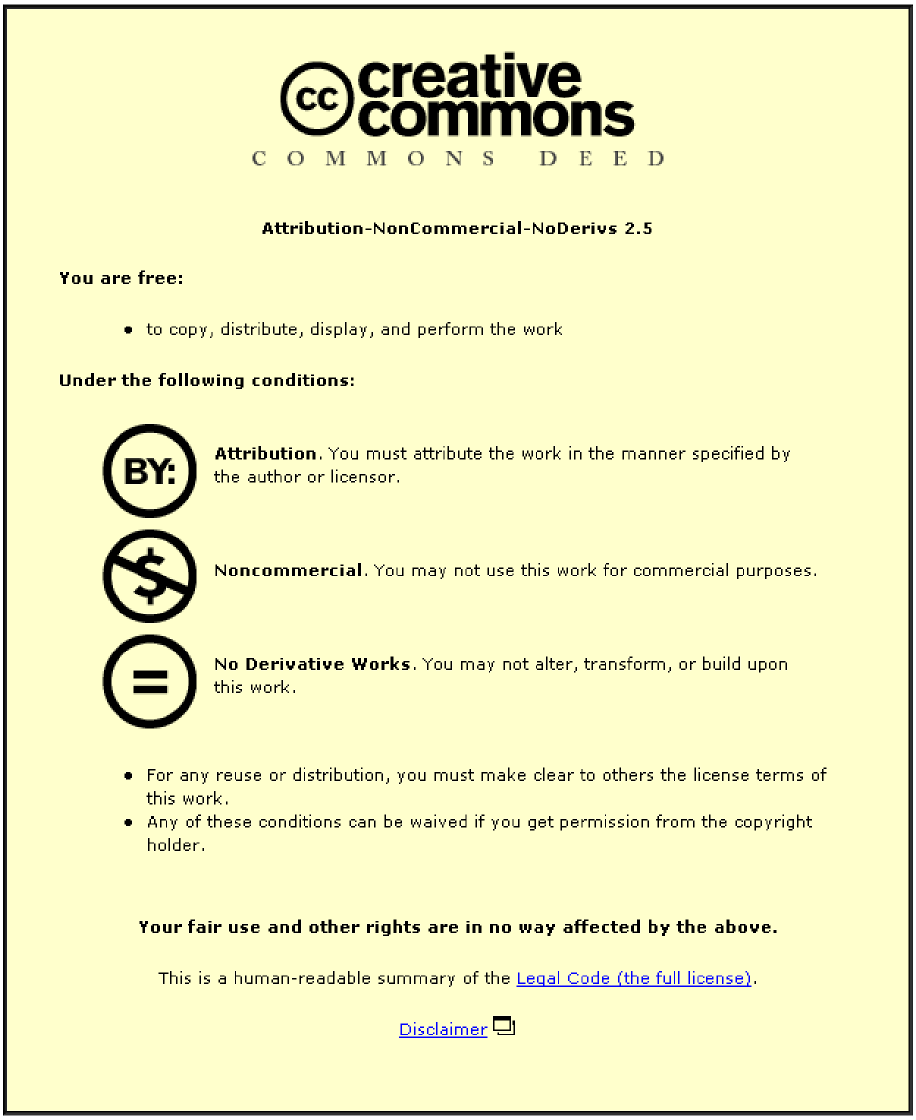

Holocene environmental change in coastal Denmark: interactions between land, sea and society

by

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Abstract

In this study a multiproxy approach (including sedimentary techniques, diatoms, molluscs, foraminifera, sedimentary pigments, isotopes, pollen and plant macrofossils) has been adopted to assess environmental change over the last ~9,000 years at three Danish coastal sites (Kilen, Norsminde Fjord and Korup Sø). Particular focus has been placed on periods of intense human coastal occupation, identifiable in Denmark's rich coastal archaeological record (i.e. shell midden accumulation periods), to test critically, hypotheses that changes in the marine environment were contemporary with major cultural and societal changes over the last ~9,000 years. For example, it has been proposed that a decrease in salinity was responsible for the widespread oyster decline, apparent in the Neolithic layers of a number of Danish shell middens. This hypothesis, however, remains speculative to date, lacking any high-resolution and quantitative salinity data covering the Mesolithic-Neolithic transition. Inside the agricultural era, two more phases of shell midden accumulation occur (i.e. during the Pitted Ware/Single Grave cultural period and the Iron Age), suggesting that people must have returned to the sea at these times for increased exploitation of its resources. A diatom-based salinity transfer function (WAPLS-C3 model, $r^2_{boot} = 0.923$, RMSEP= 0.36 square root salinity units) based on a trans-Baltic training set has been applied to fossil diatom datasets from each site for quantitative assessment of salinity change over the study period. The multiproxy results presented in this study demonstrate a close connection between environmental change and human exploitation of marine resources over the Holocene. This relationship, however, is complex, with the individual fjord systems often exhibiting spatially different responses (i.e. variations in the sedimentary regime, salinity, productivity and nutrient status) to changes in key forcing mechanisms such as sea level change, climate change and human impact upon the catchment (following the introduction of agriculture). Environmental hypotheses for cultural change are reviewed on the basis of the evidence presented in this study. Diatom-environmental relationships have also been modelled (using multivariate techniques) at Korup Sø and Norsminde Fjord using proxy data as 'predictor' variables for changes in the terrestrial and marine environment. These results suggest that a variety of marine, climatic, human and catchment related processes are important in explaining a proportion of the variation in the fossil diatom datasets, but these influences tend to vary temporally throughout the profile (e.g. human impact becomes important after ~3,900 BC).

Key words: coastal, Denmark, Holocene, diatoms, sedimentary pigments, isotopes, palaeosalinity, productivity, environmental change, archaeology.

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Chapter 1. Introduction, background and aims

1.1 Introduction

Paramount to our understanding and management of highly productive and commercially important biomes such as the coastal zone, is a firm grasp of internal dynamics (e.g. the impacts of changes in salinity, nutrient, productivity and temperature upon ecological communities) within local and regional systems, both at present and in the past. Coastal shelves currently account for 20 % of all marine total primary productivity despite only making up 10 % of the total area of the ocean floor (Scourse *et al.*, 2004). However, ecosystem functioning and the impacts of climate change (e.g. sea level change, saltwater intrusion) and human impact (e.g. nutrient enrichment, dredging, overfishing) upon the various different coastal habitats (e.g. estuaries, coral reefs, tropical mangroves, lagoons) remain poorly understood. Estuaries are extremely common shelf habitats, being found on all continents and at all latitudes across the world. An estuary is a partially enclosed body of water formed where freshwater from rivers or streams flows into the oceans, mixing with the seawater (McLusky and Elliot, 2004). They can originate in a variety of ways (e.g. drowned glacial troughs, drowned river valleys) and are often distinguished by the degree of mixing between the fresh and salt water masses (e.g. salt-wedge estuaries; Mississippi River mouth (USA), well mixed tidal estuaries; Severn Estuary (UK), Delaware Bay (USA), partially mixed estuaries; San Francisco Bay (USA), Mersey River, (UK)), though these states can alter throughout the year.

Estuaries tend to be dynamic habitats characterised by continuously fluctuating temperatures, nutrient concentrations and salinities, steep horizontal and vertical salinity gradients, as well as being subject to variations in sedimentation patterns and rates of river input. They are ecologically stressful habitats and therefore only species which can tolerate such variable conditions tend to reach high abundances (Yonge, 1960; Barnes, 1974; Wiley, 1976; Day *et al.*, 1989; Attrillf and Rundle, 2002). Estuaries also have immense societal importance both at present and in the past. Today many of the largest metropolitan areas border estuarine areas (e.g. London, New York, Tokyo, Shanghai) and in the past some of the earliest civilisations developed along them (e.g. coastal occupation extends to as early as ~5,500 BC in the Yangtze Delta and Nile Delta; Stanley and Warne, 1993; Daniel and Chen, 1996; Stanley and Chen, 1996; Stanley and Warne, 1997; Zhang *et al.*, 2005). Their very high primary and secondary productivity helps supports commercial fisheries, with a large number of commercial fish species dependent on estuaries for at least part of their life cycle (Claridge *et al.*, 1986; MacKenzie *et al.*, 2002). In heavily human impacted systems, eutrophication, pollution and over-fishing may also have serious detrimental effects on coastal ecosystems (Cooper and Brush, 1991; Jørgensen and Richardson, 1996; Blaber *et al.*, 2000; Jackson *et al.*, 2001; Conley *et al.*, 2002b; Md. Islam and Tanaka, 2004; Scheffera *et al.*, 2005; Clarke *et al.*, 2006; Lehtonen and Schiedek, 2006; Smith, 2006; Andersen and Conley, 2009).

In glaciated regions, fjords and fjards are the most common types of estuary originating from glacially scoured troughs being drowned by the sea. Fjords tend to be U-shaped over-deepened basins, whereas fjards usually form in extensive continental shelf regions and tend to be shallower, avoiding over-deepening due to isostatic rebound of the continent. Both are particularly characteristic features along Scandinavian coastlines, due to former ice sheets having completely covered the entire Baltic region in the past (Petersen, 1985; Mangerud *et al.*, 1996). For example, in Denmark alone, over 80 estuaries and fjords exist (Conley *et al.*, 2000). The east coast of Denmark and the inner Danish coastal waters are both particularly interesting and important as these systems open out into largely enclosed micro-tidal (tidal range <2m; Davies, 1964, Danish tides as low 20-30 cm) brackish seas (e.g. Kattegat, Limfjord and western Baltic). The salinity of these seas is very closely connected to the degree of connection with North Sea (strongly related to sea level) and the climate over the Baltic catchment area (Svansson, 1975; Rodhe, 1992; Kristensen *et al.*, 1995; Christensen *et al.*, 2004). There is some evidence to suggest that periods of salinity maxima might coincide with high sea level stands (Iversen, 1937; Mikkelsen, 1949), but this remains critically untested (discussed in more detail below).

The ability of estuaries to record both environmental change from both the land and the sea, make them extremely important archives of palaeoenvironmental change. They are natural sediment traps that often enable good temporal resolution to be obtained due to high sedimentation rates, that have been known to exceed 10 m kyr^{-1} (Jennings and Weiner, 1996; Maslin *et al.*, 1998; Maslin and Swann, 2006). The presence of terrestrial macrofossils in estuarine sediments also enables radiocarbon dating of terrestrial material, thereby avoiding marine reservoir effect problems often encountered when dating marine material (e.g. foraminifera, molluscs) from coastal sediments or deep sea cores (Bard *et al.*, 1987; Bard *et al.*, 1991; Lowe and Walker, 2000; Larsen *et al.*, 2002; Ascough *et al.*, 2005). Such systems also support dense human populations both at present and in the past, offering the rare opportunity to assess directly how marine changes affected human communities living at or near coastal sites.

The extremely rich Danish archaeological record indicates intense coastal/estuarine habitation and heavy dependence on marine resources by cultures and societies living in Denmark during the Holocene (Rasmussen, 1968; Andersen, 1989; Enghoff, 1989; Andersen and Rasmussen, 1991; Holm, 1991; Andersen, 1995a, 2000; Holm and Bager, 2002; Andersen, 2007; Andersen, 2008a). The presence and availability of marine resources tend to be sensitively regulated by environmental parameters such as salinity, sea-level, sedimentary regime, nutrient status, primary productivity, temperature and oxygen concentrations (e.g. Jensen and Spärck, 1934; Rasmussen, 1958; Yonge, 1960; Holm and Bager, 2002; Bailey and Milner, 2008). This has subsequently led to a number of environmental hypotheses suggesting that changes in the archaeological record (see below) might have been forced by changes in the natural marine environment (e.g. Rowley-Conwy, 1984). A clear understanding of the impacts that sea level and climate change have had upon past Danish estuarine systems (e.g. salinity, nutrient status, ecosystem structure, marine resources) is particularly important for predicting how projected future changes in sea level, climate and nutrient status (e.g. Solomon *et al.*, 2007; HELCOM; BACC, 2006) might affect marine resources and influence people living in coastal regions both today and in future. Understanding the internal dynamic of such systems is also paramount to our understanding of the global nutrient and biogeochemical cycling of major elements (e.g. carbon, nitrogen, iron and silica).

This study uses high resolution, palaeoenvironmental techniques (i.e. diatoms, sedimentary pigments and isotopes) in order to reconstruct changes in salinity, productivity, nutrients and sedimentary conditions over key archaeological events to critically assess environmental hypotheses for cultural change (e.g. decline in oysters, Rowley-Conwy, 1984; see Section 1.6). This chapter introduces the major research aims of the current study and provides background information for, and current understanding of the past and present, Baltic Sea, Kattegat and Danish coastal waters. The history of human exploitation of marine resources evident in the Danish archaeological record is also summarised. Specifically this chapter aims to address the following issues/questions:

1. What is the present day oceanographic system (e.g. currents, salinity gradients) operating in the transitional waters (i.e. Skagerrak, Kattegat, Belt Sea, Øresund and the

adjacent inner Danish coastal waters) between the North Sea and Baltic Sea and how did the present oceanographic systems evolve?

2. How has the marine environment of Baltic Sea, Kattegat and inner Danish waters changed over the Holocene, particularly in response to changing sea level and climate?
3. What factors influence sea level, salinity and productivity in the modern day system and how have palaeoecological techniques been used to reconstruct long term changes in these parameters?
4. How well are Holocene changes in these environmental parameters currently understood for the Danish Baltic region and what are the key outstanding issues/ uncertainties?
5. How has human resource exploitation changed/evolved over the Holocene as detailed by the Danish archaeological record?
6. What environmental hypotheses have been proposed to explain some of the cultural and societal changes identifiable in the archaeological records?
7. What are the specific aims of the present study and what techniques have been chosen (and why) in order to attempt to answer these research questions?

1.2 The modern Baltic Sea, Kattegat and inner Danish coastal waters

Today circulation in the Kattegat and Skagerrak is mainly driven by the input of cold saline water (30-35‰) from the North Sea via the North Atlantic, Central North Sea and North Jutland currents (NJC), counteracted by intermittent inputs of cold brackish water (8-12 ‰) from the Baltic Sea via Baltic currents (Svansson, 1975; Rodhe, 1987; Figure 1.1). The NJC is formed at the North western tip of Jutland, where the South Jutland Current (SJC) meets central North Sea and Atlantic water masses (e.g. the dominant South Trench current (STC) and the lesser Tampen Bank current (TBC)). The SJC is a product of English Channel water masses mixing with southern North Sea water and flows northwards along the sand-dominated coast of west Jutland, amassing high concentrations of suspended particles (Gyllencreutz *et al.*, 2005). The combined NJC then flows into the Skagerrak along the northern coast of Jutland until it mixes with Baltic surface currents at the entrance to the Kattegat and flows off in a northerly direction, along the Swedish and Norwegian coasts becoming the Norwegian coastal current (NCC). Further mixture with the north Atlantic currents forms an anticlockwise gyre in the deep northeastern region of the Skagerrak and results in reduced velocity of currents and sedimentation of fine grained materials at rates of up to 1 cm yr^{-1} (Gyllencreutz *et al.*, 2005; Gyllencreutz *et al.*, 2006). Part of the NJC current also enters the Kattegat at depth to counteract the lower salinity Baltic water (or Baltic current) flowing out from the Baltic Sea at the surface through the Danish straits (i.e. the three channels – Store Bælt, Lille Bælt and Øresund, connecting the Baltic Sea to the North Sea through the Kattegat and Skagerrak). This outflow produces a north-south salinity gradient ranging from approximately 30 ‰ in the north to 15 ‰ in the south (Falt, 1982; Jiang *et al.*, 1998; Figure 1.1).

Reconstruction of palaeo-currrent patterns in the Skagerrak and Kattegat indicate that this modern circulation system began approximately 6,550 BC (all years are given as calendar years unless otherwise stated) following a series of hydrographic shifts associated with the opening of the English channel and the Danish straits (Gyllencreutz, 2005; Gyllencreutz *et al.*, 2005; see below). The events leading up to the establishment of the modern circulation pattern and the subsequent evolution of current strength and modern day sedimentation regime are summarised in Figure 1.2 and Table 1.1. This stratigraphy is based on chirp sonar data, mineral magnetic and grain size analyses from the high resolution MD99-2286 core (Figure 1.1) collected from the north-eastern Skagerrak and dated via 27 AMS ^{14}C dates on mollusc shells or foraminifera tests (with marine reservoir effect correction applied; Gyllencreutz, 2005; Gyllencreutz *et al.*, 2005; Gyllencreutz *et al.*, 2006; Gyllencreutz and Kissel, 2006).

Whilst these major events appear to be correlative across the North Sea, Skagerrak and Kattegat, precise dating has proved somewhat problematic. For example, the opening of the English Channel has been previously placed somewhere between 7,050-5,750 BC (Jelgersma, 1979; Conradsen and Heier-Nielsen, 1995; Lambeck, 1995; Jiang *et al.*, 1997) and the Danish straits between 7,050-6,250 BC (Björck, 1995; Conradsen, 1995; Lambeck, 1999).

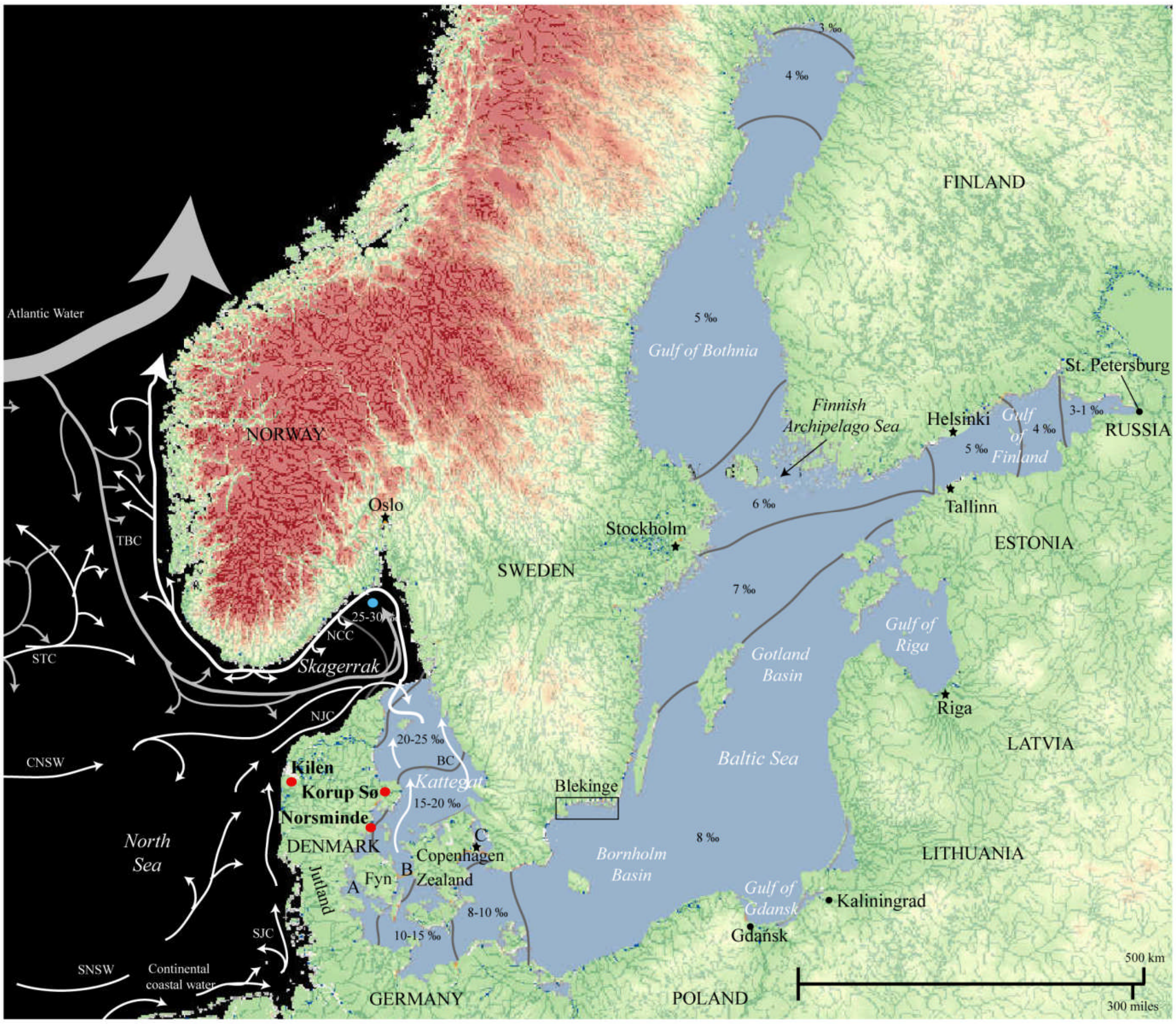


Figure 1.1 - Map of the Baltic Sea, with regional modern day average surface salinities and present current circulation system for the North Sea, Skagerrak and Kattegat (after Gyllencreutz *et al.*, 2005 and Longva and Thorsnes, 1997). Sites analysed in this thesis are indicated by red dots. The blue dot signifies the location of core MD99-2286 collected from the Skagerrak (Gyllencreutz *et al.*, 2005). Abbreviations: A = Lillebælt, B = Storebælt, C = Øresund, CNSW = central North Sea water, AW = Atlantic water, NCC = Norwegian coastal current, TBC = Tampen Bank current, SJC = south Jutland current, NJC = north Jutland current, BW = Baltic water, BC = Baltic current, CNSW = central North Sea water, SNSW = southern North Sea water. Map downloaded from HELCOM: http://maps.helcom.fi/website/mapservice/index.html

Marine Event	References	^{14}C age BP	Age cal. BP	Age cal. BC
Transgression of the southern North Sea	Stabell and Thiede, 1986; Lambeck, 1995	10,000	11,600-11,300	9,650-9,350
End of IRD dep. in Skagerrak	van Weering, 1982b, 1982a	10,000	11,600-11,300	9,650-9,350
Aker IMZ, Oslo area	Gjessing and Spjeldnaes, 1979; Sørensen, 1979; Gjessing, 1980; Andersen *et al.*, 1995	9,800-9,600	11,300-10,800	9,650-8,850
Marine limit, Oslo area	Hafsten, 1983	9,700	11,200-10,800	9,250-8,850
Glomma drainage event	Longva and Bakkejord, 1990; Longva and Thoresen, 1991	9,100	10,400-10,200	8,450-8,250
Otteid-Stenselva strait closing	Björck, 1995; Lambeck, 1999	9,100	10,400-10,200	8,450-8,250
English Channel opening	Nordberg, 1991	8,000	9,000-8,700	7,050-6,750
	Conradsen and Heier-Nielsen, 1995	7,600	8,500	6,550
	Jiang *et al.*, 1997	7,700	8,600	6,650
	Björklund *et al.*, 1985	8,000-7,000	9,000-7,700	7,050-5,750
	Jelgersma, 1979; Lambeck, 1995	8,000-7,000	9,000-7,700	7,050-5,750
Danish straits opening	Björck, 1995	8,200	9,300-9,000	7,350-7,050
	Conradsen, 1995	8,000	9,000-8,700	7,050-6,750
	Lambeck, 1999	7,500-7,800	8,700-8,200	6,750-6,250
Isolation of Dogger Bank	Lambeck, 1995	8,000	9,000-8,700	7,050-6,750
Skagerrak–Kattegat hydrographic shift	Conradsen, 1995; Conradsen and Heier-Nielsen, 1995	5,500	6,200	4,250
	Jiang *et al.*, 1997	5,100	5,900	3,950
	Nordberg and Bergsten, 1988; Nordberg, 1991	4,000	4,600-4,300	2,650-4,350

Table 1.1 – Summary of the major events leading up to the development of the modern day circulation and sedimentation system in the North Sea, Skagerrak and Kattegat (Modified from Gyllencreutz, 2005). Abbreviations: IRD = ice rafted debris, IMZ – ice marginal zone.

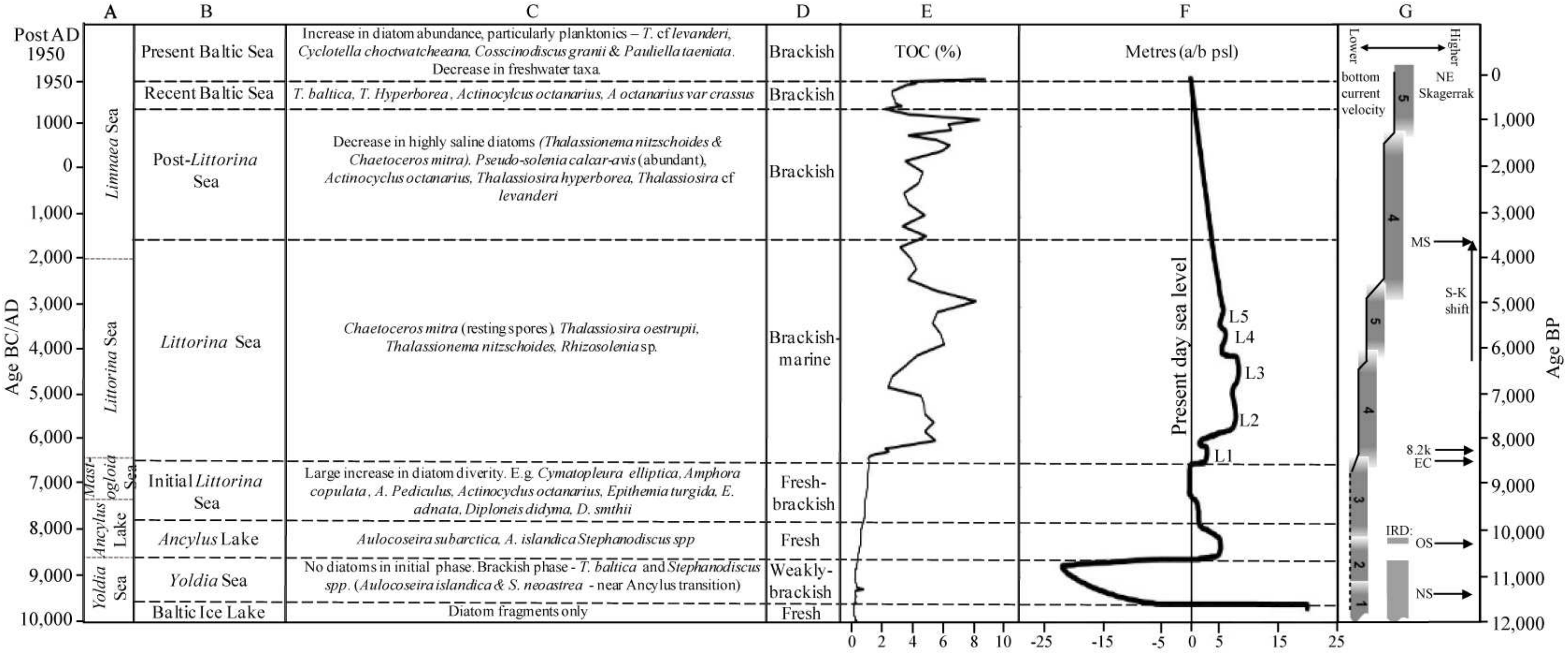


Figure 1.2 - Summary of Baltic Sea level, salinity and total organic carbon (TOC) history. A= Traditional division of the Baltic Sea's development stages (cf. Hyvärinen, 1988). B= Current Baltic Sea stage divisions (after Andrén *et al.*, 2000; Berglund *et al.*, 2005; Andrén *et al.*, 2007b). C= Representative diatom assemblages from open waters (after Andrén *et al.*, 2000). T. = *Thalassiosira* (column C). D= Salinity stages of the Baltic Sea's history. E= Total organic carbon (TOC) curve from the Gotland Basin (Andrén *et al.*, 2000). F= Blekinge Sea level curve based on shore displacement (from Blekinge SE Sweden, Berglund *et al.*, 2005; Andrén *et al.*, 2007b); L1-L5 refer to *Littorina* transgressions as recorded in coastal sediments from the Blekinge region of south-eastern Sweden (after Berglund *et al.*, 2005). G. Summary of major marine event and main sediment source and transport pathways into the Skagerrak (core MD99-2288) over the last 12,000 years (after Gyllencreutz and Kissel, 2006). 1. Glacial marine sediment from meltwater discharge (source: southern Norway and SW Sweden). 2. Glacial marine sediment from meltwater outflow (source: Baltic Sea via the Vanern basin). 3. Marine sediment (source: eastern North Sea via the North Jutland Current). 4. Marine sediment (source: chiefly from the southern North Sea and the Atlantic Ocean, transported by the North and South Jutland current, but also from the Baltic Sea and reworked coastal sediments in Sweden and Norway transported by the Baltic Current and currents along the coasts of western Sweden and southern Norway). 5. Marine sediment (source: chiefly from the southern North Sea and the Atlantic Ocean, transported by the North and South Jutland). IRD = Ice rafted debris. NS = opening of Närke Strait, OS = closing of Otteid Stenselva Strait. EC = opening of the English channel (along with increased flow of Atlantic water, opening of the Danish straits and beginning of the modern circulation system). 8.2k = large scale cooling event recorded in the Greenland ice cores (Alley *et al.*, 1997; Rohling and Pälike, 2005). S-K shift = Skagerrak-Kattegat hydrographic shift, MS = beginning of modern sedimentation in the Skagerrak. Bottom current velocity is based on median sortable silt in the MD99-2288 core with dotted line referring to less reliable inference due to low volumes of silt (Gyllencreutz and Kissel, 2006). * (a/b psl) = above/below present sea level.

1.3 Sea Level and Salinity change during the *Littorina* Sea Stage

In semi-enclosed seas such as the Baltic Sea, Kattegat and Limfjord, fluctuations in sea level and salinity are common on both short and long timescales, and have important propagative impacts upon other environmental variables such as nutrient cycling and marine productivity (Bianchi *et al.*, 2000; Zillén *et al.*, 2008). They also have clear implications for humans living in the coastal environment and exploiting marine resources both at present (MacKenzie *et al.*, 2002; BACC, 2006) and in the past (Rowley-Conwy, 1984; Enghoff, 1999; Holm and Bager, 2002; Andersen, 2007; Enghoff *et al.*, 2007). For example, successful cod spawning in the Baltic Sea requires regular renewal of oxygenated, saline water from the North Sea, while prolonged periods of oxygen depletion in the deep waters inhibit the successful development of cod eggs (Nissling and Westin, 1997; MacKenzie *et al.*, 2002; Köster *et al.*, 2005). This deep-water hydrographic exchange is driven by sea level and atmospheric pressure differences between the southern Kattegat and western Baltic Sea. Surface salinity, and perhaps to some extent sea level, is further affected by freshwater inputs from the large catchment draining into the Baltic Sea. This section considers how salinity, sea level and hydrography have changed since the beginning of the *Littorina* Sea stage (~7,850 BC-present; Figure 1.2), and how such changes have affected the hydrographic and ecological systems of the Baltic Sea, Kattegat and inner Danish coastal waters (e.g. Limfjord).

1.3.1 The *Ancylus* Lake-*Littorina* Sea transition (7,850-6,550 BC)

Prior to the *Littorina* Sea stage, the Baltic Sea underwent a series of freshwater and brackish phases (Figure 1.2) caused by complex interplays between ice retreat, global sea level rise and isostatic land adjustment (see Björck, 1995; Andrén *et al.*, 2000; Björck, 2008 and references therein for a full review). Marine waters reached the Skagen area of northern Jutland by 13,050 BC (Knudsen, 1994) and inundated the Kattegat between 11,550-11,050 BC (Petersen, 1985), but the southern Baltic Sea remained freshwater, isolated via a land bridge connecting eastern Jutland to south west Sweden. Brackish-marine conditions in the Baltic Sea were not established until between 6,550-6,150 BC (Andrén *et al.*, 2000; Bennike *et al.*, 2004; Björck, 2008). However, the earliest pulse of saline water through the Great Belt is now believed to have occurred as early as 7,850 BC (Andrén *et al.*, 2000). This, probably combined with several other intermittent intrusions of salt water, resulted in a very weakly brackish phase, recently termed the Early *Littorina* Sea Stage from 7,850-6,550 BC (Berglund *et al.*, 2005; Björck, 2008) or *Mastogloia* Sea Stage (Sundelin, 1922). The impact of these intermittent intrusions probably only affected the most southern and western regions of the Baltic Sea with only very minor changes in the diatom flora being detectable in the Gotland Basin (Andrén *et al.*, 2000) and southern Baltic coast (Saarse *et al.*, 2009; Witkowski *et al.*, 2009).

The major post-glacial marine transgression beginning at 6,550 BC and complete by 6,150 BC, often termed the Initial *Littorina* transgression, is documented across the entire Baltic Sea, Kattegat and Limfjord (Mörner, 1976; Petersen, 1981; Winn *et al.*, 1988; Petersen, 1993; Björck, 1995; Winn *et al.*, 1998; Andrén *et al.*, 2000; Hoffman and Winn, 2000; Bennike *et al.*, 2004; Berglund *et al.*, 2005; Miettinen *et al.*, 2007; Rasmussen, submitted), and was caused by global sea level rise associated with the final stages of deglaciation of the northern hemisphere ice sheets (Berglund *et al.*, 2005). This eventually drowned the Øresund and Danish Great Bælt, creating a permanent connection between the Skagerrak/Kattegat and Baltic Sea, thus beginning the *Littorina* Sea Stage as we know it today. Also associated with this global sea level rise was the opening of the English Channel (~7,050-5,750 BC; Nordberg, 1991; Conradsen and Heier-Nielsen, 1995; Lambeck, 1995; Table 1.1), the initiation of the Jutland current and the beginning of the modern Skagerrak-Kattegat current circulation system (see Figure 1.1) (Svansson, 1975; Rodhe, 1987; Otto *et al.*, 1990; Rodhe, 1996; Jiang *et al.*, 1998; Gyllencreutz and Kissel, 2006). This transgression also drowned low lying areas of northern Jutland connecting the Limfjord to the Skagerrak, though the exact date for this inundation remains uncertain (Petersen, 1981). It is likely that the Limfjord remained detached from the North Sea to the west, due to the existence of an emergent deposit ~15 km off the current coastline, known as the Jutland Bank (e.g. Leth, 1996). It is believed the Jutland Bank was eventually drowned in the mid-Holocene (~4,250 BC; Leth, 1996) as sea levels continued to rise in the North Sea (cf. Behre, 2007).

The *Littorina* transgression caused the salinity of the southern Baltic to rise up to levels similar to present day (10 ‰), and the regular input of nutrient-rich sea water at depth from the Kattegat created a vertical halocline and initiated cyclic blooms of nitrogen-fixing cyanobacteria (Bianchi *et al.*, 2000; Borgendahl and Westman, 2007). North Sea water entering the Kattegat and Baltic Sea at depth is highly saline, nutrient-rich and oxygenated, and mixes poorly with the overlying freshwater surface inflow from the catchment (mainly via river runoff and glacial meltwater). The depth of this halocline is both spatially and temporarily variable, being dependent upon weather and climatic factors (e.g. precipitation, atmospheric pressure) that regulates the amount and periodicity of saline water input into the Baltic Sea basin (Gustafsson and Andersson, 2001; BACC, 2006).

A minor regressive phase which has been dated to approximately 6,250-6,150 BC in the Blekinge records southeastern Sweden appears to have interrupted the *Littorina* transgression (Berglund et al. 2005). This is broadly correlative with the widespread "8.2 k" (~6,250 BC) cooling event (Alley *et al.*, 1997; Rohling and Pälike, 2005) and has been linked to a possible deterioration in weather conditions and alterations to the atmospheric pressure systems (e.g. North Atlantic Oscillation (NAO), Hurrell, 1995) operating over northern Europe (Berglund *et al.*, 2005). For the recent past, it has been demonstrated that sea level variations in the Baltic Sea and Kattegat are closely connected with predominant wind patterns operating over the Baltic Sea area that are in turn governed by atmospheric pressure cells (Hanninen *et al.*, 2000; Andersson, 2002; Hunicke and Zorita, 2006; Hunicke *et al.*, 2008). For example, stronger winter westerly winds associated with a positive NAO can cause a rise in sea level in some areas of the Baltic Sea (Andersson, 2002; Woolf *et al.*, 2003; Jevrejeva *et al.*, 2005) by piling up sea water. It is tenable that atmospheric systems will have impacted Baltic Sea level over much longer timescales and that a global climate anomaly such as the 8.2 k event will have caused disruptions to various regional atmospheric systems. The exact causes of this apparent sea level regression, however, remain largely unknown and un-investigated.

1.3.2 The *Littorina* Sea (~6,550 BC – present)

Sea level during the *Littorina* Sea stage was somewhat dynamic, with the western and southerly areas of the Baltic exhibiting very different sea level curves to those in the northern Baltic Sea. Sites south of the Stockholm-south Finland region and in the Kattegat and Danish/west Swedish coastal waters document a series of step-like sea level transgressions occurring up until ~3,050 BC, with maximum Holocene sea level being reached between 4,750-4,150 BC during the late Atlantic period (Iversen, 1937; Troels-Smith, 1942; Mikkelsen, 1949; Berglund *et al.*, 2005; Björck, 2008). Coincidental with highest sea levels, salinity is believed to have reached its Holocene maximum, as suggested by diatom and molluscan assemblages, isotopic analyses (Mikkelsen, 1949; Punning *et al.*, 1988; Winn *et al.*, 1988; Petersen, 1993; Winn *et al.*, 1998; Andrén *et al.*, 2000) and in natural shell fossil beds where high-salinity demanding species such as oysters (*Ostrea edulis*, requiring a salinity of at least 23-25 ‰; Jensen and Spärck, 1934; Yonge, 1960) penetrated much further south than their present day limits (Nielsen, 1938).

What caused this series of sea level changes during the early *Littorina* Sea stage, ~6,550-3,050 BC (e.g. five transgressive phases in the Blekinge region; Berglund *et al.*, 2005, Figure 1.2) remains uncertain, but eustatic changes must be predominately responsible, as prior to the *Littorina* stage, regional isostatic rebound had practically ceased in southern Scandinavia (Mörner, 1979; Björck, 1995). Continued global eustatic sea level rise, caused by the melting of the Northern hemisphere and Antarctic ice sheets (Fairbanks, 1989; Peltier, 2002; Carlson *et al.*, 2007), most likely explains the transgressional peaks prior to ~4,000 BC. For example, the second *Littorina* transgression (L2 of Berglund *et al.*, 2005) is documented in the Blekinge area of southeastern Sweden by a complex two-stage rapid sea level rise between 5,850-5,350 BC (Berglund *et al.*, 2005; Yu *et al.*, 2007). Yu *et al.*, (2007) attributed the rapid flooding of six elevated basins (between 4.0-7.2 m above present sea level) in the region at 5,550 BC to rapid sea level rise, caused by the abrupt decay of the Labrador section of the Laurentide ice sheet (e.g. Carlson *et al.*, 2007). A similarly rapid sea level rise around this time is broadly correlative across the Baltic Sea (e.g. Bennike *et al.*, 2004) and Kattegat (e.g. Mörner, 1976; Christensen, 1995) as well as being evident in other parts of the world, supporting global sea level rise (Blanchon and Shaw, 1995; Siddall *et al.*, 2003).

After ~4,550 BC, global sea level rise slowed down (Fairbanks, 1989; Bard *et al.*, 1996). In the southern and western Baltic, sea level begins to decrease at 4,250 BC and the magnitude of subsequent transgressions appears to decrease before eventually ceasing by ~3,050 BC (Björck, 2008). Ingólfsson and Hjort, (1999) suggest that between ~5,050-3,050 BC global sea level rise is predominately of Antarctic origin. Alternatively, as indicated above, shorter term variations in atmospheric pressure systems and storminess over northern Europe can also have impacts upon sea level and salinity within the Baltic and Kattegat region (e.g. Hurrell, 1995; Hanninen *et al.*, 2000; Yu *et al.*, 2003) and may offer a more localised explanation for the later more minor variations in sea level.

In the northern Baltic (approximately just north of Stockholm in the west and south Finland in the east), only the initial *Littorina* transgression (L1; Berglund *et al.*, 2005) is evident, with none of the subsequent transgressions being recorded in sedimentary records from the region

(Björck, 1995; Jantunen and Donner, 1996; Miettinen *et al.*, 2007; Björck, 2008). In the more northerly areas of the Baltic Sea, the rate of isostatic readjustment (rise) remained high enough to either exceed or at least keep pace with eustatic sea level rise (Björck, 2008) thereby overriding the changes associated with global sea level rise identifiable in the southern and western Baltic Sea.

By ~3,050 BC, the whole Baltic Sea was in a state of regression as sea level gradually fell to that of present day (Andrén *et al.*, 2000; Berglund *et al.*, 2005). Whilst some evidence exists for several minor, and short lived sea level high stands occurring in the more recent past (e.g. Roman period and Viking Age transgressions; Miller, 1982; Miller and Hedin, 1988), none appear to be as large in magnitude as those from early *Littorina* Sea. Three further stages (Post *Littorina* Sea; ~1,750-550 BC, Recent Baltic Sea; ~550-1950 BC and Present Baltic Sea; AD 1950-present) have been identified in the Baltic Sea history by Andrén *et al.*, (2000), but these are related to changes in core sedimentology, chemistry or biological communities as a response to climate, oxygen levels and nutrient status rather than sea level.

1.4 Drivers of Salinity Change in the Baltic

The salinity of the Baltic Sea, Kattegat and inner Danish coastal waters is influenced by a wide range of interrelated factors including sea level, ocean currents, climate change, atmospheric pressure, and freshwater inputs from the surrounding catchment area. A surface salinity gradient exists both within the Baltic and Kattegat, driven by distance to the source of saline water input (Figure 1.1). In the Kattegat, highest salinities are found in the north (25-30 ‰) where it joins the Skagerrak, and salinity decreases in a southerly direction to 10-15 ‰ to the north of the Øresund and Danish Straits. On the eastern side of the straits, the south-western Baltic Sea has a present day salinity of ~10 ‰, which decreases in a northerly direction. The Gulf of Bothnia, the northernmost arm of the Baltic Sea, experiences almost freshwater conditions (1-2 ‰).

A vertical halocline also exists in the Kattegat and Baltic Sea, created by the input of highly saline, oxygenated (and nutrient rich) North Sea water entering the Kattegat and Baltic Sea at depth, and overlying freshwater sourced from the catchment via rivers and run-off entering at the surface. The depth of this halocline is governed by relative sea level and local climatic and hydrographic conditions (Kullenberg, 1981; Wulff *et al.*, 1990). The shallowness of the

Øresund and Storebælt sills create further complications, as they limit the amount of North Sea water that can enter the western Baltic Sea. Only through major inflow events will significant amounts of highly saline, oxygenated water enter, which is critical for the maintenance of suitable salinity levels to support the current ecosystem (Hanninen *et al.*, 2000; MacKenzie *et al.*, 2002; BACC, 2006). Unfortunately, this replenishment to the Baltic Sea deep water only occurs at irregular intervals, and therefore periods of lowered salinity and hypoxia (low oxygen conditions) occur between saline pulses (Conley *et al.*, 2002a; Feistel *et al.*, 2003; Zillén *et al.*, 2008). In recent decades, under fears that deep water replenishment of the Baltic Sea was becoming less frequent, much effort has been focussed in trying to understand fully what drives the inflows of saline-oxygenated water into the Baltic Sea at annual-decadal scales (Matthäus and Franck, 1992; Lehmann and Hinrichsen, 2000a, 2000b; Winsor *et al.*, 2001; Feistel *et al.*, 2003; Meier *et al.*, 2006; Matthäus *et al.*, 2008). Regional variations in atmospheric pressure systems are important seasonally, while the North Atlantic Oscillation (NAO) appears to have a key role on a sub-decadal scale (Hurrell, 1995; Hanninen *et al.*, 2000; Lehmann and Hinrichsen, 2000b; Feistel *et al.*, 2006; Matthäus *et al.*, 2008).

1.5 Palaeosalinity records from the Baltic Sea, Kattegat and coastal waters

The classic salinity model for the Kattegat and Baltic Sea is summarised in Figure 1.2. This is characterised by the establishment of fully marine conditions by ~6,550-6,050 BC (Björck, 2008), increasing salinity up until ~4,250 BC (Berglund *et al.*, 2005) where salinity reached maximum Holocene levels (Punning *et al.*, 1988), followed by a gradual decrease since the late Atlantic down to the present day level. This model mirrors sea level change, as studies have generally dealt with salinity indirectly as a response to changes in sea level during the *Littorina* Stage (Mikkelsen, 1949; Miller and Hedin, 1988; Petersen, 1993; Andrén *et al.*, 2000; Berglund *et al.*, 2005). However, as already indicated above, the complexity of salinity dynamics causes variability on very short timescales (e.g. both seasonal and decadal variation associated with atmospheric systems such as the NAO, (Hurrell, 1995; Hanninen *et al.*, 2000; Zorita and Laine, 2000) for which very little palaeoenvironmental data is available.

Studies that have more directly considered salinity have largely been qualitative (Petersen, 1993; Huckriede *et al.*, 1996; Sohlenius *et al.*, 1996; Westman and Sohlenius, 1999), with low temporal resolution, which often does not pick up short-term variation. Where marine shells or sediments have been used for dating, these studies also suffer from chronological uncertainty due to the marine reservoir effect (Heier-Nielsen *et al.*, 1995; Hedenström and Possnert, 2001; Olsen *et al.*, 2009). Only a handful of studies have attempted to quantify salinity change (Punning *et al.*, 1988; Winn *et al.*, 1988; Jiang *et al.*, 1998; Winn *et al.*, 1998; Emeis *et al.*, 2003; Ryves *et al.*, 2004; Widerlund and Andersson, 2006;Burman and Schmitz, 2005 A.L. Clarke *et al.* unpublished data). Some of these are discussed in more detail below, though it is stressed that the aim here is not to review all palaeosalinity reconstructions from the Baltic Sea and Kattegat region, but rather to highlight current knowledge, progress and limitations.

As quantitative reconstruction from biological assemblages require the creation of large, modern-day training sets (Birks, 1998), isotopic signals (including oxygen, carbon and strontium) from shells and tests of marine organisms or from bulk sediment have been more commonly used (Punning *et al.*, 1988; Winn *et al.*, 1988; Winn *et al.*, 1998). The pioneering Baltic Sea isotope work of Winn *et al.*, (1988) on the foraminifera *Ammonia beccarii*, and Punning *et al.*, 1988 on *Cerastoderma glaucum* and *Lymnaea baltica*, produced the first estimates of salinity change for the Holocene, particularly over the *Littorina* transgression, and shed new light on the establishment of marine conditions in the Baltic Sea. However, probably the most detailed of these isotope-based salinity studies (accompanied by diatom analysis and sedimentological expression of anoxia) comes from the central Baltic Sea using the ^{13}C isotopic ratios from sedimentary bulk organic carbon (Emeis *et al.*, 2003). This study contradicts previous salinity estimates (e.g. Punning *et al.*, 1988; Winn *et al.*, 1988; Widerlund and Andersson, 2006) by suggesting that salinity in the Baltic Sea never exceeded present day values even during highest *Littorina* sea levels (~6,550-4,250 BC; Berglund *et al.*, 2005). According to Emeis *et al.*, (2003), the period between 6,050 BC and 3,050 BC was subjected to pronounced changes in salinity before starting a downward trend at ~3,150 BC. For the late Holocene, Emeis *et al.*, (2003) suggest that the environmental conditions within the Baltic Sea were closely linked to the evolution of climate, with variations in salinity being closely connected to patterns of reconstructed summer temperatures and winter precipitation in the southern Scandinavia. For example, the expansion of Scandinavian glaciers associated with the Neoglacial cooling (since ~2,050 BC; Nesje and Dahl, 1994) represent wetter winter conditions in the Baltic (1,750-1,150 BC) and therefore greater freshwater input (from rivers and glacial meltwater during spring-summer ablation period) and lower salinities. A renewed salinity increase also coincides with the beginning of the Roman Warm Period temperature increase, and over the Little Ice Age, lower winter temperatures produce parallel step-like changes in salinity (Emeis *et al.*, 2003).

Despite the use of these novel geochemical techniques (e.g. isotopic analysis and diagenetic mineralogy; Huckriede *et al.*, 1996), much of our knowledge concerning palaeosalinity of the Baltic Sea and Kattegat is based on qualitative biological assemblage data, most notably molluscs and diatoms. Molluscs have been the subject of a particularly long history of research in the Baltic Sea (e.g. Lindström, 1886; Munthe, 1894, 1910 with some of the original sea stages being named after the molluscs found present in the characteristic sediments (e.g. the *Littorina* Sea stage after finds of common periwinkle *Littorina littorea*; Lindström, 1886). Detailed reviews of the molluscan histories of the Baltic Sea, Kattegat and Danish waters can be found in (Petersen, 1981, 1986, 1993; Petersen, 2004; Petersen *et al.*, 2005). Though here we do stress the importance of these studies, particularly to our early understanding of relative sea-level and salinity changes in the Baltic region, to date molluscan studies have been seriously limited in terms of temporal resolution and the quantification of fossil assemblages in terms of salinity values. As large amounts of sediment are often needed to obtain enough molluscs to produce reliable inferences, it is difficult to identify the more subtle shifts in salinity that might operate on decadal-centennial timescales (e.g. smaller-scale changes in freshwater inputs and atmospheric forcing, including patterns in the NAO). Furthermore, though palaeosalinity estimates have often been inferred from molluscan assemblage data (e.g. Munthe, 1894; Petersen, 1993; Kristensen *et al.*, 1995; Christensen *et al.*, 2004; Rasmussen *et al.*, 2007), salinity values are generally given as a range based upon minimum and maximum salinity tolerances for the species present (e.g. Sorgenfrei, 1958; Funder *et al.*, 2002; Petersen, 2004).

Diatom analysis has the potential to obtain palaeoenvironmental data at far higher resolution than molluscs as they require very small samples (~0.1 g wet weight; Battarbee, 1986; Battarbee *et al.,* 2001; Jones, 2007). Diatom analysis is a widely used tool in palaeosalinity reconstructions (Juggins, 1992; Jiang *et al.*, 1998; Reed, 1998; Parsons *et al.*, 1999; Ryves *et al.*, 2002; Ryves *et al.*, 2004; Wachnicka *et al.*, 2010), due to their sensitivity to changes in salinity, ubiquitous nature, global distribution, and a growing taxonomic and ecological literature in coastal systems (e.g. Snoeijs *et al.*, 1993-1998; Witkowski, 1994; Witkowski *et al.*, 2000). While poor preservation is an issue complicating diatom analysis in marine and coastal sedimentary environments, many coastal and pelagic sites with good preservation can be found in the Baltic and Kattegat area (especially in deep, anoxic basins, and during periods of lower salinity) (Thulin *et al.*, 1992; Westman and Sohlenius, 1999; Andrén *et al.*, 2000; Ryves *et al.*, 2004; Berglund *et al.*, 2005; Ellegaard *et al.*, 2006). As a result there is a rich history of diatom research from the Baltic Sea tracking salinity change since the retreat of the Weichselian ice sheet and subsequent formation of the Baltic Ice Lake and throughout the Holocene (e.g. Iversen, 1937; Mikkelsen, 1949; Andrén *et al.*, 2000; Berglund *et al.*, 2005). The majority of these studies have remained qualitative, with diatom assemblages being classified with respect to their salinity tolerances (e.g. Figure 1.2) based on halobian systems such as Kolbe, 1927; Hustedt, 1957; van der Werff and Huls, 1958-1974 and the Baltic Sea intercalibration guides (Snoeijs *et al.*, 1993-1998). Again diatom analysis has largely been used either together with sea level reconstructions (Mikkelsen, 1949; Berglund *et al.*, 2005) or to examine salinity in relation to sea level change (Andrén *et al.*, 2000).

The highest abundance of marine diatoms in the Baltic Sea and Kattegat again occurs during the late Atlantic prior to ~4,000 BC (Figure 1.2) and is believed to represent maximum Holocene salinity. Some diatom-studies have even tracked multiple transgressions throughout the early *Littorina* stage (7,850-3,050 BC) (e.g. Mikkelsen, 1949; Berglund *et al.*, 2005). However, only two studies to date, have attempted to quantify salinity change in the Baltic Sea, Kattegat or inner Danish coastal waters using a diatom-based transfer function. Jiang, (1996) developed a modern environment-diatom training set from 39 surface samples collected from the Kattegat and Skagerrak and found that the diatom distribution was heavily associated with summer surface salinity (explaining ~30 % of the variation in the diatom dataset). A subset of 19 sites was applied to a core (Skagen-3) collected from northern Jutland (Figure 1.3), which demonstrated that the salinity of the Kattegat and Skagerrak is primarily driven by the climate of the circum Baltic area but complicated by variations in local current patterns and hydrodynamics of the Baltic Sea. Whilst this study has provided important information concerning palaeocurrents in the Skagerrak-Kattegat region (e.g. Table 1.1), its geographic location makes it unrepresentative of salinity conditions and changes in the Baltic Sea, Kattegat and inner Danish coastal waters. Furthermore the training set is based on deep sea sites from the Skagerrak and Kattegat and therefore cannot be used for sedimentary reconstructions from lower salinity waters such as the Baltic Sea or in the coastal waters of Denmark due to low species representation in the model (particularly the absence of wide range of benthic diatoms common in fjord and estuarine environments).

The other quantitative diatom-salinity reconstruction is from the Limfjord, where Ryves *et al.*, 2004 applied a diatom-based salinity transfer function to three fossil diatom sequences from Vejlerne (Han Vejle, Glombàk and Lund Fjord; Figure 1.3) covering the last ~150 years. The diatom-salinity model was produced from 27 brackish lakes and 9 fjords across Jutland spanning a salinity range of 0.2-31.1 ‰. The fjord samples were selected from the larger MOLTEN data set (Clarke *et al.*, 2003; Clarke *et al.*, 2006; Andrén *et al.*, 2007a), added in order to extend the salinity gradient into fully marine conditions. Salinity was most important environmental variable uniquely explaining 6% the variation in the lake-fjord diatom dataset, with depth and total phosphorus accounting for 3.9% and 3.5% of the total variation respectively. The final diatom-salinity model was a two component weighted averaging-partial least squares (WAPLS) model with a relatively high predictive power (r^2_{jack} = 0.887, RMSEP = 0.246 log salinity ‰, under internal validation). In all cores, the species coverage present in the model dataset was very good (ranging between 83-100% for all samples) and on the whole salinity predictions appear to be reasonable and potentially accurate (accompanied by, and generally in agreement with qualitative multiproxy inferences from cladoceran and foraminiferal data; Amsinck *et al.*, 2003), highlighting the potential for using diatom-salinity models in Danish coastal environments. However, the model did appear to struggle at the high end of the salinity gradient (i.e. towards fully marine conditions) where low abundances and poor preservation of diatoms become more of a problem (e.g. r^2 = 0.33-0.74 (positive) relationship between salinity and diatom dissolution as measured by the diatom **F** index). The inclusion of both fjord and lake sites in the training set might also have weakened the model, dismissing potentially important natural differences between these two types of site which might also drive diatom distributions (e.g. stratification, natural seasonal variations in salinity, ice cover, silica cycling biological structure, presence of mussels in fjords but not in lakes).

The long-term palaeosalinity history of the Limfjord is also poorly understood due to the lack of records extending back further than the last ~2,500 years. The few longer-term studies that do exist suffer from poor chronological control (i.e. dating problems associated with marine reservoir effect and/or sediment reworking) and low temporal resolution (Petersen, 1981; Andersen, 1992; Heier-Nielsen, 1992). For example, Petersen, (1981) has estimated that the marine transgression of the Limfjord occurred between ~7,000-6,000 BC (from 14C dating of a peat layer and several marine molluscs from overlying layers at Vust, northern Limfjord), though the chronology of this study has been seriously questioned (e.g. Christensen, 2001b) and new, preliminary evidence also exists suggesting that the Limfjord was inundated earlier than ~7,000 BC (P. Rasmussen et al. unpublished data). Burman and Schmitz, 2005 attempted to quantitatively estimate salinity and temperature during the Mesolithic period using $\delta^{18}O$ and $\delta^{13}C$ analysis, performed on two Littorina littorea (common periwinkle) present in the Ertebølle *locus classicus* shell midden (Andersen and Johansen, 1986). Using end member estimates from modern *L. littorea* shells sampled from the Limfjord, Burman and Schmitz, 2005 suggest that salinity (31 ‰) and summer surface water temperature (22°C) in the central Limfjord region were higher than today in the late Mesolithic (5,100-4,100 BC), suggesting greater water exchange/connection with the North Sea/Skagerrak than at present. This is in good accord with estimates that the Limfjord was more exposed during the mid-Holocene (e.g. sea level 3 m higher than present day by ~5,000 BC; Petersen, 1981) with an opening into the Skagerrak existing in the north (e.g. Hellesen and Tuxen, 1988). Aside from these spot estimates, salinity change within the Limfjord remains uncertain and largely un-investigated up until the late pre-Roman period (~500 BC).

During the late Holocene (~2,500 years), several palaeoenvironmental studies have demonstrated that the Limfjord has been subject to large shifts in salinity associated with variable connection with the North Sea and Skagerrak (Kristensen *et al.*, 1995; Christensen *et al.*, 2004; Ryves *et al.*, 2004). For example, a shift to brackish conditions is believed to have occurred some time during the late pre-Roman period as the Limfjord became totally closed off from the North Sea (Kristensen *et al.*, 1995; Christensen *et al.*, 2004), probably due to lowering of sea level (e.g. Behre, 2007). A return to high salinity conditions occurred during the Germanic or Viking period, before the Limfjord was closed off again sometime between ~AD 1100-1200. With the exception of several marine breakthroughs during severe storms possibly associated with the Little Ice Age (Kristensen *et al.*, 1995), the Limfjord remained closed up until ~AD 1825, when it was breached for the final time through Agger Tange and has remained open since.

The palaeoenvironmental records are also supplemented by documentary evidence (e.g. Petersen, 1976; Kristensen *et al.*, 1995; Holm and Bager, 2002) and archaeofaunal analyses (Enghoff, 1999). For example, the much debated Snorres Saga suggests that the Danish King

Knud the Great returned through the Agger sound from an expedition to England in AD 1027, indicating that the west was open in the late Viking period (Petersen, 1976). An opening in the North is also likely to have existed in AD 1061, through which Norwegian King Harald Haardraade is thought to have escaped from a superior Danish fleet led by King Sven Estridsson (Petersen, 1976). Documentary fishing records from the Limfjord have also enabled detailed information concerning the herring industry in the Limfjord throughout historical times (post ~AD 1200; Rasmussen, 1968; Holm and Bager, 2002). The herring fishery prospered in the Limfjord following the switch to more brackish conditions after ~AD 1100-1200. On several occasions (between AD 1200-1825), the herring fishery was suppressed due to marine breaches following storm events (see above; Kristensen *et al.*, 1995; Holm and Bager, 2002), prior to its final decline in AD 1825 (Holm and Bager, 2002). Archaeofaunal records from Iron Age sites situated within the Limfjord have extended the fishing records back to the fifth century, when high salinity conditions supported large plaice and eel fisheries (Enghoff, 1999). Whilst the late Holocene history of the Limfjord (i.e. post ~AD 1200) is relatively well constrained (i.e. due to documentary evidence), unfortunately the longer term studies extending back to the pre-Roman period (i.e. Kristensen *et al.*, 1995; Ellegaard, 2000; Christensen *et al.*, 2004) are again subject to dating uncertainties, associated with the marine reservoir effect. Therefore, improved chronological control is needed in order to fully understanding the timings of earlier openings and closings of the Limfjord and to assess how climate and sea level change might have driven these changes (e.g. Christiansen *et al.*, 1990; Tanner, 1993; Behre, 2007).

1.6 Cultural change in Denmark

Reconstructing long-term environmental change in the Baltic Sea is important not only in terms of ecological and hydrographic variability, but also in relation to humans inhabiting the coastal zone. The historical and archaeological record along the Danish coast is especially rich, and attests the importance of marine resources to societies living in coastal regions of Denmark throughout the Holocene (Andersen, 1995a; Enghoff, 1995; Holm and Bager, 2002; Andersen, 2007; Enghoff *et al.*, 2007). Prehistoric cultures and societies (e.g. Kongemose, Ertebølle) were often heavily dependent upon the dynamic Baltic Sea, yet our knowledge of this complex interplay remains fragmentary. For example, changing sea levels throughout the *Littorina* Sea stage (e.g. Iversen, 1937; Berglund *et al.*, 2005 and see above) is likely to have forced changes in important environmental parameters such as substrate, sediment accumulation rate, salinity, temperature, productivity and nutrient status at both local (e.g. within individual fjords) and regional scales (e.g. the entire Baltic basin). This is likely to have disrupted ecosystem structure and functioning within coastal waters and potentially affected the presence and availability of important marine resources (e.g. Rowley-Conwy, 1984). Unfortunately, changes in these environmental parameters in response to sea level (and climate change or human impact) remain poorly understood to date.

The estuaries and the inner coastal waters of Denmark are very attractive settlements for a variety of reasons. Firstly, they are highly productive systems, supporting both high primary (e.g. macrophyte and algal communities) and secondary productivity (e.g. zooplankton, bacteria) which is extremely important for supporting good commercial fishing industries. Furthermore resources within estuaries are often easily accessible and tend to vary seasonally rather than spatially. For example, eel grass, oyster and mussel beds commonly reside in the inner fjords and estuaries as they offer penetration of solar energy required for them to grow as well protection from damaging ultraviolet light. Inflowing rivers and streams also offer settlers a source of fresh water and alternative fresh water resources for a diverse diet. Estuaries can also offer protection from storms, be easily defensible, offer good trade routes both along rivers and fjords and provide easy access to the open seas. The attractiveness of such sites was clearly identified and exploited by communities living in Denmark throughout the Holocene (Andersen, 1995a; Enghoff, 1995; Enghoff, 1999; Andersen, 2007; Enghoff *et al.*, 2007). Archaeological evidence is diverse, ranging from single finds and small shell deposits through to larger sites and settlements such as burial mounds, graves and cemetries (Albrethsen and Brinch Petersen, 1977; E. Brinch Petersen, unpub.), submerged sites (Andersen, 1980, 1987) and large shell middens (Andersen, 2000, 2007; Andersen, 2008a). In the more recent past (last ~1,000 years) this is further supported by historical and documentary evidence (Holm and Bager, 2002; Poulsen *et al.*, 2007).

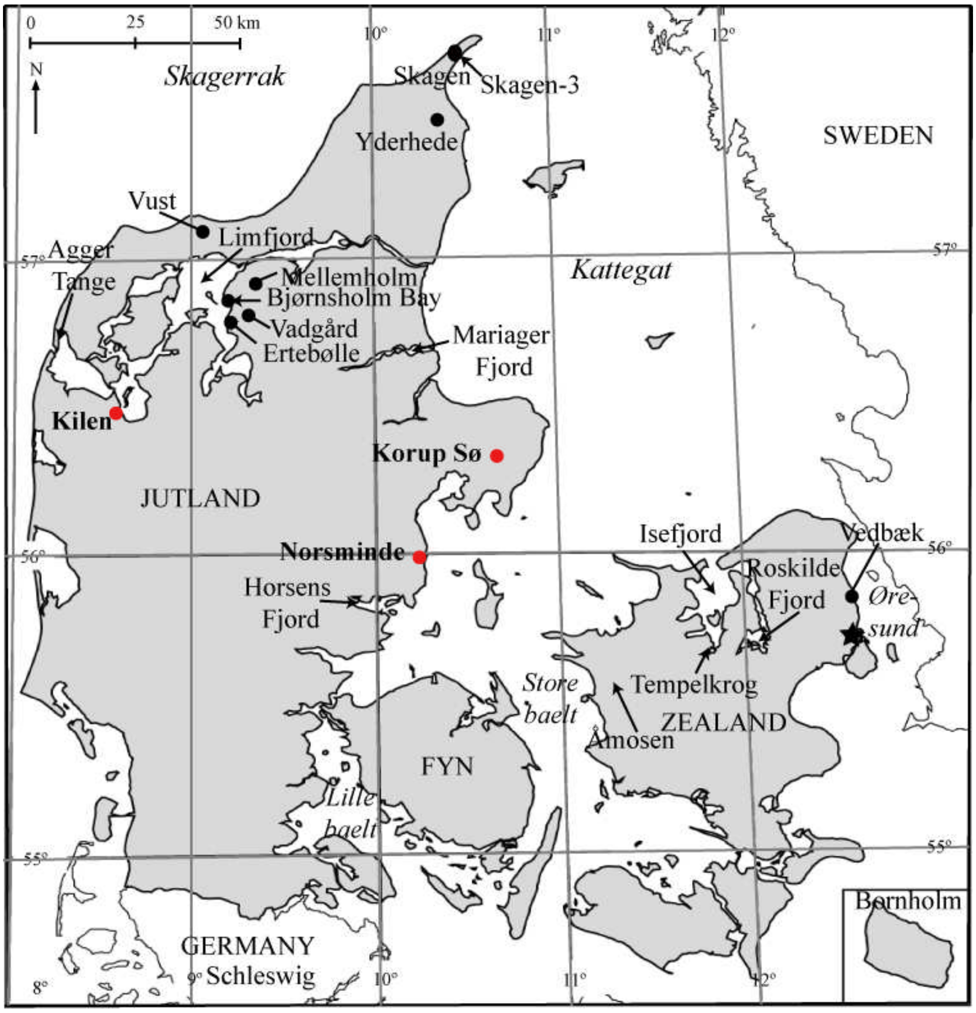


Figure 1.3 - Location map for Danish archaeological sites referenced in the text.

This study particularly focuses on reconstructing natural environmental change in coastal Denmark over the mid-Holocene with three important phases of shell midden ('køkkenmøddinger' in Danish, Figure 1.4) accumulation evident in the archaeological records:

1. Ertebølle-Funnel Beaker culture (~5,400-2,800 BC) with focus on the Mesolithic-Neolithic transition (~3,900 BC).

2. The period of the Pitted-Ware Culture and Single Grave Culture (~2,800-2,400 BC).

3. The Iron Age (500 BC-AD 1050).

Denmark is one of the classic areas for studies of prehistoric shell middens, a tradition that extends back into the early 19th century (e.g. Madsen *et al.*, 1900; Andersen, 2000; Fischer and Kristensen, 2002 and references therein). Danish shell middens are defined as a deposit that contains over 50 % shells or shell fragments and forms an undisturbed horizon at least 10 m^2 (Andersen, 2000, 2007). Shell bearing sites that fail to meet both of these criteria are simply referred to as shell deposits or shell heaps. Shell midden sites tend to be found within close ‘economic’ distance to natural shell banks and with good fishing access (Andersen, 2000). The shell middens are always found together with coastal settlement sites and were likely used for a wide range of daily activities (including processing, consuming and disposal of shell fish and other food). Over the last ~150 years a large number of shell middens (e.g. Andersen, 2000, 2007) have been discovered and detailed (large-scale) excavations have taken place in order to reconstruct changes in the practices and behaviour of past cultures and societies living in coastal areas of Denmark over the Holocene. The wealth of information now available from these shell middens (and other coastal sites) provides an excellent and rare opportunity to assess (using modern palaeoenvironmental techniques) how environmental change might have influenced (or even forced changes in) former coastal dwelling cultures and societies, which relied heavily upon marine resources.

1.6.1 The Late Mesolithic and the accumulation of Stone Age shell middens

The oldest known coastal sites in Denmark belong to the Maglemose culture (9,000-6,800 BC), though these are few in number and are largely confined to the Øresund (Andersen, 1995a; Pedersen *et al.*, 1997). Evidence for complex flint artefacts used to make arrow heads and various wooden fishing tools such as nets from pine bark and wooden spears indicate that the Maglemose culture heavily relied on fishing (e.g. pike, perch) and hunting (e.g. aurochs, elk, wild boar) to support their communities (Jensen, 2001; Schmölcke *et al.*, 2006). This culture mainly inhabited inland sites often positioning themselves on wind-sheltered plateaus in close proximity to rivers and lakes (Grøn, 1995), where they exploited fish from these freshwater sources. No coastal sites extending back into the earliest Mesolithic and late Palaeolithic have been found in Denmark to date, an absence most likely attributed to geographical or geological factors (i.e. below present day sea level having been drowned by the *Littorina* transgression; Christensen, 1995).

Far more numerous in Denmark are sites from the Kongemose culture (~6,800-5,400 BC, named after the type site ‘Kongemose’), which commonly inhabited coastal areas and showed far greater affinity to marine resources than the preceding Maglemose culture (e.g. Andersen, 1970a; Brinch Petersen, 1973; Andersen, 1985; Price, 1985; Larsson, 1990; Brinch Petersen, 1993; Enghoff *et al.*, 2007). Again, the Kongemose were skilled hunters and fishers that produced complex hunting tools such as arrowheads, scrapers, drills, awls and toothed blades from flint (e.g. Larsson, 1990; Vang Petersen, 1999). Their economy was based around the hunting of elk, aurochs, wild boar, red deer, roe deer, small fur animals as well as exploiting marine resources, with the earliest shell middens being dated to the late Kongemose period (Andersen, 1970a).

The Kongemose was succeeded by another hunter-fisher-gatherer culture named the Ertebølle, after excavation of the type-site ‘Ertebølle’ situated near Ertebølle, in the Limfjord region of northern Jutland (Madsen *et al.*, 1900, Figure 1.3). The Ertebølle people commonly lived along (or near) the coastline of fjords and estuaries (e.g. large number of Ertebølle shell middens; Andersen, 2000, 2007), relied heavily on a wide range marine resources (including fish, molluscs, marine birds and mammals), exhibiting a predominately marine diet (e.g. Tauber, 1981; Bratlund, 1991; Andersen, 1995a; Enghoff, 1995; Andersen, 2007; Enghoff *et al.*, 2007; Andersen, 2008a). They often amassed high population densities at coastal localities and again developed various complex tools in which to exploit these resources (Andersen, 1995a; Enghoff *et al.*, 2007). Whilst they appear to have adopted pottery from their close neighbours, the Ertebølle culture did not adopt agricultural practices until much later (i.e. introduction of agriculture in Denmark ~3,900 BC, but in northern Germany ~4,600 BC; Hartz *et al.*, 2002) and this ‘delay’ has sparked over 150 years of intense debate (Fischer and Kristensen, 2002 and papers therein).

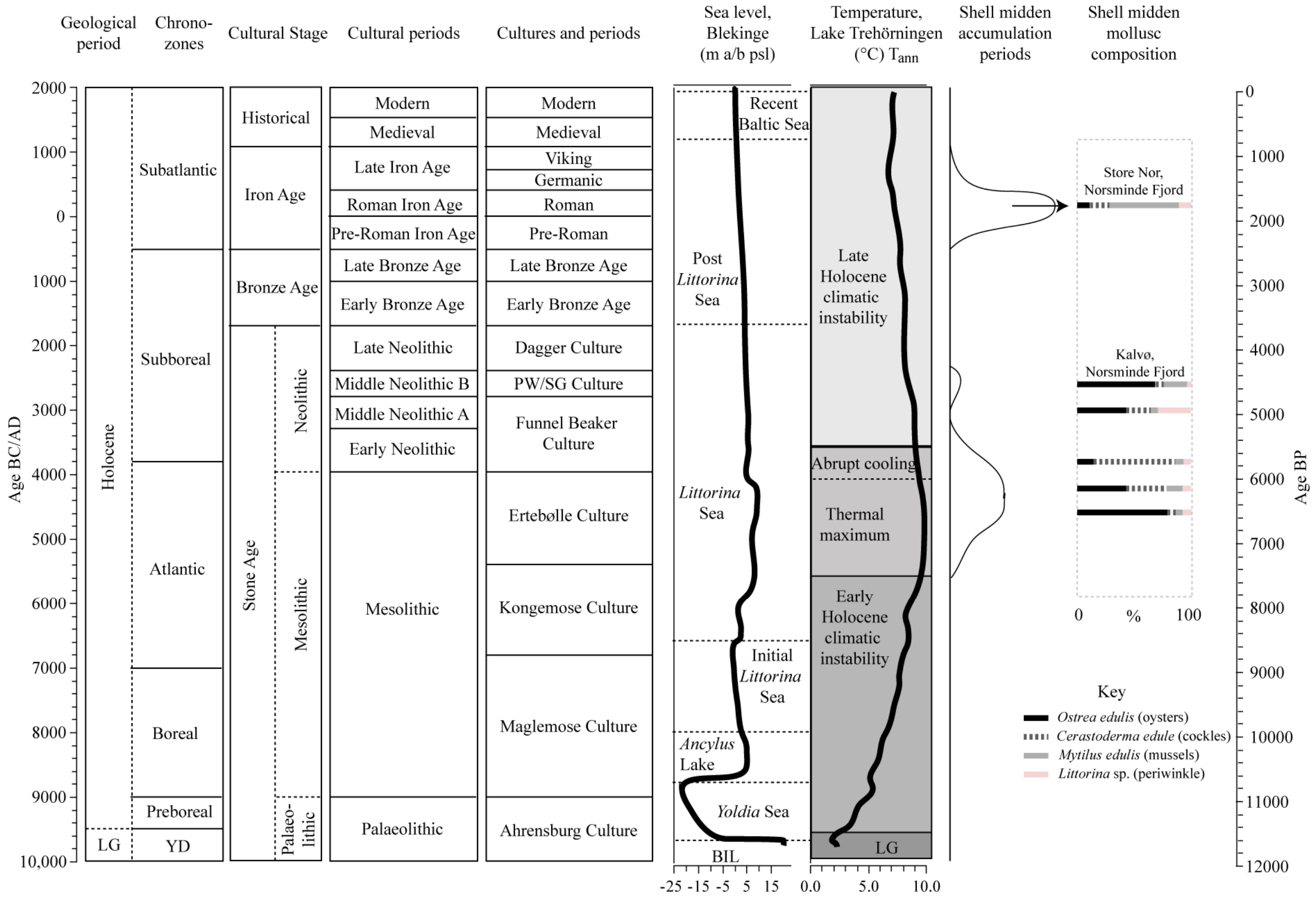


Figure 1.4 - Summary of the geological and cultural periods (including shell midden accumulation phases and shellfish composition data) referred to in the text, along with regional sea level and climate data for the Baltic Sea area. The sea level curve is from Blekinge, southeastern Sweden (Berglund *et al.*, 2005) and the (annual mean) temperature record is from Lake Trehörningen, western Sweden (Antonsson and Seppä, 2007). The Baltic Sea stages (sea level column) are based on the stratigraphy suggested by Andrén *et al.*, 2000; Berglund *et al.*, 2005; Andrén *et al.*, 2007b. The climatic periods (temperature column) follow divisions in Snowball *et al.*, 2004. The percentage shellfish composition data is based on data from Figure 3 in Andersen, 2007. The quoted values for the Iron Age midden at Store Nor, Norsminde Fjord are also from the reference. The date for the Single Grave Culture midden at Kalvø is from Andersen, 1983. Abbreviations: BIL = Baltic ice lake, LG = late-glacial, YD = Younger Dryas. *m a/b psl = metres above/below present sea level.

The number of (stratified) shell middens increases dramatically in the Ertebølle period, becoming common features of the late Mesolithic coastal landscape, only exceeded in size, number and volume during the Iron Age (e.g. Andersen, 2007). Whilst the majority of Stone Age middens belong to the Mesolithic Ertebølle culture (particularly the period ~4,600-4,400 BC; Andersen, 2007), some contain Neolithic layers (belonging to the Funnel Beaker Culture (TBK); e.g. Andersen, 1989) and cover the Mesolithic-Neolithic transition, thereby providing important archives of both cultural development and marine environmental change (Andersen, 1989; Andersen and Rasmussen, 1991). Excavation and analysis of a large number of late Stone Age middens has generated substantial information concerning the diets (i.e. through quantification of mollusc, mammal and fish bone compositions), technology and hunting/fishing practices (i.e. analysis of pottery, tools, artefacts and cooking utensils) of the Ertebølle and Funnel Beaker Cultures.

These analyses have been supplemented by state of the art techniques such as $\delta^{13}C$ and $\delta^{15}N$ isotopic analysis of teeth and bones from humans and dogs (Tauber, 1981; Milner *et al.*, 2004; Fischer *et al.*, 2007) and the analysis of incremental growth lines from the hinges of molluscan shells present within the middens (Milner, 2001a, 2001b, 2002). The former (i.e. isotopes) reveal information concerning the dietary composition (i.e. different group exhibit different isotopic signatures) and the latter enable establishment of shellfish collection patterns (e.g. season of collection) and resource pressure (i.e. the age and size of the molluscs at the time of collection). It is beyond the scope of the present study to provide a detailed review of the archaeological composition of Stone Age shell middens (e.g. differences/similarities between all sites, detailed assessment of pottery, tools and artefacts of these cultures), and therefore in the following section just some of the key characteristics, relevant to this study are highlighted. More detailed reviews can be found in the following references: Price, 1985; Larsson, 1990; Price, 1991; Stafford, 1999; Andersen, 2000; Fischer and Kristensen, 2002; Andersen, 2007; Andersen, 2008a.

One of the most striking characteristics of the late Danish Stone Age middens is the high abundances of large oysters (*Ostrea edulis*) present in the Mesolithic layers (Andersen, 1995a, 2007; Andersen, 2008a; Bailey and Milner, 2008; Bailey and Milner, submitted), sometimes accounting for up to 60-80 % of the total shell composition (e.g. Bjørnsholm Bay and Ertebølle *locus classicus*; Andersen and Rasmussen, 1991). Where midden accumulation persists into the Neolithic (e.g. Bjørnsholm Bay, Norsminde Fjord), there is a subsequent decline in oysters, with other less salt-demanding species such as *Cerastoderma edule* (cockles) and *Mytilus edulis* (mussels) becoming far more dominant (Andersen, 1989; Andersen and Rasmussen, 1991; Andersen, 2008a; Figure 1.4). One of the most classical examples of this (i.e. oyster decline) is the stratified Norsminde shell midden (occupied between ~4,500-3,200 BC; Figure 2.4). Oysters account for ~40-50 % of the total shell accumulation in the Mesolithic layers, but drop to less than 10 % in the late Mesolithic (after ~4,300 BC) and remain low thereafter (Andersen, 1989; Andersen, 2007; Bailey and Milner, 2008; Bailey and Milner, submitted). Several theories (including environmental, demographic and cultural hypotheses) have been proposed to explain this oyster decline, but to date none offer a completely satisfactory explanation or currently lack the sufficient supporting evidence (e.g. environmental 'salinity' hypothesis) to prove/discard the theory. In this study particular attention is placed on critically testing environmental-hypotheses (i.e. salinity, productivity and sedimentary regime; discussed in more detail below) by reconstructing changes in key environmental parameters over the Mesolithic-Neolithic transition (discussed below in more detail).

Also apparent in shell middens and other coastal sites are a large number of fish bones, particularly in the Mesolithic layers, indicating the importance of fishing to the Ertebølle culture (Enghoff, 1989, 1991, 1995; Enghoff *et al.*, 2007). For example, Enghoff, 1989 analysed almost 9,000 fish bones from the Norsminde shell midden, yielding over 25 different species, almost all of which were brackish or marine taxa (flounder most abundant, 57 %). Flounder and gadids (e.g. cod and saith) are generally most abundant in settlements from the Vedbæk area, whilst in one of the northern most Ertebølle settlements (Yderhede) flatfish and spurdog bones dominate the assemblage (Enghoff, 1995). Other sites, however, have demonstrated that fishing was not strictly a marine activity, with many fish species present from freshwater and fluvial sources. For example, 71 % of identified fishbones from Ertebølle *locus classicus* midden belonged to freshwater species (mostly roach) with 12 % saltwater taxa (gadids most abundant) and 17 % migratory fish (eel). At Bjørnsholm Bay, eel bones were most abundant (56 %), with saltwater taxa accounting for just 22 % of fish bones.

The variability in fish bone composition between sites is somewhat contrasting to the relatively consistent shellfish composition of Ertebølle sites. Whilst fishing was clearly an important activity, particularly in the summer months (Enghoff, 1991, 1995; Enghoff *et al.*, 2007), it appears that the catch (i.e. species exploited) was more strongly governed by localised factors, such as the topography, proximity to freshwater inlets or sites (e.g. rivers or lakes), water depth, protection, and availability of suitable breeding grounds (e.g. eel-grass beds, macrophytes). For example, it has been suggested that eel might have been readily caught at Bjørnsholm Bay, at the mouths of small streams running into the former fjord inlet, whilst at Norsminde, the shallow waters of the former Kysing Fjord might have provided ideal breeding grounds for flounder (Enghoff, 1989, 1991, 1995).

Whilst fish bones are often abundant in Mesolithic coastal shell middens and deposits, they appear to be very scarce in Neolithic layers (e.g. Bjørnsholm Bay, Mesolithic: Neolithic fish bone ratio, 11,490: 252 bones (Enghoff, 1991, 1995), Norsminde Fjord, all bar one bone were found in the Mesolithic layers; Enghoff, 1989). The reason for this remains uncertain, particularly as species composition and relative proportions (e.g. Bjørnsholm Bay), do not appear to change markedly from the Mesolithic layers (Enghoff, 1986; Enghoff, 1994; Enghoff, 1995; Andersen, 2007), suggesting similar fishing patterns (i.e. methods and targeted species). It is plausible that the fish catch was substantially reduced due to introduction of agriculture and reduced dependence on marine resources (e.g. Tauber, 1981; Milner *et al.*, 2004), though this does not satisfactorily explain the almost entire absence of fish bones from sites such as the Norsminde Fjord shell midden. Site abandonment is ruled out due the clear continuation of numerous middens (and other sites bearing coastal resources) in the Neolithic (including some sites, dated entirely to the Neolithic period). It remains possible that fish bones were deposited elsewhere, but this is hard to believe considering the fact that both molluscs and bones from other animals were deposited on these sites. Perhaps the most likely explanation is accelerated decomposition of fish bones in the Neolithic layers, due to a reduction in shell midden accumulation rates (as marine resources became less important) or changes in the physical environment (e.g. increased aerial exposure, weathering and erosion).

Shell middens (and other coastal sites) remain present for the first ~400-500 years of the Neolithic, but gradually become smaller in size and more sporadic (particularly between ~3,600-3,300 BC; Andersen, 2008a) before practically stopping in the middle Neolithic (Andersen, 2007). This early Neolithic period (i.e. ~3,900-3,600 BC) therefore probably marks a transitional phase from a marine-dependent community in the Ertebølle period, to a predominately terrestrial-based resource economy in the Neolithic period. This is supported by the continued collection of marine resources (most notably cockle and mussels, but also smaller amount of oysters in the early Neolithic) and the presence of some human skeletal remains (dated to the early Neolithic) that still exhibit isotopic signatures indicative of a predominately (or at least substantial) marine diet (Bailey and Milner, 2008). In some early

TBK layers, oysters are present at relatively high abundances (e.g. Krabbesholm and Havnø), though these sites appear to form the exception rather than the normal (Andersen, 2005, 2008b; Nielsen, 2008). After ~3,600 BC the pollen records also generally show strong signs of large scale agricultural impacts (e.g. reduction in arboreal pollen, continuous curves for arable and pastoral grassland plants such as *Plantago lanceolata* and *Rumex acetellosa* and regular presence of cereal pollen; Iversen, 1941; Aaby, 1986; Andersen, 1995b; Rasmussen, 2005).

1.6.2 Salinity decline at the Mesolithic-Neolithic transition; a long outstanding issue

A popular hypothesis for the delay of agriculture is that the marine environment was productive enough during the late Mesolithic not to necessitate agricultural innovation (e.g. Rowley-Conwy, 1984). Higher sea levels in the Ertebølle period (~5,600-4,200 BC; Iversen, 1937; Troels-Smith, 1942; Mikkelsen, 1949; Petersen, 1981; Christensen, 1995; Berglund *et al.*, 2005) is believed to have created a wider connection between the Kattegat and Skagerrak, causing greater water exchange with the North Sea. This resulted in increased inputs of high salinity, oxygenated and more nutrient rich North Sea water, important for maintaining a healthy and productive ecosystem (e.g. by reducing the likelihood of stagnation and hypoxia of deeper waters, particularly in the deeper basins). It also enabled high-salinity demanding organisms (and popular food stuffs) such as oysters to extend into southern areas of Kattegat and penetrate the inner (more human accessible) parts of the fjord (Nordmann, 1903). Some studies have also alluded to a larger tidal amplitude existing in the Mesolithic (Nielsen, 1938; Petersen, 1993; Troels-Smith, 1995; Petersen *et al.*, 2005). In particular, Nielsen, (1938) proposed that larger tides caused greater mixing of water masses and increased the ‘delivery’ of high salinity water into the innermost part of estuaries and fjords, though this hypothesis is more speculative than fact, lacking any good physical supporting evidence. Furthermore, recent modelling of tidal and tide dependent changes for NW European shelf seas, however, suggest that over the last ~8,000 years (i.e. the period in which sea level has been closest to its present level), tidal changes have been generally small (Uehara *et al.*, 2006). The Ertebølle period also coincides with the mid-Holocene climatic optimum (~5,550-3.550 BC; Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Seppä *et al.*, 2009), meaning higher temperatures might also have been important for increasing the productivity of the marine environment at this time (e.g. longer growing/breeding seasons, reduced freezing and thawing events in winter). Abundant shell fish populations in close proximity to good fishing sources (e.g. marine and freshwater resources available within in the same fjord) offered a reliable, diverse and rich marine-based diet for the Ertebølle culture. In accord with the palaeodietary evidence (see below, Figure 1.5), it is therefore feasible that the marine biotope was more attractive than agricultural innovation at this time.

There are two main lines of palaeodietary evidence that indicate marine resources were predominant in the Ertebølle diet: 1, δ^{13}C and δ^{15}N isotopic analyses of human skeletons (most notably bones and teeth). 2. Abundance volume and composition (of marine shellfish, fish and mammals see above) of shell middens and other coastal sites. The principle underlying isotope based palaeodietary techniques such as δ^{13}C and δ ^{15}N isotopic analysis is that skeletal bones are made from what the animal eats (or drinks) (Hedges *et al.*, 2005). Therefore carbon isotope ratios (from bones and teeth) can reveal the predominant source of food eaten (i.e. terrestrial or marine organism exhibit different isotopic signatures), whilst nitrogen isotope can provide further information concerning the trophic level of organisms exploited (generally quoted as 3-5 ‰ enrichment at each trophic level; Schoeninger *et al.*, 1983). Mesolithic skeletons from Denmark exhibit δ^{13}C values in the expected range for a predominately marine based diet (e.g. generally 10-15 ‰; Tauber, 1981; Milner *et al.*, 2004; Fischer *et al.*, 2007), whilst the δ^{15}N values (average ~14 ‰; Schoeninger *et al.*, 1983; Richards and Hedges, 1999; Fischer *et al.*, 2007) and δ^{13}C: δ^{15}N ratios demonstrate the importance of resources such as shell fish to the Ertebølle culture. In the Neolithic period, there is a shift towards a predominately terrestrial-based (Figure 1.5), with substantial consumption of agricultural resources (Tauber, 1981; Schoeninger *et al.*, 1983; Fischer *et al.*, 2007). Whilst substantial debate has taken place about the real meaning of these isotopic results (e.g. Hedges, 2004; Milner *et al.*, 2004, 2006; Richards and Schulting, 2006b; Richards and Schulting, 2006a), it seems there is a general consensus that over time the proportion of terrestrial protein consumed in the human diet has increased (Bailey and Milner, 2008).

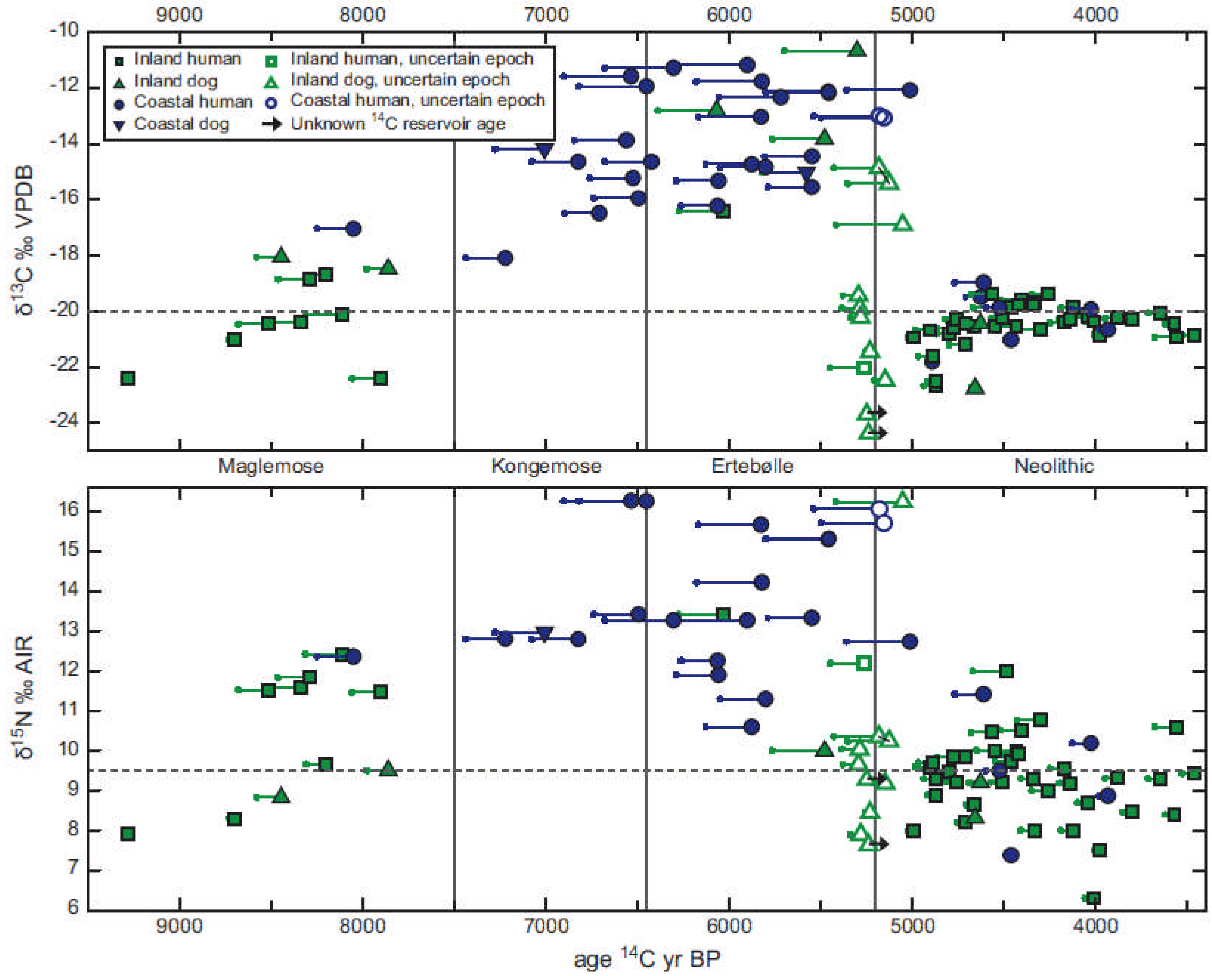


Figure 1.5 – δ^{15}N and δ^{13}C values for bone collagen from human and dogs for the Mesolithic and Neolithic period. Open symbols refer to individuals with ambiguous cultural belonging, being found at the Mesolithic-Neolithic transition. One-sided error bars represent marine reservoir correction and black arrows indicate uncertain reservoir effect. The dotted line for carbon isotopes (i.e. at 20 ‰) marks the limit above which there is an indication of a non-negligible marine diet, though individuals below this line may have consumed small quantities of marine food. The dotted line for nitrogen isotopes (i.e. at 9.5 ‰) marks the limit above which individuals must have consumed aquatic food regularly, through the same may apply to individuals below the line if substantial parts of this dietary component derived from low trophic level organisms such as shellfish (source Fischer *et al.*, 2007).

Shell analyses (i.e. for size, age and seasonality) from the Norsminde shell midden indicate that molluscan populations (particularly *Ostrea edulis*) were subject to less pressure in the Ertebølle period, when they formed a greater part of the cultural diet (Bailey and Milner, 2008; Bailey and Milner, submitted). The age and size of oysters is stable in the Mesolithic layers indicating a consistent collection pattern (in late winter/early spring), taken from a large healthy population, with no signs that exploitation was ever damaging to the availability of this resource. In the Neolithic, however, the collection season lengthens (i.e. collection season extends into the summer) and the age and size at the time of collection fluctuates suggesting variable exploitation, at time even exerting pressure upon the oyster populations.

Bailey and Milner, 2008 suggest several potential reasons for this including exaggeration of the importance of marine resources (i.e. potential inaccuracy associated with the isotope results, see above), reduced availability of marine resources (due to environmental change), increased population in the Neolithic, changes in the yearly resource schedule (associated with the introduction of agricultural resources) and less care taken by Neolithic cultures to preserve marine resources, due to their wider resource base.

It has even been suggested that the oyster decline might have been an important factor behind the neolithisation (i.e. introduction of agriculture) of Denmark (Rowley-Conwy, 1984). Peter Rowley-Conwy suggests, that for the majority of the year, oysters were probably exploited as a supplement or relish (probably forming less than 10% of the total Ertebølle diet at most; Bailey, 1975; Clark, 1975), being far less attractive than the majority of other food stuffs due to their low productivity per hour of work (based on tentative calculations for calories per hour work; Rowley-Conwy, 1984). The high abundance of oysters in the Ertebølle layers of the shell middens (Andersen, 2000, 2007; Andersen, 2008a), however, indicates that at some point during the year exploitation must have been intensive (i.e. late spring/early summer; later confirmed by Bailey and Milner, 2008; Bailey and Milner, submitted). As oysters are available all year-round (and were likely more accessible during the Ertebølle), have relatively low seasonal variation in fat content compared with other resources and are a good source of carbohydrate, their appeal is likely to have increased with the diminishment of other resources (Rowley-Conwy, 1984). It is never suggested that during the late winter/early spring, that the Ertebølle culture lived entirely on oysters, but rather that due to the reduction in other resources in these months, oysters became a more important part of the diet. When this resource become under pressure and less abundant and/or reliable due to changing environmental conditions (most notably lower salinity), it is hypothesized that humans might have adopted agricultural techniques to fill this gap in the seasonal resource cycle (Rowley-Conwy, 1984).

The decrease in salinity is believed to have been caused by declining sea levels in the late Mesolithic (e.g. Iversen, 1937; Berglund *et al.*, 2005) and a reduction in the tidal amplitude (Rowley-Conwy, 1984). In the majority of the inner Danish coastal waters today, salinity is below the minimum 23-25 ‰ required (Jensen and Spärck, 1934; Yonge, 1960) to support breeding oyster populations, except for in some areas of the Limfjord and in the higher salinity waters of the Kattegat. Nordmann, (1903) suggested that even in the Ertebølle period, when sea levels and temperatures were higher, the extension of oysters into the southern Kattegat (as far south as northern Zealand) and inner Danish coastal waters meant they were near to their lower salinity tolerance limits and even small magnitude changes in salinity might have serious implications for oyster populations. It is important to note that oysters never completely disappear from the Neolithic layers of Danish shell middens (Andersen, 2007; Andersen, 2008a, but rather appear to become exploited less and subject to greater stress (Bailey and Milner, submitted), perhaps related to reduced spawning years. Cockles are able to continue spawning at lower salinities and temperatures (6 ‰ salinity and 9°C or less temperature) and can survive better in softer muddier substrates (Sorgenfrei, 1958; Petersen, 2004; Bailey and Milner, submitted). However, despite the high abundance of cockles (and mussels) in the Neolithic layers, it is suggested that these taxa were not suitable nutritional substitutes for oysters at this time of year (i.e. late winter/early spring). Cockles and mussels do not reach their peak nutritional levels (i.e. highest meat: shell ratio) until later in the year (i.e. the summer months; Rowley-Conwy, 1984; Bailey and Milner, submitted).

The ‘salinity’ hypothesis (e.g. Rowley-Conwy, 1984) remains speculative to date, lacking any high quality salinity data from the inner Danish waters over the late Mesolithic/early Neolithic period. Furthermore, whilst sea level is likely to be an important driver of salinity change, in settings such as estuaries and fjords, salinity is also likely to be influenced by a number of other variables (e.g. climate change, terrestrial hydrology, changes in basin morphology, stratification, and marine exchange, all of which might independently be influenced by sea level). In addition to salinity, it has been proposed that changes in temperature and substrate could also explain the decline in oysters in the Danish middens (Bailey and Milner, 2008; Bailey and Milner, submitted). Regional temperature records suggest decreasing temperatures after ~4,000 BC following the mid-Holocene thermal maximum (Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Seppä *et al.*, 2009). As *Ostrea edulis* only spawns when water temperatures reach ~15°C or above (Yonge, 1960), then decreased temperatures might have suppressed the spawning season, possibly even preventing spawning in some years. *Ostrea edulis* is also known to struggle with large volumes of fine sediment (Yonge, 1960), preferring harder, sandier substrates, with little vegetation cover (Rasmussen, 1958). Some evidence exists for increased accumulation rates in the Neolithic (e.g. Petersen, 1993), which might have therefore further stressed oyster population particularly in the shallower areas where large volumes of material might have been settling out (both temperature and substrate are discussed in more detail in Chapter 6).

The idea of a rapid Mesolithic-Neolithic transition has also been questioned and a number of sociological arguments have been made for the introduction of agriculture forming part of a long-term social transition, with large regional differences (e.g. Petersson, 1996-97; Stafford, 1999; Johansen, 2006; Larsson, 2007). It is beyond the scope of this thesis to review all (non-environmental) arguments concerning the Mesolithic-Neolithic transition in southern Scandinavia. Therefore, here just some of the key counter arguments are highlighted, focussing on those relating to the interpretation of shell midden evidence.

Petersson, (1996-97) has suggested that current interpretation of the archaeological shell midden evidence might be erroneous, and that even within the shell middens there is evidence for a gradual transition, with regional variability and potential cultural dualism. Petersson draws attention to the mixture of Ertebølle and TBK artefacts in some middens (e.g. one third of the pottery presented in the Norsminde shell midden is in fact Neolithic; Petersson, 1996-97), for which no completely satisfactory explanation has been given. Based on the stratified shell middens (e.g. Norsminde, Bjørnsholm Bay), these mixed cultural layers are generally interpreted as artefacts of geological processes such as re-working and re-deposition rather than possibly reflecting cultural dualism (e.g. Petersson, 1996-97 and references therein). Furthermore, the shell middens are dated via ^{14}C dating on marine shells present (most notably *Ostrea edulis* and *Cerastoderma* spp.; Andersen and Johansen, 1986; Andersen, 1989; Andersen and Rasmussen, 1991; Andersen, 2000) and therefore it is argued that the ecological shift has been dated rather than the cultural shift (e.g. Petersson, 1996-97). In sites such as Bjørnsholm Bay, this ecological shift occurs prior to any clear signs of a cultural shift (e.g. pottery) within the midden (Andersen and Rasmussen, 1991) and some 200 years before the transition at Norsminde. Petersson, 1996-97 cites climate change as a possible cause for the ecological shift, but what is dismissed here, is that if environmental change forced the Mesolithic-Neolithic transition, then an ecological shift would be expected to pre-date the cultural shift. The delay at Norsminde is contradictory to the idea of a rapid and regional shift towards agriculture and in better agreement with finds at sites such as Ǻmosen (Zealand) which indicate a clearer gradual Ertebølle-TBK transformation (Fischer, 1993).

Johansen, 2006 has also critiqued the proposed model for Ertebølle settlement at larger fjords (e.g. Bjørnsholm Bay, Mariager Fjord), suggesting that the Ertebølle culture were not a stable sedentary culture, inhabiting big year round coastal sites, but rather comprised of numerous small groups rotating between sites on a seasonal basis. The ‘constellation’ settlement model is based around the findings at larger fjords such as Bjørnsholm Bay and Roskilde Fjord, where one large settlement (containing resources from all seasons) is surrounded by several seasonal smaller extraction camps (Andersen, 1995a; Pedersen *et al.*, 1997). Johansen, 2006 argues that the smaller camps are not seasonal extraction camps, but rather represent the day to day living activities of small groups, being not dissimilar in content to the large 'year round camps'. Johansen also highlights that the evidence for year round occupation of the larger sites is equivocal. Winter resources are poorly represented (i.e. only birds and small fur animals) in the Ertebølle *locus classicus* and Bjørnsholm Bay middens, with the majority of resources representative of spring to autumn based activities. It is also suggested that the larger middens are palimpsests of countless smaller shell heaps, being returned to on a regular basis due to the large resource base (e.g. large natural fossil beds) present in close proximity to these sites. This is based on evidence within the middens displaying sections of intense exploitation interspaced by periods with little deposition (e.g. Andersen and Johansen, 1986). Whilst there are clearly still many gaps in the archaeological record, high resolution palaeoenvironmental analysis is vitally important for understanding better, the physical environment in which the agricultural transition took place in, and to provide proper environmental context for future debates on this issue.

1.6.3 The Pitted Ware and Single Grave Culture (~2,800-2,400 BC)

During the Pitted Ware and Single Grave cultural period, there is a renewed increase in the number of shell middens and shell banks present along Danish coastlines (Andersen, 2007). There is also a shift from cockle (*Cerastoderma edule*) dominated middens in the earlier Neolithic, back to oyster (*Ostrea edulis*) dominated middens, which might reflect another period of higher salinity (Andersen, 1983, 2007). Whilst the skeletal isotopic analyses from humans suggest that terrestrial resources continue to dominate the diet throughout this period (Tauber, 1981; Bailey and Milner, 2008, Figure 1.5), there is certainly an increase in the importance and exploitation of marine resources. It is possible that a shift in diet is only (or more) apparent in the coastal-dwelling communities, as Pitted Ware cultural sites tend to most commonly be found along the former coastlines, often found near or with Ertebølle remains. This has led to suggestions that sea level was near to Ertebølle levels at this time (Becker, 1951), but sea level (and salinity) change remains relatively poorly understood in this part of the record. Several studies exhibit potential sea level peaks (i.e. transgressions) broadly coinciding with the Pitted Ware Cultural period (e.g. Miller, 1982; Miller and Hedin, 1988; Christensen, 1995), but problems with dating (e.g. marine reservoir) prevent any confident links being made. The Pitted Ware culture (PWC) is seen as an independent group distributed across southern Sweden and the north/north-east of Denmark (around the Kattegat, on islands and by fjords including the Limfjord; Rasmussen, 1993a). The PWC, is however, synchronous with the Single Grave culture (and possibly the Swedish Battle Axe culture) but is treated separately due to the lack of evidence for continuity between these other cultures (Becker, 1951; Rasmussen, 1993a). Coastal sites belonging to the Single Grave Culture demonstrate that this culture also exploited marine resources and the composition of shell midden and shell sites is often similar in mollusc composition to sites from Pitted Ware Culture. For example a Single Grave culture layer at the top of the Kalvø shell midden, (located on the north-eastern coast of the former island within Norsminde Fjord) contains 68 % oysters, 20 % mussels, 4.5 % cockles and 2.5 % periwinkles (*Littorina* spp.) (Andersen, 1983).

1.6.4 The Bronze Age

Collection of marine shells seems to decrease in the Bronze Age (1,700-500 BC), with no shell middens present, but rather just thin layers or small heaps of shells present at various settlements (e.g. rich Bronze Age burial sites in the Limfjord beach ridges at Hverrehus and Mellemholm; Andersen, 2007 and references therein). Mussels (*Mytilus edulis*) are the most abundant species in Bronze Age deposits, but oysters remain present at most sites but generally in small numbers (e.g. a small heap at Vadgård; Rasmussen, 1993b). Fish, seal and whale bones and fishing tools (e.g. fish hooks) are also found, further suggesting that coastal resources played at least some role in the Bronze Age economy (at least in the coastal regions), though the diet appears to be predominately terrestrial based (Tauber, 1981; Andersen, 2007).

1.6.5 The Iron Age (~500 BC-AD 1050) and the re-appearance of coastal shell middens

In the Iron Age, there is a large increase in the number of shell middens and other settlements containing marine resources (molluscs and fish-bones) or fishing tools present along Danish coastlines (Poulsen, 1978; Andersen, 2007). The majority of these belong to the late pre-Roman/early-Roman period (~300 BC-AD 200; Poulsen, 1978; Andersen, 2007). These shell middens are predominately composed of *Mytilus edulis* (generally most abundant), *Cerastoderma edule* and *Littorina littorea,* but again *Ostrea edulis* shells are regularly present. The larger shell midden sites appear to be confined to the eastern coast of Jutland and the northern coasts of Fyn and Zealand, but with middens extending further south into the

Baltic, to the east coast of Schleswig (Anger, 1973; Harck, 1973; Poulsen, 1978; see Figure 1.3). Some of the Iron Age middens are extremely large, with a cubic content of up to 15,000 m^3 and sometimes extend up to 1 km long with just short interruptions (e.g. Norsminde Fjord and Mariager Fjord), thereby far exceeding the size of the largest Stone Age shell middens (Andersen, 2007).

Interestingly, there are no Iron Age shell middens right on the shoreline of the Limfjord, though a number of sites are found several kilometres inland, containing heaps of marine molluscs or smaller shell middens (Mikkelsen, 1994; Andersen, 2007). The reason behind this absence of coastal middens in the Limfjord is uncertain, but evidence exists for both climatic/environmental changes (e.g. lower salinity, higher productivity, higher temperatures; Lamb, 1982; Kristensen *et al.*, 1995; Christensen *et al.*, 2004) and cultural/behavioural changes (e.g. molluscs transported inland prior to shelling; Poulsen, 1978; Ringtved, 1992) occurring in the Iron Age. Kristensen *et al.*, 1995 and Christensen *et al.*, 2004 suggest that the salinity of the Limfjord might have dropped as low as 7-20 ‰ during the pre-Roman/Roman period, indicating either partial connection or complete closure from the North Sea and/or Skagerrak at this time (though these salinity estimates are qualitative and dating remains uncertain; see above). It is believed the western and/or northern entrances to the Limfjord were re-established (or widened) by the Viking period (AD 750-1050) with foraminifera and molluscan records suggesting that high salinities occurred in this cultural period (22-30 ‰; Kristensen *et al.*, 1995; Christensen *et al.*, 2004, probably due to rising sea level (e.g. Viking sea level high stand identifiable in Tanner, 1993; Christensen *et al.*, 2004).

In the Kattegat and western Baltic Sea, it is believed that the marine environment was rich and prosperous during the late pre-Roman/early-Roman period and that salinity probably exceeded its present day levels (Petersen, 1985; Hebbeln *et al.*, 2006; Rasmussen *et al.*, 2007; Zillén *et al.*, 2008). For example, *Ostrea edulis* extends into the southern Kattegat and Baltic Sea, well beyond its present day limits (e.g. Harck, 1973; Poulsen, 1978), suggesting greater exchange of water with the North Sea (and subsequent increased input of high salinity-oxygenated water) must have existed at this time (Emeis *et al.*, 2003; Rasmussen *et al.*, 2007). Sea level is the most likely cause for this increased exchange of water, with one or more transgressions identifiable in regional sea level records covering the pre-Roman/Roman periods (e.g. Miller, 1982; Miller and Hedin, 1988; Behre, 2007).

No shell middens are found later than the Viking Age, but archaeofaunal evidence (e.g. fish and marine mammal bones) and documentary records (e.g. fish catches) demonstrate that marine resources remained an important dietary component throughout the Medieval and Historical times (Enghoff, 1999; Holm and Bager, 2002; Poulsen *et al.*, 2007). For example, along the Baltic and Danish Kattegat coastline, cod, haddock and flatfish formed an important part of the economy over the last ~1,500 years (e.g. Enghoff, 1999; Holm and Bager, 2002) and as already indicated above plaice, eel and herring fisheries were present and exploited at different times in the Limfjord (see Section 1.5). It has also been shown that the development and prosperity of these fisheries is often closely related to climatic and environmental conditions (e.g. Kristensen *et al.*, 1995; Nissling and Westin, 1997; Enghoff, 1999; Holm and Bager, 2002; MacKenzie *et al.*, 2002; MacKenzie *et al.*, 2007), again highlighting the need for long-term records of ecosystem response to climatic and environmental change in order to maintain and manage important commercially important marine resources.

1.7 Project aims and proxies

This project aims to reconstruct changes in key environmental parameters (most notably salinity, nutrient status, productivity and the sedimentary regime) at three coastal Danish sites over the Holocene, in order to assess how land, sea and society have interacted through time and assess critically environmental hypotheses for cultural change (e.g. Rowley-Conwy, 1984). To achieve this, three nearshore coastal sites (Kilen, Norsminde Fjord and Korup Sø; Figure 1.1, Figure 1.3), situated in close proximity to important archaeological sites and settings have been chosen for detailed multiproxy palaeoenvironmental analyses. At each site, diatom analysis has been performed for quantitative reconstruction of salinity change (with increased resolution over key periods of archaeological change, particularly shell midden accumulation phases) using a salinity transfer function, based upon a trans-Baltic training set (derived from sites sampled during the MOLTEN and DEFINE project; Andrén *et al.*, 2007a: http://craticula.ncl.ac.uk/Molten/jsp/). At Kilen, isotopic analysis of single-species foraminifera tests has also been employed as a secondary proxy for salinity change, and high resolution sedimentary pigment analyses for assessment of changes in productivity and/or nutrient status over the study period. Physical sedimentary analyses (including lithological descriptions, organic matter, carbonate and minerogenic matter; Troels-Smith, 1955; Dean, 1974; Birks and Birks, 1980; Bengtsson and Enell, 1986) have also been performed to provide important information concerning the physical environment, sedimentary regime and substrate at each site over the study period.

In addition to the proxies analysed as part of this thesis, mollusc (all sites), foraminifera (Kilen and Norsminde Fjord), pollen (Norsminde Fjord and Korup Sø) and plant macrofossil (Kilen and Norsminde Fjord) datasets have been used as part of this multiproxy investigation. It is important to note that these additional datasets have not been analysed by the author, but all interpretations are the author's work, unless otherwise referenced. The pollen and plant macrofossil analyses are particularly important as they provide information concerning changes in the terrestrial environment and human impacts upon the catchment. These records are vital for examining interactions between the land, sea and society and are important for verification of the age models (e.g. *Ulmus* decline at ~3,900 BC; Andersen and Rasmussen, 1993). Where possible, site chronologies have been based on AMS ^{14}C dating of terrestrial plant macrofossil material to avoid problems associated with the marine reservoir effect (Heier-Nielsen *et al.*, 1995; Olsen *et al.*, 2009).

This project specifically aims to address the following questions:

1. How did the Danish coastal environment (salinity, nutrients, productivity and sedimentary regime) change during the Holocene at each site?
2. How do the changes identifiable in the independent sedimentary records compare with other local and/or regional records of climate and environmental change (e.g. terrestrial and lacustrine records, sea level changes), and can local events be separated from regional (or global) change?
3. What were the drivers behind the changes detectable in the environmental records and are they synchronous across coastal Denmark?
4. To what extent can events be attributed to the same causal mechanisms or are there different drivers of environmental change detectable in the proxy records from the different sites?)
5. Do these events occur within the same time frame as other known and well-dated marine (e.g. sea level), terrestrial or archaeological (e.g. introduction of agriculture) changes?
6. To what extent are changes in the societal record (e.g. shell midden accumulation periods, the oyster decline in the Neolithic shell middens) driven by changes in the marine (or terrestrial) environment?

7. Can a successful salinity signal be obtained for the inner Danish coastal waters using a diatom-based salinity transfer function and what other environmental variables (e.g. nutrients, sedimentary regime, preservation), are important in explaining variation in the diatom (and foraminifera or molluscan) dataset?
8. How have land, sea and society interacted throughout the Holocene?

1.7.1 Proxies

Diatoms are unicellular photosynthetic protists that have dominated marine primary productivity since the mid-Cretaceous (Round *et al.*, 1990; Sinninghe Damste *et al.*, 2004). They currently account for approximately 40% of modern day total primary marine productivity (Treguer *et al.*, 1995; Field *et al.*, 1998), a figure that can increase to 75-90% in highly productive areas such as coastal shelves, estuaries and areas of intense upwelling (Nelson and Smith, 1986; Nelson *et al.*, 1995; Maslin and Swann, 2006). Diatoms are very sensitive indicators of a number of environmental parameters, tend to preserve well in sediments due to their siliceous frustules, are identifiable to species level and are globally distributed (Battarbee, 1986; Battarbee *et al.*, 2001; Jones, 2007). In coastal and estuarine systems they have been used to reconstruct changes in productivity (Muylaert *et al.*, 2000) sea-surface temperatures (Birks and Koç, 2002), sea-level (Dellwig *et al.*, 1999; Zong and Horton, 1999; Gehrels *et al.*, 2006), salinity (Juggins, 1992; Jiang *et al.*, 1998; Parsons *et al.*, 1999; Ryves *et al.*, 2004) and nutrient status (Clarke *et al.*, 2003; Weckström *et al.*, 2004; Clarke *et al.*, 2006; Weckström, 2006).

In order to quantify salinity, a diatom-based salinity transfer function will be applied to the fossil diatom records from each site, based on a trans-Baltic diatom-environmental training set. The training set is derived from sites analysed during three projects (2001-2006): MOLTEN (Monitoring long-term trends in eutrophication and nutrients in the coastal zone), DEFINE (Defining reference conditions for coastal areas in the Baltic Sea for the Water Framework Directive) (http://craticula.ncl.ac.uk/Molten/jsp/; Andrén *et al.*, 2007a) and DETECT (Developing a tool for assessing ecological reference conditions in the coastal zone of the Baltic Sea; http://www.helsinki.fi/bioscience/ecru/projects/detect.htm). In total the MOLTEN project incorporates modern surface sediment diatom assemblages from 206 sites from Denmark (91), Sweden (35), Finland (55) and Netherlands (25). The DEFINE (2004-2006) project extended this training set into the Bothnian Sea (both Swedish and Finnish coasts), the Baltic states, the North German coast and Norway, and the DETECT project added more sites from the Finnish Archipelago Sea. To date, this training set have primarily been used for monitoring long-term nutrient trends (particularly total nitrogen, TN) in the Baltic Sea over the last ~250 years (Clarke *et al.*, 2003; Weckström *et al.*, 2004; Weckström, 2005; Clarke *et al.*, 2006; Weckström, 2006; Andrén *et al.*, 2007a). The combined training set, however, also offers great potential for quantitative inference of salinity change over the Holocene across the Baltic, from coastal to pelagic areas (i.e. spanning a salinity gradient of 0-31 ‰ (measured at almost all sites), with salinity independently explaining the largest amount of variation (5.2 %) throughout the entire MOLTEN/DEFINE dataset (Andrén *et al.*, 2007a).

Whilst a wide range of statistical techniques are available for examining species-environment relationships, and a series of measures can be taken to reduce the impact of other environmental variables, such influences can never be completely eliminated. Diatom based salinity models have been shown to perform reasonably well (i.e. produce accurate inferences of salinity) in limnic and coastal systems (Juggins, 1992; Gasse *et al.*, 1995; Ryves *et al.*, 2004) but confidence in diatom-based reconstructions can be further increased by using a multiproxy approach (e.g. Pienitz *et al.*, 2000; Birks and Birks, 2006; Ellegaard *et al.*, 2006; P. Rasmussen, D.B. Ryves *et al.* unpublished data). Mollusc, foraminifera and isotopic analysis ($\delta^{18}O$ and $\delta^{13}C$ on single species of foraminifera) have been included in this study to provide independent salinity data for comparison with and/or verification of the diatom inferred salinity estimates over the study period. These proxies have been regularly used in coastal and estuarine studies as palaeosalinity indicators (Petersen, 1993; Kristensen *et al.*, 1995; Burman and Schmitz, 2005; Kristensen and Knudsen, 2006) and a large amount of literature is available detailing the salinity requirements/tolerance limits of a large number of mollusc and foraminifera taxa (e.g. Sorgenfrei, 1958; Knudsen, 1971; Alve and Murray, 1995, 1999; Petersen, 2004).

Sedimentary pigments have been used in this study to assess independently marine productivity and nutrient status (Leavitt and Hodgson, 2001; McGowan, 2007) and help identify periods when these variables might be driving changes in the diatom community, thereby potentially skewing diatom inferred-salinity estimates. This technique is increasingly being used for assessment of nutrient status in coastal systems (Chen *et al.*, 2001; Hodgson *et al.*, 2003; Reuss, 2005a; Reuss *et al.*, 2005; Ellegaard *et al.*, 2006; Rabalais, 2007), being much less labour intensive than analysis of microfossils such as diatoms and dinoflagellates. Pigment samples are also relatively cheap and fast to process making it feasible for high temporal resolution to be obtained at suitable sites. Careful storage and treatment procedures must be followed to avoid degradation of pigments (Jeffrey *et al.*, 1997; Leavitt and Hodgson, 2001; Reuss, 2005b, 2005a), with recommended storage below -20° C, out of natural light and freeze-drying only immediately prior to analysis (Reuss, 2005b). Pigments are produced by a large number of organisms (e.g. bacterial group, higher plants, diatoms, chrysophytes) and under suitable sedimentation conditions can provide information regarding past marine productivity, nutrient status, and (where taxonomically specific pigments are preserved) can provide information regarding relative abundances of particular groups. For example, chlorophyll-*a* is produced by all photosynthetic algae and higher plants and therefore can provide a general overview of marine productivity, whereas pigments such as alloxanthin (Cryptophytes), diatoxanthin (diatoms, dinoflagellates, and chrysophytes), okenone and rhodopinal (purple sulphur bacteria) are specific to these groups. For more details on the affinity of major pigments preserved in sediments from aquatic environments such as lakes, estuaries and coastal environments, readers are referred to Goodwin, 1980; Leavitt, 1993; Jeffrey *et al.*, 1997; Leavitt and Hodgson, 2001; McGowan, 2007.

Interpretation of pigments must be treated with care, and consideration must be given to the depositional environment in which they accumulate. It is well known that only a small percentage (often <5% of all pigments, McGowan, 2007) actually get preserved in sediments with most being degraded in the water column, and therefore absolute abundances of pigments should not be interpreted too strictly (Reuss, 2005a). At sites with poorer preservation, any productivity or environmental record is likely to be overridden by this preservation signal (e.g. Reuss, 2005a; Reuss *et al.*, 2005). Nevertheless, under favourable conditions, sedimentary pigments can produce excellent records of past changes in marine productivity and nutrients status and have been demonstrated to work well in Danish fjord and estuarine systems (Reuss *et al.*, 2005; Ellegaard *et al.*, 2006; P. Rasmussen, D.B. Ryves *et al.* unpublished data).

Chapter 2. Study sites and methodology

2.1 Introduction

In this chapter the study sites and methodologies employed to answer the major research questions outlined in Chapter 1, section 1.7 are described. This chapter aims to address the following questions:

1. What criteria must each site meet in order to be able to address the main research questions, outlined in Chapter 1, section 1.7?
2. To what extent do the selected sites meet the specified criteria and why are these factors so important?
3. What methods have been used to obtain chronological control at all three sites?
4. What palaeoenvironmental/palaeoecological techniques were employed to reconstruct changes in marine environment (most notably salinity, nutrients and productivity) of each site over the Holocene?
5. What statistical/numerical methods were used to quantify changes in salinity at each site?
6. What palaeoenvironmental/palaeoecological techniques were used as proxies for change in the terrestrial environment (i.e. the catchment of the fjord)?
7. What statistical/numerical techniques were used explore the major trends in the biological datasets and to assess the interaction between the terrestrial and marine environment over the Holocene?

2.2 Study Area

2.2.1 Site selection

Three sites were analysed as part of this Ph.D. (Figure 2.1), all situated in the nearshore coastal region of Denmark and carefully selected to match several specific criteria:

1) The presence of terrestrial macrofossils for AMS ^{14}C dating.
2) Good chronological control.
3) High temporal resolution.
4) Situated in close proximity to important archaeological sites and settlements.
5) Good preservation of fossil diatoms.

As serious problems exist with the marine reservoir effect in Danish coastal regions (e.g. Heier-Nielsen et al., 1995; Olsen et al., 2009), terrestrial plant macrofossils were favoured over marine plants and shells for ^{14}C dating at both Kilen and Norsminde. The marine reservoir effect skews radiocarbon dates from marine material generally making them appear older than terrestrial samples of equivalent age (typically set at ~400 years; Stuiver and Braziunas, 1993) due to differences in exchange between carbon reservoirs. For the world oceans, it is caused by delays in carbon exchange between the atmospheric and oceanic bicarbonate together with the dilution effect (mixing of ‘young’ surface waters with very old deep waters in upwelling regions) (Mangerud, 1972). The magnitude of this effect is not the same in all locations and a large amount of research has gone into producing standard corrections errors for different areas (e.g. Marine reservoir correction database; http://radiocarbon.pa.qub.ac.uk/marine/).

Within this database, a number of standard correction dates exist for the Danish coastal regions (including the Limfjord), but temporal variability appears to be a major problem when trying to create accurate age-depth models (Heier-Nielsen *et al.*, 1995; Olsen *et al.*, 2009, submitted; B. Phillipe and J. Olsen, unpub.). For example, comparison between parallel terrestrial and marine dates from three Danish fjords (Horsens Fjord, Isefjord & Skælskør Nor) yielded a reservoir effect of -425 to +620 years (Olsen *et al.*, 2009). Olsen *et al.*, (2009) suggest this non systematic change between terrestrial and marine derived dates is due to variations in local hydrographic and hydrologic configuration (e.g. precipitation affecting residence time), catchment characteristics and connection with inner Danish seas. More recently, a preliminary assessment of the marine reservoir effect for Kilen (Limfjord) has also revealed significant temporal variability (B. Phillipesen, J. Olsen et al. unpublished data). It is therefore clear that in these systems a standard 400 year correction for the reservoir effect cannot be applied, and that the chronologies presented in previous studies using ^{14}C dates based on marine material might be erroneous (Petersen, 1981; Andersen, 1992; Heier-Nielsen, 1992; Kristensen *et al.*, 1995; Christensen *et al.*, 2004).

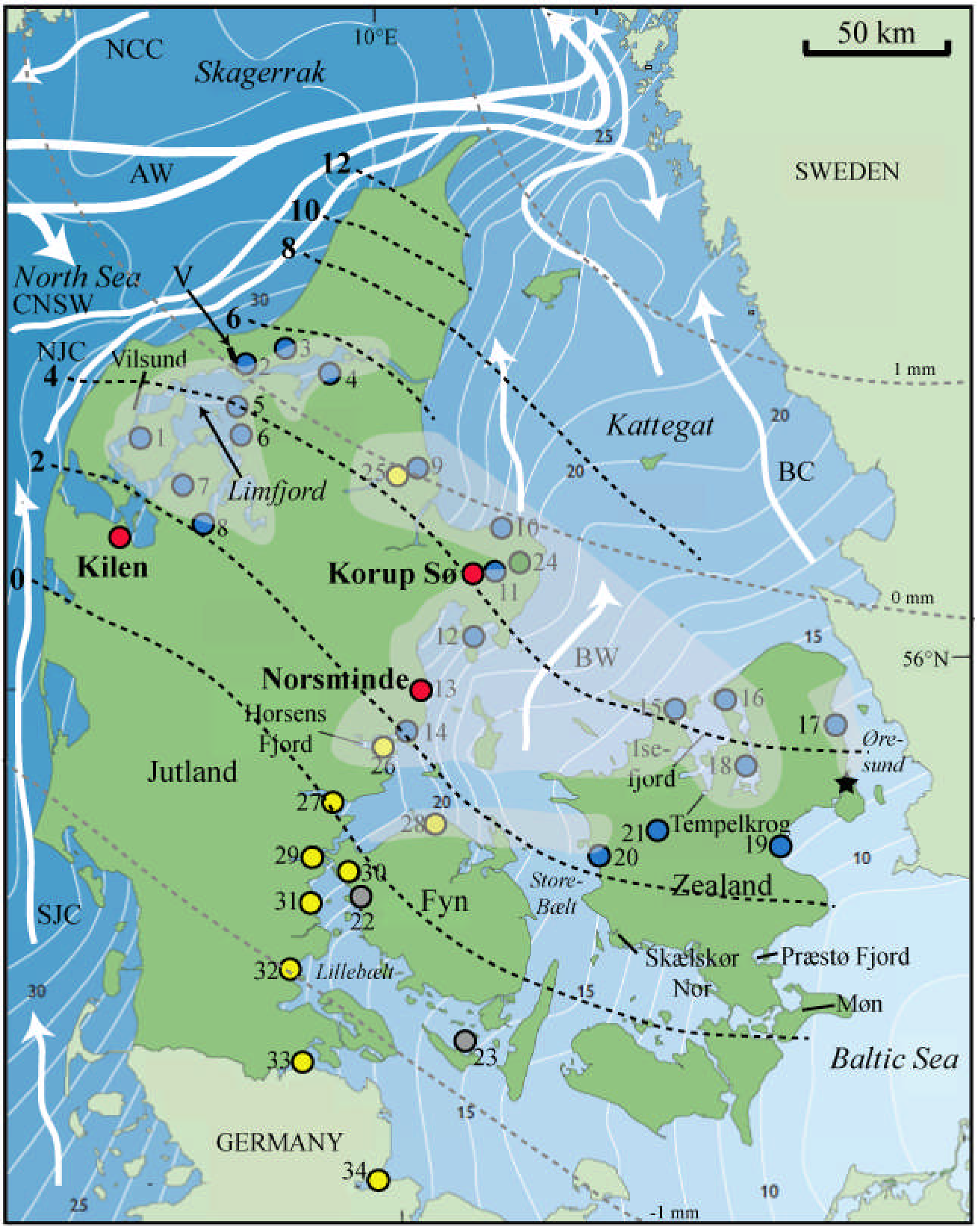


Figure 2.1 – Map of Denmark with the location of the three sites analysed in this thesis (Kilen, Norsminde Fjord and Korup Sø), plus other important sites referenced in the text. Present day isostatic adjustment (dotted grey lines; after Ekman and Mäkinen, 1996) and isobases for the highest level of the Littorina Sea in Denmark (bold numbers and black dotted lines; in metres after Mertz, 1924; Christensen, 2001a), modern day annual mean surface salinities (solid white lines; ‰) and sea current systems for the Kattegat and Skagerrak region (after Gyllencreutz *et al.*, 2006) are also included. Abbreviations: CNSW = central North Sea water, AW = Atlantic water, NCC = Norwegian coastal current, SJC = south Jutland current, NJC = north Jutland current, BW = Baltic water, BC = Baltic current. Numbered sites refer to selected important archaeological localities and settings. Shaded area highlights the areas in which middle and late Stone Age shell midden (in Danish: køkkenmødding) sites are found. Mesolithic Ertebølle and early Neolithic sites (blue dots): 1. Øster Jølby, 2. Aggersund, 3. Brovst (includes Kongemose layers), 4. Nørholm, 5. Bjørnsholm Bay and Åle, 6. Ertebølle *locus classicus*, 7. Hjerk Nor, 8. Krabbesholm, 9. Visborg and Havnø, 10. Meilgård, 11. Nederst, 12. Vængesø, 13. Norsminde Fjord and Kalvø (includes Ertebølle, Funnel Beaker culture (TBK), Single Grave culture and Iron Age sites), 14. Mosegården , 15. Klintesø, 16. Sølager, 17. Vedbæk (includes Kongemose, Ertebølle and TBK sites), 18. Lollikhuse, 19. Ølby Lyng, 20. Mullerup (important Maglemose site), 21. Kongemose *locus classicus*. Submerged Ertebølle sites (grey dots): 22. Tybrind Vig, 23. Møllegabet. Single grave and Pitted Ware cultural sites (green dots): 24. Kainsbakke and Kirial Bro. 25-34 (yellow dots) Iron Age shell midden sites. V. location of Vejlerne wetlands (Ryves *et al.*, 2004). Star – Copenhagen (capital city). Map modified from Dahl *et al.*, 2003 and Rasmussen *et al.*, 2007.

An advantage of the coastal zone over more open sea localities is the potential presence of terrestrial organic material for dating and therefore problems associated with the marine reservoir effect can be avoided or reduced. Sufficient plant macrofossil material was present at both Norsminde and Kilen and therefore these stratigraphies are based upon AMS ^{14}C dates from terrestrial plant material. Korup Sø is an exception as the site was sampled and core material processed in the 1980s for molluscan, pollen and diatom analysis (Petersen, 1993, H. Krog, unpub.) with the original stratigraphy being based upon ^{14}C dates from shell material (see Table 5.2). Too little material remained from the original investigation to produce a stratigraphy based on terrestrial macrofossils. However chronological confidence has been increased (at all three sites), by comparing independently derived age-depth models to the palynological records to test the 'fit' of well-dated stratigraphical markers (e.g. *Ulmus* decline at ~3,900 BC; Andersen and Rasmussen, 1993 (H. Krog, unpub.). The age-depth models are discussed in more detail in section 2.5. In addition, there are always difficulties encountered when comparing records from different sites or archives (e.g. Lowe and Walker, 1997; Lowe *et al.*, 2001) often sampled at different resolutions and perhaps even dated via different methods. In some situations these difficulties can be overcome to some degree by using incremental age methods (e.g. tree rings), widely spread (i.e. regional) chronostratigraphic events or markers such as tephra (Turney and Lowe, 2001) or well dated ecological markers (e.g. *Ulmus* decline in Denmark, Andersen and Rasmussen, 1993).

High precision chronological control and high temporal resolution are essential for assessment of ecological and cultural responses to 'rapid' environmental changes (e.g. climate and sea level change) that might happen on decadal to centennial scales. For example, based on the rapidity of changes in the Danish shell middens (e.g. oyster decline and cultural transition from Ertebølle-Funnel Beaker Culture layers), it has been suggested that the Mesolithic-Neolithic transition might have taken place over as little as ~50-200 years (Rowley-Conwy, 1984; Andersen, 2008a). This is based on the clustering of dates between 4,000-3,800 BC for the Mesolithic-Neolithic transition in numerous shell middens (Andersen, 2008a) and the appearance of agricultural indicators in the pollen records after ~3,900 BC. Therefore, it is essential that the 'selected site' (and proxy) offers high enough temporal resolution to examine these events at the necessary temporal level. In this study, a sampling resolution of 25-32 years is maintained for each site in the period ~4,000-3,800 BC.

Close proximity to archaeological sites and settlements is extremely important for ‘direct’ assessment of how changes in the local and regional marine environment might have forced cultural/societal changes in the past (e.g. marine resource exploitation). Whilst deep sea cores (or sites distant from archaeological settings) might provide an overview of the major regional changes in the marine environment over the Holocene (e.g. Mörner, 1976; Jiang *et al.*, 1998; Gyllencreutz *et al.*, 2006), uncertainties will still exist as to how these changes affected conditions within the inner coastal areas, where the people actually lived and laboured. For example, many regional sea level (Mörner, 1976; Christensen, 1995; Berglund *et al.*, 2005; Behre, 2007) and temperature curves (Seppä and Birks, 2001; Antonsson and Seppä, 2007; Seppä *et al.*, 2009) now exist, but it is still poorly understood how environmental variables such as the sedimentary accumulation rate, salinity and productivity (which subsequently exert pressure on marine resources such as fish and shellfish) respond to rising/falling sea levels (or temperatures). Furthermore, changes in the local marine environment, perhaps driven by terrestrial forcing factors (e.g. reduced run off, natural vegetation restructuring, human impact) might also be important, but are unlikely to be detectable in distant, deep sea cores. Therefore, all sites analysed in this study are located nearshore and within the same marine system (i.e. Kilen within the Limfjord system, Korup Sø connected to the Kolindsund system; see Figure 2.2 and Figure 2.5) as important archaeological sites and settlements (and always less than 50 km away from a major shell midden; see Section 2.2 for more details). Norsminde is particularly advantageous as it contain a high quality sedimentary record (up until ~1,700 BC; Knudsen and Konradi, submitted; Rasmussen, submitted; P. Rasmussen, unpublished data) in the immediate vicinity of palaeosettlements (Andersen, 1983; Andersen, 1989; Enghoff, 1989).

For quantitative reconstruction of salinity using a diatom-based transfer function, presence and good preservation of fossil diatoms were also key pre-requisites for each site. Sites with poor preservation may exhibit a strong taphonomic signal, which will inevitably weaken inferences from the diatom salinity-model or potentially even override diatom response to important environmental changes (in more extreme conditions). As Kilen was the only new site sampled during the course of this thesis, a preliminary diatom preservation assessment was made on all core sections (at ~20 cm resolution) prior to any other core processing. The presence (and state) of diatoms in Norsminde and Korup Sø sediments had been assessed prior to this project and deemed suitable for analysis (D.B. Ryves pers comm., H. Krog, unpub). Preservation was monitored throughout the course of the diatom analysis at all sites via assessment of diatom valve dissolution (see Section 2.7.1) expressed as the sample **F** index (Flower, 1993; Ryves *et al.*, 2001; Ryves *et al.*, 2009).

2.2.2 Site descriptions

2.2.2.1 Kilen

Kilen (56º30’005”N 08º34’089”E) is today an ‘S’ shaped shallow eutrophic lake situated in the Limfjord region of northern Jutland in the Danish commune of Struer (Figure 2.1). Kilen is classified as a brackish water lake (~6 ‰), since being almost entirely cut off from the Limfjord by the building of a road and rail embankment in 1856 (AD), connecting the small towns of Bremdal and Knarbjerg to the north with the municipal capital Struer to the south. A connection with the Limfjord is retained via a small stream in the south-east corner called the Kilerkanal that is important for maintaining salinity levels. Since 1952, Kilen has been protected as a nature reserve and today is an important breeding ground for birds. The lake is approximately 5 km long, 1 km wide with a total area of 3.34 km^2, an average depth of 2.9 m and a max depth of 6.5 m (Windolf *et al.*, 1996; Jensen *et al.*, 2006). The deepest part of the lake is a small hollow situated very close to the southeastern shoreline, deemed unsuitable for palaeoenvironmental analysis after a preliminary core contained a record of less than 2 m before reaching underlying sands and gravels. For this reason a secondary test coring was made upon a plateau area (~4 m deep; see Figure 2.2) in the north-central part of the lake (water depth 3.9 m) and a much longer Holocene record was obtained (see section 2.4.1).

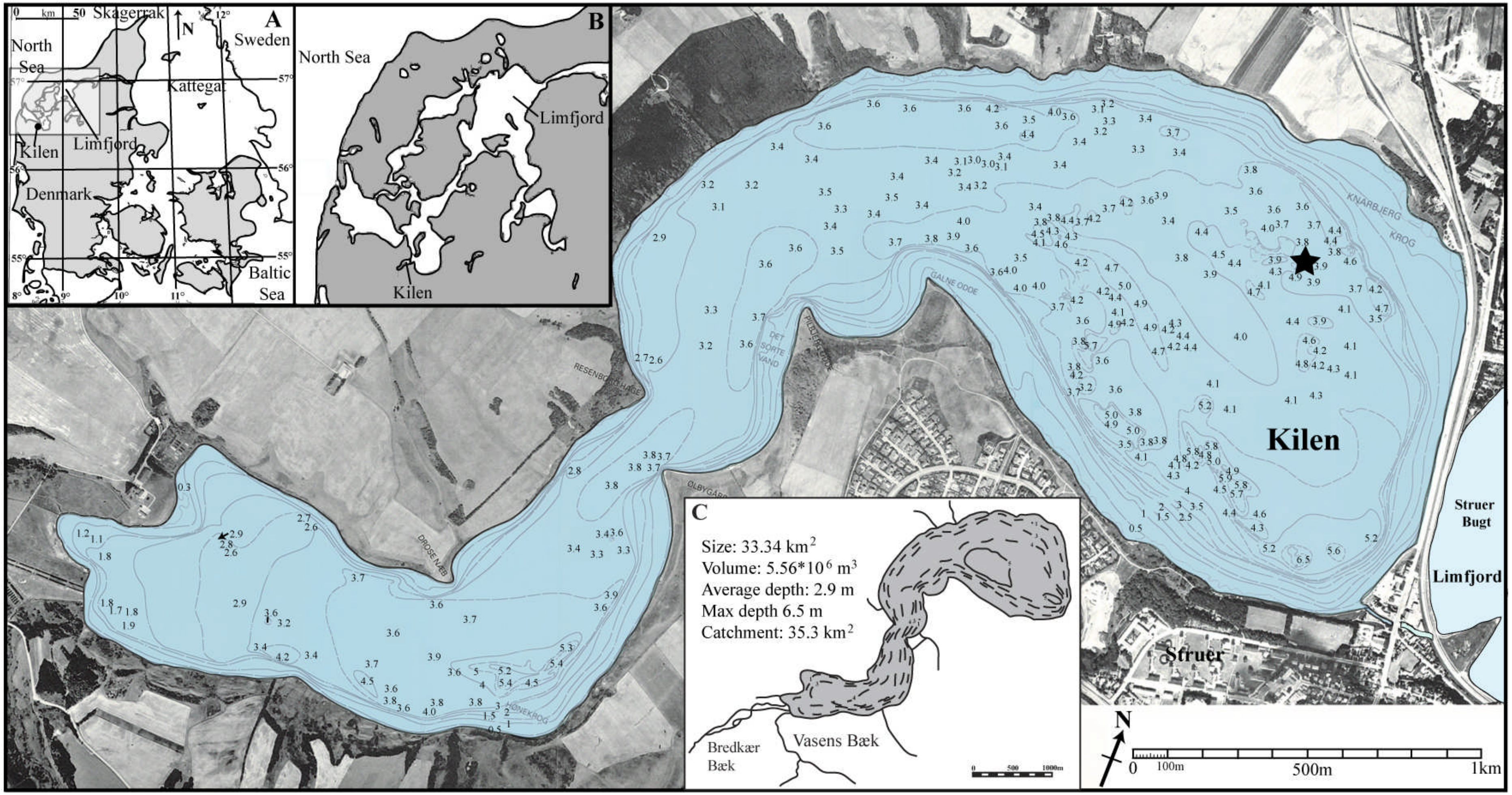


Figure 2.2 - Map of the Kilen sedimentary basin with bathymetry (originally produced by Thorkild Høy, 1989). Insets: A. Location map of Kilen and the Limfjord, B. Current configuration of Kilen and the Limfjord. C. Present fluvial inflows and basic limnological data for Kilen (Jensen *et al.*, 2006). Coring location is marked with a black star.

The Kilen catchment is relatively small (35.2 km^2) with Bredkær Bæk (Bredkær Brook), the main freshwater input entering the lake in the southwestern corner (and draining ~15.9 km^2 of the catchment). In addition, Vasens Bæk (Vasens Brook) and a couple of other minor watercourses also supply freshwater (Figure 2.2C). Despite the freshwater input, brackish conditions (6-7 ‰) are maintained in the lake via water exchange with Veno Bugt through the Kilerkanal in south-eastern corner of the lake. The introduction of a new sluice in 1986 increased the salinity in Kilen from ~2.5-4.5 ‰ (early 1980s) to ~7 ‰ in 1986-87, by improving the efficiency of water exchange. This new sluice system can be artificially regulated (i.e opened or closed) if necessary to protect the inner fish farms (Ringkjøbning Amtskommune, 1988). The catchment consists of agricultural land along its northern and south western shores, a spruce forestry plantation (Klosterhede) to the west and urban land lining the south east (Struer) and north east (Bremdal) shoreline. Substantial nutrient inputs from the surrounding agricultural land and fish farms together with long water retention times (266 days; Jensen *et al.*, 2006) has led to nutrient enrichment of the lake, culminating in Kilen being monitored between 1989-2003 (Table 2.1), as part of the Danish Nationwide Monitoring Programme (Kronvang *et al.*, 1993; Windolf *et al.*, 1996; Jensen *et al.*, 2006).

Reference	Water retention time (days)	Inlet P (annual) (mg L^{-1} TP)	Lake P (annual) (mg L^{-1} TP)	Lake P (summer) (mg L^{-1} TP)	Chlorophyll-*a* (summer) (µg L^{-1})	Secchi depth (summer) (m)
Jensen *et al.*, 2006	266	0.15	0.168	0.239	154	0.5

Table 2.1- Summary of Kilen monitoring data (averages for the period 1989-1996.

Previous palaeoenvironmental research or archaeological analyses

No previous palaeoenvironmental research has been undertaken at Kilen itself and for the entire Limfjord, only a handful of studies extend back beyond Roman times (Petersen, 1981; Andersen, 1992; Heier-Nielsen, 1992; Burman and Schmitz, 2005). This site was, however, included in a calibration dataset for a diatom-based salinity transfer function (Ryves *et al.*, 2004) which involved analysis of diatoms present in the surface sediment and the measurement of associated environmental conditions (salinity, depth, total nitrogen and total phosphorus) at the time of sampling or time averaged between AD 1996-2000. In terms of archaeology, Kilen does not contain any shell middens from the Mesolithic, Neolithic or Iron

Age, but is situated in relative close proximity to some of the key Limfjord shell middens (e.g. Krabbesholm and Hjerk Nor both less than 50 km away; Figure 2.1). Documentary sources (Holm and Bager, 2002; Poulsen *et al.*, 2007) and archaeofaunal fishbone evidence (Enghoff, 1986; Enghoff, 1991; Enghoff *et al.*, 2007) is also available for Limfjord sites in the close vicinity of Kilen.

2.2.2.2 Norsminde Fjord

Norsminde Fjord (56 º01'018"N, 10 º14'049"E) is an elongated basin situated along the east coast of Jutland about 20 km south of the city of Århus. It was formed as a valley during the last advances of the Weichseilian ice sheet by glacial meltwater erosion and inundated by seawater in the early Holocene ~6,800-6,900 BC (Rasmussen, submitted). Norsminde Fjord is a relatively small estuarine system (~6 km long, 250 m wide) predominately fed by one major inflow (Rævs Å or Rævs River) that drains a catchment area of ~101 km^2 (Nielsen *et al.*, 1995). Rævs Å (fed by several tributaries) is responsible for draining 85 % of the catchment before entering the fjord in the south-west at its innermost section (Figure 2.3). This creates a salinity gradient within the fjord with salinities of 0 ‰ freshwater occurring at the mouth of the Rævs Å, gradually increasing up to 24 ‰ at the Kattegat entrance (with most of the fjord around 18-20 ‰). The fjord lies in a SW-NE direction, entering Århus Bay and the Kattegat Sea through an artificially-maintained narrow (40-50 m) sluice established in 1964. The fjord has a surface area of ~1.68 km^2, an average depth of 0.6 m with a maximum depth of 2 m (see Figure 2.3) and a tidal range of 0.2 m (Nielsen *et al.*, 1995). It is estimated that during the higher sea level stands of the Atlantic period (7000-4000 BP: see Figure 1.2), the fjord was significantly larger (~10 km long, 2.7-3 km wide and ~500 m at the Kattegat mouth, Andersen, 1989; Rasmussen, submitted; Figure 2.4).

The subsequent lowering of sea level (Figure 1.2), together with the continuous build up of sand banks and beach ridges at its mouth (Andersen, 1989), has led to the isolation of previously claimed land (e.g. 'Kysing Fjord'; Figure 2.3 and Figure 2.4) and a reduction in the overall surface area of the fjord. The majority of the present day catchment is agricultural land, making Norsminde a moderately eutrophic fjord (primary productivity of plankton 200 g C m^{-2} yr^{-1}; Jørgensen and Sørensen, 1985) subject to occasional toxic planktonic blooms and suppression of natural seagrass (*Ruppia* spp.) and Charophyte populations by eutrophic macroalgae (e.g. *Ulva lactuca*). Nitrogen and phosphorus inputs have been reduced by 40 %

since the 1980s due to reduced run-off from agricultural land and improved waste water purification (Ellegaard and Skovgaard, 2007), though nutrient loads remain variable, particularly during years of higher precipitation (e.g. 1994). Norsminde is currently being monitored by the Integrated Protection of Surface and Groundwater in Agricultural Regions Plan (AGWAPLAN) and scenarios are in place for further reductions of nutrient inputs by 2012 and improvement of ecological status.

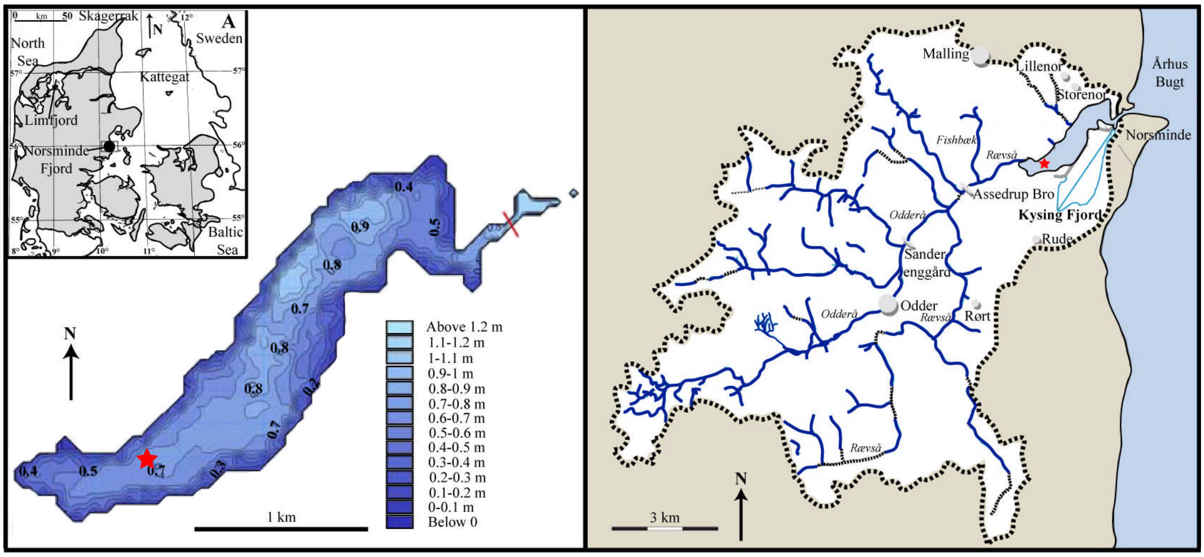


Figure 2.3 – Inset A: Location map for Norsminde Fjord. Norsminde Fjord bathymmetry (left) and present day hydrological catchment (right). The red star indicates the coring site. The reclaimed part of the fjord, named 'Kysing Fjord', is also indicated. Bathymmetric map from Ellegaard and Skovgaard, 2007, catchment hydrologcial map modified from Århus Amt, Natur og Miljø 1998.

Previous palaeoenvironmental or archaeological analyses

Prior to 1996, no palaeoenvironmental or palaeoecological research had taken place at Norsminde. The site has, however, been subject to extensive archaeological research which began in 1945 and continued intermittently up until 1989 (Andersen, 1989). Norsminde is now one of the most important (and thoroughly researched) Stone Age sites in Denmark, with its shorelines hosting a whole series of Mesolithic and Neolithic aged settlements (e.g. kitchen middens) supplemented by a large number of individual archaeological finds (e.g. artefacts, cooking implements, fishing tools) which clearly document the importance of the fjord and marine environments to Stone Age cultures and societies. It was during the latest major excavation (1972-1989) that the characteristic 'køkkenmødding' containing both Mesolithic and Neolithic layers (see section 1.6.1) was unearthed (Andersen, 1989). Prior to this, 'normal' kitchen midden sites (i.e. without the characteristic change in shell assemblage found across the Mesolithic-Neolithic transition) were uncovered at Flynderhage (Gabrielsen, 1953) and Norslund (Andersen and Malmros, 1966) of Ertebølle age, and at Kalvø, containing layers dated to the Neolithic Single Grave culture (Andersen, 1983).

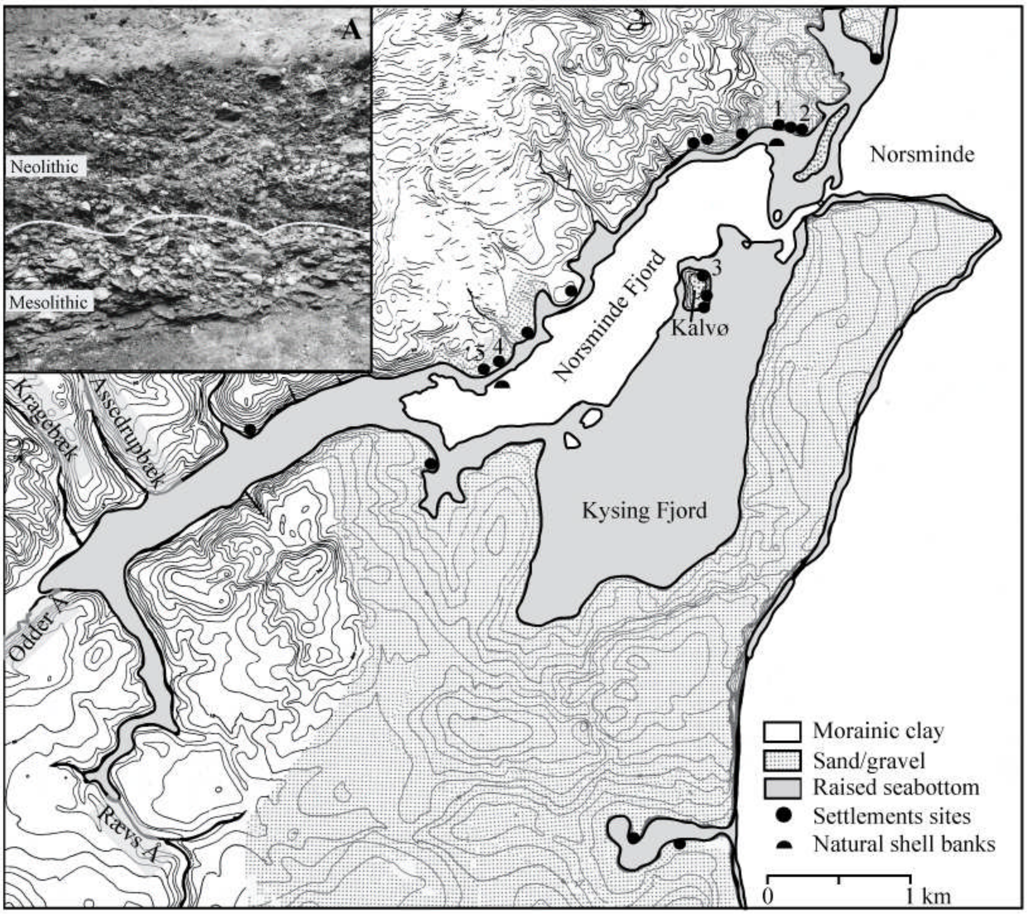


Figure 2.4 – Norsminde Fjord geology, topography and archaeological map (modified from Andersen, 1989). Numbered sites refer to key archaeological sites: 1. Norsminde køkkenmødding, 2. Frederiks Odde, 3. Kalvø, 4. Norslund, 5. Flynderhage. Inset A: section through the Norsminde kitchen midden with Mesolithic and Neolithic layers indicated (photo courtesy of Søren Andersern, Mosegard Museum).

2.2.2.3 Korup Sø

Korup Sø is situated in the central part of the Djursland peninsula, east Jutland, ~27 km northeast of the city of Århus (Figure 2.5). Today the site is terrestrial, existing as agricultural land positioned ~3 m above sea-level. The history of the site is somewhat complex, involving a series of limnic and brakish-marine phases influenced by eustatic sea level change, glacio-isostatic rebound and artificial draining. Korup Sø is a southward extension of the former Kolindsund fjord system (Figure 2.5), which was probably formed by meltwater and ordinary fluvial erosion during the Late Weichselian and Early Holocene. The Korup Sø basin itself was likely formed by a block of dead ice left by a retreating glacier (Pedersen and Petersen, 1997). The Littorina transgression eventually drowned Korup Sø thereby connecting it to the Kattegat via the Kolindsund to the east and Randers Fjord to the north-west whereby the northern part of Djursland became an island separate from Jutland (Figure 2.5; Jessen, 1920; Petersen, 1993; Pedersen and Petersen, 1997). It is estimated that the Kolindsund was a ~50 km long strait when the sea level was at its highest with the Korup Sø area being positioned along its southward arm in a relatively central position (Jessen, 1920; Petersen, 1993; Pedersen and Petersen, 1997). Kolindsund and Korup Sø remained brackish-marine until at least the end of the Stone Age but how long this lasted is unknown. Some historical records exist, suggesting that Viking ships might have entered Kolindsund around AD 1060 (Flou *et al.*, 1980). Due to the combination of a declining sea level during the Late Holocene and continued isostatic rebound, the connection between Kolindsund/Korup Sø and the sea was cut, and the former sound was transformed into a lake (area ~1.5 km^2), sometimes named Djurssø. At this stage (of unknown age) the Korup Sø basin became a freshwater lake again (known as Korup Sø). Between AD 1872 and 1880, Kolindsund/Djurssø including Korup Sø was reclaimed and turned into agricultural land (Flou *et al.*, 1980).

The aquatic phase of Korup Sø's history has received much attention being first investigated as early as the 1930's by Johannes Iversen who analysed a sediment sequence for pollen, diatoms and the so-called "Hystrix" (hystrichosphaerids, resting cysts for dinoflagellates). At the base of a 8.5 m long sequence Iversen identified Late Glacial limnic deposits superseded by limnic gyttja, referred to the late Boreal chronozone, followed by brackish-marine deposits (Iversen, 1937). The site was revisited in the 1980s by Harald Krog and Kaj Strand Petersen (both at DGU/GEUS) who analysed pollen, diatoms (H. Krog, unpublished data) and molluscs (Petersen, 1993) in two sediment sequences. Pollen and molluscan results from one of these corings (Well no. 1; Petersen, 1993) are reported in this thesis (see section 5.5.2 and 5.5.3) as well as new diatom analyses from raw material (Harald Krogs counts were unusable and the original slides were unavailable). Whilst no archaeological excavations have taken place from within the Korup Sø basin itself, the site is located close to a number of important archaeological sites and settlements (Figure 2.5), most notably Meilgård, Nederst, Dryholmen and the important Pitted Ware cultural sites of Kainsbakke and Kirial Bro (Troels-Smith, 1942; Rasmussen and Richter, 1991; Rasmussen, 1993a; Enghoff, 1994; Andersen, 2007).

Figure 2.5 - The Djursland peninsula with the location of Korup Sø and Kolindsund. The present day coastline (dotted black line) and the estimated coastline during the *Littorina* Sea stage (dark blue shading) are indicated (modified after Jessen, 1920; Rasmussen and Richter, 1991). Sites not in italics refer to important archaeological sites referenced in the text.

2.3 Material and Methods

This project incorporates a range of physical, biological and chemical techniques for high-resolution assessment of coastal environmental change at three sites in Denmark over the Holocene. Table 2.2 provides a complete record of all proxies analysed, indicating clearly the author's contribution. In this section, field work, core treatment and subsampling details are provided together with methodologies for all proxies included in this thesis. All stratigraphic diagrams presented in this thesis, detailing proxy results were produced using C2, version 1.6.2 (Juggins, 1991-2009).

2.4 Field work, core treatment and sub-sampling

2.4.1 Kilen

Two parallel boreholes (~1 m apart and vertically offset by 50 cm, labelled R3 and R4) were cored from the nearshore area of Kilen (see Figure 2.2) between 26th - 29th April 2007. For each borehole 1 m cores sections (15 for each borehole) were individually collected using Russian D-section peat samplers (Jowsey, 1966) with diameters of 10 cm and 5 cm diameter (both with 100 cm chambers). The water depth at the point of coring was 390 cm. A HON-Kajak sampler was also used to collect the very topmost section (390–426 cm below present sea level; bpsl) of the profile, which was sliced up into ½ cm intervals on site. The R3 borehole (starting at 400 cm bpsl) consisted of 15 core sections extending down to a maximum depth of 1,700 cm bpsl (Figure 2.6). The uppermost 9 core sections (400-1,200 cm bpsl) were retrieved using the 10 cm diameter Russian peat sampler and the remainder collected using the 5cm diameter sampler. The R4 borehole (starting at 450 cm bpsl) also consisted of 15 core sections extending the depth down to 1,950 cm bpsl. For the R4 borehole, the uppermost 10 core sections (450-1,450 cm bpsl) were collected using the 10 cm diameter Russian sampler and below this (1,450-1,950 cm bpsl) using the 5cm diameter sampler. All core sections were briefly visually assessed on site and any distinguishing features (e.g. shell layers, macrofossils) were recorded before being carefully wrapped up for storage in the GEUS cold store (at 4°C).

Site	Proxy	Analyst	No. of samples	Resolution
Kilen	Diatoms	**J. Lewis (LU)**	110	Average 68 years (M-N transition ~39 years, PWC ~60 years, PR/P ~70 years, Limnic ~7 years)
(~5,500 BC- present)	Sedimentary pigments	**J. Lewis (LU)**	777	Average ~10 years
	Isotopes (foraminifera - $\delta^{18}O$, $\delta^{13}C$)	**J. Lewis (LU)**	44	Average ~100 years, (M-N transition ~38 years, PWC ~46 years)
	Sedimentary physical parameters*	**J. Lewis (LU)**, B. Stavngaard (GEUS)	1,249 (JL) 150 (BS)	Average ~9 years
	Foraminifera	P. Kristensen (AU)	15	Average ~378 years
	Macrofossils (plants and other marine organisms)	P. Rasmussen (GEUS), **J. Lewis (LU)**	611	Average ~12 years
	Molluscs	K.S. Petersen (GEUS)	611	Average ~12 years
Norsminde	Diatoms	**J. Lewis (LU)**	76	Average ~66 years (LT ~18 years, M-N transition ~45 years, PWC ~43 years)
(~7,000- 1,700 BC)	Molluscs	K.S. Petersen (GEUS)	75	Average ~69 years
	Foraminifera	P. Konradi (GEUS)	77	Average ~65 years
	Pollen	P. Rasmussen (GEUS)	62	Average ~85 years
	Plant macrofossils	P. Rasmussen (GEUS)	74	Average ~70 years
	Sedimentary physical parameters *	B. Stavngaard (GEUS)		Average ~10 years
Korup Sø	Diatoms	**J. Lewis (LU)**	54	~100 years (M-N transition ~70 years, PWC ~76 years)
(~7,600-2,300 BC)	Molluscs	K.S. Petersen (GEUS)	65	Average 81 years
	Pollen	H. Krog (GEUS)	110	Average ~49 years

Table 2.2 - Details of all proxies analysed for Kilen. Norsminde and Korup Sø. Abbreviations: LT – Littorina transgression (7,000-6,600 BC), M-N =Mesolithic-Neolithic transition (4,500-3,500 BC), PWC = Pitted ware culture (2,400-2,800 BC), PR/R = Pre-Roman/Roman period (500 BC – AD 400), LU = Loughborough University, GEUS = Geological Survey of Denmark and Greenland, AU = Aarhus University. JL = Jonathan Lewis, BS = Beth Stavngaard. *= Includes organic matter, carbonate content, minerogenic matter, water content and dry mass.

All core sections were cleaned thoroughly on their outer surface and their lithologies described using a modified Troels-Smith system as described in (Troels-Smith, 1955; Birks and Birks, 1980). The Russian core sections from the R4 borehole were divided up into 2 cm depth intervals and separate ~1 cm^3 aliquots were subsampled for physical analyses (including water, organic matter and carbonate content), microfossil (pollen and diatom), and sedimentary pigments analyses, using a 1 cm^3 syringe. The material sub-sampled for physical analyses was placed directly into pre-weighed crucibles for drying (see Section 2.6.1), the diatom/pollen aliquots were freeze-dried to improve preservation (particularly of pollen grains) and the sedimentary pigment samples were placed straight into the freezer and stored at -20ºC until ready for analysis. The remainder of the material was retained for wet sieving, primarily to search for appropriate material for AMS ^{14}C dating but also for other analyses such as plant macrofossils, molluscs, foraminifera and ostracods (Table 2.2). Only two core sections from the R3 borehole (R3-C1; 400-500 cm & R3-C3; 500-600 cm) were used for the final analyses. Based on the distinctive lithological progression at the top of the profile (see Figure 3.1), R3-C1 and the upper half of R3-C3 (500-550 cm) were sliced up into 1cm intervals and the bottom half of R3-C3 (550-600 cm) into 2 cm intervals. The remaining overlapping R3 core sections (R3-C4 – R3-C14) were also sliced up into 2 cm intervals, but only the central 30 cm overlapping sections (with parallel R4 corings) were subsampled for physical (microfossil and pigment) analyses to attempt core correlation between the two boreholes. For diagrammatic representation see Figure 2.6.

In total, 625 samples (1-2cm thick, 610 of which were included in final depth profile), were wet sieved through 2 different mesh sizes (500 μm & 100 μm). The volume of sediment for each sample was quantified using the water displacement method, where prior to wet sieving the sediment was dropped in a known volume of water and the displacement volume was recorded. The recorded volumes for Kilen samples were between 1.5-45 ml, mean 21.1 ml. After wet-sieving, the two fractions were transferred into Petri dishes and potential AMS ^{14}C dating material (see below), molluscs (both fractions), ostracods (fine fraction) and any other plant or animal remains were picked out by hand.

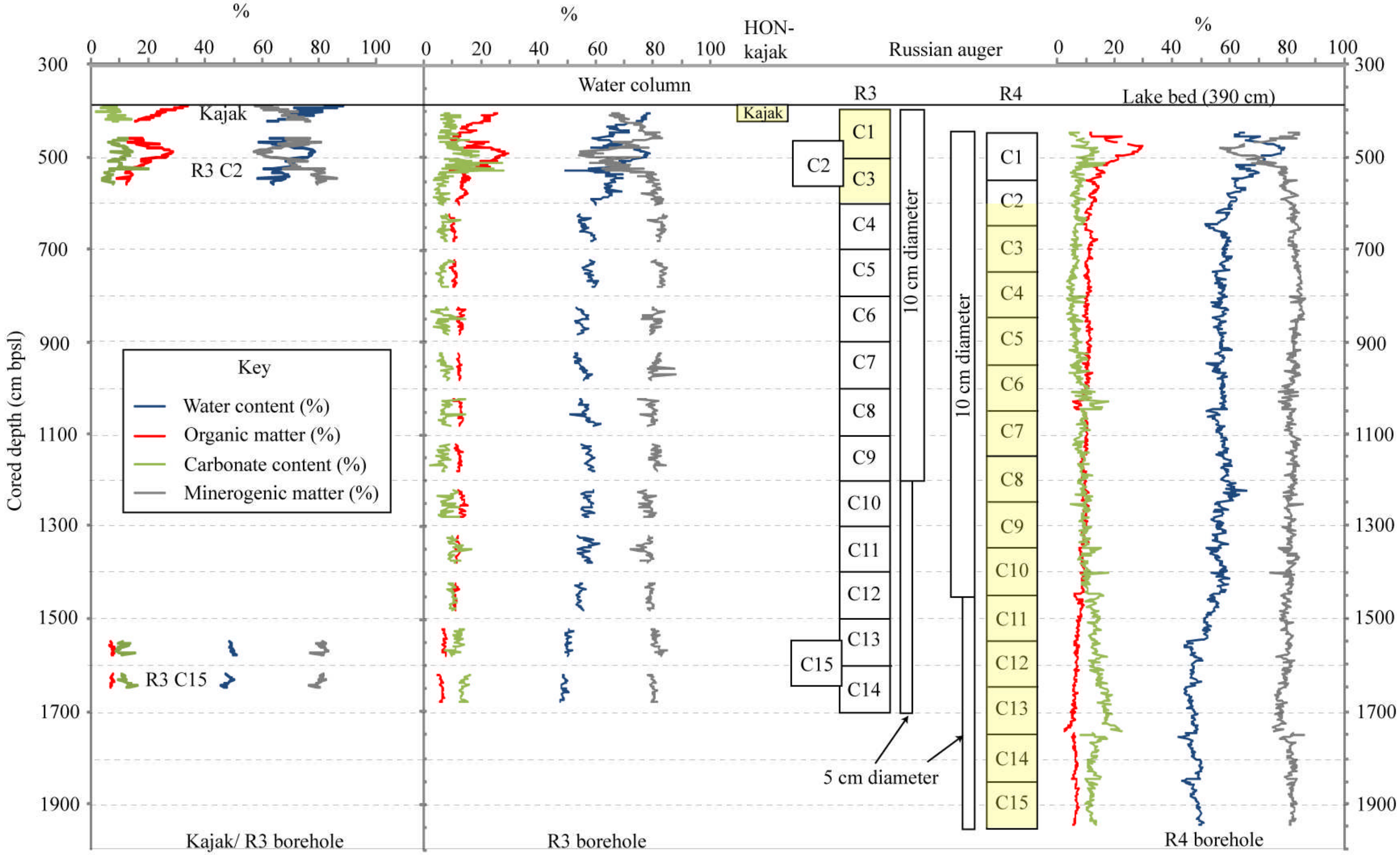


Figure 2.6 - Kilen coring details. The final sedimentary sequence (i.e. after core correlation) includes all core sections shaded yellow. Cores were correlated (where possible) using the sedimentary physical parameters (included in this diagram for all core sections and expressed percentages; see text for more details).

2.4.2 Norsminde

In 1996, two overlapping boreholes, ~1m apart were cored at Norsminde by Peter Rasmussen and Ole Bennike (GEUS). In total 25 cores sections (12 and 13 from boreholes 1 and 2 respectively) were collected using 2 variable-width Russian peat samplers (100 cm chambers, 5 cm and 10 cm diameters), together with a HON-Kajak corer to retrieve the uppermost section of the profile (31 cm). Alternating core segments from each borehole was cored at overlapping depth intervals (~20 cm) to try to account for any depth discrepancies. Cores were subsampled at 2 cm intervals for physical analyses (including water content, organic, carbonate and minerogenic using methods described below) and microfossil analyses (pollen and diatoms) by Beth Stavngaard (GEUS), before being sliced up for macrofossil analysis. The physical analyses were also performed by Beth Stavngaard using methods outlined below and overlapping core segments were correlated using water content and LOI data (P. Rasmussen pers. comm.).

In total, 74 samples (8-10 cm thickness, 79-393cm^3 volume; determined by water displacement method) were wet-sieved through 2 different mesh sizes (500 µm & 200 µm) and then analysed for plant macrofossil (and mollusc) remains. All identifiable plant macrofossils were recorded and any terrestrial material deemed suitable for AMS ^{14}C dating was picked out and frozen for preservation. Molluscan remains at Norsminde were very scarce, but those present were sent to Kaj Strand Petersen (GEUS) for identification (see Section 2.7.2). Any additional identifiable remains such as cladocerans, echinoid remains, barnacles, fish remains or ostracods were also picked out (and/or recorded). The finer fraction was retained for foraminifera analysis (performed by Peter Konradi, GEUS; see Section 2.7.3).

2.4.3 Korup Sø

Two boreholes were drilled in 1982/83 (labelled Well no. 1 and Well no. 2; Petersen, 1993). All analyses involved in this thesis (see Table 2.2 and Section 2.7) have been performed on material from Well No. 1, which extends back to ~7,600 BC (~9,550 BP) with a maximum depth of 880 cm. The age model is discussed in detail in section 5.3. The second borehole (Well no. 2) extended ~3 m further down (1,251 cm maximum depth), offering greater temporal resolution (Petersen, 1993), but shorter longevity than Well No. 1, yielding a maximum age of ~7,400 BC. Core correlation was not attempted between the parallel drillings due to the significant distance between boreholes and the absence of physical sedimentary analyses (i.e. water content, organic, carbonate and minerogenic content). It is important to note that some analyses (molluscs and pollen) were performed on samples from Well no. 2, together with the production of an age model based upon 7 ^{14}C dates on marine molluscs (see section 5.3) and the pollen stratigraphy.

2.5 Core stratigraphy, dating and chronology

2.5.1 ^{14}C dating; Kilen, Norsminde and Korup Sø

All samples for ^{14}C dating were submitted to the AMS ^{14}C Dating Centre at Aarhus University. The terrestrial plant macrofossil samples were pre-treated by 1M HCl at 80ºC for 24 hours, prior to being left for one to seven days in 1M NaOH at 80ºC. Samples were then treated in 1M HCl, before being converted to CO_2 by combustion (in sealed evacuated quart tubes containing CuO). Shell samples were subjected to thorough washing and leaching (outer 25% leached by HCl) prior to dehydrated phosphoric acid treatment (100%) for 8 hours at 25º C. Following liberation of CO_2, all samples were converted to graphite using the H_2 reduction method and AMS ^{14}C measurements were made using an EN tandem accelerator (under the supervision of Jan Heienemeier, Aarhus University). Results were reported as conventional radiocarbon dates (^{14}C yr BP, before 1950 AD; Stuiver and Polach, 1977) and calibrated into calendar years using the IntCal04 calibration curve (Reimer *et al.*, 2004) with a 400±50 year marine reservoir correction applied to shell samples (from Korup Sø).

2.5.2 ^{210}Pb and ^{137}Cs dating; Kilen

For dating of the uppermost section of the Kilen profile, a total of 23 freeze-dried samples (all 1cm homogenised slices selected from the Kajak core) were sent to Thjorbjørn Andersen at the Gamma Dating Centre, University of Copenhagen for analysis of ^{210}Pb, ^{137}Cs and ^{226}Ra activity. The activity was measured via gammaspectrometry, carried out on a Canberra low-germanium well-detector. ^{210}Pb was measured via its gamma-peak at 46.5 keV, ^{226}Ra via the granddaughter ^{214}Pb (peaks at 295 and 352 keV) and ^{137}Cs via its peak at 661 keV (T. Andersen, pers. comm.). Dates were calculated using a modified constant rate of supply model (Appleby, 2001). The final Kilen age-depth model was based on 13 AMS ^{14}C dates, together with a ^{210}Pb, ^{137}Cs and ^{226}Ra stratigraphy for the uppermost section of the profile using an OxCal 4.1 depositional model (Ramsey, 2008). To account for changes in accumulation rate, boundaries are inserted at 524, 1,055 and 1,740 cm. The position of the boundaries are based on large changes in the $CaCO_3$ content (J. Olsen, pers. comm. Figure 3.2).

2.5.3 Core Correlation

Kilen

In an attempt to eradicate any depth discrepancies during the coring process, physical parameters (dry mass/water content (%), organic matter (%), carbonate (%) and minerogenic matter (%); see Section 2.6) analysed on both the master profile and on overlapping sections of the parallel drive, were compared using wiggle-matching for best fit (Dearing, 1986). For Kilen, cores were correlated, where possible, based on organic matter (%) and dry mass/water content (%). This was only possible for the top section of the record (cored depths 390-650 cm), as below core R4-C2 (550-650 cm) only small magnitude (and non-systematic) changes occur in these parameters. The lithology of this section (i.e. 650-1,950 cm bpsl) is marked by homogenous marine clay gyttja (see section 3.2). Despite the presence of some shell layers throughout the profile, these appear randomly distributed, often not even extending across the whole diameter of an individual core, and therefore shell layers were not deemed suitable for cross-correlation.

In total, 13 samples (all terrestrial plant macrofossils; Table 3.2) from the Kilen profile were sent for AMS ^{14}C dating (at the AMS ^{14}C Dating Centre, Aarhus University). Preliminary pollen analysis (5 samples) was undertaken by Peter Rasmussen (GEUS) to identify the approximate location of the *Ulmus* (elm) decline (see Chapter 3). In some samples, the absolute number of pollen grains counted was low (282-500 grains) due to very low pollen concentration, probably as a result of the high sedimentation rate (~0.2 cm yr^{-1}). There was, however, sufficient evidence suggesting that the elm decline (~3,900 BC; Andersen and Rasmussen, 1993) fell between 1,337-1,361 cm bpsl. Accordingly the age of the elm decline in Kilen at 1,337-1,361 cm bpsl is dated to 3840-3910 BC and provides an independent check on the age model.

Norsminde

Overlapping core sections from Norsminde Fjord were correlated (by wiggle-matching) by Peter Rasmussen (GEUS) based upon the minerogenic profiles, measured at 2 cm intervals (see Section 2.4.2) on all core sections (Rasmussen, submitted).

Korup Sø

No attempt was made to correlate the core sections from the two Korup Sø wells, based on the large distance between the wells (i.e. 100 m apart), the large differences in sedimentation rate and the absence of data for any sedimentary parameters (i.e. water content, organic, carbonate and minerogenic content).

Age-depth models

Following the core correlation, ^{14}C dating and calibration processes, age-depth models were produced (by Jesper Olsen, Aarhus University, now at Queens University Belfast) using Oxcal 4.1 (for Kilen and Korup Sø; Ramsey, 2008) and B-peat 1.0 (for Norsminde; Blaauw and Christen, 2005). Both Bpeat and Oxcal incorporates Bayesian statistical modelling (allowing for sedimentation rate changes) to identify the most probable match for the complete sequence of dates by taking into account the stratigraphic information (e.g. changes in the sedimentary physical parameters) in the age calibration (Blaauw and Christen, 2005; Ramsey, 2008, 2009). Following age-depth model production and verification (using elm decline marker for example), sedimentary accumulation rates (expressed as cm yr^{-1}) for each site were calculated over the sediment sequences, together with sediment bulk density measurements (these form the basis for areal accumulation (flux) rates for individual sedimentary (i.e. organic matter, minerogenic matter, carbonate) and biological components (i.e. diatoms and sedimentary pigments).

2.6 Laboratory Analysis

2.6.1 Physical analyses: Kilen and Norsminde

The physical parameters analysed at both Kilen and Norsminde include dry weight (for calculation of water content; %), organic matter (%), carbonate content (%) and minerogenic matter (%). For Kilen, these analyses were performed on all samples from the R4 borehole and core sections R3-C1 and R3-C3 from the R3 borehole. For the remaining core sections from the parallel R3 borehole, only overlapping parts (Figure 2.6) were analysed for purposes of core correlation (see Section 2.5.3; Figure 2.6). Measurements of dry mass (for water content), organic matter and carbonate content for Norsminde Fjord follow the same methodology (outlined below), but were not undertaken by the author (see Table 2.2).

Unfortunately, no physical data exist for Korup Sø sequences, and insufficient material remained on which to perform the analyses.

After subsampling (see Section 2.4), the ~1 cm^3 aliquots of sediment were left to dry overnight in a standard laboratory oven at 105°C (in pre-weighed and pre-numbered crucibles). Samples were then cooled and re-weighed with the dry-weights (DW) recorded for calculation of percentage water content (Equation 1). For organic matter (%), samples were ignited in a stepwise manner (see Table 2.3 for ignition program) up to 550°C (Dean, 1974) at which temperature they were held for 60 minutes. Samples were then again left to cool at room temperature before being re-weighed and the remaining ignition residue weights were recorded and percentage organic matter calculated (Equation 2). For carbonate content, the samples were again ignited in a stepwise manner (Table 2.3) up to 925°C (Dean, 1974; Bengtsson and Enell, 1986) and held at the temperature for 15 minutes. Cooled samples were then weighed and recorded for calculation of calcium carbonate (%) content assuming all weight loss is CO_2 from the decomposition of $CaCO_3$ (Equation 3). The recorded weight of the remaining residue makes up the minerogenic component of the sample (converted to percentage of total weight, as in Equation 4).

$$\textbf{Eq. 1 - WC (\%) = ((WW – DW}_{105}\textbf{)/WW)*100}$$
$$\textbf{Eq. 2 – OM (\%) = ((DW}_{105}\textbf{ – DW}_{550}\textbf{)/DW}_{105}\textbf{)*100}$$
$$\textbf{Eq. 3 – CC (\%) = ((DW}_{550}\textbf{–DW}_{925}\textbf{)/DW}_{105}\textbf{)*2.27*100}$$
$$\textbf{Eq. 4 – MM (\%) = 100-(OM+CC)}$$

Where WC = water content, WW = wet weight OM = organic matter, CC = calcium carbonate content, MM = minerogenic matter, DW_{105} = Dry weight after heating to 550°C, DW_{925} = Dry weight after heating to 925°C.

Relative abundances for each component are expressed as percentages of the total wet weight (for water content and dry mass) or dry weight (for organic, carbonate and minerogenic components). Accumulation rates for each component have also been calculated (using the age-depth models, see Section 2.5 (and 3.3, 4.3, 5.3) and are expressed as mg cm^{-2} yr^{-1}.

Proxy	Stage	Rate	Temperature	Rest
Organic matter content	1	500 °C/hour	100°C	10 mins
(OM)	2	500°C/hour	300°C	10 mins
	3	500°C/hour	550°C	60 mins
	4	Cool		
Calcium carbonate	1	700 °C/hour	500°C	10 mins
(CC)	2	700 °C/hour	700°C	10 mins
	3	700°C/hour	925°C	15 mins
	4	Cool		

Table 2.3 Organic matter and carbonate heating programmes for loss-on-ignition. Resting phases are 10 minutes long unless otherwise specified.

2.7 Biological Proxies

2.7.1 Diatom analysis: Kilen, Norsminde & Korup Sø

All diatom slides from Kilen were prepared by the author at Loughborough University using the method outlined below (Battarbee and Kneen, 1982; Renberg, 1990). For Norsminde and Korup Sø, some of the diatom slides were prepared by Beth Stavngaard (GEUS), using the same method. Diatom samples were prepared at carefully selected depth intervals from the aliquots of material specifically subsampled for microfossil analysis (see section 2.4). For each sample, approximately 0.01 g (0.1 g for wet samples; several HON-Kajak core samples) of freeze-dried sediment was weighed out into labelled test tubes and left overnight in ~5 ml of concentrated hydrogen peroxide (30 % H_2O_2) at room temperature (~20°C). Samples were then prepared using hot hydrogen peroxide (30 % H_2O_2) and HCl (5-10 %) digestion and washing techniques as described by Renberg, (1990). For absolute diatom concentrations, microspheres were added to aliquots of each sample (Battarbee and Kneen, 1982).

A minimum of 300 valves were counted per sample (together with microspheres) from fields of vision along random transect lines using a Leica Leitz Laborlux S microscope at x1000 and under oil-immersion phase contrast. Diatoms were identified to species level where possible, using a range of taxonomic literature including: van der Werff and Huls, 1958-1974; Hendey, 1964; von Tynni, 1978; Krammer and Lange-Bertalot, 1986-1991; Snoeijs *et al.*, 1993-1998; Witkowski, 1994; Hasle and Syvertsen, 1996; Witkowski *et al.*, 2000; Lange-Bertalot, 2001; Levkov, 2009. Diatoms from the genera *Chaetoceros* and *Rhizosolenia* were recorded but excluded from the total count due to difficulties with identification to species level. This was due to *Chaetoceros* cysts being predominately found in their resting stage and *Rhizosolenia* valves as just remnant spines. Diatom taxonomy was also harmonised to MOLTEN/DEFINE concepts (http://craticula.ncl.ac.uk/Molten/jsp/) for application of the salinity transfer function to fossil diatom data (see below).

In order to assess the taphonomic process of dissolution, diatom valves were placed into one of two dissolution categories where possible; pristine or dissolved (following Ryves *et al.*, 2001). Valves in the pristine category were excellently preserved, showing very little or no signs of dissolution under light microscopy. Alternatively, valves exhibiting clear signs of the dissolution process were placed in the dissolved category. A number of valves could not be assessed accurately and were subsequently placed into an unassessed category. This category typically included valves that were partly covered up by other material (or other diatoms) in the slide, valves appearing in girdle view, or in the case of complete frustules in valve view, difficulties were often encountered assessing the state of the lower valve.

For *Paralia sulcata*, a 4-stage dissolution assessment was made on all valves that appeared in valve view for each site following Ryves *et al*., (2009). *P. sulcata* was deemed suitable due to its clear 4-stage dissolution process (see appendix A), general robustness (making it a good indicator for particularly poor preservation conditions) and its regular presence in sediments from Danish coastal sites (e.g. Mikkelsen, 1949; Ryves *et al.*, 2004; H. Krog, unpub., A.L.Clarke et al. unpub.). Additionally, the diameters of all *P. sulcata* valves were measured and then placed into one of two different size categories (<20 µm and >20 µm) as suggested by McQuoid and Nordberg, 2003. In McQuoid and Nordberg's observations, greater abundances of the smaller valves were found where water columns are vertically mixed and have more upwelling of nutrient-rich high salinity water. In contrast, larger sized valves were more abundant under lower nutrient (phosphorus) conditions, often in stratified waters with low surface water salinity.

2.7.2 Molluscan analysis: Kilen, Normsinde and Korup Sø

For Kilen and Norsminde Fjord, molluscan remains were picked out of all of the samples selected for wet sieving (from both the coarse and fine fractions; see Section 2.4) and were subsequently identified by Kaj Strand Petersen (GEUS). Molluscan nomenclature follows Petersen, (2004). For the Korup Sø profile, all molluscs were picked and identified from thick sediment slices (3-30 cm thick; 166-3,475 cm^3), covering the entire sequence (precise sieving details, including mesh sizes remain unknown). The Korup Sø molluscan assemblages have previously been reported in Petersen, (1993).

2.7.3 Foraminifera analysis; Kilen and Norsminde

Foraminifera samples (i.e. remainder of the finer 0.1-0.5 mm fractions following wet sieving and removal of molluscan and plant macrofossil material; see Sections 2.4.1 and 2.4.2) selected for analysis were concentrated by floatation in a heavy liquid (specific weight 1.8 g cm^3; cf Knudsen, 1998), prior to counting using standard techniques described in Feyling-Hanssen *et al.*, (1971) and Knudsen, (1998). A minimum of 300 tests were picked identified and counted where possible. Foraminifera nomenclature follows Ellis and Messina, (1949 and supplements).

2.7.4 Pollen and plant macrofossil analysis: Kilen, Norsminde and Korup Sø

Pollen samples were prepared using standard techniques described in Fægri and Iversen, (1995), from ~0.1 g aliquots of freeze-dried material. Kilen and Norsminde samples were counted by Peter Rasmussen (GEUS) using a Leitz Ortholux microscope at x320 or x800 magnification (phase contrast), and Korup Sø samples were counted by Harald Krog (GEUS). For plant macrofossils (Norsminde and Kilen only), all distinguishable remains (present from the wet sieved fractions; see above) were recorded and, where possible, identified by either the author (Kilen only) or Dr. Peter Rasmussen (GEUS; Norsminde and Kilen). Pollen and plant macrofossil nomenclature follows Tutin *et al.*, (1964-1980) (except for *Poaceae* = *Gramineae*). *Triticim* and *Hordeum* (cereal pollen) were distinguished using criteria outlined in Andersen, (1979).

2.8 Chemical analyses

2.8.1 Sedimentary Pigments: Kilen

Pigment analysis was performed by the author at the University of Nottingham, School of Geography. All sedimentary pigment procedures were carried out under subdued lighting to limit the degradation of pigments during the preparation process. Weighed aliquots of freeze-dried samples (~0.2 - 0.4 g) were placed overnight (in a freezer, -20° C) in 500 ml of extraction solvent (80 % methanol, 15 % methanol, 5 % deionised water) before being filtered to remove any remaining solid material. Samples were then dried under a nitrogen stream and dissolved in a known quantity (most commonly 500 µl) of HPLC injection solvent (70 % acetone, 25 % IPR stock solution (Table 2.4), 5 % methanol) and placed in small vials ready for injection into an Agilent 1200 series separation module with Quaternary pump (HPLC machine). Samples were loaded into an autosampler in batches (normally between ~15-25 samples, always including 2 ‘Green standards’; see Table 2.4). Each HPLC run lasted 52 minutes which included several separation phases using a modified method from that described in Chen et al., 2001; separation conditions for each phase are summarised in Table 2.5). Eluted pigments were measured by passing through a photo-diode array detector which detects UV-visible spectral data every 2 seconds. Pigment concentrations were calculated using the package Agilent Chemtstation and output as chromatogram reports which were interpreted manually. Pigments were identified based on their retention time and absorbtion spetra compared with published literature (e.g. Jeffrey et al., 1997) and authentic standards. Sedimentary pigment concentrations are expressed as nmol/g organic matter and pigment accumulation rates as nmol cm^{-2} yr^{-1}.

Solution	Details	Phase	Method
Extraction solvent	80 % acetone, 15 % methanol, 5 % deionised water	Pigment Extraction phase	This solvent is to be made in small batches and kept in a screw cap bottle. Use HPLC-grade solvents (acetone and methanol) and nanopure deionised water and combine in the stated ratios.
Injection Solution	700 ml acetone, 50 ml methanol, 250 ml IPR stock solution (see below)	Injection phase	In a 1L volumetric flask combine a 70:25:5 mixture of acetone, IPR, and methanol. Transfer the completed solution to a foil wrapped 1L ground glass reagent bottle and store in the freezer. De-gas the solution in a sonicator before using in the HPLC machine.
HPLC solution A	80 % methanol, 20 % of 0.5 M ammonium acetate, aq. pH 7.2	Mobile phase A	Dissolve 7.708 g ammonium acetate in 200 ml deionised (HPLC-grade) H_2O. Add the 200 ml 0.5 M ammonium acetate to a 1 L volumetric flask and top up to 1 litre with HPLC-grade methanol. De-gas for 5 minutes in a sonicator before use.
HPLC solution B	90 % acetonitrile, 10 % deionised water	Mobile phase B	In a 1 L measuring cylinder, measure out 900 ml acetonitrile and top up with 100 ml deionised HPLC-grade H_2O. De-gas for 5 minutes in a sonicator before use.
HPLC solution C	100% HPLC-grade ethyl acetate		De-gas for 5 minutes in a sonicator before use.
Ion-pairing reagent (IPR) stock	To make 250 ml: 1.875 g of tetra butyl ammonium acetate, 19.25 g of sigma grade ammonium acetate, 250 ml volumetric flask with HPLC-grade H^2O	Injection phase: used for production of the injection solution	This solution must be made fresh each time it is used. Weigh out 0.75g of tetra butyl ammonium acetate and 7.7g of Sigma-grade ammonium acetate and rinse into a 100 ml volumetric flask with HPLC-grade H_2O. The chemicals are hygroscopic and temperature sensitive and therefore must not be allowed in contact with the air for any length of time. Reseal the containers and return to the freezer immediately. Swirl the flask to completely dissolve the chemicals. Now add enough H_2O to bring the volume up to the 100 ml mark.
Green standard			Grind a handful of grass or other green plant and a few ml of extraction solvent in a mortar and pestle. Filter the collected material using the procedure outlined in the extraction of pigments section. Dry down the concentrated pigment under nitrogen. Add injection solution to the dry pigments until the colour turns a light green. Label the vial as “Green Standard” and store in the freezer.

Table 2.4 – Summary of the stock solutions used during the extraction and preparation stages of sedimentary pigment analysis. (Modified from HPLC laboratory handbook, Dr. Suzanne McGowan, University of Nottingham).

Time (minutes)	Solvent A (80:20 methanol: 0.5 M ammonium acetate) - %	Solvent B (9:1 acetonitrile: water) - %	Solvent C (ethyl acetate) - %	Flow (ml min^{-1})
0	100	0	0	1
4	0	100	0	1
38	0	25	75	1
39	0	25	75	1
43	100	0	0	1
52	100	0	0	1

Table 2.5 - Separation condition used in the analytical procedure (method modified from Chen *et al.*, 2001). Stationary phase (column) consisting of a Thermo Scientific ODS Hypersil column (205*4.6 mm; 5 µm particle size).

2.9 δ^{18}O & δ^{13}C isotope analysis of single species of foraminifera: Kilen

Isotopic analyses were performed at the NERC Isotope Geosciences Laboratory (NIGL), Keyworth, UK. Material was once again prepared using exactly the same method as above (Section 2.4.1), but with the selected isotope samples being immediately filter dried after wet sieving. Isotopic analyses were performed on tests of *Elphidium excavatum* fo. *selseyensis*, chosen due its regular presence and often high abundance throughout the profile, particularly through some of the key periods of archaeological and environmental change (e.g. the Mesolithic-Neolithic transition, ~3,900 BC; see Chapter 3). *Elphidium excavatum* fo. *selseyensis* has also been used in other isotopic studies (e.g. Scheurle and Hebbeln, 2003; Kristensen and Knudsen, 2006; Schweikhardt *et al.*, 2010) which have demonstrated its ability to record changes in salinity and/or temperature sensitively.

From selected samples (44 in total), between 5-20 individuals of *Elphidium excavatum* fo. *selseyensis* were picked out and placed into labelled micropaleontology slides. Tests were cleaned as much as possible by hand during the picking process with any attached matter visible being removed. Specimens from each sample were then individually transferred to small Eppendorf vials until a desired weight was achieved (30-100 µg). Samples were analysed using a GV IsoPrime mass spectrometer plus Multiprep device and Isotope values (δ^{13}C, δ^{18}O) are reported as parts per mille (‰) deviations of the isotopic ratios (^{13}C/^{12}C, ^{18}O/^{16}O) calculated to the VPDB scale using a within-run laboratory standard calibrated against NBS standards. Analytical reproducibility of the standard calcite (KCM) is < 0.1 ‰ for δ^{13}C and δ^{18}O (M.J. Leng pers. comm.).

2.10 Numerical analyses

2.10.1 Statistical zonation of sedimentary and biological datasets

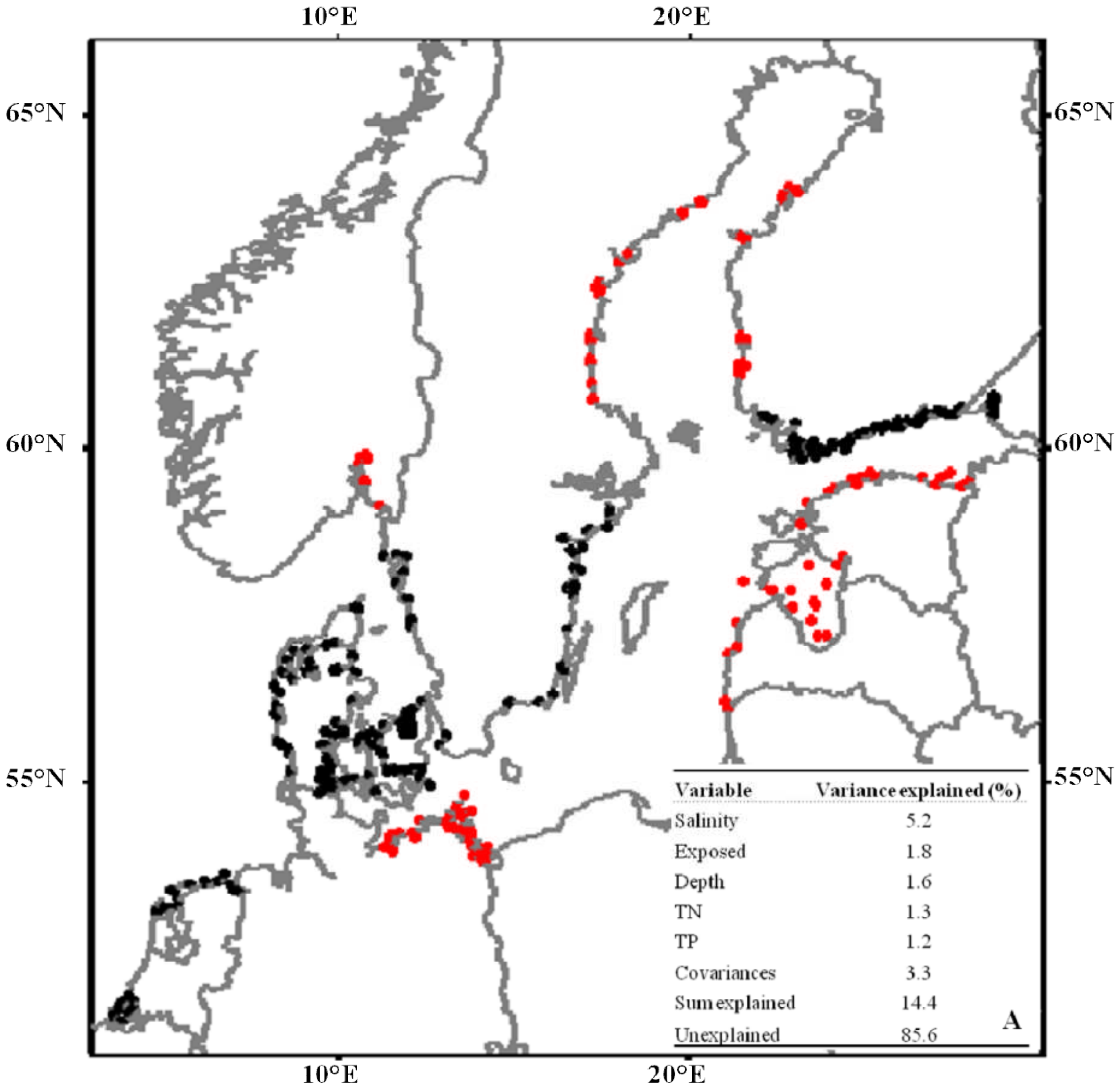
All biological (and sedimentary pigment) zones were determined using the package PsimPoll 4.27 (Bennett, 2003-2009) using an optimal splitting method based on an information content measure of total variance. This splitting method was chosen over agglomerative methods (e.g. CONISS, CONSLINK; Grimm, 1987) because of its ability to deal with transitional phases better. Significant splits were identified using a broken stick model (Bennett, 1996). However, it is important to note that when using optimal splitting techniques, the broken stick method can only help with identification of zones that are definitely significant. It cannot, positively determine if a split is non-significant (Bennett, 1996). Therefore, in some cases, additional optimal splits (not deemed significant by the broken stick method) were included as sub-zones, but only if they did not change the preceding significant split boundaries. In these circumstances, additional sub-zones were introduced if they aided the ease of interpretation (subjective measure).

2.10.2 Quantitative reconstructions of salinity: the MOLTEN salinity transfer function

Salinity at each site was inferred quantitatively using a diatom-based transfer function produced by Dr. Annemarie Clarke (Newcastle University, GEUS and APEM Ltd) and Dr. David Ryves (Loughborough University) from a subset of sites sampled during the MOLTEN and DEFINE projects (Clarke *et al.*, 2006; Andrén *et al.*, 2007a; http://craticula.ncl.ac.uk/Molten/jsp/). A combined total of 341 sites (i.e. 217 MOLTEN sites and 124 DEFINE sites; Andrén *et al.*, 2007a) were analysed for diatoms present in surface sediment samples together with measurement of associated environmental conditions (including water depth, Secchi depth (m), salinity (‰), ammonium, nitrate, nitrite, chlorophyll *a*, total nitrogen (TN), total phosphorus (TP), phosphate (PO4-P) and silica (all μg L^{-1}). Surface sediments were collected using a HON-Kajak corer (Renberg, 1991) with the top 1 cm being prepared for diatom analysis. A minimum of 500 valves were counted per sample, excluding *Chaetoceros* cysts and *Rhizosolenia* spines.

Salinity is the single most significant environmental variable explaining the distribution in the full MOLTEN/DEFINE diatom dataset (i.e. all sites, Andrén *et al.*, 2007a; Figure 2.7). The geographic expanse of the dataset, however, meant a number of other environmental gradients were also captured, and that other environmental variables are also important in explaining small amounts of the variation in the full diatom dataset (most notably depth, total nitrogen (TN), total phosphorus (TP) and exposure; Figure 2.7). The variation explainable by salinity alone was increased by constructing a smaller training set, removing sites more strongly associated with other environmental parameters (e.g. heavily eutrophic sites). The main gradients in the diatom species dataset were explored using a series of cluster and ordination techniques (e.g. detrended correspondence analysis (DCA) and non-metric multidimensional scaling (NMDS); A.L. Clarke *et al.* unpublished; and see Andrén *et al.*, 2007a). Principal component analysis (PCA) was used for investigation of environmental gradients and canonical correspondence analysis (CCA) and variance partitioning was used for examining relationships between the diatom distributions and the environmental variables (CCA deemed appropriate due to unimodal distribution of species data, as determined by DCA; DCA axis 1 for species dataset >2 SD units). Within CCA, forward selection was applied to identify significant variables (i.e. $p < 0.05$; 999 permutations), with Bonferroni corrections (for the nth variable to be accepted $p_n = p_1/n$, where $p_1 = 0.05$) until variables fell below the corrected significance value and were subsequently omitted from further analyses. Forward selection identified salinity, depth, TN and TP to be the most important (and significant) variables. The final salinity subset includes 318 diatom species and 210 sites spanning a salinity gradient of ~0-31 ‰ (A.L. Clarke, *et al.* unpub.), with salinity the most significant variable, explaining 6.58 % percent of the variation in the diatom species data.

Prior to application of the transfer function (for salinity reconstruction) to the fossil diatom datasets, the predictive power of a series of weighted average, weighted average-partial least squares (WAPLS) and modern analogue technique (MAT) models were tested. The predictive power of these models were evaluated using r^2_{boot} and RMSEP scores, following internal validation (via bootstrapping x1000 cycles) using the modern training set. The WAPLS model performed best and final diatom inferred (DI) salinity reconstructions are based on a component-3 model (r^2_{boot} =0.923 and RMSEP =0.36 square root salinity units. The component-3 model was selected because it significantly (p=0.001) improved the RMSEP (2.6 %), compared to the simpler component-1 and component-2 models.



Variable	Variance explained (%)
Salinity	5.2
Exposed	1.8
Depth	1.6
TN	1.3
TP	1.2
Covariances	3.3
Sum explained	14.4
Unexplained	85.6

Figure 2.7 - Map of MOLTEN/DEFINE sites including both water chemistry and diatom counts. Red dots indicate MOLTEN sites and black dots indicate DEFINE sites. Inset A: Variance partitioning of the full MOLTEN/DEFINE dataset (i.e. 341 sites) using CCA. (From Andrén *et al.*, 2007a).

2.10.3 Multivariate analysis and environmental-species relationships

Major trends in the biological datasets, relationships between the different proxies and land-sea interactions (i.e. modelling predictor-response scenarios) have been explored using a series of multivariate ordination techniques (including detrended correspondence analysis (DCA), principal component analysis (PCA), redundancy analysis (RDA) and canonical correspondence analysis (CCA)) using CANOCO v4.5 (ter Braak and Šmilauer, 2002). For all ordinations using linear methods (i.e. PCA and RDA), as the response (species) data were % abundances Aitchison’s log-ratio analysis (Aitchison, 1986) was applied, with samples and species data both centred. For unimodal methods (i.e. DCA, CCA) rare species were always downweighted. Prior to ordination analyses, the frequency distributions of variables included within the environmental predictor datasets were examined using Calibrate 0.81 (Juggins and ter Braak, 1997). In cases where distributions were positively skewed (generally with a few large values), the gradient length was deemed too long, data were log transformed to reduce the effect of high value outliers. Further details are given in the appropriate sections of results chapters.

2.11 Summary

- In this study, three coastal sites (Kilen, Norsminde Fjord and Korup Sø) were selected for palaeoenvironmental analyses on the basis that they matched several important criteria: 1. the presence of terrestrial macrofossils for AMS ^{14}Cdating and/or a detailed pollen record for verification of the age-depth model, 2. good chronological control, 3. offered high temporal resolution, 4. situated in close proximity to important archaeological sites and settlements, and 5. good preservation of fossil diatoms.
- Kilen is a small brackish lake (presently eutrophic) situated near to the town of Struer at the Limfjord, north-west Jutland. Up until ~AD 1856, it was directly connected to the Limfjord, but has since been cut off by the building of a road and rail embankment. A narrow connection still exists in the south-east corner (termed the Kilerkanal), through which water exchange between Venø Bugt and Kilen occurs, maintaining brackish water salinities (~6 ‰). In 2007, a ~15 m long sediment sequence was retrieved at a water depth of 390 covering the period from ~5,500 BC-present day (dated via AMS ^{14}C on terrestrial plant material and ^{210}Pb/^{137}Cs dating techniques, with age model produced in Oxcal v. 4.1).
- Norsminde Fjord is a small, shallow fjord system (average depth 0.6 m, maximum depth 2 m) situated on the east coast of Jutland. Today it has an average salinity of ~18-20 ‰, (gradient from 0 ‰ in the west at the mouth of Rævs Å, to 24 ‰ in the east where it opens into the Kattegat Sea). In 1996, a 1,180 cm core sequence was retrieved covering the time period ~6,900-1,700 BC (dated via AMS ^{14}C dating using terrestrial plant macrofossils, with age model produced in B-peat 1.0), with the most recent ~3,700 years lost due to a sedimentary hiatus.
- Korup Sø is a former fjord system (previously forming a southward extension of the Kolindsund Fjord) situated in Djursland, eastern Jutland. Today, the Korup Sø basin is agricultural land, lying ~3 m above sea level. In 1982 and 1983, two core sequences were retrieved, the longest of which (in temporal terms and used in this study) covers

the period ~7,600-2,300 BC (880 cm sediment sequence, dated by ^{14}C dating on marine molluscs with age model produced in Oxcal v. 4.1), with the most recent ~4,300 years again lost due to a sedimentary hiatus.

- A series of palaeoenvironmental techniques have been employed to reconstruct environmental changes over the study period (with higher resolution focus upon the archaeological shell midden accumulation phases). These include sedimentary physical parameters (lithological descriptions, organic matter, carbonate and minerogenic matter; Kilen and Norsminde Fjord), diatoms (all sites), isotopes from the tests of *Elphidium excavatum* fo. *selseyensis* (Kilen), sedimentary pigments (Kilen), foraminifera (Kilen and Norsminde Fjord), molluscs (all sites), pollen (Norsminde Fjord and Korup Sø), plant macrofossils (Kilen and Norsminde Fjord). The methodologies for all these techniques have been outlined in this chapter (and the analysts involved clearly referenced).
- A diatom-based salinity transfer function has been developed from the MOLTEN and DEFINE training set (consisting of 210 sites, 318 species spanning a salinity gradient 0-31 ‰). A three-component WAPLS model performed best under internal validation (via bootstrapping x1000 cycles) and has been applied the fossil diatom data for each site, to quantitatively reconstruct changes in salinity.

Chapter 3. Results: Kilen

3.1 Introduction

In this section results from multiproxy palaeoenvironmental analyses performed on sediment cores from the Limfjord site, Kilen, are presented. The analyses are presented in logical progression, starting with lithology, chronological control and proxies reconstructing the physical and sedimentary environment before progressing onto the biological and chemical remains of past communities living within the fjord over the last ~7,500 years. The chapter finishes with a synthesis and discussion section, integrating the results from the different proxies in order to critically assess how the Kilen environment has changed over the study period. Specifically this chapter aims to address the following questions:

1. How has chronological control been achieved at this site and how accurate is the dating?
2. How has the physical and sedimentary environment at Kilen changed over the last ~7,500 years?
3. How has the salinity, productivity and nutrient status of the fjord changed over the last ~7,500 years?
4. How has the Kilen diatom flora changed since the building of the road/rail embankment in AD 1856 and its transition from a marine to limnic environment?
5. What are the most important environmental parameters driving the changes in the biological communities and can their contributions be quantified?
6. Can these different environmental parameters be completely isolated from one another to produce accurate and independent reconstruction for these variables?
7. To what extent do the key environmental variables driving the changes in the biological communities interact (e.g. co-vary or drive one another) and how do these relationships vary over time?
8. To what extent do the different proxies employed to reconstruct the natural environment of Kilen agree/disagree with one another?

3.2 Lithology

The Kilen lithological profile is summarised in Figure 3.1 and the lithological descriptions based on a modified Troels-Smith system (Troels-Smith, 1955; see also Birks and Birks,

1980) are provided in Table 3.1. No description is available for the section collected using the HON-Kajak gravity corer (390-425 cm bpsl) as this was sliced up into half centimetre intervals on site, straight from the coring tube. For this section, information from a correlated parallel core is also provided in Table 3.1 to give an indication of the lithology, though exact boundaries must be treated with caution. No overlap is available for the uppermost 10 cm of the profile (390-400 cm bpsl).

Below this, the profile is entirely composed of marine gyttja containing varying amounts of silt, clay, plant matter and shell material. Slight colour changes are identifiable throughout the profile and occasional narrow shell layers are visible (see Table 3.1). The uppermost sections are generally browner in colour probably due to the higher organic content (see Figure 3.5) and occasional intervals containing dy (e.g. top-434 cm, 476-517 cm) and more humified material (e.g. 400-417 in parallel core). Below 434 cm the sediment becomes more clayey and below 524 cm there are higher levels of silt. In all cases, the transition between layers is very gradual (i.e. Limes superior = 0 according to the Troels-Smith system).

The abundance of visible shell fragments increases below 465 cm, with the uppermost section being quite barren (very few shell fragments between 425 and 434 cm and no visible shell fragments between 434 and 465 cm). The shell density is particularly high between 497 and 524 cm with some shells (e.g. cockles) even found sitting in living position. Below this, the shell concentration decreases again but scattered shells (both whole and fragmented) remain present throughout, occasionally in clusters or sometimes even in narrow layers (see Table 3.1, Figure 3.1). Superimposed upon this general trend is another short-lived phase where shell density increases between 1033-1041 cm. Below 1,753 cm, there is a further drop in concentration. Plant fragments are less abundant than shells, generally occurring sporadically throughout the profile (e.g. twig fragment at 540 cm). An exception to this occurs between 476 and 517 cm, where plant material becomes quite abundant for a short period of time.

Depth (cm bpsl)	Lithology Description
0-390	Water.
390-425	No description.
425-434	Dark brown gyttja containing dy. No shells visible. Limus detrituosus (Ld)^{2}3. Sand, silt, clay (SSC) - 1. Limes superior (LS; upper boundary) - 0 (i.e. boundary area >1 cm).
434-443	Dark brown greenish grey clay gyttja (lighter than above layer and more minerogenic). Very few shell fragments. Ld$^{1-2}$ 2. SSC – 2. LS - 0.
443-465	Dark brownish grey clay gyttja (again lighter than above layer). Very few shell fragments. Ld$^{1-2}$ 2. SSC - 2. LS - 0.
465-476	Dark brownish grey clay gyttja (darker than above layer and more shell fragments). Ld$^{1-2}$2. SSC - 2. LS - 0.
476-497	Dark-brow blackish clay gyttja containing dy and many plant fragments and more dispersed shell fragments. Gets more clayish towards top. Below 486 cm – more pure gyttja. Concentration of plant material higher. Ld^{2}3, Detritus granosus (Dg) 1. LS - 0.
497-517	Dark brown blackish clay gyttja containing dy, some plant fragments and many shell fragments (including whole shells sitting in living position. *Cerastoderma* spp. (cockle) in living position at boundary to layer below. Ld^{2}3. SSC - 1, Dg ++. LS - 0.
517-524	Dark brown blackish clay gyttja (lighter than above layer) containing many dispersed shells and more minerogenic matter. Transitional layer. Ld$^{1-2}$2-3. SSC - 1-2. LS - 0.
524-597	Dark grey homogenous silty clay gyttja with a few scattered shells (whole and fragments). Twig fragment at 504 cm. Ld^{1}1-2. SSC - 2-3, Dg +. LS - 0.
597-653	Dark grey (slightly lighter than above layer) homogenous silty clay gyttja. Scattered shells (whole and fragments). Probably more minerogenic material. Ld^{1}1-2. SSC - 2-3. LS - 0
653-753	Dark grey-brown homogenous silty clay gyttja with a few scattered shells (whole and fragments). Argilla steatodes (As) +, Argilla granosa (Ag) - 3, Ld 1. LS - 0.
753-853	Dark grey-brown homogenous silty clay gyytja with a few scattered shells (whole and fragments). As +, Ag - 2-3, Ld 1-2. LS - 0.
853-953	Dark grey-brown homogenous silty clay gyttja with a few scattered shells (whole and fragments). As +, Ag - 3, Ld 1. LS - 0.
953-1,253	Dark grey homogenous silty clay gyttja with a few scattered shells (whole and fragments). Shell fragments more abundant between 1033-1041cm. Shell layers at 1,101-1,103cm and 1,149-1,151cm. Concentration of shells at 994-996 and 1,109-1,111 cm. As + Ag 2-3, Ld 1-2. LS - 0.
1,253-1,453	Dark grey homogenous silty clay gyttja with a few scattered shells (whole and fragments). Slightly more than in cores RC11-RC15 (1,453-1,952 cm bpsl). Shell layers at 1,324-1,326, 1,363-1,365, 1,407-1,408 cm. Pebble at 1,425-1,426 cm As +, Ag – 3, Ld 1. LS - 0.
1,453-1,753	Dark grey homogenous silty clay gyttja with a few scattered shells (whole and fragments). As +, Ag - 3 Ld 1. LS - 0.
1,753-1,953	Dark grey homogenous strongly silty clay gyttja with very few scattered shells (whole and fragments). As +, Ag - 3, Ld 1. LS - 0.
Correlated to 400-417 cm (From core ref. R3 C1).	Dark brown-blackish gyttja containing dy with a little sand, silt and clay. No shells visible. Ld^{2}4, SSC +++. LS – 0.
Correlated to 417-425 cm (From Core ref R3 C1)	Dark brown gyttja (lighter than above layer – slightly greener) containing dy and a little sand, silt and clay. No shells visible. Ld^{2}3, SSC – 1. LS – 0.

Table 3.1 - Lithological descriptions for the Kilen sedimentary profile. Descriptions are based on a modified Troels-smith system (Troels-Smith, 1955; see also Birks and Birks, 1980)

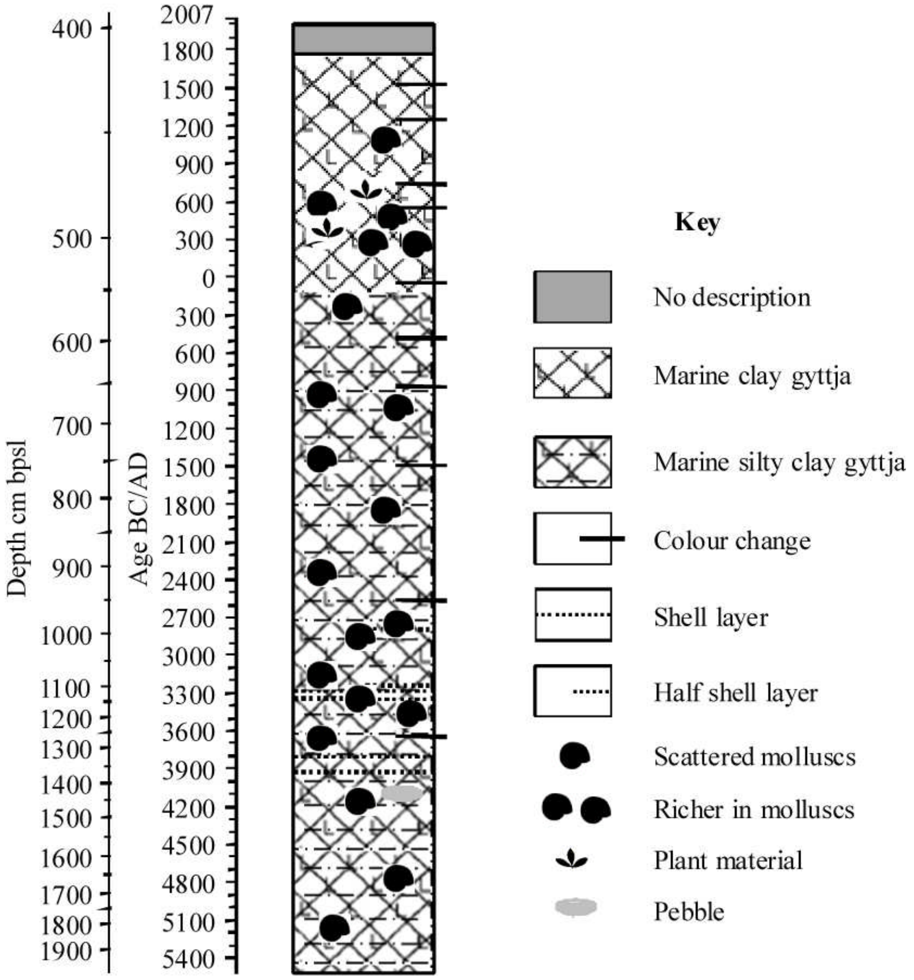


Figure 3.1 - Kilen lithological profile.

3.3 Chronology and Age Model

The Kilen chronology is based on 13 AMS ^{14}C dates together with a lead (^{210}Pb) and caesium (^{137}Cs) stratigraphy for the uppermost section of the record (since ~AD 1920). All AMS ^{14}C dates are derived from terrestrial plant macrofossil material (Table 3.2) thereby eliminating the problematic marine reservoir effect (Heier-Nielsen *et al.*, 1995; Olsen *et al.*, 2009). Preliminary analyses at Kilen suggest an extremely variable reservoir effect of between -150 and +320 years (J. Olsen and B. Phillpesen, unpublished data). All 13 dates were included in the final age model (produced by J. Olsen Aarhus University using Oxcal 4.1, k=10, A=99.4%). Six samples were analysed for pollen (by Peter Rasmussen, GEUS) in order to identify the approximate location of the *Ulmus* (elm) decline (see Figure 3.3). Accordingly the age of the elm decline in Kilen is dated to 3,840-3,910 BC and provides an independent check on the age model.

Unsupported ^{210}Pb activity is detectable from background levels down to a depth of 30 cm. The concentration generally decreases exponentially with depth from its surface concentration of 241 bq kg^{-1} down to ~25 cm, after which irregularities occur in the profile (Figure 3.4). Supported ^{210}Pb activity is detectable in all samples, remaining almost constant throughout. Sample sediment ages were derived using a constant rate of supply (CRS) model Appleby, 2001 and results are displayed in Figure 3.2. The Pb-derived chronology suggests an average accumulation rate of ~0.35 cm yr^{-1} over the duration of the lead profile (beginning ~AD 1920; see also Figure 3.5). These results have been verified to some degree by ^{137}Cs, detectable in all samples. ^{137}Cs levels also decrease with depth down until ~AD 1920 (as indicated by the ^{210}Pb stratigraphy). This pre-dates its release into the atmosphere (~1945) suggesting a degree of sediment mixing or downward diffusion of ceasium, and therefore Pb-based ages must be considered as minimum ages as mixing will result in the derivation of ‘younger’ ages. The maximum ^{137}Cs activity occurs at 8 cm dated by ^{210}Pb as 1984, which can be considered broadly consistent with fallout from the Chernobyl incident in 1986. Despite some clear signs of sediment mixing and irregularity in lead activity towards the bottom, the exponential decrease with depth in the upper section gives confidence in the results and therefore the dating can be considered to be reasonably accurate (T. J. Andersen, pers. comm.).

Both the AMS ^{14}C (Figure 3.2, Table 3.2) and ^{210}Pb/^{137}Cs stratigraphies (Figure 3.2, Figure 3.4) are deemed robust chronologies being independently verified by the *Ulmus* decline (~3,900 BC; Andersen and Rasmussen, 1993) and caesium fallout peaks (e.g. Chernobyl, AD 1986; Figure 3.4). There are, however, dating uncertainties in the section between the top of radiocarbon profile (depth 459-460; ~AD 800) and the ^{210}Pb/^{137}Cs stratigraphy (depth 420-321 cm bpsl; ~AD 1920). The period is characterised by extremely low accumulation rates (~0.03 cm yr^{-1}), low sedimentary pigment and diatom concentrations and absence of molluscs and foraminifera. This might reflect low productivity and subsequent sediment mixing and poor preservation conditions. Alternatively a sedimentary hiatus might exist in the late Holocene. Broad estimates have been made for the stratigraphic positions of key events such as the closure of the Limfjord (dated to ~1,100-1,200 BC according to Petersen, 1976; Kristensen *et al.*, 1995; Christensen *et al.*, 2004) and the and the building of the road/rail embankment (AD 1856-57) based on the available proxy evidence (see section 3.11). These estimates must, however, be treated with caution until the chronology has been improved for this part of the sequence (i.e. ~AD 800-1920).

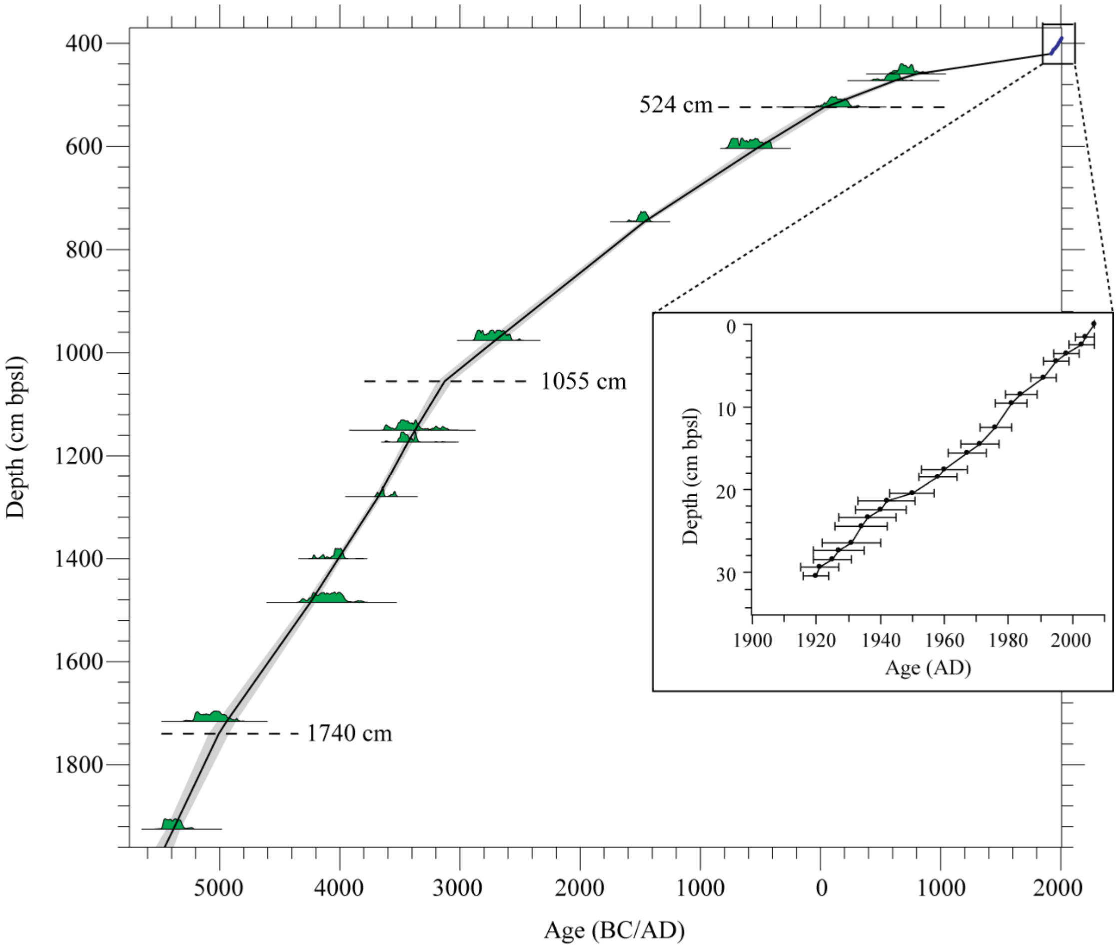


Figure 3.2 – Age-depth model and ^{210}Pb stratigraphy (blue line and inset) for the Kilen profile. Model created by Jesper Olsen (Aarhus University) using Oxcal v.4.1. Dotted lines refer to depths for which allowances for shifts in the accumulation rate were made based on changes in the carbonate curve (see Figure 3.5).

Laboratory number	Depth (cm bpsl)	Material dated	^{14}C age (BP)	Model age (± 2 SD AD/BC)	δ^{13}C (‰ VPDB)
AAR-12150	459-460	1 leaf fragment, 1 *Betula* sp. fruit, 2 bud scales, 1 *Cyperaceae* seed	1315± 65	672-894 AD (95.4%)	*-27
AAR-12151	472-473	1 flower from *Quercus* sp.	1440±65	433-495 AD (7.7%), 504-685 AD (87.7%)	*-27
AAR-12152	523-524	1 *Scirpus maritimus/lacustris* seed	1880±60	20-12 BC (0.5%), 1BC-257 AD (93.5%) 300-318 AD (1.4%)	*-27
AAR-12142	603-605	1 leaf fragment	2464±40	761-682 BC (25.4%), 671-413 BC (70.0%)	-26.87
AAR-12143	745-747	1 twig fragment with bark	3215±35	1,605-1,576 BC (4.3%), 1,536-1,415 BC (91.1%)	-28.38
AAR-12144	975-977	1 twig fragment with bark; presumably *Salix* sp.	4132±45	2,874-2,581 BC (95.4%)	-27.27
AAR-12145	1,149-1,151	1 leaf fragment	4635±75	3,634-3,553 BC (11.2%), 3,541-3,311 BC (68.8%) 3,295-3,286 BC (0.6%), 3,275-3,265 BC (0.7%) 3,239-3,105 BC (14.1%)	-28.19
AAR-11463	1,171-1,175	206 leaf fragments, 3 *Betula* sp. fruits, 1 bud, 1 bud scale	4635±40	3,621-3,611 BC (1.0%), 3,521-3,347 BC (94.4%)	-28.36
AAR-12146	1,279-1,281	1 *Alnus* sp. twig with bark	4848±42	3,709-3,626 BC (65.2%), 3,598-3,526 BC (30.2%)	-28.61
AAR-12147	1,399-1,401	97 leaf fragments	5225±40	4,228-4,201 BC (6.1%), 4,170-4,127 BC (10.5%) 4,119-4,097 BC (3.1%), 4,080-3,961 BC (75.7%)	-27.52
AAR-12148	1,483-1,487	11 leaf fragments	5260±90	4,331-3,942 BC (94.0%), 3,856-3,844 BC (0.6%) 3,836-3,821 BC (0.8%)	*-27
AAR-12149	1,715-1,717	43 leaf fragments	6120±65	5,222-4,881 BC (93.5%), 4,871-4,848 BC (1.9%)	*-27
AAR-11464	1,923-1,927	25 leaf fragments	6405±60	5,170-4,848 BC (95.4%)	*-25

Table 3.2 - Details of the samples dated by AMS ^{14}C from the Kilen profile (* estimated value).

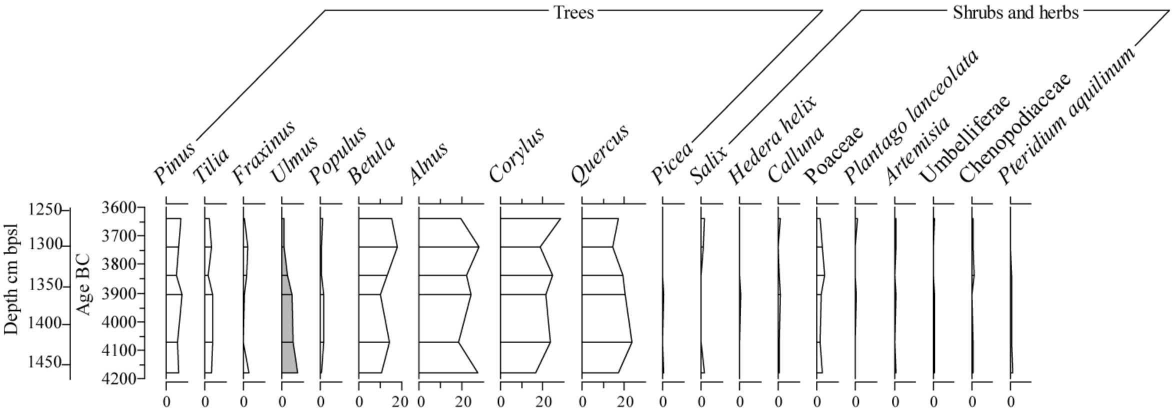


Figure 3.3 – Percentage values for selected taxa in the Kilen pollen record spanning ~4,180-3,640 BC. The *Ulmus* curve (shaded grey) exhibits a decline between 1,338 and 1,362 cm bpsl with the age ~3,900 BC, which is in good agreement with the general dating of the classic mid-Holocene 'elm decline' in Denmark (Andersen and Rasmussen, 1993). Analyst: Peter Rasmussen (GEUS).

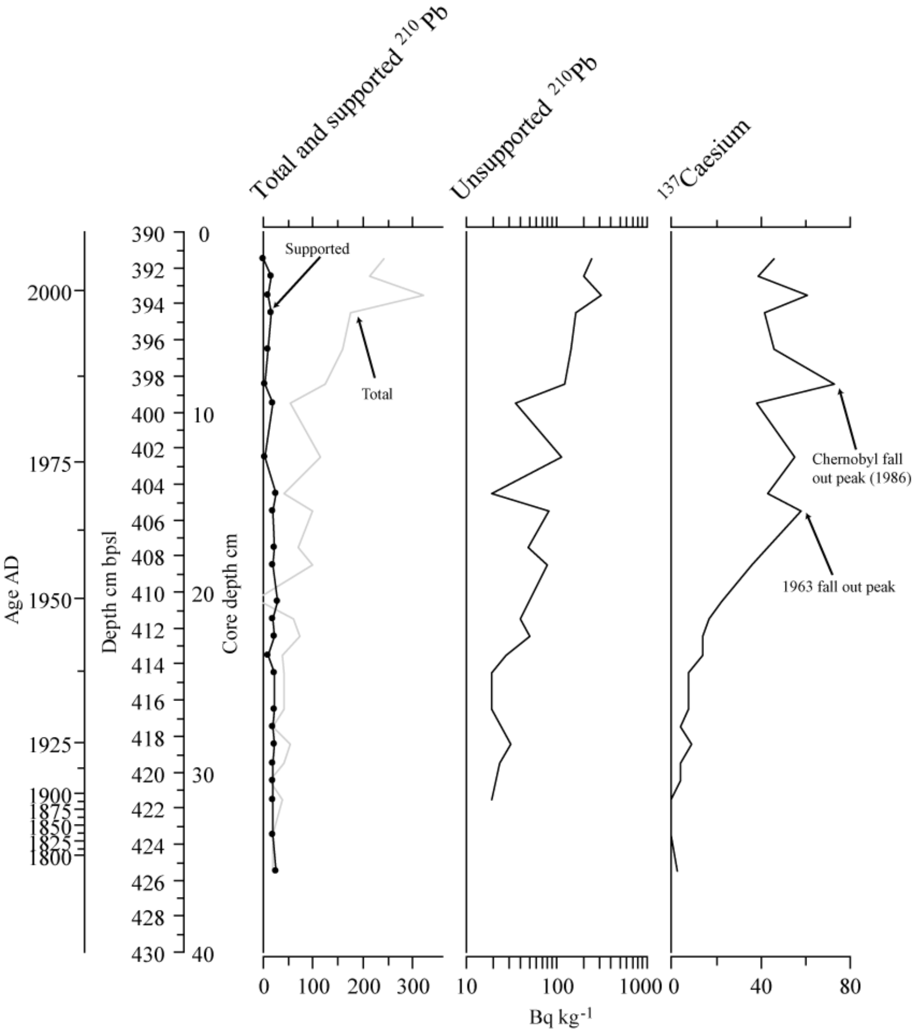


Figure 3.4 - Depth profiles of ^{210}Pb and ^{137}Cs for Kilen.

3.4 Physical Parameters

Results of all physical parameters measured at Kilen profile are provided in Figure 3.5. In total 1,369 samples from the Kajak and two boreholes (R3 and R4) were measured for organic, carbonate, minerogenic and dry mass/water contents, with the final profile consisting of 860 samples. The physical sedimentary parameters for Kilen have been divided up into four (statistically significant) zones determined by optimal splitting (see section 2.10.1) and are described below in relation to these zones. The accumulation rate for the Kilen profile and flux rates for the organic, carbonate and minerogenic components are also plotted in Figure 3.5 (calculated using the age model presented in Figure 3.2).

KIL-S1 (~5,500-3,130 BC) – At the bottom of the profile the overall sedimentation rate is high (0.5 cm yr^{-1}), with all three components (i.e. organic, carbonate and minerogenic matter) exhibiting their highest fluxes for the marine phase of the profile (i.e. pre-1856). However, these fluxes decrease at ~5,100 BC, causing a drop in the overall accumulation rate (~0.33-0.43 cm yr^{-1} between 5,100-3,130 BC). After ~5,100 BC, the carbonate and mineogenic fluxes steadily decrease and the organic flux gradually increases again. Only minor changes occur in the relative abundance within the sedimentary matrix, such as a slight rise in the carbonate content between ~5100-5,000 BC and the very gradual increases in both the organic and minerogenic contents after ~5,030 BC.

KIL-S2 (~3,130 BC- AD 30) – This zone begins with another drop in the accumulation rate (down to 0.19 cm yr^{-1}), largely driven by decreases in both the organic and minerogenic fluxes. There is very little change in the sedimentary regime between ~3,120 BC – AD 30, with the exception of a small and gradual decrease in the minerogenic flux (and subsequently the overall accumulation rate; 0.14-0.15 cm yr^{-1}) after ~1,500 BC and an increase in the carbonate flux at the very top of this zone (at ~AD 60). The relative abundance of organic matter in the sedimentary matrix gradually increases between ~3,120-100 BC (~7 % increase). In contrast, the overall carbonate content decreases up until ~200 BC and the minerogenic component decreases after ~800 BC.

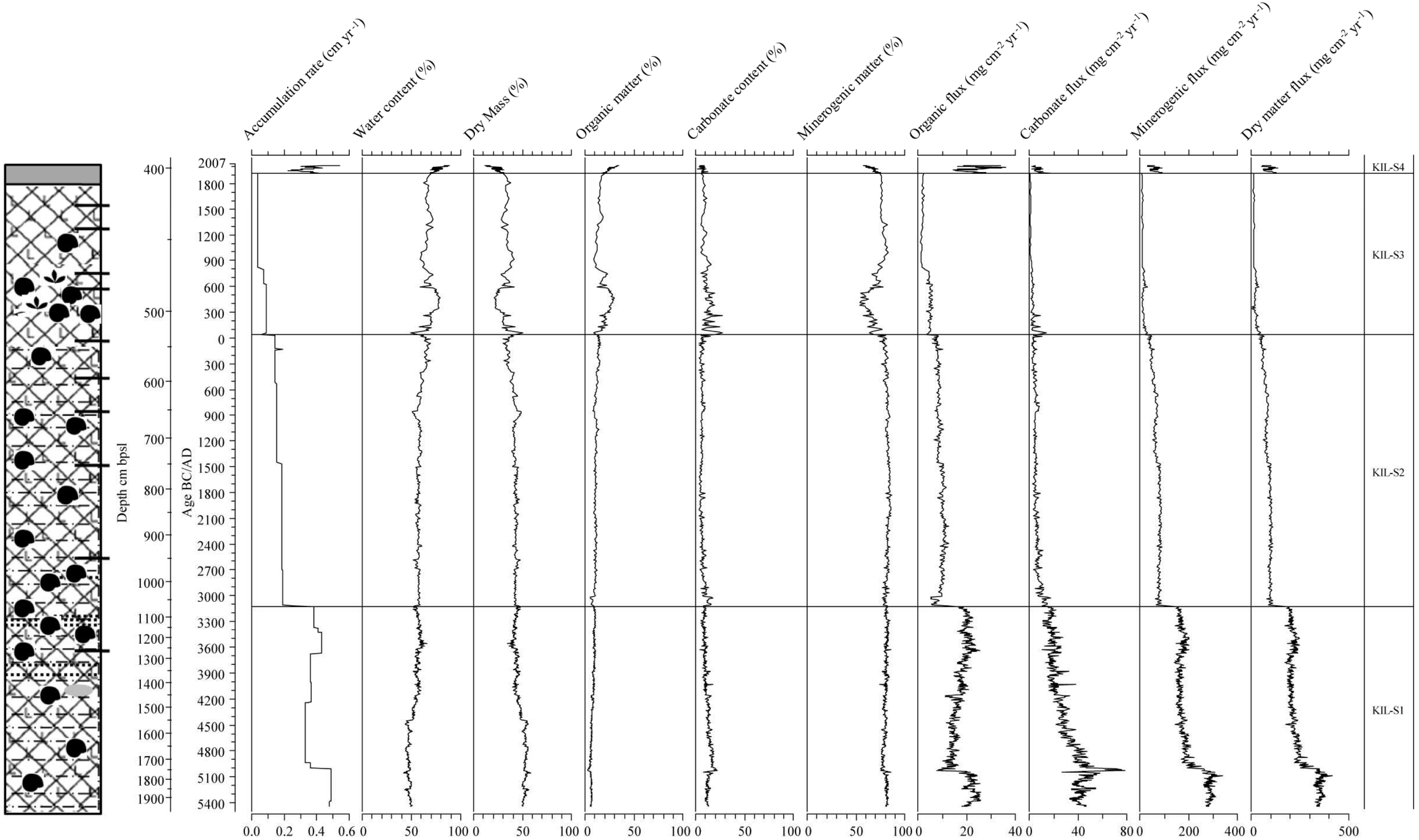


Figure 3.5 - Kilen sedimentary physical parameters, including overall accumulation rates (cm yr^{-1}), water content and dry mass expressed as percentages of the total wet weight of the sediment, percentages of the dry mass that are organic matter, carbonate and minerogenic matter (following loss-on-ignition), plus fluxes (mg cm^{-2} yr^{-1}) for organic matter, carbonate, minerogenic matter and dry mass. A magnification of Kil-S4 is provided in Figure 3.6. (NB. Accumulation rate and flux may be erroneous between ~AD 800-1920 due to uncertainties with the age-depth model (see sections 3.3 and 3.11)).

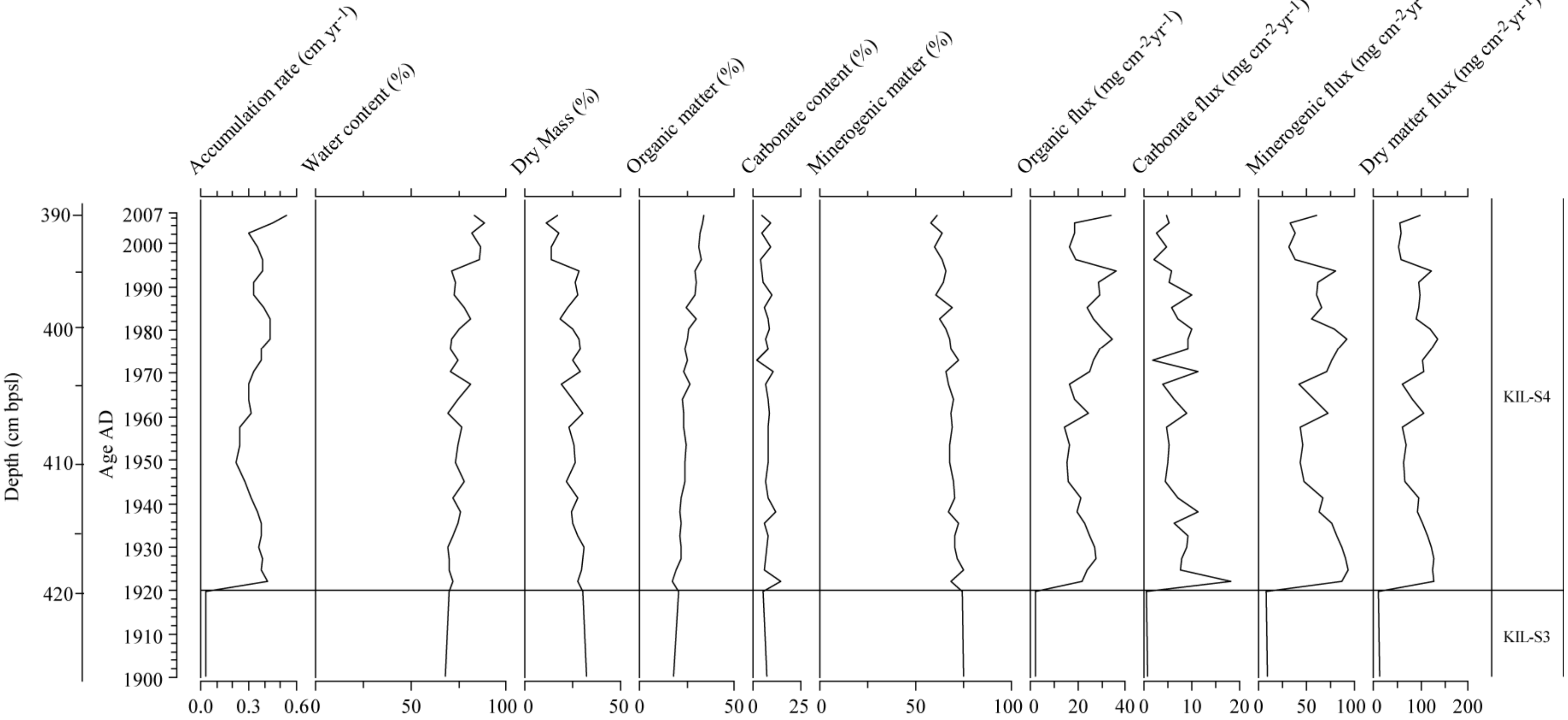


Figure 3.6 - Kilen sedimentary physical parameters focussing on Kil-S4. For full details see Figure 3.5 caption. (NB. A lithology column is omitted in this figure due to the absence of any lithological description for this section of the profile. Also note that the scales for the individual parameters in this figure have been adjusted accordingly (from Figure 3.5) for presentation purposes).

KIL-S3 (~AD 30-1920) – A further drop in the overall accumulation rate occurs at the beginning of this zone (i.e. at ~AD 30) with both the organic matter and minerogenic flux exhibiting decreases. After ~AD 30, both remain relatively constant (~0.07-0.09 cm yr^{-1}) and there is subsequently, little change in the accumulation up until ~AD 800. The carbonate flux differs slightly as it fluctuates up until ~AD 200, before also becoming constant. Problematic dating and a potential sedimentary hiatus suggests likely errors in the overall accumulation rate (0.03 cm yr-1) and sedimentary flux calculations between ~AD 800-1920. If a continuous accumulation of sediments did continue between ~AD 800-1920 (i.e. no hiatus), then an extremely low accumulation rate must have ensued, but currently this remains equivocal. Large changes occur in the sedimentary matrix in KIL-S3, as increased percentages of both organic matter (between ~AD 20-700) and carbonate content (between ~90 BC-AD 500) result in decreased abundance of minerogenic material between ~90 BC-AD 900. Relative abundances of these sedimentary parameters continue to fluctuate in the upper section of this zone (i.e. ~AD 900-1920), but changes are lower in magnitude.

KIL-S4 (~AD 1920-present) – This zone begins with an apparent large, but likely erroneous increase in the overall accumulation rate and fluxes of all three sedimentary components. This date marks the beginning of the lead stratigraphy and therefore increased accumulation rates (and fluxes) are just an artefact of the linear interpolation between the uppermost radiocarbon date and the lowermost ^{210}Pb/^{137}Cs date (see section 3.3 and 3.11). Nevertheless, in the post AD 1920 period, the accumulation rates remains high but subject to fluctuations. The predominant pattern, however, is a decrease in the accumulation of sediments up until ~AD 1950, followed by a general increase in post ~AD 1950 period. The individual fluxes also drop in this period, but increase again after ~AD 1960 (with more pronounced increases for organic and minerogenic components than for carbonates). In terms of the sedimentary matrix, the organic content of the sediment gradually increases throughout KIL-S4, with the minerogenic content decreasing and the carbonate content remaining relatively constant.

3.5 Diatoms

A total of 110 diatom slides have been counted over the ~7,500 year record with increased resolution over the Mesolithic-Neolithic transition, Pitted Ware or Single Grave cultural period, Roman period and the recent limnic phase (see Table 2.2). 96 genera were recorded with a total of 332 taxa identifiable to species level. The full diatom record and associated DI-salinity reconstruction are provided in Figure 3.7 (and for zones KIL-D4-KIL-D6, Figure 3.9) and diatom-associated metrics are illustrated in Figure 3.8 (and Figure 3.9). These include the total sum of diatoms, the centric/pennate ratio, benthic: planktonic: tychoplanktonic (B:P:T) ratio, DCA axis 1 scores (for entire and reduced datsets; see section 3.10), diatom concentration and valve accumulation rate Hill's N_2 diversity, dissolution **F** index, *Paralia sulcata* dissolution index and *P. sulcata* valve size. Six statistically significant zones were determined via optimal splitting with one additional split added to aid interpretation (i.e. sub-zone KIL-D1b).

The salinity reconstruction presented in Figure 3.7 is inaccurate in the uppermost section of the record (since ~AD 1830), due to increasing numbers of diatoms indicative of enriched nutrient status appearing in the record. This suggests that nutrients (rather than salinity) are the main environmental variable driving the changes in the diatom assemblage. Therefore, a diatom-inferred total nitrogen (DI-TN) reconstruction for the last ~200 years is also presented in Figure 3.9 using a TN-training set based on 91 MOLTEN sites from Denmark and Western Sweden (Juggins, 2004). The final TN-reconstruction utilises a WAPLS-component 2 model ($r^2_{\text{jack-knifing}} = 0.74$, $RMSE_{\text{jack-knifing}} = 0.14$). Reconstructions using the MOLTEN weighted averaging model and modern analogue technique (not shown) were also tested and yielded largely similar results. Application of similar TN-models (i.e derived from the MOLTEN training set) in other Danish coastal sites (Roskilde Fjord, Mariager Fjord) over the recent past has yielded relatively accurate reconstructions of past nutrient status (e.g. Clarke *et al.*, 2003; Clarke *et al.*, 2006; Ellegaard *et al.*, 2006).

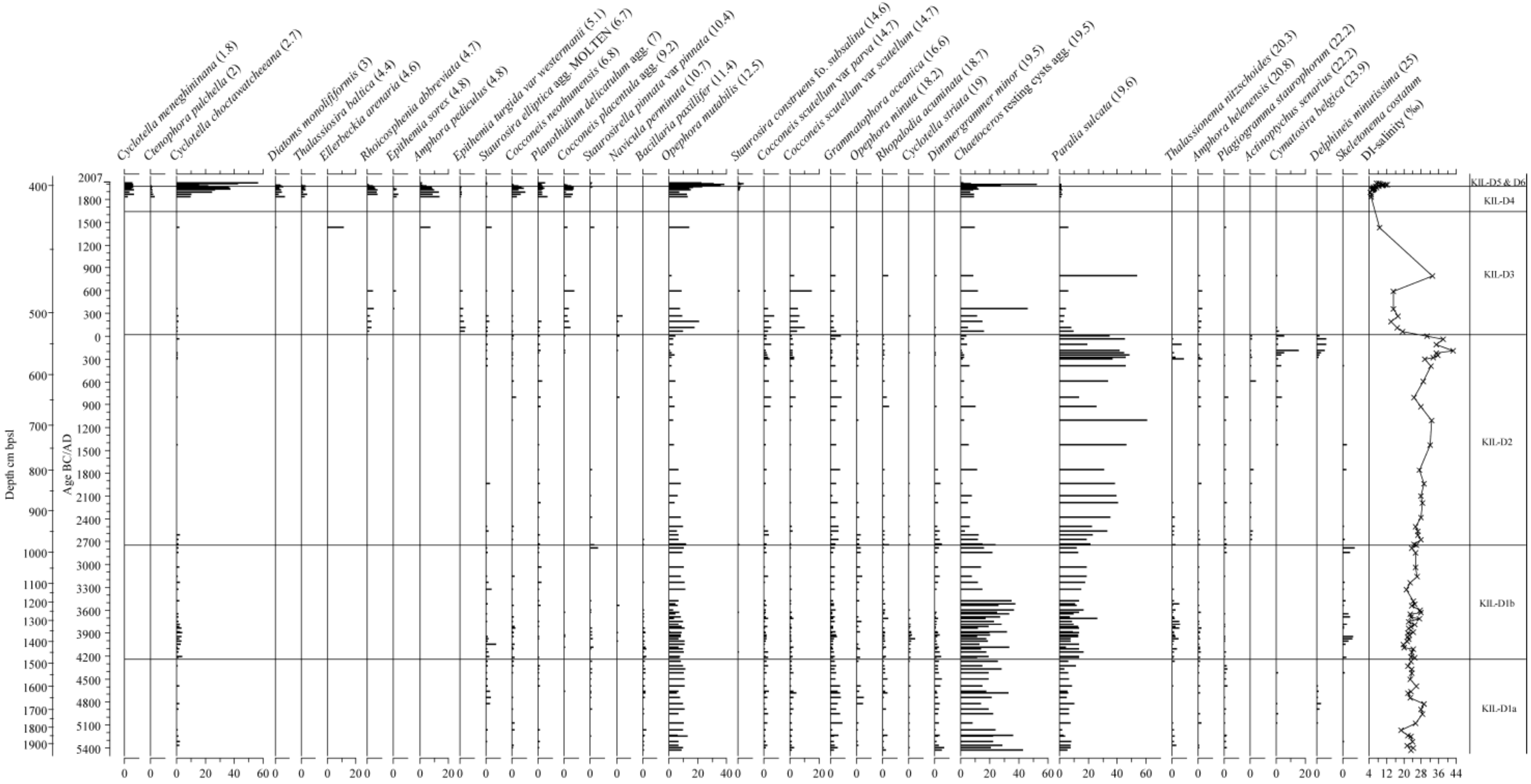


Figure 3.7 – Kilen percentage diatom record (taxa >3 %). Numbers in brackets refer to the salinity optima in the MOLTEN training set used for the quantitative reconstruction of salinity. Analyst: Jonathan Lewis.

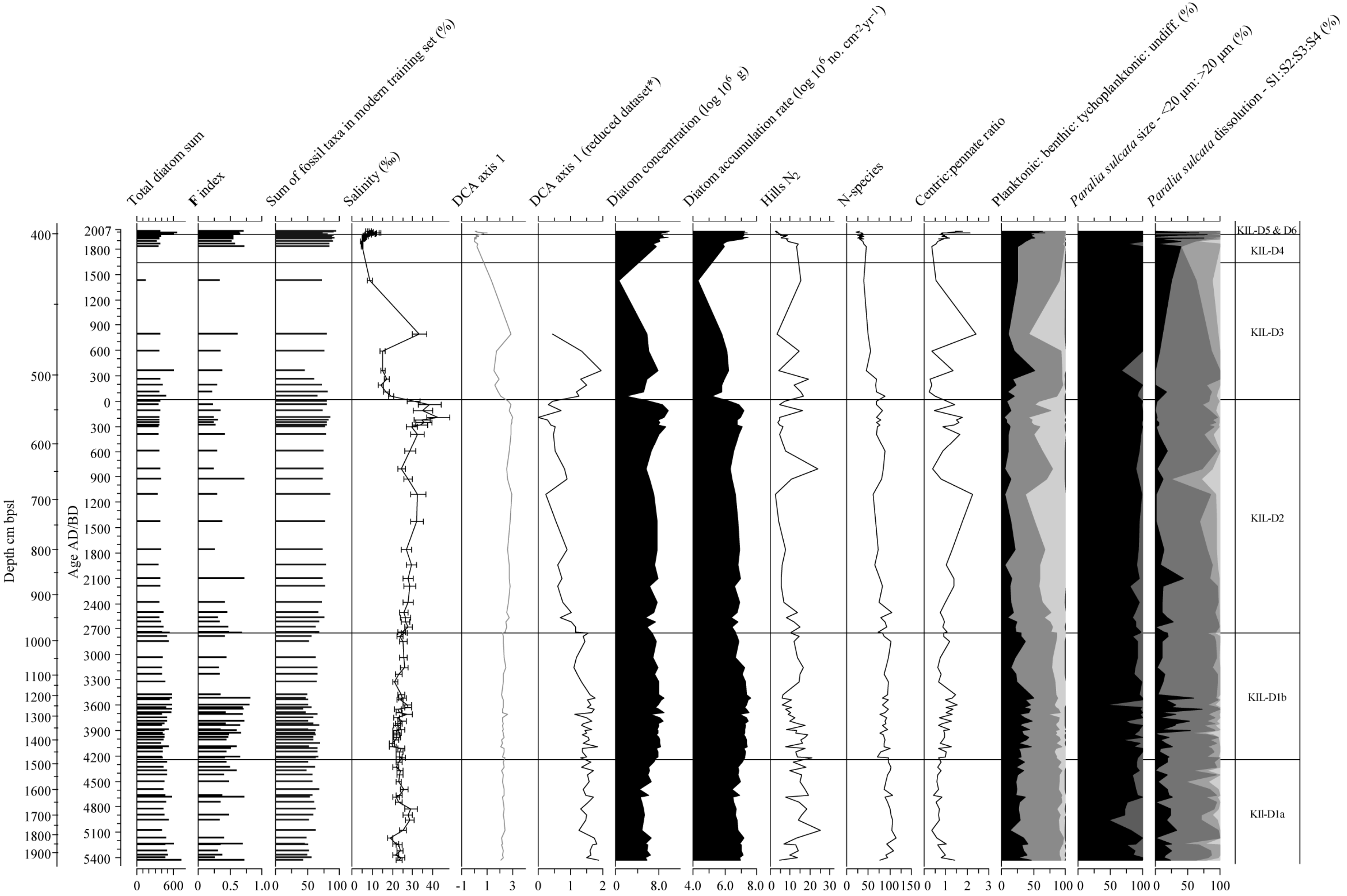


Figure 3.8 - Kilen diatom associated metrics. * Sample scores for DCA analysis on reduced diatom dataset (i.e. between ~5,430 BC-AD 800; lower 82 samples).

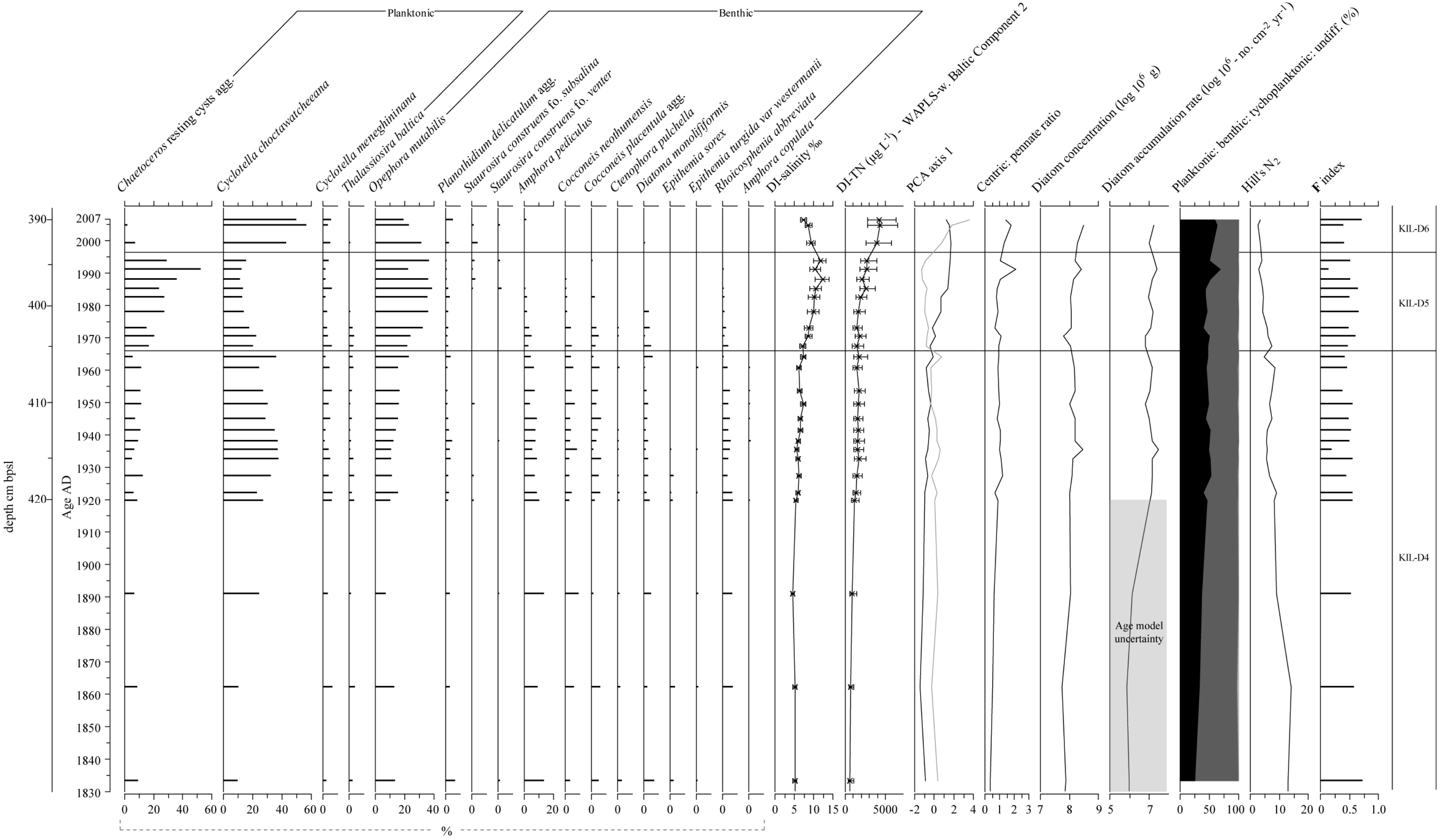


Figure 3.9 - Kilen diatom record for the uppermost section of the profile (KIL-D4-KIL-D6; ~AD 1830-present).

KIL-D1a (~5,500-4,250 BC) - *Chaetoceros* cysts the most abundant group of species, fluctuating intensely throughout, but supplemented by relatively constant percentages of *Opephora mutabilis* (most abundant benthic; ~3-13 %), *Cocconeis scutellum* agg. (*scutellum* and *parva* varieties), *Paralia sulcata, Dimerogramma minor, Bacillaria paxillifer, Plagiogramma staurophorum, Amphora helenensis* and *Rhopalodia acuminata*. The DI-salinity reconstruction suggests salinity is generally around 23-24 ‰ for the majority of the zone but with occasional fluctuations (e.g. ~25-29 ‰ between ~5,080-4,830 BC). High salinity demanding species such as *Delphineis minutissima* and *Cymatosira belgica* briefly appear in the record between (5,080-4,670 BC). *Cyclotella choctawatcheeana* numbers also fluctuate which might suggest variable nutrient status and/or water depth.

KIL-D1b – (4,250-2,740 BC) - *Chaetoceros* cysts continue to fluctuate, but exhibit a general increase from ~18 % at ~4,150 BC to ~33-35 % between ~3,650-3,480 BC (most abundant taxa up until ~3,480 BC). There is also an increase in several other planktonic taxa in the lowermost section of this zone, most notably *Skeletonema costatum*, *C. choctawatcheeana*, *P. sulcata* (tychoplanktonic), *Thalassionema nitzschoides* and *Cyclotella striata*. However, planktonic diatoms decrease again after ~3,520 BC (e.g. decrease in B:P-:T and centric:pennate ratios), largely driven by a ~20 % decrease in *Chaetoceros* cysts (down to ~15 % by 3,330 BC) but also by much smaller reductions of *T. nitzschoides, C. striata* and *S. costatum*. *P. sulcata* increases after ~3,540 BC and with it the DI-salinity rises, reaching 26 ‰ by ~3,160 BC. The benthic community remains relatively unchanged with *O. mutabilis* (3-11 %) still the most abundant benthic species, followed by relatively constant numbers of *G. oceanica*, *O. minuta, C. scutellum*, R. *acuminata* and *D. minor*.

KIL-D2 (~2,740 BC-AD 20) - *P. sulcata* is the most abundant species throughout, accounting for ~15-60 % of assemblage, and *Chaetoceros* cysts further decrease at the beginning of this zone, and thereafter fluctuate between ~1-12 %. DI-salinity generally increases throughout, with the exception of a drop between ~930-600 BC (25-29 ‰), at which point, *P. sulcata* drops to ~15 % of the assemblage. However by ~600 BC, DI-salinity is increasing again and at ~390 BC, *Cymatosira belgica* (high salinity demanding diatom) begins to increase in abundance. This is followed shortly after by *T. nitzschoides* (~310 BC) and *Delphineis minutissima* (~280 BC). *O. mutabilis* is still the most abundant benthic species in the lower section of KIL-D2, but numbers drop after ~1,760 BC. *D. minor* numbers also drop at this time and *G. oceanica* becomes the most abundant benthic taxon in most samples between ~1430-0 BC, generally accounting for ~1-8 % of the community. *C. scutellum* (both *parva* and *scutellum* varieties) and *R. acuminata* increase in abundance in the upper section of this zone (after ~930 BC).

KIL-D3 (~AD 20-1640) - A large drop in salinity occurs at the onset of this zone as *P. sulcata* number fall dramatically and *C. belgica, D. minutissima* and *T. nitzschoides* practically disappear from the record. *Chaetoceros* cysts, *O. mutabilis* and *Cocconeis scutellum* var. *scutellum* all increase in abundance and the DI-salinity fluctuates between ~14-19 ‰ in the period ~AD 50-600. A series of brackish water benthic species also appear in the record early in this zone, most notably *Rhoicosphenia abbreviata, Epithemia turgida, C. placentula* and *Achnanthes amoena* (single peak at ~AD 260; not shown in figure 3.4a). A large peak in *Chaetoceros* cysts also occurs at ~AD 360. DI-salinity briefly rises back up to ~33 ‰ at ~AD 800, driven mainly by a large increase in *P. sulcata* (which accounts for ~54 % of the assemblage at this time). However, the salinity declines in the uppermost sample (~AD 1,430) as conditions turn brackish with *O. mutabilis* (~14 %), *Ellerbeckia arenaria* (12 %, potential oligotrophic indicator), *Chaetoceros cysts* (10%) and *Amphora pediculus* (7%) the most abundant taxa.

KIL-D4 (~AD 1640-1965) (see Figure 3.9) - Salinity is lower in this zone as brackish, nutrient-demanding taxa dominate the assemblage. DI-TN increases up until ~AD 1930, largely driven by the increasing numbers of *C. choctawatcheeana* (most abundant species between ~AD 1890-1960; 23-38 %). However, *C. choctawatcheeana* percentages are decreasing again from after the 1930s. *Chaetoceros* cysts (6-13 %), *Cyclotella meneghinina* (2-7 %) and *Thalassiosira baltica* (1-4 %) are all ever-present in the plankton. *O. mutabilis* (7-23 %) and *A. pediculus* (4-14 %) are the most abundant benthic species with smaller abundancs of *P. delicatulum, C. neothumensis, Diatoma moniliformis, C. placentula* and R. *abbreviata*. *O. mutabilis* increases after ~AD 1935, but most other benthic taxa exhibit reduced percentages in the post AD 1920 period (e.g. decreasing B:P:T ratio; Figure 3.9).

KIL-D5 (~AD 1965-1996) (see Figure 3.9) - *C. choctawatcheeana* continues to decrease, being replaced by *Chaetoceros* cysts as the dominant planktonic species at ~AD 1970. *O.*

mutabilis is again the most abundant benthic species in KIL-D5 and continues its increase up until ~AD 1978, after which it remains relatively constant as it co-dominates the assemblage with *Chaetoceros* cysts and represents ~36 % of the assemblage. There is also a sharp change in both the centric:pennate and B:P:T ratios in the late 1980s, as in addition to the increase in *Chaetoceros* cysts, several benthic species experience a decrease, most notably *A. pediculus, C. placentula, C. neothumensis, D. moniliformis* and *R. abbreviata*. There is also a reduction in *T. baltica,* while *Staurosira construens* fo. *subsalina* and *S. construens* fo. *venter* appear in the record.

KIL-D6 (~AD 1996-2007) (see Figure 3.9) - *Chaetoceros* cysts experience a major decrease at the beginning of this zone dropping by over 20 % and almost disappearing from the record by the uppermost (most recent; AD 2007) sample. DI-TN (and the centric:pennate ratio) continues to increase suggesting further intensification of eutrophic conditions. A dramatic increase in *C. choctawatcheeana* occurs (~27 % rise), which again replaces *Chaetoceros* cysts as the main component of the plankton. *O. mutabilis* (~19-31 %) progressively declines, but still remains the main benthic representative, with smaller amounts of *S. contruens* fo. *subsalina* and *P. delicatulum* also present.

Diatom Preservation

Overall diatom preservation is good for the majority of the core, with an average **F** index of 0.45 for the entire profile (range = 0.14-0.82). Despite this, there are distinct phases when preservation condition appears to either improve/deteriorate. **F** index values gradually rise (improve) between ~5,500-3,500 BC, with exceptional preservation conditions persisting between ~4,200-3,600 BC (lowermost section of KIL-D1b). This phase coincides with lower DI-salinities (Figure 3.7) and higher productivity (see below, Figure 3.10, Figure 3.11). Exceptional preservation is further supported by the substantial numbers of the weakly silicified diatom *S. costatum* between 4,100-3,650 BC) and the increase in numbers of stage 1 classified *P. sulcata* valves. The **F** index fluctuates between ~0.3-0.8 in the succeeding phase, prior to a period of poorer preservation between ~1,760 BC-AD 110 (**F** index = 0.2-0.3). This period of poorer preservation coincides with highest DI-salinities, which might be expected under fully marine conditions. **F** index values increase again in the early Roman period (after ~AD 190) and become very good (**F** index =~0.5-0.7) in the uppermost section of the profile (post ~AD 1830).

In almost all samples the majority of *P. sulcata* valves fall into dissolution stages 1 or 2, with stage 2 generally being the most common category. Two exceptions to this occur at ~930 BC and ~AD 1830, but the latter of these is based on too fewer *P. suclata* valves (5 valves) for an accurate assessment. At ~930 BC, the **F** index score (~0.7) suggests that the overall preservation in this sample is good and therefore poorer *P. sulcata* preservation appears an anomaly. This is perhaps related to dissolution in water column, specific habitat characteristics or being washed in from external sources (e.g. older sediments from the catchment or deeper waters of Limfjord/North Sea).

3.6 Sedimentary pigments

The Kilen sedimentary record is presented as concentration (nmol g/OM) in Figure 3.10 and as pigment accumulation rates in Figure 3.11. The final sedimentary pigment profile consists of 777 samples, giving a temporal resolution of ~4-58 years (average 9 years). Twelve identifiable pigments were present in the profile (see Figure 3.10), with one undifferentiated pigment (labelled carotenoid X) and three ultraviolet (UV) absorbing compounds (labelled compound A, B and C). The uppermost section of the profile is absent (~1800-present; depth 390-424 cm bpsl), due to unsuitable treatment and storage conditions of pigment samples from the HON-Kajak core (upper 35 cm of profile). Superimposed upon the major trends described below, are regular minor fluctuations in pigment abundances, which occur throughout the entire profile. Sedimentary pigment affiliation (quoted in brackets after the first mention) are based on affiliations quoted in (Leavitt and Hodgson, 2001; McGowan, 2007). Chlorophylls refer to chlorophyll *a*, chlorophyll *b*, pheophytin-*a* and pheophytin-*b*, and carotenoids refer to all other pigments excluding compounds A, B and C (UV-absorbing pigments).

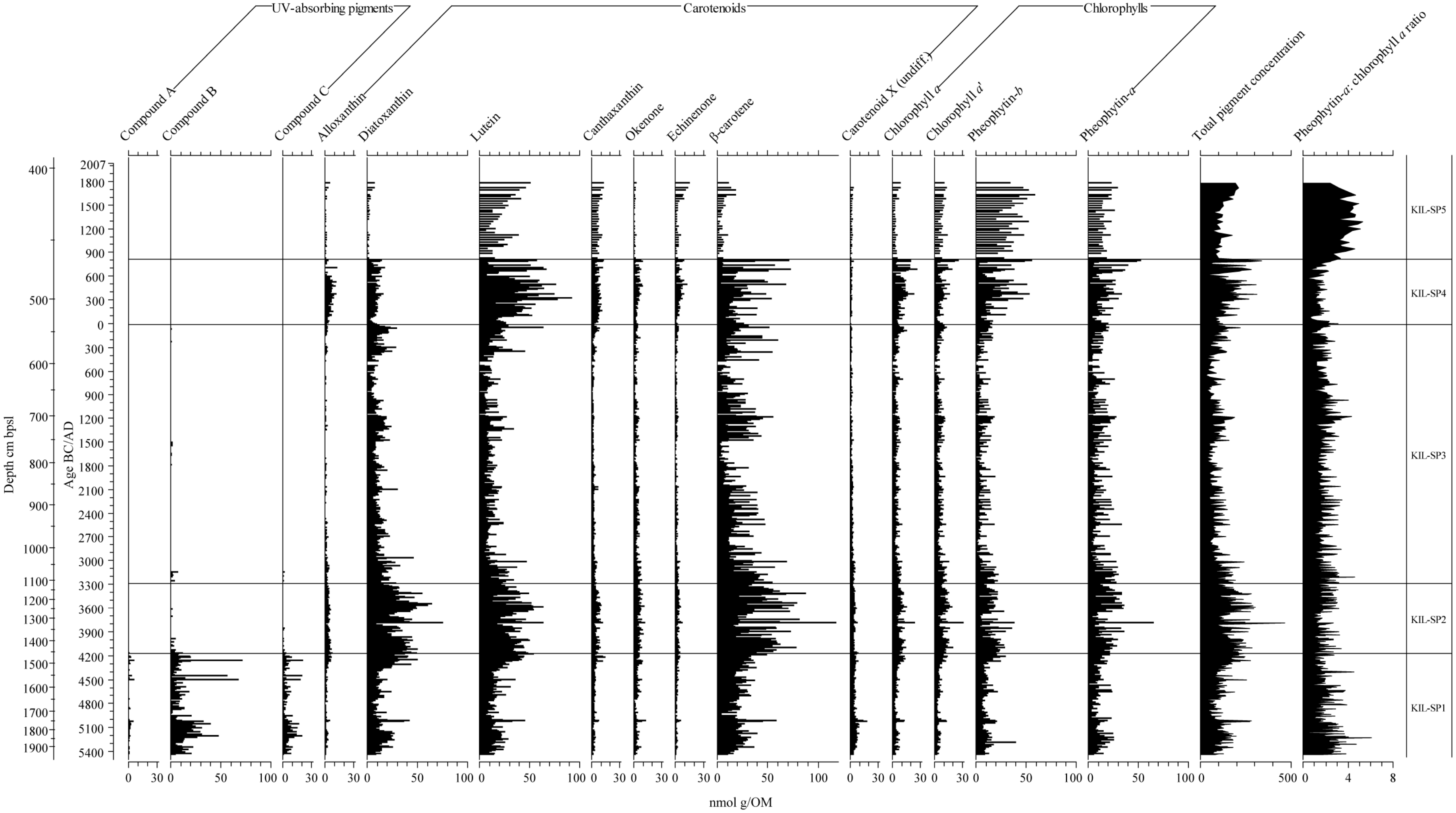


Figure 3.10 - Sedimentary pigment record from Kilen expressed as concentrations; nmol g OM. Analyst: Jonathan Lewis

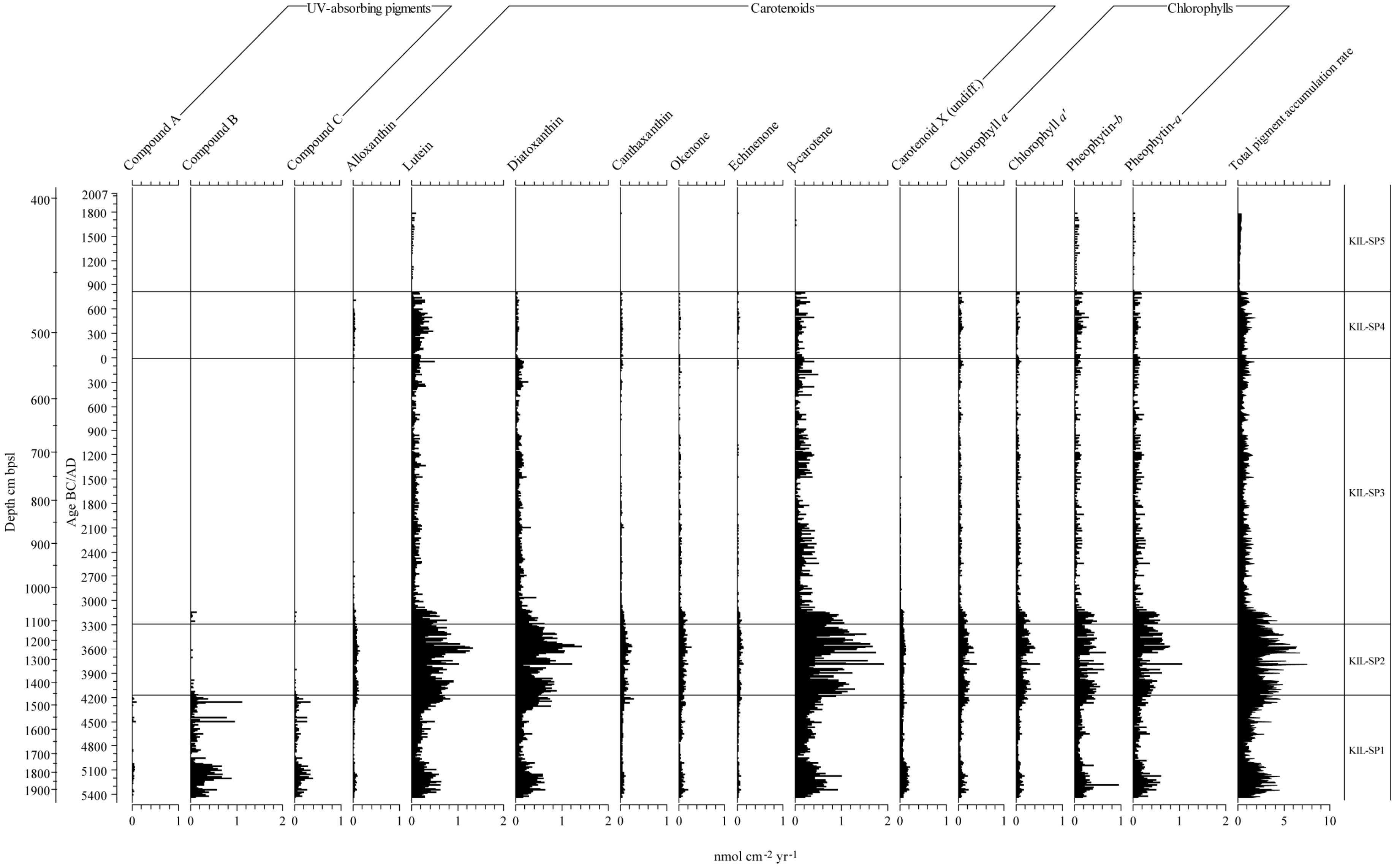


Figure 3.11 - Kilen sedimentary pigment record expressed as accumulation rates; nmol cm^{-2} yr^{-1}.

KIL-SP1 (~5,500-4,170 BC) - At the bottom of the profile, the overall abundance of pigments is relatively high, with diatoxanthin (from diatoms, dinoflagellates and chrysophytes), lutein (higher plants, green algae, euglenophytes), β-carotene (most algae and plants), pheophytin-*a* (chlorophyll *a* derivative) and UV-absorbing compound B (primarily cyanobacteria) the most abundant pigment types. Carotenoid X and the UV-absorbing pigments are both at their highest levels for the entire profile. The majority of pigments are either stable or increasing between ~5,500-5,200 BC, but most tend to exhibit a brief fall in concentrations somewhere between ~5,200-4,800 BC. When plotted as accumulation rates, there is a particularly noticeable decrease after ~5,000 BC, but this decrease is relatively short lived and by ~4,800 BC the accumulation rates (and concentrations) of most pigments are increasing again. There are particularly notable increases in okenone (purple sulphur bacteria), chlorophyll a (higher plants, all photosynthetic algae), pheophytin-*a* and pheophytin-*b* (chlorophyll *b* derivative) at this time. Later in the zone (~4,500-4,400 BC) all pigments begin to increase in concentration, which culminate in a high productivity phase (including high accumulation rates) that lasts throughout the succeeding zone (i.e. KIL-SP2; see below). The UV pigments are generally present in lower abundances after ~5,060 BC, but subject to fluctuations and high concentrations in occasional samples.

KIL-SP2 (~4,170-3280 BC) - The overall concentration of sedimentary pigment are high in KIL-SP2, with almost all most pigments (except for the UV-absorbing compounds) abundant throughout, following their increases at the top of the previous zone (KIL-SP1). The UV pigments practically disappear at the onset of this zone (by ~4,000 BC), and for the remainder of the profile only reappear sporadically. Lutein, diatoxanthin and β -carotene remain the most abundant pigments, with diatoxanthin present at its highest concentrations for the entire profile. The highest accumulation rates for all chlorophylls and carotenoids (for the entire profile) occur between ~4,300-3,200 BC suggesting this period is the most productive in the maritime section of the Kilen record presented here (i.e. pre-AD 1856).

KIL-SP3 (~3,280-10 BC) – The UV-absorbing pigments (compounds A and B) reappear briefly in small concentrations between 3,250-3,150 BC, but all other pigments decrease in the lowermost section of KIL-SP3 (~3,300-3,000 BC) as overall productivity decreases. Thereafter pigment concentrations remain relatively low, but subject to fluctuations throughout, largely driven by the abundance of β-carotene (and the pheophytins). Noticeable increases in β-carotene (and overall concentration) occur between ~2,750-2,090 BC and ~1,500-900 BC, with a period of low concentration between phases (i.e. β-arotene <10 nmol g OM; ~2,090-1,500 BC). During the latter of these phases some other pigments also exhibit short term increases in concentrations, most notably diatoxanthin (~1,490-1,180 BC) and lutein (~1,360-1,180 BC). After ~500 BC all pigments begin to increase in abundance. The accumulation of pigments is low after 3,100 BC, with very minor rises visible during the periods of elevated concentrations (listed above).

KIL-SP4 – (~ 10 BC - 810 AD) – Overall pigment concentrations are generally high, with several pigments, most notably alloxanthin (cryptophytes), lutein, canthaxanthin (colonial bacteria, herbivore tissue) exhibiting large increases in abundance near the beginning of this zone and remaining high throughout. Most other pigment gradually increase, peaking towards the middle-upper section of this zone (~AD 300-810). Diatoxanthin is an exception, exhibiting a decrease in abundance at the onset of KIL-SP4 and remaining relatively low throughout. In this zone, the pigment accumulation rates suggest that the apparent increase in productivity might be exaggerated due to a decrease in the sediment accumulation rate (Figure 3.5). The total pigment accumulation rate actually suggests only minor changes in marine productivity in this zone, with overall productivity increasing only slightly after ~AD 200. Minor increases in the accumulation of alloxanthin and canthaxanthin occur slightly earlier (~200 BC-AD 30).

KIL-SP5 (~AD 810-1800) – Almost all pigments decrease in concentration at the beginning of this zone, with the exception of canthaxanthin and pheophytin-*b*, both of which are abundant throughout. The high levels of Pheophytin-*b* and high pheophytin-*a*/chlorophyll *a* ratio suggest increased degradation of pigments. The rest of the pigments remain at lower concentrations up until ~AD 1,500 before increasing again in the uppermost section of the pigment profile (i.e. after ~AD 1800). Most notable increases occur in alloxanthin, lutein, canthaxanthin and echinenone. Diatoxanthin concentrations drop to their lowest levels for the entire profile in KIL-SP5. The accumulation rates for all pigments also drop to their lowest levels for the entire profile in this zone, particularly between ~AD 810-1,100. β-carotene, lutein and both pheophytins (*a* and *b*) exhibit the highest accumulation rates. It is important to highlight that pigment accumulation rates must be treated with caution and may even be erroneous in this section of the record due to dating uncertainties (see section 3.3 and 3.11).

3.7 Macrofossils

A total of 611 samples were analysed for plant and animal macrofossils at Kilen at an average resolution of 12 years (max. 57.6 years, min. 2.2 years). The volume of sediment analysed ranged between 1.5-45 cm^3 (average 21 cm^3) taken from 1-2 cm thick slices (see section 2.4). Macrofossils present included molluscs, terrestrial and aquatic plant material, ostracods, cladocerans, echinid remains, fish bones and scales, charcoal, crustacea and foraminifera. Due to the relatively low abundance of macrofossils present in the cores (particularly plant material), zonation is based on the entire macrofossil dataset (species abundance >3%) rather than being split into its different components (e.g. molluscs, plant macrofossils). Foraminifera were excluded for the zonation process due to the low number of samples analysed (15 in total). Presence/absence data (see Figure 3.12-Figure 3.15) were also removed. Ten statistically significant zones have been determined using an optimal splitting technique (see section 2.10.1).

3.7.1 Molluscs

The Kilen molluscan assemblage is presented in Figure 3.12. All species are expressed as concentration per 50 ml of wet sediment with the exception of *Ophistobranchia* spp. fragments and *Mytilus* spp. fragments, which are expressed as presence/absence data. Overall, the molluscan concentration (maximum concentration 125 per 50 ml at depth 504 cm bpsl; ~AD 1692) was low throughout the profile, with a total of 424 samples (out of 611) containing molluscan remains. A total of 33 genera were recorded, yielding 38 species (17 gastropods and 21 bi-valves).

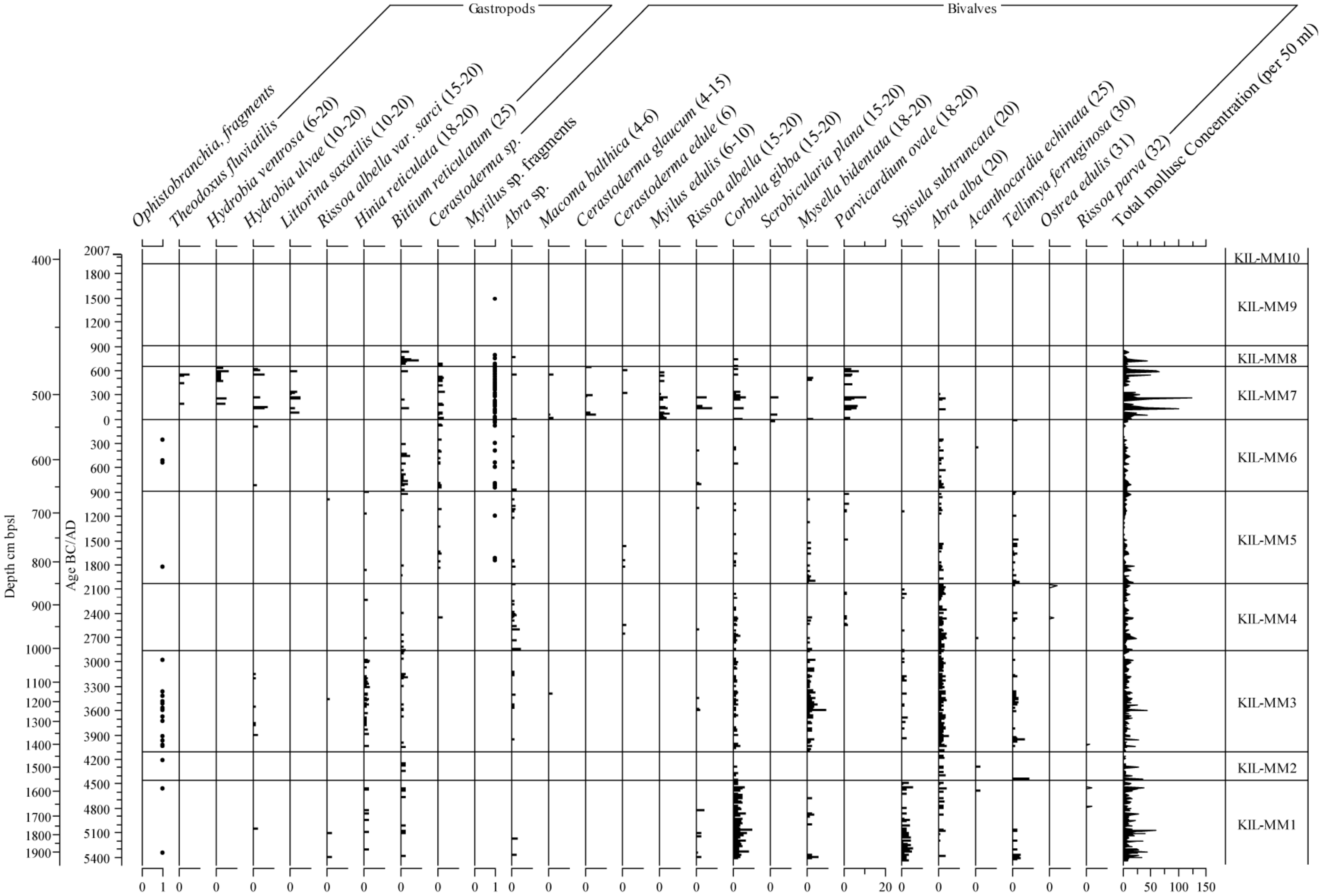


Figure 3.12- Kilen molluscan record (as concentration per 50 ml fresh sediment, with square root transformation applied). *Ophistobranchia* fragments and *Mytilus* sp. Fragments expressed as presence/absence data. Numbers in brackets refer to minimum salinity tolerances according to Sorgenfrei, (1958). Analyst: Kaj Strand Petersen (GEUS).

KIL-MM1 (~5,500-4,450 BC) - *Corbula gibba* and *Spisula subtruncata* are the most abundant and consistently present species throughout. There are also fairly regular occurrences of *Hinia reticulata* and *Abra alba* and several appearances by *Bittium reticulatum* and *Rissoa albella*. The mollusc concentration is comparatively high (only exceeded in KIL-MM7), particularly in the lowermost section (up until ~5,050 BC), due to the slightly higher concentrations of *C. gibba* and *S. subtruncata*. The high salinity demanding *Tellimya ferruginosa* is also present in some of the lowermost samples, but temporarily disappears from the record around ~5,090 BC.

KIL-MM2 (~4,450-4,105 BC) - Molluscan diversity and concentration decrease as *S. subtrunctata* and *C. gibba* practically disappear from the record. *A. alba* is the only consistently present mollusc in this zone, supplemented by small and sporadic occurrences of high salinity demanding species, most notably *B. reticulatum* but with individual peaks of *Acanthocardia echinata* and *T. ferruginosa*. The latter of these achieves its highest abundance for the entire profile at ~4,440 BC.

KIL-MM3 (~4,105-2,855 BC) - Mollusc concentration and diversity increase, with *A. alba* almost ever-present and common occurrences of *H. reticulata*, *Mysella bidentata, C. gibba* and *S. subtruncata*. A distinct peak in molluscan concentration occurs at ~3,590 BC where *M. bidentata* reaches its highest levels for the entire profile. *B. reticulatum* is also present in numerous samples in this zone which together with occasional finds of *T. ferruginosa* suggests relatively high salinity. Fragments of *Ophistobranhcia* spp. (marine sea slug) also become more regular.

KIL-MM4 (~2,855-2,030 BC) - *A. alba* is again the most consistent mollusc species, continuing at similar concentrations to the previous zones (i.e. KIL-MM2 and KIL-MM3). This is supplemented by *Abra* spp. (undifferentiated to species level) particularly in the lower section of this zone (up until ~2,460 BC), after which it is only present in a few samples and only at very low concentrations. All other species present in this zone make sporadic appearances, including the first recordings of the brackish-water edible cockle *Cerastoderma edule,* in an otherwise medium-high salinity fauna.

KIL-MM5 (2,030-890 BC) - *A. alba* decreases in abundance in this zone and even disappears for a short period between ~1,600-1,180 BC. *T. ferruginosa* and *M. bidentata* are present in several samples in the lower section of this zone and there are also occasional finds of brackish water species such as undifferentiated mussel (*Mytilus* spp.) and cockle (*C. edule* and *Cerastoderma* spp.) fragments and *C. edule* throughout. Overall, the molluscan concentration drops in this zone, particularly between (1,530-1,230 BC), where concentrations fall to their lowest pre-Viking (AD 750-1,050) levels (see KIL-MM8).

KIL-MM6 (890-10 BC) - Mollusc concentration increases again in this zone, due to increased concentrations of *A. alba* and *B. reticulatum*. Despite this, the molluscan fauna remains relatively limited, with just several samples containing undifferentiated fragments of *Cerastoderma* and *Mytilus* species and supplemented by occasional appearances from several other species. The upper section of this zone (~280-40 BC), is almost devoid of molluscs, as *B. reticulatum* and *A. alba* disappear. After ~40 BC, *S. plana* and *Retusa truncata* make their first appearances in the profile and *T. ferruginosa* reappears briefly at ~25 BC.

KIL-MM7 (10 BC- AD 650) - There is a large increase in mollusc concentration and diversity in this zone, driven by an increase/arrival of a number of species indicative of low (e.g. ~ 0-10 ‰; *Macoma baltica, M. edulis, Ceratoderma glaucum* and later *Theodoxus fluviatalis*) and medium salinity (e.g. 10-20 ‰; *Hydrobia ventrosa, Hydrobia ulvae, Littorina saxatalis, C. gibba, Scrobicularia plana, R. albella* and *Parvicardium ovale*). Molluscan concentration is variable with two distinct phases of higher concentrations (~10 BC-AD 260 and ~AD 460-650) separated by a short phase (~AD 260-460) where molluscan concentration drops very low.

KIL-MM8 (~AD 650-910) - This zone is marked by a drop in both molluscan concentration and diversity, driven by the disappearance of most of the low-medium salinity tolerant species characteristic of KIL-MM7. *B. reticulatum* becomes the most abundant species, suggesting increased salinity, supported by occurrences of *Tapes* spp. and *Gibbula tumida* (not shown in Figure 3.12). *B. reticulatum* is also the only species consistently present up until ~AD 830, after which molluscs remains almost disappear from the record. Some undifferentiated mussel fragments, however, do remain present up until ~AD 670.

KIL-MM9 (AD ~910-1930) - There are very few molluscs present in this zone, with just one sample (at ~AD 1400) containing some unidentifiable bi-valve fragments and another sample containing *Mytilus* (mussel) fragments (~AD 1490), but only identifiable to the genus level.

KIL-MM10 (AD 1930-2007) - This zone is completely devoid of molluscs.

3.7.2 Plant macrofossils

Plant material is generally scarce throughout the record with only 154 samples (out of 611) containing any plant material distinguishable to genus/species level (maximum concentration 171 per 50 ml). This number rises to 213 samples if samples containing undifferentiated fragments of wood, deciduous leaves, bark, buds and twigs are included. The main purpose of plant macrofossil analysis at Kilen was to find terrestrial material suitable for AMS ^{14}C dating (for which sufficient material was found; see above and Table 3.2). However, all identifiable terrestrial and marine plant macrofossil material was recorded and the results are illustrated in Figure 3.13 and summarised below in relation to the statistically determined macrofossil zones (labelled KIL-MP1 – KIL-MP10).

KIL-MP1 (~5,500-4,450 BC) - Some general tree taxa remains are present belonging to *Betula*, *Quercus* and *Alnus* trees, and several samples contain undifferentiated deciduous leaf, bark or twig fragments. There are also some sporadic finds of moss stems and one sample containing *Juncus* seeds (~5,320 BC),

KIL-MP2 (~4,450-4,105 BC) - KIl-MP2 is almost devoid of plant macrofossils, with the exception of a few levels containing deciduous leaf fragments, an individual *Alnus* female cone at ~4,420 BC and an undifferentiated bud scale fragment at ~4,290 BC.

KIL-MP3 (~4,105-2,855 BC) - *Betula* fruits are commonly present in the record after ~2,700 BC, together with a few *Betula* catkin scales. Moss stems are also sporadically present throughout and *Juncus* seeds are occasionally recorded between ~3,500-3,400 BC.

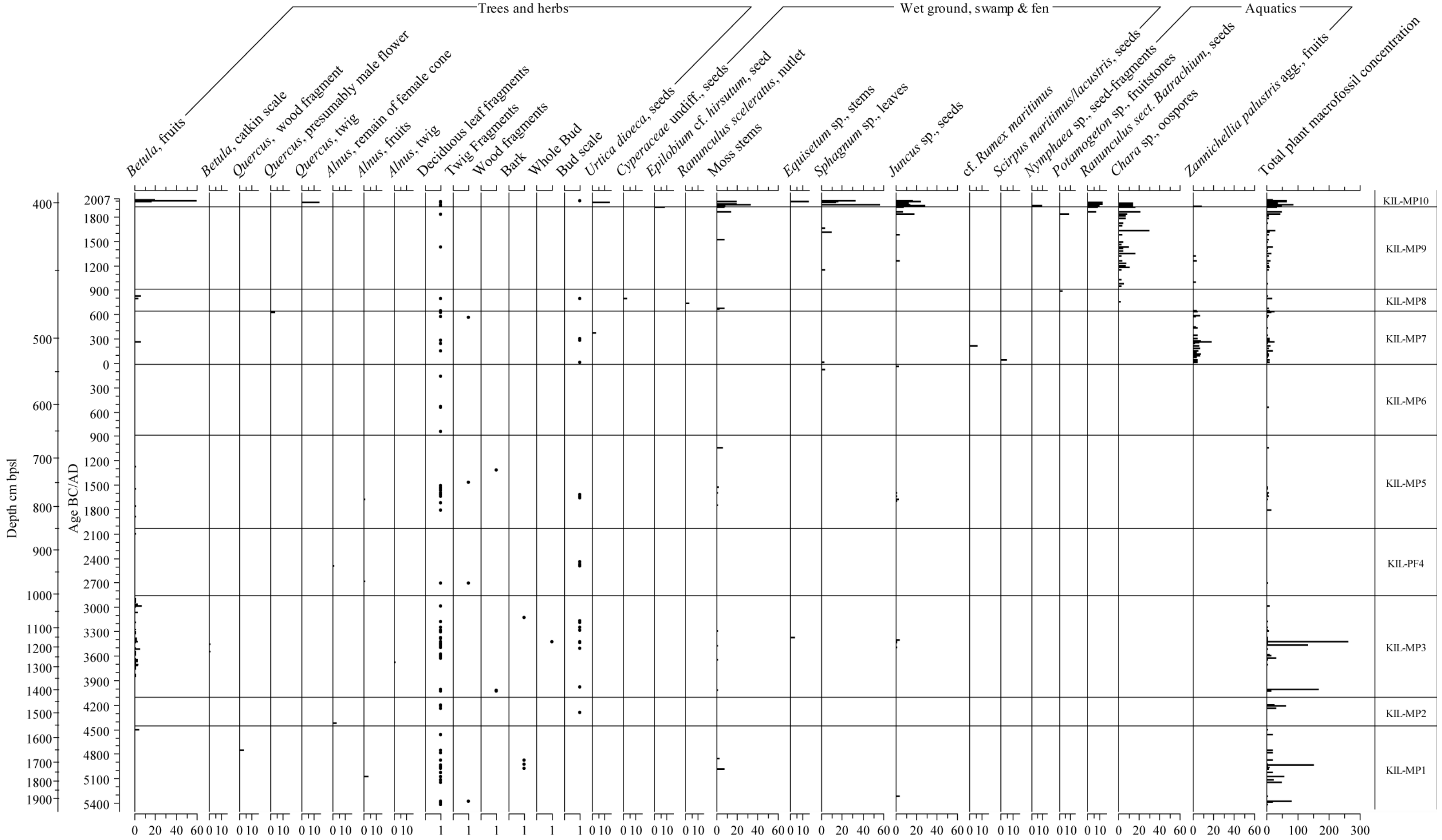


Figure 3.13 - Kilen plant macrofossil record (concentration per 50 ml fresh sediment, unless indicated as presence/absence data by black dots). Analysts: Peter Rasmussen (GEUS) and Jonathan Lewis.

KIL-MP4 (~2,855-2,030 BC) - *Betula* fruits disappear again by ~2,900 BC, and in the phase that follows, the sediment is again almost devoid of macrofossils, except for a few scattered tree remains (e.g. deciduous leaf fragments, alder fruit and female flower).

KIL-MP5 (2,030-890 BC) - Macrofossil concentration remains low in both the lower (~2,030-1,080 BC) and uppermost (after ~1,530 BC) sections, with just occasional *Betula* fruits present throughout and a few moss stems at ~1,045 BC. There is, however, a slight increase in plant macrofossils between ~1,530-1,080 BC as deciduous leaf fragments become more common, and some remains from wet ground taxa (*Juncus* seeds and moss stems) appear in the record.

KIL-MP6 (890-10 BC) - Throughout the majority of this zone, just a few scattered finds of deciduous leaf fragments are recorded. At the very top (after ~75 BC), some swamp/reed taxa appear, most notably *Sphagnum* leaves and *Juncus* seeds.

KIL-MP7 (10 BC- AD 645) - Small amounts of swamp/reed taxa also continue to appear in the record early in KIl-MP7 with finds of *Scirpus* spp. (*maritimus/lacustris*) seeds and cf. *Rumex maritimus* fragments. However, the most pronounced change in this zone is the large expanse of fruits from the aquatic horned pondweed *Zannichellia palustris*, This expanse occurs at the onset of this zone and is regularly present (and at fairly consistent concentrations; generally 3-8 per 50 ml, max. 18.75 at AD ~260) up until ~AD 270. Above this (i.e. after ~AD 270), it becomes more sporadic, causing the overall plant macrofossil concentration to drop briefly.

KIL-MP8 (~AD 645-910) - *Z. palustris* disappears after the first sample in KIL-MP8 (~AD 650), leaving this zone containing just several recordings of *Betula* fruits (~AD 795-825) and some individual remains of several taxa (including *Cyperaceae* seed, *Ranunculus sceleratus* nutlet, moss stems, *Potamogeton* fruitstone and *Chara* spp. oospores). Some undifferentiated leaf and bud remains are also present.

KIL-MP9 (AD ~910-1930) - *Chara* spp. oospores appear at ~ AD 940 and fluctuate thereafter and *Z. palustris* is recorded several times in the lowermost section (~AD 1000-1320). After ~AD 1830, wetland taxa increase in abundance, most notably *Juncus* spp. seeds at ~AD 1830 and slightly later *Ranunculus* sect. *Batrachium* (water crowfoot) seeds and moss stems (both at ~1860 AD). The overall plant macrofossil concentration also increases at the top of this zone (i.e. after ~AD 1830).

KIL-MP10 (AD 1930-2007) - *Chara* spp. oospores remain relatively abundant in the early part of KIL-MP10 (see below) but disappear from the record after ~AD 1970. Wet ground, swamp and reed taxa continue to increase in abundance, with *Juncus* spp., moss stems and *Sphagnum* leaves fairly regularly recorded throughout. *Betula* fruits are present in a few of the uppermost samples (after ~AD 1994) and deciduous leaf fragments become more regular between ~AD 1980 and 1994. Occasional seeds from the macrophyte *Ranaunculus* sect. *Batrachium* continue up until ~AD 1983.

3.7.3 Foraminifera

The foraminifera assemblages present in 15 samples from Kilen are presented in Figure 3.14. The results are displayed as percentages with a minimum of 300 foraminifera counted per sample (by Peter Kristensen, Aarhus University). Fourteen species from seven genera were identified in total.

KIL-MF1 (~5,500-4,450 BC, 5 samples) - *Elphidium incertum* (~24-28 %) is initially high but decreases in abundance (down to ~8-12 %) after ~5,250 BC. *Haynesina germanica* numbers generally fluctuate (~5-28 %), briefly being the most abundant species at ~5,250 BC. *Ammonia beccarii* and *Elphidium excavatum* fo. *selseyensis* gradually increase throughout reaching ~46 % and 27 % respectively by the uppermost sample (~4,510 BC). *Elphidium magellanicum* is also initially quite abundant (~17 %), but briefly drops to ~2 % at ~5,250 BC, before fluctuating between ~5-11 % thereafter. Foraminfera diversity is greatest in this lowermost zone.

KIL-MF2 (~4,450-4,105 BC, 1 sample) - High numbers of *E. incertum* (~32 %), *E. excavatum* (~28 %) and *H. germanica* (~28 %) indicate high salinity at ~4,210 BC. *A. beccarii* accounts for just 6 % of the foraminiferal assemblage in this sample.

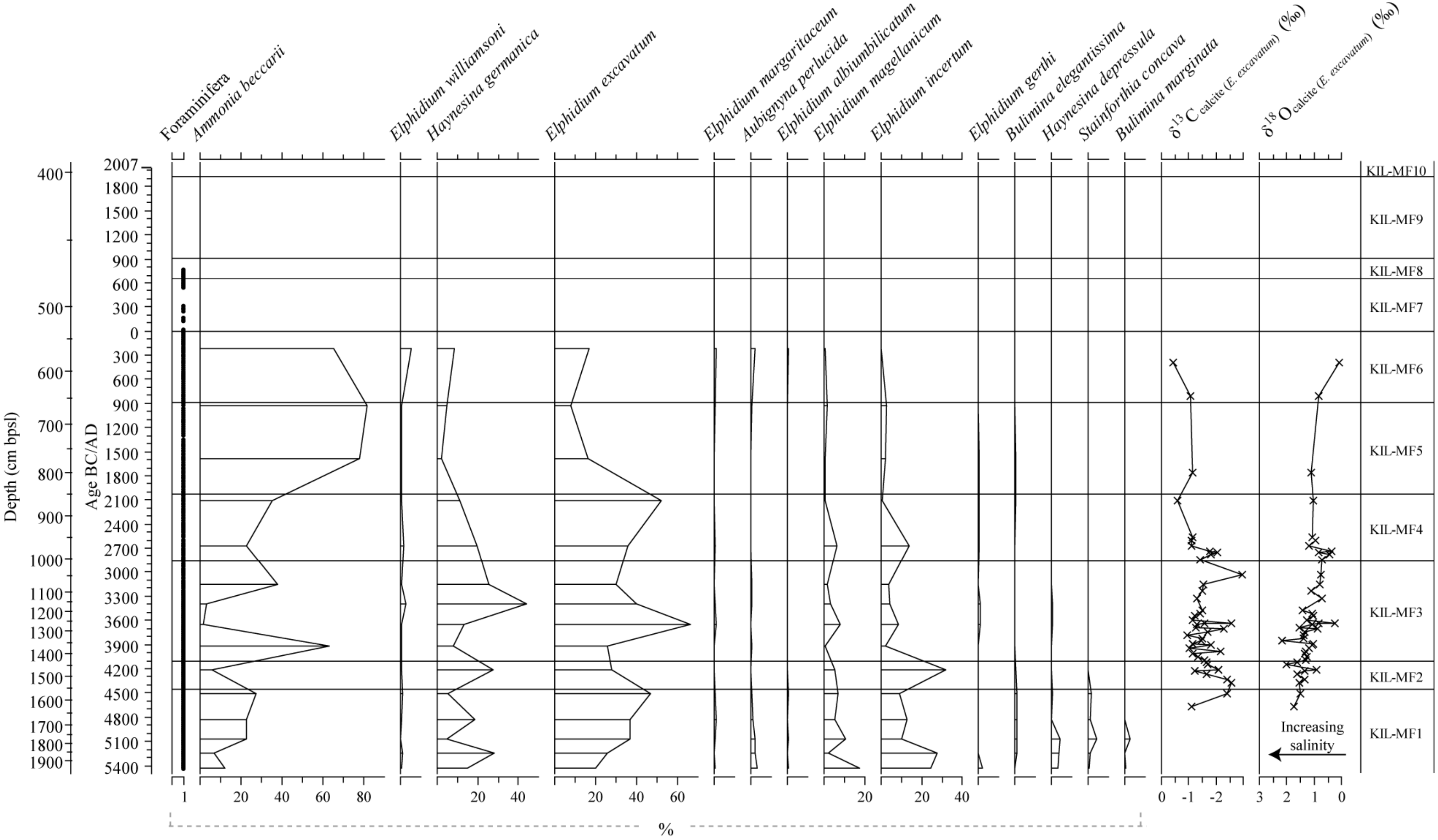


Figure 3.14 – Kilen percentage foraminifera record and $\delta^{18}O_{calcite}$ and $\delta^{13}C_{calcite}$ isotopes based on test of the benthic foraminiferal species *Elphidium excavatum* fo. *selseyensis*. (Analysts – Foraminifera Peter Kristensen Aarhus University, Isotopes: Jonathan Lewis)

KIL-MF3 (~4,105-2,855 BC, 4 samples) - At ~3,920 BC (lowermost sample of KIL-MF3) *A. beccarii* is the dominant species accounting for ~60 % of the assemblage. *E. excavatum* (26-66 %) and *H. germanica* (8-44 %) are the next most abundant species, briefly replacing *A. beccarii* between 3,650-3,400 BC, which drops to less than 4 % of the assemblage. There are also minor increases of *E. incertum* and *Elphidium magellanicum* in this period. However, by ~3,150 BC (upper sample), *A. beccarii* (~38 %) has recovered, and again co-dominates the assemblage (with *E. excavatum*; 30 % and *H. germanica*; 26 %).

KIL-MF4 (~2,855-2,030 BC, 2 samples) – The lowermost sample (~2,680 BC) is dominated by a mixture of *A. beccarii* (~23 %), *H. germanica* (~20 %) and *E. excavatum* (~36 %), but with elevated percentages of *E. incertum* (14 %) and *E. magellanicum* (6 %). *A. beccarii* (~35 %) and *E. excavatum* (52 %) increase substantially in the upper sample, as *H. germanica* decreases to 11 % of the assemblage and *E. magellanicum* and *E. incertum* almost disappear from the record (both <1 %).

KIL-MF5 (2,030-890 BC, 2 samples) - At ~1,590 BC (lower sample) *A. beccarii* completely dominates, accounting for ~80 % of the assemblage. *E. excavatum* is the next most abundant species in both samples, but exhibits a decrease of ~8 % between the two (i.e. between ~1,585-930 BC), whilst *H. germanica* numbers are initially very low (2 % at ~1,585 BC), but increases slightly in the upper sample (reaching ~5 % at ~930 BC).

KIL-MF6 (890-10 BC, 1 sample) - *A. beccarii* (~65 %) is still the dominant species (at ~220 BC), followed by *E. excavatum* (17 %) and *H. germanica* (9 %). There is also an increase in *Elphidium williamsonii* (~5 %), whereas *E. incertum* disappears from the record.

No foraminifera samples have been counted in macrofossil zones 7-10. However, a preliminary assessment of 138 samples (foraminifera presence/absence recorded) from these zones indicates a series of phases in KIL-MF7 where foraminifera are absent. These phases occur between ~AD 12-100, ~AD 165-225 and ~AD 315-520. Towards the middle of KIL-MF8 (after ~AD 755), foraminifera disappear and remain absent throughout KIL-MF9 and KIL-MF10.

3.8 Isotopes from *Elphidium excavatum* fo. *selseyensis*

The $\delta^{18}O_{calcite}$ and $\delta^{13}C_{calcite}$ results determined from tests of the benthic foraminifer *Elphidium excavatum* fo. *selseyensis* are illustrated in Figure 3.14. In total 44 measurements were made for both $\delta^{18}O_{calcite}$ and $\delta^{13}C_{calcite}$, covering the period ~4,670-390 BC. Isotopic values are uncorrected for any vital effect that may take place during synthesis of the foraminiferal test, which can skew the isotopic composition out of equilibrium with the surrounding water (Grossman, 1987). It is assumed that the vital effect for any single species will be consistent and stationary over the timescale investigated in this study (Kristensen and Knudsen, 2006).

The $\delta^{18}O_{calcite}$ isotopic values are highest at the bottom of the profile and exhibit an erratic but overall decreasing trend up until ~2,740 BC, falling by ~1.3 ‰ over this period. Soon after ~2,740 BC, the isotopic values increase again and thereafter fluctuate between ~ 0.8-1.2 ‰ up until ~810 BC, before experiencing a large drop in the uppermost isotope sample (390 BC). In this setting, salinity, rather than temperature is likely to be the most important factor influencing the $\delta^{18}O$ water composition of the seawater (K. Knudsen pers. comm.). This is due to the much larger magnitude shifts in salinity over the study period, compared to temperature (e.g. salinity gradient from brackish to fully marine; Kristensen *et al.*, 1995; Christensen *et al.*, 2004, temperature changes ~3°C over the entire Kilen profile; Antonsson and Seppä, 2007). Whilst the modern calcite-salinity and calcite-temperatures relationships are unknown for *Elphidium excavatum* fo. *selseyensis* in the Limfjord, estimates on modern periwinkle shells (*Littorina littorea*) exhibit a 0.25 ‰ change for every salinity unit (i.e. 1 ‰) and a 0.22 ‰ per °C change in temperature (Burman and Schmitz, 2005). Therefore the $\delta^{18}O_{calcite}$ profile suggests that salinity decreases up until ~2,740 BC (i.e. less negative values infer higher salinity), prior to a relatively sharp increase at ~2,680 BC, and thereafter remains relatively constant for most of the upper section of the isotope record (~2,740-810 BC).

The $\delta^{13}C_{calcite}$ isotopic curve also exhibits an erratic pattern in the lower section of the record, with large fluctuations up until ~3,600 BC. In this phase, values generally fluctuate between -1 to -2.5 ‰. Most of the lower $\delta^{13}C_{calcite}$ values (i.e. < -2 ‰ occur in individual samples (e.g. 3,630 BC, 3,710 BC and 3,980 BC), with the exception of a distinct phase between 4,510-4,330 BC. This ensues after a ~1.5 ‰ decrease in the $\delta^{13}C_{calcite}$ isotopic composition shortly after ~4,670 BC (i.e. immediately above the lowermost isotope sample which yielded a $\delta^{13}C_{calcite}$ isotopic ratio of -1.09 ‰). The $\delta^{13}C_{calcite}$ isotopic composition becomes more stable for a brief period between 3,600-3,155 BC, prior to a ~1.5 ‰ drop at ~3,040 BC, where it falls to its lowest value for the entire profile. However, this minimum appears to be very short lived, and in the sample immediately above (i.e. by ~2,850 BC), isotopic values increase again, a trend that largely continues until ~2,100 BC. Above this, the sampling resolution is very low, but it appears that the $\delta^{13}C_{calcite}$ values remain relatively high (between -1.2 and -0.4 ‰) for the upper part of the isotopic record (i.e. ~390 BC).

3.9 Other organisms

Any additional remains present in either the diatom slides or wet sieved samples were recorded and results are illustrated in Figure 3.15. In numerous diatom slides, small numbers of chrysophytes were found as well as occasional recordings of the marine silicoflagellate *Distephanus speculum* and the ebridian species *Ebria tripartita*. Additional macro-organisms (present in the wet sieved samples) include ostracods, fish (bones and scales), echinoids (spines), barnacles (*Balanus* spp.) and zooplankton remains (*Daphnia* spp. ephippia).

Chrysophytes, expressed as concentrations, after methods described in Battarbee and Kneen, (1982), are present for much of the profile appearing in 62 of the 110 diatom samples. Highest concentrations are found in the uppermost, more bracksih section of the profile (after ~AD 1930), but two other phases of apparent elevated concentration occur between ~4,150-3,330 BC and in the late pre-Roman/early-Roman period (~280 BC-AD 260). *Distephanus speculum* is regularly present up until ~2, 380 BC, with the exception of a ~600 year disappearance between ~3,330-2,740 BC. After ~2,380 BC, it is only recorded four more times, all of which occur between ~310 BC-AD 600.

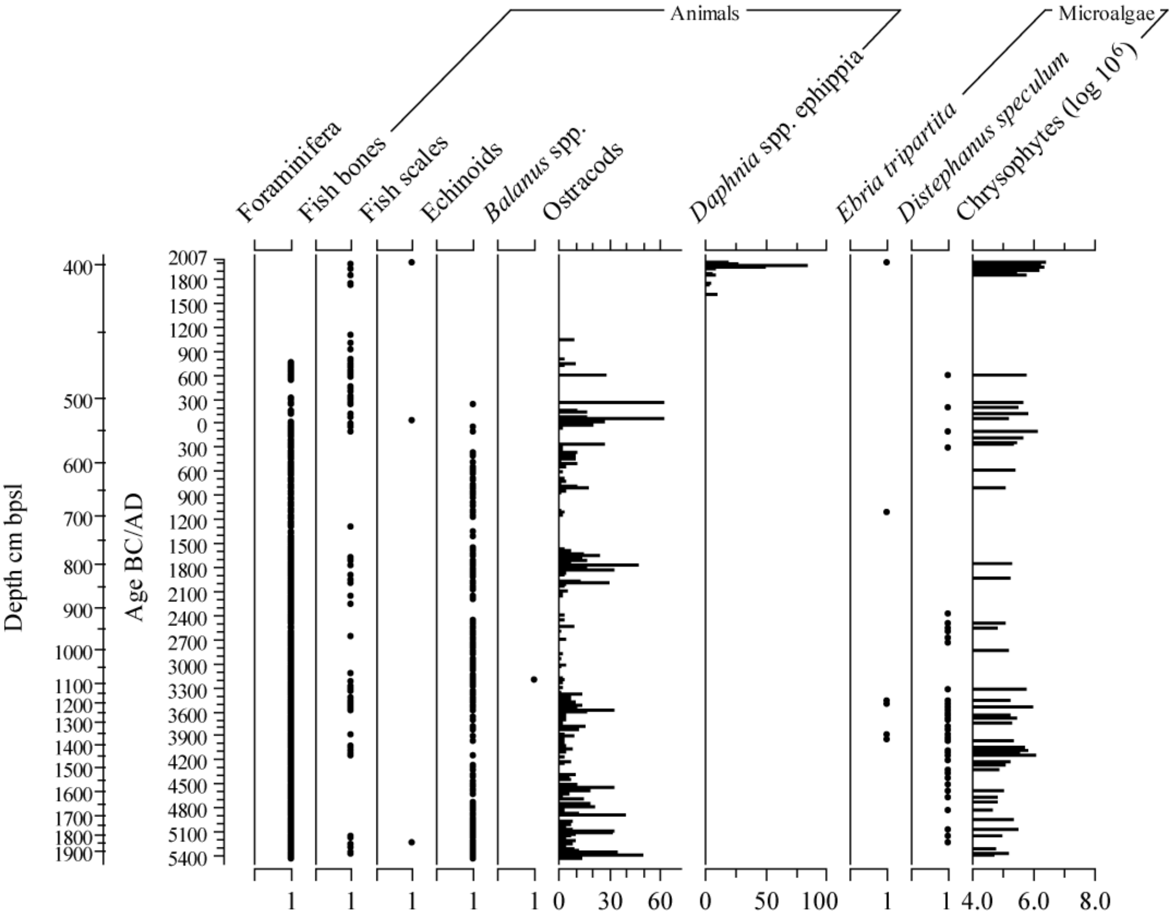


Figure 3.15 - Additional organisms present at Kilen. Ostracods and *Daphnia* spp. (Cladocera) are expressed as concentration data per 50 ml fresh sediment. Chrysophyte concentrations are expressed as number per gram of dry weight. All other organisms are expressed as presence/absence data.

Ostracods (expressed as concentrations per 50 ml wet sediment in Figure 3.15) are consistently present in the record up until ~AD 1030. There are four apparent phases of slightly elevated concentrations; 5,500-4,550 BC, 3,910-3,400 BC, 2,170-1,580 and ~65 BC-260 AD, with highest concentration occurring during the latter of these phases. It must, however, be noted that even during these four phases, absolute numbers of ostracods are still relatively low (maximum 32 specimens per sample), and therefore results must be treated with caution. Furthermore, without identification to species level it is impossible to derive any information regarding salinity (and/or temperature) from just their presence within the profile (as species differ widely in their salinity and temperature requirements). Nonetheless, in conjunction with the other proxies, they may be able to offer some insight into calcite preservation and/or aquatic productivity.

Like ostracods, undifferentiated fish bones provide little information concerning the environment and are always low in abundance, but again can potentially provide some insight into changing calcite preservation conditions. Periods where fish bones are more regular include ~5,500-5,170 BC, ~4,150-3,330, ~2,250-1,725 BC, ~120 BC-AD 1085 and ~AD 1720-1990. Echinoid remains are fairly regular up until ~AD 240, after which they disappear from the record. *Daphnia* spp. ephippia (zooplankton) only appear in the uppermost section of the record, first appearing at low concentrations around ~AD 1600. They exhibit a substantial increase in the early 1930s (KIL-MP10) and thereafter remain both common and relatively abundant up until ~AD 1996. *Daphnia* spp. eggs disappear again in the very recent past (i.e. post-AD 1996).

3.10 Statistical analyses: DCA

Major trends in the diatom dataset were explored using detrended correspondance analysis (DCA) and principal component analysis (PCA) using Canoco, v4.5 (ter Braak and Šmilauer, 2002). For the entire diatom dataset (i.e. 110 samples), DCA (unimodal) was deemed the most appropriate technique, due to the axis 1 gradient length exceeding 2 SD units (Table 3.3). The first DCA axis is responsible for explaining 26.8 % of the variation in the Kilen diatom dataset, with very little variance in comparison being explained by any of the additional axes (Table 3.3). DCA axis 1 appears to be driven by salinity, which is perhaps expected for this site when considering the substantial shifts in salinity evident over the last ~7,500 years. The majority of species appear to be strongly associated with axis 1, with those present in the uppermost (limnic) section of the profile exhibiting the lowest (i.e. most negative) score on DCA axis 1. Most notably (i.e. >3 % abundance in one or more slides) these include *Cyclotella choctawatcheeana, Cyclotella meneghiniana, Ctenophora pulchella, Thalassiosira baltica* and *Rhoicosphenia abbreviata*. At the other end of the gradient (i.e. positive values along axis 1) are the high-salinity tolerant species such as *Cymatosira belgica, Delphineis minutissima, Actinoptychus senarius* and *Paralia sulcata*.

DCA axis 1 sample scores are illustrated in Figure 3.8 and Figure 3.9 as they provide a summary of major compositional shifts in the diatom assemblage throughout the profile. The shifts in the DCA axis 1 sample scores tie in with the major salinity shifts from the DI-reconstruction. This adds further support for salinity as a major factor driving the change in the diatom assemblages. Only minor compositional changes occur in the lower section (~5,500-2,740 BC) of the profile, but after ~2,750 BC, DCA sample scores begin to increase as higher-salinity demanding species begin to dominate the profile. Highest DCA axis 1 sample scores generally occur between ~400-40 BC, prior to the major drop at the beginning the Roman period (KIL-D3; ~AD 20). A brief return to higher scores occurs at ~AD 800, where *P. sulcata* dominates the profile again, before dropping down to lowest sample scores for the entire profile in the uppermost, limnic section of the record.

Dataset	No. of samples	No. of species	Total inertia	Axis	1	2	3	4	Technique
Diatoms (%)	110	516	2.026	Eigenvalues	0.543	0.076	0.056	0.042	DCA
				Length of gradient	3.028	1.791	1.223	1.062	
				% variance explained	26.8	3.7	2.8	2.1	
	27	169		Eigenvalues	0.379	0.137	0.069	0.054	PCA
				% variance explained	37.9	13.7	6.9	5.3	
	82	477	1.345	Eigenvalues	0.208	0.115	0.047	0.039	DCA
				Length of gradient	1.956	1.555	1.005	1.032	
				% variance explained	15.4	8.6	3.5	2.9	

Table 3.3 - Summary of DCA and PCA results for the Kilen diatom dataset

As the whole-core variation is driven by the contrast between two very different aquatic ecosystems (limnic and marine, the diatom dataset was subsequently split into two smaller datasets to further explore the major trends within both the limnic and marine phases at Kilen. The limnic phase (i.e. post AD 1856) is most likely to be represented by the uppermost 27 slides (see section 3.3 and 3.11 for dating uncertainties) and a preliminary DCA on this reduced dataset suggested linear analyses (i.e. axis 1 = <1.5) were the more appropriate. Following PCA analysis (applying Aitchison's log ratio analysis; Aitchison, 1986), axis 1 explained 37.9 % of the variation with a further 13.7 % explainable by axis 2. PCA axis 1 sample scores (Figure 3.9) exhibit an overall increasing trend, which accelerates in the 1970s/early 1980s, resulting in highest overall scores occurring at the very top of the profile (~AD 1985-2007). The major gradient in the upper section of the profile appears to be predominately driven by nutrients, but possibly also influenced by water depth, sediment type and the abundance of macrophytes.

Species scoring lowest (i.e. most negative) on PCA axis 1 are mainly benthic species, which are most abundant in the pre-1920 period. These include *Amphora pediculus, Cocconeis neothumensis, C. placentula* agg., *R. curvata*, *D monoliformis, Ctenophora pulchella* and *E. sorex* (Table 3.4). These species gradually decline between ~AD 1920-1980 and almost disappear from the record after ~AD 1985, likely due to increasing salinity and/or increased nutrient status (as indicated by increasing salinity, nitrogen and phosphours concentrations; Ringkjøbning Amtskommune, 1988). The planktonic diatom, *Thalassiosira baltica* provides an exception to this by scoring negatively on PCA axis 1, but exhibits a similar trend to the low scoring benthics and practically disappears from the record at ~AD 1970, possibly being outcompeted by *Chaetoceros* spp. (24-53 % between ~AD 1980-1994).

Species scoring highest on PCA axis 1 are generally more brackish diatoms, most notably *O. mutabilis* and *S. construens* fo. *subsalina*, but to a lesser extent *Chaetoceros* resting cysts. These diatoms reach their maximum abundances between ~AD 1980-1994 (particularly ~AD 1988-1994 for *Chaetoceros* cysts) following the substantial reduction in both the abundance and diversity (e.g. reduction in Hills N_2; Figure 3.9) of the benthic community. The diatom-salinity model performs poorly in this section of the record (see Section 3.11), but salinity data measured between AD 1972-1987 shows that salinity increased after AD 1984, due to increased water exchange (i.e. greater input of higher salinity water) through the Kilerkanal, following the installation of a new sluice system (Ringkjøbing Amtskommune, 1988). The disappearance of diatoms such as *C. neothumensis* and *C. placentula* agg., both of which are commonly present throughout the marine phase of the profile, suggests that PCA axis 1 does not represent a salinity gradient. The new sluice system (built in AD 1984) increased phosphorus-release from the sediments following marine water input, causing regular decreases in oxygen concentration of the bottom waters (Ringkjøbing Amtskommune, 1988). Furthermore, DI-TN continued to gradually increase (as does the planktonic: benthic and centric: pennate ratios) between ~AD 1970-2000, which combined with elevated phosphorus concentrations, probably forced the reduction in the benthic community, by enabling planktonic diatoms (and other organisms) to bloom, resulting in increased water turbidity and reduced light penetration (Cederwell and Elmgren, 1990; Bonsdorff *et al.*, 1991; Cooper, 1995a; Clarke *et al.*, 2003; Ellegaard *et al.*, 2006). This might suggest a degree of co-variation between nutrients and salinity, with increasing nutrient status responsible for the elevated abundance of planktonic diatoms (and supressed benthic community) and *Chaetoceros* spp. perhaps outcompeting *C. choctawatcheeana* as the dominant component of the phytoplankton under higher salinities (i.e. between ~AD 1970-1994).

Increased numbers of *O. mutabilis* (and *S. construens* fo. *subsalina*) might also result from sandier sediments and/or decreased benthic competition following a reduction in the benthic community after ~AD 1980. Other benthic plants (e.g. submerged macrophytes) might also have been affected by increased numbers of phytoplankton (i.e. being shaded out), leading to a reduction in the availability of habitats for some species of diatom (e.g. epiphytic diatoms such as *R. abbreviata* and *C. placentula*). Furthermore, it is also clear that *O. mutabilis* has an affinity for, or at least is able to tolerate relatively high nutrient conditions being relatively abundant (7-40 %) in all slides in the upper section of the Kilen diatom record, again supporting co-variation between salinity and nutrient concentration in the upper section.

Taxa	**PCA axis 1**
Opephora mutabilis	0.82
Staurosira construens fo. *subsalina*	0.64
Chaetoceros resting cysts	0.29
Cyclotella choctawatcheeana	-0.01
Cyclotella meneghiniana	-0.17
Planothidium delicatulum agg.	-0.36
Epithemia sorex	-0.62
Ctenophora pulchella	-0.64
Diatoma moniliformis	-0.72
Rhoicosphenia curvata	-0.79
Thalassiosira baltica	-0.79
Cocconeis placentula agg.	-0.84
Cocconeis neothumensis	-0.85
Amphora pediculus	-0.96

Table 3.4 - PCA axis 1 species scores for taxa >3 % abundant in one or more slides in the uppermost section of the Kilen diatom record (27 samples; post AD 1856).

For the reduced dataset, spanning the marine section of the profile (i.e. 86 samples between ~5,500 BC-AD 800), species scoring lowest (most negative) are the high salinity species (*C. belgica, D. minutissima, P. sulcata* and *A. senarius*), which are particularly abundant between ~400-0 BC. The majority of other species (present at >3 % abundance) score positively, with absolute highest scores exhibited by species (*Chaetoceros* cysts, *Bacillaria paxillifer, S. costatum, Achnanthes amoena*) which are most abundant during the lower DI-salinity phases (e.g. KIL-D1b ~AD 50-600). This again suggests that salinity is the most important variable driving DCA axis 1 and the associated variation in the diatom assemblage. Inclusion of floristically unusual the sample at depth 437 cm bpsl (~AD 1,430) does not affect the main gradient of change, with salinity still being the main driving factor. Species abundant in this particular sample (*E. areanaria*, *E. frickei* and *A. pediculus*), however, appear to score very high on axis 1.

The DCA sample scores (Figure 3.7) again provide a good summary of the major compositional shifts in the diatom record, with highest DCA sample scores associated with lower salinity periods, and sample scores decreasing as salinity increases. Sample scores fluctuate (between 1 and 2) at the bottom of the record, before decreasing after ~3,500 BC, driven mainly by the increase in higher salinity diatoms, but also by the decrease in planktonic diatoms (most notably *Chaetoceros* cysts) up until ~2,700 BC. Sample scores generally decrease up until the pre-Roman period (~500-0 BC), at which point almost fully marine conditions are inferred by the diatom record (i.e. highest DI-salinities; Figure 3.7) and DCA sample scores reach their lowest values. A shift back to higher scores occurs in the Roman/early Germanic period (~AD 0-600), where DI-salinity suggests more brackish-marine conditions (i.e. 14-19 ‰), before another low score in the sample at ~AD 800 (DI-salinity of ~33 ‰).

DCA axis 2 species scores are also provided in Table 3.5, and habitat preferences stated in Snoeijs *et al.*, 1993-1998, Sabbe and Vyverman, 1995, Hasle and Syvertsen, 1996 are given. Highest sample scores are generally recorded between ~5,500-0 BC prior to a major drop at ~0 AD/BC, after which low values prevail up until ~AD 600. This axis explains 8.6 % of the variation in the diatom dataset and is most likely driven by substrate and habitat preference, with epiphytic species being most abundant between ~AD 0-600 BC scoring lowest. This period coincides with the regular appearance of *Z. palustris* in the plant macrofossil record (Figure 3.13), likely indicating increased presence of macrophytes providing greater habitat availability for epiphytic diatoms. For taxa >3 % abundant, pelagic or sediment-dwelling (e.g. epipsammic, epilithic and epipelic) diatoms score highest (see Table 3.5), but the majority of taxa also appear to score positively on DCA axis 2. This is probably due to reduced macrophyte populations during the higher salinity (more marine) phase of the Kilen profile (i.e. ~5,500-0 BC), which likely favoured sediment-dwelling (benthic) diatoms.

Taxa	**DCA axis 1**	**Taxa , habitat**	**DCA axis 2**
Chaetoceros resting cysts agg.	2.66	*Skeletonema costatum* , pelagic	3.57
Melosira moniliformis	2.57	*Opephora minuta*, epipsammic, Sabbe and Vyverman, 1995	2.50
Bacillaria paxillifer	2.33	*Dimmergramma minor*, in or on sediments	2.49
Epithemia turgida var *westermanii*	2.31	*Plagiogramma staurophorum*, in or on sediments	2.17
Staurosirella pinnata var *pinnata*	2.15	*Bacillaria paxillifer*, epipelic and epilthic	2.16
Skeletonema costatum	2.10	*Thalassionema nitzschoides*, pelagic	2.14
Achnanthes amoena	2.07	*Cyclotella striata*, pelagic, Hasle and Syvertsen, 1996	2.08
Opephora mutabilis	1.77	*Fallacia* sp. 1, unknown, probably epipelic	1.96
Dimmergramma minor	1.72	*Rhopalodia acuminata*, epipelic and epilithic	1.94
Navicula cf. *oestrupei*	1.70	*Staurosirella pinnata* var. *pinnata*, epiphytic, Hendey, 1964	1.85
Cyclotella striata	1.61	*Thalassiosira decipiens*, pelagic, Hasle and Syvertsen, 1996	1.67
Cyclotella choctawatcheeana	1.59	*Paralia sulcata*, in or on sediments or pelagic	1.62
Fallacia sp. 1	1.58	*Cyclotella choctawatcheeana*, perlagic, sometime epilithic	1.62
Staurosira elliptica agg. MOLTEN	1.47	*Fragilaria gedanensis*, epipsammic	1.51
Rhoicosphenia abbreviata	1.43	*Chaetoceros* cyst agg., pelagic	0.97
Thalassiosira decipiens	1.37	*Grammatophora oceanica* , epiphytic	0.94
Cocconeis placentula agg.	1.31	*Actinoptychus senarius*, in and on sediments, occasionally in plankton	0.92
Thalassionema nitzschoides	1.20	*Cocconeis neothumensis*, epipsammic	0.39
Rhopalodia acuminata	1.13	*Staurosira elliptica* agg. MOLTEN, epipsammic	0.32
Cocconeis scutellum var *scutellum*	1.08	*Opephora mutabilis*, epipsammic	0.25
Plagiogramma staurophorum	1.04	*Cymatosira belgica*, epipsammic, Hendey, 1964	0.24
Cocconeis neothumensis	1.02	*Delphineis minutissima*, epipsammic, Hendey, 1964	0.21
Grammatophora cf *oceanica*	0.95	*Amphora helenensis*, epipelic	-0.02
Opephora minuta	0.90	*Cocconeis scutellum* var. *parva*, epiphytic	-0.04
Amphora helenensis	0.85	*Cocconeis scutellum* var. *scutellum*, epiphytic	-0.64
Cocconeis scutellum var *parva*	0.73	*Navicula perminuta*, epilithic	-0.69
Navicula perminuta	0.58	*Cocconeis placentula* agg., epiphytic or epilithic	-1.38
Fragilaria gedanensis	0.49	*Rhoicosphenia abbreviata*, epiphytic or epilithic	-1.72
Paralia sulcata	-0.27	*Navicula* cf. *oestrupei*, unknown, probably epilithic	-2.18
Actinoptychus senarius	-0.42	*Epithemia turgida* var. *westermanii*, epiphytic or epilithic	-2.32
Delphineis minutissima	-0.86	*Achnanthes amoena*, epiphytic or epilithic	-2.75
Cymatosira belgica	-1.38	*Melosira moniliformis*, epiphytic or epilithic	-3.52

Table 3.5 - DCA axis 1 species scores for taxa >3 % abundant in one or more slides in the uppermost section of the Kilen diatom record (82 samples; ~5,430 BC-AD 800). Habitat preferences are based on Snoeijs *et al.*, 1993-1998, unless otherwise referenced.

If the dataset is further reduced to exclude the brackish water phase between ~AD 50-600 (75 samples), DCA axis 2 appears to be either driven by nutrient status or productivity. Planktonic species more abundant in KIL-D1b (e.g. *Skeletonema costatum*, *C. choctawatcheeana*, *C. striata*, *Rhizolsolenia* spp.; ~4,250-2,740 BC) generally score lowest, with benthic taxa, more abundant in other parts of the sequence scoring highest (most notably *C. belgica*, *D. minutissima*, *C. scutellum*, *D. minor* and *G. oceanica*). Despite excluding the major salinity shift after ~0 AD/BC, salinity still drives DCA axis 1.

3.11 Synthesis of results and site discussion

This section provides a synthesis of results derived from the independent proxy records analysed for the Kilen sedimentary profile (described above) and a site discussion, assessing how the marine/limnic environment has changed over the last ~7,500 years at Kilen. A summary of all proxy results is provided in Figure 3.16 plotted against regional environmental parameters such as sea level (Behre, 2007) and temperature (Antonsson and Seppä, 2007). Based on relatively good agreement between the independent proxy zones (see Figure 3.16), the Kilen profile has been split up into five environmental phases (labelled KIL-E1-KIL-E5). Environmental phase markers have been positioned in places where two or more independent proxy zone markers (determined via optimal splitting) fall within close temporal proximity to one another. The environmental history of Kilen is discussed below in relation to these five phases.

KIL-E1 (~5,500-4,200 BC) - *The late Atlantic phase* - In the lowermost phase, DI-salinities (range 18-29 ‰, average 24 ‰) are generally lower than the present day salinity of the Limfjord (average 25 ‰; 27 ‰ for the Struer region), which suggests a reduced connection with the North Sea and/or the Skagerrak. This might be considered contradictory to other studies (e.g. Petersen, 1981; Andersen, 1992) which suggest the Limfjord was more open during the late Atlantic period, due to higher sea levels (4-6 m; Petersen, 1981). This is believed to have created entrances to both the North Sea, in the west, and to the Skagerrak, in the north (Figure 3.17b).

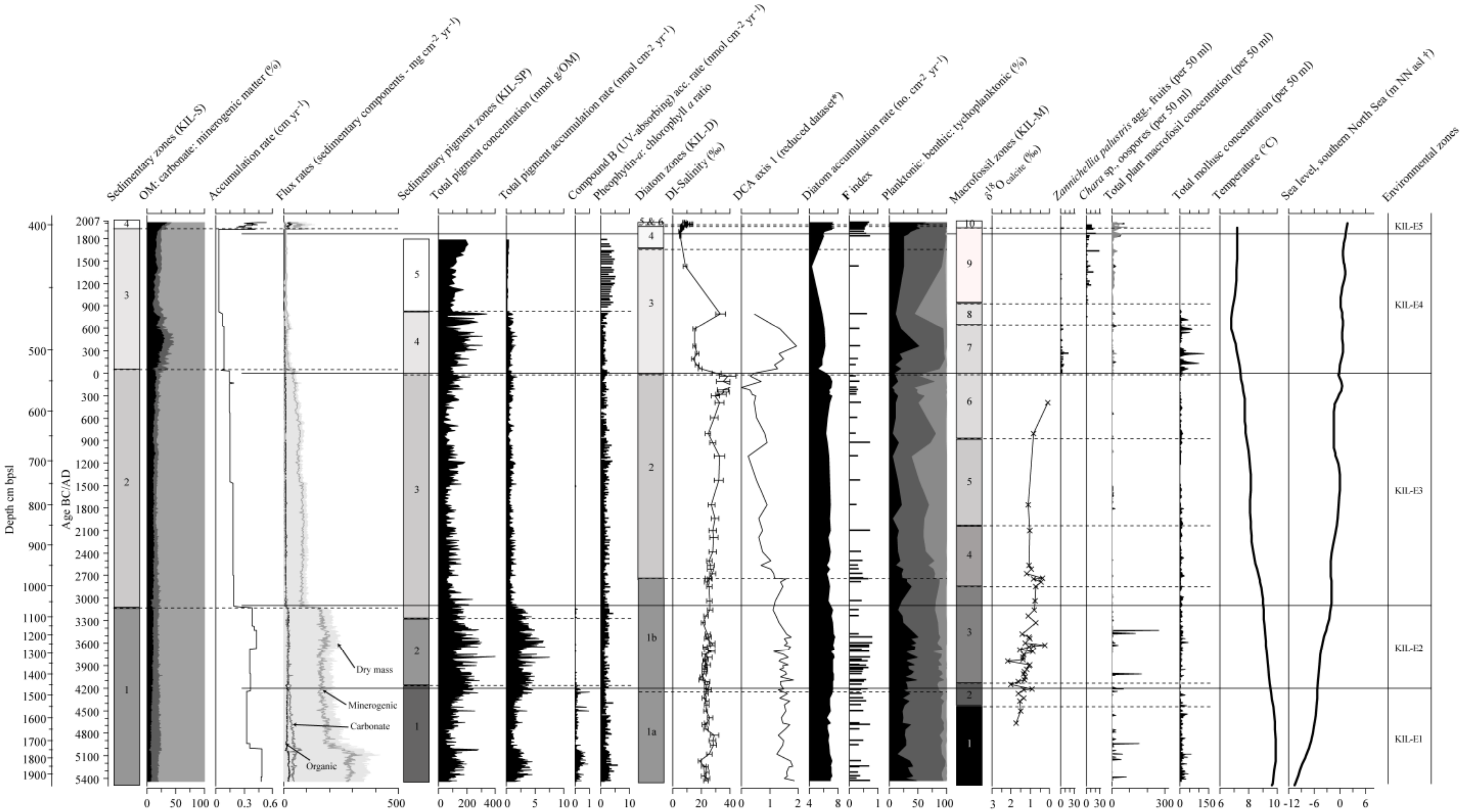


Figure 3.16 - Summary diagram of the major changes occurring at Kilen (~5,500 BC-AD 2007). Regional (pollen-based) temperature curve from Lake Trehörningen (Antonsson and Seppä, 2007) and (mean high water) sea level curve from the southern North Sea (Behre, 2007). *DCA axis 1 reduced dataset sample scores excludes the uppermost 28 diatom samples (see text for details). †m NN asl = metres relative to German ordnance datum above present sea level.

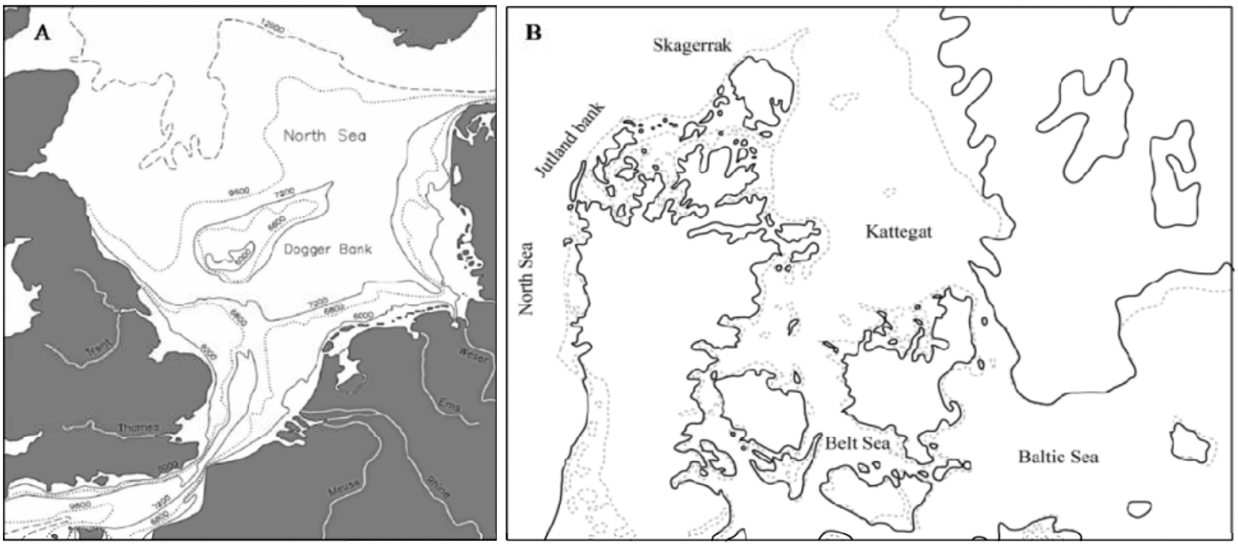


Figure 3.17 - A. Palaeo-shorelines of the North Sea (in years BC) as proposed by Behre, (2007). B. Map of Denmark and the western Baltic during the late Atlantic Ertebølle period (~5,400-3,900 BC) indicating present shoreline (dotted line) and proposed palaeo-shorlines (solid line). Map re drawn and modified from Enghoff *et al.*, (2007) after Hellesen and Tuxen, (1988). Note the more open Limfjord with openings to both the north and west.

The history of the Limfjord, in terms of the timings of former opening and closings of these entrances, is poorly constrained, particularly before ~AD 1200. It is believed that the northern entrance remained open until sometime between ~500-0 BC (Petersen, 1976, 1981; Kristensen *et al.*, 1995; Christensen *et al.*, 2004; discussed in more detail below). According to Burman and Schmitz, (2005) this resulted in almost fully marine salinity (30-33 ‰) in the central and northern areas of the Limfjord during the late Mesolithic. Based on foraminiferal evidence from Vilsund, Andersen, (1992) suggested that the west was open ~8,000 ^{14}C years BP and remained open up until late pre-Roman times (0 AD/BC). The Vilsund site is situated almost equidistant from both openings, probably making it unsuitable for isolation of the predominant marine inputs during the Holocene. Furthermore, all these studies are based on molluscan or foraminferal analyses, dated using marine shells, and therefore most likely subject to chronological problems due to the site and time-variable marine reservoir effect (Heier-Nielsen *et al.*, 1995; Olsen *et al.*, 2009).

Today, a salinity of ~27 ‰ exists in the Struer region due to a narrow opening (~1 km wide) to the North Sea in the western part of the Limfjord (Figure 2.2). For salinities to be lower than ~27 ‰ in the Struer area between ~5,400-2,800 BC, it is likely that the western part the Limfjord was still either totally closed off, or at least only connected to the North Sea via a narrow connection, perhaps located even further north than its current latitude (e.g. Figure

3.17). The emergent Jutland bank (Leth, 1996), might have largely blocked off the western entrance to the North Sea, prior to its drowning sometime around ~4,200 BC (~6,200 yrs BP; Leth, 1996; Gyllencreutz, 2005). The fluctuations in salinity in this early phase (e.g. higher salinity between ~4,960-4,830 BC) are probably related to short term variations in sea level, varying degrees of exposure with the Skagerrak and North Sea and/or variable changes in the strength of the North Sea and Skagerrak current systems.

More localised factors (e.g. precipitation, surface run-off, vegetation change) might also be important for explaining fresher conditions than today before ~2,800 BC. Unfortunately, the absence of a full pollen record (and poor plant macrofossil record) means information concerning changes in the terrestrial environment is limited. More localised sediment banks or beach ridges might have reduced Kilen's connection to the Limfjord during this early period in this section, though no evidence exists for this to date. It is also possible that the salinity model is under-estimating salinity in the lowermost section of the record. The representation of fossil taxa in the model is generally low (40-66 %) between ~5,500-2,800 BC, due to both the high diatom diversity (~70-114 species per sample and high Hills N_2 diversity; Figure 3.8) and relatively low number of training set sites at the high end of the salinity gradient. The high numbers of *Chaetoceros* cysts present between ~5,500-3,480 (8-43 %) are also partly responsible for the low species representation. *Chaetoceros* spp. are predominately a marine genus, but include species that can also live in low salinity waters (Rines and Hargraves, 1988). They are present in all parts of the Kilen profile, including the limnic phase, are unfortunately extremely difficult to differentiate between species and therefore have limited values as salinity indicators.

The common occurrence of high salinity-demanding species in the molluscan record (Figure 3.12) and foraminiferal record (Figure 3.14) might also indicate higher salinity than the diatom inferences suggest. *Bittium reticulatum* and *Tellimya ferruginosa* are present in the molluscan record throughout this phase (i.e. ~5,500-2,800 BC), supplemented by occasional appearances by *Rissoa parva* and *Achanthocardia echinata.* In the foraminiferal record, the presence of *E. incertum* (8-32 %) *E. magellanicum* (5-17 %) and *Haynesina depressula* (2-4 %) suggest salinities of at least 25 ‰ (e.g. Alve and Murray, 1999), particularly before ~4,200 BC. However, in both cases the most abundant taxa are generally medium-high salinity demanding species (i.e. *A. alba*, *M. bidentata*, *C. gibba*, *S. subtruncata* for molluscs, *A. beccarii*, *E. excavatum* fo. *selseyensis* and *H. germanica* for foraminifera). This might suggest the higher salinity-demanding species (listed above) were living in the more exposed outer reaches of the fjord. The isotope record also suggests that salinity was actually highest in this lowermost phase, and decreasing thereafter up until ~2,740 BC. There might, however, be some dissolution issues affecting the isotopic signal, particularly after ~3,200 BC (see below; KIL-E3).

Despite the above possibilities, the bulk of evidence suggests lower salinity than today in this section of the record (i.e. between ~5,500-2,700 BC) due to a likely narrow more northerly opening or complete closure of the west at this time (e.g. Figure 3.17). The relatively low percentages of the heavily silicified marine diatom *Paralia sulcata*, which is often over-represented in coastal sediments (Ryves *et al.*, 2004) might also support this. Relative abundances of this diatom appear to increase markedly during other phases of high salinity in the Kilen profile (e.g. ~2,700-0 BC). Furthermore, DI-salinities appear to agree with other published records (e.g. Kristensen *et al.*, 1995; Christensen *et al.*, 2004) in the uppermost section of the profile (~2,500-1,050 BC), which suggests that the diatom salinity model is accurately predicting salinity change. However, our understanding of the history and timings of marine inputs into the Limfjord will remain fragmentary until long-term, well dated records are available from the northern Limfjord.

High pigment concentrations, accumulation rates and organic matter content between ~5,500-5,000 BC suggest productivity is relatively high at the bottom of the profile. The slightly elevated pheophytin-*a*/chlorophyll *a* ratio might also suggest higher rates of degradation of pigments and therefore productivity might even be under-estimated at the very bottom of the profile. There is, however, a brief decrease in pigment abundances and accumulation rates between ~5,000-4,800 BC, which coincides with increased DI-salinities (28-29 ‰; see above). A decrease in the overall accumulation rate and the organic, carbonate and minerogenic flux rates also occurs between ~5,100-5,000 BC. This suggests that this site is more productive and accumulation of sediments is faster when there is reduced exposure to the Skagerrak and North Sea, but with a regular input of marine water. Linkages between sedimentation, productivity, salinity and the configuration of the Limfjord are also evident in other parts of the profile (see below). After ~4,800 BC, pigment abundances appear to gradually increase again, culminating in a high productivity phase between 4,300-3,150 BC (discussed in more detail in KIL-E2).

High abundance of UV-radiation absorbing pigments are also present in this lowermost phase, suggesting increased depth penetration of UV-radiation (Leavitt *et al.*, 1997). These pigments are primarily produced by benthic sediment dwelling organisms (e.g. cyanobacteria) as photo-protectants when subjected to increased UV-exposure (Leavitt *et al.*, 1997; McGowan, 2007). This might suggest shallower water existed within the fjord or reduced levels of plankton (i.e. benthic diatom dominated system, ~60-80 % between ~5,500-4,200 BC) and dissolved organic carbon (DOC), important in aquatic systems for absorbing or shading out UV radiation. Shallower water is, however, deemed less likely due to the depth of the Kilen basin, with sediment being deposited between ~1,500-1,900 cm bpsl at the same time as sea level reached close to its present day levels (e.g. Petersen, 1981; Behre, 2007). Reduced DOC might be related to lower input from terrestrial sources or from reduced production from phytoplankton and bacterioplankton, with the latter again likely to be influenced by shallower water and subsequent reduced habitat availability for planktonic organisms.

KIL-E2 (~4,200-3,100 BC) - *High productivity phase* - A substantial increase in marine productivity occurs in this phase, marked by higher concentrations and fluxes of organic matter and increased accumulation of sedimentary pigments and diatoms (including an increased presence of planktonic taxa). There is also a striking disappearance of the UV-absorbing pigments at ~4,000 BC, most likely due to internal loading of DOC from phytoplankton blooming (Mannino and Harvey, 2000) and shading by phytoplankton reducing the UV-penetration within the fjord. This productivity increase is likely to be the result of complex interactions between salinity and/or nutrients, most likely caused by changes in the marine environment, rather than changing inputs from terrestrial sources. As the beginning of this productivity increase pre-dates the introduction of agriculture (~3,900 BC; Andersen and Rasmussen, 1993) it is unlikely to be human induced. There is, however, a secondary increase in sedimentary pigment concentrations around 3,700 BC, which might suggest a later role for terrestrial input of nutrients following human activities (e.g. induced by forest clearances) as evident in numerous terrestrial pollen records (Iversen, 1941; Aaby, 1986; Andersen, 1995b; Rasmussen, 2005).

The beginning of this high productivity phase (i.e. ~4,300 BC) broadly coincides with the strengthening of the south Jutland current between ~4,250-2,850 BC, which ultimately causes key hydrographic changes in the North Sea and Skagerrak-Kattegat region (Conradsen and Heier-Nielsen, 1995; Leth, 1996; Jiang *et al.*, 1997; Jiang *et al.*, 1998; Gyllencreutz *et al.*, 2005; Gyllencreutz *et al.*, 2006). The beginning of this hydrographic shift appears to be synchronous with the final drowning of the Jutland Bank at ~4,250 BC (Leth, 1996; Gyllencreutz and Kissel, 2006), which enabled increased flow of North Sea water into the Skagerrak. Increased grain size of sediments in Skagen and the Skagerrak, higher energy conditions (i.e. stronger currents) and the greater transport of robust North Sea diatoms (e.g. *Delphineis surirella* in the Skagen-3 core; Jiang *et al.*, 1997; Jiang *et al.*, 1998) all support strengthening of the Jutland current (most notably the southern component) around this time (Conradsen and Heier-Nielsen, 1995; Jiang *et al.*, 1997; Jiang *et al.*, 1998; Gyllencreutz, 2005; Gyllencreutz *et al.*, 2006). The shift has been linked to a climatic cooling event at ~4,200 BC (Conradsen and Heier-Nielsen, 1995; Jiang *et al.*, 1997), but a more recent revision by Gyllencreutz, (2005), suggests a continuous hydrographic transition lasting over ~2,000 years.

The Jutland current is an erosive current, flowing over sandy sediments off western Denmark, and is responsible for eroding, transporting and re-depositing large amounts of sediment (Eisma and Kalf, 1987; Longva and Thorsnes, 1997). An increase in the strength of this current is likely to result in greater erosion along the western coast and increased sediment deposition along the northern coast of Jutland (e.g. Skagen; Jiang *et al.*, 1997; Jiang *et al.*, 1998). Sediment accretion might have even intermittently reduced the northern opening of the Limfjord, temporarily reducing the marine connection and subsequently lowering salinity (e.g. 4,100-3,900 BC; Figure 3.18). Care must be taken with the interpretation of salinity change in this phase of the record due to difficulties with isolating a clear direct salinity signal from other parameters (e.g. productivity, nutrients) which are also exhibiting intense changes (e.g. Anderson, 2000).

The direct influence that greater exposure to, or shelter from, the North Sea and Skagerrak has on the trophic status of the Limfjord is less clear than for salinity. It is possible that nutrients are retained longer when the Limfjord's connection with the North Sea and Skagerrak is reduced, causing higher productivity and enabling nutrient-demanding planktonic organisms (e.g. diatoms; *Chaetoceros* cysts, *C. choctawatcheeana, C. striata, S. costatum*) and bacterioplankton to bloom. This might in turn induce greater production of DOC and reduce UV-light penetration. In systems such as the Baltic Sea, phytoplanktonic blooming can lead to stagnation of bottom waters due to hypoxia (Sohlenius and Westman, 1998; Sohlenius *et al.*, 2001; Zillén *et al.*, 2008; Weckström *et al.*, in review). A prolonged period of high productivity in the Limfjord might be sustainable due to a regular input of oxygenated water through the northern and possibly western openings. Alternatively, regular renewal events might be required, possibly associated with short term sea level fluctuations and phases of increased erosion (e.g. climatic cooling and storm events). Renewal events might be separated by sediment accretion phases, reducing the Limfjord-marine connection, and again enabling phytoplanktonic blooming and high turnover of plants and algae.

This could explain both prolonged high productivity and the fluctuating DI-salinities between 4,100-3,150 BC (Figure 3.18). For example, lower DI-salinity exists between ~4,100-3,900 BC BC (~20-22 ‰; Figure 3.18), followed by numerous fluctuations between ~3,900-3,150 BC, with salinity reaching as high as 27-28 ‰ in some samples. This also suggests significant inputs of saline water through the west. The low salinity between 4,100-3,900 BC is particularly interesting, occurring just prior to the Mesolithic-Neolithic transition and potentially limiting marine resources important to Mesolithic/early Neolithic cultures and societies (e.g. oyster decline; Rowley-Conwy, 1984; Andersen, 1989, discussed in more detail in section 3.11). There also appears to be exceptional preservation in this section of the record, a likely response to reduced marine influence, lower salinity and high sediment accumulation rates. This is demonstrated by consistently high **F** index scores and relatively high abundances of the weakly silicified diatom species *Skeletonema costatum* (Figure 3.7).

Deeper water might provide an alternative explanation for a reduction in UV- penetration (and increased planktonic diatoms), but is deemed unlikely with respect to the current evidence. The most plausible explanation for increased water depth would be higher sea levels in the Skagerrak and North Sea after ~4,300 BC (Behre, 2007; Pedersen *et al.*, 2009) and therefore greater input of saline water into the Limfjord. Under such circumstances, a prolonged and probably substantial increase in salinity would be expected, this is not apparent in any of the salinity records presented here (i.e. diatom, isotopes, foraminifera, or molluscan records). According to the southern North Sea level curve, sea level is rising at this point, but the low DI-salinities suggests that that the western Limfjord had not been breached at this time (possibly until ~3,700 BC or later). In the other regional sea level records (e.g. Mörner, 1976; Christensen, 1995; Berglund *et al.*, 2005), sea level decreases shortly after ~4,300 BC associated with the late Atlantic regression (or late Mesolithic regression; L3, 4,200-3,800 BC in Berglund *et al.*, 2005). There is also no evidence in the Skagen and Skagerrak record for increased water depth at this time (Conradsen and Heier-Nielsen, 1995).

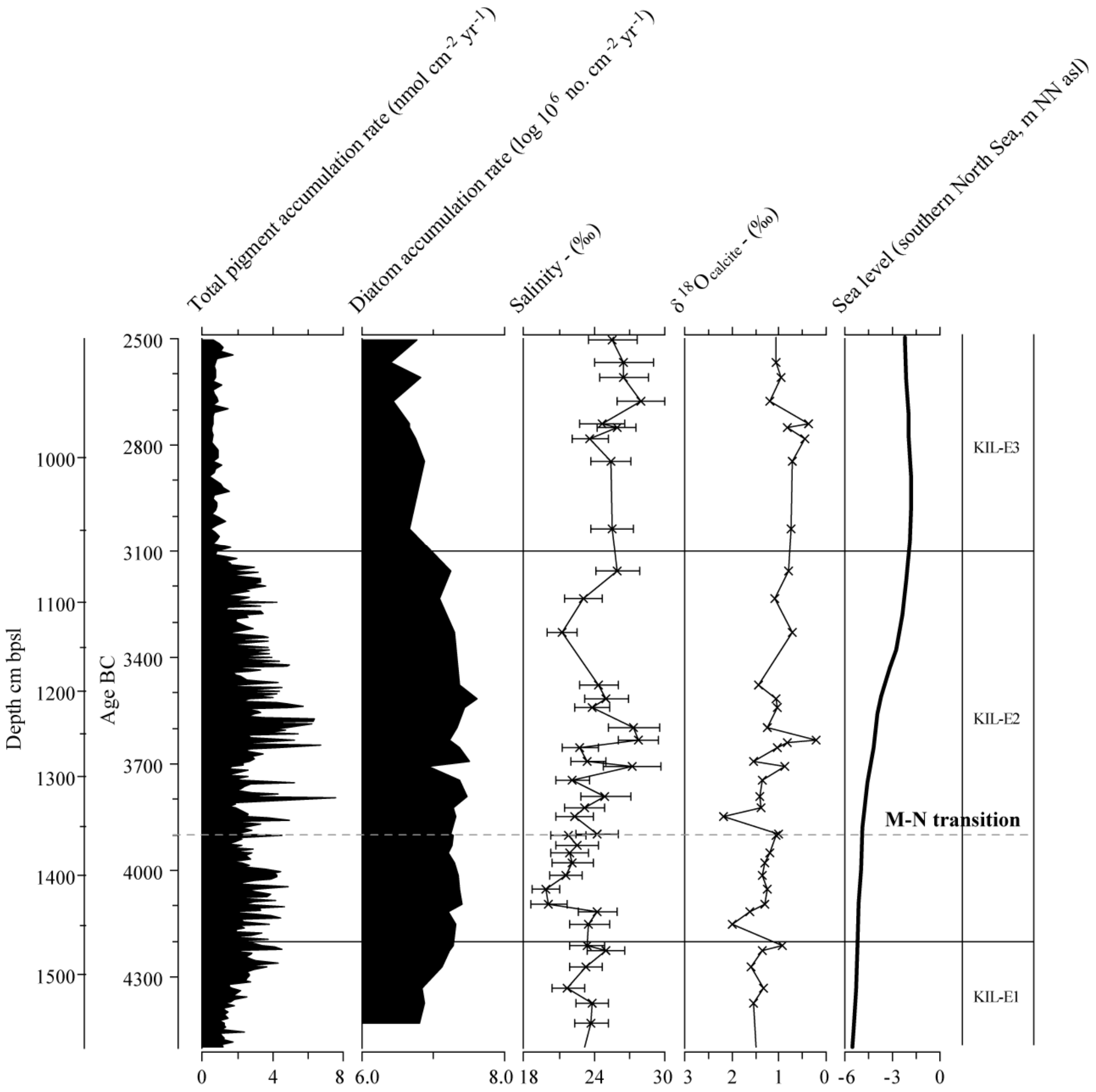


Figure 3.18 - Salinity and productivity changes in the Kilen profile between ~4,500-2,500 BC (covering the Meoslithic-Neoltihic transition). Sea level from Behre, (2007) for southern North Sea. *m NN asl = metres realtive to German ordnance datum above sea level.

In terms of terrestrial change, the increased regularity of *Betula* fruits in the plant macrofossil record after ~3,850 BC probably corresponds with the early Neolithic *Betula* expansion as recorded in the Danish pollen diagrams (e.g. Iversen, 1941; Aaby, 1986; Rasmussen, 2005). This is probably a consequence of human-induced land-use changes associated with the introduction of agriculture (Iversen, 1941). The palynological analyses spanning ~4,180-3,640 BC suggest a mixed deciduous forest dominated by *Corylus, Quercus* and *Alnus,* with *Ulmus* probably being quite abundant initially. The low percentages of *Tilia, Fraxinus* and *Pinus* pollen suggest that these tree taxa were either very sparse or were not growing in the immediate vicinity of Kilen. Agricultural indicators are also sparse in the short record. *Plantago lanceolata* appears in the record at ~3,900 BC, but a continuous curve does not begin until at least ~3,640 BC. *Poaceae* increases slightly after ~3,900 BC, but no cultivated cereal crops are present in the record (Figure 3.3).

The absolute disappearance of the UV-absorbing pigments broadly ties in with the *Ulmus* decline (Figure 3.3) and the introduction of agriculture, which might also suggest a more important role for externally sourced DOC. A reduction in forest density and a more open catchment might have increased inputs of DOC into the Kilen basin from terrestrial sources. An increase in minerogenic matter might be expected with greater input of terrestrial material, but there is no change in the minerogenic flux rate at this time.

A secondary increase in productivity is identifiable in the sedimentary pigment record at ~3,700 BC, which coincides with a substantial increase in the accumulation rate (~0.7 cm yr^{-1} increase from 0.36-0.43 cm yr^{-1}). This is caused by a small increase in both the organic and minerogenic flux rate. In the palynological record, elevated *Poaceae* pollen after ~3,900 BC, and an appearance by *Plantago lanceolata* at ~3,640 BC, probably marks the first signs of major agriculture-based land-use changes in the Kilen area. This interpretation must, however, be treated with caution due to the limited pollen analyses, but would be in good agreement with other Danish palynological studies (Aaby, 1986; Rasmussen, 2005). Nevertheless, it remains plausible that increased input of nutrients (and sediment) from the surrounding catchment, following agricultural land-use change and forestry removal, caused, or at least contributed to the rise in marine productivity after ~3,700 BC.

KIL-E3 (~3,100-0 BC) - *Maximum marine phase* - DI-salinity begins to rise after ~3,330 BC and reaches its modern day levels for the Struer area (27 ‰) by ~2,680 BC and almost fully marine conditions (>30 ‰) by ~1,430 BC. The rise broadly coincides with increased sea level in the southern North Sea (Behre, 2007; Figure 3.16) and suggests that a widening (or opening) of the western Limfjord occurred somewhere between ~3,330-2,680 BC. The relatively good agreement between DI-salinity change and sea level rise in the southern North Sea (Behre, 2007) and western Jutland (Gehrels *et al.*, 2006; Pedersen *et al.*, 2009) continues throughout this phase, suggesting that the connection with the North Sea is driving salinity between ~3,100-0 BC. The southern North Sea record is included in Figure 3.16 as it is one of the most comprehensive sea level records for the region, being based on 118 sea level tie points from the German North Sea coast (Behre, 2007), and forms an extension of the Calais-

Dunkirk system, which is applicable throughout the entire southern North Sea (Dubois, 1924; Bateman, 1981; Ervynck *et al.*, 1999).

A substantial drop in the accumulation rate occurs at ~3,100 BC, caused by marked decreases in the flux of organic and minerogenic matter. The intensified marine conditions that ensued (between ~3,150-0 BC) probably lowered the accumulation rate (~0.15-0.18 cm $yr^{-1)}$, due to increased energy (from stronger currents and increased exposure) in the system, greater transportation, re-working and re-deposition of sediments and poorer preservation (e.g. lower plant macrofossil and mollusc concentration, lower diatom **F** index). Greater distance to the coring location might also partly explain the decreased concentration of terrestrially derived material (e.g. plant macrofossils), likely due to deeper water associated with increasing sea levels. The reduction in organic matter and accumulation rates of sedimentary pigments suggest a decline in marine primary productivity also occurred around this time. In addition, there is little change in the pheophytin-*a*/chlorophyll *a* ratio, which suggests poorer preservation is not responsible for this drop in pigment concentrations. After ~3,100 BC, the sedimentary pigment concentration and accumulation rates generally remain relatively low for the remainder of this phase, but subject to minor fluctuations, perhaps related to variable inputs of high salinity oxygenated water from the North Sea, or from changing trophic status, perhaps due to nutrients being washed in from land following land-use changes (e.g. forest clearance for agricultural purposes and settlements).

The foraminiferal samples in this phase are dominated by *E. excavatum* and *A. beccarii* after ~1,590 BC, with relatively low numbers of the higher salinity species (e.g. *E. incertum*, *E. magellanicum*) present in the lower part of the sequence (Murray, 1991; Alve and Murray, 1999; Knudsen and Konradi, submitted). However, the foraminiferal record might be heavily affected by dissolution, with lower concentrations of foraminifera between ~3,000-200 BC, generally smaller tests and higher quantities of coarser grained material (probably sand and quartz) present within the wet sieved samples. The $\delta^{18}O_{calcite}$ isotope signal suggests that salinity increases at ~2,680 BC, after which it remains relatively stationary up until ~800 BC, prior to a substantial drop in the uppermost isotope sample. An increase at ~2,680 BC is consistent with DI salinity, but after this it becomes contradictory, by suggesting that salinity was lower between ~2,680-400 BC period than it was in the pre-3,200 BC period. This may be the result of alteration of the isotopic signal due to increased dissolution of the foraminiferal tests under intensified marine conditions. This is in contrast to observations at both Horsens Fjord and Norsminde Fjord, where foraminifera appear to preserve better under higher salinity conditions (Kisum, 2007; Knudsen and Konradi, submitted; D.B. Ryves *et al.* unpublished data). It is, however, plausible that preservation only deteriorates when near fully marine conditions are reached, with foraminiferal preservation at Kilen being very good between ~5,500-3,100 BC, when DI-salinity ranges between ~18-25 ‰. Factors such as increased current energy, greater transport of material and coarser sediments associated with greater exposure to the North Sea might all be affecting foraminiferal preservation.

After increasing up until ~1,500 BC, a sea level regression phase begins which culminates in a sustained period (~1,500-500 BC) of lower sea levels in the southern North Sea (Behre, 2007). DI-salinity for the Kilen record decreases slightly later than this, but salinities of ~24-28 ‰ are inferred between ~930-600 BC, which actually coincides with the lowest reconstructed sea levels for the last ~4,000 years (i.e. labelled regression phases 2-3 and Dunkirk 1a in the Calais-Dunkirk system; Behre, 2007).

This regressive phase is followed by a sharp rise in sea level rise between ~400-150 BC, which caused the abandonment of settlements along the clay districts of northern Germany (Behre, 2007). A high sea level stand is also identifiable in the Skagen region at ~500 BC (Clemmensen *et al.*, 2001). In the late pre-Roman period, sea level in the southern North Sea reached its highest pre-Roman Iron Age levels and probably marks the maximum connection of the Limfjord with the North Sea and Skagerrak. The Kilen diatom record is dominated by high salinity demanding species such as *P. sulcata*, *C. belgica* and *D. minutissima* between ~390-0 BC, which drives exceptionally high reconstructed salinities, even exceeding present day average salinities for fully marine conditions (i.e. > 35 ‰) in several samples (35-43 ‰). Salinities up to ~38 ‰ are often found in open oceans (e.g. Pacific Ocean, Sargasso Sea in the North Atlantic) and can exceed in this range in highly evaporative basins (e.g. Persian Sea, Red Sea). Even during this high sea level stand, the salinity of the Limfjord is unlikely to exceed 35 ‰ and therefore salinity has been over-estimated, probably resulting from a mathematical artefact incorporated into the deshrinking tool of the transfer function (e.g. Yuan, 2005). Almost fully marine salinities in the late pre-Roman times are also consistent with foraminiferal, dinoflagellate and mollusc records form the Bjørnsholm Bay region (Kristensen *et al.*, 1995; Ellegaard, 2000; Christensen *et al.*, 2004) and high salinity inferred from diatoms and molluscs at Horsens Fjord and Tempelkrog (D.B.Ryves *et al.* unpublished data).

KIL-E4 (~AD 0-1856) - *Limfjord opening and closing phase* - This is a dynamic phase subject to large scale shifts between marine and brackish water conditions, driven by the degree of connection to the North Sea and/or the Skagerrak. A marked shift to brackish conditions occurs at the onset of this phase, with DI-salinity falling by ~10 ‰ between ~AD 0-50. The high salinity diatoms such as *C. belgica* and *D. minutissima* practically disappear from the record, and *P. sulcata* abundance drops substantially, as brackish-marine species such as *O. mutabilis, Chaetoceros* resting cysts and *Cocconeis scutellum* become more abundant in the record. Between ~AD 0-630, more brackish water taxa appear in the mollusc record (e.g. *Mytilus edulis*, *Cerastoderma glaucum, H. ventrosa*, *H. ulvae, Cerastoderma* spp.) and foraminifera become more sporadic. *Zannichellia palustris* also becomes commonly present in the plant macrofossil record, which is likely to have provided an important habitat for epiphytic diatoms (see section 3.10 and Table 3.5), which also increase in this phase.

There is also some evidence for increased marine productivity with increased concentrations of certain sedimentary pigments (most notably alloxanthin, lutein and canthaxanthin; after ~AD 40), increased mollusc and ostracod concentrations and an increase in the organic and carbonate content of the sediment. The majority of other pigments exhibit increases in abundance slightly later in this phase, with the exception of diatoxanthin (from diatoms, dinoflagellates and chrysophytes), which decreases at the onset of this zone and thereafter remains relatively low throughout. A clear difference between the high relative abundances and the low flux rate of organic matter and accumulation of sedimentary pigments suggests, that overall productivity is actually being overestimated in the early Roman period. Exaggerated concentrations probably result from the drop in accumulation rate driven by substantial decrease in the minerogenic flux and to a lesser extent the organic flux. Consistent with the decrease in diatoxanthin is a drop in the diatom accumulation rate in the early Roman period (Figure 3.8).

The cause of this salinity decline is likely to be a closing, or at least severe narrowing, of both the northern (Kristensen *et al.*, 1995; Christensen *et al.*, 2004) and western openings of the Limfjord, probably due to accretion of sediments and a lowering of sea level (e.g. Behre, 2007). The salinity decrease at Kilen (~AD 0-50) begins later than the drop in sea level in the southern North Sea (~150 BC-AD50), suggesting that temporal sedimentary accretion due lower sea level conditions (causing reduced currents and lower energy) played an important role in the eventual closure of the Limfjord. Alternatively, this may be due to dating discrepancies between profiles, slight time lags between regions or variation in other factors that might also drive salinity change at Kilen (and the Limfjord). The reduction in the accumulation rate and flux of minerogenic material probably results a reduced input of coarse grained material, eroded form the west of Denmark and being redeposited in the lower energy waters of the Limfjord when the Limfjord entrances were open (or wider).

Shortly after ~AD 600, salinity appears to increase again causing the lower salinity-indicating molluscs and *Z. palustris* to disappear from the macrofossil record, and subsequently the reappearance of *B. reticulatum* in several samples between ~AD 590-830. DI-salinity at ~AD 800 suggests a return to almost fully marine conditions, perhaps due to a re-opening of the northern entrance (e.g. Kristensen *et al.*, 1995) together with an opening to the west (see section 6.4.2.2). This coincides with a period of higher sea levels in the Limfjord during the late Iron Age (Christiansen *et al.*, 1990; Tanner, 1993). The diatom sampling resolution in this section of the record and the absence of molluscs after ~AD 830, prohibits detailed examination of salinity change throughout the Germanic and Viking periods (AD 400-1050). Sedimentary pigments remain relatively abundant in sediments up until ~AD 780, after which most exhibit a marked drop in concentration, with the exception of the pheophytins. An increase in the ratio of the pheophtyin-*a*: chlorophyll *a* after ~AD 750 suggests greater degradation of pigments. A further drop in the concentration of diatoxanthin is consistent with the generally low concentration of diatoms found between depth ~427-450 cm bpsl (following preliminary scanning of diatom samples; results summarised in Figure 3.19), but dating of sediments become particularly problematic in this section of the profile.

It is unclear how long after ~AD 800 (uppermost radiocarbon date) the slow accumulation phase or sedimentary hiatus began (see section 3.3). Kristensen *et al.*, (1995) suggests that the Limfjord remained connected to the North Sea up until sometime between ~AD 1,100-1,200. Based on the much-debated Old Norse Kings' sagas in the Heimskringla of Snorri Sturluson, it has been suggested that the Norwegian King Harald Haardraade might have fled from the Danish fleet under King Sven Estridssøn in AD 1061 through the northern entrance of the Limfjord (Petersen, 1976). A notable increase in the regularity and abundance of *Chara* oospores (freshwater-low salinity tolerant charophytes) occurs above 454 cm (estimated date ~AD 1,140), which might mark the closure of the Limfjord. Increased abundance of *Chara* oospores was also evident in the Bjørnsholm Bay core between ~AD 1,200-1825 (Kristensen *et al.*, 1995). Additionally, marine diatoms remain present in the Kilen profile up until at least depth 450 cm bpsl (currently dated at ~AD 1030; Figure 3.19). This might suggest relatively good dating throughout the Viking period and up until the end of the marine phase (~AD 1,100-1,200; Kristensen *et al.*, 1995).

In contrast, a diatom sample at depth 437 cm bpsl (currently dated at ~AD 1430), can almost certainly be placed within the low accumulation/low productivity or sedimentary hiatus phase. This sample is extremely low in diatom concentration ($6.2*10^6$ g), mainly containing low salinity diatoms and substantial numbers of *Ellerbeckia arenaria*. This diatom grows on sandy sediments in low salinity alkaline or oligotrophic waters in North America (Spaulding and Edlund, 2009) and the Baltic Sea (Snoeijs *et al.*, 1993-1998). The combined presence of marine (e.g. *Paralia sulcata*, *Grammatophora oceanica*), oligotrophic (e.g. *Ellerbeckia arenaria*) and nutrient-favouring diatoms (e.g. *A. pediculus* and *C. choctawatcheeana*), suggests re-working and re-deposition of sediments. These conditions might be expected under slow accumulation and low productivity or during a sedimentary hiatus. The **F** index (0.35) is surprisingly high for either circumstance, which suggests relatively good valve preservation. There does, however, appear to be a high percentage of heavily silicified diatoms whose valves might be able to survive the re-working/re-deposition process.

KIL-E5 (~AD 1856- present) - *The limnic phase* - The building of the road and rail embankment in AD 1856 almost entirely cut the Kilen basin off from the Limfjord and began the limnic phase of Kilen history. Isolating the exact marker for this transition in the sedimentary record, however, has proved difficult due to lack of dating points between depths 420-460 cm bpsl (i.e. radiocarbon and ^{210}Pb/^{137}Cs stratigraphies), a slow accumulation rate (~1,000 years in 40 cm), very low concentration of diatoms, likely mixing/disturbance of sediments (or potential hiatus) and the complexity of the recent (last ~1000 years) history of the Limfjord (Kristensen *et al.*, 1995; Ellegaard, 2000; Holm and Bager, 2002; Christensen *et al.*, 2004; Poulsen *et al.*, 2007).

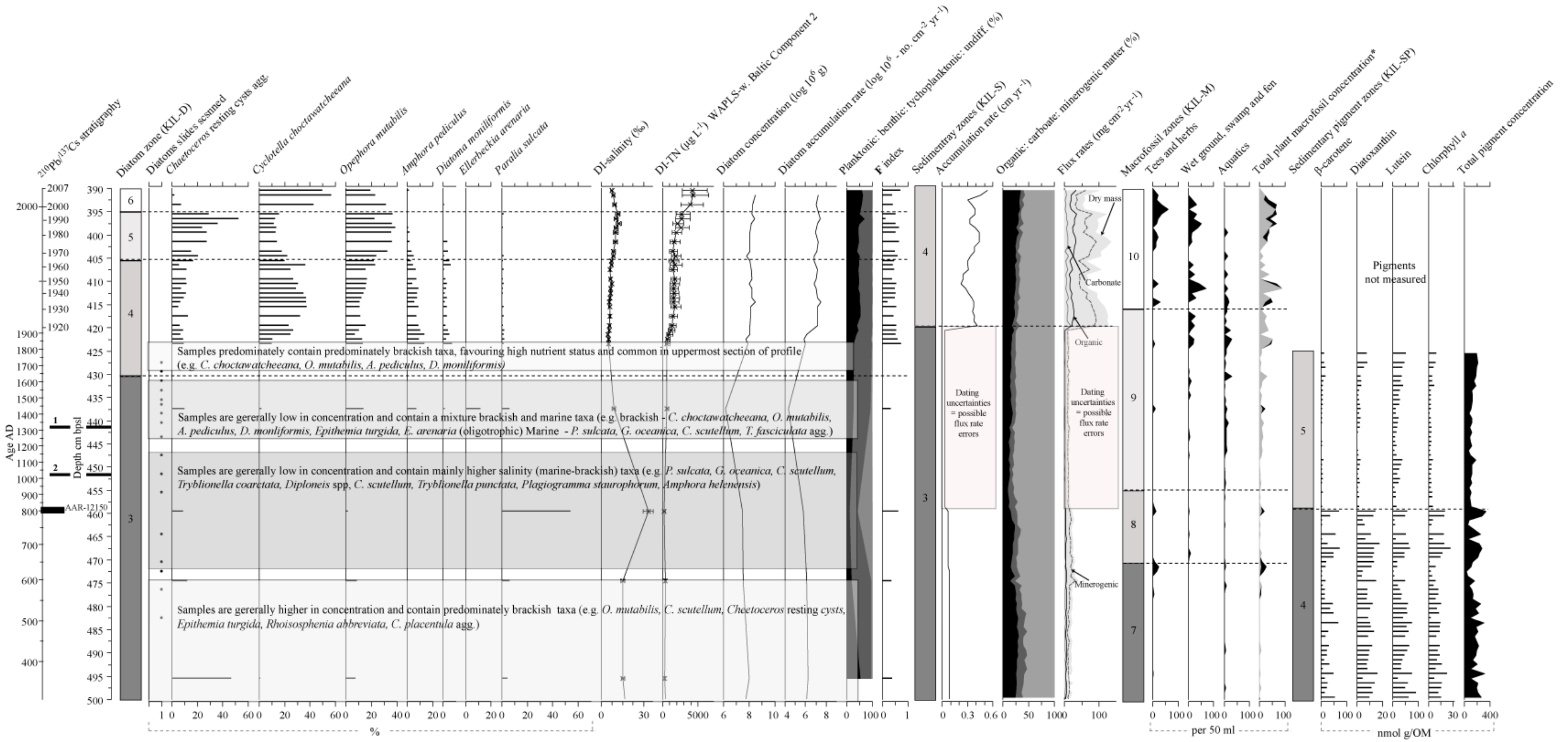


Figure 3.19 – Summary diagram of the upper section (~AD 300-present; top 110 cm) of the Kilen profile and associated dating problems. *Grey shading = total plant macrofossil concentration excluding ambiguous remains (i.e. deciduous leaf fragments, buds, wood, bark and twigs), black shading = total plant macrofossil concentration. 1 = approximate depth of AD 1856 using a straightforward extrapolation of the 210Pb/137Cs stratigraphy. 2 = approximate depth of AD 1825 using a straightforward extrapolation of the ^{210}Pb/^{137}Cs stratigraphy. AAR-12150 = uppermost radiocarbon date (depth 459-460 cm bpsl, AD 672–894 (95.4 %); Table 3.2, Figure 3.2).

The lowermost diatom sample in Figure 3.9 (at depth 417 cm bpsl in sequence; Figure 3.19) is dated to ~AD 1830, but the flora is almost identical to that in the samples immediately above it (i.e ~AD 1862-2007; depths 390-423 cm bpsl), characterised by diatoms favouring low salinity and relatively high nutrient status (e.g. *C. choctawatcheeana*, *O. mutabilis*, *A. pediculus*). It is therefore likely that this sample is post AD 1856 and that the dating in this section is erroneous. Prior to AD 1856-1857, Kilen was connected to the Limford and under marine conditions, due to an opening to the North Sea in the west. This western entrance was re-opened by a storm in AD 1825, and molluscan and foraminifera analyses at Bjørnsholm Bay suggest salinity was between 22-30 ‰ in the succeeding period (Kristensen *et al.*, 1995; Christensen *et al.*, 2004). Due to its close proximity to the western entrance, salinity at Kilen is likely to have been higher than Bjørnsholm Bay at this time, probably at or close to its present day value for the Struer area (~27 ‰).

A series of preliminary diatom samples (18 in total) between depths 427-482 cm bpsl (Figure 3.19) were scanned in an attempt to identify markers for ~AD 1825 and AD 1856-1857 via expected shifts in the diatom flora. Unfortunately, no obvious stratigraphic markers can be identified for these two transitions (via scanning alone), due to extremely low diatom concentrations and co-existence of both marine and brackish taxa (i.e. probable re-working of sediments). Therefore, at present the AD 1856 transition can only be broadly estimated at best, but no estimate is possible for ~AD 1825. Here it is suggested that ~AD 1856 falls somewhere between depths 442 cm bpsl (linear extrapolation of the ^{210}Pb/^{137}Cs stratigraphy, assuming constant accumulation which is, however, unlikely) and 429 cm bpsl, and most likely between 436-429 cm bpsl.

The latter depths are based on the likely, pre-AD 1856 position of the diatom sample at depth 437 cm bpsl (see above) and an increase in sedimentary pigment concentrations (and decrease in pheophytin-*a*: chlorophyll *a* degradation ratio) above 433 cm bpsl. An increase in productivity and/or improved pigment preservation might be expected, due to increased trapping of nutrients (and other material) within the Kilen basin and a faster accumulation rate. A diatom sample at depth 429 cm bpsl appears to be post-AD 1856, containing a similar diatom flora similar to the uppermost section of the profile (i.e. between 417-390 cm bpsl). This sample, however, again appears to be subject to some degree of sediment mixing, though numbers of marine and oligotrophic diatoms are much reduced. This might place it in close proximity to (i.e. soon after) the ~AD 1856 transition.

In the succeeding phase (i.e. above 429 cm bpsl and likely post-AD 1856), the diatom record (and associated metrics) clearly indicates increasing nutrient status and productivity at Kilen. An increase in plant macrofossil concentration (most notably wet ground, swamp and fen taxa) also occurs above 423 cm bpsl, which is again likely to represent either increased trapping of material within the basin, or improved preservation of material, perhaps due an increase in the accumulation rate. By ~AD 1930 (412 cm bpsl), planktonic diatoms account for over half the diatom assemblage, with *C. choctawatcheeana* being the most abundant taxon (accounting for ~23-38 % of the diatom assemblage between ~AD 1890-1960). This diatom is a strong indicator of enriched nutrient conditions, often found in modern coastal systems suffering from cultural eutrophication (Prasad *et al.*, 1990; Håkansson, 1996; Clarke *et al.*, 2003; Ryves *et al.*, 2004). This drives the relatively high DI-TN values (>1100 $\mu g\ L^{-1}$), indicating eutrophic conditions within the basin. These conditions also appear to affect other organisms, such as *Daphnia* spp., which increase markedly after ~AD 1930. In contrast, there is a reduction in aquatic taxa (most notably *Chara* spp. and *Ranunculus sect. batrachium*) after ~AD 1920, and by AD 1945 they become quite sporadic. Eutrophic conditions are often detrimental to macrophyte communities, with increased nutrient/sediment loading and algal blooms often causing increasing water turbidity and reduced light penetration (Duarte, 1995; Brush and Hilgartner, 2000; Middleboe and Sand-Jensen, 2000; Nielsen *et al.*, 2002).

In the 1960-90s there appears to be a slight a reduction in *Cyclotella choctawatcheeana* and an increase in the abundance of *Chaetoceros* resting cysts and *O. mutabilis*. Despite these taxa generally demanding higher-salinities than *C. choctawatcheeana*, measured salinity data (available for the period AD 1972-1987) suggest that higher salinity cannot explain this shift (Figure 3.20). These data show that salinity actually dropped between AD 1978-1984 due to decreased exchange of water between Kilen and Venø Bugt (Ringkjøbing Amtskommune, 1988). A preliminary comparison (low resolution, only four overlapping data points) of DI-salinity estimates (Figure 3.20) with salinity measurements (Ringkjøbing Amtskommune, 1988) suggest that the DI-salinity model performs poorly (i.e. incorrect pattern and overestimation of salinity; DI-salinity suggests a gradual increased up until ~AD 1988) in environments heavily impacted by humans and changing nutrient status. In Figure 3.20, the DI-TN model estimates are also compared against measured TN concentrations for the period 1976-1987. The DI-TN model tends to slightly overestimate the TN concentrations, but overall, this preliminary assessment suggests that the TN-model performs relatively well.

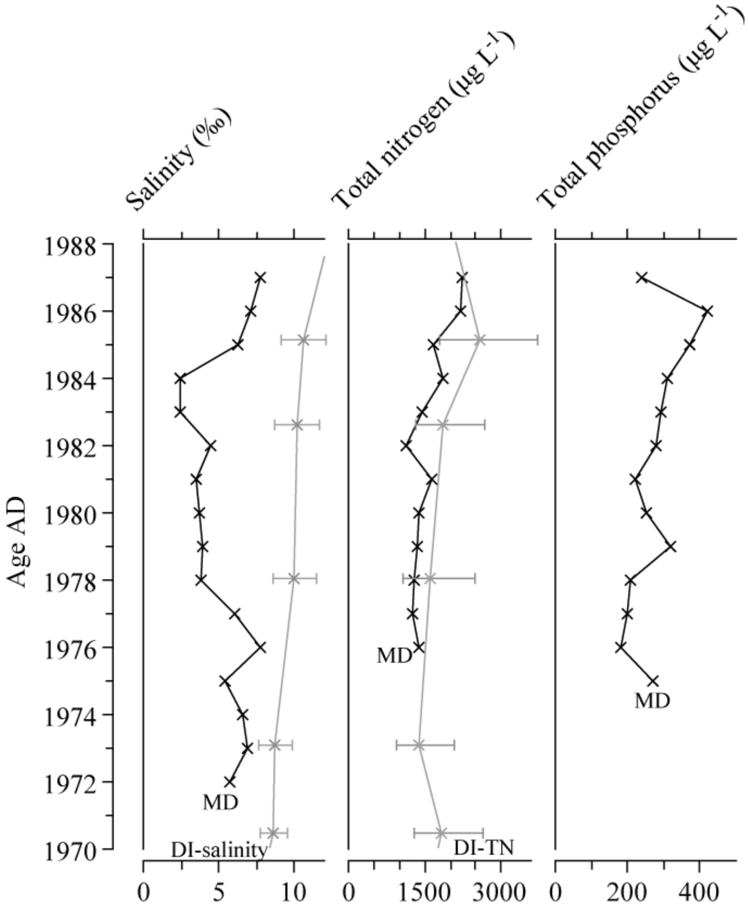


Figure 3.20 - Comparison of DI-salinity and DI-total nitrogen estimates with monitoring data for salinity, TN and TP, measured over the period AD 1972-1987.

Phosphorus is also likely to be important at Kilen over the recent past, with evidence for both increased TP concentrations in the 1960s following the building of four fish farms and rising TP levels over the monitored period AD 1976-1987 (Figure 3.20). The reduction of TP in the late 1970s (Figure 3.20) is believed to have been related to poor water exchange with the Limfjord and reduced TP loading from these fish farms. This resulted in an improved passage (or sluice) being installed in 1984, which successfully improved water exchange with the Limfjord and subsequently led to increased salinities and increased sediment release of phosphorus due to occasional salt water pockets forming above the sediments (Ringkjøbing Amtskommune, 1988). Low oxygen conditions can often prevail in the bottom waters after phosphorus release from the sediments. The importance of phosphorus in Danish limnic systems has previously been demonstrated in a number of studies (e.g. Anderson and Odgaard, 1994; Bradshaw *et al.*, 2002; Bradshaw *et al.*, 2005a; Bradshaw *et al.*, 2006) and certainly future potential exists for testing how well TP training sets (limnic and/or marine) perform in this environment (though higher resolution diatom data is required, over the period for which monitoring is available) and the potential long-term effects of TP at Kilen.

According the DI-TN reconstruction, the rate of TN loading drops between ~AD 1950-1980, but the nutrient status remains relatively high with *C. choctawatcheeana* still accounting for between ~10-20 % of the diatom assemblage (between ~AD 1970-1990), *Chaetoceros* cysts increasing in abundance and *C. meneghiniana* percentages remaining largely unchanged. As indicated above, some species of *Chaetoceros* can live in low salinity waters (Rines and Hargraves, 1988) and are often also affiliated with high nutrient status (Andrén *et al.*, 2000; Leśniewska and Witak, 2008). In marine systems, cyst production often follows high productivity events (marked by high numbers of vegetative cells which rarely preserve in sediments), due to subsequent nutrient depletion (Leveneter *et al.*, 1996). Despite the drop in *C. choctwatcheeana* percentages, this is unlikely to be the case here, due to likely regular input of nutrients, continued presence of these nutrient favouring diatoms (e.g. *C. choctawatcheeana*, *C. meneghiniana*), continued increase of DI-TN and high and increasing TP (e.g. annual mean phosphorus concentrations between AD 1989-1996 = 0.239 mg L^{-1}; Jensen *et al.*, 2006, Figure 3.20). Difficulties associated with taxonomy of resting cysts under light microscopy precludes identification down to species level for *Chaetoceros* cysts in this study, and therefore it is difficult to interpret, what environmental variables (e.g. salinity change, nutrients) they are reacting to (and when) throughout the Kilen profile.

Nevertheless, increased numbers of planktonic diatoms since the mid 1980s appear to have caused a reduction in both the abundance and diversity of the benthic community. With the exception of *O. mutabilis*, most other benthic species practically disappear from the record by ~AD 1985. The TN begins to increase rapidly after ~AD 1980 and Kilen becomes extremely eutrophic (measured TN > 2000 μg L^{-1}, DI-TN ~1800-4300 μg L^{-1}) with high phosphorus concentrations (Jensen *et al.*, 2006). This probably caused blooming of planktonic species, and subsequent decreasing water clarity and light penetration, often considered classic symptoms of eutrophication (Cederwell and Elmgren, 1990; Rosenberg *et al.*, 1990; Cooper and Brush, 1991; Nixon, 1995; Cornwell *et al.*, 1996; Clarke *et al.*, 2003). Planktonic blooms are further supported by high diatom concentrations (Figure 3.8), and both high measured summer chlorophyll *a* content in water column and shallow Secchi depths between AD 1989-1996 (Jensen *et al.*, 2006).

The high numbers of *O. mutabilis*, however, suggest that benthic conditions remained suitable for diatoms, possibly more confined to the shallower parts of the basin, and therefore light penetration alone cannot explain this reduction. There is a noticeable reduction in benthic (commonly epiphytic) diatoms (most notably *Ctenophora pulchella, Cocconeis placentula,* *Epithemia sorex*, *E. turgida, Diatoma moniliformis, Rhoicosphena abbreviata*) at this time, which might have been due to the disappearance of macrophytes. No remains of aquatic macrophytes are recorded in the plant macrofossil record after ~AD 1970. It is possible that the remaining benthic taxa (e.g. *A. pediculus*) were outcompeted by *O. mutabilis* under higher nutrient and possibly salinity conditions. Sandier sediments might also have favoured *O. mutabilis* over other benthic taxa (e.g. *A. pediculus*), though this might be considered unlikely due to the high accumulation rates, likely depositing large volumes of fine organic content (Figure 3.19).

After ~AD 1999, *C. choctawatcheeana* becomes the most abundant species present in the profile again as eutrophication appears to be intensifying. There is an increase in *Betula* fruits between ~AD 1990-2005, which might reflect improved preservation conditions within the basin due to increased bottom water anoxia (e.g. reported after periods of phosphorus release from the sediments after inputs of salt water from Venø Bugt through the Kilerkanal). *Daphnia* spp. also disappear after ~AD 1996 which is possibly related to low oxygen conditions within the lake and the reduction in macrophyte abundances offering shelter from predation (e.g. Lauridsen *et al.*, 1998; Burks *et al.*, 2001).

3.12 Summary and conclusions

- The level of connection with the North Sea and Skagerrak (predominately governed by sea level, erosion and sedimentary accretion) is extremely important in driving changes in the marine environment (i.e. salinity, productivity, nutrient status and sedimentary regime) at Kilen.
- DCA results for the entire diatom dataset demonstrates a strong salinity gradient on axis 1, due to the presence of weakly brackish taxa at the top of the profile (in the limnic section), through to fully marine diatom taxa present in the pre-Roman period (~500-0 BC). For the limnic section only (i.e. since ~AD 1856), PCA axis 1 is predominately driven by nutrients, but with depth, sediment type and macrophyte cover probably also being important. For the marine section of the profile only (i.e. pre AD 1856), a salinity gradient is again apparent on DCA axis 1, with the diatom flora spanning from fully marine in parts of the core sequence (most notably pre-Roman period) through to brackish water taxa in the Roman period. For the marine dataset, DCA axis 2 appears to be driven by substrate type and habitat preference, with several of the epiphytic species being abundant during the Roman period (when there is

evidence for increased macrophyte abundances, particularly *Zannichellia palustris*) scoring negatively. Pelagic and other benthic diatoms (i.e. living on the sediments or attached to sand or stones) score more positively on this axis.

- The DI-salinity record suggests that between ~5,500-2,750 BC, salinity was lower (generally fluctuating between 18-25 ‰, but with occasional higher peaks) than at present for the Struer region (~27 ‰). It is therefore likely that the western Limfjord was either completely, or at least partially closed off at this time, probably due to lower sea levels in the North Sea, and the presence of the Jutland Bank (up until ~4,200 BC). Prior to the opening of the western Limfjord, the main connection to the North Sea (and source of high salinity water) was probably to the north, through the Skagerrak.
- Marine productivity is relatively high at the bottom of the profile, but subject to fluctuations up until ~4,400 BC, when a pronounced high marine productivity phase begins. Higher primary productivity appears to coincide with lower salinity conditions, possibly due to greater nutrient retention within the fjord, though it is likely that inputs of marine water are important for replenishing oxygen and nutrients levels and perhaps also, inputs of nutrients from the catchment. Productivity declines again after ~3,200 BC as salinity begins to increase within the fjord.
- An upward trend in salinity begins around ~3,300 BC, likely due to the opening of the western Limfjord, thereby establishing a direct connection with the North Sea. It is likely that the Northern Limfjord was also open at this time. Near fully marine conditions being established by ~1,500 BC (>30 ‰) and maximum salinities occur in the late pre-Roman period (~250-0 BC). Also associated with the intensification of marine conditions (i.e. higher salinity, stronger currents, likely coarser sediments and a more open system for transport of sediment, nutrients and remains out of the basin) are decreases in the sedimentary accumulation rate and marine productivity (both after ~3,200 BC).
- A large magnitude drop in salinity occurs after ~AD 0 (~10-15 ‰ decrease in ~50 years) suggesting that Limfjord became closed off from the North Sea/Skagerrak. Brackish-marine (~15-17 ‰) conditions persist up until ~AD 600, before a transition back to almost fully marine conditions throughout the late Germanic and Viking periods. Increased abundances of lower-salinity demanding molluscs and macrophytes, and slightly increased accumulation rates for several sedimentary pigments suggest that a more productive marine environment existed between ~AD 0-600 (i.e. Roman and early Germanic period).

- Poor temporal resolution or a sedimentary hiatus (and uncertain dating) prevents detailed assessment of the marine environment throughout medieval and historical times (i.e. ~AD 1110-1920, beginning of ^{210}Pb/^{137}Cs stratigraphy).
- Following the building of a road and rail embankment in AD 1856, Kilen became a brackish water limnic system that has been subject to increasing eutrophication for the last ~150 years. Symptoms include increased nitrogen and phosphorus (up until AD 1986) concentrations, increased summer chlorophyll-*a* and diatom concentrations, increased abundance of planktonic diatom taxa, a reduction in benthic diatom diversity (particularly after the late 1970s) and reduced water clarity/light penetration. Comparison with environmental monitoring data suggests that the diatom-based salinity model performs poorly in environments heavily affected by nutrients, but a preliminary assessment suggests that the DI-TN inferences are reasonably accurate.

Chapter 4. Results: Norsminde

4.1 Introduction and aims

In this section results are presented from palaeoenvironmental analyses performed at Norsminde Fjord. This includes a range of physical (lithology, water content, organic matter, carbonate content) and biological proxies (diatoms, molluscs, foraminifera, pollen, plant macrofossils) analysed from a sedimentary sequence spanning the time period ~6,900-1,700 BC. The chapter aims to address the following research questions:

1. What are the main lithological and physical changes that occur throughout the Norsminde sedimentary profile?
2. How has chronological control been attained for the Norsminde Fjord profile and how accurate is the age-depth model?
3. How have the biological communities (diatoms, foraminifera and molluscs) living within Norsminde Fjord changed over the study period?
4. How has the salinity, productivity and nutrient status of Norsminde Fjord changed between ~7,000-1,700 BC, as inferred by these biological communities?
5. To what extent can salinity change be accurately and quantitatively reconstructed using a diatom-based salinity transfer function and by what magnitude has salinity changed over time period spanned by the profile?
6. How has the vegetation of the Norsminde catchment area changed over the study period?
7. What are the main environmental factors driving the changes in the diatom community?
8. Are taphonomic signals evident in the record and to what extent do they obscure/override true signals of environmental change at this fjord site?
9. Is there any evidence for human impact upon the fjord and surrounding landscape from the physical and biological proxies analysed here?

4.2 Core Lithology

The Norsminde core lithology is summarised in Figure 4.1 and the original lithological descriptions (using a modified Troels-Smith system; Troels-Smith, 1955; Birks and Birks, 1980) are provided in Table 4.1. The upper 64 cm of the profile represents the water column, with this being the water depth at the coring location (Figure 2.3). The majority of the underlying sedimentary profile (64 - 1,156 cm bpsl) is composed of a marine gyttja, containing varying amounts of sand, silt and clay. Scatterings of mollusc and plant material are generally found throughout, though in some sections these components can become more abundant or scarce. For example, increased plant material is present between 74-586 cm bpsl and 1,128-1,148 cm bpsl, and increased mollusc fragments between 861-866 cm bpsl and 1,096-1,114 cm bpsl. In contrast, plant material is particularly scarce between 509-861 cm bpsl and mollusc material between 590-804 cm bpsl and 900-1,096 cm bpsl. The uppermost sedimentary layer (64-74 cm bpsl) is dark brown in colour, loosely compacted with scattered findings of *Hydrobia* spp. and *Mya arenaria*. It is likely that this uppermost section has been regularly disturbed (churned up and re-deposited) since this shallow fjord reached its sedimentation limit (~3,700 years ago; see Section 4.11 for further discussion).

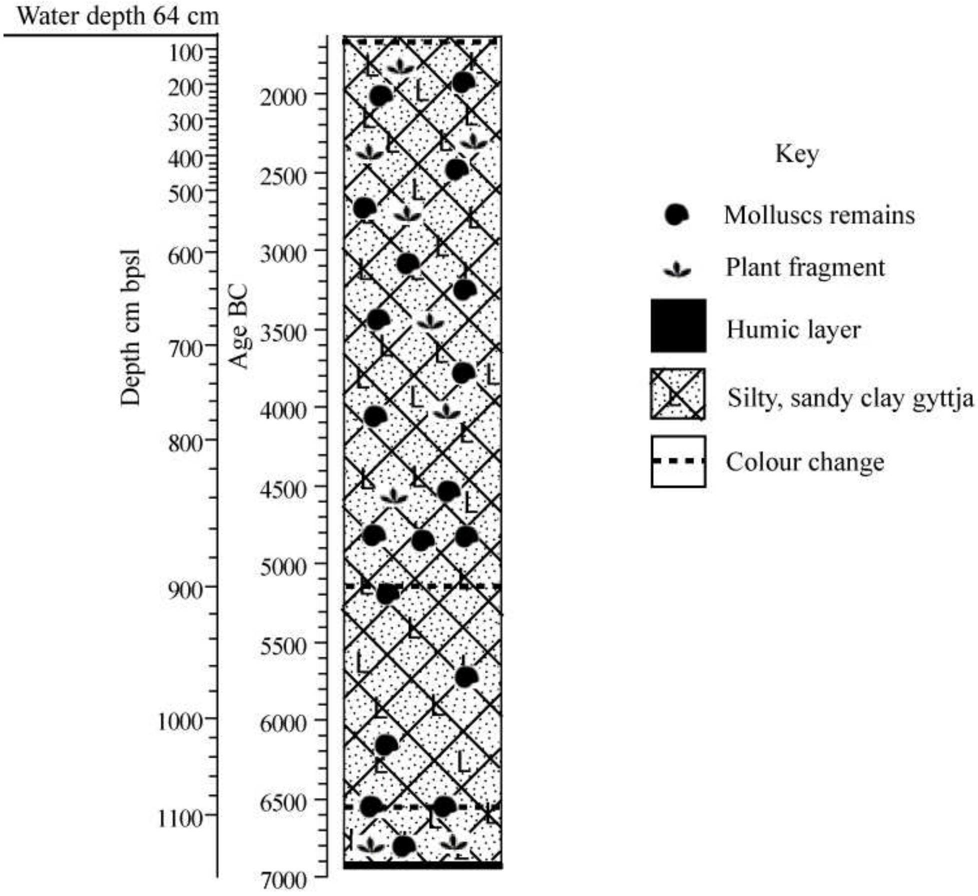


Figure 4.1 - Lithology of the Norsminde profile based on the original descriptions (Table 4.1) by Peter Rasmussen (GEUS), using a modified Troels-Smith system for characterisation of unconsolidated sediments (Troels-Smith, 1955)

Depth (cm bpsl)	Sediment description
0–64	Water
64–74	Dark-brown loose silty, sandy clay-gyttja with scattered molluscs (among others *Hydrobia* sp. and *Mya arenaria*). Ld$^{1–2}$ 2–3, Clay and silt 1–2, Sand +++, Test. moll. +, Part. test. moll. +.
74–594	Dark grey-green silty, sandy clay-gyttja with scattered molluscs and plant remains. Between 74 and 79 cm many *Hydrobia* sp. In 584 cm a larger piece of charcoal (0,5 cm). Ld$^{0–1}$ 2, Clay and silt 2, Sand +++, Dg +, Test. moll. +, Part. test. moll. +. Limes superior 0.
594–598	Sand with some clay and gyttja and very few mollusc fragments. Sand 2, Ld 1, Clay and silt 1, Part. test. moll. +. Limes superior 0–1.
598–875	Dark grey-green silty, sandy clay-gyttja with few mollusc and plant fragments. From 804 cm and downwards more molluscs and probably increasing content of minerogenic matter. Ld$^{0–1}$ 2, Clay and silt 2, Sand +++, Dg +, Part. test. moll. (+). Limes superior 2–3.
875–880	Dark grey-green silty, sandy clay-gyttja with relatively many molluscs (among others *Ostrea edulis*). A few small stones 3–4 mm in size. Ld 1–2, Clay and silt 2–3, Sand +++, Test. moll. +++ (1). Limes superior 0.
880–914	Dark grey-green silty, sandy clay-gyttja with mollusc fragments (among others *Cerastoderma* sp.). Ld 1, Clay and silt 3, Sand +++, Part. test. moll. ++. Limes superior 0
914–1,100	Dark grey silty, sandy clay-gyttja with very few molluscs. Ld 1, Clay and silt 3, Sand +++, Part. test. moll. (+). Limes superior 0 (primarily indicated by more molluscs in the layer above).
1,100–1,118	Dark grey-green silty, sandy clay-gyttja with more coarse minerogenic material and more molluscs (among others *Cerastoderma* sp.) than in the layer above. Probably also a little more organic matter than in the layer above. Ld^{0} 1, Clay and silt 3, Sand +++, Test. moll. +. Limes superior 0.
1,118–1,132	Dark grey-green silty, sandy clay-gyttja with scattered molluscs (among others *Cerastoderma* sp.); the layer is darker than the layer above. Quite some coarse minerogenic material (sand/coarse sand). Ld^{0} 1–(2), Clay and silt (2)–3, Sand/coarse sand +++, Test. moll. +, Part. test moll. +. Limes superior 0.
1,132–1,152	Dark grey-green sandy, silty clay-gyttja with quite some coarse minerogenic matter and scattered plant remains (among others fruits of *Potamogeton pectinatus*). The layer gets darker downwards. Ld$^{0-1}$ 1–(2), Clay and silt (2)–3, Sand/coarse sand +++, Dg +, Limes superior 0.
1,152–1,156	Black-brown strongly sandy humic layer. Limes superior 0.

Table 4.1- Original lithological descriptions of the sediments from the Norsminde profile, using a modified Troels-Smith system (Troels-Smith, 1955).

Below 74 cm bpsl, the sediment is predominately dark-greyish green in colour (see Table 4.1), changing only to a more greyish colour between 900-1,096 cm bpsl. Increasing amounts of coarse, minerogenic material (sand/coarse sand) are found below 1,096 cm bpsl, together with elevated numbers of cockle (*Cerastoderma* spp.; 1,096-1,114 cm bpsl) fragments. The sediment gradually gets darker below 1,128 cm bpsl as more organic material becomes visible (particularly *Potamogeton pectinatus* fruits). At the very bottom of the profile (1,152-1,156 cm bpsl), a blackish brown, sandy humic layer underlies the marine gyttja, which is likely to represent a non-marine (limnic) environment, prior to the marine incursion into the Norsminde Fjord basin.

4.3 Age Model

Details of all AMS ^{14}C dated samples (17 in total) are provided in Table 4.2. The final age model (Figure 4.2) was produced by Jesper Olsen (Aarhus University) using the age model program, B-peat v 1.0 (Blaauw and Christen, 2005), and consists of 14 AMS ^{14}C dates derived from terrestrial plant material (see Table 4.2). Two dates were omitted from the final age model; AAR-3530 and AAR-3531. The ages generated for both samples were clearly too 'old' (i.e. outliers) when compared with the other AMS ^{14}C dated samples in the profile. AAR-3531 was derived from a strongly corroded twig fragment, which may have been subject to substantial transportation, re-deposition or chemical alteration prior/during its burial. This may also be true for AAR-3530, which is derived from a wood fragment, though no visible signs of corrosion were evident.

In the uppermost section of the profile (post ~1,700 BC), it appears that a large hiatus exists, spanning the last ~3,700 years. This hiatus probably began when the shallow fjord (64 cm water depth at coring location; Figure 2.3) reached its sedimentation limit, and thereafter material has been constantly churned up, decomposed and possibly re-deposited elsewhere. The uppermost AMS ^{14}C date (AAR-4266) of 1911-1675 cal. BC (2 SD) at 161 cm bpsl appears to fit well with the age-depth model, suggesting that this part of the sequence is reliably dated (Olsen *et al.*, submitted). Therefore, Norsminde Fjord must have been extremely shallow as early as ~3,700 years ago.

Laboratory Code	Depth (cm bpsl)	Material dated	^{14}C age (yr. BP)	Calibrated age BC (± 2 SD)	δ^{13}C (‰ VPDB)
AAR-4266	156–166	15 leaf-fragments, 10 budscale-fragments	3475 ± 50	1932BC (95.4%) 1682BC	-25.1
AAR-3529	304–308	1 *Quercus* twig with bark and bud	3805 ± 50	2459BC (93.1%) 2132BC, 2082BC (2.3%) 2060BC	-25.8
AAR-4267	378–384	70 leaf-fragments	3990 ± 65	2850BC (3.2%) 2812BC, 2742BC (0.6%) 2729BC 2694BC (0.3%) 2687BC, 2680BC (91.3%) 2293BC	*-25
AAR-4268	428–434	224 leaf-fragments	3885 ± 50	2476BC (95.4%) 2204BC	-25.4
AAR-4024	484–494	132 leaf-fragments, 1 budscale-fragment, 1 Chenopodiaceae seed	4010 ± 60	2856BC (5.1%) 2812BC, 2747BC (1.5%) 2724BC 2698BC (85.4%) 2391BC 2385BC (3.5%) 2346BC	-25.6
AAR-4269	528–538	67 leaf-fragments, 13 budcale-fragments, 1 Chenopodiaceae seed-fragment	4215 ± 60	2920BC (95.4%) 2620BC	-25.5
AAR-3530	540–550	1 wood fragment	5635 ± 55	4592BC (95.4%) 4352BC	-25.3
AAR-4270	574–584	98 leaf-fragments, 6 budscale-fragments	4250 ± 50	3010BC (2.7%) 2978BC, 2958BC (0.5%) 2950BC 2942BC (51.4%) 2835BC, 2816BC (40.8%) 2668BC	-26.1
AAR-3531	608–618	1 twig-fragment (strongly corroded)	4720 ± 65	3637BC (56.2%) 3484BC, 3475BC (39.2%) 3370BC	-28.4
AAR-4271	652–662	90 leaf-fragments, 4 budscale-fragments, 1 *Phragmites communis* caryopsis, 2 *Betula* sp. fruits	4585 ± 50	3514BC (24.8%) 3424BC, 3404BC (0.3%) 3399BC 3384BC (35.3%) 3263BC, 3244BC (35.0%) 3100BC	-25.7
AAR-4025	712–716	305 leaf-fragments, 2 budscale-fragments, 1 *Epilobium* cf. *hirsutum* seed	4885 ± 60	3798BC (83.1%) 3625BC, 3600BC (12.3%) 3524BC	-26.7
AAR-4272	782–786	105 leaf-fragments, 2 budscale-fragments, 1 *Alnus* sp. Fruit	5255 ± 65	4254BC (95.4%) 3960BC	-26.1
AAR-4026	824–834	187 leaf-fragments, 5 budscale-fragments, 1 bud, 3 Chenopodiaceae seed-fragments, 1 Poaceae caryopsis	5555 ± 60	4520BC (93.7%) 4325BC, 4286BC (1.7%) 4269BC	-19.8
AAR-3532	876–878	2 twig-fragments	5980 ± 60	5007BC (95.4%) 4720BC	-30.8
AAR-3533	948–956	231 leaf-fragments	6770 ± 65	5788BC (95.4%) 5558BC	-27.7
AAR-3534	1150–1152	15 wood-fragments	7995 ± 55	7061BC (91.9%) 6734BC, 6726BC (3.5%) 6699BC	-27.5

Table 4.2 - Details of all AMS ^{14}C dated samples for the Norsminde profile. Samples in grey were omitted from the final age model (see text). *= estimated value.

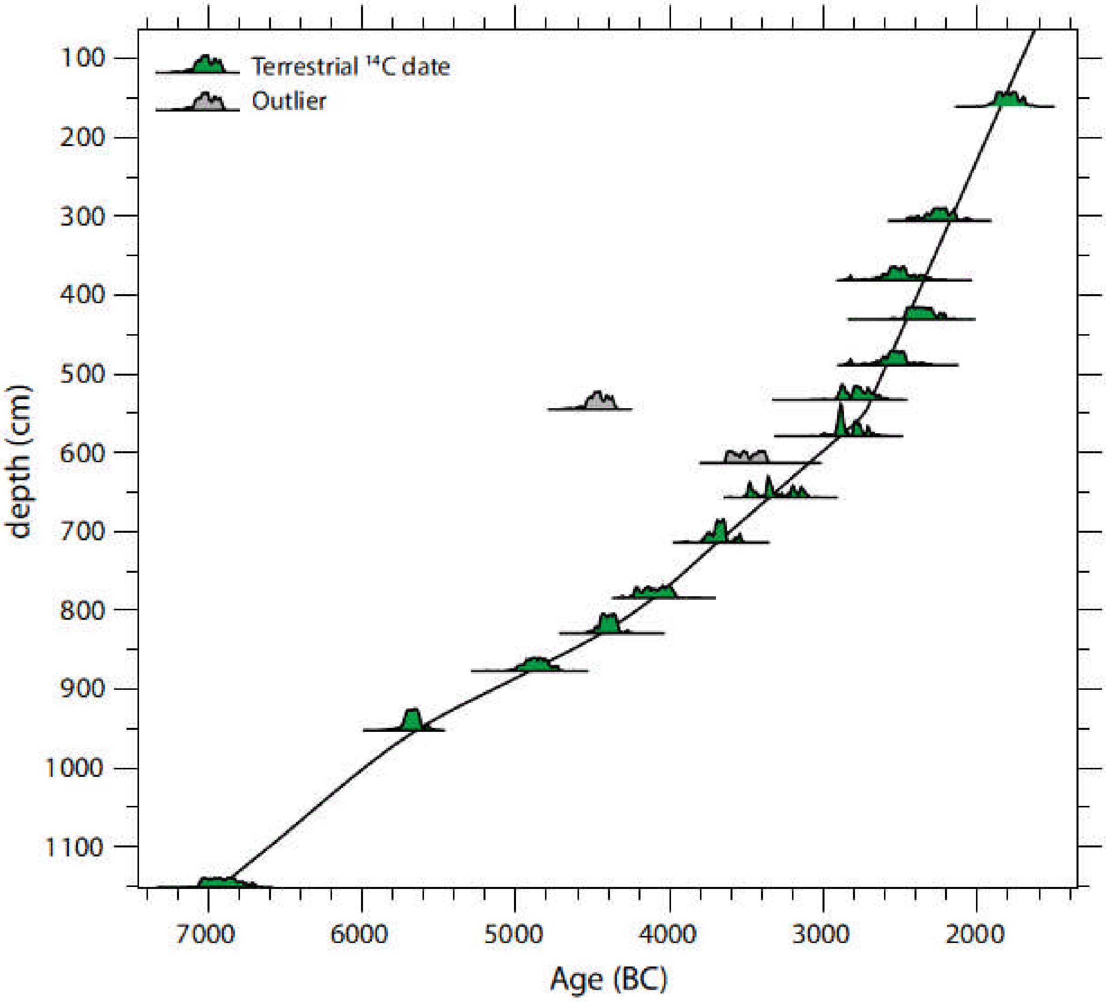


Figure 4.2 - Norsminde age model (produced by Jesper Olsen, Aarhus University, using B-peat (Blaauw and Christen, 2005).

Prior to ~1,700 BC, this ^{14}C based age-depth model appears to be reliable and robust. The date of the *Ulmus* decline at Norsminde (~4,000 BC) appears to fit well with datings of this event at many Danish lake sites (~3,900-4,000 BC; e.g. Andersen and Rasmussen, 1993). The date for the marine transgression of the Norsminde basin around 6,900-6,800 BC at 1,152 cm bpsl, is in good accord with regional estimates (e.g. Århus Bay, ~6,800 BC, (Jensen and Bennike, 2008) discussed in more detail in section 4.11).

4.4 Physical parameters

Physical sedimentary analyses (including overall accumulation rate, water content, organic matter, carbonate and minerogenic content and flux rates) were performed on 530 samples (average resolution 10 years, range 4-20 years) from the Norsminde profile and results are displayed in Figure 4.3. Results are described below in relation to four statistically significant zones (labelled NRS-1-NRS-4) determined using an optimal splitting technique (see section 2.10.1).

NRS-S1 (~7,000-6,300 BC) – At the bottom of the record, the overall accumulation rate is relatively high (~0.15-0.17 cm yr^{-1}), despite the flux of all three components (organic matter, carbonate and minerogenic matter) decreasing throughout. The only real exception is a brief increase in the minerogenic accumulation rate between ~6,760-6,580 BC. In terms of the sedimentary matrix, at the bottom of the profile the minerogenic component accounts for approximately 73-78 % of the total dry mass, but this rises to ~84-86 % after ~6,670 BC, causing both the organic and carbonate components to decrease in relative abundance.

NRS-S2 (~6,300-4,100 BC) - After remaining relatively constant in NRS-S1, the accumulation rate starts to decrease early in phase NRS-S2 (~6,300-4,100 BC). This decrease begins shortly after ~6,300 BC and continues steadily up until ~5,400 BC. Lowest accumulation rates for all parameters are recorded between 5,400-4,700 BC, with overall accumulation rate dropping as low as ~0.01 cm yr^{-1}. This changes after ~4,700 BC, as the flux of organic and minerogenic material steadily increases, causing the overall sedimentation rate to rise (which continues for the remainder of NRS-S2). The organic matter and minerogenic accumulation rates change in an almost identical manner throughout NRS-S2. The carbonate accumulation rate, however, is more variable, exhibiting a slight increase between ~5,400-4,250 BC and then becoming more erratic thereafter (e.g. two distinct peaks at ~4,930 BC and 4,410 BC).

NRS-S3 – (4,100-2,700 BC) - Early in this zone, the overall accumulation rate stops increasing (at ~4,000 BC), and remains constant (0.17 cm yr^{-1}) up until ~2,800 BC. This is largely true for each individual component, with the exception of a short lived peak in minerogenic flux between 3,070-2,900 BC and a steady decrease in carbonate accumulation after ~3,350 BC. This minerogenic peak features prominently in the relative percentage data for these components, with both the organic matter and carbonate content decreasing markedly, as increased minerogenic matter is input into the basin. At the very top of this zone (i.e. between ~2,800-2,700 BC), there is a sharp increase in the flux rate of all parameters, which subsequently causes the overall accumulation rate to rise.

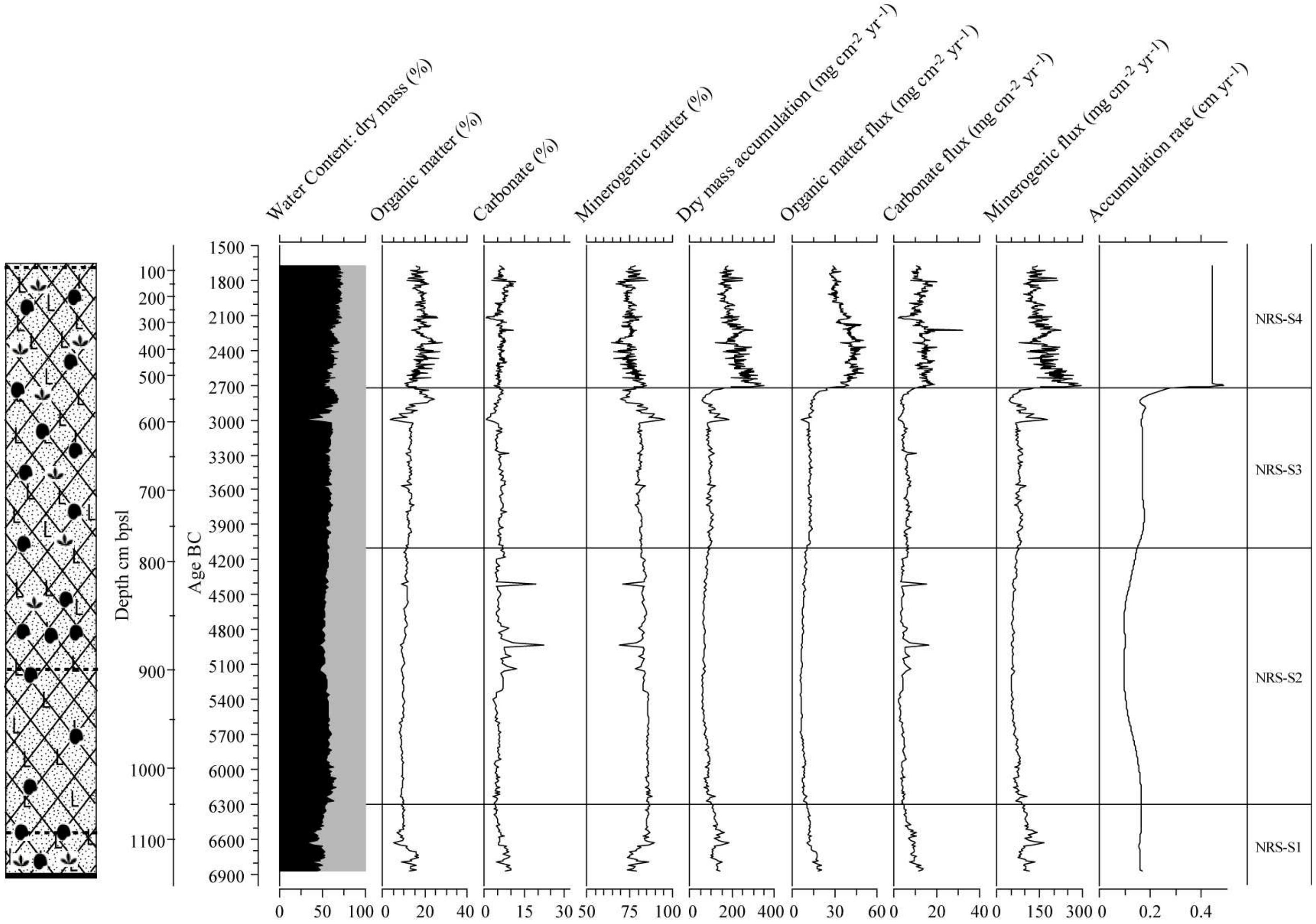


Figure 4.3 - Sediment physical parameters, overall accumulation rate and flux rates for organic, carbonate and minerogenic matter for the Norsminde profile.

NRS-S4 (2,700-1,700 BC) – The major increase in flux rates at the top of NRS-S3 results in an accumulation rate of ~0.44 cm yr^{-1} in this uppermost zone (NRS-S4; 2,700-1,700 BC), far exceeding the rate at any other time in the profile. Flux rates for each component also reach their highest levels for the entire profile in this zone but also become much more variable, exhibiting more regular and greater magnitude fluctuations than in any of the preceding zones. The only systematic change in the sedimentary matrix (i.e. percentage data) is an increase in the organic component and subsequent drop in minerogenic matter between ~2,700-2,400 BC.

4.5 Diatoms

A total of 69 diatom samples were counted, with a further 7 samples assessed for diatom presence (in which indicator species were recorded, where possible). The diatom record spans the time period ~6,900-2,020 BC, with elevated resolution over the marine transgression (6,900-6,700), the Mesolithic-Neolithic transition (~4,200-3,800 BC) and between ~3,000-2,300 BC which spans the Pitted Ware cultural period (~2,800-2,400 BC). A total of 89 different genera were recorded, with 263 taxa identified to species level. 38 species were present at >3 % abundance in one or more samples. The diatom assemblage is shown in Figure 4.4 and Figure 4.5 and associated metrics are presented in Figure 4.6. Results are described below in relation to four statistically significant biological zones (determined via optimal splitting; see section 2.10.1).

Two quantitative diatom-based salinity reconstructions are also provided in Figure 4.5. The first reconstruction includes all diatom taxa present in both the Norsminde profile (excluding *Chaetoceros* spp. and *Rhizosolenia* spp.) and the MOLTEN training set. The second reconstruction tests if DI-salinity inferences change/becomes clearer if *Fragilaria* sensu lato (s.l.) are removed. These species are often euryhaline and have great ability to colonise quickly, and survive in a broad range of unstable brackish/marine environments (e.g. Eronen, 1974; Stabell, 1985; Shennan *et al.*, 1993; Witkowski, 1994; Sabbe and Vyverman, 1995; Seppä *et al.*, 2000; Witkowski *et al.*, 2000; Miettinen *et al.*, 2007), which might in turn lead to some important shifts in salinity passing undetected. Furthermore, in the Norsminde profile *Fragilaria* s.l. percentages tend to exhibit great variability throughout (Figure 4.5, and see below), which probably drive some of the short term fluctuations in DI-salinity (Figure 4.5). These small salinity shifts, therefore, might just be the result of minor changes in community structure, possibly related to other forcing factors (e.g. water depth, nutrients) operating within the fjord. Overall, there is general agreement (in terms of salinity change) between both reconstructions over the majority of the record. There are, however, two main conflicting points: 1. the absolute values for salinity differ throughout the record, with generally higher inferences when *Fragilaria* s.l. are removed, and 2. very different reconstructed values for salinity around ~5,400 BC (see NRS-D3). The salinity transfer function and the accuracy of the presented salinity reconstructions are discussed in more detail in section 6.2.

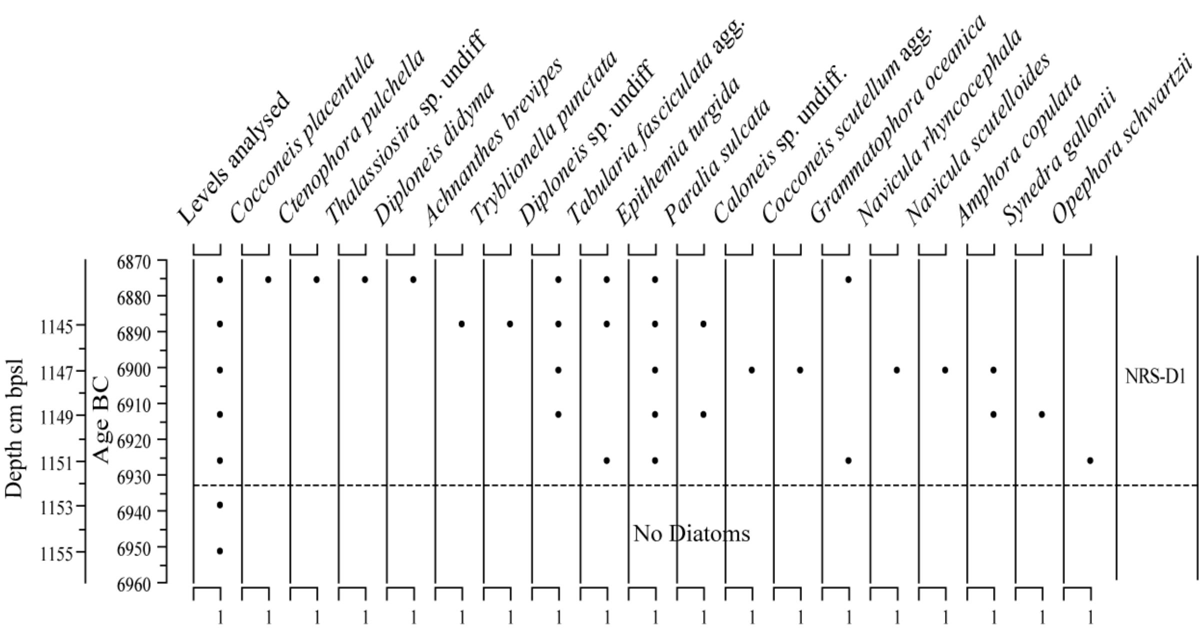


Figure 4.4 - Norsminde diatom record; ~7,000-6,870 BC. Data expressed as presence/absence.

Diatom metrics include the diatom concentration and accumulation rate, dissolution indices (**F** index and 4 stage dissolution assessment of *Paralia sulcata* valves), benthic: planktonic: tychoplanktonic ratio (B:P:T), centric:pennate ratio, DCA generated axes (discussed in section 4.10.1), Hills N2 measure of diversity and *P. sulcata* valve size. There is very little change in the predominant size of *P. sulcata* valves, with the smaller variety being more abundant throughout. In addition to diatoms, Chrysophytes and *Distephanus speculum* (marine silicoflagellate; not shown) were occasionally recorded.

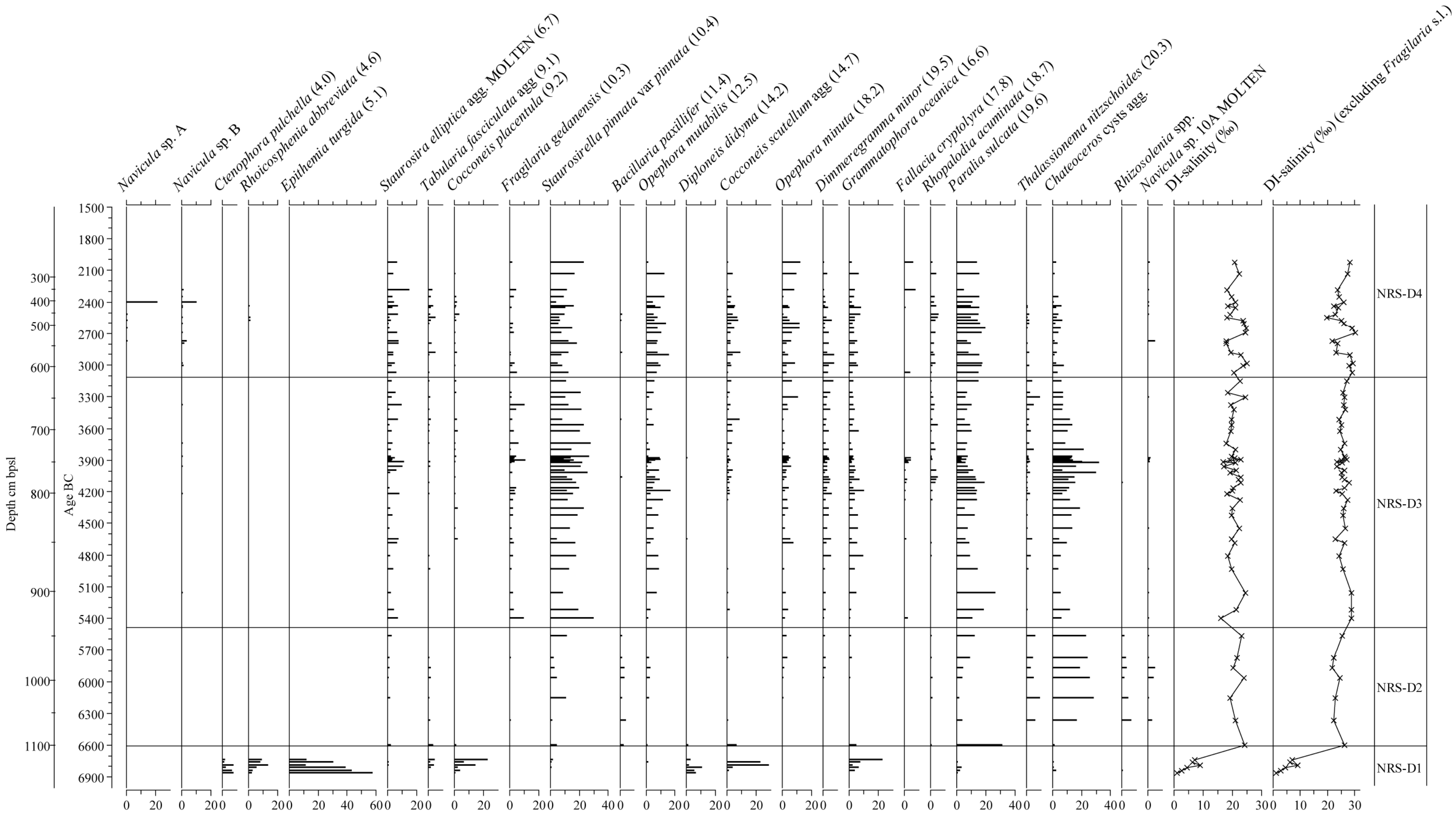


Figure 4.5 – Norsminde percentage diatom record. (>3 % abundance in one or more slides). Values in brackets refer to salinity optima in the MOLTEN salinity training set (see section 2.10.2).

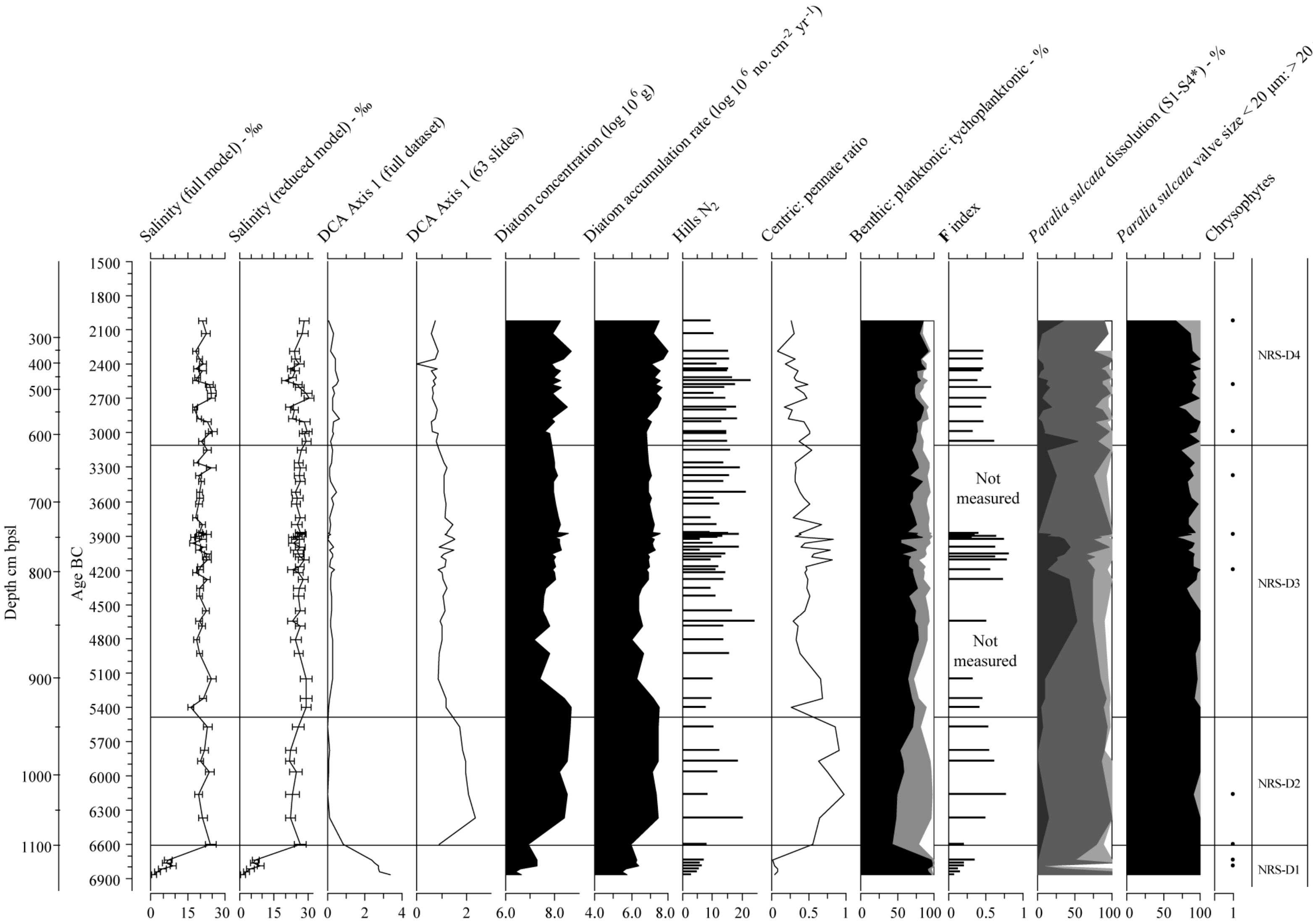


Figure 4.6 - Norsminde diatom associated metrics. Full model refers to diatom-inferred salinity including *Fragilaria* s.l. and for the reduced model *Fragilaria* s.l. were removed for reconstruction of salinity. Chrysophytes expressed as presence/absence data.

NRS-D1 (~7,000-6,600 BC) – No diatoms were present in the two samples analysed before 6,930 BC (Figure 4.4), and only a few valves (almost always fragmented) of heavily silicified robust diatoms (e.g. *Epithemia turgida*, *Tabularia fasciculata* agg. and *Diploneis* spp.) were present between ~6,930-6,870 BC, due to extremely poor preservation conditions. Extremely low diatom concentration prevented full assemblage counts (i.e. minimum 300 valves) up until ~6,800 BC. Low salinity, robust benthic diatoms, dominate the early assemblage with *E. turgida*, initially (6,860 BC) accounting for ~58 % of the assemblage, supplemented by smaller numbers of *Diploneis didyma* (7 %), *Ctenophora pulchella* (8 %) and *Rhiocosphenia abbreviata* (3 %). As the preservation conditions improve (rising **F** index; Figure 4.6) and the salinity increases, *E. turgida* and *C. pulchella* percentages fall and *Cocconeis scutellum*, *Cocconeis placentula* and *Grammatophora oceanica* all increase in abundance. Planktonic taxa are generally scarce (see B:P:T ratio and low centric:pennate ratio) in this zone.

NRS-D2 – (~6,600-5,490 BC) – There is a large increase in tychoplanktonic and planktonic diatoms in this zone, the latter of which accounts for ~40-50 % of the total assemblage between ~6,600-5,780 BC. An initial peak of the tychoplanktonic diatom, *P. sulcata* (31 %) at ~6,600 BC, marks the switch to higher salinity (~24 ‰), and probably deeper water, briefly replacing the benthic species, *C. scutellum* and *G. oceanica*, as the most abundant species. Above this, truly planktonic diatoms become more abundant with *Chaetoceros* cysts (17-30 %), *Thalassionema nitzschoides* (3-10 %) and *Rhizosolenia* spp. (2-7 %), as the main components of the phytoplankton. *P. sulcata* numbers drop drastically by ~6,360 BC (<5 %), but steadily begins to increase again after ~6,160 BC. The most abundant benthic raxa are *Bacillaria paxillifer* and *Staurosirella pinnata* (after ~6,360 BC), though a number of other species (*O. mutabilis, Opephora minuta, Navicula* spp., *Navicula perminuta* and *Dimmeregramma minor*) increase slightly after ~6,160 BC. After the initially high reconstructed salinity, associated with the *P. sulcata* peak, salinity appears to fall again up until ~6,160 BC, before starting a secondary rise in the upper section of this zone.

NRS-D3 – (~5,490-3,110 BC) – Planktonic taxa decrease again in NRS-D3, as large numbers of small *Fragilaria* s.l. (most notably *S. pinnata*, *S. elliptica* agg. *O. mutabilis*, *O. minuta* and *Fragilaria gedanensis*) become present in the record. These species all tend to exhibit irregular (non-systematic) fluctuations. A peak of *Fragilaria* s.l. (*S. pinnata* and *F. gedanensis*) at ~5,400 BC, may indicate that salinity drops as low as ~16 ‰ for a short period of time, but this is one of the major contrasting points between the two different DI-salinity reconstructions (see section 4.11). However, both reconstructions, suggest high salinity between 5,320-5,160 BC, largely driven by the marine centric diatom, *P. sulcata* (representing 19-27 % of the assemblage). As *P. sulcata* begins to drop, *Chateoceros* cysts increase (after ~4,930 BC), reaching their highest abundances for this zone between ~4,360-3,800 BC. A number of benthic species also exhibits increases throughout this zone, most notably *G. oceanica* (after ~5,160 BC) and *D. minor* (after ~4,810), followed by *O. minuta* at ~4,690 BC and slightly later (after ~4,280 BC), by *C. scutellum* and *Rhopalodia acuminata*.

NRS-D4 (~3,110-1,800 BC) – A further reduction of the planktonic community occurs in NRS-D4, as *Thalassionema nitzschoides* numbers decrease (practically disappearing after ~2,460 BC) and *Chateoceros* cysts remain low throughout. For the tychoplanktonic taxa, *P. sulcata* numbers are relatively high throughout (5-20 %) and *Fragilaria* s.l. (most notably *S. pinnata*, *O. mutabilis*, *O. minuta* and *S. elliptica* agg.) continue to dominate the increasingly important benthic community. *R. acuminata*, *C. scutellum* and *D. minor* are the main non-*Fragilaria* s.l. benthic taxa, together with smaller numbers of *Cocconeis placentula*, *Tabularia fasciculata* agg., *Navicula ramossisma* and *Navicula perminuta* in an overall diverse benthic community. This is supplemented by occasional peaks of several other Naviculoid species (e.g. *Navicula* spp. 10A, at ~2,780 BC, *Navicula* spp. A, *Navicula* spp. B, both at ~2,400 BC, and several sporadic peaks of *F. cryptolyra*). Salinity continues to fluctuate throughout, with two peaks between ~3,100-2,500 BC.

4.5.1 Diatom Preservation

With the exception of NRS-D1 (see above), overall diatoms preservation is fairly good for a marine environment (**F** index = 0.3-0.5, after ~6,600 BC; Ryves *et al.*, 2004) and therefore dissolution is deemed unlikely to be a major problem for interpretation of natural environmental change at this site. The majority of *P. sulcata* valves also fall into the first two dissolution categories.

4.6 Foraminifera

The Norsminde foraminiferal assemblage spanning ~6,860-1,860 BC is presented in Figure 4.7. Due to low abundance and poor preservation of foraminiferal tests over large parts of the profile, foraminifera are expressed as concentration data (per 20 ml of wet sediment), rather than as relative abundances (i.e. percentage data). Zonation has also been avoided for these reasons and therefore the foraminiferal assemblage results are described below in relation to the diatom-generated statistical zones (see section 0). 77 samples were analysed (using methodology outlined in section 2.7.3), containing a total of 10 species of calcareous foraminifera (from 6 genera), 6 agglutinated species (from 5 genera) and some undifferentiated planktonic species. Some re-deposited, pre-Quaternary foraminifera were recorded in the upper section of the profile (after ~3,500 BC) and are believed to represent changes in catchment depositional processes (Knudsen and Konradi, submitted).

NRS-DF1 (~7,000-6,600 BC) – Only very small concentrations of foraminifera are present in this zone with only three species recorded (*Ammonia beccarii*, *Haynesina germanica* and *Elphidium albiumbilicatum*), indicating shallow, brackish water conditions.

NRS-DF2 – (~6,600-5,490 BC) – There is a large increase in the overall concentration of calcareous foraminifera early in this zone (~6,380 BC), mainly driven by increased numbers of *H. germanica* and the arrival of *Elphidium excavatum* fo. *selseyensis*. These two species are however, supplemented by smaller concentrations of *A. becarrii* and *Elphidium williamsonii. Elphidium magellanicum* and *E. albiumbilicatum* are also present in a few samples and both exhibit a large peak at ~6,130 BC. *E. excavatum* decreases after ~6,130 BC. Variable numbers of calcareous forams throughout this zone causes the overall concentrations to fluctuate intensely.

NRS-DF3 – (~5,490-3,110 BC) – Foraminifera preservation is very poor in the lowermost section of this zone (pre ~5,160 BC), but conditions seem to improve markedly between ~5,160-4,340 BC. During this phase, calcareous foraminifera concentration (and diversity) reaches its highest levels for the entire profile. *H. germanica* and *E. excavatum* fo. *selseyensis* are again the most abundant species, but *A. beccarii, E. williamsonii* and *Elphidium incertum* also reach their maximum abundances in this period. The presence of *E. incertum* suggests relatively high salinity, particularly between ~5,160-4,910 BC, where highest concentrations for this species occur. This is further supported by small numbers of *Haynesina depressula* and *Elphidium voorthuysenii,* present only in this period. Agglutinated foraminifera briefly increase (most notably *Ammoscalaria runiana*; not shown in Figure 4.7) between ~4,630-4,340 BC. After ~4,340 BC, calcareous foraminifera concentration becomes highly variable with dissolution of tests, again, heavily affecting the assemblages. In samples where foraminifera are present, *H. germanica* and *E. excavatum* dominate the assemblage, with just sporadic recordings of *E. williamsonii, E. magellanicum* and *Aubignyna perlucida*. Both *A.beccarii* and *E. incertum* practically disappear from the record after ~4,340 BC.

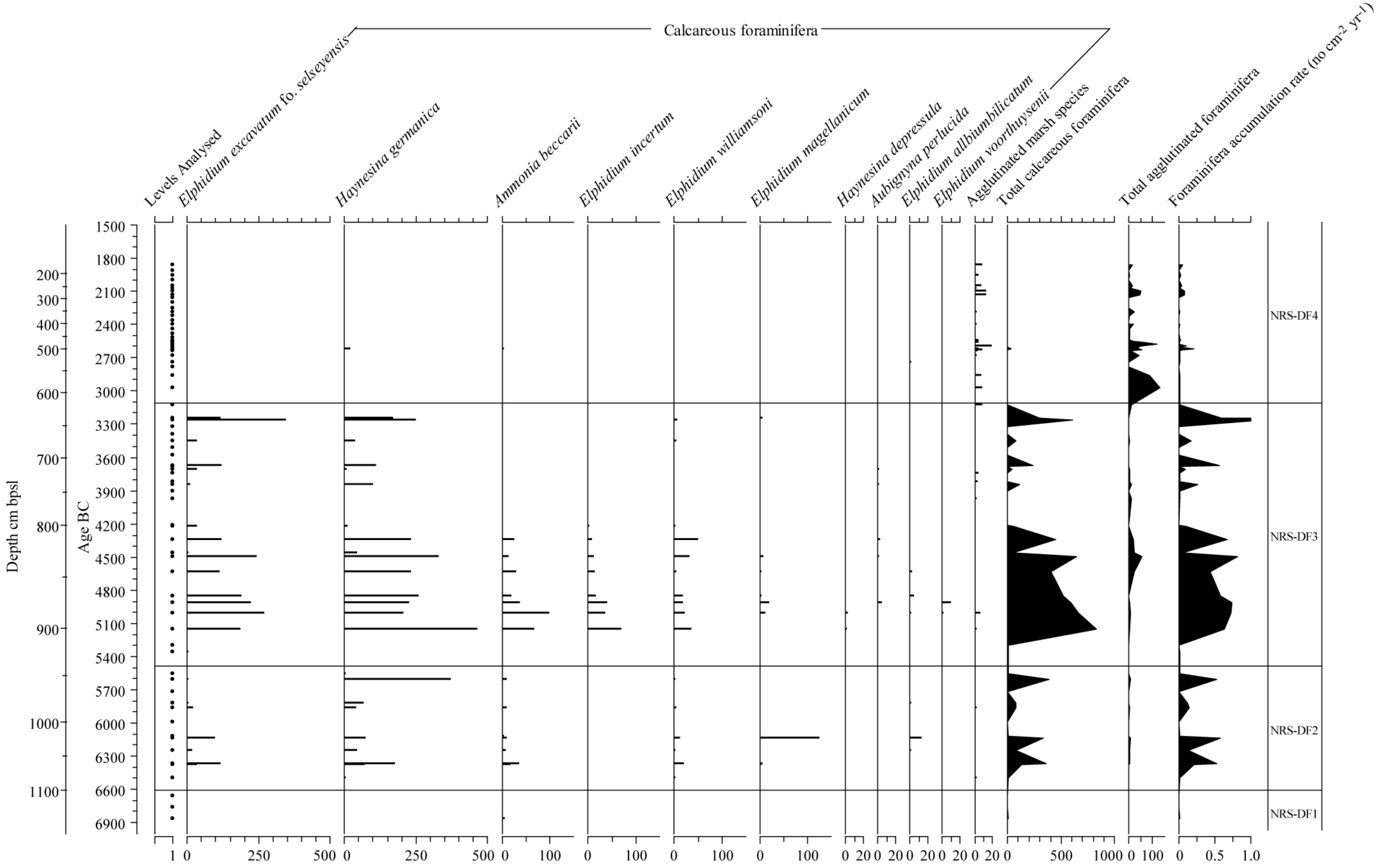


Figure 4.7 - Norsminde foraminifera record (analyst Peter Konradi, (GEUS). Data expressed as concentrations per 20 ml of wet sediment.

NRS-DF4 (~3,110-1,800 BC) – With the exception of a small peak of *H. germanica* at ~2,620 BC, this zone is almost completely devoid of calcareous foraminifera. Agglutinated foraminifera, however, become far more abundant, reaching their highest concentrations for the entire profile. The most common agglutinated foramanifer is the brackish water species, *A. runiana*, but also of note is the presence of some common marsh species (*Miliammina fusca*, *Trochammina macrescens* and *Trochammina inflata*) after ~3,130 BC.

4.7 Molluscs

A total of 74 samples were analysed for molluscs from Norsminde Fjord spanning ~6,900-2,000 BC, and this record is provided in Figure 4.8. A total of 13 genera were recorded, containing 12 taxa identifiable to species level. Overall preservation of molluscs was poor, resulting in very low absolute abundances throughout. Data are therefore expressed as presence/absence in Figure 4.8 and divided according to the diatom-derived zones (see section 0). This also means that palaeoenvironmental interpretation is limited and inferences must be treated with caution, but can provide some insight, particularly when considered in conjuntion with other proxy data (e.g. diatoms, foraminifera; see section 4.11).

NRS-DM1 (~7,000-6,600 BC) – At ~6,870, there is an individual occurrence of the gastropod *Lymnaea peregra* which can live in both freshwater and brackish-water up to ~ 8 ‰. By ~6,850 BC, undifferentiated cockle (*Cerastoderma* spp.) fragments are present, and slightly later (at ~6660 BC), *Mytilus edulis*, *Macoma balthica* and *Cerastoderma edule* all appear, which probably reflects increasing salinity.

NRS-DM2 – (~6,600-5,490 BC) – In the lowermost section of this zone (~6,600-6,520 BC), there are two recordings each of *Mytilus edulis, Scrobicularia plana* and undifferentiated *Cerastoderma* fragments. The presence of *S. plana* and an individual specimen of the mud snail *Hydrobia ulvae* (at ~6,520 BC) suggests that salinity was still rising throughout this zone. This is also further supported by finds of higher-salinity demanding species (e.g. *Hinia reticulata* at ~6,020 BC and echinid remains at ~5,890 BC) later in NRS-DM2. Molluscs become particularly scarce after ~6,400 BC, with *M. edulis*, the only species recorded in more than one sample.

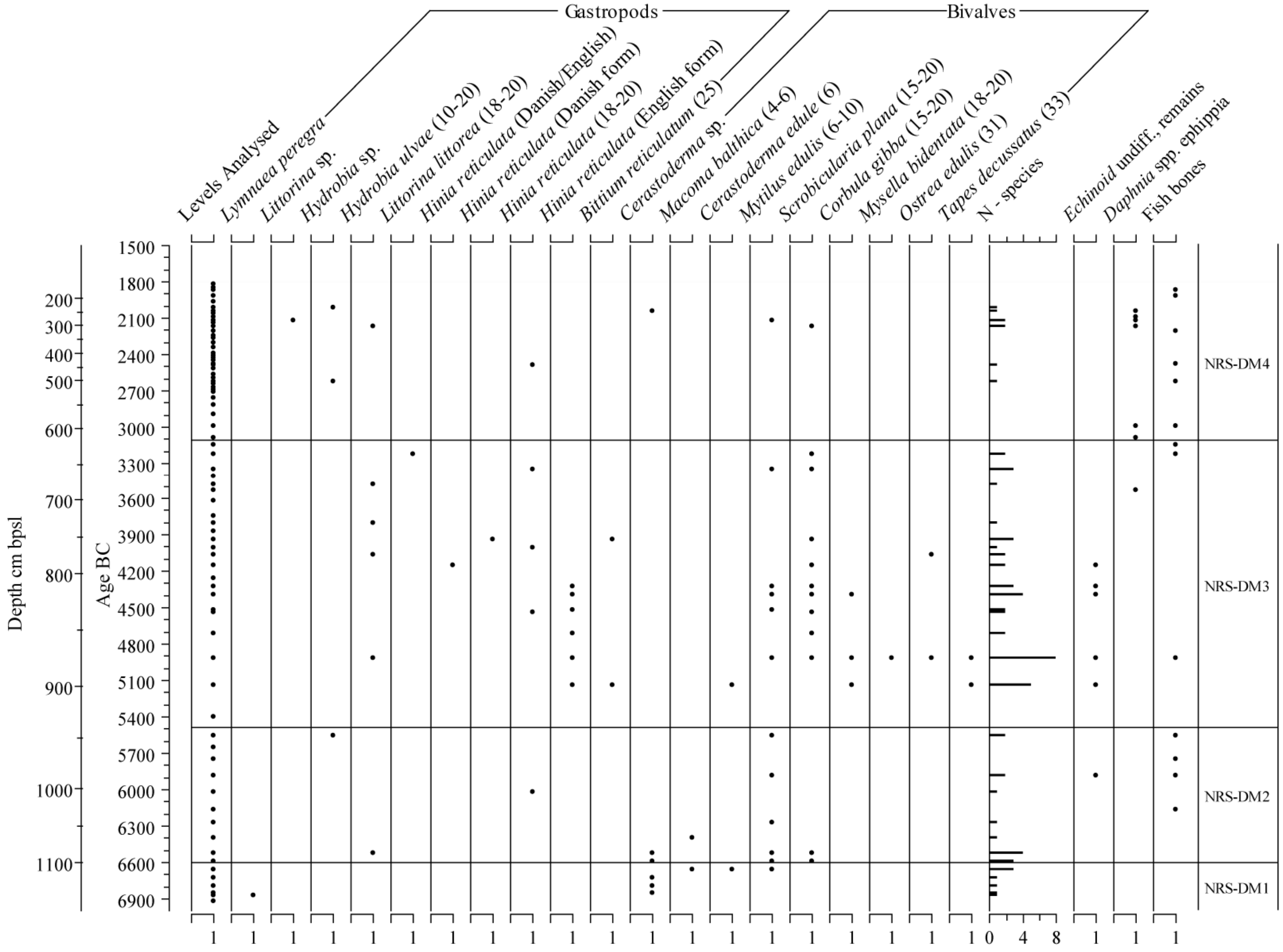


Figure 4.8 - Norsminde molluscan record and other marine animals (i.e. zooplankton, echinoderms and fish). The values in brackets refer to the lowest mean salinity inhabited by the taxa (according to Sorgenfrei, 1958). Analysts Kaj Strand Peteresen and Peter Rasmussen, GEUS). All mollusc species, zooplankton, echinoderm and fish bone data are expressed as presence/absence.

NRS-DM3 – (~5,490-3,110 BC) – Molluscs become more regular and diverse for much of this zone, despite the overall abundance remaining low. *S. plana* is the most regularly recorded taxon throughout, whereas *M. edulis, H reticulata* and *Corbula gibba* appear several times in the lowermost section of this zone (pre-4,320 BC). Higher salinity is indicated between 5,140-4,910 BC, by the presence of several higher-salinity demanding species (including *Ostrea edulis, Tapes decussates, Bittium reticulatum* and echinid remains). After ~3,800 BC, molluscs become scarce again and with the exception of *S. plana*, no species are recorded more than once for the remainder of NRS-D$_{M}$3.

NRS-DM4 (~3,110-1,800 BC) – No molluscs are recorded in the lowermost section (~3,110-2,630 BC) of this zone, but recordings of *H. reticulata* at ~2,490 BC, and *S. plana* and *H. ulvae* at ~2,170 BC, suggests salinity has not changed dramatically from NRS-D$_{M}$3. Above this, there is an individual occurrence of *M. edulis* together with an undifferentiated *Littorina* spp. at ~2,210 BC and some undifferentiated fragments of *Cerastoderma* spp. and *Hydrobia* spp. at ~2,040 BC and 2,020 BC respectively. No more molluscs are recorded after this date.

4.8 Pollen

The Norsminde pollen record is provided in Figure 4.9 (selected taxa) and aquatic plants are presented in Figure 4.10, along with associated pollen metrics (pollen sum and PCA axis 1 and 2; see section 4.10.1). A total of 62 pollen samples were counted (by Peter Rasmussen, GEUS) spanning the period ~7,000-1,700 BC. For each sample, 499-1,365 terrestrial pollen and spores were counted, containing a total of 88 terrestrial genera (31 taxa identifiable to species). A further six swamp and fenland taxa and four aquatic taxa were also recorded. Pollen types have been divided into ten groups (see Figure 4.10). Terrestrial taxa (i.e groups from “trees” to “wet ground”; Figure 4.10) are expressed as percentages of the terrestrial pollen sum (TPS). Swamp, fenland and aquatic taxa are excluded from the TPS, but their percentages are expressed relative to the TPS. Overall, 19 taxa are present at >1 % abundance in one or more samples. The stratigraphic profile has been divided up into 6 statistically significant zones (using terrestrial taxa with >5 % abundance) using an optimal splitting technique (by information content) and results are discussed in relation to these divisions.

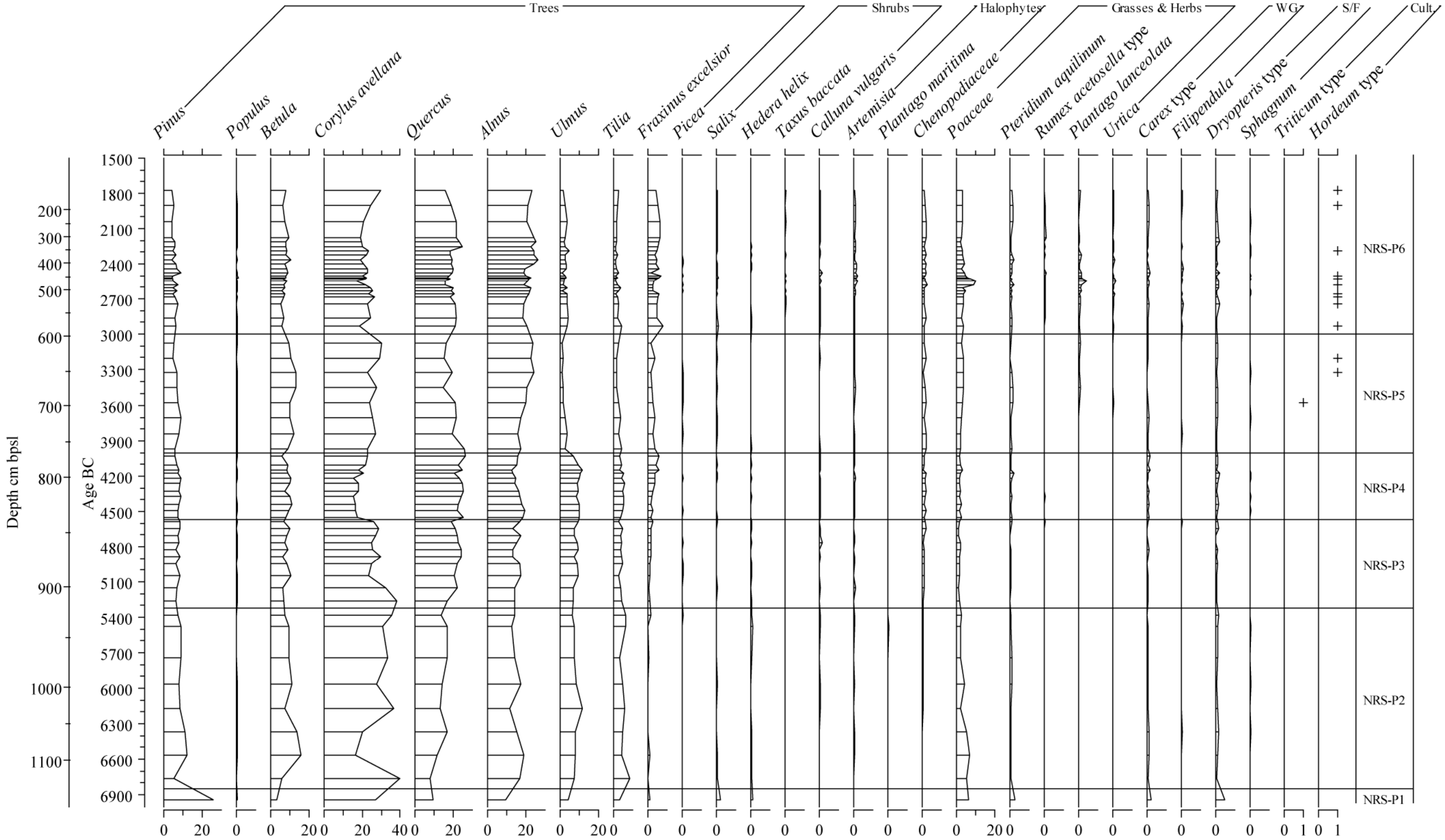


Figure 4.9 - Norsminde percentage pollen record (selected taxa). WG = wet ground taxa, S/F = swamp and fen taxa, Cult. = Cultivated taxa. (Analyst Peter Rasmussen GEUS). Data are expressed as percentages of the terrestrial pollen sum (TPS) with the exception of cultivated taxa which are expressed as presence/absence data. NB. Swamp and fen taxa are not included in the TPS but are expressed as percentages relative to the TPS.

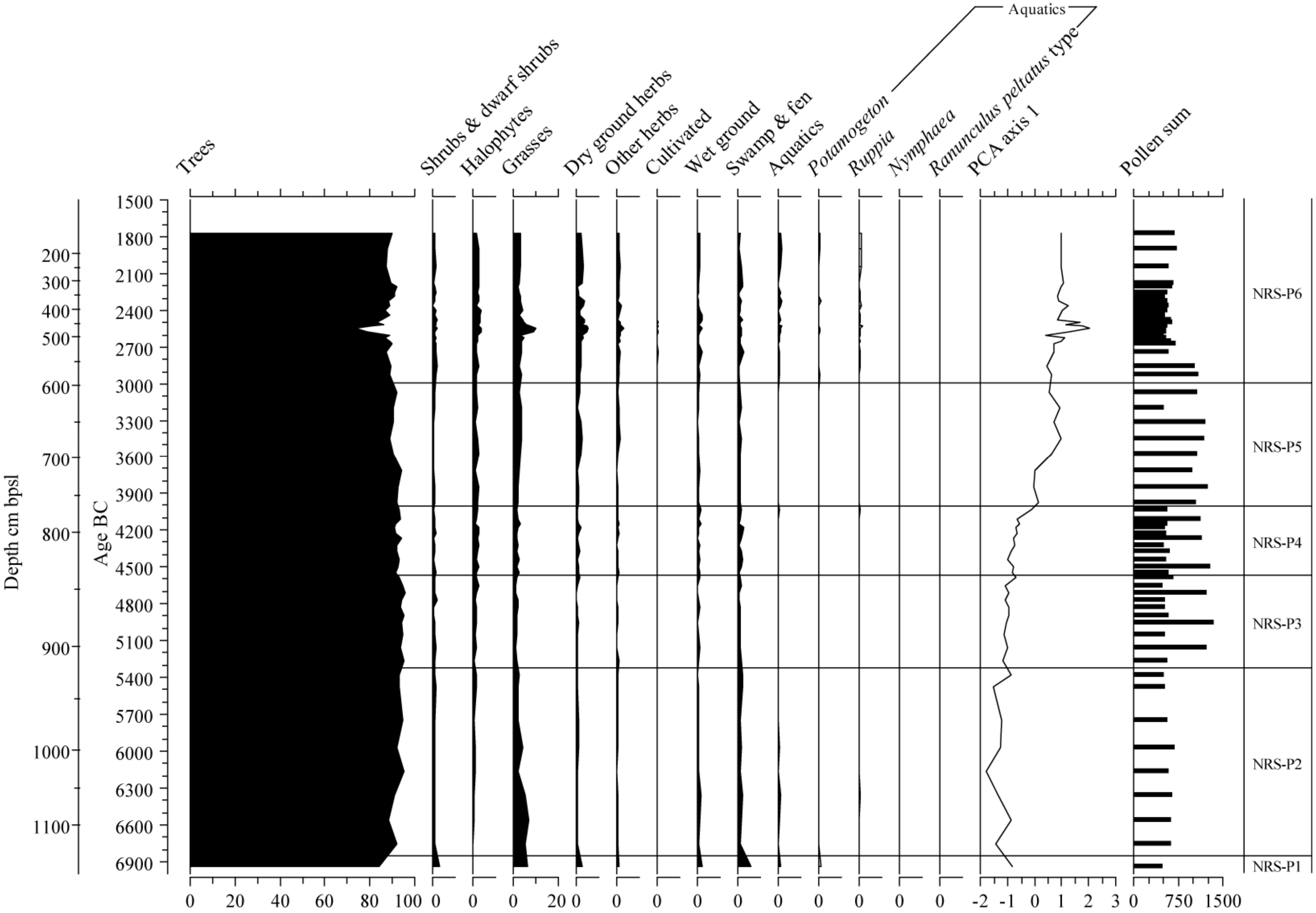


Figure 4.10 - Norsminde pollen groups, aquatic taxa and pollen associated metrics (i.e. PCA axis 1 and 2 scores and pollen sum. (Analyst Peter Rasmussen GEUS, PCA analysis Jonathan Lewis). Pollen groups and aquatic group taxa are expressed as percentages relative to the terrestrial pollen sum (i.e. trees to wet ground).

NRS-P1 (~7,000-6,850 BC) – This zone consists of just one pollen sample (at ~6,940 BC), in which *Pinus* (~26 %) and *Corylus avellana* (~27 %) pollen are the most abundant tree taxa, followed by *Quercus* (~10 %) and *Alnus* (~9 %) and supplemented by smaller numbers of Betula (3 %), *Ulmus* (~4 %) and *Tilia* (~3 %). Poaceae is the most abundant of the non-arboreal pollen (NAP) taxa, with minor occurrences of *Pteridium aquilinum*, and *Carex* type. Swamp and fen taxa are at their highest levels for the entire profile, being predominately represented by *Dryopteris* type spores (~5 %, relative to the TPS). *Salix*, which likely grew on damp ground has its highest abundance in the entire profile.

NRS-P2 (~6,850-5,300 BC) – *Pinus* pollen abundance drops at the beginning of this zone as *Alnus, C. avellana, Betula, Ulmus* and *Tilia* all increase in importance. *C. avellana* is the most abundant pollen taxon at ~6,760 BC (~40 %) and again after ~6,170 BC (for the upper section of NRS-P2; 27-37 %), but its abundance drops dramatically (by ~over 20 %) between ~6,560-6.360. *Alnus* and *Betula* increase up until ~6,560 BC, at which point *Betula* reaches it highest level in the entire profile. Both *Alnus* and *Betula* continue to fluctuate in the upper part of this zone. A short term increase in the frequency of *Pinus* also occurs at ~6,560 BC, before turning relatively stable for the remainder of the profile. *Quercus* generally increases throughout and *Tilia* (4-6 %) and *Ulmus* (6-8 %) are relatively stable. For the NAP taxa, Poaceae accounts for ~5-7% of the TPS, but its frequency drops after ~6,360 BC (thereafter ~2-4 % of the TPS). *P. aquilinum*, *Carex* type and *Dryopteris* type are ever-present, but always in very low abundances (<2 %), following decreases at the beginning of this zone.

NRS-P3 (~5,400-4,570 BC) – *C. avellana* is initially high (~39 % at ~5,260 BC), but decreases in the lowermost section of NRS-P3 (5,260-5,050 BC), as *Betula, Quercus, Alnus* and *Ulmus* all exhibit increases. *Quercus* continues to increase up until ~4,830 BC, but steadily drops off again above this. *Corylus* begins to rise steadily after ~5,050, whereas *Betula, Alnus* and *Ulmus* all stop increasing around this time. Little change occurs in the NAP taxa, with Poaceae remaining marginally the most abundant taxon, but in reduced numbers compared to the previous zone. There is a minor increase in Chenopodiaceae pollen at ~4,650 BC and an overall further reduction in *Dryopteris* type spores. Most other taxa appear sporadically.

NRS-P4 (~4,570-4,000 BC) – Another large drop in *Corylus* (by >10 %) occurs at the onset of this zone, and again *Quercus, Alnus* and to a lesser extent *Betula* synchronously exhibit minor increases. After this initial drop, *C. avellana* gradually increases throughout (along with *F. excelsior*), whereas *Betula* and *Quercus* fluctuate and *Alnus* numbers fall again (up until ~4,180 BC). *Pinus*, *Tilia* and *Ulmus* are relatively stable in NRS-P4 with the exception of a drop in *Ulmus* numbers in the upper sample (~4,030 BC) of this zone. Again there is little change in the NAP pollen assemblage compared to the previous zone (NRS-P3), with the exception of very slightly elevated numbers of Chenopodiaceae pollen and *Drypoteris* type spores.

NRS-P5 (~4,000-3,000 BC) – The onset of NRS-P5 broadly coincides with the widespread elm decline (Andersen and Rasmussen, 1993; Peglar, 1993; Peglar and Birks, 1993; Rasmussen, 2005) where *Ulmus* pollen numbers drops to ~2 % of the terrestrial pollen sum (at Norsminde) and remains low (2-3 %) for the rest of NRS-P5. At this point several taxa increase, most notably *Betula*, *Corylus* and *Alnus*. Between ~3,850-3,320 BC, *Alnus* and *Corylus* continue to increase throughout, with the latter representing ~30 % of the terrestrial pollen sum by ~3,070 BC. *Quercus* fluctuates, but overall exhibits a decreasing trend, as does *Tilia* and *F. excelsior* but to a lesser extent. For the non-arboreal taxa, higher percentages of Poaceae and Chenopodiaceae pollen are present. An overall drop in arboreal pollen (~4-6 % decrease) occurs after ~3,710 BC, as grass pollen continues to increase and is joined by slightly elevated occurrence of *Pteridium aquilinum* pollen and slightly later (~3,580 BC) by *Plantago lanceolata*. *Tritiucm* type first appears at ~3,580 BC and *Hordeum* type is regularly recorded in the record after ~3,320 BC indicating arable farming within the Norsminde catchment area.

NRS-P6 (~3,000-1,700 BC) – *Corylus*, *Quercus* and *Alnus* are the most abundant taxa, but are all subject to minor fluctuations, particularly in the lower part of this zone (up until ~2,300 BC). Above this, minor rises in *Alnus*, (~5 % increase between ~2,500-2,400 BC), and *Quercus* (by 5 % between 2,330-2,250 BC) occur, whereas *C. avellana* rises by ~10 % in the uppermost section (2,370-1,770 BC; as *Quercus* numbers fall again). *Ulmus* and *F. excelsior* both increase at the onset of this zone, the latter of which reaches its highest levels for the entire profile. Poaceae is again the most abundant non-arboreal taxon, generally accounting for ~2-4 %), but peaking between ~2,850-2,530 BC (along with *P. lanceolata* and coinciding with a substantial drop in arboreal pollen). Chenopodiaceae, *P. aquilinum* and *Artemisia* are all ever-present, joined by regular appearances from several other non-arboreal taxa (including *Rumex acetellosa, Urtica, Carex* type and *Filipendula*). *Hordeum* type (cereal) pollen is also regularly present throughout this zone, again indicating the importance of agriculture within the Norsminde catchment ares. The aquatics *Ruppia* and *Potamogeton* both reappear in the record.

4.9 Plant Macrofossils

The Norsminde plant macrofossil record is presented in Figure 4.11, generally as presence/absence profiles. A total of 74 samples (spanning 6,920-1,810 BC) were analysed (by Peter Rasmussen, GEUS) from 8-12 cm thick core slices, with the volume of sediment analysed generally ranging between 78-117 cm^3. Larger volumes of sediment (392 cm^3) were analysed in the uppermost 13 samples (after ~2,210 BC). Sufficient material was present for AMS ^{14}C dating (see section 4.3), but results are limited in terms of palaeoenvironmental inferences, due to the overall low concentration (0-85 per 50 ml, average 2.9 per 50 ml) of plant macrofossil material identifiable to genus/species level. In the 59 samples which contained identifiable material, 18 different genera were recorded, but only 5 taxa were identifiable to species level. Charcoal was commonly recorded throughout. The plant macrofossil results are discussed below in relation to the vegetation zones, as determined from the pollen record.

NRS-PM1 (~7,000-6,850 BC) – Undifferentiated bud, wood and deciduous leaf fragments are the only remains recorded from arboreal taxa in NRS-PM1. Swamp, reed and aquatic taxa are the most abundant, with the lowermost sample containing high concentrations of *Typha* seeds, *Equisetum* stem fragments and *Nymphaea* seeds. In the upper sample, *Chara* oospores, *Ruppia* fruits, *Potamogeton* and *Zanichellia palustris* fruits appear, suggesting low salinity at this time. *Typha* seeds however disappear, whereas *Equisetum* and *Nymphaea* remain present, but both decrease in concentration. Charcoal fragments are present in both samples, but the concentration is higher in the uppermost (i.e. most recent; 6,860 BC) sample of NRS-PM1.

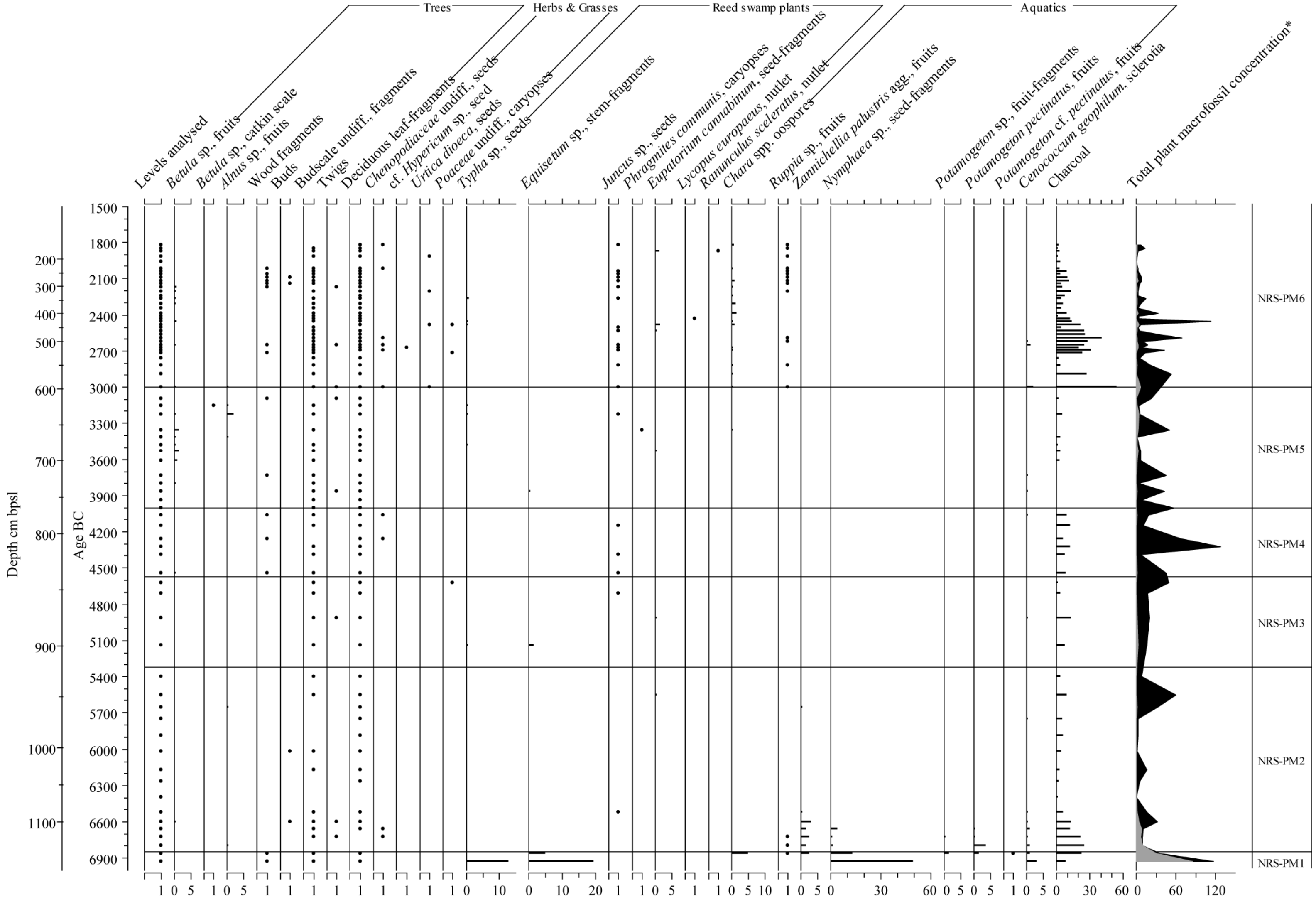


Figure 4.11 - Norsminde plant macrofossil record. Groups are organised according to pollen groupings. (Analyst: Peter Rasmussen GEUS). Data expressed as concentration per 50 ml wet sediment (bars) or as presence/absence (circles). *= Gray silhouette refers to plant macrofossil concentration excluding undifferentiated leaf, bud, wood and twig fragments.

NRS-PM2 (~6,850-5,300 BC) – Macrofossils from some aquatic taxa (*Ruppia, Potamogeton, Nymphaea* and *Z. palustris*) remain present in the lowermost section of this zone, but all disappear by ~6,520 BC. There are also a few levels with Chenopodiaceae remains and one level containing preserved *Juncus* seeds. Above ~6,520 BC, only undifferentiated buds and deciduous leaf fragments are regularly recorded. Charcoal concentration is relatively high between 6,800-6.600 BC, before dropping down to it lowest levels for the Mesolithic section of the Norsminde profile (i.e. pre 3,900 BC). However, it does begin to rise again shortly after ~6,170 BC.

NRS-PM3 (~5,400-4,570 BC) – Undifferentiated buds and deciduous leaf fragments are the only regularly recorded plant macrofossils in this zone, as overall concentration is low. Individual finds of a Poaceae caryopsis, seed remains from *Eupatorium cannabinum, Juncus* and *Typha* seeds, stem fragments from *Euqisetum* and *Cenococcum geophilum* sclerotia are the only supplements to this record. Charcoal is present in all samples, increasing up until ~4,910 BC, before dropping off again in the upper part of this zone (after ~4,910 BC).

NRS-PM4 (~4,570-4,000 BC) – There is little change in the plant macrofossil record in NRS-PM4. Deciduous leaf and bud remains are again present in most levels, supplemented by several scattered findings of wood fragments, *Juncus* seeds and Chenopodiaceae seeds. A slight increase in charcoal concentration occurs at the beginning of this zone, after which it remains stable.

NRS-PM5 (~4,000-3,000 BC) – *Betula* fruits appear in the record at ~3,800 BC and thereafter are fairly regularly recorded up until ~3,220 BC. Some *Alnus* remains are also found (after ~3,410 BC). *Cenococcum geophilum* also makes a couple of appearances in the lower part of the zone (up until ~3,740 BC) and several other taxa are present in individual samples (e.g. *Equisetum* stems, *Urtica dioeca* seeds, *Phragmites* caryopsis and *Chara* oospores). Buds and deciduous leaf fragments are regularly recorded throughout. However, the charcoal concentration drops to its lowest levels for the entire record, almost disappearing between ~4,000-3,610 BC.

NRS-PM6 (~3,000-1,700 BC) – There is a substantial increase in both the concentration and the diversity of plant macrofossils in this zone. The only identifiable arboreal remains are *Betula* fruits, which become common between ~2,650-2,200 BC. There are, however, remains from several different species of grass and herb taxa present, including multiple recordings of Chenopodiaceae, *U. dioeca* and Poaceae caryopses. Representing the reed swamp taxa, are several appearances of *Juncus* and *E. cannabium* seeds and for the aquatics, *Ruppia* fruits and *Chara* oospores reappear in the record, being regularly recorded throughout NRS-PM6. Deciduous leaf fragments and bud remains are present in most samples, and several samples also contain wood and twig fragments. There is a large increase in charcoal concentration at the onset of this zone, and with the exception of a couple of samples between (~2,820-2,760 BC), charcoal remains high up until ~2,390 BC, after which the concentration decreases again.

4.10 Statistical analysis

4.10.1 Trends in species datasets

The major trends in the diatom dataset were assessed using the unimodal technique of DCA (axis 1 gradient length >2 SD; Table 4.3). For the entire diatom dataset (69 slides with assemblage counts), 26 % of the variation was explained on the first DCA axis, but only an additional 6 % was explained by the second axis. Axis 1 appears to be driven by the change in salinity (r=0.87, p<0.05; DCA axis 1 sample scores vs. DI-salinity; full model) and probably to some degree, valve preservation (r=0.72, p<0.05; DCA axis 1 sample scores vs. **F** index) associated with the transition from brackish to marine conditions within the fjord between ~6,900-6,600 BC. The low/medium salinity, robust, benthic diatoms (*E. turgida, R. curvata, D. didyma, C. placentula C. scutellum;* >3 % abundance), that dominate the NRS-D1 assemblage (see Section 0) score highest on DCA axis 1 (Table 4.4). The majority of other species, present during the marine phase of the profile, generally have weakly-positive to weakly-negative scores (i.e. after ~6,600 BC), with substantial clustering around zero (~-0.4 - 0.4). This explains why DCA axis 1 sample scores for the entire dataset are high in NRS-D1, but decrease almost as low zero by ~6,360 BC (early in NRS-D2; Figure 4.6), and remain low for the rest of the profile.

Dataset	No. of samples	No. of species	Total inertia	Axis	1	2	3	4	Technique
Diatoms (%)	69	392	2.194	Eigenvalues	0.571	0.136	0.079	0.063	DCA
				Length of gradient	3.389	2.268	2.055	1.38	
				% variance explained	26	6.3	3.6	2.8	
	63	385	1.387	Eigenvalues	0.169	0.085	0.048	0.037	DCA
				Length of gradient	2.349	1.602	1.232	1.017	
				% variance explained	12.2	6.2	3.4	2.7	
Pollen (%)	62	88	1	Eigenvalues	0.406	0.148	0.075	0.053	PCA*
				% variance explained	40.6	14.7	7.5	5.3	

Table 4.3 - Summary of Norsminde Fjord diatom and pollen DCA/PCA results. Analyses performed on entire datasets for both pollen and diatoms and a reduced dataset (63 samples) for diatoms (see text for details; section 4.10.1.). PCA performed only on entire pollen dataset, applying Atchisons log ratio (*) for PCA linear data.

To examine more subtle gradients of change within the dataset, during the marine phase, a secondary DCA was performed on a reduced diatom dataset, excluding the lowermost 6 'brackish' samples (NRS-D1; ~7,000-6,600 BC; Section 0). Only ~12 % of variance was captured by the new DCA axis 1 (Table 4.3), with again ~6 %, explained on the second axis. The low percentage variance explained by both axis 1 and 2, is probably due to the great variability within the dataset, with a wide range of species regularly exhibiting short term, minor fluctuations in abundance, probably caused by a variety of factors (see Section 4.11). In the reduced dataset (63 samples), species scoring highest (at >3 % abundance) are the planktonic (and some benthic) species present during NRS-D2, most notably *Rhizosolenia* spp., *Chaetoceros* cysts, *Thalassionema nitzschioides* and *Cyclotella* cf *choctawatcheeana* (Table 4.4). Species scoring lowest (>3 % abundance) include the *Navicula* species, that appear sporadically in the uppermost diatom zone, NRS-D4 (i.e. *Navicula* sp. A, *Navicula tripunctata* and *Naviucla* sp B). Other benthic diatoms, commonly present throughout the rest of profile (e.g. *C. scutellum, O. mutabilis, O minuta, C. placentula, O. horstiana* and *A. helenensis*; Table 4.4) but in lower numbers in the NRS-D2 score close to zero on DCA axis 1.

Therefore the main compositional shifts captured by DCA axis 1 in the reduced dataset are not related to salinity (r=0.12, p>0.05 (insignificant); DCA axis 1 sample scores reduced dataset vs. DI-salinity, full model) but to changes in the relative abundances of benthic/planktonic species and is therefore probably driven by water depth within the fjord. Despite only 12 % of the variance being explained by axis 1, changes in DCA axis 1 sample scores therefore, do appear to provide a good summary of the major compositional shifts during the marine phase of the Norsminde profile. This is characterised by high sample scores in NRS-D2 ~6,360-5,570 BC, when an almost evenly mixed benthic/planktonic community existed within the fjord (~6,600-5,780 BC). Sample scores drop in NRS-D3 as the number of planktonic valves decrease, prior to a further reduction in NRS-D4, as the fjord begins to fill up with sediment and conditions become somewhat restricted (e.g. shallower water, reduced habitat, increased competition for nutrients with benthic diatoms).

The Norsminde pollen data exhibit a linear gradient (DCA, axis 1 gradient = 0.93; Table 4.3), and therefore major trends in the terrestrial pollen sum (excluding swamp, fen and aquatic taxa) were examined using the linear technique of PCA. Over 40 % of the variation was explained by axis 1, with a further ~15 % captured by the second axis. This suggests that axis 1 and 2 sample scores (Table 4.5) provide a good summary of the major compositional changes in the pollen dataset. Axis 1 appears to be predominately driven by a light-shade gradient, with closed canopy (shade-giving) forest dominating up until ~3,900 BC, after which it becomes a culturally influenced landscape containing higher abundances of light-demanding plants. *Betula*, *Pinus*, *Corylus*, *Ulmus* and *Tilia* all exhibit negative PCA axis 1 scores, and therefore drive the low PCA sample scores in NRS-P1 and NRS-P4 (~7,000-4000 BC). The negative scores for *Pinus* and *Betula* (pioneer trees) are due to their exaggerated abundance at the bottom of the profile, probably resulting from catchment erosion and subsequent inwash.

Quercus is an is at its most abundant in the pollen record during the late Atlantic, but due to its highest abundance in the post 3,900 BC period it scores zero. The relatively high PCA axis 1 score for *Quercus* (compared to most other closed canopy trees) is misleading, because it causes the PCA axis 1 sample scores to rise in NRS-P3 and NRS-P4 (together with a synchronous decrease in *Corylus*, between ~5,380-4,380 BC; PCA axis score -0.3). In this instance, rising PCA scores do not indicate a more open landscape, as at this time the Norsminde catchment on the high ground was covered by a dense closed canopy forest, dominated by shade-trees (most notably *Tilia*, *Ulmus*, and *Corylus*). Absolute lowest (negative) scores are exhibited by *Ulmus* and *Tilia*, whose numbers drop dramatically at

~4,180-3,970 BC, causing a shift from negative to positive PCA sample scores at the beginning of NRS-P5 (~3,970 BC).

P. lanceolata, *Alnus*, *F. excelsior* and *Urtica* exhibit the highest (most positive) species scores (for taxa >1 % abundance), all of which are most abundant in the Neolithic phase of the record (post 3,900 BC). Other taxa scoring relatively high on PCA axis 1 are Chenopodiaceae and *Artemisia*, which are again more abundant in the post 3,900 BC period, together with *Rumex acetellosa* (important on pastoral lands; Rasmussen, 2005) and *Urtica*, both of which are practically absent from the Mesolithic section of the pollen record. The positive scores for *Alnus* and *Fraxinus* could be considered somewhat anomalous, as both are shade trees, but are both more abundant in the post ~4,000 BC period, suggesting that other factors might also be important in driving axis 1 (e.g. climate changes, local water table changes, edaphic properties). The early Neolithic increases in *Alnus* and Poaceae (the latter due to forest clearances for agricultural purposes) pollen largely explain the gradual increase in sample scores, between (~3,980-3,450 BC; NRS-P5). After ~3,450 BC, PCA sample 1 scores are consistently high (above 0.4), but with a phase of elevated scores occurring over the short term deforestation phase (between 2,580-2,530 BC; see section 4.11), predominately driven by increases in *P. lanceolata*, Poaceae and *Alnus* and decreases in *Quercus*, *Corylus* and *Pinus*.

Taxa	Axis 1	Taxa	Axis 1	Taxa	Axis 1	Taxa	Axis 1
Epithemia turgida	3.70	*Staurosira elliptica* agg.	0.05	*Cyclotella* cf. *choctawatcheeana*	4.94	*Rhoicosphenia curvata*	0.33
Amphora copulata	3.69	*Glyphodesmis cf distans*	0.04	*Rhizosolenia* spp.	4.09	*Opephora minuta*	0.29
Ctenophora pulchella	3.60	*Amphora helenensis*	0.00	*Navicula salinicola*	3.36	*Grammatophora oceanica*	0.21
Diploneis stroemii	3.54	*Navicula* sp. B	-0.01	*Bacillaria paxillifer*	2.79	*Amphora helenensis*	0.09
Navicula rynchocephala	3.47	*Bacillaria paxillifer*	-0.04	*Thalassionema nitzschioides*	2.61	*Cocconeis placentula*	0.05
Diploneis didyma	3.27	*Opephora horstiana*	-0.05	*Navicula sp.* 10A MOLTEN	2.58	*Opephora mutabilis*	0.02
Rhoicosphenia curvata	3.12	*Opephora minuta*	-0.11	*Chaetoceros* cysts	2.54	*Navicula perminuta*	0.00
Cocconeis placentula	2.88	*Chaetoceros* cysts	-0.12	*Amphora graeffeana*	2.39	*Cocconeis scutellum*	-0.01
Cocconeis scutellum	2.60	*Naviucla ramossisima*	-0.14	*Navicula ramossisima*	1.55	*Opephora horstiana*	-0.01
Tabularia fasciculata agg.	2.15	*Dimmeregramma minor*	-0.15	*Fallacia tenera*	1.30	*Cocconeis neothumensis*	-0.31
Grammatophora oceanica	2.08	*Staurosirella pinnata*	-0.16	*Tabularia fasciculata* agg.	1.29	*Navicula* sp. B	-1.14
Amphoras gacialis	0.93	*Fallacia tenera*	-0.25	*Staurosirella pinnata*	1.26	*Navicula tripunctata*	-1.88
Navicula sp. A	0.81	*Navicula palpebralis*	-0.26	*Fragilaria gedanensis*	1.16	*Navicula* sp. A	-3.10
Paralia sulcata	0.31	*Thalassionema nitzschioides*	-0.28	*Fallacia cryptolyra*	1.10		
Navicula perminuta	0.29	*Fragilaria gedanensis*	-0.36	*Glyphodesmis* cf *distans*	1.10		
Rhizosolenia spp.	0.24	*Amphora graeffeana*	-0.38	*Dimmeregramma minor*	0.71		
Cocconeis neothumensis	0.19	*Navicula* sp. 10A MOLTEN	-0.38	*Amphora gacialis*	0.50		
Navicula tripunctata	0.06	*Navicula salinicola*	-0.44	*Staurosira elliptica* agg.	0.47		
Rhopalodia acuminata	0.06	*Fallacia cryptolyra*	-0.48	*Navicula palpebralis*	0.40		
Opephora mutabilis	0.06	*Cyclotella* cf. *choctawatcheeana*	-0.82	*Paralia sulcata*	0.39		
	A				**B**		

Table 4.4- Norsminde Fjord diatoms DCA axis 1 ranked species scores (taxa >1 % abundance only). A = Full dataset (69 samples; 6,900-2,000 BC), B= Reduced dataset (63 samples; 6,600-2,000 BC).

Taxa	**Axis 1**
Plantago lanceolata	0.8
Alnus	0.7
Fraxinus excelsior	0.7
Urtica	0.7
Artemisia	0.7
Rumex acetellosa	0.6
Poaceae	0.5
Chenopodiaceae	0.4
Pteridium aquilinum	0.2
Salix	0.2
Carex type	0.0
Quercus	0.0
Calluna vulgaris	0.0
Betula	-0.2
Corylus avellana	-0.3
Pinus	-0.6
Ulmus	-0.9
Tilia	-0.9

Table 4.5 - Norsminde Fjord pollen PCA axis 1 ranked species scores (taxa >1 % abundance only) for the entire dataset (i.e. 62 samples).

4.10.2 Environmental-species relationships

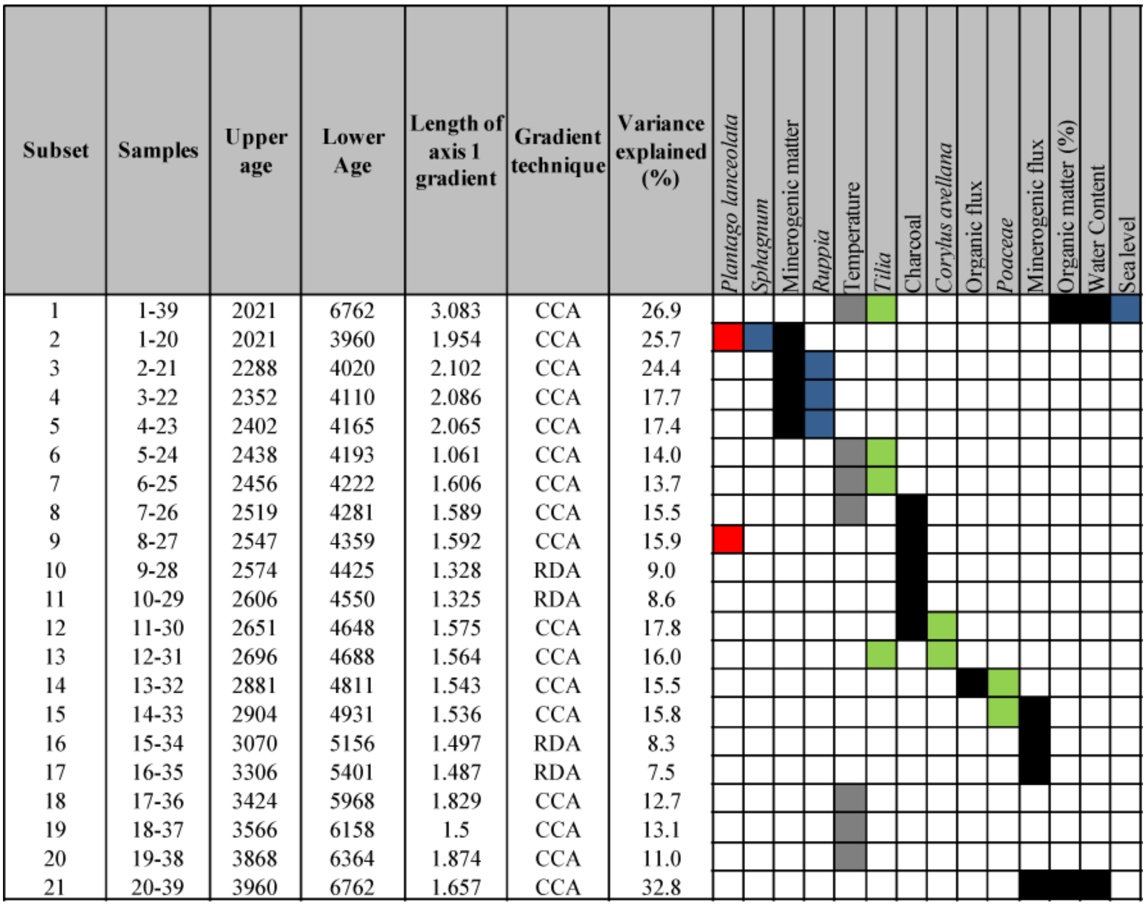
A top-down approach (i.e. lowest trophic level = response dataset, e.g. Anderson *et al.*, 2008) was implemented to explore the relationships between environmental parameters and change in the Norsminde diatom assemblage. The environmental ‘predictor’ dataset included palynological analyses (as predictors for changes in vegetation, catchment and land-use), arbitrary charcoal units (see below; for catchment/land use changes), organic matter, carbonate and minerogenic matter (abundance and fluxes), water content, total accumulation rate (as proxies for productivity, sediment load/type and catchment erosion), regional temperature (Lake Trehörningen, Sweden; Antonsson and Seppä, 2007) and sea level data (Blekinge, Sweden; Berglund *et al.*, 2005). Only 39 samples contained both palynological data and diatom assemblage data, at either corresponding stratigraphical levels or within a maximum of 25 years of one another. Therefore, environmental-species relationships were only explored using this reduced subset (39 samples), spanning 6,760-2,020 BC. Sediment age was factored out (i.e. included as a co-variable) to eliminate purely time-related processes.

To reduce long gradient lengths and reduce the effect of high value outliers, log (base 10) transformations were applied to the following environmental variables: sea level, temperature, water content, organic matter, minerogenic matter, organic flux, carbonate flux and minerogenic flux. Charcoal was included in the predictor dataset as a dummy variable, being assigned a value 1 or 0 (for each predictor sample), with 1 representing high charcoal concentration (>15 per 50 ml fresh sediment) and zero for low concentration (<15 per 50 ml fresh sediment). In cases of 'missing' overlaps for diatom and charcoal samples, the assigned value is based on the closest stratigraphic representative. Species-environment relationships were explored for both the entire dataset, and over 20 'moving window' analyses (after Bradshaw *et al.*, 2005b), to examine changing temporal trends. Each subset consisted of 20 samples, with the uppermost sample being removed and a lower sample added (i.e. samples 1-20, 2-21, 3-22…etc) for each progressive run. For each reduced subset, DCA was performed on the species data to establish the gradient length of the first axis, and appropriate linear/unimodal techniques were selected accordingly (i.e. $\text{RDA}_{\text{linear}}= <1.5$, $\text{CCA}_{\text{unimodal}} >1.5$). For CCA analyses, rare species were downweighted and for RDA analyses, species data was log transformed (after Aitchison, 1986) and centred by samples and species.

Using CCA/RDA analyses, variables (from the predictor dataset) were manually selected using a Monte Carlo (n=999 permutation) forward selection process. The significance of the resultant variables (ranked in order of importance, following CCA/RDA analysis) were tested prior to selection, and a model was progressively built up to include only those that significantly explained variation in the species dataset ($p <0.05$, with Bonferroni corrections applied for selection of each additional variable). The selection process was stopped when no further variables could significantly improve the model. Minority variables, present in less than 4 samples per subset (or sporadically occurring throughout) were omitted from selection, independent of their significance level. After selection, the significance of each model (i.e. subset of selected variables) was independently tested (i.e. PCA/RDA re-run, containing only those variables, with sediment age, again, factored out as a co-variable). All models illustrated in Figure 4.12 were significant ($p <0.05$).

For the whole dataset (i.e. subset 1, all 39 samples; 6,760-2,020 BC), five environmental variables (sea level, temperature, *Tilia*, organic matter (%) and water content) significantly explained 26.9 % of the variation in the diatoms dataset. Sea level is an important determinant of water depth within the fjord and the degree of connection with the Kattegat (and therefore salinity, saline water input, energy). The water content/dry mass suggests that over the entire dataset, sedimentary processes (probably supplying both allocthonous and autochthonous material), water table changes and perhaps flooding of sediments (e.g. increasing water content, particularly towards the bottom of the profile during the marine transgression) are important in driving changes in the diatom community. In the individual subsets, these factors are broken down more, with a series of individual sedimentary predictors appearing (organic matter (%), minerogenic flux and minerogenic matter (%); Figure 4.12 and see below). It is important to note, that as the physical parameters are expressed as percentage data, relative to one another (and therefore are often co-linear), apparent changes in one of these parameters might also/actually be related to change in another sedimentary physical parameter. This inevitably complicates the extraction of true environmental signals. The other three significant predictors (*Tilia*, temperature and organic matter (%)) in the first subset, also appear in the moving window analyses and are discussed in turn below.

For the moving window analyses, overall, climate-related processes seem to be important at the bottom of the profile (i.e. subsets 18-20; 6,360-3,430), catchment processes predominate in the middle section (subsets 10-17; 5,400-2,570 BC) and a variety of catchment, climatic and human related process in the upper section (subsets 2-9; 4,360-2,020 BC). The selected variables for each model were significant in explaining between 7.5-32.8 % (generally, 7-20 %, and average 16 %) of the variation in the diatom dataset. Maximum variance (32.8 %) is explained in the lowermost subset (subset 21; 6,760-3,960 BC), by minerogenic flux, organic matter (%) and water content. High organic content might act as a predictor for extensive bedding of aquatic macrophytes as suggested by the high abundance of aquatic plants present in the plant macrofossil record between ~6,930-6,600 BC (e.g. *Ruppia* spp., *Nymphaea* spp.). Macrophyte bedding could also explain the high abundances of benthic taxa, commonly found living as epiphytes (e.g. *Epithemia turgida*, *Cocconeis placentula, Rhoicosphenia abbreviata*, *Grammatophora oceanica*, *Ctenophora pulchella*), present at the bottom of the record (~6,860-6,740 BC). The high organic matter content might also indicate large quantities of plant matter being washed in from the surrounding catchment (e.g. high frequencies of wetland grass and *Carex* type pollen and *Dryopteris* type spores, some coarse minerogenic matter apparent in the lithology) and/or changing productivity within the fjord, perhaps due to variable nutrient status. This might be further supported by the relatively high (70-170 mg cm^2 yr^{-1}) and more variable flux of minerogenic material at the bottom of the profile, likely to be due to increasing sea levels and marine inundation of low lying lands, subsequently causing inwash of terrestrial material and erosion from the catchment. Shallow water, regular mixing up of sediment and high minerogenic loads might explain the high (and likely exaggerated) abundances of heavily silicified diatoms (particularly *Epithemia turgida* and *Diploneis didyma*) in the lower section of the profile and could explain the poor diatom preservation at the bottom of the profile (i.e. lower **F** index values at the bottom of the profile; generally **F**= 0.1-0.2 in the period 6,900-6,600 BC).



Subset	Samples	Upper age	Lower Age	Length of axis 1 gradient	Gradient technique	Variance explained (%)	*Plantago lanceolata*	*Sphagnum*	Minerogenic matter	*Ruppia*	Temperature	*Tilia*	Charcoal	*Corylus avellana*	Organic flux	*Poaceae*	Minerogenic flux	Organic matter (%)	Water Content	Sea level
1	1-39	2021	6762	3.083	CCA	26.9					▨	🟩						⬛	⬛	🟦
2	1-20	2021	3960	1.954	CCA	25.7	🟥	🟦	⬛											
3	2-21	2288	4020	2.102	CCA	24.4			⬛	🟦										
4	3-22	2352	4110	2.086	CCA	17.7			⬛	🟦										
5	4-23	2402	4165	2.065	CCA	17.4			⬛	🟦										
6	5-24	2438	4193	1.061	CCA	14.0					▨	🟩								
7	6-25	2456	4222	1.606	CCA	13.7					▨	🟩								
8	7-26	2519	4281	1.589	CCA	15.5					▨		⬛							
9	8-27	2547	4359	1.592	CCA	15.9	🟥						⬛							
10	9-28	2574	4425	1.328	RDA	9.0							⬛							
11	10-29	2606	4550	1.325	RDA	8.6							⬛							
12	11-30	2651	4648	1.575	CCA	17.8							⬛	🟩						
13	12-31	2696	4688	1.564	CCA	16.0						🟩		🟩						
14	13-32	2881	4811	1.543	CCA	15.5									⬛	🟩				
15	14-33	2904	4931	1.536	CCA	15.8										🟩	⬛			
16	15-34	3070	5156	1.497	RDA	8.3											⬛			
17	16-35	3306	5401	1.487	RDA	7.5											⬛			
18	17-36	3424	5968	1.829	CCA	12.7					▨									
19	18-37	3566	6158	1.5	CCA	13.1					▨									
20	19-38	3868	6364	1.874	CCA	11.0					▨									
21	20-39	3960	6762	1.657	CCA	32.8											⬛	⬛	⬛	

Figure 4.12 - Environmental predictor-species response relationships for Norsminde Fjord (6,760-2,020 BC). Length of axis 1 gradient refers to DCA analysis on reduced diatom dataset (if DCA axis 1 gradient length > 1.5 s.d. then unimodal CCA employed to explore trends. If DCA axis 1 <1.5. s.d. then linear RDA technique used). Subset 1 = entire dataset (39 samples). Subsets 2-21 = moving window analyses (20 samples per window). Shadings: Red = Anthropogenic indicator, Black = sedimentary regime/catchment erosion, Green: vegetation change predictors, Grey = climatic and Blue = fjord depth and level of connection with Kattegat.

Temperature is an important factor up in subsets 18-20 (i.e. between ~6,360-3,420 BC). The direct role of temperature on controlling diatom communities has previously been critiqued

(Anderson, 2000), with indirect temperature-related processes likely to have greater impact upon diatoms communities than temperature itself (e.g. Kilham *et al.*, 1996; Rippey *et al.*, 1997). Here, it is likely that temperature rise (and later decrease; see below) influenced the entire catchment (e.g. Anderson *et al.*, 2008), causing changes in the overall vegetation composition and structure, influencing edaphic processes, affecting net evaporation and run off volumes from the catchment, and affecting the abundances of grazing organisms such as molluscs and foraminifera (e.g. Yonge, 1960). Lower temperatures at the bottom of the profile might also have affected winter snow cover on the catchment and subsequent thawing events, or even sea ice formation, though the latter is unlikely due to the absence of sea ice diatoms in the profile. Thermal stratification in the late spring/summer (assuming the fjord was deeper in Mesolithic and early Neolithic period than it is today, due to higher sea levels and minimal isostatic change) might also have influenced diatom communities, particularly as temperatures warmed during the Holocene thermal optimum. Temperature is again important between ~4,280-2,440 BC (subsets 6-8). In this period, pollen-inferred air temperatures for the Lake Trehörningen area (Antonsson and Seppä, 2007) suggest decreasing temperatures and, around this time, the thermophilous species *Hedera helix* disappears (~3,850 BC) from the record (P. Rasmussen, unpublished data and see below). Again cooler temperatures (following the thermal optimum) in this section of the record might have caused greater catchment snow/ice cover. Large inputs of cold, fresh water following thawing events (during more extreme winters) might explain some of the variation in the diatom data and perhaps some of the salinity fluctuations in the DI-reconstruction.

The presence of *Corylus* (in subsets 12 and 13) and perhaps *Tilia* (significant in subsets 6-7 and 13), as environmental predictors suggest that at different times, restructuring of the forest had propagative effects upon the Norsminde Fjord ecosystem. Such changes in the forestry structure might affect the pH of the fjord (though unlikely due to large amount of saline water input from the ocean acting as a buffer for pH), nutrient inputs to the fjord, the speed of sediment delivery, sediments loads and subsequently water turbidity and depth. *Corylus* is dynamic in the phase in which it becomes a significant variable for change in the diatom dataset (i.e. between ~4,690-2,650 BC, subsets 12 and 13). The fall in *Corylus* pollen between ~5,260-4,380 BC (Figure 4.9), might be related to climate change and/or edaphic factors (see section 4.11), whilst its increase again after ~4,380 BC, probably reflects a change in forest structure.

Tilia, on the other hand is a climax tree, belonging to high ground and favouring well-drained soils (Iversen, 1969; Andersen, 1978). *Tilia* appears to be the most common tree of the primeval forest, being underestimated in pollen records due to its lower pollen production (Andersen, 1970b). Both *Corylus* and *Tilia* were prominent trees in the Danish landscape during the mid-Holocene, and therefore changes in these two species are likely to result in largescale changes across the entire catchment. *Tilia* might also act as a proxy for human-induced land-use changes and forest clearance for agricultural purposes. A reduction in *Tilia* frequencies (together with *Quercus* and *Ulmus*), with simultaneously increasing *Betula* and *Corylus* are characteristic features of Iversen's landnam (Landnam is an old Norse term for 'taking land') (Iversen, 1941; Iversen, 1973). Signs of a landnam phase are evident in the pollen diagram between ~3,900-3,000 BC (Figure 4.9, see also section 4.11). Again, this is likely to alter catchment run off rates, sediment loads and nutrient inputs into the fjord and might explain some of the variations in the diatom data (e.g. fluctuations in *Fragilaria* abundances due to increased instability of the marine environment, caused by short term removal events and/or other human impacts upon the catchment).

P. lanceolata, important in two subsets (2 and 9), is another anthropogenic proxy, considered to be a common indicator of pastoral farming, growing on pastures and meadows (Iversen, 1941; Behre, 1981; Gaillard, 2007). The lowermost of these two subsets (i.e. subset 9, 4,360-2,550 BC) incorporates the time when the species attains a continuous presence in the record, implying the beginning of pastoral farming in the landscape. This suggests that either the clearance of land for animal husbandry or that free-grazing of animals influenced the marine environment (and subsequently the diatom community). Therefore, a variety of factors might be involved, including a more open catchment, increased surface run off, greater nutrient input (from natural fertilisers) and possibly less direct human pressure placed upon the marine ecosystem following the introduction of agriculture (e.g. Andersen, 1989). Subset 2 (3,960-2,020 BC) incorporates the peak in *P. lanceolata* values between ~2,580-2,530 BC, which is synchronous with a large decrease in arboreal pollen, signifying a short-term deforestation phase. It is important to note, however, that the pollen source area is likely to be far greater than the scale of the Norsminde Fjord catchment area and therefore changes evident in the vegetation record, might to some extent have taken place outside the catchment. Care must therefore be taken with the interpretation of pollen data and the potential effects that vegetation might have on the Norsminde Fjord diatom communities as coincidental changes cannot be fully ruled out, particularly with pollen taxa low in abundance for which their presence in the local catchment is more uncertain.

The minerogenic flux significantly explains a small amount of the variation in the diatom dataset between ~5,400-2,900 BC (subsets 15-17), and is the only significant explanatory variable in subsets 15 and 16. As the minerogenic flux and charcoal concentration curve are closely related throughout the profile, it is likely that the predominant source of minerogenic matter is from the surrounding catchment (see below). In fact, later, charcoal itself becomes significant in several subsets (subsets 8-12; 4,650-2,520 BC) and, therefore both are likely to act as predictors for changes in the sedimentary supply of material from the catchment. Specifically, charcoal probably reflects land clearance and subsequent catchment/edaphic processes (discussed in more detail in section 4.11).

The mineorgenic flux appears to be more important when sediment accumulation rate is lowest, whereas the overall percentage minerogenic content is more important in the later subsets (i.e. subsets 2-5 ~4,170-2,020 BC), that incorporate the large scale rise in sedimentation (after ~2,800 BC). The minerogenic flux might also affect the diatom assemblage, with minimum diatom concentrations (~5,160-4,360 BC), occurring in the same period as the low minerogenic fluxes. This suggests that the minerogenic flux might play an important role in optimising the habitat for the diatoms, perhaps reducing abundance of mollusc beds and subsequently filter-feeding, though this is unlikely to disrupt the benthic diatom community. This period also coincides with maximum foraminifera and mollusc abundance (Figure 4.7 and Figure 4.8), and archaeological shell midden accumulation in the fjord (Andersen, 1989) suggesting a favourable environment for grazing organisms. Alternatively, it might just be a coincidental link, with minerogenic matter being reduced due to the coring location being more distant from the shoreline, reducing terrestrial input of minerogenic matter. Sea level is also at its highest during this phase (i.e. ~5,650-4,250 BC; Berglund *et al.*, 2005), and therefore, by this point, the sea might have already claimed all low-lying land, lowering catchment erosion rates and causing the input of minerogenic matter to slow down and perhaps with it, the input of key nutrients required for the diatom community to thrive (e.g. drop in *Chaetoceros* cysts between ~5,600-4,650 BC might indicate reduced spring and/or summer diatom blooms and cysts production when nutrients become limited). These results seem to highlight the important influence that the sedimentary regime operating within the fjord has upon ecosystem structuring. Changes in the sedimentary matrix may influence the diatom community in several ways, including reduced light penetration with increased turbidity of the water column (with increased sediment loads), variations in the substrate, coarsening of sediments and changes in higher trophic communities (e.g. zooplankton, molluscs). Some species of molluscs (and foraminifera) might also find it difficult to establish and survive under conditions of rapid sedimentation (e.g. Rasmussen, 1958; Yonge, 1960). The large increase in minerogenic matter flux after ~2,800 BC, and subsequent infilling of the fjord appears to have reduced the abundance of the planktonic diatom assemblage (Figure 4.6). Increased run-off rates from the surrounding catchment are also likely to have resulted in coarser sediment loads, which might explain the slight increase in abundances of epipsammic diatoms (e.g. *Opephora* species; Figure 4.5) after ~3,400 BC (see section 4.11).

Towards the top of the profile, *Ruppia* spp. becomes significant (in subsets 3-5; ~4,170-2,290 BC). This is probably a predictor for shallower water and (potentially lower salinity) as living conditions become more restricted and difficult (increased competition for space and nutrients, potentially increased grazing pressure, greater sediment disturbance in shallower water) in the fjord, as it begins to reach its sedimentation limit (see section 4.11). Specifically, beds of *Ruppia* spp. might also become important as a new epiphytic habitat for benthic diatoms (e.g. relatively high abundances of *S. pinnata*, though no increase in epiphytic diatoms is apparent), as well as providing protective habitats for grazers (e.g. increase in *Daphnia* spp. in upper sediments; Figure 4.8) and reducing light penetration to the bed of the fjord. *Sphagnum* is also a significant predictor in the uppermost subset (i.e. subset 2; 3,960-2,020 BC), which might represent an increase in swamp/fenland vegetation surrounding the fjord, probably due to decreasing sea level. Increased sediment load (after ~2,800-2,700 BC) and lower water levels probably caused silting up the fjord sides, causing the fjord to contract, resulting in the emergence of wet swampy land. Therefore, *Sphagnum* is probably another predictor for increased sedimentation rate, shallower water and restricted living conditions.

4.11 Synthesis of Results and site discussion

In this section, the results from the individual proxy analyses (described in sections 4.2-4.9) are bought together to provide a synthesis and discussion of the Norsminde Fjord environment and surrounding catchment area over the early-mid Holocene (~7,000-1,700

BC). In Figure 4.13, a summary diagram is provided of the major changes exhibited by these proxies (see Figure caption for details). Temperature (from Lake Trehörningen; Antonsson and Seppä, 2007) and sea level curves (from Blekinge, southeastern Sweden; Berglund *et al.*, 2005) have also been included to examine changes in relation to regional climate and marine environmental change. Five environmental zones have been included to aid interpretation labelled NRS-E1, NRS-E2, NRS-E3a, NRS-E3b and NRS-E4. An environmental zone has been assigned where at least two of the original optimally-split proxy zones (see above) fall within close temporal proximity to one another. The boundaries for NRS-E1/NRS-E2, NRS-E2/NRS-E3a and NRS-E3b/NRS-E4, have been determined by close agreement between the diatom and pollen derived zones. The boundary split for NRS-3a and NRS-3b, marks close correspondence between changes in the pollen and physical parameters profiles. Changes within Norsminde Fjord and its surrounding catchment (spanning ~7,000-1,700 BC) are discussed below, in relation to these environmental zones.

NRS-E1 (6,900-6,730 BC) – The marine transgression phase – Preservation of both micro- and macro- fossils is very poor at the bottom of the record, which therefore limits palaeoenvironmental interpretation. The lowest recorded marine microfossils occur at ~6,930 BC and belong to three brackish-marine diatom species (*E. turgida*, *G. oceanica* and *T. fasciculata* agg.), suggesting that the fjord was either already under marine influence at this time, or at least had received an input of marine water (e.g. a short term pulse). This is further supported by finds of brackish-water foraminifera (*A. beccarrii* and *H. germanica*) and *Ruppia* spp. remains at ~6,860 BC, the latter of which only tolerates salinities down to 3 ‰ (Moeslund *et al.*, 1990). The first recorded mollusc at ~6,870 BC, is a well-preserved specimen of *Lymaeae peregra,* which can live in both freshwater and brackish water with salinities up to 8 ‰ (Mandahl-Barth and Bondesen, 1949), suggesting that the fjord might still have been under freshwater conditions at this date. The first unambiguous brackish water molluscs appear at ~6,850 BC (*Cerastoderma* spp.) Establishing the exact timing of the limnic-brackish transition is difficult due to some probable signs of re-working and re-deposition of sediments. Almost all diatom valves present between (6,930-6,870 BC; Figure 4.4) are heavily fragmented (and scarce) and the only identifiable valve fragments belong to robust diatoms, which to some degree, might be able to survive the re-deposition process.

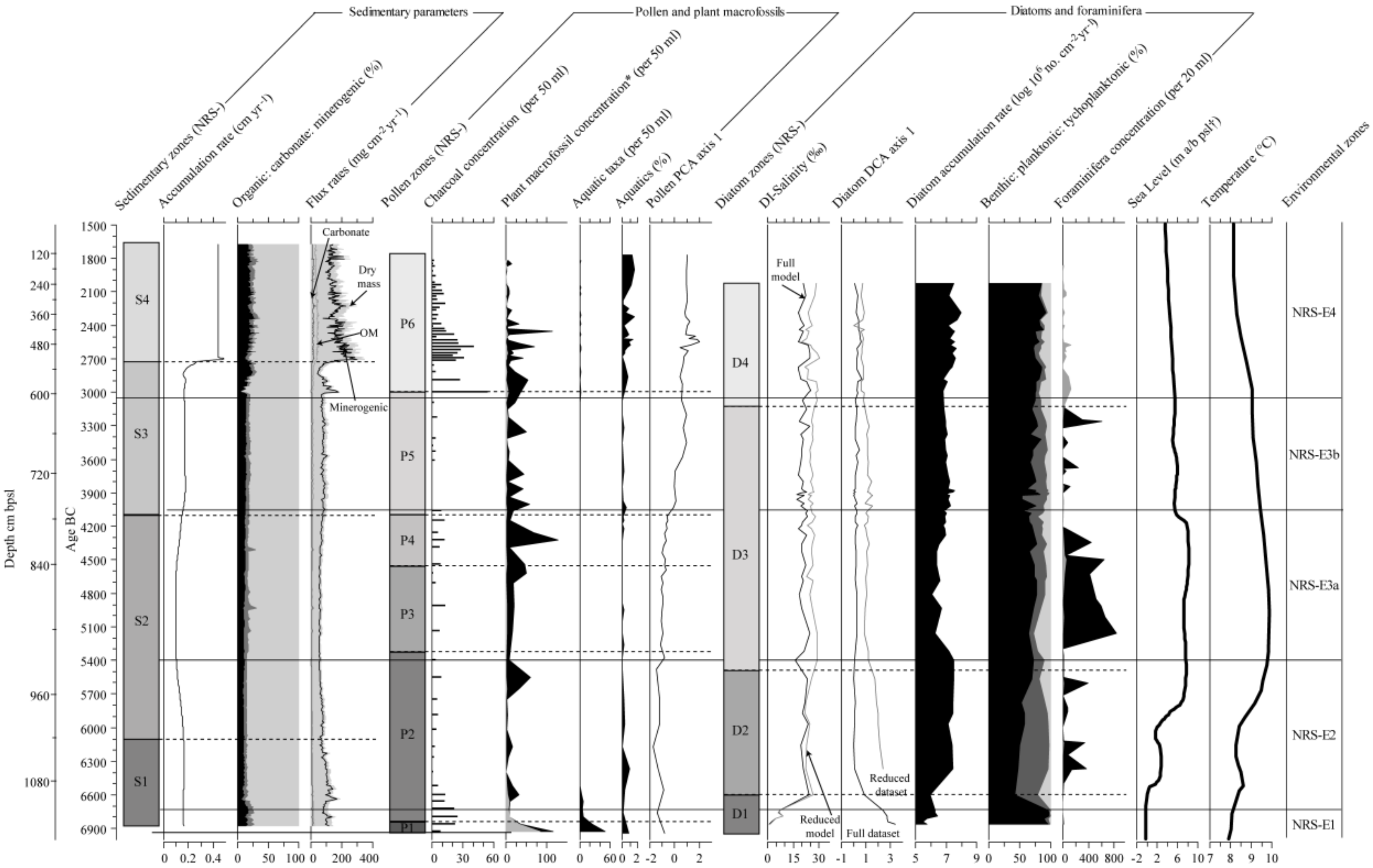


Figure 4.13- Summary diagram of the major changes occurring at Norsminde Fjord (~7,000-1,700 BC). *= Gray silhouette refers to plant macrofossil concentration excluding undifferentiated leaf, bud, wood and twig fragments. Full model, refer to Di-salinity including all fossil taxa, reduced model refers to diatom-inferred salinity excluding *Fragilaria* s.l. from the fossil dataset. Regional sea level curve from Blekinge, SE Sweden (Berglund *et al.*, 2005) and pollen-based temperature record from Lake Trehörningen, Sweden (Antonsson and Seppä, 2007). †m a/b/ psl = metres above/below present sea level.

Nevertheless, the bulk of evidence suggests that some time between ~6,930-6,870 BC the first inundation of marine water into the Norsminde Fjord basin occurred, and that a full marine connection was definitely established before ~6,800 BC. This agrees well with other local estimates for the area such as Århus harbour (*in-situ* pine stump dated to ~7,200 BC; Dencker, 1999) and Århus Bay (oldest marine mollusc ^{14}C dated to ~6,800 BC; Jensen and Bennike, 2008). Prior to this, it is likely that a lake or a stream existed at the coring point, but the exact date of the transition from limnic-brackish conditions cannot be further constrained, due to both poor preservation and/or reworking of sediments. This transition might have begun as intermittent pulses of saline water, but again this is difficult to establish. This would, however, provide an alternative explanation to re-deposition/dissolution, for findings of high-salinity tolerant diatoms, such as *P. sulcata* and *G. oceanica*, pre-dating (and after co-existing with) low-salinity fresh/weakly brackish water diatoms (e.g. *N. scutelloides*, *N. rhyncocephala, A. copulata, E. turgida*; Figure 4.4). It is important to note, that although this community almost certainly reflects low salinity conditions (e.g. 1 ‰ at ~6,860 BC), the precise DI-estimates of salinity in the lowermost samples must be treated with caution. The accuracy of these inferences is questionable due to the low concentration of diatoms and dominance of robust diatoms (e.g. *E. turgida,* 58 % at ~6,860 BC). It is highly likely that dissolution of the more brittle and less silicified taxa has resulted in exaggerated representation of some taxa in this zone skewing the inferred salinity.

Above this (i.e. after ~6,800 BC), the salinity appears to increase rapidly, as the Norsminde basin begins to develop as a marine system and sea level continues to rise. The diatom flora increases in diversity as the salinity rises and preservation improves, and brackish-marine diatoms become more abundant, replacing the low-salinity, robust diatoms characteristic of the bottom few samples (pre-6,600 BC). Molluscs are low in number and foraminifera absent in the upper section of this zone, probably due to poor preservation, low salinity and difficult living conditions (e.g. high sediment loads). The overall accumulation rate is relatively high (0.15 mg cm^{-2} yr^{-1}) in NRS-E1, which is largely driven by the input from the surrounding catchment. This is suggested by the increase in minerogenic matter between 6,680-6,650 BC, and general agreement between the minerogenic flux and charcoal concentration curve (both within this zone and throughout the profile). Erosion of the catchment and drowning of low lying land during the marine transgression probably best explains this.

At the very bottom of the record, most major tree taxa are present (except possibly *F. excelsior*), but with *Pinus* and *C. avellana* being the most abundant pollen taxa. *C. avellana* pollen accounts for 25-40 % of the TPS in this zone, suggesting that it featured prominently in the local landscape, together with *Pinus* and *Alnus* and to a lesser extent *Quercus*, *Ulmus* and *Tilia*. The arrival of *Tilia* in western and northwestern Jutland has been dated to ~7,000 BC, but it may have arrived earlier than this in the Norsminde area, as it already account for ~3 % of the TPS in the lowermost sample (~6,940 BC). Whether *F. exclesior* was present in the Norsminde area before ~5,380 BC, is uncertain due to low pollen frequencies. The particularly high numbers of *Pinus* (e.g. 26 % at ~6,940 BC), suggests that it was locally abundant, probably growing on wet soils or peats, but the high pollen percentages could be due to over-representation due to buoyancy of the saccate (vesiculate) pollen grain (P. Rasmussen, unpublished data). *Salix* is also at its highest numbers in this lowermost zone, probably also growing in these damper areas.

The high abundance of wet ground and marsh/fenland plants, present in both the pollen and plant macrofossil assemblages, suggest that the fjord was immediately surrounded by marsh/wetlands. The relative abundances of these taxa, however, may be exaggerated, due to catchment erosion and subsequent washing in of sediment from the sides of the fjord. *Carex* type pollen (~2 %) and *Dryopteris* type spores (~5 %) are the main indicators in the pollen record of damp/marshy conditions, but remains of *Typha* seeds and *Equisetum* stem fragments are also found in the plant macrofossil record. The majority of the grass pollen before ~5,750 BC, is most likely from wetland species such as *Phragmites australis* and *Phalaris arundinaceae*. Remains of aquatic plants are also relatively high in the earliest phase of the record, particularly *Nymphaea* spp., *Chara* spp., Z*anichellia palustris* and *Potamogeton* spp.. These taxa can tolerate low salinity and live in freshwater, and therefore, they cannot provide any further information regarding the timing of the marine transgression.

NRS-E2 (~6,730-5400 BC) – The early marine phase – Salinity continues to increase rapidly and by ~6,600 BC, DI-salinity (in both models) exceeds 20 ‰ (~15 ‰ rise between 6,760-6,600 BC). Water depth also appears to increase, resulting in a large increase in the proportion of planktonic diatoms present, which account for approximately half the diatom community between ~6,600-5,780 BC. According to the DI salinity reconstruction (with all taxa), a further increase in salinity occurs after ~5,880 BC, which broadly coincides with the second major sea level transgression documented in the Blekinge stratigraphy (i.e. L2 transgression, ~5,750-5,250 BC, Berglund *et al.*, 2005; Figure 4.13). Under the full model, this rise is interrupted by a peak of *Fragilaria* s.l. (most notably *S. pinnata* and *F. gedanensis*) at ~5,400 BC, which causes DI-salinity to fall to ~16 ‰ (discussed in more detail below;

NRS-E3). DI-salinity under the reduced model does not show this drop (Figure 4.13). Mollusc abundance remains very low in this zone, but as the water depth and salinity increases, preservation conditions for foraminifera appear to improve. Foraminifera become more common after ~6,500 BC, but abundances of individual species are somewhat variable (most notably *H. germanica* and *E. excavatum* fo. *selseyensis*), which might indicate fluctuating salinity. A peak of *E. magellanicum* at ~6,130 BC suggests a relatively open connection with the Kattegat at this time (Knudsen and Konradi, submitted).

The rise in minerogenic matter between ~6,680-6,650 BC coincides with rapidly increasing salinity (as evident in both DI-reconstructions; see Figure 4.5) within the fjord. As indicated above, this is probably due catchment erosion, as water level increases, further claiming low-lying land and washing material into the fjord. This is supported by the relatively high numbers of Poaceae pollen (most likely from wetland species) and wetland taxa (*Carex* type and *Dryopteris* type), which are more abundant in the record, up until ~6,170 BC. The accumulation rate begins to fall after ~6,180 BC. Interestingly, the entire phase of lowest accumulation throughout the Norsminde profile broadly coincides with highest sea levels (Mörner, 1976; Berglund *et al.*, 2005) and temperatures (Antonsson and Seppä, 2007) for the southern Baltic area (i.e. ~6,000-4,200 BC; Figure 4.13). When sea level is higher, a more open connection with the Kattegat might result in more material being lost to the open sea and increased energy within the system might lead to greater decomposition of material. This suggests that sea level (affecting accommodation space and level of connection with the Kattegat), climatic factors and later, human impact all influence the sedimentary regime of Norsminde Fjord (Rasmussen, submitted).

There is little change in the forest tree species present, with *C. avellana* still being the most abundant pollen type for the majority of this zone. The curves for *C. avellana, Betula* and *Pinus* all exhibit fluctuations in the lower section of this zone (e.g. *C. avellana* pollen drops by ~20 % (of the TPS) between 6,760-6,560 BC, with *Betula* and *Pinus* pollen simultaneously increasing), before stabilising after ~6,200 BC. This coincides with the reduction in both the flux of minerogenic matter and a drop in the overall accumulation rate. This probably means lower rates of catchment erosion, and therefore might indicate that the earlier fluctuations (between ~6,760-6,170 BC) in *Corylus, Betula* and *Pinus* abundance, were caused by erosion and re-deposition of sediments (P. Rasmussen, unpublished data), rather than large-scale changes in the forest composition. *Quercus* pollen frequencies rise throughout, suggesting that it became more prominent in the landscape, accompanied by relatively extensive occurrence of *Alnus, Ulmus* and *Tilia*.

In terms of non-arboreal taxa, the low porportion of grass pollen (the majority of which after ~6,000 BC, most likely originate from dry ground taxa) and dry ground herbs (most abundant being *P. aquilinum*), suggests a closed-canopy forest with a sparse field-layer vegetation. Two halophyte groups (Chenpodiaceae and *Artemisia*) become more common in NRS-E2, probably growing on shore meadows along the edge of the fjord. The latter of these taxa is ever-present after 6,500 BC, which broadly coincides with the establishment of fully marine conditions in the fjord, and probably the promotion of shore-meadow plant communities (P. Rasmussen, unpublished data). Both summer and winter temperatures at this time were probably warmer than today, perhaps up to 2.5˚C warmer in summer, and able to support the growth of the thermophilous species, *Hedera helix* and *Viscum album* within the catchment (Iversen, 1944; Antonsson and Seppä, 2007; Seppä *et al.*, 2009; Brown *et al.*, submitted; P. Rasmussen, unpublished data). Pollen and spores from wet habitats taxa (most notably *Carex* type and *Dyropteris* type) are less abundant than in NRS-E1, but remain present, probably still important in the nearshore area of the fjord. Due to lower rates of catchment erosion and greater distance away from the coring location under higher sea level conditions, they feature less prominently in the profile. This is again supported by a drop in the minerogenic flux in this zone. Aquatic species practically disappear from the plant macrosfossil record somewhere between 6,600-6,500 BC, consistent with the transitions to high salinities (>20 ‰ under both models), though minor quantities of *Ruppia* spp. and *Potamogeton* spp. are found up until ~6,360 BC in the pollen record (likely originating from fresher, upstream parts of the fjord).

NRS-E3a (5,400-4,050 BC) – The middle maximum phase – High salinity is indicated by all three marine proxies in the lowermost section of this zone. There is, however, some difference between the two diatom reconstructions, regarding the nature of salinity change in this part of the record. If *Fragilaria* s.l are removed, DI-salinity appears to rise steadily from 5,870-5,160 BC. However, a peak in *Fragilaria* species (*S. pinnata* and *F. gedanensis*) at ~5,400 BC results in a low-reconstructed salinity (~16 ‰) under the full model, but it is difficult to assess whether this is a true salinity decline, or just an opportunistic increase in *Fragilaria* s.l. (often abundant in unstable systems such as fjord/estuarine systems or recently isolated/transgressed basins (Eronen, 1974; Stabell, 1985; Seppä *et al.*, 2000; Miettinen, 2002). The period between ~5,550-5,360 BC is almost devoid of foraminifera, which might support a low salinity inference at this time (see below). However, it is clear that the

Norsminde Fjord environment is dynamic between ~6,000-5,000 BC, with the sedimentation rate decreasing and planktonic abundance (most notably, *T. nitzschoides* and *Chaetoceros* cysts, after ~5,570 BC) dropping, perhaps due to increased instability in the fjord (e.g. rise in Fragilaria s.l abundances), changing nutrient status and or increasing salinity, greater marine connection with increased sea level and increased inwash of robust marine diatoms of *P. sulcata* (e.g. increasing *P. sulcata* abundances between ~5,570-5,160 BC) . Therefore, changes in other parameters might also be influencing these communities.

While it remains unclear as to whether, salinity began rising at ~5,800 BC, or rose quickly after ~5,400 BC, all three proxies document high values between ~5,160-5,140 BC (~DI-salinity of 24 ‰ with full model; 29 ‰ under reduced model). High-salinity demanding molluscs appear (e.g. *O. edulis, T. decussatus, B. reticulatum*) at ~5,140 BC. *E. incertum* is the main foraminiferal indicator of high salinity (requiring at least 20-25 ‰; Knudsen and Konradi, submitted), particularly between (~5,160-4,900 BC) where it reaches its maximum abundance. It is further supported by small numbers of *Elphidium voorthuysenii* and *Haynesina depressula* (found only at salinities >24 ‰ in marginal areas of Kattegat and Skagerrak; Alve and Murray, 1999). This might also suggest that foraminifera preserve better under higher salinity conditions at Norsminde Fjord, which has also been observed at Horsens fjord, situated ~25 km south of Norsminde (Kisum, 2007). The maximum salinity phase appears to persist for longer in both the molluscan and foraminifera records (up until ~4,340 BC in the foraminiferal record than the diatom record, but both the DI-reconstructions suggest salinity has fallen again by 4,940 BC). In the succeeding phase (after ~5,160 BC), evidence from both diatoms (characterised by high and fluctuating percentages of *Fragilaria* s.l.), and later (after ~4,340 BC) the foraminifera (fluctuating *H. germanica* and *E. excavatum* concentrations) suggest variable salinity (and/or other environmental parameters).

A late peak in *Chaetoceros* cysts occurs between (~4,300-3,800 BC), possibly due to the increased sedimentation rate and increased input of nutrients from the catchment. This might result from both falling sea levels (see Figure 4.13; Berglund *et al.*, 2005) and later, the marked changes in forest structure associated with the *Ulmus* decline at Norsminde (beginning around ~4,100 BC) (see NRS-E3b) and/or human disturbance of the catchment. An alternative explanation might be that nutrient status is actually higher in Norsminde fjord under lower sea level, with shallower water perhaps preventing nutrient from the catchment from being washed out to sea. This might also be supported by the lowest diatom concentrations between ~5,400-4,300 BC, broadly coinciding with maximum sea levels and lowest accumulation of sediments (Figure 4.13). The reduced input of sediment (particularly minerogenic matter) from the catchment during higher sea levels (i.e. ~6,200-3,900 BC) might also be important for providing stable habitats for molluscs and foraminifera, thereby placing added grazing pressures upon the diatom community.

Salinity might also play a role, with some evidence for lower DI-reconstructed salinities occurring simultaneously with high *Chaetoceros* cyst numbers, though there are uncertainties attached to the DI-reconstruction at this time (see below; NRS-E3b). Dissolution might also explain lower diatom concentrations, but neither the **F** index, or *P. sulcata* dissolution index suggest any shift towards poorer preservation. Diatom concentration increases again after ~4,430 BC, broadly coinciding with peak values for *Chaetoceros* cysts (and decreasing sea level; Figure 4.13) and thereafter remains relatively high.

Arboreal pollen is at its most abundant in phase NRS-3a and rising numbers of *Quercus*, together with high numbers of *Ulmus, Alnus* and a marked increase in *F. excelsior* (after ~5,380 BC) suggest that the Norsminde catchment is dominated by a dense forest over this period. *Corylus* pollen is still abundant but decreases throughout this zone, possibly being out-competed by *Quercus* and other climax trees. Alternatively climatic and/or edaphic factors may have contributed to its decrease. The very low percentages for grasses and dry ground herbs (1-4 %) suggest a continued densely forested landscape in this phase. There is very little change in the abundances of *Carex* type and *Dryopteris* type pollen, and minerogenic flux rates are at their lowest. Aquatics are practically absent in this phase of the record.

NRS-E3b (4,050-3,050 BC) – The late marine phase – The foraminifera indicate lower salinity in this phase, but this is not evident in the diatom flora or DI-salinity reconstruction (Figure 4.14). Calcareous foraminifera become much sparser (and less diverse), as the higher salinity species disappear from the record and those that remain (most notably *H. germanica* and *E. excavatum*), are subject to greater fluctuations in numbers. Molluscs also become fewer in number, but there is, however, little change in species composition compared to the uppermost section of NRE-3a. There is also very little change in the diatom flora, compared to the preceding phase. *Fragilaria* s.l. taxa continue to fluctuate, which inevitably cause fluctuations in the DI-salinities, making it difficult to isolate clear patterns of salinity change.

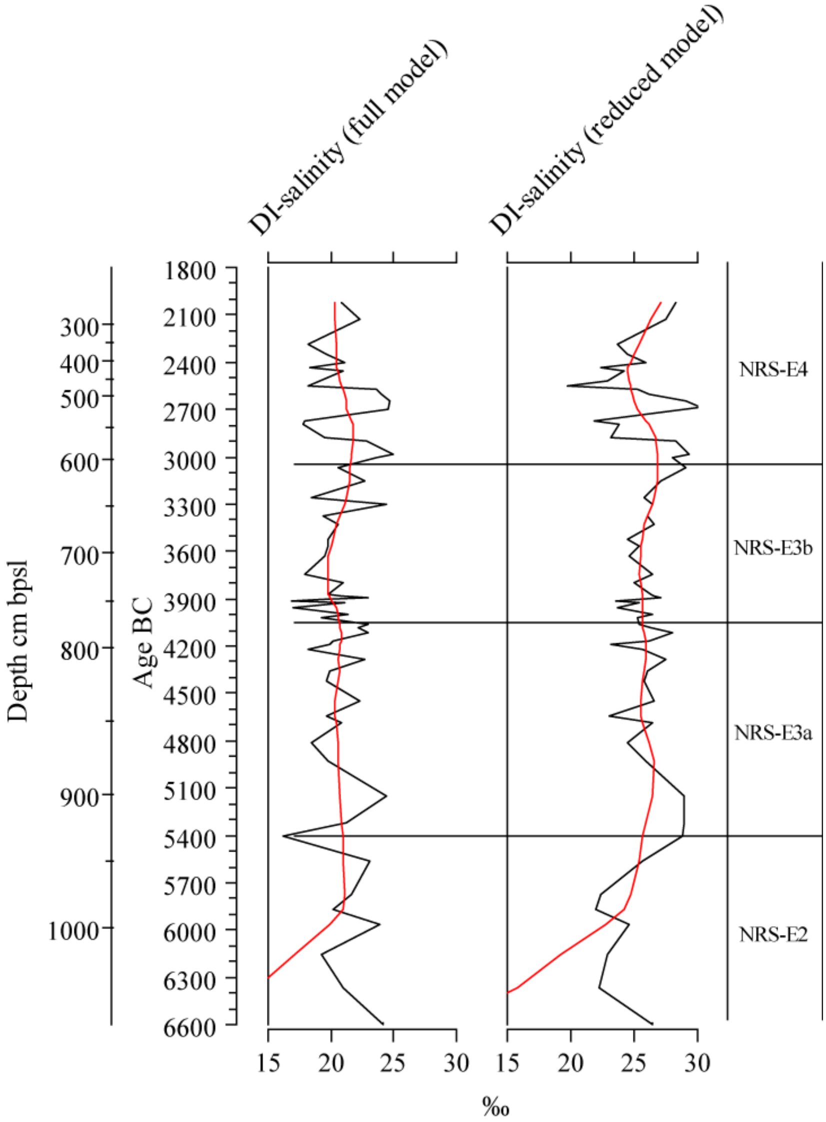


Figure 4.14 - DI- reconstructions with added loess smooth (0.25 span) for Norsminde Fjord between ~6,600-2,100 BC.

Therefore, on the basis of the multiproxy evidence presented here, lower salinity in the early Neolithic can only be considered equivocal at best. Furthermore, interpretation must be viewed in the light of likely large short-term variability (e.g. decadal-centennial scale fluctuations in DI-salinity possibly up to 5 ‰ in magnitude) and must consider the dynamic nature of the region at this time. Lowering of sea level, elevated sedimentation rate, forest re-structuring (following the elm decline) and human impact upon the landscape all occur (see below) in this phase. Therefore, again salinity may not be the only factor influencing the diatom assemblage at this time, and therefore the apparent short term shifts in salinity, might actually be driven by small scale restructuring of the community, driven by other forcing factors (e.g. weak salinity relationship with DCA axis 1 for the reduced diatom dataset; 6,600-2,000 BC). In terms of the diatom community, perhaps one of the most important additional factors is water depth (likely to be indirectly exerting pressure through depth related process; e.g. Weckström and Juggins, 2006, and see section 6.2), as the shallowing of the fjord probably accelerated in this zone, driven by the higher accumulation rates (~0.17 mg cm^{-2} yr^{-1} after ~4,000 BC) and the drop in sea level between 4,200-3,800 BC (see Figure 4.13). Benthic diatoms gradually increase in number throughout this phase, as both the tychoplanktonic and planktonic community are reduced (*P. sulcata* decreases after ~4,080 BC and *Chaetoceros* cysts after ~3,800 BC). Variable nutrient status again, might also be influencing the diatom community (see NRE-3a).

In NRS-3b, large scale changes occur in the pollen record, with *Ulmus* values dropping down to as low as ~2 %, accompanied by increases in *Betula*, *Alnus* and *Corylus*. This suggests a re-structuring of the forest, resulting from the reduction of elm in the catchment. Disease and/or human impact upon the landscape are the most commonly cited causes for the decline of elm in northern Europe (Troels-Smith, 1960; Ten Hove, 1968; Perry and Moore, 1987; Peglar and Birks, 1993; Parker *et al.*, 2002; Rasmussen, 2005), but it is difficult to isolate the exact cause. At Norsminde, there is no direct evidence for human interference from the palaeoenviromental records, at the time of the elm decline. *Poaeceae* remains low and unambiguous agricultural indicators (e.g. cereals, *P. lanceolata*) do not appear until later in the record. Charcoal concentrations are also at their lowest for the entire profile between ~4,060-3,610 BC, which might suggest little activity in terms of fire-induced forest clearance by humans at this time. The archaeological evidence from surrounding sites (e.g. shell middens) provide evidence for changes in cultural practice (Andersen, 1989; Enghoff, 1989; Enghoff *et al.*, 2007) at the time of the Mesolithic/Neolithic transition, but again nothing that directly suggests forest clearances between ~4,180-3,710 BC. This probably means disease is most likely to be responsible for the elm decline at Norsminde.

A gradual increase in Poaceae pollen frequency occurs throughout, which is probably related to both a natural reduction in forest density associated with the elm decline and the introduction of agriculture. The first hard evidence for agriculture appears shortly after ~3,600 BC, when a continuous curve for *P. lanceolata* begins, signifying pastoral farming (Iversen, 1941; Behre, 1981; Gaillard, 2007). Pollen from cultivated crops also appears around this time; *Triticum* type at ~3,580 BC and slightly later, but much more commonly thereafter, *Hordeum* type pollen appears at ~3,320 BC.

Signs of Iversen's classic landnam (Iversen, 1941; Iversen, 1973) are identifiable between the elm decline and ~3,000 BC, with decreasing *Ulmus*, *Tilia, Quercus* and *Fraxinus* values paralleled by rising *Betula* and *C. avellana* values accompanied by increases in Poaceae and *P. lanceolata* and the presence of cereal pollen. These events are inferred as the cumulative effects of many short-term clearance episodes followed by arable and pastoral farming. Whether or not the catchment was frequently disturbed by fire after ~3,500 BC is inconclusive. *Betula* remains high throughout, but changes in *Populus* and *Pteridium aquilinum* are minimal. Charcoal exhibits a minor increase at ~3,610 BC, but overall remains relatively low, probably mainly being washed in from the catchment, rather than air-borne (see below).

NRS-E4 (3,050-1,700 BC) – The infilling phase – The fjord almost certainly becomes shallower in this phase as it approaches its sedimentation limit and living conditions become somewhat restricted (reduced habitat, increased competition for nutrients amongst benthic diatoms, increased grazing pressure). This is suggested by both the reduction in planktonic diatom ratio (which begins decreasing at ~3,150 BC, and again after ~2,400 BC) and the rise in agglutinated marsh foraminifera (*T. inflata*, *M. fusca* and *T. macrescens*). Despite sea level records suggesting that sea level is slowly decreasing throughout this zone (Figure 4.13), rapid infilling with sediment is the main cause of this shallowing, particularly after ~2,800-2,700 BC, when the average accumulation rate almost trebles within the basin (from 0.17 to 0.44 cm yr^{-1}). This probably explains why salinity (as inferred by both diatom reconstructions) does not really decrease in this upper section (i.e. after ~3,000 BC) of the profile, which would be expected under rapidly falling sea levels. In fact, DI-salinity (especially under the reduced model) suggests two phases of increased salinities between ~3,100-2,500 BC, which might also explain the re-accumulation of oyster dominated middens at Norsminde over the Pitted Ware cultural period (~2,800-2,400 BC; Andersen, 1983, discussed in section 6.3.2.1). Likely fluctuating salinities and sporadic peaks of benthic diatoms suggest living conditions are highly variable within the fjord after ~3,000 BC.

What exactly caused these large increases in the flux of organic matter, carbonate and minerogenic material, in just over 100 years (from 2,850-2700 BC), remains equivocal. Rasmussen, (unpublished) suggests that anthropogenic processes and subsequent catchment soil erosion best explains the contemporaneous increase between the minerogenic and organic components. Catchment erosion is further supported by the more regular appearance of pre-Quaternary foraminifera after ~3,000 BC (Knudsen and Konradi, submitted), which must have been washed in from the surrounding catchment following erosion from pre-Quaternary sediments. In addition, a more open catchment (and subsequent higher run-off rates), a shallower fjord, narrower Kattegat entrance (perhaps due to silting up of the entrance and development of beach ridges; Andersen, 1989), sea level change, high productivity (e.g. high diatoms concentration, high organic content), extensive bedding of macrophytes trapping sediments and climate change might all, also have contributed either directly or indirectly to the infilling of the Norsminde basin.

Calcareous foraminifera are almost absent, and the few remains that are found do not shed light on the salinity of the fjord. Mollusc remains also become even sparser in this phase, which probably reflects both poor preservation and poor living conditions, most notably, shallower water, higher sediment loads into the fjord and perhaps coarser sediments. An increased number of epipsammic diatoms, such as *O. minuta* and *O. mutabilis* (Sabbe and Vyverman, 1995) might also suggest a coarsening of sediments, which is consistent with shallower water and likely more physical turbulence and wave action. Despite this, there is little change in the molluscan faunal composition, compared to the previous phase. The presence of *H. reticulata* at ~2,490 BC and *S. plana* at ~2,170 BC, however, both suggest that medium-high salinity existed within the fjord at these dates, which supports the diatom inferred salinity (under both models).

Quercus, Alnus and *Corylus* are again the most abundant pollen taxa, but the latter of these experiences a drop at ~3,070 BC, as another overall shift in the forest structure occurs in this phase. *Betula* decreases, while *Ulmus*, *F. excelsior* and *Tilia* abundance all increase, indicating a change in land-use pattern and forest density. The persistent occurrence of relatively high percentage values for Poaceae and *P. lanceolata*, being joined by *R. acetellosa* after ~2,930 BC, indicates a continued presence of grassland. Arable farming is documented by sporadic occurrences of *Hordeum* type pollen. Between ~2,580-2,530 BC there is a substantial increase in Poaceae (~5-7 %) and *P. lanceolata*. At the same time, there is a distinct drop in the percentage of arboreal pollen (~10 % decrease), which signifies a marked and short term episode of deforestation. High concentrations of charcoal are present in the plant macrofossil record in this phase (particularly between 3,000-2,530 BC), which might reflect forest clearance (via burning) for agricultural purposes.

However, increased charcoal concentrations correspond with increased minerogenic fluxes into the fjord, which rather suggests that the likely cause is inwash from the catchment, as (see Figure 4.13). Therefore, care must be taken with the interpretation of the charcoal record, as this is not a direct proxy for fire, but could rather represent catchment erosion, changes in input loads or source area. There is no direct evidence for air-borne charcoal input (e.g. no correspondence exists between the non-arboreal pollen curve and the charcoal record at Norsminde), which acts as a more direct proxy for fire, and therefore it can only be speculated that the contemporaneous high charcoal concentrations and vegetation signals of human-induced forest clearance and land-use change are related. Aquatic plants increase again in this phase, providing more evidence for infilling and shallowing of the fjord, enabling species such as *Chara* spp. (Figure 4.11) and *Ruppia* spp. (Figure 4.10, Figure 4.11) to spread across large areas of the fjord. After ~1,700 BC, Norsminde Fjord became too shallow to accommodate any regular (stratigraphic) accumulation pattern and since then sediments have been continuously disturbed, thereby preventing palaeoenvironmental analysis of the late Holocene at this site.

4.12 Summary and conclusions

- Norsminde was first inundated around ~6,900 BC, with a lake or stream existing at the coring point prior to this date. Salinity rises rapidly in the succeeding phase, reaching DI-salinities values of above 20 ‰ by ~6,600 BC. Exaggerated abundances of wet ground and swamp and fen taxa in the pollen and plant macrofossil records probably indicated catchment erosion and inwash from surrounding wetlands immediately surrounding the fjord as sea level rise claims low lying lands.
- A further increase in salinity is apparent in the reduced DI-salinity model between ~5,800-5,100 BC, which broadly coincides with rapid sea level rise in other parts of the Baltic basin and possibly global sea level rise. Highest salinities for the Mesolithic period are inferred in the period between ~5,400-4,850 BC (i.e. in the DI-reduced salinity model, foraminifera and molluscan records).
- Salinity appears to drop again by ~4,800 BC and thereafter, both DI-salinity models suggest salinity fluctuates (generally between ~17-22 ‰ in the full model, with the exception of two pronounced peaks between ~3,000-2,900 BC and 2,700-2,570 BC) throughout the remainder of the Norsminde Fjord sequence. Fluctuating salinity is particularly apparent in the full model, due to the high and variable abundances of *Fragilara* s.l. taxa, which are opportunistic species, able to survive in unstable habitats (i.e. subject to constant changes in important environmental parameters such as

salinity and nutrient status). This is likely to have implications for the accuracy of DI-salinity reconstructions in some sections of the sequence.

- Decreased rates of sedimentary accumulation appear to coincide with maximum sea levels (i.e. lowest rates in the period ~6,000-3,900 BC), suggesting that sea level exerts an important control on the sedimentary regime at Norsminde, prior to human disturbance of the landscape (after ~3,900 BC).
- Multivariate 'moving window analyses' (using CCA and RDA techniques) suggest that climate, sea level rise and associated processes/responses (e.g. minerogenic input from both marine resources and from the catchment as land is claimed) significantly explain some of the variation in the diatom dataset in the bottom section of the profile (i.e. 6,760-2,900 BC).
- The first indicators of large-scale agricultural impact upon the landscape appear in the pollen record at ~3,700-3,600 BC (increase in Poaceae pollen, beginning of a continuous curve for *Plantago lanceolata* and appearance of cereal crops). Multivariate analysis (CCA and RDA) suggests that human-related catchment changes (e.g. *Plantago lanceolata*, *Corylus avellana*, *Tilia*, Poaceae and charcoal) are important in explaining a significant proportion of the variation in the diatom dataset in the 'moving window' analyses covering the time period ~4,930-2,020 BC.
- DCA results for the full diatom dataset suggests that DCA axis 1 is driven by salinity, but this changes to depth and when the low-salinity samples at the bottom of the profile are removed (i.e. samples between ~7,000-6,600 BC). For the pollen dataset, PCA (linear data) axis 1 appears to be predominately driven by a light-shade gradient, with shade-giving trees dominating in the lowermost section of the record (i.e. ~6,900 -3,900 BC), while higher abundances of light-demanding plants are present in the record as the catchment become culturally disturbed.
- After ~3000 BC, benthic diatom taxa increase in abundance and submerged macrophyte taxa (most notably *Ruppia* spp and *Chara* spp.) reappear in the pollen and plant macrofossil records as the fjord begins to shallow due to sedimentary infilling and declining sea level. This infilling accelerates after ~2,800 BC, following a large increase in the accumulation rate, most likely due to land-use change and subsequent catchment soil erosion (supported by increase in charcoal and pre-Quaternary foraminifera from the catchment). Predictors of shallower water conditions (e.g. *Ruppia* spp.) also appear to explain some of the variation in the diatom dataset in the uppermost section (i.e. moving window analyses covering the period 4,160-2,020 BC) of the Norsminde Fjord record.

Chapter 5. Results: Korup Sø

5.1 Introduction and aims

This section presents the results from a multiproxy palaeoenvironmental investigation at Korup Sø, aimed at reconstructing changes in both the terrestrial and marine environment of the Djursland area over the early-mid Holocene (~7,600-2,200 BC) years. The sedimentary analyses described in this chapter include core lithology and sedimentation history and palaeoecological analyses include diatoms (analysed as part of this thesis), pollen (H. Krog, unpublished data) and molluscs (Petersen, 1993). Salinity has been reconstructed quantitatively using both a diatom-based salinity transfer function (see section 2.10.2) and a weighted-average minimum salinity molluscan transfer function. This chapter specifically aims to address the following questions:

1. How was chronological control achieved at Korup Sø in order to be able to consider changes in the environmental record in their proper context?
2. How does the sedimentary environment at Korup Sø change over the mid-Holocene as reflected by the sedimentation rates?
3. What are the main changes in the structure and composition of the marine biological communities (diatoms and molluscs) present at Korup Sø over the mid-Holocene?
4. How does the salinity at Korup Sø change over the mid-Holocene as inferred by diatoms and molluscs?
5. How does vegetation of the surrounding area change over the mid-Holocene?
6. To what extent do changes in the terrestrial (e.g. vegetation, land-use) and marine (e.g. sea level change, accumulation rates, productivity) environments affect the diatom community?
7. What are the most significant variables driving the changes in the marine biological communities and do these variables change throughout different sections of the record?

5.2 Core Lithology

Descriptions of the lithology were made by Harald Krog and Bent Odgaard (Geological Survey of Denmark, now Geological Survey of Denmark and Greenland). These original descriptions (for Well no. 1) are presented in Table 5.1 and a summarised lithology column is provided in Figure 5.1. Due to drainage and modern agricultural use of the Korup Sø area, the last ~4,200 years are not present in the sediment record. The top 50 cm of the record are disturbed by drying up of the former basin and agricultural activities. Below this, the majority of the profile is marine clay gyttja, with varying amount of shells (occasional layers) and other macrofossils. Between 727-753 cm (5,510-5,940 BC), the sediment was particularly rich in shells (notably *Ostrea edulis, Mytilus edulis* and Barnacle spp.) exhibiting characteristics similar to a natural shell bank. Slight colour changes occur towards the bottom of the profile and between 753-849 cm (5,940-7,080 BC), small dark lumps of peat material were present. There is eventually a shift to a limnic white-grey calcareous gyttja at 865 cm (~7350 BC), which was deposited before the Korup Sø was inundated by the sea (see also Iversen, 1937).

Depth (cm)	Lithological description
0-50 cm	Disturbed
50-727 cm	Olive-grey-green, clay-gyttja with shells of *Cerastoderma edule* and *Mytilus edulis*. Scattered *Ruppia* spp. fruit-stalks until 150 cm depth.
727-753 cm	Olive-grey, clay-gyttja with more clay and sand than in the layer above. Rich in shells and has the character of a natural shell bank with shells of among others *Ostrea edulis* and *Mytilus edulis* plus barnacles.
753-846 cm	Predominantly dark olive-brown, clay-gyttja with a relatively high content of peat with woody material. *Cerastoderma* shells until 765 cm depth. Shell layer between 842 and 846 cm depth. Lower boundary slanting and irregular.
846-849 cm	Dark olive-grey, clay-gyttja with dark lumps of peat material like in the layer above. Shells of *Cerastoderma edule* and *Mytilus edulis*. Lower boundary irregular and gradual (not sharp).
849-853 cm	Dark olive-grey, clay-gyttja with few shells. Lower boundary gradual.
853-856 cm	Dark olive-green, clay-gyttja, with scattered small snails.
856-865 cm	Light olive-green, clay-gyttja with scattered small snails. Lower boundary sharp.
865-881 cm	Light white-grey calcareous gyttja.

Table 5.1 - Original lithological desciptions of the sediments from well no. 1 Korup So, (H. Krog and B. Odgaard, Geological Survey of Denmark, now Geological Survey of Denmark and Greenland).

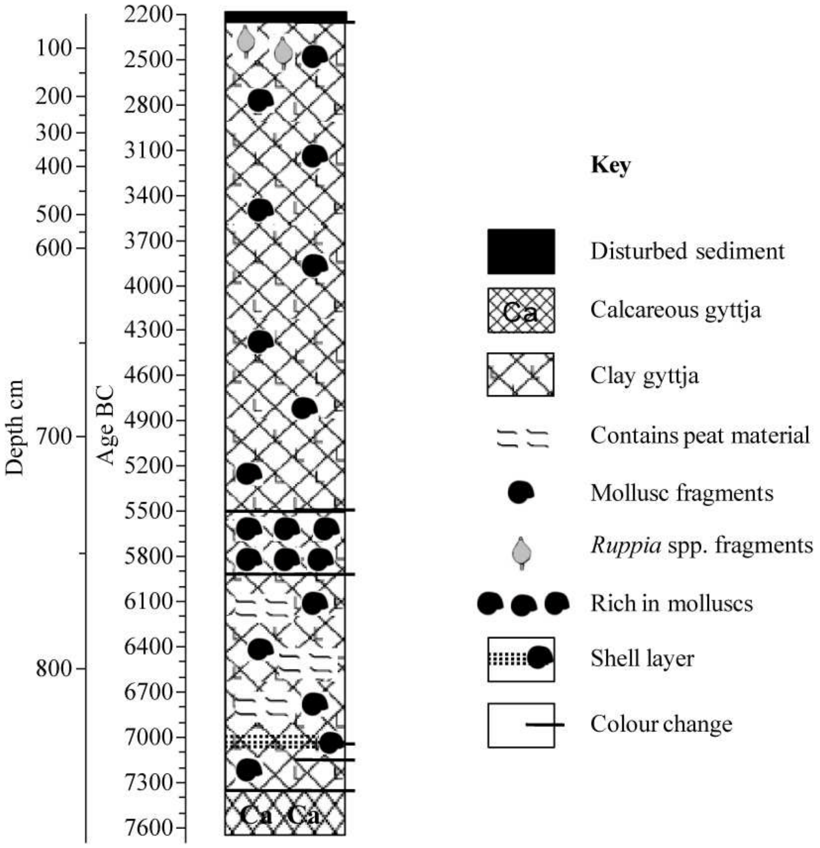


Figure 5.1 - Summary of Korup Sø lithological profile (based on original description by Harald Krog and Bent Odgaard (Geological Survey of Denmark, now Geological Survey of Denmark and Greenland). For colour changes, see Table 5.1.

5.3 Age Model

Details of all radiocarbon dates obtained at Korup Sø are provided in Table 5.2. Twelve shell samples (5 from well no. 1 and 7 from well no. 2) were dated at Copenhagen Carbon-14 Dating Laboratory and results are reported in (Petersen, 1993). Four additional samples were dated more recently from Well no. 1 at the AMS ^{14}C Dating Centre at Aarhus University (Table 5.2) using methods described in 2.5. The final Well no. 1 Age model (produced by J. Olsen, Aarhus University; Figure 5.2) includes 8 (out of the 9) ^{14}C dates. Date AAR-4569 (645 cm depth, *Hinia reticulata*) was excluded from the model because of a low agreement index with the applied model (J. Olsen pers. comm.). The age-depth models for core 1982 and core 1983 are constructed using depositional models in OxCal 4.1 (Ramsey, 2008, 2009). The ^{14}C samples are calibrated to calendar years using the IntCal04 calibration curve (Reimer *et al.*, 2004) with a ^{14}C reservoir age correction of 400±50 yrs. The availability of pollen records for both Korup Sø wells (H. Krog unpub.) enabled eight unambiguous (i.e. clearly visible in both profiles) pollen makers Table 5.3 to be identified. These pollen markers have been built into the OxCal depositional model as additional chronological constrains on the age model. The depositional model has been constructed using a k value of 10 yielding model agreement indices A_{model} of 58.2% (J. Olsen pers. comm.).

All lithological and biological (diatom, molluscan and pollen) analyses described below are based on sedimentary material from Well no. 1, but the Well no. 2 age model is also presented (Figure 5.2, Table 5.2) here as it provides spatial information about the sedimentary environment at Korup Sø over the study period (~7,500 -2,200 BC). The Well no. 2 age model is based on 7 AMS ^{14}C dates.

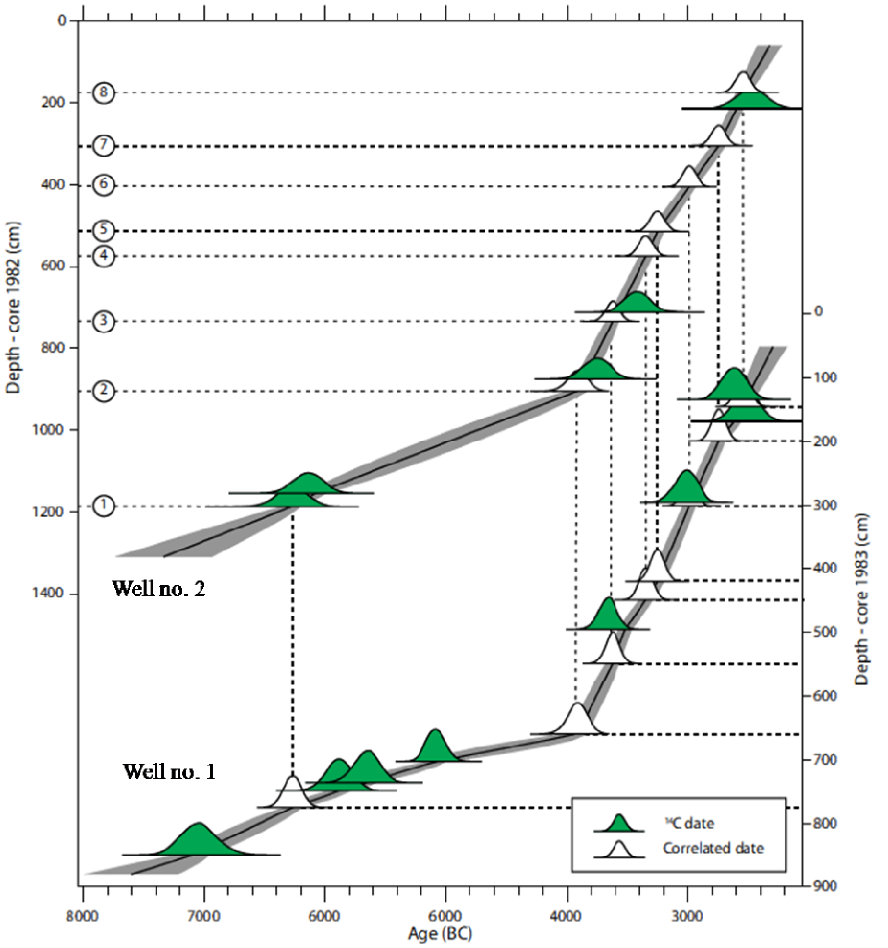


Figure 5.2 - Korup Sø age models. Correlated dates refer to the stratigraphic pollen links (Table 5.3) identifiable in both wells and incorporated into the depostional models (J. Oslen, P. Rasmussen, unpub.). Analyses performed as part of this thesis are all from core-1983 sediments (well no. 1).

Laboratory Code	Depth cm	Material dated (marine molluscs)	Age* ^{14}C BP	Age (cal. BC)	δ13C ‰ VPDB	δ18O ‰ VPDB
Well No. 1	Korup Sø 1983					
K-4233	134	*Cardium edule*	4369 ± 65	2,644-2,395 (95.4 %)	-3.3	
K-4234	168	*Cardium edule*	4290 ± 65	2,736-2,492 (95.4 %)	-2.7	
AAR-4567	295	*Cerastoderma* sp.	4275 ± 50	3,082-2,867 (95.4 %)	-1.3	-2.7
AAR-4568	495	*Cerastoderma edule*	4840 ± 50	3,630-3,405 (95.4 %)	-0.3	-2.3
AAR-4569	645	*Hinia reticulata*	5325 ± 50	Excluded	-1.0	-1.2
AAR-4570	702.5	*Mytilus edulis*	6140 ± 55	5,226-4,921 (95.4 %)	-1.0	-1.0
K-4235	736	*Mytilus edulis*	7161 ± 100	5,845-5,480 (95.4 %)	-1.3	
K-4236	749	*Ostrea edulis*	7402 ± 100	6,057-5,686 (95.4 %)	-0.4	
K-4237	849.5	*Mytilus edulis, Cardium edule*	8401 ± 110	7,355-6,809 (95.4 %)	-1.6	
Well No. 2	Korup Sø 1982					
K-3681	212.5	*Cardium edule*	4271 ± 85	2,708-2,478 (95.4 %)	-1.0	
K-3984	215	*Cardium edule*	4259 ± 85	2,712-2,582 (95.4 %)	-3.6	
K-3986	712.5	*Mytilus edulis, Cardium edule*	5012 ± 90	3,690-3,459 (95.4 %)	-0.6	
K-3987	875	*Mytilus edulis, Cardium edule, Ostrea edulis, Macoma baltica*	5321 ± 90	4,010-3,702 (95.4 %)	-1.0	
K-3988	1155	*Corbula gibba*	7652 ± 110	6,254-5,900 (95.4 %)	-0.3	
K-3989	1187.5	*Cardium edule*	7779 ± 110	6,426-6,125 (95.4 %)	-3.2	
K-3990	1187.5	*Corbula gibba*	7790 ± 110	6,418-6,120 (95.4 %)	-2.5	

Table 5.2 - Details of ^{14}C dated samples from the two Korup Sø boreholes (well no. 1 & well no.2). *The dating results are reported according to international convention (Stuiver and Polach, 1977) as conventional ^{14}C dates in ^{14}C yr BP (before AD 1950) based on the measured ^{14}C/^{13}C ratio corrected for the natural isotopic fractionation by normalizing the result to the standard δ^{13}C value of -25‰ VPDB Andersen *et al.*, 1989).

Pollen link	Core 1982 Depth (cm)	Core 1983 Depth (cm)	Pollen event
8	175	145	Late Hystrix/Dinoflagellate peak
7	305	199	Upper *Corylus* peak/*Quercus* dip
6	405	301	Lower *Corylus* peak/*Quercus* dip
5	515	420	Second *Plantago lanceolata* peak
4	575	449	Major Hystrix/Dinoflagellate peak
3	735	549	First *Plantago lanceolata* peak
2	905	660	*Ulmus* peak (before decline)
1	1155	775	*Pinus* peak (before decline)

Table 5.3 – Unambiguous pollen stratigraphic marker levels incorporated into the depositional age model (J. Olsen, P. Rasmussen, unpublished data).

5.4 Sediment accumulation rate

The Korup Sø sedimentation rates for both wells are shown in Figure 5.3. The predominant pattern for both wells is low accumulation of sediments up until ~3,900 BC, at which point the rate increases sharply and remains high for the upper section of each profile. Superimposed upon this general pattern are a number of smaller shifts, which are described below for Well no. 1 only (analysed as part of this thesis). Between ~9,000-3,900 BC (Mesolithic or Atlantic section of the record), the Well no.1 accumulation rate ranges between ~0.04-0.09 cm yr^{-1} (average 0.07 cm yr^{-1}), with highest rates (i.e. ~0.09 cm yr-^{1}) occurring between ~7250-6,300 BC. Lowest values for the entire Well no. 1 profile occur in the period between ~5,000 - 3,900 BC.

The accumulation rate in the Neolithic (i.e. post ~3,900 BC) period is far more variable, though this may be explained to some extent by the higher number of dating/tie points for this period. Highest accumulation rates (~0.5 cm yr^{-1}) for the whole Well no. 1 profile occurs between ~3,600-3,500 BC, after which there is relatively sharp decrease down to ~0.25-0.3 cm yr^{-1} (up until ~3,200 BC). Between ~3,200 BC-2,800 BC the accumulation rate is ~0.40-0.45 cm yr^{-1}, prior to a short term drop (~300 years) down to its lowest (~0.23 cm yr^{-1}) Neolithic/Subboreal level (i.e. post ~3,900 BC). Another minor rise occurs in the uppermost section of the profile (0.37 cm yr^{-1}; 2,500-2,300 BC). The overall pattern of sedimentation rate is very similar for both wells situated ~300 m apart, but absolute sedimentation rates are higher in Well no. 2.

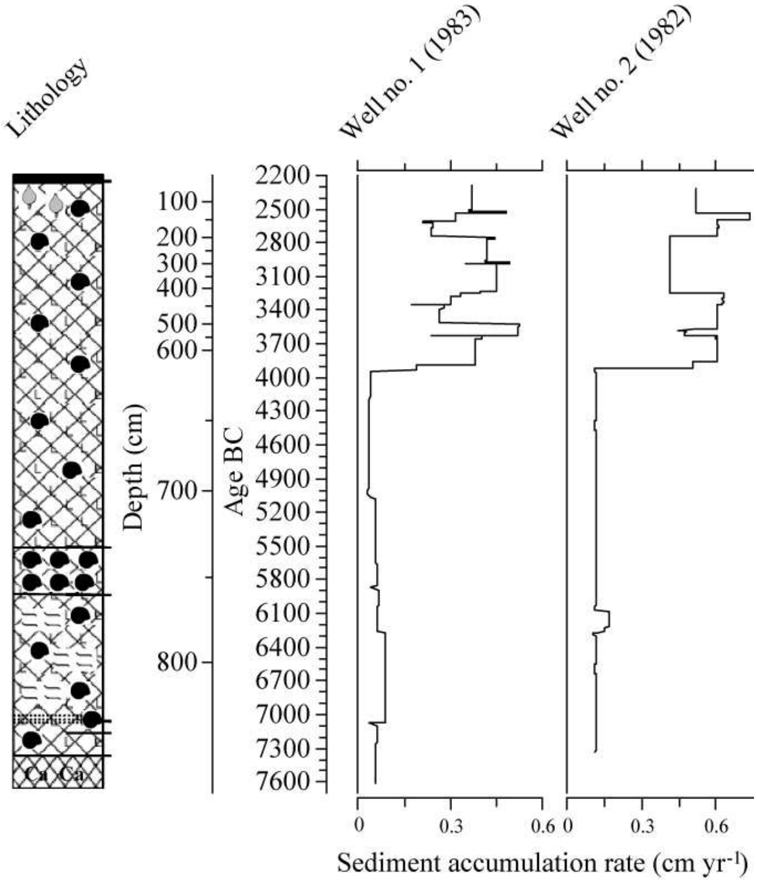


Figure 5.3 – Sediment accumulation rates at Korup Sø (for both the1982 and 1983 wells). Lithology column included for the 1983 well (see Table 5.1 and Figure 5.1 for key).

5.5 Biological Analyses

5.5.1 Diatoms

A total of 52 diatom slides were counted spanning ~2200-7,500 BC at an average temporal resolution of ~100 years with increased resolution over the Mesolithic-Neolithic transition and the period of the Pitted Ware and Single Grave Culture see Table 2.2). A total of 201 taxa were identifiable to species level (from 71 genera), with 31 species having an abundance of >3% in one or more slides. The Korup Sø diatom record (>3 % abundance) for is presented in Figure 5.4 together with a diatom-inferred quantitative salinity reconstruction. Figure 5.5 includes a range of associated metrics, including diatom concentration, diatom dissolution (**F** index), benthic: planktonic: tychoplanktonic and centric: pennate ratios, percentage of taxa incorporated into the salinity model and species richness (Hills N_2 diversity). Three statistically significant diatom assemblage zones have been identified (using taxa >3 %;

Figure 5.4) using optimal splitting, together with one further sub-zone (added for ease of interpretation).

KRP-D1 (~7,330-6,930 BC) – *Fragilaria* sensu latu (s.l.) dominate the zone, with *Pseudostaurosira brevistriata* (~60 %) the most abundant species at ~7,330 BC, but accompanied by smaller numbers of *Staurosira construens* fo *venter* and *Staurosira elliptica* agg. The presence of *Chaetoceros* resting cysts and the occasional valves of *Paralia sulcata* and *Dimeregramma minor* indicate some degree of saline water input into the system at this time. *P. brevistriata* quickly disappears and the upper sample of KRP-D1 (~7,180 BC) is dominated by *S. construens* (*venter* and *subsalina* varieties), *Staurosirella pinnata* and *Rhopalodia constricta*, yielding a DI-inferred salinity of ~9 ‰. Very few planktonic species (e.g. low centric:pennate ratio and high B:P:T ratio) are present and diatom concentration and diversity are low.

KRP-D2 (~6930-5830 BC) – Diatoms indicating higher salinity (e.g. *Paralia sulcata, D. minor*) become more abundant, and overall diversity and concentration increase. Benthic diatoms still dominate with high percentages of *Fragilaria* s.l. (*S. pinnata* & *S. elliptica* agg.) and *Opephora* (*mutabilis* and *minuta* agg.) species present. Planktonic diatoms increase in abundance, however, around ~6,300 BC, with an increase in *Chaetoceros* cysts and a peak of *Cyclotella striata* at ~6,270 BC. Small amounts of the high nutrient favouring diatom, *Cyclotella choctwatcheeana* are also present between ~5,970-5,930 BC. A drop in DI-salinity occurs at ~6,000 BC, driven mainly by the decrease in *P. sulcata* and *D. minor* and the increase of *Fragilaria* S.L. (*S. pinnata, S. elliptica* agg., *O. mutabilis*, and *.O minuta* agg.).

KRP-D3a (~5830-5490 BC) - Salinity reaches it highest levels (between 19-27 ‰) in this zone, driven by the highest percentages of the marine diatom *P. sulcata* (salinity optima 19.6 ‰ in training set), for the entire profile. *S. pinnata* (~13-26 %) and *S. elliptica* agg. (~5-10 %) are also abundant, but both *O. minuta* and *O. mutabilis* decline. Small amounts of D. minor are also present in the benthos. *Chaetoceros* cysts are the most abundant of the true planktonics (~3-8 %), as *C. choctawatcheeana* disappears. Overall diatom concentration and diversity decreases, whereas the centric: pennate ratio and tychplanktonic abundances reaches their highest levels (due to the high percentage of *P. sulcata* valves).

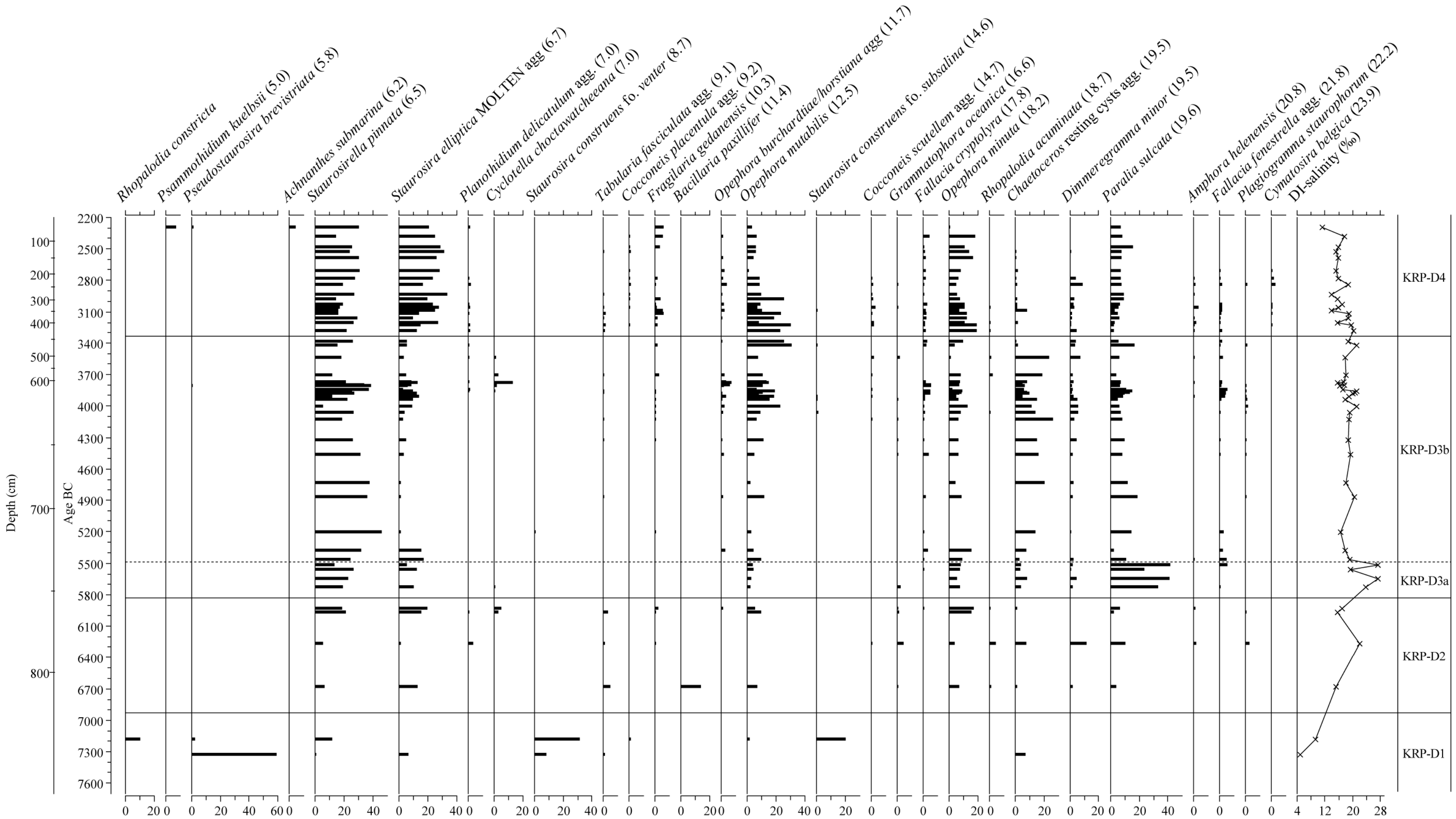


Figure 5.4 - Korup Sø percentage diatom record (>3% abundance). Taxa sorted by salinity optima in diatom training set (values quoted in brackets). Error bars for DI-salinity reconstruction plotted in Figure 5.5.

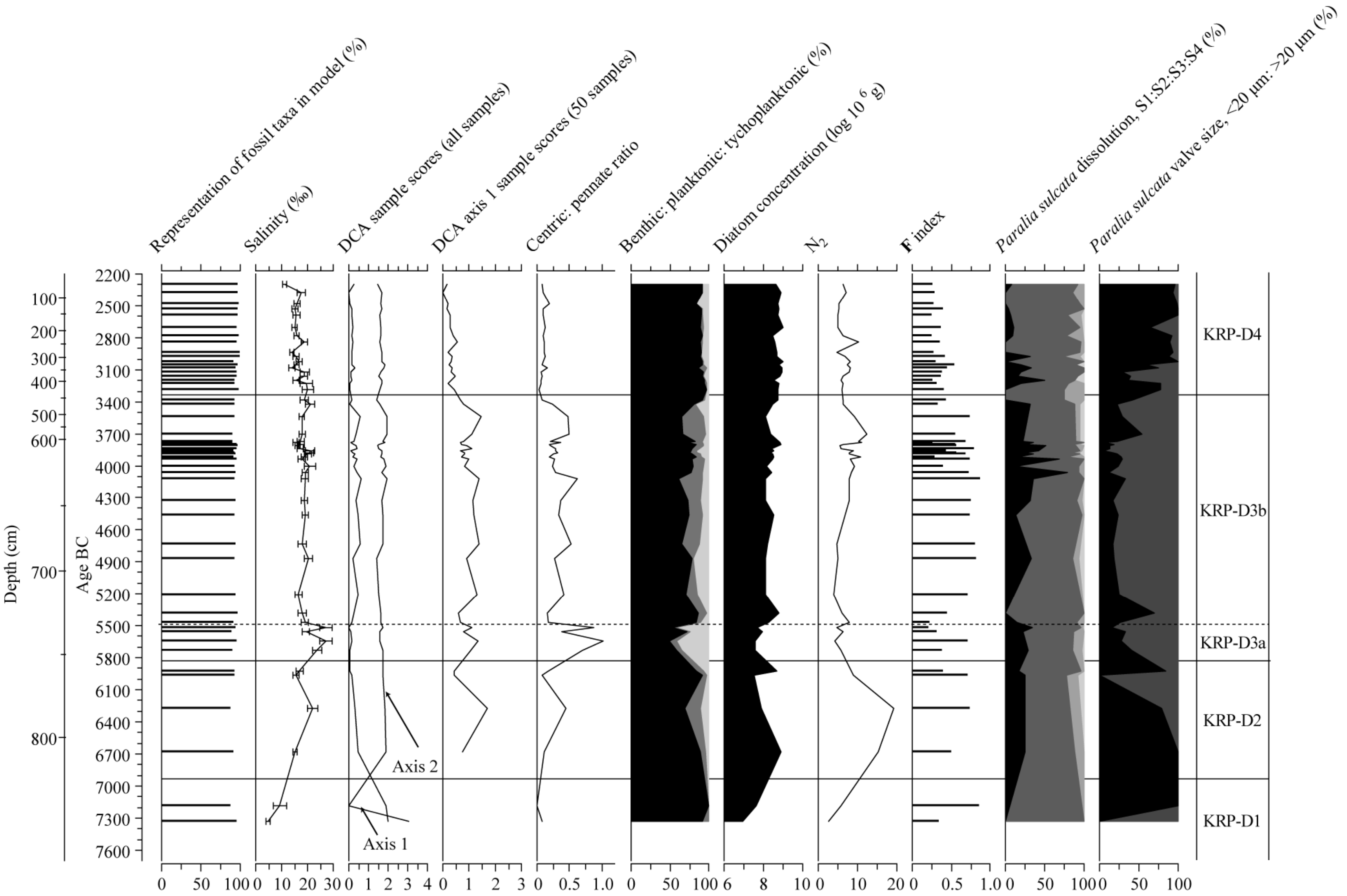


Figure 5.5 - Diatom associated metrics.

KRP-D3b (5,490-3,330 BC) – *P. sulcata* drop markedly at the beginning of KRP-D3b and with it the DI-salinity also falls, thereafter fluctuating between 18-20 ‰ for the majority of the zone. One exception to this is a short phase between ~3,840-3,770 BC, where it drops to as low as ~16-17 ‰. *S. pinnata* is the generally most abundant species, but percentages do tend to fluctuate. *Chaetoceros* cysts increase at the beginning of this zone (>15 % between ~5,470-4,730 BC) and remain relatively high until ~3,530 BC, after which the decline sharply (down to <2%). *O. mutabilis* and *S. elliptica* agg. both increase after ~4,060 BC. Several species also exhibit minor increases between (4,000-3,800 BC), including *Fallacia fenestrella* agg, *Fallacia cryptolyra, C. choctwatcheeana, S. pinnata, P. sulcata* and *Opephora horstiana* agg. A further increase in *O. mutabilis* and *O. minuta* occurs at ~3,430 BC.

KRP-D4 (3,330-2,290 BC) – There is a further reduction in the planktonic community in this zone as *Chaetoceros* cysts decrease. Benthic diatoms account for ~83-97 % of the assemblage, with high numbers of *Fragilaria* s.l. present throughout (most notably *S. elliptica* agg. and *S. pinnata*, which co-dominate). Marine benthic species such as, *F. fenestrella* agg., *D. minor,* and *A. helenensis* are present in low abundances and begin to disappear (successively) after ~2,780 BC. DI-salinity fluctuates throughout, but overall exhibits a downward trend in this zone, dropping from ~20 ‰ at ~3,280 BC, to ~11 ‰ by the uppermost sample (~2,290 BC). At this point a series of more brackish diatoms appear in the record, most notably *Psammothidium kuelbsii* (~6 %) and *Achnanthes submarina* (~4 %). Superimposed upon this general trend are two DI higher-salinity peaks at ~2,840 BC (19 ‰) and ~2,380 BC (18 ‰), both characterised by a decrease in *Fragilaria* s.l and minor increases of higher salinity-tolerant diatoms (e.g. *Achnanthes brevipes* and *D. minor* at ~2,840 BC, *O. minuta* and *F. cryptolyra* at ~2,380 BC).

Diatom Preservation

Valves preservation is generally good (**F** index > ~0.3 for the majority of the profile and range 0.22-0.86), but **F** index values decrease (i.e. greater dissolution) towards the top of the profile as the fjord gets shallower (0.25-.0.54 after ~3,400 BC, discussed in section 5.7). In all samples, the majority of *Paralia sulcata* valves fall into dissolution categories 1 and 2.

5.5.2 Molluscs

Molluscan assemblages from 65 samples (volume 166-3475 cm^3, 3-30 cm slices) were analysed by Kaj Strand Petersen (Petersen, 1993) spanning the time period ~2,300-7,500 BC with an average temporal resolution per sample of ~81 years (26 – 398 yrs). A total of 25 species were recorded (from 20 different genera), 11 of which were bi-valves and 14 gastropods. The full molluscan record (and supplemented statistics) is shown in Figure 5.6. Supplemented statistics include mollusc concentration, bivalve-gastropod ratio, DCA axis 1 sample scores and a weighted-averaged minimum salinity reconstruction (MI-salinity). Molluscan abundances are expressed as concentrations per L^{-1}, with square root transformation for improved presentation of data (in Figure 5.6). The following section describes the Holocene molluscan assemblages sequentially based on the implementation of statistical zoning, application of a new age model and application of a weighted average-salinity transfer function (see section 2.10). Five statistically significant zones have been determined (via optimal splitting by information content, on all species with >5 % abundance; see section 2.10.1) with three additional zones, added to aid interpretation (additional zones do not affect the significant optimal splits).

KRP-M1 (7,500-7,300 BC) - This zone consists of just 2 samples, both of which are low in concentration and yield a combined total of 2 species; *Cerastoderma edule* (edible cockle), present in both samples and *Omalogyra atomus* (atom snail) at ~7,350 BC. Reconstructed MI-salinity in this zone is low (~6 ‰).

KRP-M2a (7,300-6820 BC) – MI-minimum salinity increases in this zone (~10-14 ‰) as *O. atomus* is replaced by brackish-marine taxa, most notably *Hydrobia ventrosa* and *Hydrobia ulvae*. These *Hydrobia* species reach particularly high concentrations between ~7,060-7,010 BC (*H. ventrosa* ~2200 per L^{-1} at ~7,060 BC and *H. ulvae* ~420 per L^{-1} at ~7,010 BC). Diversity also increases at this time as smaller numbers of *Mytilus edulis, Onoba vitrea, Hinia reticulata, Retusa trunculata* and *Rissoa membrancea* are present in the record. *H. ulvae* remains the most abundant mollusc species in the upper part of the zone, but overall molluscan concentration drops dramatically after ~7,010 BC.

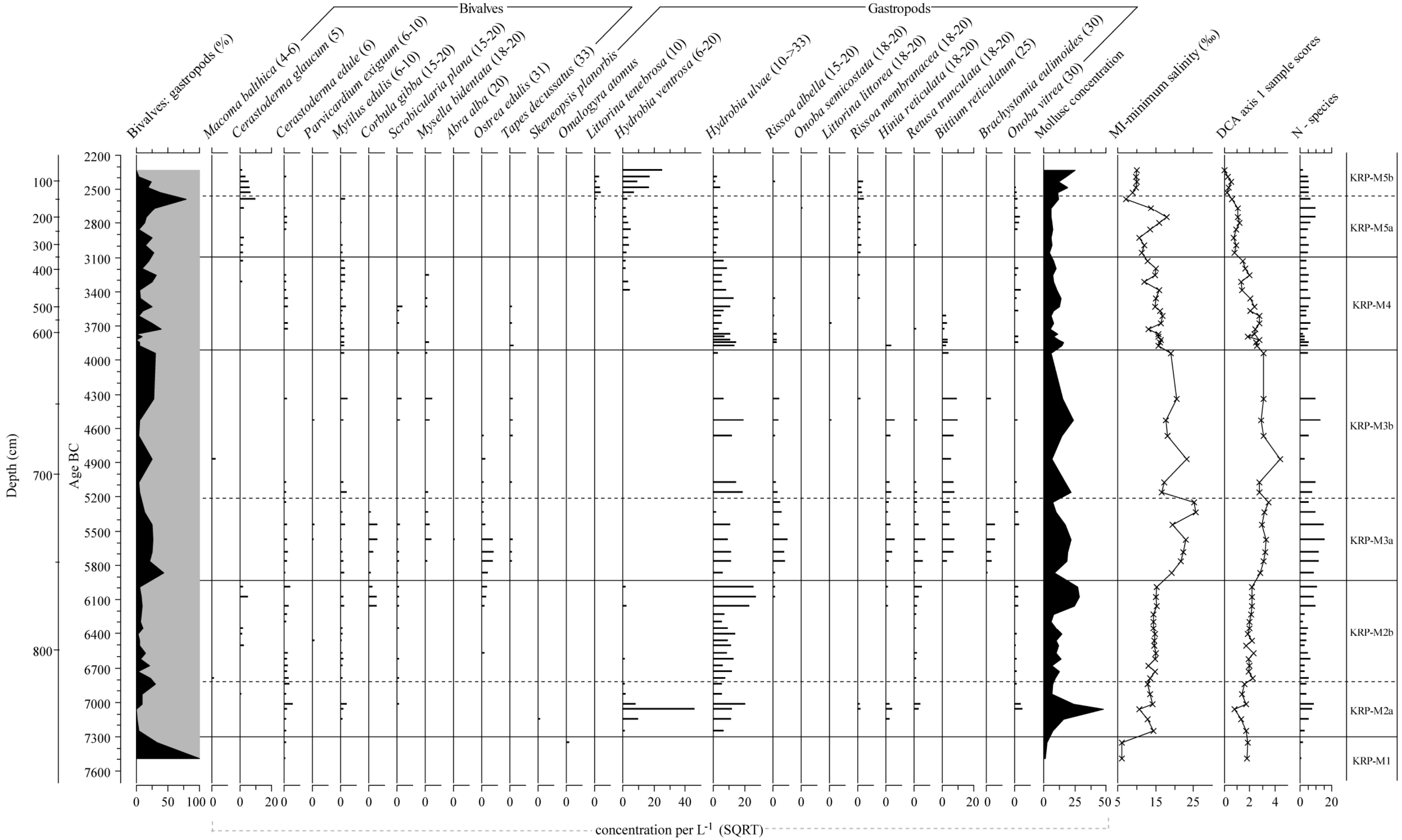


Figure 5.6 Korup Sø molluscan record (from data originally published in Petersen, 1993). Taxa sorted by minimum salinity based on values stated in Sorgenfrei, 1958. Mollusc abundances and total concentration expressed as concentration per L^{-1} (with square root transformation applied to the data for presentation purposes). Dotted lines indicate optimally split sub-zones. It is important to note that sub-zones have added to aid interpretation, and are therefore not necessarily statistically significant splits (see section 2.10.1).

KRP-M2b (6,820-5,930 BC) – Throughout the majority of this zone (~6,800-6,150 BC), molluscan concentration and diversity are low. *H. ulvae* is the most abundant throughout, supplemented by regular appearances from *C. edule, M. edulis, O. vitrea* and *R. trunculata*. A large increase in molluscan concentration occurs at ~6,150 BC, predominantly driven by a ten-fold increase in *H. ulvae* concentration. A number of higher salinity species also appear at this time including *R. trunculata, Ostrea edulis* and *Corbula gibba*. The MI-minimum salinity estimates range between ~13-15 ‰.

KRP-M3a (5,930-5,240 BC) – *O. edulis* and *R. albella* continue to increase, together with the first appearances of a number of other high-salinity demanding species; *Bittium reticulatum, Brachystomia eulimoides* and *Tapes decussates*. *H. ulvae*, remains the most abundant species throughout, and *C. gibba* and *R. truncalata* reach their maximum abundances (for the entire profile) in this zone. MI-minimum salinity is high throughout (~19-26 ‰) reaching its maximum level for the whole record between ~5,430-5,250 BC. Molluscan concentration is variable, rising up until ~5,600 BC, prior to dropping off again in the uppermost section of KRP-M3a.

KRP-M3b (5,240-3,910 BC) – *H. ulvae* concentrations fluctuate intensely, causing both the MI-minimum salinity (16-23 ‰) and molluscan concentration to fluctuate accordingly (i.e. low salinity and high concentration between ~5,160-5,070 BC and 4,670-4,530 BC). *B. reticulatum* is abundant throughout, suggesting relatively high salinity and is supplemented by several sporadic appearances of *H. reticulata* and *R. albella*.

KRP-M4 (3,910-3,100 BC) - *H. ulvae* is again the most abundant species throughout this zone, with *M. edulis*, the only other ever-present, and the rest only making brief or sporadic appearances (e.g. *C. edule, H. ventrosa*). The higher-salinity tolerant species, *B. reticulatum* and *T. decussates* both disappear by ~3,530 BC, but *O. vitrea* remains present throughout. MI- minimum salinities are lower in this zone (fluctuating between ~12-17 ‰) and generally exhibit a decreasing trend. Molluscan diversity and concentration are also relatively low throughout.

KRP-M5a (3,100-2,560 BC) - *H. ventrosa* and *H. ulvae* are the generally most abundant species, with the exception on the uppermost sample (*C. glaucum* most abundant at~2,590 BC). MI-minimum salinity rises between ~2,930-2,750 BC (reaching ~19 ‰ at ~2,750 BC), predominately driven by the reappearance of *O. vitrea* in the record. *R. membranacea* is ever-present and *M. edulis* and *C. glaucum* are present in most samples. *C. edule* also re-appears at ~2,850 BC and *Littorina tenebrosa* (lagoon periwinkle) makes is first appearance at ~2,750 BC. After ~2,600 BC, the MI-minimum salinity drops again, falling as low as ~7 ‰ by ~2,590 BC.

KRP-M5b (2,560-2,330 BC) – Molluscan concentration steadily increases, due mainly to the increasing numbers of *H. ventrosa* (most abundant species in KRP-M5b). *C. glaucum* steadily decreases throughout, and small numbers of *O. vitrea* and *R. membranacea* are present in the lowermost section, but disappear by ~2,480 BC and ~2,430 BC respectively. *L. tenebrosa* and *H. ulvae* disappear slightly later at ~2380 BC. Both salinity and diversity are low throughout, with the MI-minimum salinity initially rising, up to ~10 ‰, where it stabilises for the rest of the zone.

5.5.3 Pollen

A total of 110 pollen samples were counted from Well No.1 at an average resolution of 50 years. Between 737 and 1257 terrestrial pollen grains were counted per sample, containing a total of 89 terrestrial taxa (9 identifiable to family level only, 72 genera with 32 taxa identifiable to species level), 5 swamp and fen taxa (from 5 genera, 2 of which were identifiable to species level a further 7 aquatic taxa (7 genera, 2 identifiable to species level). Besides the pollen and spores *Pediastrum* spp. (freshwater green algae), dinoflagellates and hystrichosphaerids were recorded in the pollen samples. Percentage abundances for these organisms are calculated relative to the pollen sum (see Figure 5.8 caption). Data for these are however expressed independently, relative to the terrestrial pollen sum. A pollen diagram including selected terrestrial taxa is provided in Figure 5.7 and Figure 5.8 includes percentage curves for major plant groupings (as for Norsminde; see section 4.8), PCA sample scores for axis 1 and a summary of the aquatic pollen, and other aquatic organisms (e.g. *Pediastrum*, Hystrix, Dinoflagellates) present in the palynological slides.

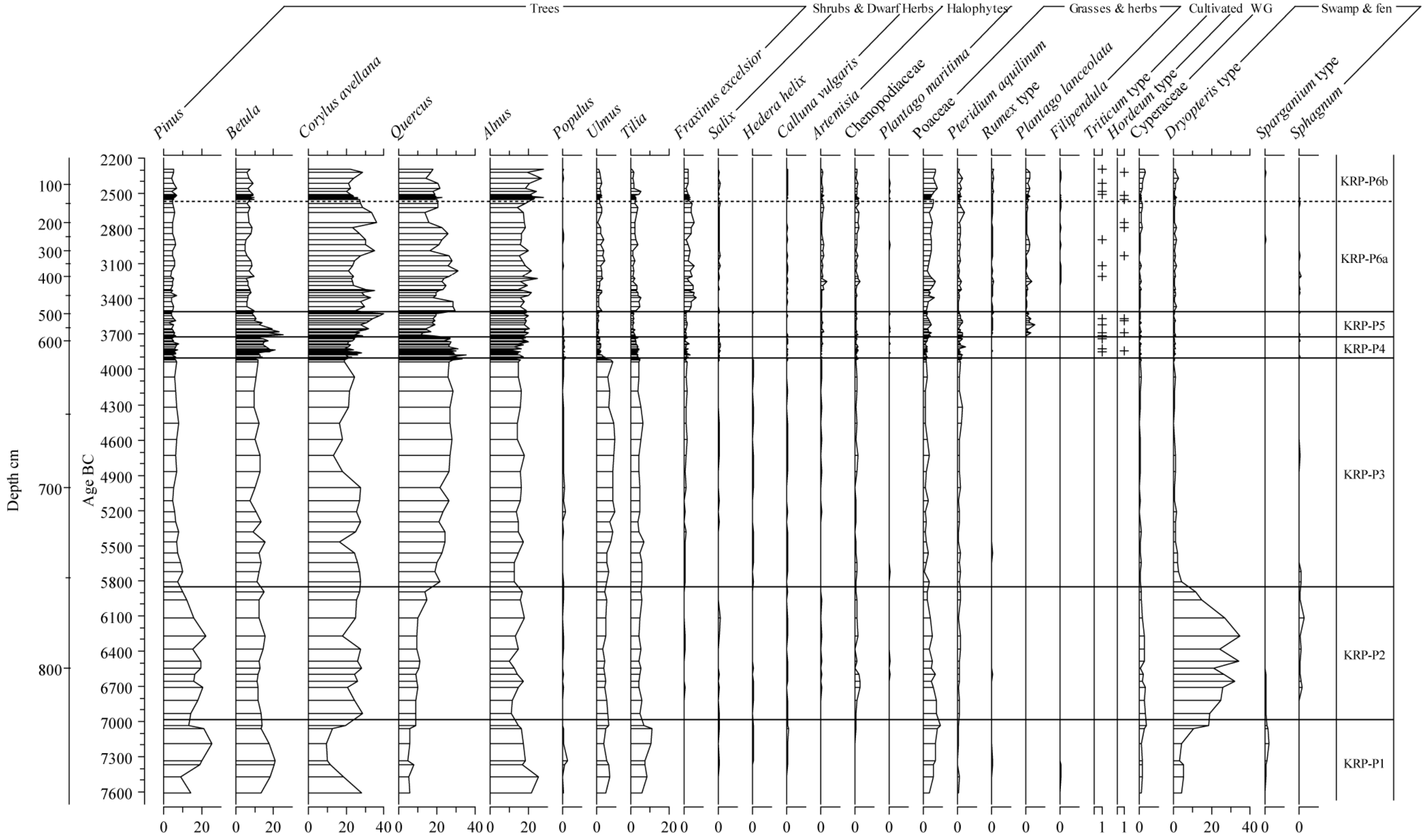


Figure 5.7 - Korup Sø percentage pollen record (selected taxa). Solid zone lines indicate significant zones (determined using the broken stick method; Bennett, 1996). Analyst: Harald Krog, GEUS. (Abbreviations: WG = wet ground). Data are expressed as percentages of the terrestrial pollen sum (TPS) with the exception of cultivated taxa which are expressed as presence/absence data. NB. Swamp and fen taxa are not included in the TPS but are expressed as percentages relative to the TPS.

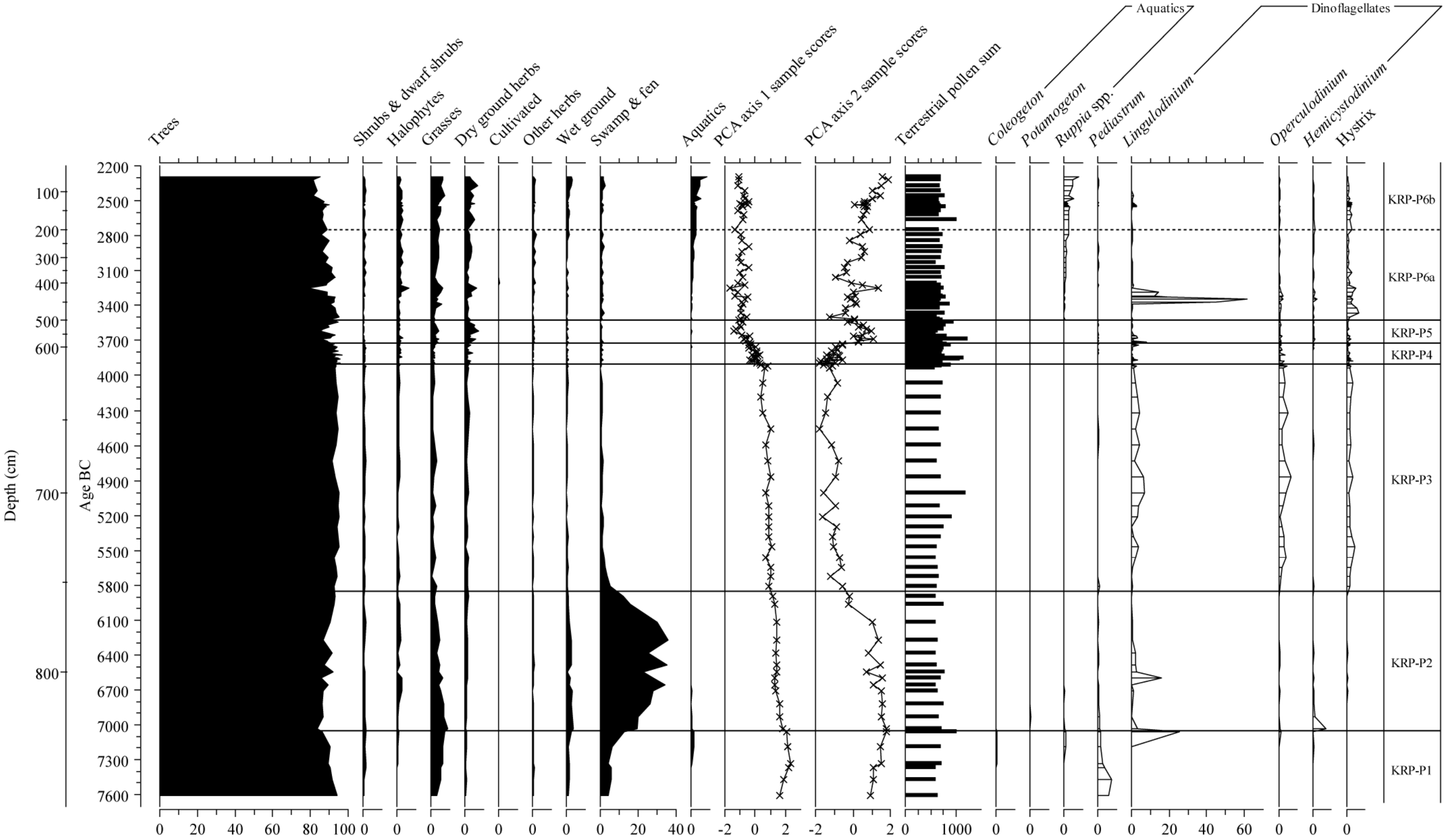


Figure 5.8- Major plant groupings percentage pollen diagram for Korup Sø and aquatic species (including aquatic plants, Pediastrum, dinoflagellates and hystrix; all recorded during palynological analyses and expressed relative to the terrestrial pollen sum. NB. The percentage terrestrial pollen sum is only made up of (all) taxa from the following groups: Trees. Shrubs and herbs, halophytes, grasses, dry ground herbs, cultivated, other herbs and wet ground. Percentage abundances for all other taxa (i.e. swamp and fen, aquatics, Pediastrum and dinoflagellates) are calculated relative to the terrestrial pollen sum.

KRP-P1 (7,600-6,980 BC) – *Corylus* (10-28 %) and *Alnus* (14-25 %) are the most abundant pollen taxa at the bottom of the record, with the latter decreasing throughout and *Corylus* fluctuating. *Pinus* (9-25 %) and *Betula* (13-21 %) also fluctuate and *Tilia* pollen (5-10 %) increases up until ~7,060 BC. *Quercus* and *Ulmus* both independently represent ~5-7 % of the terrestrial pollen sum (TPS) and very small amounts of *Populus* pollen (up to 3 %) are also present. The shrubs, dwarf and terrestrial herbs are represented by small and generally sporadic appearances of *Salix, Hedera helix, Calluna vulgaris* and *Pteridium aquilinum*. A number of species indicative of wet conditions are found in the profile, most notably *Poaceace, Cyperaceae, Dryopteris* and *Sparganium*. *Poaceae* pollen (probably derived from wetland grasses at this time) increases throughout the zone, representing between 4-9% of the TPS. *Pedisastrum* numbers are also initially high, but decline after ~7,470 BC and small amounts of *Ruppia* and *Coleogeton* appear towards the top of the zone (after ~7,330 BC). There is also a *Lingulodinium* spp. (dinoflagellate) peak (~20 %) at ~7,060 BC.

KRP-P2 (6,980-5,850 BC) - *Corylus* is the single most abundant species in zone KRP-P2 increasing substantially at the very start (up to >24%) and then fluctuating throughout. *Alnus* (10-17 %) and *Pinus* (9-22 %) also fluctuate, though the latter of these begin to decreases after ~6,270 BC. There is generally little change in the pollen frequencies of *Betula, Ulmus* and *Tilia*, whereas *Quercus* remains relatively constant up until ~6,120 BC, after which it begins to increase. For the non arboreal taxa, increased percentages of *Chenopodicieae* (up tp ~3 %) and *Pteridium aquilinum* (up to ~2 %) pollen are present in this zone and *Calluna vulgaris* is regularly recorded. High numbers of pollen and spores from local wetlands remain present, particularly *Dryopteris* (11-35 %), Poaceae (2-7 %) and Cyperaceae (1-3 %). In fact, Drypoteris spores increase dramatically between ~7,180 (in KRP-P1) - 6,650 BC (~30 % increase), before declining again in the upper section of KRP-P2. The aquatics species (*Coleogeton* and *Ruppia*) disappear at the start of this zone and *Pediastrums* never exceeds 1 %. There are two distinct dinoflagellate peaks during this zone; *Hemicystodinium* spp. at ~7,035 BC and *Lingulodinium* species at ~6650-6700 BC.

KRP-P3 (5,850-3,900 BC) – *Quercus* (19-20 %) and *Ulmus* (5-9 %) both increase in this zone, the latter of which reaches its highest Holocene frequencies in the period between 5,200-4,500 BC. *Corylus* is initially high (~28 %) but declines after ~5,000 BC, at which *Quercus* takes over as the most abundant pollen type. *Betula* (7-16 %) fluctuates throughout and small numbers of *Fraxinus excelsior* (ash) pollen are also present in the record. There is very little change in the frequencies of *Pinus* (5-10 %), *Alnus* (12-18 %) and *Tilia* (3-7 %) in this zone. For the non-arboreal taxa, *P. aqulinum* and *Chenopodiciae* are ever present and *C. vulgaris* and *Hedera helix* are present in most samples, but always at very low numbers. *Poaceae*, *Cyperaceae* and *Dryopteris* percentages all decrease in abundance and *Sphagnum* disappears (by ~6,600 BC) from the pollen record. The aquatics also disappear, and both hystrix and dinoflagellate frequencies increase.

KRP-P4 (3,900-3,720 BC) - There is substantial change in the arboreal populations in KRP-P4, with increases in *Quercus*, *Betula*, *Corylus* and *Alnus*, synchronous with a very pronounced and rapid *Ulmus* decline (~5 % decrease between ~3,940-3.880 BC. *Pinus* continues its steady decrease and *Tilia* also experiences a small drop in pollen frequencies. The first agricultural signs appear in KRP-P4, as pollen from cultivated cereals such as *Triticum* (wheat) and *Hordeum* (barley) appear in the record for the first time (at ~3,850 BC). Swamp, fen and aquatic representation on the pollen record remains very low and dinoflagellate and hystrix numbers both decrease.

KRP-P5 (3,720-3,510 BC) – There is a large decrease in *Quercus* pollen (~15 % decrease between ~3,740-3,690 BC) at the beginning of this zone, with Betula and to a lesser extent, *Corylus* simultaneously increasing in abundance. This increase is relatively short lived for *Betula*, which declines after ~3,660 BC, and falls by ~13% in the upper section of this zone (3,680-3,550 BC). After its initial drop *Quercus* gradually increases again, along with *Corylus* between ~3,700-3,550 BC. There is little change in the pollen frequencies of the other major tree types (*Alnus* ~18%, *Pinus* ~5%, *Fraxinus* <1%, *Tilia* 2%. and *Ulmus* <1%) or for shrub and dwarf herb species. *Poaceace* (probably now largely from dry grassland species) pollen increases slightly and continuous curves for both *P. lanceolata* and *Rumex* begin in this zone, reflecting the first signs or pastoral farming (Iversen, 1941; Behre, 1981; Gaillard, 2007) in the Djrursland area. Cultivated crops (*Triticum* and *Hordeum*) also remain regularly present in the record. There are no aquatic taxa present in the pollen record and both hystrix and dinoflagellate numbers are very low, despite a slight increase in *Lingulodinium* spp. at the onset of KRP-P5.

KRP-P6a (3,520-2,570 BC) – *Quercus* (13-31 %) and *Corylus* (21-36 %) are the most abundant pollen types and their frequencies fluctuate synchronously in a cyclical fashion (i.e. as one taxa increases, the other simultaneously decreases). *Alnus* percentages remain high

(~17-25 %) and *Tilia*, *Ulmus* and *Fraxinus excelsior* all increase in abundance, the latter of which reaches its highest percentages for the entire profile in this zone. *Pinus* and *Betula* are relatively stable throughout. *Poaceae* (1-6 %) continues to increase along with *Pteridium aquilinum*, and *P. lanceolata* and *Rumex* remain ever-present. Slight increases of *Chenopodiciae* and *Artemisia* also occur after ~3,400 BC, and a number of terrestrial herbs (and cultivated crops) make sporadic appearances throughout this zone, the most regular of these being *Filipendula*. A short lived, but distinct decrease in arboreal pollen, together with a rise in halophyte (*Artemisia*, *Chenopodiaceae*), grass and herb pollen (*Poaceae*, *P. lanceolata*) also occurs between ~3,320-3,230 BC. *Ruppia* reappears in the record in this zone (at ~3,460 BC) and *Pediastrum* is occasionally recorded. Overall, dinoflagellate percentages remain low, with the exception of a peak in *Lingulodinium* spp. between ~3,390-3,250 BC. Increased numbers of hystrix are present between ~3,460-3,120 BC.

KRP-P6b (2,570-2,290 BC) – *Corylus, Quercus* and *Alnus* are again the most abundant pollen types, with *Quercus* and *Alnus* fluctuating throughout and *Corylus* increasing after an initial drop. *Pinus* (~5%) and *Betula* (~8%) remain unchanged from the previous zone and small percentages of Ulmus (1- 4 %), Tilia (1-5 %) and Fraxinus (1-6 %) are also present. The amount of open grassland appears to increases as overall tree pollen falls to as low ~82% of the TPS, with both *Poaceae* (4-7 %) and *P. lanceolata* (1-2 %) pollen frequencies increasing. Additionally, the continued presence of cultivated cereals (*Triticum* and *Hordeum*) indicates the importance of arable farming on the catchment. Towards the top of the profile *Cyperaceae* and *Dryopteris* increase slightly (after ~2,500 BC) and for the aquatics, *Ruppia* continues to rise, reaching its highest level for the whole profile in the uppermost sample (8% at ~2,300 BC). Hystrix numbers experience a slight increase at the very start of the zone, before decreasing again towards the top (after ~2,550 BC) of KRP-P7 and very few dinoflagellates are present.

5.6 Statistical analysis

5.6.1 Trends in species datasets; DCA and PCA analyses

Major trends in the diatom dataset were explored using detrended correspondance analysis (DCA) and the results are summarised in Table 5.4. The variation in the entire diatom dataset captured by the first two DCA axes is relatively even explaining 20.1 % and 15.4 %

respectively. Axis 1 is driven by salinity with the early low-salinity benthic species (e.g. *Epithemia sorex*, *Staurosira construens* varieties, *Rhoicosphenia curvata*, *Pseudostaurosira brevistriata*) characteristic of the lowermost (oldest) samples yielding the highest species scores. The lower/negative end of the gradient is broadly characterised by higher salinity species. However, due to the large magnitude salinity increase associated with the marine transgression (compared to more the subtle shifts in salinity later in the record) and high diatom diversity, but often low abundance (i.e. 201 species in dataset, 37 species at >3% abundance) the individual species scores are noisier (with respect to salinity) and more compressed at the low end of the gradient. As a result there is little change in DCA axis 1 gradient sample scores after KIL-D1 (Figure 5.5).

DCA was also performed on a slightly reduced dataset, excluding the lowermost two samples due to their very different nature to the rest of the profile (i.e. very low salinity). In summary, in this reduced dataset, DCA axis 1 sample scores fluctuate up until ~3,400 BC, after which they fall and remain low for the upper section of the profile. Salinity still appears to be important in explaining part of the variation in the diatom dataset (e.g. $r=0.52$, $p<0.05$ (significant); DCA axis 1 sample scores vs. DI-salinity), with the more brackish taxa present in the uppermost section of the profile (e.g. *Psamothidium kuelbsii*, *Achnanthes submarina*, *Staurosira elliptica* agg. and *Fragilaria gedanensis*) scoring lowest (most negatively) on DCA axis 1 and the majority of other higher-salinity demanding taxa scoring positively (Table 5.5). This axis, however, cannot be explained by salinity alone as some ‘high’ salinity demanding species such as *Cymatosira belgica* and *Opehora minuta* agg. also score low on DCA axis 1. Both *C. belgica* and *O. minuta* are benthic species, which reach their highest abundances in the upper part of the profile (after ~3,400 BC; see section 5.5.1) when the fjord is beginning to shallow, DI-salinity is decreasing and brackish-marine benthic diatoms generally dominate the profile. As *C. belgica* and *O. minuta* agg. diatoms have higher salinity demands, then other processes are also likely to be important on DCA axis 1, most likely sedimentary regime, water depth and indirect depth related processes (e.g. light penetration, sediment accumulation rate, competition for space, sediment disturbance). At the other end of the gradient, highest scoring species (>3 % abundant in one or more slides) are all planktonic (i.e. *Chaetoceros* cysts, *Skeletonema costatum*, and *Cyclotella striata*), rather than being the highest salinity demanding species, further supporting the case that water depth and indirect depth related processes (in addition to salinity) are also important on DCA axis 1.

Dataset	No. of samples	No. of species	Total inertia	Axis	1	2	3	4	Technique
Diatoms (%)	52	242	1.653	Eigenvalues	0.333	0.253	0.12	0.09	DCA
				Length of gradient	1.979	3.042	1.754	1.902	
				% variance explained	20.1	15.4	7.2	5.5	
	50	236	1.104	Eigenvalues	0.167	0.099	0.056	0.032	DCA
				Length of gradient	1.702	1.587	1.124	1.037	
				% variance explained	15.1	9	5	2.9	
Pollen (%)	110	89	1	Eigenvalues	0.381	0.188	0.095	0.050	PCA*
				% variance explained	38.1	18.9	9.5	5	
Molluscs (conc. per L with SQRT tarnsformation)	65	25	1.104	Eigenvalues	0.586	0.178	0.095	0.072	DCA
				Length of gradient	4.426	2.246	1.779	1.684	
				% variance explained	29.9	9.1	4.9	3.6	
Molluscs (conc. per L with SQRT tarnsformation)	63	24	1.789	Eigenvalues	0.586	1.175	0.088	0.072	DCA
				Length of gradient	4.42	2.042	1.586	1.272	
				% variance explained	32.7	9.8	4.9	4	

Table 5.4 - Summary of DCA/PCA results for the full pollen, mollusc and diatoms datasets.

It is difficult to determine precisely what is driving DCA axis 2, but it appears to be at least partly influenced by specific living conditions and dissolution. For example species scoring highest are *P. kuebsii*, *A. submarina*, and *P. sulcata*, all of which are most abundant when **F** index scores are low (0.2-0.4). Furthermore, both *A. submarina* and *P. kuelbsii* are present in the upper sample of the profile, when the fjord is very shallow and sediment input is high, which combined with lower salinity, likely increased turbidity and increased presence of macrophytes (e.g. *Ruppia* spp.) probably produced quite unique habitats and therefore favoured diatoms well adapted to living in these specific conditions. Absolute lowest sample scores occur in the period when *P. sulcata* is most abundant (~5,730-5,510 BC) and DI-salinity is highest, suggesting maximum marine conditions and probably poorer preservation, increased energy and deeper water. Species scoring highest on DCA axis 2 are *Bacillaria paxillifer*, *Nitzschia frustulum* agg. and *T. fasciculata* which are all most abundant at ~6,680 BC, when conditions within the fjord were still rapidly changing (i.e. increasing water level and salinity, low but increasing nutrient status). The ability of these taxa to survive in such a wide range of conditions (e.g. widely euryhaline; Hendey, 1964; Snoeijs *et al.*, 1993-1998), probably enabled them to reach their highest percentages at this time. A relatively high score for *S. costatum* (an indicator of good preservation) on DCA axis 2, also further supports a role for dissolution in explaining the variation on this axis.

DCA was also applied to the square-root transformed mollusc dataset (Table 5.4). The variation explained by DCA axis 1 was relatively high (29.9 %) and therefore axis 1 sample scores (Figure 5.6, Figure 5.7) provide good summaries of the major compositional changes in the mollusc dataset. Species scores for almost all molluscan species plotted along DCA axis 1, which appears to be a salinity driven gradient. Species tolerant of high salinities generally scored higher (e.g. *O. edulis*, *B eulimoides, B. reticulatum*) with lower/negative scores represented by species indicative of more brackish (e.g. *H. ventrosa, L. tenebrosa, C. glaucum*) conditions (Table 5.6). DCA was also performed on a reduced dataset, excluding the lowermost 2 samples. In this case, 32.7 % of the variation was explained by axis 1, but there was very little change in the axis species or sample scores (see Table 5.6) and again a strong salinity gradient was apparent.

PCA axis 1 of the pollen dataset tends to exhibit a strong temporal pattern and is therefore probably defined by the interaction of a number of environmental variables influencing the vegetation (e.g. long-term temperature, edaphic conditions and human impact) rather than representing a single gradient. There appears to be a Mesolithic-Neolithic divide, with species only found in the Mesolithic exhibiting the highest positive PCA axis 1 species scores and *vice versa* (i.e. negative for species only present in the Neolithic). Of the major tree species, *Pinus*, *Tilia*, *Ulmus* and *Poplulus* (more abundant pre ~3,900 BC) exhibit positive axis 1 species scores, whereas *Fraxinus* increases after ~3,500 BC and therefore scores negatively on axis 1. *Alnus* (positive), *Corylus* and *Quercus* (negative) have scores close to zero being similarly abundant both pre- and post- 3,900 BC, whereas *Betula* remains positive despite its sharp increase following the elm decline due to its particularly high abundance in the earliest part of the record. Poaceae also scores weakly negative, being abundant between ~7,600-6,100 and again after ~3,500 BC (in the Neolithic). It is likely that before 6,100 BC, that the high numbers of Poaceae pollen was probably derived from wetland grasses, being washed in following catchment erosion, whereas the second increase is most likely from dryland grasses associated with agricultural land-use changes. Other plants commonly associated with agricultural activities (e.g. *Rumex*-type, *P. lanceolata, Ononis*-type, *Ranunculus acris, Triticum* type and *Hordeum* type) also exhibit negative species scores.

Taxa	**Axis 1**	**Taxa**	**Axis 2**
Cyclotella striata	5.0554	*Bacillaria paxillifer*	4.5473
Skeletonema costatum	2.6895	*Nitzschia frustulum*	4.1442
Chaetoceros cysts agg.	2.5807	*Tabularia fasciculata* agg.	2.5029
Grammatophora oceanica	2.5413	*Skeletonema costatum*	2.418
Rhopalodia acuminata	2.4459	*Rhopalodia acuminate*	2.4033
Plagiogramma staurophorum	2.0523	*Cyclotella choctawatcheeana*	2.1537
Dimmeregrammer minor	1.9333	*Cocconeis scutellem* [*parva + scutellum*]	1.9935
Paralia sulcata	1.5128	*Grammatophora oceanica*	1.9022
Bacillaria paxillifer	1.4949	*Chaetoceros* cysts agg.	1.768
Fallacia fenestrella agg.	1.3374	*Opephora mutabilis*	1.6531
Cyclotella choctawatcheeana	1.0012	*Plagiogramma staurophorum*	1.5343
Staurosirella pinnata	0.9024	*Fragilaria gedanensis*	1.4882
Planothidium delicatulum agg.	0.8008	*Amphora helenensis*	1.3704
Cocconeis scutellem [*parva + scutellum*]	0.5668	*Dimmeregrammer minor*	1.3598
Nitzschia frustulum agg.	0.5445	*Opephora burchardtiae/horstiana* agg.	1.3254
Tabularia fasciculata agg.	0.493	*Opephora minuta* agg.	1.292
Opephora burchardtiae/horstiana agg.	0.4251	*Planothidium delicatulum* agg..	1.1064
Fallacia sp. undiff.	0.3862	*Cyclotella striata*	1.0311
Amphora helenensis	0.3824	*Cymatosira belgica*	0.8415
Opephora mutabilis	0.3337	*Staurosira elliptica* agg.	0.7296
Fallacia cryptolyra	0.251	*Fallacia cryptolyra*	0.3516
Opephora minuta agg.	-0.0218	*Staurosirella pinnata*	0.1054
Fragilaria gedanensis	-0.6719	*Fallacia fenestrella* agg.	-0.1046
Staurosira elliptica agg	-0.8279	*Paralia sulcata*	-0.8848
Cymatosira belgica	-1.3495	*Fallacia* sp. undiff.	-0.9515
Psammothidium kuelbsii	-3.0609	*Psammothidium kuelbsii*	-1.8202
Achnanthes submarina	-3.0638	*Achnanthes submarina*	-1.8525

Table 5.5- Summary of DCA results for the reduced diatom dataset (50 samples)

A further 18.9 % of the variation in the pollen dataset is explained by PCA axis 2 (Table 5.4). This axis is likely to be driven, initially, by inwash and catchment erosion as suggested by exaggerated percentages of Poaceae, Cyperaceae and *Pinus* pollen probably from local wetlands (see section 5.5.3), but later by forest density as catchment erosion is reduced and shade trees such as *Quercus, Corylus* and *Ulmus* become more abundant in the catchment. High PCA axis 2 sample scores at the bottom of the profile (~7,600-6,100 BC; Figure 5.8) are caused by the high, and exaggerated percentages of Cyperaceae and Poaceae pollen (both of which score positively on PCA axis 2; Table 5.7). These taxa give the impression that the Korup Sø catchment was relatively open, but these wetland sedges and grasses probably mainly grew in low-lying areas in the immediate vicinity of the fjord together with some

Betula, *Pinus* and *Alnus*. The rest of the catchment was dominated by a dense forest consisting predominately of *Tilia*, *Corylus* and *Quercus*. The lowest PCA axis 2 scores occur between ~6,000-3,500 BC, when a dense closed-canopy forest dominated by *Tilia*, *Corylus* and *Quercus* (up until ~3,900 BC) covered the Korup Sø catchment. The lowest (most negatively) scoring species (abundant at >1 % in one or more slides) on axis 2 are *Quercus* and *Ulmus*, the latter of which reaches its highest percentages between ~5,560-3,900 BC. *Corylus* is relatively abundant throughout the whole record and therefore scores close to zero.

Taxa	**DCA axis 1 (all samples)**	**DCA axis 1 (63 samples)**
Macoma baltica	6.87	6.86
Abra alba	6.51	6.51
Brachystomia eulimoides	4.96	4.95
Bittium retulatum	4.40	4.39
Ostrea edulis	3.94	3.93
Tapes descussatus	3.82	3.81
Rissoa albella	3.62	3.61
Littorina littorea	3.56	3.55
Parvicardium exiguum	3.50	3.50
Mysella bidentata	3.50	3.49
Corbula gibba	3.24	3.24
Srobicularia plana	2.99	2.98
Hinia reticulata	2.87	2.86
Retusa truncata	2.69	2.68
Omalogyra atomus	2.12	Absent
Hydrobia ulvae	2.06	2.05
Mytilus edulis	1.93	1.92
Cerastoderma edule	1.78	1.78
Onobo vitrea	1.22	1.23
Skeneopsis planorbis	0.86	0.87
Onobo semicostata	0.61	0.61
Rissoa membranacea	0.44	0.44
Cerastoderma glaucum	0.28	0.28
Hydrobia ventrosa	-0.02	-0.02
Littorina tenebrosa	-0.77	-0.77

Table 5.6 - DCA species scores for the molluscan dataset

PCA axis 2 sample scores rise again in the Neolithic, as overall arboreal pollen decreases and elevated percentages of Poaceae (derived from dryland grasses) and other open ground taxa (most notably *P. lanceolata* and *Rumex* type) become present in the record. These taxa all score positively on PCA axis 2, with Poaceae exhibiting the highest score (Table 5.7). Axis 2 sample scores generally increase throughout the Neolithic period, primarily driven by the continually increasing percentages of Poaceae pollen and the increase in wet ground taxa (e.g. Cyperaceae) after ~2,800 BC, the latter probably caused by the expansion of wetlands in the Korup Sø vicinity as the fjord begins to dry up. Superimposed upon this general trend are occasional high scores (e.g. ~3,250 BC) which might represent short-term episodes of forest clearances for agricultural purposes. Fluctuations in arboreal pollen also result (e.g. increase in *Quercus* pollen at ~3,160 BC; ~31 %) in occasional lower sample scores in the post-3,900 BC period.

Taxon	**Axis 1**	**Taxon**	**Axis 2**
Pinus	0.84	Poaceae	0.90
Tilia	0.81	Cyperaceae	0.73
Ulmus	0.76	*Plantago lanceolata*	0.46
Betula	0.55	*Rumex* type	0.39
Populus	0.39	*Pinus*	0.39
Cyperaceae	0.34	*Alnus*	0.26
Salix	0.02	*Artemisia*	0.23
Poaceae	-0.04	Chenopodiaceae	0.13
Pteridium aquilinum	-0.28	*Corylus avellana*	-0.02
Chenopodiaceae	-0.31	*Tilia*	-0.13
Alnus	-0.43	*Salix*	-0.13
Quercus	-0.50	*Fraxinus excelsior*	-0.14
Corylus avellana	-0.53	*Betula*	-0.16
Rumex type	-0.56	*Pteridium aquilinum*	-0.19
Artemisia	-0.69	*Ulmus*	-0.24
Plantago lanceolata	-0.73	*Populus*	-0.40
Fraxinus excelsior	-0.82	*Quercus*	-0.78

Table 5.7 - Summary of PCA axis 1 and 2 species scores for the Korup So pollen record (taxa > 1 % abundance in one or more slides)

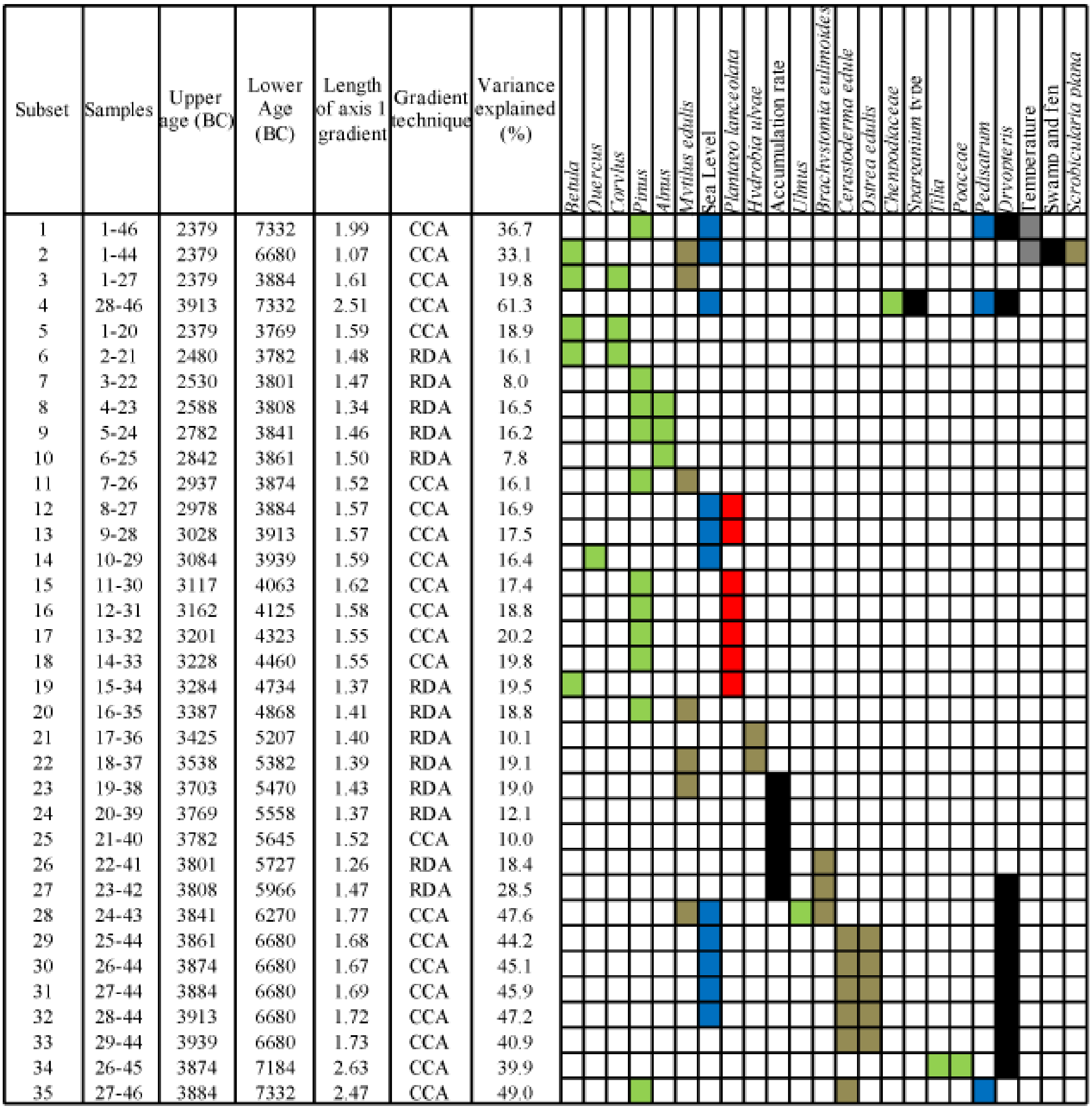
5.6.2 Environmental-Species relationships

Environmental-species response relationships were examined using a top-down approach with diatoms (the lowest trophic level) used as the response dataset. Palynological analyses were used as predictors for vegetation, catchment and land-use changes, and molluscs (and dinoflagellates) for changes in the marine environment and grazing pressure. Independent sea level data (from Blekinge, Sweden; Berglund *et al.*, 2005) and temperature data (pollen-based quantitative temperature reconstruction from Lake Trehörningen, Bohuslän, southwest Sweden; Antonsson and Seppä, 2007) were also included in the environmental dataset. Both of these reconstructions were considered to be good regional estimates for these parameters and applicable to the Danish area (P. Rasmussen, pers, comm.; Brown *et al.*, submitted). Lastly, sedimentation rate data (from Well No. 1, see section 5.3.) was included as another predictor of biotic change in the marine environment. For the predictor dataset, the accumulation rate data was expressed as cm yr^{-1} and sea level as metres above present sea level (as in Berglund *et al.*, 2005). The pollen and mollusc data were expressed as percentages, but rare or sporadically occurring species (i.e. less than four occurrences and/or only sporadically present through the profile) were removed prior to analysis. In addition, temperature was log (base 10) transformed to reduce the relatively long gradient length exhibited by this variable.

Differences in the sampling strategy between proxies meant that a reduced subset of 46 samples was used for predictor-response analyses. Each sample contained pollen, mollusc and diatom data either from the same stratigraphic level or within a maximum of ~25 years of one another (as determined by the age model). Coarser sampling resolution for molluscs (analysed for the entire profile) meant that several pollen and diatom samples often corresponded to one mollusc sample. In these circumstances the time-averaged mollusc samples were duplicated. Rare pollen and mollusc species were eliminated if they occurred less than 4 times in the 46-sample environmental dataset and only at sporadic intervals and low abundances. Sediment age was included as a co-variable to partial out variance explainable by time-related processes (e.g. ontogeny, succession). Moving window analyses were again performed to examine the possible influence of different environmental parameters at different times throughout the profile (after Bradshaw *et al.*, 2005a and see Chapter 4). Subsets of 20 samples were analysed starting at the top of the record and working backwards, eliminating the upper sample and adding a lower sample each successive run (e.g. 1-20, 2-21, 3-22…etc). Each time the datasets were reduced, DCA was re-performed on the species subset for selection of the appropriate ordination technique. Axis 1 gradient lengths ranged between 1.07-2.63 s.d. units and therefore both canonical correspondence analysis (CCA) and redundancy analysis (RDA) techniques were employed (Figure 5.9). For each CCA/RDA analysis, significant variables ($p < 0.05$; with Bonferroni corrections applied for each subsequent selection) were chosen using a forward selection process (999 unrestricted permutations). After this process, CCA/RDA analysis was performed using just the selected variables to test the significance ($p < 0.05$) of the reduced model and time factored out as a covariable.

CCA and RDA results are presented in Figure 5.9. For each subset only significant variables are included, together with the amount of variance explained. The length of the DCA axis 1 gradient together with the type of analysis performed is also included. CCA on the full response-predictor dataset produced a statistically significant model ($p < 0.01$) with five independent predictor variables (*Pediastrum*, *Dryopteris*, *Pinus,* sea level and temperature; discussed in detail below) explaining 36.7 % of the variation in the diatom dataset. In the moving window analyses ~7.8-49 % of variation is explained by the selected variables with all models proving statistically significant ($p < 0.05$). The greatest variation is explained in the lower subsets (subsets 27-35; Figure 5.9), where the system is at it most dynamic and the diatom dataset appears to be responding strongly to changes in the marine environment, particularly salinity, water depth and nutrients, all probably related to rising sea level (discussed in more detail below). Above this (i.e. subsets 5-26; ~5,730-2,380 BC), individually selected environmental predictors generally only explain relatively small amounts (<20%) of the variation with the diatom dataset. In subsets 21-35 (~7,330-3,400 BC) predictors for marine environmental change (e.g. sea level, sedimentation rate, changing mollusc faunas) are generally more common, whereas catchment processes/vegetation changes (probably heavily driven by human disturbance after ~3,900 BC) are more important in the upper section of the record (subsets 5-20; 4,870-2,380 BC).

Sea level is important in explaining some of the variation in the diatom dataset at two separate times throughout the profile (i.e. subsets 28-32; ~6,680-3,840 BC, subsets 12-14; ~3,930-2,980 BC) as well as being a significant proxy for the entire dataset (subsets 1 and 2). The lowermost block of subsets that sea level is important in (i.e. subsets 28-32) covers the most dynamic phase of the *Littorina* Sea (i.e. L1-L3 transgressions; Berglund *et al.*, 2005). Sea level is likely to influence salinity, marine exposure (i.e. currents and waves) catchment erosion and inwash (see below), water depth (and light penetration) and macrophytic vegetation. For example, aquatic taxa (*Ruppia* spp. and *Coleogeton*) practically disappear from the pollen record after ~7,000 BC as sea level (water depth) and salinity increase, altering the benthic habitat and enabling planktonic diatom taxa to increase in abundance (Figure 5.5). Furthermore, the DI-salinity reconstruction appears to be closely connected with sea level rise, in the lowermost section of the sequence (~7,300-5,500 BC). In this period, several floral changes are identifiable in the diatom record, as salinity rises and marine conditions intensify. This also includes a peak of planktonic diatoms at ~6,270 BC possibly related to increasing water depth and maybe higher nutrient status, perhaps being washed in from the catchment as low lying lands were sub-merged. In the uppermost block of subsets (~3,930-2,980 BC), declining sea levels (together with high sedimentation rates) causes the basin to get shallower and salinity to decline (after ~3,400 BC, according to the DI-salinity record). This directly affects the diatom community as benthic diatom taxa increase in abundance and macrophyte vegetation (e.g. *Ruppia* spp.) appear again in the pollen record.



Subset	Samples	Upper age (BC)	Lower Age (BC)	Length of axis 1 gradient	Gradient technique	Variance explained (%)	*Betula*	*Quercus*	*Corylus*	*Pinus*	*Alnus*	*Mytilus edulis*	Sea Level	*Plantago lanceolata*	*Hydrobia ulvae*	Accumulation rate	*Ulmus*	*Brachystomia eulimoides*	*Cerastoderma edule*	*Ostrea edulis*	*Chenopodiaceae*	*Sparganium* type	*Tilia*	*Poaceae*	*Pediastrum*	*Dryopteris*	Temperature	Swamp and fen	*Scrobicularia plana*
1	1-46	2379	7332	1.99	CCA	36.7				🟩			🟦												🟦	⬛	▨		
2	1-44	2379	6680	1.07	CCA	33.1	🟩					🟫	🟦														▨	⬛	🟫
3	1-27	2379	3884	1.61	CCA	19.8	🟩		🟩			🟫																	
4	28-46	3913	7332	2.51	CCA	61.3							🟦								🟩	⬛			🟦	⬛			
5	1-20	2379	3769	1.59	CCA	18.9	🟩		🟩																				
6	2-21	2480	3782	1.48	RDA	16.1	🟩		🟩																				
7	3-22	2530	3801	1.47	RDA	8.0				🟩																			
8	4-23	2588	3808	1.34	RDA	16.5				🟩	🟩																		
9	5-24	2782	3841	1.46	RDA	16.2				🟩	🟩																		
10	6-25	2842	3861	1.50	RDA	7.8					🟩																		
11	7-26	2937	3874	1.52	CCA	16.1				🟩		🟫																	
12	8-27	2978	3884	1.57	CCA	16.9							🟦	🟥															
13	9-28	3028	3913	1.57	CCA	17.5							🟦	🟥															
14	10-29	3084	3939	1.59	CCA	16.4		🟩					🟦																
15	11-30	3117	4063	1.62	CCA	17.4				🟩				🟥															
16	12-31	3162	4125	1.58	CCA	18.8				🟩				🟥															
17	13-32	3201	4323	1.55	CCA	20.2				🟩				🟥															
18	14-33	3228	4460	1.55	CCA	19.8				🟩				🟥															
19	15-34	3284	4734	1.37	RDA	19.5	🟩							🟥															
20	16-35	3387	4868	1.41	RDA	18.8				🟩		🟫																	
21	17-36	3425	5207	1.40	RDA	10.1									🟫														
22	18-37	3538	5382	1.39	RDA	19.1						🟫			🟫														
23	19-38	3703	5470	1.43	RDA	19.0						🟫				⬛													
24	20-39	3769	5558	1.37	RDA	12.1										⬛													
25	21-40	3782	5645	1.52	CCA	10.0										⬛													
26	22-41	3801	5727	1.26	RDA	18.4										⬛		🟫											
27	23-42	3808	5966	1.47	RDA	28.5										⬛		🟫								⬛			
28	24-43	3841	6270	1.77	CCA	47.6						🟫	🟦				🟩	🟫								⬛			
29	25-44	3861	6680	1.68	CCA	44.2							🟦						🟫	🟫						⬛			
30	26-44	3874	6680	1.67	CCA	45.1							🟦						🟫	🟫						⬛			
31	27-44	3884	6680	1.69	CCA	45.9							🟦						🟫	🟫						⬛			
32	28-44	3913	6680	1.72	CCA	47.2							🟦						🟫	🟫						⬛			
33	29-44	3939	6680	1.73	CCA	40.9													🟫	🟫						⬛			
34	26-45	3874	7184	2.63	CCA	39.9																	🟩	🟩		⬛			
35	27-46	3884	7332	2.47	CCA	49.0				🟩									🟫						🟦				

Figure 5.9 - Summary of 'moving window' results from CCA/RDA analyses assessing predictor-response (diatoms) relationships through time. Length of axis 1 gradient refers to DCA analysis on reduced diatom dataset (if DCA axis 1 gradient length > 1.5 s.d. then unimodal CCA employed to explore trends. If DCA axis 1 <1.5. s.d. then linear RDA technique used). Subset 1 = entire dataset (46 samples). Subsets 2-35 = moving window analyses. Shadings: Red = Anthropogenic indicator, Black = sedimentary regime/catchment erosion, Green: vegetation change predictors, Grey = climatic and Blue = fjord depth and level of connection with Kattegat, Brown = molluscs (grazers).

Average temperature has changed by ~3° C (Figure 5.10) over the course of the Korup Sø record, rising in the early Holocene and reaching a 'thermal maximum' between ~5,550-3,550 BC, but beginning to decline again after ~4,000 BC. Changes in temperature can both directly and indirectly affect the diatom community (see section 4.10.2). For example, cooler temperatures (particularly before and/or after the thermal maximum) might have affected the severity of winters and the volume of snow and ice on the catchment. This might have induced large thawing events inputting significant volumes of cold-fresh water into the fjord and subsequently causing short-term fluctuations in both temperature and salinity (e.g. numerous salinity fluctuations are evident in the DI-salinity reconstruction; Figure 5.4). Sea ice might also have developed in extremely cold winters, though there is no evidence for any major sea ice event in the diatom record (e.g. complete absence of sea ice diatoms).

Interestingly, temperature is absent from the reduced 'moving window' subsets at Korup Sø but appears to explain some of the variation when all samples are included (i.e. present in subsets 1 and 2; 2,380-7,330 BC). This is likely to result from temperature more strongly exerting an indirect influence upon the diatom community, by driving changes in other important variables (and at different times) such as hydrology, vegetation structure, edaphic processes and grazing communities (see also Chapter 4; Norsminde). For example, reproduction in grazing organisms such as molluscs and foraminifera is often temperature-dependent (e.g. oysters require 15° C temperatures for spawning, whilst cockles (~9° C) and mussels (9-12° C) can tolerate slightly lower temperatures; Yonge, 1960; Petersen, 2004) and if temperatures fall below the minimum required, the length of the spawning season can be reduced or even prevented entirely (Yonge, 1960; Johannesson *et al.*, 1989). In prolonged cold periods, molluscan and foraminifera populations might become severely reduced, thereby potentially reducing grazing pressure upon diatoms. Alternatively, a change in species composition might occur (i.e. colonisation by a colder water mollusc or foraminiferal fauna) which might alter the habitat for benthic diatoms and/or alter ecosystem structure (e.g. feeding patterns, predators).

It is also possible that the influence of temperature on the diatom communities is purely coincidental, as for example maximum Holocene temperatures (i.e. during the Holocene climatic optimum, ~5,550-3,550 BC; Snowball *et al.*, 2004; Antonsson and Seppä, 2007) broadly coincide with the highest abundances of dense closed canopy tree taxa (e.g. *Quercus*, *Ulmus*) in the pollen record (~6,000-3,900 BC; Figure 5.7). Denser woodland is likely to influence catchment run off rates and inputs of sediment (e.g. lowest accumulation rates broadly coincide with highest temperatures), nutrients and minerals. This is particularly identifiable later in the profile when human impact upon the catchment (i.e. forestry clearances, crop-growing and expansion of agricultural grasslands) becomes more important in driving changes in the diatom record (see below). Summer thermal stratification might also have occurred under higher temperatures (e.g. during the thermal optimum), though this is probably unlikely due to the depth of the fjord (probably relatively shallow even in the late Mesolithic when sea level was higher). The lack of evidence for a halocline (e.g. no freshwater planktonic diatoms present in the profile) suggests that the fjord was probably well mixed at all/most times, though short-term stratification events might have occurred in particularly warm years. Such events might offer an alternative explanation (to water depth or productivity) for the high abundances of *Chaetoceros* cysts between ~5,380-3,540 BC, with cyst production events following spring/summer planktonic blooms when nutrients in the surface layers became depleted.

Pediastrum (present in the lowermost subset and significant for the entire dataset; i.e. subset 1, 46 samples) is relatively abundant (>3 %) in the record between ~7,600-7,360 BC and is a likely predictor for fresh to weakly brackish conditions, prior to the marine transgression (after ~7,400 BC). At this time the diatom record is also dominated by fresh/brackish water diatoms (e.g. *Pseudostaurosira brevistriata*, *Staurosira construens* and some freshwater *Cyclotella* species). *Pediastrum* percentages drop as salinity in the fjord rises, and higher salinity diatoms appear in the record (i.e. <1 % after ~7,000 BC). It reappears again (sporadically and at low abundances) towards the top of the record (after 3,100 BC) as the salinity in the fjord decreases. *Cerastoderma edule* is also important in the lowermost subsets (and in subsets 29-33), probably acting as another predictor for increasing salinity at the very bottom of the profile. *Ostrea edulis* is important in subsets 29-33 (~6,680-3,860 BC), covering the time period in which this taxon reaches its highest abundances (~6,000-5,570 BC). Maximum *O. edulis* abundances coincide with maximum DI-salinities (~24-27 ‰, between ~5,730-5,510 BC) and high MI-salinities (19-23 ‰ between ~5,570-5,510 BC) suggesting that *O. edulis* is a predictor for high salinity conditions (and greater marine influence on the fjord). This is probably also true of *Brachystomia eulimoides* (high-salinity demanding; Sorgenfrei, 1958; Petersen, 2004), which also achieves its highest abundance in this period (~5,870-5,440 BC) and is important in subset 26-28; (~6,270-3,800 BC). Varying molluscan abundances might also directly exert grazing pressures upon the diatom communities, though this is difficult to isolate in a system subject to such rapidly changing environmental conditions.

The vegetation predictors in the lower subsets (i.e. *Pinus*, subset 35, Poaceae, subset 34, *Dryopteris*, subsets 27-34) probably represent catchment erosion and inwash at this time, associated with rising sea levels and submergence of low lying lands. The relatively high abundances of Poaceae pollen (before ~6,000 BC) is likely to be predominately derived from wetland grass species, whilst extremely high percentages of *Dryopteris* between ~7,060-5,890 BC suggest that the Korup Sø area either was surrounded by, or situated in close proximity to, extensive swamp and fenlands (also *Sparganium* present between ~7,470-6,600 and *Sphagnum* between ~6,700-5,650 BC). Along with inputs of plant material, catchment erosion and submergence of terrestrial swamps and wetlands is likely to have resulted in greater inputs of nutrient and minerals into the fjord. Several shifts in the diatom flora are identifiable between ~7,330-5,930 BC (Figure 5.4) and, whilst the diatoms appear to be primarily responding to sea level and salinity change at this time, it is possible that additional factors might also be important. For example, *C. choctawhatcheeana* (high-nutrient favouring diatom) appears in the record between ~5,970-5,930 BC, *Cyclotella striata* and *Skeletonema costatum* exhibit a peak at 6,270 BC *and Chaetoceros* cysts fluctuate throughout, which might suggest that nutrient inputs either from the catchment following changes in vegetation structure or from inputs of nutrient rich water from the North Sea (through the Kattegat) are also important in driving changes in the diatom dataset.

It is important to stress that the pollen catchment area is likely to be far greater than the hydrological catchment area of the fjord, and therefore care must be taken with the interpretation of predictors of vegetation change. Nevertheless, it remains likely that local vegetation changes within the catchment will exert pressure on fjord diatom communities. Despite changes in the marine environment dominating in the lower section of the record, some proxies for vegetation change are also significant in the Mesolithic (e.g. *Tilia* is important in subset 34; 7,180-3,870 BC). The decrease in *Tilia* and *Pinus* pollen at ~7,060 BC (Figure 5.7) might reflect a change in the size of the pollen catchment due to increasing sea level around this time (i.e. a larger pollen catchment meaning a change from a more local to regional signal). Whilst a change in the size of the pollen catchment is unlikely to directly affect diatom communities, in the moving window analyses, *Tilia* (in subset 34) and *Pinus* (in subset 35) might be acting as predictors for a change in the size (and/or configuration) of the hydrological catchment area as this probably also increased under rising sea levels. A larger hydrological catchment might have affected factors such as surface run-off, sediment and nutrient supply to the fjord and perhaps increased water turbidity. Later, the establishment of a closed canopy forest is likely to have reduced inputs of material (including nutrients) from the catchment, perhaps partly explaining the low sediment accumulation rates in the Mesolithic section of the record (i.e. pre 3,900 BC).

Ulmus is important in subset 28 (~6,270-3,840 BC), covering the period of maximum elm abundance in the catchment (according to the Korup Sø pollen record) and its rapid decline at ~3,900 BC. The elm decline triggered a large restructuring of the woodlands in the early Neolithic (~3,900-3,550 BC), which in turn is likely to have contributed to the large increase in the sedimentary accumulation rate (important in subsets 23-27: 5,970-3,700 BC) after ~3,900 BC (Figure 5.3). During this woodland transition, *Betula* (important in subset 19, 4,730-3,280 BC), *Corylus* and *Quercus* (important in subset 14; 3,900-3,080 BC) increase rapidly (Figure 5.7). The reduction of elm trees and shift in forest structure might also have released nutrients into the fjord (e.g. Bormann *et al.*, 1974; Bayley *et al.*, 1992), possibly explaining the peak in diatom concentration around ~3,800 BC and increased abundances of *Cyclotella choctawhatecheeana* at ~3,780 BC. There is also a general increase in *Fragilaria* s.l. around this time which might further reflect changing (or variable) nutrient status and/or instability (perhaps associated with increased sediments loads) within the fjord. Lower sea levels and human impact (e.g. cereal pollen first appears at ~3,850 BC; Figure 5.7) might also have influenced the Korup Sø sedimentary regime at this time, whilst climate change and edaphic factors might have exerted some pressure upon the vegetation.

Mytilus edulis (subsets 20, 23-23 and 28; 6,270-3,380 BC) and *Hydrobia ulvae* (subsets 21-21; 5,380-3,430 BC) appear as significant predictors for environmental change in subsets covering the late Mesolithic/early Neolithic period. *Hydrobia ulvae* is abundant for much of the profile, but exhibits a brief drop in the late Mesolithic. After a substantial increase in the early Neolithic, it becomes the dominant species as overall mollusc concentrations drop. Whilst the decrease in *Bittium reticulatum* and general decline in high-salinity demanding taxa might be interpreted as a salinity decline, the DI-reconstruction and the continued presence of *Onoba vitrea* and *Tapes decussates* (albeit at low concentrations) suggest that salinity does not decline at this point. Petersen, (1993) suggested that the decline in molluscan concentration is caused by an increase in the sedimentary accumulation rate (i.e. at

~3,900 BC; Figure 5.3). Therefore, it is likely that these two taxa (i.e. *M. edulis* and *H. ulvae*) are proxies for a change in the sedimentary regime and substrate (likely driven by forest restructuring and later human impact), a decrease in mollusc concentration and a shift in grazing pressure. These taxa have the ability to survive living in muddy conditions, with higher sediment loads and finer substrates, which other molluscs might be unable to survive (e.g. oysters; Rasmussen, 1958; Yonge, 1960).

Plantago lanceolata (important in subsets 12-13, ~3,900-2,980 BC and 15-19, ~4,730-3,100 BC) is a likely predictor for expanding grasslands (Iversen, 1941; Behre, 1981; Gaillard, 2007) in the Korup Sø catchment area. This taxon is consistently present in the pollen record (Figure 5.7) after ~3,730 BC, marking the replacement of areas of woodland with agricultural land for both pastoral and arable farming (Iversen, 1941; Troels-Smith, 1960). Again, this is likely to have affected rates of surface run off and inputs of sediment and nutrient loads (e.g. animal waste and other natural fertilisers) from the catchment. Variable sediment and nutrient loads probably explain some of the fluctuations in *Fragilaria* s.l. and *Chaetoceros* cyst abundances after ~3,700 BC. In the uppermost section of the profile (i.e. subsets 5-11; 2,380-3,870 BC), changes in the major tree taxa (i.e. *Pinus*, *Betula*, *Corylus* and *Alnus*) significantly explain some of the variation (8-19 %) in the diatom record. Whilst climatic (e.g. decreasing temperatures; Antonsson and Seppä, 2007) and edaphic factors might also be important, fluctuations in these tree taxa are likely to be heavily associated with anthropogenic disturbance, particularly rapid forest clearance events (e.g. decrease in arboreal pollen at ~3,250 BC). Features such as short term increases in *Betula* followed by expansions of *Corylus* are described in Iversen's landnam model (Iversen, 1941; Iversen, 1973; Edwards, 1993) further supporting the case here, that fluctuations in the major tree taxa are likely predictors of anthropogenic disturbance. *Corylus* nuts might also have been an important resource, and therefore this tree might have been directly coppiced by Neolithic (and late Mesolithic) cultures.

5.7 Synthesis of results and discussion of site

This section attempts to tie together the results from the independent proxy records to build up a more complete picture of environmental change in and around the Korup Sø area during the mid-Holocene. Figure 5.10 provides a summary of the major environmental changes described in the above sections, together with sea level and temperature curves, included to examine changes in a more regional context. For ease of interpretation, five environmental zones (labelled KRP-E1 to E5) have been identified, based on relatively good agreement between the independently assigned (optimal split) proxy zones. An environmental zone has been assigned at points where independent zone splits for all 3 proxies fall in close temporal proximity to one another. There is one exception to this rule; the split between KRP-E3 and KRP-E4 is based on agreement of only 2 proxies (molluscs and pollen). This has been added due to the almost identical timing of the split in both proxies and the substantial changes exhibited between zones. The multiproxy changes are discussed in relation to these environmental zones.

KRP-E1 – Early brackish phase (~7,600-6,900 BC) – The high abundance of *Pediastrum* (freshwater algae recorded in the pollen samples) at the bottom of the profile (i.e. ~7,600 BC), together with a variety of freshwater diatoms (preliminary assessment only) suggests that Korup Sø was probably limnic at this time. By the start of the diatom (~7,330 BC) and molluscan (*C. edule* recorded at ~7,490 BC) records presented here, there was certainly a marine influence. Both DI- and MI-inferred salinities (~4-6 ‰ between ~7,500-7,300 BC) suggest brackish conditions, but at this point it is difficult to determine the degree of connection with the sea. Korup Sø may have been isolated from the sea, but breached by intermittent pulses of saline water, perhaps during storm conditions. Alternatively, it may have been only partially connected at this time (i.e. ~7,500-7,300 BC). The molluscan fauna at this point is very poor, consisting of just two species (*O. atomus* and *C. edule*) which could potentially live under both scenarios. If completely cut off, regular inputs of saline water would be required to support these species (and maintain this salinity) and therefore at the very least, we must assume that the Korup Sø was in very close proximity to the Randers/Kolinsund Fjord systems at this time.

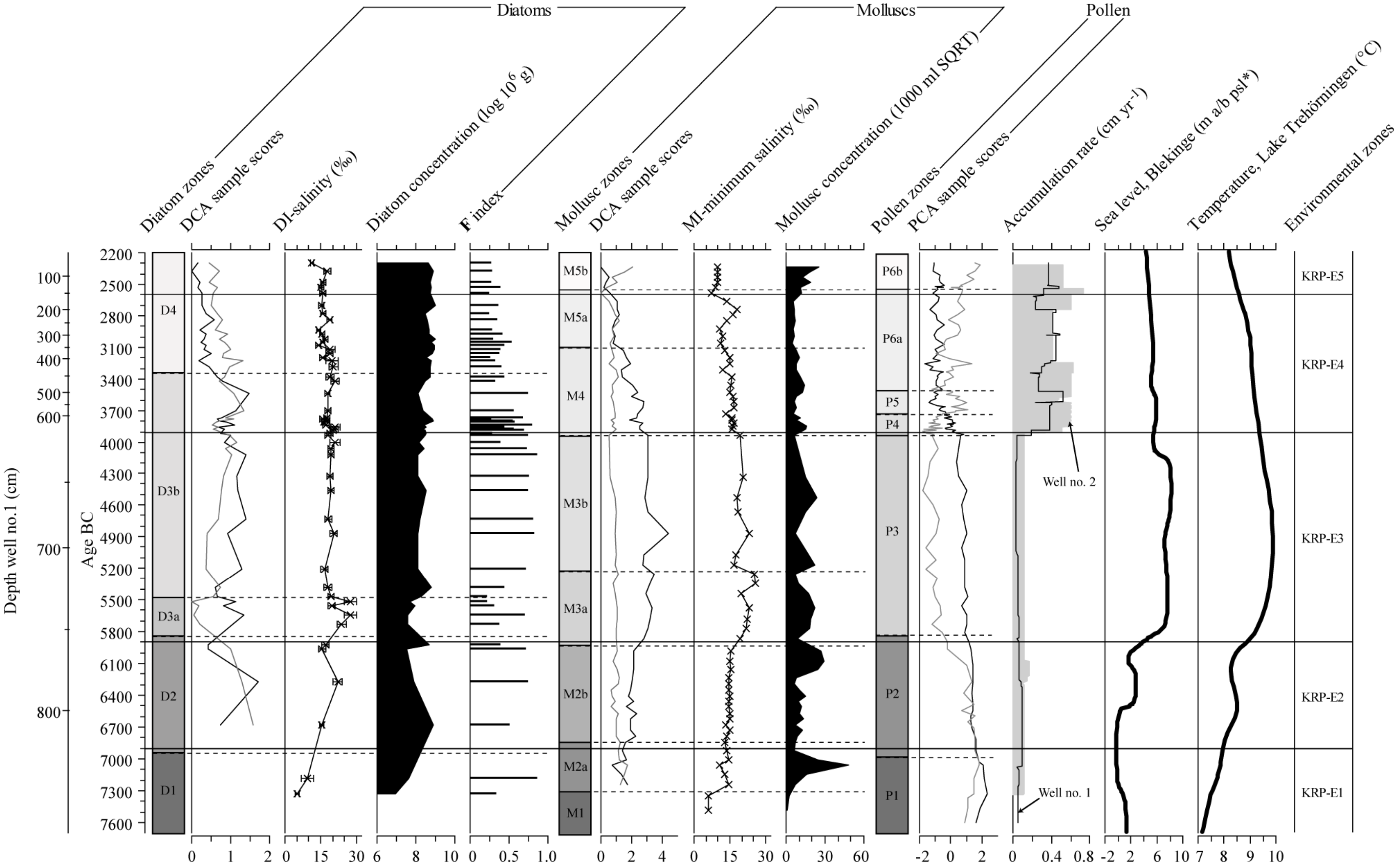


Figure 5.10 - Summary diagram for all proxies analysed at Korup Sø, together with regional temperature (Antonsson and Seppä, 2007) and sea level (Berglund *et al.*, 2005) reconstructions. Well no. 1 sedimentation rate is based on latest age model calculations (J. Olsen, P. Rasmussen, unpublished data). *m a/b psl = metres above/below present sea level.

In the succeeding phase, salinity increases quite rapidly in response to global sea level rise (e.g. Mörner, 1976; Berglund *et al.*, 2005; Yu *et al.*, 2007), and by ~7,200 BC a permanent marine connection is definitely established. The molluscan fauna increases both in diversity and concentration, characterised by high numbers of the brackish water gastropods *Hydrobia ventrosa* and *Hydrobia ulvae* but supplemented by several other brackish/marine species. The diatom flora at this point is still dominated by opportunistic *Fragilaria* s.l. that are highly competitive in unstable environments such as transitional zones (Eronen, 1974; Stabell, 1985; Seppä *et al.*, 2000; Miettinen, 2002), but the overall shift in the flora indicates strengthening of brackish conditions and the DI-salinity of ~9 ‰ (at ~7,200 BC) would probably be unachievable by intermittent intrusions alone. Dinoflagellates also appear in the record around this time and *Pediastrum* numbers fall (after ~7,370 BC). Small amounts of *Ruppia* and *Coleogeton* (both living in weakly brackish water systems) are also present probably occupying the more littoral areas, but both disappear around ~7,000 BC, likely due to rising salinity and an increase in water depth.

The pollen record suggests that the Korup Sø basin (and probably large areas of low-lying Kolindsund coastal land) were surrounded by extensive wetlands (including swamp and fenlands) at this time. The high and exaggerated percentages of Poaceae (probably derived from wetland grass species at this time), Cyperaceae pollen and *Dryopteris* spores in the lower part of the pollen record are most likely to be the result of catchment erosion and inwash, associated with rising sea levels and gradual submergence of the lower lying wetlands. At this time the catchment was covered by a mixed deciduous forest, with *Tilia*, *Corylus*, *Ulmus* and *Quercus* as the main taxa on the high ground and *Betula*, *Pinus* and *Alnus* likely to be prevailing on low-lying moist ground (Iversen, 1969; Andersen, 1978; Aaby, 1983). The relatively high abundances of *Pinus*, *Betula* and *Alnus* at this time might suggest that their representations in the pollen record are also affected by catchment erosion and inwash under rising sea levels.

KRP-E2 - The early Korup Sø marine phase (~6,900-5,900 BC) – The diatom sampling resolution is low in this phase (4 samples), but the diatom flora changes several times in this period due to rapidly changing conditions. The lowermost sample is completely dominated by benthic species (particularly *Bacillaria paxillifer* and *Fragilaria* s.l.), but by ~6,270 BC, higher-salinity planktonic/tychoplanktonic diatoms (e.g. *Paralia sulcata*, *Chaetoceros* cysts, *Cyclotella striata*) become more abundant in the profile due to increasing salinity, water depth and nutrient status, driven by rising sea levels. A drop in salinity and a shift back to a

Fragilaria s.l. dominated system occurs at ~5,970 BC, possibly driven by a sea level regression phase (identifiable in the Blekinge stratigraphy between 6,250-6,050 BC; Berglund *et al.*, 2005), which has been tentatively linked to the 8.2 k cooling event (Alley *et al.*, 1997; Rohling and Pälike, 2005).

In contrast to the diatoms, there is little change in the molluscan faunal composition in this phase, which remains dominated by *Hydrobia ulvae* throughout, with a minimum MI-salinity of ~13-15 ‰. However, despite the huge number of *H. ventrosa* snails in the upper section of KRP-E2 (after ~6150 BC), salinity appears to be increasing as a number of higher-salinity demanding species (*Corbula gibba, Retusa truncatula, Ostrea edulis, Onoba vitrea*), appear in the record. There is no evidence for decreased salinity associated with a marine regression (6,250-6,050 BC) in the molluscan record, though this is more likely to go undetected due to the coarser sampling resolution (5-10 cm thick slices in KRP-M2; 50-120 year resolution), absolute dominance of *H. ulvae* and generally lower sensitivity of molluscs (when compared to diatoms) to more minor salinity shifts.

Pinus are still frequent in the pollen record, perhaps related to catchment erosion and inwash. *Tilia*, *Ulmus*, *Corylus*, *Quercus*, *Betula* and *Alnus* all remain relatively abundant in the catchment, with *Quercus* increasing slightly after ~6,100 BC. High percentages of wet ground, swamp and fen taxa (Figure 5.8) are present in the pollen record, most notably *Dryopteris* spores (14-35 %), but also Cyperaceae, *Sphagnum* and Poaceae (from wetland grass species), again indicating large scale catchment erosion and drowning of extensive low lying swamp and marsh areas. The aquatic taxa disappear from the record after ~6,700 BC as the salinity rises (and water depth increases) and only small percentages of *Pediastrum* remain present in this phase.

KRP-E3 – The Korup Sø marine maximum (5,900-3,900 BC) - The beginning of KRP-E3 broadly coincides with the second phase of the Littorina marine transgression (L2; Berglund *et al.*, 2005) and marks the beginning of the maximum marine phase at Korup Sø. Rapid sea level rise appears to seriously disrupt the early Holocene system at Korup Sø, causing DI-salinity to rise sharply, reaching its highest inferred levels in the whole record between ~5,700 -5,200 BC. This salinity rise causes a substantial change in the mollusc fauna with the appearance of a number of high salinity tolerant species and a reduction in *H. ulvae* numbers (though it remains the dominant species). Several of the high-salinity demanding molluscs practically disappear from the record again shortly after ~5,200 BC, but the continued presence of *B. reticulatum* suggests salinity remained high. The high DI-salinity values (between ~5,700-5,200 BC) result from a large increase in *Paralia sulcata* valves, perhaps present in exaggerated numbers due to poorer preservation (e.g. **F** index drops below 0.4 in several samples between ~5,700-5,500 BC and decreased diatom concentration). Its abundance drops off again by ~5,470 BC and *Fragilaria* and *Opephora* species dominate the record thereafter (*Fragilaria* s.l.). The decrease in DI-salinity after ~ 5,510 BC might be caused by a minor sea level regression (~1 m drop in sea level between the L2 and L3 transgression in the Blekinge stratigraphy; Berglund *et al.*, 2005) and a subsequent drop in water level, enabling *Fragilaria* s.l and other benthic taxa to increase in abundance.

There is no renewed increase in DI-salinity associated with the L3 transgression (~4,850-4,250 BC; Berglund *et al.*, 2005), probably due to the high abundances of euryhaline *Fragilaria* s.l. present in the record at this time. Variable abundances of *Fragilaria* s.l. suggests that this phase is characterised by general instability (e.g. Eronen, 1974; Stabell, 1985), likely being subject to fluctuations in salinities, nutrient status, abundances of grazing populations (e.g. molluscs and foraminifera) and variable currents, perhaps being driven by rising sea level and greater exposure to the Kolindsund. In such circumstances, several environmental variables might be influencing the diatom assemblage and perhaps skewing the DI-salinity estimates. Alternatively, salinity rise associated with higher sea levels might be counteracted by freshwater inputs. Cooler temperatures at the end of the Holocene thermal optimum (e.g. Antonsson and Seppä, 2007; Seppä *et al.*, 2009; Brown *et al.*, submitted) might have increased winter ice cover on the catchment, leading to larger inputs of freshwater (and nutrients) in the melt season.

Chaetoceros cysts are also abundant through KRP-E3, probably indicating high nutrient status and high productivity, at least in the spring and summer months. Higher productivity probably results from increased input of nutrient rich water from the Kolindsund during the Mesolithic high sea level stand (Berglund *et al.*, 2005) and perhaps nutrient inputs from the catchment, particularly during the late winter/early spring. The dominance of the larger sized *P. sulcata* valves (> 20 µm, more abundant in nutrient poor conditions) might be explained by nutrient depletion in winter (when it reaches its highest abundances; Hobson and McQuoid, 1997), following likely spring/summer *Chaetoceros* blooms. At the top of this phase, the largest Littorina regression occurs at ~4,200 BC (~3 m decline in sea level in the Blekinge stratigraphy; Berglund *et al.*, 2005), but there appears to be little change in the biological records until ~3,900 BC (see below; KRP-E4).

In this phase the woodland is dominated by *Tilia*, *Ulmus*, *Corylus*, *Quercus*, *Alnus* and *Betula* while *Pinus* and especially *Fraxinus* are more sparsely represented. The catchment probably closely resembles the Atlantic woodland structure suggested by pollen analyses from small hollows and soils in Denmark with *Tilia* prevailing on dry well-drained soils, together with *Corylus* and *Quercus*, whilst *Ulmus*, *Alnus* and *Fraxinus* could be found on damp or wet areas, presumably mixed with some *Corylus* and *Quercus* (e.g. Iversen, 1969; Andersen, 1978; Aaby, 1983). The presence of *Hedera helix* in the pollen record is an indicator of warmer conditions than today (Iversen, 1944) associated with the Holocene thermal maximum (5,550-3,550 BC; Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Seppä *et al.*, 2009; Brown *et al.*, submitted). Poaceae pollen decreases in abundance in this phase, as does *Dryopteris* (less than 2 % after 5,560 BC) and the other wet ground, swamp and fen taxa, suggesting that rising sea levels (Berglund *et al.*, 2005; Yu *et al.*, 2007) had probably drowned most of the low lying swamp and marsh areas surrounding the fjord by the late Kongemose/early Ertebølle period.

KRP-E4 – The late Korup Sø marine phase (3,900-2,600 BC) – The MI-reconstruction suggests salinity decreases at the onset of KRP-E4 (~3 ‰ decrease at ~3,900 BC, largely driven by the decrease in *B. reticulatum* and increase in *H. ulvae*) and continues to decline steadily throughout this phase. There is, however, little change in DI-salinity up until ~3,400 BC, with the exception of a relatively short-lived decline between ~3,850-3,750 BC. This short drop coincides with a peak in *C. choctawatcheeana* (~12 % at ~3,780 BC) suggesting higher nutrient status within the fjord, probably resulting from increased inwash of nutrients (and sediment) from the catchment, following the *Ulmus* decline and subsequent restructuring of the woodland (Figure 5.7 and see below). Unfortunately, changing nutrient status and increased sediment loads might weaken the DI-salinity estimates in this section of the record (e.g. DI-salinity model performs poorly under high nutrient conditions in the upper section of the Kilen record, see section 3.11). The bulk of evidence presented here suggests that there was no major decline in salinity at the Mesolithic-Neolithic transition and/or lower salinity throughout the early Neolithic period (as proposed by Rowley-Conwy, 1984). Despite the apparent decrease in the MI-salinity reconstruction, salinity remains high enough to support small (but sporadic) concentrations of *B. reticulatum R. albella* and *T. decussatus* (up until ~3,400 BC, all high salinity-demanding molluscs; Figure 5.6) and occasional finds of *O. vitrea* up until ~2,750 BC (which might be considered to be in better agreement with the diatom record). Salinity fluctuations might have been important potentially disrupting or preventing spawning of oysters and other mollusc populations on yearly to decadal scales (but likely to be at too fine a resolution to be picked up by the diatom and molluscan records presented here). The DI-salinity reconstruction provides some evidence for fluctuating salinities (on decadal to centennial scales) between ~4,100-3,700 BC; Figure 5.10), but as indicated above, it is difficult to be certain whether or not these are true salinity variations, or just changes in the diatom flora associated with other environmental parameters (most notably changing nutrient and sedimentary regime).

Both molluscan diversity and concentration decrease in this phase, probably in response to the rapid increase in the sedimentary accumulation rate (~0.4 cm yr^{-1} increase) at ~3,900 BC. Petersen, (1993) previously attributed this sedimentary shift to the lowering of the tidal range in the early Neolithic period (e.g. Nielsen, 1938), but this is rejected on the basis of the lack of evidence supporting this hypothesis. It is more likely to be associated with increased sediment loads from the catchment following the *Ulmus* decline and subsequent restructuring of the woodland (discussed in more detail below). Thereafter, the accumulation rate continues to fluctuate throughout the remainder of the profile, probably in response to a mixture of land use changes (affecting sediment loads) and variations in the marine environment (e.g. sea level variation, productivity, sediment inwash from the Kolindsund perhaps under storm conditions).

By ~3,400 BC, both proxies indicate decreasing salinity, which is further supported by the reappearance of *Ruppia* in the pollen record. In fact *Ruppia* continues to gradually increase in abundance throughout KRP-E4 as the fjord begins to get shallower, due to decreasing sea levels and the high sediment accumulation rates within the basin (see below). Both the molluscs and the diatoms indicate a short term rise in salinity in the upper part of KRP-E4 (~2,800-2,700 BC), with *O. vitrea* reappearing in the mollusc record, albeit at low numbers, and small increases in marine benthic diatoms such as *Cymatosira belgica* (likely washed in from the outer exposed parts of the fjord) and *D. minor*. This coincides with the beginning of the Pitted Ware Cultural period (2,800-2,400 BC), and the reappearance of oyster-dominated coastal shell middens in some parts of Denmark (Andersen, 1983; Rasmussen and Richter, 1991; Rasmussen, 1993a; Andersen, 2007).

In KRP-E4 there is also a substantial shift in the vegetation, driven by a restructuring of the forest following the rapid *Ulmus* (elm) decline and the introduction of agriculture (Figure 5.7). *Betula, Quercus* and *Corylus* (all prolific pollen producers) quickly increase at the onset of KRP-E4 causing a brief rise in the total arboreal pollen (2-3 % higher between ~3,940-

3,800 BC), before plants indicative of open land, related to agricultural impacts, become more abundant in the profile. The elm decline in the Korup Sø catchment is likely to be predominately driven by an elm pathogen, though the early signs of agriculture in the pollen record (i.e. *Triticum* and *Hordeum* appear around ~3,850 BC) might suggest that humans also played some role (e.g. Peglar, 1993; Peglar and Birks, 1993). The appearance of cereal pollen is followed shortly after by increased frequencies of Poaceae pollen (predominately from terrestrial grassland taxa at this time) and plants characteristic of pastoral and arable farming such as *Plantago lanceolata* and *Rumex*-type (Iversen, 1941; Behre, 1981; Gaillard, 2007) both appear around 3,750 BC.

The gradual reduction in *Tilia* pollen between ~3,900-3,550 BC further reflects woodland clearances, supported by the increased abundances of secondary woodland taxa (i.e. *Betula* and *Corylus*), apophytes (most notably Poaceae, *P. lanceolata* and *Artemisia*) and anthropochores (*Triticum*-type and *Hordeum*-type; Iversen, 1941; Iversen, 1973; Rasmussen, 2005). After ~3,550 BC, *Tilia* re-establishes itself, accompanied by increases in *Quercus*, *Ulmus* and *Fraxinus*, perhaps reflecting reduced cultural pressure on the landscape at this time. This might be supported by more sporadic appearances of cereal and *Rumex*-type pollen between ~3,500-2,580BC and low percentages of *P. lanceolata* (except for occasional peaks, see below). The slight increase in grass pollen at this time might be exaggerated by the reduction in prolific pollen producing trees (e.g. *Betula* and *Corylus*). This might also suggest that the pollen record underestimates the area of open land in the earliest part of the Neolithic (i.e. 3,900-3,500 BC) due to the abundance of prolific pollen producers (e.g. *Betula* abundant between ~3,900-3,700 BC and *Corylus* increase up until ~3,500 BC). It is also possible that in the later part of the record larger proportions of pollen from wetland grass taxa are being washed into the fjord, as low lying wetlands begin to emerge, as high sediment accumulation rates cause the basin to shallow and sea levels decline (e.g. Mörner, 1976; Berglund *et al.*, 2005).

The four-six fold increase in the sediment accumulation rate (identifiable in both wells) around ~3,900 BC (Figure 5.3) is probably the most characteristic change of this phase and therefore it is important to expand upon it here. The timing of the accumulation rate increase at Korup Sø suggests that it is driven by forest restructuring and anthropogenic disturbance associated with the introduction of agriculture, leading to increased rates of soil erosion and surface run off into the Korup Sø basin. However, as the chronology of the Korup Sø profile is based on radiocarbon dates from marine shells (Table 5.2) then it is possible that errors associated with the marine reservoir effect are inherent in the age model, though it is argued here that these errors are likely to be relatively small in magnitude for several reasons. Firstly the sequence of radiocarbon dates appears to fit well with known palynological events particularly the *Ulmus* decline (~3,900 BC; Andersen and Rasmussen, 1993). This is particularly rapid in the Korup Sø pollen record (decrease of 5-6 % in ~50 years; Figure 5.7) and can therefore be considered a precise dating marker in the Korup Sø stratigraphy. Whilst there is no specific regional date for the introduction of *Plantago lanceolata*, *Rumex* and the first appearance of cereal cultivation (*Triticum* and *Hordeum*), these 'events' are in good accord with other Danish coastal, limnic and mire pollen records (e.g. Troels-Smith, 1960; Aaby, 1983; Odgaard, 1994; Odgaard and Rasmussen, 2000; Rasmussen, 2005; Rasmussen and Olsen, 2009, P. Rasmussen, unpublished data). Secondly, dates generated from the two boreholes are largely in agreement and can be tied via events in the pollen stratigraphy (Figure 5.2; H. Krog, unpublished data). Thirdly, all dates appear in stratigraphic order. The latter two points might be considered unlikely if the marine reservoir effect was as problematic at Korup Sø as in other Danish fjord sites such as Tempelkrog and Horsens Fjord (Olsen *et al.*, 2009). Fourthly, an increase in sediment accumulation rate is often recorded in northwestern Europe after the *Ulmus* decline, subsequent forest restructuring and introduction of agriculture in the early Neolithic period (Odgaard, 1989; Edwards and Whittington, 2001; Rasmussen and Bradshaw, 2005; Odgaard, 2006; Rasmussen and Olsen, 2009) and it is therefore consistent with, rather than remarkable that it is recorded at this site.

It is, however, acknowledged that even minor errors (i.e. ±0-200 years) in the age model could affect the rate and overall pattern of sedimentary accumulation in the Korup Sø profile. As calculations of sediment accumulation rates are determined by chronological fix points (either ^{14}C dates or other tie points such as palynological markers) sedimentary dynamics between dates/tie points remains unknown and therefore assumed constant (i.e. linearly interpolated). Whilst it is argued here that there is a true increase (of substantial magnitude) in the sedimentary accumulation rate following the *Ulmus* decline and the introduction of agriculture, there is potential scope for variable sedimentation, particularly in the Mesolithic section of the record. For example, as sea level begins to decline (at ~4,200 BC; Berglund *et al.*, 2005), a slight increase in the sedimentary accumulation rate might be expected (e.g. as seen at Norsminde Fjord; see section 4.11) due to a factors such as decreased size and depth of the basin, lower velocity currents, reduced marine connection and less output of sediment from the fjord/estuary into the open sea. The nearest date to the *Ulmus* decline (precise dating marker) in the Mesolithic section of the Korup Sø Well no. 2 profile is located ~50 cm below this (at depth 702.5 cm; Table 5.2) and dated to ~5,070 BC (~1,100 years earlier). Therefore, with the potential error in the dating model it must be considered plausible that accumulation rates might have started rising earlier than ~3,900 BC, perhaps in response to sea level change in the late Mesolithic period (~4,200 BC; Berglund *et al.*, 2005, Figure 5.10).

Consideration must also be given as to whether a four to six fold increase in the sediment accumulation rate is possible around ~3,900 BC (under the current age model). This large magnitude increase occurs at a time when no reduction in overall arboreal pollen (despite the elm decline and subsequent forest restructuring) and very little expansion of open land (i.e. Poaceae pollen only begins to increase after ~3,800 BC and *Plantago lanceolata* and *Rumex* spp. appear after ~3,740 BC; Figure 5.7). However, the lack of change in overall arboreal pollen percentages might be explained by a higher pollen capacity of the existing forest vegetation at this time (Aaby, 1988) and the expanse of light demanding (and some prolific pollen producing) trees such as *Betula*, *Corylus* and *Quercus* (Figure 5.7). Aaby, 1988 suggested that openings in the forest created by Neolithic people (by tree felling) enabled a larger part of the tree’s crown to flower, explaining the positive relationship between arboreal pollen influx and cultural activity in the early to mid Neolithic. Conversely, in the late Mesolithic period, when the forest was dense, arboreal pollen influx was lower (cf Iversen, 1973). The early appearance of cereal crops in the Korup Sø pollen record (*Triticum* appears at ~3,850 BC and *Hordeum* at ~3,840 BC) and lack of pastoral indicators might suggest that arable farming was the dominant mode of farming in the earliest part of the Neolithic. Arable land is believed to be more prone to soil erosion than permanently grass-covered pastures (Veihe *et al.*, 2003), which might partly explain why there is such a substantial (initial) increase in the accumulation rate at ~3,900 BC. This might also explain why later, larger magnitude agricultural changes such as the expansion of pastoral grasslands after ~3,800 BC (i.e. increase in Poaceae, *Plantago lanceolata* and *Rumex* spp. in the pollen) and additional rapid but short term deforestation phases (e.g. ~3,350-3,200 BC; Figure 5.7) appear to have less impact on the rate of sediment accumulation in the Korup Sø basin. Unfortunately there is little data available from other Danish coastal or fjord systems for comparison, largely due to poor dating and the lack of complete palaeo records from these environments. Much more data exists for lake or mire systems and some of these records also document an early increase in soil erosion and sediment accumulation (e.g. Odgaard, 1989; Rasmussen and Bradshaw, 2005). Whilst the accumulation rates in these systems do not increase by such a large magnitude as in Korup Sø, it is likely that factors such as a reduction in the size and depth of the basin with declining sea level, reduced currents and reduced connection with the

Kolindsund will also have contributed to the large increase in the sedimentation rate at Korup Sø.

It is also important to highlight that increased soil erosion and sediment accumulation is often heavily affected by localised factors (e.g. Edwards and Whittington, 2001) and subsequently no repeatable, synchronous regional signal is identifiable across Denmark or Scandinavia. For example, in several Danish lakes and mires there is a delay evident, with accumulation rates increasing slightly later (e.g. Aaby, 1988; Odgaard, 2006; Rasmussen and Olsen, 2009), often coinciding with the appearance of the unambiguous indicators for arable and pastoral farming (i.e. cereal pollen and *Plantago lanceolata*). The few well dated coastal records available (e.g. Norsminde Fjord (see Chapter 4), Horsens Fjord and Tempelkrog; P. Rasmussen *et al.* unpublished data) all appear to exhibit unique sedimentation records, likely being affected by factors such as response to sea level change, fjord configuration and local vegetation and catchment changes. Furthermore, Colman *et al.*, 2002 demonstrate that accumulation rates in shallow water sites in the Chesapeake Bay region reflect more localised factors, whereas the deeper sites appear to respond more consistently to larger-scale regional changes. On a global scale Dearing and Jones, 2003 identify that small basins are the most responsive to human impacts and show the greatest changes in sediment flux, typically up to 5-10 fold higher than under undisturbed conditions, though this figure can rise depending on the intensity of the impact or local environmental conditions. At the Mesolithic-Neolithic transition, Korup Sø can almost certainly be classified as a small shallow water fjord and probably extremely sensitive to catchment changes. It might, therefore, be considered unsurprising that a large increase in sedimentation rate occurs following a substantial forestry change (i.e. rapid decrease in *Ulmus* and structural changes of the forest) and with respect to clear early evidence for arable farming taking place in the immediate catchment (Figure 5.7).

Outside Denmark, other systems have shown substantial increases in the sedimentary accumulation rate associated with the introduction of (primitive) agriculture and even reduced sensitivity (i.e. little or no increase in sedimentary accumulation rates) to later, larger scale agricultural changes. For example, in a closed lake-catchment system in northern Guatemala, Anselmetti *et al.*, 2007 found that the most rapid soil loss (~1000 t/km^2 yr^{-1}) occurred during the initial land clearance when Mayan population densities were relatively low. A dramatic rise of 30-80 times the undisturbed sediment yield (used as a proxy for soil erosion) is also evident in sediments from Frains Lake, Michigan (USA) following the initial phase of deforestation by pioneer settlers (Davis, 1976). Sediment yield stabilises again after the initial disturbance peak, but post settlement levels remain higher (~10 fold) than pre-clearance levels. Both these studies demonstrate that land clearance and primitive agriculture, likely carried out by small numbers of people can have quite large effects on erosion rates and subsequent sediment delivery to aquatic systems. Additionally, lake records from a steepland catchment in New Zealand document an increase in sedimentation/erosion rates following fire-induced changes from indigenous forest to fern/scrub following Polynesian settlement (Page and Trustrum, 1997). Later, European settlement in this catchment and a change to pastoral land cause the lake sedimentation rate to further increase, reaching a rate of 8-17 times than under indigenous forest. Therefore, viewed in the light of such drastic changes in other systems, a four to six fold increase in the sedimentation rate at Korup Sø can be considered both feasible and likely with respect to the environmental and cultural change occurring within the catchment at this time.

KRP-E5 – Late brackish Zone (~2,600-2,300 BC) - In the uppermost zone, sea level continues to decrease gradually and signs of this weakening marine influence become apparent in the proxy records. With the exception of *Rissoa membranacea* (present up until ~2,430 BC), there is an absence of high-salinity demanding mollusc species. Instead, the molluscan fauna is dominated by the brackish water snail *H. ventrosa* and lower concentrations of *Littorina tenebrosa* and the lagoon cockle *Cerstoderma glaucum*. This is supported by relatively high abundances of *Ruppia* (2-8 %) in the pollen record, a decrease in the abundance of dinoflagellates (Figure 5.8) and the disappearance of some of the marine diatom species such as *Dimmeregramma minor, Fallacia fenestrella* agg. and *Chaetoceros* resting cysts. Diatom concentration continues to increase as *Fragilaria* s.l. further increase in abundance, though this probably indicates the ability of these species to colonise unstable environments (e.g. subject to sediment disturbance, reducing habitat availability, fluctuating nutrients, salinity and sediment loads) and/or perhaps shallower water at the coring location, rather than increasing productivity and/or nutrient status at this time.

Low numbers of planktonic species throughout this zone suggest shallower water and the appearance of taxa such as *Psammothidium kuelbsii, Pseudostaurosira brevistriata* and *Achnanthes submarina* in the uppermost diatom sample suggest that by ~2,300 BC the Korup Sø basin was brackish, probably maintaining only a very narrow connection with the Kolindsund (e.g. DI-salinity of ~11 ‰ at this time). The appearance of scattered *Ruppia* stalks in the sediments after ~2,500 BC, together with a 1-2 % *Ruppia* rise in the pollen record likely indicates extensive emergent bedding occupying the shallow littoral areas. Some high-salinity tolerant species remain present in the profile such as *P. sulcata* and *Fallacia cryptolyra*, though the latter of these is never particularly abundant and disappears by ~2,380 BC. Despite *P. sulcata* favouring higher salinities, its euryhaline nature, ability to cope with low nutrient conditions and tychoplanktonic lifestyle might have ensured its survival into the later stages of the fjords development. Alternatively, the relatively high percentages of *Paralia sulcata* (and continued presence other higher salinity diatoms) might be explained by greater disturbance and reworking of sediments under shallower conditions, or by inwash from the Kolindsund and/or the more exposed parts of the fjord. The sedimentation rate of the profile remains high, but continues to fluctuate, probably being affected by variable sediment loads from the catchment due to human impact (e.g. forest clearance episodes), but climate changes (or events such as storms) and minor variations in sea level might also be important.

Alnus becomes more abundant in this zone, perhaps due to expanding wetland areas, whilst *Corylus* and *Quercus* continue to fluctuate, again probably being influenced by human activity on the catchment. There is a slight decrease in *Fraxinus* pollen, whilst the other major tree taxa (*Tilia*, *Ulmus*, *Betula* and *Pinus*) remain relatively unchanged in this phase. There is a further increase in Poaceae pollen, together with a minor rise in *Plantago lanceolata* indicating increasing stretches of open catchment used for pastoral and arable farming (*Rumex* and cereal pollen also still present). A very slight increase in Cyperaceae pollen at the top of the profile probably reflects increasing wetland areas surrounding the fjord, as sea level continues to decrease and the basin begins to fill with sediment. The sediment record ends at ~2,300 BC due to continual disturbance of sediments after this date, and therefore the late Holocene history (e.g. the transition from brackish to limnic and later terrestrial land) of Korup Sø remains uncertain.

5.8 Summary

- The diatom and molluscan marine records presented here suggest that the Korup Sø basin had already been subjected to marine water input by ~7,500 BC (in accord with previous results), but a permanent connection to the Kolinsund might not have been established until after ~7,400 BC.
- In the lower section of the record (i.e. ~7,300-5,500 BC) there is a close connection between salinity and sea level change. The early Littorina transgression phases (i.e. L1-L2 in Blekinge stratigraphy; 6,550-5,350 BC) cause DI-salinity to increase rapidly between ~7,300-5,640 BC, with the exception of a brief drop (~16-17 ‰) around

~6,000 BC (likely caused by a minor sea level regression). According to the MI-salinity reconstruction, a maximum marine phase occurs between ~5,870-3,940 BC, after which salinity begins to decline. However, the DI-salinity record suggests that salinity does not begin to decline until after ~3,400 BC. Absolute maximum DI-salinities (~27 ‰) occur between ~5,730-5,510 BC, and highest MI-salinity occurs (~25-26 ‰) between ~5.340-5,250 BC.

- After ~5,500 BC, DI-salinity becomes less sensitive to changes in sea level (e.g. no response to late Mesolithic transgression), likely due the high percentages of euryhaline *Fragilaria* s.l. present in the diatom record at this time. This genus is known for its ability to tolerate a wide range of (constantly fluctuating) environmental conditions and therefore high abundances might reflect generally instability within the fjord (e.g. fluctuating salinity, nutrient status, water depth, current strength, marine exposure, substrate) and possibly even skew salinity estimates. Alternatively, other factors might also be influencing salinity (perhaps offsetting the increase expected under rising sea level, e.g. greater freshwater input) in this section of the record.
- The multivariate 'moving window' analyses suggest that in the lower section (between ~7,330-3,400 BC) predictors for marine environmental change (e.g. sea level rise, catchment erosion and inwash, sedimentation rate, changing mollusc faunas) tend to explain relatively large amounts of the variation in the diatom dataset (up to 49 %). There is, however, a shift in the upper section of the profile (i.e. ~4,870-2,380 BC) as the diatom distributions within the fjord becomes more heavily affected by catchment changes associated with human impact (i.e. deforestation and expanding agricultural land likely affecting sediment loads and nutrient inputs into the fjord).
- A large increase in the accumulation rate occurs at ~3,900 BC, probably associated with increased sediment loads from the catchment following the elm decline (probably predominately caused by an elm pathogen) and subsequent restructuring of the catchment woodland. The first clear indicators of agriculture (i.e.cereal pollen) appear in the Korup Sø pollen record at ~3,850 BC, followed shortly afterwards (~3,800 BC) by increases in Poaceae pollen and the appearance of *Plantago lanceolata* and *Rumex*. The relatively early sign of agriculture in the Korup Sø catchment might suggest that humans were partly responsible for the elm decline.
- For the full diatom dataset, DCA axis 1 is heavily driven by salinity due to the high abundance of brackish water taxa at the very bottom of the profile prior to the rapid sea level rise associated with the Littorina transgression (beginning at ~6,550 BC). When the lowermost samples are removed, salinity and water depth appear to be the

most important variables, but neither can explain all the variation on DCA axis 1. A strong salinity gradient is apparent on DCA axis 1 for the mollusc dataset. For the pollen dataset, PCA axis 1 appears to be a temporal trend, probably driven by a variety of factors including climate (particularly long-term temperature), edaphic conditions and human impact. PCA axis 2 is likely to be driven by a light-shade gradient with dense closed canopy tree taxa most abundant in the Mesolithic scoring highly and taxa indicative of more open habitats and more abundant in the Neolithic period scoring low/negative. At the very bottom of the profile (before ~7,000 BC), it is likely that erosion and inwash from the catchment is also important for explaining some of the variation in the pollen dataset captured on PCA axis 2. This is supported by the exaggerated abundances of *Drypoteris* (fern) spores, Poaceae (most likely from wetland grass species) and Cyperaceae (sedges) in the lowermost section of the pollen record (~7,600-5,800 BC), which suggests that extensive wetlands (including swamp and fenlands) either surrounded the basin or situated in close proximity (e.g. in low-lying areas of the Kolindsund Fjord catchment) to Korup Sø at this time. Erosion and later sub-mergence of these wetlands (resulting in greater inwash of material) was likely to be caused by rapidly rising sea levels characteristic of the early Holocene (e.g. particularly between ~6,550-5,800 BC associated with stages L1 and L2 in the Belkinge sea level startigraphy).

- After ~3,400 BC, salinity begins to decline and the fjord continues to get shallower in response to declining sea levels and high sediment loads (as suggested by the high accumulation rate after ~3,900 BC). This is supported by the reappearance of *Ruppia* beds, an increase in benthic diatom taxa (accounting for >83 % of the diatom after ~3,400 BC) and the appearance of low-salinity tolerant mollusc species. Continuous disturbance of the sedimentary record after ~2,300 BC prevents assessment of the late-Holocene history at Korup Sø.

Chapter 6. Discussion

6.1 Introduction

In Chapters 3 - 5, the environmental history of each study site (Kilen, Norsminde Fjord and Korup Sø) has been independently discussed, with particular focus on localised changes within the fjord or immediate catchment area. Whilst some attention has been given to regional influences (e.g. sea level change, temperature) these have largely been discussed in the context of their potential impact upon a particular site. In this chapter, the accuracy of the presented palaeoenvironmental reconstructions (e.g. DI-salinity inferences) is assessed, and results from the individual site profiles are considered in a more regional context (i.e. comparison of results between sites and with other regional records of environmental change). The environmental changes identified in this thesis are also discussed with respect to their likely impact upon cultures and societies living in Denmark over the Holocene and existing environmental hypotheses for cultural change are reviewed (e.g. Rowley-Conwy, 1984). The specific aims of this chapter are:

1. To assess critically the accuracy and quality of the palaeoenvironmental reconstructions presented in Chapters 3 - 5.
2. To compare results with other palaeoenvironmental records and establish if there are any regional patterns of environmental change (in the sedimentary regime, salinity, nutrients and/or productivity) in coastal Denmark over the Holocene.
3. To attempt to explain what caused the major regional environmental changes (i.e. forcing factors) identifiable in the sedimentary records.
4. To assess how the coastal environment of Denmark changed over key periods of cultural change and assess how influential such changes might have been in driving cultural and societal change (i.e. compare regional interactions between land, sea and society over the study period).
5. To evaluate how this study has improved our understanding of coastal environmental change in Denmark over the Holocene and identify key questions that require further research in future.

It is important to note that changes in the vegetation records (i.e. pollen and plant macrofossil records) are not discussed any further than the site summaries provided in Chapters 3 - 5. Only interactions between the terrestrial and marine environment (as detailed by these records) and any associations with cultural/societal changes are considered in this chapter. Good regional accounts of the vegetation (and climate) history for Denmark and southern Scandinavia can be found in the following references: Iversen, 1973; Odgaard, 1994; Andersen, 1995b; Snowball *et al.*, 2004; Rasmussen, 2005; Antonsson and Seppä, 2007.

6.2 Interactions between environmental variables and their impact upon biological communities

Isolating the impact of a single environmental variable such as salinity in systems where several variables are changing at once is extremely difficult (e.g. Underwood *et al.*, 1998; Zong and Horton, 1999; Anderson, 2000; Engstrom *et al.*, 2000; Muylaert *et al.*, 2000; Anderson *et al.*, 2008; Velle *et al.*, 2010). Variables often co-vary or exert indirect influences upon biological communities by driving changes in other important variables and/or processes. For example, Anderson, (2000) criticized diatom-based temperature reconstructions from freshwater lakes on the basis that indirect temperature related changes or processes (e.g. soil and vegetation changes, pH, nutrients dissolved organic carbon) can exert greater influence upon the diatom assemblage than temperature itself. A number of complex statistical techniques (e.g. multivariate techniques such as RDA and CCA, variance partitioning) are now often used to explore the major trends in past and present datasets and for isolation of the most important environmental variables driving changes in the biological dataset. This is particularly important for assessing the strength and accuracy of transfer function inferences and understanding when and where they perform well and cases where (and why) they do not.

Using transfer functions for environmental reconstructions is a popular technique, but beset with difficulties and potential sources of error, relating to both the statistical techniques and assumptions implicit (Juggins, 1992; Birks, 1998; Racca and Prairie, 2004; Telford *et al.*, 2004; Telford and Birks, 2005, 2009), and the complexity of natural biological systems (Underwood *et al.*, 1998; Anderson, 2000; Underwood and Provot, 2000; Velle *et al.*, 2010). Ryves *et al.*, (2004) highlight that often as much as 60-80 % of the variation in the dataset remains unexplained (e.g. Zong and Horton, 1999) and that a substantial amount of this information might be lost from the environmental dataset due to sampling strategies. Environmental datasets, such as the one used in this study (http://craticula.ncl.ac.uk/Molten/jsp/; Andrén *et al.*, 2007a) are often limited to a set number of carefully selected variables (e.g. ten environmental variables measured for MOLTEN sites; see section 2.10.2), which likely results in the omission of some factors that are important in explaining part of the variation in the biological dataset (e.g. iron, dissolved organic carbon, potential metal pollutants such as mercury or lead not measured in this coastal training set). Complex temporal (e.g. diurnal, seasonal) processes are often summarised by using either spot measurement taken at the time of sampling or averaged data sampled over a period of time (e.g. Juggins, 1992; Parsons *et al.*, 1999; Clarke *et al.*, 2003; Ryves *et al.*, 2004; Clarke *et al.*, 2006).

A significant proportion of the variation in biological communities might also be explained by spatial or temporal characteristics (e.g. Lapointe, 2000; Muylaert *et al.*, 2000). In systems which exhibit high spatial variability, large training sets are needed to fully capture true relationships between biological communities and key environmental parameters (e.g. Lapointe, 2000). Large datasets, however, are likely to introduce additional causes of variation (e.g. differences in site configuration, catchment lithology and vegetation cover, human impact). For a successful salinity transfer function, ideally a wide salinity gradient will be sampled, over which all other environmental factors are ‘frozen’ (i.e. exhibit minimal change throughout the dataset). It is particularly difficult to produce such ideal training sets in open coastal and estuarine systems, due the operation of a large number of factors important in driving these systems, and common co-variation of variables (e.g. salinity and nutrients; Underwood *et al.*, 1998; Ryves *et al.*, 2004). For large datasets spanning a wide geographical range, ‘freezing’ of environmental variables also becomes increasingly difficult. There is also the additional problem that modern species-environmental relationships may not hold true in the past. For example, in modern training sets there might be stressors from several environmental variables (e.g. nutrient enrichment, metal and chemical pollutants, acidification) that did not exist in the past.

There are a number of taxonomic issues which might lead to further sources of error within environmental-species training sets. Producing large training sets which capture all the necessary ecological information is labour intensive, often requiring several taxonomists (e.g. MOLTEN, DEFINE and DETECT projects) and subject to potential human error. Additionally, taxa abundant within a fossil dataset might be absent or rare in the contemporary training set for a variety of reasons (e.g. previously unidentified taxa, or taxonomic revisions in later floras, or taxa which have become locally or regionally extinct, or misidentified). In coastal system, often poor preservation and increased dissolution of diatom valves can cause further problems (e.g. difficulties with taxonomic identification, underrepresentation or overrepresentation of taxa). Taxonomic and preservation issues are discussed in more detail below, with specific reference to the environmental and fossil datasets involved in this study.

Despite all these problems, transfer functions can perform well in coastal and estuarine systems and high quality reconstructions of parameters such as salinity and nutrient status have been produced in a number of Danish and Baltic coastal systems (Clarke *et al.*, 2003; Ryves *et al.*, 2004; Weckström *et al.*, 2004; Clarke *et al.*, 2006; Ellegaard *et al.*, 2006; Weckström, 2006; Andrén *et al.*, 2007a). In some cases diatom inferences have been compared to short-term monitoring data (e.g. Ellegaard *et al.*, 2006; Andrén *et al.*, 2007a) or longer-term documentary records (Ryves *et al.*, 2004), which have verified that transfer function inference can produce accurate reconstruction in some circumstances.

Diatom-salinity model

Salinity, depth, total nitrogen (TN) and total phosphorus (TP) explain 16 % of the variation in the modern day training set presented here (based on 210 MOLTEN and DEFINE sites, salinity gradient 0-31 ‰, 318 species; A.L. Clarke, *et al.* unpub.). Salinity is the single most important variable uniquely explaining 6.58 % of the variance, when TN, TP and depth are partialled out. Depth was also important, uniquely explaining 3.26 %, whilst TN and TP accounted for 1.96 and 1.85 % respectively. These four variables have also been identified in other coastal Danish training sets (and sedimentary profiles) as important drivers of the change within diatom communities (Jiang, 1996; Jiang *et al.*, 1998; Clarke *et al.*, 2003; Ryves *et al.*, 2004; Ellegaard *et al.*, 2006).

A strong association between salinity and the distribution of biological organisms (such as diatoms, molluscs, foraminifera) has been demonstrated for the inner Danish coastal waters and the transitional waters between the Baltic Sea and North Sea (e.g. Sorgenfrei, 1958; Jiang, 1996; Petersen, 2004; Ryves *et al.*, 2004). Diatoms have been shown to be sensitive indicators of salinity, exhibiting change to osmotic stress over short and long salinity gradients in both freshwater and brackish-marine systems (e.g. Fritz, 1990; Juggins, 1992; Gasse *et al.*, 1995; Jiang, 1996; Underwood *et al.*, 1998; Parsons *et al.*, 1999; Ryves *et al.*, 2002; Ryves *et al.*, 2004). The relationship between diatom distributions and salinity is not however necessarily linear, with often small changes in freshwater or weakly brackish systems causing a similar magnitude of species turn over as large changes might in higher salinity marine systems (van der Werff and Huls, 1958-1974; Ryves *et al.*, 2004). It is also important to note that whilst salinity might be the most important variable, in this study higher salinity conditions tend to be associated with intensified marine conditions and/or large-scale environmental changes. For example, intensified marine conditions in Kilen occur between ~3,100-0 BC, driven by increased exposure to the North Sea and/or Skagerrak (see Chapter 3). Greater exposure to the North Sea is also likely to affect current systems, energy, productivity (Figure 6.3), accumulation rates, sediment type, and in some cases diatom preservation (see below). Salinity can also affect macrophyte abundances (important for epiphytic diatoms) and grazing populations such as zooplankton (e.g. Amsinck *et al.*, 2003), mollusc populations (Petersen, 2004; this study) and fish (Enghoff, 1999; Holm and Bager, 2002), all of which can exert further pressure upon diatom communities and influence the accuracy of single-parameter diatom-based salinity reconstructions. This is apparent in this study at Kilen between ~AD 0-600, with increased abundances of epiphytic diatoms coinciding with a regular occurence of *Zannichellia palustris* in the plant macrofossil record in response to lower salinity, decreased marine exposure and potentially shallower water conditions (see section 3.11).

In coastal systems, water depth might be considered as an indirect environmental influence upon the diatom communities with associated factors such as light and UV penetration, habitat availability, macrophyte cover, energy and wave (and perhaps tidal) action exerting more direct influences (Campeaua *et al.*, 1999; Wulff *et al.*, 2005; Weckström and Juggins, 2006). Increased water depth will reduce light penetration to benthic areas limiting macrophyte and benthic organisms (e.g. Carter and Rybicki, 1990; Kurtz *et al.*, 2003). In coastal and estuarine systems, changes in sea level will also likely affect the salinity, energy, nutrients and oxygen conditions within the fjord or estuary. In the modern training set, there is very little co-variation between depth and salinity (<0.5 % under variance partitioning), but depth appears to be important in the sedimentary records and is likely to be associated with salinity, particularly at the lower end of the salinity gradient (e.g. before the marine transgression and in the upper section of the Korup Sø record; see Chapters 4 and 5).

Only a small amount of variation in the diatom training set is explained by TN or TP, with the latter showing a small amount of co-variation (0.87 % using variance partitioning, Borcard *et al.*, 1992) with salinity. This is likely to result from the source of higher salinity water (i.e. from the North Sea) being also richer in phosphorus. The small covariation between salinity and phosphorus is deemed unlikely to cause major changes in DI-salinity inferences using this model. Furthermore, the modern training set spans a large salinity gradient, with salinity uniquely explaining far more of the variation in the diatom dataset and therefore probably overriding the effects of TP on the diatom community. In terms of the role played by nutrients in determining planktonic growth, TN is generally considered far more important than TP in coastal and estuarine systems (Nixon *et al.*, 1996), though TP can become a limiting factor for phytoplankton growth, particularly in the spring (Conley, 2000 and references therein). Seasonal or year round TP limitation appears to occur more often in the lower salinity areas of the Baltic Sea such as the Bothinian Bay or in close proximity to large freshwater inputs such as the Himmefjärden estuary (Andersson *et al.*, 1996; Conley, 2000). Recent coastal and estuarine systems from Denmark often contain high TN concentrations due to anthropogenic loading, but the long-term history of TN in these systems (prior to recent major anthropogenic disturbance) is unknown. Whilst the small amount of variation explained by TN (<2 %) suggests it is not particularly important in explaining the variation in the modern diatom dataset, this does not mean that TN has not been important sometime in the past (e.g. Bianchi *et al.*, 2000). Care must therefore be taken when interpreting DI-salinities in the sedimentary fossil records, particularly when potential key indicators of changing nutrient status are present (e.g. increased planktonic taxa, high abundances of *Cyclotella choctwatcheeana*) in the record. At Kilen, sedimentary pigment analysis has also been performed to examine changes independently in productivity and nutrient status.

Taxonomic problems and representation of fossil taxa in the modern training set are also key issues, which might weaken transfer function inferences. Over the course of the MOLTEN, DEFINE and DETECT projects a series of taxonomic workshops were held in order to maintain taxonomic consistency and harmonize any differences in taxonomic interpretation. In several cases, aggregate taxa were produced by clumping two or more species together (see Andrén *et al.*, 2007a), which might have resulted in the loss of some ecological information. Diatoms from the genera *Chaetoceros* and *Rhizosolenia* are also problematic, often being extremely important components of seasonal planktonic cycles in coastal and estuarine systems, yet their vegetative cells often preserve poorly in surface sediments (e.g. MOLTEN training sites) and the sedimentary record (rare at all three sites analysed here). These taxa are generally represented by resting cysts (*Chaetoceros*) or remnants of spines (*Rhizosolenia*) which are largely unidentifiable to species level, but can account for significant proportions of the diatom community. Both genera contain a large number of species, which inhabit a wide range of environments (Rines and Hargraves, 1988; Hasle and Syvertsen, 1996), making them ambiguous salinity indicators (unless they can be successfully identified to species level). For both the modern and fossil training sets, all *Chaetoceros* cysts were clumped and therefore must also be omitted from the model. In this study, there are also a large number of other taxa present in the fossil sediments that are absent from the training set. In addition, high abundances of ubiquitous euryhaline taxa (most notably *Fragilaria* s.l. in this study) make it increasingly difficult to isolate a true environmental signal (see Chapters 4 and 5).

Classically, the predictive power of a transfer function model is evaluated by its regression coefficient (r^2 for observed versus inferred values) and its root mean square error of prediction (RMSEP) following cross-validation (usually via bootstrapping or jackknifing). This criterion was used for model selection in this study with the WA-PLS model outperforming other methods tested including classic and inverse weighted average models, modern analogue technique (MAT), locally weighted-weighted averaging (LW-WA) and maximum likelihood models. An additional problem with WA-PLS models (particularly over WA techniques) is that greater care is needed in model selection and model evaluation (Birks, 1998) in order to choose the correct number of components (additional components improve the 'fit' of data by improving the r^2 and lowering the RMSEP, but can introduce 'apparent' statistics that have little or no predictive value; Birks, 1995; Birks, 1998). In statistical modelling the aim is to choose the minimum adequate model (Crawley, 1993; Birks, 1998) and therefore a common cut-off for the addition of components is if a component reduces the prediction error by 5 % or more under internal validation (ter Braak and Juggins, 1993). Whilst the WA-PLS component-3 model employed here only improved RMSEP by ~2 %, it was chosen because it was shown to 'significantly' lower the RMSEP from the WA-PLS model component 1 and 2 models under a randomised t-test ($p<0.05$) incorporated into the C2 package (Juggins, 1991-2009) and produced more realistic species estimates/coefficients.

Problems associated with model selection can be exaggerated in spatially structured environments as close similarities between neighbouring sites (e.g. similar environmental conditions and species present) can lead to spatial autocorrelation, which violates assumptions underlying many statistical tests (Telford and Birks, 2005, 2009). Spatial autocorrelation can result in over-optimistic (and therefore unrealistic) estimates of model predictive power (Telford and Birks, 2005, 2009) and subsequently lead to inappropriate model selection and unreliable inferences from fossil data. Over optimistic predictive power was demonstrated by Telford and Birks, (2009) for several different transfer function models based on calibration sets exhibiting differing degrees of spatial autocorrelation, with the RMSEP of WA-PLS (and MAT) models being substantially higher (and r^2 much lower) when 'neighbouring' sites were removed from the training sets (i.e. to reduce spatial structuring). No independent test for spatial autocorrelation was employed in this study (e.g. *h*-block: Telford and Birks, 2009) but it is argued that the selected WA-PLS component 3 model produces reasonable salinity estimates that are more reliable than the other tested models. Under preliminary assessment (via comparison with other proxy records, subjective assessment of past salinity and based on ecological knowledge) all other tested models appear to underestimate salinity, with the modern analogue techniques and maximum likelihood models performing particularly poorly in almost all circumstances. It is, however, acknowledged that a degree of spatial autocorrelation is likely to be inherent in the MOTEN-DEFINE-DETECT dataset and that in future this should be properly tested (e.g. Telford and Birks, 2009) and subsequently the RMSEP and r^2 values for different transfer function models should be re-evaluated correcting for spatial autocorrelation. The final applied DI-salinity model should then be selected accordingly based on these new estimates of predictive power. Future salinity inferences using this training set might also benefit from using a consensus reconstruction approach (e.g. Bartlein and Whitlock, 1993) based on all (reasonable) model reconstructions, as it is clear that in certain circumstances the WA-PLS-component 3 model employed here also performs poorly (e.g. overestimation of some DI-salinity values (>35 ‰) at Kilen in the late pre-Roman period, ~250-0 BC).

Nevertheless, using the ‘classical’ approach the WA-PLS component-3 model used in this study appears to have a high predictive power (i.e. performs well under internal validation by bootstrapping with 1000 cycles; $r^2_{boot} = 0.97$ and RMSEP = 0.36 square root salinity units), comparing well with other published salinity-based transfer functions (e.g. Ryves *et al.*, 2004; $r^2_{jackknife}=0.887$, RMSEP=0.246 log salinity, Wachnicka *et al.*, 2010; $r^2_{jackknife}=0.95$, RMSEP=0.39 square root salinity units). The application of the diatom based salinity model (see above and section 2.10.2) to the three fossil records analysed in this study is therefore deemed to have produced reasonably accurate reconstructions of salinity change (qualitatively confirmed by ecological knowledge) over the study period. It is important to distinguish between patterns of salinity change and precise quantitative estimates for salinity, as the latter are extremely difficult to verify, particularly in physical coastal systems, influenced by many variables (see above) and host to high abundances of euryhaline taxa (e.g. *Fragilaria* s.l.). Existing palaeosalinity records for the Limfjord are only available for the last ~2,500 years, over which period the Kilen DI-salinity record is in relatively good agreement with other published records of salinity change (Kristensen *et al.*, 1995; Christensen *et al.*, 2004; discussed in more detail below). Salinity estimates based on foraminifera and molluscan analyses (Kristensen *et al.*, 1995; Christensen *et al.*, 2004) are also in good accord with the

DI-reconstructed values for Kilen, suggesting this model performs well, at least in the upper section of the Kilen profile. For Norsminde, reconstructed DI-salinity values in the reduced model (excluding *Fragilaria* s.l.) are also generally in accord with 'expected' salinities during the mid-Holocene, placing the salinity (~19-22 ‰) in the Mesolithic at slightly higher than present day (~18-20 ‰; Muus, 1967). For Korup Sø, DI-salinity inferences might be slightly lower than previously thought (e.g. Petersen, 1993) and in disagreement with the previous suggestion that lower salinity existed in the early Neolithic period (discussed in more detail below). It is, however, highly likely that the diatom record is producing more realistic estimates, based on the location of Korup Sø (i.e. right at the back of a southward limb of the Kolindsund Fjord; Figure 2.5) and the lack of major floral shifts in the Neolithic period (at both Korup Sø and Norsminde).

The diatom-salinity model appears to perform poorly at the high end of the salinity gradient, where salinity appears to be overestimated (e.g. DI-salinity inferences at Kilen >35 ‰ ~250-0 BC; see Chapter 3). Higher salinity phases are also often marked by high (and likely exaggerated) abundances of robust *Paralia sulcata* valves which might be skewing the DI-salinity estimates (see below). Across the salinity gradient, high abundances of euryhaline *Fragilaria* s.l at both Norsminde and Korup Sø are also likely to be affecting the accuracy of the DI-reconstruction at these sites. However at Norsminde, both the full model (i.e. including eurhaline *Fragilaria* s.l.) and reduced model produce largely similar curves (though inferred values differ), which suggests that the high (and fluctuating) abundances of *Fragilaria* s.l. are not completely muting/overriding the salinity signal at this site.

6.2.1 Diatom preservation

The differential susceptibility of diatom valves to dissolution can be problematic for the creation of high quality modern day training sets that accurately detail species ecological distributions (Fritz, 1990; Reed, 1998; Ryves *et al.*, 2004; Ryves *et al.*, 2006). For example, salinity training sets from both coastal and limnic systems have demonstrated close correlation between salinity and diatom dissolution (e.g. Ryves *et al.*, 2004; Ryves *et al.*, 2006), with a general consensus that diatom preservation is poorer under higher salinity conditions. In addition to salinity, physical processes (erosion, wave and tidal action, sediment re-working, sediment transportation), water chemistry, pH and bacterial breakdown of silica and stratification pattern (e.g. meromixis) can also affect the rate of diatom dissolution in aquatic systems (Ryves *et al.*, 2009). Such processes can often be exaggerated in coastal and estuarine systems (i.e. naturally complex and high energy environments), making fossil organisms particularly susceptible to dissolution in these environments. Understanding the impact of dissolution upon a fossil diatom assemblage is paramount to assessing the quality of the dataset, and ultimately, the confidence in palaeoenvironmental reconstructions (e.g. Ryves *et al.,* 2004).

Ryves *et al.*, 2004 found a strongly positive relationship ($r^2 = 0.33$-0.74) between diatom dissolution (as measured by the **F** index) and salinity in the Vejlerne (Limfjord) sedimentary profiles. There was also a tendency towards lower **F** index values (i.e. poorer preservation) as salinity increased in the lake samples from the training set, but valve preservation was not assessed in the higher salinity fjord sites. Unfortunately, diatom dissolution was not measured for the MOLTEN, DETECT or DEFINE training sites and therefore its influence upon the diatom distribution (i.e. the amount of variance explainable by dissolution) in the modern environmental dataset cannot be assessed. Sites heavily affected by dissolution would probably be identifiable as outliers in the training set (e.g. likely contain high abundances of large heavily silicified diatoms and overall low concentration). Nevertheless, it is likely that dissolution explains at least some of the variation in the MOLTEN/DETECT derived modern training set, which might add a further minor source of error to DI-salinity inferences in some sections of the record. This is most likely to occur at the high end of the salinity gradient, where near-fully marine conditions exist (e.g. at Kilen during the pre-Roman period, ~500-0 BC).

For the fossil dataset, **F** index sample scores can be directly plotted against the DI-salinity values (assuming reasonably accurate DI-salinity reconstruction; see above) in order to assess the relationship between these variables. The relationship between DI-salinity and diatom **F** index score was very weak (and not significant) for both Kilen ($r=0.14$, $p>0.05$) and Korup Sø ($r= 0.05$, $p>0.05$, with an inverse relationship (i.e. negative correlation) between salinity and dissolution), but for Norsminde Fjord, it was only weak ($r=0.04$, $p>0.05$; not significant) after the low salinity (fresh-brackish) samples at the bottom of the profile (i.e. 6,900-6,600 BC) were removed. For the full Norsminde dataset, there is relatively strong (significant) inverse gradient ($r=0.52$, $p<0.05$), with greater dissolution (i.e. lowest **F** index) occurring under lower salinity. This is driven by poor preservation in the lowermost salinity samples (i.e. prior to the rapid salinity rise associated with the Littorina transgression (6,900-6,600 BC; see section 4.11). In this period, Norsminde Fjord is characterised by shallow water (but subjected to rising sea level and salinity), catchment erosion, sedimentary mixing, transport and re-deposition (see section 4.11), all of which are likely to increase dissolution. After ~6,360 BC, the preservation at Norsminde Fjord does not change systematically and is generally good (above **F** = 0.4), but subject to occasional fluctuations. Canonical correspondence analysis (CCA) suggests that dissolution significantly explains 10 % of the variation in the Norsminde Fjord diatom dataset (from 38 samples for which an **F** index assessment is available). When the lowermost 6 samples are removed (i.e. ~6,900-6,600 BC) the variance explained by dissolution is insignificant ($p > 0.05$; Monte Carlo test; 999 permutations).

Periods of poorer diatom preservation also exist in sections of both the Kilen and Korup Sø records, which might subsequently affect the accuracy of the DI-reconstructions (Figure 6.2). CCA analysis of the Kilen diatom record suggests dissolution is significant ($p < 0.05$) in explaining ~ 5 % of the variation in the diatom dataset. At Kilen, poorer preservation tends to exist throughout the more marine phase (i.e. lower **F** index values, generally **F** = 0.2-0.5) between ~3,500-0 BC, probably associated with higher salinity and/or greater exposure to the North Sea and subsequently greater energy and coarser sediments. Absolute lowest values coincide (**F** = 0.2-0.3) with the late pre-Roman salinity maxima, when common North Sea diatoms (e.g. *Delphineis minutissima* and *Cymatosira belgica*) become abundant in the diatom record. Therefore, salinity is probably also important, but relatively low **F** index values in the succeeding brackish period (~AD 0-600), probably weakens the apparent dissolution-salinity relationship (i.e. $r=0.34$, $p<0.05$ (significant) when brackish samples from the Roman/Germanic period are removed). Poorer preservation in the Roman/Germanic period might be related to lower accumulation rates (i.e. slower burial rates) or increased predation (e.g. higher molluscan abundances)..

CCA suggests that dissolution significantly also explains ~5 % of the variation in the Korup Sø diatom dataset. This is likely to be driven by the poorer preservation in the upper sediments (**F** index = ~0.25-0.45 after ~3,400 BC) of the Korup Sø profile, particularly as salinity begins to decrease, but perhaps more importantly as the basin begins to shallow, due to decreasing sea levels and sediment infilling. Poorer preservation probably results from greater volumes of sediment in the basin and sediment disturbance under shallower conditions (e.g. movement, mixing and reworking of valves, particularly during storm events, strong currents or sea surges). This period is characterised by extremely high percentages of benthic diatoms, particularly *Fragilaria* s.l. which are generally insensitive salinity indicators and likely reflect the dynamic, unstable nature of the Korup Sø environment at this time. There are also several samples with reduced **F** index values (**F** = < 0.4) between ~5,930-5,470 BC, which broadly ties in with increasing sea level, salinity and high abundances of *Paralia sulcata* (see below).

In this study, higher salinity phases are often associated with higher abundance of *Paralia sulcata*, which probably skews the accuracy of salinity inferences at the high end of the salinity gradient. *P. sulcata* is present at a wide range of salinities, but generally reaches it highest abundances at higher salinities (Hendey, 1964; Cooper, 1995b; Jiang, 1996; Ryves *et al.*, 2004; this study), often in winter due to being able to cope with low light conditions (Hobson and McQuoid, 1997; McQuoid and Nordberg, 2003). Therefore, in cases where *P. sulcata* is abundant, higher salinity is probably a true signal, but the salinity estimates generated by the model value might be biased by potential overrepresentation due to the robust, heavily silicified nature of *P. sulcata* valves (Ryves *et al.*, 2004; see also Chapters 3-5). This is typified by the general lack of change in *P. sulcata* dissolution indices throughout any of the sedimentary profiles, even in cases where the **F** index decreases substantially (i.e. preservation conditions deteriorate). Alternatively, confidence in DI-salinity reconstructions might be increased when weakly silicifed diatoms such as *Skeletonema costatum* or *Chaetoceros* vegetative cells (very seldom found in the present study) are preserved in the sediments.

6.3 Environmental change in coastal Denmark and interactions between land, sea and society; ~7,600-1,700 BC

Obtaining high quality palaeoenvironmental records from dynamic coastal and estuarine systems is extremely challenging as a wide range of problems must be overcome (e.g. coring logistics, chronological issues (see section 2.2.1), sediment mixing and disturbance, taphonomic problems). This study has demonstrated that well preserved, high resolution sedimentary records, securely datable via ^{14}C dating of terrestrial material, are obtainable from Danish fjord and estuarine systems. These records also offer potential for multiproxy analyses (i.e. preservation of the remains of a wide range of organisms) including proxies detailing changes in both the marine (e.g. diatoms, foraminifera) and terrestrial (pollen and plant macrofossil) environment. Such archives are therefore invaluable for testing environmental push hypotheses for cultural change (Rowley-Conwy, 1984) and for understanding how large scale changes in oceanic (and terrestrial) systems drive changes in commercially important coastal areas which continue to be foci for human settlement and activity (both at present and in the past; Holm and Bager, 2002; MacKenzie *et al.*, 2002; Andersen, 2007; MacKenzie *et al.*, 2007; Poulsen *et al.*, 2007; Andersen, 2008a).

In Figure 6.1-Figure 6.3, selected independent proxy records (analysed in this study) are plotted on common temporal scales (i.e. calendar years, BC; based on age-depth models) to compare changes in the sedimentary regime (Figure 6.1), salinity (Figure 6.2) and productivity/nutrients (Figure 6.3) at each site. The similarities (and differences) between these sedimentary records are discussed below (in chronological order), with comparisons also being made to other palaeoenvironmental records from the region (e.g. Petersen, 1981; Berglund *et al.*, 2005; Emeis, 2005; Antonsson and Seppä, 2007; Rasmussen *et al.*, 2007) and to the archaeological records documenting cultural change (e.g. Andersen, 1989; Andersen and Rasmussen, 1991; Andersen, 1995a; Enghoff, 1995; Andersen, 2005; Andersen, 2007). The broad aims of this section are two fold; firstly to assess the extent to which the results from this study improve/contradict (i.e. by comparison with other regional studies) our previous understanding of environmental change in coastal areas of Denmark over the Holocene. Secondly, to consider how changes in the marine environment might have influenced past cultures and societies living in Denmark throughout the Holocene and to review critically environmental push hypotheses for cultural change (e.g. Rowley-Conwy, 1984) on the basis of the evidence presented here.

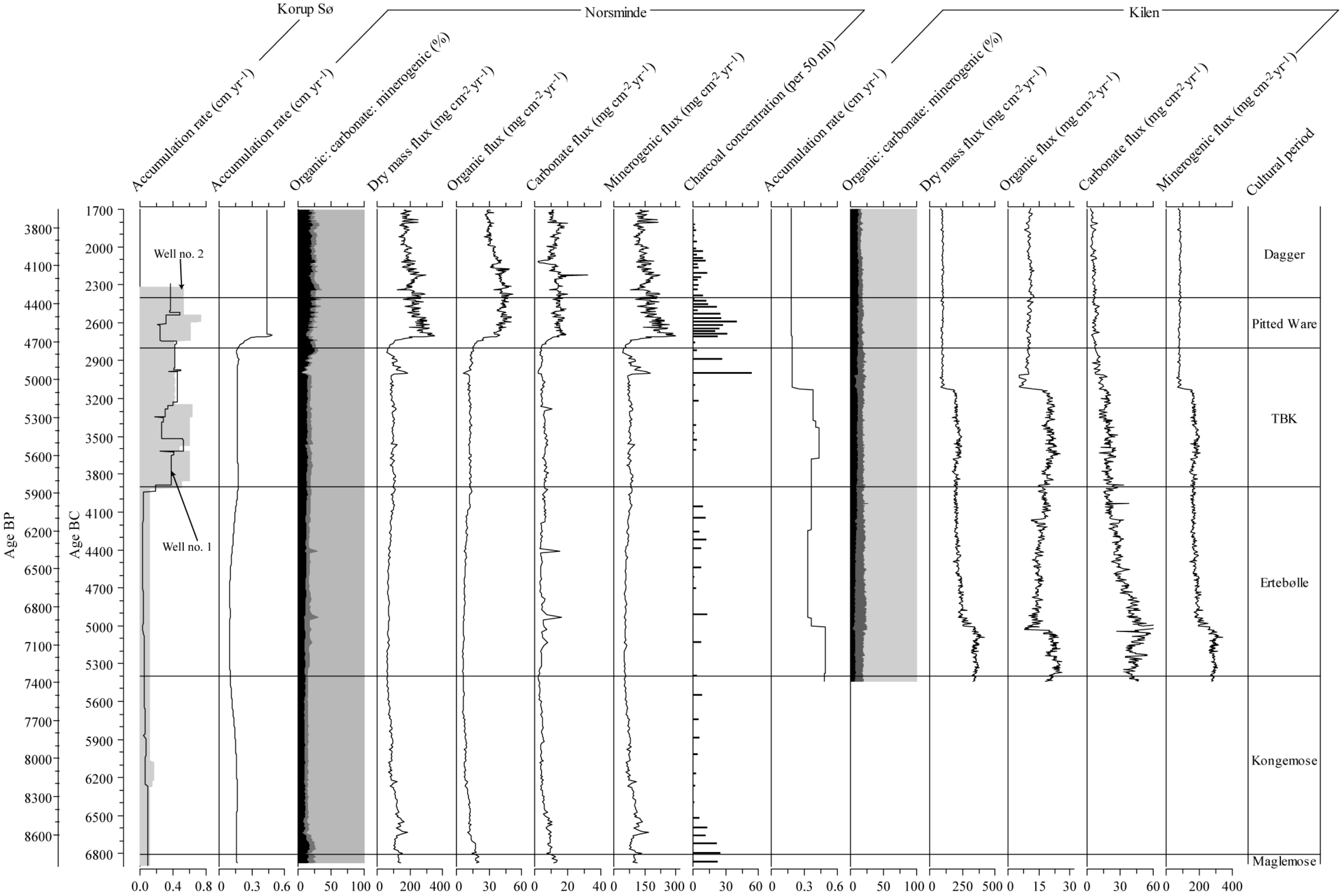


Figure 6.1- Sedimentary physical parameters for all three sites (Korup Sø, Norsminde Fjord and Kilen) plotted on a common temporal scale (~6,900-1,700 BC). Cultural division are based on those given in Hvass and Storgaard, 1993.

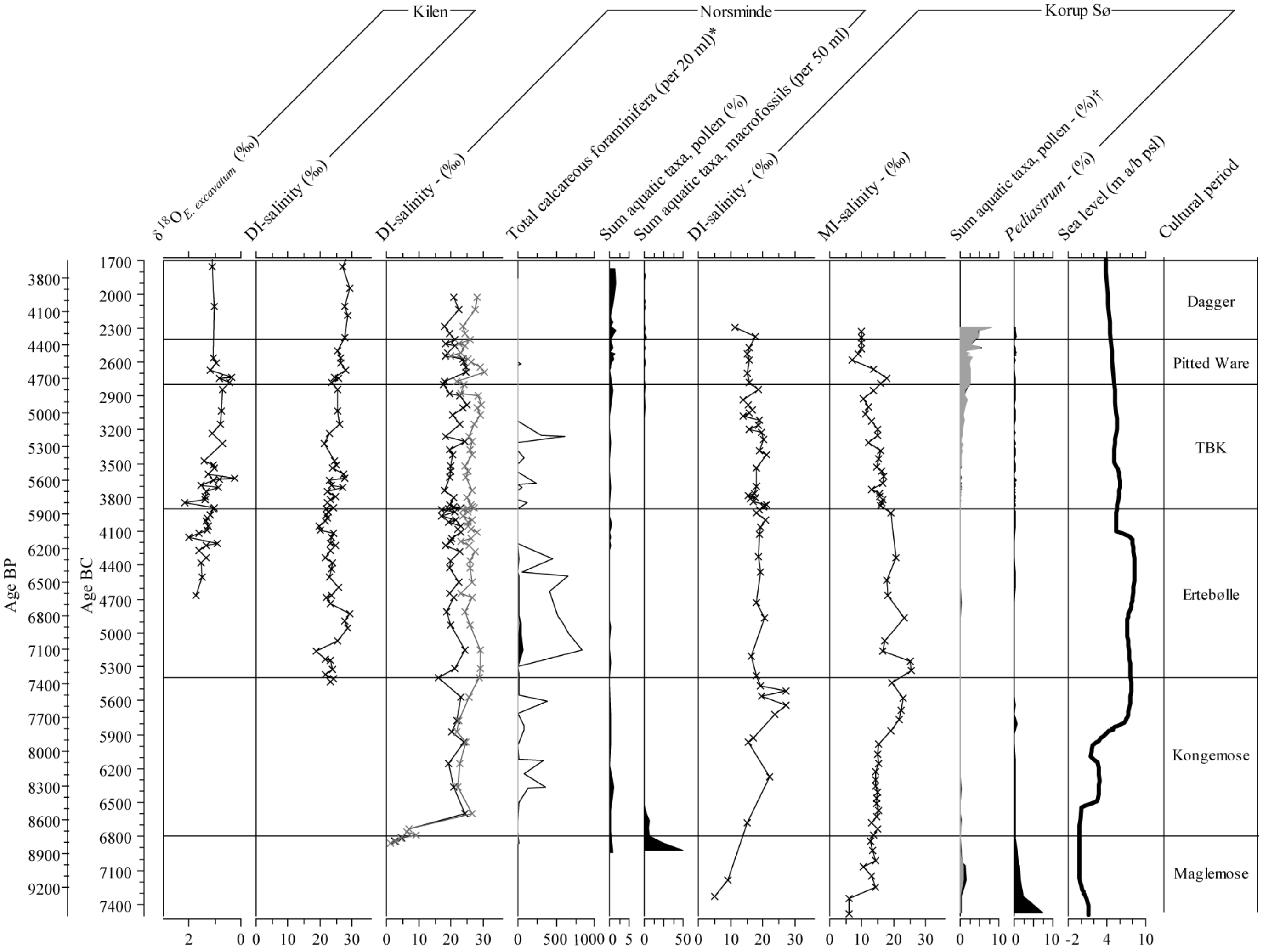


Figure 6.2- Selected parameters providing information about salinity at all three sites (Kilen, Norsminde Fjord, Korup Sø; ~7,400-1,700 BC). Cultural division are based on those given in Hvass and Storgaard, 1993. Regional sea level curve from Blekinge, SE Sweden (Berglund *et al.*, 2005, m a/b psl = metres above/below present sea level.). *Includes the high salinity demanding foraminifer *Elphidium incertum* (shaded black). † = Includes *Ruppia* spp. (shaded grey). DI-salinity = diatoms inferred salinity, MI-salinity = mollusc-inferred salinity.

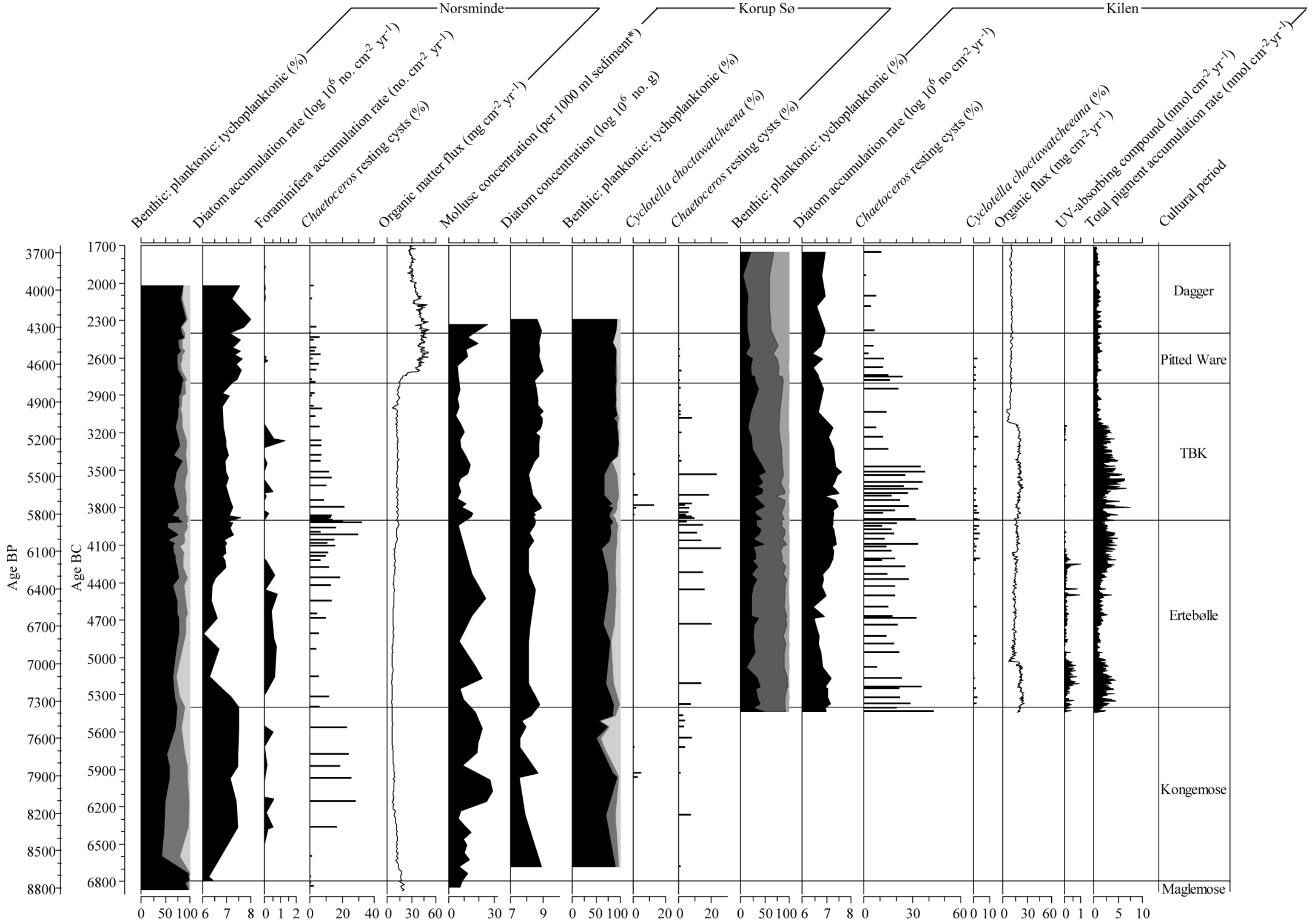


Figure 6.3 - Selected parameters providing information about productivity and/or nutrient status at all three sites (Kilen, Norsminde Fjord, Korup Sø; ~6,900-1,700 BC). Cultural division are based on those given in Hvass and Storgaard, 1993. *= Mollusc concentration data square root transformed for Korup Sø.

6.3.1 The marine transgression and the early Mesolithic Maglemose (9,800-6,800 BC) and Kongemose (6,800-5,400 BC) cultures

The ingression of marine water into Norsminde Fjord and Korup Sø appear to differ by ~200-300 years. Korup Sø was inundated first (between ~8,000-7,500 BC; Iversen, 1937; Petersen, 1993; this study), with a limnic phase identifiable, prior to its full submersion during the Littorina transgression (Iversen, 1937; Petersen, 1993, H. Krog unpublished data). Sediments containing lacustrine molluscs (e.g. *Anodonta* sp.) were originally dated to the Boreal period (8,000-7,000 BC) based on pollen analyses (H. Krog, unpub.). This is further supported in this study by the occurrence of a number of freshwater diatoms present in the lowermost sediments (identified via a preliminary scanning of slides). The first ingression of seawater into Århus Bay (Dencker, 1999; Jensen and Bennike, 2008) and Norsminde Fjord (Rasmussen, submitted; and this study) occurred approximately 6,900-6,800 BC, with again a likely freshwater (limnic or fluvial) phase prior to submersion. The Kilen record (presented here) does not extend back as far as the marine transgression and so the date of inundation can only be estimated based on regional records. Petersen, (1981) suggests a marine transgression of the Limfjord occurred between ~7,000-6,000 BC and that by ~5,000 BC, sea level in the Limfjord was ~3 m above its present day level. More recently a preliminary date from deeper sediments (~2,900 cm bpsl; not analysed as part of this thesis) collected from the Kilen basin indicate marine conditions (i.e. contain marine molluscs and diatoms) at ~6,930 BC. This suggests that a marine transgression occurred earlier than ~7,000 BC, at least in the Struer area of the Limfjord.

Prior to the Littorina transgression, the Maglemose culture inhabited Denmark and Scandinavia, but the extent to which this culture exploited coastal resources remains uncertain. The majority of Maglemose sites are based inland, often near rivers and lakes (Grøn, 1995), portraying them as an inland hunter gatherer culture, exploiting freshwater fish species (particularly pike and perch) and hunting for elk, aurochs, red deer, roe deer, wild boar and small fur animals. The Maglemose coastline was very different to that today, and most of the coastal sites belonging to the Maglemose culture were probably drowned during the Littorina transgression. During this transgression (beginning somewhere between ~6,700-6,400 BC), sea level rose by approximately ~28 m in the Kattegat and Storebælt, and ~15 m in the southern Baltic Sea in just six centuries (e.g. Mörner, 1976; Petersen, 1981; Christensen, 1995; Schmölcke *et al.*, 2006). This completely transformed the Baltic and

Kattegat coastlines, drowned the Øresund and Storebælt, creating a permanent marine connection to the Baltic Sea, and resulting in brackish-marine salinities within the Baltic basin (Winn *et al.*, 1998; Andrén *et al.*, 2000; Bennike *et al.*, 2004; Berglund *et al.*, 2005). Based on the DI-salinity (and mollusc inferred-salinity, MI-salinity) records presented here, by ~6,600 BC, the salinity was relatively high at both Norsminde Fjord (~25 ‰) and Korup Sø (~15 ‰), suggesting they were quickly inundated, very early in this rapid phase of sea level rise.

The succeeding Kongemose culture certainly exploited marine resources as detailed by the numerous coastal sites (including the earliest kitchen middens) belonging to this cultural group (e.g. Andersen, 1970a; Larsson, 1990). Therefore, in archaeological terms, the Littorina transgression really marks the transition between the Maglemose and Kongemose cultures (e.g. Christensen, 1995). The Littorina transgression completely changed the landscape for humans, reducing the total land area, cutting off the land-bridge between Denmark and Sweden and likely submerged or re-configured many formerly important freshwater bodies (lakes and fluvial systems). As the marine waters poured into low lying areas such as former glacially-scoured troughs and basins, shallow accessible fjords and estuaries were created. Increasing temperatures (Iversen, 1944; Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Brown *et al.*, submitted) and salinity (Mikkelsen, 1949; Petersen, 1993; this study) enabled edible marine fish and molluscs (e.g. mussels and cockles; Figure 5.6) to colonise these waters providing valuable resources for the Kongemose. It appears that the Kongemose quickly recognised the prosperity of the marine environment, more regularly inhabiting coastal areas and increasingly gearing their economy towards its resources (e.g. Andersen, 1970a, 2007).

6.3.2 From Ertebølle (~5,400-3,900 BC) to Funnel Beaker Culture (TBK; ~3,900-2,800 BC) and the Mesolithic-Neolithic transition (~3,900 BC)

The beginning of the Ertebølle period (~5,400 BC) broadly coincides with an acceleration in sea level rise in the Baltic Sea, Kattegat (Mörner, 1976; Christensen, 1995; Berglund *et al.*, 2005; Brenner, 2005; Yu *et al.*, 2007) and other globally distributed sea level records (Blanchon and Shaw, 1995; Blanchon *et al.*, 2002; Siddall *et al.*, 2003; Sloss *et al.*, 2005; Behre, 2007; and see section 1.3). It is likely that this rapid rise in sea level (between ~5,600-5,000 BP) triggered the large increase in shell midden accumulation in the late

Kongemose/Ertebølle period. Several factors associated with this sea level rise might have been responsible for increased cultural exploitation, including increased salinity (initially), changing sedimentary conditions (e.g. lower accumulation rates between ~6,000-3,900 BC at Norsminde Fjord and 6,200-3,900 BC at Korup Sø; Figure 6.1) and an overall increase in marine productivity. Increasing temperature is also likely to have been important as Scandinavia entered the Holocene thermal optimum (Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Brown *et al.*, submitted).

The poorly constrained sea level history of the Limfjord makes it difficult to understand how this period of rapid sea level rise affected the Limfjord region. Petersen, (1981) suggests that by ~5,000 BC sea level was ~3 m above the present day sea level, but the dating of this study has been has been criticized by Christensen, (2001) and no other detailed sea level studies for the Limfjord cover this period. Based on the timings of marked changes at Korup Sø (i.e. salinity rise at ~5,600 BC; this study) and in the Baltic Sea (e.g. Brenner, 2005; Yu *et al.*, 2007), it is likely that rapid sea level rise would have impacted the Limfjord before ~5,400 BC, and most likely through its northern opening at this time (see section 3.11). This also ties in with the submergence of the oldest known Kongemose site from the Limfjord (~5,600 BC; e.g. Brovst, layer 11, Andersen, 1970a, 2007) and slightly later, the appearance of the kitchen middens from the Ertebølle culture (~5,400 BC; Andersen, 1995a). This almost entirely pre-dates the Kilen record presented here and so changes associated with this sea level rise at Kilen cannot be investigated. In the lowermost section of the Kilen record (i.e. between ~5,500-5,000 BC), there is very little change in DI salinity, suggesting that any potential impact upon salinity had either ceased by ~5,500 BC or by this point too small in magnitude to influence the Kilen profile. Cores collected more recently from Kilen (May 2008) offer future potential to investigate environmental changes at this site between ~6,930 BC and 5,500 BC according to the new AMS ^{14}C dates of this deeper part of the record, but this lies beyond the scope of this study.

Both the Korup Sø DI- and MI-salinity reconstructions suggest a relatively close connection between sea level rise in the Baltic Sea (Berglund *et al.*, 2005) and salinity change between ~6,000-5,200 BC (Figure 6.2). Increased salinity between ~5,880-5,160 BC is also apparent in the reduced Norsminde DI-salinity model, but not in the full model due to DI-salinity decline at ~5,400 BC (see section 4.11). The sea level-salinity connection, however, appears to weaken in the mid-Ertebølle period (after ~5,500 BC for Korup Sø and ~4,900 BC BC for

Norsminde Fjord under the reduced model), particularly highlighted by inferred maximum salinities occurring in the late Kongemose/early Ertebølle period (somewhere between ~5,600-4,900 BC; see below and Figure 6.2), several hundred years before the maximum sea levels of the late Atlantic period (4,800-4,200 BC; Berglund *et al.*, 2005). Whilst all proxies analysed in this study place the salinity maximum in this period (i.e. somewhere between ~5,600-4,900 BC), there are considerable differences between these estimates (Figure 6.2). At Korup Sø, both the DI and MI-salinity reconstruction suggest salinity begins to increase almost simultaneously (after ~6,000 BC), but highest DI-salinities (23-27 ‰; ~5,730-5,520 BC) occur before maximum MI-salinities (~5,340-5,250 BC). The earlier maximum in the DI-reconstruction is largely driven by high abundances of *Paralia sulcata* and therefore estimates might be exaggerated due to the robust nature of this diatom and often over representation in marine sediments (Ryves *et al.*, 2004). A drop in the **F** index is apparent in several samples between ~5,930-5,470 BC might further support this, but dissolution of *P. sulcata* valves remains unchanged (see section 5.7). After ~5,520 BC, *P. sulcata* percentages drop as *Fragilaria* s.l become more abundant, suggesting that preservation might have improved and that the system is subject to instability, perhaps caused by variable salinity, nutrients and productivity (see section 5.7).

At Norsminde, highest DI-salinities under the reduced model (~5,400-5,160 BC) also occur before the high-salinity demanding species of foraminifera (particularly *Elphidium incertum* and *Elphidium voorthuysenii*) reach their maximum abundances (~5,160-4,900 BC). It is difficult to explain the drop in salinity after ~4,900 BC, perhaps being related to the minor drop in sea level between transgressions (i.e. minor regression between L2 and L3 transgressions in Berglund *et al.*, 2005) or the drowning of low lying shallow marsh areas which might form productive habitats for diatoms, giving a fresher water signal, than the ‘higher-salinity’ conditions in the main part of the fjord itself. If the latter argument was true, however, then higher abundances of wet ground/marsh plants/pollen might be expected, but there is very little evidence of this in the pollen or plant macrofossil records. Alternatively, the lower salinities after ~4,900 BC might suggest that other factors exert greater influence upon salinity than sea level in the mid-late Ertebølle period. Isolating a clear salinity signal in this section of the records is extremely difficult due to the dynamic nature of the marine environment at this time, with apparent variations in salinity (Figure 6.2), nutrient status, productivity (Figure 6.3) and sedimentation rates (Figure 6.1) all occurring. All these factors might drive changes in the diatom communities and have implications for producing accurate estimates of salinity over this period. This is exemplified by the regular salinity fluctuations (particularly in the Norsminde Fjord record; Figure 6.2) and the high numbers of *Fragilaria* s.l. present in both the Norsminde Fjord and Korup Sø records, which are highly adapted to living in unstable environments (Eronen, 1974; Stabell, 1985). It is important to note that despite decreasing in the mid-late Ertebølle period, all salinity proxies suggest that salinity remained higher (i.e. more marine) than in the earlier Magelmose/early Kongemose periods (i.e. ~7,200-5,900 BC). This probably explains the continued presence of higher-salinity demanding organisms (absent in the earlier periods), higher marine productivity and continued intense human exploitation of marine resources throughout the entire Ertebølle phase (e.g. Andersen, 1989; Andersen, 1995a, 2007; Andersen, 2008a).

In the Kilen record, maximum DI-salinity for the Ertebølle period occurs between ~5,100-4,800 BC, which likely reflects an opening or widening of the western entrance (e.g. North Sea drift diatoms such as *C. belgica* appear in the record). Whilst salinity in the preceding (i.e. before ~5,500 BC) period remains uncertain, the Kilen record suggests that DI-salinity was generally lower throughout the Ertebølle period than in the Neolithic (see Chapter 3, Figure 6.2). Reduced marine connection (and lower salinity), however, appears to have driven higher marine primary productivity (Figure 3.16) in the Limfjord throughout the Ertebølle and early Neolithic period. This probably provided sufficient food for higher organisms (e.g. molluscs and fish), enabling them to reach high abundances, subsequently providing an environment rich in marine resources for cultures and societies inhabiting the Limfjord catchment (Andersen and Johansen, 1986; Andersen and Rasmussen, 1991; Enghoff, 1991; Andersen, 1995a; Andersen, 2005).

6.3.2.1 The oyster decline and the Mesolithic-Neolithic transition

The large scale shift from oyster dominated middens in the Ertebølle period to cockle dominated middens in the Neolithic (Funnel Beaker Culture) has generated substantial interest regarding the role of environmental change, particularly salinity change (e.g. Rowley-Conwy, 1984), in potentially forcing the introduction of agriculture into Denmark (see section 1.6.2). Contrary to the “salinity hypothesis”, there appears to be no synchronous, regional decrease in salinity, at or near the Mesolithic-Neolithic transition, or any convincing evidence to suggest that salinity was markedly lower in the early Neolithic period. In fact, salinity actually increases in the Neolithic period at Kilen (see section 3.11) whilst at

Norsminde Fjord, it remains relatively high until the fjord reaches its sedimentation limit (at ~1,700 BC). At Korup Sø, a shift towards lower DI-salinities occurs later in the record (after ~3,300 BC), when the fjord basin begins to shallow. There is, however, some disagreement between the Korup Sø DI- and MI-inferences (Figure 6.2), with the latter supporting a decrease in salinity at ~3,900 BC and subsequent lower inferred salinities in the Neolithic period. This transition broadly coincides with a large increase in the accumulation rate (Figure 6.1), which might have driven the changes in mollusc record, rather than salinity (Petersen, 1993; see section 5.7 for more detailed discussion). Elsewhere, preliminary diatom analyses from two other Danish coastal sites (Tempelkrog in Isefjord and Horsens Fjord; Figure 2.1) also fail to document any marked change in salinity over the Mesolithic-Neolithic transitions (A. Clarke, unpublished data, K. Jensen unpublished data). Furthermore, in the NW Baltic Sea salinity continued to increase up until~3,000 BC (Emeis *et al.*, 2003), despite the drop in sea levels (Berglund *et al.*, 2005). Bailey and Milner, (unpub.) also dismiss salinity as causal factor for the quantitiative compositional changes in the Norsminde shell midden due to the continued presence of both *Ostrea edulis* and *Tape decussatus* (high-salinity demanding species; Jensen and Spärck, 1934; Yonge, 1960) throughout the Neolithic shell layers.

The lack of any long-term salinity decline does not necessarily mean that salinity change had no impact whatsoever upon mollusc abundances and/or distributions. Despite some uncertainties with the DI-salinity signal (see above) it remains highly likely that salinity was subject to fluctuations (magnitude 0-5 ‰ in the three DI-salinity reconstructions; Figure 6.2) at, or around the Mesolithic-Neolithic transition. It is, however, extremely difficult to assess the impact that short term salinity ‘events’ might have on mollusc abundances due to the low numbers of molluscs generally present in sedimentary archives and the coarse resolution of archaeological shell midden deposits. Whilst it is difficult to establish any firm linkages between short term salinity events and molluscan change, it certainly remains plausible that non-breeding populations (e.g. when salinity drops below 23-25 ‰ for *O. edulis*; Jensen and Spärck, 1934; Yonge, 1960) might become stressed in relatively short time periods (e.g. 1-20 years), particularly if regular and/or intensive human exploitation continues. Following deterioration, molluscan populations can also exhibit slow re-generation times (e.g. after the Limfjord re-opened to the North Sea in AD 1825, oysters did not appear until 24 years later (AD 1849; Bailey and Milner, submitted), which might further affect human resource scheduling and encourage the introduction of alternative resources (e.g. Rowley-Conwy, 1984). The most striking DI-salinity decrease occurs at Kilen between ~4,100-4,000 BC, with average salinity appearing to remain low up until ~3,900 BC (Figure 6.2). This might be related to a narrowing of the northern entrance of the Limfjord (see section 3.11) and it is likely that lower salinity exerted at least some pressure upon oyster abundances in the nearshore areas of the Limfjord in the late Mesolithic (e.g. the documented oyster decline in the shell midden at Bjørnsholm Bay; Andersen and Rasmussen, 1991).

Based on the data presented in this study, it is unlikely that any single environmental factor alone, can explain the widespread oyster decline apparent in various Danish shell middens (e.g. Andersen, 2007). There is, however, plenty of evidence suggesting that both the marine and terrestrial environment changed substantially over the late-Mesolithic/early Neolithic period (e.g. Mikkelsen, 1949; Petersen, 1993; this study). This period is also marked by deteriorating climate (e.g. decreasing temperatures; Antonsson and Seppä, 2007) associated with the end of the Holocene climatic optimum (Seppä and Birks, 2001; Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Brown *et al.*, submitted). It is argued here, that environmental change did play an important role in the oyster decline and the changing prosperity of the marine environment over the Mesolithic-Neolithic transition, with both sea level and climate change (particularly temperature) being the predominant drivers. The precise changes (i.e. response to sea level and/or climate change) often differ between sites and therefore the manner in which changes manifest within individual fjords and estuarine systems appear to be particularly important. For example, sea level is an important driver of changes in salinity (see section 1.4 and above), sedimentary regime (including substrate and accumulation rate), water depth, productivity and/or nutrient status, all of which are likely to have implications for community dynamics and societies exploiting marine resources. Climate and temperature change can influence catchment processes (e.g. precipitation, run off, winter ice cover and subsequent thawing events), vegetation change, coastal sea ice extent, marine productivity (e.g. Andrén *et al.*, 2000; Emeis *et al.*, 2003) and sea level (Berglund *et al.*, 2005), as well as directly affecting the reproduction and metabolism of marine organisms (e.g. oysters; Jensen and Spärck, 1934; Yonge, 1960; see below).

The sedimentary changes at both Norsminde and Korup Sø, suggest that the substrate and sedimentary regime might be more important than previously thought for explaining/forcing the shell midden oyster decline (e.g. Andersen, 1995a, 2007; Andersen, 2008a; Bailey and Milner, 2008). The period of lowest accumulation at Norsminde and to a lesser extent Korup

Sø broadly coincides with the maximum sea levels of the late Atlantic period (~5,600-4,200 BC), as evident in Baltic Sea and Kattegat sea level curves (Mörner, 1976; Christensen, 1995; Berglund *et al.*, 2005) and with maximum oyster abundances in Danish shell middens (Andersen, 1989; Andersen, 1995a, 2007; Andersen, 2008a). At Korup Sø, this relationship with sea level is less obvious as the accumulation rate is very low (0.03-0.09 cm yr^{-1}) throughout the entire Mesolithic section of the profile (Figure 6.1). Whilst absolute lowest rates do coincide with the maximum sea levels of the late Atlantic, the accumulation rate does not really increase until after ~3,960 BC, thereby post-dating the late Atlantic regression (~4,200 BC as dated by Berglund *et al.*, 2005). Whilst sea level fall might be a contributory factor to the rise in accumulation rate after ~3,960 BC (by reducing the strength of currents, thereby lowering the energy of the system and encouraging deposition of finer sediments within the fjord), catchment changes (e.g. forestry restructuring after the *Ulmus* decline) and increased marine productivity are also likely to have been important factors (see section and 5.7). Nevertheless, the site clearly demonstrates a large shift in the sedimentary regime, near the time of the oyster decline (Andersen, 2007; Bailey and Milner, 2008).

Ostrea edulis struggles with higher volumes of fine sediment, being less efficient at ejecting continuously accumulating sediment from its mantle cavity than species such as *Cerastoderma edule*, which are better adapted to living in finer sediments (Yonge, 1960; Bailey and Milner, 2008; Bailey and Milner, submitted). The gradual increase in accumulation rate at Norsminde (after ~4,700 BC) might have eventually been sufficient to stress *O. edulis* populations, particularly within the inner more human accessible parts of the fjord, where often less turbulent (or even still) waters might have led to greater deposition of fine material. Lower salinity might be also expected in conjunction with lower sea level, due to a likely reduction in the connection with the Kattegat (and Skagerrak/North Sea). The lack of any systematic salinity decline (see above) might be explained by other factors, such as increased terrestrial input from the land being more important for the change in the sedimentary regime (e.g. particularly for Korup Sø; see below), or that the sedimentary regime is more sensitive to changes in fjord (e.g. sea level) configuration than salinity. Alternatively, records from the Skagen and Skagerrak region document an increase in the strength of the Jutland current and a reduction in the outflow of Baltic water at the surface, and subsequent decrease in the strength of the Baltic current beginning around 4,300 BC (~6,300 BP in Conradsen and Heier-Nielsen, 1995; Jiang *et al.*, 1997; Gyllencreutz *et al.*, 2006). These current systems might have helped maintain the salinity within the Kattegat and inner Danish coastal waters, despite the drop in sea level and subsequent reduction in the size of the fjord entrances (plus likely siltation).

Unfortunately, there is a lack of suitable records available for the inner Danish waters or Kattegat sites with which to compare changes in the sedimentary regime over the early-mid Holocene. This is partly due to poor dating, as a robust age model is essential for determining changes in the accumulation rate at a site, but only two other sites from eastern Denmark (Tempelkrog and Horsens Fjord; P. Rasmussen and D.B. Ryves unpub., Figure 2.1) have long-term records with chronologies based on AMS ^{14}C dates on terrestrial material. At Tempelkrog, a sharp peak in accumulation rate occurs somewhere between ~4,400-3,900 BC, which broadly coincides with the late Mesolithic regression in the Blekinge sea level stratigraphy (~4,200-3,800 BC; Berglund *et al.*, 2005). This peak also succeeds a relatively low accumulation phase throughout the majority of the late Mesolithic period at Tempelkrog. The Horsens Fjord record, however, does not exhibit any increase in the accumulation rate in the late Mesolithic or early Neolithic period. This might be due to the sheer size of Horsens Fjord, being much larger than Tempelkrog or any of the other sites analysed in this study. It is likely to have maintained a large, open connection with the Kattegat, even during the late Mesolithic sea level regression (Berglund *et al.*, 2005), and therefore relatively minor changes in sea level might have had little impact upon the sedimentary regime.

For the other systems, however, a more open connection with the Kattegat and intensified marine conditions under high sea level conditions (between ~5,600-4,200 BC) is likely to result in increased energy within the system (stronger currents), greater transport of material out of the fjords and possibly greater breakdown and decomposition of material. It is also important to note here, that whilst sea level appears to influence the sedimentary regimes at Norsminde, Korup Sø and Tempelkrog, particularly during the late Mesolithic period, it is not a straightforward relationship, as all three sites exhibit quite different accumulation patterns. This might be related to physical differences between sites (e.g. site configuration, exposure to currents, depth, topography) and/or additional factors (e.g. productivity, catchment vegetation, human impact), potentially influencing the independent sedimentary regimes.

The idea that substrate was perhaps more important than salinity in driving the distribution of oysters in the Mesolithic has previously been suggested by Rasmussen, (1958). In the Mesolithic *Tapes pullastra* and *O. edulis* exhibit practically identical distributions

(Rasmussen, 1958), but their salinity tolerances today appear to be quite different. T. *pullastra* occurs at 18 ‰ or maybe even at lower salinities (Rasmussen, 1958), whereas salinities of at least 23-25 ‰ are needed to support breeding populations of *O. edulis* (Jensen and Spärck, 1934; Yonge, 1960). Rasmussen, (1958) suggests that their affinity to a similar substrate better explains their identical distributions, with both species preferring hard bottom substrates (i.e. sand and coarse sediments) with sparse vegetation cover. Such conditions were more likely in the late Mesolithic when the sea was more open (e.g. Berglund *et al.*, 2005) and deeper water limited vegetation cover (e.g. eel-grass beds).

In addition to sea level, a number of other factors might also influence the sedimentary regime of a particular site. For example, increased productivity at Norsminde Fjord between 4,550-3,510 BC is likely to have contributed to the higher accumulation rates in the late Mesolithic. Higher productivity is indicated by marked increases in both the diatom concentration (particularly planktonic diatoms from the genus *Chaetoceros*) and organic matter flux (Figure 6.3). A period of increased marine productivity is also evident at Kilen between ~4,400-3,100 BC (see Chapter 3). In contrast, the short term drop in *Chaetoceros* cysts at Korup Sø between ~3,900-3,770 BC (Figure 6.3) might represent a decrease in productivity. However, the diatom concentration does not decrease, due to higher abundances of benthic taxa, probably as a response to lower sea levels and subsequently shallower water between ~4,200-3,800 BC (Berglund *et al.*, 2005; Figure 6.2). It is therefore possible that reduced habitat availability and increased competition for resources is driving the *Chaetoceros* decline rather than marine productivity. Alternatively the *Chaetoceros* decline might be a data artefact, with the *Chaetoceros* abundances actually remaining the same, but being diluted by an increase in benthic diatoms due to shallower water.

The large increase in accumulation rate after ~3,960 BC at Korup Sø appears to be unexplainable by sea level alone, occurring towards the end of the late Mesolithic/early Neolithic regression phase. Whilst sea level is probably important (e.g. rise in accumulation rate occurs during likely shallower water associated with sea level regression phase; ~4,200-3,800 BC; Berglund *et al.*, 2005), the timing of this sedimentary change (i.e. rise) is very close to the sharp *Ulmus* decline at Korup Sø (Figure 5.7), possibly suggesting a role for increased input of terrestrial material and nutrients following vegetation change. The pronounced increase in *Quercus*, *Betula* and *Corylus* pollen immediately after the *Ulmus* decline suggest large scale restructuring of the catchment woodlands (see section 5.7) which certainly might have influenced the volume and rate of sediment input. High accumulation rates were likely maintained by human impact upon the landscape (particularly forest clearances for arable and pastoral farming) with the first signs of agriculture appearing relatively early in the Korup Sø pollen record (e.g. first cereal pollen present at ~3,850 BC). A possible increase in marine productivity occurs shortly after ~3,700 BC (e.g. increase in *Chaetoceros* cysts and *C. choctawhatcheeana*), probably in response to both rising sea level (i.e. L4 transgression, Berglund *et al.*, 2005; Figure 6.2) and increased nutrient input from the catchment associated with forest restructuring and land use change.

The very different sedimentary regime at Kilen suggests that this site is either heavily driven by more localised changes, or that mechanisms important in governing accumulation and fluxes in the Limfjord and/or northwestern Jutland region differ from those operating along the eastern coast of Jutland (see section 3.11). The accumulation rate at Kilen is very high (0.33-0.49 cm yr^{-1}) between ~5,500-3,150 BC, far exceeding the rates at both Korup Sø (before ~3,900 BC; 0.03-0.09 cm yr^{-1}) and Norsminde Fjord (0.09-0.17 cm yr^{-1}). Such high accumulation rates might explain why shell midden sites are absent, to date, from the western reaches of the Limfjord (e.g. Figure 1.3 - Location map for Danish archaeological sites referenced in the text.). Fast accumulation of fine sediments in the far southern and western reaches of the Limfjord might have limited the oyster’s ability to survive there (see above) there, particularly in the inner more accessible areas. The accumulation rate at Kilen drops after ~3,200 BC, which coincides with increasing salinity, likely coarsening of sediments due to stronger currents meaning less settling out of finer material (both suggesting intensified marine conditions) and a decrease in marine productivity (Figure 6.1). This might suggest that conditions actually became more suitable for oysters in the western Limfjord after ~3,200 BC as a result of higher salinity and reduced accumulation of finer substrates. This might also help explain the high numbers of oyster present in some of the Neolithic layers of the Krabbesholm shell midden (Andersen, 2005; Nielsen, 2008), which contrasts with many of the other shell middens because it does not exhibit a ‘conventional’ oyster decline (e.g. Andersen, 1989; Andersen and Rasmussen, 1991).

As suggested by Petersen, (1993), the case for sea level as a predominant driver of changes in the accumulation rate and substrate type might be further strengthened by the lack of change in the sedimentary regime in other archives such as lakes (e.g. Odgaard, 1989; Odgaard, 1992; Rasmussen and Bradshaw, 2005). For example, if large scale catchment changes (and associated processes) were responsible, then increased accumulation rates might be expected in these other archives. Additionally, the reduction in the strength of the Baltic current after ~4,300 BC (Conradsen and Heier-Nielsen, 1995; Gyllencreutz *et al.*, 2006; see above) might also have further reduced energy within the Kattegat system and connected coastal waters, and subsequently led to greater deposition of fine sediments.

It is also likely that temperature exerted an additional stress on oyster populations in the late Mesolithic. Spawning of *O. edulis* occurs in the summer months when temperature reaches above ~15°C (Yonge, 1960). Scandinavian temperature records suggests that temperature was declining following the Holocene climatic optimum (~5,500-3,500 BC) and that by the Mesolithic-Neolithic transition, average annual temperatures had fallen by almost ~1°C (Seppä and Birks, 2001; Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Seppä *et al.*, 2009). There is evidence suggesting that temperature might have decreased more rapidly during the last 500 years of this thermal optimum (~4,000-3,500 BC; Snowball *et al.*, 2004), possibly coinciding with one of the North Atlantic ice rafting events (Bond *et al.*, 1997). At this timer there was increased activity in Swedish and Norwegian glaciers (Karlén, 1988; Karlén and Matthews, 1992; Matthews and Karlén, 1992; Dahl and Nesje, 1994; Dahl and Nesje, 1996), a shift to colder and drier winters in western Norway (Nesje *et al.*, 2001) and a distinct fall in mean July temperatures in Finland (Korhola *et al.*, 2002). More recently, quantitative temperature inferences have been extended into Denmark (Brown *et al.*, submitted), exhibiting similar patterns to other Fennoscandian records (e.g. Seppä and Birks, 2001; Antonsson and Seppä, 2007). A reduction in temperature has been shown to have affected the distribution of other organisms such as *Emys orbicularis* (European pond turtle), which became extinct from Denmark and southern Sweden in the early Subboreal period (3,750-1,750 BC; Sommer *et al.*, 2007). There is also an southward expansion of *Picea abies* in southern Sweden (Giesecke and Bennett, 2004) and a general decline in the abundance of thermophilous vegetation (e.g. *Hedera helix*, *Viscum album*) as the climate cooled (Antonsson and Seppä, 2007; Brown *et al.*, submitted).

Oysters are likely to have been more sensitive to short term decreases in temperature (and/or salinity) than cockles in this period, being close to both their temperature and salinity tolerance limits (Jensen and Spärck, 1934; Yonge, 1960). The shift to cooler summers probably suppressed the length of the oyster spawning season, whilst colder drier winters and likely increased irregularity and frequency of extreme cold events might have led to elevated oyster mortality. Temperature is actually considered to be more important than salinity for oyster reproduction (Korrigna, 1952; Yonge, 1960) and in some areas of Norway and Sweden, colder temperatures prevent oyster from spawning at all in some years (Johannesson *et al.*, 1989). Carbotte *et al.*, 2004 have also previously linked the demise of the eastern oyster (*Crassostrea virginica*) from the Hudson River estuary to cooler climates and increased sediment accumulation rates. Prolonged cold weather (temperatures around 0°C) throughout the winter months of 1939-1940 is also believed to have destroyed more than half the native population and almost the entire imported population of *O. edulis* in the Thames estuary (Yonge, 1960). Rapid thawing events in Denmark might also have temporarily lowered salinity within the coastal waters, which might have further increased mortality. As with the salinity events (see above), it is extremely difficult to pick up short-term temperature changes in sedimentary records and directly assess their impact upon mollusc populations. Rapid thawing events might also explain some of the fluctuations in DI-salinity throughout this period (Figure 6.2).

Temperature decrease certainly offers a plausible explanation for the oyster decline yet continued presence and exploitation of oysters into the Neolithic, as evident in the shell middens (Andersen, 2007). *Ostrea edulis* can survive in substantially lower temperatures and salinities than it requires for successful spawning (Yonge, 1960). Oysters can also live up to 20 years in oyster beds and therefore, even during prolonged periods of colder weather (and/or lower salinity), oysters probably survived in the inner waters. Despite this, shorter spawning periods and possibly years without any reproduction would have inevitably impacted oyster populations, leading to greater stress, particularly within the nearshore accessible populations, if exploitation continued at the same rate in the early Neolithic as in the Mesolithic.

The suggestion that a decline in oysters alone forced the neolithisation of Denmark is certainly oversimplifying a complex transition. However, the decline in oysters appears to be a symptom of general change in the marine environment and it is increasingly likely that a change played at least some role in the transition from a Mesolithic to a Neolithic economy. Carbon (δ^{13}C) and nitrogen (δ^{15}N) isotopic evidence from human bones and teeth suggests greater consumption of marine resources in the Mesolithic with a switch to a more terrestrial diet in the Neolithic (Tauber, 1981; Milner *et al.*, 2004; Andersen, 2007; Andersen, 2008a), though this evidence remains equivocal to date (e.g. Milner *et al.*, 2004, 2006; Richards and

Schulting, 2006a). The isotopic shift, however, fits well with a change in the exploitation of marine resources (e.g. reduced oyster collection and change in seasonality of shell gathering; Bailey and Milner, 2008; Bailey and Milner, submitted) and the appearance of agricultural indicators in the Danish pollen records (Iversen, 1941; Behre, 1981; Odgaard, 1994; Andersen, 1995b; Rasmussen, 2005). Whereas the changes in both the isotope ($\delta^{13}C$ and $\delta^{15}N$) and the pollen records, could potentially be explained by cultural factors alone (e.g. environmentally-unforced human resource shift), substantial evidence exists to suggest that the oyster decline and the shift in marine resources is synchronous (within the constraints of dating shell midden deposits) across Denmark and strongly related to environmental change (see above).

In this study it is suggested that changes in the substrate and sedimentary regime (Figure 6.1) probably made it more difficult for oysters to survive, particularly in the innermost, more accessible areas of the fjord where shallower, still water might have increased fine mud deposition. In addition, short term stresses associated with salinity, productivity and temperature events, probably driven by sea level, vegetation and climatic changes (see above) are likely to have prevented spawning in some years and even caused widespread mortality, suppressing oyster populations. More variable abundance, combined with continued human exploitation would have placed greater pressure on oyster beds, probably making them less of a dependable resource, causing their decline in the middens as humans subsequently shifted towards a more terrestrial-based diet (e.g. Tauber, 1981; Andersen, 2007; Andersen, 2008a).

There is still substantial work to be done before a clear and certain link can be made between changes in the marine environment forcing changes in the shell middens. Thorough investigation of the accumulation rates and changes in the substrate from more localities across the inner coastal waters of Denmark and even eastern Sweden is essential to understand fully changing sedimentary conditions, particularly in response to sea level changes over the late Mesolithic and early Neolithic period. Such inferences can be strengthened by use of proper sedimentological techniques (e.g. grain size analyses, magnetic susceptibility), in conjunction with the more standard sedimentary parameters and descriptive lithological techniques employed in this study.

More long term records of salinity change are also required, particularly from the Limfjord region, where there is evidence to suggest lower salinities might have existed in the late

Mesolithic period. Future diatom counts should involve larger census counts (i.e. 300-500 valves excluding *Fragilaria* s.l, *Chaetoceros* and *Rhizosolenia* species), so that salinity models can be tested excluding euryhaline indicators, or taxa that cannot be differentiated below the genus level. This is particularly important for east Jutland and Kattegat coastal sites which contain particularly high abundances of *Fragilaria* s.l., likely due to regular disturbances, instability and variable salinity, nutrient concentrations and productivity. Exclusion of *Fragilaria* s.l. is likely to overestimate salinity inferences, but comparison of full and reduced models might help to establish periods where transfer functions are performing well from sections where other factors might be driving changes in the diatom distributions (i.e. disagreement in patterns of salinity change between models). Future diatom inferences would be strengthened by high resolution foraminifera, molluscan (where possible) and ostracod records for comparison of salinity inferences and to begin to understand better under what circumstances these independent proxy records agree or disagree with one another. Other proxies not considered here, can also be used, for example, dinoflagellates (abundant at Korup Sø; Figure 5.8). Cyst morphology (particularly process length of dinoflagellates) has been shown to be closely associated with salinity change, making it another potential palaeosalinity indicator (Ellegaard *et al.*, 1993; Ellegaard, 2000; Ellegaard *et al.*, 2002; Mertens *et al.*, 2009).

High resolution records of water temperature (e.g. alkenone records or Mg-Ca records) from fjord sites might help to improve our understanding of the role of temperature in driving the changes in oyster (and cockle) abundances. Such records would also provide a useful comparison with the quantitative temperature records (particularly regional pollen- or chrionomid based studies; Seppä and Birks, 2001; Antonsson and Seppä, 2007; Seppä *et al.*, 2009), which might be subjected to temperature lags due to the slower response of vegetation to climatic changes (e.g. Pennington, 1986; Williams *et al.*, 2002). Pollen-based temperature records are also less likely to be able to pick-up any short-term extreme temperature events and might also be affected by human impact too (particularly after the introduction of agriculture in Denmark; e.g. Hannon *et al.*, 2000; Odgaard and Rasmussen, 2000; Brown *et al.*, submitted).

Furthermore, as suggested by Bailey and Milner, (2008), it still remains likely that a mixture of environmental and cultural factors explain the changes in the shell middens (and the introduction of agriculture). In addition to changes in the substrate and temperature, they speculate that population dynamics (i.e. a larger Neolithic population), changes in economic scheduling and/or a reduction in conservation measures (i.e. more careful measures taken to protect marine food source in the Mesolithic, when it was more important to their subsistence) might have placed more pressure on marine resources (such as cockles and oysters) in the Neolithic. In terms of the introduction of agriculture, the evidence presented in this study supports the idea that environmental change was important in forcing the oyster decline and subsequent shift in shell midden composition, but more work is needed before the agricultural question can be confidently addressed and non-environmental hypotheses can be dismissed (e.g. Rowley-Conwy, 1985; Klassen, 1996-97; Petersson, 1996-97; Johansen, 2006). For example, more archaeological work is also needed from the middens to understand better the rapidity of midden changes (both ecological and cultural) and regional differences in speed and timing of these changes. Whilst timing and rapidity is beset with difficulties, particularly when using chronologies based on ^{14}C dating, further complications might be commonly encountered in shells middens associated with discontinuous accumulation and sedimentary gaps. A better understanding of the physical environment that the shell midden formed in is also required and greater focus placed on understanding exactly why Mesolithic and Neolithic artefacts are sometimes found in mixed layers (e.g. reworking or cultural dualism; Petersson, 1996-97). Isotopic, age and seasonality analyses (e.g. Milner, 2001a, 2001b; Bailey and Milner, 2008) of the shells present from the actual shell middens and parallel palaeoenvironmental investigations in the immediate vicinity of other major shell middens might help answer some of these questions.

Oyster abundance generally remains low throughout the entire TBK section of Danish shell middens (i.e. up until ~2,800 BC), as marine resources appear to become less important in the TBK diet (Tauber, 1981; Andersen, 2007; Andersen, 2008a; see section 1.6.2). A period of lower and more variable abundance of oysters (and other marine resources) probably ensued, as sea level and (probably) salinity (see above) continued to fluctuate and air temperature further decreased (Snowball *et al.*, 2004; Antonsson and Seppä, 2007; Seppä *et al.*, 2009; Brown *et al.*, submitted). Large scale human impacts upon the catchment also become more visible in the sedimentary records in the TBK period, as seen for example at Norsminde and Korup Sø (see section 4.11 and 5.7). Episodes of forest clearances and land-use change are probably the main causes of the fluctuations in the accumulation rate at these sites, though sea level, climate and natural vegetation change probably all contributed. The impact of humans upon the Kilen catchment area, and the level of interaction between the terrestrial and marine environment, is more ambiguous, due to lack of information regarding the local terrestrial environment (i.e. no real pollen record or poor plant macrofossil record; Figure 3.3 and Figure 3.13). There is a minor increase in the accumulation rate between 3,700-3,400 BC, which broadly coincides with an appearance by *Plantago lanceolata* and increasing Poaceae pollen. This is possibly related to increased input of nutrients and sediment from the catchment, causing the organic and minerogenic matter to increase and helping to maintain high pigment concentrations/accumulation at this time. It is difficult, however, to be certain that terrestrial input is the source with such little information available regarding changes in the surrounding catchment. After ~3,200 BC, accumulation rate drops substantially, and changes in the marine environment (i.e. greater exposure to the North Sea, higher energy and higher salinity) appear to control the sedimentary regime at this site.

6.3.3 The Pitted Ware and Single Grave cultural period (~2,800-2,400 BC)

The increase in the number of shell middens and shell banks in the Pitted Ware/Single Grave cultural period (hereafter PW-SG), and the switch back to oyster-dominated middens, might suggest either a return to higher marine productivity, intensified marine conditions (e.g. increased salinity, coarser sediments) or a change in cultural resource exploitation. Temperature reconstructions (e.g. Antonsson and Seppä, 2007) suggest a minimal change (gradually decreasing) throughout the PW-SG cultural period and is therefore unlikely to be able to explain shell midden patterns, although occasional warm temperature events might have been important (see above). In the Limfjord region, salinity increases over the PW/SG periods, almost reaching its present day value for the Struer region (24-27 ‰ according to the Kilen DI-reconstruction). This is probably due to opening and/or widening of the western entrance (see section 3.11) in response to rising sea levels (Behre, 2007). Higher salinities and a change in the sedimentary regime (i.e. drop in the accumulation rate at Kilen and likely coarsening of sediments) after ~3,100 BC might have resulted in a more suitable environment for oysters and enabled them to penetrate into the inner reaches of the fjords/estuaries. It is important to note, that despite decreasing, the accumulation rate at Kilen remained relatively high (0.15-0.18 cm yr^{-1}) and therefore oysters probably might still have been confined to the more exposed, outer reaches of high-accumulating fjords such as Kilen.

There is also some evidence suggesting that higher salinities existed at Norsminde Fjord and Korup Sø, either prior to, or during the PW-SG cultural period. At Norsminde, the DI-salinity reconstruction suggests that higher salinity might have existed between ~3,070-2,900 BC and ~2,700-2,570 BC. Under the full model, average DI-salinities reach 23-25 ‰ in these two phases, suggesting that in some areas of the fjord spawning populations of oysters might have existed (i.e. requiring salinities above 23-25 ‰). Again, oysters were probably still confined to the more exposed parts of Norsminde Fjord due the persistence of relatively high accumulation rates throughout the TBK period and the large increase in accumulation rate at ~2,800 BC probably driven by local catchment changes (Figure 6.1 and see section 4.11). For example, the Kalvø site (a shell midden dated to the Single Grave culture, where oysters make up 68 % of shellfish remains; Andersen, 1983) is located along the north eastern coast of the former island within Norsminde Fjord (Figure 2.4), placing it relatively close to the Kattegat opening and potentially subject to greater currents and coarser sediments than in the inner reaches of the fjord. Alternatively, sediment focussing might have resulted in higher accumulation rates at the coring location, but perhaps not in the shallower reaches of the fjord, oysters might have been able to survive in these locations. Their confinement to the Kalvø site (at Norsminde Fjord) might therefore be related to salinity (i.e. greater exposure to higher salinity Kattegat) rather than sediment accumulation.

At Korup Sø, the main indicators of higher salinity over the PW-SG period are increased abundances of *Hinia reticulata* and *Onoba vitrea* between ~3,000-2400 BC, which causes the MI-salinity to peak (~14-18 ‰) between ~2,800-2,660 BC. There is little change in the DI-salinity reconstruction over the PWC/SGC cultural period with the exception of a possible brief rise at ~2,840 BC. The accumulation rates for both Korup Sø core profiles (i.e. Well no. 1 and Well no.2) remains relatively high, despite a drop in Well no. 1 shortly after ~2,800 BC, which together with increasing *Ruppia* beds and salinities below the required 23-25 ‰ for oysters to spawn (Jensen and Spärck, 1934; Yonge, 1960) might have prevented oysters from living in the immediate vicinity of Korup Sø. Oysters are, however present in the Djursland area at this time, being found at the Pitted Ware culture sites Kirial Bro and Kainsbakke (Rasmussen and Richter, 1991; Rasmussen, 1993a). Both these sites are located north east of Korup Sø (Figure 2.5) and closer to the Kattegat entrance, where higher salinity and lower sediment accumulation rates probably existed.

Higher sea level also might be able to explain some of the changes evident in the DI- and MI-salinity records at Norsminde Fjord and Korup Sø (Figure 6.2), but as discussed above, there is no simple relationship between rising sea level and salinity in this period. Furthermore, the sea level history is generally poorly understood throughout the Funnel Beaker and PWC/SGC periods, but it remains likely that minor fluctuations in sea level continued throughout this period (e.g. Mörner, 1976; Miller, 1982; Berglund *et al.*, 2005; Behre, 2007). To date, no sea level transgression has been specifically linked to the PW-SG period but it has been suggested that sea level might have been at a similar level to the Ertebølle period (e.g. Becker, 1951).

This study demonstrates that environmental change might have provided better conditions for oysters during the PW-SG period in coastal Denmark, but more work needs to be done before strong links between environmental change, oyster abundance and human exploitation can be made. Coastal sites from the PWC/SGC are still quite rare compared to the Mesolithic and Iron Age (e.g. Andersen, 2007) and despite demonstrating signs that marine resources were important to these two cultures (Becker, 1950; Andersen, 1983; Rasmussen and Richter, 1991; Rasmussen, 1993a; Andersen, 2007), the full extent of their marine exploitation remains uncertain. Better dated sea level records and more information (i.e. more spatially distributed records and higher resolution analysis of several proxies), particularly concerning salinity change and the sedimentary regimes of coastal sites, are required to improve our understanding of the marine environment over this period of marine exploitation. There is also a clear need to undertake palaeoenvironmental analyses in the immediate vicinity of PWC/SGC settlements containing marine resource deposits (e.g. shell layers or banks). It is likely that abundant, breeding oyster populations were confined to smaller, specific areas during the PWC/SGC period, than in the late Atlantic (i.e. 5,600-3,900 BC; see above), only existing (and exploited) where very specific environmental conditions were met (e.g. accessible areas of high salinity, more exposed to the open sea and with lower accumulation rates and perhaps only in warmer years).

It is also again likely that no single factor will explain the shell midden patterns observed during the PWC/SGC, and that both environmental and cultural factors are important. For example, increased abundance of oysters in the Kalvø shell midden broadly coincides with a large increase in accumulation rate and charcoal concentrations in Norsminde Fjord (Figure 4.13). As this likely reflects land-use change and subsequently greater input of material from the catchment area, then it is possible that greater pressure was placed on both terrestrial and marine resources at this time, perhaps to support a population increase or a diminishment in terrestrial resource for some reason (e.g. poor growing seasons perhaps due to decreasing temperatures; Antonsson and Seppä, 2007, death of livestock due to disease). Increased exploitation of marine resources might therefore have been more essential to supplement shortfalls in the terrestrial diet. Marine resources might always have been present, but more heavily exploited when alternative resources were subject to greater pressure. Alternatively, marine resources might have been both more abundant and accessible during the PWC/SGC period due to more favourable environmental conditions.

6.4 Environmental change in the Limfjord region; ~1,700 BC – present

6.4.1 The Bronze Age (1,700-500 BC)

The DI-record from Kilen suggests that high salinity persisted throughout the Bronze Age and that an open connection existed between the Limfjord and the North Sea and Skagerrak throughout the period. Salinity generally increases up until the late Bronze Age (~1,000 BC), at which point there is a brief drop, probably driven by a sea level regression (e.g. regression 2 in southern North Sea beginning at ~1,500 BC and followed by persistently lower sea levels up until ~500 BC; Behre, 2007; see Figure 3.16). Additionally, the sedimentary pigment record suggests that marine primary productivity is relatively low throughout this period. Whilst the archaeological record demonstrates that coastal resources were still exploited in the Bronze Age (e.g. occasional shell layers and fishing tools and artefacts; see section 1.6.4), marine resources appear to be much less important to the subsistence of Bronze Age cultures and societies. It is possible that the higher salinity conditions, greater currents, coarser sediments and generally lower marine productivity confined natural shell beds to specific areas of the Limfjord. The absence of coastal shell middens indicate that shell fish were not collected in such large numbers, unless there was a change in the preparation method during the Bronze Age (e.g. transported elsewhere prior to processing) or greater erosion of coastal sites.

6.4.2 The Iron Age shell midden accumulation period (~500 BC-AD 1050)

The relationship between sea and society in the Limfjord region appears to be anomalous with much of Denmark over the Iron Age period. Numerous shell and kitchen middens are found along the coastlines of eastern Denmark and north eastern Germany (predominately dating to the Roman period and largely consisting of *Mytilus edulis* (blue mussel), *Cerastoderma edulis* (cockle) and *Littorina littorea* (periwinkle); Rasmussen *et al.*, 2007) but are absent from the shorelines of the Limfjord (between ~500 BC-AD 1050; see section 1.6.5). The results presented in Chapter 3 (and previous palaeoenvironmental analyses; Kristensen *et al.*, 1995; Christensen *et al.*, 2004), suggest that the Iron Age was a dynamic phase in Limfjord (and north-western Danish) history. Large scale shifts in salinity occur as a result of varying degrees of connection of the Limfjord to the North Sea, predominately driven by changes in sea level (Christiansen *et al.*, 1990; Tanner, 1993; Behre, 2007; Pedersen *et al.*, 2009) and sedimentary processes (i.e. sedimentary accretion and erosion) along the northern and western coasts of the Limfjord (see section 3.11).

6.4.2.1 Pre-Roman (500 BC-AD 0) and Roman period (AD 0-400)

The most important changes over the Iron Age section of the Kilen profile are presented in Figure 6.4, together with regional curves for sea level change (Behre, 2007), and temperature (Antonsson and Seppä, 2007). In summary (for detailed discussion, see section 3.11) Kilen undergoes a major shift from high salinity (30-35 ‰) in the pre-Roman period, to brackish salinities (14-20 ‰) in the Roman period (~AD 50-600). This salinity decline is sharp (occurring over ~50 years) and is accompanied by changes in both primary (e.g. increase in the accumulation rate of several sedimentary pigments) and secondary productivity (e.g. increase in mollusc and ostracod concentrations). The molluscan record also exhibits a shift in faunal composition, with low-medium salinity demanding species (*Cerastoderma* sp., *Mytilus edulis*, *Hydrobia ventrosa*, *Hydrobia ulvae*, *Corbula gibba*, *Littorina saxatilis* and *Parvicardium ovale*) becoming more abundant. In the plant macrofossil record, the brackish horned pondweed, *Zannichellia palustris*, becomes regularly present between ~AD 0-600. The sedimentary regime also changes in the late pre-Roman/early Roman period (after ~AD 40) driven largely by a decrease in the minerogenic flux (decreasing between ~AD 0-500), which causes the overall accumulation rate to fall (see section 3.11 for more detailed discussion).

In Figure 6.5, the results from this study are compared with three previous late-Holocene palaeosalinity studies from the Limfjord (Kristensen *et al.*, 1995; Ellegaard, 2000; Christensen *et al.*, 2004). There is generally good agreement between salinity estimates between these four studies (i.e. including Kilen), but dating problems (e.g. marine reservoir effect) inherent in all previous studies have prevented a precise dating of the various late

Holocene changes (~500 BC - present). For example, previous studies document a large decrease in salinity (i.e. a shift from marine to brackish conditions) towards the end of the pre-Roman period, but the date of this closure has previously only been broadly estimated to sometime between ~500-0 BC (Petersen, 1976, 1980; Christensen *et al.*, 2004). This study suggests that the Limfjord was closed off from the North Sea and Skagerrak at ~0 AD/BC, probably following the late pre-Roman drop in sea level (~150 BC-AD 50; Behre, 2007) and likely increased sedimentary accretion and dune activity under lower energy conditions (Christiansen *et al.*, 1990; Tanner, 1993). In the succeeding Roman period, the DI-salinities (~14-17 ‰; ~AD 50-600) at Kilen are in good agreement with the microfossil estimates (~15 ‰) of Kristensen *et al.*, 1995, and the upper macrofossil estimates of Christensen *et al.*, (2004).

Increased productivity of the marine environment (particularly of marine molluscs) during the Roman period occurs throughout other parts of Denmark, the Baltic Sea and Skagerrak (Andrén *et al.*, 2000; Hebbeln *et al.*, 2006; Rasmussen *et al.*, 2007; Zillén *et al.*, 2008; D.B. Ryves *et al.* unpublished data). The Limfjord, however, appears to differ from other localities, in the sense that lower salinity prevails throughout this period (Figure 6.4). For example, molluscan analyses from Horsens Fjord (east Jutland) and Tempelkrog (southern Isefjord, see Figure X; sites map) document up to two periods of increased molluscan concentration throughout the Roman period, during which higher-salinity demanding species such as *Ostrea edulis* and echinoid remains are present in the record (Rasmussen *et al.*, 2007). In fact, *O. edulis* is found in shell middens as far south as Flensburg Fjord and Eckernförder Bucht (in northern Germany) during the Iron Age (Anger, 1973). Quantitative salinity reconstructions (using diatoms) from two sedimentary cores from the Skagerrak also suggest higher surface salinity in the Roman period, particularly between ~AD 0-280 (Hebbeln *et al.*, 2006), and higher DI-salinity is apparent over this period at both Tempelkrog and Horsens Fjord (D.B. Ryves *et al*. unpublished data).

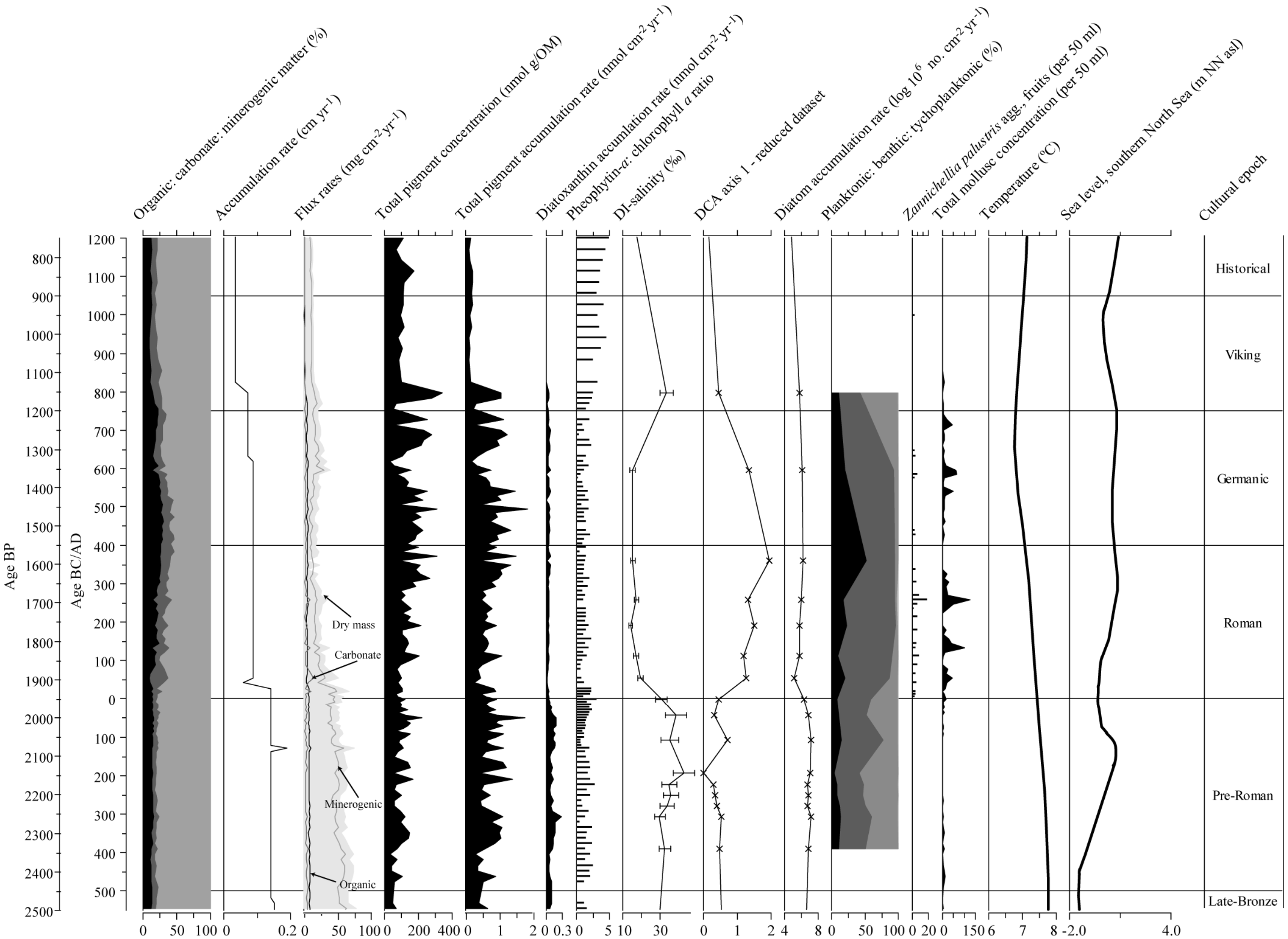
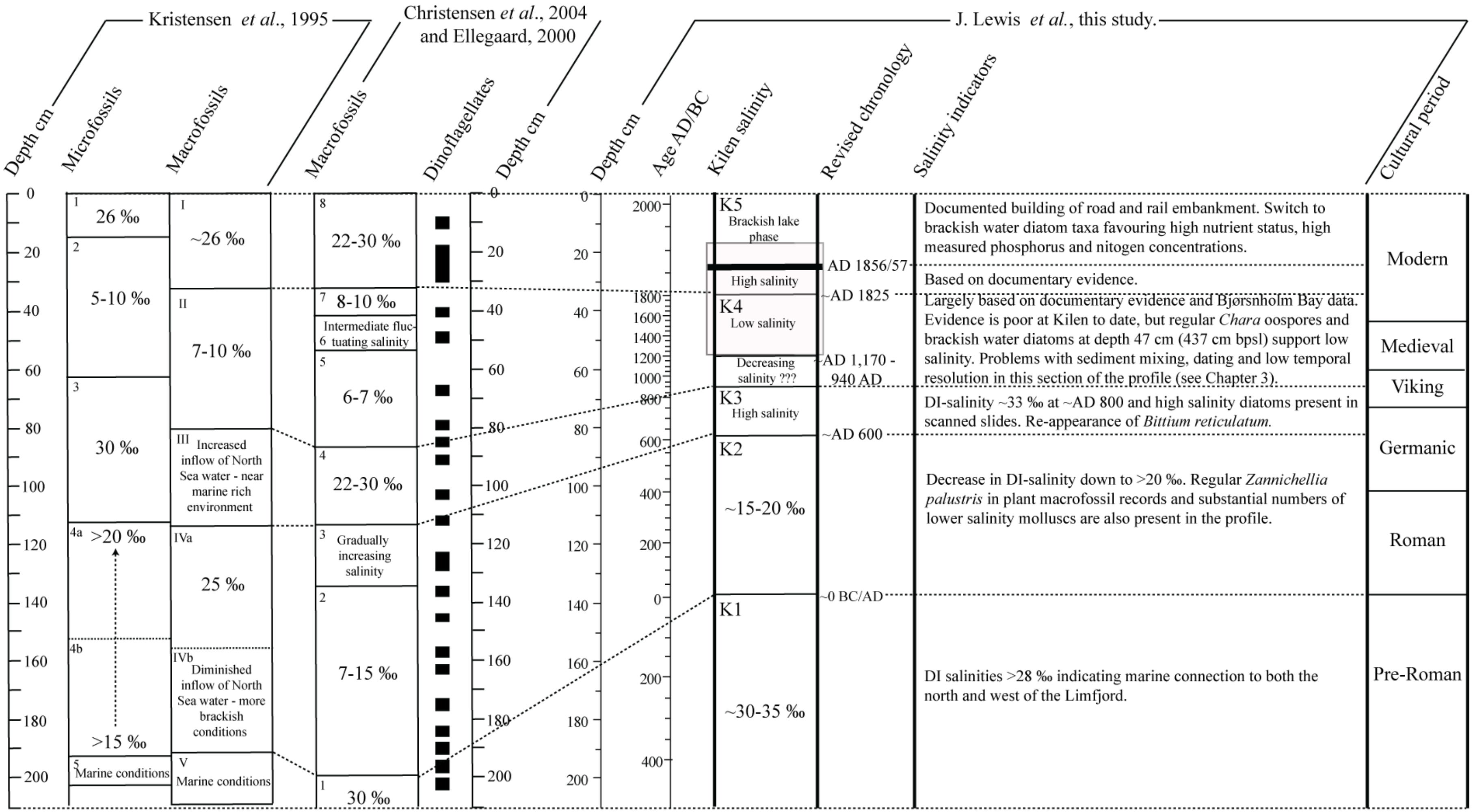


Figure 6.4 - Summary of key changes occurring over the Iron Age period (~500 BC-AD 1050) at the Limfjord site Kilen. Regional temperature (from Lake Trehörningen; Antonsson and Seppä, 2007) and sea level curves (southern North Sea; Behre, 2007) have also been included. DCA axis 1 reduced dataset excludes the uppermost 28 diatom samples (see section 3.10 for more details). *m NN asl = metres in relation to German ordnance datum above/below present sea level.



Documented building of road and rail embankment. Switch to brackish water diatom taxa favouring high nutrient status, high measured phosphorus and nitogen concentrations.

Based on documentary evidence.

Largely based on documentary evidence and Bjørsnholm Bay data. Evidence is poor at Kilen to date, but regular *Chara* oospores and brackish water diatoms at depth 47 cm (437 cm bpsl) support low salinity. Problems with sediment mixing, dating and low temporal resolution in this section of the profile (see Chapter 3).

DI-salinity ~33 ‰ at ~AD 800 and high salinity diatoms present in scanned slides. Re-appearance of *Bittium reticulatum*.

Decrease in DI-salinity down to >20 ‰. Regular *Zannichellia palustris* in plant macrofossil records and substantial numbers of lower salinity molluscs are also present in the profile.

DI salinities >28 ‰ indicating marine connection to both the north and west of the Limfjord.

Figure 6.5 – Comparison of Limfjord palaeosalinity records for the last ~2,500 years (i.e. Kristensen *et al.*, 1995; Ellegaard, 2000; Christensen *et al.*, 2004; this study). An updated chronology is proposed based on the AMS ^{14}C dates from terrestrial material in this study. The box in the 'Kilen salinity' column covers the period of dating uncertainty in the age-depth model for the Kilen profile (see Chapter 3).

In the central Baltic Sea, salinity estimates (based on $\delta^{13}C_{TOC}$ analyses) suggest a shift towards increasing salinity after ~1,100 BC, which culminates in a severe reduction in the number of freshwater diatom taxa (being replaced by brackish taxa) in the early Roman period (Emeis *et al.*, 2003). In this Gotland basin core, the shift towards increasing salinities is also accompanied by the formation of laminated sediments (beginning at ~2,600 BP; Emeis *et al.*, 2003). Periods of increased productivity, salinity, warmer temperatures (leading to thermal and haline stratification) and higher total organic carbon (Andrén *et al.*, 2000; Zillén *et al.*, 2008) appear to amplify hypoxia, and laminations form due to the absence of large benthic fauna, which cannot live without oxygen. These organisms are responsible for the bioturbation of sediments and in their absence, alternating layers of diatomaceous (formed mainly during the spring bloom) and clay-rich authigenic mud (Zillén *et al.*, 2008) or calcium rich rhodochrosite form (common in the Gotland sediments; Huckreide and Meischener, 1996; Sohlenius *et al.*, 1996). In the period between ~AD 0-1200 laminations appear to form across the Baltic basin (Andrén *et al.*, 2000; Emeis *et al.*, 2003; Zillén *et al.*, 2008), which coincides with periods of increased marine productivity (e.g. high TOC; Figure 1.2 Andrén *et al.*, 2000) and possibly warmer climate associated with the Roman warm period (~AD 100-400; Lamb, 1982; Seppä *et al.*, 2009) and, later, the Medieval warm period (Briffa *et al.*, 1990; Briffa *et al.*, 1992; Broecker, 2001).

The importance of salinity in driving changes in the Baltic Sea in the early Roman period, however, remains uncertain. Whilst it is likely that higher salinity existed in the Roman period (as suggested by Emeis *et al.*, 2003), due to greater exchange of saline water through the Danish straits, not all salinity studies document an increase at this time. For example, Gustafsson and Westman, (2002) employed an oceanographic model to investigate the exchange of water and salt through one of the Danish straits, and their results suggest that salinity briefly dropped ~2,000 years ago and has remained relatively constant since. It is difficult, however, to perceive how oysters were able to extend in the northern reaches of the Baltic Sea (e.g. Flensburg Fjord and Eckernförder Bucht, where surface salinity is today ~15 ‰; see above, Figure 2.1) without greater exchange of saline water through the Danish straits and Øresund, and hence higher salintiy. Furthermore, evidence from south-eastern Sweden suggests an early Roman sea level transgression (Miller, 1982; Miller and Hedin, 1988), increasing water exchange with the North Sea, and subsequently higher salinity within the Baltic Sea. Further work is clearly required to fully understand the nature of salinity change in the Baltic Sea over the late Holocene.

The spatial extent of evidence for high marine productivity across Denmark and the Baltic Sea clearly demonstrates that the Iron Age (and particularly the early Roman period) was a productive and diverse period. It appears, therefore, that the large number of shell middens dated to the Iron Age (but concentrated in the Roman period; Poulsen, 1978) is no coincidence and that humans responded to the richer marine biotope by heavily exploiting its resources. Based on the size and cubic content of some of the Iron Age middens located in Denmark and Germany, Andersen, (2007) has suggested marine resources (particularly mussels) might have been exploited almost industrially, perhaps for widespread distribution.

The absence of Iron Age shell middens right on the shoreline of the Limfjord is difficult to explain. Molluscan records from the Limfjord (e.g. Figure 3.12) all document that *Mytilus edulis* (generally the most common molluscan species present in Iron Age middens (Poulsen, 1978; Andersen, 2007) was abundant during the Roman period. The high abundance and widespread spatial distribution of mussels during the Iron Age, might have meant that site selection was more heavily determined by additional resources available to these cultures. If mussels could be farmed at any location then, primary importance might have been placed on alternative food sources such as oysters or particular fish species, potentially absent from the Limfjord due to lower salinity waters (and/or unsuitable substrate perhaps due to the regular presence of macrophytes such as *Zannichellia palustris*) in the Roman period. Alternatively mussels might have been collected more sporadically (perhaps as a dietary supplement), from different localities, preventing the temporal accumulation of shell middens. Existing data indicate that in the Limfjord area relatively small amounts of shells have been transported to more inland settlement sites prior to shelling (Poulsen, 1978; Ringtved, 1992).

6.4.2.2 Germanic (AD 400-750) and Viking (AD 750-1050) periods

The low salinity and high productivity conditions at Kilen extend into the Germanic period, with higher salinity organisms re-appearing around ~AD 600 (e.g. *Bittium reticulatum*). This date broadly agrees with Kristensen *et al.*, 1995, after which they suggest conditions in the Limfjord follow a gradual shift back towards higher salinity, associated with steadily rising sea level (e.g. Christiansen *et al.*, 1990; Tanner, 1993). The rapidity of the shift is unknown for the Kilen profile as no diatom samples have been analysed between ~AD 600 (still low salinity; ~15 ‰) and ~AD 800 (~33 ‰). Unlike, the anomalous Roman period, high salinity in the Limfjord during the Viking period is consistent with high surface salinity stands in the Skagerrak (Hebbeln *et al.*, 2006) and possibly in the Baltic Sea (Emeis *et al.*, 2003). Evidence for a high sea level stand during the Viking period is also apparent in a number of spatially distributed sea level records from the North Sea and, into the Baltic region (Mörner, 1976; Miller, 1982; Miller and Hedin, 1988; Christiansen *et al.*, 1990; Tanner, 1993; Behre, 2007). At least for part of the Viking period, both the north and west entrances were probably open, enabling salinity to reach near fully marine values (Kristensen *et al.*, 1995; Christensen *et al.*, 2004; this study, Figure 6.4). The exact dates for the various openings and closing of these entrances throughout the Germanic, Viking and Medieval period, however, remain open to debate.

The fall in sedimentary pigment accumulation rates (and concentrations) in the Kilen profile at ~AD 800 might mark the beginning of a maximum salinity phase in the Viking period (e.g. DI-salinity of 33 ‰). The intensified marine conditions (e.g. greater energy and coarser sediments) possibly caused pigment preservation conditions to deteriorate, as indicated by the increase in the pheophytin-a/chlorophyll *a* (degradation) ratio. At ~AD 900, productivity (and surface salinity) within the Skagerrak also appears to decrease, which has been attributed to increased Baltic outflow of low salinity water (Hebbeln *et al.*, 2006). This decrease in productivity, however, appears to be confined to the Limfjord-Skagerrak region, as the Baltic Sea records indicate higher productivity (Andrén *et al.*, 2000; Zillén *et al.*, 2008), likely associated with increased sea level (Miller, 1982), greater water exchange and warmer temperatures associated with the Medieval warm period (~AD 750-1200; Briffa *et al.*, 1990; Briffa *et al.*, 1992; Seppä *et al.*, 2009).

In terms of commercial value, the archaeofaunal evidence suggests that the Limfjord remained an important fishery throughout the Viking Age with flounder, garfish and whitefish the most abundant catches (Enghoff, 1999). Plaice and eel were also important, probably during the maximum salinity phase (Enghoff, 1999; Holm and Bager, 2002). High resolution diatom analyses are required to fully examine changes in salinity and the Limfjord environment over the last ~2,000 years. This remains possible at Kilen, providing the late-Holocene dating issues can be overcome, but additional analyses from other Limfjord sites would also offer information regarding spatial variability and regional differences. Further research in this area is essential for making more confident linkages between sea and society, and understanding why shell middens are absent from the Limfjord throughout the Iron Age.

6.4.3 The Limfjord during the Historical Period (AD 1050-present)

Despite the absence of shell middens in the post Iron Age period, documentary evidence suggests that humans continued to rely on marine resources from the Limfjord throughout historical times (Enghoff, 1999; Holm and Bager, 2002; Poulsen *et al.*, 2007). For example, the important plaice and eel fisheries of the Viking period (Enghoff, 1999) were replaced by a rich herring fishery after ~AD 1200 (Holm and Bager, 2002; Poulsen *et al.*, 2007), following the closure of the Limfjord (see section 3.11). These lower salinity waters provided ideal feeding grounds for herring, but the prosperity of this fishery remained very closely connected with environmental conditions, most notably salinity and exposure to the North Sea. On several occasions this fishery was disrupted, particularly in the 1600s when storms breached through the western sand deposits (e.g. AD 1624), enabling highly saline marine water from the North Sea to enter and salinity to rise (Holm and Bager, 2002; Ryves *et al.*, 2004). The fishery was able to recover from these temporary openings several times, prior to its final collapse shortly after ~AD 1825, when marine water broke through the Agger Tange sand dunes and created a permanent marine connection with the North Sea, kept open in modern times by active dredging of the channel (Holm and Bager, 2002). Unfortunately, the poor dating in this section of the Kilen profile (i.e. ~AD 1200-1920), prevents detailed examination of interactions between sea and society throughout this brackish phase of the Limfjord's history. Furthermore, it is unlikely that short term storm events and shifts in salinity would be discernible from the record due to the poor resolution and mixing of sediments in this section (though until full diatom counts are made this remains uncertain).

6.4.4 The limnic phase of Kilen (AD 1856-present)

The uppermost section of the Kilen profile represents a contemporary problem in many Danish lake/coastal sites (Conley *et al.*, 2000; Conley *et al.*, 2002b; Clarke *et al.*, 2003; Clarke *et al.*, 2006; Ellegaard *et al.*, 2006), where anthropogenic activities (e.g. agriculture, sewage treatment, industrial effluents) are causing nutrient enrichment of aquatic environments, which subsequently affects the ecological status of these systems (see Chapter 3). Unfortunately, Kilen does not offer the potential to explore baseline nutrient conditions (before the large-scale anthropogenic disturbances associated with the industrial revolution) due to its history as a marine environment, prior to the building of the road and rail embankment in AD 1856. The establishment of baseline conditions has been important issue in many nutrient-based transfer function studies in order to be able to understand the natural conditions of a lake/fjord, and to assess fully the extent of human impacts (Andrén *et al.*, 2007a). Furthermore, baseline conditions must be treated with caution, particularly in Denmark, where the landscape has been culturally impacted since the Neolithic period. Kilen does, however, provide a good example of how rapidly humans can influence a system after isolation/reclamation from the marine environment. There is also potential for future comparison of the diatom stratigraphy with contemporary monitoring data available for Kilen between AD 1972-1997, sampled as part of the Danish Nationwide monitoring project (Ringkjøbing Amtskommune, 1988; Kronvang *et al.*, 1993). This will be extremely useful for testing the transfer function inferences (e.g. Figure 3.20) and for a better understanding the interactions between recent diatom communities, key environmental variables and potentially help manage heavily human impacted systems such as Kilen.

6.5 Conclusions

- The diatom-salinity model used in this study (WAPLS component-3 model (r^2_{boot} = 0.923 and RMSEP = 0.36 square root salinity units) is deemed to have produced reasonably accurate reconstructions of palaeosalinity (qualitatively confirmed by ecological knowledge) at each site over the study period. However, there are some circumstances when the diatom-salinity model appears to perform poorly, such as at the high end of the salinity gradient, where salinity appears to be overestimated (i.e. DI-salinity inferences at Kilen >35 ‰ ~250-0 BC). The salinity-model also performs poorly in circumstances where other environmental variables (particularly high nutrient conditions) appear to exert pressure upon the diatom assemblage (e.g. poor agreement between DI-salinity estimates and measured salinities in the uppermost section of the Kilen profile under eutrophic conditions).

- Diatom preservation was generally good at each site (generally **F** = > 0.4 for the majority of the three core sequences), though some exceptions to this were apparent (e.g. lowermost section of the Norsminde Fjord diatom sequence, F= < 0.3 between ~6,900-6,600 BC). Whilst high salinity probably affected valve preservation at certain times in the record (e.g. lower **F** = 0.2-0.4 at Kilen during the pre-Roman period, under high salinity condition, DI-salinity >28 ‰), overall, the salinity-dissolution relationship was weak, with increased dissolution appearing to be more strongly affected by factors such as depth, accumulation rate, sediment re-working, transport and deposition.

- Korup Sø was likely first inundated by the sea sometime between 8,000-7,500 BC, but probably remained a limnic system (subject to the occasional breaches) up until the second half of the Boreal chronozone (~7,500-7,200 BC). After a permanent connection to the Kolindsund was eventually established (i.e. between ~7,500-7,200 BC) salinity increased rapidly in the fjord due to rising sea levels, reaching >15 ‰ by ~6,700 BC. Norsminde Fjord was inundated slightly later than Korup Sø, with a marine connection being established between ~6,900-6,800 BC, also following a limnic or fluvial phase, and by ~6,600 BC was again under high salinity conditions (DI-salinity >20 ‰). The date at which Kilen was inundated by the sea remains uncertain but a recent AMS ^{14}C date (at which time marine diatoms and molluscan are present) suggests that the Limfjord was already under marine conditions by ~6,900 BC, likely predating earlier estimates (i.e. 7,000-6,000 BC).

- The earliest Danish coastal shell middens are dated to the late Kongemose period (i.e. ~5,600 BC), thereby coinciding with an abrupt (and potentially global) rise in sea level. At this time there is also evidence for increased salinity (initially), decreased sedimentary accumulation rates (at Norsminde Fjord), and an increase in molluscan and foraminiferal concentrations. It is likely that this rapid rise in sea level (between ~5,600-5,000 BP) was the main driver behind these (environmental and ecological) changes and potentially triggered the accumulation of shell middens (i.e. increased exploitation of marine resources) in the late Kongemose/Ertebølle period. Increasing temperatures was most likely an important factor (e.g. longer spawning seasons, reducing winter freezing and thawing) and at this time Scandinavia entered the Holocene thermal optimum (5,550-3,550 BC).

- Maximum salinity at both Korup Sø and Norsminde appears to have occurred somewhere between ~5,700-4,900 BC (rising after ~6,000 BC), but estimates differ slightly between proxies. At Korup Sø, highest DI-salinities (23-27 ‰) occur between ~5,730-5,520 BC, thereby predating the maximum MI-salinities (~25 ‰ between ~5,340-5,250 BC). At Norsminde Fjord, highest DI-salinities occur before (~5,400-5,160 BC under the reduced model) the highest-salinity demanding foraminiferal species reach their maximum abundances (~5,160-4,900 BC). This suggests that salinity and sea level change are not linearly related, and that other factors might be important in driving salinity change at different times in the record, (e.g. freshwater inputs from the catchment, reduced marine connection/exposure due to sediment accretion). Furthermore, biological communities (in which the salinity inferences are based on, e.g. diatoms, molluscs and foraminifera) generally respond to a wide range environmental variables (e.g. changes in nutrients, sediment, depth), some of which might exert greater (or additional) pressure on these organisms at different times, thereby potentially skewing salinity estimates (as demonstrated at Kilen). In this study, the high abundances of *Fragilaria* s.l. at both Norsminde Fjord and Korup Sø (probably indicating general instability within these fjord systems) often made it more difficult to isolate a true salinity signal.

- Contrary to the salinity hypothesis for the oyster decline, there is no convincing evidence from the sites analysed in this study, for a regional shift to lower salinity in the Neolithic period (including a widespread decline at the Mesolithic-Neolithic transition). There is, however, plenty of evidence suggesting that both the marine and terrestrial environmental changed over the late Mesolithic/early Neolithic period and that several environmental factors, probably contributed to the oyster decline. Here it is argued that both regional sea level and climate changes were important drivers, but that the response to these forcing mechanism differed between sites, due to being heavily affected by more localised characteristics (e.g. vegetation cover and change (possibly both climatic and human forced, site configuration, exposure etc).

- The oyster decline in the shell middens was likely to be driven by a complex series of events including changes in the sedimentary regime (i.e. increased sedimentary accumulation rates evident at Norsminde Fjord and Korup Sø) and short term stresses associated with salinity (i.e. evidence for a short term decline at Kilen and fluctuating DI-salinities in both the Norsminde Fjord and Korup Sø record), productivity and temperature events (i.e. decreasing temperatures and less stable climate after the Holocene thermal maximum). Higher sedimentary accumulation rates (associated with lower sea level and catchment vegetation changes) might have pushed oysters out of the inner more human accessible parts of the fjord. Lower temperatures and fluctuating salinities might have reduced/ prevented spawning in some years and even caused widespread mortality in extreme winters, suppressing oyster populations. Subsequent human exploitation would have placed greater pressure on oyster populations, reducing their dependability as a resource and therefore culminating in their decline in the middens, as humans shifted towards a more terrestrial-based diet.

In terms of the introduction of agriculture it is likely that the oyster decline formed part of a complex transition driven by a mixture of environmental and cultural factors.

- There is some evidence for higher salinity over the Pitted Ware and Single Grave cultural period (~2,800-2,400 BC) from the records presented here. At Kilen, DI-salinity increases throughout the course of the Pitted Ware/Single Grave cultural period (24-27 ‰). At Norsminde Fjord, higher DI-salinity exists between ~3,070-2,900 BC and ~2,700-2,570 BC (23-25 ‰ under the full model) and at Korup Sø, increased abundances of *Hinia reticulata* and *Onoba vitrea* are present between ~3,000-2400 BC. Higher salinity (and perhaps changing sedimentary conditions) might have increased oyster abundances and/or enabled them to re-colonise the inner more accessible parts of the fjord at this time. However, before a positive connection between environmental change, oyster abundance and human exploitation can be made more work needs to be done (including better dated sea level records covering this period and more information concerning salinity change and the sedimentary regime at a variety of coastal sites).

- The Iron Age (~500 BC-AD 1050) is a dynamic phase in Limfjord history characterised by large scale shifts in salinity and productivity, resulting from varying degrees of connection with the North Sea (to the west) and the Skagerrak (to the north), and heavily affected by sea level and sedimentary processes (i.e. sedimentary accretion and erosion). The large number and size of shell middens (generally dominated by *Mytilus edulis*) dating to the early Iron Age (particularly ~300 BC-AD 200) clearly demonstrates that marine resources were important to cultures and societies living in Denmark at this time. Near fully marine salinities in the late pre-Roman period (i.e. >30 ‰) suggest that the Limfjord was heavily exposed to the North Sea and Skagerrak, likely being connected in both the west and north in this period. This is succeeded by a rapid decline in salinity in the early Roman period, driven by a sea level regression in the North Sea (~150 BC-AD 50) and subsequent accretion of sediments, blocking off the Limfjord entrances and isolating it from the North Sea and Skagerrak. Consistent with other Danish and Baltic studies, increased marine productivity is apparent (particularly macrophytes and molluscs, including *Mytilus edulis*) at Kilen over the Roman period, but the contrasting point is that this occurs under lower salinity conditions in the Limfjord region (rather than high salinity in these other regions). The reasons for the lack of shell middens being present along the shoreline of the Limfjord remains uncertain. Several possible reasons have been suggested here, (e.g. primary importance placed on alternative food sources absent from the Limfjord (e.g. oysters), sporadic collection of mussels from different localities, preventing the temporal accumulation of shell middens) but these remain speculative.

- Modelling species-environment relationships (using multivariate techniques) has demonstrated that a wide range of terrestrial and marine changes can influence the diatom distributions present in Danish coastal systems. The most important variables driving changes in the diatom datasets also appear to change over time as these complex coastal systems develop both naturally, and in response to humans altering of the landscape. At Norsminde Fjord, climate and climate-related processes (e.g. indicated by temperature), together with predictors for marine environmental change (e.g. macrophyte bedding, catchment inwash and erosion probably associated with sea level rise) tend to be most important at the bottom of the profile (~6,360-3,430 BC). Catchment processes (as indicated by charcoal, minerogenic fluxes and predictors of vegetation change) are the most significant variables in the middle part of the profile (5,400-2,570 BC), and a series of catchment (e.g. minerogenic input), climatic (decreasing temperatures) and human related processes (forest clearances and expansion of agricultural lands) dominate in the upper section (4,360-2,020 BC). Similarly, at Korup Sø, in the lower section (between ~7,330-3,400 BC) predictors for marine environmental change (e.g. sea level, sedimentation rate, changing mollusc faunas) are generally more common (and tend to explain large amounts of the variation, up to 49 %), whereas catchment processes/vegetation changes (probably heavily driven by human disturbance after ~3,900 BC) are more important in the upper section of the record (~4,870-2,380 BC, generally explaining 8-20 % of the variation in the diatom dataset).

Chapter 7. References

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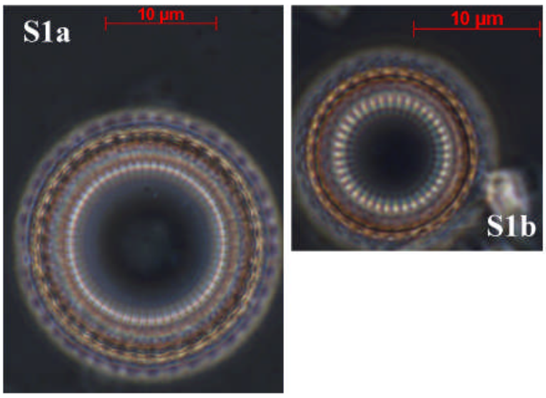
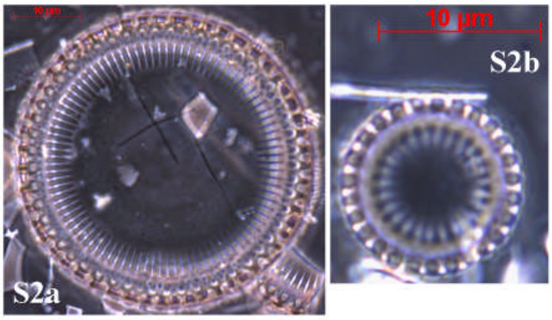
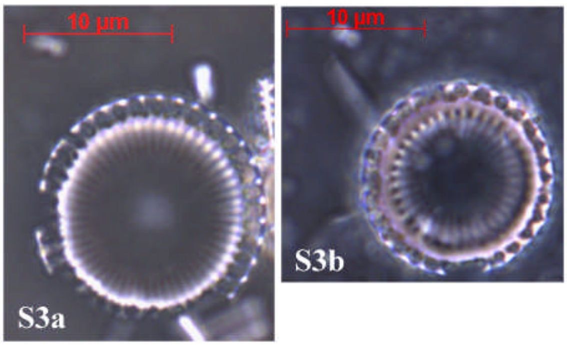
Appendix A. *Paralia sulcata* dissolution stages



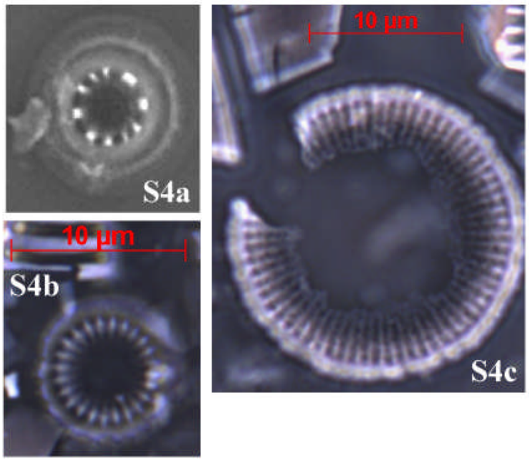
Stage 1 - Valves in pristine condition showing very little or no signs of dissolution.



Stage 2 - Valves clearly showing all of the main identification features visible under light microscopy but exhibiting some preliminary signs of dissolution. This might include slight wearing of the marginal rim and/or signs of the radial markings beginning to dissolve.



Stage 3 - Valves showing clear advanced signs of dissolution being in a relatively poor state. Marginal rim showing substantial dissolution. Radial markings generally visible, but in some specimens can be poorly preserved appearing as irregularly shaped silica blotches on the valve surface.



Stage 4 - Valves in very poor state with outer margin (almost) completely dissolved away and radial markings commonly showing clear signs of dissolution, more regularly appearing as silica blotches on the valve surface (e.g. S4a). In some valves radial markings might even be absent. (NB. Scale unavailable for photo S4a).

*All *Paralia sulcata* dissolution images are taken from fossil material present in Danish coastal sediments.