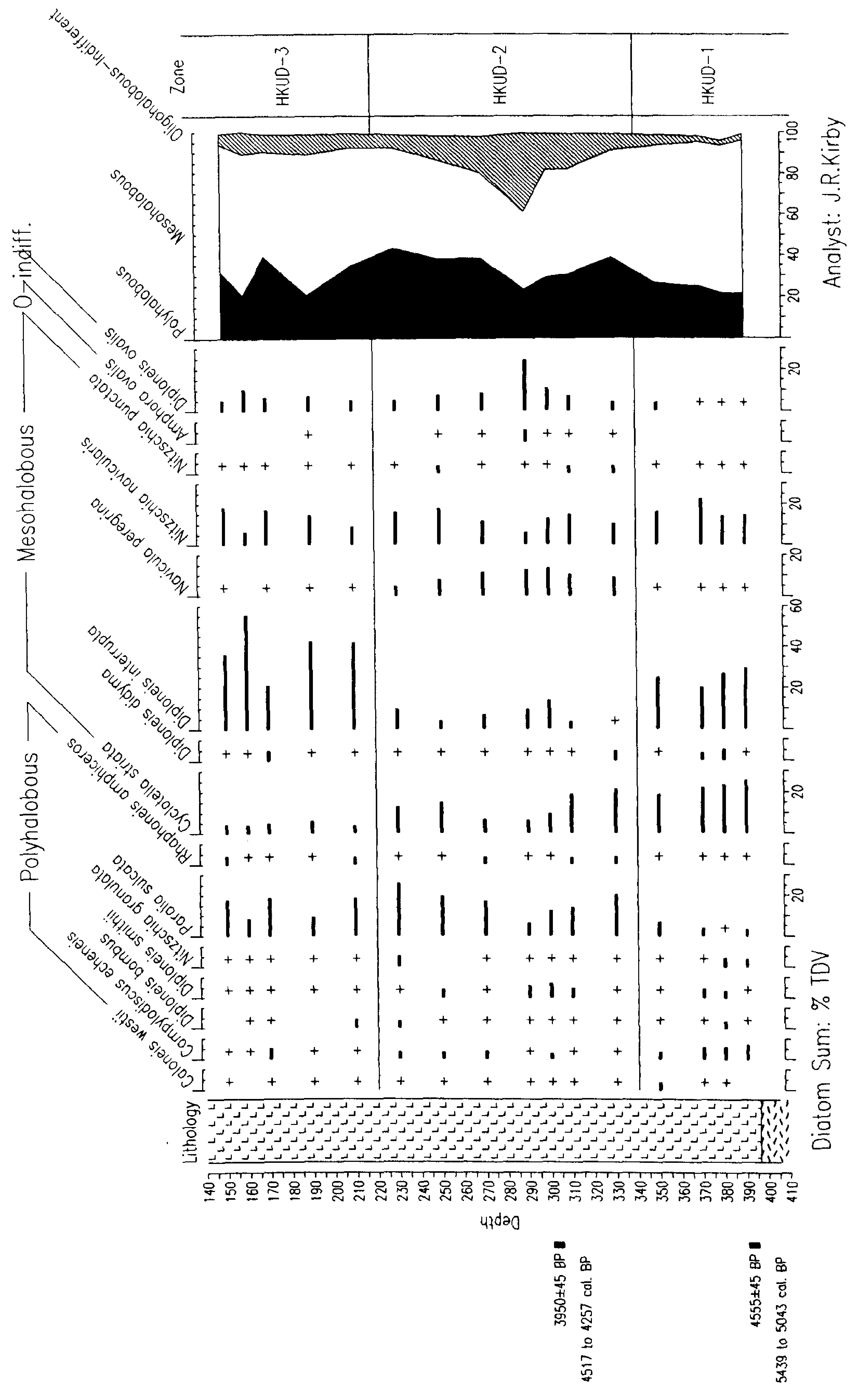


Figure 9.8 Hook Lane upper percentage diatom diagram based on salinity (+ represents frequencies <3%)

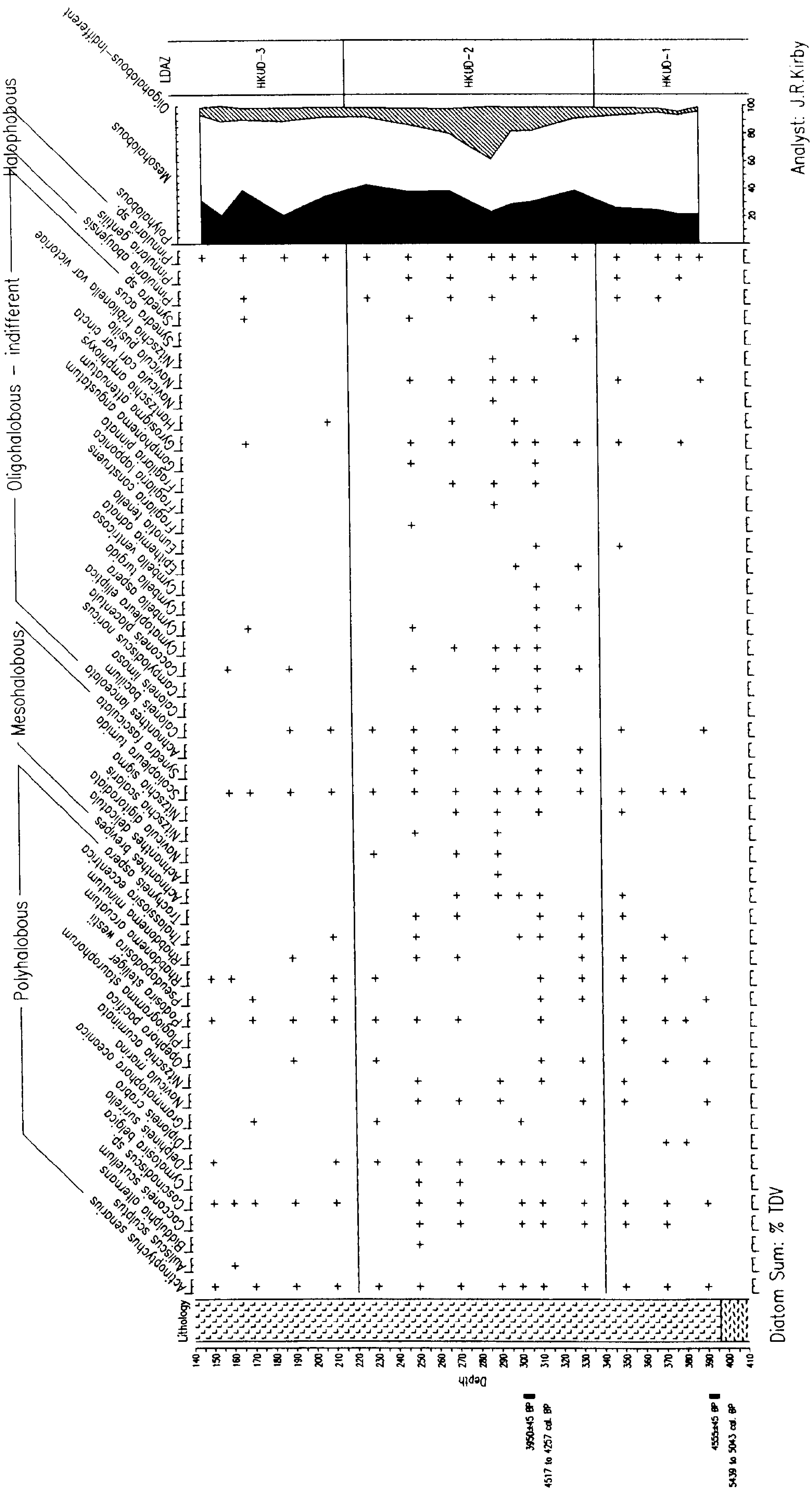


Figure 9.9 Hook Lane upper percentage diatom diagram (minor taxa) based on salinity (+ represents frequencies <3%)

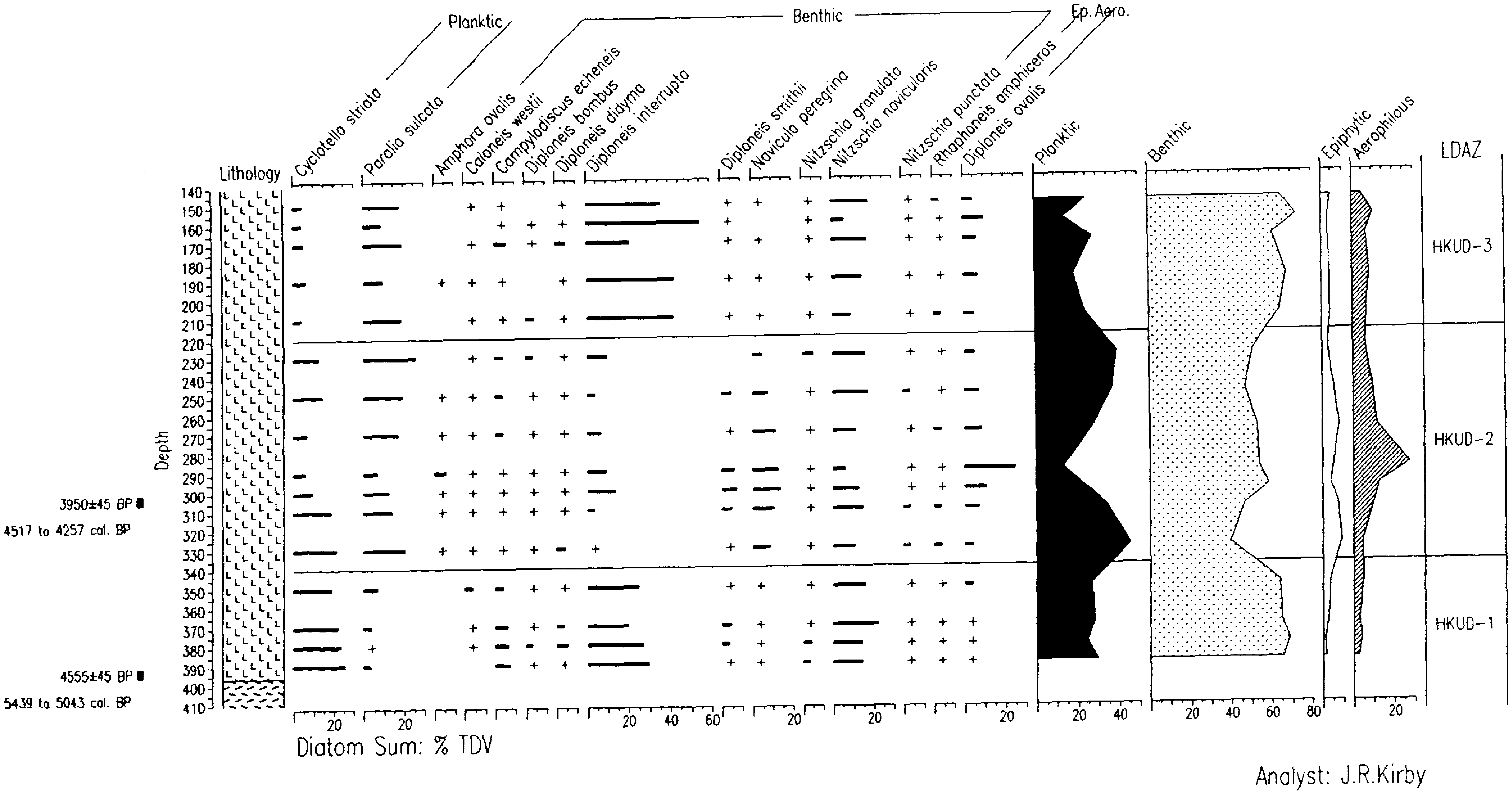


Figure 9.10 Hook Lane upper percentage diatom diagram based on life-form (+ represents frequencies <3%)

HKLD-1 925 to 880 cm <*c.*7000 BP (<*c.*7800 cal. yrs BP)

Figures 9.5, 9.6 and 9.7

This ldaz is initially dominated by oligohalobous-indifferent taxa (87 to 71%) and benthic diatoms (54 to 45%). The main species are *Gyrosigma attenuatum* which declines from 40 to 14% and *Amphora ovalis* which increases throughout the zone from 8 to 18%. Less important oligohalobous-indifferent taxa include aerophilous taxa such as *Achnanthes lanceolata* (2 to 12%), and *Diploneis ovalis* (0 to 6%) and epiphytic species such as *Cocconeis placentula* (3 to 7%). *Pinnularia* sp. is moderately abundant at 9% in the lowest sample but then declines to 1% at the close of the ldaz.

Frequencies of oligohalobous-indifferent diatoms decline throughout the zone and are replaced by mesohalobous species which increase in abundance from 2 to 28%. This is attributable to the increase in abundance of the planktonic species (2 to 27%) such as *Cyclotella striata*, which increases from 0 to 25% at the close of the zone. Epiphytic (11 to 15%) and aerophilous (15 to 19%) taxa are moderately well represented during this ldaz whereas polyhalobous species are insignificant (<1%).

HKLD-2 880 to 685 cm <*c.*7000 to>*c.*6000 BP (<*c.*7800 to >*c.*6800 cal. yrs BP)

This zone is dominated by the same diatom species. However, fluctuations in the representation of less abundant taxa enable the separation of two subzones.

HKLD-2a 880 to 770 cm <*c.*7000 to >*c.*6000 BP (<*c.*7800 to >*c.*6800 cal. yrs BP)

Mesohalobous (41 to 59%) and planktonic (39 to 59%) diatoms are generally the most abundant taxa in this subzone. Oligohalobous-indifferent species (37 to 55%) are well represented although generally decline in frequency towards the close of the subzone. Polyhalobous taxa are poorly represented, although they increase from *c.*1 to 5% throughout the subzone.

The mesohalobous, planktonic species *Cyclotella striata* (36 to 54%) dominates this ldaz. Although benthic taxa (24 to 36%) are less well represented than in HKLD-1, a range of mesohalobous (e.g. *Gyrosigma spencerii*, 1 to 6%), and oligohalobous-indifferent benthic species (e.g. *Amphora ovalis*, 1 to 14%, *Gyrosigma attenuatum* 5 to 13%, and *Nitzschia triblionella var vicatriae*, 1 to 5%), are moderately well represented.

The frequency of oligohalobous-indifferent taxa declines during the lower part of the ldaz owing to the decrease in abundance of taxa such as *Amphora ovalis*, *Achnanthes lanceolata* and *Diploneis ovalis*. *Gyrosigma attenuatum* maintains a significant presence, despite being less well represented than in the previous ldaz. Epiphytic taxa (8 to 17%) such as the oligohalobous-indifferent species *Cocconeis placentula* (4 to 8%) are slightly more abundant in HKLD-2a. Aerophilous taxa (6 to 9%) are not important in this subzone.

HKLD-2b 770 to 685 cm >*c*.6000 BP (>*c*.6800 cal. yrs BP)

This ldaz is dominated by mesohalobous taxa (48 to 68%) and there is a further reduction in the proportion of diatoms from the oligohalobous-indifferent group (17 to 36%) compared to the lower ldaz. Polyhalobous species rise in frequency throughout the subzone from 3 to 19%. *Cyclotella striata* (43 to 61%), a mesohalobous, planktonic diatom, remains the most abundant species. Due to the high frequencies of *Cyclotella striata*, the planktonic group (49 to 67%) dominates the subzone, although several polyhalobous diatom species increase in abundance, such as *Paralia sulcata* (0 to 6%) and *Thalassiosira eccentrica* (0 to 4%), which also contribute to this total.

Benthic diatoms (21 to 29%) are relatively abundant in this subzone. This is largely attributable to the presence of mesohalobous species such as *Nitzschia navicularis* (1 to 9%) and oligohalobous taxa such as *Gyrosigma attenuatum* (5 to 10%). Several polyhalobous benthic taxa, such as *Campylodiscus echeneis* (0 to 6%) and *Diplonies bombus* (0 to 3%), increase in frequency towards the close of the subzone.

Epiphytic taxa (5 to 6%) are less important than the previous ldaz, largely due to the decrease in frequency of *Cocconeis placentula* (1 to 3. Aerophilous taxa (5 to 6%) also occur at low frequencies with species such as *Diploneis ovalis* (1 to 2%) less well represented than previous ldazs. The decrease in abundance of these species also contributes to the lower frequencies of oligohalobous-indifferent taxa in this ldaz. Frequencies of *Pinnularia* spp. diatoms increase to 13% at the close of the subzone.

HKUD-1 390 to 340 cm *c*.4600 to *c*.4200 BP (*c*.5200 to *c*.4700 cal. yrs BP)

Figures 9.8, 9.9 and 9.10

This ldaz is dominated by mesohalobous diatom taxa (52 to 75%) such as *Cylotella striata* (19 to 25%), *Diploneis didyma* (2 to 5%), *Diploneis interrupta* (20 to 29%), and *Nitzschia navicularis* (14 to 22%). Species such as *Diploneis interrupta* and *Nitszchia navicularis* contribute significantly to the high proportion of benthic diatoms recorded in this zone (64 to 69%), along with polyhalobous species such as *Campylodiscus echeneis* (4 to 7%), *Diploneis smithii* (2 to 4%), and *Nitzschia granulata* (2 to 4%). Planktonic species (25 to 30%) are also well represented, with *Cyclotella striata* and the polyhalobous species *Paralia sulcata* (1 to 7%) the main taxa.

Frequencies of oligohalobous-indifferent species (3 to 5%) are low, with *Diploneis ovalis* (2 to 4%) the main species. Epiphytic (1 to 4%) and aerophilous (3 to 5%) taxa are not important components of this ldaz.

HKUD-2 340 to 220 cm *c*.4200 to <*c*.4000 BP (*c*.4700 to <*c*.4400 cal. yrs BP)

Polyhalobous (23 to 43%) and oligohalobous-indifferent diatoms (7 to 38%) occur with greater frequencies during this ldaz, whereas mesohalobous taxa (38 to 53%), whilst still dominant, are generally less abundant. Planktonic taxa are generally better represented in this zone due to the increased frequencies of the polyhalobous species *Paralia sulcata* (6 to 26%) and high values for *Cyclotella striata* (6 to 21%). However, the proportion of polyhalobous and mesohalobous planktonic taxa are markedly reduced in the centre of the zone due to the increased frequency of oligohalobous-indifferent species such as *Diploneis ovalis* and *Amphora ovalis*, which peak at 290 cm with values of 24% and 5% respectively.

Benthic taxa (40 to 59%) are well represented during this ldaz, despite the reduction in frequency of *Diploneis interrupta* (2 to 13%) which are lower compared to the previous zone. Several other mesohalobous benthic diatoms such as *Navicula peregrina* (4 to 13%) and *Nitzschia navicularis* (6 to 17%) are abundant in HKUD-2, along with a range of polyhalobous species, including *Campylodiscus echeneis* (*c*.1 to 4%), *Diploneis smithii* (1 to 7%), and *Nitzschia punctata* (*c*.1 to 3%). Aerophilous taxa (5 to 28%) are better represented in this zone, and are particularly abundant in the centre of the zone due to the high frequencies of *Diploneis ovalis*. Epiphytic taxa (4 to 9%) maintain relatively low percentage values although they are recorded at higher frequencies than in HKUD-1.

HKUD-3 220 to 150 cm <*c*.4000 BP (<*c*.4400 cal. yrs BP)

This zone is marked by an increase in the abundance of mesohalobous (58 to 69%) and benthic (61 to 72%) taxa, despite the decrease in frequency of *Navicula peregrina* (<*c*.1 to 2%). This is attributable to a dramatic rise in the frequency of *Diploneis interrupta* (21 to 55%), which dominates the zone. *Nitzschia navicularis* (6 to 18%) is also an important mesohalobous, benthic species.

Polyhalobous (21 to 39%) and planktonic (14 to 28%) taxa are generally less well represented than the previous ldaz. This is partly attributable to the lower frequencies of *Paralia sulcata* (8 to 18%) and *Cyclotella striata* (4 to 6%) within this zone. Oligohalobous-indifferent (5 to 11%) and aerophilous (5 to 10%) taxa are also less abundant than in HKUD-2 due to the lower frequencies of *Diploneis ovalis* (5 to 10%). Epiphytic taxa (3 to 5%) are not a significant component of this ldaz.

9.5 Inferred Local Vegetation History and Depositional Environment

HKP-1 947 to 929.5 cm *c*.7000 BP (*c*.7800 cal. yrs BP)

Figure 9.3 and 9.4

The base of this lpaz is dated to 7030±45 BP (7914 to 7699 cal. yrs BP) as paludification ensues and frequencies of *Alnus glutinosa* increase. High percentages of *Alnus glutinosa* pollen within this zone indicate the presence of this taxon locally. A range of fen herbs occur at the base of the zone, such as *Lysimachia vulgaris*-type, Cyperaceae, and *Sparganium erectum*, which indicates wet and relatively open conditions close to the site during the early stages of paludification. The frequency of these wetland herbaceous taxa decline as *Alnus glutinosa* frequencies increase to *c*.63% (TLP+*Alnus*).

Representation of arboreal taxa from possible dry land sources, such as *Pinus sylvestris*, *Ulmus*, *Quercus*, and *Tilia* is low, suggesting considerable filtration of pollen from extra-local sources due to the abundance of local fen carr vegetation (Tauber, 1965; Jacobson & Bradshaw, 1981). Percentages of *Corylus avellana*-type pollen are relatively high suggesting that this taxon was prevalent close to the site, or was perhaps colonising drier parts of the fen carr with other trees such as *Quercus* that can tolerate a certain degree of waterlogging (see Brown, 1997; Rodwell, 1991 and section 7.5).

HKP-2 929.5 to 666 cm <*c*.7000 to *c*.6000 BP (<*c*.7800 to *c*.6800 cal. yrs BP)

The zone boundary between HKP 1 and 2 is placed where vegetation changes occur, associated with a shift in sedimentary environment, from a fen carr to a lagoonal palaeoenvironment. Extensive flooding of water is indicated by the influx of inorganic sediments and also by inferred responses of the surrounding arboreal vegetation. Frequencies of *Alnus glutinosa* decline, suggesting that the local area became too wet for the survival of this species. *Quercus* and *Corylus avellana*-type also show a slight decline during the early stages of this lpaz, suggesting that some of the pollen from these taxa recorded during HKP-1 may have originated from trees colonising the drier parts of the fen carr which became flooded. *Ulmus* and *Tilia* pollen percentages are considerably higher, which may reflect a greater extra-local input of pollen from these taxa subsequent to the removal of fen carr locally. *Ulmus* is not considered to be under-represented in the pollen record (Andersen, 1970), and average frequencies of 11%, along with *Tilia* (7%) and *Quercus* (30%), are high enough to suggest that the dry areas close to the site were dominated by these tree taxa.

Unequivocal evidence for the presence of a permanently standing open body of water is provided by the diatom data (see section 9.6). There is also pollen evidence to support this interpretation. Throughout the zone, aquatic taxa which occur in shallow water (Haslam *et al.*, 1975; Stace, 1997), such as *Menyanthes trifoliata* and *Sparganim erectum* are recorded, and the occurrence of floating macrophytes such as *Nymphaea alba* and *Myriophyllum alterniflorum*, along with *Potamogeton natans*-type, attests to the occurrence of open water bodies. It is interesting to note that *Typha latifolia* is recorded only in the bottom half of the zone, and is apparently replaced by *Sparganium emersum*-type pollen (which includes *Typha angustifolia* pollen as well as *S. emersum*, Bennett, 1994a), which is confined to the upper part of the lpaz. According to Grime *et al.* (1988), *Typha latifolia* is common at the edges of shallow water bodies at depths not exceeding 600 mm, although Grace & Wetzel (1981) suggest that at water depths exceeding 150 mm, it tends to give way to *Typha angustifolia*. Haslam *et al.* (1975) suggest that *Typha latifolia* can grow in water depths between 0.5 and 1 m, whereas *Typha angustifolium* and *Sparganium emersum* are capable of colonising lakes up to 2 m deep. This is consistent with the depth range for aquatic taxa presented by Harrison & Digerfeldt, (1993), which suggests that water depth increased throughout HKP-2.

Another possibility is that the water salinity increased. Grime *et al.* (1988) suggest this would cause *Typha latifolia* to be replaced by *Phragmites australis*. However, this suggestion remains unsubstantiated because despite the increased occurrence of polyhalobous diatoms towards the top of this unit (see section 9.6, Figure 9.5), the apparent increase in Poaceae pollen occurs at 840 cm, which is before the disappearance of *Typha latifolia* pollen. The effect of increasing marine influence in the perimarine zone would be to pond back river water draining seawards, which would cause water levels to rise in addition to a possible salinity increase. Therefore, increasing water depths and salinity may both be influential factors in the demise of *Typha latifolia*.

Average values for *Alnus glutinosa* pollen of 42% during this zone suggests that a fringe of wetland trees surrounded the edge of the water body. Although the frequencies of *Salix* are low (<c.1 to 2%), this taxa is often under-represented in the pollen record (Bradshaw, 1981a; Huntley & Birks, 1983, Janssen, 1984; Bunting *et al.*, 1998a; 1998b), and these percentage values are enough to suggest that this species was a component of this carr fringe (*cf.* Waller, 1994c; Long *et al.*, 1998c). Identification of *Salix* spp. wood and leaf macrofossils (M. Waller, pers. comm.) within this unit confirms this supposition, and considering the prevalence of *Salix* sp. leaf detritus throughout, it is tempting to suggest that *Salix* may have formed a substantial component of the wetland community close to the site. The higher frequencies of Poaceae pollen are likely to originate from an ecotone adjacent to the carr, possibly consisting of *Phragmites* reeds colonising the shallow water.

Four zonal vegetation communities can be envisaged here. Floating and emergent aquatics occur in the deepest water around the coring site, with reedswamp aquatic taxa colonising the shallow water margins. Fen carr dominated by *Salix* and *Alnus glutinosa* are likely to form a fringe beyond the reedswamp community, with dryland taxa such as *Ulmus*, *Tilia* and *Quercus* growing beyond the fen carr on the valley sides.

LOI values are low within this zone reflecting the increased input of minerogenic sediment from river water washing into the lagoon, and the lack of a local peat forming vegetation community. Pre-Quaternary spores are higher in this zone, which may also be related to the input of secondary pollen from river sources. Frequencies of Pteropsida (monolete) indet. spores are also higher, and show a similar pattern of distribution to pre-Quaternary spores which suggests they too may be of secondary riverine origin. It has been suggested that these factors were contributory in the anomalously old age for the dated sample (SRR-6394) from 763 to 773 cm from this unit.

The sedimentary and environmental conditions during this lpaz, and the timing of events are remarkably similar to those described by Van der Woude (1983; 1984) in the perimarine area of the Rhine-Meuse system in the SW Netherlands. These similarities are summarised here.

From *c.*7400 BP (*c.*8400 to *c.*8000 cal. yrs BP) in the Molenaarsgraaf area, local groundwater levels began to rise under the influence of rising sea level and increasing river discharge. This resulted in the formation of small shallow freshwater lakes depositing organic gyttja surrounded with areas of *Phragmites* peat accumulation. As the water level rise continued, large freshwater lagoons spread out over the entire perimarine area, intersected with channel levees standing above the water, supporting *Quercus* and *Ulmus* woodland (Van der Woude, 1983; 1984). For this palaeoenvironment, Van der Woude (1983; 1984) proposed the term 'fluviolagoon'. Sediments deposited in these basins were dominated by clay, with a high humus content perhaps where the water was shallower and there was sparse marsh herb vegetation.

As at Hook Lane, Van der Woude (1983; 1984) also argued that these lagoons were characterised by permanently open areas of water for several reasons. The basin clays were soft, and showed no evidence for oxidation, suggesting they were not exposed to the air. Also, the plant remains (mainly leaf and wood fragments) were perfectly preserved in thin laminae, suggesting suballuvial deposition. These humic laminae were thought to relate to seasonal variations in the rate of basin clay sedimentation (Van der Woude, 1983; 1984). Also, very low herb pollen values were recorded which obviated the presence of dense vegetation, which was restricted to shallow areas.

Based on detailed palaeoecological analysis of pollen and seeds, Van der Woude (1984) suggested that the basins were colonised by *Nymphaea* and *Nuphar*, the shallow lagoon shores were colonised by *Typha* and *Sparganium*, and fringing this was a *Phragmites* dominated swamp and *Alnus* carr. Van der Woude (1983) suggested the river dunes that dissected and surrounded the lagoons were dominated by *Quercus*, *Ulmus* and *Fraxinus excelsior* in the moister areas, with *Tilia* in dry areas, and an undergrowth of *Corylus* *avellana*. This scenario is similar to the interpretation proposed for the extra-local pollen in this study. Fluviolagoonal sedimentation ceased in The Netherlands *c*.6100 BP (*c*.7200 to *c*.6700 cal. yrs BP) which (Van der Woude, 1983) suggests was related to an apparent decrease in the rate of relative sea-level rise (see section 10.5 for full discussion). This is closely correlated with the time at which the 'fluviolagoon' silts up at Hook Lane at the close of this lpaz (*c*.6000 BP, *c*.7200 to *c*.6600 cal. yrs BP).

HKP-3a 666 to 632.5 cm *c*.6000 to *c*.5800 BP (*c*.6800 to *c*.6600 cal. yrs BP)

This zone boundary coincides with the lithological change from humic clay to clayey wood peat and the transition back to a fen carr community. A date of 5965±45 BP (6889 to 6719 cal. yrs BP) has been obtained for the onset of wood peat deposition. Frequencies of *Alnus glutinosa* pollen increase, and peak at 90% mid-zone, indicating the establishment of *Alnus glutinosa* dominated carr locally, with perhaps a *Salix* component. Scalariform perforation plates, which occur within the vessels of *Alnus glutinosa* wood, were also consistently recorded, indicating local stands of *Alnus glutinosa* (*cf.* Van der Wiel, 1982).

Frequencies of extra-local pollen types such as *Quercus*, *Ulmus*, *Tilia*, and *Corylus avellana*-type remain moderately high, suggesting the presence of these tree taxa close to the coring site. *Quercus* pollen percentages increase towards the centre of the zone which may suggest the colonisation of some of these tree taxa in the locally drier areas of fen. *Ulmus* pollen declines coincident with a band of clay in the stratigraphy which may be caused by a dilution in pollen concentrations, associated with the input of riverine pollen and minerogenic sediment. Wheeler (1995) has suggested that pollen concentration can be affected by dilution as a result of clay being washed onto a site. Pollen concentrations appeared lower on the pollen slides although the calculation of absolute pollen values would be required to fully explore this change. Clearly, the pollen taphonomy is complex at this point.

The palaeoenvironmental conditions inferred in this lpaz are similar to those immediately following the 'fluviolagoonal' phase in the Rhine/Meuse perimarine area of The Netherlands (Van der Woude, 1985), when at *c*.6100 BP (*c*.7200 to *c*.6700 cal. yrs BP), the floodbasins silted up and gave way to extensive swamp forests (*Alnus* carr), and to a lesser degree, *Phragmites* marsh (Van der Woude, 1993; 1985). The initiation of the swamp forest period would have been facilitated by a reduced rate (or an actual drop) in relative sea-level rise (Van de Plassche, 1982) and fast sedimentary filling in of the pre-existing floodbasins (Van der Woude, 1985, see section 10.5 for full explanation). This forest persisted for 2000 radiocarbon years, owing to continued (but slower) rises in water level, and deposited *Alnus* wood peat throughout the perimarine area. *Ulmus* and *Quercus* forest apparently persisted during this phase, despite the dramatic rise in *Alnus glutinosa* pollen (Van der Woude, 1985). A comparable scenario is evident at Hook Lane in this lpaz as it is suggested that *Alnus glutinosa* simply re-expanded over the wet floodbasin sediments, with the influx of extra-local pollen still being recorded at the site despite some possible filtration.

HKP-3b 632.5 to 587.5 cm *c*.5800 to *c*.5700 BP (*c*.6600 to *c*.6400 cal. yrs BP)

This subzone boundary has been drawn at the point where *Ulmus* pollen frequencies decline from *c*.12%, to values of *c*.5%. At the opening of HK-3b, *Ulmus* pollen percentages are reduced, but even in HKP-3c, *Ulmus* frequencies are as high as 4%, which suggests that whatever vector caused the *Ulmus* to decline, it was not drastically detrimental to the survival of this taxon, as although populations were diminished, they were apparently able to survive after the main decline.

This event has been dated to 5440±45 BP (6304 to 6216 cal. yrs BP) at Eskamhorn *c*.7 km upstream. However, linear interpolation between the radiocarbon dates above and below this lpaz boundary suggests an age of *c*.5800 BP (*c*.6900 to *c*.6400 cal. yrs BP) for this horizon (see Figure 9.2). Although ages for the *Ulmus* decline have been obtained which are as old as this (e.g. Bennett, 1983b; Hofstede *et al.*, 1989; Waller, 1994c), and some time transgression between sites may be expected, a difference of *c*.400 radiocarbon years for the same event at sites *c*.7 km apart requires discussion. The apparent contrasting nature of the *Ulmus* decline event at Eskamhorn and Hook Lane may be explained by the local conditions at Hook Lane and the problems associated with radiocarbon (and interpolated ages). These two factors are considered below.

The slow nature of the fall in pollen percentages implies that the classic *Ulmus* decline at this site was not strongly manifested, perhaps due to the filtration of pollen from extra-local sources caused by dense *Alnus glutinosa* dominated carr. Indeed, the floodplain community would have been expanding up the valley sides as peat accumulated at this time which may have been a contributory factor in reducing some of the *Ulmus* populations by waterlogging. The increase in *Fraxinus excelsior* pollen coincident with the *Ulmus* decline may be related to the increased light availability, which would encourage flowering of this species, but may also be because ash is capable of colonising damp areas (*cf.* Wardle, 1981; Iremonger & Kelly, 1988). *Corylus avellana*-type pollen percentages also increase towards the end of the lpaz and a *Corylus avellana* nut is recorded from 615 cm in the core. This may suggest that conditions close by became suitable for the colonisation of this taxon or may also be evidence to suggest that *Corylus avellana* populations were able to expand after the reduction in *Ulmus* trees. Weakly expressed or hard to define *Ulmus* decline events have been noted from other lowland floodplain sites (e.g. Long *et al.*, 1998c; Waller, 1998 and see also Brayshay & Dinnin, 1999) which suggests that such contexts are not ideal for studying changes in the dryland environment, especially where there are changes occurring in the local vegetation that influence the pollen taphonomy.

Whilst there is little evidence to suggest that the age estimate for the *Ulmus* decline at Hook Lane is erroneous, it has been derived from interpolation which is not a particularly accurate means of defining this horizon. Therefore, whilst it may actually be a true representation of the wide spatial and temporal variations that can occur when studying events within dynamic vegetation systems, the potential overlap in ages means a possibly more synchronous event may be 'smeared' by the imprecise nature of radiocarbon dating (*cf.* Baillie, 1991).

Pollen of *Alnus glutinosa* dominates this lpaz and the values have a characteristic spiky distribution, indicating local *Alnus* carr communities at the site (Devoy, 1979; Thorley, 1981; Scaife, 1982; Van der Wiel, 1982; Smyth, 1986; Brown, 1988; Waller, 1993; 1994a, 1994c). A range of herbs are also recorded at low values, such as *Filipendula* and Poaceae, which may represent the carr ground flora, or indicate open areas of fen meadow close to the coring site.

HKP-3c 587.5 to 477.5 cm *c.*5700 to *c.*4800 BP (*c.*6400 to *c.*5500 cal. yrs BP)

There is evidence for changing local conditions in this lpaz. Frequencies of *Alnus glutinosa* are lower towards the close of the zone which suggests some opening up of the carr community. *Quercus* percentages are relatively high, which may suggest a corresponding increased input of extra-local pollen. However, a *Quercus* wood macrofossil between 520 and 527 cm and a *Fraxinus excelsior* branch between

544 and 547 cm suggest that some (perhaps an increased proportion) of the pollen from these taxa is from trees growing locally (see Table 9.3). This may suggest the local environment was becoming drier.

The opening of the zone has been dated to 5665±45 BP (6528 to 6313 cal. yrs BP) and is associated with significant changes in the fen shrub and herb flora. Frequencies of *Ilex aquifolium* are unusually high, which is also consistent with opening up of the fen woodland. *Ilex aquifolium* has been reported as a component of the drier parts of fen woodland (Peterken & Lloyd, 1967; Wheeler, 1980b; Rodwell, 1991) and can tolerate soils which are permanently waterlogged except for the summer months (Peterken & Lloyd, 1967), although it is not often abundant in the pollen record as it is an entomophilous low pollen producer with poor dispersal (Godwin, 1975a; Huntley & Birks, 1983). Therefore, the values recorded of up to 7% probably indicate a local presence. Pollen production of such taxa is usually low during shaded conditions, either because plants do not reach reproductive maturity, or because shading leads to inhibition of flowering (Bunting *et al.*, 1998b). Therefore, any canopy opening would allow an increase in pollen dispersion through the trunk space (*cf.* Tauber , 1965) and also promote increased flowering.

Associated with this increase in *Ilex aquifolium* pollen during the lower part of this lpaz are higher percentages of pollen from *Caltha palustris*-type, *Filipendula*, *Plantago lanceolata*, and Poaceae (which may be entirely of fen origin (Wheeler, 1980b). This may also be interpreted as additional evidence in support of an opening up of the fen woodland. One tentative explanation that is considered here is that the observed change in pollen may be consistent with the effects of human activity (i.e. pastoral agriculture). Barber & Brown (1987) recorded exceptionally high frequencies of *Ilex aquifolium* (up to *c*.25%) from wet floodplain deposits of (*cf.*) Ipswichian age in Hampshire. They suggest open scrub conditions were maintained by herbivory.

Scaife (1987; 1988) interprets high *Ilex aquifolium* percentages (up to 11%) from post-*Ulmus* decline age deposits on the Isle of Wight as indicative of the openness of the landscape within a pastoral context. Increases in *Ilex aquifolium* pollen have been related to the presence of free ranging cattle and pig foraging in a woodland environment (Grigson, 1982). *Ilex aquifolium* is particularly responsive to clearance (Godwin, 1948; Godwin, 1975a), and 'capable of repeatedly regenerating new shoots under heavy browsing pressure...even small seedlings show remarkable tenacity in the face of repeated nibbling'

(Peterken & Lloyd, 1967; 849). Grazing pressure on a shrub layer may also promote *Ilex aquifolium* due to its unpalatability (Scaife, 1988). Similarly, *Plantago lanceolata* is common in heavily trampled habitats, and is palatable to grazing animals (Iverson, 1941; Sagar & Harper 1964; Grime *et al.*, 1988). Frequencies of *Plantago lanceolata* would be expected to increase as light conditions improve, and it is a particularly well adapted ruderal herb, capable of persistence within disturbed habitats (Grime *et al.*, 1988).

It is therefore possible that the changes in the pollen record observed within HKP-3c are associated with deliberate clearing activities for pastoral activities, or increasing grazing pressure. Considering the context, such areas would perhaps be suited to marginal farming activities such as this. However, some of the evidence is somewhat contradictory and because such areas are naturally dynamic the evidence for animal browsing has another plausible explanation. For example, the frequencies of *Alnus glutinosa* are lowest at the close of the zone, not at the beginning, when the increased pollen production from herbs in apparent. This suggests that any manipulation of the fen canopy by humans to facilitate grazing animals is not apparent. Also, if the local area of fen carr was being browsed by animals, frequencies of *Hedera helix* pollen, which is particularly palatable to animals, would be expected to be reduced considerably (Martyn Waller, pers. comm.). However, such a trend is not apparent.

Brown (1997b) has warned against over-interpreting possible evidence for human activity in pollen diagrams where the nature of the evidence (as in this case) could simply be the result of natural occurrences, such as wind-throw, which may create openings for the benefit of shrubs and herbs. Also, as is apparent in this case, the spatial resolution of pollen diagrams is not good enough to pick out the size of isolated clearings or elucidate the difference between events that may be related to the subtle effects of small scale human activity or natural woodland ecology (Brown, 1997b). Therefore, the evidence for cultural activities remains equivocal at best and such changes in local vegetation are as likely to result from wetland communities shifting position across the floodplain over time (Waller, 1998; Long *et al*, 1998c). This explanation maybe more appropriate given the length of time of the over which the increase in *Ilex aquifolium* pollen percentages takes place. The point at which frequencies of *Ilex aquifolium* pollen drop away is dated to 4850±45 BP (5655 to 5478 cal. yrs BP) and marks the end of the zone. This lpaz represents *c*.800 radiocarbon years, which is more compatible with natural ecological change than deliberate human alteration of the vegetation, although such an interpretation cannot be ruled out.

HKP-4 477.5 to 381.5 cm *c*.4800 to *c*.4500 BP (*c*.5500 to *c*.5200 cal. yrs BP)

Alnus glutinosa and *Quercus* pollen frequencies are generally low in this lpaz despite a spike in the centre of the zone. Frequencies of *Salix* are higher which might indicate local conditions became suitable for this shrub. One explanation for this may be that the floodplain environment opened up due to wetter conditions, which would cause local tree taxa such as *Alnus glutinosa* (and possibly *Quercus*) to be die away because of excessive waterlogging. Evidence in support of this is forthcoming from changes in several other morphotypes.

Frequencies of extra-local pollen such as *Tilia*, and regional pollen such as *Pinus sylvestris*, increase which may be the result of increasingly open conditions within the carr community. Relatively high regional input of *Pinus sylvestris* is a recorded at Eskamhorn (EKP-3) and Hirst Courtney (HCP-1) from *c*.5000 BP (*c*.5700 cal. yrs BP) and *c*.4200 BP (*c*.4700 cal. yrs BP) respectively. Smith (1985a) recorded high *Pinus sylvestris* frequencies during his HHL/A rpaz from the Humberhead Levels which dates from *c*.4300 BP (*c*.5200 to *c*.4600 cal. yrs BP). *Pinus sylvestris* wood has also been uncovered from the basal sandy deposits of Thorne Waste which suggests that this taxa was common prior to extensive paludification of the Humberhead Levels area (Smith, 1985a). This combined evidence confirms the earlier suggestion by Smith (1958a) that *Pinus sylvestris* was a relatively important component of this region at this time.

However, an alternative source for the *Pinus sylvestris* pollen in this lpaz is likely on litho- and biostratigraphic evidence. LOI values are lower during this lpaz, averaging just 29%, which suggests increased clay deposition associated with flooding of river water prior to the stratigraphic change at the top of the zone (*cf.* Devoy, 1979). Also, other indicators of river deposition, such as pre-Quaternary spores and possibly *Polypodium* and Pteropsida (monolete) indet., (Walker *et al.*, 1998) occur with higher frequencies during this zone which indicates that at least some of the increase in *Pinus sylvestris* may be attributable pollen floatation (*cf.* Hopkins (1950) and derived from riverine or estuarine sources (e.g. Robinson, 1993; Waller, 1994c).

Therefore, there is evidence during this zone for increased flooding, which created conditions too wet locally for extensive *Alnus glutinosa-Quercus* carr woodland. The fen woodland composition would have changed according to these wetter conditions with greater amounts of *Salix* present locally, with perhaps *Quercus*, *Corylus avellana*, and some of the *Alnus glutinosa* colonising less wet areas of floodplain. This is testified by the presence of two large *Salix* branches identified from between 396 and 426 cm (see Table 9.3) towards the close of the zone. The encroachment of brackish conditions is indicated by the increase in pollen of Chenopodiaceae in this lpaz, suggesting the development of saline marshes within the vicinity. Sediment from close to the end of this zone is dated to 4555±45 BP (5439 to 5043 cal. yrs BP).

HKP-5a 381.5 to 305.5 cm *c.*4500 to *c.*4000 BP (*c.*5200 to *c.*4400 cal. yrs BP)

Poaceae pollen percentages rise at the opening of this zone, *Salix* pollen is better represented and pollen percentages from *Alnus glutinosa* decline. This indicates the development of open areas of saltmarsh at the coring site, possibly with *Phragmites* reedswamp close by, and the further marginalisation of *Alnus glutinosa* so that the carr community represents pollen from extra-local sources, with *Salix* significant perhaps at the wetter edges at the back of the saltmarsh.

The lithostratigraphic change to organic clay sediments is also coincident with the deposition of palynomorphs indicating flooding by estuarine water. Frequencies of pre-Quaternary spores are particularly high at the opening of the zone, and significant proportions of *Pinus sylvestris* and foram test linings are indicative of inwashing of brackish water prior to marine inundation (Devoy, 1979; Wheeler, 1995). In addition to these derived microfossils, exotic pollen grains of *Picea* and *Abies* are frequently encountered on the pollen slides, and have been reported from other studies in coastal floodplain areas (e.g. Smith, 1958a; Waller, 1993). Palynomorphs such as these may be derived from processes of long distance transportation (pollen grains can be transported thousands of miles by marine currents, Muller, 1959; Rossignol, 1961; Stanley, 1969) or from the reworking of deposits of a substantially earlier age, such as the erosion of river terraces and interfluves (Scaife & Burrin, 1992). Whatever the source, the presence of pre-Quaternary spores and exotic pollen grains are typical within wave affected estuarine and marine sediments (Devoy, 1979; 1980; Farr, 1989; Tipping, 1994; Waller, 1994c; Long *et al.*, 1999).

Saltmarsh development is indicated either at or close to the site by the occurrence of halophytic herbs such as Chenopodiaceae, *Plantago maritima*, and *Artemisia*-type. The cereal pollen grains of *Hordeum*-type size class recorded in this context are probably derived from wild maritime wetland grasses such as *Elymus*.

HKP-5b 305.5 to 245 cm <*c*.4000 BP (<*c*.4400 cal. yrs BP)

The opening of this subzone is dated to 3950±45 BP (4517 to 4257 cal. yrs BP). The palaeoecological development of the local environment is difficult to interpret with certainty here due to problems concerning the pollen and diatom taphonomy (see section 9.6), and the absence of pollen concentration data. This also reflects the difficulty in identifying the autochthonous microfossil community in marginal river/estuary environments and, the lack of contemporary information that can assist in reconstructing such palaeoenvironments.

The development of a more freshwater river marsh locally is indicated by the increase in *Filipendula* pollen. *Ranunculus acris*-type, *Plantago lanceolata*, and Cyperaceae pollen may also belong to this fen meadow community. The high frequencies of *Filipendula* pollen suggest that there were conditions surrounding the coring site that were not strongly saline. However, the continued presence of foram test linings and pollen of Chenopodiaceae, *Plantago maritima*, Asteraceae (Cardueae/Asteroideae), and *Artemisia*-type indicates saltmarsh with strong brackish influence locally at the coring site, although the lack of epiphytic diatoms may suggest the vegetation cover was limited (see section 9.6, Figure 9.10).

Pollen concentrations would be of use in this situation, as it is probable that a decrease in the amount of local pollen, and perhaps an increase in sediment accumulation rates would cause pollen concentrations to fall. This would be evidence to suggest the high frequencies of *Filipendula* pollen are derived from high freshwater marshes beyond estuarine influence and not from the immediate site. The increased extra-local input of *Alnus glutinosa* and *Salix* pollen from marginal carr communities during this zone may also be attributable to this decrease in input of local pollen. Therefore, mud flat and saltmarsh communities are tentatively inferred at the site, with freshwater conditions prevailing at the higher marsh areas away from direct marine influence. Analysis of macrofossil remains in the core may also reveal seeds which may clarify the situation. The identification of the forams in the core at this point may also help elucidate the exact nature of the local depositional conditions providing a source for the *Filipendula* pollen.

A decrease in tidal flooding is suggested by the decline in pre-Quaternary spores and *Pinus sylvestris* during this lpaz. The may suggest the stabilisation of the saltmarsh environment (perhaps a relative rise in elevation if sedimentation rates are faster than the water level rise) or a change in the source area of river-borne pollen. If the river estuary environment has stabilised (LOI values increase slightly perhaps suggesting more vegetation), then it might be expected that input of secondary pollen, from reworking of older river sediments or marine sources, would decline. Input from freshwater dominated wetland areas may then increase proportionately, which may explain the higher *Filipendula* frequencies in this lpaz. Freshwater may also be draining towards the river from the adjacent freshwater wetlands, through the saltmarshes and mudflats, which will transport pollen from this community onto the lower brackish areas. Brown (1996) has suggested that the waterborne pollen spectra probably more closely reflects the vegetation of such marginal wetland areas than the wind-blown component.

Extra-local pollen from dryland taxa such as *Ulmus*, *Tilia*, and probably *Quercus* and *Corylus avellana*-type are relatively low, suggesting an increasing distance between these communities and the coring site or perhaps some clearance of the surrounding valley sides. A spike in *Fraxinus excelsior* pollen at the opening of the zone suggests the presence of this taxon in close proximity to the coring site.

HKP-5c *c.*245 to 150 cm <*c.*4000 BP (<*c.*4400 cal. yrs BP)

The pollen stratigraphy within this lpaz again indicates the apparent presence of two marsh environments in close proximity. Saltmarsh is indicated by the abundance of Chenopodiaceae along with *Plantago maritima*, *Artemisia*-type, Asteraceae (Cardueae/Asteroideae), foram test linings and perhaps Poaceae. Freshwater communities are on the other hand, indicated by the presence of taxa such as *Filipendula* and *Iris* and probably some Poaceae, with Cyperaceae increasing in abundance towards the top of the lpaz. *Alnus glutinosa* and *Salix* pollen suggest the presence of extra-local carr woodland surrounding the marshes. Increased washing-in of estuarine microfossils such as *Pinus sylvestris*, pre-Quaternary spores and *Picea*/*Abies* is also evident.

Pollen taxa such as *Filipendula* and Cyperaceae increase in abundance as aerophilous diatoms dominate the ldaz (see section 9.6) which indicates supratidal conditions, although occasional flooding by brackish water is still evident. It is feasible that estuarine conditions withdrew, and the freshwater dominated environment close to the coring site encroached on the saltmarsh community and spread closer to the coring site.

Quercus and *Corylus avellana*-type pollen are abundant. This suggests these taxa were growing within the surrounding extra-local or regional dryland community, with openings perhaps indicated by better representation of *Taxus baccata*.

9.6 Palaeohydrological Interpretation

Diatom diagrams are presented from the laminated humic clay towards the base of the sequence (Figures 9.5, 9.6, and 9.7) and the organic clays towards the top of the core (Figures 9.8, 9.9, 9.10).

HKLD-1 925 to 880 cm <c.7000 BP (<c.7800 cal. yrs BP)

Figure 9.5, 9.6 and 9.7

Oligohalobous-indifferent diatom taxa such as *Gyrosigma attenuatum* and *Amphora ovalis* dominate this zone. The lack of aerophilous taxa along with these epipelic species indicate submerged freshwater, muddy conditions. The rising frequencies of allochthonous planktonic taxa such as *Cyclotella striata* indicate increasing brackish conditions in the lagoon due to input of estuarine water. Epiphytic diatoms such as *Cocconeis placentula* (and perhaps *Achnanthes lanceolata*, Vos & de Wolf, 1993a) indicate shallow water with emergent aquatics close by. Broken valves of *Pinnularia* sp. are abundant only in the basal level which suggests the marginalisation of fen peat forming environments prior to flooding and lagoon formation (*cf.* Verbruggen & Denys, 1989; Denys, 1994; Vos & de Wolf, 1994). A shallow water, muddy environment is envisaged, with emergent aquatics close by and fen peat communities in the vicinity.

HKLD-2a 880 to 770 cm <*c*.7000 to >*c*.6000 BP (<*c*.7800 to >*c*.6800 cal. yrs BP)

The allochthonous component dominates this ldaz as the planktonic species *Cyclotella striata* increases to high frequencies. This suggests frequent tidal flooding of brackish river water into the lagoon environment. Frequencies of oligohalobous-indifferent species such as *Amphora ovalis* decrease in response to the increase in salinity, whereas more euryhaline taxa such as *Cocconeis placentula* (which colonises fresh/brackish lagoons, Vos & de Wolf, 1988) and *Gyrosigma attenuatum* continue to prevail due to their broad salinity tolerance (Vos & de Wolf, 1993a). An assemblage with species such as *Gyrosigma attenuatum*, *Gyrosigma acuminatum* and *Nitzschia tryblionella* has close similarities with the *Nitzschia debilis* group identified by Vos & de Wolf (1988). This group of diatoms colonises clayey sediments in the shallow subtidal lagoons where water is sufficiently clear for photosynthesis (Vos & de Wolf, 1988).

This ldaz suggests permanently submerged lagoonal conditions with fresh-brackish conditions indicated by increasing input of estuarine water. This may have the effect of ponding back freshwater, and increasing the water depth in the lagoon.

HKLD-2b 770 to 685 cm >*c*.6000 BP (>*c*.6800 cal. yrs BP)

Polyhalobous taxa increase in this ldaz indicating an increase in estuarine influence. Brackish riverine inputs are dominant, indicated by the continued abundance of the estuarine tidal channel species *Cyclotella striata*. Planktonic taxa such as *Paralia sulcata* and *Thalassiosira eccentrica* are allochthonous, and suggest marine conditions are expanding into the lower parts of the valley systems comprising the inner Humber. Brackish epipelic species such as *Nitzschia navicularis*, which can colonise clayey substrates around subtidal brackish lagoons (Vos & de Wolf, 1988) increase in frequency, and suggest a further salinity increase. The inferred increase in salinity is likely to have caused the decline in *Cocconeis placentula*, whereas *Gyrosigma attenuatum* is capable of surviving higher salinity conditions (Vos & de Wolf, 1993a).

The increase in fragments of unidentified *Pinnularia* spp. diatoms at the close of the subzone along with aerophilous species such as *Caloneis bacillum*, is suggestive of the silting-up of the lagoon and the reestablishment of aerial semi-terrestrial peat forming communities at the site (*cf.* Verbruggen & Denys, 1989).

The sequence of events described in these three ldazs is very similar to the lagoonal conditions described from the 'Velsen layer' of the 'Calais deposits' in the lower Schelde river valley in the Dutch Province of Zeeland, The Netherlands (Vos & de Wolf, 1994; 1997). Vos & de Wolf (1997) suggest these lagoons formed from *c.*8000 BP (c.9200 to 8500 cal. yrs BP) and were sheltered from the sea initially by sand flats (indicated by the occurrence of epipsamic diatoms which live attached to sand grains). Within the lower part of this unit, strongly dissolved *Pinnularia* sp. are present, suggesting the presence of fen peat environments (Vos & de Wolf, 1994). High frequencies of freshwater diatoms are recorded at the base, with a combination of benthic and epiphytic groups indicating a permanently submerged lagoonal environment, and low salinity after the drowning of the peat (Vos & de Wolf, 1994; 1997). Euryhaline epiphytic taxa such as *Amphora ovalis* and *Cocconeis placentula* are prominent (as at Hook Lane) in the lower lagoon sediments and frequencies of *Cyclotella striata* gradually increase, suggesting a small input of planktonic taxa from the coastal zone. The relative abundance of brackish and marine planktonic diatoms increase towards the top of the layer suggesting increasing marine influence. Finally, at the top of the layer, Vos & de Wolf (1994; 1997) record an increase in benthic marine taxa such as *Nitszchia navicularis*, suggesting increasing salinity and silting up of the floodbasin, with intertidal conditions prevailing before a return to fen carr conditions. These factors influencing the formation of these lagoons is discussed in section 10.5, and it is suggested that similar processes can be invoked to explain the evolution of this sedimentary environment at Hook Lane.

HKUD-1 390 to 340 cm *c.*4600 to *c.*4200 (*c.*5200 to *c.*4700 cal. yrs BP)

Figure 9.8, 9.9 and 9.10

The change to organic clay sedimentation is coincident with the regular flooding of estuarine water onto the site, indicated by the high frequencies of the brackish planktonic species *Cyclotella striata* (Vos & de Wolf, 1993a). High saltmarsh conditions are suggested by the prevalence of benthic aerophilous taxa such as *Diploneis interrupta* (Denys & Baeteman, 1995), although limited vegetation cover is perhaps indicated by the low abundance of epiphytic taxa and the frequency of marine and brackish epipelic species such as *Campylodiscus echeneis, Diplonies didyma, Nitzschia granulata,* and *Nitzschia* *navicularis* (Vos & de Wolf, 1993a). Polyhalobous and mesohalobous diatoms dominate the assemblage suggesting saline marshes are locally present with freshwater environments peripheral to the actual coring site.

HKUD-2 340 to 220 cm *c.*4200 to <*c.*4000 (*c.*4700 to <*c.*4400 cal. yrs BP)

Polyhalobous planktonic species such as *Paralia sulcata* increase in frequency at the opening of this zone suggesting the expansion of marine conditions into the inner estuary and lower Aire valley. *Cyclotella striata* remains abundant at the opening of the zone suggesting the continued input of allochthonous brackish tidal channel diatoms onto the site.

Mesohalobous epipelic species such as *Navicula peregrina* and *Nitzschia navicularis* are common in this ldaz. These are likely to be part of the autochthonous component which suggests the presence of intertidal or lower supratidal mudflats close to or at the site. Frequencies of the aerophilous high saltmarsh species *Diploneis ovalis* increase in the centre of the zone as polyhalobous, planktonic species decline in abundance. This suggests the development of supratidal marshes close to the site, which is consistent with the increase in abundance of *Filipendula* and decrease in flooding of estuarine palynomorphs at site shown in the pollen diagram during HKP-2b (Figure 9.3). The local brackish conditions suggest mudflat and saltmarsh environments at the site, and freshwater marshes at higher localities marginal to the site.

An increase in marine flooding is suggested by the rise in *Paralia sulcata* towards the end of this ldaz. This also coincides with an increase in frequency of marine palynomorphs at the opening of HKP-5c. This is more likely to be the result of a proximal tidal channel moving closer to the site than an increase in sea-level rise.

HKUD-3 220 to 150 cm *c.*4200 to <*c.*4000 (*c.*4700 to <*c.*4400 cal. yrs BP)

Supratidal saltmarsh and mudflat conditions are indicated in this ldaz as the fresh-brackish aerophilous taxa *Diploneis interrupta* and *Diploneis ovalis* increase to prominence along with epipelic species such as *Nitzschia navicularis*. This indicates that the duration and frequency of marine inundation is less and the surface of the marsh and mudflat is subaerially exposed for large parts of the tidal cycle (Denys & Baeteman, 1995; Vos & de Wolf, 1997). Marine influence is still indicated by washing in of allochthonous valves of *Paralia sulcata*, although generally polyhalobous and planktonic diatoms are less frequent in this ldaz.

This is consistent with the pollen evidence for proximal freshwater marshes close to the site and increasingly dry and freshwater conditions encroaching on the site.

9.7 Hook Lane Palaeoenvironmental Summary

The patterns and processes of wetland development at Hook Lane inferred from pollen and diatom evidence are summarised in this section and in Figure 9.11. Paludification of the valley bottom has been dated to 7030±45 BP (7914 to 7699 cal. yrs BP) and indicates rising water tables in the lower part of the lower Ouse and Aire river valleys at this time. Wetland and aquatic herbaceous communities with Cyperaceae and *Sparganium erectum* are indicated locally, which are soon replaced by *Alnus glutinosa* dominated carr vegetation.

This fen carr phase is relatively short lived and sometime after *c*.7000 BP (<c.7800 cal. yrs BP), the *Alnus glutinosa* woodland is flooded and replaced by a 'fluviolagoonal' environment. Occasional freshwater aquatic pollen is recorded although the diatom evidence suggests relatively brackish conditions. Reedswamp conditions are suggested at the shallow lagoon margins, with *Alnus* carr beyond this wetland fringe. The removal of local vegetation due to the permanently submerged conditions enables the extra-local input of the surrounding dryland vegetation, which is dominated by *Ulmus*, *Quercus*, *Tilia* and perhaps *Corylus avellana* to be recognised.

Increasing water depth is perhaps indicated during this lagoon phase as water levels continued to rise and estuarine conditions pushed further into the lower Aire valley. An increase in polyhalobous diatoms during HKLD-2b towards the end of the lagoon conditions is accompanied by greater frequencies of marine and brackish epipelic taxa, indicating intertidal conditions as the lagoon began to silt up. This eventually enabled *Alnus* carr communities re-invade the site. This transition to semi-terrestrial wetland and carr communities is dated to 5965±45 BP (6889 to 6719 cal. yrs BP).

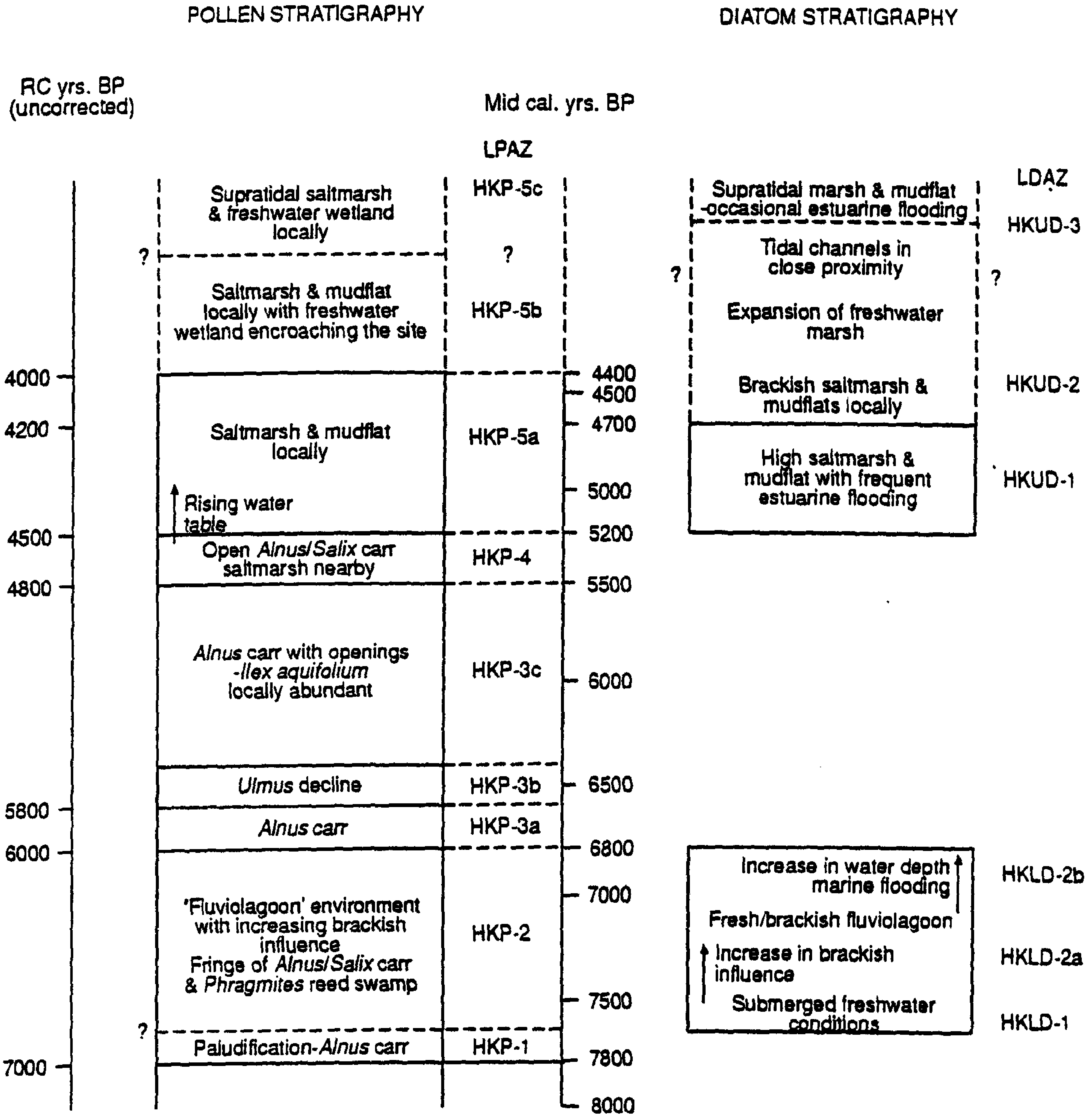


Figure 9.11 Hook Lane palaeoenvironmental summary diagram

A decline in *Ulmus* pollen at the opening of HKP-3b is estimated at *c*.5800 BP (*c*.6600 cal. yrs BP) although the cause is uncertain. Diachroneity of this event in the lower Aire valley is suggested by the difference in dates at Eskamhorn and Hook Lane, but this may be due to the poor resolution of the interpolated age estimation, and the actual age may overlap with the date at Eskamhorn. The weak manifestation of the traditional mid-Holocene *Ulmus* decline event, at Hook Lane may be due to the overrepresentation of local fen carr pollen taxa and expansion of the floodplain edge, which may also blur the exact point at which *Ulmus* pollen begins to decline in the biostratigraphic record.

A shift in composition of the local carr community is indicated within lpaz HKP-3c as *Ilex aquifolium* and a range of fen herbs become more abundant. Opening up of the local carr community is inferred to explain the increased pollen production of these taxa. This phase is dated between 5655±45 BP (6528 to 6313 cal. yrs BP) and 4850 BP (5655 to 5478 cal. yrs BP). Pastoral grazing activities are tentatively suggested as one possible cause of this shift in vegetation, although the spatial resolution of pollen diagrams is inadequate to support an anthropogenic cause with certainty, and natural local wetland vegetation dynamics could also be invoked to explain this sequence of changes (*cf.* Brown, 1997b).

Increasing flooding and higher water tables are suggested in lpaz HKP-4, prior to the transition to saltmarsh vegetation communities in HKP-5a, which has been dated to 4555±45 BP (5439 to 5043 cal. yrs BP). *Alnus glutinosa* and *Salix* dominated carr communities are probably present, fringing the saltmarsh areas. Expansion of freshwater communities is indicated close to the site as *Filipendula* increases in HKP-5b, dated to 3950±45 BP (4517 to 4257 cal. yrs BP). Diatom evidence suggests both saltmarsh and exposed areas of mudflat are present locally in HKUD-2, and the frequency of marine flooding is reduced.

Supratidal conditions are suggested in HKUD-3 although marine inundation is still in evidence in both the diatom assemblage and the pollen in HKP-5c. Drier conditions prevail, with freshwater wetland vegetation communities including *Filipendula* and Cyperaceae occurring close to the site or even beginning to invade the local saltmarsh and mudflat conditions. This is tentatively interpreted as a withdrawal of marine conditions at the top of the sequence.

Chapter 10 Discussion

10.1 Introduction

This chapter compares and combines the environmental changes at the three sites studied in order to reconstruct the mid- to late-Holocene evolution of the lower Aire valley. On- and off-site vegetation dynamics are discussed and the wider implications of the palaeobotanical results are considered. The major controls on sedimentary palaeoenvironments within the perimarine zone are discussed in terms of the relative importance of catchment and estuarine processes. The evidence for tendencies of sea-level movement within the lower Aire valley over the Holocene are analysed and compared with other dated events from elsewhere in the estuary. A graph showing the timing and altitude of increases and decreases in water table from this study is plotted and compared against the previous database of sea-level data from the Humber (Long *et al.*, 1998a and references therein; Neumann, 1998). Finally, the effects of sea-level change on valley sedimentation within the lower Aire perimarine zone are evaluated and the balance between relative inputs from the fluvial and estuarine system is discussed.

10.2 Vegetation Dynamics

Three palaeobotanical records covering the mid-Holocene period have been presented from sites in the lower Aire valley (Chapters 7, 8 and 9). A summary of local vegetation changes from these sites is presented in Figure 10.1, which illustrates the chronological sequence of events. This summary of the sedimentological and vegetational evolution of the lower Aire valley from the mid-Holocene, forms the basis of the following consideration of autogenic and allogenic controls on wetland development.

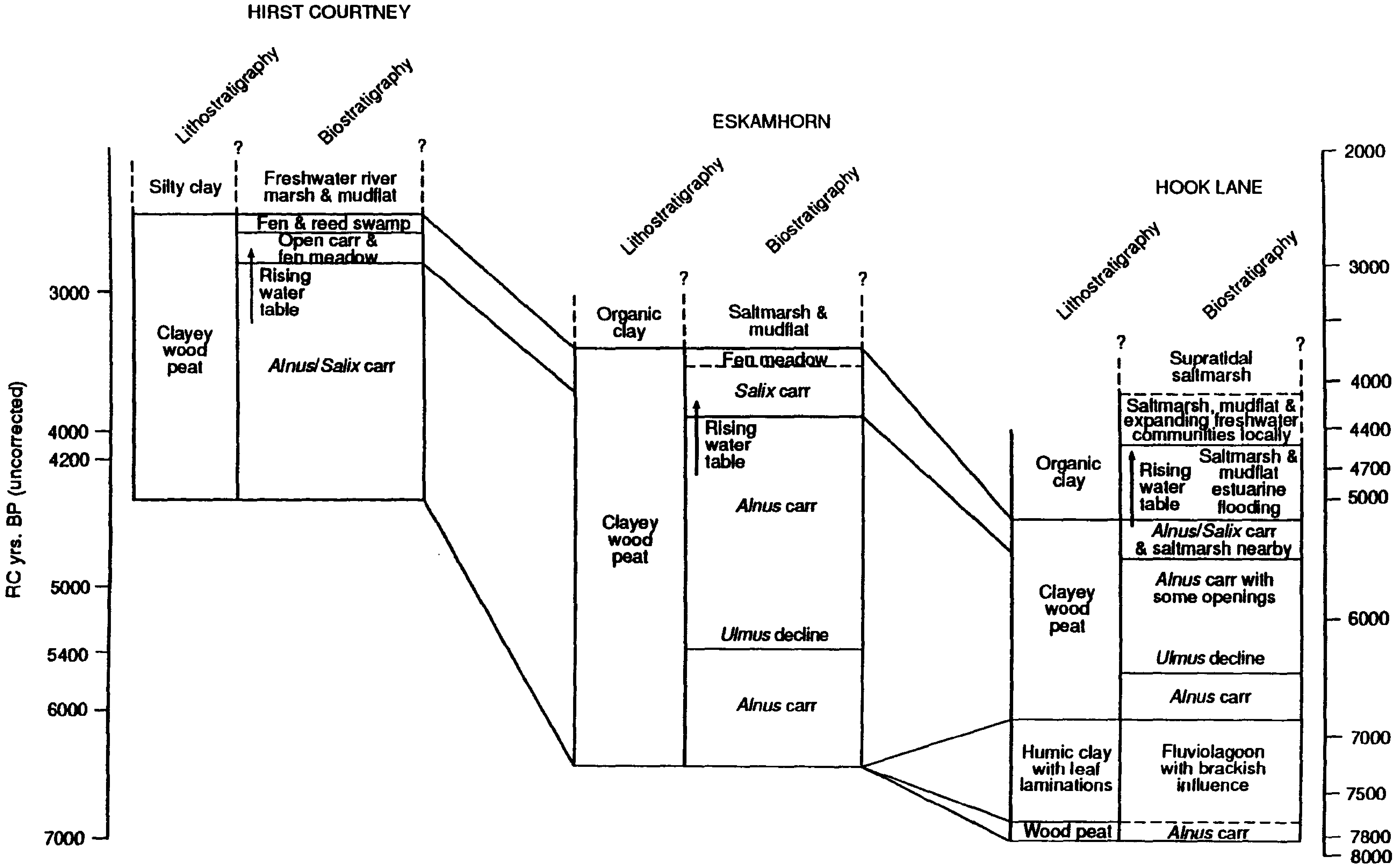


Figure 10.1 Summary of vegetation changes in the lower Aire Valley

10.2.1 Local Vegetation Communities

The focus of this research is on-site vegetation dynamics, as they provide insights into the processes controlling vegetation change on the floodplain. The pollen record from the floodplain sediments and backswamp peats at each of the study sites indicates for much of the Holocene, the dominant community is fen carr dominated by *Alnus glutinosa* (see Figure 10.1).

The Holocene history and palaeoecology of *Alnus glutinosa* in the British Isles has been a subject of considerable debate (McVean, 1956b; Smith, 1965; Smith & Pilcher, 1973; Huntley & Birks, 1983; Smith, 1984; Chambers & Price, 1985; Bush & Hall, 1987; Waller, 1987; Brown, 1988; Birks, 1989; Chambers & Elliot, 1989; Bennett & Birks, 1990; Tallantire, 1992; Day, 1995), although there is agreement over the time transgressive nature of the early-Holocene rise in *Alnus* pollen. Therefore, despite the fact that no information is available in this study to contribute to the debate on the Lateglacial/early-Holocene status of *Alnus glutinosa*, the results presented in Chapters 7, 8 and 9 provide an insight into the mechanisms resulting in the dramatic spread of this species.

The nature of the spread of *Alnus glutinosa* populations is erratic both in space and time, with dates for its expansion ranging several millennia between *c.*8500 and *c.*5000 BP (*c.*9800 to *c.*5500 cal. yrs BP, Bennett & Birks, 1990). Only two dates have been obtained for the *Alnus* rise in the Humber basin prior to this study. In the river Trent valley at Bole Ings, Dinnin & Brayshay (1994) and Brayshay & Dinnin (1999) provide a relatively late age for the expansion of *Alnus glutinosa* populations in the valley bottom at 6290±70 BP (7361 to 7012 cal. yrs BP). Gearey & Lillie (1999) provide an earlier date of 7720±50 BP (8557 to 8372 cal. yrs BP) from Askham Bog in the Vale of York, although the low $\delta^{13}C$ values suggest that this age estimate may be too old owing to a hard water error (Bayliss *et al.*, 1999) and therefore provides only a *terminus post quem* for this event. However, this date is similar to the age obtained for the *Alnus glutinosa* rise in the Vale of Pickering of 7640±85 BP (8551 to 8189 cal. yrs BP) by Day (1995).

The earliest date for the presence of *Alnus glutinosa* in the lower Aire valley is 7030±45 BP (7914 to 7699 cal. yrs BP) and comes from the basal sample at Hook Lane. *Alnus glutinosa* pollen values are already high (46% TLP+*Alnus*) in this basal level, so this age provides a *terminus ante quem* for *Alnus* expansion at this site (see Chapter 9).

The spread of *Alnus glutinosa* in Britain has been attributed to a range of causal factors. Godwin (1940a; 1975a) proposes that it is a response to increasing climatic wetness, whereas various later authors suggest that Mesolithic human activity (e.g. woodland clearance and fire) may have facilitated the establishment of *Alnus* by reducing local competition and promoting waterlogging (Smith, 1984; Chambers & Price, 1985; Edwards & MacDonald, 1991; Tipping, 1995; Leah *et al.*, 1997). Chambers & Elliot (1989) propose that beaver activity may provide an agency for the spread of *Alnus* although Brown (1988) opposes this view, suggesting that changing local edaphic conditions (i.e. substrate stability and decreased drainage due to changing catchment controls) may have enabled *Alnus glutinosa* to dominate, by out-competing other floodplain species such as *Salix*. Bennett & Birks (1990) note that the erratic expansion of *Alnus glutinosa* communities reflects the sporadic availability of suitable habitats due to rising sea levels, hydroseral succession and floodplain development.

The pattern of *Alnus glutinosa* carr establishment and basal floodplain peat development along the river Aire is time transgressive in nature, occurring later up the valley, and it is most likely that these *Alnus glutinosa* communities are able to develop as a result of an increase in habitat availability, due to waterlogging of the floodplain backswamp areas as fresh river water was ponded back ahead of estuarine conditions, which continued to push into the lower Ouse and Aire valleys from *c.*7000 BP (*c.*7900 to 6600 cal. yrs BP, see section 10.4.5).

Such communities persist, largely uninterrupted for *c.*1500 radiocarbon years at Hirst Courtney, *c.*2700 at Eskamhorn, and *c.*1500 at Hook Lane. Brown (1988; 1996; 1997) suggests that the stability and limited diversity of *Alnus glutinosa* woodland is attributable to waterlogging and substrate stability, which probably made these communities non-invadable. Once established, these fen carr communities seem to have been robust and capable of withstanding minor environmental perturbations due to their considerable inertia (*cf.* Smith, 1965). Nevertheless, as the floodplain community evolves over time, or is forced into change due to local channel avulsion or external processes such as windthrow (Brown, 1997b), the mosaic of wetland carr vegetation may change in composition, although *Alnus glutinosa* usually remains the dominant species. Along the lower Aire valley, it appears likely that *Corylus avellana* was a significant component of the fen carr community, with other tree and shrub taxa surviving, such as *Fraxinus excelsior*, *Quercus*, *Betula*, *Taxus baccata*, *Salix*, *Vibernum opulus*, *Euonymous europaeus*, *Frangula alnus*, *Crataegus*, *Sorbus*, *Prunus*, and *Ilex aquifolium*, depending on local conditions. The palynological representation of the ground flora is likely to be suppressed by lack of flowering due to the shaded conditions, although a range of fen herbs are indicated by the presence of pollen types such as such as Poaceae, Cyperaceae, *Filipendula*, *Caltha palustris*-type, *Ranunculus acris*-type, *Lysimachia vulgaris*-type, *Rubus*-undif., and Rubiaceae, which are recorded in the pollen diagrams at low frequencies.

The longevity of this fen carr community however, is not consistent with previous theories of ecological change. The successional pathways proposed by Tansley (1939) and Walker (1970) suggest such environments are seral stages which naturally dry out, and will be replaced by *Quercus/Betula* woodland or continue into poor fen and bog. Other authors have noted a lack of internal regeneration within alder carr communities (e.g. McVean, 1953; 1956c; 1956d; Vinther, 1983; Grime *et al.*, 1988; Bennett & Birks, 1990). However, McVean (1953) suggests rising water levels will maintain the stability of the vegetation by allowing regeneration. The likelihood of autogenic vegetation changes being affected by allogenic processes in coastal areas is high (Waller *et al.*, 1999). Therefore, the reason for the persistence of *Alnus glutinosa* in perimarine areas and floodplain corridors, which suggests eutrophic fen carr communities can form stable vegetation communities, is given consideration here.

The persistence of *Alnus glutinosa* carr in coastal hinterlands has been linked to continued sea-level rise, which maintains a high water table (Kidson & Heyworth, 1973; Smith & Morgan, 1989; Waller, 1993; 1994a; 1994c). However, recent research in the Romney Marsh area of southern England suggests that, for this to be the sole control, the rate of sea-level rise would have to maintain the same pace as peat accumulation in the alder carr for *c.*2000 years, which seems unlikely (see Long *et al.*, 1998c; Waller, 1993; Waller *et al.*, 1998; 1999). Nevertheless, the dominance of *Alnus glutinosa* communities during the mid-Holocene at many peripheral coastal locations (including the lower Aire valley) does occur against a background of generally rising water levels. Although the maintenance of a high water table due to rising sea levels is important in explaining the persistence of alder carr in perimarine areas, Waller *et al.* (1998; 1999) suggest that other factors are also likely to be important, such as cycles of competition between

Alnus glutinosa and *Salix* (which may be masked due to the over and under-representation of these respective pollen types), and variations in the production and decomposition of organic material.

The decline in alder woodland in the lower Aire valley seems to be related to hydrological change, as opposed to extensive human deforestation, land drainage for agricultural purposes, or autogenic processes. At each site, the vegetation response suggests a continued rise in water table, which outstrips the peat accumulation rate and drowns the local *Alnus* carr. This community is then replaced either by *Salix* carr communities (see Chapter 7) which are more tolerant of high water levels (Dupont, 1987) or fen meadow communities (dominated by *Filipendula*, see Chapter 8), suggesting allogenic influence. However, as at floodplain sites in the Severn (Brown, 1982; 1988) and Combe Haven valley (Smyth & Jennings, 1988; 1990), this decline is highly diachronous in the lower Aire valley, from as early as 4850±45 BP (5655 to 5478 cal. yrs BP) at the lowest site, Hook Lane, to 2665±45 BP (2850 to 2740 cal. yrs BP) at Hirst Courtney in the upper part of the study reach. It seems likely that in perimarine river valley contexts, *Alnus* carr will persist until the rate of sea-level rise overtakes the rate of organic sediment accumulation, which leads to a shift in ecotones on the floodplain, with marsh and mudflat environments expanding at the expense of areas of former fen carr.

The herbaceous wetland communities that existed within the lower Aire subsequent to the marginalisation of fen carr due to the rising water table are poorly understood. Along the margins of middle and outer estuarine areas, the progressive succession may be expected to change through seres characterised by wood fen, into reedswamp, before succumbing to saltmarsh conditions as marine-brackish water floods the site (see Shennan, 1986b; Waller, 1994c, and Figure 2.1), although Wheeler (1995) records a different transitional, permanently submerged palaeoenvironment, that existed landward of saltmarsh deposits, from Holocene deposits in the Fenland.

Vegetation communities in perimarine areas such as the lower Aire respond to the rising water levels but the nature of this transition and the resulting palaeoenvironment is different. High frequencies of *Filipendula* are characteristic of this changing environment, but there is a lack of contemporary information from modern analogues regarding the environment reflected by such pollen assemblages, and the precise controls on this vegetation community. *Filipendula ulmaria* is common in a narrow band of fen meadow, often on silty substrates, along rivers in Broadland (Wheeler, 1984) but the non-tidal nature of these environments suggests they develop as a response to a different control. In the upstream freshwater area of Hirst Courtney (see section 7.5), *Salix* supplants this fen community until the site apparently becomes sufficiently wet, that aquatic reedswamp marsh communities prevail, dominated by Poaceae, Cyperaceae, and *Sparganium erectum*. Similar conditions are recorded in the lower Trent valley at Bole Ings at the same time as these changes at Hirst Courtney (Brayshay & Dinnin, 1999), so these new results may preclude the anthropogenic cause the authors suggest (see below).

At Eskamhorn (see section 8.5), this fen meadow vegetation is interpreted as a result of freshwater ponding behind the advancing estuarine tide. The apparent demise of this freshwater wetland community is related to increasing environmental salinity, as it becomes replaced by saltmarsh conditions, which represents an ecotone shift. The continued existence of freshwater wetland communities consisting of taxa such as *Filipendula* after the establishment of saltmarsh probably represents vegetation growing higher on the floodplain where flooding of brackish water is infrequent. At the most seaward site, Hook Lane, the registration of rising water table is not associated with any increase in herbaceous fen communities but the *Alnus* carr community thins below the sedimentary contact, and the vegetation appears to change straight into saltmarsh, with no evidence of an intermediate transitional phase (see section 9.5). Although the switch to transitional fen or reedswamp community may be relatively rapid, the pollen sampling interval of 2 cm over this horizon (representing *c*.20 calibrated years) should be sufficiently high resolution to detect any subtle or sudden change in vegetation (assuming no erosion). This direct pathway of vegetation change has also been recorded by Waller (1994a) in the Brede valley, East Sussex.

Whilst this may be an artefact of uneven sedimentation owing to the presence of two large pieces of *Salix* wood which interrupt the biostratigraphic record just below the transition to saltmarsh, possible alternative reasons for the absence of an intermediate freshwater wetland include: that the rate of sea-level rise was more rapid than the sites upstream (although there is no evidence to support such an assertion), or that the site, owing to its seaward location, was influenced by brackish conditions as soon as water levels began to rise which inhibited the development of *Filipendula* dominated marshes. The subsequent inferred encroachment of *Filipendula* communities onto the site in lpaz HKP-5b may be related to the establishment of high marsh conditions with high freshwater input (*Diploneis ovalis* is abundant) and a reduction in marine flooding. This suggests the existence of fen meadow communities away from direct and frequent estuarine flooding during the period of saltmarsh deposition (see sections 9.5 and 9.6).

It is interesting to note the lack of acidification in these floodplain peats, considering the original ideas of Tansley (1939) and Walker (1970), the presence of raised mires within the Humberhead Levels, and the fact that many coastal peat communities can be shown palynologically to develop into ombrotrophic bog conditions (Godwin, 1943; 1945). For example, peat sequences in the Somerset Levels, Gwent Levels, Cardigan Bay, Fenland, Romney Marsh, and the Belgian and Dutch coastal plain become isolated from the groundwater and develop ombrogenous vegetation (Beckett & Hibbert, 1979; Wilks, 1979; Housely, 1988; Smith & Morgan, 1989, Allermersch, 1991; Pons, 1992; Waller, 1994c; Rippon, 1996; Waller *et al.*, 1998; 1999). Acidophilous bog communities are absent from other inner estuarine locations such as the Thames, Solent, and Severn (Devoy, 1979; 1980; Long & Tooley, 1995; Hewlett & Birnie, 1996). It is likely that in such areas, the continual input of eutrophic waters into the rivers and streams draining the uplands, was high enough that the vegetation communities along the floodplain corridor never became sufficiently isolated from base rich inputs for the transition from eu- to meso-, and ultimately oligotrophic vegetation to occur (*cf.* Giller & Wheeler, 1988).

It has been suggested that the observed shift to bog communities in the Romney Marsh area towards the late Holocene (from *c.*4000 cal yrs BP), may have been induced by considerable slowing down in rate of sea-level rise. This would have promoted vertical isolation of the sediment surface from the base rich water in areas away from the adjacent uplands (Waller *et al.*, 1999). In the Humber estuary at this time, water levels appear to be rising, and positive tendencies in sea-level are recorded both in the outer estuary areas (Long *et al.*, 1998a) and the inner areas such as the lower Aire valley. These conditions would be sufficient to maintain eutrophy, and are ideal for promoting allogenic changes from fen carr through to fen meadow and saltmarsh (or freshwater riverine marsh). This situation is not conducive to the development of acidophilous vegetation.

10.2.2 Dryland Vegetation Signal

The character of the dryland woodland community surrounding the floodplain corridor of the lower Aire valley is difficult to ascertain due to the dense local vegetation, and the overrepresentation of pollen from floodplain sources. The pollen source area from such backswamp floodplain sites is generally believed to reflect local and extra-local zones (Jacobson & Bradshaw, 1981; Brown, 1996; Moores *et al.*, 1999). In addition, the precise nature and composition of the vegetation of the upland environment is difficult to reconstruct due to the considerable overlap in ecological niche of many of the tree species. For example, tree and shrub taxa such as *Quercus*, *Alnus glutinosa*, *Fraxinus excelsior*, *Corylus avellana*, *Ilex aquifolium*, and *Taxus baccata* could all occur in both dry and wetland vegetation systems (Waller, 1994c; Rodwell, 1991). A similar drawback is associated with the palynological analysis of herbaceous plant types, many of which cannot be identified below family or genus level. Many of the pollen types that may be used to infer open disturbed areas in upland regions, often associated with human activity, have members that occur naturally in fen systems and saltmarshes (Wheeler, 1980a, 1980b; Rodwell, 1995). Therefore, palaeovegetational reconstructions from such peats, and in particular, identifying anthropogenic impact within such contexts, is fraught with pitfalls (Jones, 1988; Waller, 1993).

Owing to these problems inherent in studying dryland vegetation change from floodplain contexts, the construction of rpazs is not attempted in this study. However, some indication of the nature of the extra-local vegetation growing on the surrounding valley sides, and further afield within the regional landscape can be obtained from the pollen record, when the fen carr community opens up slightly, or before extensive paludification ensues.

Although the picture of off-site vegetation is clouded and fragmented, it is possible to make comparisons with existing palynological research from the region. Data from the lower Aire valley is compared with the regional pollen assemblage zonation scheme proposed by Smith (1985a) for the Humberhead Levels, based on extensive palynological work on Thorne and Hatfield Moors (see section 4.3). A comparison is also made between the Holocene vegetation history characteristic of the contrasting lithologies of the sand and clay glacial outwash deposits of the Humberhead Levels and the chalk deposits of the Holderness and Yorkshire Wold area (Beckett, 1981; Bush, 1986; 1993; Bush & Ellis, 1987; Bush & Flenley, 1987). This overview is presented in the following four sections.

Pre-peat vegetation

The basal pollen samples from the cores provide information concerning the nature of the dryland vegetation prior to *Alnus* carr domination. Because closed carr was absent, the area of pollen recruitment is likely to be larger, which may represent the vegetation colonising the dry valley areas and perhaps beyond in the surrounding landscape. The earliest record is from Hook Lane where a date of 7030±45 BP (7914 to 7699 cal. yrs BP) has been obtained from the peat/basal soil contact (see section 9.5). *Quercus*, *Ulmus*, and *Corylus avellana*-type pollen are relatively high indicating these taxa were prevalent, at least within the extra-local source area during this time. *Tilia* pollen is less well represented although this cannot be interpreted as a lack of *Tilia* trees in the vicinity considering the generally underrepresented nature of this taxa due to the poorly dispersed nature of its pollen grains and the expanding wetland area at this time (Andersen, 1970; Huntley & Birks, 1983; Greig, 1982a; Waller, 1994b). *Pinus sylvestris* pollen is present at low frequencies, and interpreted as coming from regional pollen representing vegetation growing beyond the immediate catchment area.

At Eskamhorn, the onset of peat formation is dated to 6385±45 BP (7380 to 7202 cal. yrs BP). Pollen samples were taken from the basal sand deposits below the peat, which was probably deposited in the context of a small stream environment prior to extensive paludification (see section 7.7) and expansion of *Alnus glutinosa* communities. A woodland community dominated by *Quercus*, *Tilia*, *Ulmus*, and *Corylus avellana* is suggested at this time, along with *Pinus sylvestris* pollen, possibly from regional sources, which is in close agreement with the situation downstream at Hook Lane.

At Hirst Courtney, *Quercus*, *Tilia*, and *Corylus avellana* are also important dryland taxa within the basal soil which has been dated to 4150±45 BP (4831 to 4525 cal. yrs BP), with *Pinus sylvestris* again consistently represented from perhaps more distant sources. The relatively high frequencies of *Tilia* pollen (*c.*20% TLP-*Alnus*) within the basal samples at Eskamhorn and Hirst Courtney add to the to the well established body of evidence for the dominance of *Tilia* in woodlands during the mid-Holocene (Birks *et al.*, 1975; Godwin, 1975b; Girling & Greig, 1977; Moore, 1977; Greig, 1982a; 1982b; Bradshaw, 1981b; Birks, 1989).

Elsewhere in the Humber basin, Beckett (1975) notes that *Tilia* is an important component of the woodland around Gransmoor in northern Holderness during the mid-Holocene, but its distribution is likely to be locally variable as it is apparently not an important component of the woodland surrounding other mere sites (Beckett, 1981; Flenley, 1987). Clark & Godwin (1956) and Beckett (1975) suggest *Tilia* was important on the glaciofluvial sands and gravels of east Holderness, perhaps reflecting their better drainage (Dinnin, 1995). Greig (1982a; 1982b) has also recorded significant quantities of *Tilia* pollen from coastal peats at Ingoldmells and Butterbump in Lincolnshire. Since the pollen productivity of *Tilia* is low, he estimates that about 90% of the woodland cover in this area would have been comprised of *Tilia*. Finds of the *Tilia* dependant beetle *Ernoporus caucasicus*, associated with relatively high frequencies of *Tilia* pollen in the Holocene deposits of the Trent floodplain at Bole Ings and Misterton in the Humberhead Levels (Dinnin & Brayshay, 1994; Dinnin, 1997), also suggest a predominance of *Tilia* throughout the region. This evidence supports the earlier suggestion for an abundance of *Tilia* in the Humberhead Levels area until at least the late-Bronze Age (Smith, 1958a; Turner, 1962; Smith, 1985a).

Although the record is fragmented, the situation on the chalk uplands is different, with open woodlands indicated from *c*.4000 BP (*c*.4800 to *c*.4200 cal. yrs BP) in the Yorkshire Wolds (Bush & Ellis, 1987) and *Tilia* disappearing by *c*.3400 BP (*c*.3900 to *c*.3400 cal. yrs BP), reflecting the different edaphic conditions and higher levels of human activity (Bush & Flenley, 1987; Flenley, 1990). Moore (1977) and Huntley & Birks (1983) suggest that *Tilia* has a preference for base rich mull soils and it is particularly sensitive to browsing by animals and disturbance (Rackham, 1980). It is also likely that *Tilia* was abundant on the dry floodplain edge in locations like the base of valley slopes, and river levee/terrace features, where its growth may be promoted by seepage (Godwin, 1975a; Brown, 1988; Waller, 1994b). Therefore, it is likely that *Tilia* was regionally important, and prevalent on the drier soils around the floodplain corridor in the lower Aire valley.

Pre-*Ulmus* decline vegetation

The bio- and chronostratigraphic record from Hirst Courtney does not extend back as far as the *Ulmus* decline, so the following account is based solely on the downstream sites at Hook Lane and Eskamhorn. The on-site vegetation at Hook Lane is flooded *c*.7000 BP (*c*.7900 to *c*.7600 cal. yrs BP) and replaced by 'fluviolagoonal' conditions (*sensu* Van der Woude, 1983). The disappearance of the local vegetation under these conditions enables the composition of the surrounding woodland to become clearer. It is apparent that *Tilia*, *Ulmus*, *Quercus*, and *Corylus avellana* are particularly well represented dryland taxa during lpaz HKP-1 and HKP-2 (see section 9.5 and Figure 9.3), at least within the extra-local range of pollen recruitment (*sensu* Jacobson & Bradshaw, 1981).

Evidence in support of this inference is available from the predominance of these taxa in the pollen diagram from Eskhamhorn (see section 8.5 and Figure 8.4) in lpazs EKP-1 and 2a, between *c*.6400 and *c*.5400 BP (*c*.7400 to *c*.6200 cal. yrs BP), although a clear signal is obscured by the presence of dense *Alnus* carr. The distinction between dry and wetland vegetation also a problem. The identification of *Quercus* wood macrofossils in the sediment core (Table 8.3), and the abundance of *Quercus* pollen may suggest that at least some of the source trees were growing either in the drier parts of the carr, or at the floodplain edge ecotone between upland and floodplain. Similarly, finds of *Corylus avellana* wood and nuts suggest that at least some of the source of this pollen type was shrubs within the fen carr community. The relative proportions of these taxa within the upland woodland community, along with *Tilia* and *Ulmus*, is difficult to ascertain and it is likely that most of the pollen contributing to these lpazs at Eskamhorn is derived from the adjacent extra-local community. It is also likely that such communities were spatially variable and therefore, the extrapolations as to the nature of the regional vegetation from this evidence not attempted.

The *Ulmus* decline

The *Ulmus* decline had not been dated in this area of the Humber basin prior to this study, although Beckett (1975) dates this event at Gransmoor Quarry in northern Holderness (see Figure 3.2) to 5099±50 BP (5936 to 5730 cal. yrs BP). Two dates are now available for the mid-Holocene decline in *Ulmus* pollen in the lower Aire valley; a radiocarbon date of 5440±45 BP (6304 to 6216 cal. yrs BP) from Eskamhorn, and a date from linear interpolation of *c*.5800 BP (*c*.6900 to *c*.6400 cal. yrs BP) from Hook Lane. Although the date from Hook Lane must be treated as tentative due to the imprecise nature of interpolated ages and the changing local environmental conditions at that time, this evidence suggests that the *Ulmus* decline was relatively early in the river valleys of the Humber lowlands. Brayshay & Dinnin (1999) provide an approximate age for the *Ulmus* decline in the lower Trent valley (derived from linear interpolation) of *c*.5200 BP (*c*.6300 to *c*.5700 cal. yrs BP) although it is apparent from the pollen diagram that the actual decline horizon could be placed earlier.

The nature of the decline in *Ulmus* pollen at both Hook Lane and Eskamhorn is a gradual fall in pollen percentages. However, the closed canopy *Alnus* carr conditions that existed at this time are likely to filter out much of the extra-local and regional pollen from the canopy component and through the trunk space (*sensu* Tauber, 1965) which may cause a lag in pollen recruitment at the site (see section 9.5). This may spread the *Ulmus* decline in the biostratigraphic record. The filtration effect may also explain why there is only limited evidence for an increase in open ground from ruderal herbaceous pollen types (e.g. Poaceae, *Plantago lanceolata*). The main beneficiaries of the decline in *Ulmus* populations, at both Eskamhorn (EKP-2b) and Hook Lane (HKP-3b), appears to be *Fraxinus excelsior*, *Quercus*, and perhaps *Corylus avellana*.

Taylor (1995) records *Ulmus* pollen percentages of only *c*.1% from a spot sample dated to 5140±43 BP (5891 to 5753 cal. yrs BP - average age of GU-5477 and GU-5478) in southern Holderness and Long *et al.* (1998a) also record low *Ulmus* percentages in the Foulness valley from a peat deposit dated to 5615±45 BP (6484 to 6301 cal. yrs BP), suggesting that *Ulmus* pollen values may have already declined by this time (although this may be a product of the TLP pollen sum used in situations where *Alnus glutinosa* pollen abundant, which swamps subordinate pollen types). These early *Ulmus* decline dates are consistent with many other dates from eastern England (Bennett, 1983b; Simmons & Innes, 1987; 1988; Simmons *et al.*, 1993; Waller, 1994c). Although the *Ulmus* decline is usually considered to be of a relatively synchronous nature across Britain (e.g. Smith & Pilcher, 1973; Huntley & Birks, 1983), a broad spread of elm decline dates, over approximately a millennium, suggests some degree of local diachroneity. However, the dates in this study may be 'smeared' (*cf.* Baillie, 1991) owing to the use of an estimated interpolated age and the poor resolution of radiocarbon dates. The evidence presented in this thesis perhaps suggests the *Ulmus* decline was related to the slow spread of a pathogen, facilitated by the movements of prehistoric people in the area at this time (e.g. Peglar & Birks, 1993).

Post-*Ulmus* decline vegetation

At Hook Lane and Eskamhorn, the surrounding dryland vegetation of *Quercus*, *Tilia*, *Corylus avellana*, and even *Ulmus* is still inferred from the post-*Ulmus* decline lpazs. This time period is of particular interest to archaeologists, as it is usually associated with evidence for major impacts of human activities on the natural vegetation. However, the pollen data from the sites in the lower Aire valley may be masking significant changes in the regional upland vegetation due to the dominance of pollen from extra-local communities. The valley side woodland is also likely to be contain trees which are also growing in the local fen carr and in the regional landscape (e.g. *Corylus avellana*, *Fraxinus excelsior*, *Quercus* etc.). This is particularly problematic, considering these floodplain and river valley areas were probably the last places to be cleared by human activity (Flenley, 1990), which may give a biased picture of continuity of relatively undisturbed regional vegetation. Despite these problems, Turner (1962) and Smith (1985a) have obtained similar results to those in this study, from pollen analysis of basal peat deposits from Thorne Moors. Smith (1985a) dated the opening of his rpaz HHL/A from 4545±75 BP (5452 to 4878 cal. yrs BP), characterised by *Quercus*, *Corylus avellana*, *Tilia*, *Ulmus*, and *Fraxinus excelsior*, which suggests this mixed woodland was prevalent throughout this region during this time.

All of the evidence in the lower Aire valley, that may be taken to indicate human interference with the vegetation, can be explained by natural processes of vegetation dynamics or are associated with changing local conditions within the floodplain wetland. For example, frequencies of *Fraxinus excelsior* pollen are generally higher in the post *Ulmus* decline assemblages at all three sites. This may be interpreted as the expansion of *Fraxinus excelsior* into increasingly open areas, forming secondary woodland (*cf.* Tinsley, 1981). However, this is not the only explanation for the occurrence of *Fraxinus excelsior* in this context.

At Hook Lane, *Fraxinus excelsior* pollen is relatively well represented after *c.*5800 BP (*c.*6900 to *c.*6400 cal. yrs BP) in lpazs HKP-3b and 3c (see section 9.5 and Figure 9.3). During HKP-2b, this may, in part, be related to *Fraxinus excelsior* colonising gaps in the surrounding woodland left by dead *Ulmus* trees, during HKP-3c. However, it may also be related to increased flowering of trees growing locally in the drier areas of fen carr. *Fraxinus excelsior* wood macrofossils are recorded during this zone which suggest local origin, and shade intolerant taxa such as *Ilex aquifolium* are also represented locally. Although it has been suggested (see section 9.5) that this sequence of pollen changes may be related to Neolithic people utilising the fen carr for grazing of livestock, not all of the changes in the pollen spectra are consistent with such an interpretation. It is perhaps more likely that these changes are related to shifting fen communities over the floodplain area with time (*cf.* Waller, 1998), with taxa such as *Fraxinus excelsior* and *Ilex aquifolium* closer to the coring site at this time or due to the expansion of these taxa associated with a local clearing of *Alnus* carr due to processes such as windthrow (*cf.* Brown, 1997b).

Fraxinus excelsior pollen is also well represented at Eskamhorn during lpaz EKP-3 in association with wood macrofossils, where they are explained as a response to local reduction in *Alnus* carr (section 8.5). At Hirst Courtney, the spiky nature of the *Fraxinus excelsior* pollen curve, and coincidence of wood macrofossils, also suggests the local growth of these trees in the drier areas of fen carr (section 7.5). Such an interpretation is supported by the ecological status of *Fraxinus excelsior* (*cf.* Wheeler, 1980b; Wardle, 1981; Rodwell, 1991), which is capable of withstanding waterlogging (Iremonger & Kelly, 1988) and can be found growing within moist peat forming habitats (Godwin, 1978a; 1978; Wardle, 1981; Van der Wiel, 1982; Rodwell, 1991; Wiegers, 1992; Waller, 1994c).

Small openings in the *Alnus* fen carr can also be invoked to explain the increased occurrence of *Pinus sylvestris*, although such an interpretation is made with caution owing to the dual input of *Pinus sylvestris* pollen from the regional wind blown canopy component, and a possible waterborne, tidal input. At Eskamhorn, during EKP-3, after *c.*5000 BP (*c.*6000 to 5500 cal. yrs BP), increased regional input of *Pinus sylvestris* is suggested subsequent to a slight opening up of the fen carr locally (section 8.5). Later, above the transgressive contact dated to 3440±45 BP (3828 to 3572 cal. yrs BP), *Pinus sylvestris* pollen is particularly abundant, probably from a combination of both airborne and waterborne sources. Similarly, at Hook Lane, as the fen carr opens up during the opening of lpaz HKP-4 from 4850±45 BP (5655 to 5478 cal. yrs BP), there is an increase in *Pinus sylvestris* pollen abundance, although again, it is uncertain what proportion of this pollen is derived from aerial sources or from estuarine transport, due to the increase in water table and possible greater frequency of tidal flooding at this time (section 9.5). Nevertheless, this evidence points to the presence of significant stands of *Pinus sylvestris* growing in the Humberhead Levels area at this time.

Relatively high percentages of *Pinus sylvestris* pollen are also recorded later at Hirst Courtney, during lpaz HCP-1 between *c.*4200 and *c.*3900 BP (*c.*4800 to *c.*4300 cal. yrs BP). This suggests that *Pinus sylvestris* was regionally important up until at least the end of this lpaz, when it is suggested the closure of *Alnus* fen carr caused the input of *Pinus sylvestris* from the wind blown canopy component to be effectively filtered out at this site.

Smith (1958a) records high values for *Pinus sylvestris* in the base of his pollen diagram from Hatfield Moors, which is confirmed by later work (Smith, 1985a), when dates from the base of this sequence of *c*.4200 BP (*c*.5000 to *c*.4400 cal. yrs BP) were obtained. Smith (1985a) records *Pinus sylvestris* pollen percentages of *c*.10% (but up to 30% TLP) in basal samples at Thorne Moors (rpaz HHL/A), which suggests some of this pollen is from *Pinus sylvestris* growing locally. Wood macrofossils of *Pinus sylvestris* in peat marginal to Thorne Moors suggest that *Pinus sylvestris* was established on the higher ground surrounding the mire prior to waterlogging (Smith, 1985a). Dinnin (1997a) suggests that *Pinus sylvestris* colonised drier areas of the peat surface and also on the sandy soils of the Sherwood Sandstone (e.g. the Lindholme 'island') and extensive blown sands in the region, with mixed deciduous woodland consisting of *Quercus*, *Tilia*, *Corylus avellana*, *Fraxinus excelsior*, and *Ulmus* growing on the clayey soils derived from the Mercia Mudstones and fine grained '25-Foot Drift' glaciolacustrine and glaciofluvial sediments. Long *et al.* (1998a) also record relatively high frequencies of *Pinus sylvestris* (*c*.15% TLP) at Sandholme Lodge in the Foulness Valley after 4170±45 BP (4835 to 4534 cal. yrs BP). Therefore, there is considerable evidence for the presence of stands of *Pinus sylvestris* between at least *c*.5000 and *c*.3400 BP (5900 and 3600 cal. yrs BP) in the Humberhead and Vale of York region. This is in contrast to the regional pollen rain from Holderness during this time (*Alnus*/*Quercus* rpaz) where taxa such as *Quercus* and *Pinus sylvestris* are relatively rare in sites such as Roos Bog (lpaz, RB7, Beckett, 1981).

The decline in *Pinus sylvestris* (interpreted as filtration due to closure of the fen carr in section 7.5) at the close of lpaz HCP-1 at Hirst Courtney, *c*.3900 BP (*c*.4600 to *c*.4000 cal. yrs BP), is of particular interest due to the simultaneous decline in *Tilia* pollen values (see Figure, 7.3). Smith (1985a) identified a decline in *Tilia* and *Pinus sylvestris* pollen percentages in his work on Thorne and Hatfield Moors, which he used to mark the opening of his rpaz HHL/B from *c*.3600 BP (*c*.4200 to *c*.3600 cal. yrs BP). *Quercus* and *Ulmus* pollen frequencies also decline at some sites, and 'clearance herbs' rise, which lead Smith (1985a) to suggest the impact of human activity as a cause, temporarily creating clearings in the surrounding landscape. The *Pinus sylvestris* decline is dated to 3715±70 BP (4262 to 3847 cal. yrs BP) on Thorne Moors, and 3685±65 BP (4225 to 3835 cal. yrs BP) on Hatfield. Charcoal layers have been recorded at the HHL/A-B boundary, dated to 3715±70 BP (4262 to 3847 cal. yrs BP) at Thorne and 3570±70 BP (4080 to 3647 cal. yrs BP) at Hatfield, and at Crowle Moors, a burnt and chopped *Pinus sylvestris* trunk is dated to 3545±70 BP (4048 to 3634 cal. yrs BP). Smith (1985a) used this as evidence for deliberate tree clearance by Bronze Age people, perhaps using *Pinus sylvestris* stands growing on the Lindholme outcrop, a raised island in the middle of Hatfield Moor.

Dinnin & Brayshay (1994) and Brayshay & Dinnin (1999) correlate a similar decline in *Tilia* and *Pinus sylvestris* from floodplain deposits at Bole Ings in the lower Trent valley with this proposed phase of Bronze Age anthropogenic activity in the Humberhead Levels (Smith, 1985a) and an undated sequence at Butterbump in the Lincolnshire Marsh (Greig, 1982b). However, there are problems associated with the evidence for human activity during this time at these sites. For example, the sites on the periphery of Thorne Moors record apparently clearer phases of disturbance (e.g. the 'Trackway' site), whereas at the central sites such clearance episodes are not obvious (Turner, 1965; Smith, 1985). Whilst this may be, in part, reflecting the limited dispersal range of the pollen taxa recorded (e.g. Tauber, 1965; Jacobson & Bradshaw, 1981), the evidence for human activity is equivocal. At the Trackway site, Buckland & Kenward (1973) and Buckland (1979) record 14% cereal pollen and 30% Poaceae, which they use as evidence to suggest agricultural activity associated with the trackway, which is dated to 2980±110 BP (3393 to 2851 cal. yrs BP). However, Smith (in Buckland, 1979) casts doubt on this interpretation and has identified *Glyceria* from this site which would account for the cereal pollen, suggesting that the source of the Poaceae pollen is likely to be from *Phragmites* reeds growing around the mire Smith (1985a).

At Bole Ings (Dinnin & Brayshay, 1994; Brayshay & Dinnin, 1999), there are also problems associated with the nature of the decline in *Tilia* and *Pinus sylvestris* which is dated to 3570±70 BP (4080 to 3647 cal. yrs BP). *Alnus glutinosa* pollen frequencies expand to *c*.80% TLP concomitant with this decline. The likely effects of this are to swamp the pollen represented by other taxa, due to the method of percentage calculation (*cf.* Janssen, 1959), and the *in situ Alnus* carr is also likely to filter out much of the pollen recruited from extra-local sources. There is no clear evidence in support of an anthropogenic cause for this decline from the beetle evidence in this work, which suggests natural floodplain processes may be invoked as a reasonable alternative explanation.

Whilst the declines in *Tilia* and *Pinus sylvestris* at Hirst Courtney which are similar to other pollen diagrams in the region may relate to the effects of deliberate human activity, this interpretation cannot be made with any degree of certainty in the lower Aire valley. The decline in *Tilia* at the boundary between lpaz HCP-1/2a has been interpreted as due to changing local conditions and expanding wetland area (*cf.* Waller, 1994b). This is related to the suggested abundance of *Tilia* on the floodplain edge ecotone (Brown, 1988), and the encroachment of peat onto such river terrace and levee features prominent on river floodplains, as water levels continue to rise promoting vertical floodplain development.

Turner (1962) identifies a 'primary' *Tilia* decline, which was dated to 3179±115 BP (3680 to 3078 cal. yrs BP) and 2942±115 BP (3375 to 2782 cal. yrs BP) on Thorne Moors and attributed to the effects of Bronze Age clearance activity. However, this was not such a pronounced feature of the diagrams constructed by Smith (1985a) in his analysis of Thorne and Hatfield Moors. This suggests that this was not a regional phenomena, and if it was related to human activities at all, they were of a localised nature. It is also possible that the *Tilia* and *Pinus sylvestris* decline used by Smith (1985a) to define the opening of rpaz HHL/B are related to expanding mire conditions at this time, which may explain the weakly expressed nature of the *Tilia* decline. This is supported by the discovery of *Pinus sylvestris* stumps in the base of the peat.

The first unequivocal evidence for human impact on the regional vegetation in the Humberhead Levels area comes from the 'secondary' *Tilia* decline identified by Turner (1962), which is supported with abundant evidence by Smith (1985a). Six dates are available for this event from Thorne and Hatfield Moors, the average of which is *c.*2250 BP (*c.*2500 to *c.*2000 cal. yrs BP). It is perhaps more likely that this event may be registered regionally, although despite the asynchronous nature of the *Tilia* decline, it probably occurred too late to be registered at any of the sites presented in this research. The lack of evidence for human activity in the lower Aire valley is in agreement with Brown (1982) and Flenley (1990) who argue that wet river valleys were the last areas to be cleared by human activity. Although it is possible that upland clearance may have been sufficient to change patterns of sediment transport and river/estuarine behaviour (*cf.* Brayshay & Dinnin, 1999), there is no evidence from the pollen record to support this (see section 10.5).

The occurrence of *Taxus baccata* pollen in the peats in the lower Aire valley is interesting, considering the scarcity of finds of this pollen taxa in previous studies. This may in part be related to differential preservation and/or recognition, as *Taxus baccata* is not believed to be an underrepresented taxon (Bradshaw, 1981a). *Taxus baccata* is recorded throughout the profile at Hook Lane and Hirst Courtney (see section 7.5 and 9.5), and is particularly well represented at Hook Lane in lpaz HKP-5c after *c*.4000 BP (*c*.4800 to *c*.4200 cal. yrs BP), and at Hirst Courtney lpaz HCP-3, 4 and 5 after *c*.2700 BP (*c*.3000 to *c*.2500 cal. yrs BP), where it reached values up to (8% TLP-*Alnus*). *Taxus baccata* is usually associated with chalk or limestone woodland (Rodwell, 1991). However, wood macrofossils have been found within fen peats in the Fenland (Godwin, *et al.*, 1935; Godwin & Clifford, 1938) and the occurrence of *Taxus baccata* at a large number of Fenland sites suggests it must have formed part of the fen woodland (Waller, 1994c). Therefore, it is possible that the *Taxus baccata* pollen recorded in this study represents trees growing in the drier areas of fen and along the floodplain edge. The better representation of *Taxus baccata* towards the top of the sequences at Hook Lane and Hirst Courtney is likely to be related to increased representation of pollen from extra-local communities following a reduction in local fen carr due to rising water tables.

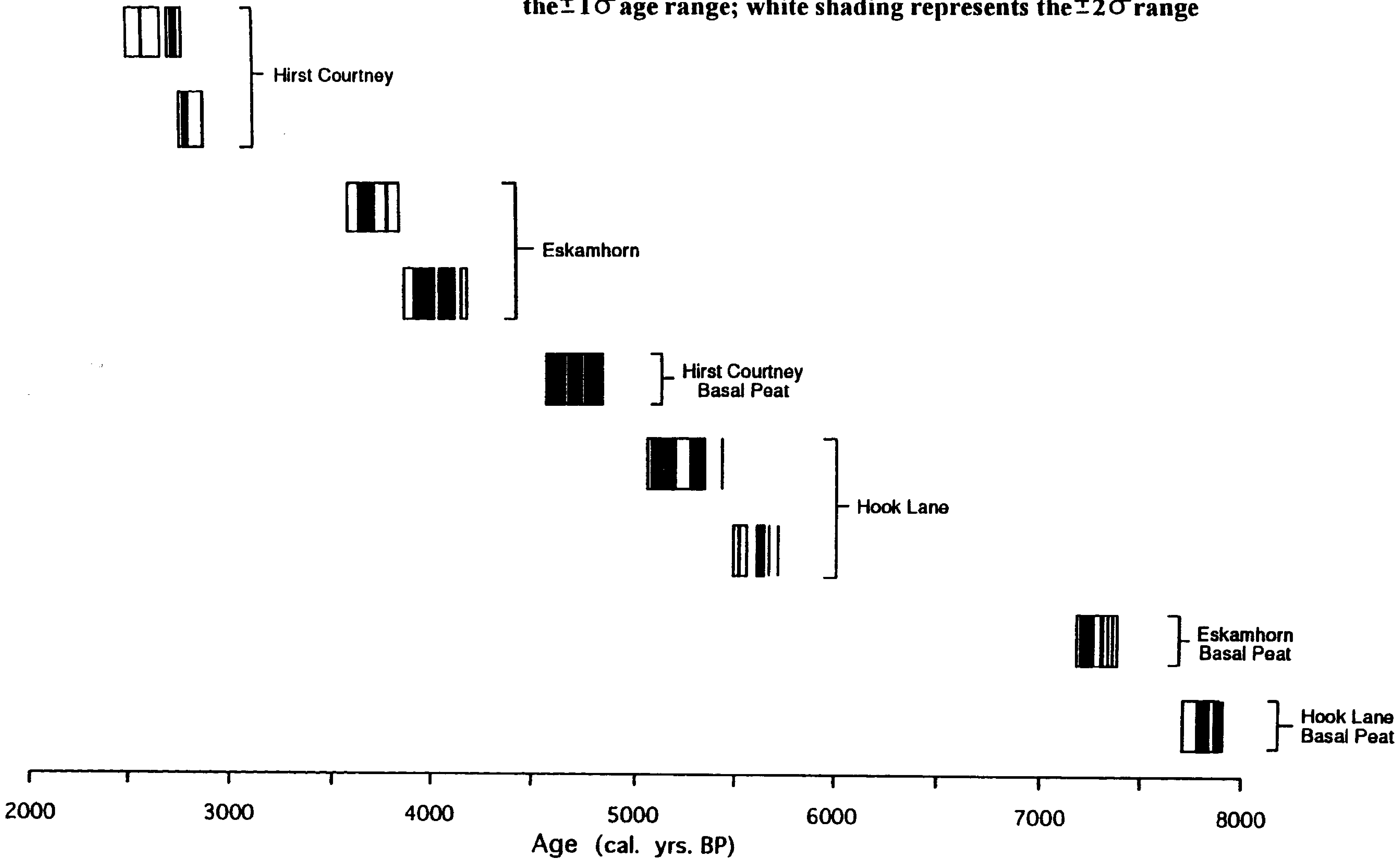
10.3 Tendencies of Sea-Level Movement

Figure 10.2 shows a plot of the tendencies of sea-level movement from the lower Aire valley. These results are combined with the sea-level data obtained from other studies (Smith *et al.*, 1981; Switsur, 1981; Crowther, 1987; Millett & McGrail, 1987; Dinnin & Lillie, 1995; Long *et al.*, 1998a; Neumann, 1998), which includes archaeological information, and shown as a diagram of tendencies of sea-level from the Humber estuary using the mid cal. ages of the dates (Figure 10.3).

The data collected from the lower Aire valley (Figure 10.2) exclusively record positive tendencies between *c*.7900 and *c*.2500 cal. yrs BP, although evidence for a reduction of estuarine conditions is sporadically recorded in other parts of the estuary during this time (Figure 10.3). The registration of negative tendencies of sea level does not necessarily imply an actual altitudinal fall in RSL (Shennan, 1982; 1986b; Shennan *et al.*, 1983), but it is possible that these negative tendencies perhaps reflect at least a reduction in the rate of sea-level rise. It is likely that once *Alnus* fen carr conditions become established within the river valley, the robustness and inertia of such vegetation systems means that they are insensitive to small falls in the rate of sea-level rise (see section 10.2.1). The continued input of freshwater from inland sources may therefore be sufficient to maintain carr communities during periods of negative sea-level tendency. This also suggests that the rate of sediment accumulation in such vegetation communities is not a good proxy indicator for the rate of RSL rise, but only indicates general underlying trends. Nevertheless, once water levels begin a sustained rise above the rate of organic sedimentation in the fen carr, the vegetation records a definite response to the wetter conditions, and provides good proxy data for rising water levels and phases of estuarine expansion.

Basal dates at Hook Lane and Eskamhorn relate to rising water tables, associated with rising tide levels seawards, between *c*.7900 and *c*.7400 cal. yrs BP (Figure 10.2). Freshwater conditions are replaced by estuarine conditions at Hook Lane between *c*.5700 and *c*.5000 cal yrs BP. Positive tendencies are also recorded at Kilnsea Warren on the coast of the outer estuary between *c*.5200 and *c*.5000 cal. yrs BP (Long *et al.*, 1998a) and paludification associated with rising water levels is recorded throughout the Ancholme valley at this time (Neumann, 1998). It is evident that positive tendencies of sea-level are widespread in the Humber estuary during this period (Figure 10.3).

Figure 10.2 Calibrated radiocarbon ages of positive sea-level tendencies in the lower Aire Valley during the Holocene. Dates are calibrated using the intercepts method of Stuiver & Reimer (1993). Black shading represents the ± 1σ age range; white shading represents the ± 2σ range



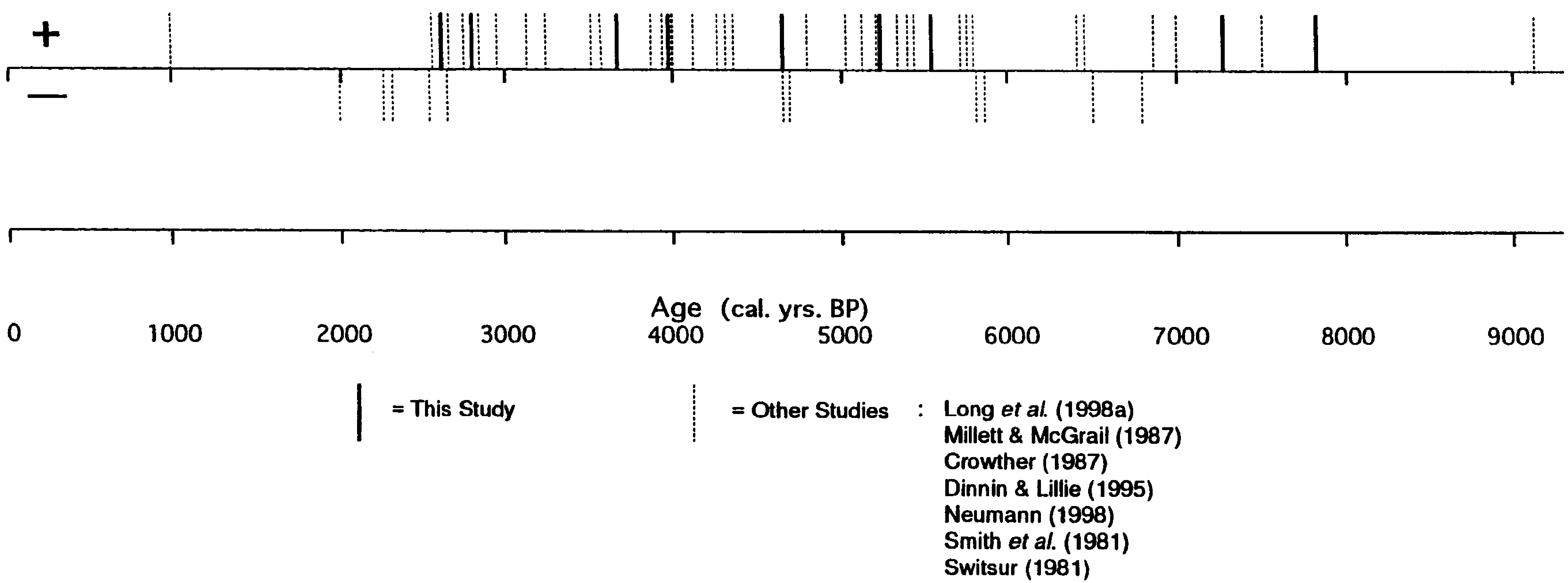


Figure 10.3 **Chronology of positive and negative sea-level tendencies in the Humber Estuary during the Holocene. The mid cal. ages (*sensu* Bennett, 1994) of the calibrated age range (calibrated using Stuiver & Reimer, 1993) are plotted using the method explained in section 5.10**

Peat development is initiated by rising water levels at Hirst Courtney between *c*.4800 and *c*.4500 cal. yrs BP. However, evidence from Sandholme Lodge in the Foulness Valley (Long *et al.*, 1998a) and Halsham Carrs in Holderness (Dinnin & Lillie, 1995), within the inner and outer estuary respectively, suggests negative tendencies and falling water levels during this time (Figure 10.3). This suggests that paludification at Hirst Courtney, and continued basal peat formation in the Ancholme valley at this time (Neumann, 1998), reflects a generally rising trend in sea level (although an increase in freshwater flooding can also be invoked to explain this), with negative sea-level tendencies being registered at some sheltered sites. This contradiction may be attributable to local effects such as sediment supply, site morphology and varying terrestrial and fluvial inputs.

Positive sea-level tendencies are indicated at Eskamhorn between *c*.4100 and *c*.3600 cal. yrs BP (Figure 10.2). This coincides with evidence for rising sea-level in the middle estuary at East Clough, Melton, between *c*.4300 and *c*.3800 cal. yrs BP (Long *et al.*, 1998a) and continued basal peat formation in the Ancholme valley (Neumann, 1998).

Positive sea-level tendencies are again recorded at Hirst Courtney between *c*.2900 and *c*.2500 cal. yrs BP. Although increasing marine influence is recorded between *c*.3100 and *c*.2800 at Newton Marsh, Tetney, in the outer estuary (Long *et al.*, 1998a), and some areas of the Ancholme valley (Neumann, 1998), all other evidence from the middle and outer estuary (Brigg, Thirtle Bridge and Barrow Haven) points to widespread removal of estuarine conditions from *c*.2900 to *c*.1900 cal. yrs BP (Gaunt & Tooley, 1974; Dinnin & Lillie, 1995; Long *et al.*, 1998a). This overlap in dates may be the result of the radiocarbon plateau 'smearing' ages during this time (*cf.* Pilcher, 1991; Van Geel *et al.*, 1996), or may be due to the location of Hirst Courtney at the top of the system, where these effects may be experienced later. The onset of a retraction of estuarine conditions is inferred at the top of the sequences at both Eskamhorn and Hook Lane, although these episodes are undated (see section 8.6 and 9.6).

Although the spatial and temporal dataset used in these comparisons is limited, the data collected in this study supports the main findings of Long *et al.* (1998a). Ponding back of freshwater initiated waterlogging and peat development beyond direct marine influence in the Aire valley at Hook Lane and Eskamhorn between *c*.7900 and *c*.7400 cal. yrs BP. The continuation of this trend until between *c*.2900 and *c*.2500 cal. yrs BP is inferred by the expansion of estuarine conditions which caused the inundation of freshwater communities at

Hook Lane and Eskamhorn between *c*.5700 and *c*.3600 cal. yrs BP. This is also indicated by the waterlogging and subsequent inundation of saltmarsh and fen communities at Kilnsea Warren, East Clough, Newton Marsh (Long *et al.*, 1998a), and elsewhere in the Humber (Gaunt & Tooley, 1974; Smith *et al.*, 1981; Jordan, 1987). Freshwater peat accumulation occurred, beyond marine influence, at Hirst Courtney in the Aire valley, and in the Foulness and Ancholme valleys during this time (Long *et al.*, 1998a; Neumann, 1998). Local variations in sediment supply and site exposure (and perhaps small fluctuations in RSL) are responsible for local episodes of temporary peat accumulation indicating negative tendencies at this time in protected settings such as the Roos valley in south Holderness (Dinnin & Lillie, 1995; Long *et al.*, 1998a).

Widespread changes in sea-level tendency have been identified between *c*.3300 and 1900 cal. yrs BP in the Humber estuary (Long *et al.*, 1998a). The lack of sea level index points from this time means that much of the information is derived from archaeological evidence, although new data from Hirst Courtney adds to the limited stratigraphic information. The synthesis of data in this study suggests that positive sea-level tendencies are recorded throughout the estuary from *c*.4500 cal. yrs BP until some time between *c*.2900 and *c*.2500 cal. yrs BP. A date of 2597±117 BP (2983 to 2349 cal. yrs BP) has been obtained from organic sediment at the base of estuarine clay within which the Brigg raft is entombed, along with a range of other artefacts with broadly similar ages (Smith *et al.*, 1981; Switsur, 1981). This has been used as evidence to suggest that the maximum extent of estuarine conditions within the Ancholme valley occurred at *c*.2700 cal. yrs BP, 22 km inland at Waddingham Holmes (Fletcher, 1981). Dates on thin basal peats at South Ferriby also suggest inundation of peat forming communities some time after *c*.2700 cal yrs BP in the Ancholme valley (Neumann, 1998). Other evidence, such as dated wooden artefacts buried within marine clays on the foreshore at Melton and North Ferriby, provide minimum ages for this positive tendency of between *c*.3400 and *c*.2900 cal. yrs BP (Crowther, 1987; Dinnin & Lillie, 1995). The freshwater transgressive contact at Hirst Courtney and inundation of Newton Marsh between *c*.3100 and *c*.2500 cal. yrs BP represent the culmination of this expansion in estuarine conditions and the maximum extent of marine conditions in the Humber estuary (Long *et al.*, 1998a). A slightly younger date of 2530±70 BP (*c*.2800 to *c*.2400 cal. yrs BP) for the end of peat formation comes from the Foulness valley (Millett & McGrail, 1987). This coincides with estuarine conditions and peat formation indicating waterlogging in the Hull valley (Van de Noort & Davies, 1993).

This expansion of estuarine conditions in the Humber estuary was first proposed by Smith (1958a; 1958b), and is supported by subsequent research (Gaunt & Tooley, 1974; Tooley, 1978a; 1982; 1990; Fletcher, 1981 and Long *et al.*, 1998a). From *c.*2900 cal. yrs BP, negative sea-level tendencies are widespread in the Humber (see Figure 10.3), based on dated regressive sequences at Barrow Haven and in the Roos valley in south Holderness (Dinnin & Lillie, 1995; Long *et al.*, 1998a). No dated evidence has been obtained during this study to add to this, although diatom evidence from all three sites reported here suggests a retraction of riverine or estuarine influence towards the top of the sequences. This evidence is consistent with similar trends observed in alluvial sediments containing the Hasholme logboat and the Brigg raft (Jordan, 1987; Smith *et al.*, 1981), which register fresher conditions. Lithostratigraphic evidence which may support this switch to a period of shoreline advance is perhaps provided by the presence of an upper peat within the Ancholme valley (Smith, 1958a) which is now unfortunately largely non-existent due to oxidation (Neumann, 1998). Although this unit remains undated, remnants of an old peat surface in a stratigraphically similar position have been recorded in the lower Aire valley which is also in a poor state of preservation (see section 6.3). This may tentatively support a regressive event following the well documented phase of estuarine expansion (e.g. Long *et al.*, 1998a).

This study adds further support to this model of coastal change during the late Bronze Age/Early Iron Age transition. Long *et al.* (1998a) draw attention to the similarity in timing of this change in coastal development with the climatic deterioration and RSL fall reported from The Netherlands and elsewhere during this time (Van Geel *et al.*, 1996). This climate change has been attributed to an abrupt increase in atmospheric ^{14}C during the Subboreal/Subatlantic transition in northern Europe which Van Geel *et al.* (1996) suggest caused a switch to a more oceanic climate, steric contraction of the ocean, and/or a decrease in the velocity of the Gulf Stream, an increase in terrestrial water storage in ice, soils, and bogs, and ultimately a fall in RSL.

Although this climate change is broadly contemporaneous with the observed coastal changes in the Humber, there are problems associated with the dating of this shift in climate due to the radiocarbon plateau at this time (the radiocarbon interval between 2750 and 2450 radiocarbon years BP lasts only 45 calendar years, Van Geel *et al.*, 1996). Long *et al.*, (1998a) note that the registration of such changes in the stratigraphic record will be considerably offset by lags in the sedimentary response to such perturbations within the coastal system. In addition, the effects of such a northwestern Hemispheric climate change may be expected to be registered by similar coastal changes, at least on a regional scale. However, this is not the case, as there is little agreement between the coastal changes recorded in the Humber between *c*.3300 and 1900 cal. yrs BP and those occurring within the Fenland during this time. All existing evidence from the East Anglian fens between *c*.3000 and 2750 cal. yrs BP actually points to progressively freshwater conditions, and between c.2600 and 1700 cal. yrs BP, the stratigraphic and archaeological data suggest an asynchronous phase of marine inundation and positive tendency (Waller, 1994d). This is opposite to the situation in the Humber during this time. Long *et al.* (1998a) suggest that, although there are differences in crustal movements between the Humber and Fenland, with the latter experiencing higher rates of crustal subsidence (Shennan, 1989), the magnitude of these differences is too small (<1 mm yr^{-1}) to be significant. This suggests that the interplay of local depositional processes are important in controlling the sedimentary regimes in at least one of these regions (Long *et al.*, 1998a).

Interestingly, Smith (1985a) records a recurrence horizon (R.Y.HHL/IV) indicating increasing surface bog wetness at Thorne and Hatfield Moors, dated to 2525±60, 2470±65, and 2345±65 BP (between 2756 and 2156 cal. yrs BP), which he attributes to the phase estuarine expansion rather than a climatic deterioration. This is followed by a period of surface dryness on the bog which may be related to the subsequent reduction in estuarine extent. The relationship between mire conditions and regional groundwater levels must be regarded as tenuous, and therefore a direct relationship between the climate changes as identified by Van Geel *et al.* (1996) and coastal change is not substantiated further. However, this recurrence surface may not be related to sea-level change (further dating of this sequence of coastal change indicates that R.Y.HHL/IV occurs as marine conditions are waning), but actually to the climatic deterioration identified by Van Geel *et al.* (1996), which may provide further independent evidence for a period of climate change at this time.

A lack of stratigraphic data covering the period from *c*.2000 cal. yrs to present means that this time period remains poorly covered by sea-level data. A single date from a transgressive contact at Barrow Haven (Long *et al.*, (1998a) of 1080±40 (1063 to 927 cal. yrs BP) is the only information for this time, so RSL trends during this period are uncertain.

10.4 Age-Altitude Data

10.4.1 Introduction

In this section, water table movements inferred from biostratigraphic evidence in the lower Aire valley are plotted on a graph of RSL (see Figure 10.6), and compared with the SLIPs published by Long *et al.* (1998a). The index points used by Long *et al.* (1998a) all have a clearly defined indicative meaning and reference water level (see Appendix C) and are corrected to MHWST using the procedures outlined in Shennan (1982) and refined in Horton (1997). Data from other studies are included (e.g. Smith *et al.*, 1981; Millett & McGrail, 1987; Dinnin & Lillie, 1995; Neumann, 1998) only where the indicative meaning can be validated (see Appendix C). The information in Gaunt & Tooley (1974) is not used in this analysis because of uncertainty regarding the quality of this data, which is collected from a range of palaeoenvironments, including woody and monocotyledonous peats, intertidal molluscs, and detrital wood fragments (*cf.* Dinnin & Lillie, 1995; Long *et al.*, 1998a).

To allow comparison between different areas and different types of dated material, samples must be related to a reference water level (Shennan, 1982; 1986b; Horton, 1997). The data in Long *et al.* (1998a) are based on the rise of MHWST in the Humber, but graphs constructed on this basis cannot be used to compare samples related to different tide levels. To allow for tidal variability, it is necessary to reduce all the SLIPs to MSL, by subtracting the appropriate value for the reference water level (e.g. MHWST measured at the nearest tide gauge) from the observed values. Therefore, in this study, relative sea-level graphs (relative to MHWST) are produced in which all data are related to this common datum (Figures 10.4, 10.5, 10.5 and 10.7). This is based on present day tide gauge data from Goole and Grimsby, and assumes no change in tidal regime over the Holocene. The altitude of MHWST is 4.30 m at Goole and 3.10 m OD at Grimsby, and MTL varies from 1.2 m at Goole to 0.15 m OD at Grimsby (Admiralty Tide Tables, 1998). Considering the likely changes in coastal morphology and estuary bathymetry as the estuary infills during the Holocene, this assumption is a large one, but a constant palaeotidal range has to be accepted at this stage of the research, due to the lack of published information to suggest otherwise. Therefore, all the altitudinal sea-level data is reduced to RSL using the methods described below.

Many of the data points obtained from this study can only be used as 'limiting dates'. For instance, at all three sites, a combination of litho- and biostratigraphic evidence demonstrates rising water levels initiated paludification and basal peat formation in the Aire valley. However, these basal samples contain no indicators of marine influence, and appear to have formed in a freshwater environment, above any contemporary sea-level datum. They therefore differ from traditionally used 'basal dates' in which saltmarsh conditions are indicated, which therefore can be directly related to former sea-level altitudes (see Long *et al.*, 1998a; Horton *et al.*, 1999c for examples).

A precise relationship with former sea level cannot be calculated from freshwater basal samples such as these, due to the lack of information regarding the relationship between such peat forming communities and tide-levels (Tooley, 1978a). However, they are still of considerable use, as they provide limiting age and altitude values which constrain the possible past position of sea level. Horton *et al.* (1999c; 1999d) demonstrate the use of such limiting data in their analysis of relative sea-level change in Northumberland.

The limiting basis dates in this study are useful because they provide maximum altitudes, below which sea level must have been at that date. The indicative meaning (defined in terms of a reference water level and indicative range, see Chapter 2) for a limiting date on a basal peat sample is calculated as follows (after Shennan, 1982; 1986b; Horton, 1997).

1) Reference Water Level

- $\frac{\text{MHWST} + \text{MTL}}{2}$

2) Indicative Range

- MHWST - MTL

The index points are calculated as:

- ALTITUDE - REFERENCE WATER LEVEL

The calculation of the vertical error terms (e) is based on the square root of the sum of the squares of the following parameters (x): $e = \sqrt{\sum x^2}$

- Indicative range
- Levelling error
- Sample thickness
- Tide table accuracy

The error terms associated with a limiting dates are large because of the uncertainty regarding the indicative range of the sample, which in this study is over 3 m. Such a wide vertical range is to be expected because variations in local palaeogeography, terrestrial and fluvial input, and distance to active tidal channels have a large influence on the level of the groundwater (Shennan, 1986b; 1994). Shennan *et al.* (in press) obtain similar data points from peat/clay contacts, which cannot be directly related to a past tidal level because of erosion, but can be used as limiting values, because at the time of deposition, freshwater peat was forming, and tidal influence must have operated at or below that altitude. Therefore the indicative range for these dates is the same as for freshwater basal peats, and is from above MHWST, where freshwater peat formation is governed by tidal influence, to just below MTL, where groundwater is the controlling factor (Godwin, 1940a; Shennan, 1982; 1986b).

The freshwater clay overlying the peat at Hirst Courtney cannot be regarded as a trangressive overlap because the sedimentary facies does not record a positive shift in marine tendency (Shennan *et al.*, 1983). However, if the data from this site is to be directly compared with RSL, it is necessary to assign a datum to this sedimentary contact to enable it to be compared with a specific point in the tidal range. Heyworth & Kidson (1982) suggest that the lower limit of freshwater vegetation lies between MHWST and HAT. *Alnus glutinosa* can be found towards the lower limit of this range (Ranwell, 1972; 1974; Heyworth, 1978; 1986; Walker *et al.*, 1998). Shennan (1982) suggests that monocotyledonous peat overlying fen wood deposits represents a reference water level of MHWST-10 cm. Hewlett & Birnie (1996) use this data to suggest that similar freshwater transgressive sedimentary contacts in the Severn estuary equate to a little more than MHWST-10 cm. However, because estuarine conditions do not occur above this transition at Hirst Courtney, a reference tide level cannot be assigned to these samples with confidence, so the same rules must apply to these SLIPs as to the basal limiting dates (see above).

Transgressive episodes are traced from top of the floodplain fen wood peats into the overlying freshwater fen meadow deposits (*cf.* Long, 1991; Long & Shennan, 1994). According to Shennan (1982; 1986b), the calculation of the indicative meaning in these circumstances is the same as the example above, except that the reference water level is close to MHWST and the indicative range is ±20 cm. Transgressive contacts are calculated using a reference water level of MHWST-20 cm and an indicative range of ±20 cm, and are a more accurate indicator for palaeo- sea-level reconstructions.

10.4.2 Analysis of Basal Peat Formation

Figure 10.4 is based on the data in Appendix C, and reflects the upward trend in sea-level shown by progressive paludification of landsurfaces ahead of rising sea-level, as well as some dates from basal saltmarsh deposits (Long *et al.*, 1998a). A limiting date from the base of freshwater fen wood deposits at Union Dock, Grimsby, indicates that estuarine conditions were occurring below -11.67 m between 9250 and 8981 cal. yrs BP (Long *et al.*, 1998a). The trangressive top of this basal peat provides the first reliable age estimate of marine conditions in the vicinity, between 7544 and 7392 cal. yrs BP, although due to compaction, the altitude of the sample is probably too low, and an extended upper vertical range has been added accordingly (Long *et al.*, 1998a). The limiting basal dates from paludification of the valley floor of the Aire valley at Hook Lane and Eskamhorn between 7878 and 7202 cal. yrs BP plot above this SLIP, as expected. Basal limiting dates from the Ancholme valley at Brigg (Neumann, 1998) and the Foulness valley at Hasholme (Millett & McGrail, 1987) show the continuation of this upwards trend.

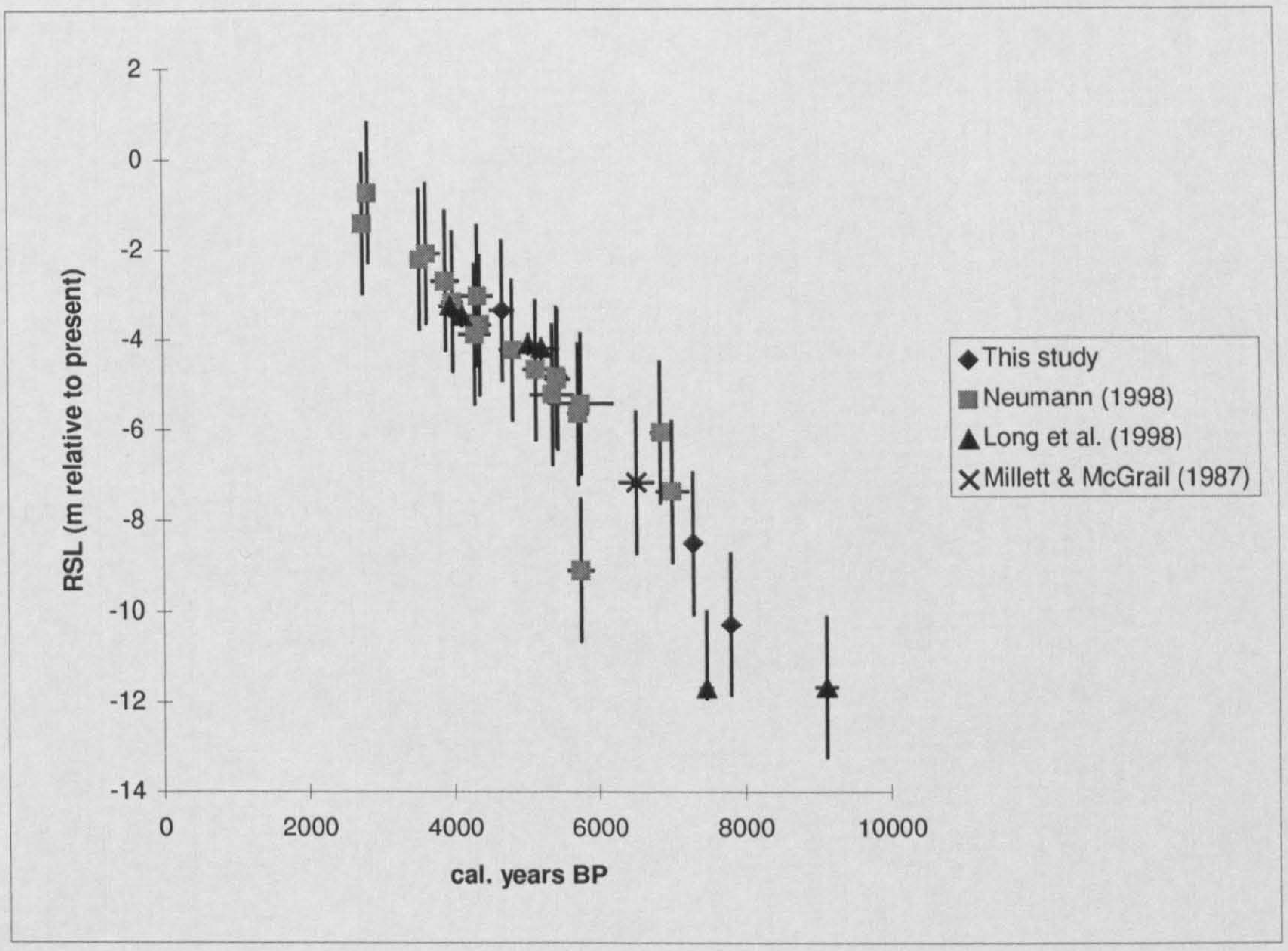


Figure 10.4 Age-altitude graph of RSL showing the limiting and transgressive basal peat dates from the Humber Estuary (see Appendix C)

A date from the base of a peat from Brigg (Neumann, 1998) plots as a distinct outlier on the time-altitude graph at *c*.5800 cal. yrs BP. This data point has been removed from subsequent data analysis due to possible problems with the reliability of the age of this sample within its stratigraphic position (see section 4.4). The remaining limiting data from the Ancholme valley at Redbourne, Brigg and South Ferriby plot very closely with the basis peats from Long *et al.* (1998a) where estuarine conditions are directly demonstrated. This suggests that freshwater basal peat formation in the Ancholme valley was occurring very close to the contemporary level of MHWST in the Humber estuary at that time. The basal limiting date from Hirst Courtney plots towards the top of the trend of RSL rise between 4831 and 4525 cal. yrs BP, which probably reflects the location of this site well beyond direct marine influence in upper reaches of the lower Aire valley.

10.4.3 Analysis of Intercalated Dates from the Humber Estuary

Figure 10.5 shows the trend of RSL derived from SLIPs from the contacts between interbedded clay and peat sequences (see Appendix C for raw data).

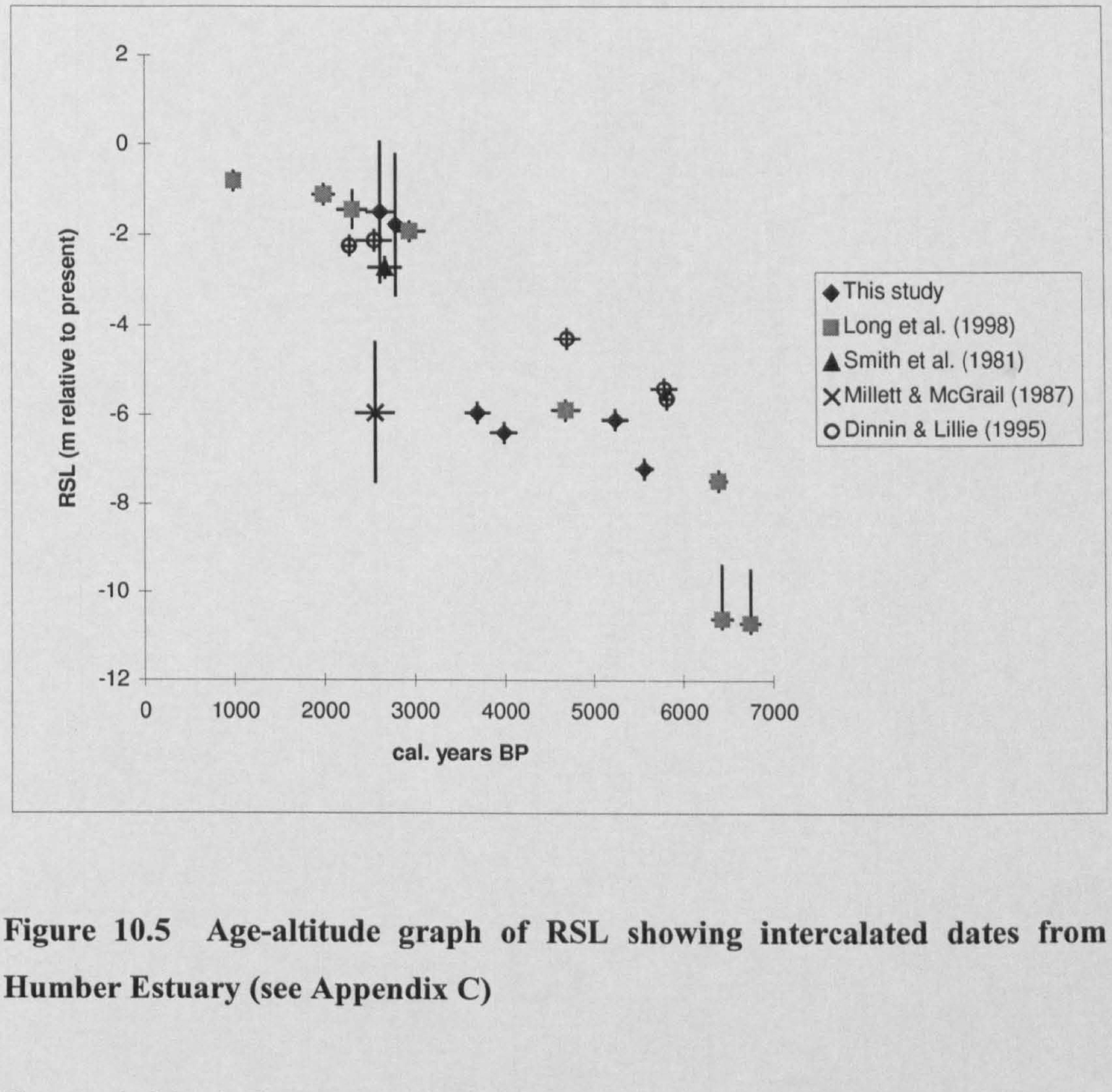


Figure 10.5 Age-altitude graph of RSL showing intercalated dates from the Humber Estuary (see Appendix C)

The spread of data in Figure 10.5 is much wider than the trend shown in Figure 10.4 (based on the basal dates). The date from the peat/clay contact at Hasholme in the Foulness valley (Millett & McGrail, 1987) is a distinct outlier in this plot and is not included in the subsequent analysis of the data for several reasons. The clay is of presumed marine origin, although microfossil support for this is weak (Jordan, 1987), and the contact was abrupt indicating erosion of the peat during inundation. This data point is therefore not an accurate reflection of RSL at that time. The spread of data may be expected due to the problems associated with sediment compaction, whereby different types and thicknesses of sediment are affected by compressional processes by differing amounts (see 10.4.5). This effect is clearly shown in Figure 10.6, which shows all the sea-level data from the Humber Estuary.

10.4.4 Analysis of all SLIPs from the Humber Estuary

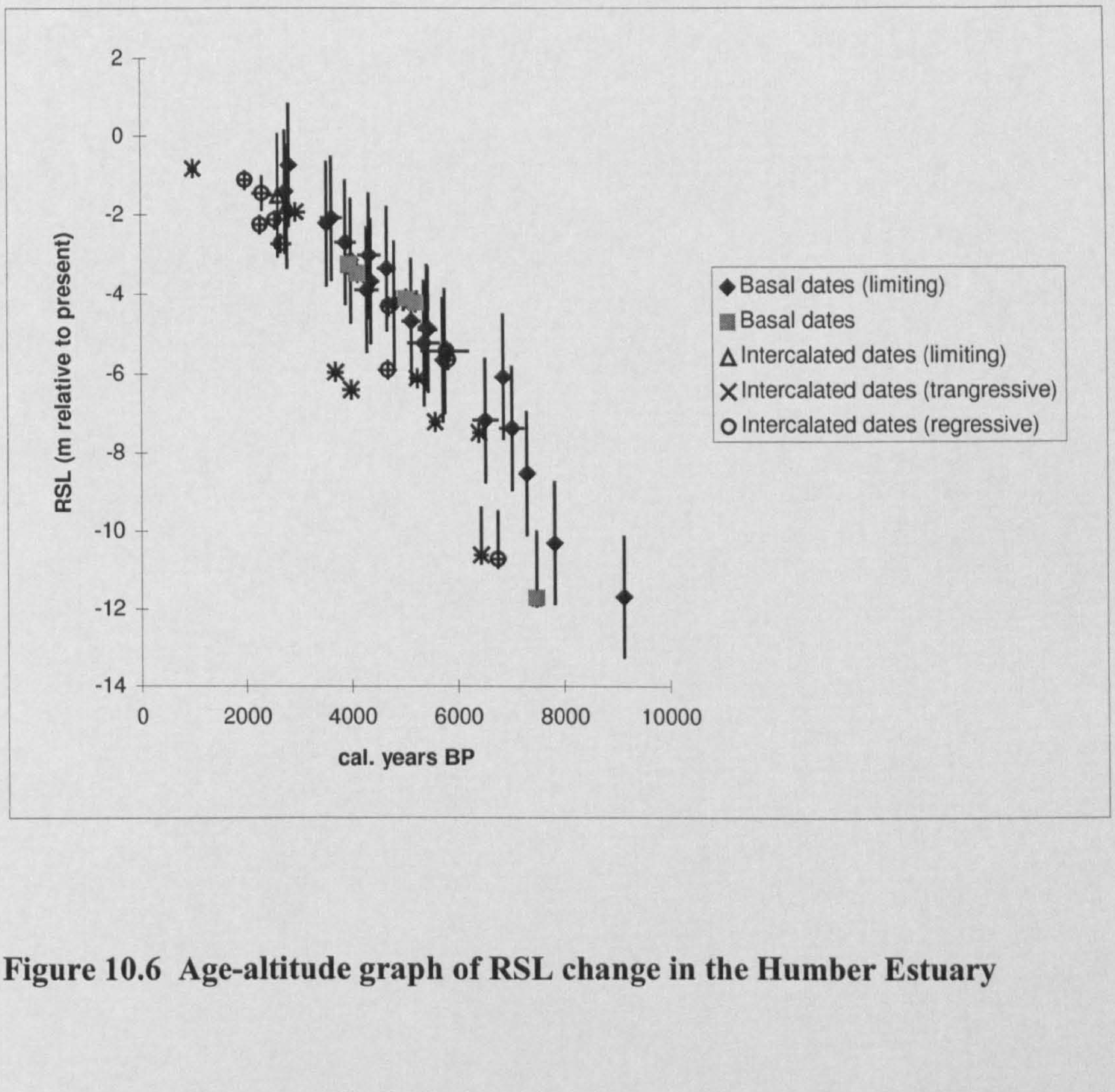


Figure 10.6 Age-altitude graph of RSL change in the Humber Estuary

Figure 10.6 shows that generally the limiting basal dates plot above the main band of SLIPs, reflecting accumulation under freshwater conditions above direct tidal influence. However, some of these limiting data points, particularly from the Ancholme valley (Neumann, 1998), plot closely with the other SLIPs, which suggests these basal dates are a good indicator of rising sea-levels. The basal peat dates, which demonstrate estuarine conditions, also lie towards the top of the band reflecting their relatively compaction free nature.

It is also apparent that the spread of data is generally greater the older the age of the SLIP. This may be related to increased compaction associated with deeper, older samples (e.g. Union Dock data, Long *et al.*, 1998a), and perhaps the development of basal limiting peats not related to contemporary sea level (Kiden, 1995). However, such convergence trends are apparent from other areas (Louwe Kooijmans, 1974; Devoy, 1979; 1982; van de Plassche, 1980; 1982; Kiden, 1989; 1991, van Dijk *et al.*, 1991). Devoy (1979; 1982) observes that the gap between the sea-level curves from the inner and outer Thames estuary (where data points from the inner estuary plot higher) is progressively reduced towards the late-Holocene. This is explained by a combination of differential compaction, tidal range changes, river discharge changes, and differential land subsidence. The collection of further sea-level data from south-east England (Long, 1992; Long & Innes, 1993), and following a re-investigation of the Thames data, Long & Shennan (1993), Long (1995), and Haggart (1995) have shown that differential crustal movements between the inner and outer estuary are unlikely. They suggest that this discrepancy is related to the processes of sediment compaction and tidal differentials. Therefore, the assumption of a constant palaeotidal regime throughout the Holocene is likely to be unacceptable. Research addressing this issue in the Humber and North Sea region is one aim of the LOEPS work currently ongoing (Shennan *et al.*, in press).

Van de Plassche (1980; 1982) cites a reduction in river gradient effect over time as the estuary infills and the gradient levels off as the reason for the dissipation in the gap between tidal curves from the Rhine-Meuse delta (see section 2.5.7). This has been confirmed by later research (van Dijk *et al.*, 1991). This 'gradient effect reduction curve' (van de Plassche, 1980) may be evident from the Humber sea-level data, although perhaps not as clearly as the Rhine-Meuse data due to differences in the morphology and nature of the respective perimarine areas.

10.4.5 Analysis of Sea-Level Trends from the Lower Aire Valley

Figure 10.7 summarises how the data collected in this study compare with the band of sea-level data obtained from elsewhere in the Humber.

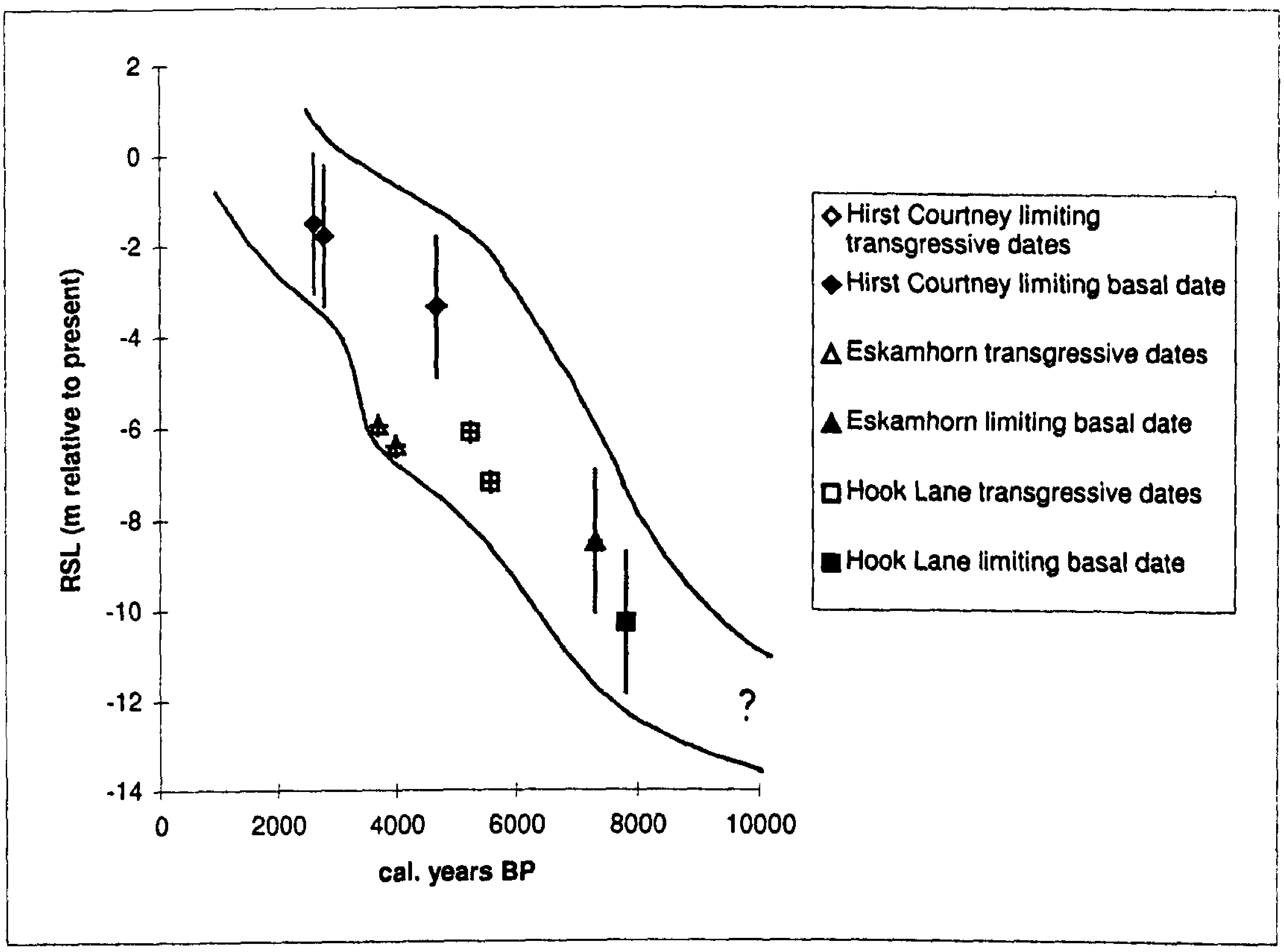


Figure 10.7 Age-altitude graph of RSL change in the lower Aire valley

The limiting basal dates from Hook Lane and Eskamhorn suggest the influence of rising sea level on sedimentation in the perimarine zone by initiating peat formation above contemporary tidal levels between 7914 and 7218 cal. yrs BP upwards of -10.28 m. The intercalated dates from Hook Lane and Eskamhorn show that estuarine conditions penetrated into the lower Aire valley progressively between 5567 and 3700 cal. yrs BP. This phase of positive sea-level tendency initiated freshwater peat formation upstream at Hirst Courtney due to ponding back of freshwater drainage. The transgressive SLIPs from Hook Lane and Eskamhorn both plot towards the base of the sea-level envelope in Figure 10.7. This may be attributable to the 'floodbasin effect' (Van Veen, 1950; Zonneveld, 1959), whereby the amplitude of the tidal wave is lowered in an upstream direction due to frictional dissipation of tidal energy as it passes into wide shallow estuaries, or to the increased storage effect as the tidal flood volume becomes absorbed upon entering into floodbasins within estuaries (van de Plassche, 1980; 1984; Kiden, 1989; 1991; van Dijk *et al.*, 1991, see section 2.4.7). The floodbasin effect has been proposed to explain the drowning of peat deposits in the inner Severn estuary, where a decrease in height of sedimentary contact is observed upstream, suggesting rapid flooding of perimarine storage basins (Hewlett & Birnie, 1996).

Whilst this phenomenon is perhaps to be expected within perimarine areas, the intercalated nature of the SLIPs from the lower Aire valley means that sediment compaction can equally be invoked to explain the position of these points within the Humber sea-level trend. Compaction and consolidation can arise due to many mechanisms, such as weight of overburden and the effects of dewatering (Paul & Barras, 1998), and can substantially alter the elevation of the original peat surface (Jelgersma, 1961; Hawkins, 1984; Smith, 1985b; Allen, 1999). However, it is extremely difficult to quantify due to variability in sediment thickness and composition, and the diagenetic and burial history of the sequences (Smith, 1985b; Greensmith & Tucker, 1986).

If compaction is to be used to explain the position of these data points, it is necessary to explain why the SLIPs from the lower Aire are more adversely affected by compaction than other intercalated dates in the other areas of the estuary. The SLIPs sampled from the lower Aire valley overlie considerable thicknesses of largely organic floodplain peat deposits, whereas the data used by Long *et al.* (1998a) are mainly from intercalated clastic and organic deposits, which may not be as badly affected by compression. The resulting pattern is one of uneven subsidence of stratigraphic contacts in Holocene coastal lowlands (Greensmith & Tucker, 1986). This may, in part, explain the lower position of sea-level data points from the lower Aire compared to those seawards from the Humber estuary area.

The freshwater transgressive contacts at Hirst Courtney represent the maximum culmination of sea-level rise in the Humber estuary, and possibly mark the Holocene marine limit in the lower Aire valley. The palaeoecological data indicates freshwater conditions (apart from some probably allochthonous brackish diatoms) and therefore the points may be expected to plot well above the general trend, but they seem to plot within the middle of the band (Figure 10.7). Considering the inland location of this site, their relative position could perhaps be expected to be elevated further due to the river gradient effect. The SLIPs do plot slightly above the estuarine transgressive contacts (Figure 10.5) but it is possible that their lower than expected position may be the result of the disappearance of the river gradient effect and the upstream migration of the floodbasin effect (van Dijk *et al.*, 1991) as Holocene sea levels reached their maximum extent (Long *et al.*, 1998a). Compaction may also be a factor in their relatively low position. A similar freshwater transgressive contact may also be inferred from the contact between floodplain peat and upper clay in the lower Trent valley at Bole Ings, dated to 2690±100 BP (2985 to 2496 cal. yrs BP, Brayshay & Dinnin, 1999) which supports this sea-level maximum. However, Brayshay & Dinnin (1999) do not consider this interpretation for the observed change, the litho-, bio-, and chronostratigraphic evidence suggests this is a likely explanation.

10.5 Controls on Sedimentation

Mechanism of Floodplain Development

The stability of the *Alnus* carr vegetation, the clayey organic nature of the floodplain substrate, and lack of stratigraphic evidence for significant shifting of the river channel position, suggests a resistance to channel migration and change (*cf.* Törnqvist, 1993b) within the lower Aire valley. (Brown, 1996) suggests that from *c.*6000 BP, limited channel change was restricted to patterns of avulsion, and such stability is common throughout lowland Britain during this time (Macklin & Needham, 1992; Macklin & Lewin, 1993). This pattern of floodplain development in the lower Aire is similar to the stable-bed, aggrading-banks (SBAB) model by Brown & Keough (1992a) which allows for continuing floodplain aggradation with no change in bed height if flood magnitudes increase or sediment availability is maintained. In low energy floodplain systems such as those in the lower Aire valley, two principal factors can be invoked to explain floodplain aggradation and stability. Fine sediment delivery from terrestrial and estuarine sources may deposit clay levees along the banks of the river (which may be maintained by dense vegetation), and continued *Alnus* carr regeneration due to sea-level rise leads to the accumulation of carr peats. In effect, the river channel becomes deeper, due to the disequilibrium between bed aggradation and floodplain deposition (Brown & Keough, 1992a), but accretion and stability of the floodplain is maintained by high sediment inputs, carr vegetation, and continued sea-level rise.

Lagoonal conditions at Hook Lane

After *c.*7000 BP (<*c.*7800 cal. yrs BP) a suite of humic clays are recorded at Hook Lane and along the lower tract of the Rawcliffe long profile transect (Figures 6.6 and 6.13), which are finely laminated by leaf and wood detritus. The litho- and biostratigraphical evidence suggests this unit formed in a permanently submerged open body of water with a brackish tidal influence. The striking similarities between the palaeoenvironmental characteristics of this deposit, and the 'fluviolagoon' palaeoenvironment described by Van der Woude (1983; 1984; 1985) from the Rhine/Meuse delta in The Netherlands, and 'lagoonal' conditions in the Dutch Province of Zeeland (Vos & de Wolf, 1994; 1997), has led to these terms being used in this study. The origins of tidal floodbasins such as this have received no attention in Britain owing to this palaeoenvironment being hitherto unrecorded, and therefore, the conditions necessary for their formation and the appropriateness of the term 'fluviolagoon' and 'lagoon' is discussed here.

Shennan (1986b) and Wheeler (1995) refrained from using the term 'lagoon' or 'lagoonal sediments' to describe the transitional low energy quiet water environment in the inter-creek areas behind saltmarshes in Fenland Holocene sediments due to the common association of these terms with coastal barriers. Since there is no evidence for a coastal barrier during any time in the Holocene in the Wash or Fenland region, Shennan (1986b) advised against the use of such terms to avoid confusion with such back barrier sedimentary systems. A closed coastal barrier is not suggested in this study. The use of the term lagoon however, does imply usually the presence of a coastal barrier system, which may or may not have an opening to the sea (Reineck & Singh, 1973; Carter, 1988; Cooper, 1994). It is argued here that the expansion of peat forming fen carr communities in the lower Aire valley from *c.*7000 BP (<*c.*7800 cal. yrs BP) may have created topographical highs, which impeded drainage to some extent, and the combination of estuarine and fluvial processes outlined below may be capable of forming 'lagoonal' type conditions in the perimarine zone.

In addition to the 'fluviolagoonal' clays described from the Rhine-Meuse delta (Van der Woude, 1983), which began forming *c.*7400 BP (*c.*8400 to *c.*8000 cal. yrs BP), Van Straaten (1957) and Bennema & Pons (1957) describe similar laminated humic clays from the Dutch Province of Zeeland, around the perimarine area of the lower river Schelde in the southwest Netherlands. In this area, the formation of basal peat is suggested as the first indirect evidence for the effects of sea-level rise, and brackish lagoon formation reflects the first direct evidence for sea-level rise. Vos & van Heeringen (1997) discuss the processes associated with the formation of this lagoonal facies, which is defined lithostratigraphically as the 'Velsen layer', within the lower 'Calais deposits'. Vos & de Wolf (1994; 1997) carried out diatom analyses on the Velsen layer sediments and concluded that the environment of deposition was a permanently submerged lagoonal environment with a moderate but increasing tidal influence, which was sheltered from the sea by intertidal sand flats. No radiocarbon dates are available for the exact time of formation but Vos & de Wolf (1997) suggest lagoons existed from *c*.8000 BP (*c*.9200 to *c*.8600 cal. yrs BP), inferred from correlation with the time-depth position on the sea-level graph. Vos & van Heeringen (1997) suggest the following factors are important in the formation of these 'lagoons', which are be considered appropriate analogies for the initiation of equivalent environments in the lower Aire valley.

1) Rapid sea-level rise. During the early- to mid-Holocene, rates of sea-level rise were higher than later periods, and were sufficiently rapid to outstrip the rate of sedimentation.

2) Low sediment accumulation rates. The transport capacity of tidal currents at this time appears to have been low, since fine grained clayey sediment accumulated in the lagoon basins, rather than sand.

3) Low tidal range. Vos & van Heeringen (1997) suggest that the tidal range was low during this time (but increased by up to 1 m during the later stages of the Holocene).

4) Dampening of the tides caused by frictional effects. The effect of bottom friction in the tidal basin, and an increase in storage capacity in the 'lagoons', results in the dissipation of tidal energy or the 'floodbasin effect' (*cf.* Van Veen, 1950; Zonneveld, 1959; van de Plassche, 1980, see section 2.4.7). Therefore, the lagoons never completely filled or drained during the tidal cycle. In riverine deltaic parts of coastal plains, the surface area occupied by such lagoonal floodbasins tends to widen in a downstream direction (Allen, 1965). This may explain why the 'fluviolagoonal' environment is geographically limited in extent to the lower portion of the study area. The 'fluviolagoon' palaeoenvironment may be therefore be regarded as a result of this downstream increase in the floodbasin surface and partly due to sea-level rise (Van der Woude, 1983). Due to the likelihood of a considerable floodbasin effect the relationship between lagoon sediments and contemporary tide levels is uncertain.

It seems likely, that these lagoons are related to the meeting of rising sea waters and downflowing river drainage in the perimarine zone. The freshwater drainage that used to flow freely out into the estuary is impeded higher up estuary, in the lower reaches of the Aire valley, causing ponding of water. The possibility that the fen carr conditions also created organic 'barriers', which constricted the river inlet but allowed tidal penetration, in conjunction with the above factors (Vos & van Heeringen, 1997), is probably influential in creating lagoonal and possibly deltaic conditions, with large ponded areas of brackish water, creeks and levees in the lower Aire from *c*.7000 BP (*c*.7800 cal. yrs BP). The increase in marine influence recorded in the diatom evidence from these humic clays in this study (see section 9.6) and from the Velsen layer in The Netherlands (Vos & de Wolf, 1994; 1997), reflect the increasing proximity of estuarine conditions as sea level continues to rise.

The demise of this 'lagoonal' phase coincides with an apparent increase in marine influence into the basin in the lower Aire valley. Therefore, it is apparent that a fall in the rate of RSL rise cannot alone be invoked to explain the demise of lagoonal conditions *c*.6000 BP (*c*.7200 to *c*.6600 cal. yrs BP). This is very similar in timing to the events culminating in the disappearance of corresponding palaeoenvironments in the Rhine-Meuse delta (Van der Woude, 1983; 1984; 1985) and in the Province of Zeeland (Vos & van Heeringen, 1997) which also disappear *c*.6000 BP (*c*.7200 to *c*.6600 cal. yrs BP). Again, Vos & van Heeringen (1997) provide the an explanation which may also apply to the lower Aire valley. They suggest that this increase in penetration of marine water inland, brought with it an increase in sediment, which enabled sediment accumulation rates in the basin to outstrip the rate of sea-level rise, which caused the fluviolagoon to silt up *c*.6000 BP (*c*.6800 cal. yrs BP). Such a scenario is termed an 'accretionary surplus' and leads to gradual reduction in volumetric capacity of tidal basins, and the encroachment of marsh and swamp conditions (Nichols, 1989). The occurrence of intertidal mudflat dwelling diatoms at the top of the lagoon sediments at Hook Lane (see section 9.6) suggests these conditions were influential in the silting-up of the lagoons in the lower Aire valley. The close correspondence between events in the lower Aire valley, both in timing and nature, suggests the possibility of a regional effect across the North Sea basin. The possible causal factors concerning the demise of the lagoons are described below (from Vos & van Heeringen, 1997):

1) A slowing down in the rate of RSL rise (relative to sedimentation). This may have caused the balance between rates of water level rise in the perimarine area and sedimentation rates to gain equilibrium (van der Spek & Beets, 1992; Beets *et al.*, 1992). This is supported by diatom evidence from this Hook Lane

and in the lower Schelde deposits in Zeeland (Vos & de Wolf, 1997) which suggest sedimentation largely within or above the intertidal range towards the end of lagoon conditions.

2) An increase in tidal asymmetry (i.e. the difference between the magnitude and duration of the flood and ebb tide). This change will affect both the amplitude and velocity of the tidal wave, causing an increase in inshore tidal velocities, which results in a net increase in sediment input into the tidal basin (van der Spek & Beets, 1992). The ebb tides are no longer strong enough to flush sediment out of the lagoons, which begin to fill in. This tidal asymmetry is caused by the changing morphology of the North Sea basin as sea-level rises, which causes distortion of the tidal wave during propagation (van der Spek & Beets, 1992; Vos & van Heeringen, 1997).

3) A change in tidal range. Vos & van Heeringen (1997) present evidence for an increase in tidal range during the period of lagoon infill. The change in the velocity of the tidal wave, and tidal range changes, are functions of decreasing frictional effects due to increasing water depths associated with continued sea-level rise. This effect may be inferred in the lower Aire valley. Although MSL appears to be rising as the lagoon silts up (marine diatoms are increasingly abundant), this could the misleading expression of increasing tidal range. In this scenario, MTL remains the same, but the level of MHWST increases, giving the impression of a rise in sea-level. It is hoped that new research by Shennan *et al.* (in press), addressing palaeotidal change in the Humber estuary, will shed further light on the validity of this explanation.

The silting-up of the 'fluviolagoons' in the lower Aire valley lead to the expansion of *Alnus* carr conditions back over the coring site at Hook Lane. Fen carr conditions then persisted at all three sites until organic sedimentation was replaced by clastic deposits. The riverine and estuarine sediments overlying the floodplain wood peats are associated with a time transgressive rise in sea level up the lower Aire valley from 4555±45 BP (5439 to 5043 cal. yrs BP) to 2560±45 BP (2757 to 2480 cal. yrs BP). This extension of the estuary is likely to have influenced paludification within the shallower valleys of the Humberhead Levels. Peat initiation is dated at the river Don at Thorne Waterside to 4230±100 BP (4865 to 4575 cal. yrs BP), at Crowle to 4230±70 BP (4874 to 4560 cal. yrs BP), and at the base of the Idle floodplain at Misterton to 4330±120 BP (5296 to 4560 cal. yrs BP, Buckland & Dolby, 1973; Buckland &

Sadler, 1985; Smith, 1985), and on Thorne Waste (Smith, 1985a) from 4545±75 BP (5452 to 4878 cal. yrs BP).

Late-Holocene Sediment Delivery

The biostratigraphic record is complete up until the top 1.5 m or so in all cores. Some of the upper metre or so of sediment is probably derived from recent warping practices. Therefore, the assumption that the upper oxidised clastic sediments in the Humberhead levels area are derived from Roman and post-Roman land clearance and agricultural activities (Buckland & Dolby, 1973; Samuels & Buckland, 1978; Buckland & Sadler, 1985; Riley *et al.*, 1995) in not entirely upheld on this evidence. Although the original assertion was based on localised stratigraphic investigations in conjunction with studies of archaeological sites, such as the massive influx of eroded sediment which buried the Roman site at Sandtoft, Dinnin (1997b) has recorded similar orange-brown oxidised deposits within the upper floodplain sediments of the Old River Don and the Torne and Idle rivers. Dinnin (1997b) suggests these deposits are derived from the local Humberhead Levels catchments, and record the combined effects of increasing soil erosion (during the Roman and post-Roman period) and impounding of freshwater run off due to RSL rise (Buckland & Sadler, 1985).

The correlation of deposits based on their oxidised colour (see above) is a highly inappropriate means of categorisation, as it can lead to spurious chronostratigraphic correlations where no independent support for the time of deposition is available (such as Roman pottery), since it merely reflects the local redox conditions in relation to the height of the water table. For instance, Buckland *et al.* (1990) tentatively relate the occurrence of a reddish brown oxidised alluvium above peat at North Ferriby to the suggested Roman and post-Roman phase of soil erosion. It is probable that the impact of Roman activity did have a considerable influence on soil erosion and sedimentation in the Humberhead Levels. However, it must be recognised that local variations are important, so that each site must be interpreted on the basis of a combination of litho- bio- and chronostratigraphic evidence to avoid ambiguous extrapolations.

Much of the upper sediments, at least in the lower tract of the Aire and Ouse valleys are brackish, contrary to the suggestion that much of the alluvial sediments in this area are freshwater in derivation (Lillie, 1997), and were deposited during a time transgressive marine incursion well before the arrival of the

Romans. If detailed microfossil analyses of the upper sediments within the perimarine zone of other rivers in the Humberhead Levels are carried out, they may reveal a similar estuarine origin for the sediments as in the lower Aire valley.

The source of these clastic sediments is difficult to ascertain. The sediment itself could either be supplied from offshore sinks or from catchment erosion (Pethick, 1990) and although the methodology adopted in this study is not designed to recognise specific sediment sources, the possible cause of sediment delivery may be inferred from the palaeoecological record. Limbrey (1978) and Brown (1996) suggest that as floodplain woodland became deforested, river bank resistance to erosion decreased, which led to increased sedimentation across the floodplain. Similar evidence has been proposed to explain an acceleration in valley floor alluviation from the Bronze Age, and particularly the Iron Age, onwards, in other British river valleys such as the Severn-Avon (Shotton, 1978; Brown, 1982; 1983; Brown 1987a; Brown & Barber, 1985) and those draining the Weald in southern England (Scaife & Burrin, 1983; 1985; 1987; Burrin, 1985; Burrin & Scaife, 1984).

There is very little evidence for human interference with woodland (see section 10.2), either dryland or wetland, in the lower Aire valley, although the sites chosen for study were not selected to maximise the chance of uncovering anthropogenic activity. This suggests that the sediments washed into the carr floodplain by high tides or flood events, are more likely to be associated with sediment transported up the estuary from offshore sources, and are perhaps not derived from extensive soil erosion within the immediate catchment.

The upper levels of the pollen diagrams from Hook Lane and Eskamhorn do not record significant woodland losses other than those locally attributable to flooding, although they do receive an increase in inorganic sediment input from *c*.4900 BP (*c*.5600 to 5500 cal. yrs BP) commensurate with an increase in frequency of tidal flooding. This lack of evidence for anthropogenic activity is in agreement with other pollen diagrams from the Humber lowlands at sites such as Union Dock, Sandholme Lodge, Kilnsea Warren, and East Clough (Long *et al.*, 1998a), which all show high arboreal pollen percentages indicating extensive *Quercus* fenwoods and *Alnus* carr between 8170±45 and 3640±45 BP (9259 to 3832 cal. yrs BP). This supports other palaeoecological studies from the region which suggest a lack of significant anthropogenic clearance during the Neolithic (Smith, 1958a; Turner, 1962; 1987; Beckett, 1981; Gilbertson, 1984a; Flenley, 1990). It is possible that some sediment was delivered to the estuary due to soil erosion from the upland areas. Bush & Ellis (1987) and Bush (1993) provide evidence for open areas of chalk grassland related to human activities during the Mesolithic at Willow Garth in the Great Wold Valley, and Ellis (1990) and Flenley (1990) suggest that the lighter soils of the Yorkshire Wolds were largely cleared of woodland by *c.*5000 BP (*c.*6000 to *c.*5500 cal. yrs BP), although the impact of Neolithic people in Holderness is rather limited. This evidence alone is not significant enough to infer regional changes in sediment regime associated with such clearance.

Evidence for woodland clearance which is likely to release stored sediment into the rivers and estuary is widespread from *c.*3500 BP (*c.*4100 to 3500 cal. yrs BP) and by *c.*2600 BP (*c.*2900 to 2400 cal. yrs BP) large areas of mid-Holocene woodland were cleared (Smith, 1958a; Turner, 1962; 1987; Beckett, 1981; Gilbertson, 1984a; Smith, 1985; Flenley, 1987; 1990). Pollen records from Newton Marsh and Barrow Haven (Long *et al.* (1998a) suggest at least a depleted local woodland cover by this time, and after the flooding of carr woodland at Hirst Courtney (lpaz HCP-5), the frequency of tree pollen averages just 11%. Devoy (1980) suggests that such estuarine zones would have formed a centre for Bronze Age hunting, fishing, and gathering. Therefore, it may be possible that the sedimentation associated with the coastline advance which occurred between *c.*2600 and 1900 cal. yrs BP in the Humber (Long *et al.*, 1998a) was, in part, linked with increased soil erosion associated with this landscape clearance. However, the marine transgression and phase of inorganic sedimentation was well under way by this time suggesting other sediment sources are likely.

Preece & Robinson (1984) record direct stratigraphic evidence for local colluviation associated with woodland clearance at Castlethorpe in the Ancholme valley dated to 3410±80 BP (3846 to 3465 cal. yrs BP), and the inorganic unit that seals the organic mere deposits in Holderness perhaps relates to extensive colluviation as a result of Bronze and Iron Age deforestation (see Dinnin, 1995). However, the extent to which the sparse tree cover at Hirst Courtney contributed to soil destabilisation and erosion, and the inorganic sedimentation that ensued subsequent to the inundation of the site, is difficult to quantify. The presence of green-grey clastic sediments at this site, which were absent from all other sites in the lower Aire valley, may suggest that the source of this sediment may be from the catchment soils and is terrestrial in origin. However, this colour of sediment may relate to the geochemistry of the water, which is likely to be different in the freshwater fluvial area. Sediments recorded from upper reaches of lowland rivers in southern England are also characterised by green-grey clays whereas lower reaches contain a sequence of blue-grey clays (Burrin & Scaife, 1984; Burrin, 1988; Waller *et al.*, (1988).

A direct link between catchment woodland clearance due to human activities and rates of floodplain sedimentation and patterns of coastal change is yet to be unequivocally demonstrated in the Humber basin. A major limitation is that the sediments end up mixed in the Humber estuary which receives drainage from one fifth of England (Pethick, 1990) as well as considerable amounts of sediment from the eroding Holderness cliffs (Al-Bakri, 1986). The possible source of sediments is effectively so large, that comparisons between the local palaeoecological record and increased sediment input is not entirely justified. The relative input of sediment from terrestrial and offshore sources is likely to have changed throughout the Holocene and new sedimentological techniques in conjunction with a revised methodology are required to investigate the link between sediment sources and sinks.

No evidence has been gathered which relates to climate change in the lower Aire valley during the mid-Holocene period of organic sediment accumulation. Therefore, the extent to which phases of vegetation change, sedimentation, and floodplain development reflect possible climatic events is not known. It is possible however, that the change in estuarine dynamics between *c.*3300 and *c.*1900 cal. yrs BP (Long *et al.*, 1998a) may be linked with a period of northern Hemispheric climate change suggested by Van Geel *et al.*, (1996) although independent evidence from the Humber area is not available to substantiate this inference.

Chapter 11 Conclusions

11.1 Précis

A database of over 100 boreholes has been used to establish the nature and spatial variability of sedimentary environments in the perimarine zone of the lower river Aire during the mid- to late-Holocene. Floodplain aggradation has been shown to be time transgressive up the valley, reflecting waterlogging ahead of rising sea level. The earliest date for peat formation in the valley bottom has been dated to *c*.7000 BP (*c*.8000 to *c*.7600 cal. yrs BP) at Hook Lane. Organic sediment accumulation took place under *Alnus* carr conditions, which were replaced by lagoonal sediments. Several factors have been invoked to explain this change in sedimentary environment (*cf.* Vos & van Heeringen, 1997). It is likely that a combination of continued sea-level rise, low sediment accumulation rates, and increased tidal frictional effects, led to the ponding of fresh/brackish water, and resulted in the flooding of the *Alnus* carr in the lower Aire valley. The widespread accumulation of organic deposits in the valleys of the inner Humber estuary during this time may have impeded the seaward drainage of freshwater, and contributed to the formation of 'fluviolagoons' (*sensu* Van der Woude, 1983).

Palaeoecological analysis of the lagoon sediments has shown that they were tidally influenced, permanently submerged, quiet water environments, surrounded by fringing reedswamp and fen carr. Diatom evidence suggests a gradual increase in marine influence during the period of lagoonal sedimentation, which points to the continued penetration of marine conditions up the valley. Intertidal depositional conditions at the final stages of lagoon conditions, indicate an 'accretionary surplus' in the floodbasin (*sensu* Nichols, 1989). The silting-up of the 'fluviolagoon' at *c*.6000 BP (*c*.7200 to *c*.6600 cal. yrs BP) was perhaps linked with a slowing down in the rate of sea-level rise relative to sedimentation. Van der Spek & Beets (1992), and Vos & van Heeringen (1997) suggest an increase in tidal range due to a reduction in tidal friction, and an increase in tidal asymmetry, may result in a net input of sediment in to the lagoon.

Progressive waterlogging of lower Aire valley further inland, against a background of generally rising water levels, resulted in paludification of the valley floor at Eskamhorn and Hirst Courtney, dated to 6385±45 (4831 to 4525 cal. yrs BP) and 4150±45 BP (7380 to 7202 cal. yrs BP) respectively. The dominant vegetation on the lower Aire floodplain during the period of organic sedimentation was *Alnus glutinosa* dominated fen carr, with *Corylus avellana* and *Salix* probably important components. This vegetation community persists for several millennia at all sites. This persistence suggests internal regeneration of the *Alnus glutinosa* community. This was facilitated by continued sea-level rise, which maintained high groundwater levels, and also, possibly, competitive cycles between *Alnus glutinosa* and *Salix*, and variations in the production and decomposition of organic material, which created a suitable substrate (Waller *et al.*, 1999). Fluctuations in the pollen representation of *Alnus glutinosa* are recorded at all sites, although these are likely to be related to local changes in the spatial distribution of wetland vegetation on the floodplain over time (Waller, 1998), perhaps initiated in some cases by the effects of windthrow (Brown, 1997b).

The presence of clayey organic floodplain substrates and dense carr vegetation has helped to maintain a relatively stable channel position during the mid- to late-Holocene (*cf.* Törnqvist, 1993b). The model of floodplain formation within the lower Aire valley is analogous to the SBAB model by Brown & Keough (1992a).

The dryland vegetation appears to have been dominated by a mixture of *Tilia*, *Quercus*, *Ulmus*, and probably also *Fraxinus excelsior* and *Corylus avellana* during the mid-Holocene. *Pinus sylvestris* is also shown to be present within the regional landscape for much of this period. No unequivocal evidence for human activities has been identified in the palynological record, although as Sagan (1977) notes, 'absence of evidence is not necessarily evidence for absence'. It is likely that humans exploited such wet environments but did not attempt to modify them significantly until much later in prehistory. The floodplain corridor perhaps provided suitable areas for hunting and fishing activities (Devoy, 1980), which are not detectable in the pollen record. A marked decline in *Tilia* and *Pinus sylvestris* abundance is recorded at Hirst Courtney, and dated to *c.*3900 BP (*c.*4600 to *c.*4000 cal. yrs BP) by linear interpolation. This event has been recorded in other pollen diagrams from the Humberhead Levels, and river Trent floodplain (Smith, 1985a; Brayshay & Dinnin, 1999), albeit somewhat later, and has been attributed to deliberate tree clearance by Bronze Age people. Although an anthropogenic explanation cannot be discounted, it has been suggested that expanding wetland communities during this time (*cf.* Waller, 1994b), and closure of the *Alnus* carr, may be invoked to provide an equally plausible explanation for the decline in these taxa at Hirst Courtney.

At all three sites, wood peat is overlain by an organic silty clay unit. Biostratigraphic evidence shows that a decline in *Alnus glutinosa* abundance accompanied this lithostratigraphic change, which was attributable to a time transgressive expansion of marine conditions up the lower Aire valley. At Hook Lane, rising water table caused the *Alnus glutinosa* vegetation to thin, and freshwater fen carr conditions were replaced by saltmarsh at 4555±45 BP (5439 to 5043 cal. yrs BP). At Eskamhorn, rising water levels are registered by 3670±45 BP (4130 to 3852 cal. yrs BP) by the replacement of *Alnus* carr by *Salix* dominated carr communities. Increased environmental wetness is indicated by the decline in *Salix*, and abundance of *Filipendula* dominated fen meadow vegetation. This is replaced by saltmarsh at 3440±45 BP (3828 to 3572 cal. yrs BP). Finally, upriver at Hirst Courtney, wetter conditions on the floodplain are suggested by an increase in *Filipendula* at 2665±45 BP (2850 to 2740), followed by a decline in *Alnus glutinosa*. A sedimentary contact below silty clay is dated to 2560±45 BP (2757 to 2480 cal. yrs BP), which indicates the flooding of the site, with freshwater reedswamp communities prevalent during this time.

This environmental change in the lower Aire valley adds further evidence for an expansion of the estuary limits from the Neolithic period, culminating with the widespread registration of positive sea-level tendencies within the Humber estuary between *c.*3300 and *c.*2500 cal. yrs BP (see Long *et al.*, 1998a). The flooding of Hirst Courtney with freshwater represents the maximum extension of the estuary during the Holocene in the lower Aire valley. Similar conditions have been recorded from inland at Waddingham Holmes in the Ancholme valley (*c.*2700 cal. yrs BP, Fletcher, 1981) and at Bole Ings in the Trent (*c.*2700 cal. yrs BP), although Brayshay & Dinnin (1999) suggest an anthropogenic interpretation for this event.

Long *et al.* (1998a) provide evidence for a subsequent contraction of estuarine conditions after *c.*2800 cal. yrs BP), with negative sea-level tendencies identified at various sites around the estuary. Although a chronology is lacking from the lower Aire valley to substantiate any suggestions concerning environmental conditions during this time, there is tentative diatom evidence for drying from each of the sites studied, which may relate to the onset of marine regression, and a period of shoreline advance. Lithostratigraphic evidence for the remnants of an upper peat within the lower Aire valley, also provide equivocal evidence to support such an assertion.

The evidence presented from the lower Aire valley suggests that the phase of upper inorganic alluviation was diachronous up the valley, and that the change in sediment type was in fact a transgressive overlap. Much of the upper clastic facies recorded in the valley fill was deposited within intertidal conditions during a marine transgression, which was first registered at Hook Lane in the late-Neolithic, and reached as far as Hirst Courtney by the early-Iron Age. Although it is likely that this period of estuarine expansion coincided with the increasing impact of human activity on the environment, which may have resulted in increased sediment delivery to the estuary (Brown, 1987a), the balance of evidence suggests that the deposition of this sediment was related to an estuary wide change in coastal dynamics, and coincides with a period of climate change (Van Geel *et al.*, 1996). The origin of the upper oxidised facies which seals the floodplain sediments has been attributed to increased alluviation due to soil erosion in the Humberhead Levels during Roman and post-Roman times (*cf.* Samuels & Buckland, 1978; Buckland & Sadler, 1985; Riley *et al.*, 1995). The widespread applicability of this model needs reassessing in the light of this evidence.

The research by Samuels & Buckland (1978), and Buckland & Sadler (1985), clearly shows the burial of Roman archaeology at Sandtoft by substantial thicknesses of sediment. This clearly indicates the adverse effect that Roman and post-Roman occupation had on the local environment at this site. However, it has been shown that much of the sediment in the lower Aire valley, at least, was deposited by estuarine processes prior to this time. The extrapolation of the Roman/post-Roman alluviation hypothesis (*cf.* Buckland & Sadler, 1985) to other sites in the Humberhead Levels, and elsewhere in the Humber region, (see Buckland *et al., 1990*; Dinnin & Brayshay, 1994; Lillie & Grattan, 1994; Dinnin, 1997b), to explain the cause of late-Holocene sediment delivery within the region cannot be supported without better dating control. The identification of Roman and post-Roman alluviation simply on the occurrence of orange-brown oxidised clay deposits is highly inappropriate. It is recommended henceforth that this lithostratigraphic subdivision is no longer used without substantiating chronostratigraphic evidence in support of the assumed Roman/post-Roman age for the deposit.

The combination of sedimentological and palaeoecological data from floodplain environments in the perimarine zone demonstrates the important interactions between geomorphology, vegetation dynamics, and sea-level change. The application of pollen and diatom analyses to the organic and inorganic sediments has enabled the dominant controls on floodplain development to be established. In perimarine areas, the spatial patterns in composition of floodplain corridor vegetation are strongly influenced by the influence of both freshwater drainage and estuarine dynamics. These factors were apparently more important in shaping floodplain evolution in the lower Aire valley than climate or human interference.

11.2 Appraisal of Palaeoenvironmental Techniques

The main evidence for the environmental reconstructions in this thesis has been based on pollen and diatom evidence. However, the precise nature of some of the vegetation communities identified is unclear, partly through a lack of modern analogues, but also because of the taphonomic limitations of palynology and the overlap between important wetland and dryland species. For example, the local vegetation community prevailing during the transition from fen carr to fen and saltmarsh at Eskamhorn and Hook Lane is difficult to interpret due to uncertainty in identifying local pollen types. The application of plant macrofossil analysis in these circumstances may help resolve this matter. For example, *Filipendula* seeds may be characteristic of the fen stage and it may be possible to identify at which point this freshwater community is replaced by saltmarsh, or indeed whether some specimens are able to survive within it. The interpretations of the saltmarsh environments may also be improved by analysing the foraminifera within the sediments. Evans *et. al.* (submitted) have shown that in some coastal situations, the forams are a more powerful tool for palaeoenvironmental reconstructions than diatoms due to taphonomic uncertainties regarding the latter.

Furthermore, the origin of the minerogenic sediments remains unresolved. This means that it has not been possible to quantify the amount of sediment derived from catchment or estuarine/offshore sources. This is because the micropalaeontological data collected (diatoms) only provide evidence on the depositional environment itself and are little help in establishing characteristic sediment types or dominant transport pathways. This important issue needs to be addressed using geochemical analysis of the sediments, in conjunction with particle size analysis and mineral magnetic studies. These sedimentological techniques may help characterise sediments from different sources to enable the link between sediment sink and source to be investigated more directly.

11.3 Recommendations for Future Research

Interpretation of pollen diagrams constructed from floodplain peats in the perimarine zone is problematic due to the lack of pollen rain data from equivalent modern environments. Although the saltmarsh-fen transitional environment has largely been destroyed in the British Isles, collection of contemporary pollen information from alder carr and fen systems could still be very valuable. This will enable quantification of the pollen-vegetation relationship, enabling numerical models to be constructed, which can be used to aid interpretation of past fen environments. It is recommended that modern pollen data are collected from the following areas; the transition between closed alder woods and open areas, alder carr at back of a saltmarsh, alder carr bordering fen, and the transition from alder carr into dry woodland. Furthermore, plant macrofossil analysis should be undertaken to help resolve the uncertainties in recognising the local vegetation communities. Macrofossil analyses should be targeted so that sampling is made over the critical horizons (i.e. transitional vegetation communities).

The sea level data collected from the lower Aire valley in this study add to the large database from the Humber and east coast collected during the LOEPS project. Subsequent to the publication of these results, the data presented in this thesis can be used to enhance inner-outer estuary comparisons of data, using statistical methods. This will enable the relationship between estuarine and perimarine sedimentation to be evaluated with a more substantial dataset.

Further research effort in the lower Aire valley should be concentrated on collecting closely spaced borehole data in the vicinity of areas where possible remnants of a possible upper peat deposit were recorded. This organic sediment is rapidly decaying and, if it can be sampled, will provide important information regarding the period of negative sea-level tendency in the Iron Age.

The application of a multiproxy palaeoecological approach to the reconstruction of the environmental history of the lower Aire valley has enabled a chronological story of valley development to be established. Palaeoenvironmental reconstructions are more robust from organic sediments, where pollen and radiocarbon dating enable reliable estimates of local conditions to be made. Palaeoecological study of clastic sediments is less secure. Although diatom analysis provides information on the nature of the local depositional environment, inferences concerning the causal mechanisms of inorganic sedimentation, and the source of sediments remain subjective. It is recommended that a better methodology is developed, in order to differentiate between changes that occur as a result of increased sediment availability (due to human activity) and those that take place in response to changes in magnitude and frequency of transporting events (e.g. floods due to climate change), Brown (1997). This partly depends on obtaining a better dating resolution, and more precise dating methods, to provide an absolute chronology for inorganic alluvial sequences. Difficulties arise due to the mixing of marine and terrigenous sediments in coastal river valleys (e.g. Pethick, 1988; Long *et al.*, 1998a), and future work will need to concentrate on the application of sediment tracing (geochemistry), to effectively establish the provenance of alluvial deposits.

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Appendix A

Pollen Preparation Procedure

(Based on Moore *et al.*, 1991)

Use *c.* 1cm^3 (equivalent to 1g/1ml of sample)

1. Potassium Hydroxide Treatment

Disaggregation of material and removal of soluble organic compounds (humic acids).

- Place sample in plastic centrifuge tube and 1/3 fill tube with 10% KOH. Heat in a pre-heated water bath at *c.*90°C for 15 minutes. Break up with a plastic rod and stir occasionally.
- Pass/wash through a metal sieve (mesh size *c.*150 µm with distilled water.
- Centrifuge for 5 minutes at 3000 rpm and discard the supernatant.
- Repeat wash until supernatant is clear.

2. Hydrochloric Acid Treatment

Removal of carbonates.

- 1/2 fill tube with 10% HCl, stir, centrifuge and decant.

3. Hydrofluoric Acid Treatment

Removal of silica (clays).

- 1/3 fill tube with HF in a fume cupboard and heat in water bath at 100°C for 30 minutes stirring occasionally with a plastic rod, centrifuge and decant carefully if sample is not gritty.
- If sample is still gritty/silty repeat HF treatment immediately.
- 1/2 fill tube with 10% HCl (to remove any silicofluorides which cloud sample) and heat in a water bath for *c.*20 minutes. Stir with plastic rod, centrifuge and decant.
- Transfer sample to 15ml conical centrifuge tubes.
- Wash with distilled water to remove any traces of HF, stir, centrifuge and decant.
- Repeat wash with distilled water, stir, centrifuge and decant.

4. Acetolysis

Evacuation of unaltered lignin and cellulose.

- Add glacial acetic acid (this removes water from the sample which reacts with the acetolation mixture) , stir, centrifuge and decant.
- Carefully make up the acetolation mixture in a cool water bath of 1 part concentrated sulphuric acid and 9 parts acetic anhydride. Slowly add 10 mls of sulphuric acid to 90 mls of acetic anhydride in a dry beaker stirring vigorously to make a clear or very pale yellow/green mixture. If the mixture turns red/brown in colour, remix the chemicals adding sulphuric acid to the acetic acid more slowly.

- 1/3 fill tubes with acetolation mixture and heat in a water bath at 100°C for 3-4 minutes (no longer) and stir the sample. If 4 samples are prepared, after stirring the last tube, the first one will almost be ready to be removed.
- Top up tubes with glacial acetic acid to quench the reaction, centrifuge and discard the supernatant.
- Add glacial acetic acid, stir, centrifuge and decant.
- Wash in distilled water, centrifuge and decant.

5. Nitric Acid Treatment

Removal of excess lignin and iron sulphide (pyrite).

- Add 10% nitric acid and heat (but do not boil) in a water bath for *c.*3 minutes, (this process also clears darkened microfossils). Stir, centrifuge and decant.
- Wash with distilled water, centrifuge and decant.

6. Staining

- Wash sample with distilled water, adding 2 drops of 10% KOH, (this will assist the staining process).
- Add 3 drops 0.2% aqueous Safranin solution and mix well. Top up with 90% alcohol, stir, centrifuge and decant.
- Wash sample with absolute alcohol, stir, centrifuge and decant.
- Wash the remaining contents of the centrifuge tube into labelled glass vials using the organic solvent TBA.
- Cap the vials and wrap them upright in tissue paper.
- Centrifuge the vials for only 2 minutes then decant the TBA.
- Cover sample with a small amount of silicon oil to prevent drying (usually a layer the same thickness as the sample) and stir with a cocktail stick.
- Evaporate off the remaining TBA in an oven at 40-50°C overnight.

Appendix B

Diatom Habitat and Life Form Definitions

The ecological coding used in this report is based on several sources, principally Denys (1991), de Wolf (1993), van Dam *et al.* (1994), and van de Werff and Huls, (1958-1974). All classifications are a simplification and many diatom species are tolerant of intermediate stages of any classification and across a broad spectrum of classes.

Salinity

The salinity data is drawn up according to the salinity classification by Hustedt (1957) where:

Polyhalobian:	marine (optimum range of 30‰)
Mesohalobian:	brackish (30-2‰)
Oligohalobian-halophile:	requires some low salinity
Oligohalobian-indifferent:	can tolerate low salinity
Halophobes:	fresh water only

Life form

Life form is the simplest form of ecological classification and follows de Wolf (1993).

Planktic:	Aquatic: lives in the water column
Benthic:	Aquatic: lives on or in the sediments
Epiphytic:	Aquatic: is attached to plants, rocks, sand or any substrate
Aerophile:	Aquatic but can withstand occasional exposure to air
Eu-terrestrial:	Mainly terrestrial

Trophic Conditions

Trophic conditions follow the classification of Denys (1991).

Eutrophic:	High supplies of nutrients and high rates of primary productivity
Mesotrophic:	Slightly less nutrient rich with less primary productivity
Oligotrophic:	Low levels and supplies of at least one major nutrient, low primary productivity

Dystrophic:	Indicates higher productivity in acidic bog pools with peat filled margins. Generally water of negligible nutrient content; acidic and usually stained brown

Currents

Water currents indicate if the diatoms require the presence of flowing or still water and follows the classification in de *Wolf (1993)*.

Rheophile:	requires running water
Indifferent:	indifferent
Limnophile:	requires still water

Appendix C

1. Freshwater Basal (limiting) Dates

Site	Lab Code	Age (^{14}C yr BP ± 1σ)	Mid Cal. Age (cal. yrs BP ± 2σ)	Altitude (m OD)	RSL (m)	Indicative Range (m)	Tendency	Stratigraphic Context	Reference
Hirst Courtney	SRR-6271	4150 ± 45	4678 ± 153	-0.60	-3.35	1.55	Positive	Base of basal peat	This study
Eskamhorn	SRR-6222	6385 ± 45	7291 ± 89	-5.75	-8.50	1.55	Positive	Base of basal peat	This study
Hook Lane	SRR-6385	7030 ± 45	7807 ± 108	-7.53	-10.28	1.55	Positive	Base of basal peat	This study
Ancholme (R'bourne)	OxA-7064	*4494 ± 31	5140 ± 152	-1.92	-4.67	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (R'bourne)	GU-5701	3940 ± 70	4343 ± 193	-0.27	-3.02	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (R'bourne)	OxA-7065	3595 ± 50	3896 ± 173	0.06	-2.69	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (R'bourne)	GU-5702	3370 ± 60	3638 ± 173	0.67	-2.08	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (Brigg)	OxA-7137	4990 ± 75	5752 ± 161	-6.34	-9.09	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (Brigg)	GU-5700	4730 ± 100	5369 ± 283	-2.47	-5.22	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (Brigg)	OxA-7136	6170 ± 90	7014 ± 212	-4.61	-7.36	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (Brigg)	GU-5699	5040 ± 170	5760 ± 427	-2.67	-5.42	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (Brigg)	OxA-7090	4300 ± 50	4820 ± 161	-1.48	-4.23	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (Brigg)	OxA-7091	3940 ± 45	4374 ± 138	-0.92	-3.67	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	OxA-7057	6000 ± 50	6857 ± 126	-3.31	-6.06	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	GU-5707	4700 ± 80	5441 ± 156	-2.13	-4.88	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	OxA-7054	4960 ± 40	5726 ± 124	-2.89	-5.64	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	OxA-7055	4650 ± 50	5428 ± 140	-2.07	-4.82	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	GU-5706	3890 ± 60	4299 ± 204	-1.13	-3.88	1.55	Positive	Base of basal peat	Neumann (1998)

Site	Lab Code	Age (^{14}C yr BP ± 1σ)	Mid Cal. Age (cal. yrs BP ± 2σ)	Altitude (m OD)	RSL (m)	Indicative Range (m)	Tendency	Stratigraphic Context	Reference
Ancholme (S.Ferriby)	OxA-7052	3670 ± 40	3994 ± 126	-0.40	-3.15	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	OxA-7053	3310 ± 40	3544 ± 86	0.54	-2.21	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	OxA-7066	2635 ± 45	2755 ± 35	1.34	-1.41	1.55	Positive	Base of basal peat	Neumann (1998)
Ancholme (S.Ferriby)	OxA-7067	2690 ± 70	2832 ± 106	2.02	-0.73	1.55	Positive	Base of basal peat	Neumann (1998)
Union Dock 4 (Grimsby)	SRR-4747	8170 ± 45	9116 ± 135	-8.92	-11.67	1.55	Positive	Base of basal peat	Long *et al.* (1998a)
Hasholme	HAR-7007	5710 ± 100	6518 ± 220	-4.41	-7.16	1.55	Positive	Base of basal peat	Millett & McGrail (1987)

* = average age

2. Basal Dates

Site	Lab Code	Age (^{14}C yr BP ± 1σ)	Mid Cal. Age (cal. yrs BP ± 2σ)	Altitude (m OD)	RSL (m)	Indicative Range (m)	Tendency	Stratigraphic Context	Reference
Union Dock 3 (Grimsby)	SRR-4746	6645 ± 45	7468 ± 76	-8.79	-11.69	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
Kilnsea Warren 1	UB-3901	4384 ± 54	5041 ± 203	-1.20	-4.10	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
Kilnsea Warren 2	UB-3900	4562 ± 59	5221 ± 227	-1.30	-4.20	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
East Clough 1	SRR-4748	3640 ± 45	3958 ± 126	-0.35	-3.25	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
East Clough 2	SRR-4749	3770 ± 45	4126 ± 143	-0.57	-3.47	0.20	Positive	Transgressive contact	Long *et al.* (1998a)

3. Intercalated Dates

Site	Lab Code	Age (^{14}C yr BP ± 1σ)	Mid Cal. Age (cal. yrs BP ± 2σ)	Altitude (m OD)	RSL (m)	Indicative Range (m)	Tendency	Stratigraphic Context	Reference
Hirst Courtney	SRR-6268	2560 ± 45	2619 ± 138	1.25	-1.50	1.55	Positive	Freshwater transgressive contact	This study
Hirst Courtney	SRR-6269	2665 ± 45	2795 ± 55	0.97	-1.78	1.55	Positive	Fen peat	This study
Eskamhorn	SRR-6218	3440 ± 45	3700 ± 128	-1.85	-5.95	0.20	Positive	Transgressive contact	This study
Eskamhorn	SRR-6218	3670 ± 45	3991 ± 139	-2.10	-6.40	0.20	Positive	Fen peat	This study
Hook Lane	SRR-6390	4555 ± 45	5241 ± 130	-2.00	-6.10	0.20	Positive	Transgressive contact	This study
Hook Lane	SRR-6391	4850 ± 45	5567 ± 89	-2.90	-7.20	0.20	Positive	Fen peat	This study
Sandholme Lodge 5	SRR-4894	5615 ± 45	6393 ± 92	-3.37	-7.47	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
Sandholme Lodge 24	SRR-4743	4170 ± 45	4685 ± 151	-1.79	-5.89	0.20	Negative	Regressive contact	Long *et al.* (1998a)
Union Dock 1 (Grimsby)	SRR-4744	5665 ± 45	6429 ± 113	-7.69	-10.59	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
Union Dock 2 (Grimsby)	SRR-4745	5990 ± 45	6745 ± 106	-7.79	-10.69	0.20	Negative	Regressive contact	Long *et al.* (1998a)
Barrow Haven	SRR-4897	2040 ± 40	1995 ± 113	1.78	-1.12	0.20	Negative	Regressive contact	Long *et al.* (1998a)
Barrow Haven	SRR-1373	1080 ± 40	995 ± 68	2.08	-0.82	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
Barrow Haven	SRR-1374	2325 ± 60	2310 ± 157	1.45	-1.45	0.40	Regressive	Whole peat dated	Long *et al.* (1998a)
Newton Marsh Tetney	RCD-1598	2840 ± 70	2948 ± 162	0.97	-1.93	0.20	Positive	Transgressive contact	Long *et al.* (1998a)
Brigg	CAR-60	2625 ± 65	2675 ± 176	1.37	-2.73	0.20	Negative	Regressive contact	Smith *et al.* (1981)
Hasholme	HAR-7005	2530 ± 70	2560 ± 205	-1.85	-5.95	1.55	Positive	Transgressive contact	Millett & McGrail (1987)

Site	Lab Code	Age (^{14}C yr BP ± 1σ)	Mid Cal. Age (cal. yrs BP ± 2σ)	Altitude (m OD)	RSL (m)	Indicative Range (m)	Tendency	Stratigraphic Context	Reference
Thirtle Bridge	GU-5480 GU-5479	*2310 ± 36	2278 ± 76	0.65	-2.25	0.20	Negative	Regressive contact	Dinnin & Lillie (1995)
Thritle Bridge	GU-5489 GU-5490	*2505 ± 36	2556 ± 187	0.77	-2.13	0.20	Negative	Regressive contact	Dinnin & Lillie (1995)
Halsham Carrs	GU-5476 GU-5475	*4195 ± 36	4704 ± 134	-1.40	-4.30	0.20	Negative	Regressive contact	Dinnin & Lillie (1995)
Roos Drain West	GU-5483 GU-5484	*5060 ± 60	5791 ± 137	-2.51	-5.41	0.20	Negative	Regressive contact	Dinnin & Lillie (1995)
Roos Drain West	GU-5477 GU-5478	*5140 ± 43	5822 ± 69	-2.73	-5.63	0.20	Negative	Regressive contact	Dinnin & Lillie (1995)

* = average age