The polar and subpolar North Atlantic during the last five glacial-interglacial cycles

Der polare und subpolare Nordatlantik während der letzten fünf glazial-interglazialen Klimazyklen

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Die vorliegende Arbeit ist die leicht unveränderte Fassung einer Dissertation, die 2001 an der mathematisch-naturwissenschaftlichen Fakultät der Christian-Albrechts-Universität zu Kiel vorlag.

Abstract

The main goal of this study was to determine Late Quaternary climatic variations of the high northern latitudes on glacial-interglacial as well as on millennial time-scales using a range of paleoceanographic proxy records, i.e., sediment color, carbonate content, ice-rafted debris (IRD), stable oxygen isotopes, and carbonate preservation, from deep-sea sediments from the Northeast Atlantic Ocean and the Nordic Seas.

Carbonate preservation analysis of deep-sea sediments from the Nordic Seas revealed a distinct glacial-interglacial carbonate preservation pattern during the last five climate cycles: Foraminiferal tests showed good glacial preservation and increasing corrosion during interglaciations. This pattern is similar to Pleistocene records from the deep Pacific Ocean, but seems to be out of phase with the rest of the Atlantic Ocean. The generally good carbonate preservation suggests well-ventilated deep water at the study areas in the Norwegian and Iceland Seas throughout most of the last 12 marine isotope stages (MIS), regardless of the climatic mode. There are two mechanisms that could have caused this distinctive Nordic Seas carbonate preservation pattern: (1) The enhanced calcite corrosion during interglacial periods may be due to high regional planktic productivity with an increased downward flux rate of organic material to the sea floor and, thus, to more corrosive bottom water due to carbondioxide enrichment. (2) It may be linked to global scale variations in the marine carbonate system, which caused a change in ocean alkalinity, i.e., altered the depth of the calcite saturation horizon.

In order to use color data for further paleoclimatic investigations the factors that control sediment lightness and red-green color in the North Atlantic were specified. Analysis of a Northeast Atlantic deep-sea sediment core proved that during the last five climate cycles changes in sediment lightness were controlled mainly by fluctuations of the fine (<20 μm) carbonate fraction. Additionally, changes in carbonate corrosion also have a strong influence on total sediment lightness, as stronger corrosion of the carbonate fraction will lead to increased sediment lightness values. This lightness increase of the entire sediment is caused by corrosional effects on the surface structure of foraminifera and, probably, coccoliths. Changes in the red-green color of the sediment are caused by variations in the iceberg-rafted input of terrigenous, reddish iron-containing minerals. Spectral analysis of the red-green color record revealed persistent millennial-scale climate variability during the last 500,000 years. It seems that maxima of climate variability were tied to times of changes in ice mass and coincided to times when sea level was 40% below present-day global sea level indicating threshold behaviour. This threshold controlled the amplitude of climatic variations on millennial time-scales during the Late Quaternary. Intervals of relatively stable ice volume, i.e., peak

interglacial and also peak glacial periods, showed reduced short-term variability of the climate system.

Proxy records from the Northeast Atlantic and the Nordic Seas unveiled that, in contrast to the more frequent occurrence of warm interglacial conditions in the subpolar North Atlantic, full interglacial conditions in the polar North Atlantic occurred during MIS 1, 5e, and 11 only. A comparison of planktic $\delta^{18}O$ from the North Atlantic suggests colder sea-surface temperatures during MIS 11 than during other peak warm periods, i.e., MIS 1, 5e, and 9. IRD data indicate a major temperature gradient in the polar North Atlantic and less advection of warm Atlantic surface water into the Nordic Seas during interglacial MIS 11 than at present. IRD input in the Northeast Atlantic and the Nordic Seas increased almost coherently during glaciations and terminations. A general lack of synchroneity in the recurrence periods of IRD maxima points at differences in the timing of iceberg discharges between Northern Hemisphere ice sheets. However, the subpolar and polar North Atlantic were synchronously affected by the mechanisms that caused variability of the glacial climate system on millennial time-scales.

These new results from Northeast Atlantic and Nordic Seas deep-sea sediments point at distinct climatic variability in the Northern Hemisphere on longer, glacial-interglacial as well as on shorter, millennial time-scales. The investigations revealed partly synchronous forcing of the glacial climate system in the polar and subpolar North Atlantic. In contrast, peak interglacial conditions in the polar North Atlantic were less frequent than in the subpolar North Atlantic region indicating pronounced differences in the paleoclimatic and paleoceanographic conditions between the two areas during these intervals.

Zusammenfassung

Das Hauptanliegen dieser Arbeit war es, Schwankungen im Klimasystem der hohen nördlichen Breiten während des späten Pleistozäns sowohl auf glazial-interglazialen als auch auf kürzeren, tausendjährigen Zeitskalen zu erfassen. Zu diesem Zweck wurde an Tiefseesedimenten des Nordost-Atlantiks und des Europäischen Nordmeeres ein breites Spektrum paläo-ozeanographischer Untersuchungsparameter (im einzelnen: Sedimentfarbe, Karbonatgehalt, Anteil eistransportierten, terrigenen Materials, stabile Sauerstoffisotope, Grad der Karbonaterhaltung) bestimmt und ausgewertet.

Die Analysen zur Karbonaterhaltung an Sedimenten des Europäischen Nordmeeres ergaben für den Zeitraum der letzten fünf Klimazyklen ein spezifisches glazial-interglaziales Muster in der Erhaltung des karbonatischen Materials: Die untersuchten Foraminiferenproben zeigten eine gute Erhaltung in glazialen Phasen und zunehmende Anlösungserscheinungen während der Interglaziale. Diese Ergebnisse entsprechen dem Muster pleistozäner Karbonaterhaltung im Pazifischen Ozean, stehen aber im Gegensatz zum Karbonaterhaltungsmuster wie es aus anderen Gebieten des Atlantischen Ozean bekannt ist. Die generell gute Karbonaterhaltung der Proben deutet darauf hin, daß es unabhängig von der Ausprägung des Klimas während der letzten 12 marinen Isotopenstadien mit Ausnahme kurzer Phasen eine ausreichende Durchlüftung des Tiefenwassers in den Untersuchungsgebieten in der Norwegen- und Island-See gegeben hat. Zwei Mechanismen könnten das Muster der Karbonaterhaltung im Europäischen Nordmeer verursacht haben: (1) Eine erhöhte Anlösung des Karbonates während interglazialer Phasen könnte auf eine hohe regionale planktische Produktivität zurückzuführen sein, die ein verstärktes Absinken organischen Materials zum Meeresboden und damit korrosiveres Bodenwasser durch eine erhöhte Freisetzung von Kohlendioxid zur Folge hat. (2) Die Karbonaterhaltung im Europäischen Nordmeer wird möglicherweise durch Veränderungen im globalen marinen Karbonatsystem gesteuert, wobei diese eine veränderte Alkalinität der Ozeane und damit Schwankungen in der Tiefe des Sättigungshorizontes von Kalzit im Ozean bewirkt.

Um eine paläo-ozeanographische Interpretation von Sedimentfarbdaten aus dem Nord-Atlantik durchführen zu können, wurden die Faktoren, die in diesem Gebiet die Helligkeit des Sedimentes sowie dessen rot-grüne Farbe kontrollieren, bestimmt. Untersuchungen an einem Sedimentkern aus dem Nordost-Atlantik belegten dabei, daß die Sedimenthelligkeit während der letzten fünf Klimazyklen im wesentlichen durch Fluktuationen in der feinen (<20 µm) Karbonatfraktion gesteuert wird. Darüber hinaus haben Veränderungen in der Karbonaterhaltung einen starken Einfluß auf die Helligkeit des Sedimentes, da eine erhöhte Anlösung der Karbonatfraktion mit einem Ansteigen der Sedimenthelligkeitswerte einhergeht, Dieser Anstieg in der Helligkeit des

gesamten Sedimentes ist auf Anlösungseffekte an der kalzitischen Oberfläche von Foraminiferen und - wahrscheinlich - auch von Coccolithen zurückzuführen. Veränderungen in der rot-grünen Farbe des Sedimentes sind durch Schwankungen im eisbergtransportierten Eintrag von roten, eisenhaltigen Mineralien in das Sediment verursacht. Eine Spektralanalyse des rot-grünen Farbkanals indiziert für den Zeitraum der letzten 500.000 Jahre eine permanente klimatische Variabilität auf Zeitskalen von 1.000 bis 10.000 Jahren. Dabei sind die Maxima klimatischer Variabilität an Phasen gekoppelt, in denen sich das Volumen der kontinentalen Eismassen verändert hat. Die Maxima fallen dabei mit Zeiten zusammen, in denen der Meeresspiegel 40% unter dem heutigen globalen Wert gelegen hat. Das Auftreten der verstärkten klimatischen Variabilität ist an das Überschreiten dieses Schwellenwertes gekoppelt, der somit die Amplitude von klimatischen Veränderungen auf kurzen Zeitskalen kontrolliert. Intervalle mit einem vergleichsweise stabilen Eisvolumen, d.h. die ausgeprägten warmen und kalten Phasen der Interglaziale und Glaziale, zeigten dagegen für die kurzfristigen Zeitskalen eine reduzierte Variabilität innerhalb des Klimasystems.

Ein Vergleich geochemischer und sedimentologischer Daten aus dem Nordost-Atlantik und dem Europäischen Nordmeer, die den Zeitraum der letzten 500.000 Jahre abdecken, konnte zeigen, daß es voll ausgeprägte interglaziale Bedingungen im polaren Nord-Atlantik nur in drei Phasen der marinen Isotopenstadien 1, 5e und 11 gegeben hat, wohingegen solche Bedingungen im subpolaren Nord-Atlantik regelmäßiger vorgeherrscht haben. Die planktischen $\delta^{18}O$-Werte aus dem Nordost-Atlantik lassen vermuten, daß die Meeresoberflächen-Temperaturen im Isotopenstadium 11 geringer waren als während anderer ausgeprägter Warmphasen der jüngeren Isotopenstadien 1, 5e und 9. Die Untersuchungen an terrigenem, eisverfrachteten Material indizieren für das Stadium 11 einen deutlichen Temperaturgradienten im polaren Nord-Atlantik, wobei es vermutlich einen geringeren Einstrom warmen Oberflächenwassers aus dem Atlantik in das Europäische Nordmeer gegeben hat als rezent. Der Eintrag eisverfrachteten Materials in den Nordost-Atlantik und das Europäische Nordmeer erfolgte weitestgehend zeitgleich während der Glaziale und in den anschließenden Terminationen, den Übergangsphasen von Glazialen zu Interglazialen. Die in den untersuchten Regionen zeitlich asynchronen Perioden der Eintragsmaxima in den Zeiten erhöhter Anlieferung von eisverfrachtetem Material sind ein Hinweis dafür, daß die Hauptkalbungsphasen der Eisberge an den einzelnen Eisschilden der nördlichen Hemisphere zu unterschiedlichen Zeitpunkten erfolgten. Generell wurden der subpolare und der polare Nord-Atlantik jedoch scheinbar synchron durch Mechanismen beeinflußt, die zu einer rapiden Variabilität im glazialen Klimasystem geführt haben.

Die hier präsentierten neuen Erkenntnisse aus Untersuchungen an Sedimenten des Nordost-Atlantiks und des Europäischen Nordmeeres zeigen deutlich, daß die nördliche Hemisphere in den letzten 500.000 Jahren durch charakteristische klimatische Variabilität sowohl auf glazial-

interglazialer Ebene als auch auf kürzeren Zeitskalen geprägt war. Die Untersuchungen belegten dabei vergleichbare Auswirkungen des glazialen Klimasystems auf die Verhältnisse im polaren und im subpolaren Nord-Atlantik. Im Gegensatz dazu zeigten die interglazialen Warmphasen, die im polaren Nord-Atlantik weit weniger häufig auftraten als im subpolaren Sektor, daß es in diesen Abschnitten deutliche Unterschiede in den paläoklimatischen und paläo-ozeanographischen Bedingungen beider Untersuchungsgebiete gegeben haben muß.

Chapter I - Introduction

I.1. Main objectives

The major task of this study was to determine distinct climatic variability in the Northern Hemisphere during the past 500,000 years by applying several sedimentological, micropaleontological, and geochemical analyses to deep-sea sediments from the Northeast Atlantic and the Nordic Seas. Today, the high northern latitudes are the world ocean's major place of deep-water formation, i.e., North Atlantic Deep Water (NADW) is an integrated a part of the global conveyor belt. During Late Quaternary intervals of repeated Northern Hemisphere ice-sheet waxing and waning this modern-type thermohaline circulation pattern declined or evolved. The strong differences in the climatic conditions of glacial and interglacial intervals in the high northern latitudes have been imprinted on deep-sea sediments from this region. Accordingly, much paleoceanographic research of the last decades has concentrated on the North Atlantic and the Nordic Seas. However, few data of long, continuous high-resolution sediment records exist that document both long-scale glacial-interglacial climate change, and rapid climatic variations on millennial time scales. In this study several records from deep-sea sediments with varying temporal resolution were used in order to obtain detailed knowledge about orbitally induced mechanisms that effect the climate on Milankovitch time-scales as well as into climatic variability on shorter, millennial time-scales. This approach was to unveil new insights into the forcing and the feedback mechanisms of climate fluctuations in the Northern Hemisphere during glacial and interglacial times.

Changes in the marine carbonate system are documented in deep-sea carbonate preservation records, because they are closely linked to paleoceanographic conditions, i.e., surface productivity, ice cover, ventilation of the bottom water, and alkalinity of the oceans (Berger, 1992; Archer and Maier-Reimer, 1994; Henrich, 1998). Glacial-interglacial carbonate dissolution cycles from the Atlantic and Pacific Oceans have been studied in great detail (e.g., Crowley, 1983; Farrell and Prell, 1989; Karlin et al., 1992). However, investigations on carbonate dissolution in the northernmost Atlantic region have so far been limited to qualitative descriptions of coarse carbonate preservation (Henrich, 1986; Henrich et al., 1989; Huber et al., 2000). A new method to analyse carbonate preservation will be presented in order to estimate the total carbonate loss in the Nordic Seas caused by dissolution during the last five climate cycles. This will give new insights into one of the current problems of climate research: The large amplitude glacial-interglacial change in the atmospheric CO_2 concentration documented by air bubbles trapped in Greenland and Antarctic ice cores (Sowers et al., 1991; Jouzel et al., 1993, Petit et al., 1997). The terrestrial biosphere and soil carbon reservoirs are too small to trigger the preanthropogenic atmospheric carbondioxide increase of about 80 ppm known since the last glacial period. Hence, changes in the marine carbonate system are believed to

be responsible for these fluctuations (Sarmiento and Toggweiler, 1984; Broecker and Peng, 1993; Broecker et al., 1999; Archer et al., 2000).

Color is one of the fundamental sediment properties. The relationship between sediment color, i.e., lightness L* and the carbonate component of a Northeast Atlantic core site will be investigated in detail. In general, color of glacial-interglacial deep-sea sediments from the North Atlantic is characterized by an alternating light-dark-pattern of carbonate-rich sediments and sediments rich in detrital particles. Previous studies of North Atlantic sediments have shown a good correlation between sediment lightness (or comparable measurements like gray scale or brightness) and carbonate content (Nagao and Nakashima, 1992; Cortijo et al., 1995). Therefore, lightness has been used to estimate sediment carbonate content (e.g., Ortiz et al., 1999). However, it will be shown that the relationship between sediment color and carbonate content is not that straightforward. It is crucial to determine the effects of the main sediment components on total sediment color to avoid misleading results when interpreting color records of late Quaternary sediments for paleoceanographic purposes.

High-resolution North Atlantic sediment records and Greenland ice core data covering the last climate cycle both document climate instability on millennial time scales during the last glaciation, i.e., the periodic occurrence of so called Heinrich and Dansgaard-Oeschger events (Broecker et al., 1992; Bond et al., 1992; Dansgaard et al., 1993; Grootes and Stuiver, 1997). Sediment color records can provide a much higher resolution than conventinal proxy methods (Andrews and Freeman, 1996), consequently they are suitable to investigate rapid climate fluctuations (e.g., Cortijo et al., 1995). However, sediment color has so far not been used to document the evolution of millennial-scale climatic variability during the late Pleistocene. Thus, a Northeast Atlantic sediment color record will be studied by means of time series analysis to investigate millennial-scale climate change during the past five glacial-interglacial cycles. The color record will help to reveal distinct patterns of short-term climate instability and be used to identify mechanisms that can force the climate system to undergo fluctuations on short time scales

Since first detailed studies from the polar and subpolar North Atlantic showed that severe glacial-interglacial climatic variability was imprinted on deep-sea sediments (Ruddiman and McIntyre, 1976; Kellogg, 1977; 1980) much paleoceanographic research in the high northern latitudes concentrated on the conditions of specific glacial and interglacial intervals during the late Pleistocene (e.g., Boyle and Keigwin, 1987; McManus et al., 1994; Oppo et al., 1997; Vidal et al., 1999). In general, it could be documented that complex climatic interactions appeared during each glacial and interglacial period. However, so far glacial and interglacial sediment records from the polar and subpolar North Atlantic region have not been compared in detail. In this study several

sedimentological and geochemical parameters from Northeast Atlantic and Nordic Seas sediment records will be used to compare the climate characteristics of interglacial and glacial times and to reveal possible, distinct synchronous or asynchronous trends between the areas under study. This can help to decipher the complex forcing and feedback mechanisms that drove the Quaternary climate system at high northern latitudes.

I.2. Regional hydrography of the study area

Three sites (obtained on several cruises of the german research vessels *Meteor* and *Polarstern*) from the Northeast Atlantic (M23414) and the Nordic Seas (M23352 and PS 1243) were selected for this study (Fig. I-1, Fig. I-2).

At present, warm and salty surface water, i.e., the North Atlantic Current (NAC), advects from the North Atlantic into the eastern Norwegian-Greenland Sea (NGS). As Norwegian Current (NC) it forms the water of the Atlantic Domain (AND) that flows along the Norwegian and Svalbard shelf margins into the eastern Arctic Ocean (Swift, 1986; Fig. I-1). The Polar Front separates these Atlantic inflow from cold and less salty Arctic surface water of the East Greenland Current (EGC) that makes up the Polar Domain (POD) and conveys south along Greenland and through Denmark Strait into the North Atlantic (Swift, 1986). Polar and Atlantic domains are separated by mixed waters of the Arctic Domain (ARD), which is the main location of deep water formation in the Nordic Seas (Fig. I-1).

The formation of deep water in the Nordic Seas is generated by cooling and evaporation of warm surface waters leading to a vertical overturn of these waters to form dense and cold waters in the Greenland and Iceland Seas (Broecker and Denton, 1990). The outflow of these waters into the North Atlantic across the Greenland-Scotland Ridge (Aagaard and Carmarck, 1994) contributes to the formation of cold and well ventilated NADW that flows south through the Atlantic and Indian Oceans into the Pacific (Broecker, 1991). A major part of the cold deep water upwells in the northern Pacific and returns as a warm and less saline surficial flow into the North Atlantic, increases temperature and salinity in the Carribean and conveys as Gulf Stream heat and moisture into the high northern latitudes. Any disturbance of these flow pattern in the past may have led to a shallower and weaker deep circulation and to a southward migration of Atlantic deep water production with strong influence on the climatic conditions of the high northern latitudes, because of less heat transfer to the north (Rahmstorf, 1994).

At present, Site M23352 from the Iceland Plateau is located in the ARD near the recent Polar Front (Fig. I-1, Fig. I-2), whereas Site PS1243 from the western part of the southern Norwegian Sea is located at the western edge of the warm NC near the Arctic Front in the AND (Fig. I-1, Fig. I-2).

During the last five climate cycles the severe differences between glacial and interglacial climate modes in the high northern latitudes, as documented by repeated south- and northward movement of the Polar and Arctic Fronts, caused highly variable surface and bottom water mass conditions in the entire Nordic Seas and at the study sites in the Norwegian and Iceland Seas (e.g., Labeyrie et al.,

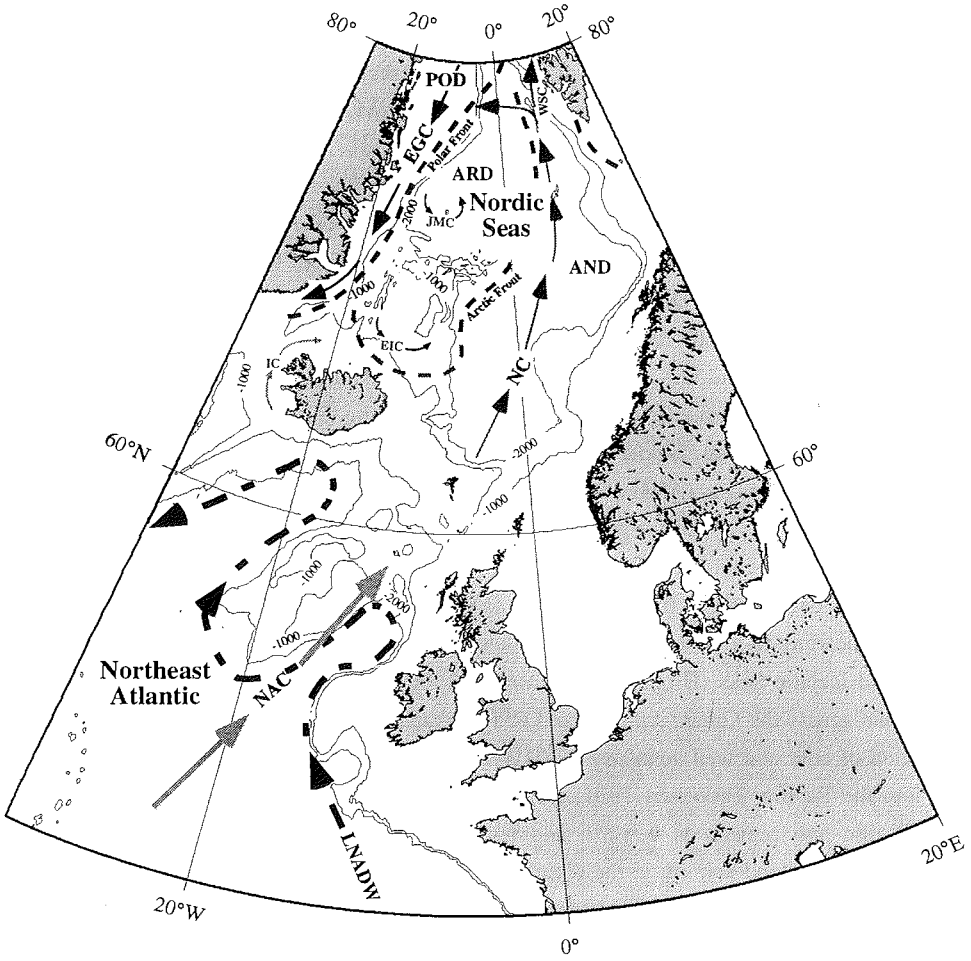


Fig. I-1: Simplified modern surface water circulation of the Northeast Atlantic and the Nordic Seas and the major oceanographic regimes of the later (Swift, 1986); POD (Polar Domain); ARD (Arctic Domain), AND (Atlantic Domain); Dotted lines indicate Polar (black line) and Arctic (gray line) Fronts. Gray and black arrows denote warmer and cooler surface currents, respectively; NAC (North Atlantic Current), NC (Norwegian Current), EGC (East Greenland Current), WSC (West Spitsbergen Current), JMC (Jan Mayen Current), EIC (East Iceland Current), IC (Irminger Current). Water circulation of the Lower North Atlantic Deep Water (LNADW) is denoted also (thick, dashed black line). Water depth is in m.

1987; Dokken and Hald, 1996; Hebbeln and Wefer, 1997; Henrich, 1998; Bauch et al., 2000a).

Today, study Site M23414 on the southern Rockall Plateau in the Northeast Atlantic at about 2200 m water depth is under influence of the Lower North Atlantic Deep Water (LNADW; Fig. I-1, I-2). LNADW is a mixture of dense overflows from the Nordic Seas and Labrador Sea Water (McCave and Tucholke, 1986), which in the Northeast Atlantic is located between 2000 and 3500 m water depth (Venz et al., 1999). The deepest parts of the Rockall area at the Plateau slope below 3500 m water depth are occupied by Southern Ocean Water, a nutrient-rich and oxygen-depleted water mass consisting of modified Antarctic Bottom Water that generates from the Antarctic (Manighetti and McCave, 1995). It is believed that during glacial times the production of NADW ceased, whereas

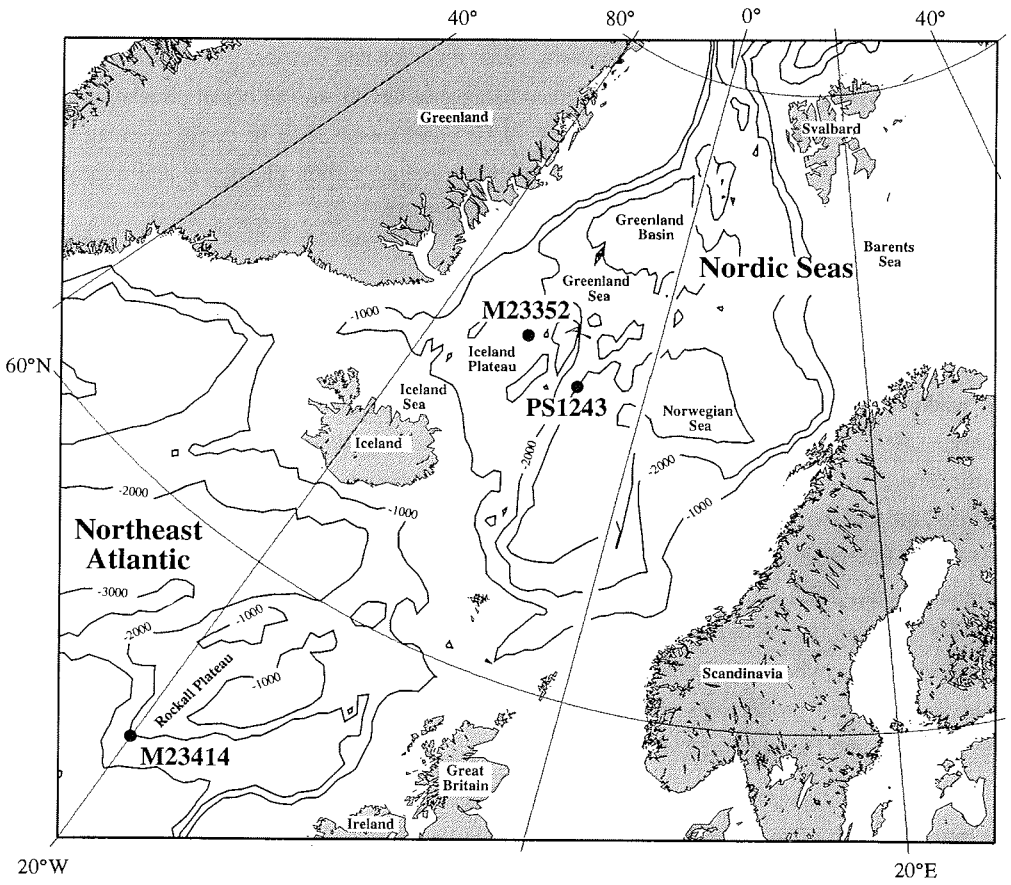


Fig. I-2: Geographical position of investitgated core sites in the Northeast Atlantic and the Nordic Seas. Water depth is in m.

the production of intermediate water, i.e., the Glacial North Atlantic Intermediate Water (GNAIW) was enhanced (Boyle and Keigwin, 1987; Oppo and Fairbanks, 1987; Venz et al., 1999). During the late Pleistocene Site M23414 was bathed by the well ventilated water masses of either LNADW during interglacial or GNAIW during glacial times (Venz et al., 1999). At present, the Rockall Plateau area is directly underlying the NAC (Fig. I-1).

I.3. The marine carbonate system

Calcium carbonate is a major component of deep-sea sediments and a large and reactive carbon reservoir. Accordingly, its accumulation and dissolution on the seafloor plays an important role in the global carbon cycle. The distribution of calcite in deep-sea sediments is determined by the kinetics of carbonate dissolution competing with the rate of carbonate burial, i.e., by the balance between biogenic production of $CaCO_3$ in the supersaturated surface waters and its dissolution in the undersaturated deep waters (e.g., Takahashi, 1975; Emerson and Bender, 1981; Farrell and Prell, 1989; Archer, 1991). Changes in the global mass balance of $CaCO_3$ and in ocean circulation largely control variations in carbonate preservation. As these mechanisms lead to cycling among several carbon reservoirs, they have strong influence on both oceanic carbon chemistry and atmospheric CO_2 content over orbital time scales (Broecker and Peng, 1987).

The recent carbonate sedimentation in the Nordic Seas is closely connected to the surface current pattern and its influence on carbonate flux (Honjo, 1990; von Bodungen et al., 1995; Henrich et al., 1995). Carbonate-rich sediments in the eastern NGS reflect the inflow warm Atlantic water by the NC, whereas the cold surface waters of the EGC govern extremely low carbonate sedimentation (Henrich, 1998; Huber et al., 2000). Additionally, the recent carbonate distribution pattern of the Nordic Seas is influenced by varying input of terrigenous material. Close to the continental margins of Greenland and Norway, the low carbonate contents of the sediments are likely to reflect dilution due to input of terrigenous material from gravitative downslope transportation, icebergs, and sea-ice (Baumann et al., 1993; Johannessen et al., 1994). In general, high carbonate contents in the Nordic Seas are believed to be related to the influence of Atlantic water masses and high surface productivity rates, whereas low carbonate values seem to be connected with cold Arctic surface waters and reduced productivity as well as enhanced dilution by terrigenous material (Henrich, 1989; Henrich and Baumann, 1994).

The overall pattern of recent carbonate preservation in the Nordic Seas reveals also the differences between the Polar and the Atlantic domains: Well-preserved carbonate tests occur in sediments underlying ventilated Atlantic surface waters (Henrich, 1998), as the pore waters of these carbonate-rich sediments will be supersaturated with respect to calcite (Huber et al., 2000). In contrast, modern

carbonate tests that are influenced by the EGC show increased dissolution (Helmke, 1996). Here, ice cover in the Polar Domain throughout most of the year causes low surface water production and low carbonate rain to the seafloor (Honjo, 1990), consequently the pore waters are undersaturated with respect to calcite. Enhanced recent carbonate dissolution is also reported from sites near the continental margins (Huber et al., 2000) and in the deep Greenland Basin. An enhanced dissolution close to continental margins can be explained by increased supply of organic matter and thus high release of respirative CO_2, whereas carbonate tests from the deep Greenland Basin were deposited close to the calcite saturation horizon in the Nordic Seas (Huber et al., 2000). In general, carbonate dissolution of deep-sea sediments in the Nordic Seas can be triggered by significant changes in the chemistry of deep-waters, i.e., by a reduction of deep-water renewal. A disturbance of deep-water renewal would decrease the formation and transport of oxygen-rich water masses in the NGS and imply reduced ventilation and aging of bottom water masses in the Nordic Seas (Henrich, 1986). Both study sites from the Nordic Seas (water depth about 1800 and 2700 m) were positioned above the calcite saturation horizon, i.e., the calcite lysocline, during the last five glacial-interglacial cycles. Thus, the late Pleistocene carbonate preservation mode of M23352 and PS1243 will also be influenced by supralysoclinal processes, e.g., by glacial-interglacial changes in regional surface bioproductivity and differences in the rain ratio between organic carbon and carbonate particles that reach the seafloor.

I.4. Color of marine sediments

Much information about oceanographic and sedimentological conditions at the time of sediment deposition is recorded directly in the color of fresh marine sediments. As optical scanning of sediment cores provides a fast and non-invasive paleoceanographic method, much effort has been taken into the development of technical systems to determine sediment color variations (Mix et al., 1992, 1995; Schaaf and Thurow, 1994; Merrill and Beck, 1995; Ortiz and Rack, 1999).

Color measurements have the potential to monitor a wide variety of sedimentological environments: For example, the ratio between goethite and goethite + hematite, which serves as a proxy for continental climate, as it increases with increasing precipitation, was recorded with sediment color measurements from Ocean Drilling Program (ODP) Leg 154 and used to characterize late-Pleistocene wet periods in the Amazon lowlands (Harris and Mix, 1999). Diffuse reflectance spectra analysis of Quaternary sediments from the Atlantic and Pacific Oceans suggest that sediment color can be used as a proxy for estimating organic carbon and opal contents (Balsam et al., 1996). However, as color records are strongly dependent on the sedimentary system of the area under study, interpretation of sediment color is not a stand-alone method, but has to be used together with further sedimentological proxy data.

The hand-held, Minolta CM-2002 spectrophotometer (Minolta Corporation, 1994)used in this study is a rather inexpensive, easily maintained instrument that provides extreme mobility. ODP, the French research vessel Marion Dufresne, as part of the IMAGES program, and several German research vessels operate with the Minolta CM-2002. When used carefully to avoid measurement error during hand-held operation color data from the Minolta CM-2002 can resolve information at the 1-cm scale indicating that century-scale and even decadal-scale fluctuations are likely to be investigated (Chapman and Shackleton, 1998).

I.5. Individual studies

This thesis comprises four manuscripts (Chapter III-VI) which have been submitted to peer-reviewed scientific journals. Together with the references from Chapters I and II the references from each of this manuscripts have been combined into one reference list:

Chapter III:

Glacial-interglacial carbonate preservation records in the Nordic Seas

A new method to investigate carbonate dissolution was used to determine the glacial-interglacial carbonate preservation pattern of deep-sea sediments from the Nordic Seas during the last 500,000 years. This new methodical approach is a combination of weight and color measurements as well as Scanning Electron Microscope analyses on planktic foraminiferal tests. The results were discussed with carbonate preservation records from the Atlantic and Indo-Pacific Oceans.

Chapter IV:

Glacial-interglacial relationship between carbonate components and sediment reflectance in the North Atlantic

The purpose of this manuscript was to qualify the factors that govern the relation between carbonate content and sediment lightness in the North Atlantic during glacial-interglacial times. Therefore, a detailed analysis of the coarse (>20 μm) and the fine (<20 μm) carbonate fraction from a Northeast Atlantic sediment core was carried out. Moreover, the influence of changing carbonate preservation on sediment lightness was discussed.

Chapter V:

Sediment-color record from the Northeast Atlantic reveals patterns of millennial-scale climate variability during the last 500,000 years

For the first time, a marine sediment color record was used to give insight into the evolution of climate change on millennial time scales throughout the last five climate cycles. Spectral analysis was applied to a red-green color record from Northeast Atlantic sediments to reveal variations in the

amplitude of millennial-scale variability during the last 500,000 years. The patterns of variability were then linked to changes in continental ice mass.

Chapter VI:
Comparison of glacial and interglacial conditions between the polar and subpolar North Atlantic Region over the last five climatic cycles

Sedimentological and geochemical proxy records of sediment cores from the Northeast Atlantic and the Nordic Seas were compared to evaluate the glacial and interglacial conditions between the areas under study during the last five climate cycles. The records were used to reveal synchroneity or asynchroneity in climatic behaviour between the subpolar and polar North Atlantic.

Chapter II - Material and strategy

II.1. Sample material

Seven sediment cores (Tab. II-1, Fig. I-2) from the Northeast Atlantic Site M23414 (cruise Meteor 17; Suess & Altenbach, 1992) and the Nordic Seas Sites PS1243 (cruise Arktis 2; Augstein et al., 1984) and M23352 (cruise Meteor 7; Hirschleber et al., 1988) were investigated:

Tab. II-1:

Core	Corer	Latitude	Longitude	Water Depth (m)	Recovery (cm)
PS1243-2	Trigger box core	69°22,5´N	06°31,3´W	2710	49
PS1243-1	Gravity core	69°22,3´N	06°32,4´W	2712	767
M23352-2	Trigger box core	70°00,5´N	12°25,5´W	1819	35
M23352-3	Kasten core	70°00,4´N	12°25,8´W	1822	826
M23414-6	Trigger box core	53°32,2´N	20°17,4´W	2201	36
M23414-8	Gravity core	53°32,3´N	20°17,5´W	2199	1339
M23414-9	Kasten core	53°32,2´N	20°17,3´W	2196	908

In the course of previous investigations sediment samples of the trigger box core M23414-6 and the upper 650 cm of the kasten core M23414-9 were taken at sampling intervals of 1 and 2.5 cm, respectively (Nees, 1997; Didié and Bauch, 2000). The lower about 2.5 m of the kasten core (from 651 cm core depth down to the core base) and the lower about 6 m of the gravity core M23414-8 (from 720 cm core depth down to the core base) were sampled at 2.5 cm intervals. Sediment lightness and carbonate records were taken to produce a composite record of Site M23414 (Fig. II-1; M23414-8 is corrected by +75 cm with respect to M23414-9). All sediment samples were freeze-dried and washed over a 63 µm sieve. After drying of the >63 µm residues they were sieved into the subfractions 63-125, 125-250, 250-500, and >500 µm for further sedimentological, micropaleontological, and geochemical investigation.

Both cores from the Nordic Seas have previously been sampled at intervals between 1 and 10 cm (Birgisdottir, 1991; Henrich, 1992, 1998; Bauch et al., 2000a, 2000b).

II.2. Stratigraphy

Stratigraphy of the Northeast Atlantic site is based on the planktic $\delta^{18}O$ record of *Globigerina bulloides* (data partly from Jung, 1996). In Chapter III benthic oxygen isotopes of *Cibicidoides wuellerstorfi* were used for stratigraphy of core M23414 (data from Jung, 1996). Due to the good agreement between planktic isotopes and sediment lightness at the Rockall Plateau site the

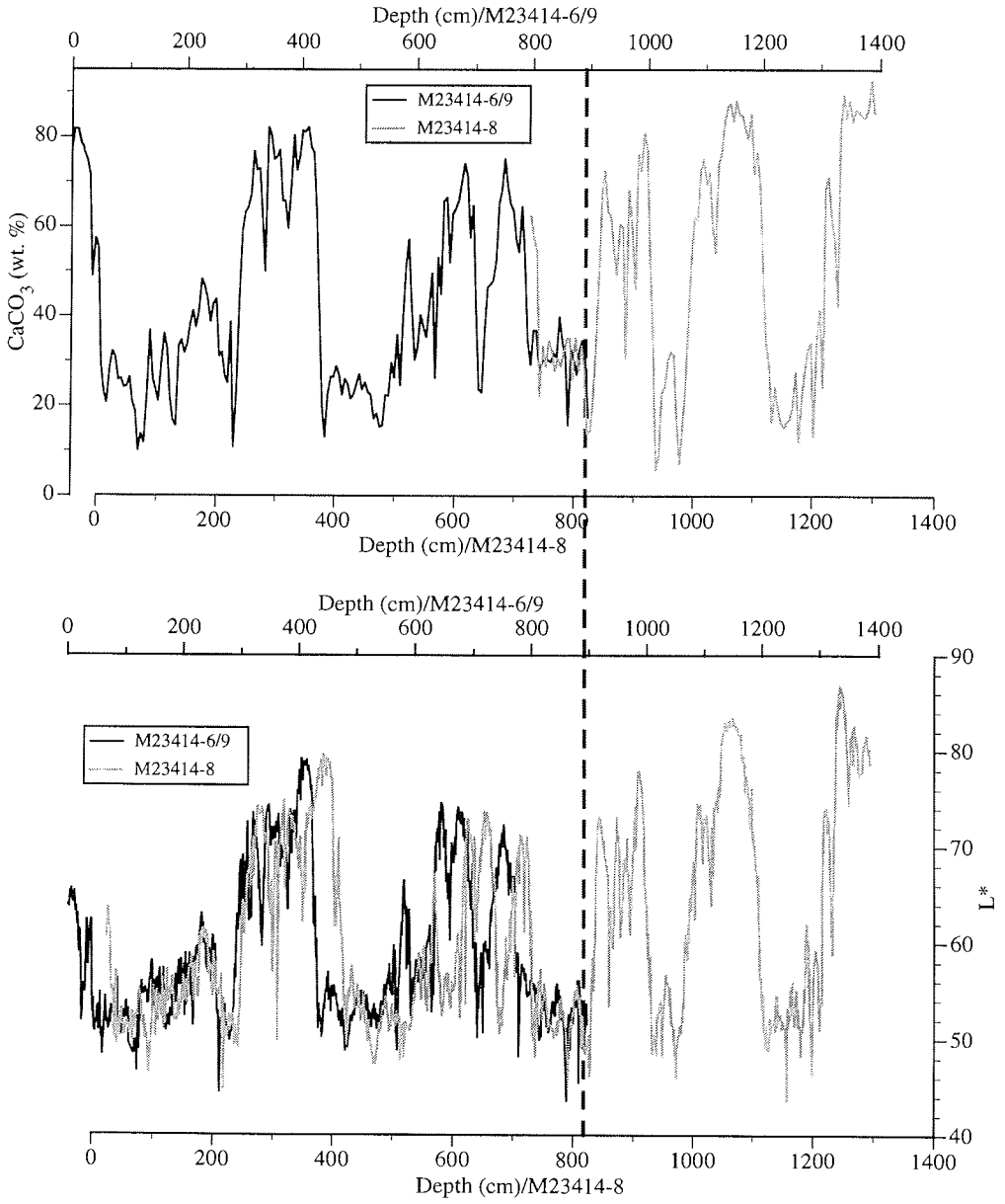


Fig. II-1: Correlation between the carbonate (weight% of bulk samples, upper panel) and the L* (lower panel) records of composite core M23414–6/9 (black line; original depth of M23414-9 is corrected by + 6 cm with respect to trigger box core M23414-6) and M23414-8 (gray line). Dotted dark gray line denotes point of correlation between the two cores. In composite core M23414 the original depth of M23414-8 (from 820 cm to the core base) is corrected by + 75 cm with respect to M23414-6/9.

stratigraphy of M23414 can be refined with the higher resolved, centimeter-sampled L* data (see Chapter V, Fig. V-1 and Chapter VI, Fig VI-2). At Sites PS1243 and M23352 the polar planktic foraminifer *Neogloboquadrina pachyderma* sinistral (sin.) was used for stratigraphy (data taken from Bauch et al., 2000a; 2000b; see Fig. IV-5). Apparently, Site M23414 goes back to marine isotope stage (MIS) 13, whereas the Nordic Seas cores both penetrate MIS 12 making the study sites suitable to investigate paleoclimatic changes during the last 5 glacial-interglacial cycles.

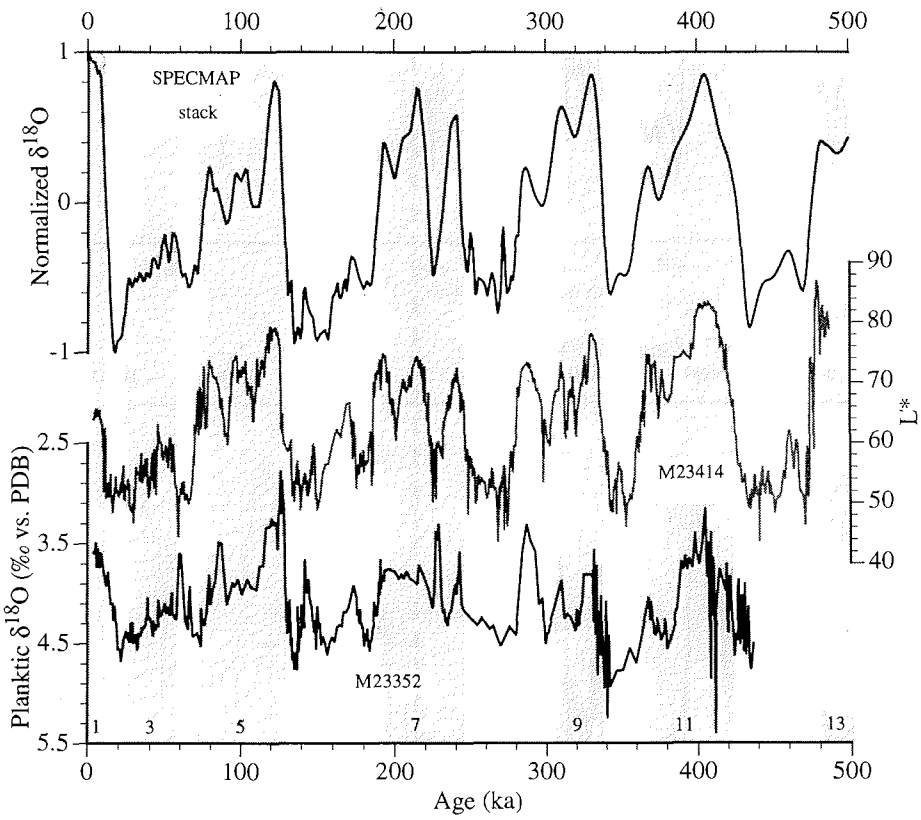


Fig. II-2: Age models of sediment lightness L* from M23414 (middle panel) and planktic oxygen isotopes from M23352 (*N. pachyderma* (sin.), lower panel). Both records were produced by alignment to the SPECMAP-stack chronology (upper panel).

Ages were assigned to the composite core M23414 by correlation of the centimeter-sampled lightness record to the stacked SPECMAP oxygen isotope chronology (Imbrie et al., 1984; Martinson et al., 1987) using the software AnalySeries (Fig. II-2). Between the stratigraphic correlation levels, the sedimentation rate was assumed to be constant and linear interpolation was

applied to produce the depth-age relations of the M23414-proxy data lightness L*, red-green color a*, $CaCO_3$, and ice-rafted debris (IRD) (see Chapters V and VI). In the upper part of the record, 4 AMS ^{14}C ages (C. Didié, unpubl.), which were converted into calendar ages using CALIB 4.1.2. (Stuiver et al., 1998) and the calibration model of Voelker et al. (1998), were included. Ages of Heinrich events H 1-6 (according to Sarnthein et al., in press) were also used to refine the uppermost part of the record. The positions of H 1-6 at Site M23414 have been previously identified by determining the concentration of lithic grains in the dry residues of size fraction >250 µm (Didié and Bauch, 2000). Average sedimentation rates of M23414 range between 0.25 and 14.5 cm/ky; Accumulation rates range between 0.4 and 24 g/cm²/ky (Fig. II-3).

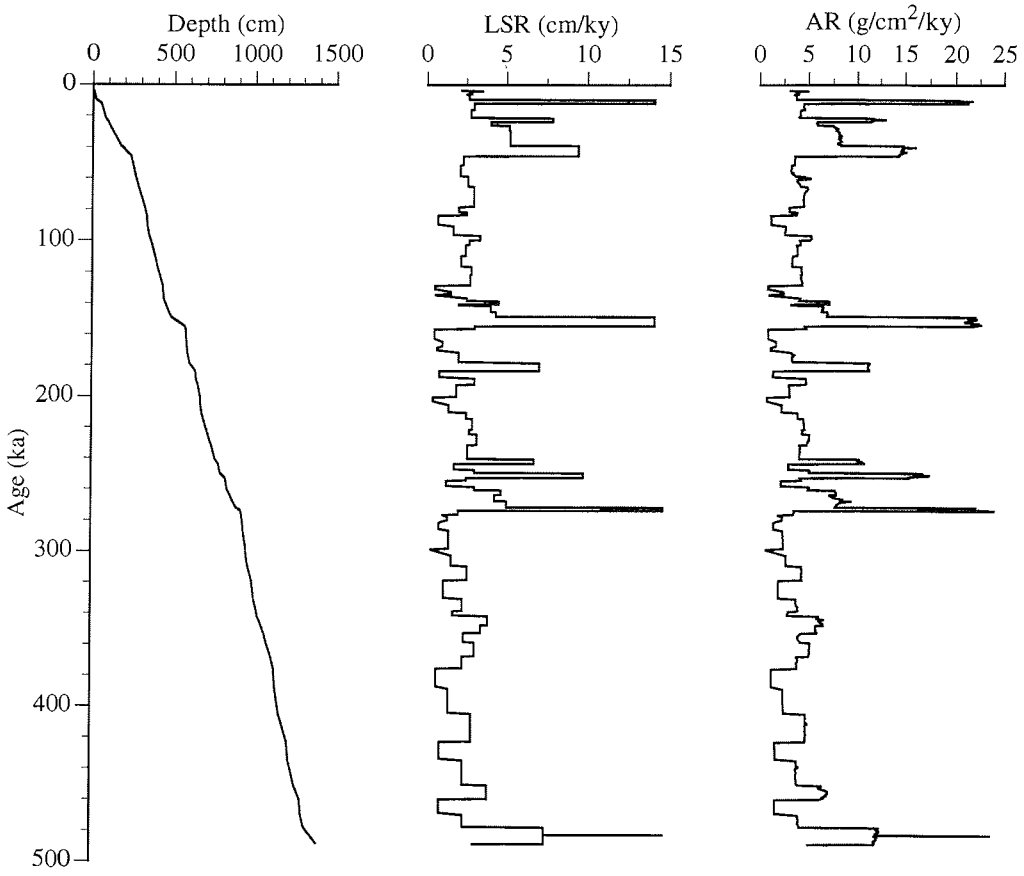


Fig. II-3. Age/depth control, linear sedimentation rates (LSR), and accumulation rates (AR) of Site M23414.

The age model for the oxygen isotope record of M23352 (sample resolution 1-4 cm) was produced likewise by synchronization to the SPECMAP stack (Fig. II-2). The age model of the younger part was refined by 7 AMS ^{14}C ages (C. Didié, unpubl.; converted using CALIB 4.1.2.) and by including the age of H 6 (according to Sarnthein et al., in press; identification of H 6 as described for core M23414) as well as the age of the Vedde ash layer (following Voelker, 1999; position of the ash layer according to Bauch, unpubl.). Prior to the alignment, the data were smoothed with a 14-point least square smooth. This was necessary to avoid misinterpretations of the lowermost part of the M23352 record, which is characterized by massive IRD-input and reveals atypical large scale fluctuations in isotope values. As for M23414, linear interpolation was used to establish the age models for the records of $CaCO_3$ and IRD (see Chapter VI).

II.3. Color measurements

Distinct sediment color measurements of core M23414 were carried out at centimeter steps using a hand-held Minolta CM-2002 spectrophotometer with a 8-mm-diameter measurement spot. The three-dimensional L*a*b* color space, which was established by the Commision International de l'Eclairage (CIE) in 1976, was recorded. Previously, this color description system has been used for various paleoceanographic investigations on glacial-interglacial sediments (e.g., Nagao and Nakashima, 1992; Balsam et al., 1999). The illumination system of the spetrophotometer was set at an angle of 2° and the "normal light" D_{65}-mode was chosen.

Both sediment lightness L* (with L* = 0 corresponding to black and L* = 100 corresponding to white) and red-green chromaticity a* (with a* = +60 corresponding to red and a* = -60 corresponding to green) were recorded.

The Minolta CM-2002 spectrophotometer calculates the variables L* and a* as follows:

$$L^* = 116\,(Y/Y_n)^{1/3} - 16 \qquad \text{when } (Y/Y_n) > 0.008856$$
$$L^* = 903.29\,(Y/Y_n) \qquad \text{when } (Y/Y_n) \leq 0.008856$$

where Y is tristimulus value Y (for 2° Standard Observer) of the specimen and Y_n tristimulus value Y (for 2° Standard Observer) of a perfect reflecting diffuser.

$$a^* = 500\,[(X/X_n)^{1/3} - (Y/Y_n)^{1/3})] \qquad \text{when } X/X_n \text{ and } Y/Y_n > 0.00856$$
$$a^* = 500\,[7.787\,(X/X_n) + 16/116 - 7.787\,(Y/Y_n) + 16/116] \qquad \text{when } X/X_n \text{ and } Y/Y_n < 0.00856$$

where X and Y are tristimulus values X and Y (for 2° Standard Observer) of the specimen and X_n and Y_n tristimulus value Y (for 2° Standard Observer) of a perfect reflecting diffuser.

Repeated measurements of the white calibration standard of CM-2002 demonstrates that the spectrophotometer provides a precise tool for measuring lightness and red-green chromaticity. Mean and standard deviation σ of 50 measurements were L* 96.63 ±0.01 and a* -0.11 ±0.01. The variations in the measured color of the sediments obtained L* values between 43.52 and 86.98 and a* values between -0.52 and 4.19, thus, the signal/noise ratio of the instrument is large. To avoid long-term instrumental drift (Chapman and Shackleton, 1998) the spectrophotometer was recalibrated every 200 measurements.

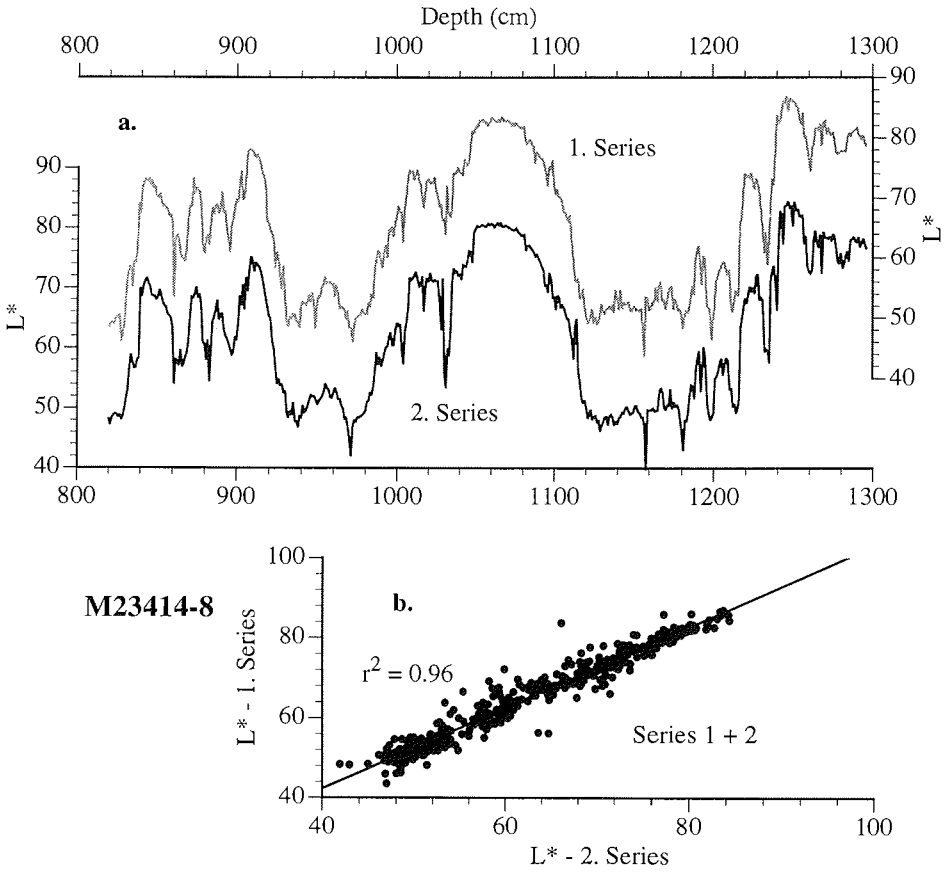


Fig. II-4: (a.) Downcore lightness variations of core M23414-8 (from 820 cm to the core base). Series 1 and 2 follow separate tracks. (b.) Linear correlation of Series 1 and 2.

Additionally, the precision of the sediment color measurements was tested using a set of 10 repeated measurements at 5 single positions in the core. The mean standard deviation of these sample sets is 0.06 for L* and 0.01 for a*. Two separate tracks of downcore lightness measurements on the lower 5 m of core M23414-8 show negligible differences between the records and reveal a correlation coefficient of 0.96 (Fig. II-4). Two shorter tracks of red-green chromaticity (Series 1 and 2 followed separate tracks) also obtain a high correlation coefficient of 0.95 (Fig. II-5). These comparisons suggest that lateral downcore color changes, e.g., caused by bioturbation, have a rather small effect on the measurements and that the centimeter-sampled color data reflect real features of the sediment record rather than analytical noise.

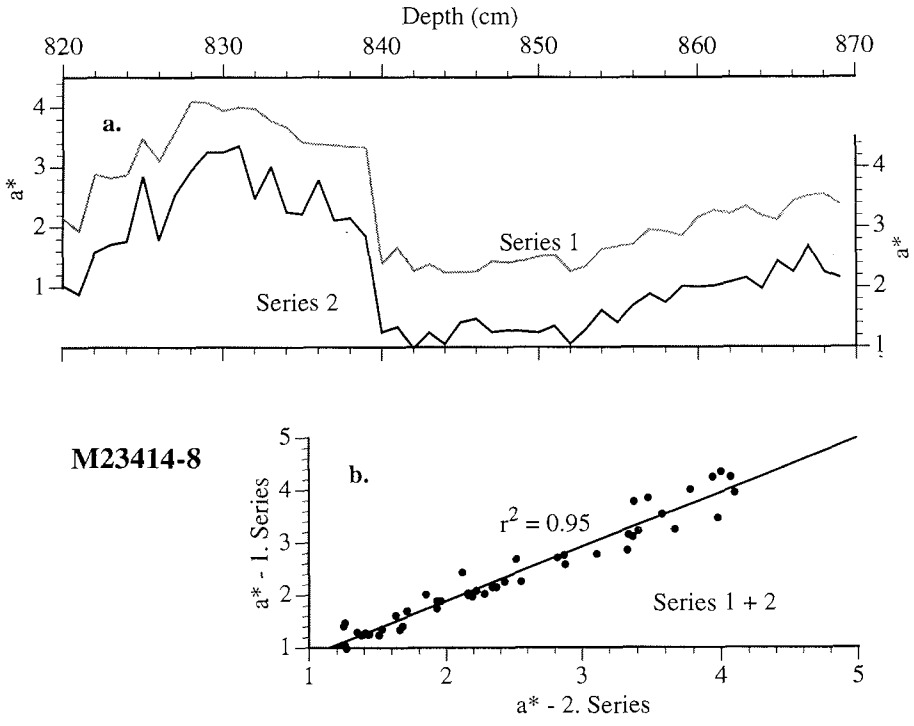


Fig. II-5: (a.) Downcore red-green chromaticity variations of core M23414-8 (from 820-870 cm). Series 1 and 2 follow separate tracks. (b.) Linear correlation of Series 1 and 2.

II.4. Carbonate and total organic carbon

Carbonate and total organic carbon (TOC) content of all three sediment cores from Site M23414 were analyzed at a sampling interval of 5 cm. All dry sediment residues were grinded using a hand-held pestle and mortar to provide a well homogenized sample for the quantitative elemental determination. For the TOC analyses the samples had to be decalcified prior to measurement. For

this purpose a few drops of hydrochlorid acid were added several times to the samples until the whole calcium carbonate was removed an no further reaction took place. To accelerate the reaction the samples were placed on a warm plate. Each sample was measured twice to reduce measurement errors by minimizing deviations, as the results were rejected if two measurements of the same sample differed by more than 0.2%.

As the biogenic component of the >20 µm fraction consists mainly of foraminifers while the <20 µm fraction is composed mostly of coccoliths, these fractions will be refered to as coarse and fine, respectively. It is recognized that the <20 µm fraction consists partly of detrital carbonate. In order to measure the coarse and fine carbonate content parts of the sampled sediments had to be wet-sieved with a 20 µm sieve (previous treatment of sediment samples as described above).

Bulk, coarse, and fine carbonate as well as TOC content were measured using a LECO C-200 carbon determinator at GEOMAR, University of Kiel. The carbonate content was calculated by subtracting the TOC content of decalcified samples from the inorganic carbon content (IC):

$CaCO_3 = IC - TOC * 8.33.$

II.5. Stable oxygen and carbon isotopes

Thirty individual tests of the planktic foraminifer *G. bulloides* from the size fraction 125-250 µm were picked from core M23414-8 (sampling between 800 cm and the core base) with a sample interval of 2.5 cm. Only complete tests were selected to reduce the risk of picking a wrong species.

Oxygen and carbon isotope values were measured with an automated carbonate reaction device, coupled with a Finnigan MAT-251 mass spectrometer at the Leibniz Laboratory for Age Determinations and Isotope Research, CAU Kiel. Stable oxygen and carbon isotopes are given as per mil deviations relative to the Pee Dee Belemnite standard (PDB; The standard is defined from the rostrum of the belemnite *Beleminitella americana*, B-1, from the Pee Dee Formation of South Carolina) and were calibrated using the NBS-19 standard (-2.2‰ $\delta^{18}O$). The analytical standard deviation is ±0.06‰ and ±0.03‰ for $\delta^{18}O$ and $\delta^{13}C$, respectively.

II.6. Ice-rafted debris (IRD)

To determine the amount of IRD at Site M23414-8 lithic grains in the size fraction >250 µm were counted with a spacing of 2.5 cm from 820 cm core depth down to the core base and are expressed as lithic grains per gram dry sediment. Investigations on the amount of IRD in the entire trigger box

and kasten cores have previously been carried out using the identical sampling scheme (Didié and Bauch, 2000).

II.7. Scanning Electron Microscope (SEM)

Surface structures of planktic foraminiferal tests using *N. pachyderma* (sin.) were investigated with the SEM at GEOMAR, University of Kiel to characterize glacial-interglacial carbonate preservation changes in cores PS1243 and M23352. In the Nordic Seas the left-coiled variety of *N. pachyderma* is the only species that occurs in both glacial and interglacial periods and therefore available for carbonate preservation analyses. Two principal surface textures of *N. pachyderma* (sin.) have to be distinguished: The reticulate form with a microcrystalline ultrastructure representing an earlier stage of calcification and the crystaline form with a secondary calcified crust that leads to a coarse crystalline ultrastructure. Most foraminiferal test are transition forms between these two types with the youngest chamber being still reticulate while the older chambers are crystalline. To avoid ambigous results only the oldest chambers on the umbilical side of the tests, which were always completely calcified, were analyzed. Previous SEM investigations on the ultrastructure of planktic foraminiferal tests from the Nordic Seas (e.g., Henrich, 1986, 1998) indicated good to excellent carbonate preservation during glacial and interglacial times, with increased corrosion being limited to short episodes, which were linked to diamiction horizons in the sediment. These results are partly in contrast to the SEM records presented in this study (see Chapter III). This may be due to the differences in the preselection of the examined tests: In contrast to this study, the results of Henrich (e.g., 1986) are a compilation of investigations on foraminiferal tests using both the reticulate and the crystalline form. Moreover, these previous studies have a lower average sample resolution of only 10 cm.

Based on the planktic oxygen isotope stratigraphy of PS1243 and M23352 (Bauch et al., 2000a; Bauch et al., 2000b) all samples were selected from depth intervals that respresent full glacial or interglacial periods, i.e., MIS 1, 2, 5.5, 6, 10, and 11. About 30 tests were randomly picked from various dry residues of the size fraction 125-250 µm and investigated with the SEM. The SEM analyses of corrosionial features on the ultrastructure of foraminiferal tests were used in combination with weight and color measurements to record downcore changes in carbonate preservation. The following successive steps in corrosion of the calcite crystals could be distinguished :

Step 1: Tests with sharp-edged, well preserved calcite crystals.
Step 2: First signs of corrosion of the coarse skalenoedra fabric with slight rounding of the crystal edges.

Step 3: Strong corrosion of the skalenoedra with loosening of the densley packed fabric of prisms and rounding of the crystal edges. Tests are still complete but can be easily disintegrated at this stage.

In general, carbonate corrosion at Sites PS1243 and M23352 is to low to produce ghost structures of the tests skalenoedra or even test fragmentation.

II.8. Weight and color measurements of foraminiferal tests

Weight and lightness of planktic foraminiferal tests of *N. pachyderma* (sin.) from cores PS1243 and M23352 were measured at sampling intervals between 1 and 10 cm. Weight measurements on varying amounts of foraminiferal specimens from size fractions 125-250 and 224-250 µm were determined using a microbalance (standard deviation is ±0.03 mg). The surface of all selected tests had no visible contamination of the surface with silt material. The tests were picked regardless of their surface texture, i.e., reticulate or crystalline morphotype.

Color of the previously weighed foraminiferal tests was measured using a Minolta CM-2002 spectrophotometer and was recorded as lightness L* in the L*a*b* color system. A cone-shaped depression (0.3 cm diameter) in a black metallic standard (mean and standard deviation of 50 measurements were L* 19.87 ±0.05) attached to the illumination system of the spectrophotometer was used to carry out the color measurements of foraminiferal tests. Lightness of the foraminiferal tests (500 foraminifers, 224-250 µm) and the black standard varies between 35.63 and 47.31, hence, as described above for the sediment measurements, the signal/noise ratio is large. Precision of the foraminiferal test measurements was calculated using 10 repeated measurements of 5 randomly selected tests (500 foraminifers, 224-250 µm). Mean standard deviation of the lightness from these set of tests is 0.11.

II.9. X-ray-fluorescence-spectrometry

This method can be used to give insight into the elemental composition of sediment samples. In this study it was used exclusively to investigate changes in the iron content of the sediments from Site M23414.

All analyses were carried out on about 1 gram of fine (<20 µm) and coarse (>20 µm) fraction sediment samples using the PW 1480 X-ray-flourescence-spectrometer at GEOMAR, University of Kiel. Prior to the measurements the samples were homogenized by grinding with a hand-held pestle and mortar and than oven dried at 110°. In a next step, the samples were vitrifed at a maximum temperature of 1150° C, additionally about 3.6 gram of flux component had to be added to the

sediment samples to lower the melting points of some of the elements the samples contained. After cooling of a melted sample flourescence of particular elements was stimulated with the X-ray source of the PW 1480. Whether the sample consisted a certain element could be identified by seting an internal crystal at the diffracted angel of the deflection peak of specific elements. The intensity of the diffracted angel of a X-ray peak was used to quantify the total amount of a specific element in the bulk sample. The analytical standard deviation for the Fe-analyses is ±0.4%.

II.10. Spectral analysis

In this study spectral analysis was applied to the sediment color record of Site M23414. Spectral analysis was performed using the software REDFIT (Schulz and Mudelsee, subm.), which allows direct processing of unevenly spaced time series. This program is based on the Lomb-Scargle Fourier transform (Lomb, 1976) for unevenly spaced time series in combination with a Welch-Overlapped-Segment-Averaging procedure (Welch, 1967). With this method a shortcoming of spectral analysis programs for evenly spaced time series, e.g., the Blackman-Tukey method (e.g., Jenkins and Watts, 1968), can be avoided: Interpolation of the unevenly time series may lead to an underestimation of the high frequency components in a spectrum, i.e., a 'reddening' of the spectrum (Horowitz, 1974, Schulz and Stattegger, 1997).

Time-dependent changes in the variance of the red-green-chromaticity color time sèries from Site M23414 was quantified using a sliding rectangular window, after high-pass filtering the unevenly spaced time-series (Rybicki and Press, 1995). The data were filtered with a cut-off frequency of 1/12 ky^{-1}. This is close to the accuracy of the age model from Site M23414, which is at least half the period of the precessional cycle. The width of the sliding rectangular window was 8 ky, thus offering a good compromise between statistical and systematic errors.

Chapter III: Glacial-interglacial carbonate preservation records in the Nordic Seas

III.1. Abstract

A combination of weight and lightness measurements as well as Scanning Electron Microscope (SEM) analyses on planktic foraminiferal tests from two sites in the Nordic Seas were used to investigate the pelagic carbonate preservation during the last 5 glacial-interglacial cycles. In general, a pattern showing good preservation during glacial times and enhanced corrosion during interglacial times can be observed. Marine Isotope Stage 11 (MIS 11) reveals the strongest corrosional features with estimated 45% total loss of foraminiferal carbonate before shell fragmentation. One reason for the enhanced interglacial corrosion may be a high regional surface productivity during these intervals, which led to increased dissolution rates in the deep sea driven by metabolic carbon dioxide. Yet, the carbonate preservation changes may also be linked to global changes in the marine carbonate system. Although the reason for the observed dissolution pattern in the Nordic Seas remains speculative, it is in phase with the rhythm of glacial-interglacial carbonate preservation in the Pacific Ocean but out of phase with the rest of the Atlantic. The data further support the hypothesis that much of the glacial decrease in atmospheric CO_2 may be attributed to changes in the alkalinity of the oceans.

III.2. Introduction

One key problem in interpreting past variations in the global carbon cycle is the massive decrease in atmospheric carbon dioxide during the last glacial period (Neftel et al., 1982; Barnola et al., 1987; Stauffer et al., 1998). The ocean is a large and sensitive carbon reservoir and, therefore, many studies on the origin of the glacial-interglacial atmospheric CO_2 changes have been focused on the marine carbonate system (e.g., Volk and Hoffert, 1985; Broecker and Peng, 1989, 1993; Archer and Maier-Reimer, 1994; Sanyal and Bijma, 1999; Broecker et al., 1999).

The pelagic $CaCO_3$ distribution in sediments is controlled by the balance between biogenic production in surface waters and dissolution in the deep sea (Catubig et al., 1998). Productivity processes determine the input of calcium carbonate but changes in the dissolution rate are mainly responsible for variations in calcium carbonate preservation. The dissolution rate of marine carbonate is dictated by the ratio of the in situ CO_3^{2-} ion concentration to the calcite saturation CO_3^{2-} ion concentration of sea water. Carbonate dissolution increases with increasing water depth because of the effects of hydrostatic pressure on the solubility of CO_2 (Hawley and Pytkowicz, 1969). Shallower water depths are supersaturated with respect to $CaCO_3$ and show only minor dissolution, whereas water masses of greater depth are undersaturated with respect to $CaCO_3$ and show complete dissolution. The crossover between these depth sections is called transition zone (e.g., Broecker and Peng, 1993). The top of the transition zone is defined by the lysocline, the shallowest depth where effects of carbonate dissolution are clearly visible, and the bottom by the Calcite Compensation Depth (CCD), the shallowest depth where the $CaCO_3$ is being completely dissolved.

The carbonate component of deep-sea sediments reveals a characteristic Pleistocene glacial-interglacial dissolution pattern. The Pacific and Indian Oceans show good glacial and poor interglacial carbonate preservation, with an estimated fluctuation in the water depth of the Pacific lysocline in the order of 400-800 m (Peterson and Prell, 1985; Farrell and Prell, 1991; Bassinot et al., 1994). Several investigations of carbonate preservation in the Atlantic (e.g., Crowley, 1983) showed evidence for stronger dissolution during glacial times. Estimates about the shallowing of the Atlantic lysocline during these dissolution pulses are still uncertain and vary between 300-500 m (Crowley, 1983; Howard and Prell, 1994) and more than 1000 m (Balsam, 1983). However, the interpretation of the data points to an out-of-phase preservation pattern between the Atlantic and Indo-Pacific ocean basins. Among the proposed explanations for a mechanism that can dramatically alter the alkalinity profile of the ocean and change the depth of the calcite saturation horizon by several hundred meters is the so called "coral reef hypothesis" (Berger, 1982; Opdyke and Walker, 1992). According to this scenario weathering and deposition of shallow water reef carbonate vary as a function of sea level, and trigger the glacial-interglacial input of alkalinity to the deep ocean.

Previous carbonate preservation analyses from the Nordic Seas (Henrich, 1998), the northermost part of the Atlantic Ocean, described good preservation within the past 300 kyr with peak interglacial carbonate preservation and dissolution pulses that were limited to short events during glacial times. Here we present new carbonate preservation records from the Nordic Seas using a new methodological approach in order to analyse the general Late Quaternary dissolution pattern. These records will then be discussed with carbonate preservation data from the Atlantic and Indo-Pacific Oceans.

III.3. Methods

Previously developed methods to describe carbonate dissolution involve establishing of a fragmentation index of planktic foraminifers or calculating the ratio between benthic and planktic foraminifers (Thunell, 1976; Diester-Haass, 1985; Le and Shackleton, 1992). Such methods work well for analyzing sediments with a strongly dissolved carbonate component but to estimate carbonate corrosion prior to foraminiferal test fragmentation other methods are needed.

We investigated two sediment cores from the Nordic Seas (Fig. III-1), PS1243 (69°22´N, 6°32´W, 2715 m water depth) and M23352 (70°0´N, 12°25´W, 1819 m water depth), each going back to MIS 12 (representing the past 5 glacial-interglacial cycles). Previous investigations on carbonate preservation in the Nordic Seas are based on corrosion indices of planktic foraminiferal tests using Scanning Electronic Microscope (SEM) analyses (Henrich, 1986, 1989; Baumann et al., 1996). We present here a new method to investigate carbonate corrosion of relatively well preserved foraminiferal specimens. This method is a combination of weight and lightness measurements as well as SEM analyses of planktic foraminiferal tests of the polar species *Neogloboquadrina pachyderma* (sin.). We used *N. pachyderma* (sin.) because this is by far the most common foraminiferal species in the Nordic Seas during the investigated time period.

The lightness of foraminiferal tests was measured using the Minolta CM-2002 spectrophotometer. Measurements were carried out on the level L* (%) in the L*a*b*-color space. All lightness measurements of foraminiferal tests were carried out using a small (0.3 cm diameter) cone-shaped depression in a black metallic standard that was attached to the illumination system of the spectrophotometer. For all lightness measurements the mean of five measurements was taken.

For 5 peak glacial and interglacial depth intervals 5000 specimens from a small size range (224-250 µm) were picked and weighed. In addition, 500 foraminifers from each sample were measured for lightness. For both study sites complete downcore records of foraminiferal test weight and

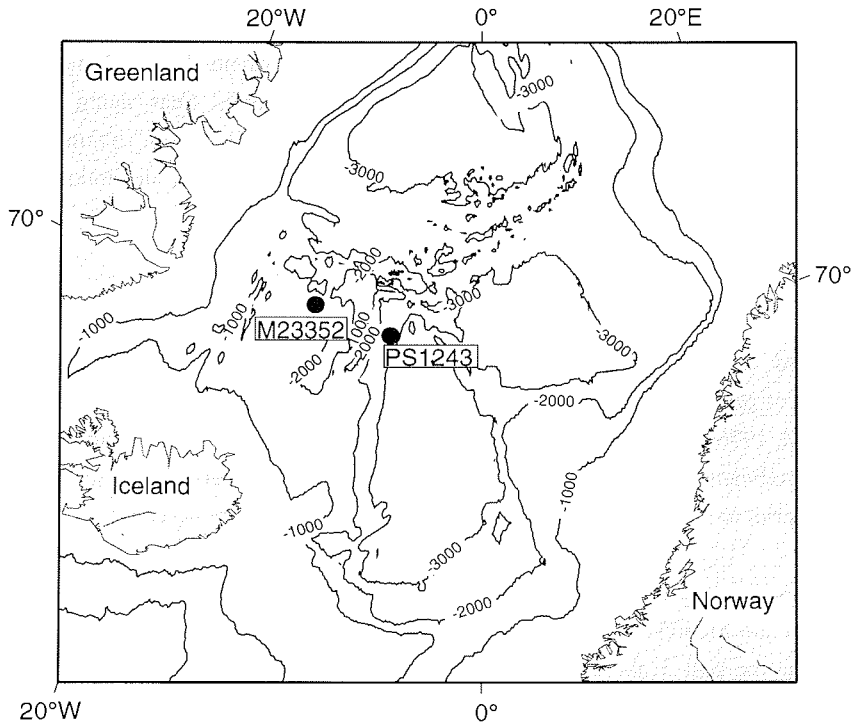


Fig. III-1: Overview of work area and position of studied sites (PS1243: 69°22´N, 6°32´W, 2715 m water depth and M23352: 70°0´N, 12°25´W, 1819 m water depth). Water depth is in m.

lightness were produced using 500 foraminifers from the size fraction 125-250 µm. These measurements have a sample interval of 1-5 cm in M23352 (with gaps between 526-586 and 672-737 cm, where the samples are barren or almost barren of *N. pachyderma* sinistral) and 5-10 cm in PS1243. Because variations in carbonate corrosion cause changes in the crystalline ultrastructure of foraminiferal tests, tests from different glacial and interglacial depth sections were also analysed with the SEM.

III.4. Results

III.4.1. Records of foraminiferal carbonate preservation

Carbonate corrosion essentially removes calcite from foraminiferal shells. Therefore, the loss in calcite should show up during weight measurements. In general, both weight records shown in

Figure III-2 reveal a remarkably good agreement, if one considers the distance between the two sites and the difference in water depth of about 1000 m. The heaviest tests occur within glacial MIS 2 and

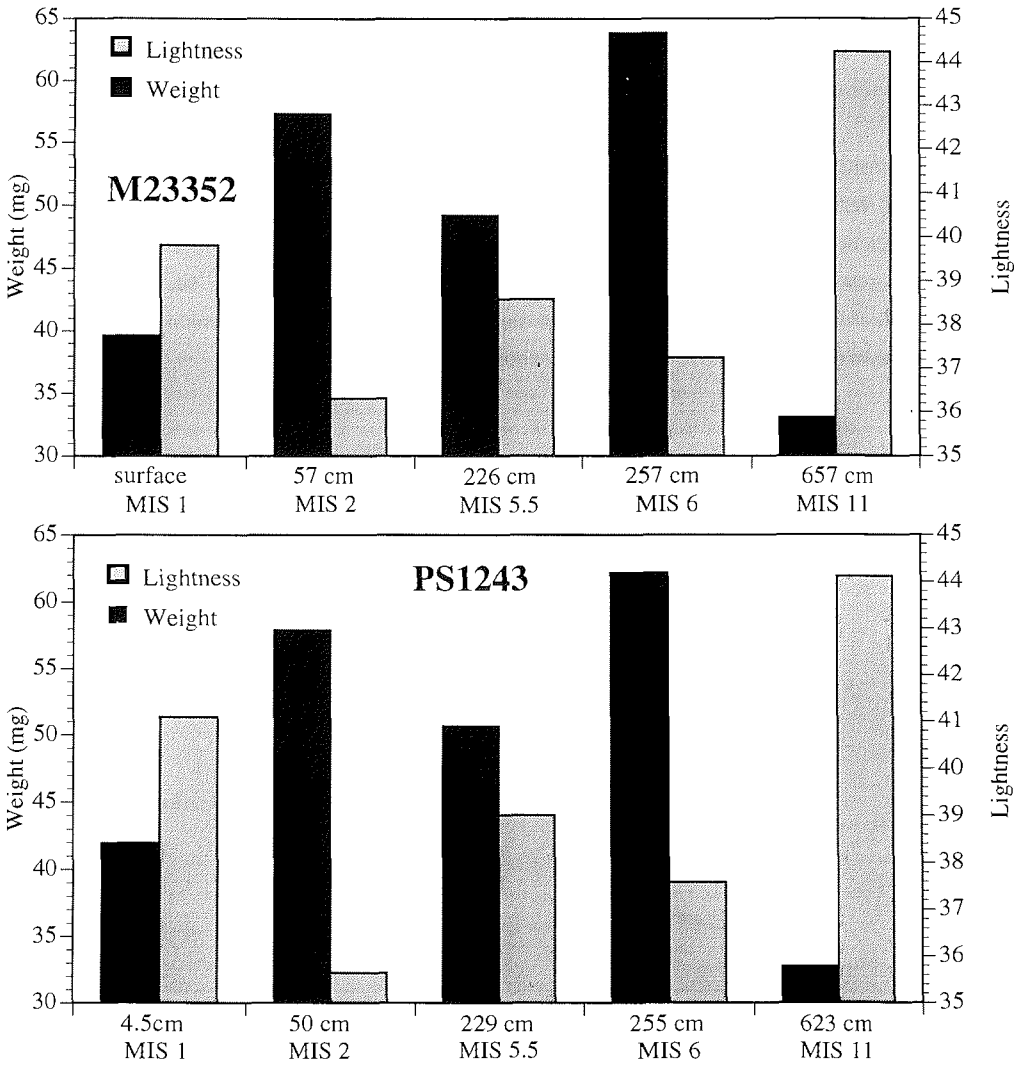


Fig. III-2: Comparison of weight and lightness data of tests of *N. pachyderma* sinistral (224-250 µm) from 5 glacial and interglacial stages. The black columns show the weight results (mg) of 5000 specimens, the columns the lightness results (L*%) of 500 specimens. The well preserved glacial tests (MIS 2 and 6) are characterized by highest weight and lowest lightness values. In contrast, the enhanced carbonate corrosion of MIS 11 is documented by lowest weight and highest lightness. Slightly corroded tests of the interglacials MIS 1 and 5.5 are somewhere between glacial tests and MIS 11. Estimates about the foraminiferal carbonate loss indicate a maximum value of 45% in comparison of MIS2/6 and MIS 11.

6. By comparison with MIS 2/6, the tests within interglacial MIS 1, 5.5, and 11 show mean weight losses of 32.5%, 17.5%, and 45%, respectively. The corresponding test lightness results reveal inverse trends. In other words, the glacial tests show lowest lightness values, whereas MIS 11 is characterized by highest test lightness values. This implicates that carbonate corrosion leads to changes in the surface structure of the calcite crystals and has a profound influence on the reflectivity of foraminiferal tests.

To test the sensitivity of the weight analyses, the 5000 shells of the 5 glacial-interglacial intervals were split into various smaller fractions (Fig. III-3). This exercise shows that a sample size of 100 shells is sufficient to trace the absolute glacial-interglacial weight changes in both cores. Moreover, shell weight results can be used to estimate calcite loss because effects such as possible contamination of the foraminiferal shell with silt material or varying degrees of calcification seem to be constant through time and have a minor influence on shell weight.

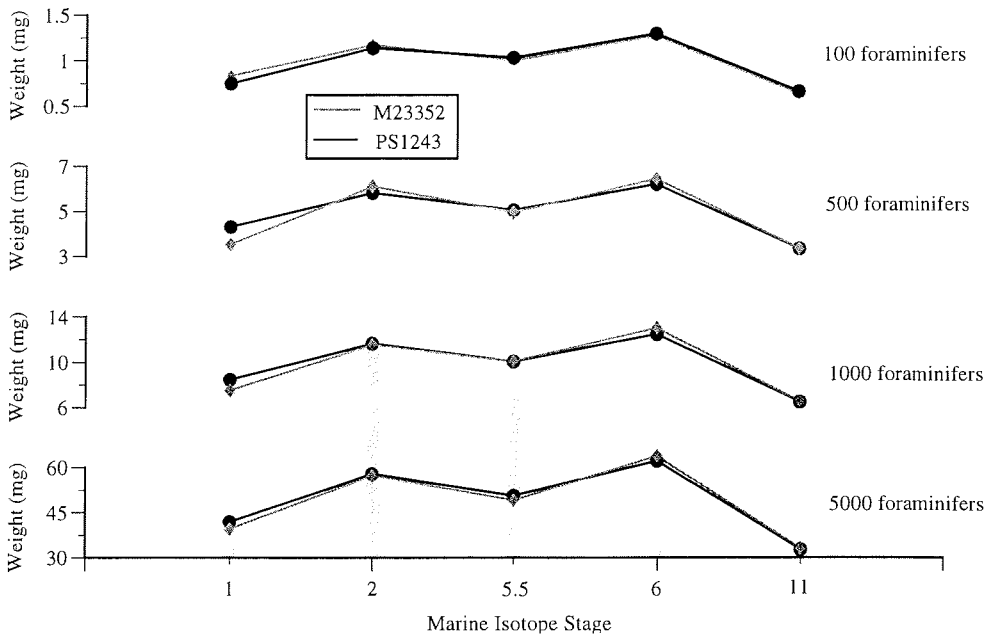
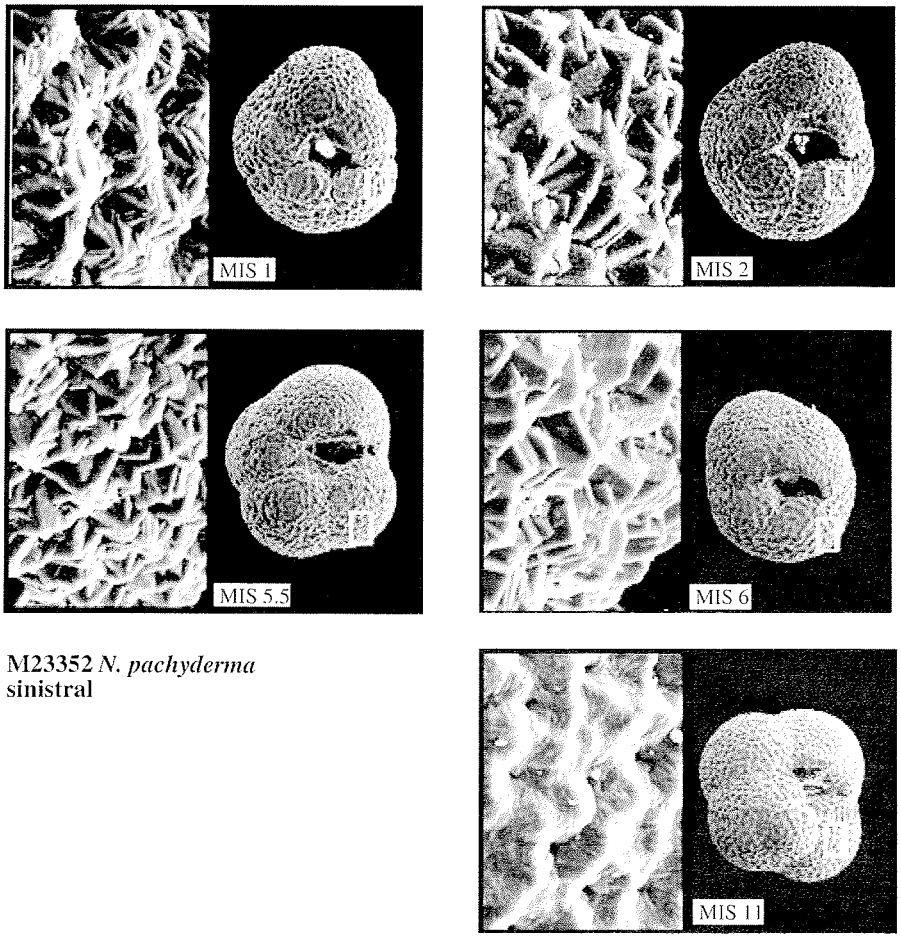


Fig. III-3: Various weight measurements (mg; note different scaling of value axis) of planktic foraminiferal tests (224-250 µm) from 5 isotopic stages using tests of *N. pachyderma* (sin.). The results of this split of tests reveals that the absolute foraminiferal test weight changes between different climatic intervals can be documented with a sample size of 100 foraminiferal specimens.

Tests of *N. pachyderma* (sin.) from glacial MIS 2/6 as well as from the interglacial MIS 1, 5.5, and 11 were examined with a SEM. In both cores the SEM analyses reveal a pattern of better glacial

than interglacial preservation (Fig. III-4). The glacial foraminiferal tests seem to be unaffected by corrosion and show perfect calcite rhombohedra, whereas the interglacial tests (from MIS 1 and 5.5) are characterized by early signs of carbonate corrosion, i.e., dissolution along the rims of the crystal edges with rounding of the crystal edges. Strongest corrosional features can be observed in MIS 11, where all calcite crystals are rounded and preferential dissolution around the chamber pores is visible. In MIS 11, the foraminiferal shells are very fragile and can be easily broken by any slight mechanical attack but the chamber structure is still complete. This weak chamber structure of



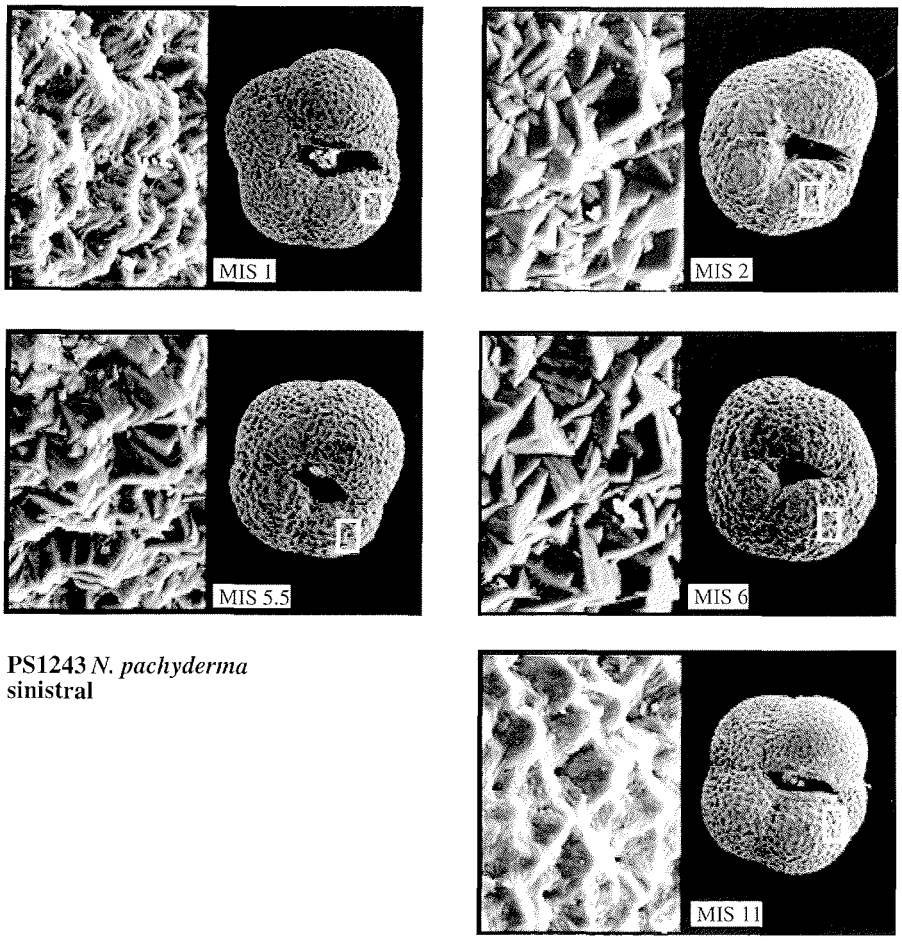
M23352 *N. pachyderma* sinistral

Fig. III-4: Comparison of various surface structures of *N. pachyderma* (sin.) tests from different depth sections taken from cores (a) M23352 and (b) PS1243. White rectangle indicates the part of the foraminiferal chamber that is magnified. Sharp-edged, well preserved calcite crystals are found on tests from glacial periods, whereas rounded crystals are observed on tests from MIS 11, indicating enhanced calcite corrosion. Tests from MIS 1 and 5.5 are characterized by first slight corrosional features.

specimens from MIS 11 verifies the observed strong weight loss in calcite during this period.

III.4.2. Weight and lightness of foraminiferal tests

Figure III-5 shows the downcore records of foraminiferal test weight, lightness and oxygen isotopic composition of *N. pachyderma* (sin.). Both weight records are marked by a decrease intest weight from MIS 6 to 1 with MIS 6 showing the highest weight results of the entire records. Two



PS1243 *N. pachyderma* sinistral

Fig. III-4b.

low-weight intervals, one in MIS 1 and the other in MIS 5.5, also occur. From MIS 11 to 6 the records reveal a general trend of increasing weight, with MIS 11 being characterized by the lowest total foraminiferal test weights. Inverse to the minimum test weight values, MIS 11 shows the

maximum test lightness values. In addition, there are high lightness values in MIS 1 and 5.5 and low values in MIS 2 and 10 that are inverse to the weight data of these periods. Besides the low-weight intervals of MIS 1, 5.5, and 11, several other interglacial low-weight intervals occur in MIS 7 and 9 but they do not always correlate between cores. High foraminiferal test weights occur throughout MIS 2-4, 6, and 10, although some tests of these glacial periods show low test weight comparable to the interglacial range. In general, however, the test weight data reveal a pattern of pronounced carbonate preservation during glacial periods and enhanced carbonate corrosion during interglacial periods.

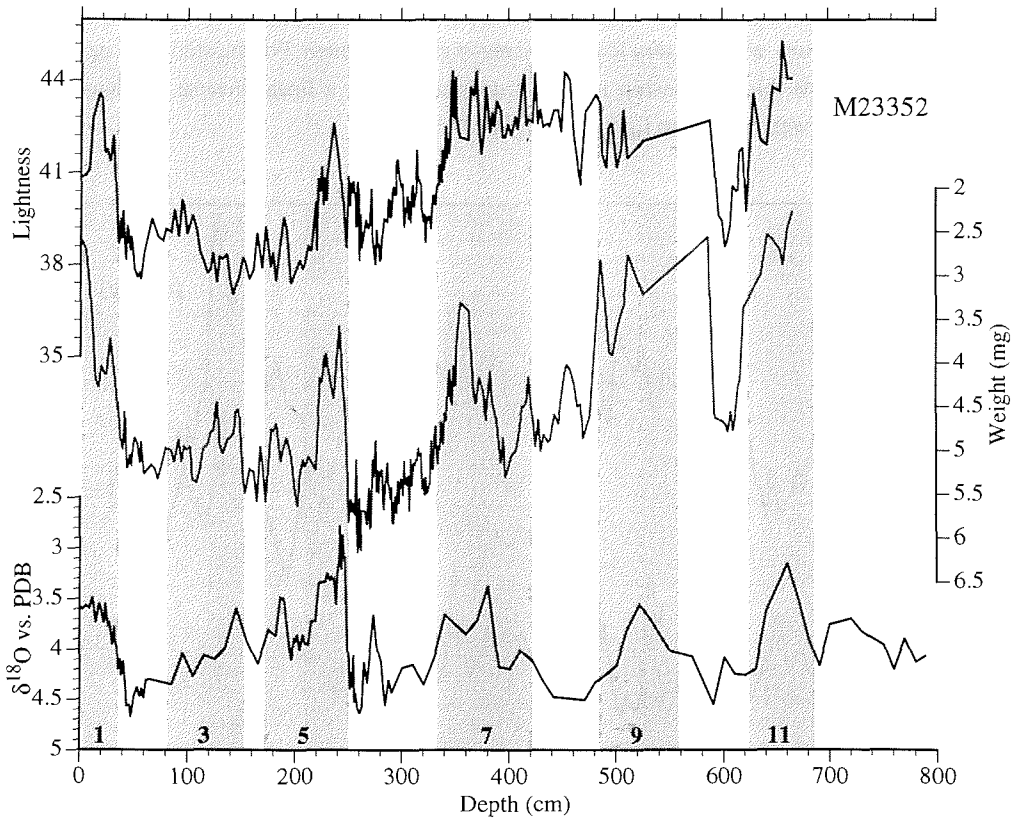


Fig. III-5: Records of test lightness, weight and $\delta^{18}O$ of *N. pachyderma* sinistral for (a) M23352 and (b) PS1243 (the data of the planktic oxygen isotopic curves are taken from Bauch et al., 2000a and Bauch et al., 2000b; Odd numbers indicate marine isotope stages (MIS) for reference and are shaded in gray). High test weight and low test lightness indicates good test preservation, whereas low test weight and high test lightness characterizes periods of enhanced corrosion. Despite some deviations between the weight and lightness data the records reveal glacials (MIS 2-4, intervals of MIS 6 and 8) as times of good carbonate preservation and peak interglacials (MIS 1, 5.5, intervals of MIS 7 and 9, and especially, MIS 11) as times when increasing carbonate corrosion prevailed.

When comparing the test lightness and weight records, it is notable that the inverse relationship between these variables, as observed for the peak glacial and interglacial periods (Fig. III-2), does not extend over the entire downcore records (Fig. III-5). For example, the maximum weight values of MIS 6 do not correspond with minimum test lightness. Also, the high lightness results of the entire MIS 7 in core M23352 are not in accordance with the weight record in this interval.

III.5. Discussion

Recent investigations from the Nordic Seas identified a relationship between carbonate preservation and foraminiferal test lightness (Bauch and Helmke, 1999). Their study indicated that microstructural changes cause the foraminiferal crystal surfaces to become less transparent with increasing dissolution, which in turn leads to increasing test lightness. For the studied 5 peak glacial and interglacial tests this relationship is clearly documented by a linear inverse relationship (r^2 = 0.84) between test weight and test lightness (Fig. III-6).

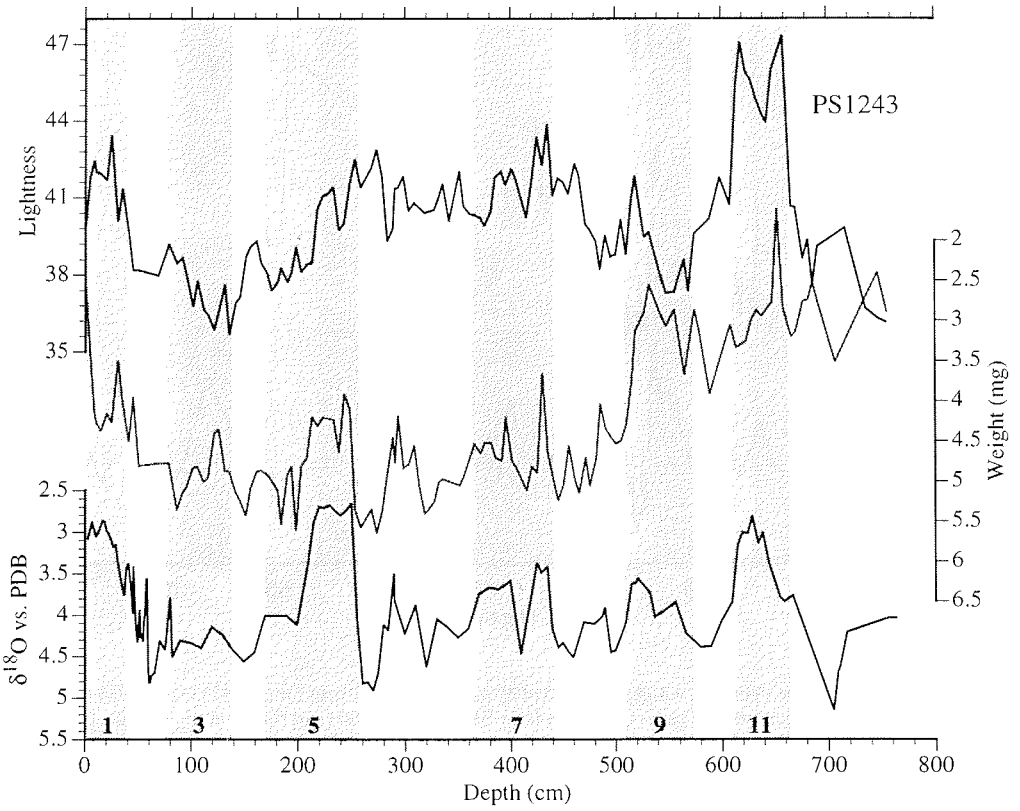
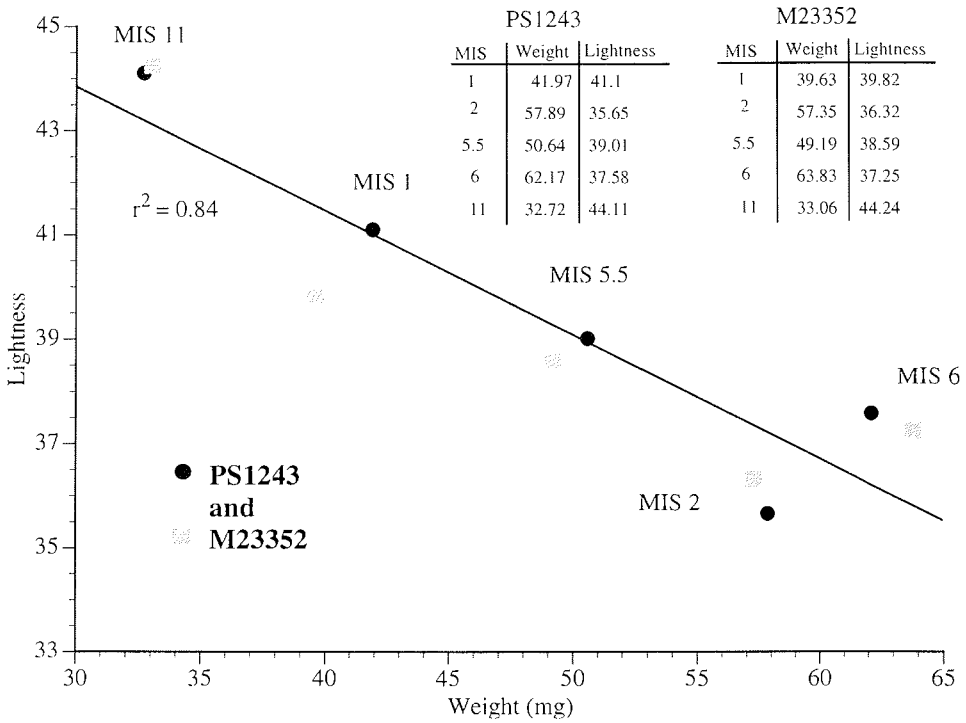


Fig. III-5b.

Previous studies of carbonate preservation in the Nordic Seas established qualitative foraminiferal dissolution indices (e.g., Henrich, 1986). Our analyses represent a direct approach to estimating the total carbonate loss caused by Late Quaternary dissolution in this region. At our two study sites, the coarse carbonate fraction commonly dominates the bulk carbonate content throughout the past 12 marine isotope stages, and within most core intervals the coarse carbonate is composed mainly of *N. pachyderma* (sin.). For these reasons we used foraminiferal test weight of *N. pachyderma* (sin.) to estimate carbonate loss. This method is limited to regions and time intervals where carbonate preservation is generally good. In this respect, the investigated sites are suitable because throughout both cores the shells of *N. pachyderma* (sin.) show no fragmentation.

Estimates about glacial-interglacial fluctuations of the water depth of the lysocline in the North Atlantic (Crowley, 1983) indicate that the studied Nordic Seas sites should have been situated above the lysocline during the entire investigated time interval. Both cores reveal a maximum difference in shell weight of 45% between glacial MIS 2/6 and interglacial MIS 11 (Fig. III-2). This exceeds



PS1243

MIS	Weight	Lightness
1	41.97	41.1
2	57.89	35.65
5.5	50.64	39.01
6	62.17	37.58
11	32.72	44.11

M23352

MIS	Weight	Lightness
1	39.63	39.82
2	57.35	36.32
5.5	49.19	38.59
6	63.83	37.25
11	33.06	44.24

Fig. III-6: Foraminiferal test weight versus test lightness of the previously described 5 glacial-interglacial samples (see Fig. III-2). The comparison of the two parameters reveals an inverse linear relationship ($r^2 = 0.84$) between test weight and lightness.

previous estimates of carbonate loss above the lysocline as reported by Peterson and Prell (1985), who observed maximum values of 30% in the Pacific. In the North Atlantic, shell density analyses of *N. pachyderma* (sin.) were used to calculate total foraminiferal carbonate loss (McManus and Lohmann, 1998). For the last glacial-interglacial cycle these authors observed the highest carbonate loss (more than 30%) in MIS 4, and found this dissolution pulse associated with foraminiferal shell fragmentation. In contrast, our analyses reveal that in the Nordic Seas the carbonate loss of *N. pachyderma* (sin.) can be much higher without any shell fragmentation occurring. In this respect it has to be considered that the left coiled variety of *N. pachyderma* is most resistant to carbonate dissolution among planktic foraminifers (Berger, 1968; Malmgren, 1983).

Many studies have focused on carbonate preservation in marine sediments, most of them emphasizing the out-of-phase dissolution pattern of the Indo-Pacific and Atlantic Oceans. It is believed that during glacial times carbonate dissolution was lesser in the Pacific (Farrell and Prell, 1989) and greater in the Atlantic (Berger et al., 1983; Crowley, 1983). However, certain preservation and dissolution events of the Indo-Pacific and Atlantic Oceans seem to be in phase (Crowley, 1983). Today the formation of well-ventilated North Atlantic Deep Water (NADW) makes the Atlantic Ocean less corrosive to calcite than the rest of the world´s ocean (Archer, 1991). A weakening in the strength of NADW formation during glacials (Mix and Fairbanks, 1985) is thought to be one major reason for decreasing carbonate preservation in the Atlantic (Crowley, 1983; Diester-Haass, 1985). It is still an open discussion as to whether the good glacial Pacific carbonate preservation can be attributed to a source of well-ventilated deep water in the Pacific (Moore et al., 1978; Keigwin, 1987; Curry et al., 1988). Our study confirms previous investigations that indicated generally good carbonate preservation in the Nordic Seas (Henrich, 1998). The preservation records point to well-ventilated deep water in the Nordic Seas throughout most of the last 12 marine isotope stages, regardless of the climatic mode.

One mechanism that might have caused the enhanced interglacial carbonate corrosion is high surface productivity. In the Nordic Seas, the interglacial MIS 11 was probably a time of high productivity (Bauch et al., 2000a). Increased plankton productivity leads to a high vertical flux of organic carbon to the sea floor, and consequently more carbon dioxide is released into the bottom water by decay of organic matter, lowering the pH and increasing carbonate dissolution. High carbonate accumulation in shallow water depth and enhanced deep sea carbonate dissolution in MIS 11 is also reported from other parts of the world ocean (Droxler et al., 1996; Howard, 1997). However, in all probability core M23352 is way above the saturation horizon, and it would need massive interglacial organic rain to trigger carbonate corrosion. This is not verified by the generally low TOC concentration (the mean less than 0.5%) of the bulk sediment (Henrich, 1992), yet certain benthic foraminifers indicate enhanced vertical flux of TOC in MIS 11 (Struck, 1997). Though

increased surface productivity seems to be an appropriate explanation for a generally stronger pelagic interglacial carbonate corrosion and the observed dissolution event during MIS 11 in the Nordic Seas, it has to be considered that high surface productivity has been invoked to explain both higher carbonate preservation and dissolution (Berger, 1970; Thunell, 1976; Arrhenius, 1988; Archer, 1991).

With the present data set we are not able to quantify which part of the foraminiferal test corrosion may be caused by porewater undersaturation with respect to calcite. In a scenario with effects of porewater chemistry on carbonate preservation, a higher relative rate of organic carbon rain decreases the porewater saturation with respect to calcite and would lead to a proportionally greater dissolution in the sediment driven by respiratory carbon dioxide (Berelson et al., 1990; Archer, 1991; Jahnke et al., 1997). A recent study on carbonate preservation in sediments from the Norwegian-Greenland Sea (Huber et al., 2000) also observes supralysoclinal calcite dissolution in the Nordic Seas. In this context the authors emphasize the importance of the molar ratio between organic carbon and carbonate particles that reach the sea floor for triggering dissolution above the calcium saturation horizon. An increase in the organic carbon/calcite ratio of material reaching the sediments was postulated for the last glacial (Archer and Maier-Reimer, 1994). This would lead to an increase of the carbonate ion in the bottom water and, through the mechanism of $CaCO_3$ compensation, would decrease the atmospheric carbon dioxide content. However, the sites investigated in this study were probably situated above the North Atlantic lysocline throughout the Late Quaternary, and in situ measurements of porewater by several authors are in contrast to each other: While Jahnke et al. (1994) found no evidence for dissolution driven by metabolic carbon dioxide above the saturation horizon, Hales et al. (1994) observed dissolution.

Another proposed explanation for a higher carbonate ion concentration in the ocean, and hence, lower atmospheric CO_2 content during glacials, is the "coral reef hypothesis" (Berger, 1982; Opdyke and Walker, 1992). According to this hypothesis the coral growth rates were lower during glacials when the sea level had dropped and the continental shelves were exposed. This would imply a shift of $CaCO_3$ deposition to the deep sea and therefore would have led to a deepening of the oceans saturation horizon and better carbonate preservation. However, this hypothesis calls for a global ocean increase in the saturation state (Opdyke and Walker, 1992; Archer and Maier-Reimer, 1994), and cannot be invoked to explain the observed preservation difference between the Nordic Seas and the rest of the Atlantic or between the Atlantic and the Indo-Pacific, respectively.

The observed Late Quaternary preservation cycles are consistent with estimated glacial-interglacial pH changes in the Atlantic and Pacific (Sanyal et al., 1995) wherein the deep water of both oceans

had a higher pH during the last glacial period. Accordingly, both the Atlantic and Pacific Oceans had a higher glacial deep-sea carbonate ion concentration. However, it is very likely that the boron isotope-based carbonate ion change will not be equivalent to the amount of change in preservation, as the boron based paleo-pH estimates need to be coupled with dissolution driven by metabolic CO_2 so as not to have an unrealistic change in lysocline depth of several kilometers.

At present, the glacial-interglacial changes in the bottom water saturation state of the Nordic Seas cannot be exactly quantified. The two cores have a depth difference of about 1000 m which would translate to a difference in the saturated CO_3^{2-} ion concentration in the bottom water of about 15 µmol/kg. However, no significant differences show up in the lightness data of the two cores. Therefore, future work on the exact relationship between the carbonate ion concentration and test lightness is needed in order to further quantify the changes in the depth of the Nordic Seas lysocline during the Pleistocene.

Given that the Nordic Seas data are out of phase with the rest of the Atlantic Ocean, changes in $CaCO_3$ production seem to be a more likely explanation for the observed preservation pattern than mechanism with effects on a global scale. Regardless of the mechanism that drives the changes in glacial-interglacial carbonate preservation, our results indicate that the Late Quaternary carbonate dissolution and preservation episodes occurred synchronously throughout the Nordic Seas and Indo-Pacific ocean basins, with better preservation during glacial periods. A better deep-sea carbonate preservation implies a high deep-sea carbonate-ion concentration, which in turn would lead to a lowered glacial atmospheric carbon dioxide content. Therefore, our data support previous investigations (Archer and Maier-Reimer, 1994; Sanyal, et al., 1995) that attributed much of the glacial drop in atmospheric carbon dioxide content to global changes in the marine carbonate system between the climatic modes.

III.6. Conclusions

The carbonate preservation of two intermediate water depth sites from the Nordic Seas were investigated using weight measurements as well as lightness and SEM analyses on planktic foraminiferal tests.

Within the last 12 marine isotope stages a generally good carbonate preservation with better glacial than interglacial preservation is indicated for this region. Although the foraminiferal tests are not fragmented, estimates about the foraminiferal carbonate loss reveal maximal values of 45% in comparison between peak glacial stages and interglacial MIS 11.

As a glacial-interglacial pattern with stronger interglacial dissolution is also known from the Indo-Pacific Ocean, our data point at an in-phase carbonate preservation pattern of the Nordic Seas and the Indo-Pacific Ocean. However, it has to be considered that the main phases of dissolution in the Nordic Seas seem to be out of phase compared to the rest of the Atlantic.

Whether regional processes like stronger surface productivity during interglacials or global changes in the marine carbonate system are responsible for the observed preservation record from the Nordic Seas is still uncertain. However, a late-Pleistocene carbonate dissolution pattern with better glacial preservation supports scenarios that attribute much of the drop in atmospheric carbon dioxide during the last glacial to changes in the alkalinity of the oceans.

Chapter IV: Glacial-interglacial relationship between carbonate components and sediment reflectance in the North Atlantic

IV.1. Abstract

A high resolution composite sediment record from intermediate water depth in the North Atlantic that dates back to marine isotope stage (MIS) 13 was investigated in order to determine the relationship between sediment reflectance (lightness) and carbonate content (weight %). For this purpose, a detailed analysis of the coarse (>20 μm) and fine (<20 μm) carbonate component was carried out to define which of the two carbonate components drives glacial-interglacial changes in sediment lightness. The results indicate that the bulk carbonate component is clearly dominated by the fine carbonate fraction regardless of glacial or interglacial climatic mode, suggesting that the sediment lightness is usually controlled by fluctuations of the fine carbonate content. However, a comparison of MIS 1 and 5.5 indicates that, besides this difference in the content of the two carbonate size fractions, changing modes in carbonate preservation, i.e., the preservational state of planktic foraminifera, may also have a profound influence on total sediment reflectance.

IV.2. Introduction

In recent years, the North Atlantic region has been the focus to study rapid climatic oscillations during the late Pleistocene (e.g. Heinrich, 1988; Bond et al., 1992, 1993; Oppo et al., 1998). To be able to investigate centennial to millennial-scale climatic fluctuations using marine sediment records, it is crucial to produce data series with densely sampled resolution. Sediment color analyses can be a serviceable tool, both to investigate such rapid fluctuations and to make estimates of varying sediment components. For example, Mix et al. (1995) measured reflectance spectra of Pacific Ocean sediments to estimate biogenic calcite and opal contents, whereas Grousset et al. (1993) used sediment grey level as a proxy for identifying time intervals with increased iceberg rafting in the North Atlantic, so-called "Heinrich events" (after Heinrich, 1988).

However, to use sediment color variations for paleoceanographic reconstructions requires information on the main sediment components of the general area under study. In the North Atlantic, sediments are mainly composed of biogenic carbonate which is predominantly deposited during warmer periods, and detrital terrigenous components such as iceberg-rafted debris (IRD) which are deposited during colder periods (Ruddiman and McIntyre, 1976; Smythe et al., 1985). Previous studies have shown a good correlation between bulk carbonate content and sediment reflectance as a result of changes in biogenic carbonate productivity and IRD input (Nagao and Nakashima, 1992; Cortijo et al., 1995; Ortiz et al., 1999). Increased carbonate deposition marks warmer climatic periods and contributes to higher grey-level values, whereas low carbonate contents and increased inputs of terrigenous material during colder periods yield lower grey-level values. However, this simple correlation may not always be that straightforward. For instance, the data from the North Atlantic (Cortijo et al., 1995) clearly reveal a marked mismatch between bulk carbonate content and corresponding grey-level values when comparing the present and previous interglacial periods, MIS 1 and 5.5. Despite similarly high bulk carbonate contents in both of these warm intervals, the sediment grey-level values in MIS 1 are significantly lower. In fact, they are even lower than in the colder substages MIS 5.1 and 5.3 when total carbonate content is less than in MIS 1

Recent investigations on sediments from the Nordic Seas using lightness data (Bauch and Helmke, 1999) demonstrated that the particular preservational mode of foraminiferal tests can have a significant influence on total sediment lightness reflectance. Consequently, to determine the effects of carbonate on sediment lightness, measurements of bulk, fine, and coarse carbonate as well as analyses of foraminiferal preservation changes may be necessary. The purpose of this paper, therefore, is to qualify the factors which govern the relationship between carbonate content and sediment lightness in the North Atlantic during the late Pleistocene.

IV.3. Methods

Sediment lightness and carbonate content were investigated using core M23414 (53°32´N, 20°17´W, 2196 m water depth; Fig. IV-1) from the northeast Atlantic.

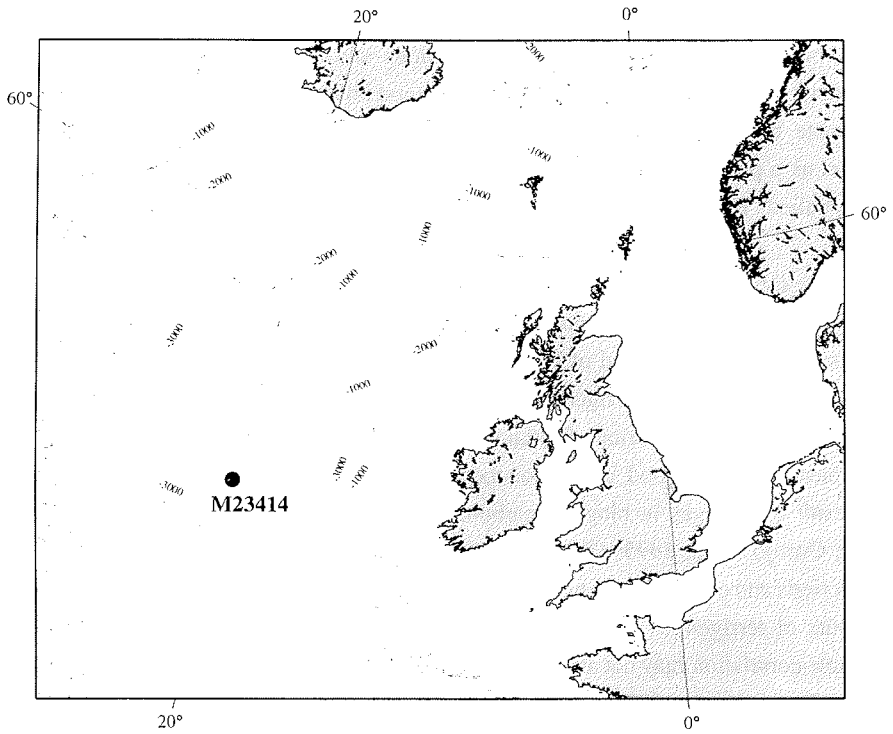


Fig. IV-1: Overview of work area and position of studied site where sediment cores were obtained (53°32´N, 20°17´W, 2196 m water depth). Water depth is in m.

Color measurements were obtained with a hand-held Minolta CM 2002 spectrophotometer. For all measurements the illumination system of the spectrophotometer was set at an angle of 2° and the "normal light" D_{65}-mode was chosen. All measurements were carried out at centimeter steps and the lightness value L* in the spherical L*a*b* color space was recorded. In this color space L* is defined as the brightness-axis with L* = 0% corresponding to black and L* = 100% corresponding to white.

With the exception of the trigger box core M23414-6, which was measured every 1.5 - 5 cm, the carbonate content was analyzed every 5 cm using a LECO-200. To better distinguish between the principle carbonate components and their influence on sediment color the samples were wet-sieved

over a 20 μm mesh. It is assumed in this study that the fine (<20 μm) and coarse (>20 μm) carbonate fractions are mainly composed of coccoliths and foraminifera, respectively. The terms coarse and fine carbonate are used here to refer to the two size fractions. However, we recognize that the size fraction <20 μm may also partly consist of detrital carbonate.

Both lightness and carbonate data were then taken to produce a composite record of site M23414 (Fig. IV-2, Fig. IV-3). The upper 31 cm are based on the trigger box core M23414-6 which was used because of its undisturbed surface section. The dashed line in Figure IV-2 indicates the main point of correlation between the lightness records of Kasten-core M23414-9 and the stratigraphically much longer piston core M23414-8. The shorter Kasten-core M23414-9 was also used in this case because it already had an oxygen isotope record (Jung, 1996). Based on the lightness data and the $\delta^{18}O$ record we were able to reconstruct a single record with a much longer stratigraphic range.

IV.4. Results

IV.4.1. Stratigraphy

Stratigraphic control in core M23414-6/9 (Fig. IV-2) is based on stable oxygen isotopes of the benthic foraminifera *Cibicidoides wuellerstorfi* which were measured at an average sample resolution of 2.5 cm (Jung, 1996). The corresponding lightness and carbonate records are in good agreement with the isotopic record, and the lightness results can be used to construct a detailed stratigraphic framework down to oxygen isotopic substage levels. The entire record extends to MIS 8 (approx. the past 250 ka).

Although no stable isotopes were measured in core M23414-8, the good agreement between lightness and stable isotopes in M23414-6/9 allowed to reconstruct an oxygen isotope stratigraphy also for this core dating back to MIS 13 (approx. the past 500 ka).

IV.4.2. Lightness and carbonate data

The bulk carbonate record is characterized by strong glacial-interglacial variations (Fig. IV-3). Peak interglacial values fluctuate between 70 and 90%, whereas glacial values can decrease down to less than 10%, although the latter mostly remain between 20 and 30% during colder intervals. A comparison of the fine and coarse carbonate fractions reveals that for most core intervals the bulk carbonate content is dominated by the fine component. Especially during the warm interglacial periods between MIS 5 and 13, up to 80% of the total carbonate is composed of fine carbonate with maximum values occurring in MIS 13. Through most of the glacial periods the bulk carbonate

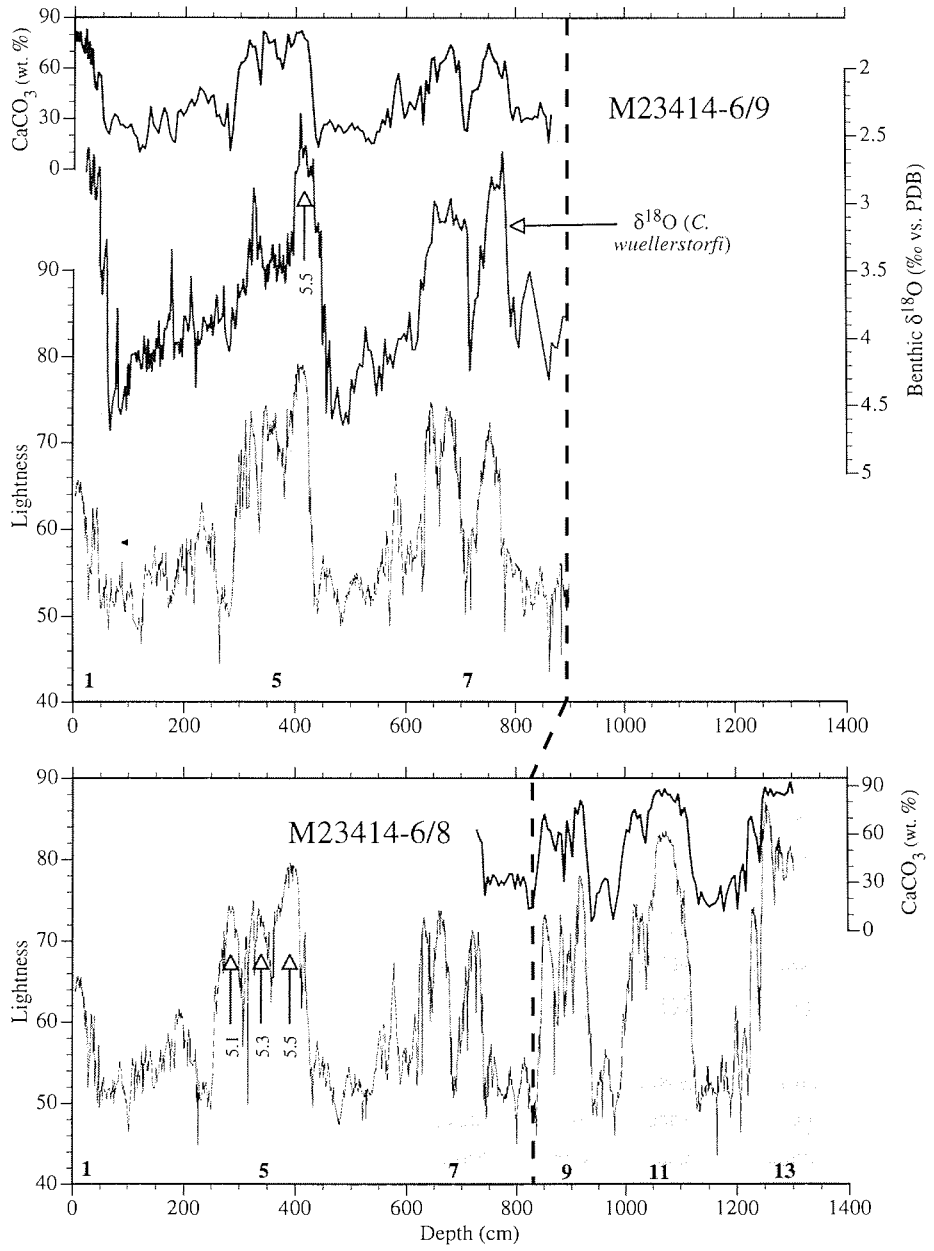


Fig. IV-2: Sediment lightness measurements and $CaCO_3$ contents for the composite cores M23414-6/9 and M23414-6/8 (some carbonate data of M23414-6 were taken from Andruleit, 1995). The benthic oxygen isotopes of *C. wuellerstorfi* from M23414-9 are also shown. Dashed line indicates depth of splicing. Interglacial marine isotope stages are shaded in grey and labeled for reference.

content is dominated by fine carbonate. At least for the interval stages 13 to 7 there seems to be a decreasing trend in peak interglacial values. A similar trend of decreasing interglacial carbonate can be traced from the records of ODP Site 982 in the North Atlantic (Venz et al. 1999). Lowest fine carbonate contents, ranging between 1 and 3%, are documented in MIS 2 and 12. However, even in some glacial intervals the values of fine carbonate contents may be as high as 20%

The coarse carbonate contents remain much lower than the corresponding fine carbonate values. The coarse carbonate fraction exceeds 40% within some short intervals represented by spikes which cannot be directly related to the peak interglacial periods. The coarse carbonate content of stage 1 clearly differs in this respect from the other interglaciations by showing a consistent increase to about 50 % after MIS 2 which is also much higher than, for instance, those found in MIS 5.5 and 11. Lowest coarse carbonate values occur in glacial stages 2 and 12, and in late interglacial stage 7 (values between 2 and 3%).

The linear regression between bulk carbonate content and lightness gives a correlation coefficient of r = 0.9 (Fig. IV-4a), thus confirming the results of previous studies in the North Atlantic where the sediment grey level is basically controlled by the relative proportion of carbonate content. The correlation coefficent between fine carbonate content and lightness is r = 0.91 (Fig. IV-4b), whereas the coarse carbonate to lightness correlation coefficient is only r = 0.35. This indicates that fluctuations in fine carbonate content are responsible for most of the variability in sediment lightness observed at this study site.

The present interglaciation is characterized by bulk carbonate contents of more than 80% (Fig. IV-3), but the corresponding peak lightness data reveal values of only 66%. Three comparable carbonate maxima of more than 80% are also found in MIS 5, but here the corresponding lightness values are much higher ranging between 70 and 80% (Fig. IV-3). This comparison indicates that sediment lightness is not responding to fluctuations in carbonate content alone but that additional mechanisms may also contribute to the observed lightness values. Investigations from the Nordic Seas have shown that enhanced carbonate corrosion of foraminiferal tests leads to an increase in sediment lightness (Bauch and Helmke, 1999; Helmke and Bauch, 1999). To test whether the lightness anomaly between MIS 1 and 5.5 may be linked to different degrees of carbonate corrosion, samples from MIS 1 (depth 10 and 17.5 cm) and MIS 5.5 (depth 380 and 385 cm) were closely analysed. For this purpose two pairs of samples with rather similar bulk carbonate contents and likewise similar fine to coarse carbonate ratios were selected to demonstrate the general lightness discrepancy between MIS 1 and 5 (Fig. IV-5a). In order to investigate whether carbonate corrosion is responsible for the lightness discrepancy, the dry sediment residues from the size

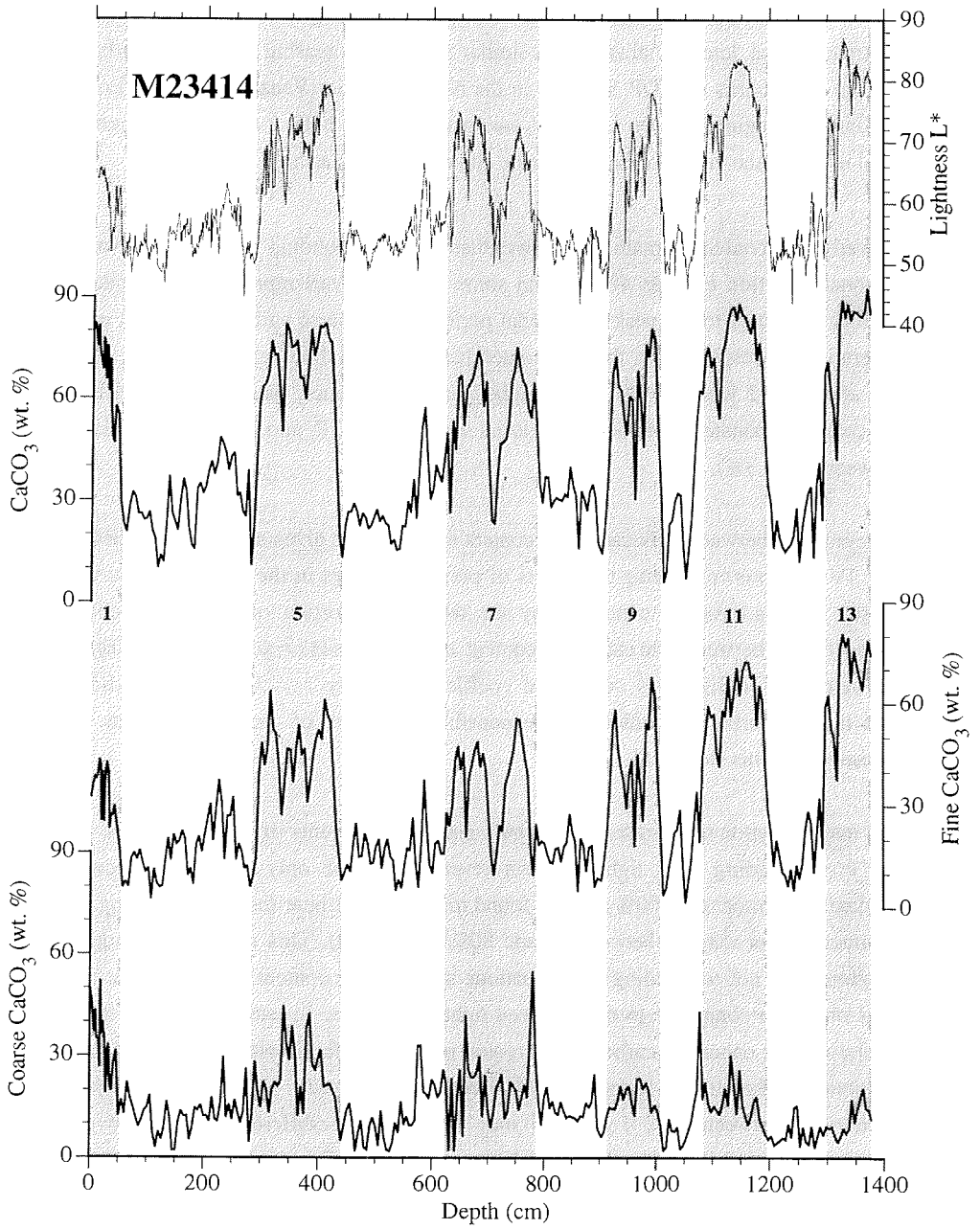


Fig. IV-3: Records of lightness, bulk as well as fine and coarse $CaCO_3$ content in composite core M23414. Interglacial marine isotope stages are shaded in grey and labeled for reference.

fractions 250-500 µm and <20 µm were analyzed and compared with reflectance analysis of 200 specimens of the planktic foraminifera *Globigerina bulloides* picked from the 250-500 µm fraction. All of the resulting lightness values clearly show a lower reflectance for MIS 1 than for MIS 5.5

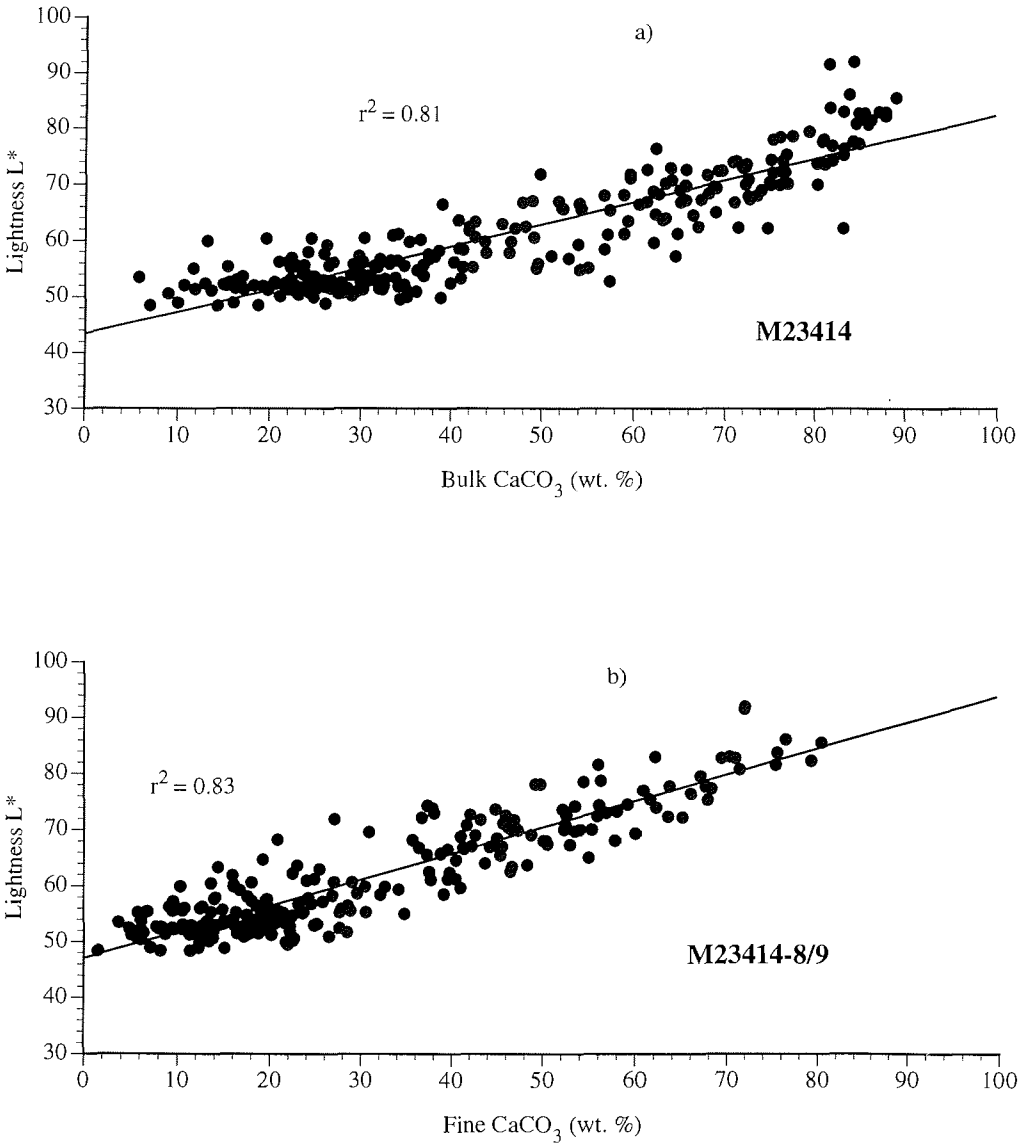


Fig. IV-4: Best fit linear regressions to illustrate the relationship between (a) sediment lightness and bulk carbonate content for core M23414 and between (b) sediment lightness and fine carbonate content for core M23414-8/9.

(Fig. 5b) regardless of the fine to coarse ratio of the carbonate component. This indicates that the higher sediment lightness values found in the latter interval were probably caused by enhanced corrosion of the carbonate component.

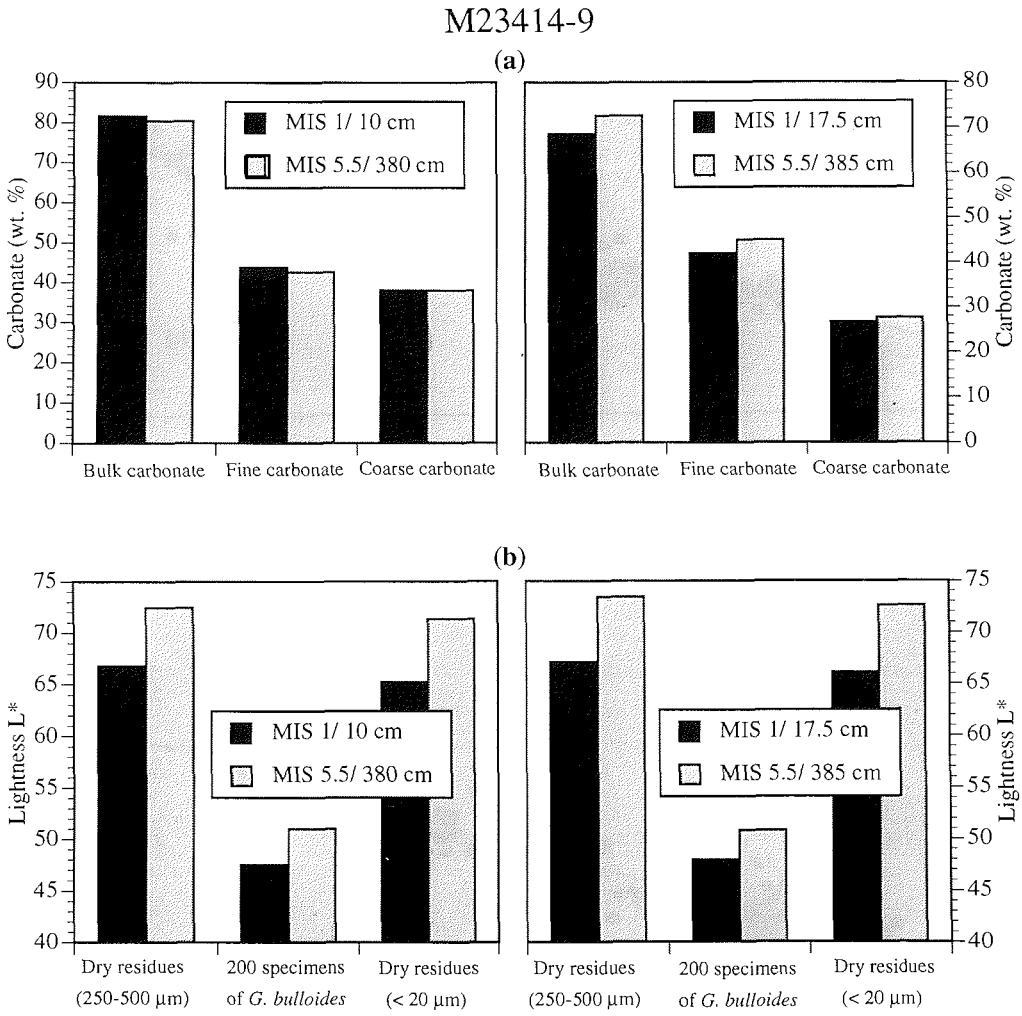


Fig. IV-5: Comparison of (a) bulk, fine, and coarse carbonate and (b) lightness values of dry sediment residues from size fractions <20 µm and 250-500 µm as well as of 200 picked specimens of *G.bulloides* (250-500 µm) for two samples from MIS 1 (depth 10 and 17.5 cm) and 5.5 (depth 380 and 385 cm).

IV-5. Discussion and conclusions

It is well known that coccolith accumulation rates in the North Atlantic are strongly coupled to glacial-interglacial climate oscillations (Ruddiman and McIntyre, 1976). In the North Atlantic peak interglacial periods typically correspond to high fine carbonate values, whereas glacial and cold interglacial/interstadial periods correspond to intervals of low fine carbonate (also van Krefeld et al., 1996; Lotoskaya et al., 1998). However, in core M23414 the fine carbonate on average dominates the bulk carbonate content in interglacial and glacial intervals (Fig. IV-6). Except for a few core sections, variations in the bulk carbonate are therefore controlled by fluctuations in the fine carbonate content. This is corroborated by other studies from further south in the northeastern North Atlantic which also show that fine carbonate contents generally dominate the bulk carbonate content during glacial as well as interglacial periods (Lotoskaya et al., 1998). As it can be assumed that the fine carbonate component is mainly composed of coccoliths, temporal changes in North Atlantic sediment lightness records are thus strongly coupled to variations in coccolith contents.

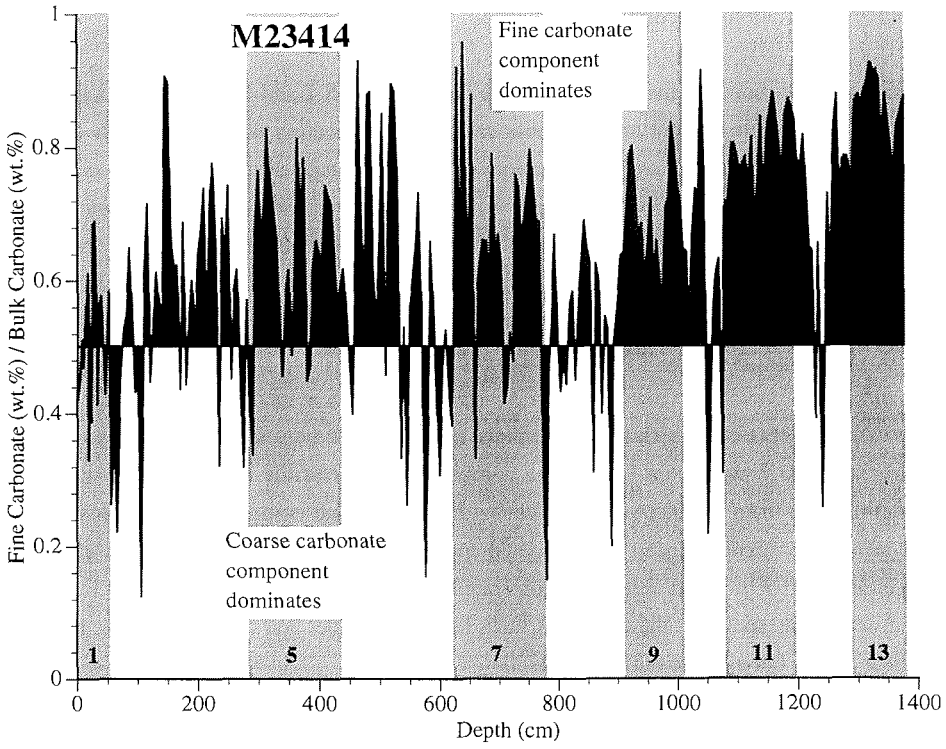


Fig. IV-6: Ratios of fine carbonate content to bulk carbonate content, indicating the depth sections where the bulk carbonate content of M23414 is dominated by the <20 µm fraction (values >0.5) and the >20 µm fraction (values <0.5), respectively. Interglacial marine isotope stages are shaded in grey and labeled for reference.

Fluctuations of biogenic carbonate that are driven by glacial-interglacial climate change, i.e., the south- and northward movement of polar water masses (Ruddiman and McIntyre, 1976), are clearly imprinted on North Atlantic sediments. Due to varying inputs of clastic sediments via iceberg-rafting during cold periods, alternating with time intervals showing high carbonate bioproductivity and no IRD deposition, the North Atlantic sediments have an excellent potential for high-resolution color analyses. Since in this region sediment color is basically controlled by productivity changes of carbonate-bearing microfossils, several authors have used reflectance values to estimate carbonate content (Harris et al., 1997; Ortiz et al., 1999) and to infer climatic changes (Cortijo et al., 1995).

Recent investigations (Balsam et al., 1999) indicate that changes in the non-carbonate sediment fraction may significantly alter the reflectance of sediment sequences with constant carbonate content. The content of total organic matter in core M23414 does not exceed 0.5%. Moreover the relative proportions of additional non-carbonate material deposited at the study site did not change significantly during the investigated time interval, indicating that such sediment components have a minor influence on variations in sediment reflectance.

Our comparison of sediment lightness and carbonate data from MIS 1 and 5.5 shows a stage 1/5.5-anomaly between carbonate content and lightness values, also indicating that the lightness changes are not necessarily directly controlled by fluctuations in total carbonate content alone. Based on the analysis of fine and coarse carbonate, it is evident that MIS 1 has the highest content of foraminifera of all peak interglaciations. What this means in terms of paleoceanography remains unclear at present, since it would require further investigations with different methods. However, the MIS 1/5.5 comparison suggests that the differences in sediment reflectance are not necessarily attributed to changes in the relative proportions of fine to coarse carbonate.

One other possible mechanism which may explain the observed sediment lightness discrepancy is a difference in carbonate preservation. An earlier comparison between the reflectance of sediments and foraminiferal tests from sediment sections having high carbonate contents has indicated that an increase in sediment reflectance can be attributed to enhanced corrosion of foraminiferal tests (Bauch and Helmke, 1999). It was further demonstrated that increased corrosion leads to changes in the shell microstructure, such that the crystal surfaces become rougher and opaque, making not only the tests appear whiter but also the entire sediment.

During the late Pleistocene the North Atlantic switched from a deep convection mode (production of NADW = North Atlantic Deep Water) during interglacial periods to an intermediate one (production of GNAIW = Glacial North Atlantic Intermediate Water) during glacial times (Oppo and Fairbanks, 1987; Venz et al., 1999). Throughout glacial times the boundary between GNAIW

and SOW (Southern Ocean Water) is suggested to have been near 2200 m (Zahn et al., 1987; Oppo et al., 1997), indicating that during these periods the study site was located at the boundary between two water masses. As a consequence, the core location was characterized by well ventilated water masses during interglacial times. This suggests that the observed variations in interglacial carbonate corrosion are more likely caused by some regionally limited changes in carbonate preservation due to variations in pore- and bottom-water chemistry (e.g. remineralization of organic matter).

Although most investigations of corrosional effects on calcite shell microstructure were carried out on foraminifera (e.g., Henrich, 1986, 1989), the manner in which foraminifera and coccoliths are transported downwards to the sea floor is rather different (Honjo, 1982; Pilskaln and Honjo, 1987). Because coccoliths are encapsulated into coproliths and thus buried, corrosional effects on the sea floor may differ from those of foraminifera. However, the lightness measurements on the coarse and fine fraction as well as on the foraminifera show a similar trend between MIS 1 and 5.5 (Fig. IV-5b), suggesting that corrosion affects both coarse and fine carbonate components. The decreasing trend observed in the fine carbonate during interglacial intervals since MIS 13 - the values in MIS 5 being the exception - may thus be caused by some unknown chemical changes rather than the result of variations in carbonate productivity. Recognizing this effect of carbonate corrosion on total sediment lightness may therefore be important when interpreting glacial-interglacial records for paleoceanographic purposes.

Chapter V - Sediment-color record from the Northeast Atlantic reveals patterns of millennial-scale climate variability during the last 500,000 years

V.1. Abstract

A 500,000-year long deep-sea sediment color record from the Northeast Atlantic was investigated to reconstruct the evolution of late-Pleistocene climate variability on millennial time scales. Variations of the red-green color intensity are probably caused by climatically induced changes in the ice-rafted input of red-colored, iron-bearing terrigenous material to the core site. The resolution of the age model impedes the detection of distinct spectral features at sub-Milankovitch periods. Hence, millennial-scale climate variability is quantified as time-dependent variance of the high-pass filtered color time series. The course of the estimated variance shows distinct patterns, which can be linked to continental ice mass. During the past 500,000 years, large-amplitude millennial-scale climate variability occurs only if continental ice mass exceeds a threshold level, equivalent to sea level at approximately 40% of the drop during the last glacial maximum and is tied to times of change in ice mass. For sea level below the threshold, reduced climate variability is associated with short time intervals of relatively stable phases of sea level.

V.2. Introduction

Climate proxy data from various archives document large-amplitude climate fluctuations during the last glacial period on millennial time scales: On longer time scales these include "Heinrich events" (e.g. Bond et al., 1992) with an average recurrence time of approximately 7 thousand years (ky) (Sarnthein et al., 2001) and on shorter scales "Dansgaard-Oeschger events" (Dansgaard et al., 1993; Grootes and Stuiver, 1997) with an average spacing of about 1.5 ky (Bond et al., 1997, 1999). Heinrich events are thought to be linked to internal oscillations of the Laurentide ice sheet (MacAyeal, 1993), whereas the origin of the Dansgaard-Oeschger events remains controversial, ranging from internal oscillations of the ocean-atmosphere system (Broecker et al., 1990; Winton, 1993; Sakai and Peltier, 1997) to periodic calving of the Greenland ice sheet (van Kreveld et al., in press), and to external forcing mechanisms (van Geel et al., 1999; Keeling and Whorf, 2000). Furthermore, it was suggested that millennial-scale climate variability, although with smaller amplitude, is not restricted to the last glacial period but is a pervasive feature of the recent interglacial, the Holocene (Bond et al., 1997; deMenocal et al., 2000).

The possible interference of anthropogenic climate perturbations with natural climate fluctuations on millennial time scales poses a challenging task for predicting future climate change. Previous studies demonstrated the suitability of color records from marine sediments to quantify millennial-scale climate change with high temporal resolution (Cortijo et al., 1995; Ortiz et al., 1999). However, marine sediment color records have so far not been used to document the evolution of millennial-scale climate fluctuations throughout the Late Quaternary. To gain better insight into these natural variations we analyse high-resolution color records from Northeast Atlantic deep-sea sediments, which document climate variability down to millennial time-scales over the last five glacial-interglacial cycles.

V.3. Data and stratigraphy

Sediment color measurements of composite core M23414 from the Rockall Plateau in the Northeast Atlantic (53°32´N, 20°17´W, 2200 m water depth; details of core splicing in Helmke and Bauch, in press.) were carried out at centimeter steps, using a hand-held Minolta CM 2002 spectrophotometer. We use lightness L* and red-green chromaticity a* of the spherical L*a*b* color space (b* denotes blue-yellow chromaticity). Precision of the downcore color measurements was investigated using a set of 10 repeated measurements at 3 spots of the core. The mean standard deviation for these measurements is 0.06 for L* and 0.01 for a* resulting in typical signal-to-nosie ratios of ~500. Additionally, the lightness and red-green color variations of 2 separate tracks down the core were measured. The differences are negligible with correlation coefficients of r = 0.98 (L*) and r = 0.97 (a*). These results clearly indicate that the spectrophotometer color data can faithfully be used

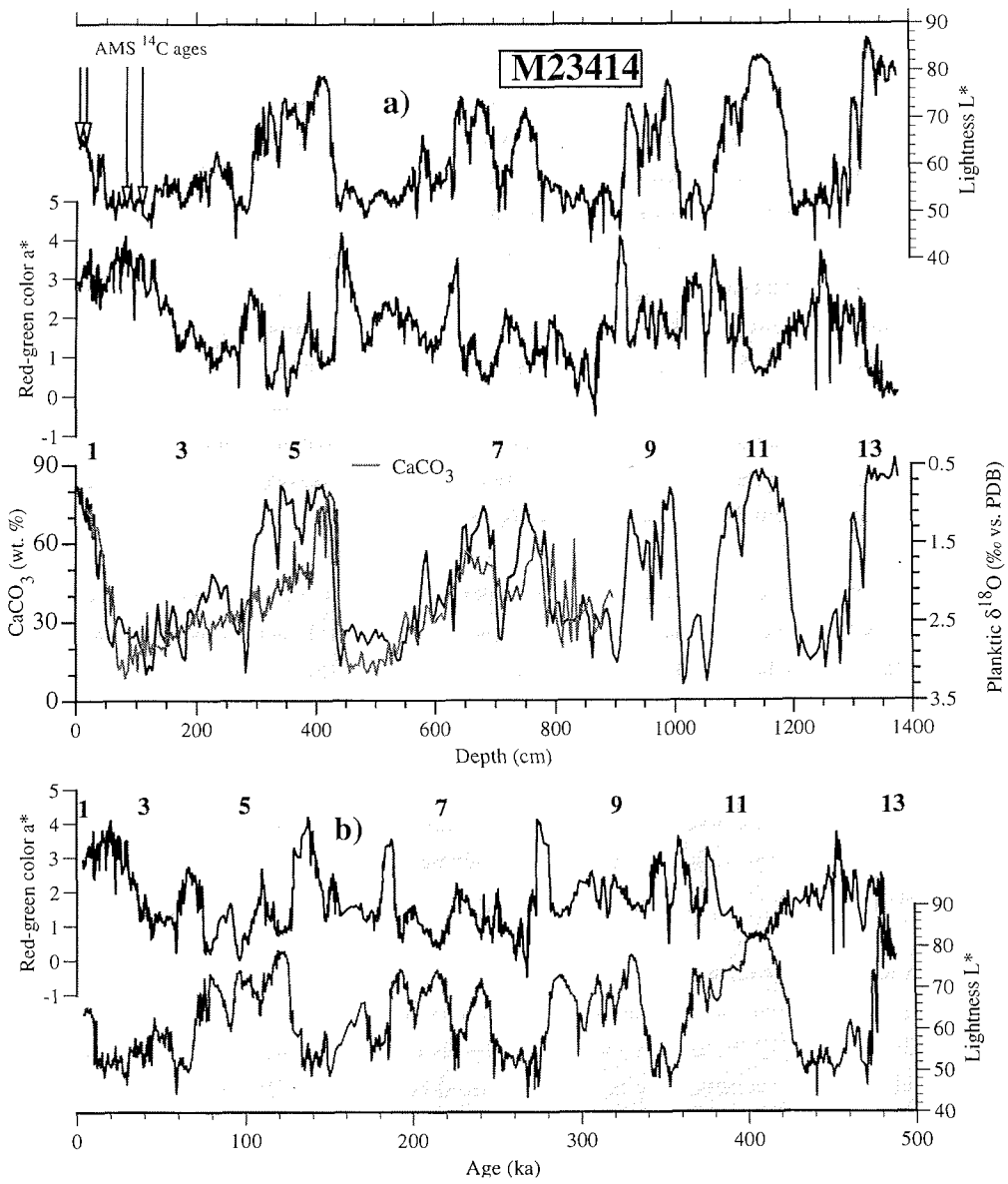


Fig. V-1: (a) Downcore records of lightness L*, red-green chromaticity a*, and $CaCO_3$ content (weight % of bulk sediment) from composite core M23414. Oxygen isotopes values of planktic foraminifer *Globigerina bulloides* from M23414-9 (gray line) cover only the upper 900 cm. $CaCO_3$ content and oxygen isotopes were measured at average sample resolutions of 5 cm (Helmke and Bauch, in press.) and 2.5 cm (Jung, 1996), respectively. Arrows mark depth sections with AMS ^{14}C ages. Odd numbers indicate marine isotope stages for reference and are shaded in gray. (b) Red-green chromaticity a* (top) and lightness L* (bottom, gray line) *vs.* age. The age model is based on the spliced SPECMAP time scale (Imbrie et al., 1984; Martinson et al., 1987).

to determine the downcore color fluctuations at the study site. Sediment lightness is affected by several processes such as carbonate productivity and preservation, concentrations of organic matter and ice-rafted debris (Bond et al., 1999). Red-green chromaticity in North Atlantic deep-sea sediments is primarily controlled by the deposition of iron- and manganese-bearing components (Nagao and Nakashima, 1992). All these effects on sediment color are closely linked to climate variations, hence, changes in sediment color at this core location can be used as climate proxy.

The good agreement between planktic oxygen-isotope values, lightness L* and carbonate concentrations over the last 250 ky indicates that the latter two proxies can be used with confidence to identify the glacial and interglacial periods for the entire record (Fig. V-1a). Ages were assigned to downcore L* and a* color data by synchronizing the lightness record to the standard SPECMAP chronology (Imbrie et al., 1984; Martinson et al., 1987), and assuming constant sedimentation rates between the correlated stratigraphic levels (Fig. V-1b). Four AMS ^{14}C ages and published ages of six Heinrich events (Sarnthein et al., 2001) are included to refine the age model in the uppermost part of the record. The ^{14}C ages were converted into calendar ages using the software CALIB 4.1.2 (Stuiver et al., 1998) and following Voelker et al. (1998). Heinrich events were identified based on the concentration of lithic grains in the size fraction >250 μm (Didié and Bauch, 2000).

V.4. Quantifying climate variability

The visual alignment of the L* record to the SPECMAP timescale rests on the identification of congruent patterns in both records. In the SPECMAP time series, the shortest patterns are associated with the precessional cycle of the Earth's orbit, having a periodicity of about 21 ky (Imbrie et al., 1984). Since it is possible to identify and align half-precessional cycles in both records, the accuracy of the resulting age model will at least be half the period of the precessional cycle (relative to the SPECMAP chronology). On the other hand, the centimeter-sampled color data have an average resolution of approximately 360 yr (sample resolution ranges between 70 yr and 2 ky), making them, in principle, suitable to resolve millennial-scale climate variability, i.e., variability between 10^3 and $< 10^4$ years. Spectral analysis is commonly employed to detect and quantify the contribution of millennial-scale variance to the total variance of the series under consideration. However, the uncertainty of the age model used makes it unlikely that any distinct spectral peaks in the millennial band, which may have been recorded by the time series, will show up in a spectrum.

To nevertheless quantify climate variability in the millennial band, we first filtered the red-green color time series with a high-pass filter (Rybicki and Press, 1995) using a cut-off frequency of 1/12 ky^{-1}, which is approximately half the precessional period, i.e., the shortest patterns in the SPECMAP time series (Fig. V-2). The filtered time series thus reflects climate variability from millennial down

to multi-centennial time scales (limited by the average temporal resolution of the color record). In a second step, the time-dependent variance of the filtered time series was estimated by means of a 8-ky wide sliding rectangular window. This width offers a good compromise between statistical and

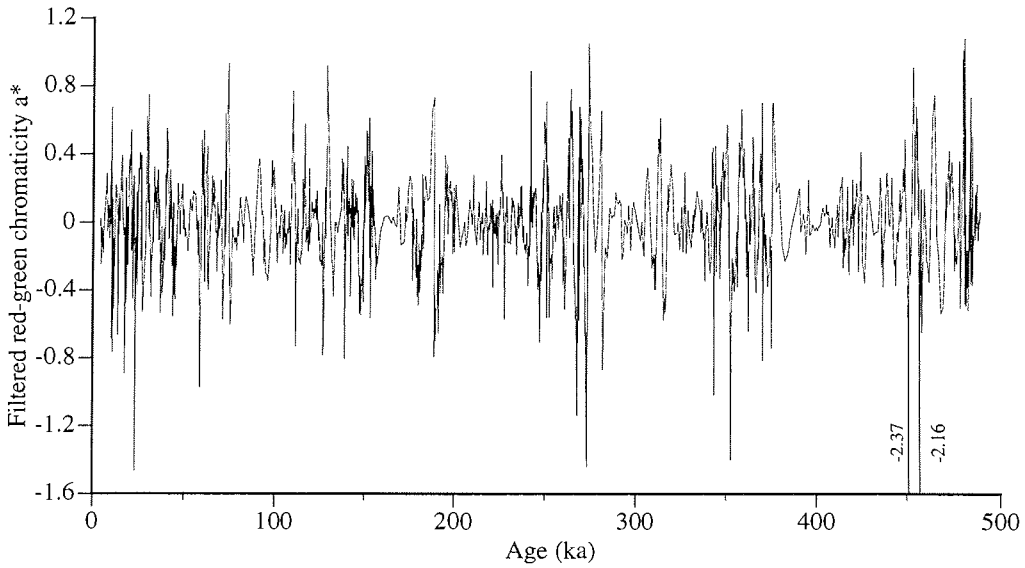


Fig. V-2: High-pass filtered red-green color time series (see Fig. V-1b). Cutoff frequency is 1/12 ky^{-1}.

systematic errors. Since variance and the area under a spectrum measure the same quantity (Parseval's theorem; Bendat and Piersol, 1986), this approach indirectly estimates spectral power in the entire millennial-to-multi-centennial band as function of time. Both computational steps were performed on the unevenly spaced time series to prevent any bias in spectral character caused by interpolation to an even sampling interval (Horowitz, 1974; Schulz and Stattegger, 1997).

V.5. Patterns of climate variability

Time-dependent variance of the red-green color shows that millennial-scale fluctuations with different amplitudes persisted during the past half a million years. It is most likely that variations in the concentration of red-colored, iron-bearing minerals are responsible for the observed changes in the red-green chromaticity of the sediment. Several studies at sites in the vicinity of M23414 have indicated that the deposition of a red-colored mineral, i.e., hematite, in the Northeast Atlantic during the last 40 ky is linked to millennial-scale climate variations (Bond and Lotti, 1995; Bond et al., 1997, 1999). These authors proposed that hematite-stained lithic grains stemming from continental red beds were transported via ice-rafting to the core sites. X-ray-flourescence analysis of samples

from M23414 prove that iron content correlates with changes in the a* color record, i.e., a higher iron content goes along with higher a* values and vice versa (Fig. V-3a).

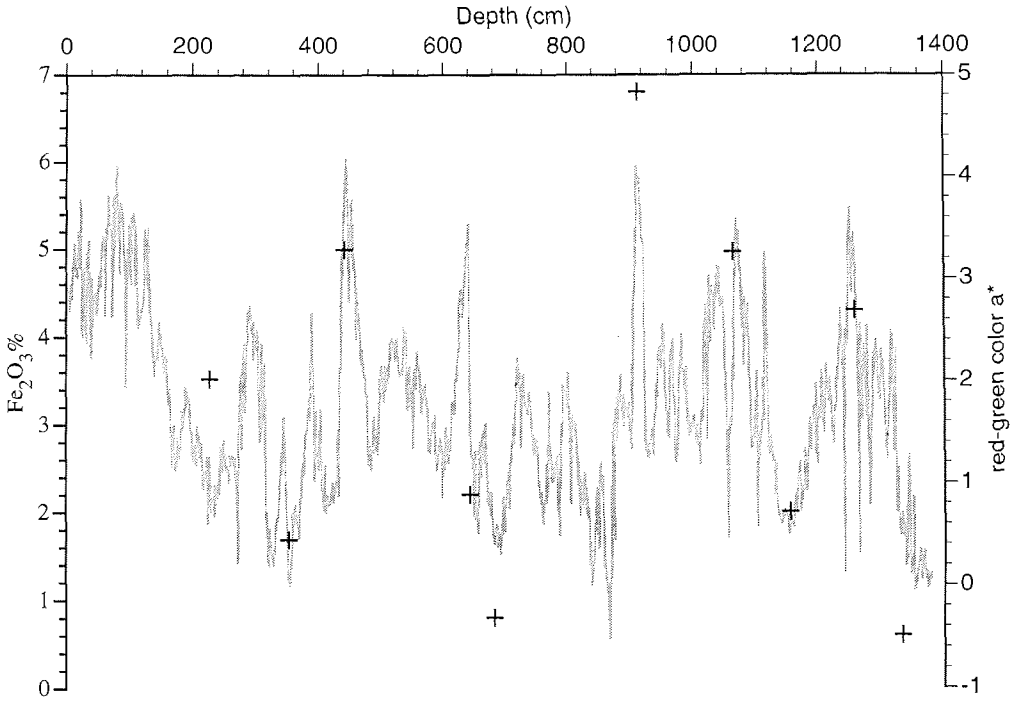


Fig. V-3a: Red-green color (gray line) and corresponding Fe_2O_3 content (weight % of bulk sediment) for a series of 10 sediment samples (data points denoted as +) from M23414.

The correlation is significant on the basis of Spearman´s rank-order correlation coefficient (ρ = 0.89, $p < 0.001$, n = 10; Fig. V-3b). However, iron concentrations are too low to identify iron-bearing minerals using the X-ray-diffraction method.

A comparison of red-green color and ice-rafted debris (IRD) reveals a generally good agreement between the patterns of variability (Fig. V-4). Especially the younger part of the record (Fig. V-4b) strongly suggests that changes in a* are linked to changes in the concentration of IRD. However, a dependance of red-green color from the concentration of IRD cannot be verified for the entire record (Fig. V-4a). This may be due to various sources of IRD input on the Rockall Plateau during the last 500 ky. A correlation of red-green chromaticity and total organic matter (TOC) shows a very weak correlation coefficient (ρ = 0.23, $p < 0.001$, n = 268), thus it is unlikely that organic matter should have a strong influence on the red-green color record. Without knowledge about the pore

water chemistry at the study site an influence of sediment geochemistry on the red-green color record cannot be ruled out completely. However, an oxidation of sediment components containing Fe^{3+} into red-colored components containing Fe^{2+} during intervals of low sediment accumulation seems to be unlikely: A comparison of accumulation rates and a* during the last 60 ky indicates that a* is not linked to changes in accumulation rate (correlation is not significant: $\rho = -0.05$, $p = 0.911$, $n = 8$).

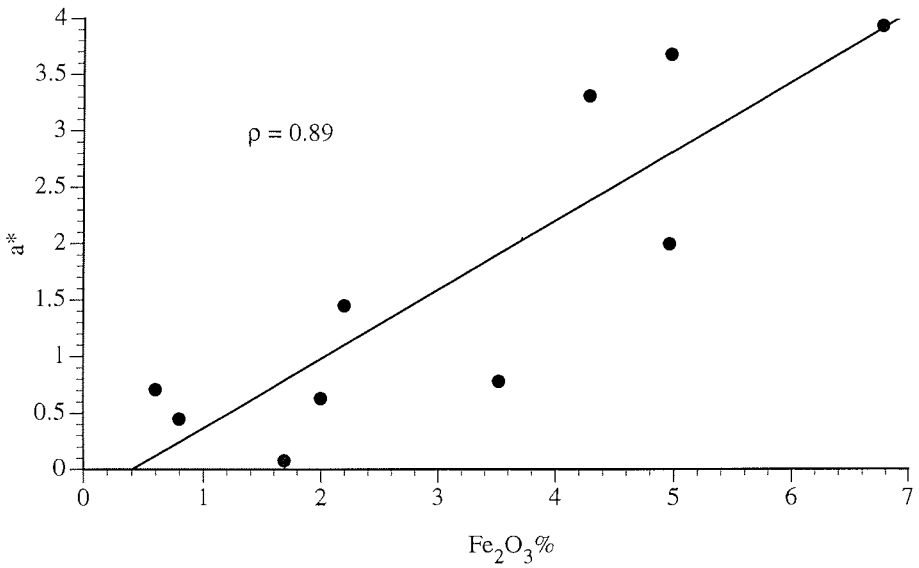


Fig. V-3b: Correlation of Fe_2O_3 (weight % of bulk sediment) and red-green chromaticity a* for a series of 10 sediment samples from M23414.

Hence, we surmise that variations in red-green chromaticity over the last 500 ky are mainly controlled by fluctuations in the input rate of reddish iron-bearing minerals into the sediment, which in turn is linked to climate variations. In contrast, the interpretation of the lightness record is hampered by the fact that sediment lightness is affected by several mechanisms that may even counterbalance each other during the course of a climatic perturbation (Bond et al., 1999). To avoid ambiguous interpretations, we restrict our analysis to the red-green time series.

The comparison of the time-dependent variance of the red-green color record with the SPECMAP oxygen-isotope record, which is a proxy for continental ice mass and, therefore, global sea level (Imbrie et al., 1984), reveals that nine variance maxima occurred during time intervals of ice-sheet

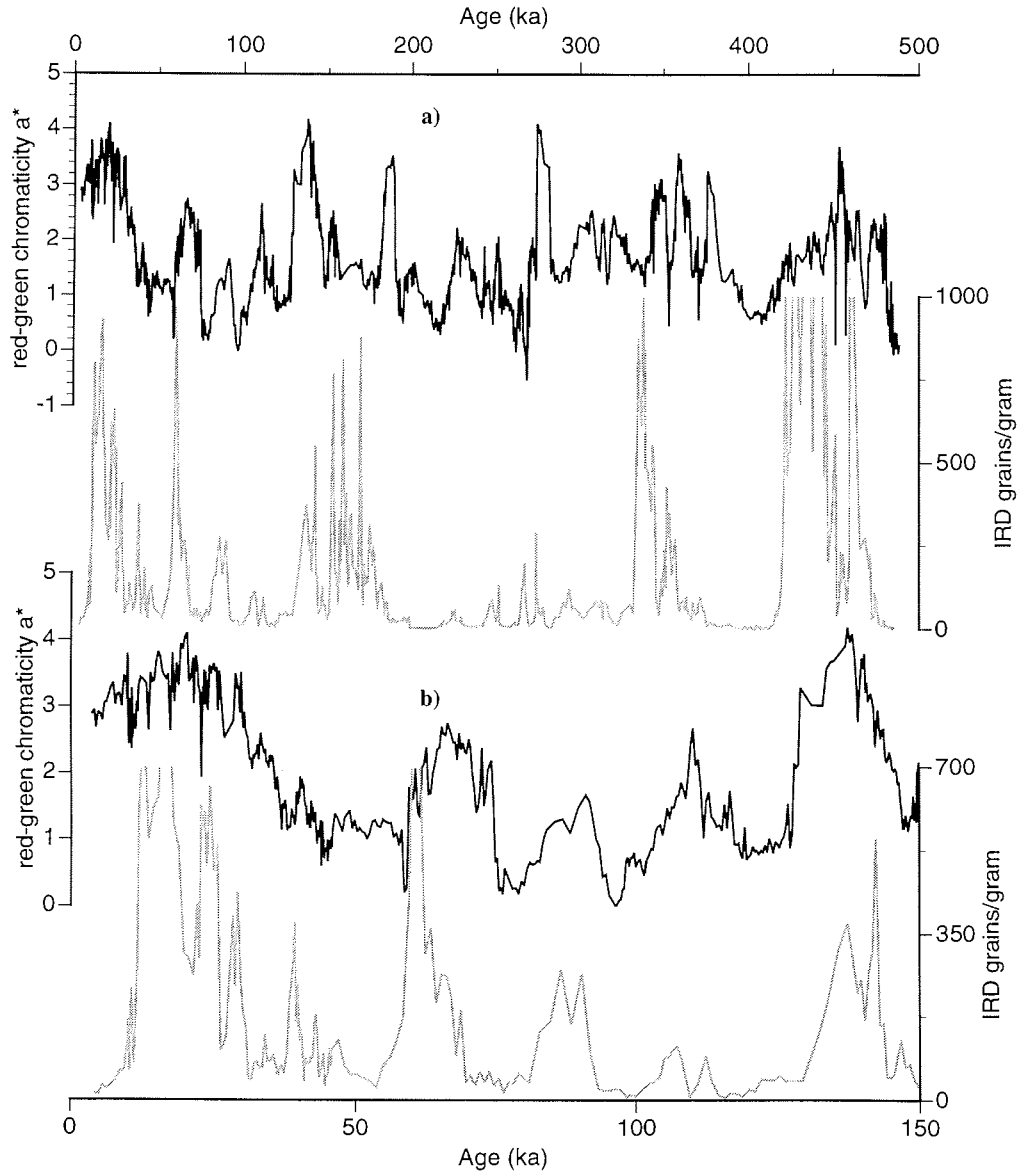


Fig. V-4: (a) Records of red-green chromaticity a* and IRD (lithic grains per gram) from M23414 *vs.* age. (b) Same records than in (a) for the upper 150 ky. Note that IRD is cut off at a) 1000 and b) 700 lithic grains per gram.

growth, while three maxima go along with times of ice-sheet decay (Fig. V-5). Moreover, these variance maxima coincide with times when sea level was 40% of ist maximum drop during the last

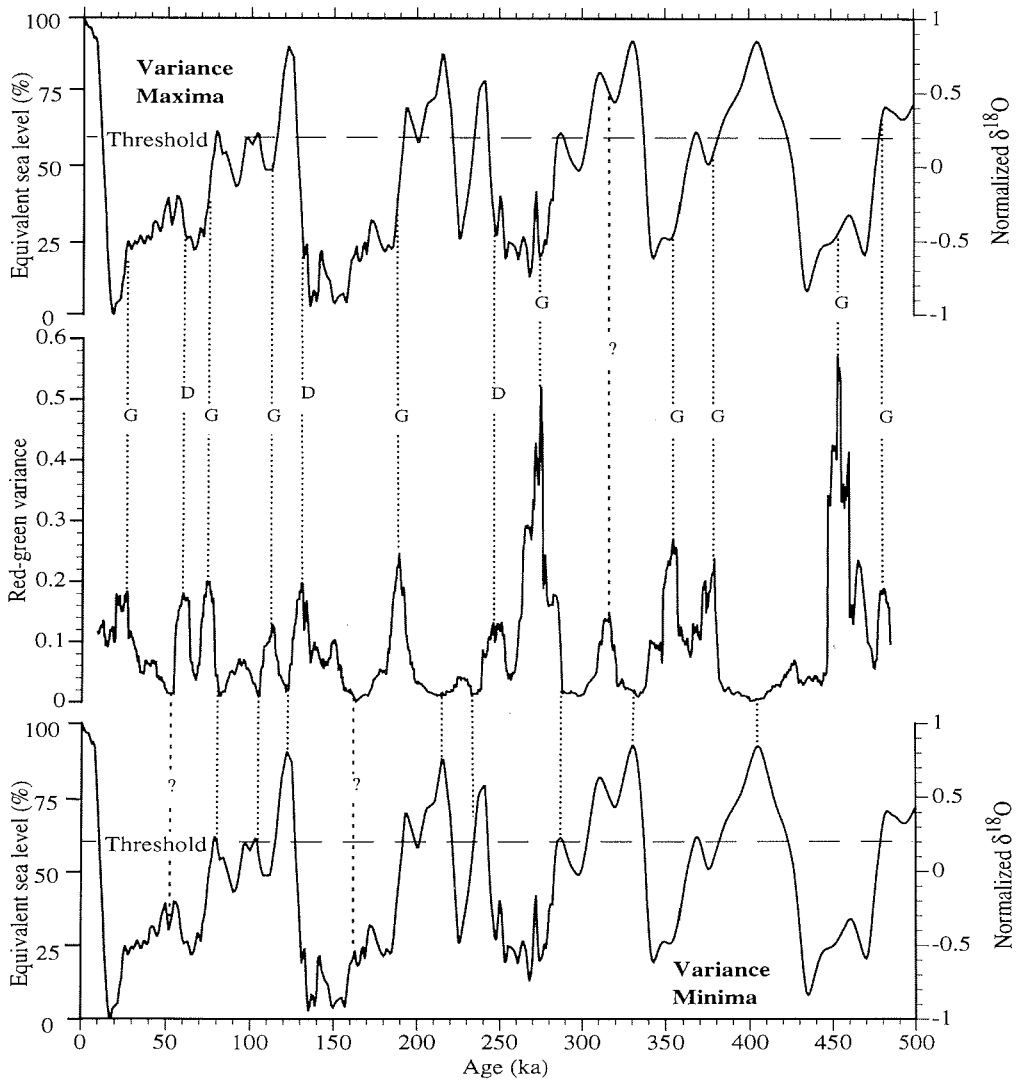


Fig. V-5: Ties between millennial-scale variance (centre) of the red-green color time series and sea level (%) (top/bottom). Sea-level (%) is derived from the normalized SPECMAP (Imbrie et al., 1984; Martinson et al., 1987) oxygen-isotope record (top/bottom). Variance maxima are aligned to the upper panel, variance minima to the lower one. Variance maxima coincide with times of either sea-level rise (ice decay; denoted by D) or sea-level drop (ice growth; denoted by G), as indicated by the dotted lines between the two upper records. An exceptional maximum occurs at 315 ka (short dashed line), when SPECMAP-derived sea level is probably overestimated (see text for details). Variance minima are tied to times with sea level above a threshold of approximately -40% (long dashed line), as indicated by the dotted lines between the two lower records. Two apparent exceptions occur at 53 and 164 ka (short dashed lines), when SPECMAP-derived sea-level is probably underestimated (see text for details).

glacial maximum, i.e., -50 ± 4 m (Fig. V-5), indicating threshold behavior. The range for this threshold results from the uncertainty of sea level at the last glacial maximum (between 121 ± 5 m, Fairbanks, 1989 and 130-135 m, Yokoyama et al., 2000) and are in accordance with those derived from ice-core data for the last 100 ky (Schulz et al., 1999). The variance maximum at 315 ka is seemingly at odds with the inferred SPECMAP based threshold. However, according to Berger et al. (1996) sea-level at this time may have been 18 m lower than indicated by SPECMAP, bringing this variance maximum in line with our estimated threshold value. Of the five glacial terminations only Terminations III and II, the glacial-to-interglacial transitions from marine isotope stages 8 to 7, and 6 to 5, respectively, are characterized by variance maxima. Although this suggest different dynamical behavior across glacial-to-interglacial transitions, future studies are needed to confirm this pattern.

Prominent variance minima, indicating low-amplitude millennial-scale climate variability, are generally associated with peak interglacials (Fig. V-5), that is, intervals with small continental ice mass. This agrees with previous studies, which have suggested that a minimum amount of continental ice is required to sustain prominent climate fluctuations on millennial time scales (Oppo et al., 1998; McManus et al., 1999; Schulz et al., 1999). The two variance minima at 53 and 164 ka seem to contradict our estimated threshold (Fig. V-5). However, radiometrically dated corals (Chappell et al., 1996) indicate that sea level at ~ 50 ka was indeed in the range of our threshold. If the same holds true for the relatively warm interval within glacial marine isotope stage 6 at 164 ka remains speculative but might deserve further investigations. During glacial maxima, millennial-scale climate variability is low, although higher than during peak interglacials, which is consistent with the observations of McManus et al. (1999) and Schulz et al. (1999).

V.6. Conclusions

A red-green color record from Northeast Atlantic deep-sea sediments reveals variations in the amplitude of millennial-scale variability during the last 500 ky. Changes in red-green color intensity are probably linked to varying input of reddish terrigenous material by ice rafting, which in turn is closely connected to climate variations on sub-Milankovitch time scales.

Using sea level as proxy for continental ice mass, our results indicate that large-amplitude climate variations are restricted to times when continental ice volume exceeds a threshold, equivalent to sea level at approximately 40% of ist last glacial maximum value (or -50 ± 4 m). This value is almost indistinguishable from an earlier estimate, derived for the last 100 ky based on ice-core data. It is remarkable and favorable for future modelling studies that a single threshold value apparently controlled the cessation of large-amplitude millennial scale climate variations over the past five glacial-to-interglacial cycles.

Once continental ice mass exceeds the threshold value, maxima of millennial-scale climate variability are tied to times of changes in ice mass. In contrast, relatively stable ice volume during glacial maxima goes along with considerable dampening of climate variability. The inferred relationship between ice volume and "agility" of the climate system to undergo short-term fluctuations holds throughout the past 500 ky.

Chapter VI - Comparison of glacial and interglacial conditions between the polar and subpolar North Atlantic Region over the last five climatic cycles

VI.1. Abstract

A multiparameter analysis of two sediment cores from the Northeast Atlantic and the Nordic Seas, i.e., the subpolar and polar North Atlantic, suggests that the Nordic Seas have experienced only three intervals of full interglacial conditions during the last 500,000 years, i.e., marine isotope stage (MIS) 11, 5e, and 1. In contrast, the Northeast Atlantic has experienced such conditions more frequently. A comparison of planktic $\delta^{18}O$ from the Northeast Atlantic reveals colder SSTs during MIS 11 than during other peak warm periods (MIS 9, 5e, and 1). The IRD record from the Nordic Seas points at a major temperature gradient in the polar North Atlantic and a reduced advection of warm surface water into the Nordic Seas during interglacial MIS 11. Maximum input of ice-rafted debris (IRD) during the final phase of glaciations and into early terminations was almost coherent between both sites indicating that the subpolar and polar North Atlantic were synchronously affected by the mechanisms that triggered a persistent variability of the glacial climate system. Millennial-scale recurrence periods of IRD maxima at the study areas point at differences in the timing of iceberg discharges between the Laurentide Ice Sheet and European ice sheets.

VI.2. Introduction

A large portion of more recent paleoceanographic research using high northern sediment cores investigated the climatic and environmental conditions of specific glacial and interglacial periods focussing, in particular, on the last glacial period (e.g., Labeyrie et al., 1992; Sarnthein et al., 1995; Fronval and Jansen, 1996; Bauch, 1996; McManus et al., 1998). Comparative analyses of North Atlantic marine sediment records and Greenland ice cores demonstrated that this last glaciation was not a stable climatic regime but punctuated by two modes of abrupt millennial-scale climate change, the so-called "Heinrich events" and "Dansgaard-Oeschger cycles" (Broecker et al., 1992; Bond et al., 1992; Dansgaard et al., 1993). Some of the cold events seem time-coeval between the subpolar North Atlantic and the Nordic Seas, i.e., the polar North Atlantic, even for longer time intervals such as the last interglacial-glacial cycle (McManus et al., 1994; Fronval and Jansen, 1997). But so far, no detailed comparison of climatic conditions for glacial and interglacial periods older than the last peak interglaciation, marine isotope stage (MIS) 5e, has been done for these two regions (Cortijo et al., 1994; Fronval and Jansen, 1997; Bauch et al., 2000b).

Nearly 25 years ago, a first paleoclimatic reconstruction of the Nordic Seas suggested that during the last 450,000 years only MIS 5e and 1 were characterized by a similar strength of warm Atlantic surface water inflow into the Norwegian and Greenland Seas resulting in comparable interglacial conditions in this region (Kellogg, 1976, 1977). More recently, comparisons of peak interglaciations in the Nordic Seas revealed significant differences in the overall water mass pattern, for instance colder conditions during MIS 11 when compared to MIS 5e and 1, suggesting that boundary conditions of peak interglacial climates must have been dissimilar throughout the upper Quaternary (Bauch et al., 1999; 2000a). In contrast, sea-surface temperatures (SST) estimates from about 55°N indicated that interglaciations with comparable peak warm conditions prevailed more frequently in the subpolar North Atlantic over the last five climatic cycles (Ruddiman et al., 1986; McManus et al., 1999).

Taking all the published evidence together implies that major differences in the surface ocean conditions of peak interglacial periods must have existed between the polar and subpolar North Atlantic Region. The purpose of this study is, therefore, to determine the climatic conditions of glacial and interglacial periods over the last five climate cycles in order to unveil possible synchroneities or asynchroneities in the behavior of the glacial-interglacial climate system at high northern latitudes. Using a multiparameter approach, sediment cores were selected that are located in oceanographically crucial areas of the Northeast Atlantic and the Nordic Seas.

VI.3 Core material and methods

Site M23352 from the northwestern Iceland Plateau in the Nordic Seas and Site M23414 from the southern edge of the Rockall Plateau in the Northeast Atlantic were investigated for this study (Fig. VI-1). Site M23352 was selected, as it is located close to the recent Polar Front, which is expected to

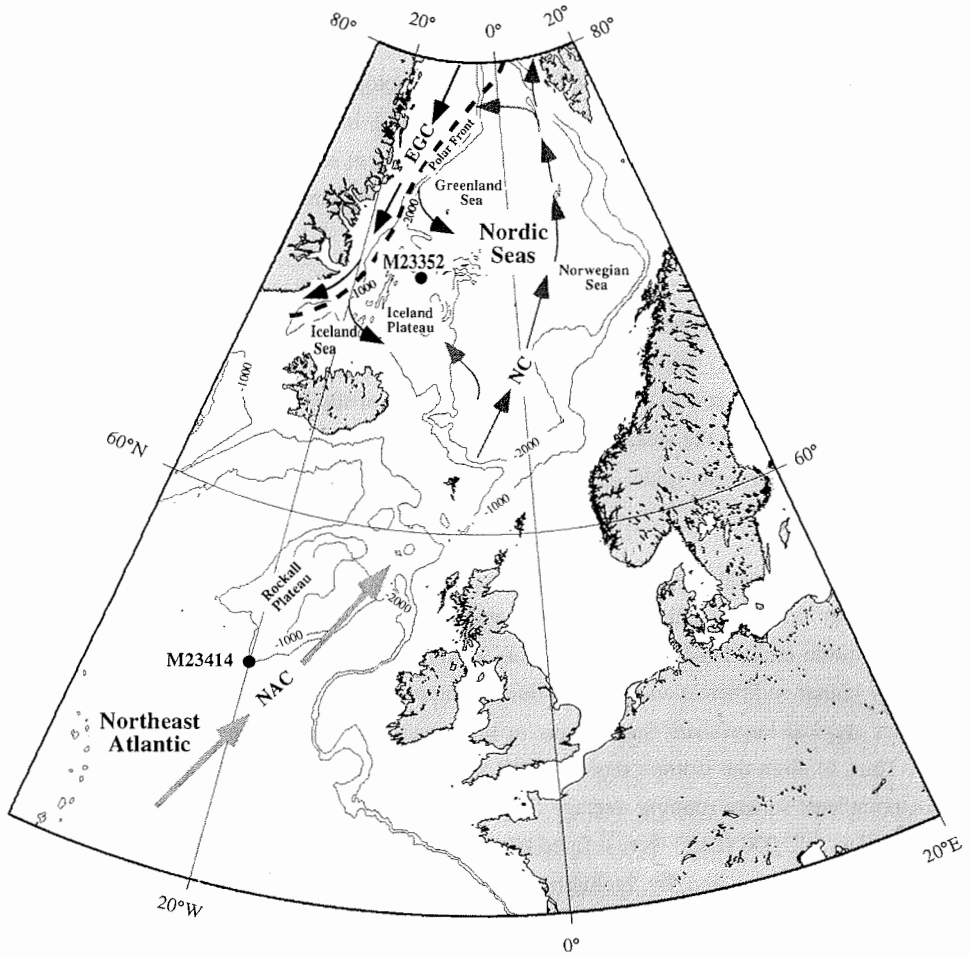


Fig. VI-1: Geographical position of study sites on the Iceland Plateau (M23352: spliced record of trigger box core M23352-2 and kasten core M23352-3; 70°00´N, 12°25´W, water depth 1820 m) and the Rockall Plateau (M23414: spliced record of trigger box core M23414-6, kasten core M23414-9 and gravity core M23414-8; 53°32´N, 20°17´W, water depth 2200 m) and simplified modern surface water circulation of the Northeast Atlantic and the Nordic Seas (Swift, 1986): Gray and black arrows denote warmer and cooler surface currents, respectively; NAC (North Atlantic Current); NC (Norwegian Current), EGC (East Greenland Current). Dashed black line indicates recent position of the Polar Front in the Nordic Seas. Water depth is in m.

have shifted repeatedly over the Iceland Plateau and further south during colder climatic intervals than at present. Accordingly, the Iceland Plateau sediment record should reveal the contrasts of glacial-interglacial climatic conditions in the polar North Atlantic. Site 23414 is located directly below the North Atlantic Current. This region has demonstrated its suitability for recording both the warm water advection that is mostly pronounced during the warmer climate periods and the major expansions of iceberg drift that characterizes extreme conditions during glacial times.

The stratigraphic frameworks are based on stable oxygen isotope records measured on the planktic foraminifers *Neogloboquadrina pachyderma* sinistral (sin.) in core M23352 (between 1-3 cm steps) and on *Globigerina bulloides* in core M23414 (at 2.5 cm steps). All stable isotope analyses on these two cores were conducted using multiple-specimen samples and standard laboratory techniques (Bauch et al., 2000b).

Sedimentologic investigations were carried out by various means and at different depth intervals. The Sediment color was measured at discrete 1 cm steps using a Minolta CM-2002 spectrophotometer and the results are given as lightness L* of the L*a*b* color space. Ice-rafted debris (IRD) was counted in the mesh size > 250 μm with an average sample interval of 2.5 cm, and is expressed as lithic grains per gram. The bulk carbonate content (weight %) was measured every 5 cm.

Age models were constructed by synchronizing the planktic oxygen isotope record of M23352 and M23414 together with the sediment lightness record of M23414 to the standard SPECMAP chronology (Imbrie et al., 1984; Martinson et al., 1987). Massive IRD input in MIS 11-12 and a consequently higher sedimentation rate than during intervals with enhanced pelagic sedimentation led to rather atypical large-scale fluctuations in the lowermost part of the isotope record at site M23352. Thus, to align the isotope record to the SPECMAP chronology, the data were smoothed with a 14-point least-square running average (Fig. VI-2). The age models for the younger sections of each core (i.e., <40,000 years) were refined based on AMS ^{14}C ages (C. Didié, unpubl.) measured on *N. pachyderma* (sin.) and by including the ages of Heinrich events 1-6. Identification of Heinrich layer deposition was based on the concentration of lithic grains (see Didié and Bauch, 2000) and by comparison with many other records from this region.

VI.4. Results

VI.4.1. The Rockall Plateau record

At Site M23414 at the Rockall Plateau all peak interglacial warm periods are characterized by

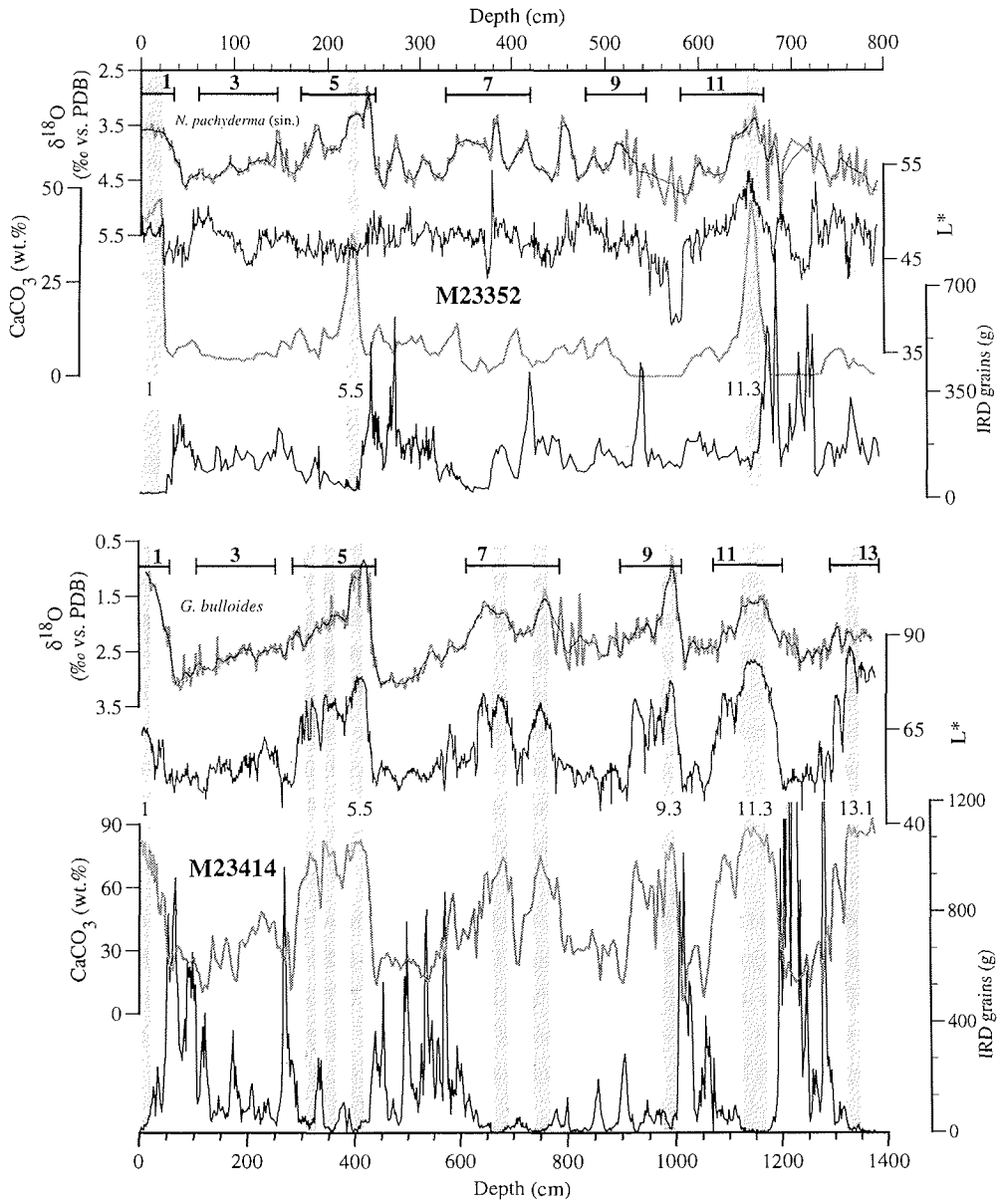


Fig. VI-2: Downcore records of planktic $\delta^{18}O$, sediment lightness, $CaCO_3$ content, and IRD content from the Iceland Plateau (upper panel, data from Henrich, 1998; Bauch and Helmke, 1999; Bauch et al., 2000; Bauch, unpubl.) and the Rockall Plateau (lower panel, data partly from Jung, 1996; Didié and Bauch, 2000), with odd numbered interglacial MIS indicated for reference. Thin black line underlying both isotope records (gray line) represents 14-point least square running average of raw isotopic data. Shaded areas denote full-interglacial conditions during MIS 1, 5, 7, 9, 11 and 13. Note that the maximum IRD amount at Termination V (2554 grains (g) at 1220 cm core depth in M23414) is cut off at 1500 grains (g) to enlarge low amplitude fluctuations of IRD input.

minima in planktic oxygen isotope values, no IRD deposition, and maxima in bulk carbonate content ranging between 50 and 90%. These high carbonate values correlate with maxima in sediment lightness, except in MIS 1. Despite the different sample resolution applied, there is a strikingly good correlation between carbonate content and the much higher-resolved sediment lightness record (Fig. VI-2). MIS 13 reveals both the highest carbonate content and corresponding sediment lightness of the entire record. Peak interglacial carbonate contents seem to show a decreasing trend between MIS 13 and MIS 7 but increases again in MIS 5 and 1 (Fig. VI-2). Although this trend is naturally reflected also in the sediment lightness, there is a notable mismatch between carbonate content and sediment lightness for the Holocene.

A pronounced glacial-interglacial difference is observed in IRD record, showing little or no IRD input during the peak interglacial warm phases of MIS 13, 11, 9, 7, 5, and 1. However, it received measureable amounts of IRD during cooler interglacial substages, such as within MIS 7 and 5 (Fig. VI-2). IRD was deposited continuously at the Rockall Plateau during glacial periods (i.e., MIS 12, 10, 6, and 4-2), with millennial-scale recurrence periods of maxima in glacial IRD input (Fig. VI-3). Distinct IRD peaks in the last two glacial periods have been identified as Heinrich events and Heinrich-like IRD events, respectively (Didié and Bauch, 2000). All glacial IRD maxima at the Rockall Plateau, including those of low-amplitude such as in MIS 8, correspond with minima in the sediment lightness and carbonate records (Fig. VI-3). By far the highest input of IRD during the last 500,000 years occurred in late MIS 12 with a distinct double peak that reaches into the ensuing deglacial transition, i.e., Termination V. Nearly all later glacial intervals and terminations are characterized by IRD input of rather similar amplitudes. Only MIS 8 stands out by showing extreme low glacial IRD depositional rates (Fig. VI-3).

Planktic isotope values of *G. bulloides* at the Rockall Plateau show significant differences between the peak interglacial periods MIS 11, 9, 5e, and 1 (Fig. VI-2). MIS 9 and 5e are characterized by values of ~0.9‰, values in MIS 1 and 11 are heavier by 0.3 and 0.7‰, respectively. In fact, planktic isotope values of MIS 11 are heavier than those of all later peak interglaciations but lighter than the preceding interglaciation MIS 13, which reveals values of ~2.1‰.

VI.4.2. The Iceland Plateau record

When compared to the Rockall Plateau, the core sequences on the Iceland Plateau contains generally lower amounts of carbonate (less than 10%), with increased values (30-50%) occurring only in MIS 11, 5e, and 1 (Fig. VI-2). The two younger peak interglacial intervals reveal minimum isotope values associated with strongly reduced or no IRD, as at the Rockall Plateau. However, MIS 11 is characterized by significant IRD (Fig. VI-3). In contrast to the Rockall study site the Iceland Plateau

record does not show any strong correlation between carbonate content and sediment lightness. This is due to the generally low carbonate content of the sediment (Bauch and Helmke, 1999).

The Iceland Plateau received an almost continuous input of IRD, regardless of glacial or interglacial climate mode (Fig. VI-2). Besides MIS 5e and 1, strongly reduced IRD input and IRD-free intervals occurred episodically only in late MIS 7. In comparison with MIS 5e and 1, the reduced IRD deposition during late MIS 7 is associated with relatively low carbonate content (up to 15%) and, because of faunal evidence, seems not representative of a significant warm phase (Bauch, 1997). In general, the glacial IRD record from the Iceland Plateau shows a pattern of millennial-scale variability during MIS 12, 10, 6, and 4-2, but we could not discern a definite recurrence period (Fig. VI-3). Like at the Rockall Plateau, the end of MIS 12 and the ensuing Termination V are marked by highest IRD deposition and a distinctive double peak. After Termination V, IRD maxima are of comparable amplitude during the remaining glacial intervals with steep and short-term increases occurring during Terminations IV-II. In contrast to the Rockall Plateau, the IRD input on the Iceland Plateau during MIS 8 shows an amplitude that is quite comparable to the later glaciations (Fig. VI-3).

Planktic oxygen isotope values of *N. pachyderma* (sin.) show only minor differences between MIS 11, 5, and 1: Mean minimum values of the warm substages during these three interglaciations range between 3.5‰ and 3.2‰ (Fig. VI-2). The extremely light $\delta^{18}O$ spike in early MIS 5e (at ~240 cm core depth) has been identified as meltwater event, and does not represent interglacial conditions (Bauch et al., 2000b). However, meltwater overprints in the planktic $\delta^{18}O$ are a common feature in the Nordic Seas, particularly, below MIS 7, making the establishing of a chronology based on $\delta^{18}O$ alone more difficult if no other means, such as $\delta^{13}C$ records, sedimentologic and faunal evidence, are taken into account (Bauch, 1997).

VI.5. Discussion and conclusions

In accordance with previous results our stable isotope, carbonate, and IRD records from the Iceland Plateau demonstrate that during the last 500,000 years full-interglacial conditions with a significant northward advection of warm Atlantic surface water into the Nordic Seas prevailed only in MIS 11, 5e, and 1 (e.g., Kellogg, 1980; Bauch, 1997; Henrich, 1998). Foraminiferal assemblage data from the Norwegian Sea indicate that SST were significantly lower in MIS 11 than in MIS 1 (Bauch et al., 2000a). The continuous occurrence of IRD on the Iceland Plateau corroborates this finding. The Norwegian Sea revealed a 10,000 years long IRD-free interval during MIS 11 (Bauch et al., 2000a). This implies that a strong frontal system separated the cold water masses in the Greenland and Iceland seas characterized by icebergs from a warmer water mass marked by ice-free conditions further east in the Norwegian Sea. Hence, compared with the present day situation the data would

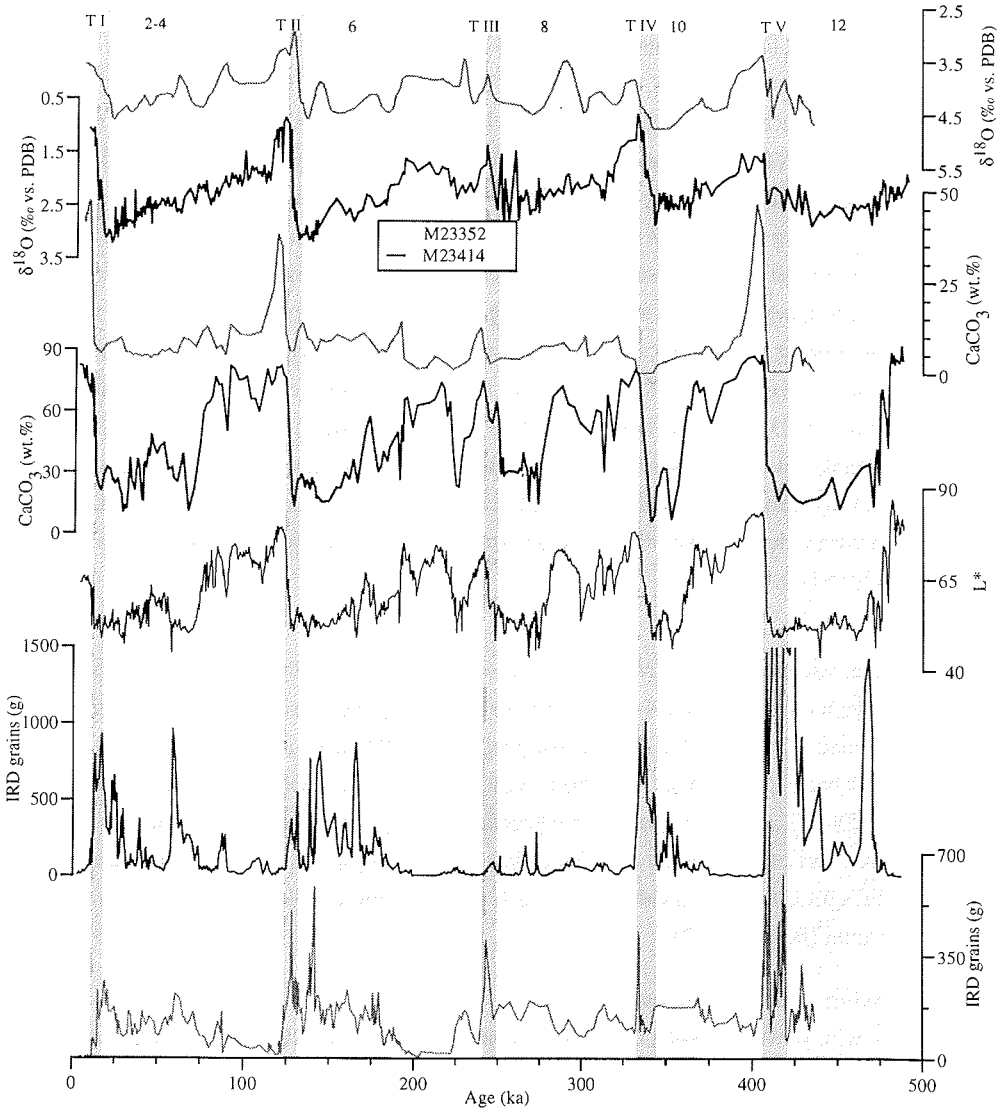


Fig. VI-3: Some records from Figure 2 plotted *vs.* age. Age models are based on spliced standard SPECMAP chronology (Imbrie et al., 1984; Martinson et al., 1987). Glacial MIS and Terminations I-V (T I-V) are shaded in gray and labeled for reference.

suggest a different circulation pattern during MIS 11 compared with MIS 5e and 1 (Bauch et al., 1999; Bauch et al., 2000a). This may point at an eastern expansion of the polar water mass with its

low-saline surface waters that could have caused a reduced advection of warm Atlantic water and a weakening of the thermohaline circulation in the Nordic Seas (Fig. VI-1) as well as further south.

The various peak interglacial episodes during MIS 13, 11, 9, 7, 5, and 1 that are marked by low or no IRD input but high contents of carbonate are clearly recognized by us, in accordance with others (Ruddiman et al., 1986; McManus et al., 1999), as periods of low global ice volume and relatively warm SST. In agreement with our planktic isotope record, benthic $\delta^{18}O$ data from nearby ODP Site 980 (northwest Rockall Plateau) show highest full-interglacial values in MIS 13 (McManus et al., 1999), indicating a larger ice volume than during the later interglaciations. During peak MIS 11, average benthic $\delta^{18}O$ values at Site 980 are about 0.7‰ lower than during MIS 13 but comparable to MIS 9, 5e, and 1, thus suggesting ice volumes in these warm periods of similar magnitude. However, our planktic foraminifer *G. bulloides* yields much higher $\delta^{18}O$ values in MIS 11 than later in MIS 9, 5e, and 1 (Fig. VI-2). As a face value, this indicates that during the last 450,000 years the coldest full-interglacial SSTs in the North Atlantic occurred in MIS 11. This is contrary to isotopic results from Site 980 on the basis of the right-coiling planktic foraminifer *Neogloboquadrina pachyderma* dextral (dex.) as a proxy for SST, which indicate comparable maximum SSTs during all full-interglacial intervals since MIS 12 (McManus et al., 1999). This discrepancy in the isotopic data may be explained by the ecology of the two foraminifers in the North Atlantic. *N. pachyderma* (dex.) is frequently found during spring and summer, living mainly beneath the mixed layer. In contrast, *G. bulloides* lives preferentially in surface waters, i.e., in and above the thermocline, and its abundance in the subpolar North Atlantic seems to coincide with the spring plankton bloom (Hemleben et al., 1989; Duplessy et al., 1991, Ottens, 1992; Schiebel et al., 1997; Schiebel and Hemleben, 2000). SST estimates based on these two species likely reflect the differences in temporal and spatial distribution. Moreover, it is recently suggested that the $\delta^{18}O$ isotopes of *N. pachyderma* (dex.) records mean annual SST in the Northeast Atlantic more accurately than other planktic foraminifers (e.g., Oppo et al., 1998). Regardless, if our *G. bulloides* data from MIS 11 represent a SST signal that more likely reflects a seasonal thermocline signal, it obviously was not recorded by *N. pachyderma* (dex.). In this respect it is certainly interesting to note that our data accord with the results obtained from the Vostok ice core in Antarctica, which also shows the lowest peak interglacial temperatures of the last 450,000 years in MIS 11 (Petit et al., 1997).

The maximum input of IRD towards the end of MIS 12 at both study sites may reflect unusually large discharges of icebergs from the Laurentide and European ice sheets into the polar and subpolar North Atlantic alike. That these high IRD input towards the end of MIS 12 reflects the terminal stage of a strong glaciation is further supported by benthic oxygen isotope data from ODP Sites 980 and 982 (McManus et al., 1999; Venz et al., 1999) and Pleistocene sea-level lowstand calculations (Rohling et al., 1998), both indicating that MIS 12 was probably characterized by the

largest global ice volume during the past 500,000 years. The IRD maxima indicate large amplitude climate variations, although insolation forcing during Termination V was less than during later

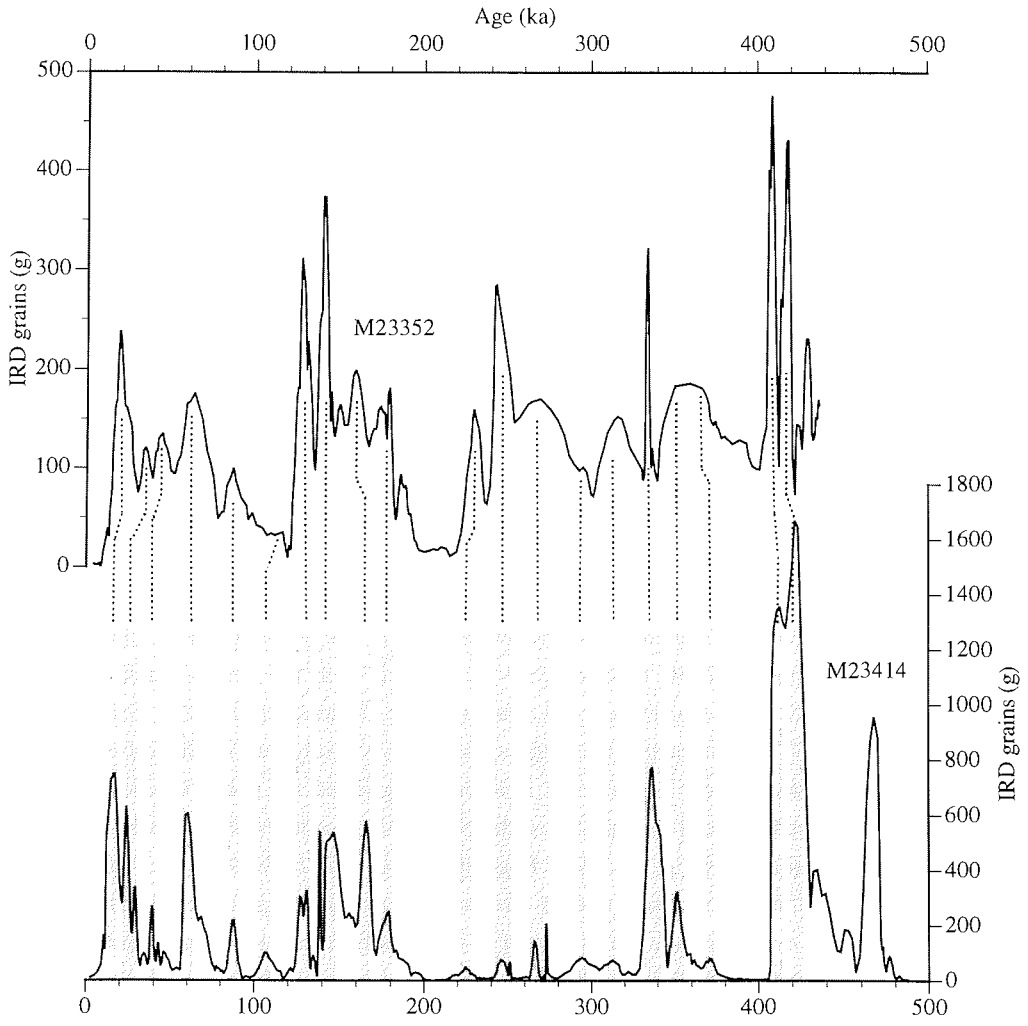


Fig. VI-4: 9-point least square running average of IRD records from M23352 (upper panel) and M23414 (lower panel) *vs.* age. Shaded areas denote major IRD events at the Rockall Plateau during the last 450 ka. Dashed lines indicate corresponding IRD events on the Iceland Plateau.

terminations (Berger, 1978). A maximum input of IRD during Termination V is also reported from Late Quaternary sediments of ODP Site 982 at 57°N (Venz et al., 1999). These authors linked their

IRD record to glacial deep-water production and characterized MIS 12 as an interval of perennial sea ice cover in the Nordic Seas.

Based on the low IRD input to the Rockall Plateau during MIS 8, the iceberg discharge from northern ice sheets is expected to have been less when compared to previous and later glaciations. This finding is in good agreement with benthic isotopes from the Rockall area (McManus et al., 1999; Venz et al., 1999) and with sea-level lowstand estimates that characterize MIS 8 as the glacial period with the smallest ice volume of the past 500,000 years (Rohling et al., 1998). The obvious differences in IRD input between the Iceland and Rockall Plateaus during this time can be explained by the geographical position of the study sites with respect to potential glaciated areas: Because the Laurentide ice sheet is believed to be the major source of massive IRD input into the North Atlantic during the last glacial period (e.g., Broecker, 1994; McManus et al., 1998), the low IRD deposition at the Rockall Plateau during MIS 8 probably indicates a reduced glacier activity and iceberg drift pattern along the northwest Atlantic margin. In contrast, the Iceland Plateau record gives evidence of a continuous supply of icebergs. The reason behind it is most likely the vicinity to a glaciated landmasses, but may be also due to a different behavior of North American and European ice sheets in general, as previously reported for the last climate cycle (Snoeckx et al., 1999; Grousset et al., 2000).

The "cold" climate proxy IRD shows a geographically remarkable coherent pattern of variability between the Iceland and Rockall Plateaus during the past 500,000 years, despite the differences in the total amount of IRD deposited at each core site (Fig. VI-2, Fig. VI-4). At the Rockall Plateau distinct IRD maxima of the last climate cycle have been identified as Heinrich events (Didié and Bauch, 2000), whereas the contemporary Iceland Plateau IRD record reveals millennial-scale variability without a definite periodicity (Didié et al., subm.). Prior to the last glaciation, both records show variations in IRD on millennial time-scales, but due to the uncertainty of SPECMAP-derived age models clear millennial-scale recurrence periods could not be determined. However, the data suggest that comparable short-term climate fluctuations with iceberg calving from the northern hemisphere ice sheets and large discharges of IRD on the Iceland and Rockall Plateaus also occurred during earlier glaciations (Fig. VI-4). Differences in the intervals as well as in periodic recurrence of increased IRD input between the study sites can be explained by the vicinity of the Iceland Plateau to the recent Polar Front (Fig. VI-1). As it can be expected, the Polar Front with its cold and low-saline waters as well as icebergs, shifted repeatedly towards the southeast during glacial periods. Therefore, the Iceland Plateau received an almost continuous IRD sedimentation during the last five climate cycles. The lack of synchroneity in the periods of millennial-scale IRD input between the subpolar and polar North Atlantic indicates differences in ice sheet dynamics between the Laurentide Ice Sheet, the major source of North Atlantic Heinrich events (Bond and

Lotti., 1995), and ice-sheet outlet glaciers draining into the Nordic Seas and iceberg drift on the Iceland Plateau (Dowdeswell et al., 1999, Grousset et al., 2000). Because ice-sheet basins of varying size respond asynchronously to any single external forcing (Johannessen et al., 1989), there is local variability in the drift tracks and sediment-release histories of groups of icebergs (Dowdeswell and Murray, 1990). Thus, the geographical position of our study sites with respect to the northern hemisphere ice sheets during the upper Quaternary caused differences in the recurrence periods of polar and subpolar North Atlantic IRD on millennial time-scales. However, the IRD patterns in the two cores demonstrate that unstable climatic modes with rapid, short-term fluctuations prevailed during glacial intervals, reflecting the overall dynamic nature of the upper Quaternary climate system at high northern.

We conclude that the Nordic Seas have experienced full interglacial conditions only three times during the last 500,000 years (i.e., MIS 11, 5e, and 1), whereas full interglacial periods were more frequent in the Northeast Atlantic. By comparison, planktic $\delta^{18}O$ from the Northeast Atlantic indicate strong discrepancies among interglacial SSTs, i.e., coldest SSTs were found for MIS 11 compared with other peak warm periods (MIS 9, 5e, and 1). The IRD record from the Nordic Seas points at a major temperature gradient in the North Atlantic and a reduced advection of warm surface water into the Nordic Seas during interglacial MIS 11. It would be intriguing to speculate about the effects the reduced heat transfer into the Nordic Seas would have had on the deeper water circulation in the Nordic Seas and the climatic conditions on land. At both sites IRD input was highest during the final phases of glaciations (including early terminations) revealing an almost coherent pattern of IRD deposition between the two study areas with differences in millennial-scale recurrence periods, i.e., in the timing of iceberg discharges between the Laurentide Ice Sheet and European ice sheets. However, the IRD records point at mechanisms that synchronously affected both the Nordic Seas and the North Atlantic and that led to millennial-scale variability of the glacial climate system.

VII - Summary and conclusions

The rationale of this study was to give insight into Late Quaternary climatic variations of the high northern latitudes on glacial-interglacial as well as on millennial time-scales using a wide range of sedimentological, micropaleontological, and geochemical proxy data from Northeast Atlantic and Nordic Seas deep-sea sediments. Four manuscripts that make up the body of this thesis discussed some aspects of the upper Quaternary climate system over the subpolar and polar North Atlantic regions during the last five glacial-interglacial cycles. The paleoceanographic and paleoclimatic conclusions obtained from the results of these manuscripts are summarized below:

In order to determine changes in the Late Quaternary marine carbonate system of the high northern latitudes the carbonate preservation of sediments from the Nordic Seas was studied **(Chapter III):**

- ⇨ A new method to investigate carbonate corrosion was presented. This method is a combination of weight and lightness measurements as well as SEM-analyses of planktic foraminiferal tests.
- ⇨ By using this new approach foraminiferal tests from the Nordic Seas revealed distinct glacial-interglacial changes in carbonate preservation with good glacial preservation and increasing corrosion during interglaciations. This pattern is similar to the Late Quaternary carbonate dissolution pattern from the deep Pacific Ocean, but seems to be out of phase with the rest of the Atlantic Ocean.
- ⇨ Strongest carbonate corrosion was observed in interglacial MIS 11.
- ⇨ With only few exceptions the samples showed relatively good carbonate preservation. This indicates that deeper waters in the Nordic Seas were almost persistently well-ventilated during last 500,000 years, regardless of the climatic mode.
- ⇨ The mechanisms that led to the Nordic Seas carbonate preservation pattern cannot be determined definitely: The enhanced calcite corrosion during interglaciations may have been caused by high planktic productivity with an increased downward flux rate of organic material to the sea bottom and, consequently, to more corrosive bottom water due to carbondioxide enrichment. Yet, it may also be linked to global scale variations in the marine carbonate system, e.g., to changes in the alkalinity of the oceans.

Climate variability in the subpolar North Atlantic during the last five climate cycles was investigated using sediment color records. Thus, in a first step it was necessary to unveil the factors that control sediment color **(Chapters IV and V):**

- ⇨ A detailed investigation of the carbonate fraction from a Northeast Atlantic deep sea sediment core indicated that during the last 500,000 years changes in sediment lightness were controlled mainly by fluctuations of the fine (<20 μm) carbonate fraction.

- ⇨ There is a strong influence of carbonate corrosion on total sediment lightness: Enhanced corrosion has an effect on the surface structure of the coarse (>20 μm) and, probably, also of the fine carbonate fraction. An increase in carbonate corrosion goes along with an increase in sediment lightness.
- ⇨ No straightforward correlation between sediment lightness and carbonate content was recognized. Changes in the preservational mode of the carbonate fraction need to be considered before using sediment color data for estimates of carbonate content.
- ⇨ Changes in the sediment red-green chromaticity during the past half a million years are most likely driven by variations in the iceberg-rafted terrigenous input of reddish minerals containing iron.
- ⇨ Spectral analysis of the red-green color record revealed persistent millennial-scale climate variability during the upper Quaternary.
- ⇨ Maxima of climate variability were tied to times of changes in ice mass and coincide to times when sea level was 40% below present-day value indicating threshold behaviour. This single threshold controlled the amplitude of climatic variations on millennial time-scales throughout the investigated period.
- ⇨ Short-term fluctuations of the climate system were reduced during peak interglacials and also during peak glacial intervals of relatively stable ice volume.

Sedimentological and geochemical proxy records form both study regions were compared in detail to determine climatic conditions of the polar and subpolar North Atlantic during the last five climate cycles (**Chapter VI**):

- ⇨ In contrast to the more frequent occurrence of warm interglacial conditions in the subpolar North Atlantic full interglacial conditions in the polar North Atlantic occurred only during MIS 1, 5e, and 11.
- ⇨ A comparison of planktic $\delta^{18}O$ from the Northeast Atlantic suggests colder SSTs during MIS 11 than during other peak warm periods, i.e., MIS 1, 5e, and 9.
- ⇨ IRD records indicate a major temperature gradient in the polar North Atlantic and a reduced advection of warm surface water into the Nordic Seas during interglacial MIS 11.
- ⇨ IRD input in the Northeast Atlantic and the Nordic Seas increased almost coherently during glaciations and terminations. A general lack of synchroneity in the reccurrence periods of IRD maxima points at differences in the timing of iceberg discharges between Northern Hemisphere ice sheets.
- ⇨ Obviously, the subpolar and polar North Atlantic were synchronously affected by the mechanisms that caused a millennial-scale variability of the glacial climate system.

These new results from Northeast Atlantic and Nordic Seas sediments indicate highly dynamic climates during the Late Quaternary with distinct variability on longer, glacial-interglacial as well as on shorter, millennial time-scales. The investigations revealed synchronous influences on the glacial climate system in the polar and subpolar North Atlantic but also pronounced differences in the paleoclimatic and paleoceanographic conditions between the two areas during the last five climate cycles. The findings could shed some light onto the complex interactions of the forcing and feedback mechanisms that commonly are thought to drive climatic variations in the high northern Atlantic region.

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