Climate development and its effect on the North Sea environment during the Late Holocene

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Abstract

The climate development of the Late Holocene (last 2,700 years) is generally characterized by several changes, e.g. the decrease and increase of (air) temperature. Accordingly, in comparison with the present temperature average, this time interval can be divided in cooler phases, such as the Subatlantic Pessimum, the Migration Period and the Little Ice Age, which alternated with time intervals in which the climatic conditions were ameliorated, i.e. with the Roman Period, the Medieval Warm Period and the present Modern Optimum.

In this study Late-Holocene climatic changes as well as effects on the environment within the North Sea region were investigated in the two main accumulation areas of the North Sea: the Helgoland mud area (German Bight, southern North Sea) and the Skagerrak (eastern North Sea). Sediment cores from either area served to reconstruct the Late-Holocene development within the North Sea area. The methods used included analyses on grain-size distribution, stable isotopes and foraminifera.

While in the geological past primarily natural components were forcing climate change, mankind is increasingly influencing the system since the early 20th century. Human impact on the coastal environments around the North Sea, however, dates much further back in time. For instance, dike building began in the late 11th century. Another example of an anthropogenic effect on the environment is the extensive deforestation since the 6th century leading to soil erosion in Central Europe, with its consequences for sediment loads transported by the rivers. Such human-induced effects on the environment of the North Sea are discussed, focusing on possible factors that leave an imprint on the more recent sedimentary history of the German Bight. As one of the most important natural processes affecting the marine environment of the southern North Sea the storm-flood activity is considered, with special attention to the conditions during the Late Middle Ages and the Little Ice Age. Albeit, the disintegration of the island of Helgoland seems to play a key role in the earlier depositional history, leading to relatively high sedimentation rates (>13 mm/year) prior to AD ~1250.

Only within the course of the Modern Optimum, technical progress and the development of adequate instruments allowed to obtain trustworthy measurements of meteorological and hydrographical parameters. Sea-surface temperature and sea-surface salinity (SSS) data, measured at Helgoland Reede, are available for the last 100 years. From these data sets we were able to evaluate and calibrate a stable oxygen isotope ($\delta^{18}O$) record of a sediment core from the Helgoland mud area. The analyses reveal a clear dependence of the $\delta^{18}O$ signal on SSS, a parameter which is mainly driven by the Elbe discharge. Since the $\delta^{18}O$ series is covering the previous 800 years, a high-resolution reconstruction of the parameters salinity and discharge further back in time became

possible as well. Within this time span of 800 years, estimated paleosalinity values range between 29 and 33.9 psu, and the estimated paleodischarge varies between 100 and 1375 m^3/s. An increased freshwater input via the Elbe River, resulting in a lowered SSS signal in the near-coastal area, is a consequence of relatively higher precipitation rates in the large Elbe catchment area. The variation of precipitation in Central Europe is thus also mirrored in the Helgoland isotope data.

The onshore precipitation pattern also appears to be displayed in the sediments of the Skagerrak, but on a different temporal resolution. There, the relatively cool climatic periods, such as the Migration Period, are characterized by high productivity, which may have resulted from pronounced precipitation supplying nutrients via the main rivers of Northern and Central Europe into this area. To investigate this coherence, stable oxygen and carbon isotopes, which reflect mainly temperature conditions and nutrient availability, were analyzed in addition to benthic and planktonic foraminifera data.

Since the end of the Migration Period, at around 1200 years before presence, there is foraminiferal evidence of a reduction of North Atlantic water inflow into the Skagerrak. The change in Atlantic influence is most likely due to a change in the atmospheric circulation pattern. As a consequence, the sediments along the southern slope of the Skagerrak basin are remobilised and the grain-size distribution in the Skagerrak gradually changed to coarser sediments.

Zusammenfassung

Die Klimaentwicklung wurde während des späten Holozäns (der letzten 2,700 Jahre) von mehreren Veränderungen, wie z.B. dem Abfallen und Ansteigen der (Luft-)Temperatur, geprägt. Dementsprechend kann dieser Zeitabschnitt, im Vergleich zum derzeitigen Temperaturmittel, in kühlere Phasen, wie dem Subatlantischen Pessimum, der Völkerwanderungszeit sowie der Kleinen Eiszeit unterteilt werden, die sich mit den klimatisch günstigeren Zeitintervallen, der Römerzeit, der Mittelalterlichen Wärmeperiode und dem heutigen Modernen Optimum abwechseln.

In der vorliegenden Arbeit wurden die spätholozänen Klimaänderungen und Effekte auf die Umweltbedingungen speziell in der Nordseeregion untersucht. Dazu wurden an Sedimentkernen, die aus dem Helgoländer Schlickgebiet (Deutsche Bucht, südliche Nordsee) sowie aus dem Skagerrak (östliche Nordsee) stammen, den beiden wichtigsten Akkumulationsräumen der Nordsee, Korngrößen-, Isotopen- und Foraminiferen-Analysen durchgeführt. Mit deren Hilfe konnte ein wichtiger Beitrag zur Rekonstruktion der spätholozänen Entwicklung im Nordseeraum geleistet werden.

In der geologischen Vergangenheit waren vorwiegend natürliche Faktoren für Klimawechsel verantwortlich. Seit dem frühen 20. Jahrhundert greift allerdings der Mensch in zunehmendem Maße in das System ein. Der menschliche Einfluss auf die küstennahe Umgebung der Nordsee begann jedoch schon viel früher. Zum Beispiel begann man im späten 11. Jahrhundert mit dem Deichbau. Ein weiteres Beispiel anthropogen bedingter Umwelteffekte ist die extensive Abholzung, die in Mitteleuropa seit dem 6. Jahrhundert zu Bodenerosion und den damit verbundenen Konsequenzen auf die Sedimentfracht in den Flüssen geführt hat. Derartige vom Menschen verursachte Effekte auf die Umweltbedingungen der Nordseeregion werden diskutiert, insbesondere im Hinblick auf Faktoren die die jüngere Sedimentationsgeschichte in der Deutschen Bucht prägten. Als einer der wichtigsten natürlichen Prozesse auf die marine Umwelt der südlichen Nordsee gilt die Sturmflutaktivität, vor allem während des späten Mittelalters und der Kleinen Eiszeit, wenn auch in früherer Zeit (vor AD ~1200) die Erosion der Insel Helgoland eine Schlüsselrolle im Sedimentationsgeschehen zu spielen scheint, indem sie zu relativ hohen Sedimentationsraten (>13 mm/Jahr) geführt hat.

Der technische Fortschritt und die Entwicklung adäquater Messgeräte im Verlauf des Modernen Optimums ermöglichen seit dieser Zeit verlässliche Messungen meteorologischer und hydrologischer Größen. So stehen von der Station Helgoland Reede Messreihen der Oberflächenwassertemperatur und -salinität (SSS) für die letzten 100 Jahre zur Verfügung. Durch diese Datensätze war es möglich, stabile Sauerstoffisotope ($\delta^{18}O$) eines Sedimentkerns aus dem Helgoländer Schlickgebiet zu evaluieren und zu kalibrieren. Die Analysen zeigen deutlich eine

Abhängigkeit des $\delta^{18}O$-Signals zur Salinität, die hauptsächlich über den Abfluss der Elbe gesteuert wird. Da die $\delta^{18}O$-Zeitreihe insgesamt die letzten 800 Jahre abdeckt, ist zudem eine hochauflösende Rekonstruktion der Parameter Salinität und Abfluss möglich. In dieser Zeitspanne von 800 Jahren variieren die Werte der abgeschätzten (Paleo-)Salinität zwischen 29 und 33.9 psu und die Abschätzung der (Paleo-)Abflüsse ergibt Werte zwischen 100 und 1375 m^3/s. Verstärkte Süßwasserzufuhr über die Elbe, aus der wiederum geringere Salzgehalte im küstennahen Bereich herrühren, ist das Resultat stärkerer Niederschläge im ausgedehnten Einzugsgebiet der Elbe. Daher können ebenso mitteleuropäische Niederschlagsänderungen aus den Helgoländer Isotopendaten rekonstruiert werden.

Das landseitige Niederschlagsmuster wird auch in den Sedimenten des Skagerraks deutlich, wenn auch in anderer zeitlicher Auflösung. Dort sind relativ kühle Klimaphasen, wie z.B. die Völkerwanderungszeit, durch höhere Produktivitäten gekennzeichnet, die wahrscheinlich auf stärkere Niederschlagsereignisse, welche Nährstoffe über die großen Flüsse Nord- und Mitteleuropas in die Region bringen, zurückzuführen sind. Um diesem Zusammenhang nachzugehen, wurden zusätzlich zu Informationen aus benthischen und planktischen Foraminiferen-Daten auch stabile Sauerstoff- und Kohlenstoffisotope analysiert, die die Temperaturverhältnisse bzw. die Nährstoffverfügbarkeit anzeigen.

Die Abnahme planktischer Foraminiferen seit dem Ende der Völkerwanderungszeit (etwa 1200 Jahre vor heute) deutet auf eine Reduktion des Wasserzustroms vom Nordatlantik in den Skagerrak hin. Dieser Wechsel im nordatlantischen Einfluss kann auf eine Veränderung der atmosphärischen Zirkulation zurückgeführt werden. Als Folge werden Sedimente am südlichen Hang des Skagerrakbeckens mobilisiert, wodurch sich die Korngrößenverteilung verändert; das Sediment wird grobkörniger.

1 Introduction

During the late Holocene, i.e. during the last approximately 2,700 years, climate development is characterized by several changes. While at first natural components e.g. ocean-atmosphere interactions were forcing climate change, the latest shift is being forced by i.a. enhanced 'greenhouse' warming, ozone depletion, urban growth and atmospheric pollution - each of these latter causes related to human activity (e.g. IPCC, 2001; Kalnay and Cai, 2003).

Understanding natural changes is an essential component of modern climate analyses. Only this allows them to be distinguished from anthropogenic influences and on this basis, predictions of future climate development become achievable. Against the background of the growing world's population and its consequences for human health and ecosystems such knowledge is important.

Whatever anthropogenic effects there are on climate in the future, they will be superimposed on the underlying background of natural variability, which varies on all timescales. On climate variability, its relationship to natural forcing mechanisms and to feedback effects, paleoclimatic research provides information. Such perspectives cannot be derived from the limited set of instrumental data alone. Moreover, the period for which detailed instrumental records exist coincides with the time when human activities have increasingly influenced the natural system.

Since coastal areas are among the most densely populated and intensely used areas of the world, it is indispensable to learn about these. In this thesis the focus is on the North Sea area. Marine sedimentary sequences were obtained from suitable sites, characterized by continuous, high accumulation. Among the very few of such sites, the Skagerrak and the Helgoland mud area are to name. Natural proxies were analyzed, compared with other proxy as well as instrumental data, and used to reconstruct European climatic and environmental conditions.

1.1 Motivation and main objectives

The purpose of this study was to investigate the effects of climate development on the North Sea environment during the Late Holocene. Therefore, sediment cores of two of the main accumulation areas of the North Sea providing continuous information about paleoenvironmental conditions were analyzed. The analyzed proxies give clues to variations in natural- as well as anthropogenic-induced changes, and allow a high-resolution (annual-decadal) reconstruction.

Focusing on the climate history and the associated natural processes in the North Sea area several questions arise, which are addressed in five manuscripts:

- Is it possible to distinguish climatic variations from marine proxy data of Northern Europe and link them to epochs of the Late Holocene, knowing that the latter are usually derived from terrestrial data, and possibly not valid for entire Europe?
- How practicable are comparisons with other available proxy data?
- To what extent are natural processes, such as storm-flood activity and shoreline displacement, affecting the marine environment of the North Sea?
- Which processes are dominating the marine signal of environmental variation?
- Since the Late Holocene includes the main period of colonization of the North Sea coastal area, are human impacts distinguishable from natural ones, and were the effects of human action already becoming noticeable prior to industrialization?

1.2 European climate development of the Late Holocene

Commonly, measured or reconstructed (air) temperatures are used to distinguish climatic periods. Historical events serve names and ease temporal correlation. During the Late Holocene several climatic shifts occurred (Figure 1.1). Temperature was fluctuating with an amplitude of about 1-2°C above and below the present average (e.g. Schönwiese, 1988), and accompanied to this the natural system changed significantly on all spatial scales. The earliest climatic phase which is considered in this study is called the Subatlantic Pessimum beginning around 3150 BP and ending between 2500 and 2250 BP (Lozán, 1998). This phase is followed by the Roman Period, the Migration Period, the Medieval Warm Period and the Little Ice Age (LIA). These naturally induced (e.g. by solar and volcanic forcing) climatic fluctuations are followed by the most recent climatic period, the Modern Optimum (Schönwiese, 1988), which is characterized by an increasing anthropogenic influence and is well-documented by reliable sources and measured data sets. However, there are indications that human impact, in particular emissions of the greenhouse gases CO_2 and CH_4, had already started 8000-5000 years ago – long before the industrial era (Ruddiman, 2003).

The Subatlantic Pessimum (3150-2500 BP) is supposed to represent cooler conditions, where the terrestrial temperatures were c. 2°C below the present average (Lamb, 1977). There is also evidence for more precipitation (Lamb, 1977; Fairbridge, 1987). Subsequently, glaciers advanced (Lozán, 1998), and in the North Sea area the sea level was lowered about half a meter (Freund and Streif, 2000).

According to historical sources, mild temperature conditions are attributable to the Roman Period (2150-1550 BP). However, several cooling and storm events point to rather unstable atmospheric conditions (Lamb, 1977; Lozán, 1998), and storm-flooding events displaced the shoreline of Denmark and Germany.

As a consequence of phases of intense dryness, tribes from the Russian steppe were migrating westwards and were claiming land in north-western Europe around 1550-1250 BP (Lozán, 1998). Besides this resettlement, climate deterioration in northern parts of Europe provoked a migration as well; the Germanic Migration. There might have been additional reasons for this migration, but the unfavorable climatic conditions certainly played an important role. Within the Migration Period temperatures decreased and precipitation increased at the same time.

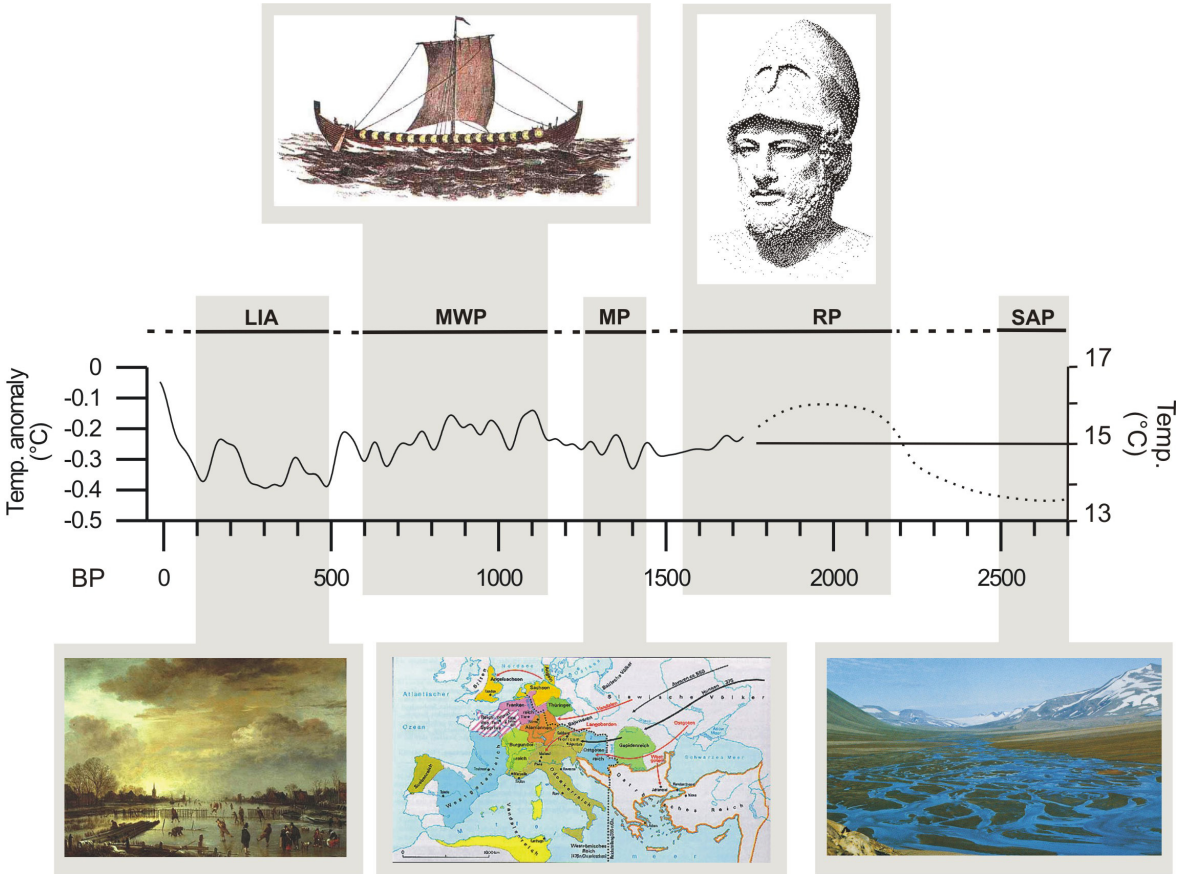

Figure 1.1: The timing of climatically differentiated periods of the Late Holocene (last 2,700 years), and Northern Hemisphere temperature anomalies (full line; left y-axis) reconstructed by Mann and Jones (2003), continuing as sketched mean temperatures (dotted line; right y-axis) in relation to the present average of 15°C modified after Schönwiese (1995). SAP, Subatlantic Pessimum illustrated by a tundra environment; RP, Roman Period illustrated by Perikles, a Grecian-Roman politician; MP, Migration Period illustrated by a map of the tribal movements; MWP, Medieval Warm Period illustrated by a Viking ship; LIA, Little Ice Age illustrated by a contemporary painting.

In the beginning still rich in precipitation, the climatic conditions became dryer during the course of the Medieval Warm Period. Mean annual temperatures were about 1.5°C above the present average, and the summer as well as the winter seasons was mild (Lamb, 1977). As a characteristic of this time period, inhomogeneous climatic effects on a regional and temporal scale are to mention. Furthermore, the Medieval Warm Period is divided into two phases: the Middle Ages (c. 1150-700 BP) and the Late Middle Ages (700-600 BP). While the Middle Ages are more temperate, the latter phase is connected with a marked cooling.

The general climatic amelioration during the Medieval Warm Period is followed by the LIA. In Northern and Middle Europe this time period lasted from 600 to 100 BP (Grove, 2002). Annually averaged temperatures were about 1°C lower than the present average. Relatively cool summers, and winters becoming more and more chilly were typical for this period. The most distinct changes, however, were linked to the seasons spring and autumn, which became significantly cooler. Besides the cooling, several phases of increased wetness were recorded.

Already since the end of the Medieval Warm Period and especially during the LIA the atmospheric circulation pattern changed. An intensified meridional component, resulting from an intensified north-south temperature difference, was in causal connection to a pronounced continental climate in Central Europe (Glaser et al., 2000). As a consequence, an increased north-south temperature difference resulted in an increased storm-flood activity in the North Sea region. Storm-flooding events like the “Große Mandränke” (1362), the October (1634) or the Christmas (1717) floods, caused huge losses of inhabitants and land (Liedtke and Marcinek, 1994; Reineck, 1994), and required enormous efforts to protect the coastline (Flemming and Mai, 1998).

With the beginning of the 20th century, the Modern Optimum set in (Schönwiese, 1988). Increasing temperatures were already recognizable around 1870 (Jones et al., 2001), and the distinct continental-climate conditions of the LIA vanished. Moreover, e.g. a change in greenhouse-gas concentrations results in the observed human-induced temperature increase of the last approximately 100 years (e.g. IPCC, 2001).

1.3 Climate reconstructions – unique possibilities in Europe

Nowadays and in the more recent past, meteorological and hydrographical measurements are used as a basis for climatologic studies. Such time series are available for the last approximately 150 years, with some extra-long instrumental records (~250 years) mainly describing the European climate history. Further back in time, climatic information has to be

reconstructed. Therefore, proxies are contributing to the knowledge of climatic development. Proxies are of two basic kinds, natural (physical or biological) and documentary (written) archives.

Natural archives, such as trees, ice cores, corals and sediments are the basis of natural proxy data. From suitable sites all over the world, such proxy records can be derived, having the potential to reach back in time much further than any documentary record. In general, proxy data allow to obtain a relatively constant quality of reconstructions of various parameters. Despite these achievements, dating and data resolution are causing difficulties.

The term "documentary archives" mainly refers to written historical observations. In particular in Europe, such sources are usable for climate reconstructions for the last millennium. Documents are valuable and important data sources for the more recent, pre-instrumental period. For this period of time, their age control is comparatively reliable. However, an accurate determination of specific parameters is problematic sometimes, and other uncertainties interpreting the data may occur additionally (see chapter 6).

In this context it must be kept in mind that most of the documentary data provides information from inland. The source areas of the various natural proxies, however, may either be terrestrial or marine. For this reason, comparisons and interpretations may imply even more details. Also from this point of view, the North Sea is a potentially interesting, promising place that can contribute to (European) climate reconstructions.

1.4 Geographic and oceanographic setting of the North Sea area

The North Sea is a semi-enclosed marginal sea of the North Atlantic, which is widely opened to the north (Figure 1.2). The comparatively narrow English Channel connects it to the North Atlantic in the southwest. The easternmost part of the North Sea, the Skagerrak, forms the transition to the Baltic Sea, with the latter being separated from the North Sea by the Kattegat. In general, the North Sea can be divided into three different regions:

1) A shallow southern part (water depth: few tenths of meters to 50 m) covering about one-fourth of the total area. The German Bight (mean depth of 22.5 m) is located in the southeastern part.
2) A northern part (water depth: 50 m to 200 m), with an abrupt increase in depth at about 100 m.
3) The glacially-eroded Norwegian Trench (water depth: 270 m to 700 m) penetrating from the Norwegian Sea and continuing along the coast of Norway to the Skagerrak.

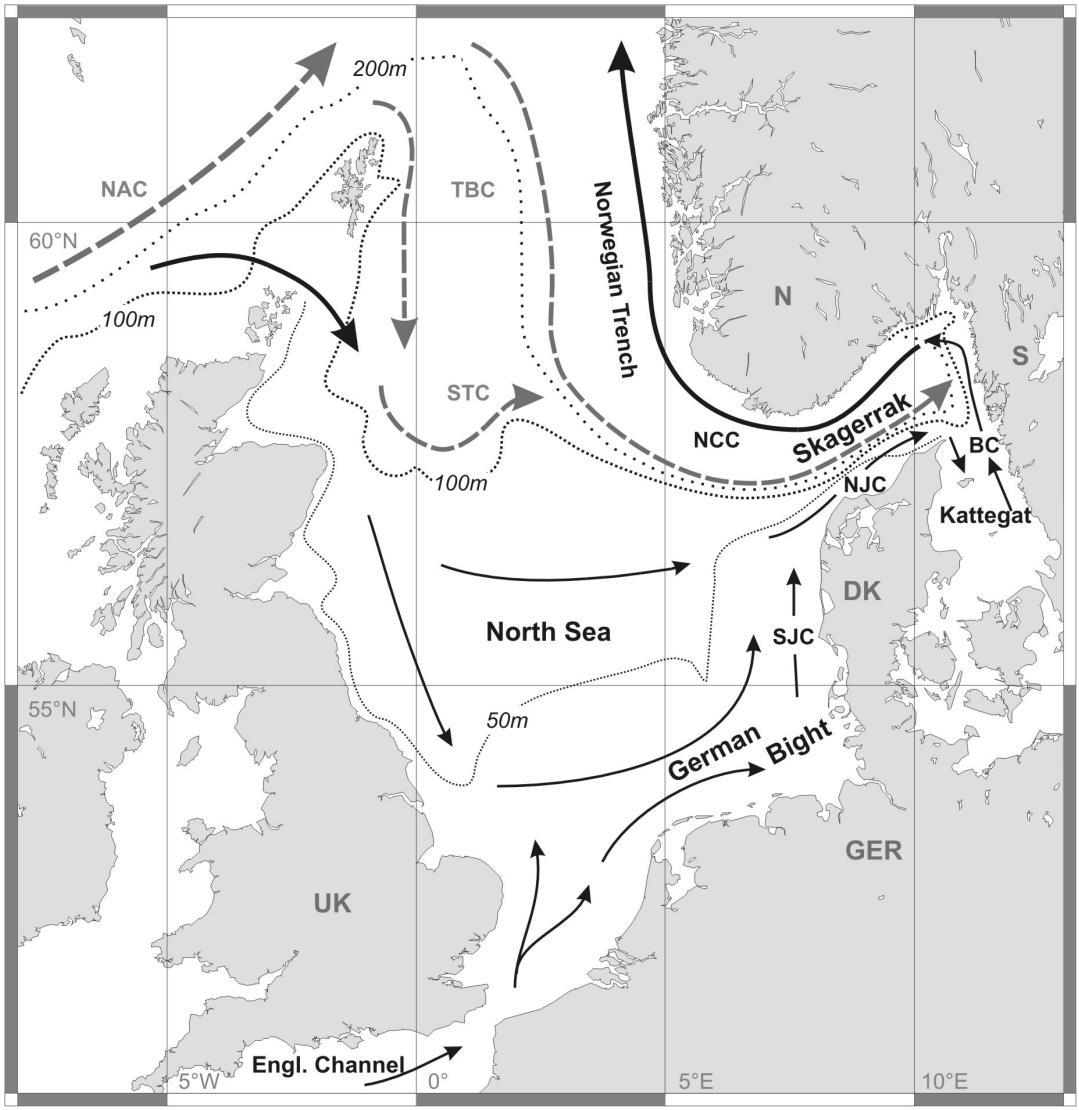

Figure 1.2: The North Sea area. The contour lines of 50, 100 and 200 meters water depth as well as the prevailing surface currents are displayed (see text). Abbreviations: BC, Baltic Current; DK, Denmark; GER, Germany; N, Norway; NAC, North Atlantic Current; NCC, Norwegian Coastal Current; NJC, North Jutland Current; S, Sweden; SJC, South Jutland Current; STC Southern Trench Current; TBC, Tampen Bank Current; UK, United Kingdom. Modified from Danielssen at al. (1991) and Nordberg (1991).

The general circulation of the North Sea runs counter-clockwise, with the most important in- and outflow concentrated in the northern areas and the Norwegian Trench (Rodhe, 1998). Saline water masses from the Atlantic Ocean (Figure 1.2: gray, stippled arrows) enter the North Sea near the Orkney and Shetland islands. It appears that the volumetrically largest amount of inflow from the northwest is guided eastwards to the Norwegian Trench by the topography along the 100-m contour. Only a small portion flows southwards along the Scottish and English coasts, and is

supplemented by a minor amount of North Atlantic water entering via the English Channel. On their way, these water masses are modified (Figure 1.2: black arrows) by e.g. freshwater input of the rivers Thames, Rhine, Weser and Elbe as main contributors affecting the chemical composition as well as the sediment load. Eventually, the water masses head towards the northeast and north, respectively, passing through the Skagerrak before leaving the North Sea along the eastern slope of the Norwegian Trench. This current pattern of the North Sea is also related to the large-scale atmospheric circulation system. For instance, westerly winds enhance the currents, whereas easterly wind directions can hamper the water mass circulation (Dooley and Furnes 1981, cited in: Rodhe, 1998). Therefore, atmospheric forcing (e.g. wind stress and heat flux) is regarded as one of the main factors determining the dynamics of the North Sea.

Due to its high tidal- and wave-energy level, the North Sea is considered as a highly dynamic shallow sea. Sediment redistribution is the predominating process in shaping the sedimentary environment. Hence, sediments are only deposited in a small number of areas (Lohse et al., 1995). Among the few areas where accumulation takes place, the Skagerrak is by far the most important one, and the Helgoland mud area in the German Bight comes a close second as a depocenter of the North Sea (Eisma et al., 1979; van Weering et al., 1987).

1.4.1 Skagerrak

The Skagerrak is part of the Norwegian Trench, and forms the north-eastern extension of the North Sea (Figure 1.2). The mean depth of the Skagerrak amounts to 210 m. The sill is only 270 m deep, whereas a maximum depth of about 700 m is reached in the central part of the Norwegian Trench (Rodhe, 1996). The bathymetry is characterized by a rather flat-floored central basin, and irregular slopes (Figure 1.3). Towards Norway the slope is steep, whereas the eastern and the southern slopes are more gentle forming shallower areas (<100 m). The shallower area is especially enlarged towards Denmark.

The current system of the Skagerrak is influenced by the large-scale atmospheric and oceanic circulation pattern as well as the outflow of the Baltic waters. The tidal motion, with a range of about 20 cm, has little influence only (Svansson, 1975; Rodhe, 1987; Nordberg, 1991; Rodhe, 1998). The hydrographical conditions in the Skagerrak are dominated by the inflow of North Atlantic water masses via the Tampen Bank Current (TBC) and the Southern Trench Current (STC) (Figures 1.2 and 1.3). Modified water masses, e.g. influenced by river discharge, mainly originate from the central and the southern North Sea, and are delivered by the Southern Jutland Current (SJC). This current runs along the western Danish coast and is considered one of the main

contributors to the North Jutland Current (NJC), which flows into the Skagerrak and the Kattegat. Close to the Swedish coast the NJC is supplemented by the outflow of low-saline Baltic waters, then turning towards northwest and generating the Norwegian Coastal Current (NCC), which runs towards the north along the Norwegian coast. Both the surface and the bottom circulation correspond more or less to this pattern, with the main difference that the velocity of the bottom current is lower than that of the surface currents (Qvale and van Weering, 1985; Rodhe, 1987). In general, higher current speeds are connected to the inflowing water masses, constrained on the southern slope.

Most of the waters entering the North Sea and following its anticlockwise movement eventually pass the Skagerrak before leaving along the Norwegian coast. Therefore, the Skagerrak functions as a main area of sediment accumulation, assisted in this role of sediment trap for suspended and traction-transported sediment from the whole North Sea region by its specific topography.

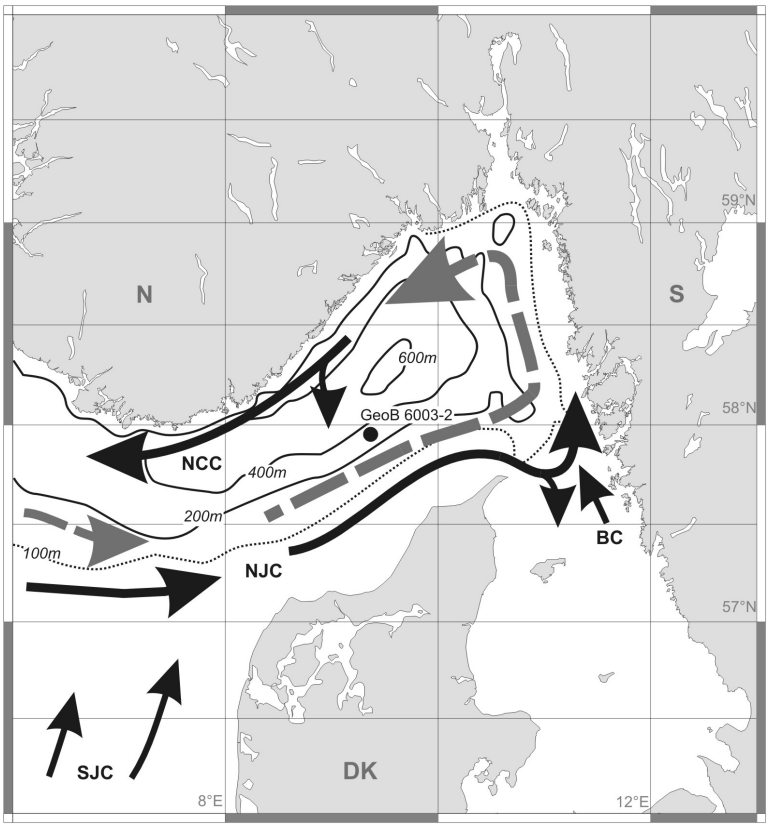

Figure 1.3: The Skagerrak basin with the contour lines of 100, 200, 400 and 600 meters as well as the present-day surface circulation pattern in this area. The location of the studied core is marked by a dot. Abbreviations: BC, Baltic Current; DK, Denmark; N, Norway; NCC, Norwegian Coastal Current; NJC, North Jutland Current; S, Sweden; SJC, South Jutland Current. .Modified from Danielssen et al. (1991) and Nordberg (1991).

1.4.2 Helgoland mud area

South of the island of Helgoland initially two halotectonic depressions were found. While the western depression (Helgoland Hole) still exists more or less today, the eastern depression, the Helgoland mud area, was filled with up to 30 m of Holocene sediments (von Haugwitz et al., 1988) (Figure 1.4). The Helgoland mud area has an extent of approximately 500 km^2 (Irion et al., 1987).

Due to a specific oceanographic setting, deposition takes place in this limited area. A small-scale, cyclonic eddy is generated by the general circulation pattern of the North Sea and is intensified by the inflowing water masses of the Elbe River (Hertweck, 1983). For this reason, the Elbe River is ascribed to play a leading role in forming the Helgoland mud area (Hertweck, 1983; Zuther et al., 2000). Sediment supply, originating from the large catchment, is transported by the river and mainly deposited in the outer parts of the Elbe Estuary. Moreover, the small-scale eddy together with high tidal dynamics (mean tidal range 3 m) mix the freshwater input by the Elbe River with the saline North Sea waters in the shallow (mean depth of about 22.5 m) inner part of the German Bight.

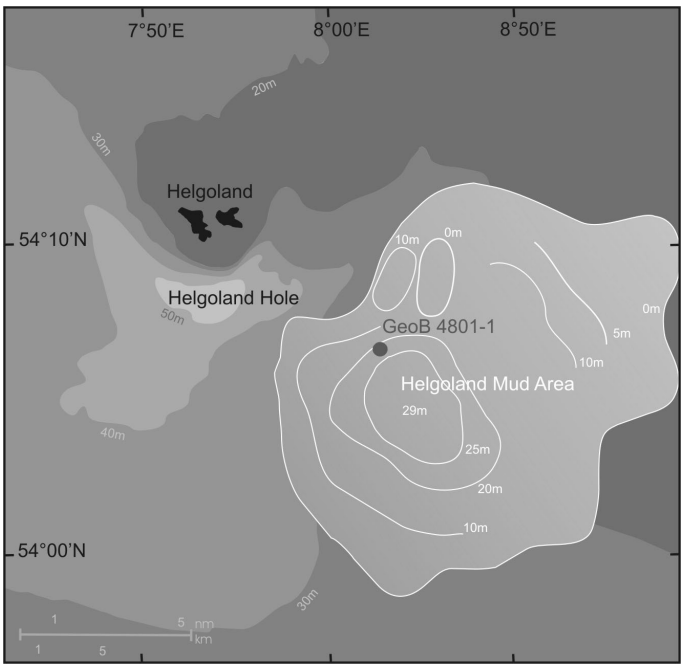

Figure 1.4: The Helgoland mud area is situated south-east of the island of Helgoland. The extent (gray shaded area) and the thickness of the Holocene sediments (white numbers) as well as the location of the core are shown. The Helgoland Hole is located south-west to the island of Helgoland.

However, the origin of the sediments of the Helgoland mud area is still under debate. Based on analyses of sediments from the rivers Ems, Weser, and Elbe, Irion et al. (1987) argue that the accumulated material has not been supplied by river input. They also state that under normal conditions, the river-transported sediments are either trapped in the estuaries themselves (Milliman and Meade, 1983; Irion et al., 1987) or, after bypassing the southern North Sea, the suspended material is deposited in the Skagerrak (Wirth and Wiesner, 1988). However, according to Dellwig et al. (2000) most of the sediments in the Helgoland mud area might be derived from the back-barrier system of the East Frisian islands.

1.4.3 Elbe catchment area

The Elbe River strongly affects the chemical composition of the German Bight, and most likely supplies sediments to the Helgoland mud area (Figure 1.5). Due to its importance concerning the investigated Helgoland mud area, the Elbe River is considered separately in this compilation.

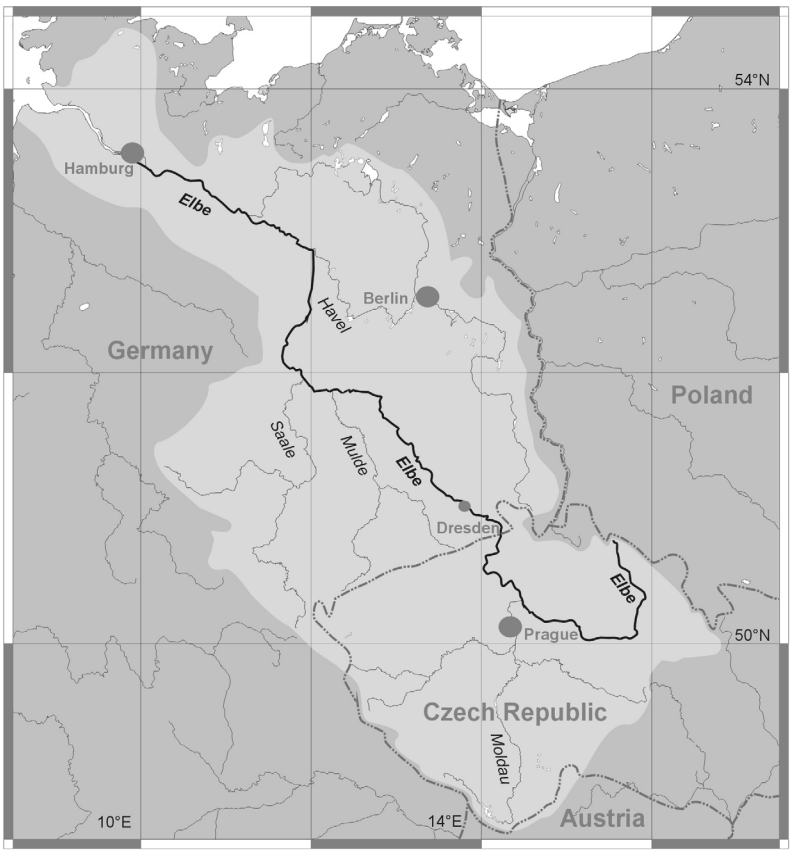

Figure 1.5: The course of the Elbe River (bold black line) and its major tributaries. The catchment area (light gray shading) and the borders of the countries sharing it are indicated.

The Elbe River discharges an area of 149,000 km^2 (Liedtke and Marcinek, 1994). About two-thirds of the catchment belong to Germany, one-third is counted to the Czech Republic (Figure 1.5). In the south, the catchment area is characterized by low-mountain ranges such as the Erz Mountains and the Bohemian Forest. These mountainous areas, relatively rich in precipitation, surround an area of less topography where also less precipitation is recorded because of a rain-shadow effect. In the northern parts of the catchment, the Elbe River runs through the middle and northern lowlands of Germany. From the spring in the Giant Mountains (1,394 m above sea level) to the river mouth at the North Sea a distance of 1,165 km has to be covered.

Today, the mean annual discharge rate amounts to 750 m^3/s (Liedtke and Marcinek, 1994). The regime of the Elbe River is pluvio-nival to pluvial, with generally high discharge rates in wintertime and a period of low discharges in summer (June to November). The snow cover persists in parts of the catchment, characterized as more continental, until March or even May. There, also heavy rainfalls occur in summer, and therefore flooding events are probable until July.

2 Material and methods

2.1 Core locations and sampling

The core material used in this study originates from the two main accumulation areas of the North Sea, the relatively shallow Helgoland mud area and the deeper Skagerrak basin. The sediment cores were obtained during the expeditions M40-0 (1997, Helgoland mud area) and M45-5 (1999, Skagerrak; Neuer and cruise participants, 2000) by the RV Meteor. Detailed site information of the investigated sediment cores are given in the following table:

Table 2.1: Site information of the investigated sediment cores in the Helgoland mud area and the Skagerrak.

	Core No.	Latitude	Longitude	Water-depth (m)	Core length (m)
Helgoland mud area	GeoB 4801-1	54°06.7´N	8°02.2´E	25	11.41
Helgoland mud area	GeoB 4802-1	54°05.7´N	8°04.4´E	25	8.01
Helgoland mud area	GeoB 4805-1	54°06.8´N	8°07.6´E	21	7.34
Helgoland mud area	GeoB 4806-1	54°03.1´N	8°00.6´E	29	7.40
Skagerrak	GeoB 6003-2	57°58.3´N	9°23.2´E	312	10.49

In order to achieve a high-resolution reconstruction of environmental conditions of the North Sea area, measurements and analyses (see below) on these gravity cores (see for a detailed description e.g. Weaver and Schultheiss, 1990) were performed in 1- and 5-cm intervals. Two sets of samples (A and B) were, therefore, taken with syringes (5 and 10 ml). Samples of series A were weighted before and after freeze-drying. With these data, the wet and dry bulk density and, consequently, the water content were determined. The second set of samples (B) was washed over a 63 and a 150 µm sieve to provide i.a. information about the grain-size distribution of the fine-middle sand fraction and to separate the coarse fraction for picking foraminifera tests for the radiocarbon and isotopic analyses.

2.2 Methods

2.2.1 Age determination

For paleoenvironmental studies a reliable stratigraphic control of the sediments investigated is necessary. Only if the sedimentary records are precisely dated, interpretation of the data can yield valid results and enable comparisons with other cores or data sets. In this study, two methods were applied: radiocarbon and ^{210}Pb dating.

Radiocarbon (^{14}C) dating

Modern radiocarbon measurements are performed using an Accelerator Mass Spectrometer (AMS) device, which measures absolute ratios of ^{14}C to ^{12}C, rather than the decay of ^{14}C. From the derived values, a radiocarbon age can be calculated assuming a constant time-independent atmospheric ^{14}C in the past. The radiocarbon ages are given in years before 1950, and referred to as before present (BP).

To date the sediments, in this thesis primarily the radiocarbon method was used. At first, a quite regular spacing (where possible every 50-200 cm) of samples for the ^{14}C datings provided a basis for temporal integration. Afterwards, several more ^{14}C datings were added to improve the age model in some core sections. In most cases, mixed benthic foraminifera were picked from the samples and used for the age determinations. In particular cases, other carbonate material such as shell fragments was selected. Approximately 10 mg carbonate was utilized for the analyses. The material was transferred into graphite at the University of Bremen, and afterwards determined at the Leibniz Laboratory for Age determinations and Isotope Research at the University of Kiel. See Nadeau et al. (1997), for more details about the method used. All ^{14}C ages were calibrated with the Calib 4.3 software (Stuiver and Reimer, 1993; Stuiver et al., 1998), using the marine model calibration curve. A reservoir correction of 400 years is built into the model (Heier-Nielsen et al. 1995b).

^{210} Pb dating

The short half-life (22.3 years) of ^{210}Pb makes this method suitable to the dating of historical age sediments (in the range of the last 150 years). The ^{210}Pb determinations, used for the age-depth models presented in this thesis, were performed at the Royal Netherlands Institute for Sea Research (NIOZ). For a more detailed description of the method see van Weering et al. (1987).

2.2.2 Gamma-ray attenuation

Geophysical properties of sediment cores were measured with a non-destructive GEOTEK Multi-Sensor Core Logger (MSCL) which allows measurements on whole cores (Weber et al., 1997). The set up of the logger is arranged such that the gamma-ray source is located above the core, and the detector (scintillation counter) for measuring the attenuation of gamma-rays is positioned below the core. The number of scintillations over a given length of time is counted giving the count rate (Gunn and Best, 1998). The intensity of the gamma-ray beam is modified after passing an air gap, the core liner and the sediment (unit: counts per second (cps)). While passing the sediment it will be attenuated primarily by Compton scattering (Ellis, 1987), a characteristic depending on the electron density of the material. Most of the sediment-forming minerals have similar values, whereas water has a relatively high electron density. Therefore, information about sediment composition (e.g. terrigenous components, bulk density and porosity) is reflected giving indications on lithological variations (Weaver and Schultheiss, 1990; Gerland and Villinger, 1995; Weber et al., 1997; Gunn and Best, 1998). In this thesis, data of four gravity cores of the Helgoland mud area are used for correlation of the cores, analysed with 5-cm spacing.

2.2.3 X-ray fluorescence core scanner

The x-ray fluorescence, XRF, core scanner (Jansen et al., 1998; Röhl et al., 2000) facility at Bremen was used for the determination of the bulk chemical composition. This scanner allows semi-quantitative analyses of the contents of major and minor elements, and is a non-destructive tool. The split sediment cores (archive halves) were scanned, and measured in 1-cm resolution. The measurements are based on x-rays interacting with the sample, where every atom/element is emitting its characteristic energy and wavelength spectrum. The total fluorescent radiation is received by an energy dispersive detector, to which a multi-channel analyzer is connected. The intensity of the element-specific fluorescent radiation reflects the concentration of this element. These comparatively fast measurements of chemical compounds might be spoiled e.g. by porosity effects and water contents. From the data obtained, only the Zinc (Zn) contents of core GeoB 4801-1 were relevant to this study (unit: cps).

2.2.4 Stable isotopes

The stable oxygen and carbon isotope measurements were performed using an automated carbonate preparation device attached to a Finnigan MAT 251 mass spectrometer. For the isotopic analysis, the powdered carbonate samples (at least 100 µg) were treated with phosphoric acid at a constant temperature of 75°C to form CO_2 gas, which evolves and is measured.

The results of the stable isotope measurements are given in the conventional δ notation relative to the Vienna Pee Dee Belemnite (VPDB) by applying the National Institute of Standards and Technology (NIST) 19 standard. The values of $\delta^{18}O$ and $\delta^{13}C$ are based on the following equations:

$$\delta^{18}O\ (‰) = \{[(^{18}O/^{16}O)_{sample} - (^{18}O/^{16}O)_{standard}] / (^{18}O/^{16}O)_{standard}\}*1000$$

$$\delta^{13}C\ (‰) = \{[(^{13}C/^{12}C)_{sample} - (^{13}C/^{12}C)_{standard}] / (^{13}C/^{12}C)_{standard}\}*1000$$

Analytical standard deviation is about ±0.07‰ VPDB for $\delta^{18}O$ and ±0.05‰ VPDB for $\delta^{13}C$ (Isotope Laboratory Bremen University).

In case of the analyses on core GeoB 4801-1 (Helgoland mud area) 10-20 shells of the benthic foraminifera *Elphidium excavatum* (Terquem)*, forma selseyensis* (Heron-Allen and Earland, 1911) were extracted from the >150 µm-fraction for each measurement. For the analyses on core GeoB 6003-2 (Skagerrak) the benthic species *Melonis barleeanus* (Williamson) and the planktonic species *Globigerinita uvula* (Ehrenberg) were picked from the >100 µm fraction. Whereas 10-20 shells of the benthic foraminifera species were used, approximately 150 individuals of the small planktonic foraminifera species were needed as sample material.

2.2.5 Organic carbon

Freeze-dried sediment samples were used for the determination of organic carbon contents. Homogenized sample material (25 mg) was weighted in a silver foil crucible. Subsequently, the samples were wetted with a few drops of Ethanol to suppress foaming during acidification, and decalcified by drop-wise addition of 6 N HCL. Afterwards, they were dried on a hot plate at 80°C. For the determination of the stable carbon isotope composition of organic matter ($\delta^{13}C_{org}$), the samples were burned in a Heraeus-CHN-analyzer. The evolved CO_2 gas was transferred via an automated trapping box into a Finnigan MAT delta E spectrometer. The $\delta^{13}C_{org}$ values are relative to the VPDB standard, and the reproducibility, which is based on repeated measurements on a laboratory sediment standard, was about ±0.1‰ VPDB (Isotope Laboratory Bremen University).

2.2.6 Foraminiferal analyses

For foraminiferal analyses the isolated 63-150 µm fraction was processed. Subsequently, the samples were dry sieved on a 100 µm sieve. This latter step was done in order to obtain better comparability with previous foraminiferal analyses in the Skagerrak area. Because of the relatively high contents of sand and organic material in the samples, the foraminifera of the size fraction 100 µm were concentrated using the heavy liquid carbon tetrachloride (CCl_4, $\rho = 1.59$ g/cm^3) (Meldgaard and Knudsen, 1979).

After the gravity separation, foraminifera containing pyrite end up in the heavy fraction and not among the concentrated foraminifera (Feyling-Hanssen et al., 1971). Therefore, also the residues (heavy fraction) were examined visually under the binocular microscope, and the considered samples (concentrated foraminifera) were assessed for being representative.

The foraminifera were picked from an extraction tray under the binocular microscope. The specimens were identified, counted, listed and transferred to a slide. In sparse samples the entire foraminiferal content was counted, whereas in rich ones, at least 300 benthic and planktonic foraminifera were picked. Agglutinated benthic species were extracted separately. The remaining part was examined to see if there is a considerable amount of other species than there is in the counted part. The volumes of the samples, which were used in this study, were enough to give statistically significant numbers.

For the analyses, the percentages of the most dominant species were used. Additionally, e.g. the total numbers of benthic and planktonic foraminifera per 100 g dry sediment weight as well as the benthic and planktonic fluxes were considered. The flux was determined with the following equation: $F = ni * \rho_d / t$ (F = foraminiferal flux (specimens/cm^2/year); ni = number of foraminifera per gram; ρ_d = dry specific density (g/cm^3); t = years per centimeter core interval).

3 List of publications

This thesis is based on data, interpretations and conclusions presented in five manuscripts. In the following an overview of the contents of these manuscripts (Chapter 4-8) is given.

Depositional history of the Helgoland mud area, German Bight, North Sea
(Dierk Hebbeln, Carolyn Scheurle and Frank Lamy, *Geo-Marine Letters*, 2003).

From different parts of the Helgoland mud area, situated south-east of the island of Helgoland, four sediment cores were obtained. In order to investigate whether sedimentary records of this area reflect disturbance by natural processes and/or anthropogenic activities or whether they provide an undisturbed, continuous archive, these cores were compared. Their physical properties correlate well, and are pointing to rather undisturbed paleoenvironmental conditions. However, more profound analyses of one of the cores (GeoB 4801-1) indicate a marked shift in the sedimentation pattern around 800 years ago, when the sedimentation rate dropped by a factor of ~10. Various processes leading to the previously high sedimentation rates are discussed - favouring the disintegration of Helgoland itself to be responsible. Within the period of low sedimentation, natural (e.g. storm-flood activity) as well as a variety of human impacts were traced.

Stable oxygen isotopes as recorders of salinity and river discharge in the German Bight, North Sea
(Carolyn Scheurle and Dierk Hebbeln, *Geo-Marine Letters*, 2003).

To decipher the environmental signal - temperature and/or salinity - recorded in $\delta^{18}O$ data of the German Bight, these data were evaluated and calibrated with instrumental records. For that purpose, a sea-surface temperature and a sea-surface salinity data set, measured at Helgoland and reaching back for 100 years, was used. The analyses revealed a clear dependence of the $\delta^{18}O$ signal on (summer) salinity. A comparison with river runoff indicates that the salinity is mainly driven by the freshwater input of the Elbe River. Therefore, the $\delta^{18}O$ data provide the possibility to reconstruct the salinity of the German Bight area as well as the discharge of the Elbe River beyond instrumental measuring.

An 800 year reconstruction of Elbe River discharge and German Bight sea-surface salinity
(Carolyn Scheurle, Dierk Hebbeln and Phil Jones, *The Holocene*, in press).

On the basis of the evaluation and calibration of the $\delta^{18}O$ data of the German Bight, discussed in chapter 5, the Elbe River runoff and its influence on the sea-surface salinity in the extended river mouth area were reconstructed. The $\delta^{18}O$ series covers the past 800 years. During this time span

climatic variations lead to significant changes in the $\delta^{18}O$ record reflecting changes in the salinity, the river discharge and consequently the precipitation in the Elbe catchment.

Late Holocene environmental changes in the Skagerrak, eastern North Sea - foraminiferal indication (Carolyn Scheurle, Karen Luise Knudsen, Dierk Hebbeln and Peter Kristensen, *Marine Micropaleontology*, submitted).

The faunal composition as well as the fluxes of benthic and planktonic foraminifera were analyzed from a gravity core (GeoB 6003-2) from the Skagerrak, and revealed significant changes in the bottom and the surface water masses. Together with the grain-size distribution these changes were interpreted as climatic variations in the area during the Late Holocene. In the records presented, the past 2,700 years are resolved, and subdivided into several time intervals, which are linked to historical epochs, such as e.g. the Migration Period. Furthermore, the benthic foraminiferal flux implies that phases of cooler (moderate) climatic conditions are associated with high (low) productivity. From the planktonic foraminiferal flux it appeared that the influx of North Atlantic water masses into the Skagerrak region was lowered since the end of the climatic deteriorated Migration Period (around 1200 cal yr BP).

Stable isotopes recording environmental conditions in the Skagerrak (North Sea) during the Late Holocene (Carolyn Scheurle, Karen Luise Knudsen and Dierk Hebbeln, Manuscript).

The aim of this study was to reconstruct the environmental conditions, namely temperature and/or salinity as well as the amount of nutrients, for the Skagerrak area during the Late Holocene. To achieve this goal, stable oxygen and carbon isotopes of benthic and planktonic foraminifera were analyzed, resolving the past 2,700 years on a temporal resolution of 30 to 40 years. Subsequently, the data sets were compared with other proxy data. It appears that the $\delta^{18}O$ signal reflects the temperature trends of Northern Europe. The $\delta^{13}C$ data are taken as a proxy for the availability of nutrients, which certainly affected the productivity in the area. Productivity is linked to environmental and climatic conditions such as e.g. temperature and precipitation. During relatively cool and precipitation-rich phases solubility may have been favoured, and the nutrients were reaching the North Sea system via the major rivers. It is, therefore, hypothesized that the nutrients mainly are originating from catchments of Central Europe.

4 Depositional history of the Helgoland mud area, German Bight, North Sea

Dierk Hebbeln, Carolyn Scheurle and Frank Lamy

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Geo-Marine Letters (2003), 23, 81-90

Abstract

The Helgoland mud area in the German Bight is one of the very few sediment depocenters in the North Sea. Despite the shallowness of the setting (<30 m water depth), its topmost sediments provide a continuous and high-resolution record allowing the reconstruction of regional paleoenvironmental conditions for the time since AD ~400. The record reveals a marked shift in sedimentation around AD 1250, when average sedimentation rates drop from >13 to ~1.6 mm/year. Among a number of major environmental changes in this region during the Middle Ages, the disintegration of the island of Helgoland appears to be the most likely factor which caused the very high sedimentation rates prior to AD 1250. According to historical maps, Helgoland used to be substantially bigger at around AD 800 than today. After the shift in sedimentation, a continuous and highly resolved paleoenvironmental record reflects natural events, such as regional storm-flood activity, as well as human impacts at work at local to global scales, on sedimentation in the Helgoland mud area.

4.1 Introduction

In the shallow North Sea, which is on average less than 100 m deep and characterised by high tidal- and wave-energy levels, continuous sediment redistribution is the predominating process in shaping the sedimentary environment. There are only a few depocenters where sustained sediment deposition occurs (Lohse et al., 1995). Besides the most prominent depocenter - the Skagerrak, the Helgoland mud area in the German Bight is particularly important in this context. Here, continuous sediment deposition is attributed to a small-scale eddy driven by the interaction of the longshore coastal current, the discharge from the Elbe and Weser Rivers, and tidal dynamics (Hertweck, 1983).

The Helgoland mud area extends over ~500 km^2, with mean water depths of ~20 m (Figge, 1981; Figure 4.1). In its western part, the sediments consist mainly of mud (clayey silt), whereas towards the eastern part the proportion of sand increases (Hertweck, 1983).

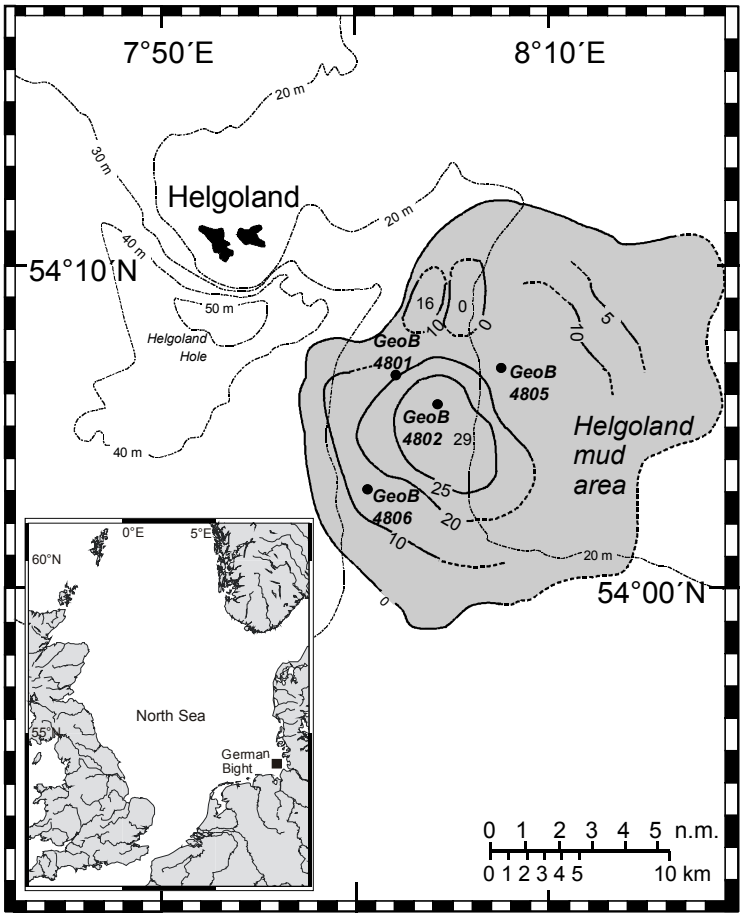

Figure 4.1: Extent and thickness (m) of the Helgoland mud area (shaded area and bold dashed lines; after von Haugwitz et al., 1988) to the southeast of the island of Helgoland in the German Bight, SE North Sea (see inset map). The closed dots mark the locations of the four investigated sediment cores. Bathymetry is indicated by the thin lines (n.m. = nautical mile).

Seismic investigations have shown that up to 30 m of Holocene sediments has accumulated in the Helgoland mud area, filling a depression otherwise similar to the still unfilled, so-called Helgoland Hole, directly south of Helgoland (von Haugwitz et al., 1988; Figure 4.1).

Estimates of sedimentation rates in the Helgoland mud area, based on sedimentological observations, vary between 2 mm/year (Reineck, 1963) and 5 mm/year (Reineck, 1969). Using ^{210}Pb dating on a sediment core from the area, Dominik et al. (1978) report an average sedimentation rate of 7.7 mm/year, with a range of 2 to 18 mm/year. According to the observations of Dominik et al. (1978), sedimentation is strongly controlled by the frequency of storm floods, the sediments being relatively undisturbed by bioturbation. For the core used by Dominik et al. (1978), Förstner and Reineck (1974) described a marked increase of trace element contents at ~120-cm core depth, indicating the onset of anthropogenic trace metal input into the region around AD 1820. The reported range of sedimentation rates most likely reflects a trend towards increasing sedimentation from west to east, indicated by a correlation of zinc (Zn) contents in a number of short cores from the Helgoland mud area (Irion et al., 1987).

In order to investigate whether the sediment record in the Helgoland mud area reflects disturbance by natural processes (e.g. storms) and/or anthropogenic activities (e.g. fisheries) or whether it provides an undisturbed, continuous archive of the depositional history in the German Bight, a set of four sediment cores has been retrieved from different parts of the depocenter. Correlation of the cores by physical properties was followed by a detailed stratigraphic analysis of the longest and best-resolved core by ^{210}Pb and accelerator mass spectrometry (AMS) ^{14}C dating. Furthermore, this core has been investigated for grain-size distribution, zinc content and stable carbon isotope composition of the organic matter in order to allow reconstructions of the paleoenvironmental setting.

4.2 Material and methods

In 1997 a set of four gravity cores was collected in the Helgoland mud area from aboard the German RV METEOR (Table 4.1). Before core splitting, all cores were analyzed at 5-cm intervals for gamma ray attenuation, using a non-destructive GEOTEK multi-sensor core logger (MSCL; Gunn and Best, 1998). The gamma ray attenuation of sediments is dependent on the density and porosity of the sediments and it often provides a first indication of lithological changes (e.g. Weber et al., 1997). In this study it was used as a tool to correlate the sediment cores investigated.

Table 4.1: Site information for the four investigated sediment cores from the Helgoland mud area taken from aboard the German RV Meteor in 1997.

Core No.	Latitude	Longitude	Water-depth (m)	Core length (m)
GeoB 4801-1	54°06.7´N	8°02.2´E	25	11.41
GeoB 4802-1	54°05.7´N	8°04.4´E	25	8.01
GeoB 4805-1	54°06.8´N	8°07.6´E	21	7.34
GeoB 4806-1	54°03.1´N	8°00.6´E	29	7.4

More detailed analyses for paleoenvironmental reconstructions focused on the 11-m-long sediment core GeoB 4801-1. This core was analyzed for chemical composition using the Bremen University X-ray fluorescence (XRF) core scanner (e.g. Jansen et al., 1998; Röhl et al., 2000). This instrument allows a semi-quantitative determination of the contents of major and minor elements by scanning split sediment cores. From the data obtained, only the Zn contents, measured at 1-cm resolution, are presented in this study.

For further analyses on core GeoB 4801-1, subsamples were taken at 5-cm intervals throughout the core. In the upper 135 cm, the sample resolution was increased to 1-cm intervals. The subsamples were washed through 63 and 150 µm sieves to provide information about grain-size distribution, and to separate out the coarse fraction from which Foraminifera shells were selected for ^{14}C analyses (see below).

The determination of the stable carbon isotope composition of organic matter ($\delta^{13}C_{org}$) was carried out on homogenized sample material which was decalcified with 6 N HCl. After drying the samples on a hot plate at 80°C, they were burned in a Heraeus-CHN-analyzer. The evolved CO_2 gas was transferred via an automated trapping box into a Finnigan MAT delta E mass spectrometer. These isotope measurements were done versus a working standard (Burgbrohl CO_2 gas) which was calibrated against PDB by using the NBS 18, 19 and 20 standards. Consequently, the isotope data given here are relative to the PDB standard. Precision calculated from repeated measurements on a laboratory sediment standard was about ±0.1‰ for $\delta^{13}C_{org}$ (Isotope Laboratory Bremen University).

The age model for core GeoB 4801-1 is based mainly on AMS ^{14}C and ^{210}Pb dating. Sedimentation rates for the uppermost part of the core were determined by measuring ^{210}Pb activity via its α particle-emitting grand-daughter ^{210}Po, following the method described by van Weering et al. (1987). Twelve AMS ^{14}C dates (Table 4.2) were determined on ~10 mg carbonate (shells of mixed

benthic Foraminifera or carbonate parts of macrozoobenthic organisms) at the Leibniz Laboratory for Age Determinations and Isotope Research at the University of Kiel (Nadeau et al., 1997).

All ages are corrected for ^{13}C and for the global mean reservoir age of 400 years (Bard, 1988), which has been confirmed for the nearby Danish marine waters (Heier-Nielsen et al., 1995). The ^{14}C ages were converted into calendar years using the Calib 4.3 software (Stuiver and Reimer, 1993) with the marine data set of Stuiver et al. (1998).

Table 4.2: Overview of the AMS ^{14}C dates in core GeoB 4801-1 from the Helgoland mud area (bf - mixed benthic Foraminifera).

Sampled core depth (cm)	Mid depth (cm)	Material	Lab ident	Raw ^{14}C age (years)		Reservoir-corr. ^{14}C age (years)	Calibrated ^{14}C age (cal. years BP)	Years (AD)
48	48	Mollusc	KIA6797	690	+/-25	290	311	1639
73	73	Echinoderm	KIA6796	765	+/-35	365	441	1509
93-98	96	bf	KIA8247	1040	+/-35	640	584	1366
108-118	113	bf	KIA13040	1085	+/-60	685	657	1293
128-132	130	bf	KIA13039	1170	+/-45	770	683	1267
292-303	297	bf	KIA8245	1465	+/-30	1065	961	989
353-358	355	bf	KIA8244	1280	+/-30	880	772	1178
398-403	400	bf	KIA8242	1655	+/-25	1255	1216	734
713	713	bf	KIA8241	1650	+/-35	1250	1216	734
823	823	bf	KIA8240	1575	+/-30	1175	1087	863
948	948	bf	KIA8239	2055	+/-40	1655	1545	405
948	948	bf	KIA8237	1660	+/-50	1260	1215	735

4.3 Results

The smoothed 9-point running averages of the gamma ray attenuation display continuous cyclic variations in the range ±1000 cps, which seem to be easy to correlate from core to core (Figure 4.2), although the uppermost part of the sequence seems to be missing in core GeoB 4805-1. These data indicate that core Geob 4801-1 contains the longest stratigraphic record with the highest sedimentation rate and, thus, the highest temporal resolution. Consequently, this core was selected for further detailed analyses.

Stratigraphic analyses on core GeoB 4801-1 focused on ^{210}Pb and AMS ^{14}C dating. The ^{210}Pb measurements reveal a continuous decrease of $^{210}Pb_{(tot)}$ from ~40 Bq/kg at the core top to a background value of 21.55 Bq/kg at ~30-cm core depth. From the decrease in $^{210}Pb_{(excess)}$ (Figure 4.3) and assuming no vertical mixing, a maximum sedimentation rate of 2.6 mm/year was modelled. Indeed, the ^{210}Pb data from the very top of the core indicate that no mixing occurs. In

addition, the correlation of the sediment data with a boxcore from the same site clearly shows that the gravity core GeoB 4801-1 contained the complete sediment sequence up to the sediment-water interface (Hebbeln, unpublished data). Therefore, the assumption that the core is not affected by mixing seems valid, and a sedimentation rate of 2.6 mm/year for the uppermost part of the core is supported.

In the upper 130 cm of the core, five AMS ^{14}C ages line up continuously back to an age of ~750 years, with an average sedimentation rate of 1.6 mm/year (Figure 4.3). Further downcore the AMS ^{14}C ages do not reveal a continuous sequence, and even age reversals (i.e. older ages following younger ones) occur. Although the data do not allow a detailed age determination, the sequence shows a general trend to younger dates towards the top of the core. Compared to the upper part of the core, calculating an average sedimentation rate for this lower core section reveals a much higher sedimentation rate of 13 to 20 mm/year (Figure 4.3).

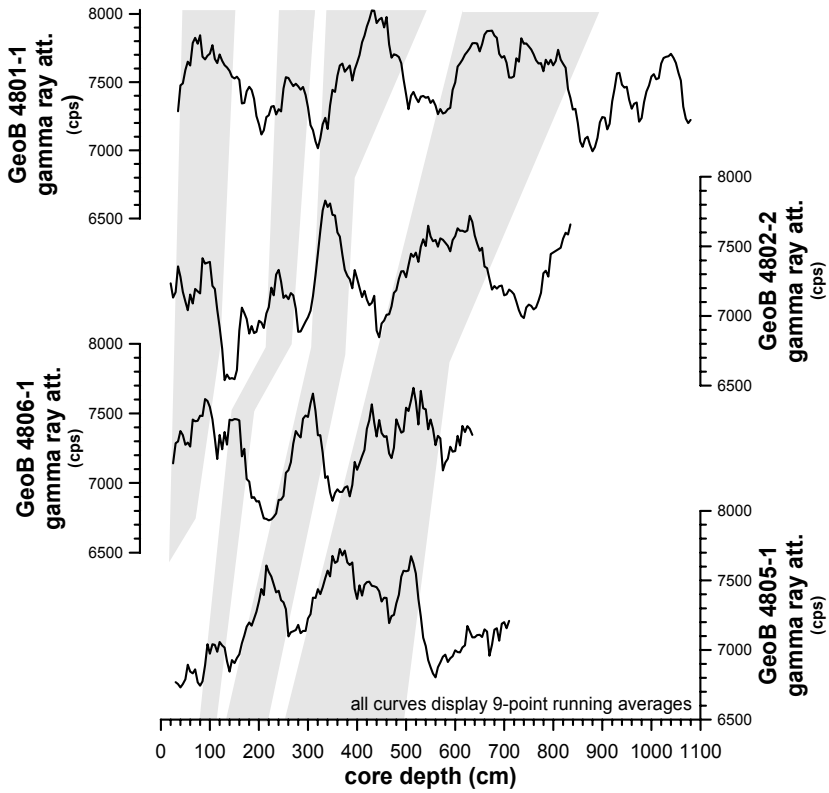

Figure 4.2: Correlation of the four sediment cores from the Helgoland mud area by gamma ray attenuation obtained with the GEOTEK multi-sensor core logger (MSCL). The shaded areas connect core sections inferred to be of the same age.

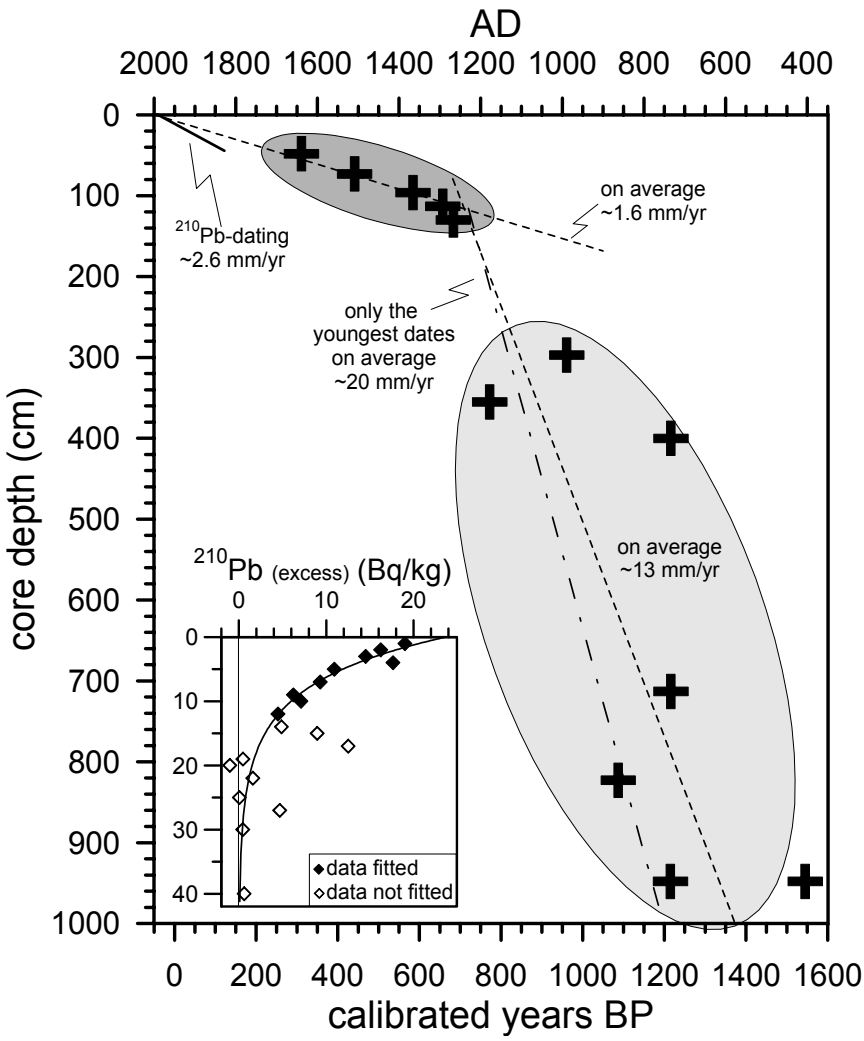

Figure 4.3: Stratigraphic framework for core GeoB 4801-1 from the Helgoland mud area. Closed crosses mark individual AMS ^{14}C dates, with ages given in years AD (top axis) and calibrated years BP (bottom axis) versus core depth. The errors of the dates, given in Table 4.2, are smaller than the width of the crosses. Light shading marks the high-sedimentation period in the lower part of the core, with an average sedimentation rate of 13 mm/year. Focusing on the youngest ages in this core section (dash-point line), the sedimentation rate may be as high as 20 mm/year. Dark shading marks the low-sedimentation period at the top of the core, with an average sedimentation rate of 1.6 mm/year. The inset shows the amount of $^{210}Pb_{(excess)}$ for the upper 40 cm of the core. Only the uppermost samples (closed diamonds) have been used for modelling the sedimentation rate of 2.6 mm/year.

Additional data obtained on core GeoB 4801-1 include coarse-fraction contents, $\delta^{13}C_{org}$ values and Zn contents (Figure 4.4). The fraction >150 µm rarely exceeds 2 wt% throughout the core. Only at the base, below 10-m core depth, the content of the >150 µm fraction is substantially higher, reaching 60 wt%. This fraction is enriched also at the very top of the core, contributing up to

8 wt%. A similar pattern is revealed by the 63-150 µm fraction which has on average markedly higher percentages. However, some values are not as high as for the >150 µm fraction, especially those at the base of the core (<25 wt%). Above 10-m core depth, core GeoB 4801-1 can be divided into three main parts. Between 7- and 10-m core depth, the coarse-fraction (63-150 µm) levels are rather low (5 wt% on average, albeit with some higher peaks), whereas the Zn contents are comparably high over this interval and down to the base of the core (15-30 cps, Figure 4.4). Between 4 and 7 m, coarse-fraction percentages (5 wt% on average) and Zn contents (8-15 cps) are low and, from 1.4 to 4 m, there is another increase in the coarse fraction (to around 10 wt%), without any change in the Zn contents. Similarly to the >150 µm fraction, the 63-150 µm fraction also shows a remarkable increase (up to 45 wt%) in the very upper part of the core.

The $\delta^{13}C_{org}$ data vary between -25 and –24‰ PDB throughout the core, without showing any marked variability (Figure 4.4). Only within the upper 20 cm of the core do the values increase sharply to >-23‰ PDB. A comparable signal is found in the Zn content (Figure 4.4) which increases from around 10 to 40 cps close to the core top.

4.4 Discussion

4.4.1 Correlation of the cores

Arguably, the first question when dealing with sediments from a highly dynamic, shallow-water area (~20 m water depth) is to which extent the available sediment record provides a coherent and undisturbed record of past environmental conditions. To answer this question, four cores from different parts of the Helgoland mud area were compared using their gamma ray attenuation data. The similarity in pattern revealed by these data in all four cores indicates that the cores contain the same sediment sequence (Figure 4.2), most likely representing the depositional history of the Helgoland mud area over a relatively long time span. Thus, by supporting earlier results from Irion et al. (1987), it appears that the sediments of the Helgoland mud area can confidently be interpreted as representing a coherent sediment record rather unaffected by local disturbances induced by, for example, storms and fisheries.

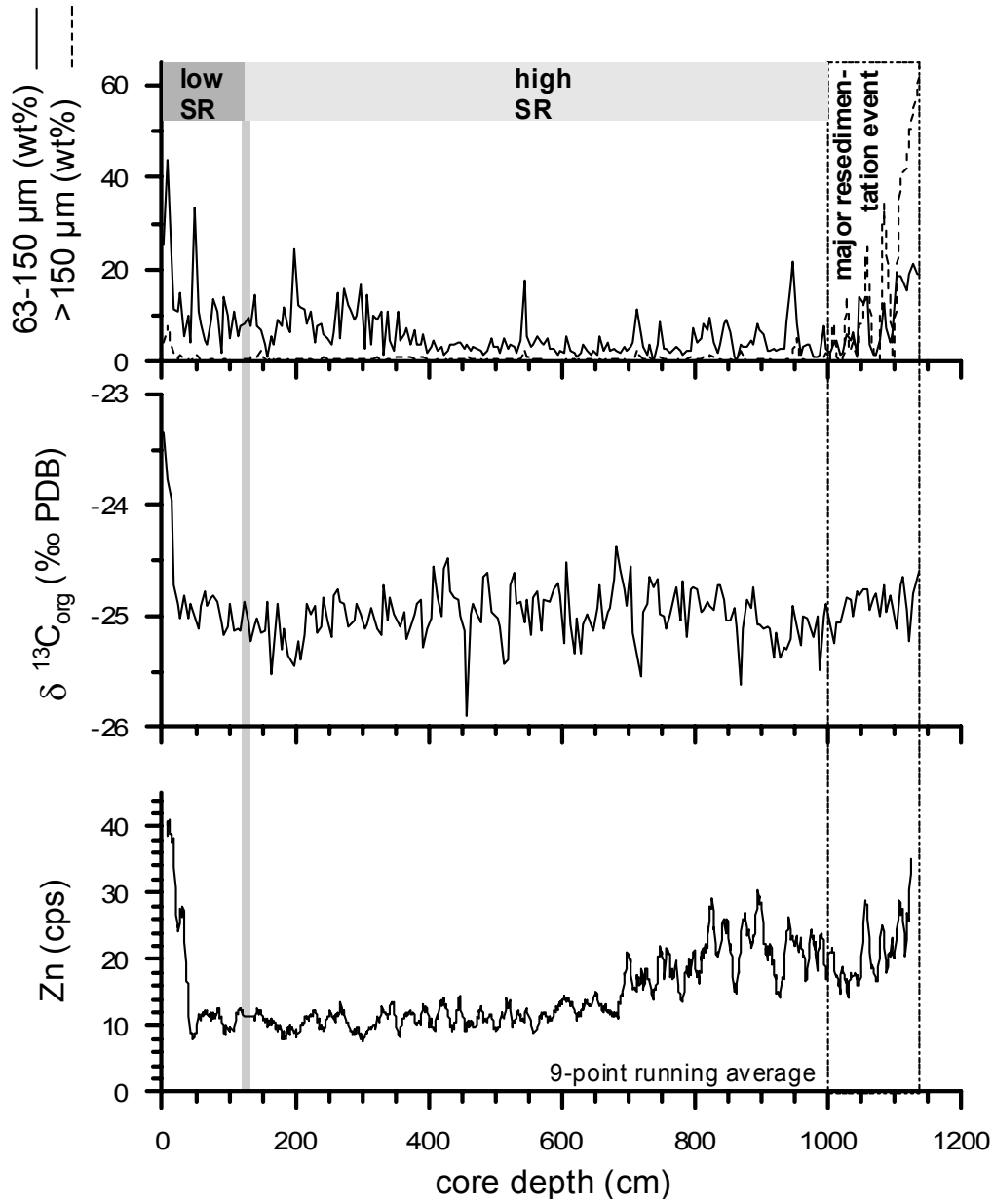

Figure 4.4: Downcore records of the coarse-fraction (63-150 µm and >150 µm) percentages, $\delta^{13}C_{org}$ values and Zn contents in core GeoB 4801-1 from the Helgoland mud area. The low and high sedimentation rate (SR) parts of the core are indicated according to Figure 4.3. The upward-fining sequence in the bottom of the core, indicated in the coarse-fraction data, reflects a major resedimentation event.

4.4.2 Age model

In order to investigate the depositional history of the Helgoland mud area in detail, the stratigraphically longest and temporally best-resolved core (GeoB 4801-1) was selected for further analyses. According to the AMS ^{14}C dates, the core can be divided into two parts: a high-sedimentation lower part, and a low-sedimentation upper part (Figure 4.3).

The uneven sequence of the seven AMS ^{14}C dates in the lower part indicates substantial sediment redistribution, with the deposition of reworked, older material and contemporary sediments at the same time. As the Helgoland mud area acts as a regional sediment trap, nothing can be said here about the source area of the reworked material. According to the AMS dates, it is inferred that the section between 3- and 9.5-m core depth was deposited between 750 and 1300 calendar years BP. Following the main trend of these dates and the trend described by the youngest dates (which are presumably least affected by reworking) reveals average sedimentation rates of ~13 and ~20 mm/year, respectively. More precise age assignments are difficult due to the obvious impact of reworked material. Nevertheless, it can be concluded that, after 1300 cal. years BP, rapid sedimentation resulted in the deposition of more than 8 m of sediment within ~500 to 600 years.

In the upper part of the core (between 0.5 and 1.3 m), by contrast, five AMS ^{14}C dates line up well, indicating an average sedimentation rate of 1.6 mm/year (Figures 4.3 and 4.5) – only about 10% of what is found further downcore. Actually, at the base of this core section the AMS ^{14}C dates indicate a major decrease in the sedimentation rate (Figure 4.5). Most likely this represents the final shift from the very high sedimentation rates downcore to the lower rates in the upper part of the core. The lining up of the AMS ^{14}C dates points to a rather continuous sedimentation without significant admixtures of reworked material. For this part of the core, ages between dated levels have been assigned by linear interpolation between the nearest AMS ^{14}C dates.

The ^{210}Pb-derived sedimentation rate of 2.6 mm/year for the uppermost part of the core is somewhat higher than the sedimentation rate indicated by the AMS ^{14}C dates slightly further below. Thus, an increase in sedimentation rate is inferred for the recent past. In terms of establishing an age model, the problem arises how to link the uppermost AMS ^{14}C date with the ^{210}Pb dating, as the first would imply a sedimentation rate of 1.4 mm/year for the uppermost core section, whereas the latter gives a value of 2.6 mm/year (Figure 4.5).

Additional stratigraphic information for such high-resolution records spanning the last couple of centuries can be drawn from the $\delta^{13}C_{org}$ data. Due to the burning of isotopically light fossil fuels since the beginning of the industrial revolution (roughly 150 years ago), the carbon isotopic composition of the atmosphere has shifted towards lighter values, in pace with the increasing CO_2 content in the atmosphere (the so-called Suess effect; Keeling, 1979). The ocean-atmosphere gas exchange transferred this signal into the ocean, where it entered the food web through the photosynthetic activity of marine phytoplankton and, eventually, the sedimentary record (e.g. Fontugne and Calvert, 1992).

Comparing the $\delta^{13}C_{org}$ record of core GeoB 4801-1 with the atmospheric CO_2 content (Mann et al., 2000) clearly shows how the core data mirror the historical increase of CO_2 in the atmosphere (Figure 4.5). For the German Bight sediments, these $\delta^{13}C_{org}$ data reveal a Suess effect of magnitude

1.7‰ PDB. By fitting the $\delta^{13}C_{org}$ data to the CO_2 record, an additional stratigraphical tie point is obtained which allows to link the ^{210}Pb and the AMS ^{14}C data (Figure 4.5). Thus, the final age model for the upper part of core GeoB 4801-1 is based on linear interpolation between the individual AMS ^{14}C dates and the $\delta^{13}C_{org}$-derived tie point (38-cm core depth, AD 1938) as well as the ^{210}Pb-derived sedimentation rate. From this it follows that the AMS ^{14}C date indicating the year AD 1267 at 130-cm core depth can be taken as the onset of the low-sedimentation period.

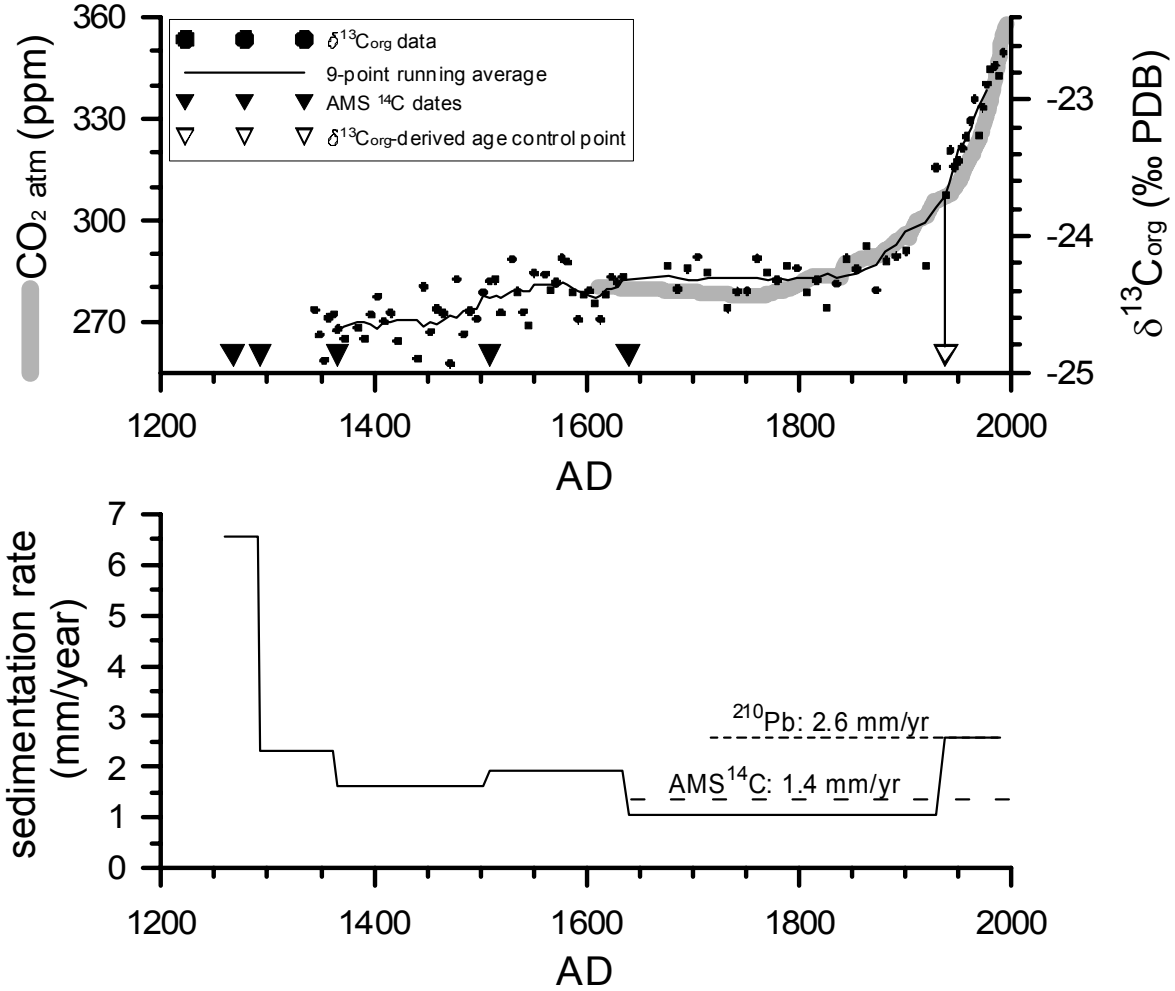

Figure 4.5: Final age model for the upper part of core GeoB 4801-1 from the Helgoland mud area. Data points result from AMS ^{14}C dating (closed triangles), a ^{210}Pb-derived sedimentation rate of 2.6 mm/year at the very top of the core, and an additional age control point (open triangle) resulting from the fitting of the $\delta^{13}C_{org}$ record to the atmospheric CO_2 content (see text). CO_2 data are taken from Mann et al. (2000). The resulting sedimentation rates decrease from >6 to <3 mm/year over the last 800 years.

4.4.3 A major resedimentation event at the beginning of the record (~1300 years BP)

The lowermost part of core GeoB 4801-1 (>10-m core depth) consists of the deposits of a major resedimentation event. Thus, the grain-size data reveal a ~1.5 m thick, upward-fining sequence at the base of the core, where the >150 µm fraction contributes >60 wt% to the total sediment (Figure 4.4). One can only speculate about the nature of the event having caused this pattern. Possibly it can be attributed to a huge storm flood, during which the wave base reached the seafloor and reworked the sediments. However, this must have been an extraordinary storm flood, as nothing even slightly comparable has happened since then (~1300 cal. years BP to present). Although there are some more upward-fining sequences further up in the core (e.g. at 9.5, 7.1 and 5.4 m), they are of very different dimensions - thicknesses of 10 to 20 cm and >150 µm fraction contents never exceeding 5 wt%.

4.4.4 The high sedimentation period at AD ~700 to ~1250

This part of the record (below 1.4-m core depth), marked by the deposition of >8 m of sediment within less than 600 years, does not provide very much information. Interpreting the grain-size data in terms of hydrodynamic energy levels at the seafloor, there seems to have been a considerable increase in energy in the upper part of this core section (above 4-m core depth) which occurred sometime before AD 1250 (Figure 4.4).

Zn contents in marine sediments are generally very low and do not provide a significant environmental signal, except for regions with either anthropogenic heavy metal input and/or extremely high terrigenous sediment contributions. Close to Helgoland, there can have been an impact from copper mining and smelting on the island, which may have lasted from the Bronze Age to medieval times (Schulz, 1981). However, the absence of any relation between the Zn and the Cu records, as well as rather low Cu contents (~20 cps; Hebbeln, unpublished data) largely exclude such an impact.

Thus, over the time span considered here (AD ~400-1200), the Zn content is most likely a proxy for the source area of the sediment. At the time corresponding to 7-m core depth, there was probably a change between two, more remote source areas (Figure 4.4). Unfortunately, relevant information about Zn contents in possible source areas is lacking. By contrast, the increase of the Zn content at the core top is not a shift back to the former source area, but rather represents a well-known anthropogenic signal related to the beginning of industrialization. The rather low variability of the $\delta^{13}C_{org}$ data around -24.5‰ PDB between AD ~400-1200 points to the deposition of a mixture of marine (~-20‰ PDB) and terrigenous (~-28‰ PDB) organic matter (Emerson and Hedges, 1988).

Of course, the main question regarding this part of the core addresses the cause of an extremely high sedimentation rate in the order of 13 to 20 mm/year. Thus, we are looking for (1) a process explaining the high sedimentation rates between AD 700 and 1250, and (2) the reason for the shift towards lower sedimentation rates around AD 1250. Interestingly, the German Bight region and its hinterland experienced a number of changes and events in or close to that time interval which may correspond to the depositional history of the Helgoland mud area. Among these are (1) increased storm-flood activity, (2) onset of dike building, (3) increased soil erosion in the hinterland, (4) sea-level changes, and (5) the partial disintegration of the island of Helgoland.

Finally, the observed change in sedimentation rate may also reflect the end of the filling-up process of the former depression which now forms the host for the Helgoland mud deposits. Although this possibility cannot be excluded, the seismic data provided by von Haugwitz et al. (1988) indicate that the mud lens to the SE of Helgoland nowadays already forms a positive seabed structure, and that the shift from filling a negative to building a positive seabed feature should be deeper than the depth of the shift in sedimentation in core GeoB 4801-1, which is only 1.4 m below the present sediment surface. Thus, the reason for this shift has to be sought among those potential factors listed above.

As mentioned above, the grain-size record as a proxy for the storm-flood activity indicates that, with the exception of the event reflected at the very base of the core, floods did not directly affect the seafloor in a marked way. However, a number of historical storm floods in the Middle Ages caused huge land losses in the German Bight. Most severe were the 14th century events which caused the opening of the East Frisian Jade Bay and flooded most of the North Frisian marshland (Behre, 2002). At that time, huge amounts of sediment were mobilized and transported from the coastal area offshore. However, the 14th century storm floods are too recent to be related to the rapidly accumulated section of the core. This also shows that even the large volumes of sediments mobilized during extensive land-loss periods, such as those occurring in the 14th century (90- to 110-m core depth), did not always leave an obvious imprint in the sedimentary record of the Helgoland mud area. Thus, any storm-flood events causing the observed high sedimentation rates must have been of extraordinary dimensions. However, although the German North Sea coast was inhabited during most of this period (Behre, 2002), there are no historic documents indicating such severe events. Thus, the high sedimentation rates in core GeoB 4801-1 before AD 1250 are probably not related to storm-flood activity.

In the 11th century the people inhabiting the lowlands along the coast of the German Bight began building dikes. At first there were only small summer dikes to protect the cattle but as early as AD 1300 there was already a rather continuous dike line around East Frisia (Behre, 2002). These dikes must have had an impact on sediment exchange between the coast and the open sea.

However, this effect would probably have been rather small compared to, for example, the huge flood-induced land losses described above. Most importantly, dike building commenced only after the sedimentation rates started to decrease. This argues against land reclamation being a major factor causing the high sedimentation rates in the Helgoland mud area.

Significant human impact on the environment was not restricted to the coastal area during these times. Extensive deforestation since the 6th century led to increased soil erosion in central Europe (Bork et al., 1998) and, consequently, to higher suspension loads in rivers. In the first half of the 14th century, extreme rainfall events caused soil erosion to increase by almost two orders of magnitude, whereby half of the total hill-slope erosion recorded since AD 650 took place during the period AD 1310-1350 (Lang et al., 2000). A large part of the material eroded during this time interval was probably delivered via rivers to the North Sea. The exceptional size of this erosion event may well have resulted in a huge sediment input to the coastal sea. Nevertheless, also this event occurred too late to be associated with the high-sedimentation signal found in the Helgoland mud area.

Investigations on historic salt marsh deposits outcropping on the seaward side of the East Frisian island of Juist point to a sudden sea-level change (rise and fall) of up to one meter during the 14th century (Freund and Streif, 2000). Such sea-level fluctuations are known to result in the mobilization of huge amounts of sediments, which could have contributed significantly to accumulation also in the Helgoland mud area. Again, however, this event occurred too late to explain the high sedimentation rates between AD 700 and 1250.

Interestingly, all of the processes discussed above, each of which could be assumed to have a significant impact on the sedimentation pattern in the German Bight, left no obvious traces in the regional sedimentary record, as indicated by the smooth sedimentation after AD 1250. Thus, the processes which resulted in the high sedimentation rates prior to AD 1250 must have been much more effective in shaping the environment than those prevailing afterwards.

We contend that the most likely candidate having caused the high sedimentation rates in the Helgoland mud area is the island of Helgoland itself. In historical times the island underwent a major reduction in size. A historical map comparing the size of Helgoland in the years AD 800, 1300 and 1649 shows a land loss of about 90% during this time interval (Figure 4.6). Although this old map, drawn up in the 17th century, i.e. 900 years after the time period in question, should be interpreted with caution, it is clear that Helgoland was much bigger around AD 800. This is substantiated by the numerous churches indicated on the land subsequently lost. Thus, it can be concluded that Helgoland experienced a major reduction in size between AD 800 and 1300, without having any information about the time before AD 800. The mobilization of this huge amount of sediment so close to the Helgoland mud area probably resulted in the deposition of at

least a substantial part of this material at the depocenter. However, ongoing disintegration of the island after AD 1300 has left no significant traces in the sediment record. Thus, the remarkable shift in sedimentation at AD ~1250 may indicate that the destruction of the island passed a threshold, at which the disintegration slowed down strongly and/or at which material provided by this process was deposited elsewhere.

Figure 4.6: Historical map (Domeier and Haack, 1963) showing the island of Helgoland around AD 800 (largest extension), around 1300 (light shading) and at 1649 (dark shading), with the last stage being very close in size to the present-day island of Helgoland. Thus, this map indicates a major reduction in size of the island after AD 800.

4.4.5 The low sedimentation period from AD ~1250 to the present

For the last 750 years, the Helgoland mud area has been characterized by continuous and rather stable sedimentation, with a rate of ~1.6 mm/year (Figs. 4.3 and 4.5). Although this rate is an order of magnitude lower than the one recorded further downcore, this part of the data set, accompanied by a detailed age model, provides an exceptional, highly resolved record for paleoenvironmental reconstructions. During this period, the coarse-fraction levels are comparably high (up to 30 wt%), and probably reflect the storm-flood activity in the region (Dominik et al.,

1978). A comparison of the coarse-fraction data with a historical record of storm-flood activity in the German Bight, compiled from numerous sources (Scheurle, unpublished data), reveals some degree of interrelationship. Although the correlation is weak (Figure 4.7), it seems that particularly periods with less storm-flood activity (e.g. AD 1500 to 1600, around 1770, and 1860 to 1900) are marked by relatively low coarse-fraction percentages, supporting a relationship between grain-size composition and storm-flood activity.

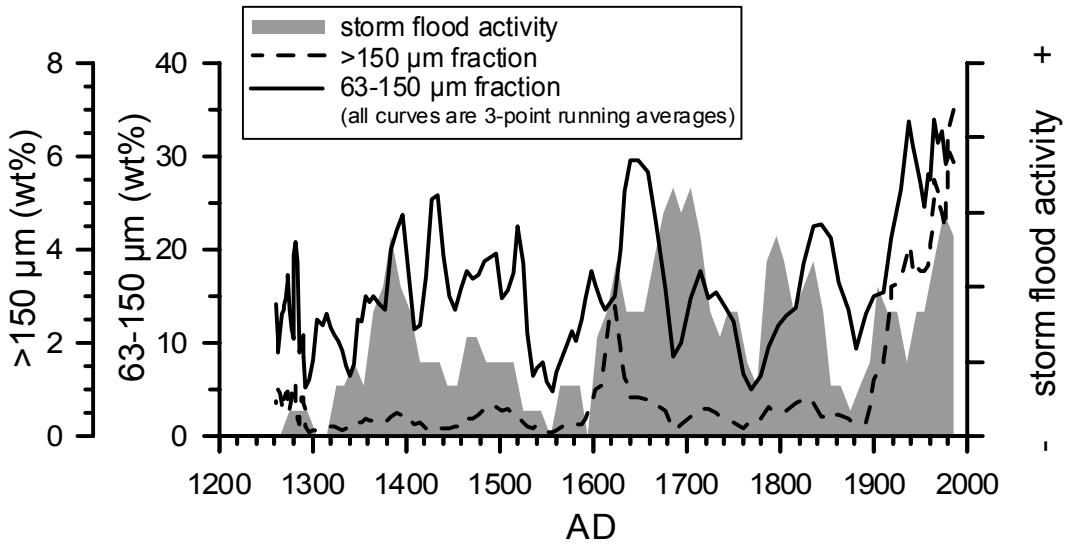

Figure 4.7: Coarse-fraction percentages in core GeoB 4801-1 from the Helgoland mud area for the last 800 years. Storm-flood activity is based on summing-up storm-flood events on a decadal scale, the events being compiled from numerous sources based on historical records (Scheurle, unpublished data). As definitions for storm floods vary widely, here the storm-flood activity is described only relatively by higher (+) or lower (-).

In the most recent part of the record, a distinct human impact is recognized. Whereas the increasing $\delta^{13}C_{org}$ values recorded since AD ~1850 reflect a global signal conveyed via the atmosphere into the ocean and the marine record (see above), increasing Zn contents and most likely also increasing coarse-fraction percentages are due rather to regional/local human influences (Figure 4.8). Thus, the Zn content increases in two steps. The first one occurred around AD ~1750 and may be related to the very early onset of industrialization and its effects on silver and zinc mining in the Harz mountains (Bartels, 1996). The Harz mountains as well as the Erzgebirge are well-known mining areas, exploited for several centuries, and both lie along the tributaries of the Elbe and Weser rivers. Even before, during medieval times, mining activities in the Harz mountains resulted in elevated Zn contents in river clays of the Weser River (Ortlam, 1989). The second increase in Zn content occurred at AD ~1900 and probably reflects increasing environmental pollution in the

course of an accelerating industrialization. Based on this observation, it is inferred that the increase in Zn content occurred well before the second half of the 20th century, to which it was confined by Irion et al. (1987).

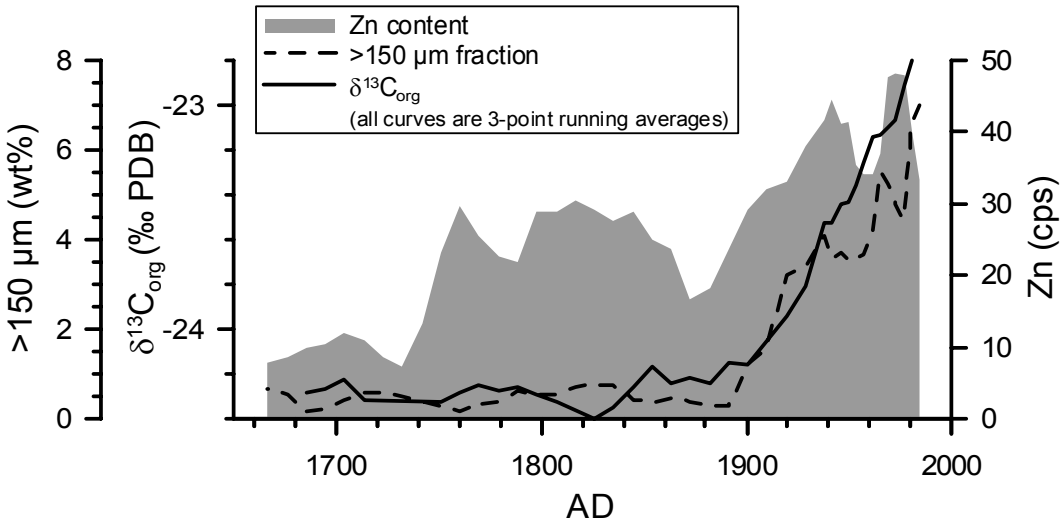

Figure 4.8: Anthropogenic impact indicators in sediment core GeoB 4801-1 from the Helgoland mud area (see text).

Of interest is also the marked increase in coarse-fraction percentages since AD ~1900 (Figure 4.7), most obvious for the >150 µm fraction (Figure 4.8). The increasing sedimentation rates over this period exclude a simple winnowing effect which would enrich the coarse fraction. As the pattern parallels the human impact markers $\delta^{13}C_{org}$ and Zn, and as the storm-flood activity during this period is well in the range of that recorded for the preceding centuries (Figure 4.7), this increase in coarse-fraction percentages is also most likely related to human activities.

Since 1885 the German fishing fleet in the North Sea has made extensive use of steam vessels, resulting in the onset of intensive beam-trawl fishing activity along the German coast (I. Heidbrink, personal communication). This new way of fishing probably contributed significantly to sediment redistribution in the area and, thus, may have increased overall sedimentation in the Helgoland mud area and the coarseness of the sediments deposited there. In addition, since the end of the 19th century systematic deepening and maintenance work in the Elbe and Weser River shipping channels (Flügel, 1987; Wetzel, 1987) probably brought substantial amounts of sediment into the water column which, to some extent, may have reached the Helgoland mud area.

This investigation shows that also shallow marine sediments from highly dynamic settings can provide undisturbed sedimentary records allowing the reconstruction of the depositional history at a given site. In addition, the high sedimentation rates usually found in such shallow waters can allow reconstructions of the paleoenvironment with utmost temporal resolution. For example, a better than decadal resolution, as obtained from core GeoB 4801-1, can be used to calibrate marine proxy data to instrumental climate records. Such calibrations open up a new dimension for the interpretation of marine proxy data and, together with the high temporal resolution, allow reconstructions of the paleoenvironment with formerly unknown detail.

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5 Stable oxygen isotopes as recorders of salinity and river discharge in the German Bight, North Sea

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Abstract

A high-resolution sediment core (sedimentation rate ~2 mm/year) from the German Bight was analysed for its foraminiferal stable oxygen isotope ($\delta^{18}O$) composition. These data were correlated with instrumental summer sea-surface temperature and salinity data from the nearby island of Helgoland, reaching back 100 years. Comparing the isotope data with the instrumental records reveals a distinct $\delta^{18}O$-salinity relationship ($\delta^{18}O = 0.34*S-9.36$; $r = 0.86$) for the German Bight, where salinity is mainly driven by freshwater input from the Elbe River. Thus, these findings provide the possibility for future regional paleosalinity and paleodischarge reconstructions for times far beyond the instrumental timescale.

5.1 Introduction

Instrumental time series of climate and environmental data are only available for the last decades or for a couple of centuries at best. As a consequence most assessments of paleoclimatic and paleoenvironmental conditions in terrestrial and marine systems are based on proxy relationships. Paleoceanographic proxies are measurable descriptors for desired (but unobservable in the past) environmental variables such as temperature, salinity, nutrient contents and many others (Fischer and Wefer, 1999). To reconstruct paleoenvironmental parameters from proxy data, the proxies need to be calibrated. This is done mostly by statistical comparison between a set of surface sediment data assumed to reflect present-day conditions and the desired environmental parameter taken from instrumental measurements at corresponding sites (Wefer et al., 1999). Probably the best-known example for this approach is the reconstruction of global sea-surface temperatures for the Last Glacial Maximum (CLIMAP-Project Members, 1976; 1981).

However, relationships between proxy and environmental parameters may be biased due to a number of reasons: (1) without stratigraphic control, the basic assumption that the surface sediments reflect the actual setting can be erroneous; (2) mixing through bioturbation can spread the environmental signal over several 100 years; (3) proxy relationships can vary on regional scales and may not be coherent on the spatial scale of the surface sediment set used for the calibration.

Proxy relationships are further complicated by the fact that most proxies do not respond to a single environmental parameter only, but to two or even more. The prime example for this is the most widely used proxy in paleoceanography, namely stable oxygen isotopes ($\delta^{18}O$) measured on calcareous shells. As pointed out by Emiliani (1955), and many others since, this proxy is influenced by changes in temperature and seawater chemistry, with the $\delta^{18}O$ composition of seawater being closely related to salinity (Craig and Gordon, 1965).

Under exceptional circumstances an alternative approach can be taken to calibrate paleoceanographic proxies, called on-site calibration. Such calibrations require very high sedimentation rates in the order of >10 cm/100 years, as well as the availability of instrumental records reaching back for several decades (e.g. Malmgren et al., 2001). Such conditions are found in the case of the German Bight, North Sea (Figure 5.1). In the so-called Helgoland mud area, sediments are continuously deposited at rates in the order of >10 cm/100 years (Hebbeln et al., 2003). In addition, instrumental records of sea-surface temperature (SST) and sea-surface salinity (SSS) are available from the island of Helgoland, reaching back for more than 100 years.

In this study we present a direct correlation of proxy data taken from the sediments (i.e. foraminiferal $\delta^{18}O$) and instrumental data (SST, SSS). Based on good stratigraphic control, this

approach overcomes most of the problems mentioned above and provides a reasonably reliable proxy calibration for the German Bight.

5.2 Regional setting

The hydrography of the North Sea is driven mainly by the influx of temperate and highly saline North Atlantic waters between Scotland and Norway as well as through the English Channel. The water masses are locally modified by river runoff and by the interaction of North Sea and Baltic Sea waters in the Skagerrak. This general pattern is well reflected in the salinity distribution of the North Sea, which clearly shows the Atlantic inflow as well as the low-salinity plumes close to the two main rivers, the Rhine and the Elbe (Schott, 1966; Becker, 1981; Figure 5.1).

The annual mean temperatures of North Sea waters are rather uniform, ranging between 9.5 and 11.5°C; temperatures of >10°C being closely related to Atlantic waters advected through the English Channel (Becker, 1981). In shallow parts of the German Bight, the seasonal temperature range can exceed 16°C, which is the largest for the entire North Sea (Sündermann et al., 1999).

Whereas the annual mean temperatures in the German Bight do not show any effect of the rivers, the steep gradient of SSS demonstrates the sensitivity of this region to the discharge of the Elbe River, and to a minor extent also of the Weser River (Figure 5.1). Annual means of SST and SSS taken at Helgoland Reede since 1900 vary between 8.4 and 11°C and between 30.8 and 33.2 psu, respectively, whereas average summer values (June-July) of SST and SSS range between 12.2 and 15.2°C and between 29 and 33.3 psu (Figure 5.2).

In the region south-east of the island of Helgoland, which is influenced by the Elbe Estuary, a local sediment depocenter is located, the so-called Helgoland mud area (von Haugwitz et al., 1988). Forced by a small-scale eddy which is induced by the interaction of North Sea waters, river discharge and tidal dynamics, mainly suspended material is deposited (Hertweck, 1983). With sedimentation rates of 1-2.6 mm/year over the last 100 years (Hebbeln et al., 2003), the sediments from the Helgoland mud area provide a high-resolution record suitable for comparison of sedimentary proxy data with the instrumental time series from Helgoland Reede.

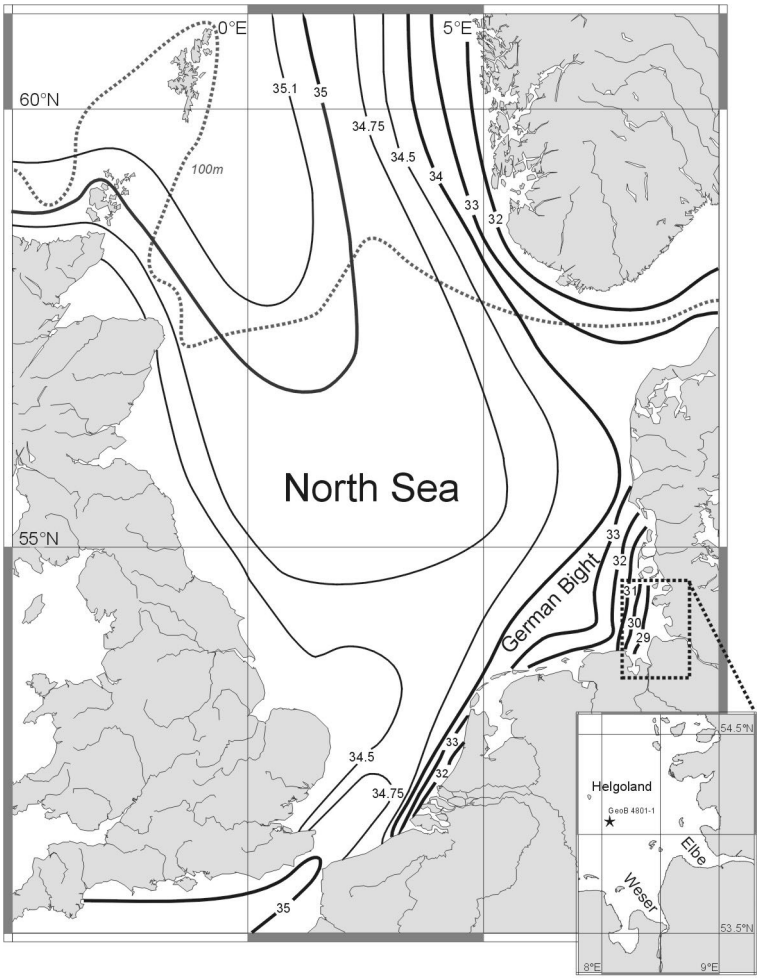

Figure 5.1: Annual mean sea-surface salinity in the North Sea (after Schott, 1966). The inset shows the inner German Bight with the site of core GeoB 4801-1 near the island of Helgoland close to the Elbe and Weser Estuaries.

5.3 Material and methods

This study focuses on sediment core GeoB 4801-1 taken at 54°06.7´N and 8°02.2´E in 25 m water depth during cruise M40-0 of the German RV METEOR in 1997 (Figure 5.1). The uppermost 20 cm of the core were examined and subsampled at 1-cm intervals. The samples were washed over a 150-µm sieve to separate the coarse fraction for Foraminifera analyses.

The stable oxygen isotope composition of shells of the benthic foraminifer *Elphidium excavatum, forma selseyensis* was analysed with a Finnigan MAT 251 mass spectrometer. Ten to twenty individual shells were picked for each measurement, all of the specimens being perfectly preserved without showing any sign of transport or reworking. The isotopic composition of the carbonate sample was measured on the CO_2 gas evolved by treatment with phosphoric acid at a constant

temperature of 75°C. For all stable oxygen isotope measurements, a working standard (Burgbrohl CO_2 gas) was used which was calibrated against Pee Dee Belemnite (PDB) by applying the NBS 18, 19 and 20 standards. Consequently, all $\delta^{18}O$ data given here are relative to the PDB standard. The analytical standard deviation is ±0.07‰ PDB (Isotope Laboratory Bremen University).

Furthermore, SST and SSS data collected at Helgoland Reede since about 1900 were provided by the German Federal Maritime and Hydrographic Agency. For the same period, Elbe River runoff data from the gauge in Neu Darchau were made available by the Local Waterways and Shipping Office, Lauenburg. These runoff data sum up hydrological influences of the catchment area (149,000 km^2; Liedtke and Marcinek, 1994) beyond the region affected by tidal processes. From all these data sets the annual mean as well as mean June-July values were used in this study. With respect to the discharge data, it is to emphasise that the calculations are based on the calendar year (not the hydrological year, Nov-Oct).

5.4 Results

According to a detailed stratigraphic analysis based on ^{210}Pb dating, and supported by ^{14}C AMS dating, the $\delta^{18}O$ record of the uppermost 20 cm of sediment core GeoB 4801-1 covers the last 100 years (Hebbeln et al., 2003). With a temporal resolution of ~4 years per sample back to 1938, and ~9 years per sample before that date, the $\delta^{18}O$ values vary between 0.9 and 1.7‰ PDB (Figure 5.2). The resolution therefore appears to be high enough to compare the $\delta^{18}O$ data with the available instrumental SST and SSS time-series data.

The instrumental time-series data collected at Helgoland Reede cover the period 1908 to 1995. However, both datasets (SST and SSS) have major gaps from 1944 to 1960 resulting from a lack of measurements after World War II. In addition, the SSS data set has another gap between 1918 and 1928 as a consequence of World War I.

The correlation of sea-surface hydrographic data and stable oxygen isotope data measured on benthic foraminifers may not appear straightforward at first sight. However, seeing that core GeoB 4801-1 was taken in a water depth of only 25 m in an area characterized by strong tidal dynamics, a well-mixed water column can be expected (Dyer, 1979). In such circumstances, the $\delta^{18}O$ data from the benthic Foraminifera *E. excavatum f. selseyensis* should correspond to the surface conditions reflected in the available hydrographic data.

5.5 Discussion

5.5.1 Proxy evaluation

As pointed out above, the $\delta^{18}O$ data from foraminiferal calcite depend on the temperature and isotopic composition of the seawater in which the foraminifers lived (Wefer et al., 1999), the latter being closely related to salinity (see Wolff et al., 1999 for a detailed discussion). In order to decipher the environmental signal - temperature and/or salinity - recorded in the $\delta^{18}O$ data from the German Bight, these data were compared with instrumental data obtained at Helgoland.

There is little information about the season when *E. excavatum f. selseyensis* calcifies, i.e. during which part of the year environmental signals such as SST and SSS are recorded in the carbonate shells of the foraminifer. However, it is assumed that *E. excavatum f. selseyensis* finds best living conditions after the spring bloom, in early summer, when food supply is highest. Thus, we assume that the main calcification period of *E. excavatum f. selseyensis* is in June and July and, consequently, we compare the $\delta^{18}O$ data (measured on *E. excavatum f. selseyensis*) with the averaged June-July SST and SSS data for the corresponding years.

In this context it must be kept in mind that the sediment data were smoothed, each data point representing 4 to 9 years - in a strict stratigraphic sense. In addition, effects of bioturbation could have had a substantial impact on the sediment depth vs. age pattern. Typical mixing depths caused by bioturbation are of the order of 10 cm or more, resulting in smoothing over several decades. If this were the case, it would impair any direct comparison between proxy and instrumental data. Fortunately, the ^{210}Pb data of core GeoB 4801-1 indicate that the uppermost part of this core is rather unaffected by bioturbation (Hebbeln et al., 2003).

In most paleoceanographic reconstructions, $\delta^{18}O$ data are used as proxies for ocean temperatures (see overview in Niebler et al., 1999). The temperature-$\delta^{18}O$ relationship is rather constant, with a change of ~4.3°C in seawater temperature corresponding to a shift of 1‰ $\delta^{18}O$ in foraminiferal calcite (Epstein et al., 1953). Thus, a quantitative interpretation of the $\delta^{18}O$ data in terms of temperature is quite straightforward, as long as any major influences of salinity can be confidently ruled out. This is mostly the case for open ocean settings, where temperature variations are usually much more pronounced than fluctuations in salinity. In near-coastal areas, by contrast, where freshwater inflow plays a major role in controlling physico-chemical conditions, the temperature signal in the $\delta^{18}O$ record can be suppressed by a strong salinity signal, as may be expected in the case of the German Bight which is characterised by a distinct salinity gradient (Figure 5.1).

Indeed, the comparison of the $\delta^{18}O$ data with the June-July SST data does not reveal a clearly correlated trend (Figure 5.2a). Only in the better resolved, younger part of the $\delta^{18}O$ record do the

data show some form of a reversed correlation between lower $\delta^{18}O$ values (i.e. higher temperatures) and lower SSTs. The SST therefore does not play a measurable role in controlling the $\delta^{18}O$ signal in the German Bight. By contrast, the $\delta^{18}O$ data match very well with the instrumental June-July SSS data. Particularly in the well-resolved, uppermost part of the core, the correspondence between the salinity and $\delta^{18}O$ data is striking, especially when considering the smoothed salinity record (Figure 5.2b).

The strong salinity gradient in the German Bight (Figure 5.1) points to a close relationship between salinity and river discharge. For the summer period the salinity data correspond closely to the discharge rates of the Elbe River (Figures 5.2b and c, Figure 5.3b), but discharge levels seem to have no direct impact on water temperature in the German Bight (Figures 5.2a and c, Figure 5.3a). In order to identify any longer than seasonal effect of river discharge on these hydrographic parameters in the German Bight, summer SST and SSS are compared with the annual mean discharges of the Elbe River (Figures 5.3c and d). These findings clearly show that also on longer timescales off Helgoland the Elbe River discharge affects only SSS but not SST.

The dependence of salinity in the German Bight on river discharge is also indicated in earlier studies (Radach and Berg, 1986; Heyen and Dippner, 1998). Heyen and Dippner (1998) show that salinity variability in the German Bight is in-phase and correlated with a lag of several months to large-scale air pressure on an annual timescale, both measures being linked by advective precipitation. These authors also found that the relationship between air pressure and salinity cannot be explained by the North Atlantic Oscillation.

5.5.2 Proxy calibration

The data clearly indicate the $\delta^{18}O$ record of core GeoB 4801-1 to be a valuable proxy for June-July sea-surface salinity in the inner German Bight which, in turn, is mainly driven by freshwater inflow via the Elbe River. However, the application of this proxy for a reconstruction of paleosalinities in the German Bight requires a calibration.

In contrast to the rather stable temperature-$\delta^{18}O$ relationship mentioned above, the salinity-$\delta^{18}O$ relation shows strong regional variations (Duplessy et al., 1991; Fairbanks et al., 1992). In addition, this is only an indirect relationship because the foraminiferal $\delta^{18}O$ ($\delta^{18}O_c$) depends on the $\delta^{18}O$ composition of the seawater ($\delta^{18}O_w$) the foraminifers live in.

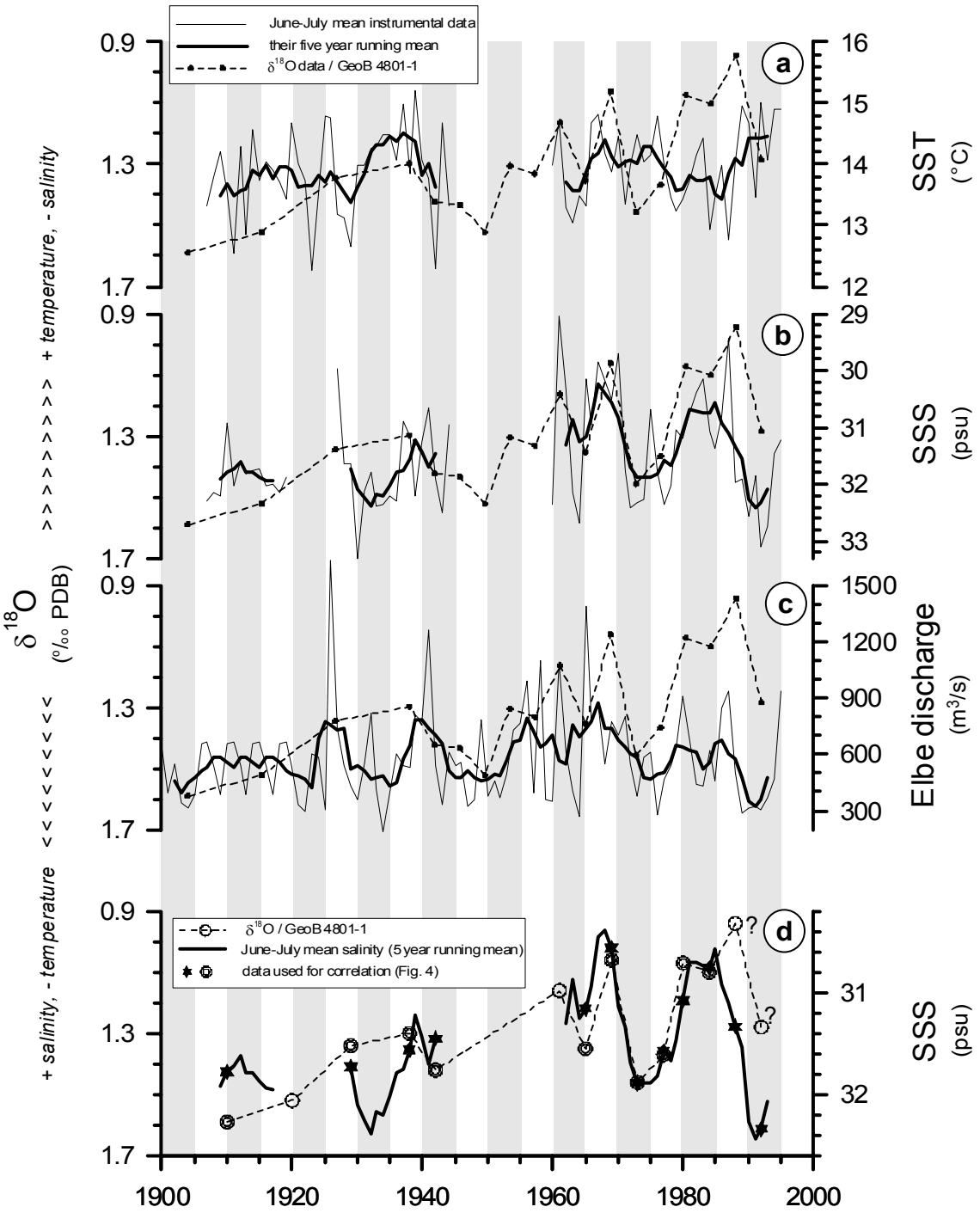

Figure 5.2: Instrumental summer time-series data for (a) sea-surface temperature (SST) at Helgoland Reede, (b) sea-surface salinity (SSS) at Helgoland Reede, and (c) the Elbe River discharge. Thin lines: mean June-July data for the respective years; thick lines: 5-year running means. All data are accompanied by $\delta^{18}O$ values measured on the benthic Foraminifera E. excavatum f. selseyensis in core GeoB 4801-1 from the Helgoland mud area (dashed line). (d) Correlation of $\delta^{18}O$ data and the smoothed mean June-July sea-surface salinity (SSS) record, as in (b). Data used for the correlation shown in Figure 5.4 are marked by stars in the salinity record and by double circles in the $\delta^{18}O$ record (see text for explanation).

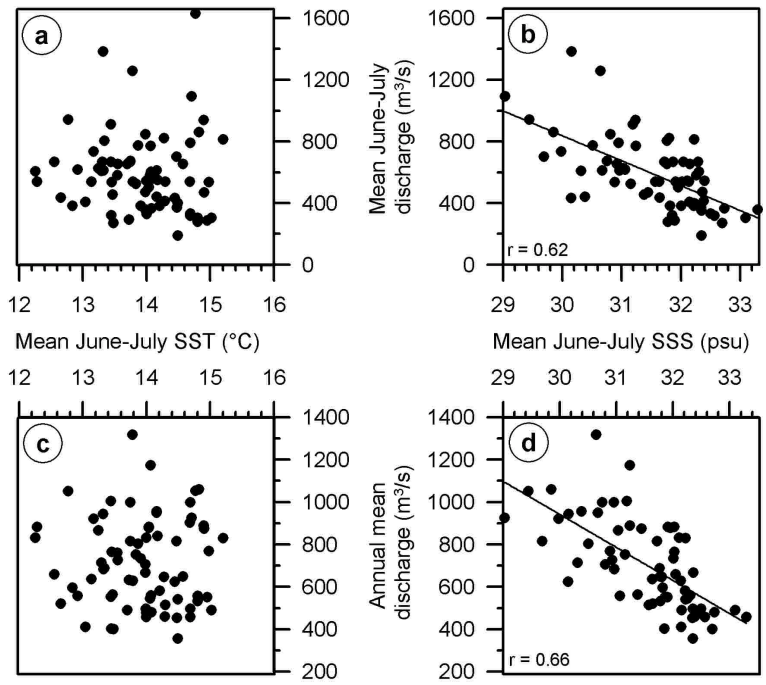

Figure 5.3: Direct comparison of instrumental data. June-July means of (a) Elbe River discharge vs. sea-surface temperature (SST) and (b) Elbe River discharge vs. sea-surface salinity (SSS; r = 0.62). Annual mean Elbe River discharge is compared with the mean June-July SST (c) and mean June-July SSS (r = 0.66) (d).

However, there is a close relationship between salinity and $\delta^{18}O_w$, as both are controlled by the evaporation/precipitation regime (Epstein and Mayeda, 1953; Craig and Gordon, 1965). This relationship becomes further complicated when water vapour is transported over long distances, because of ongoing fractionation and subsequent precipitation far from its source area, resulting in quite variable regional salinity-$\delta^{18}O_w$ relationships. However, in a regional context, the salinity-$\delta^{18}O_w$ relationship, and thus the salinity-$\delta^{18}O_c$ relationship, is assumed to remain rather constant in the long-term annual mean (Wolff et al., 1999).

Our data allow an empirical calibration between salinity and $\delta^{18}O_c$ when assuming a constant salinity-$\delta^{18}O_w$ relationship over the respective timescales. Following the basic approach on a temporal rather than the more common spatial basis, we used the $\delta^{18}O$ data of core GeoB 4801-1 and the instrumental salinity data for this calibration. In order to align the temporal resolution of both data sets, we compared the $\delta^{18}O$ data with a resolution of 4 to 9 years to the 5-year running mean of the annually resolved mean June-July salinity data (Figure 5.2d).

In the next step we correlated the $\delta^{18}O$ values assigned to a specific year by the stratigraphic framework with the smoothed mean June-July salinity data for that year (Figures 5.2d and 5.4a). Both data sets appear to be highly correlated (r = 0.86), with the exception of the uppermost two

data points. These two samples are far off the calibration and are hence assumed to have been affected either by bioturbation near the sediment surface resulting in a non-equilibrium state or by core-top disturbance in the course of coring. For this reason, they were excluded from the calibration. The resulting equation ($\delta^{18}O = 0.34 * S - 9.36$) indicates a slope of 0.34 for the $\delta^{18}O_w$-salinity relationship and a $\delta^{18}O$-value of the freshwater end member (salinity = 0 psu) of ~-9.4‰.

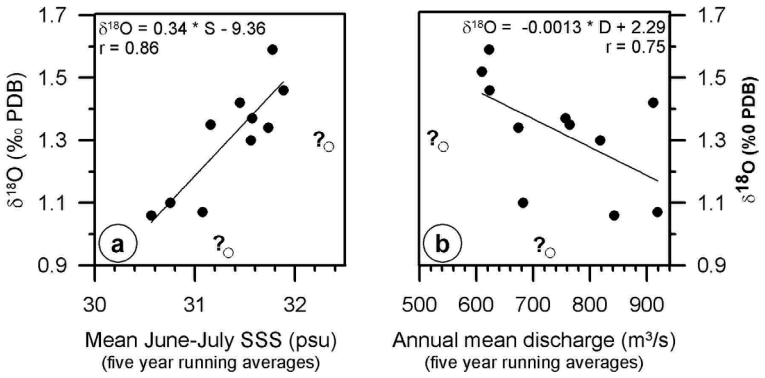

Figure 5.4: Comparison of instrumental and proxy data. (a) Smoothed mean June-July sea-surface salinity (SSS) vs. foraminiferal $\delta^{18}O$ data of core GeoB 4801-1 (compare with Figure 5.2d) excluding the two topmost samples of core GeoB 4801-1 (open circles with question-mark) as discussed in the text. The regression line of $\delta^{18}O$ vs. SSS is $\delta^{18}O = 0.34 * S - 9.36$ with $r = 0.86$. (b) Annual mean Elbe River discharges vs. foraminiferal $\delta^{18}O$ data of core GeoB 4801-1, again excluding the two topmost samples (see above). The regression line of $\delta^{18}O$ vs. discharge is $\delta^{18}O = -0.0013 * D + 2.29$ with $r = 0.75$.

On a global scale, the slope of the $\delta^{18}O_w$-salinity relationship varies between ~1 in the polar regions (Vetshteyn et al., 1974; Fairbanks et al., 1992) and much lower values of ~0.2 and ~0.1 in the western and eastern equatorial Atlantic, respectively (Fairbanks et al., 1992). For the North Sea, Israelson and Buchardt (1991) described a relationship of 4.25‰ SSS per 1‰ $\delta^{18}O$ (slope of 0.24) by referring to unpublished data without, however, giving any further details about the data. Although $\delta^{18}O_w$ measurements for the German Bight are not available, we conclude on the basis of our calibration that the slope for the German Bight is 0.34 (Figure 5.4a), thus perfectly matching the value of 0.33 provided by Mook (2001) for the Scheldt Estuary in the southernmost North Sea.

Also the end member of about -9.4‰ fits to measurements that revealed mean $\delta^{18}O$ values of ~-9.5‰ for Elbe waters (5 year range: -8.5 to -10.5‰; Trettin et al., 1999), the supposed freshwater end member for the German Bight. In addition, these data are in agreement with oxygen isotope ratios of precipitation within the Elbe catchment in Central Europe (-7 to –10‰; Hirschfeld, 2003). Actually, to obtain a reasonable oxygen isotope ratio for the freshwater end member, which for the

Elbe is set to ~-9.5‰ by the precipitation regime, the slope of the $\delta^{18}O$–salinity relation is prescribed to ~0.35, as becomes also obvious for the Scheldt Estuary which is supplied by the same precipitation regime. The correspondence between our proxy relationship and these regional settings strongly supports the assumption of a June-July calcification period for *E. excavatum f. selseyensis* as, for instance, a comparison of the measured $\delta^{18}O$ data with the annual mean salinities yields an unrealistic end member of ~-18‰.

5.5.3 Proxy discussion

The proxy calibration presented above relates the $\delta^{18}O$ signal solely to variations in salinity, although it is well known that there is also an influence of temperature (Epstein et al., 1953). However, in the multi-year calibration data set from the German Bight, temperature does not have any measurable effect on the $\delta^{18}O$ data. The poor temperature to $\delta^{18}O$ relationship, which is even partly inverse to the expected one, may be due to the stable annual mean temperature distribution in the North Sea (Becker, 1981), which is also reflected in relatively low interannual variability of the summer SSTs in the German Bight (Figure 5.2a). This does not mean that there is no temperature effect at all in the $\delta^{18}O$ data. However, if present, it is entirely suppressed by a salinity effect, probably induced by large salinity variations which are also reflected in the particularly steep salinity gradient of the German Bight (Figure 5.1).

In an empirical proxy calibration, as performed here, only those environmental parameters can be used which leave a quantifiable imprint on the selected proxy. The good agreement between the $\delta^{18}O$ record and the salinity data proves that $\delta^{18}O$ can be used as a proxy for salinity in this area. Of course, as long as it has not been substantiated for other parts of the North Sea, the present calibration is valid only for the German Bight with its specific conditions (Radach and Bohle-Carbonell, 1990). In this area, however, it can be used to reconstruct (summer) salinities back in time beyond the instrumental record. It has to be kept in mind, however, that the calibration is based on smoothed data (5-year running mean of annual mean values), having been adjusted to the resolution which can best be extracted from these sediments. Thus, all reconstructed salinities would reflect only June-July means averaged over ~5 years. Due to the close relationship between salinity and river discharge, the $\delta^{18}O$ data can also serve as a kind of semiquantitative proxy for the paleodischarge of the Elbe River.

5.5.4 Proxy application

For the reconstruction of the functioning of past ecosystems, summer paleosalinities are probably better suited than annual means as the summer season is characterised by strongest biological activity. However, for applications for which annual mean paleosalinities are needed, the close correlation between mean June-July salinities and annual mean salinities (Figure 5.5a) indicates that the reconstructed paleosalinities also reflect the pattern of long-term development of annual paleosalinities. The same holds true for the relation between mean summer and annual mean discharge rates of the Elbe River (Figure 5.5b).

For the assessment of paleodischarges as indicators for paleoprecipitation and thus paleoclimate, preferably annual mean data are to be used. Fortunately, the June-July salinity data, reflected in the foraminiferal $\delta^{18}O$ values, correspond as well to the annual mean Elbe discharges as to the mean June-July discharges (Figures 5.3b and d). Thus, by applying the resulting $\delta^{18}O$ to annual mean discharge relation (Figure 5.4b), annual paleodischarges can also be crudely estimated. A similar approach using $\delta^{18}O$ data to reconstruct paleosalinities and paleodischarges may be applicable to other estuarine areas of the North Sea, and elsewhere.

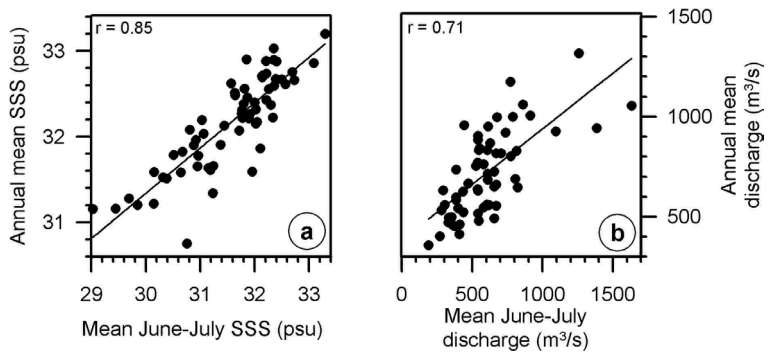

Figure 5.5: Comparison of instrumental data. Annual means vs. mean June-July (a) sea-surface salinities (SSS; r = 0.85) and (b) Elbe River discharge rates (r = 0.71).

5.6 Conclusions

This example from the German Bight demonstrates how an on-site calibration can be established between foraminifer isotopic signatures and regional physico-chemical conditions. Here, the paleoceanographic proxy $\delta^{18}O$ is closely related to instrumental summer salinity data over the last 100 years. Thus, the established correlation allows a reconstruction not only of summer paleosalinity in this region but also of the summer and even annual paleodischarge of the

Elbe River back in time far beyond the instrumental record. The calibration also demonstrates that, even under high sedimentation rates, the reconstruction of environmental data resolves the mean of several years only, a constraint often ignored in proxy calibrations.

Compared to the commonly used calibration of paleoceanographic proxies in the space domain, on-site calibration in the time domain offers a number of advantages. Besides the stratigraphic control, which provides information on the age and temporal resolution of a sample, necessary for a valid correlation to instrumental data, this kind of correlation also takes site-specific conditions into account which are probably smoothed out on a broader spatial scale. In addition, the coherence of the calibration on the given spatial scale is commonly simply assumed, but rarely shown.

Such on-site calibrations, however, are meaningful on local to regional scales only. Furthermore, they can be applied only where reasonably high sedimentation rates occur, and correspondingly long instrumental records are available. On-site calibration is therefore not always an alternative to the common spatial calibration. Nevertheless, where applicable, it can provide a distinct relationship which takes local conditions into account and, importantly, the temporal resolution which can best be extracted from the sediments, with the resulting limitation of the resolution of the reconstructed environmental parameter.

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6 An 800 year reconstruction of Elbe River discharge and German Bight sea-surface salinity

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Abstract

On the basis of stable oxygen isotopes ($\delta^{18}O$), the summer sea-surface salinity of the German Bight southeastern North Sea, was determined for the past 800 years. In this near-coastal area, salinity is mainly dependent on the freshwater input of the Elbe River discharging its large catchment, which covers an area of 149.000 km^2 of Central Europe. Therefore, a proxy for Elbe River discharge was reconstructed at the same time, and consequently the $\delta^{18}O$ record is also mirroring variations in precipitation within the entire drainage basin. Significant variations in these paleoenvironmental variables are linked to climatic changes.

Keywords: stable oxygen isotopes ($\delta^{18}O$), salinity, Elbe River discharge, precipitation, North Sea, Central Europe

6.1 Introduction

Assessing natural climate variability requires climatic records extending far beyond the period of instrumental measurements. On glacial-interglacial timescales, such extended climate records are mostly based on analyses of proxy parameters (Wefer et al., 1999). Proxies are measurable indicators for environmental variables unobservable or unobserved in the past. These sources are also used for the last millennium, but in Europe, climate reconstructions for the more recent, pre-instrumental period extensively use documentary/historical observations (Flohn and Fantechi, 1984; Lamb, 1984).

Documentary sources can provide an accurate determination of a specific parameter, but some values can be problematic. For instance, it may be difficult to determine temperature and/or precipitation conditions from chronicles documenting other variables such as crop yields. Moreover, ubiquitous uncertainties are associated with the interpretation of historic data (Brázdil, 1992a; Munzar, 1992; Glaser, 2001). Examples are (1) transcription errors, (2) misdating of the same event, causing duplicate reporting, (3) documenting of extremes only, (4) spatial allocation and extent, and last but not least (5) personal biases of the chronicler. Despite all these problems, documentary observations are a valuable and important data source for the pre-instrumental period, when used and developed with care.

Prior to instrumental measuring, the possible inaccuracy of documentary data as well as an increasing scarcity of documents makes natural proxy data useful. Such proxy records have the potential to reach much further back in time than any historic records. On the basis of careful calibrations, these paleoclimatic reconstructions also get credit for a relatively constant quality. However, making use of (marine) proxy records for high-resolution reconstructions, involving radiocarbon-dated (marine) datasets (as applied in the present study), generally perform significantly worse compared to the dating control of most documentary data sources.

In this study, the paleoceanographic proxy of stable oxygen isotopes ($\delta^{18}O$), obtained from a sediment core in the German Bight, southeastern North Sea, is used. In general, the stable oxygen isotope composition recorded in calcareous shells of marine organisms is mainly determined by the temperature and the isotopic composition of the water mass in which they calcify. As seawater's isotopic composition is closely related to salinity, $\delta^{18}O$ data are often used as proxies for both, temperature and/or salinity (Wefer et al., 1999). In this case, the evaluation and calibration of the proxy $\delta^{18}O$ revealed a dominant dependence on (summer) salinity, while any temperature effect could be neglected (Scheurle and Hebbeln, 2003). Since the location of the study area is close to the mouth of the Elbe Estuary, variations in sea-surface salinity (SSS) are primarily linked to Elbe discharges. Therefore, the proxy data presented here, allow a reconstruction of paleosalinities in the

German Bight and, thus, of Elbe River paleodischarges. As discharge is coupled with precipitation in the drainage basin, these hydrologic variables give insight about the climate of Central Europe during the past 800 years. It is well-accepted that the precipitation patterns within this drainage basin and Central Europe, respectively, respond to natural as well as to anthropogenic-forced climatic variations (e.g. Bradley et al., 1987; Brázdil, 1992b).

From a previous evaluation and calibration of the $\delta^{18}O$ data with (instrumental) SSS and river discharge measurements (Scheurle and Hebbeln, 2003), the reconstruction also allows an estimation of the range of these paleoenvironmental parameters during the past 800 years. Unfortunately, no instrumental time series of hydrologic measures from the entire Elbe catchment covering a prolonged period are available. Therefore, assessing the paleoprecipitation a reconstruction can only be achieved qualitatively.

Paleoclimate data about the region of interest is sparse, with only two series of relevance. The first is a precipitation dataset for Prague (since AD 1200) reconstructed by using historical, instrumental and proxy data (Brázdil, 1996) pointing to the importance of regional effects, like e.g. rain-shadow effects. The second is a precipitation dataset reconstructed for Central Europe (AD 1500-1997; Glaser, 2001), which is based on combined historical and instrumental data revealing a common influence of the atmospheric circulation system on this and our dataset.

The purpose of this study, however, was to follow another approach using exclusively (natural) proxy data. On the one hand, we obtained a record for comparison with already existing datasets, and were able to distinguish regional or methodological effects. On the other hand, considering paleoprecipitation, it was possible to extend the time frame (300 years), since the past 800 years could be resolved in the present study - emphasizing the Elbe catchment area and Central Europe, respectively. And for the first time, a high-resolution reconstruction of the paleoenvironmental parameters German Bight SSS and Elbe river discharge was achieved.

6.2 Regional setting

The spring of the Elbe River is situated in the Giant Mountains, northwestern part of the Czech Republic. A distance of 1165 km has to be covered until the river discharges into the German Bight, North Sea. The Czech Republic and Germany mainly share a catchment area of 149.000 km^2 (Liedtke and Marcinek, 1994; German Federal Institute of Hydrology in Koblenz; Figure 6.1). Low mountain ranges form large parts of the drainage basin.

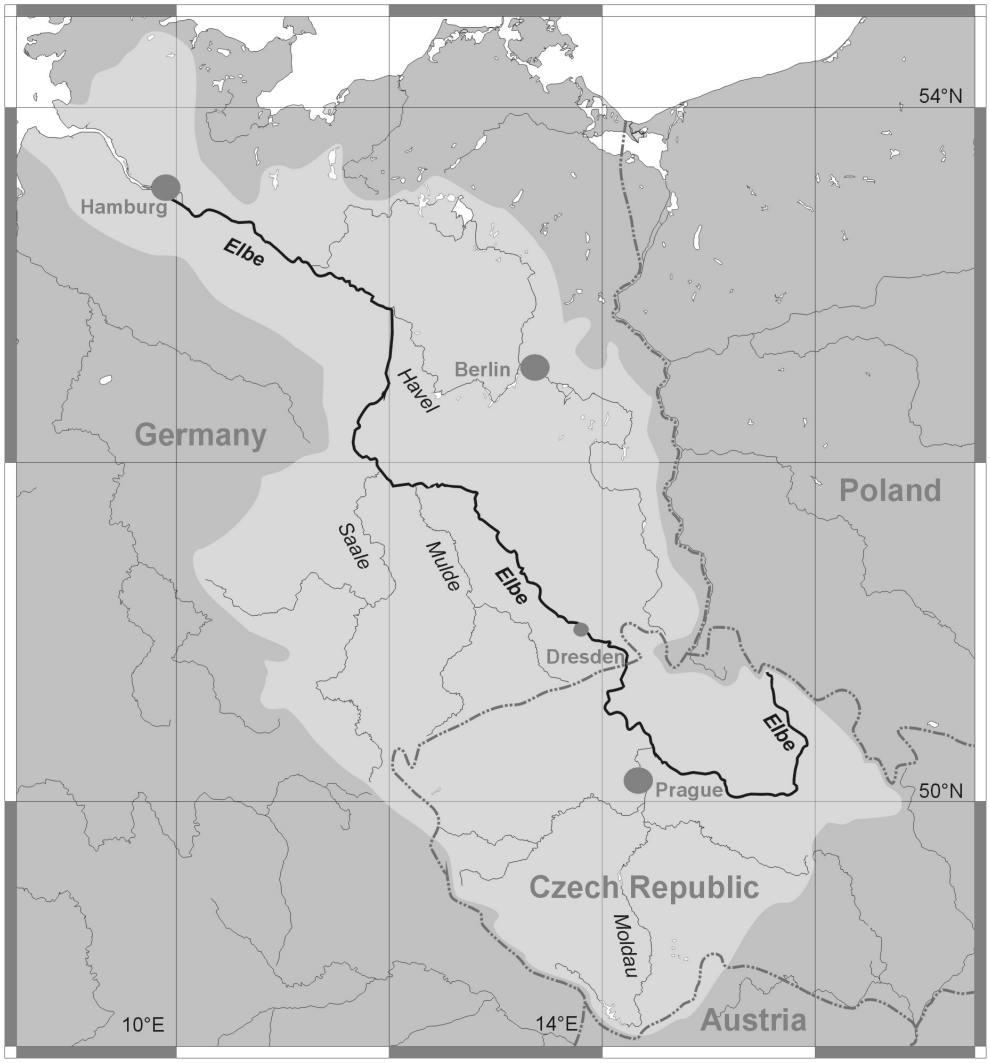

Figure 6.1: The Elbe catchment covers an area of 149.000 km^2 (shaded light grey) of which one third belongs to the Czech Republic and two thirds to Germany. The borders are indicated by the dotted lines and the course of the Elbe River by the bold black line. Additionally, the names of the major tributaries of the Elbe River are given.

The mountainous areas of the Bohemian Forest and Giant Mountains experience high precipitation. Precipitation in Central Europe is mainly influenced by cyclonic activity, carried by the dominant westerlies. Total annual precipitation is, however, primarily influenced by elevation with rain-shadow effects and coastal influences only being of local-regional importance. Precipitation is distributed throughout the year, with a summer maximum in parts of the upper Elbe catchment. Although, snowfall also constitutes a significant fraction of winter precipitation, this is not immediately available to produce runoff. Therefore, the Elbe River regime is regarded as pluvio-nival to pluvial. Whereas high discharge rates occur in wintertime, usually reaching the North Sea in the months January to March/April, in summer less runoff is recorded.

The freshwater inflow from the Elbe River, and to a minor extent from the Weser River, considerably influences the salinity of the German Bight, as indicated by a near-coastal low-salinity plume (Figure 6.2). Such a close relationship between river discharge and salinity is also described in earlier studies (Taylor et al. 1983, Radach and Berg, 1986; Heyen and Dippner, 1998). The relatively shallow German Bight (mean depth of 22.5 m) is characterized by strong tidal and wave dynamics, resulting in mixing of salt water and river-supplied freshwater. Thus, in the German Bight, the annual variability of sea-surface salinity amounts to 1-2‰ (Schott, 1966), attributable to meteorological forcing and river discharge (Becker et al., 1999).

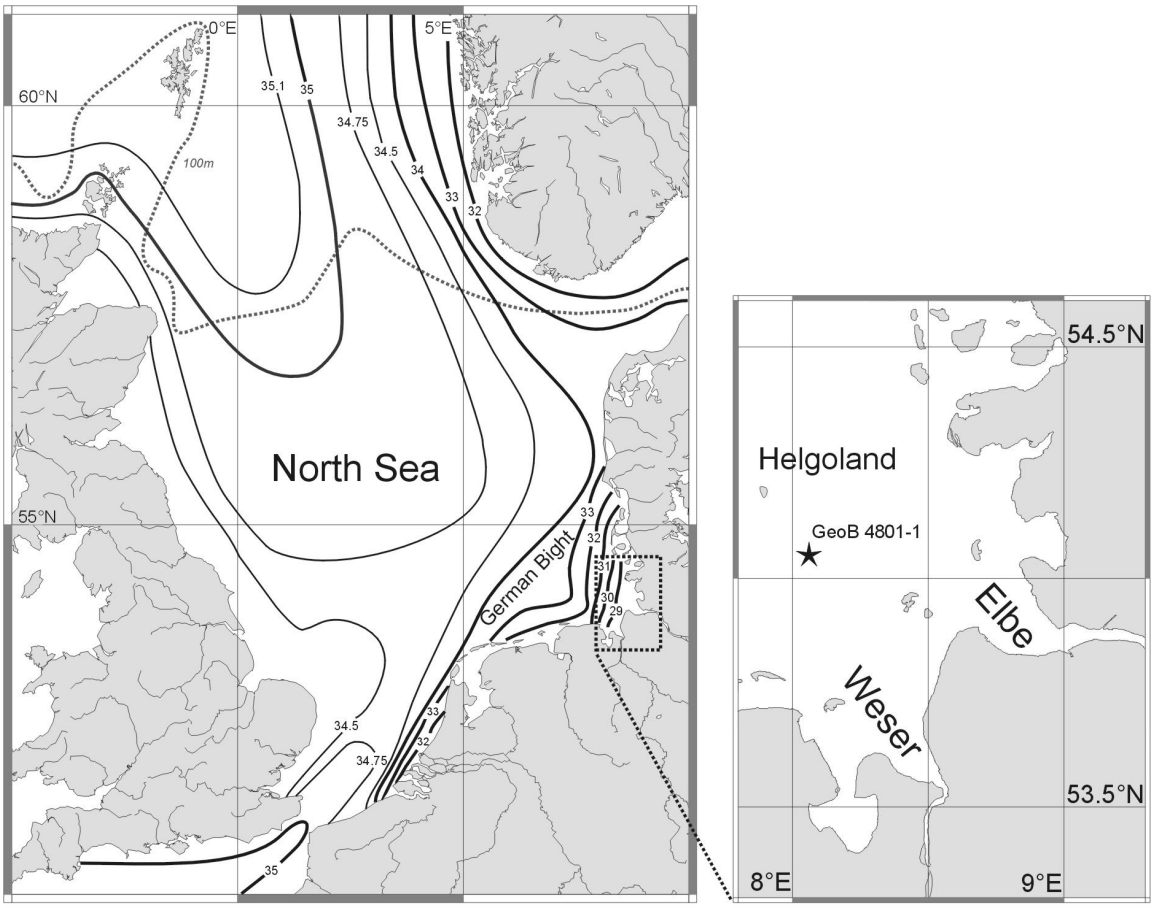

Figure 6.2: Left: The annual mean sea-surface salinity in the North Sea (after Schott, 1966). Right: The inner German Bight with the site (black star) of core GeoB 4801-1. The core location is situated southeast of the island of Helgoland and close to the Weser and Elbe Estuaries.

Like the Elbe River, several other major river systems (e.g. rivers Thames and Rhine) draining catchments in Central and Northern Europe provide runoff with high temporal variations and, therefore, are strongly affecting the salinity pattern of the North Sea (Figure 6.2). On a larger scale,

the Atlantic inflow via the English Channel as well as between Scotland and Norway determines the salinity distribution. This inflow of temperate and saline water masses is one of the main hydrographic driving forces of the North Sea.

The hydrography of the shallow North Sea (mean depth <100 m) is characterized by high tidal and wave energy levels, where re-sedimentation is a predominant process. Thus, the hydrographic conditions and also the morphology of the North Sea restrict continuous sedimentation to very few depocenters. Besides the most prominent sediment trap, the Skagerrak, there is also the Helgoland mud area. In this area, southeast of the island of Helgoland, sediments accumulate due to a small-scale eddy induced by currents, river runoff (Elbe and Weser River) as well as tidal dynamics (Hertweck, 1983).

6.3 Material, methods and stratigraphy

The investigation was carried out on sediment core GeoB 4801-1 (water depth 25 m, 54°06.7' N and 08°02.2' E, Helgoland mud area; Figure 6.2) collected during the Meteor cruise M 40/0 in 1997. Aiming to develop a high-resolution reconstruction of the paleoenvironmental history, the uppermost 1.4 m of the gravity core were sampled every centimeter.

For foraminiferal analyses the samples were washed over a 150 µm sieve to separate the coarse fraction. Subsequently, ten to twenty individual shells of the benthic foraminifera *Elphidium excavatum* (Terquem), *forma selseyensis* (Heron-Allen and Earland, 1911) were extracted. The specimens showed no signs of transport and/or reworking. The stable oxygen isotope composition of these shells was analyzed using a Finnigan MAT 251 mass spectrometer. The stable isotope measurements were calibrated against Vienna Pee Dee Belemnite (VPDB) by applying the NIST 19 standard. The analytical standard deviation is about ±0.07‰ VPDB (Isotope Laboratory Bremen University).

The age model of core GeoB 4801-1 is mainly based on ^{210}Pb data for the topmost core section and five AMS ^{14}C analyses further below (Hebbeln et al., 2003; Figure 6.3). The AMS ^{14}C-datings increase continuously with core depth and yield a sedimentation rate of 1.6 mm/yr on average. This is close to the ^{210}Pb derived sedimentation rate of 2.6 mm/yr. Thus, the radiocarbon data and the ^{210}Pb data imply that the upper 1.4 m of the core comprise a continuous sediment record covering the last 800 years (Hebbeln et al., 2003), with every datapoint giving a mean for about five years (Scheurle and Hebbeln, 2003).

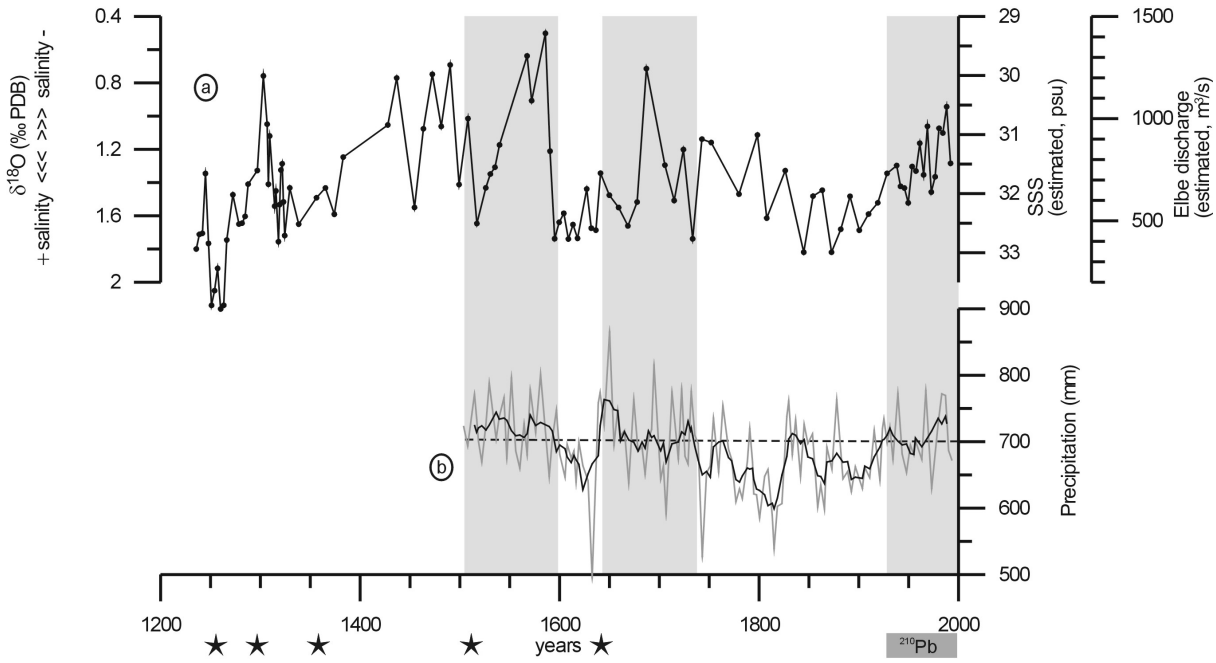

Figure 6.3: Comparison of (a) the $\delta^{18}O$ data of the Helgoland mud area and (b) an annual precipitation dataset reconstructed for Central Europe (smoothed values; Glaser, 2001). The y-axis for the $\delta^{18}O$ dataset (a) is placed on the left hand side (note the inverted scale). On the right hand side the equivalent SSS and Elbe discharge values (estimated) are indicated. Precipitation (b) values exceeding 700 mm (mean value of the time period AD 1500-1997; dashed line) are marked by light grey shading. The black stars mark the age control points, the dark grey bar on bottom of the age-axis points to the ^{210}Pb data for the topmost core section (Hebbeln et al., 2003).

6.4 Proxy application and interpretation

This study is based on a previous comparison of the $\delta^{18}O$ record (core GeoB 4801-1) with instrumental data for the last 100 years (Scheurle and Hebbeln, 2003). For this time span, instrumental sea-surface temperature (SST) and salinity (SSS) data were obtained from the island of Helgoland (Helgoland Reede). The comparison of these two datasets with the $\delta^{18}O$ data clearly demonstrated that the isotope data are dominated by the salinity effect, whereas the temperature signal is suppressed. This evaluation indicates the $\delta^{18}O$ signal to be a valuable proxy for SSS (summer/June-July). A $\delta^{18}O$ versus SSS regression line ($\delta^{18}O = 0.34 * SSS - 9.36$; $r = 0.86$) was determined.

Additionally, the SSS data measured at Helgoland Reede are closely linked to annual (and summer) Elbe discharge data. The discharge data were collected at the gauge of Neu Darchau (upstream of Hamburg), which is the last discharge gauging station inland. Thus, freshwater input via the Elbe River is having a major influence on SSS in the German Bight, and consequently Elbe discharge is also mirrored in the $\delta^{18}O$ data of core GeoB 4801-1. A $\delta^{18}O$-Elbe discharge relationship ($\delta^{18}O = -0.0013 * \text{discharge} + 2.29$; $r = 0.75$) has been established as well.

6.5 Reconstruction of paleosalinity and paleodischarge

The $\delta^{18}O$ values (Figure 6.3a) vary between 0.5 and 2.1‰ PDB with sudden changes (e.g. in the 1560-1590s) as well as slow decreases and increases (e.g. at the end of the 18th and the beginning of the 19th century). High $\delta^{18}O$ values reflect more saline conditions as a consequence of less freshwater input from river runoff. Low $\delta^{18}O$ values point to lower salinities and higher discharge rates. For instance, the 1560-90s are a period of historically documented extreme flooding events (Schmidt, 2001; Figure 6.3a).

The two relationships resulting from the comparison of the $\delta^{18}O$ data with instrumental data (see above) allow us to estimate a range for SSS and Elbe discharge variations within the last 800 years (Figure 6.3a). The estimations for the past century are in good agreement with the salinity measurements obtained at Helgoland Reede and the gauge measurements from Neu Darchau. For instance, the average discharge for the Elbe River estimated from $\delta^{18}O$ for the past ~100 years amounts to 754 m^3/s. This value corresponds well with the averaged annual Elbe River runoff given in the literature (750 m^3/s; Liedtke and Marcinek, 1994). The estimated paleodischarge ranges between 100 and 1375 m^3/s. This range appears reasonable as the instrumental data for the past 100 years indicate a range of annually averaged discharges from 300 to 1300 m^3/s (gauge data from Neu Darchau, Local Office for Waterways and Shipping in Lauenburg, displayed in Scheurle and Hebbeln, 2003).

The estimated paleosalinitiy values range between 29 and 33.9 psu (practical salinity units), which appears to be a rather broad range for a marine site. However, due to the mixing of river discharge and marine North Sea waters the present-day salinity distribution in the German Bight covers the entire range (29 psu at the mouth of the Elbe Estuary and 34 psu in the seaward parts of the German Bight; Figure 6.2). In addition, the most recent (AD 1940-2000) salinity data of about 31 to 32 psu generally fit to the annual mean salinity (31 to 32 psu) indicated for the core site in Figure 6.2.

6.6 Paleoclimatic implication

The $\delta^{18}O$ series covers the past 800 years, and climatic variations during this time period lead to significant changes in the paleoenvironmental parameters (SSS and river runoff) considered. As river discharge results from precipitation within the catchment area (Liedtke and Marcinek, 1994), our $\delta^{18}O$ record can also be regarded as a long-term precipitation record of the large Elbe catchment (Sturm et al., 2001; German Federal Institute of Hydrology in Koblenz, 2002). The paleoprecipitation record can be subdivided into four main periods. Between AD 1250 and the 1590s, there is a general trend towards lower salinities (lower $\delta^{18}O$ values), i.e. higher Elbe discharge rates. Therefore, we assume enhanced precipitation within the drainage basin. A sudden increase of salinity (high $\delta^{18}O$ values) occurs at the end of this time period. The following period, a period of climate amelioration (Schönwiese 1988; Brázdil 1994), is characterized by rather low discharges and precipitation totals, and lasts until AD 1680. After a temporary return to higher Elbe discharges and precipitation rates, the subsequent period points to a slow decrease of these hydrologic measures (AD 1680-1870). By the beginning of the 20th century this trend is reversed again, discharge as well as precipitation are gradually increasing.

By comparing the $\delta^{18}O$ data from the Helgoland mud area (Figure 6.3a) with an annual precipitation dataset reconstructed for Central Europe (Figure 6.3b; Glaser, 2001), similarities become evident. Precipitation values exceeding 700 mm (mean value of the time period AD 1500-1997) coincide with lower $\delta^{18}O$ values indicating significantly reduced salinities and more freshwater inflow from the catchment and vice versa. The best fit between these two datasets is observed for the period AD 1580 to 1640, marked by low precipitation and high salinities, and for the time after AD 1840 when precipitation decreased until AD 1900 followed by a significant increase towards the present day (Figure 6.3). However, there are other periods (e.g. AD 1740 to 1840), when there is less agreement between the two data sets. Glaser (2001) reconstructed an annual precipitation total, based on combined historical and instrumental meteorological observations scattered throughout Germany. In contrast, the Helgoland dataset reflects the precipitation history of a clearly defined region, the Elbe catchment, which represents a significant part of Central Europe. This record provides a regional integration of precipitation over the whole area. Besides, the discrepancies may be due to stratigraphical uncertainties. The stratigraphic resolution of the Helgoland record is limited by the accuracy of the AMS ^{14}C-datings having standard deviations of 25 to 50 years for the last 800 years. Shifting the Helgoland data about 20 to 40 years during the appropriate time period (AD 1640-1900, a period where we have no age control points), would result in an almost perfect fit (Figure 6.4).

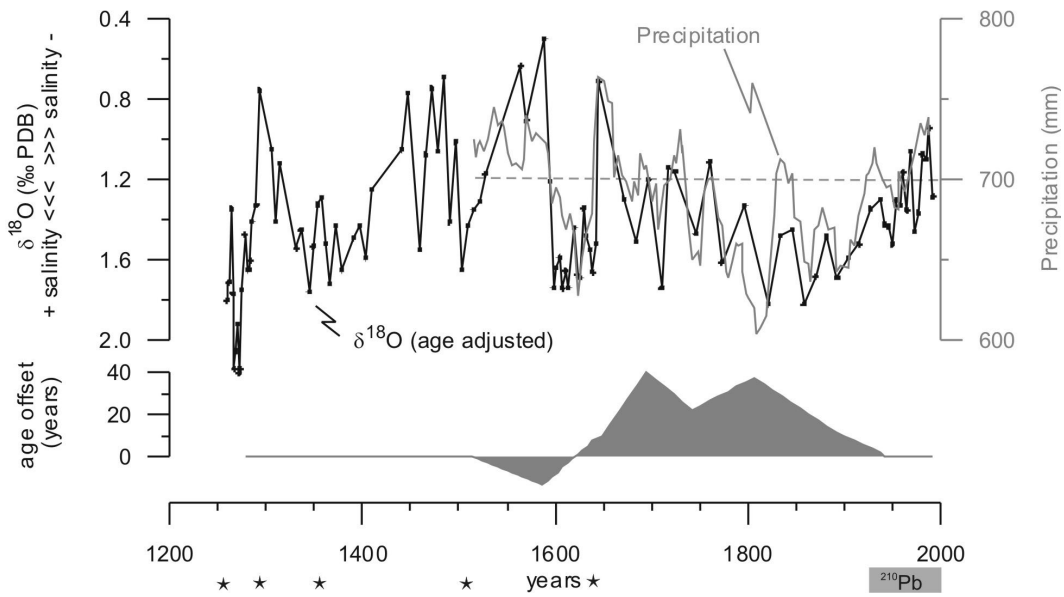

Figure 6.4: The $\delta^{18}O$ data of the Helgoland mud area (black line; age adjusted; see text) and an annual precipitation dataset reconstructed for Central Europe (grey line; smoothed values; Glaser, 2001) (see also Figure 6.3). The dashed line marks their mean value (700 mm) of the time period AD 1500-1997. Furthermore, the lower curve (dark grey shading) signals the offset in years resulting from the age adjustment of the original Helgoland $\delta^{18}O$ data to the Glaser (2001) precipitation curve. The black stars mark the age control points, the dark grey bar on bottom of the age-axis points to the ^{210}Pb data for the topmost core section (Hebbeln et al., 2003).

Regarding temporal data resolution, we assume that the dataset derived from the sediments is displaying a mean for several years rather than single events (Scheurle and Hebbeln, 2003). Even though it cannot be ruled out that extreme precipitation, probably combined with deforestation and/or land-use changes in the catchment, results in flood events leaving a pronounced imprint (Radach and Berg, 1986; Van der Ploeg and Schweigert, 2001), such meteorological events are rather unlikely to be reflected in the smoothed (about five years) record, which instead is more likely driven by long-term changes in precipitation.

Paleoclimate data based on historical information about the region of interest is sparse. Nevertheless, documentary evidence from the Czech Lands show, for instance, that while the climatic conditions in the first half of the 16th century were favorable, the latter half was characterized by cooler and wetter conditions, culminating in the 1590s (Munzar, 1992; Brázdil, 1994). Based on such historical data as well as instrumental records (since the late 18th century) and proxy data, precipitation was reconstructed for Prague (since AD 1200; Brázdil, 1996). This city is situated on the Moldau River, one of the major tributaries of the Elbe River (Figure 6.1). Although

this precipitation is contributing to the runoff entering the North Sea via the Elbe River, there is no clear relation between the precipitation record for the Prague area and the Elbe River discharge as reflected in the Helgoland $\delta^{18}O$ record. This might be due to a regional effect. The mountainous areas of the Bohemian Forest and Giant Mountains experience high precipitation, but the areas surrounding Prague, which are characterized by lower topography receive less precipitation.

Regarding the comparisons, it may be concluded that our proxy record is focusing on the entire Elbe catchment area, and provides reliable paleoprecipitation data for a longer time period than previously achieved. The results encourage further investigations following this approach.

6.7 Conclusion

For the first time, this study provides a high-resolution, 800-year long record of paleosalinity in the German Bight and paleodischarge of the Elbe River. These data, based on $\delta^{18}O$, can also be used to assess paleoprecipitation over the Elbe catchment area, and thus, provide a unique paleoclimatic record.

Though the pattern of climatic change is different in selected areas of Europe (Pfister and Lauterburg, 1992), a comparison of the Elbe catchment data with a reconstructed precipitation dataset for Central Europe (until AD 1500; Glaser, 2001; Figure 6.3) indicates a link to the overall atmospheric circulation system. Phases with more precipitation (>700 mm as the mean value of the time period AD 1500-1997) are recorded as phases with a low-salinity signal in the Helgoland $\delta^{18}O$ data, pointing to a pronounced freshwater influence via the Elbe River.

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7 Late Holocene environmental changes in the Skagerrak, eastern North Sea - foraminiferal indication

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Marine Micropaleontology (submitted)

Abstract

A high-resolution study of marine environmental conditions and changes through the Late Holocene (2,700 years) in the eastern North Atlantic realm is presented. In order to contribute new knowledge about the North Sea region, the faunal distribution, fluxes of benthic and planktonic foraminifera and grain-size data were analysed from a gravity core (GeoB 6003-2) from the Skagerrak. The data indicate environmental changes (i.e. stability and productivity) both in the bottom and surface waters. Major shifts, recorded at around 2200, 1900, 1500, 1200 and 400 cal yr BP, are interpreted as climatic changes. The resulting intervals are linked to well-known historical epochs such as the Subatlantic Pessimum, the Roman Period, the Migration Period, the Medieval Warm Period and the Little Ice Age. The benthic foraminiferal flux indicates that highly productive phases are generally connected with relatively cool climates. Moreover, the planktonic foraminiferal flux reflects a decreasing influence of North Atlantic surface water masses since the end of the Migration Period at around 1200 cal yr BP.

Keywords: foraminifera, paleoenvironment, paleoceanography, Skagerrak, Late Holocene

7.1 Introduction

Meteorological and hydrographical instrumental data recording Northern Hemisphere climate variations are only available for the last 250 years. Further back in time, useful information for the reconstruction of past climatic changes is for example preserved in marine sediments, i.a. in microfossil assemblages. Those indicators that have a close relationship to parameters such as temperature and salinity are called environmental 'proxies' for that particular parameter (Wefer et al., 1999).

Although the climate of the Holocene (last 11,500 years) is designated as rather stable, significant changes occurred within this period. The knowledge of the exact timing of the onset and the duration of these climatic changes as well as the background for the variations is, however, still sparse. Examples of climatic shifts in the Late Holocene include the Medieval Warm Period, coinciding with the Viking Period and Nordic settlement of Iceland and Greenland, and the climatic depression of the Little Ice Age, when coolness and rainfall resulted in crop failures and diseases in Europe (Lamb, 1977; Schönwiese, 1988; Lozán, 1998; Grove, 2002). Since the beginning of industrialization such natural climatic variations are increasingly influenced by mankind.

Due to a close coupling of atmospheric and oceanic circulation (Rodhe, 1987; van Weering et al., 1993a) and due to high sedimentation rates (e.g. van Weering, 1981; van Weering et al., 1987; Pederstad et al., 1993; van Weering et al., 1993 a, b), the Skagerrak area bears high potential for environmental reconstruction and appears to be a promising place to study Holocene climate changes.

A number of investigations have been carried out on recent as well as on subrecent foraminiferal data from the Skagerrak (e.g. Qvale et al., 1984; Qvale and van Weering, 1985; Corliss and van Weering, 1993; Seidenkrantz, 1993; Conradsen et al., 1994; Alve and Murray, 1995; Alve, 1996; Bergsten et al., 1996; Alve and Murray, 1997). In general, surface sediments were examined in these studies with respect to the spatial distribution of benthic foraminifera and their response to water mass properties. For the Holocene, paleoenvironmental interpretations based on several proxies were presented by Conradsen and Heier-Nielsen (1995), Knudsen et al. (1996) and Knudsen and Seidenkrantz (1998). A multidisciplinary study of a core from the northern part of the Skagerrak contributed new knowledge on the general climatic development in the area through the Late Glacial and the Holocene (Stabell and Thiede, 1985). The time resolution for the Late Holocene part in these studies is, however, not high enough for a detailed correlation with our record.

Benthic foraminiferal assemblages and the sedimentological development through the Late Holocene in the Skagerrak were described by Hass in 1993, 1994, 1996, and 1997, who investigated sites located further to the east, at water depths of about 420 m. In the present paper, we partly follow Hass's approach in adopting the links to historical epochs. Comparisons to his cores were, however, problematic, e.g. because of differences in temporal resolution and water depth.

This study is part of the HOLSMEER project aiming to reconstruct paleoenvironmental conditions in the North Atlantic realm through the last 2,000 years. Benthic and, for the first time planktonic foraminiferal distributions were used as proxies for the interpretation of environmental and climatic changes in the Skagerrak area during the Late Holocene (c. 2,700 years).

7.2 Geographic and oceanographic setting

In this study, sediment core GeoB 6003-2, which was obtained from the central Skagerrak basin at 57°58.3'N and 9°23.2'E at a water depth of 312 m (Figure 7.1), was investigated. The Skagerrak is the deepest part (700 m) of the Norwegian Trench, which has been formed by repeated glacial advances from the east and the south during the Quaternary. The trench has the topography of a large fjord (Rodhe, 1987) and runs parallel to the Norwegian coast. The asymmetric Skagerrak basin has an irregular, steep northern and a more regular, gentle southern slope (Figure 7.1). It serves as a natural sediment trap receiving input from a number of northwest European source areas and from remobilization of North Sea sediments.

The modern current system is influenced by the large-scale atmospheric and oceanic circulation pattern as well as by the outflow of Baltic waters (e.g. Rodhe, 1987). The Skagerrak surface circulation is a part of the anticlockwise movement of the North Sea waters (Figure 7.1). A large amount of North Atlantic water enters the Skagerrak directly through the Tampen Bank Current (TBC). The Southern Trench Current (STC) brings water masses from the north-western North Sea. Another contributor to the current system is the South Jutland Current (SJC) which runs parallel to the Danish west coast and continues as the North Jutland Current (NJC). In the innermost Skagerrak, the NJC and STC waters are supplemented by less saline Baltic outflow water, the Baltic Current (BC). These combined waters turn towards northwest and west and form the Norwegian Coastal Current (NCC), which continues towards north along the Norwegian coast (Figure 7.1).

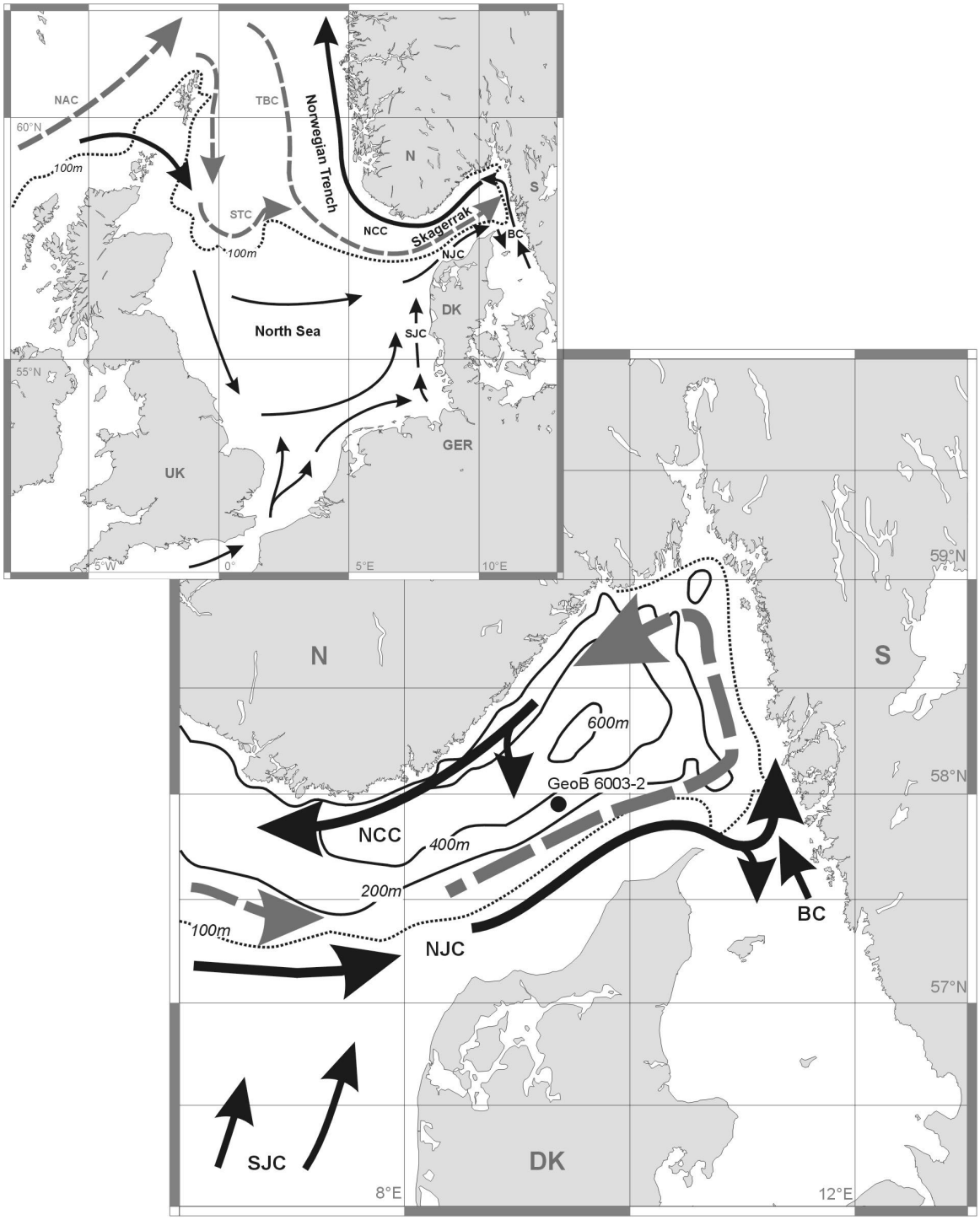

Figure 7.1: Location of core GeoB 6003-2 and the present-day surface circulation pattern in the Skagerrak and the North Sea (insert map). Partly, North Atlantic water flows directly into the Skagerrak (grey-stippled arrows), modified waters flow into the area via the southern North Sea or comes from the Baltic (black arrows). Abbreviations: BC, Baltic Current; DK, Denmark; GER, Germany; N, Norway; NAC, North Atlantic Current; NCC, Norwegian Coastal Current; NJC, North Jutland Current; S, Sweden; SJC, South Jutland Current; STC, Southern Trench Current; TBC, Tampen Bank Current; UK, United Kingdom. Modified from Danielssen et al. (1991) and Nordberg (1991).

The deep water circulation in the Skagerrak follows more or less the same pattern as the surface circulation, but the velocity of the bottom currents is lower than that of the surface currents (Dahl, 1978; Qvale and van Weering, 1985; Rodhe, 1987). A volumetrically large part of the high salinity and oxygen-rich bottom water is of North Atlantic origin. A minor amount is derived from dense water formed in the northern North Sea during very cold winters (Løjen and Svansson, 1972).

In general, the highest current speeds occur on the convex southern slope of the Skagerrak (Rodhe, 1987), to which the inflowing water masses are constrained. Due to the relatively high near-bottom current velocities, non-deposition or reworking of sediments may occur locally, especially in the shallow areas with water depths of less than 100 m (van Weering, 1981; Salge and Wong, 1988; Kuijpers et al., 1993).

7.3 Material and methods

The sediment core GeoB 6003-2 (core length 1049 cm), which consists mainly of silty clay, was obtained in 1999 during RV Meteor cruise M45-5 in the Skagerrak. Environmental proxy data were analysed within the upper 400 cm of the core at five centimetre intervals.

The sediment samples for foraminiferal analyses were wet sieved on 63 and 150 µm sieves (see grain-size distribution, Figure 7.2). In order to obtain better comparison with previous foraminiferal analyses in the area, the samples were subsequently dry sieved on a 100 µm sieve. Foraminifera in the size fraction >100 µm were concentrated by the help of the heavy liquid CCl_4 (ρ = 1.59 g/cm^3) as described by Meldgaard and Knudsen (1979). The major part of the fauna consisted of calcareous benthic foraminifera which were generally well preserved. Some agglutinated foraminifera occurred, but due to their poor preservation potential in the geological record these were excluded from the analyses. The planktonic fauna consists of only a few species. At least 300 specimens of benthic calcareous and 300 specimens of planktonic foraminifera were identified and counted in each of the 80 samples, where possible. In total, 165 different foraminiferal taxa were determined. The 15 most important taxa, 13 benthic and 2 planktonic, are listed in the Appendix.

The percentage distribution of the most important benthic species in the core was used for the subdivision of the record into intervals (Figures 7.3-7.5) and for the interpretation of environmental conditions. Moreover, the benthic and the planktonic foraminiferal fluxes were considered for the interval definitions and interpretation.

The foraminiferal flux, as a measure for productivity, was determined with the following equation: $F = ni * \rho_d / t$ (F = foraminiferal flux (specimens/cm^2/year); ni = number of foraminifera per gram; ρ_d = dry specific density (g/cm^3); t = years per centimetre core interval). For the dry specific density an average value of 2.0 g/cm^3 (Knudsen and Seidenkrantz, 1998) was inserted. The fluxes are shown in Figures 5 (planktonic) and 6 (benthic).

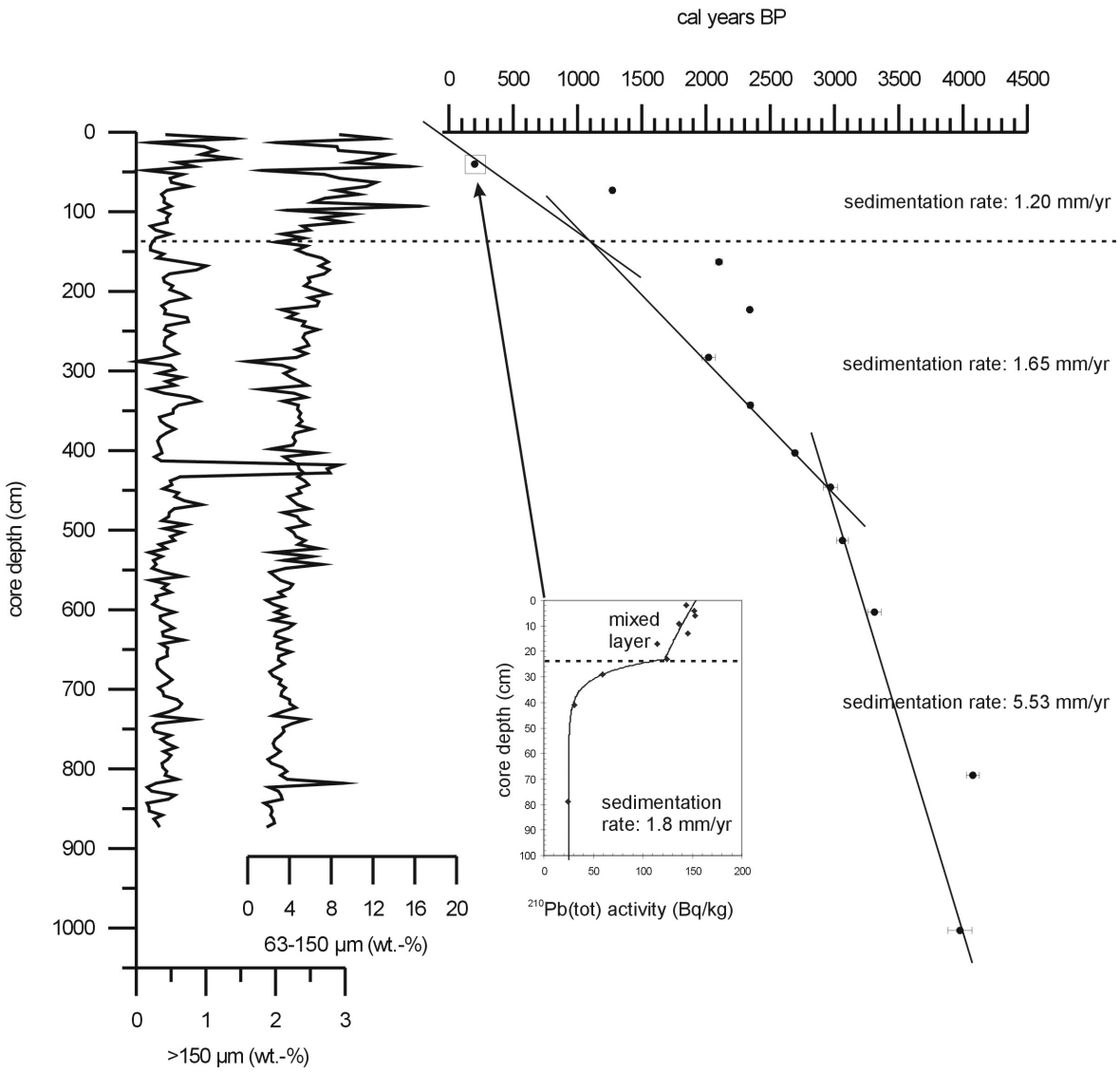

Figure 7.2: The age-depth model for core GeoB 6003-2 with the AMS ^{14}C datings (shown with 1σ error bars), the ^{210}Pb results and the sedimentation rates for each linear segment. The change in grain-size distributions (63-150 and >150 µm) at about 138 cm depth was used to support the definition of an age zone boundary at that level.

7.4 Chronology

The chronology of the entire core GeoB 6003-2 down to its base at 1049 cm is discussed in order to understand the changes in sedimentation rates throughout the record. The age-depth model was constructed on the basis of 11 AMS ^{14}C datings and ^{210}Pb measurements, taking changes in grain-size distribution into account (Figure 7.2).

The ^{210}Pb determinations for the uppermost part of the record were performed at the Netherlands Institute for Sea Research (NIOZ) by measuring ^{210}Pb activity via it's α-particles emitting grand daughter ^{210}Po, following the method described by van Weering et al. (1987). The ^{210}Pb activity in the upper 40 cm decreases significantly downcore, enabling an accurate estimation of the sedimentation rates (1.8 mm/yr) for the last 200 years (Figure 7.2).

The AMS ^{14}C datings were made at the Leibniz Laboratory for Age Determinations and Isotope Research at the University of Kiel. They were carried out on approximately 10 mg carbonate from mixed benthic foraminifera or from shell fragments, respectively (Table 7.1). All ages were corrected for fractionation, and the $\delta^{13}C$ values are listed in Table 7.1. The ^{14}C ages were calibrated with CALIB4 (Stuiver et al., 1998), using the marine model calibration curve. A standard reservoir correction of 400 years ($\Delta R=0$) is built into this model (see also Heier-Nielsen et al., 1995b).

The AMS ^{14}C datings do not all increase continuously with core depth (Figure 7.2, Table 7.1). A few of the datings on benthic foraminifera appear to give too old ages compared to the others. Reworked older material, moved by hydrodynamic transportation, might have caused such age discrepancies. Temporary intense bottom currents may thus have led to the transport and re-deposition of a certain amount of foraminiferal tests in the sediment. This is the background for considering the too old ages for the upper three AMS ^{14}C datings as a result of partly reworked assemblages (Figure 7.2). A similar process of reworking may also explain an apparently too old age for the 803 cm level. Deviations towards too old ages have previously been reported by Heier-Nielsen et al. (1995a) and by Knudsen et al. (1996) from high-energy Holocene deposits of northern Denmark.

The distribution of the grain-size fractions 63-150 and >150 µm through core GeoB 6003-2 (Figure 7.2) also indicates varying hydrodynamic conditions in the area during the Late Holocene. Presumably, a significant coarsening of the sediment in the upper 200 cm (from a mean value of about 4 to 8 weight percent of the grain-size fraction 63-150 µm) points to gradually higher current velocities in the area. The distinct grain-size change at 138 cm depth is suggested to represent the level of a change in sedimentation rate (Figure 7.2, Table 7.2). This level is, therefore, introduced as an age zone boundary between two linear segments of the age model.

Accordingly, the sedimentation rate from core top down to 138 cm amounts to 1.20 mm/yr. This is close to the sedimentation rate obtained by the ^{210}Pb measurements. A second linear segment, determined by four AMS ^{14}C datings, indicates a sedimentation rate of 1.65 mm/yr between 138 cm and a dated sample at 446 cm. A third linear segment between 446 cm and the deepest date at 1003 cm is determined by two datings on shell fragments and two on benthic foraminifera. The accumulation rate in this interval is 5.53 mm/yr. These sedimentation rates are in agreement with the results of previous studies in the area (van Weering et al., 1987; Kuijpers et al., 1993; Pederstad et al., 1993).

Thus, the suggested age model for core GeoB 6003-2 (Figure 7.2, Table 7.2) is based on a linear interpolation between the top of the core and two age zone boundaries. The age-depth model shows that the upper 400 cm of the core comprise the last 2,700 years of the Late Holocene. The temporal resolution in this study is approximately 30 to 40 years per sample.

Table 7.1: AMS ^{14}C datings used for the age-depth model for the Skagerrak core GeoB 6003-2. A reservoir correction of 400 years has been applied. The ages were calibrated with CALIB4 (Stuiver et al., 1998), using the marine model calibration curve. The 1σ intervals (min./max. range) of the calibrated ages and the δ^{13}C values used for correction of the fractionation are given.

Core depth (cm)	Lab. no.	Material	^{14}C ages (BP)	Err. (+/-)	Reservoir corrected ages (BP; R=400)	Calibrated ages (cal yr BP)	Min./Max. range 1σ	δ^{13}C
73	KIA 17063	Benthic foraminifera	1720	30	1320	1272	+17/-19	+1.8
163	KIA 17062	Benthic foraminifera	2455	30	2055	2101	+27/-47	+2.1
223	KIA 18914	Benthic foraminifera	2680	25	2280	2341	+12/-13	+0.2
283	KIA 18238	Benthic foraminifera	2400	35	2000	2020	+53/-40	-9.8
343	KIA 18236	Benthic foraminifera	2690	30	2290	2345	+22/-14	-2.6
403	KIA 13694	Benthic foraminifera	2875	30	2475	2692	+15/-30	0.1
446	KIA 18926	Shell fragments	3190	30	2790	2968	+55/-34	+1.1
513	KIA 18925	Shell fragments	3245	30	2845	3062	+47/-58	-1.2
603	KIA 13693	Benthic foraminifera	3420	70	3020	3311	+53/-103	-20.5
808	KIA 13691	Benthic foraminifera	4050	35	3650	4076	+51/-70	-2.1
1003	KIA 13690	Benthic foraminifera	3990	50	3590	3976	+94/-68	-11.6

Table 7.2: Age zone boundaries for the three linear segments in the age-depth model for the Skagerrak core GeoB 6003-2. The sedimentation rate for each interval is indicated as well as the background for determination of the age zone boundaries, including the number of datings.

Top (cm)	Base (cm)	Base (cal yr BP)	Sedimentation rate (mm/yr)	Basis (dating type, number)
0	138	1100	1.20	^{210}Pb and grain-size data
138	446	2968	1.65	AMS ^{14}C datings (4)
446	1003	3976	5.53	AMS ^{14}C datings (4)

7.5 Foraminiferal analyses and faunal interpretation

The percentage distribution of the most important benthic foraminiferal species is shown in Figures 7.3a-j and 7.4. Among those, *Cassidulina laevigata* (b), *Elphidium excavatum* (i), *Pullenia osloensis* (g) and *Stainforthia fusiformis* (j) are the most abundant. Few planktonic specimens are found, and they are generally small-sized. The flux of each of the two most common planktonic species *Globigerinita uvula* (a) and *Turborotalita quinqueloba* (b) is shown in Figure 7.5, together with the percentages of the benthic group of Miliolids (c). The environmental indication of the foraminifera is interpreted on the basis of modern distributions and ecology of the species in the Skagerrak and North Sea areas.

For the interpretation of the data, it is important to be aware of possible reworking (see above), which may have influenced the original fauna by removing or introducing additional species and/or specimens. Species like *E. excavatum* as well as some species of the group of Miliolids are common in relatively shallow, high-energy environments of the southern slope of the Norwegian Trench (Conradsen et al., 1994). These are suggested to be easily transported into the area of investigation (Qvale and van Weering, 1985; Conradsen and Heier-Nielsen, 1995). For instance, the increasing abundances of *E. excavatum* (Figure 7.3i) after around 1000 cal yr BP may be at least partly due to reworking from shallower areas. Even though three radiocarbon dates from the analysed sequences of the core appear to give too old ages, no major disturbances of assemblage compositions were clearly visible.

7.5.1 Benthic foraminifera

On the basis of major faunal changes, the record was subdivided into six intervals (Figure 7.3, A-F). The assemblages are described, and their environmental indication is interpreted in the following (from the bottom to the top). To simplify the description, the species are lettered with a-j.

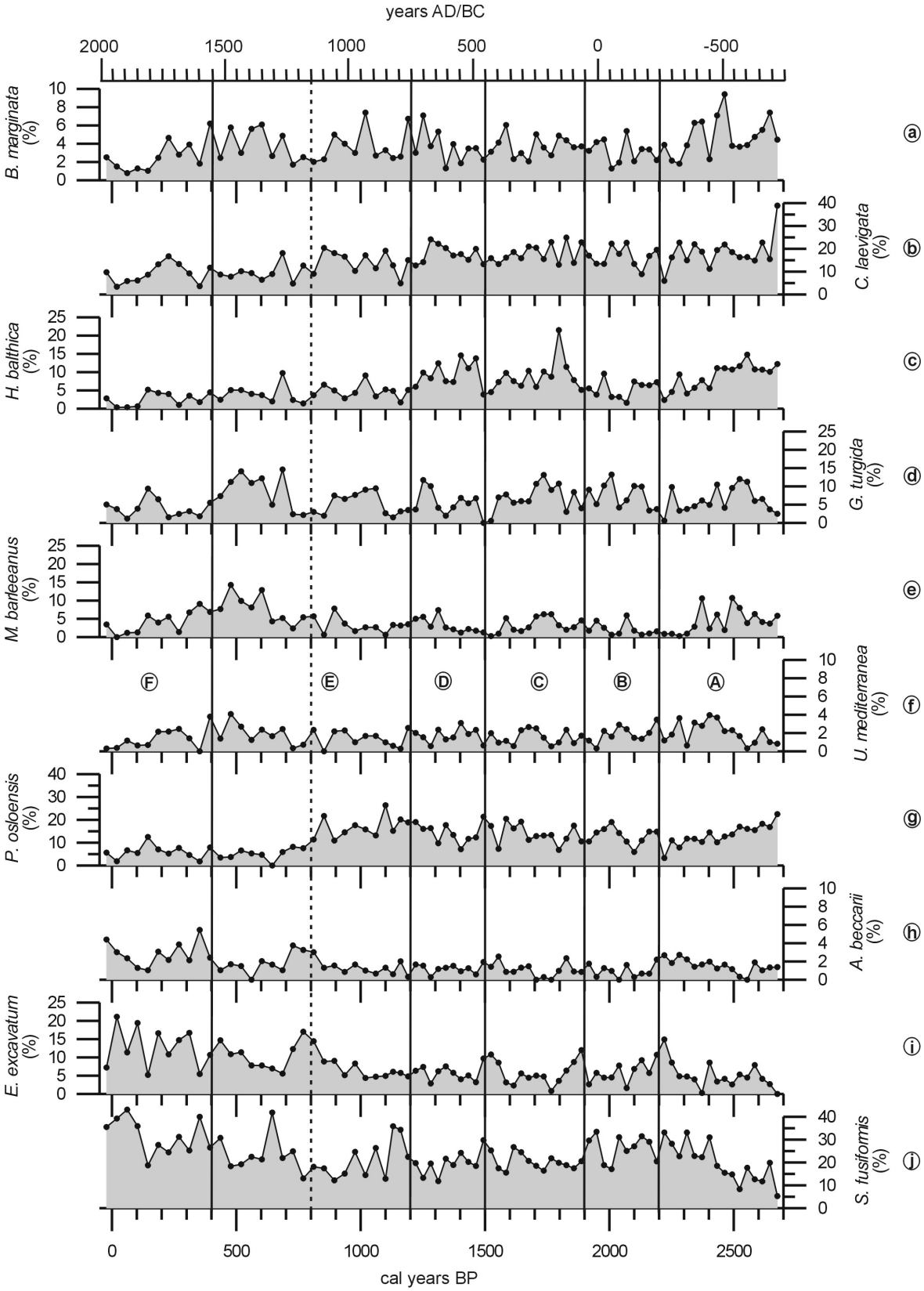

Figure 7.3: The percentages of selected benthic foraminifera in core GeoB 6003-2. The horizontal unbroken lines indicate the subdivision of the record into intervals (A-F), whereas the stippled line marks an environmental change at around 800 cal yr BP.

The time interval 2690-2200 cal yr BP (A)

The most common species in this interval are *C. laevigata* (b), *S. fusiformis* (j), *P. osloensis* (g) and *Hyalinea balthica* (c). The percentages of *C. laevigata* (b) are relatively constant. *Stainforthia fusiformis* (j) changes from low frequencies in the lower part to higher values from around 2500 cal yr BP. A drop in percentages of *H. balthica* (c) is recorded at the same level. *Pullenia osloensis* (g) decreases steadily through the entire interval. The percentages of *Bulimina marginata* (a) and *Melonis barleeanus* (e) are low, with strong variations and a general decrease through the interval. The percentages of *Cibicides lobatulus* show maximum values during the upper part (Figure 7.4).

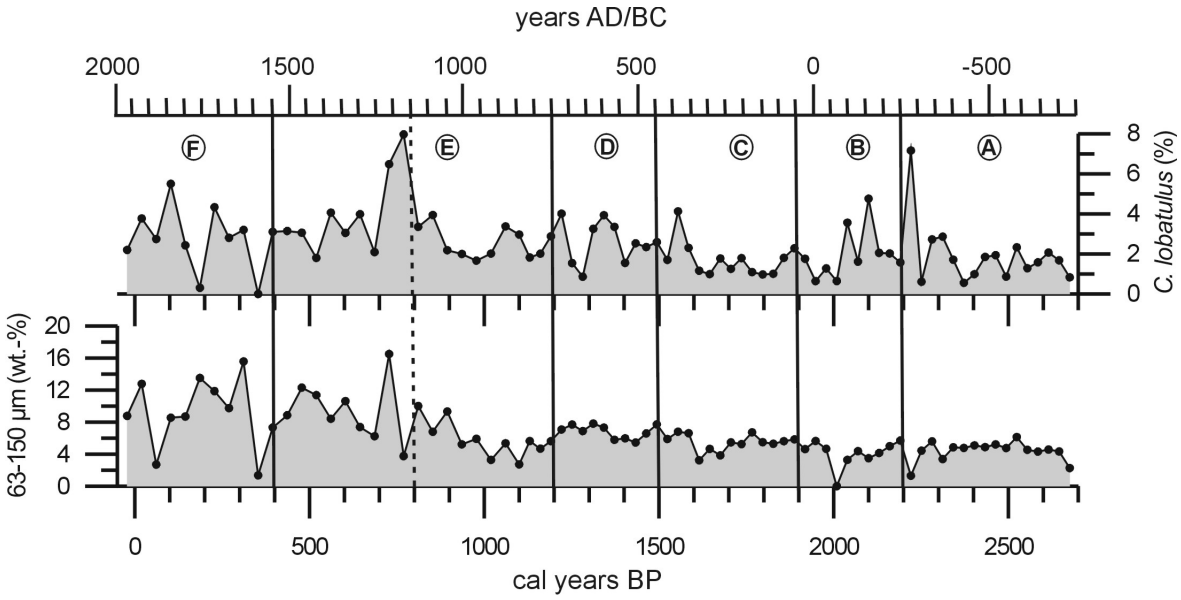

Figure 7.4: Selected indicators of bottom current activity: the percentages of the grain-size fraction 63-150 µm and the percentages of Cibicides lobatulus (for intervals see Figure 7.3).

In modern faunas of the Skagerrak area, *C. laevigata* is abundant on the southern and the northern slopes of the Norwegian Trench. This species is related to the boundary zone between the stable bottom water mass and the surface waters with more variable environmental conditions (Qvale and van Weering, 1985; Conradsen et al., 1994). The opportunistic species *S. fusiformis* is especially abundant on the shallower part of the southern slope (Alve and Murray, 1995, 1997; Bergsten et al., 1996). *Stainforthia fusiformis* and *B. marginata* are both associated with high contents of organic carbon in the sediment (Qvale and van Weering, 1985; Alve and Murray, 1997). *Pullenia osloensis* appears to be confined to the deeper part of the Norwegian Trench (Bergsten et al., 1996). *Hyalinea balthica* is most common in the deepest part as well, but mostly in small numbers, and is linked to high organic carbon contents (Qvale and van Weering, 1985). Hass (1997) considered *H. balthica*

to be an indicator of low oxygen contents in stagnant bottom waters. *Melonis barleeanus* is most abundant in the deeper part of the Norwegian Trench (Qvale and van Weering, 1985; Bergsten et al., 1996). This species has a preference for slightly altered organic material (Caralp, 1989) and is associated with areas of high organic carbon content (Murray, 1991). *Cibicides lobatulus* is an indicator of high bottom current velocities (Murray, 1991).

During interval A the foraminiferal assemblages generally point to a high input of organic matter to the area, an interpretation which is based on relatively high amounts of one or more of the species *S. fusiformis* (j), *H. balthica* (c) and *B. marginata* (a). Opposite trends in the relative amounts of such species are often seen. This may either be due to the influence of other environmental parameters than organic carbon content or to biological competition. The decrease in *H. balthica* (c) and *P. osloensis* (g) together with the increase in *S. fusiformis* (j) indicates a change from relatively stable bottom waters towards more variable conditions in the later part of the interval. The increasing trend in percentages of the less common species *E. excavatum* (i) and *Ammonia beccarii* (h) also points to gradually higher environmental variability, and *C. lobatulus* (Figure 7.4) shows increasing bottom current velocities towards the end of the interval.

The time interval 2200-1900 cal yr BP (B)

The dominant species in this interval are *S. fusiformis* (j), *C. laevigata* (b) and *P. osloensis* (g). *Globobulimina turgida* (d) is relatively common too, with frequencies close to those in the previous interval. The onset of this interval is characterised by an increase in percentages of *P. osloensis* (g) and an abrupt decrease in *A. beccarii* (h). After an abundance peak in *E. excavatum* (i) at around 2200 cal yr BP, the frequency of this species returned to similar low values as before (interval A). *Stainforthia fusiformis* (j) and *C. laevigata* (b) occur with continued high frequencies. The percentages of *P. osloensis* (g) are generally slightly higher than during the later part of the previous interval. A decreasing trend in percentages of *Uvigerina mediterranea* (f) is already apparent since around 2500 cal yr BP, while the percentages of *M. barleeanus* (e) became very low at around 2300 cal yr BP. The percentage values are continuously low through interval B.

Uvigerina mediterranea is a slope species overlapping the area of *C. laevigata* in the modern Norwegian Trench (Qvale and van Weering, 1985). It does not live in high-energy environments, but seems to require stable water conditions. The decreasing trend may point to an increase in variability through the interval. The composition of species, e.g. with high percentages of *S. fusiformis* (j) and *C. laevigata* (b), indicates continuously variable environmental conditions. This is supported by relatively high contents of *C. lobatulus* (Figure 7.4).

The time interval 1900-1500 cal yr BP (C)

Cassidulina laevigata (b) remains the dominant species together with *S. fusiformis* (j) and *P. osloensis* (g). At the transition to this interval there is, however, a decrease in the percentages of *S. fusiformis* (j). *Hyalinea balthica* (c) increases in frequency at the transition and has its maximum peak at 1800 cal yr BP. After a marked decrease, there is a slight general, but variable increase in the percentages of the accessory species *A. beccarii* (h) and *E. excavatum* (i) during the later part of the interval. These species are tolerant to variable environmental conditions, but their distribution pattern appears to reveal two periods of temporary less variability. *Globobulimina turgida* (d) and the less common *M. barleeanus* (e) show considerable variation with relative frequency minima at about 1500 cal yr BP.

In general, the environmental conditions are still highly variable, but there is faunal indication of more stable conditions than during the previous interval. This is for instance indicated by relatively high amounts of *H. balthica* (c) and the decrease in *S. fusiformis* (j). A relatively high input of organic material is indicated by the contributions of *H. balthica* (c), *G. turgida* (d) and *M. barleeanus* (e).

The time interval 1500-1200 cal yr BP (D)

The species distribution is not very different from that in the previous interval. The dominant species are the same, though with slightly decreasing percentages of *S. fusiformis* (j). After an increasing trend in *C. laevigata* (b) and generally high percentages of *H. balthica* (c) through this interval, there is a marked drop in the abundances of both species. *Melonis barleeanus* (e) and *B. marginata* (a) gradually increase in frequency through the entire interval. The amount of Miliolids (Figure 7.5c) has increased compared to the intervals before 1500 cal yr BP, but the values are generally rather low in the core. It is an indicator group for increased bottom water salinity (Murray, 1991, 1992).

During interval D, the environment may be slightly more stable, as reflected by relatively high percentages of *H. balthica* (c) and the decrease in *S. fusiformis* (j). The Miliolids point to an increased influence of North Atlantic bottom waters in the area. A hydrographic shift to more unstable conditions is indicated close to the end of the interval. This is displayed by a drop in *H. balthica* (c) and *C. laevigata* (b) and a subsequent increase in *S. fusiformis* (j). The faunal indication of organic supply to the area is not much different from that in the previous interval.

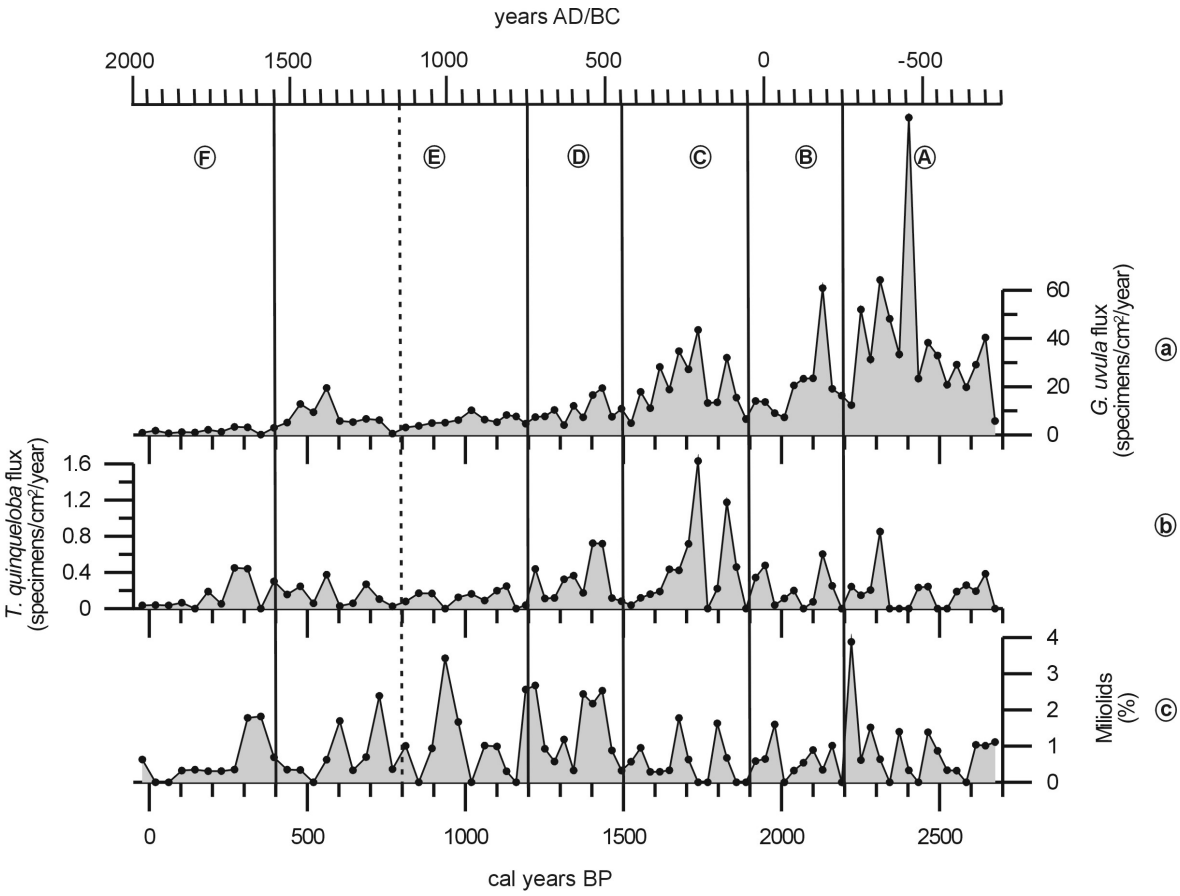

Figure 7.5: Indicators for North Atlantic water influence in the Skagerrak: the fluxes of the planktonic Globigerinita uvula and Turborotalita quinqueloba as well as the percentages of the benthic group of Miliolids (for intervals see Figure 7.3).

The time interval 1200-400 cal yr BP (E)

After 1200 cal yr BP there is an increase in *S. fusiformis* (j). A subsequent decrease between 1100 and 800 cal yr BP is again followed by gradually higher percentages. *Cassidulina laevigata* (b) and *H. balthica* (c) decrease in frequency, with especially low values at around 800 cal yr BP. Similar minima are seen for *B. marginata* (a) and *G. turgida* (d). These latter species are associated with high organic carbon contents in the sediment and tolerate low-oxygen environments (Sen Gupta, 1999). *Elphidium excavatum* (i) and *A. beccarii* (h) show a significant rise in percentages between 1200 and 900 cal yr BP, reaching maxima at around 800 cal yr BP. The most pronounced change in *P. osloensis* (g) also occurs at around 800 cal yr BP, a change which appears to have taken place within a few decades. After 700 cal yr BP, *G. turgida* (d) reaches its highest values for the entire record. In addition, *M. barleeanus* (e) becomes much more abundant, reaching maximum percentages after about 600 cal yr BP.

The environmental conditions during the early part of this interval were more unstable than during the previous interval. At 800 cal yr BP, there is a marked environmental change, and since then the assemblages gradually show even higher instability. This is for instance indicated by the decrease in *P. osloensis* (g) and by the increase in *S. fusiformis* (j) and *E. excavatum* (i). The relatively high percentages of *S. fusiformis* (j), *M. barleeanus* (e), *G. turgida* (d) and *B. marginata* (a) point to an increased input of organic matter to the area after 800 cal yr BP. The Miliolids (Figure 7.5c) continue to show periodical abundance peaks throughout interval E, indicating variations in the influence of North Atlantic water masses to the area.

The increasing abundances of *E. excavatum* (i) and *A. beccarii* (h) clearly indicate gradually more unstable conditions and lower salinities in the area (Qvale and van Weering, 1985), an interpretation which is supported by the general changes in the assemblages, e.g. increase in *C. lobatulus*, and by the change in grain-size distribution towards coarser sediments (Figures 7.2 and 7.4). The combined change in the fauna and in the sediment may, however, partly be a result of an increased input of reworked material from shallower areas of the slope. Reworking of part of the assemblage is indicated by some apparently too old ^{14}C datings within the depth range of this interval (see above).

The time interval 400 cal yr BP-present (F)

This interval is characterised by an increase in the percentages of *A. beccarii* (h), *E. excavatum* (i) and *S. fusiformis* (j), the latter species reaching particularly high values. *Ammonia beccarii* (h) and *E. excavatum* (i) fluctuate a lot in frequency. A decrease in salinity, as indicated by these two species, is supported by low percentages of the Miliolids (Figure 7.5c). Other characteristic features for this interval are continued low percentages of *P. osloensis* (g), decreased percentages of *G. turgida* (d) and a marked gradual decrease in *U. mediterranea* (f), *M. barleeanus* (e) and *B. marginata* (a).

A single distinct peak in the abundance of *Bolivina skagerrakensis* is seen at one level around 150 years ago. This species typically occurs in modern faunas of the deepest part of the Skagerrak, and it was suggested by Qvale and van Weering (1985) and Conradsen et al. (1994) to be an indicator of stable bottom water conditions. In contrast to this, Hass (1997) regarded *B. skagerrakensis* as an opportunistic species, signalling instability during periods of changing conditions. This interpretation was based on the fact that the species always occurred during transitional intervals in his records, for instance at the transition to the Little Ice Age. Since we only find one peak of *B. skagerrakensis* in our record, it is not possible to resolve whether this represents a short period of environmental change, or if it is rather a product of irregular occurrences of the species.

The assemblages in interval F indicate a decrease in salinity and a pronounced increase in the instability of the environment. This is e.g. indicated by an increase in *E. excavatum* (i) and *A. beccarii* (h). Unstable conditions are also reflected by high percentages of *S. fusiformis* (j), which additionally points to an increase in the organic carbon contents. The pronounced decrease in *U. mediterranea* (f), *M. barleeanus* (e), *G. turgida* (d) and *B. marginata* (a) can be linked with the change in grain-size to coarser material. These species are all associated to fine grained substrates (Murray, 1991).

7.5.2 Planktonic foraminifera

Globigerinita uvula totally dominates the planktonic assemblages with >80%, while *T. quinqueloba* is less common. Therefore, the *G. uvula* flux (Figure 7.5a) is used in this study as a measure for the total planktonic foraminiferal flux. The variation in flux of *G. uvula* ranges between 0 and 60 specimens/cm^2/year, with a single peak reaching 130 specimens/cm^2/year (Figure 7.5a). The maximum values occur in the intervals A, B and C, particularly between 2400-2100 cal yr BP. Minimum values are reached at around 1200 cal yr BP. Since then, the *G. uvula* flux has been close to zero with a slight peak at about 500 cal yr BP (within interval E).

The flux of *T. quinqueloba* (Figure 7.5b) is usually below 1 specimen/cm^2/year, with only two higher peaks between 1800-1700 cal yr BP (within interval C). These two major peaks as well as several other maxima in the *T. quinqueloba* flux correlate with high values of the *G. uvula* flux, indicating periods of increased influx of North Atlantic surface water masses to the area. As seen for *G. uvula*, the *T. quinqueloba* flux is very low after 1200 cal yr BP, and it became almost zero only within the last 200 years.

Studies of recent and subrecent foraminiferal distributions in the Skagerrak (Seidenkrantz, 1993; Bergsten et al., 1996) showed that planktonic foraminifera are generally very sparse in this area. However, the two planktonic foraminiferal species *G. uvula* and *T. quinqueloba* have been observed rather frequently in the outer parts of Norwegian fjords (Mikalsen, 1999; Husum and Hald, in press). By regarding the Skagerrak as a fjord-like system, one could argue that our results are in accordance with their observations. The percentage distribution map of Johannessen et al. (1994) also exhibits a branch of slightly increased values for *T. quinqueloba* off the Norwegian west coast.

7.6 Environmental changes and historical epochs

Major changes both in the benthic foraminiferal assemblages and in the benthic flux are found to occur at around 2200, 1900, 1500, 1200 and 400 cal yr BP in the Skagerrak core. The different intervals defined by the foraminiferal data correspond to well-known historical epochs such as the Subatlantic Pessimum, the temperate Roman Period and the cooler Migration Period, the Medieval Warm Period and the Little Ice Age. In general, these epochs are distinguished by differences in climatic conditions, mainly in temperature. Our results are also compared with surface temperature conditions, reconstructed by Mann and Jones (2003) for the last 1,700 years. Their combined proxy record is valid for the Northern Hemisphere (Figure 7.6).

In particular, the benthic foraminiferal flux appears to be highly sensitive to climatic variations, and a relationship between flux and $\delta^{13}C$ values has been demonstrated (Scheurle et al., 2004a). Therefore, the benthic foraminiferal flux, which varies between 0 and 90 specimens/cm^2/year through the record, will be discussed and linked to the Late Holocene climatic development below (Figure 7.6).

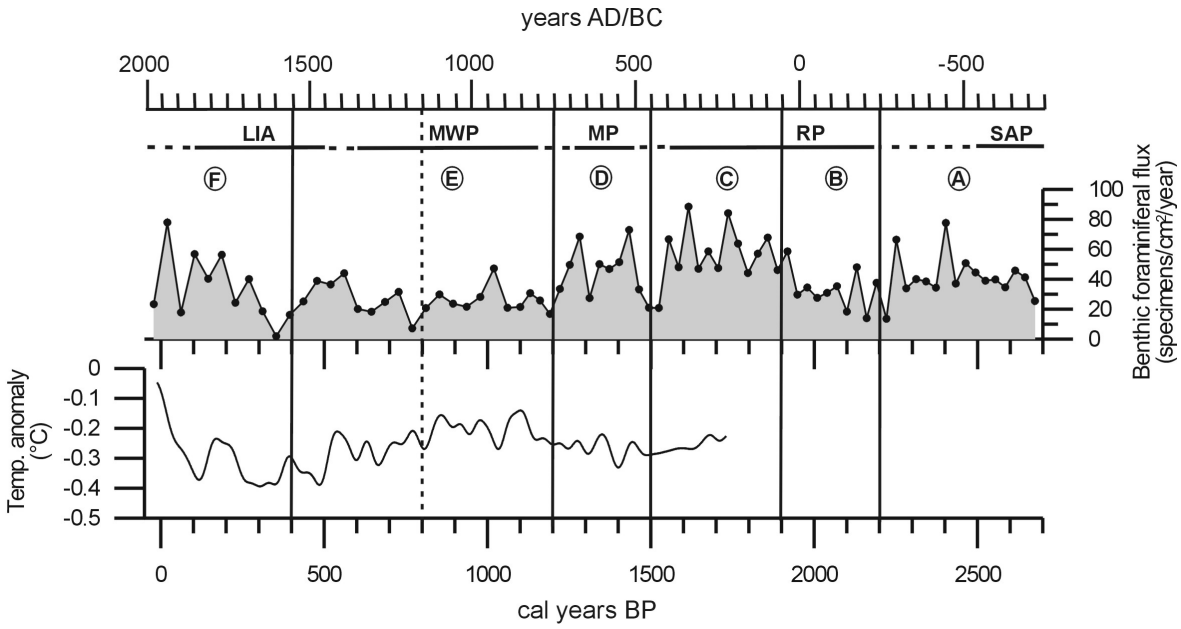

Figure 7.6: The benthic foraminiferal flux compared with the temperature anomaly curve of Mann and Jones (2003). The approximate timing of the documentary climatic history in north-western Europe is also shown. Abbreviations: SAP, Subatlantic Pessimum; RP, Roman Period; MP, Migration Period; MWP, Medieval Warm Period and the LIA, Little Ice Age (for intervals see Figure 7.3).

The time interval A (2690-2200 cal yr BP) - The Subatlantic Pessimum

During interval A, benthic foraminiferal data of the Skagerrak core indicate rather stable environmental conditions, changing to a higher variability at around 2500 cal yr BP. The benthic foraminiferal flux is relatively high (mean value 40 specimens/cm^2/year; Figure 7.6). This is associated with faunal indication of high input of organic matter to the area, increasing after 2500 cal yr BP. The entire period is characterised by a relatively strong influence of North Atlantic water masses, as reflected by maximum values of the planktonic foraminiferal flux (*G. uvula*; Figure 7.5a).

This period corresponds to the final part of the Subatlantic Pessimum in Europe, a relatively cold period with high precipitation rates (Lamb, 1977; Fairbridge, 1987). In general, the Subatlantic Pessimum is reported to end at approximately 2200 BP. Lozán (1998), however, placed the termination of this epoch at 2550 BP, followed by a transitional period. This latter timing of climatic change corresponds to a change in our faunal data.

The time interval B (2200-1900 cal yr BP) – Early part of the Roman Period

Pronounced instability in the benthic environment prevailed during interval B. Increased bottom current velocities are indicated by maximum percentages of *Cibicides lobatulus* (Figure 7.4). In particular, this interval can be distinguished from the previous one by its significantly lower benthic foraminiferal flux (mean value 33 specimens/cm^2/year). The planktonic flux still indicates a relatively strong influence of North Atlantic water to the Skagerrak, though less important than previously.

This interval is linked to the early part of the Roman Period, which is characterised by several cooling spells and storm events (Lamb, 1977; Lozán, 1998). In general, the Roman Period (2150-1550 BP; Lozán, 1998) is described as mild, but unstable (Lamb, 1977).

The time interval C (1900-1500 cal yr BP) – Later part of the Roman Period

The benthic environmental conditions were still highly variable, but with intervals of some stability. The benthic foraminiferal flux is at its maximum, pointing to high productivity (mean value 57 specimens/cm^2/year; Figure 7.6). This clearly distinguishes interval C from the previous one. A slightly increased influence of North Atlantic waters is reflected in the planktonic foraminiferal fluxes, particularly between 1850 and 1650 cal yr BP (Figure 7.5a, b).

This interval corresponds to the later part of the Roman Period. The temperature reconstruction from Mann and Jones (2003) shows a cooling trend towards the end of this time period (1700-1500

cal yr BP; Figure 7.6). In addition, both planktonic and benthic the stable oxygen isotopes from core GeoB 6003-2 signal decreasing temperatures through the entire interval (Scheurle et al., 2004a).

The time interval D (1500-1200 cal yr BP) – The Migration Period

The benthic environment became slightly more stable in interval D than in the previous one. A moderate decrease in the benthic foraminiferal flux (mean value 45 specimens/cm^2/year; Figure 7.6) indicates a slight change to lower productivity. The planktonic *G. uvula* flux (Figure 7.5a) decreases to very low numbers, and the *T. quinqueloba* flux (Figure 7.5b) remains low. The group of Miliolids (Figure 7.5c), however, points to a small increase in the advection of saline bottom waters to the area. A shift to a gradually higher variability in the environmental conditions towards the end of the interval is indicated by the benthic foraminifera.

From historical observations, the Migration Period (1550-1250 BP; Lozán, 1998) was characterised by a decrease in temperature and an increase in precipitation. This climate deterioration is suggested as one of the main initiation factors for the Germanic Migration. Relatively cool conditions are also indicated in the temperature anomaly record of Mann and Jones (2003) as well as in stable oxygen isotope data of the Skagerrak core (Scheurle et al., 2004a), both reflecting similar temperature conditions as during the previous interval C.

The time intervals E and F (1200 cal yr BP-present) - The Medieval Warm Period and the Little Ice Age

The benthic foraminiferal distribution points to increased instability in the environmental conditions during the early part of interval E (1200-800 cal yr BP). After a marked change at 800 cal yr BP, the assemblages show gradually higher instability. During interval E (1200-400 cal yr BP), the benthic foraminiferal flux is generally low (mean value 27 specimens/cm^2/year; Figure 7.6), with two distinct minima. Faunal indication of increased organic input to the area between 800 and 500 cal yr BP is not associated with increased foraminiferal flux values.

The planktonic foraminifera almost disappear (Figure 7.5a, b). This points to a noteworthy decrease in the influence of North Atlantic surface water masses. Fluctuations in the planktonic *T. quinqueloba* flux (Figure 7.5b) as well as in the percentages of benthic Miliolids (Figure 7.5c), however, still show some variations in the influx of Atlantic water masses. The change in planktonic productivity in the area at about 1200 cal yr BP appears to be closely associated with a general decrease in benthic productivity (Figures 7.5 and 7.6). A decrease in the advection of Atlantic water masses to the region was previously reported by Klitgaard-Kristensen et al. (2001) and by Mikalsen (1999).

A change in the hydrographical pattern through interval E becomes also evident from the grain-size data (Figures 7.2 and 7.4). A decreased sedimentation rate coincides with an increase in grain-size. The coarsening upward trend in the Skagerrak is most likely a result of increased coastal and bottom erosion in the North Sea together with an intensification of the erosion and transport capacity of the North Jutland Current (Jørgensen et al., 1981; Nordberg, 1991; Figure 7.1). Since the onset of the Medieval Warm Period, a 'mud transport belt' has controlled the horizontal sediment flux, which was transporting sediment from the west to the east (Hass, 1996). This was connected by Hass (1996) to periods with prevailing stormy atmospheric conditions. According to Rodhe (1987), the inflow of water into the Skagerrak is highly wind-driven.

Interval E corresponds to the so-called Medieval Warm Period and the earliest part of the Little Ice Age (i.a. Grove, 2002). Usually, the beginning of the Medieval Warm Period is associated with the Viking Period (1150-950 BP) and the Nordic settlement of Iceland and Greenland. Furthermore, winegrowing in north-western Europe is used as an indicator for an ameliorated climate during this time period (e.g. Schönwiese, 1988). In addition to the historical evidence, the proxy-based temperature reconstruction presented by Mann and Jones (2003) shows the highest temperatures from 1200-800 cal yr BP, followed by a cooling trend between 800-500 cal yr BP. The timing of that temperature change coincides with a significant change in the benthic foraminiferal composition in the Skagerrak.

At around 500 cal yr BP, the foraminiferal assemblages in the Skagerrak core changed markedly within a period of 200 years. A decrease in salinity is shown by diminishing percentages of the group of Miliolids (Figure 7.5c) and by an increase in the percentages of *A. beccarii* and *E. excavatum* (Figure 7.3h, i). These latter species contribute significantly to the final rise in the benthic foraminiferal flux (mean value 34 specimens/cm^2/year; Figure 7.6; interval F). The planktonic foraminiferal flux of *G. uvula* remains very low, and the flux of *T. quinqueloba* becomes almost zero during the last 200 years (Figure 7.5a, b). This shows that the North Atlantic surface water influence was even further reduced.

The environmental change around 500 cal yr BP can be linked to the climatic transition from the Medieval Warm Period, when climate altered (Flohn and Fantechi, 1984; Lamb, 1984) and deteriorated into the Little Ice Age. The Little Ice Age is generally considered to have lasted from about 600 to 100 cal yr BP in central and northern Europe. Due to changes in the atmospheric circulation system, the weather conditions became unstable, for example resulting in increased precipitation and storminess. The faunal indication of particularly high input of organic carbon to the sediments during the last couple of centuries may be due to anthropogenic influence in the area.

7.7 Summary and conclusion

This study contributes new knowledge about Late Holocene changes in the marine environment of the Skagerrak, which is part of the North Atlantic realm, by presenting benthic and planktonic foraminiferal analyses as well as grain-size distribution in core GeoB 6003-2. Due to the fact that there is a close connection between the atmospheric and the oceanic systems in the area, the foraminiferal indication of changes in bottom and in surface water characteristics, like e.g. stability and productivity, can be used for environmental and climatic reconstruction. During the last 2,700 years, major environmental shifts are recorded in the benthic faunal distribution as well as in the benthic foraminiferal fluxes at around 2200, 1900, 1500, 1200 and 400 cal yr BP, indicating climatic changes. The defined intervals can be linked to historical epochs such as the Subatlantic Pessimum, the Roman Period, the Migration Period, the Medieval Warm Period and the Little Ice Age.

Changes in the environmental stability at the seafloor, as indicated by benthic foraminiferal distributions and grain-sizes, are suggested to be at least partly controlled by changes in bottom current speeds. Unstable conditions may thus be related to periods of frequent storm events.

It can be concluded that the benthic foraminiferal flux, which is used as a measure of productivity in the area, is highly sensitive to climatic variations. During periods with relatively warm conditions, the benthic foraminiferal flux is generally low, while it is higher during cooler periods. It is assumed that more nutrients were available in the marine environment during relatively cool phases. This can be linked with historical evidence of increased precipitation and storminess.

A pronounced advection of North Atlantic surface waters to the Skagerrak prevailed until about 1200 cal yr BP. Subsequently, a marked decrease in the planktonic foraminiferal flux, which coincides with a general decrease in the benthic foraminiferal flux, shows that this inflow diminished. Most likely, this shift can be linked to a change in the large-scale atmospheric-oceanic circulation system.

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Appendix. List of the most important foraminiferal taxa

The original description of the foraminiferal taxa cited in the text is reported in Ellis and Messina (1949)

Benthic taxa:

Ammonia beccarii (Linné, 1758) = Nautilus beccarii, 1758

Bolivina skagerrakensis Qvale and Nigam, 1985

Bulimina marginata d'Orbigny, 1826

Cassidulina laevigata d'Orbigny, 1826

Cibicides lobatulus (Walker and Jacob) = Nautilus lobatulus Walker and Jacob, 1798

Elphidium excavatum (Terquem, 1875) = Polystomella excavata Terquem, 1876

Globobulimina turgida (Bailey, 1851) = Bulimina turgida Bailey, 1851

Hyalinea balthica (Schroeter, 1783) = Nautilus balthicus Schroeter, 1783

Melonis barleeanus (Williamson, 1858) = Nonionina barleeanum Williamson, 1858

Miliolids

Pullenia osloensis Feyling-Hanssen, 1954

Stainforthia fusiformis (Williamson, 1858) = Bulimina pupoides, var. fusiformis Williamsen, 1858

Uvigerina mediterranea Hofker, 1932

Planktonic species:

Globigerinita uvula (Ehrenberg, 1861) = Pylodexia uvula Ehrenberg, 1861

Turborotalita quinqueloba (Natland, 1938) = Globigerina quinqueloba Natland, 1938

8 Stable isotopes recording environmental conditions in the Skagerrak (North Sea) during the Late Holocene

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Abstract

On the basis of stable isotopes of planktonic and benthic foraminifera, we present a high-resolution reconstruction of marine environmental conditions and changes through the Late Holocene (last 2,700 years) in the eastern North Sea. Samples from core GeoB 6003-2 from the Skagerrak were analysed, and the results were compared with other proxy records.

The oxygen isotopes ($\delta^{18}O$) mainly reflect temperature changes in both the surface and bottom waters, whereas the carbon isotopes ($\delta^{13}C$) primarily signal the nutrient supply. The $\delta^{13}C$ data point to a significant effect of nutrient availability on the productivity (benthic foraminiferal flux). To some extent, the productivity in the area is linked to temperature and precipitation in central and north-western Europe.

Keywords: stable isotopes, $\delta^{18}O$, $\delta^{13}C$, paleoenvironment, Skagerrak, Late Holocene

8.1 Introduction

The Skagerrak is an important sediment accumulation area of the North Sea. Due to high sedimentation rates and a close coupling of atmospheric and oceanic circulation (Rodhe, 1987; van Weering et al., 1993), the Skagerrak bears high potential for environmental reconstructions. Useful information for such reconstructions is preserved in the sediments, e.g. in form of microfossils with a calcareous shell. The isotopic composition of these shells is closely related to environmental parameters such as temperature and/or salinity, and they can therefore be used as 'proxies' for these specific parameters. Observations and instrumental measurements e.g. of temperature and salinity are only available for the last decades or for a couple of centuries, respectively. Further back in time, stable isotopes can be used for the reconstruction of these environmental parameters.

Previous work on stable isotopes in the Skagerrak region is sparse. Isotopic results from the Younger Dryas and the Holocene from the area were presented by Erlenkeuser (1985), while Conradsen and Heier-Nielsen (1995), presented a Holocene record from northern Denmark. The time resolution for the Late Holocene part of these studies is, however, not high enough for a detailed correlation with our record. For the Late Holocene, stable isotope analyses from the Swedish west coast are mainly focusing on the rather specific conditions in fjord environments (Gustafsson and Nordberg, 2002).

High-resolution isotopic records from the Skagerrak were provided by Hass (1994, 1996). On the basis of a combined study of granulometric and $\delta^{18}O$ measurements, Hass (1996) presented detailed information on the temperature development and water mass changes in the Skagerrak within the last 1,500 years. He linked his paleoenvironmental reconstructions to historical epochs and interpreted them to reflect variations in the positions of the cyclonic tracks across the North Atlantic. Furthermore, variations in $\delta^{18}O$ through the last 900 years were reported by Senneset (2002) from the eastern Skagerrak. Comparison with our core was problematic, presumably due to the location of these cores further to the east and due to differences in water depth and temporal resolution.

The purpose of this study is to reconstruct environmental and climatic changes through the last 2,700 years of the Late Holocene in the central Skagerrak at high temporal resolution. A previous study from the same site was focusing on foraminiferal assemblages and fluxes as well as grain-size data (Scheurle et al., 2004b). The stable isotope results of the present study supplement the environmental interpretations of that investigation.

8.2 Regional setting

Sediment core GeoB 6003-2 was obtained from the central Skagerrak basin at the location 57°58.3'N and 9°23.2'E at a water depth of 312 m (Figure 8.1). The asymmetric Skagerrak basin, with a maximum depth of about 700 m, serves as a natural sediment trap receiving input from a number of north-west European source areas and from remobilization of North Sea sediments. The waters in the North Sea, including the Skagerrak, basically consist of two components: high salinity water from the North Atlantic and fresh water derived from precipitation and rivers. Almost 90% of the inflowing waters are of North Atlantic origin, flowing through the northern North Sea and along the southern flank of the Skagerrak (via the Tampen Bank Current and the Southern Trench Current; e.g. Nordberg, 1991; Figure 8.1). Waters from the central and southern North Sea (~9%), and from the Baltic Sea (1%) comprise the remaining waters entering the Skagerrak.

The Skagerrak current system forms an anticlockwise circulation pattern, which is influenced by the large-scale atmospheric and oceanic circulation system (Rodhe, 1987). Both, the surface and bottom currents in the Skagerrak generally follow this pattern, but the bottom current velocity is lower than that of the surface waters (i.a. Rodhe, 1987). Particularly the surface currents, but also the bottom currents in the North Sea and the Skagerrak, are affected by winds shifting in strength and direction.

Climatic variability in Europe depends upon the interaction between zonal versus meridional air circulation (Bryant, 1997). The former brings warmer temperatures, heavier rainfall and storm conditions to northern Europe, and more stable and dry weather conditions in the south. Meridional flow is conditioned by blocking of high pressure cells over the eastern North Atlantic or north-western Europe. In this situation, depressions tend to track north-south across Europe. Beneath the high pressure, climatic conditions are dry and rather stable. However, deflection of the low pressure tracks brings storm conditions beneath its path.

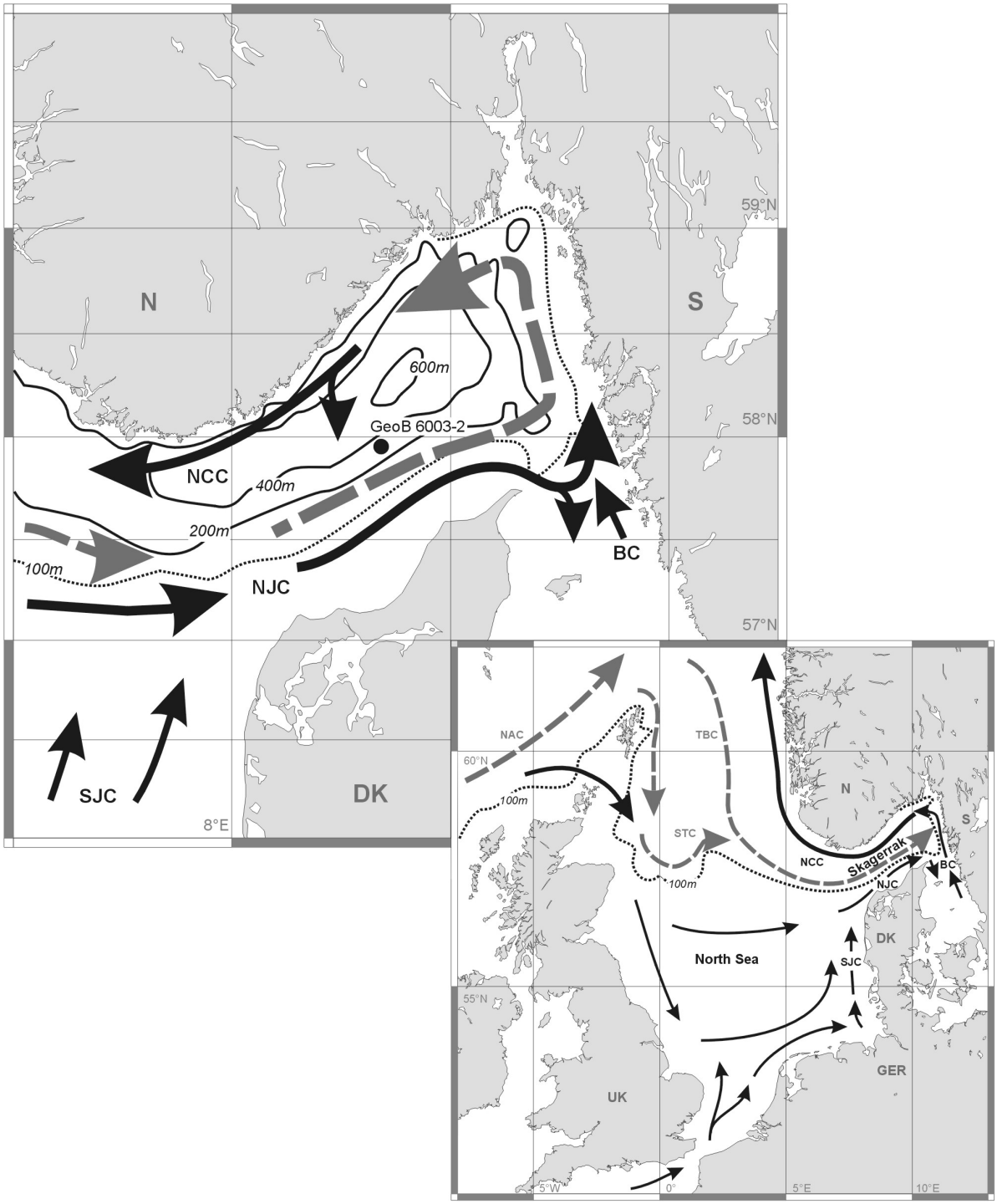

Figure 8.1: Location of the sediment core GeoB 6003-2 and the present-day surface circulation pattern in the Skagerrak and the North Sea (insert map). Partly, North Atlantic water flows directly into the Skagerrak (grey-stippled arrows), modified waters flow into the area via the southern North Sea or comes from the Baltic (black arrows). BC, Baltic Current; DK, Denmark; GER, Germany; N, Norway; NAC, North Atlantic Current; NCC, Norwegian Coastal Current; NJC, North Jutland Current; S, Sweden; SJC, South Jutland Current; STC, Southern Trench Current; TBC, Tampen Bank Current; UK, United Kingdom. Modified from Danielssen et al. (1991) and Nordberg (1991).

8.3 Material and methods

This study was carried out on sediment core GeoB 6003-2, which was retrieved in 1999 during Meteor cruise M45-5 in the Skagerrak (Figure 8.1). The upper 400 cm of the core were sampled at five centimetre intervals. The sediment samples (volume: ~10 ml) for foraminiferal and stable isotope analyses were wet sieved (63-150 µm). Subsequently, the samples were dry sieved on a 100 µm sieve, and foraminifera in the size fraction >100 µm were concentrated using the heavy liquid CCl_4 (ρ=1.59 g/cm^3) (Meldgaard and Knudsen, 1979). Calcareous benthic foraminifera, which were generally well preserved, form the major part of the fauna (Scheurle et al, 2004b). The benthic species *Melonis barleeanus* (Williamson) was chosen for the stable isotope measurements, because enough material was available in most of the samples, and because it has shown the most consistent stable oxygen isotope results compared to other benthic foraminifera in the region (Senneset, 2002). The planktonic fauna consists of small-sized specimens of only a few species (Scheurle et al., 2004b), and the most common species *Globigerinita uvula* (Ehrenberg) was used for the isotopic measurements.

The $\delta^{18}O$ and $\delta^{13}C$ compositions of the benthic and the planktonic foraminifera were analysed with a Finnigan MAT 251 mass spectrometer. About 10-20 individual shells of the benthic species *Melonis barleeanus* were used for each isotopic measurement, whereas an amount of approximately 150 individuals of the small planktonic foraminifera *Globigerinita uvula* was necessary for each sample. The isotopic composition of the carbonate samples was measured on the CO_2 gas evolved by the treatment with phosphoric acid at a constant temperature of 75°C. The stable isotope measurements were calibrated against Vienna Pee Dee Belemnite (VPDB) by applying the NIST 19 standard. Analytical standard deviation is about ±0.07‰ VPDB for $\delta^{18}O$ and ±0.05‰ VPDB for $\delta^{13}C$ (Isotope Laboratory Bremen University).

Benthic isotope measurements were carried out for the entire record, but with three data points missing in the upper part. Planktonic isotope measurements were determined for the interval between 398 and 58 cm, with only a few data points missing due to lack of enough material. It was not possible to obtain enough planktonic material for isotope measurements in the upper 58 cm.

The age-depth model of core GeoB 6003-2 was based on AMS ^{14}C datings and ^{210}Pb data, taking changes in grain-size into account (Scheurle et al., 2004b). The average sedimentation rate changes from 1.20 mm/yr in the upper 138 cm (1100 years) to 1.65 mm/yr between 138 cm and 400 cm (1100 to 2690 cal. yrs BP). The temporal resolution in this study is generally about 30 to 40 years.

For the interpretation of the data, it is important to be aware of possible reworking of benthic foraminiferal shells that may have affected the upper c. 200 cm of the core (Scheurle et al., 2004b).

Even though some ^{14}C datings appear to be influenced by reworked material, no major disturbances were clearly visible, neither in the sedimentary record nor in the foraminiferal and isotope data.

8.4 Results

8.4.1 Stable oxygen isotopes

The benthic $\delta^{18}O$ values range between 1.6 and 2.1‰ (Figure 8.2). The general pattern displays two different levels, one interval with relatively low values from 2690 to 1500 cal. yrs BP, succeeded by an interval with slightly higher values from 1500 to 500 cal. yrs BP. The transition into this latter interval seems to have happened within 400 years (1900 to 1500 cal. yrs BP), but the difference in mean values between the two intervals is not significant. There is much variation in the benthic $\delta^{18}O$ values in the upper part of the record (500 cal. yrs BP to present).

The amplitude of the planktonic $\delta^{18}O$ values is 1.1‰, with only one point out of range. The total range is 1.5‰ (between -0.67 and 0.78‰; Figure 8.2). Relatively low planktonic $\delta^{18}O$ values are observed between 2690 and 1900 cal. yrs BP. Within the following c. 300 years, the values become gradually higher with a maximum level between 1650 and 1350 cal. yrs BP. Lower values prevail from 1350 to 500 cal. yrs BP.

The benthic and planktonic $\delta^{18}O$ curves generally follow the same pattern through the record (Figure 8.2). Relatively low $\delta^{18}O$ values between 2690 and 1900 cal. yrs BP, and a subsequent change to higher $\delta^{18}O$ values between 1900 and 1500 cal. yrs BP are found in both records. There is a more pronounced variation in the planktonic than in the benthic $\delta^{18}O$ signal. Particularly increased values, as recorded for the planktonic isotopes between 1650 and 1350 cal. yrs BP, are not found for the benthic isotopes. After 1350 cal. yrs BP, the records continue to follow similar trends until 500 cal. yrs BP.

8.4.2 Stable carbon isotopes

The benthic $\delta^{13}C$ values range from 0.2 to –0.9‰, while the planktonic values vary between -0.3 and –1.5‰. A comparison of the general pattern of the planktonic and benthic carbon isotope curves (Figure 8.3) shows that there has been a close connection between surface and bottom waters, at least until about 500 cal. yrs BP, i.e. the uppermost measured level for the planktonic foraminifera. An interval in which the values tend to show different trends is only noticeable between 1900 and 1500 cal. yrs BP.

8.5 Discussion

8.5.1 Stable oxygen isotopes

At water depths of 250 m and deeper in the Skagerrak, the salinity is around 35‰ and shows little variation (Lee, 1980). Due to this relatively constant salinity, the dominant effect on changes in the $\delta^{18}O$ values of benthic foraminifera can be ascribed to temperature variations (Erlenkeuser, 1985; Hass, 1996; Senneset, 2002). Temperatures are also relatively stable at these depths, with a seasonal variability between 4 and 7°C (Larsson and Rodhe, 1979). Since foraminifera are supposed to grow mainly during summers, their isotopic signal presumably reflects summer temperatures.

According to Shackleton (1974), a change of 4.0 K in sea water per 1‰ $\delta^{18}O$ is suggested for temperatures below 16.9°C. By applying this relation, the range of the benthic $\delta^{18}O$ values of *Melonis barleeanus*, a species reflecting bottom water mass conditions, corresponds to an amplitude of 2.0 K in temperature. This temperature range is similar to that reported from a west Norwegian fjord environment during the same time period (Mikalsen, 1999).

In the upper part of the record there is much variation in the benthic $\delta^{18}O$ values (Figure 8.2). A distinct increase between 500 and 300 cal. yrs BP is succeeded by decreasing values since around 300 cal. yrs BP. The general trend towards lower values corresponds closely to that documented previously in a high-resolution core further to the east (Hass, 1994). Since there is foraminiferal indication of decreasing salinities through the uppermost part of the record (Scheurle et al., 2004b), this increase in benthic $\delta^{18}O$ may be due to salinity changes. However, a temperature anomaly reconstruction of Mann and Jones (2003) shows a clear temperature increase during this most recent interval, corresponding to the pattern displayed by the benthic $\delta^{18}O$ values. It is, therefore, not evident whether the benthic $\delta^{18}O$ in this part of the record is mainly influenced by temperature and/or salinity. With the exception of this large amplitude during the last centuries, variations in $\delta^{18}O$ are too small for a meaningful calculation of salinity variations.

The planktonic $\delta^{18}O$ record of *Globigerinita uvula* reflects the isotopic composition of surface water masses. On the basis of its carbon isotope values, it is concluded that this species lived and reproduced in the Skagerrak (see below). This finding is contradictory to the previous assumption that the species is transported into the area by the (varying) advection of North Atlantic water masses, an assumption which is based on information provided e.g. by Seidenkrantz (1993) and Bergsten et al. (1996). If this foraminiferal species had calcified in the North Atlantic waters and subsequently was transported via the currents into the Skagerrak area, it would represent open ocean conditions, and the planktonic $\delta^{18}O$ would mainly reflect temperature variations. Living in

the Skagerrak, however, the $\delta^{18}O$ values of *Globigerinita uvula* may also be influenced by variability in freshwater discharge affecting the water masses, seasonally and annually. Though, a relatively strong dependence of the planktonic $\delta^{18}O$ on variations in temperature is still expected. Due to a possible seasonal occurrence of *Globigerinita uvula* in fall, as observed for subarctic regions (Boltovskoy and Wright, 1976), however, the planktonic $\delta^{18}O$ data might bear a seasonal imprint.

Assuming a constant salinity through time, the amplitude of the planktonic $\delta^{18}O$ values of 1.1‰ corresponds to a maximum temperature variation of 4.4 K (excluding the outlier). There is a difference of 0.34‰ between the isotopic mean values for the intervals 2690-1900 and 1650-1350 cal. yrs BP. This corresponds to a temperature difference in the surface water masses of c. 1.4 K.

A maximum temperature difference in the interval 1350 and 800 cal. yrs BP, indicated by the difference between the benthic and the planktonic oxygen isotope values, appears to be due to a period of relatively high surface water temperatures, while the increased differences between 1650 and 1350 cal. yrs BP are presumably a result of relatively low surface water temperatures (Figure 8.2).

The oxygen isotope data were compared with the temperature anomaly curve for the last 1,700 years presented by Mann and Jones (2003), which is valid for the Northern Hemisphere (Figure 8.2). For this curve, the authors combined temperature reconstructions of 23 individual proxy records, most of them with high temporal resolution. The compared records display similar patterns for the overlapping time period. The similarity between the temperature anomaly record and our isotope record, in particular the benthic isotope record, may be regarded as a support of our age-depth model, which caused some difficulties due to possible reworking in the area (Scheurle et al., 2004b). Even though, there might be a lag due to a response time between the reconstruction based predominantly on terrestrial data and the marine record, the differences between our $\delta^{18}O$ record and the temperature data set of Mann and Jones (2003) amount to only about 30 years, a value which is close to the time resolution of our data points.

In summary, the benthic and the planktonic $\delta^{18}O$ values mainly reflect temporal changes in temperatures of the bottom and surface waters. However, salinity effects may have had some influence on the $\delta^{18}O$ values, particularly during the last few centuries (Scheurle et al., 2004b). Especially the benthic $\delta^{18}O$ record is in close correspondence to a proxy record reflecting temperature anomalies (Mann and Jones, 2003).

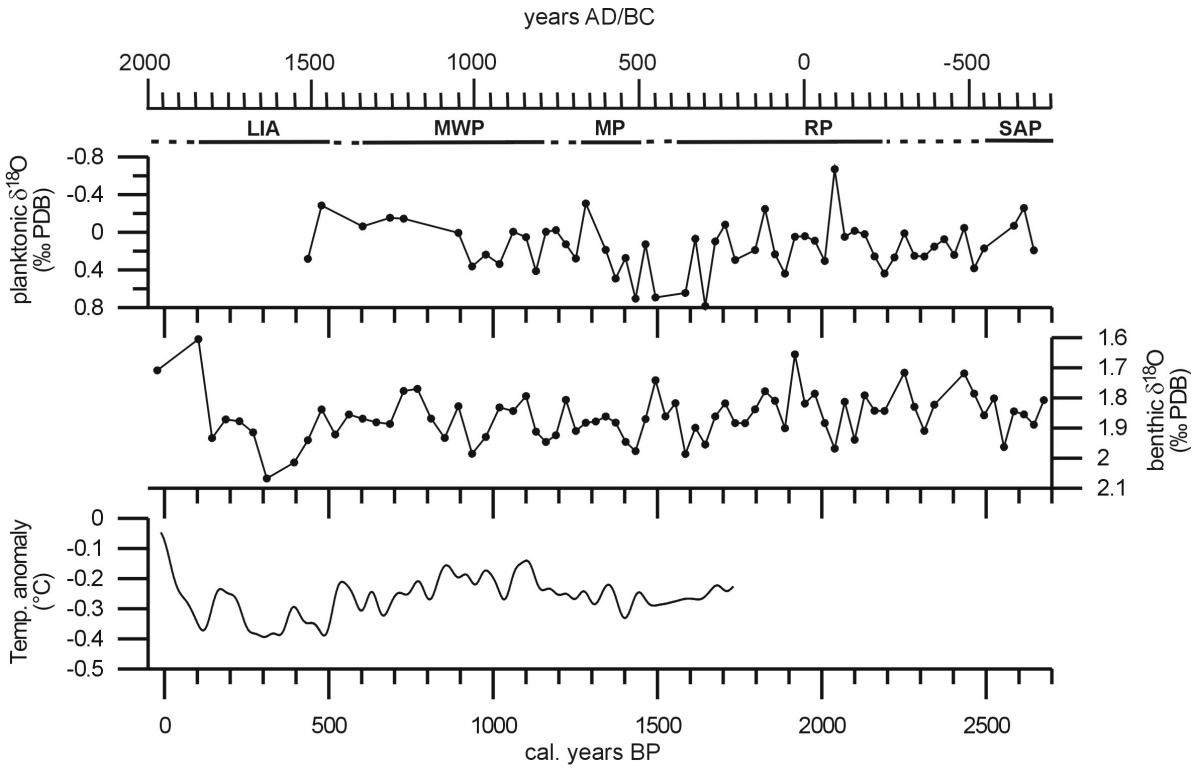

Figure 8.2: Comparison of the planktonic and benthic oxygen isotope records with a reconstruction of the Northern Hemisphere temperature anomaly (Mann and Jones, 2003). SAP, Subatlantic Pessimum; RP, Roman Period; MP, Migration Period; MWP, Medieval Warm Period; LIA, Little Ice Age.

8.5.2 Stable carbon isotopes

The infaunal benthic species *Melonis barleeanus* prefers a shallow subsurface habitat during calcification of its tests (Corliss and van Weering, 1993; Mackensen et al., 2000). The $\delta^{13}C$ values recorded from this species may be influenced by the pore water, representing modifications in carbon content compared to the overlying bottom water (Bickert, 2000). Decomposition of organic material and calcite dissolution, which are driven by the rain of organic matter to the seafloor, lower the stable carbon isotope ratio of pore waters and lead to a $\delta^{13}C$ gradient in the uppermost few centimetres below the sediment-water interface (McCorkle et al., 1985). Thus, the tests of an infaunal species, calcifying within the pore water, should reflect depleted $\delta^{13}C$ values. Indeed, negative deviations were observed e.g. by Woodruff et al. (1980), who analysed several benthic species. Such habitat effects of foraminifera are accompanied by vital effects (e.g. McCorkle et al., 1990; Spero, 1992; Mackensen et al., 2000).

An upward movement of the living foraminifera (*Melonis barleeanus*) to the sediment-water interface at different times of the year has, however, also been observed (Linke and Lutze, 1993).

In case they calcify there, bottom water conditions would be reflected in the benthic $\delta^{13}C$ signal. However, Mackensen et al. (1993) discussed possible processes of an influence on the $\delta^{13}C$ signal by the decay of organic matter, even of foraminiferal species living on the sediment surface (epifaunal).

Results of recent studies have shown that the sediment depth at which living benthic foraminifera are found, cannot be used as an indicator for the depth where they calcified (Scourse, pers. comm. 2003). The foraminifera are often transported downwards in the sediment by bioturbation after calcification, and there they presumably die. The calcification depth of several infaunal species has thus proved to be relatively shallow, i.e. within the upper 1 cm of the sediment (Scourse, pers. comm. 2003).

The benthic $\delta^{13}C$ values of *Melonis barleeanus* in the present material appear to show that this species has been shallow infaunal during calcification. If it had calcified on the sediment surface, the carbon isotope values would possibly have been positive and not slightly negative (Figure 8.3).

Carbon isotope values of recent surface water masses of the North Atlantic Current are relatively high, whereas the $\delta^{13}C$ values of the Norwegian Coastal Current (relatively low in oxygen and nutrient-rich) are extremely low (Weinelt, 1993; Simstich, 1999). The relatively low planktonic $\delta^{13}C$ values for *Globigerinita uvula* (Figure 8.3) indicate that the specimens cannot have been advected into the area from the Atlantic Ocean. They actually lived and reproduced in the area. Additionally, it can be concluded that the shell formation did not occur in the surface waters, but deeper in the water column, i.e. in the zone of organic decomposition. If these foraminifera had calcified in the surface waters, which are enriched in ^{13}C, the values of the $^{13}C/^{12}C$ ratio most likely would have been more positive.

The similar patterns of the planktonic and benthic carbon isotope curves (Figure 8.3) show that there has not been any pronounced stratification of the water column through most of the studied interval. However, while the benthic $\delta^{13}C$ values do not change, the planktonic $\delta^{13}C$ values show a continued trend towards lower values from 2200 to 1900 cal. yrs BP (Figure 8.3). As a result, the difference between the bottom and surface water values becomes slightly larger. This may be considered as an indication of gradually increasing stratification in the water column. Between 1900 and 1500 cal. yrs BP, the data tend to show different trends (Figure 8.3) and the differences reach maximum values. The benthic $\delta^{13}C$ values exhibit a change to extremely low values at around 1600 cal. yrs BP, succeeded by a rapid rise. In contrast, the planktonic $\delta^{13}C$ values continue at a stable level, only with two outliers. This presumably indicates a period of temporal stratification of the water column and a difference in the paleonutrient content of the two water masses.

The planktonic and the benthic carbon isotope data, reflecting the paleonutrient contents in the surface and bottom water masses, respectively, are compared with the benthic foraminiferal flux values, which have also been used as a measure of productivity (Scheurle et al., 2004b). A decreased benthic foraminiferal flux, and thus relatively lower productivity is accompanied with relatively high $\delta^{13}C$ values (Figure 8.3). Whereas, a faunal indication of higher productivity coincides with a depletion in $\delta^{13}C$. Such a depletion is generally connected with indication of benthic foraminiferal data of a high input of organic matter (Scheurle et al., 2004b).

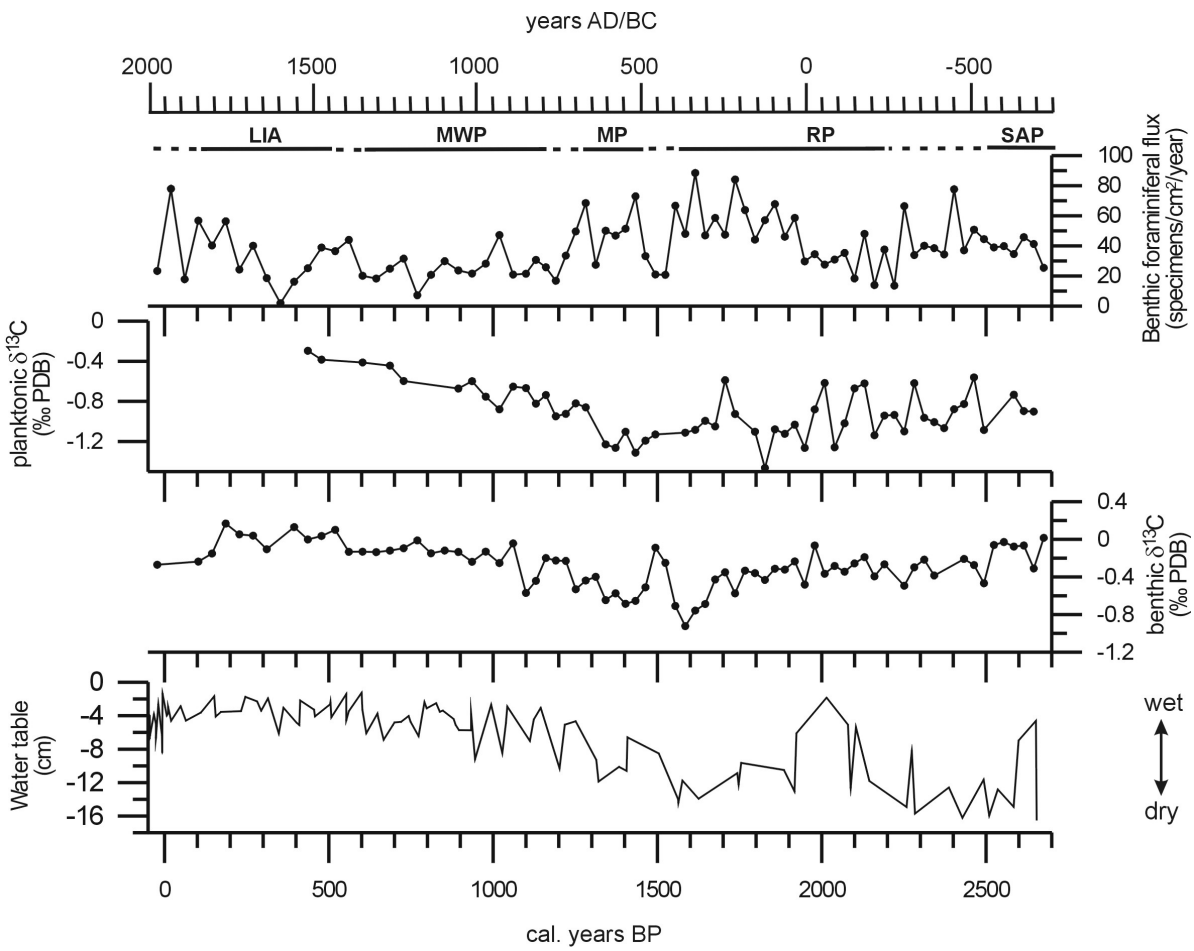

Figure 8.3: The planktonic and benthic carbon isotope records are compared with the total benthic foraminiferal flux (Scheurle et al. 2004b), and the reconstructed soil water table of a peatland, north-western Europe (Charman and Hendon, 2000). SAP, Subatlantic Pessimum; RP, Roman Period; MP, Migration Period; MWP, Medieval Warm Period; LIA, Little Ice Age.

There is a high degree of correlation between the 5-point averages of the planktonic and benthic carbon isotope values and the benthic foraminiferal flux (planktonic r^2 = -0.65 and benthic r^2 = -0.66; Figure 8.4). Five-point averages were used, because of a pronounced scatter in the original values.

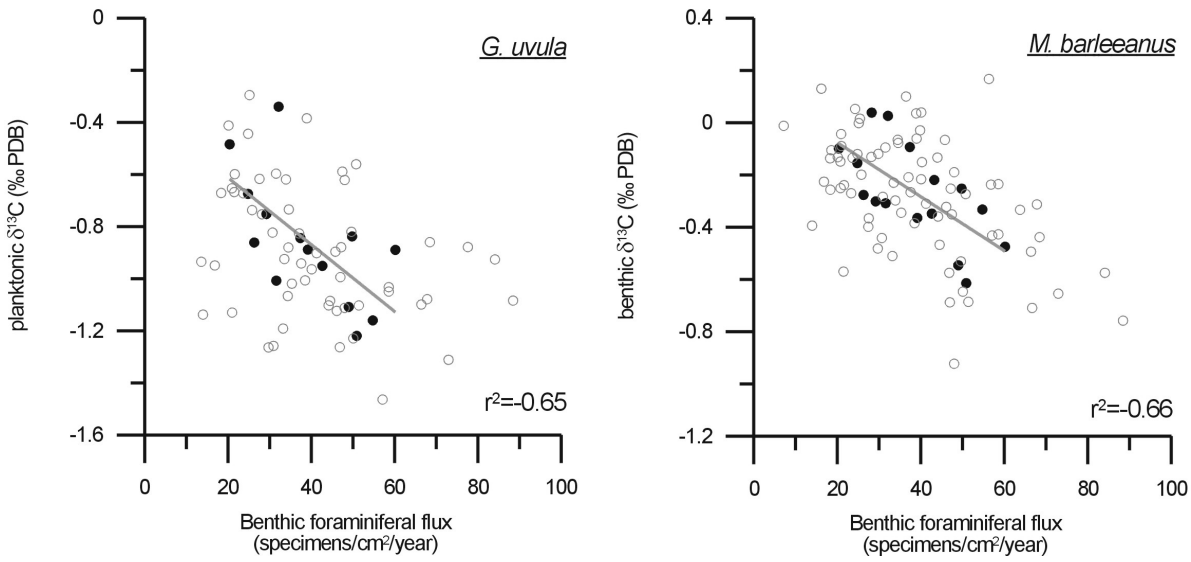

Figure 8.4: Comparison of the planktonic and benthic carbon isotope values and the benthic foraminiferal flux (grey open circles) and a correlation between their 5-point averages (black dots; planktonic r^2 = -0.65 and benthic r^2 = -0.66). Benthic foraminiferal flux data were obtained from Scheurle et al. (2004b).

In summary, a comparison of the benthic and planktonic $\delta^{13}C$ values shows that there has been a close connection between the surface and bottom waters, only with a short period of stratification in the water column between 1900 and 1500 cal. yrs BP. Furthermore, there are clear indications that nutrient availability at least partly controls the productivity. Possible problems with the use of the infaunal benthic species *Melonis barleeanus* are discussed, as well as the living and calcification depth of the planktonic species *Globigerinita uvula* in the water column of the Skagerrak.

8.6 The atmospheric circulation system and marine environments in the Skagerrak

Changes in the foraminiferal assemblages and in grain-size distribution of sediment core GeoB 6003-2 have been used for the reconstruction of changes in the marine environment of the Skagerrak through the last 2,700 years (Scheurle et al., 2004b). Major shifts occurred around 2200,

1900, 1500, 1200 and 400 cal. yrs BP. The resulting intervals were linked to historical epochs such as the Subatlantic Pessimum, the temperate Roman Period, the cooler Migration Period, the Medieval Warm Period and the Little Ice Age. In general, these time periods are mainly distinguished by the prevailing temperature conditions.

Our $\delta^{18}O$ data from the Skagerrak clearly suggest temperature changes both in the bottom and surface waters during the last 2,700 years (Figure 8.2). The assumption of mainly temperature-driven $\delta^{18}O$ values is supported by the similarity between the Skagerrak $\delta^{18}O$ record and a surface temperature record for the Northern Hemisphere (Mann and Jones, 2003; Figure 8.2).

The benthic foraminiferal flux in core GeoB 6003-2 (Figure 8.3) has been shown to be highly sensitive to climatic variations (Scheurle et al., 2004b). During periods with warm conditions, the benthic foraminiferal flux has been relatively low (i.e. low productivity), while it was higher (i.e. high productivity) during cooler conditions. However, this link is presumably not directly connected to changes in temperature, but rather to other environmental factors such as the input of nutrients into the marine area. The amount of nutrients is also reflected in the $\delta^{13}C$ data (Figures 8.3 and 8.4).

Historical sources relate the cooler phases Subatlantic Pessimum, the late part of the Roman Period, the Migration Period and the Little Ice Age to phases with increased precipitation (see e.g. compilation in Scheurle et al., 2004b). During these phases, more nutrients may have been dissolved and transported via rivers from source areas around the North Sea into the Skagerrak.

In order to obtain a more differentiated understanding of the environmental history of the area through the last 2,700 years, we have compared our data with a soil water table reconstruction of a peatland in the UK, north-western Europe (Charman and Hendon, 2000; Figure 8.3). The surface wetness of a peatland is thought to be a result of a combination of precipitation and temperature. The variability in surface wetness of a north-European peatland is primarily related to summer climate, because peatlands are either saturated or frozen in winter months. Charman and Hendon (2000) associated high water tables with wet summers and severe winters and low water tables with dry summers and mild winters. Indeed, a recent study of Charman et al. (2004) concludes that the surface wetness of a north-European peatland site can be interpreted as primarily reflecting summer precipitation variability, and that also a summer temperature signal may be present.

There appears to be a close correspondence between the supra-decadal variability in the soil water table of Charman and Hendon (2000) and the $\delta^{13}C$ record of the Skagerrak core (Figure 8.3). Since foraminifera generally reproduce and grow during the summer half-year, the carbon isotopes are also supposed to reflect summer conditions, including the nutrient supply. Additionally, there is an

inverse relationship between the soil water table record and the benthic foraminiferal flux record (Scheurle et al., 2004b; Figure 8.3). This productivity indicator is i.a. driven by the amount of nutrients, which is to some extent also determined by temperature.

From this comparison with our data set, we can address the following questions, (1) what could be the background for less nutrients and a low productivity during precipitation-rich intervals, and more nutrients and a higher productivity, when the conditions were relatively drier, and (2) why do historical sources give opposite indications, i.e. pronounced precipitation during the Subatlantic Pessimum, the late part of the Roman Period and the Migration Period.

Regarding the European paleoprecipitation pattern, the position of the atmospheric frontal zone may have played an important role (Lamb, 1969, 1977; Hass, 1994, 1996; Figure 8.5). During cold phases the temperature differences between north and south are intensified, the atmospheric frontal zone and the westerlies are strengthened and shifted to the south. This implies that the track of the cyclones is moved south, with the ’angle‘ situated north-west to Great Britain/Iceland. It is therefore hypothesized that less precipitation occurred during these cold phases in north-European areas, whereas more precipitation occurred in the continental catchment areas of central Europe (meridional type). The possibility to dissolve and transport nutrients from the large catchment areas of central Europe into the North Sea may have been favoured by this type of precipitation pattern. Such a shift in the atmospheric system may have changed on one hand the total amount of precipitation, and on the other hand the seasonal pattern.

Besides the shift in the position of the cyclonic track, also its direction may change towards north-west. Therefore, the North Atlantic influence on the Skagerrak may be intensified during cooler phases. In this context, it should be noticed that there is a clear foraminiferal indication (planktonic foraminiferal flux) that the North Atlantic influence in the Skagerrak was actually more intense during relatively cool intervals (Scheurle et al., 2004b). A change in the number of Rossby waves, which guarantee an intense exchange of energy and/or a change in the position of their ridges and troughs as well as varying amplitudes, may have led to additional climatic and environmental consequences.

During warm periods the (meridional) temperature differences are less intense, resulting in weakened cyclonic activity and westerlies. The atmospheric frontal zone is shifted to the north, and the direction of cyclonic tracks changed to west–east (zonal type). Such a change in direction may have caused less North Atlantic influence, as recorded by changes in the planktonic foraminiferal flux at our core site since the onset of the Medieval Warm Period (since 1200 cal. yrs BP; Scheurle et al., 2004b).

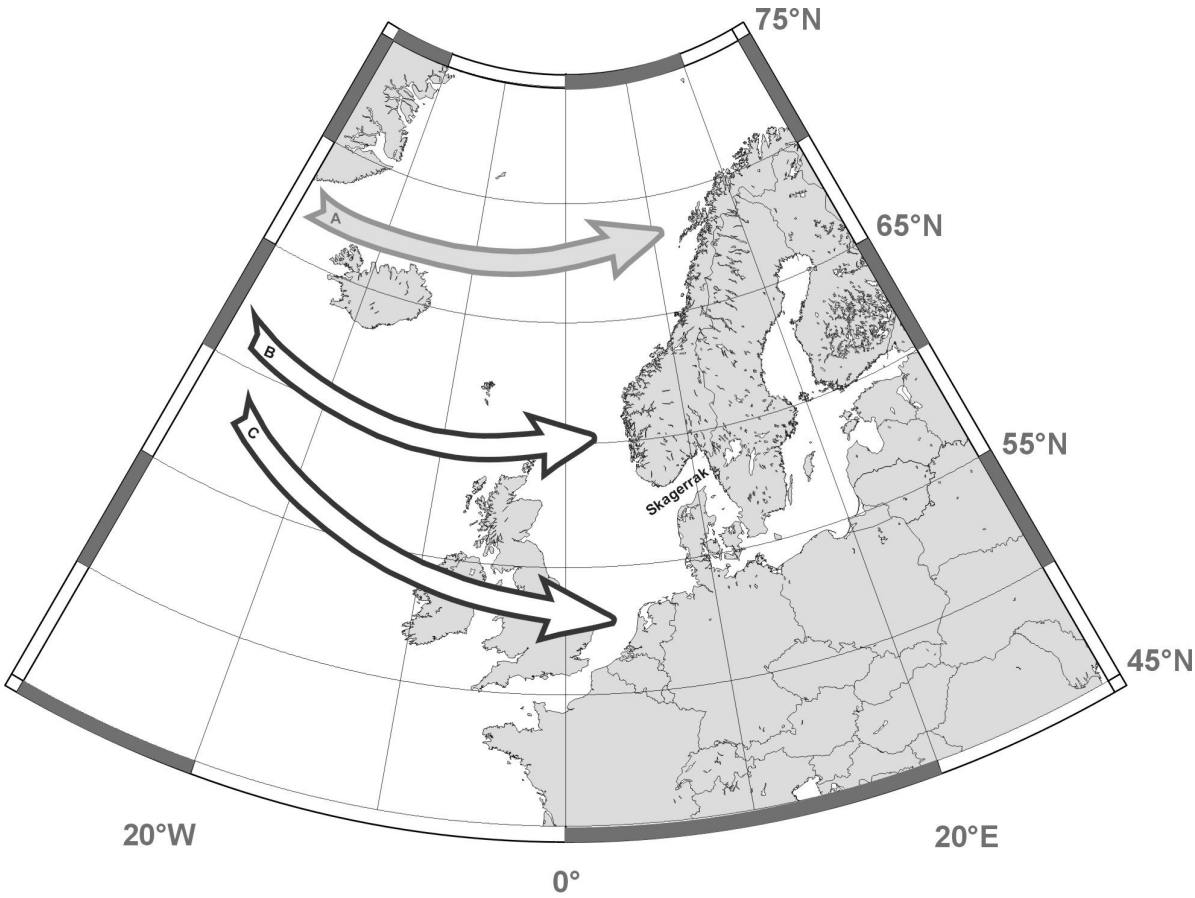

Figure 8.5: Sketch of the north-southward shift of the cyclonic track. The situation during warm phases is shown with an A. Deflection towards south during cold phases according to Lamb (1969) is indicated by a B, while the situation during cold events according to Hass (1994, 1995, 1996) is marked with a C.

Concerning the comparison of our data sets with the soil water table reconstruction, it has to be kept in mind that they represent site-specific conditions, i.e. even though the conditions most likely result from broader scale influences, they may display rather local signals only. Hendon et al. (2001), however, analysed three mires within a restricted area of northern Europe. The similarity between reconstructed soil water tables shows that the main changes identified can be related to climatic shifts. The authors stated that the magnitude of change between the three sites is more variable in the earlier part of the record than in the later part. This is probably partly due to problems with temporal resolution and age control, but, moreover, non-climatic factors are thought to have affected the peatland sites at that time.

However, there is a mismatch between the soil water table (Charman and Hendon, 2000) and the temperature (Mann and Jones, 2003) reconstruction during the Little Ice Age. The reasoning would suggest that the surface wetness should decrease during this cold period. Due to a southward shift of the cyclonic track, less low-pressure areas bringing precipitation should reach the north-

European peatland site. According to Lamb (1969), the main cyclonic tracks were positioned between 57 and 60°N, at times rather south of 57°N with a northward direction on its course, during the Little Ice Age, and not further south as indicated by the data presented by Hass (1995) (Figure 8.5, B and C). In this case, high precipitation could still have influenced the peatland site in the UK. Moreover, a change in the prevailing Rossby wave number and, therefore, wave length may have taken place (Lamb 1969). Furthermore, it may be argued that the observed wet conditions could either be a result of seasonal effects or be due to the high variability of climatic effects on a regional scale during this time interval with its rather specific conditions.

In summary, we conclude that the productivity in the Skagerrak area is partly responding to the nutrient availability, which in turn depends on the summer precipitation rates in central European cathments. Moreover, changes in the atmospheric circulation system, i.e. changes in the location of the frontal system and the cyclonic activity, as well as their direction, are suggested to play an important role, not only for the precipitation pattern in central and north-western Europe but also for the oceanic circulation pattern in the North Sea area. As previously mentioned, Hass (1996) came to a similar conclusion concerning the cyclonic activity and tracks, but on the basis on a different kind of data (i.e. on granulometric and $\delta^{18}O$ measurements).

8.7 Summary

This study presents planktonic and benthic stable isotope data, which were derived from a sediment core from the Skagerrak, the most important accumulation area of the North Sea region. The stable isotopes are recording environmental variations for the surface and bottom water masses during the Late Holocene.

Both, the planktonic and the benthic oxygen isotopes ($\delta^{18}O$) mainly display changes in temperatures during the last 2,700 years. The planktonic and the benthic $\delta^{18}O$ records generally show similar patterns, but with relatively more variation in the planktonic $\delta^{18}O$ values than in the benthic values. Especially the benthic $\delta^{18}O$ changes of the Skagerrak sequence correspond to the reconstructed temperature anomaly curve of Mann and Jones (2003).

The $\delta^{13}C$ values give indication of the habitat of the foraminiferal species *Globigerinita uvula* and *Melonis barleeanus* in the Skagerrak. A comparison of the planktonic and benthic stable carbon isotopes ($\delta^{13}C$) shows a generally mixed water column, but points to a period with stratification between 1900 and 1500 cal. yrs BP. The $\delta^{13}C$ record reflects changes in the amount of nutrients in the Skagerrak area. An increased productivity, as indicated by an increased benthic foraminiferal flux, is partly a result of high nutrient availability (low $\delta^{13}C$ values). Highly productive phases have

previous been linked to relatively cool climatic conditions (Scheurle et al., 2004b). Historical observations give indications that cool intervals, as for example the Migration Period, are also characterised by an increased precipitation. Therefore, it is assumed that more nutrients could be dissolved and transported via the rivers from source areas around the North Sea into the Skagerrak during such cool phases. In contrast to this, however, a soil water table record from the UK (Charman and Hendon, 2000) suggests relatively dry conditions in north-western Europe during the Migration Period. These opposite signals most likely reflect that the general pattern of the atmospheric circulation system, i.e. location of the frontal system, the cyclonic activity, and the direction of the cyclones entering Europe, has changed significantly during the last 2,700 years.

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9 Summary

9.1 Findings

With analyses on sediment cores from the accumulation areas situated in the southern (Helgoland mud area) and the eastern (Skagerrak) North Sea, continuous high-resolution reconstructions of paleoenvironmental parameters for the Late Holocene were achieved. Natural as well as anthropogenic effects, distinguished within the different time intervals, were considered in five manuscripts. The following concluding remarks emphasize the main results.

Helgoland mud area

- A marked shift in sedimentation is recorded around AD ~1250. This change from high (>13 mm/yr) to low (~1.6 mm/yr) averaged sedimentation rates is most likely attributable to the disintegration of the island of Helgoland.
- Concerning the depositional history, various natural processes were considered. Among these, especially storm-flood activity is supposed to alter the North Sea coastline. However, it did not primarily affect the Helgoland mud area.
- Human impacts like mining in the mountainous regions of Germany, or the burning of fossil fuels since the beginning of the industrial revolution are detectable in the sedimentary record of the German Bight. Besides these large-scale (regional-global) influences, local activities, as for example fisheries, may have had an effect on sedimentation.
- Stable oxygen isotopes ($\delta^{18}O$; benthic foraminifera) obtained from the Helgoland core, are a proxy for summer salinity. In this shallow, near-coastal area, salinity is mainly driven by Elbe River discharge, and consequently by precipitation in the catchment area.
- A reconstruction of paleosalinities in the German Bight is dating back to AD 1200. The estimated SSS values range between 29 and 33.9 psu. Correspondingly, paleodischarge rates were calculated varying from 100 to 1375 m^3/s. An estimated mean of c. 750 m^3/s over the last eight centuries, corresponds to the present annual average discharge rate.
- High precipitation rates in the Elbe catchment are signaled by high discharge rates and reduced SSS. Precipitation-rich phases seem to be linked to the overall atmospheric circulation system of Central Europe.

Skagerrak

- In the Skagerrak, environmental changes were distinguished by the help of foraminiferal assemblages as well as benthic and planktonic foraminiferal fluxes. Whereas high productivity is constrained to relatively cool time intervals, low productive phases are linked to warmer climatic conditions. The observed time intervals were linked to historical epochs within the last 2,700 years.
- The $\delta^{13}C$ data reflect nutrient availability in the Skagerrak area, which is - to a certain extent - driven by the precipitation pattern of Central Europe.
- The $\delta^{18}O$ data mainly reflect temperature conditions, and correspond to the Northern Hemisphere temperature anomaly reconstruction of Mann and Jones (2003).
- The surface water masses appear to have reached minimum temperatures between 1650 and 1350 cal yr BP, during the Migration Period.
- Although there is foraminiferal indication of fluctuations of the North Atlantic inflow throughout the Skagerrak record, the influence decreased from maximum values between 2500-2200 cal yr BP to minimum values at present; with the major change reflected at around 1200 cal yr BP.

These findings indicate that precipitation in the drainage basins of the rivers of Central Europe in particular leaves an imprint on both sites investigated in the North Sea area. On the one hand, precipitation rates significantly influence the sea-surface salinity signal in the southern North Sea. On the other hand, environmental conditions in the eastern North Sea are most likely affected by changes in mean average precipitation, leading for example to more or less weathering. Since changes in precipitation rates and their spatial distribution through time over Central Europe are driven by the overall atmospheric circulation system, we conclude that precipitation is one of the climatic key parameters controlling the (marine) environmental conditions.

On longer timescales, there are fewer reconstructions of hydrological parameters such as precipitation than e.g. of temperature. Therefore, assessments of paleoprecipitation as presented in this study seem to be a promising contribution to paleoclimatic research.

9.2 Perspectives

In the present thesis a couple of marine proxies picture the paleoenvironmental conditions in the North Sea region throughout the last 2,700 years of the Late Holocene. In this respect, especially the Skagerrak core (GeoB 6003-2) bears high potential for further analyses. The present resolution, a spacing of 5 centimeters per sample corresponding to a temporal resolution of 30 to 40 years, can easily be improved by denser sampling. In the upper part of the core, this would improve the reconstruction and facilitate a comparison with the results derived from the Helgoland mud area (GeoB 4801-1). At first sight, the two records display a similar precipitation-related pattern. With regard to the rare paleoprecipitation records available for Northern and Central Europe, this seems promising to gain more insight about the climatic-oceanographic conditions in the entire North Sea region during the overlapping time period (Modern Optimum and LIA).

On a smaller spatial scale, the knowledge about the LIA would strongly benefit from a more detailed comparison with results obtained from the Skagerrak, northeast of core GeoB 6003-2 (Hass 1994; Senneset, 2002). Up to now, a thorough comparison with these previous studies is difficult, i.a. due to the difference in temporal resolutions. Since the study from Hass (1994) is considering the last 1,500 years, even a longer period of time than the LIA can be considered and the picture of the Skagerrak system completed with the (extended) GeoB 6003-2 data. Furthermore, the comparability with other proxy data, such as the Charman and Hendon (2000) and Mann and Jones (2003) data could become more detailed.

Additionally to the improvement of the data resolution in the upper part of core GeoB 6003-2, an expansion further downcore is possible. A reconstruction of paleoenvironmental parameters at this Skagerrak site could extend to 4,000 years back in time, i.e. back until the Bronze Age.

Concerning possible forcing factors, such a higher temporal resolution and more samples, respectively, would make i.a. spectral analyses possible, allowing to define periodic climate-ocean variability on a multidecadal time scale. Moreover, additional XRF data are available which were not under consideration in this thesis.

Besides the Skagerrak core GeoB 6003-2, a set of sediment cores was taken slightly to the east of this position during the expedition M45-5. Studying theses cores would close the gap between the locations of GeoB 6003-2 site and a sediment core in the easternmost part of the Skagerrak, which is at present investigated at the Department of Earth Sciences, Aarhus. Furthermore, the outcome of the HOLSMEER (Late Holocene Shallow Marine Environments of Europe) project will provide additional, new results from the North Atlantic realm.

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