The North Atlantic Ocean: An Overview

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INTRODUCTION

The North Atlantic is one of the most important oceans for virtually any subject of scientific interest. Geophysicists investigating rifting and plate tectonics, as well as marine geologists and marine micropaleontologists, appreciate that even the birth of the North Atlantic is still recorded in the form of sediments in the eastern- and westernmost parts of the equatorial North Atlantic. Like all other oceans, much of the geological history of the North Atlantic has been reconstructed due to the success of the Ocean Drilling Program (ODP) and its predecessor the Deep Sea Drilling Project (DSDP). Since the launch of the drilling programs, many cores have recovered sediments, including subsediment crustal structures, that led to the understanding of the evolution of (not only!) the North Atlantic Ocean, its paleoceanography and the Earth's climatic development through the past 155 million years.

From its earliest stages, when it formed the westernmost part of the Tethys Ocean as early as the Early Cretaceous (DSDP Leg 44, ODP Leg 101), the Atlantic started to play a critical role in the climatic and oceanographic history of the Earth. Later on, the Atlantic Ocean became the only entirely open waterway that until today connects both polar regions, providing a „deep sea highway" for Arctic-Antarctic water masses as well as interhemispherical heat exchange. Today, the northern North Atlantic is characterized by two major surface current systems and water masses that couldn't be more different (see also Baumann *et al.*, *this volume*). In the western part, the East Greenland Current conveys cold, mainly Arctic water masses south. These are less saline and significantly colder than their counterparts on the eastern side which emerge from the Carribean, transporting temperate water masses far north via the North Atlantic and Norwegian Currents. Altogether, these currents together with the associated deep currents play a crucial role in global climate dynamics.

A further important North Atlantic aspect is the epicontinental seas that are adjacent to and part of the North Atlantic. All bear their own important evolutional histories:

The Gulf of Mexico and the Carribean Sea, presently forming the Atlantic's warm water pool, carry some of the oldest sediments found in the entire Atlantic, some of them oil-bearing (e.g. Jenkyns, 1994).

The Mediterranean turns out to be an excellent example for the impact that even a small sea, blocked by huge continents, may have on the Earth's climate: during the latest Miocene - the Earth was already in a glacial mode, and decreasing sea levels cut-off the connection between the Mediterranean and the World Ocean. Water evaporated and eventually reached the World Ocean again but the salt remained on site. Thus, the World Ocean became more prone to freezing over because of slightly decreased salinity which in turn slightly increased the Earth's Albedo and finally caused a global cooling during what is called the „Mediterranean salinity crisis" (e.g. Hodell *et al.*, 1986; Weijermars, 1988; Kastens, 1992).

Further north, the Baltic Sea forms the largest reservoir of brackish water in the world. Its oscillations between marine and freshwater stages due to iso- and eustasy at least during the past 13,000 years had a tremendous effect not only on the biology of the Baltic Sea but perhaps also on the climate of the northern North Atlantic (Bergsten & Nordberg, 1992). Its gateway to the North Sea, and thus its only connection with the World Ocean, the Skagerrak-Kattegat, bears extreme high-resolution Holocene sedimentary records (e.g. Stabell & Thiede, 1985; Hass, 1996). The North Sea basin itself, bearing the North Atlantic's most important hydrocarbon province, has a colorful tectonic history. Rifting commenced in the earliest Triassic and terminated during the Paleocene, thus leaving major structures like the Central and Viking Grabens (Ziegler & Van Hoorn, 1989).

The Norwegian Sea is situated at one of the most crucial gateways of the World Ocean circulation. It was not before the middle Miocene that the

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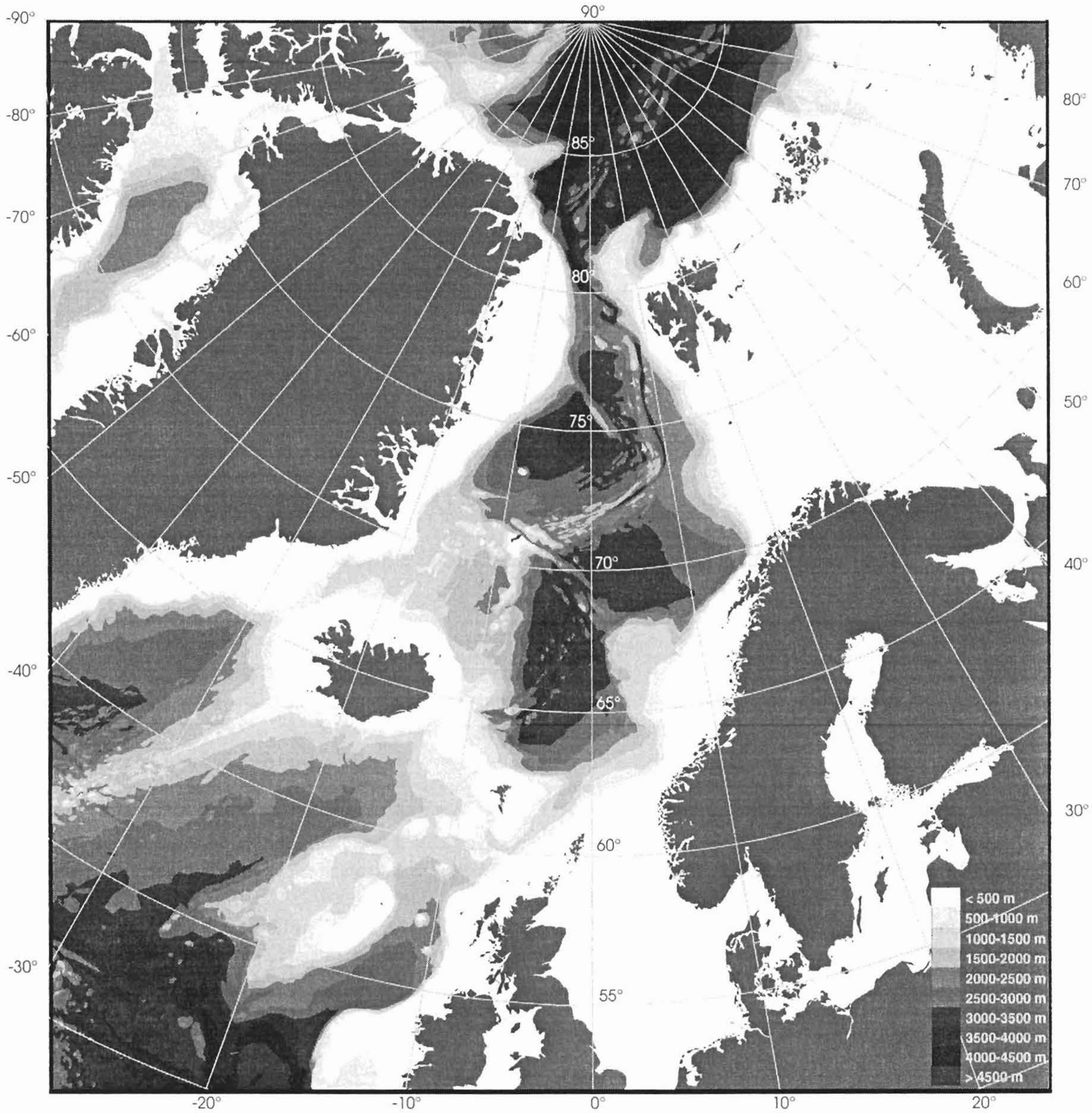


Figure 1. The northern North Atlantic (bathymetric data from GEBCO).

Greenland-Scotland Ridge submerged below sea level, thus terminating the isolation of the northern North Atlantic and providing a connection between the central Atlantic and the Arctic Ocean heat sink (Eldholm *et al.*, 1987; Berggren & Schnitker, 1983). Eventually, a proto-modern Atlantic circulation commenced as cold and dense deep water masses headed south and were replaced by warmer and less dense surface water masses from the south (Kennett, 1982); the highly climate effective Norwegian Current that passes along the Norwegian shelf and through the Norwegian Sea was born.

Regarding the present global ocean circulation and climate, the initial pacemaker of the Ocean Conveyor (Broecker, 1991) is considered to be in the North Atlantic. Cold Arctic winds cause evaporation and cooling of surface water masses and thus contribute to the sinking of huge cells of denser cooled surface water (Broecker & Denton, 1990). This produces a gigantic flow of cold and salty deep and intermediate water which then travels from the North Atlantic south around Africa and through the Indian Ocean (see also Sy *et al.*, 1997). The stability of the deep water flow is controlled by the freshwater balance of the North Atlantic basin (Rahmstorf, 1996). The major part of this cold water mass upwells in the northern Pacific due to intense wind forcing, releasing salt and incorporating freshwater and heat. According to the model, this water mass returns to the North Atlantic as a warm and less saline surficial return flow, increases its salinity and temperature in the Carribean and eventually conveys heat up to the higher latitudes in the form of the North Atlantic Current. If this process is disturbed, the deep circulation may become shallower and weaker and the areas of deep water renewal may migrate further south (Rahmstorf, 1994). As a result, less heat is conveyed north and the northern North Atlantic regions rapidly cool down (the „glacial" mode of deep circulation).

Presently, a legion of researchers is working on paleoclimatic reconstructions on Milankovitch and sub-Milankovitch time scales from Quaternary sediments in order to exactly reconstruct rapid climate change with regard to the paleoceanographic setting of the North Atlantic. While deep-sea drilling is very expensive and takes a long time for planning and organization, dozens of large and small research vessels collected sediment cores and water samples, and carried out measurements at hundreds of sites in the North Atlantic Ocean during the last 20-30 years in order to understand the present oceanographic and climatic conditions as well as those of the past few hundred thousends of years. This small armada of research ships includes vessels as small as a rubber dinghy (to take water and plankton samples from coastal areas) up to huge icebreakers such as the German „Polarstern" that are able to enter pack ice covered basins in the northernmost areas of the North Atlantic and elsewhere.

TECTONIC EVOLUTION OF THE NORTH ATLANTIC

Today, the Atlantic forms the second largest basin of the World Ocean. Through time it undertook a variety of forms, from a narrow, tropical part of the Tethyan Ocean at the time the reptiles began to rule the surrounding supercontinents to the present elongated, more than 11,000-km long basin that connects both polar regions. While the South Atlantic is slightly younger, the evolution of the North Atlantic can be traced back as far as the Triassic.

The following sections deal exclusively with the North Atlantic in general and the northern North Atlantic in particular. This chapter is based on a great number of references as listed in the reference section. Basically, where not specified, the tectonics chapter is based on papers in Ziegler (1975), Kennett (1982), Sheridan (1986), Eldholm *et al.* (1987), Srivastava & shipboard scientific party (1987), Ziegler & van Hoorn (1989), Myhre & Thiede (1995), and Scrutton *et al.* (1995).

The birth of an ocean

Long before the splitting of the African and North American continents in the middle Jurassic, early tectonic activity affected the northern North Atlantic area when late Hercynian suturing marked the future breakup of Gondwana and Laurasia.

While rifting in the Norwegian-Greenland Sea started as early as in the Namurian, the southward propagation of rifting processes was presumably inhibited by oppositely directed stress fields emerging from the peak Variscan orogenic activities.

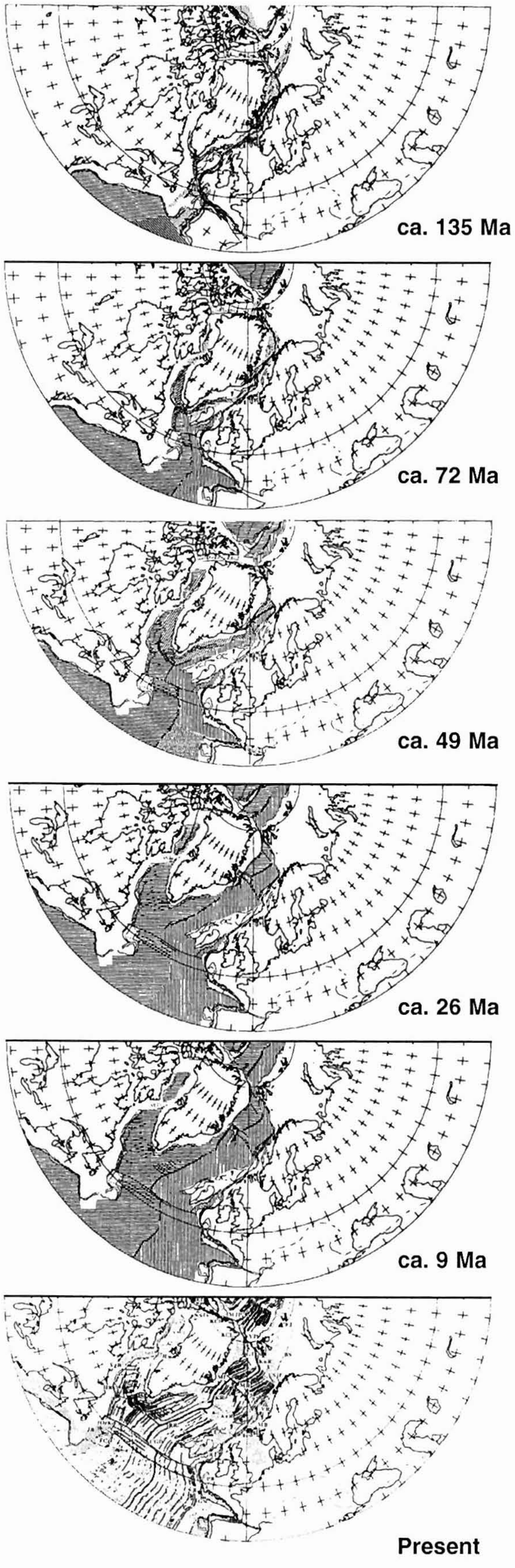
As the stress fields changed, rifting may have expanded to the south during the early Permian. During the Permo-Triassic, the northern North Atlantic area was covered with red beds that were trapped in rapidly forming graben structures. Rifting further increased and eventually formed a complicated pattern from the Norwegian Sea to the western Tethys, where rifting processes from the early Permian heralded a major reorganization of the plates. Although there were no late Paleozoic marine ingressions into the North Atlantic area, the structures that eventually led to the formation of the early North Atlantic Ocean established themselves during the Late Carboniferous at the latest.

The Mesozoic development of the North Atlantic

It was not until the early Triassic that Permo-Carboniferous fracture zones of the Mediterranean were reactivated through strong rifting processes related to the tectonic activity in the Tethys. These structures eventually enabled marine advances further west. By the end of the Triassic and at the dawn of the Jurassic, marine conditions had established themselves through rapidly subsiding basins as far as the Rockall Trough and the NW African, New England and Canadian shelves. East of the Rockall Trough, a number of basins formed; these now carry up to 6 km of clastic Triassic sediments (Ebdon *et al.*, 1995). Further advances finally led to a connection with the cyclically advancing Arctic Ocean through the Rockall Trough and the grabens of the Irish Sea during the Raethian and Hettangian. Early Lias then saw the development of open marine conditions between the Tethys, which had meanwhile advanced to the Carribean, and the Arctic Ocean.

Although this connection was suspended during a short period of extensive doming in the North Sea (Ziegler & van Hoorn, 1989), the breakup of Pangaea, and thus the development of a major ocean, did not stop.

Intensive volcanism accompanied the prograding rift system that advanced from the south through the Labrador Sea, while the main sea floor



spreading axis of the central Atlantic could not overcome the Azores fracture zone until the Early Cretaceous. Later, rifting reached Baffin Bay and the central Atlantic spreading axis advanced into the northern North Atlantic, where strong tectonic activity resulted in rapid doming and subsidence.

During the Cretaceous, crustal separation between Europe and Greenland advanced as seafloor spreading continued in the Norwegian-Greenland Sea. The North Atlantic spreading axis reached the North American continent and succeeded in separating Greenland. Until the Early Tertiary (mostly within Anomalies 24-25), major changes took place: crustal separation isolated Greenland from Norway and Rockall, crustal separation also took place in the Baffin Bay, major ridges formed in the North Atlantic and Arctic Oceans and the Eurasian Basin opened (e.g. Srivastava, 1978; Eldholm *et al.*, 1987).

Cenozoic development

Spreading continued in the Norwegian-Greenland and Labrador Seas; this eventually caused enormous compressional stress on Greenland's shelf until the early Oligocene when the Labrador-Baffin Bay spreading axis became extinct. The Greenland-Scotland Ridge, one of the most important structures for global ocean circulation (see e.g. Bott *et al.*, 1983), appears to be the track of a mantle plume that was situated underneath eastern Greenland. During the divergence of Norway and Greenland, this hot spot left a trace of up to 30 km thick igneous rocks that formed the ridge (White, 1988). Activity of this hot spot caused uplift of up to 1,500 m in the Shetland-Faeroe area (Joppen & White, 1990). Due to this process, large submarine fans formed in the North Sea and eventually became excellent hydrocarbon reservoirs (Anderton, 1993; Ebdon *et al.*, 1995).

Due to continued spreading in the Norwegian-Greenland Sea, Greenland was eventually pushed back to the North American Plate.

Until the mid-Miocene, tectonic and volcanic activitiy remained high: Jan Mayen as well as the northeastern part of the Barents Shelf separated from Greenland. Meanwhile, a connection between the North Atlantic and the Arctic spreading centers was established by the Knipovitch Ridge.

Until today, major sea floor spreading is slow to moderate (Norwegian-Greenland Sea at 0.95 cm/yr; Eldholm *et al.*, 1987).

Figure 2. Paleophysiographic evolution of the northern North Atlantic area from Early Cretaceous (Valanginian, ca. 135 Ma) to the Present according to several sources (after Rowley & Lottes, 1988). Ridge axes, different crust types, and magnetic anomalies (Present) are displayed (see Rowley & Lottes, 1988).

PALEOCEANOGRAPHIC AND PALEOCLIMATIC HISTORY OF THE NORTH ATLANTIC

The section above could only provide a glimpse of the long history of rifting prior to the eventual formation of the present day North Atlantic. In the following section a short overview of the paleoceanographic evolution of the North Atlantic and its role in global climate change will be given. Of course it can not be complete. The main emphasis will be placed on the northern North Atlantic.

Mesozoic paleoenvironments

The lack of Mesozoic glacial deposits suggests that there were no glaciated areas anywhere in the world (Frakes, 1986). Among many theories to explain the vast global warming that followed the glaciation of Gondwana, an increased greenhouse effect through volcanogenic CO_2 is only one (see Graedel & Crutzen, 1993). Towards the middle Jurassic, substantial parts of the continental margins worldwide became drowned, mostly due to eustatic sea level rise. What is Europe today consisted of widespread epicontinental seas and intercalated land areas (Ziegler, 1975). There is a reasonably good correlation of sea-level highstands and periods of fast seafloor spreading during the Jurassic and Cretaceous (Sheridan, 1986; see also Pittman, 1978).

Unlike the crucial role the North Atlantic played in the paleoclimatic evolution of the Cenozoic, Mesozoic sediments suggest that the young North Atlantic Ocean was built up of mostly restricted basins (Kennett, 1982). At least temporally, these basins must have been cut off from the World Ocean, as evidenced by evaporites of probably early or middle Jurassic age (Thiede, 1979). These oldest sediments in the North Atlantic were deposited near the spreading axis of this ca. 500 km wide, tropical ocean (Robertson & Ogg, 1986).

Favored by the warm climate and high atmospheric CO_2 concentrations, calcareous nannoplankton began to shift the major part of carbonate production from the shallow shelves to the deep sea during the late Jurassic (Roth, 1986). Towards the middle Cretaceous, when the central North Atlantic approached probably half of its present size, water mass circulation was controlled by a large cyclonic gyre. There was a gateway that opened to the Tethys Ocean and probably a narrow strait that allowed surficial water mass exchange with the Pacific through the Carribean, whereas the south was blocked by land masses and there was no connection to the northern North Atlantic until the Eocene (Myhre & Thiede, 1995; Roth & Krumbach, 1986). Deep water was thus generated in marginal evaporitic basins, which resulted in warm and saline deep and bottom waters. A very stable water column with low-saline and very warm surface waters and high saline but only slightly cooler bottom waters resulted (Roth, 1986; Saltzman & Barron, 1982); low oxygen values kept marine life low (Bralower & Thierstein, 1986). These conditions became even more pronounced during the Cenomanian transgression, when warm, saline South Atlantic water masses fueled the North Atlantic and widespread epicontinental seas contributed increasing amounts of low oxygen, high saline water to the bottom water (Roth, 1986). DSDP Leg 76 (Site 534, Blake-Bahama Basin) sediments revealed 2 major cycles of carbon-poor to carbon-rich sediments starting in the middle Jurassic; these extend as far as in the eastern Tethys (Baumgartner, 1983). These sediments may be the result of a cyclic occurrence of anoxic conditions and eustatic sea-level changes that, in turn, led to CCD fluctuations (Sheridan, 1986). Hallam (1986) discusses possible Milankovich forcing that may be the reason for obvious sedimentary cycles of 10,000 to 100,000 years superimposed on the large cycles of Cretaceous ocean ventilation.

From the Turonian, Tethyan planktic foraminifers that were found in cores from the South Atlantic (e.g. DSDP Leg 39) suggest the establishment of strong currents connecting the North and the South Atlantic oceans (Premoli Silva & Boersma, 1977).

Cenozoic paleoenvironments

The basins that formed in the Norwegian-Greenland Sea were framed by a complex tectonic pattern of ridges and fracture zones (see Myhre *et al.*, 1982; Nunns & Peacock, 1982; Eldholm *et al.*, 1987). These still prevented major water mass exchange with the World Ocean. The Greenland-Scotland Ridge remained subaerial at least until the late early Eocene, when the most easterly parts began to subside (Myhre & Thiede, 1995). Complex sedimentary patterns (sediment drifts) suggest that a large portion of the bottom waters made their way through the Faeroe channels that formed the deepest passage of the Greenland-Scotland Ridge (Andersen & Boldreel, 1995). Faunal exchange (deep water agglutinated foraminifers) between the North Atlantic and Norwegian Sea is already evident in the middle Eocene (Kaminski *et al.*, 1990). Subsidence of the ridge apparently followed the trace of the mantle plume to the west. By the middle Miocene the ocean basins had almost their modern shapes (Schnitker, 1980). The Denmark Strait section of the ridge became submerged in the late Miocene, as evidenced by changes in the benthic foraminiferal faunas in the Norwegian-Greenland Sea (Berggren & Schnitker, 1983). Cold bottom waters that had formed since the Miocene in the Nordic Seas were now able to flow across the ridge. The Earth's climate system had already entered a glacial mode during this time and a worldwide ca. 40 m sea level drop took place (Kennett, 1982). Partly due to this sea level drop, the Mediterranean became isolated from the World Ocean (see also Weijermars, 1988). As a result, evaporation transformed the Mediterranean into a series of salty inland lakes until the beginning of the Pliocene, leaving behind a huge series of evaporites. Eventually, this salt was withdrawn from the World Ocean; thus, its salinity decreased by 2 PSU. Since less saline waters freeze earlier, which in turn increases the Earth's Albedo, the isolation of the Mediterranean was the cause of a remarkable world-wide cooling during a period that is called the „Mediterranean salinity crisis" (Kennett, 1982).

Continuing deterioration of the climate led to a permanently frozen Arctic Ocean, probably from the late Miocene (Clark, 1982; however, see Wolf & Thiede, 1991; Repenning *et al.*, 1987). During the late Pliocene, sedimentation of ice-rafted debris in the North Atlantic began (Shackleton *et al.*, 1984), although first indications of an unstable ice cover or icebergs in the Labrador and Norwegian-Greenland Seas could be dated back as far as the middle/late Miocene (Wolf & Thiede, 1991; Fronval & Jansen, 1996; Thiede & Myhre, 1996; Wolf-Welling *et al.*, 1996). In the continuation of the Cenozoic, the growths and decays of the major ice sheets around the northern North Atlantic, which had another amplification around 1 MyBP, led to the deposition of huge masses of glacial sediments that were transported as far as the deep-sea basins through ice rafting or in the form of mass movements through channel systems on the slope. Shelves prograded considerably into the North Atlantic as huge submarine fans formed, such as on the West Hebrides and West Shetland shelves (Stoker, 1995) and all around the northern North Atlantic from Newfoundland to Norway (e.g. Piper *et al.*, 1990; Vorren *et al.*, 1991; Aksu & Hiscott, 1992; Elverhøi, 1992;).

The climate deteriorated stepwise during the past 55 Ma. From at least 3 MyBP, strong orbital forcing is suggested from SST records and ice volume (Myhre & Thiede, 1995; see also DSDP Leg 94 results). Maslin *et al.* (1995a) consider a sharp increase in insolation through increased obliquity and precession amplitudes the main trigger for the intensification of the Northern Hemisphere galciation between 2.75 and 2.55 MyBP. The final closure of the Panama Isthmus around 1.8 MyBP (Keller *et al.*, 1989) is thought to have either delayed the formation of ice sheets through increased heat transport to the north (Maslin *et al.*, 1995a) or stimulated ice sheet formation as a result of increased transport of moisture to the north (Mikolajewicz *et al.*, 1993). Burton *et al.* (1997) present geochemical evidence for increased North Atlantic Deep Water (NADW) flow since the closure of the central American Isthmus suggesting the onset of modern North Atlantic water mass circulation at that time.

Of the Plio-Pleistocene, which is characterized by an increasingly significant climate deterioration, the last ca. 900 ka (Mid Pleistocene Revolution 900-920 kyBP, Berger & Jansen, 1994) - i.e. oxygen isotope Stages 22-1 - and especially the last 650 ka attracted major interest as the Earth's climate system then switched from 41,000 and 23,000 into a powerful 100,000 year cycle (e.g. Pisias *et al.*, 1984; Martinson *et al.*, 1987). During this period, the advancing and retreating Laurentide and Eurasian ice sheets had their major maxima (e.g. Last Glacial Maximum, LGM, ~18 kyBP) and minima (e.g. Eemian, ~125 kyBP). Since mathematically, the glacial/interglacial cycles should have been more continuous than they obviously were, a couple of theories were established in order to explain rapid climate change (see e.g. Ruddiman & McIntyre, 1981; Boyle & Keigwin, 1987; Berger & Jansen, 1994). Most likely, variations in the formation of NADW such as the southward movement of the source area of NADW, switching to a weaker production mode or a complete shutdown of the flow of the global ocean circulation system (the Ocean Conveyor) led to changes in global heat transport which subsequently caused rapid climate change (Broecker, 1987, 1991; Broecker & Denton, 1990; see also Rahmstorf, 1994, 1996; Sarnthein *et al.*, 1994; Weinelt *et al.*, 1996). Thus, aside from orbital parameters, the northern North Atlantic in general and the NADW production cells in particular are considered to be the principal pace maker of global climate. Recently, new hypotheses (mainly based on geochemical evidence and atmospheric general circulation models) challenged this view by proposing a significant drop in *tropical* sea surface temperatures that may have led to a series of feedback mechanisms affecting atmospheric water vapour and thus the greenhouse effect finally resulting in global cooling (Webb *et al.*, 1997; Beck *et al.*, 1997; Bard *et al.*, 1997; Curry & Oppo, 1997; see also Zahn, 1994).

The production rate of NADW appears to be in close conformity with the climate variability as recorded in Greenland's ice cores (e.g. Keigwin *et al.*, 1994). Recently, North Atlantic deep-sea sediment analyses (op. cit.) suggested stable climate conditions during the Eemian interglacial as no change in NADW production could be detected, thus challenging earlier interpretations from ice cores (GRIP Members, 1993). However, Maslin

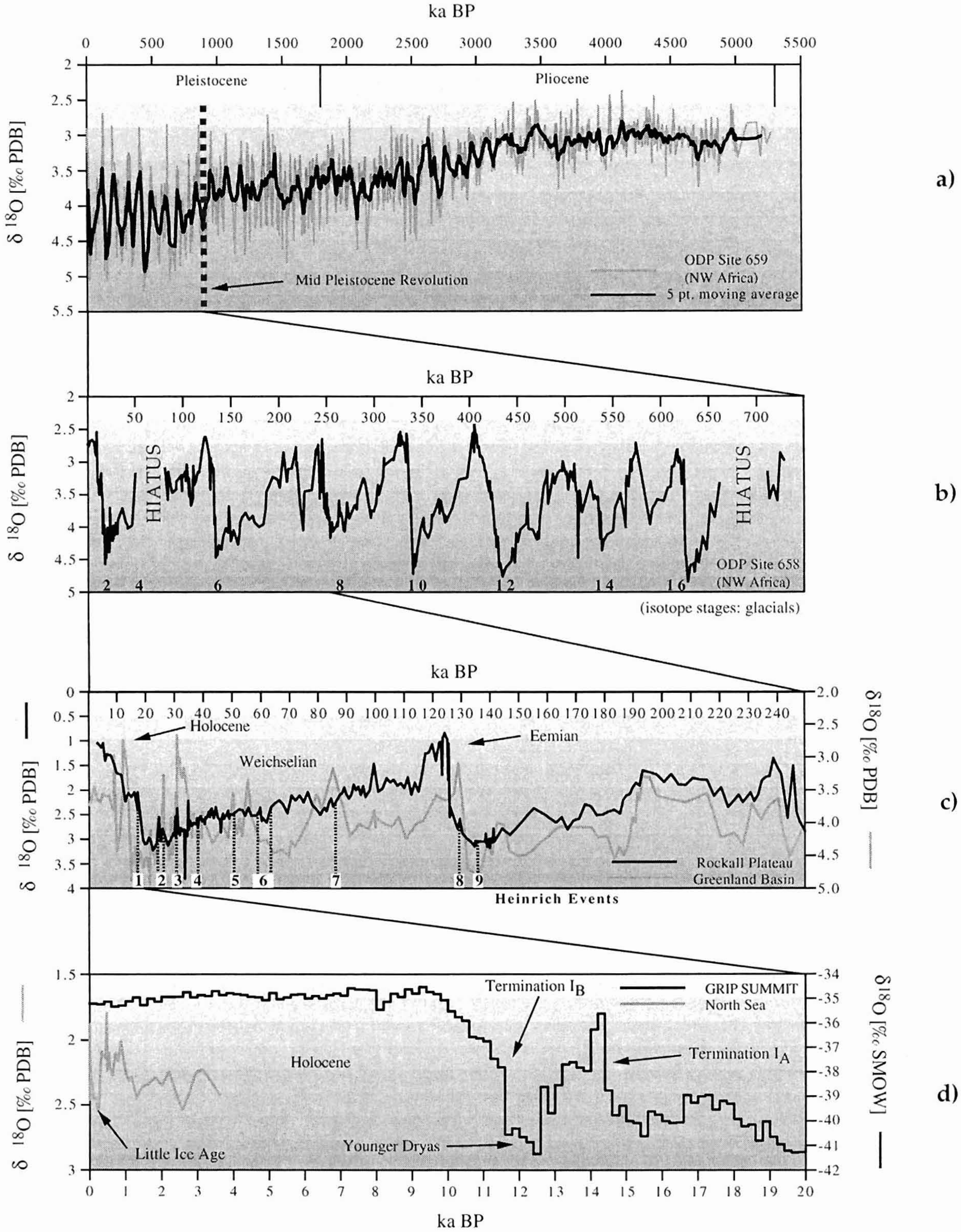


Figure 3. a) Oxygen isotope profile of Site 659 (ODP Leg 108, off NW Africa) (after Tiedemann *et al.*, 1994); b) oxygen isotope profile of Site 658 (ODP Leg 108, off NW Africa) (after Sarnthein & Tiedemann, 1989) including glacial isotope stages; c) solid line: oxygen isotope profile of Core 23414 from the Rockall Plateau including Heinrich Events (age scale: calender years) (after Jung, 1996); grey line: oxygen isotope profile of several cores from the Greenland Basin combined to a „standard" profile (after Antonow, 1995; Antonow *et al.*, *this volume*); d) black line: oxygen isotope profile of Greenland ice core GRIP SUMMIT (after Dansgaard *et al.*, 1993; data from Dansgaard, W., Johnsen, S. J., Clauson, H. B., Dahl-Jensen, D., Gundestrup, N. S., Hammer, C. U., Hvidberg, C. S., Steffensen, J. P., Sveinbjornsdottir, A. E., and J. Jouzel); grey line: oxygen isotope profile of Core III KAL, Skagerrak (after Hass, 1996; *this volume*).

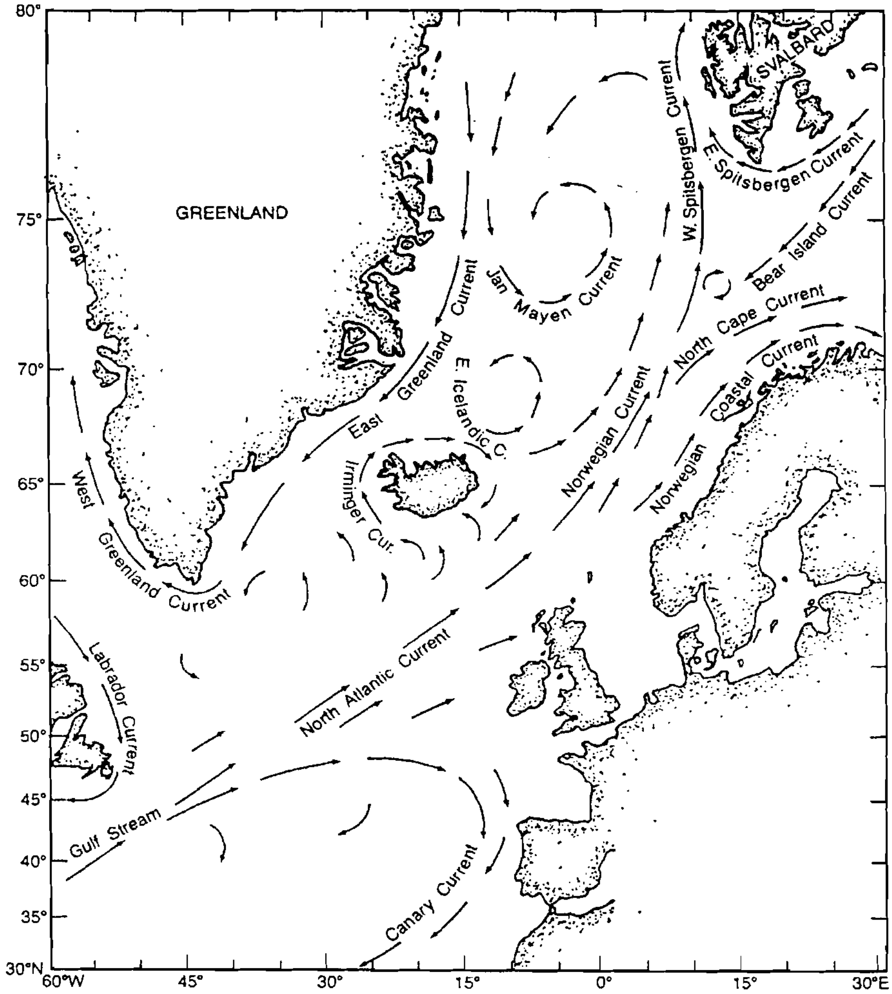


Figure 4. Present-day surface water mass circulation of the northern North Atlantic (adapted from Eldholm *et al.*, 1987, after data from Kellogg, 1975, 1976).

(1996) was able to present stable isotope evidence (ODP Site 658, eastern Atlantic) for a short term (<400 yrs) intra-Eemian shutdown of deep water formation and a synchronous cold spell. Keigwin *et al.* (1991) and Lehman *et al.* (1991) showed that during the last deglaciation (15-10 ka), multiple meltwater events were followed by decreased NADW production and cold spells. The sudden shutdown of NADW production was found to cause SST shifts of ≥ 5° C in fewer than 40 years (Lehman & Keigwin, 1992).

Apparently, the pattern of glacial/interglacial changes during the Brunhes Epoch basically repeated itself through time. As an example, the conditions from the LGM to the beginning of the Holocene in the northern North Atlantic are now shortly described.

There is much debate about the surface and deep water circulation of the northern North Atlantic during the LGM. Earlier interpretations of SSTs suggest a sluggishly cyclonically rotating, though permanently frozen, Norwegian-Greenland Sea (CLIMAP Project Members, 1976, 1981, 1984) whereas Henrich (1990, 1992) and Henrich *et al.* (1995) inferred at least seasonally open cyclonically rotating waters in the Norwegian-Greenland Sea from IRD facies, carbonate contents and carbonate preservation studies. From stable isotope measurements of planktic foraminifers, Sarnthein & Altenbach (1995) and others (e.g. Ruddiman & McIntyre, 1976; Robinson *et al.*, 1995; Weinelt *et al.*, 1996; Bauch, *this volume*) postulate a seasonally ice-free, weakly cyclonic rotating Norwegian-Greenland Sea. According to this theory, the Norwegian Sea was entirely isolated from the North Atlantic by ice sheets of Iceland, the Shetlands, and Norway. Furthermore, a persistent anticyclonic rotating meltwater lens west of Ireland not only blocked the connection to the northeast but also supported the Irminger Current that conveyed temperate water masses north across the Denmark Strait (Sarnthein & Altenbach, 1995).

Aside from unequivocally astronomically steered climate change, there is still strong and rapid climate change on a much shorter -millenial- timescale: A strange pattern of ice-rafted debris (IRD) that was deposited approximately every 12 ky during the past ca. 60 ka was discovered in the North Atlantic, stretching from Labrador to the British Islands (e.g. Bond *et al.*, 1992; Broecker *et al.*, 1992) including associated events further north (e.g. Stein *et al.*, 1994, 1996). Termed Heinrich Layers after H. Heinrich, who was the first to describe them (Heinrich, 1988), most of these layers result from massive iceberg discharge out of Baffin Bay; the icebergs were transported northeast by the cyclonic polar gyre (Grousset *et al.*, 1993). These events of massive iceberg release into the North Atlantic Ocean during the last glaciation appear to be parallel to recurring episodes of unusually cold North Atlantic surface water (Bond cycles; Bond *et al.*, 1992) and cold air temperatures over Greenland (Dansgaard-Oeschger events; Dansgaard *et al.*, 1993). Interpreting sea surface properties (SST, SSS, $\delta^{18}O$ of planktic foraminifers), Maslin *et al.* (1995b) could not substantiate a sea surface cooling prior to Heinrich events, thus adding support to the largely climate-independent free-oscillating system theory of MacAyeal and Wang (1992) as a cause for the Heinrich Events. After each Heinrich event (except H1, which occurred during the Milankovitch forced insolation rise) an episode of increased SST and air temperature over Greenland lasting up to 2,000 years is suggested before the climate trend continued further into the glacial maximum (GRIP Members, 1993; Grootes *et al.*, 1993; Maslin *et al.*, 1995b). As NADW production was possibly halted by Heinrich events H3 and H4, and there was basically no or only very weak deep water production between 30 and 13 kyBP, it was inferred that Heinrich events were at least partly responsible for amplifying glacial conditions (Broecker, 1994; Maslin *et al.*, 1995b; Sarnthein & Altenbach, 1995). Seidov & Haupt (1997) showed that NADW production was low although the glacial conveyor was rather intensive; deep-water convection was restricted to the center of the subpolar gyre. However, Oppo & Lehman (1995) postulate that NADW production was already weak prior to Heinrich events H2, 3, and 4 and that there was no additional weakening in response to the massive iceberg discharge. Although still part of scientific debate, it turns out that the Heinrich events are not only in phase with the air temperature above Greenland (Bond *et al.*, 1993) but that corresponding events left worldwide traces (e.g. Broecker, 1996, 1994; Porter & An-Zhisheng, 1995; see also Stein *et al.*, 1994; Goldschmidt, *this volume*).

Based on AMS ^{14}C and planktic $\delta^{18}O$ results, Sarnthein *et al.* (1992) dated the beginning of the last deglaciation at 14.7 kyBP peaking at 13.5 kyBP. A giant meltwater lens formed off northern Norway and cold water masses were conveyed south, whereas the influx of temperate waters into the Norwegian-Greenland Sea across the Denmark Strait persisted since the LGM (Sarnthein & Altenbach, 1995). During the recovery from the LGM, this unusual current regime in the Norwegian-Greenland Sea caused a decrease or even shutdown of the NADW production (e.g. Jansen & Veum, 1990; Keigwin & Lehman, 1994; Seidov *et al.*, 1996; Seidov & Haupt, 1997). As a result, the Oldest Dryas (equivalent to H1) cold spell forced the climate system back towards late glacial conditions at around 14.5 kyBP; $\delta^{13}C$ values of planktic foraminifers indicate the lowest NADW production rate of the past 20 ky (Keigwin & Lehman, 1994). However, insolation increased rapidly; thus, the Oldest Dryas soon saw another reorganization of water mass circulation in the North Atlantic that eventually brought the Bølling warm period at glacial Termination I_A. The maximum melting of the Fennoscandian icesheet has been dated at around 12.5 kyBP, the peak warm phase during the Bølling (Andersen *et al.*, 1995). Koç Karpuz and Jansen (1992), working on diatom SST transfer functions of high resolution cores from the southwestern Norwegian shelf, were able to distinguish a series of short-term climate episodes within the period of the meltwater peak that characterizes Termination I_A.

Still within the Allerød warm interval, around 11.2 kyBP, calculated SSTs suggest that Polar and Arctic fronts in the northern North Atlantic rapidly readvanced to near-glacial conditions. Sea surface temperatures may have dropped as rapidly by more than 6° C as the air temperatures over Greenland did (Dansgaard *et al.*, 1989; Koç Karpuz & Jansen, 1992; Lehman & Keigwin, 1992; see also Costello & Bauch, *this volume*). Sea ice drifted as far south as to the latitudes of northern Spain (Keigwin & Lehman, 1994) during the most recent extreme cooling episode of the Younger Dryas (11.2-10.2 kyBP after Koç Karpuz & Jansen, 1992). Despite the enormous impact, the role of the North Atlantic during the Younger Dryas cold spell remains enigmatic: stable isotope data suggest a regular Holocene-style meridional circulation regime, however, with SST's indicating strong seasonality of the sea ice cover (Sarnthein *et al.*, 1995) that ceased in the course of the Holocene. Around 8.5 kyBP the end of iceberg drifting in the Norwegian Sea that is suggested by disappearing IRD deposits (Bauch & Weinelt, *in press*) heralds the begin of the Holocene climate optimum (Hypsithermal) (see also Nesje & Dahl, 1993) after a few centuries of slightly dropped sea surface temperatures (Younger Dryas II, Koç-Karpuz & Jansen, 1992; Koç *et al.*, 1996). Around this time, the English Channel was already open and a powerful current system that includes the Jutland and Norwegian Coastal Currents established itself in the North Sea (e.g. Stabell & Thiede, 1985). The interplay between eustasy and isostasy in the course of the Holocene lead to marine and freshwater stages of the Baltic Sea, which may have had impact on the oceanography of the northeastern North Atlantic (Bauch & Weinelt, *in press*; see also Bergsten & Nordberg, 1992). The Skagerrak, which forms the gateway between the Baltic Sea and the North Atlantic, proved to be a high resolution archive of Holocene climate fluctuations (e.g. Stabell & Thiede, 1985; Conradsen & Heier-Nielsen, 1995; Hass, 1996, *this volume*). Periods such as the Little Ice Age (ca. 1350-1900 AD) are characterized by wind forced-fluctuations in current strengths (Hass & Kaminski, 1994).

Regarding the large scale oceanographic patterns, no exceptional longer term paleoceanographic changes occurred during the past 10,000 years in the northern North Atlantic. Moderate climate fluctuations are recorded mainly in shelf sediments (e.g. Stabell & Thiede, 1985) because limited temporal resolution mostly prevents the accurate dating of short-term oceanographic and climatic events in sediment cores from the deep sea. During the Holocene - like today - water mass circulation in the North Atlantic is characterized by the Gulf Stream that forms the northern part of the Subtropical Gyre. Towards the northeast, it forms the southern part of the Subpolar Gyre as the North Atlantic Current. It passes Scotland and eventually meets the Norwegian Coastal Current, which adds North Sea water masses, reducing its salinity from 35.1-35.3 PSU to <34.7 PSU (Johannessen, 1986). Providing heat and moisture to Europe, this current continues further northward as the Norwegian Current. Through cooling and evaporation, these northbound water masses gain density; eventually portions sink and form NADW in the Greenland and Iceland Seas, thus fuelling the Ocean Conveyor that links the oceans of the world (e.g. Veum *et al.*, 1992). Meridionally separated from Atlantic water masses by the Arctic Front, the East Greenland Current conveys cold Arctic water masses south along Greenland's coastline.

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This paper was meant to be a short introduction to the topics that are addressed by the authors contributing to this book. It took me a while to figure out that it is kind of impossible to even produce a snap-shot of the state-of-the-art of a scientific field that is so lively discussed as is the paleoceanography of the North Atlantic. Thus, snowed under papers, journals, and books I decided to finally give up although knowing quite well that there are still important papers left to add. I tried to take any care not to misrepresent conclusions drawn in the various papers referred to in this introduction. If I did anyway in the heat of finishing this book I apologize herewith.

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