QUATERNARY RESEARCH 43, 185–197 (1995)

Reflection of Scandinavian Ice Sheet Fluctuations in Norwegian Sea Sediments during the Past 150,000 Years

KARL-HEINZ BAUMANN,*,† KLAS S. LACKSCHEWITZ,* JAN MANGERUD,‡ ROBERT F. SPIELHAGEN,* THOMAS C. W. WOLF-WELLING,* RÜDIGER HENRICH,† AND HEIDEMARIE KASSENS*

*GEOMAR Research Center for Marine Geosciences, Wischhofstrasse 1-3, D-24148 Kiel, Germany; ‡Department of Geology, University of Bergen, Allégt. 41, N-5007 Bergen, Norway; and †Fachbereich Geowissenschaften, University of Bremen, Postfach 330 440, D-28334 Bremen, Germany

Received April 12, 1994

The record of glacier fluctuations in western Scandinavia, as reconstructed from continental data, has been correlated with records of ice-rafted detritus (IRD) from well-dated sediment cores from the Norwegian Sea covering the past 150,000 yr B.P. The input of IRD into the ocean is used as a proxy for ice sheet advances onto the shelf and, thus, for the calibration of a glaciation curve. The marine results generally support land-based reconstructions of glacier fluctuations and improve the time-control on glacial advances. The Saalian ice sheet decayed very rapidly approximately 125,000 yr B.P. In the Early Weichselian, a minor but significant IRD maximum indicates the presence of icebergs in isotope substage 5b (especially between 95,000 and 83,000 yr B.P.). Reduced amounts of calcareous nannofossils indicate that surface waters were influenced by meltwater discharges during isotope substages 5d and 5b. An extensive build-up of inland ice began again during isotope stage 4, but maximum glaciation was reached only in early stage 3 (58,000–53,000 yr B.P.). Marine sediments have minimum carbonate content, indicating strong dilution by lithogenic ice-rafted material. Generally, the IRD accumulation rate was considerably higher in stages 4–2 than in stage 5. A marked peak in IRD accumulation rates from 47,000 to 43,000 yr B.P. correlates well with a second Middle Weichselian ice sheet advance dated by the Laschamp/Olby paleomagnetic event. Minimum ice extent during the Ålesund interstade (38,500–32,500 yr B.P.) and several glacial oscillations during the Late Weichselian are also seen in the IRD record. Of several late Weichselian glacial oscillations on the shelf, at least four correspond to the North Atlantic Heinrich events. Ice sheet behavior was either coupled or linked by external forcing during these events, whereas internal ice sheet mechanisms may account for the noncoherent fluctuations.

INTRODUCTION

Fluctuations of the Scandinavian Ice Sheet during the last interglacial/glacial (the Eemian/Weichselian) cycle have been reconstructed from the few onshore locations at which sediments older than the Late Weichselian glacial maximum have been discovered (Mangerud, 1991a). Most of the older sediments were removed by the Late Weichselian ice sheet, leaving an incomplete terrestrial record for the earlier, and longest, part of the last glaciation. The reconstruction of ice sheet fluctuations also remains debatable because accurate dating methods are not applicable for most continental sediments older than 50,000 yr B.P.

Contrary to the continental record, deep-sea sediment cores from the neighboring Norwegian Sea contain continuous records of the last interglacial/glacial cycle. High-resolution oxygen isotope stratigraphies have been established for many sediment cores (e.g., Kellogg *et al.*, 1978; Scholten *et al.*, 1990; Vogelsang, 1990; Weinelt, 1993) and confirmed by radiometric dating (^{14}C, ^{230}Th). Climatically induced paleoceanographic changes in the Norwegian Sea have been detected in these cores by sedimentological (Kellogg, 1976; Henrich *et al.*, 1989; Kassens, 1990), and micropaleontological investigations (Gard, 1988; Baumann, 1990; Bauch, 1993) as well. However, there is only a little information available about the correlation of marine and terrestrial data, although Larsen and Sejrup (1990) have shown a correspondence between the influx of Atlantic water into the Norwegian Sea and interstades in western Norway.

In this paper, we correlate the continental record of glacier fluctuations in western Scandinavia with marine sediment records from the Vøring Plateau area in the Norwegian Sea, which was adjacent to the western flank of the Scandinavian Ice Sheet. We use the input of coarse lithogenic ice-rafted detritus (IRD) into the ocean as a monitor of glacier fluctuations, as most of this material was eroded by the Scandinavian Ice Sheet and released to the ocean by iceberg calving and subsequent melting. In turn, the good chronology of the cores can be applied for a more precise dating of the glacial history. The main result is that there is a close similarity between the glaciation curve constructed from continental records and the IRD curve.

To define relatively warm time intervals in the ocean, we use other parameters from the marine cores, such as the calcium carbonate content and abundance of cocco-

0033-5894/95 $6.00

liths. These data reflect variations in biologic productivity, which is related to environmental parameters such as water temperatures, nutrients, and ice coverage.

Today, the Norwegian–Greenland Sea is characterized by strong meridional gradients due to the warm Norwegian–Westspitsbergen Current in the east and the cold polar East Greenland Current in the west (Hopkins, 1988). The Vøring Plateau area is situated east of the Norwegian–Greenland Sea (Fig. 1) and is characterized by relatively warm (6–10°C) and saline (35.1–35.3‰) Atlantic surface water. The warm current plays an important role in the transport of moisture and heat toward the north and for the climate in Scandinavia. In the past, variations of this current certainly influenced glacier fluctuations.

CONTINENTAL RECORDS OF GLACIER FLUCTUATIONS

The glaciation curve (Fig. 2) is only slightly modified from Mangerud (1991a), and for most of the primary data and discussions we refer to that publication. Here, we will only describe some of the principles for the construction of the curve and discuss modifications from newer results.

The original diagram (Mangerud, 1991a, slightly modified in 1991b) was presented as one glaciation curve for the western, oceanward part of the ice sheet, and one curve for the eastern, continental part, because the two segments certainly developed somewhat differently. However, with our limited knowledge, we are not really able to differentiate the two before 14,000 yr B.P., and we therefore now present only one curve. Most of the information on glacial maxima is from the west coast; thus, the curve is basically a glaciation curve for the western flank of the ice sheet. The best data, indicating periods during which the central part of Scandinavia was ice-free, come from the eastern side of the watershed in Sweden. However, deglaciation of that area implies that most of the western side also was ice-free, even if there was some asymmetry in the extension of the glaciers.

Because different dating methods are used for different time periods, we will split the following discussion into three sections: (1) oxygen isotope stage 5 (130,000–75,000 yr B.P.), in which the chronology is mainly based on palynological correlations with sequences south of the glacial boundary; (2) stage 4 and the older part of stage 3 (75,000–40,000 yr B.P.), in which dating is problematic; and (3) the period after 40,000 yr B.P., in which the chronology is mainly based on radiocarbon dates. To adjust ^{14}C-dated sample ages to the calendar year time scale of the marine isotopic events, we have calibrated ^{14}C ages using a linear approximation of Bard *et al.* (1993) for the period 6000–18,000 ^{14}C yr and an extended second-order fit (E. Bard, personal communication, 1994) for the period 18,000–28,000 ^{14}C yr. All ages are given in calendar years (cal yr) B.P., except where indicated.

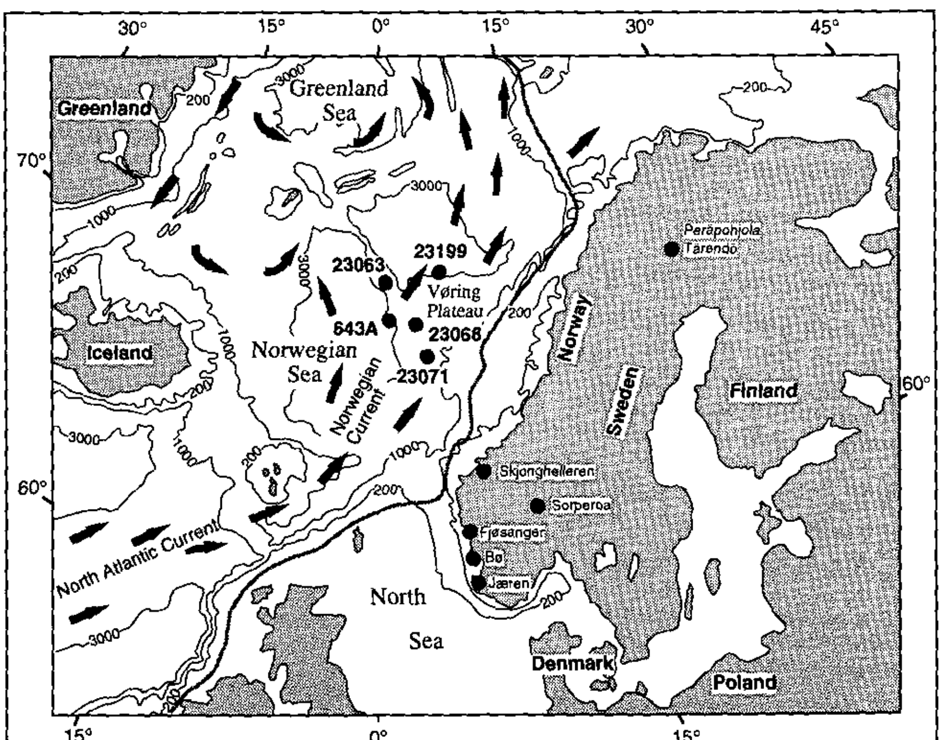

FIG. 1. Map showing bathymetry (in m), major surface water-mass circulation, sediment core locations in the Norwegian–Greenland Sea, and main localities of Weichselian deposits. The maximum extent of the Weichselian ice sheet is indicated by a hatched line (after Mangerud, 1991a).

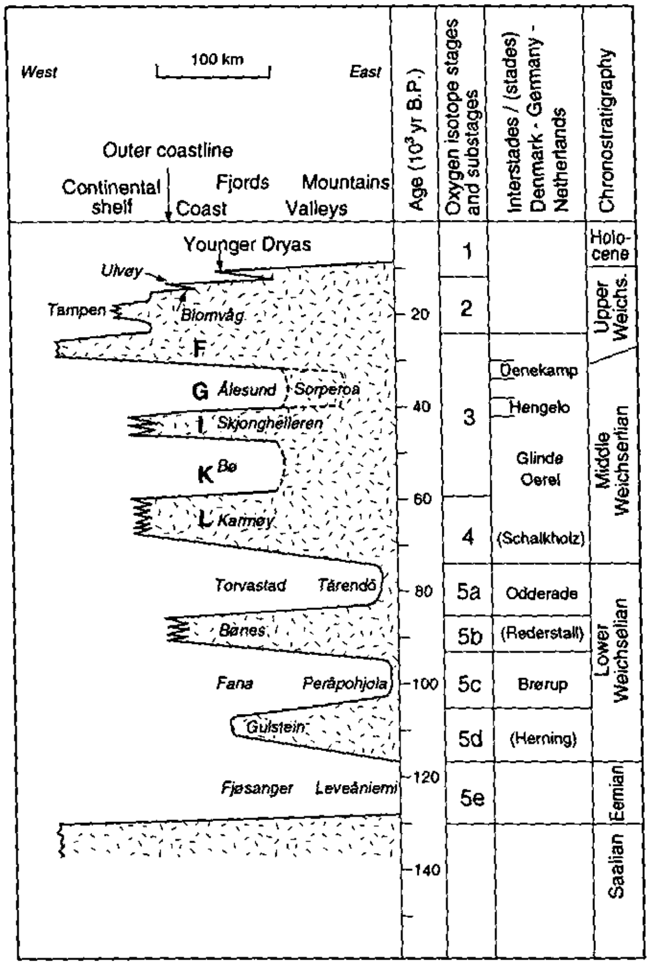

FIG. 2. Glaciation curve of western Scandinavia for the last interglacial/glacial cycle as reconstructed from continental data (modified from Mangerud, 1991a,b). The curve shows schematically the position of the ice front at different times. The time scale is in calendar yr. Radiocarbon dates younger than 18,000 yr are calibrated to calendar ages according to Bard (1993), whereas an extended second-order fit (E. Bard, personal communication, 1994) was used for the period 18,000–28,000 ^{14}C yr. The ages of the Oerel–Denekamp interstades are according to Behre and van der Plicht (1992). Ages of the isotope stage boundaries are according to Martinson *et al.* (1987).

Oxygen Isotope Stage 5

For isotope stage 5, there is a distinct pollen stratigraphy in western Europe, for which the correlation with the deep-sea isotope stratigraphy is generally accepted by most scientists (arguments summarized in Mangerud, 1989). The Eemian interglaciation correlates with isotope substage 5e (i.e., the time interval around event 5.5 in the marine nomenclature), the Brørup interstade with 5c (5.3), and the Odderade interstade with 5a (5.1) (Fig. 2). The Brørup includes the Amersfoort interstade in the first part. Thus, the Brørup, in this sense, has a short, cold interlude between an older mild period (Amersfoort) and a younger warmer period.

The Brørup and Odderade were the only interstades during the Weichselian that were warm enough to allow the growth of forests in northern Germany (Behre, 1989). In northern Sweden, Lagerbäck and Robertsson (1988) described several sites with two interstades (Peräpohjola and Tärendö) in superposition that both had birch trees. It is unlikely that trees could grow in northern Sweden unless there was forest in northern Germany, and, therefore, the two interstades in Sweden are correlated with Brørup and Odderade, respectively. This means that most of Scandinavia was ice-free during these two interstades, even though mountain glaciers certainly were larger than at present. The existence of the two interstades was recently supported from another area in central Sweden (Lundqvist and Miller, 1992).

In northern Sweden, the Eemian, Peräpohjola, and Tärendö deposits are separated by tills, demonstrating that there were glacier advances from the Scandinavian mountains during substages 5d (5.4) and 5b (5.2). How far the glacier advanced during these intervals needs to be mapped in more peripheral parts of the ice sheet, but few sites have as yet been found in those areas. At the Fjøsanger site (Mangerud *et al.*, 1981), a glacialmarine silt shows that the glacier reached within 1–3 km of the site during substage 5d. Thus, in western Norway this glaciation can be compared with the Younger Dryas in extent, and the ice front was situated in the fjords, rather than beyond the coast.

Above the glacialmarine silt are deposits of the milder Fana interstade and the thick Bønes Till, which Mangerud (1991a) correlated with substages 5c (Brørup) and 5b (Rederstall), respectively. By this interpretation, the ice reached close to, or beyond, the outer coastline for the first time during substage 5b. Larsen and Sejrup (1990) reinterpreted this chronology so that the Fana Interstade should represent only the first part of the Brørup, and the Bønes Till the cooler interval within the Brørup mentioned above. We disagree that the amino acid ratios allow such a precise age estimate of the Fana interstade; the main question, in fact, is whether it is of Brørup or older age. We also consider the assumption that the first large ice advance should have taken place during the cool inter-Brørup phase, which lasted only 400–800 yr (Grüger, 1991), unlikely, and thus we retain the interpretation of Mangerud (1991a,b).

Based on TL dating, the Godøy sandur was also considered to be evidence of 5d glaciation (Mangerud, 1991a). However, with correction for the recently discovered shallow TL traps (Mejdahl *et al.*, 1992), this sandur should be of Late Saalian (isotope stage 6) age, if the TL dating is reliable.

Oxygen Isotope Stage 4/First Part of Stage 3

Mangerud (1991a) placed the Karmøy Diamicton at the Bø site in stage 4, based on age estimates of the underlying Torvastad and overlying Bø interstades (Andersen *et al.*, 1983; Miller *et al.*, 1983; Sejrup, 1987). In their reexamination of Mangerud's (1991a,b) curve, Larsen and Sejrup (1990) argued that the amino acid epimerization was more rapid in substages 5d–5a than in stages 4–2. This will influence the amino acid age estimates, and they moved the mentioned interstades back to substages 5c and 5a, respectively. We agree that epimerization took place more rapidly (Miller and Mangerud, 1986) (Fig. 2), but neither the D/L ratios nor the temperature history are known well enough to determine that these interstades are as old as proposed by Larsen and Sejrup. Andersen *et al.* (1991) concluded that there was a major isotope stage 4 glaciation across Jæren, but again the dating results are ambiguous.

Several observations in Denmark indicate that the Old Baltic Till, providing evidence of a major glaciation of the Baltic Sea, was deposited during or soon after stage 4 (Petersen, 1985; Houmark-Nielsen, 1989; Petersen and Kronborg, 1991). In Poland, most scientists (e.g., Drozdowski and Fedorowicz, 1987; Mojski, 1992) agree that there was a glaciation ca. 60,000 to 50,000 yr B.P., mainly based on TL data, supporting the age of the Old Baltic Till in Denmark.

In the Skjonghelleren Cave, a third interstadial bed (K) was dated by a single U-series speleothem date of 56,000 ± 4000 yr (Larsen *et al.*, 1987). This age has not been reproduced by new dating and is, therefore, possibly incorrect (E. Larsen, personal communication, 1993). Interstade K was preceded by a glaciation of assumed stage 4 age, during which ice overran the cave and reached the coast (Larsen *et al.*, 1987; Mangerud, 1991a).

Above the mentioned interstade K deposits in Skjonghelleren is a key horizon for future correlation of the glaciation curve with marine cores. Ice-dammed lake sediments show that a glacier advance reached the open coast simultaneously with a paleomagnetic excursion, which is most probably the Laschamp/Olby event (Larsen *et al.*, 1987, Løvlie and Sandnes, 1987) and which is already identified in some marine cores (Nowaczyk *et al.*, 1993). The best age estimates of the Laschamp/Olby event, and thus of this glacier advance, probably are 19 K/Ar dates from Iceland and 30 from France, giving mean values of 43,000 to 47,000 yr B.P. (Levi *et al.*, 1990).

We conclude that the land data indicate a glaciation during part of stage 4 and the early part of stage 3, followed by deglaciation on the west coast and another glacier advance at the time of the Laschamp/Olby event.

40,000 yr B.P. to Present

The west coast was subsequently ice-free during the Ålesund interstade (Mangerud *et al.*, 1981). The duration of that interstade was recently supported by more than 30 AMS dates of well-preserved bones from Skjonghelleren Cave (Larsen *et al.*, 1987) which all gave ages between 32,500 and 38,500 cal yr B.P. (S. Gulliksen, M. S. Skovhus, T. W. Stafford, E. Larsen, and J. Mangerud, unpublished data). The dates show unambiguously that the outer coast was ice-free, but cannot pinpoint the period at which the ice withdrew.

Bergersen *et al.* (1991) and Bergersen (1991) have obtained four TL dates of windblown sand from Sorperoa, Gudbrandsdalen in central Norway, which all yielded ages between 37,000 and 40,000 cal yr B.P. If the dates are correct, and windblown sediments are well suited for TL dating, they show that the glaciers had withdrawn far inland at that time. The problem with the site is that the dates are not supported by lithostratigraphy, as the windblown sediments are not covered by till. Lauritzen (1991) reported three U-series speleothem dates of ca. 30,000 yr B.P. which indicate a major deglaciation of Nordland. Thus, two completely different deposits and dating methods (TL and U-series) indicate that the ice withdrew far inland in Scandinavia during the Ålesund Interstade and, accordingly, that much of the large Late Weichselian ice sheet was formed during less than 10,000 yr. This conclusion is so unexpected that we will await further confirmation before we fully accept it.

Another paleomagnetic key horizon occurs in ice-dammed lake sediments above the Ålesund interstadial deposits in Skjonghelleren Cave, showing that the glacier again reached the open coast at the time of a paleomagnetic excursion. This excursion is correlated with the Lake Mungo event (Larsen *et al.*, 1987; Løvlie and Sandnes, 1987). According to M. Barbetti (personal communication, 1993) the best radiocarbon dates for the Mungo type site give a mean age of 29,000 ^{14}C yr B.P. (32,500 cal yr B.P.), while TL dates give 33,500 yr B.P. (Barbetti, 1980). The Lake Mungo (Australia) event is probably equivalent with the Mono Lake (California) event (Løvlie, 1989); for the latter, the most recent age estimate is 28,000 ^{14}C yr B.P. (31,500 cal yr B.P.) for the midpoint and the event had a duration of 2000 yr (Liddicoat, 1992).

Sejrup *et al.* (1994) showed that the last-mentioned ice advance crossed the North Sea and met the glacier on the British Isles. They also obtained two dates above the till that demonstrate ice retreat ca. 25,500 cal yr B.P. The age of the Late Weichselian glacial maximum should be in the interval 32,500 to 25,500 cal yr B.P., and closer to the younger age.

After the first withdrawal, the glacier readvanced to a less-extensive position where it stood between 19,000 and 15,500 ^{14}C yr B.P. (Vorren *et al.*, 1988; Sejrup *et al.*, 1994). During rapid deglaciation, the main halts/readvances occurred 12,400 to 12,100 ^{14}C yr B.P. and during the Younger Dryas stade 11,000–10,200 ^{14}C yr B.P. (Mangerud, 1980; Vorren *et al.*, 1988).

MARINE RECORDS

Materials and Methods

Five sediment cores from the Vøring Plateau in the Norwegian Sea were selected for this study (Fig. 1; Table 1). A detailed comparison of the sedimentological records allows the exclusion of local features. Giant gravity cores 23063, 23065, 23071, and 23199 were collected during the RV *Meteor* Cruises 2/2 and the ARK I cruise of RV *Polarstern*. Site 643 was drilled during Leg 104 of the Ocean Drilling Project (ODP). All cores consist of Late Quaternary hemipelagic muds. Detailed core descriptions are given in Henrich *et al.* (1989) and Eldholm *et al.* (1987).

A LECO CS-125 infrared analyzer was used for measuring total carbon (TC) and total organic carbon (TOC) contents of bulk sediments. Calcium carbonate content was calculated in weight percentages of the bulk sample by $CaCO_3\% = (TC\% - TOC\%) \times 8.33$.

The sediment samples were freeze–dried, weighed, and washed through a 63-μm mesh. A split (>500 grains) of the 125–500 μm fraction was microscopically analyzed because its composition is considered most representative for changes between pelagic biogenic sedimentation (foraminifers) and the input of coarse lithic fragments (IRD). In addition, cores 23065, 23199, and 643A were investigated for abundance of coccoliths using a scanning electron microscope (SEM). The processing technique has been described by Baumann (1990).

Time control for all cores is based on high-resolution oxygen isotope stratigraphy and accelerator mass spectrometry (AMS) ^{14}C dating. Oxygen isotope measurements were carried out on the planktic foraminifera *Neogloboquadrina pachyderma* (sin) (125–250 μm fraction) (Fig. 3) All measurements were performed on Finnigan MAT 251 mass spectrometers (Kiel University and University of Bergen). Results are expressed in the usual δ-notation, and external reproducibility for $\delta^{18}O$ is 0.08‰ (Kiel) and 0.1‰ (Bergen), respectively.

Accumulation rates (AR) were calculated according to Ehrmann and Thiede (1985):

$$\begin{aligned} AR_{CaCO_3}\ (g/cm^2 \times 10^3\ yr) &= AR_{BULK} \times CaCO_3 \\ &\quad \text{weight } \% \times 10^{-2} \\ \text{IRD weight } \% &= \text{IRD grain-}\% \\ &\quad \times \text{ weight-}\% > 63\ \mu m \\ &\quad \times 10^{-2} \\ AR_{IRD}\ (g/cm^2 \times 10^3\ yr) &= AR_{BULK} \times IRD \\ &\quad \text{weight } \% \times 10^{-2} \end{aligned}$$

where

$$\begin{aligned} AR_{BULK} &= \text{accumulation rate of bulk sediment} \\ &\quad (g/cm^2 \times 10^3\ yr) = LSR \times DBD \\ LSR &= \text{linear sedimentation rate } (cm/10^3\ yr) \\ DBD &= \text{dry bulk density } (g/cm^3). \end{aligned}$$

Oxygen Isotope Stratigraphy and Dating of Marine Records

Age determinations (cal yr B.P.) of the cores presented in this paper are based on correlation with the orbitally tuned isotope record (Martinson *et al.*, 1987). Oxygen isotope data sets for cores 23063, 23065, 23071, 23199, and Site 643 were partly taken from Vogelsang (1990), Ramm (1988), and Wolf (1991). We performed more isotope analyses and partly revised their age models. Ages between identified isotope events were calculated from sedimentation rates by linear interpolation. The identification of critical events 4.2 and 3.31 was also based on planktic $\delta^{13}C$ data, which were interpreted by Vogelsang (1990; adopted for this study) in accordance with results of Labeyrie and Duplessy (1985) and Shackleton *et al.* (1983).

In all cores, a brief event is seen early in stage 3 which is characterized by extremely high amounts of terrigenous grains (98–100 grain %) and a lack of carbonate (Fig. 4). According to the initial stratigraphies, the calculated minimum and maximum duration of this event is 2000 yr in core 23071 and 8600 yr in core 23199, respectively. The

TABLE 1

Locations, Water Depths, Core Recovery, and Estimated Age of Investigated Cores from the Norwegian Sea

Core no.	Cruise	Latitude (N)	Longitude (E)	Water depth (m)	Length of core (m)	Estimated age of core base (yr)	Source of $\delta^{18}O$-Data
23199	ARK I/3	68°22.6′	05°19.5′	1968	6.36	250,000	Ramm, 1989
23063	M2/2	68°45.0′	00°00.0′	2299	8.81	450,000	Vogelsang, 1990
23068	M2/2	67°50.0′	01°30.1′	2230	6.54	185,000	Vogelsang, 1990
23071	M2/2	67°05.1′	02°54.5′	1308	7.00	135,000	Vogelsand, 1990
643A	ODP Leg 104	67°42.9′	01°02.0′	2753	13.92[a]	440,000[a]	Wolf, 1991

[a] Length/age of isotopically dated section.

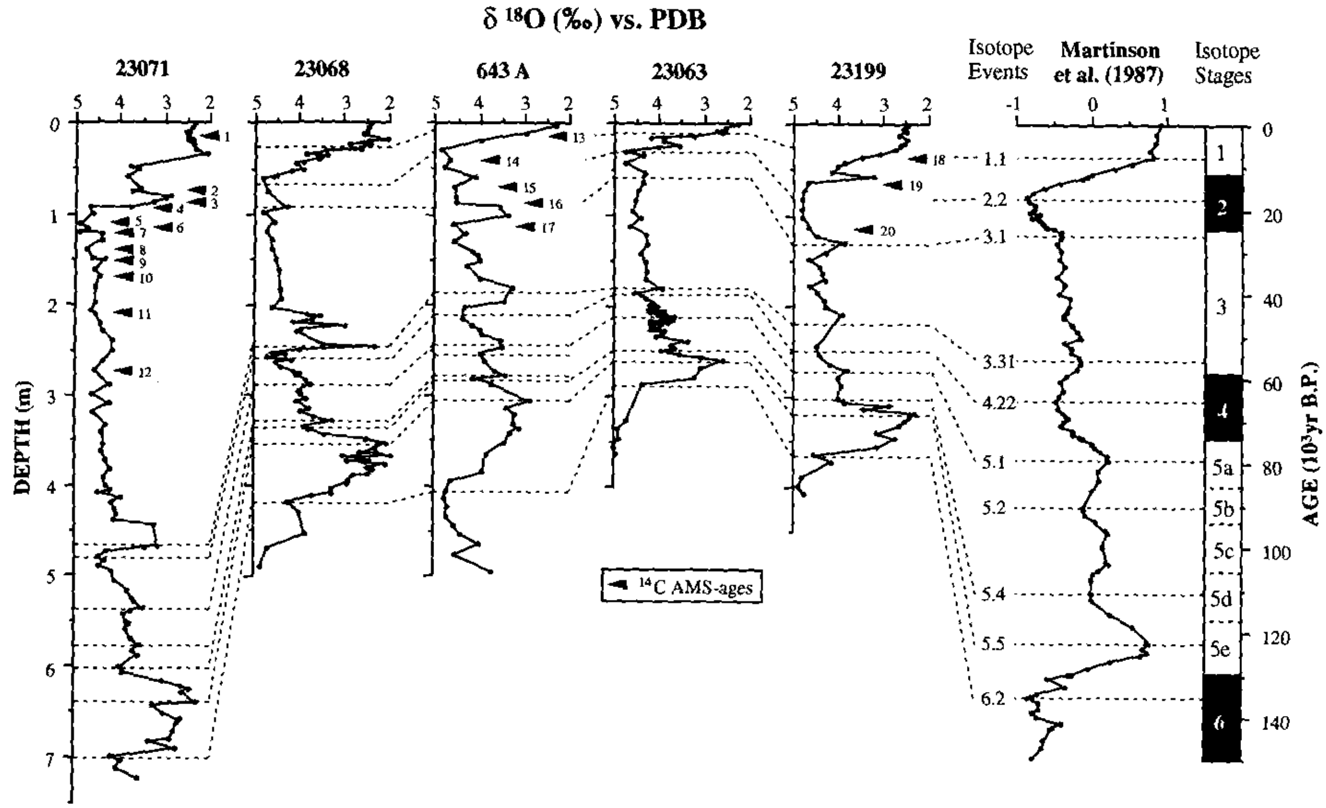

FIG. 3. Oxygen isotope records (*N. pachyderma* sin.) of analyzed cores from the Norwegian Sea correlated with the standard record of Martinson *et al.* (1987). Isotopic events 1.1 to 6.2 in our records were identified by graphic correlation to the standard record. Arrows indicate where AMS ^{14}C ages are available (Table 2).

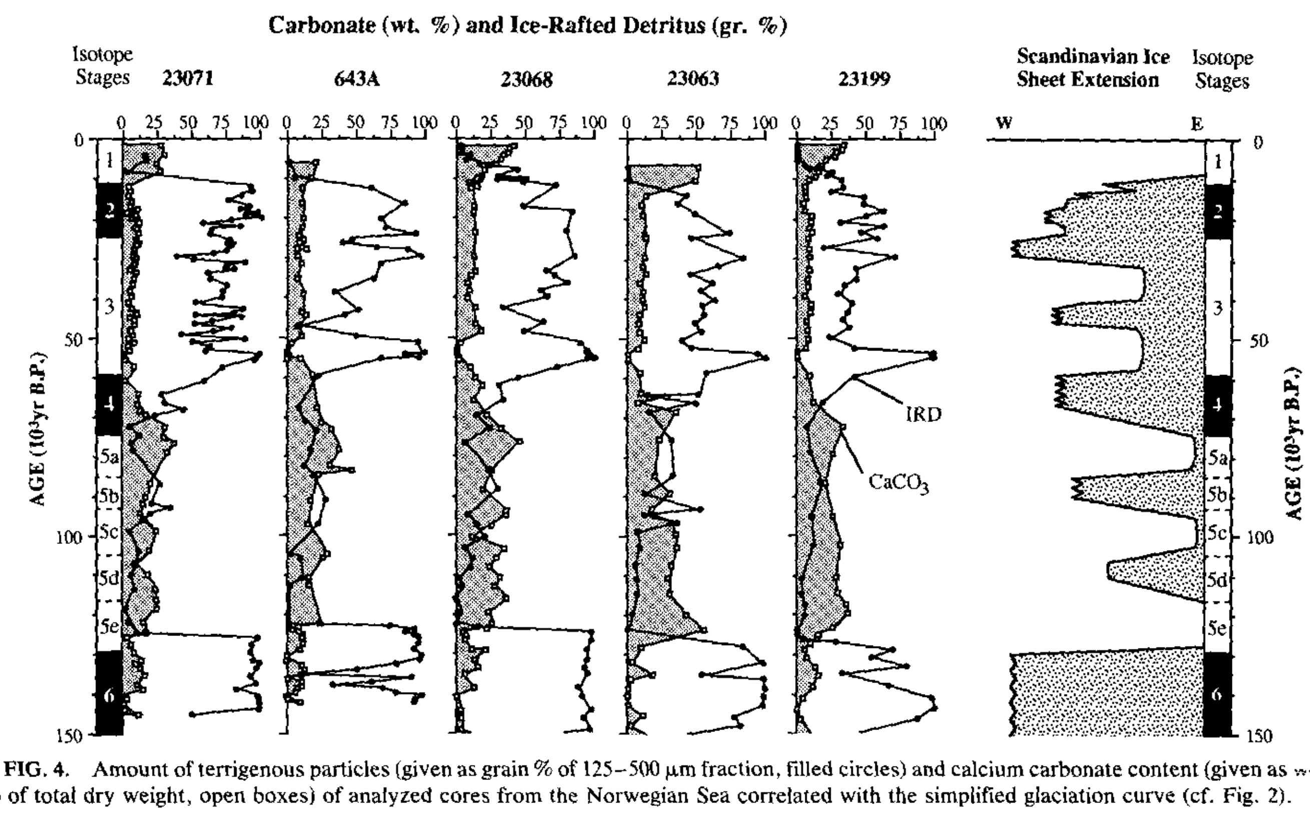

FIG. 4. Amount of terrigenous particles (given as grain % of 125–500 μm fraction, filled circles) and calcium carbonate content (given as weight % of total dry weight, open boxes) of analyzed cores from the Norwegian Sea correlated with the simplified glaciation curve (cf. Fig. 2).

given stratigraphic resolution clearly means that the carbonate-free interval correlates with global oxygen isotope event 3.31 (55,450 yr B.P.; Martinson *et al.*, 1987) which is expressed as a warming and meltwater event in the entire Norwegian Sea (Vogelsang, 1990; Weinelt, 1993). Since we postulate that the carbonate-free interval is time-equivalent in all cores, the midpoint (54,400 yr B.P.) and duration (2000 yr) in the high-resolution core 23071, which is located nearest to the continental source for IRD, were transferred to the other cores, resulting in an intercalibration. In addition, absolute age control is available from ^{14}C-AMS dates of the younger parts of sediment cores 23071, 23199, and 643 A (Table 2), which were calibrated by the procedure described above for terrestrial samples.

Unfortunately, some oxygen isotope records have gaps due to the lack of planktic foraminifers and/or sampling resolution. These gaps sometimes occur in core sequences that are critical for an accurate identification of isotopic events. Age determinations for the calculation of sedimentation rates and the timing of the paleoclimatic changes are therefore limited by the resolution between the assigned isotopic events. The transition from glacial stage 6 to interglacial stage 5 is defined by a sudden δ^{18}O decrease between isotope events 6.2 and 5.53 (Martinson *et al.*, 1987). However, we have not been able to identify 5.53, and the ages are interpolated between 6.2 and the peak of 5.5. Therefore, the duration of short-term sedimentation events, e.g., the drastic increase in IRD deposition at the end of stage 6, may have been calculated as lasting too long; the reason is that during deglaciations, meltwater produces local isotopic excursions (Jones and Keigwin, 1988; Weinelt *et al.*, 1991; Sarnthein *et al.*, 1992). We interpret the drastic δ^{18}O decrease after isotope event 6.2 in our cores as a local meltwater signal and have therefore defined only the peak of event 5.5 (123,820 yr) in the isotope records and used this for correlation and age determination.

Calcium Carbonate Records

Downcore calcium carbonate records reveal characteristic variations that can be correlated between the cores (Fig. 4). Although peak values are different from core to core, highest $CaCO_3$ contents (up to 55%) were generally measured in interglacial sediments from oxygen isotope stages 1 and 5 (especially substage 5e). In glacial sediments from stages 6 and 4–2, $CaCO_3$ contents are lowest in all cores. Most samples analyzed had contents of <15 wt%. In all cores, sections nearly free of carbonate were observed during stage 6 (ca. 145,000 yr B.P.) and earliest stage 3 (ca. 54,400 yr B.P.). Abrupt changes in $CaCO_3$ content are recorded during glacial/interglacial transitions 6/5 and 2/1.

Ice-rafted Detritus

In all cores, lithogenic particles mainly consist of monocrystalline grains (quartz, feldspar), polycrystalline

TABLE 2
Conventional Reservoir-Corrected ^{14}C Ages (yr B.P.) and Calibrated Ages of Marine Samples

No.	Core	Depth (cm)	^{14}C age (yr B.P.) (reservoir corr.)	Laboratory code	^{14}C age (cal yr B.P.)	Source of data point
1	23071	15	6680 ± 120		7443	Vogelsang, 1990
2		75	12,550 ± 220		14,722	Vogelsang, 1990
3		82	13,600 ± 300		16,024	Vogelsang, 1990
4		93	14,790 ± 190		17,500	Vogelsang, 1990
5		112	16,550 ± 200		19,682	Vogelsang, 1990
6		116	16,690 ± 230		19,856	Vogelsang, 1990
7		120	16,990 ± 220		20,228	Vogelsang, 1990
8		140	18,520 ± 240		21,929	Vogelsang, 1990
9		150	18,790 ± 240		22,246	Vogelsang, 1990
10		170	21,430 ± 300		25,294	Vogelsang, 1990
11		210	23,730 ± 360		27,883	Vogelsang, 1990
12		273	28,020 ± 560		32,548	Vogelsang, 1990
13	643A	16	9590 ± 120	AAR-1138	11,052	This study
14		44	18,240 ± 220	AAR-1139	21,600	This study
15		67	21,200 ± 250	AAR-1140	25,031	This study
16		85	22,600 ± 270	AAR-1141	26,619	This study
17		114	27,300 ± 380	AAR-1142	31,780	This study
18	23199	37	8990 ± 340		10,308	Ramm, 1988
19		65	15,420 ± 215		18,281	Ramm, 1988
20		113	21,120 ± 350		24,940	Ramm, 1988

Note. The ^{14}C ages were corrected for an ocean-reservoir age of 400 yr. Conventional radiocarbon years were converted to calendar years (cal yr B.P.) by using a linear approximation of Bard *et al.* (1993) for the period 6000–18,000 ^{14}C yr and an extended second-order fit (E. Bard, personal communication, 1994) for the period 18,000–28,000 ^{14}C yr.

fragments of igneous rocks, and fragments of sedimentary rocks (shales, sandstones). The amount of detrital carbonate rock fragments is generally negligible (Bischof, 1990). According to Bischof (1994), most of the mono- and polycrystalline particles in sediment cores from the Norwegian Sea probably originate from Scandinavia. Previous studies of marine sediments from high latitudes have shown that coarse lithogenic particles (>63 μm) can be interpreted as ice-rafted detritus if other transport mechanisms (e.g., gravity flows, boundary currents) can be excluded (e.g., Henrich, 1988; Spielhagen, 1991; Bond *et al.*, 1992).

The IRD data are presented in two ways: as the relative abundance of IRD in the coarse fraction (grain %, Fig. 4) and as IRD accumulation rates (g/cm^2 × 10^3 yr; Fig. 5). Grain % data document even slight changes in the depositional paleoenvironment by quantifying the relative importance of ice-rafting vs biogenic sedimentation. During initial cooling and ice-sheet growth, the onset of increased ice-rafting and reduced bioproduction will be visible first from an increase in relative IRD abundance. IRD accumulation rates are considered an excellent measure of the iceberg output from continental ice sheets. Because IRD accumulation rates are sensitive to the dilution of IRD sedimentation by fine-grained material, the initiation of ice-rafting during the growth of an ice sheet will not be indicated.

Large amounts of IRD (40 to 100 grain %) are found in sediments from late oxygen isotope stage 6 and from stages 2 and 3. Peak values above 75 grain % occur in all cores during the time intervals 150,000–124,000 yr B.P., 58,000–50,000 yr B.P., and 28,000–10,000 yr B.P. In sediments from stages 5, 4, and the 1, values are usually relatively low (0 to 15 grain %). However, small peaks of up to 30 grain % can be identified in all cores in sediments from 95,000–80,000 yr B.P., which roughly corresponds to oxygen isotope substage 5b (93,000–85,000 yr B.P.).

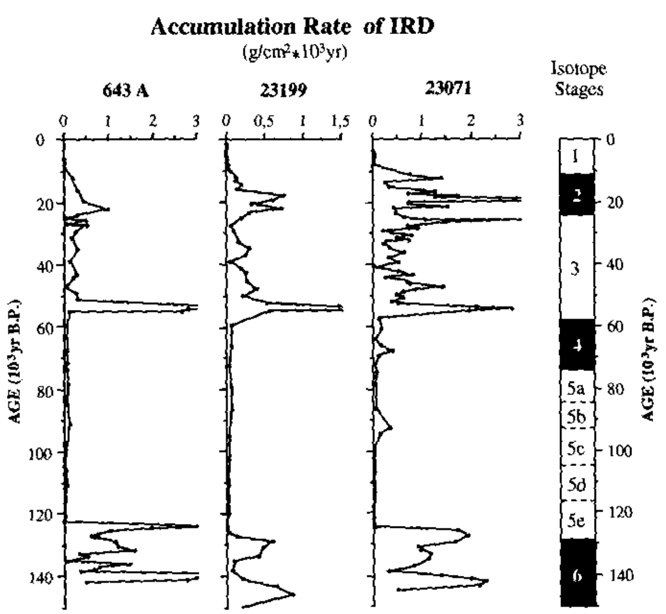

FIG. 5. Accumulation rates of IRD during the last interglacial/glacial cycle in sediment cores 643 A, 23199, and 23071 from the Norwegian Sea.

IRD accumulation rates varied between almost 0 and 3.7 g/cm^2 × 10^3 yr (core 23071); typically, high values are between 0.5 and 2 g/cm^2 × 10^3 yr in all cores (Fig. 5). Significant minima seem to correlate in cores and can be recognized at ca. 135,000 yr B.P., from 124,000 to 58,000 yr B.P. and from 42,000 to 24,000 yr B.P. The same is obvious for the pattern of maximum values that covers nearly the remaining time intervals, with distinct maxima of IRD fluxes around 145,000 and 130,000 yr B.P. in oxygen isotope stage 6 and in stages 2 and 3. Short peak events are best seen in high-resolution core 23071 at 54,400, 47,700, 26,200, 21,900, 19,700, 18,400, and 13,000 yr B.P. (Fig. 5).

DISCUSSION

Carbonate Production and Surface Water Environments in the Norwegian Sea

The difference in carbonate content of the sediments can be used to distinguish surface water masses in the Norwegian–Greenland Sea (Kellogg, 1976; Henrich *et al.*, 1989; Baumann *et al.*, 1993). High carbonate shell production (i.e., planktic foraminifers and coccolithophorids) is found in the Atlantic water masses of the Norwegian Current, whereas low production and relatively high dissolution of $CaCO_3$ characterize the ice-covered polar surface water masses (Hebbeln and Wefer, 1991; Carstens and Wefer, 1992; Samtleben and Schröder, 1992). Thus, the low carbonate contents in sediments of oxygen isotope stages 6, 4, 3, and 2 are interpreted to reflect cold, often ice-covered surface water masses. Especially around 145,000 yr B.P. and from 58,000 to 53,000 yr B.P., production was extremely low. In contrast, high calcium carbonate contents during stages 5 and 1 reflect a stronger inflow of North Atlantic surface water (Kellogg, 1976; Henrich *et al.*, 1989) and a lack of sea ice in the area of the Vøring Plateau, at least during substage 5.5 and the Holocene (Baumann, 1990; Bauch, 1993; Vogelsang, 1990).

Calcareous nannofossil abundances show strong variability and three distinct abundance peaks during oxygen isotope stage 5 (Fig. 6). The production of planktic carbonate in the Norwegian Sea was enhanced during events 5.5, 5.3, and 5.1, whereas it was drastically restricted during events 5.4 and 5.2. This pattern was observed in a number of sediment cores from the northern Norwegian-Greenland Sea and Fram Strait (e.g., Gard, 1988; Baumann, 1990). Coccolithophorids are predominantly autotropic plankton and largely restricted to the photic zo

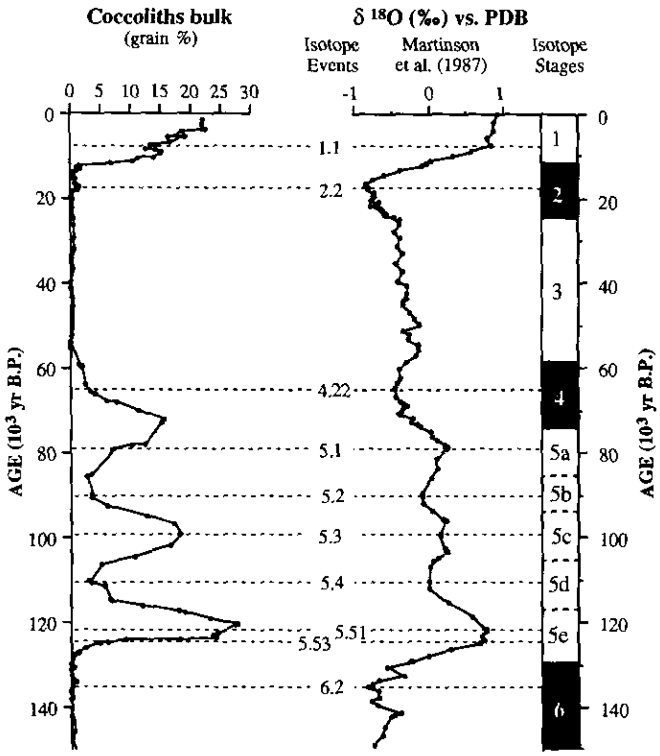

FIG. 6. Stacked curve of coccolith abundance (grain %, three-point moving average smoothed) of cores 23071, 643 A, and 23199 (from Baumann (1990) and unpublished data) correlated to the normalized oxygen isotope record of Martinson *et al.* (1987).

of the water column (0–30 m depth in the Norwegian–Greenland Sea). Therefore, the absence or drastic decrease in the amount of calcareous nannofossils in sediments from events 5.4 and 5.2 is most probably due to low temperature and salinity. During glacial stages 6, 3, and 2, planktic productivity was generally suppressed, most probably caused by an at least seasonal ice cover. Low abundances of coccoliths during oxygen isotope stages 3 and 2 (Gard, 1988; Hebbeln *et al.*, 1994; Fig. 6) suggest that some sea ice melting occurred and open waters formed intermittently, also as far north as Fram Strait.

Transport of IRD to the Norwegian Sea

The amounts of coarse lithogenic material in our cores indicate that deposition from ice rafting was the dominant sedimentation process in the Norwegian Sea during various intervals in the Late Quaternary. Because sea ice sediments are mostly fine-grained and contain little or no terrigenous grains (>63 μm; Pfirman *et al.*, 1989; Nürnberg *et al.*, 1994), icebergs from glaciers are proposed as the transport agent for the bulk of the sand-sized terrigenous material. Icebergs may contain sediments of all grain sizes, from clay to boulders; we therefore conclude that large amounts of the fine-grained terrigenous matter (<63 μm) were also ice-rafted.

There are several processes involved in the glacial erosion of a particle, its release from a floating iceberg, and its deposition on the sea floor. The processes are influenced by such variables as subglacial temperatures, glacier velocities, bedrock properties, position of the ice front relative to the coastline, calving rates, iceberg sizes, sea and air temperatures (determining melting rates), and ocean current directions determining the distribution of the icebergs. Thus, there is no simple relationship between fluctuations of the Scandinavian Ice Sheet and the flux of IRD in the area of our cores. We will not discuss the processes in detail, but will mention some of the factors.

The first icebergs from the Scandinavian Ice Sheet broke off when the glaciers advanced to the head of the Norwegian fjords, but most of them probably melted in the fjords. Iceberg release and IRD production to the open ocean, and thus to our core sites, must have increased continuously as the glaciers proceeded to the fjord mouths. A step increase occurred when the ice front reached the outer coastline. Maximum production probably occurred when a broad ice front reached the continental shelf edge (Powell, 1984). Peak values, indicating maximum output, were reached during maximum extension of an ice sheet (cf. Bond *et al.*, 1992) and during deglaciation, when large volumes of ice, accumulated earlier, were released.

We postulate that high IRD values in our cores indicate that many icebergs from the Scandinavian Ice Sheet were drifting in the Norwegian Sea because the ice sheet was large and/or breaking up. Thus, the IRD fluctuations are proposed to reflect oscillations of the ice sheet. From our results the highest numbers of icebergs in the Norwegian Sea can be inferred for the time intervals 150,000–124,000, 58,000–50,000, and 28,000–10,000 yr B.P., as indicated by maximum values of amounts and accumulation rates of coarse terrigenous particles at all sites (Fig. 7). In the Lower Weichselian, a minor but significant maximum (up to 30 grain %) indicates the presence of icebergs in the Norwegian Sea in the time interval from 95,000 to 80,000 yr B.P.

A gradual change of the paleoenvironment is documented for the early/middle Weichselian boundary where the IRD (grain %) increases (Fig. 4), whereas a drastic change in the iceberg production resulting in strong input of coarse IRD is documented for the earliest middle Weichselian when the IRD accumulation rate (g/cm^2 × 10^3 yr) suddenly increased (Fig. 5).

The Weichselian IRD input to the Norwegian Sea differs from the formation of the North Atlantic Heinrich layers. While the Heinrich layers are distinct IRD-rich layers in a sequence of otherwise pelagic sediments rich in biogenic carbonate (Bond *et al.*, 1992; Andrews *et al.*, 1994; Grousset *et al.*, 1993), Norwegian Sea sediments from 58,000 to 10,000 yr B.P. are generally characterized

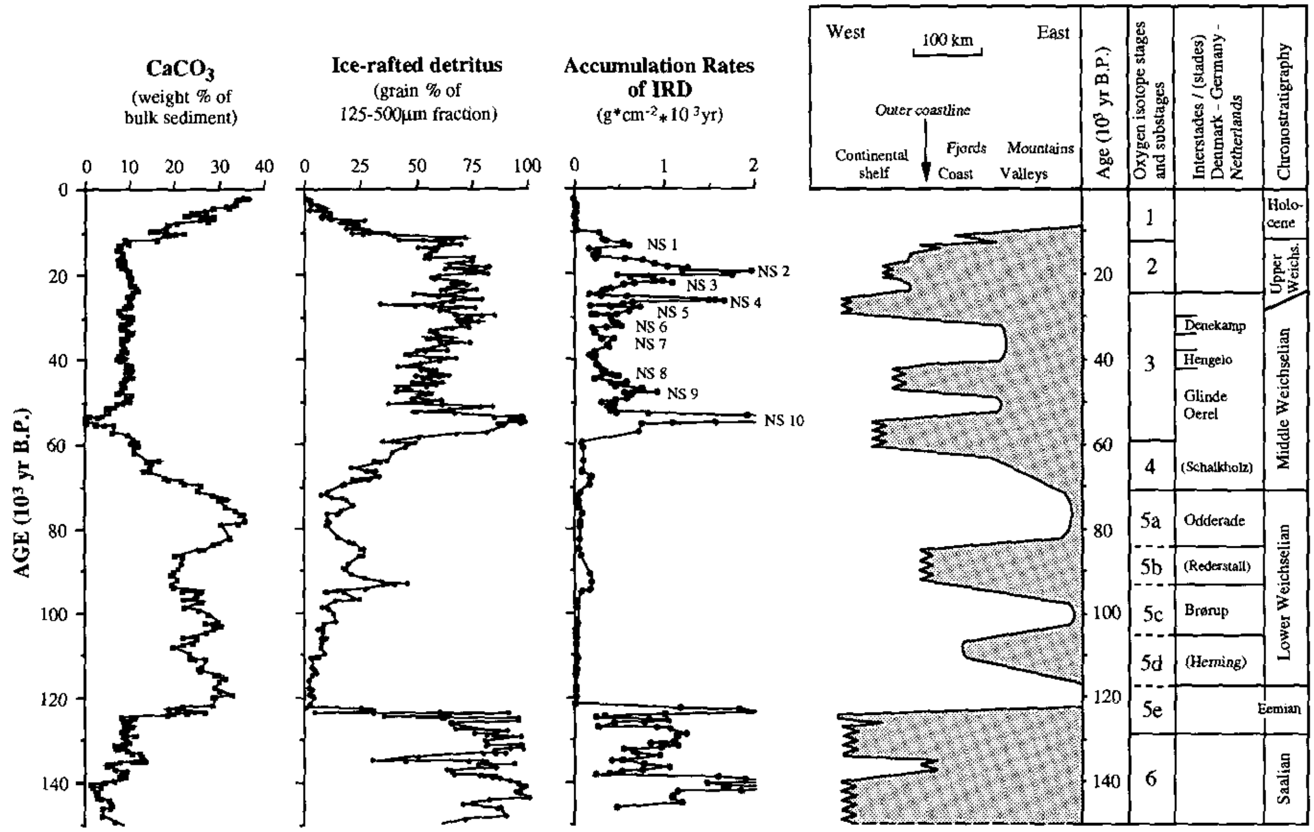

FIG. 7. Glaciation curve of western Scandinavia (right) as reconstructed from terrestrial data and time-calibrated by marine stacked records of $CaCO_3$ content, amount of IRD (all analyzed cores), and accumulation rates of IRD (cores 23071, 643 A, and 23199). A three-point moving average smoothing filter was applied to the marine data sets. NS 1–NS 10 mark sharp IRD deposition events in the Norwegian Sea during the Weichselian (Table 3).

by a high IRD content, indicative of a continuous iceberg supply. However, the distinct peaks in IRD accumulation rates in our cores (NS 1–NS 10; cf. Fig. 7) suggest short events of extreme iceberg discharge. Assuming an age uncertainty of ±2000 yr derived from the dating techniques, it is possible to correlate clearly two IRD-AR peaks (NS 4 and NS 6) with the North Atlantic Heinrich layers H 2 and H 3, respectively, while the correlations of NS 8 with H 4 and NS 10 with H 5 is more ambiguous (Table 3). Six further IRD-AR peaks seem to have no correlatives in the North Atlantic, and correlatives in Heinrich layers H 4 and H 6 are obviously not represented in the Norwegian Sea.

Oscillations of the Laurentide Ice Sheet are assumed to account for the Heinrich events (Broecker *et al.*, 1992; Bond *et al.*, 1992). The correlation of at least four Heinrich events with maximum iceberg calving from the Scandinavian Ice Sheet strongly suggests a link between the behavior of both ice sheets. Although the coincidence seems to point at a common, external forcing (e.g., eustatic sea level rise, sub-Milankovitch oscillations of the conveyor belt) for the calving events, T. Fronval *et al.* (unpublished data) suggest a coupling mechanism for ice sheet oscillations, where instability in one of the ice sheets can cause other ice sheets to behave in a coherent, phase-locked manner. In addition, the noncoherent rapid ice sheet advances may be explained by internal forcing of the individual ice sheets, e.g., the subglacial morphology and internal ice sheet mechanics.

Comparison of the Marine and Continental Glaciation Record

When comparing the reconstructed glaciation record of western Scandinavia with the IRD record of the sediment cores (Fig. 4), the general coincidence of glacial maxima and maximum IRD input is striking. This strongly implies that the events can be correlated. Minor glacier advances (e.g., 5d and 5b) are reflected mainly by reduced carbonate content, but not by strong IRD input. This may indicate a cooling event with reduced bioproduction in the Norwegian Sea. However, there is a significant difference in IRD accumulation rates between oxygen isotope substage 5d, when the glacier front did not reach the coastline, and 5b, when it passed the coastline. Here we will discuss only some major inconsistencies in the correlation of the records.

Terrestrial reconstruction indicates a rapid retreat of

TABLE 3
Comparison of IRD Events on the Vøring Plateau in the Norwegian Sea (NS 1–NS 10; This Study) with North Atlantic Heinrich Events (H1–H6)[a]

Norwegian Sea (NS) IRD events (cal yr B.P.)	North Atlantic Heinrich events (cal yr B.P.)	North Atlantic Heinrich events (^{14}C age)
	66,000 (H 6)	
54,300 (NS 10)	50,000 (H 5)	
47,500 (NS 9)		
43,750 (NS 8)	40,200 (H 4)	35,500 (H 4)
36,380 (NS 7)		
31,800 (NS 6)	31,500 (H 3)	27,000 (H 3)
28,450 (NS 5)		
26,800 (NS 4)	24,800 (H 2)	21,000 (H 2)
21,600 (NS 3)		
19,700 (NS 2)		
	16,900 (H 1)	14,300 (H 1)
13,000 (NS 1)		

[a] Ages of Heinrich events from Bond *et al.* (1993). For conversion of ^{14}C ages of H1–H4 to cal yr ages see text and Table 2.

glacier ice at the oxygen isotope stage 5/6 boundary (128,000 yr B.P.), while the IRD input to the Norwegian Sea clearly lasted until 125,000 yr B.P. In this case, the age discrepancy between both events is an artifact. Deglaciation on land was not dated directly, but was only correlated with the stage boundary. Taking into account the development during the last deglaciation, when the final deglaciation of Scandinavia occurred a few thousand years after the global oxygen isotopic transition, the delay of IRD compared to the isotope signal is not surprising.

The main discrepancy occurs in the interval from 70,000 to 50,000 yr B.P. In the terrestrial record, where this is the period with poorest age control, it is postulated that glaciers grew and reached the continental shelf during oxygen isotope stage 4 and retreated soon after. In the marine record, IRD input is relatively low from 69,000 to 63,000 yr B.P. and increased strongly to maximum values at 58,000 to 52,000 yr B.P. The peak coincides with a large (oxygen isotope) meltwater event (3.31). Thus, maximum IRD input corresponds to deglaciation in early stage 3, following short-term glaciation during stage 4. Taking the IRD record as a proxy for ice sheet advances onto the shelf and accepting that the marine chronology is more precise than the terrestrial, we have calibrated the glaciation curve for the western part of the Scandinavian Ice Sheet (Fig. 7).

CONCLUSIONS

1. Scandinavian Ice Sheet reached a position on the shelf, probably at the shelf edge, during the Saalian Glaciation. At 125,000 yr B.P., it decayed rapidly within 2000 yr or less, at the transition to the Eemian.

2. Two glacier advances occurred during the Early Weichselian (oxygen isotope substages 5d–5a). The first (substage 5d) did not reach the coastline, whereas the second (substage 5b) extended to the outer coastline in some areas and led to the release of icebergs. Generally, deposition of IRD was low, and therefore glaciation phases were brief or short-lived during entire stage 5.

3. During the Middle Weichselian, ice build-up may have started early in oxygen isotope stage 4, but according to the IRD signal, the maximum extent of the ice cap was not reached before 63,000 yr B.P., when the shelf must have been covered by glacier ice. This state of glaciation continued for a few thousand years and was terminated by a strong deglaciation event ca. 54,000 yr B.P.

4. A marked glaciation phase in the continental record correlated with the paleomagnetic Laschamp/Olby event dated at 43,000–47,000 yr B.P. and is identified as an IRD accumulation rate peak in the marine cores.

5. A partial deglaciation on land during the Ålesund interstade is well dated at 32,500–38,500 yr B.P. and identified as a low IRD accumulation rate period in the marine cores.

6. During the Late Weichselian, there were several glacial oscillations on the shelf, of which at least four correspond to the North Atlantic Heinrich events, suggesting a link between the behavior of both ice sheets.

ACKNOWLEDGMENTS

The idea for this paper came out of cooperative work in the European Science Foundation project "Polar North Atlantic Margins" (PONAM). Our work profited tremendously from previously established stratigraphies and published and unpublished data by A. Dettmer, M. Ramm, U. Struck, and E. Vogelsang, and from invaluable corrections made by J. Welling-Wolf. H. Erlenkeuser and the staff of the ^{14}C Laboratory of Kiel University performed the stable isotope measurements. J. Heinemeier (Aarhus University) conducted additional AMS ^{14}C datings and E. Bard gave valuable suggestions for the conversion of ^{14}C ages to calender ages. T. Reincke provided technical assistance in preparation of the data sets and figures. The manuscript profited strongly from thorough reviews by W. F. Ruddiman and one anonymous reviewer. This work was financially supported by the "Paläoklimaprojekt" of the Bundesministerium für Forschung und Technologie (Grant 07KF021/4), the Deutsche Forschungsgemeinschaft (Grant TH 200/10-1-2), the Sonderforschungsbereich 313 of Kiel University, and the Norwegian Research Council. We thank all persons and institutions mentioned.

REFERENCES

Andersen, B. G., Sejrup, H. P., and Kirkhus, L. (1983). Eemian and Weichselian deposits at Bø on Karmøy, SW Norway: A preliminary report. *Norges Geologiske Undersøkelse* 380, 189–201.

Andersen, B. G., Bakken, K., Dale, B., Fugelli, E., Henningsmoen, K. E., Høeg, H. I., Nagy, J., Read, A., and Tellemann, H. (1991). Weichselian Stratigraphy at Oppstad, Høgemork and Foss Eigeland, Jæren, SW Norway. *Striae* 34, 109–124.

Andrews, J. T., Erlenkeuser, H., Tedesco, K., Aksu, A. E., and Jull, A. J. T. (1994). Late Quaternary (Stage 2 and 3) meltwater and Heinrich Events, Northwest Labrador Sea. *Quaternary Research* 41, 26–34.

Barbetti, M. (1980). Geomagnetic strength over the last 50,000 years and changes in atmospheric 14C concentration: Emerging Trends. *Radiocarbon* 22, 192–199.

Bard, E., Arnold, M., Fairbanks, R. G., and Hamelin, B. (1993). ^{230}Th–^{234}U and ^{14}C ages obtained by mass spectrometry on corals. *Radiocarbon* 35, 191–199.

Bauch, H. (1993). Planktische Foraminiferen im Europäischen Nordmeer—ihre Bedeutung für die paläo-ozeanographische Interpretation während der letzten 600,000 Jahre. *Berichte Sonderforschungsbereich 313* 40, 108.

Baumann, K.-H. (1990). Veränderlichkeit der Coccolithophoridenflora des Europäischen Nordmeeres im Jungquartär. *Berichte Sonderforschungsbereich 313* 22, 146.

Baumann, K.-H., Lackschewitz, K. S., Erlenkeuser, H., Henrich, R., and Jünger, B. (1993). Late Quaternary calcium carbonate sedimentation and terrigenous input along the East Greenland continental margin. *Marine Geology* 14, 13–36.

Behre, K.-E. (1989). Biostratigraphy of the last glacial period in Europe. *Quaternary Science Reviews* 8, 25–44.

Behre, K.-E., and Plicht, J. van der (1992). Towards an absolute chronology for the last glacial period in Europe: Radiocarbon dates from Oerel, northern Germany. *Vegetation History Archaeobotany* 1, 111–117.

Bergersen, O. F. (1991). Norske mammutfunn-og kvartærgeologi. *Naturen* 6, 254–262.

Bergersen, O. F., Thoresen, M., and Hougsnæs, R. (1991). Evidence for a newly discovered Weichselian Interstadial in Gudbrandsdalen, Central South Norway. *Striae* 34, 103–108.

Bischof, J. (1990). Dropstones in the Norwegian-Greenland Sea—Indications of a late Quaternary circulation pattern? *In* "Geologic History of the Polar Oceans: Arctic versus Antartic" (U. Bleil and J. Thiede, Eds.), NATO ASI Series C (308), pp. 499–518. Kluwer Academic, Dordrecht.

Bischof, J. (1994). The decay of the Barents ice sheet as documented in nordic seas ice-rafted debris. *Marine Geology* 117, 35–55.

Bond, G., Heinrich, H., Broecker, W., Labeyrie, L., McManus, J., Andrews, J., Huon, S., Jantschik, R., Clasen, S., Simet, C., Tedesco, K., Klas, M., Bonani, G., and Ivy, S. (1992). Evidence for massive discharges of ivebergs into the North Atlantic ocean during the last glacial period. *Nature* 360, 245–249.

Broecker, W. S., Bond, G., McManus, J., Klas, M., and Clark, E. (1992). Origin of the Northern Atlantic's Heinrich Events. *Climatic Dynamics* 6, 265–273.

Carstens, J., and Wefer, G. (1992). Recent distribution of planktonic foraminifera in the Nansen Basin, Arctic Ocean. *Deep-Sea Research* 39(S2A), S507–S524.

Drozdowski, E., and Fedorowicz, S. (1987). Stratigraphy of Vistulian glaciogenic deposits and corresponding thermoluminescence dates in the lower Vistula region, northern Poland. *Boreas* 16, 139–153.

Ehrmann, W., and Thiede, J. (1985). History of Mesozoic and Cenozoic sediment fluxes to the North Atlantic Ocean. *Contribution to Sedimentology* 15, 109.

Fronval, T., Jansen, E., Bloemendal, J., and Johnson, S. Stadial-interstadial cycles in the Norwegian Sea: Evidence for coherent changes in European and North American ice sheet on millennium timescales. *Nature*, submitted for publication.

Gard, G. (1988). Late Quaternary calcareous nannofossil biochronology and paleo-oceanography of Artic and Subartic Seas. *Meddelanden Stockholms Universitets Geologiska Institution* 275, 1–45.

Grousset, F. E., Labeyrie, L., Sinko, J. A., Cremer, M., Bond, G., Duprat, J., Cortijo, E., and Huon, S. (1993). Patterns of ice-rafted detritus in the glacial North Atlantic (40–55°N). *Paleoceanography* 8, 175–192.

Grüger, E. (1991). Late Quaternary biostratigraphy in Northern Germany 150,000–15,000 years B.P. *Striae* 34, 7–14.

Hebbeln, D., and Wefer, G. (1991). Effects of ice coverage and ice-rafted material on sedimentation in the Fram Strait. *Nature* 350, 409–411.

Hebbeln, D., Dokken, T., Andersen, E. S., Hald, M., and Elverhøi, A. (1994). Moisture supply for northern ice-sheet growth during the Last Glacial Maximum. *Nature* 370, 357–360.

Heinrich, H. (1988). Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years. *Quaternary Research* 29, 143–152.

Henrich, R., Kassens, H., Vogelsang, E., and Thiede, J. (1989). Sedimentary facies of glacial-interglacial cycles in the Norwegian Sea during the last 350 ka. *Marine Geology* 86, 283–319.

Hopkins, T. S. (1988). The GIN Sea. Review of physical oceanography and literature from 1972 *Saclantcen Report*, SR-124. Saclant Undersea Research Centre, San Bartolomeo, Italy.

Houmark-Nielsen, M. (1989). The last interglacial-glacial cycle in Denmark. *Quaternary International* 3(4), 31–39.

Jones, G. A., and Keigwin, L. D. (1988). Evidence from the Fram Strait (78°N) for early deglaciation. *Nature* 336, 56–59.

Kassens, H. (1990). Verfestigte Sedimentlagen und seismische Reflektoren: Frühdiagenese und Paläo-Ozeanographie in der Norwegischen See. *Berichte Sonderforschungsbereich 313* 24, 117.

Kellogg, T. B. (1976). Paleoclimatology and paleoceanography of the Norwegian and Greenland Seas: The last 450,000 years. *Marine Micropaleontology* 2, 235–249.

Kellogg, T. B., Duplessy, J.-C., and Shackelton, N. J. (1978). Planktonic foraminiferal and oxygen isotopic stratigraphy and paleoclimatology of Norwegian Sea. *Boreas* 7, 61–73.

Labeyrie, L. D., and Duplessy, J.-C. (1985). Changes in the oceanic ^{13}C/^{12}C ratio during the last 140,000 years: High-latitude surface water records. *Palaeogeography, Palaeoclimatology, Palaeoecology* 50, 217–240.

Lagerbäck, R., and Robertsson, A. M. (1988). Kettle holes—Stratigraphical archives for Weichselian geology and palaeoenvironment in northernmost Sweden. *Boreas* 17, 439–468.

Larsen, E., and Sejrup, H. P. (1990). Weichselian land–sea interactions: western Norway–Norwegian Sea. *Quaternary Science Review* 9, 85–97.

Larsen, E., Gulliksen, S., Lauritzen, S.-E., Lie, R., Løvlie, R., and Mangerud, J. (1987). Cave stratigraphy in western Norway: Multiple Weichselian glaciations and interstadiel vertebrate fauna. *Boreas* 16, 267–292.

Lauritzen, S.-E. (1991). Uranium Series Dating of Speleothems: A Glacial Chronology for Nordland, Norway, for the Last 600 ka. *Striae* 34, 127–133.

Levi, S., Audunsson, H., Duncan, R. A., Kristjansson, L., Gillot, P.-Y., and Jakobsson, S. P. (1990). Late Pleistocene geomagnetic excursion in Icelandic lavas: Confirmation of the Laschamp excursion. *Earth and Planetary Science Letters* 96, 443–457.

Liddicoat, J. C. (1992). Mono Lake Excursion in Mono Basin, California, and at Carson Sink and Pyramid Lake, Nevada. *Geophysical Journal International* 166, 442–452.

Løvlie, R. (1989). Palaeomagnetic excursions during the last interglacial/glacial cycle: A synthesis. *Quaternary International* 3(4), 5–11.

Løvlie, R., and Sandnes, A. (1987). Palaeomagnetic excursions recorded in mid-Weichselian cave sediments from Skjonghelleren, Valderøy, W. Norway. *Physics of the Earth and Planetary Interiors* 45, 337–348.

Lundqvist, J., and Miller, U. (1992). Weichselian stratigraphy and glaciations in the Tæsjø–Hoting area, central Sweden. *Sveriges Geologiska Undersökning* C 826, 4–35.

Mackiewicz, N. E., Powell, R. D., Carlson, P. R., and Molnia, B. F. (1984). Interlaminated ice-proximal glacimarine sediments in Muir Inlet, Alaska. *Marine Geology* 57, 113–147.

Mangerud, J. (1980). Ice-front variations of different parts of the Scandinavian Ice Sheet, 13,000–10,000 Years B.P. *In* "The Lateglacial of North-West Europe" (J. J. Lowe, J. M. Gray, and J. E. Robinson, Eds.), pp. 23–30. Pergamon, New York.

Mangerud, J. (1989). Correlation of the Eemian and the Weichselian with deep sea oxygen isotope stratigraphy. *Quaternary International* 3/4, 1–4.

Mangerud, J. (1991a). The Scandinavian Ice Sheet through the last interglacial/glacial cycle. *In* "Klimageschichtliche Probleme der letzten 130,000 Jahre" (Frenzel, B., Ed.), pp. 307–330, G. Fisher, Stuttgart, New York. (This paper was presented as a lecture in 1987, accepted for press March 1988 and widely distributed. It is therefore in the literature cited with different dates 1987–1991).

Mangerud, J. (1991b). The last interglacial/glacial cycle in northern Europe. *In* "Quaternary Landscapes" (L. Shane and E. Cushing, Eds.), pp. 38–75. Univ. Minnesota Press, Minneapolis.

Mangerud, J., Gulliksen, S., Larsen, E., Longva, O., Miller, G. H., Sejrup, H. P., and Sønstegaard, E. (1981). A Middle Weichselian ice-free period in western Norway: The Ålesund interstadial. *Boreas* 10, 447–462.

Martinsson, D. G., Nicklas, G. P., Hays, J. D., Imbrie, J., Moore, T. C., and Shackleton, N. J. (1987). Age dating and the orbital theory of the ice ages: development of a high-resolution 0 to 300,000 years chronostratigraphy. *Quaternary Research* 27(1), 1–29.

Mejdahl, V., Shlukov, A. I., Shakhovets, S. A., Voskovskaya, L. T., and Lyashenko, H. G. (1992). The effect of shallow traps: a possible source of error in TL dating of sediments. *Ancient TL* 10, 2225.

Miller, G. H., and Mangerud, J. (1986). Aminostratigraphy of European marine interglacial deposits. *Quaternary Science Review* 4, 215–278.

Miller, G. H., Sejrup, H. P., Mangerud, J., and Andersen, B. G. (1983). Amino acid ratios in Quaternary molluscs and foraminifera from western Norway: Correlation, geochronology and plaeotemperature estimates. *Boreas* 12, 107–124.

Mojski, J. E. (1992). Vistulian stratigraphy and TL dates in Poland. *Sveriges Geologiska Undersökning* C826, 195–200.

Nowaczyk, N. R., Frederichs, T. W., Eisenhauer, A., and Gard, G. (1993). Stratigraphy of late Quaternary Sediments from the Yermak Plateau, Arctic Ocean: Evidence for four Geomagnetic Polarity Events within the last 170 kys of the Brunhes Chron. *Geophysical Journal International* 46.

Nürnberg, D., Wollenburg, I., Dethleff, D., Eiken, H., Kassens, H., Letzig, T., Reimnitz, E., and Thiede, J. (1994). Sediments in Artic sea-ice: Implications for entrainment, transport and release. *Marine Geology* 119, 185–214.

Petersen, K. S. (1985). The Late Quaternary history of Denmark. *Journal of Dansk Archeology* 4, 7–22.

Petersen, K. S., and Kronborg, C. (1991). Late Pleistocene history of the inland glaciation in Denmark. *In* "Klimageschichtliche Probleme der letzten 130000 Jahre" (B. Frenzel, Ed.), pp. 331–342. G. Fisher, Stuttgart, New York.

Pfirman, S., Wollenburg, I., Thiede, J., and Lange, M. A. (1989). Lithogenic sediment on Arctic pack ice: Potential aelian flux and contribution to deep sea sediments. *In* "Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport" (M. Leinen and M. Sarnthein, Eds.), Nato ASI Series C 282, pp. 463–493.

Powell, R. D. (1984). Glacimarine processes and inductive lithofacies modelling of ice shelf and tidewater glacier sediments based on Quaternary examples. *Marine Geology* 57, 1–52.

Ramm, M. (1988). A stratigraphic study of Late Quaternary sediments on the Vøring Plateau, eastern Norwegian Sea. *Marine Geology* 83, 159–191.

Samtleben, C., and Schröder, A. (1992). Living coccolithophore communities in the Norwegian–Greenland Sea and their record in sediments. *Marine Micropaleontology* 19, 333–354.

Sarnthein, M., Jansen, E., Arnold, M. Duplessy, Erlenkeuser, H., Flatoy, A., Veum, T., Vogelsang, E., and Weinelt, M. S. (1992). $\delta^{18}O$ time-slice reconstruction of meltwater anomalies at Termination I in the North Atlantic between 50 and 80°N. *In* "The Last Deglaciation: Absolute and Radiocarbon Chronologies" (E. Bard and W. S. Broecker, Eds.), NATO ASI Series I, Global Environmental Change, Vol. 2, pp. 1–344. Springer-Verlag, Berlin, Heidelberg, New York.

Scholten, J. C., Botz, R., Mangini, A., Paetsch, H., Stoffers, P., and Vogelsang, E. (1990). High resolution $^{230}Th_{ex}$ stratigraphy of sediments from high-latitude areas (Norwegian Sea, Fram Strait). *Earth and Planetary Science Letters* 101, 54–62.

Sejrup, H. P. (1987). Molluscan and foraminiferal biostratigraphy of an Eemian–Early Weichselian section on Karmøy, southwestern Norway. *Boreas* 16, 27–42.

Sejrup, H. P., Haflidason, H., Aarseth, I., King, E., Forsberg, C. F., Long, D., and Rokoengen, K. (1994). Late Weichselian glaciation history of the northern North Sea. *Boreas* 23, 1–13.

Shackleton, N. J., Imbrie, J., and Hall, M. A. (1983). Oxygen and carbon isotope record of the East Pacific core V19-30: Implications for the Formation of deep water in the late Pleistocene North Atlantic. *Earth Planetary Science Letters* 65, 233–244.

Spielhagen, R. F. (1991). Die Eisdrift in der Framstraße während der letzten 200,000 Jahre. *GEOMAR Report* 4, 133.

Vogelsang, E. (1990). Paläo-Ozeanographie des Europäischen Nordmeeres an Hand stabiler Kohlenstoff- und Sauerstoffisotope. *Berichte Sonderforschungsbereich 313* 23, 136.

Vorren, T. O., Vorren, K.-D., Alm, T., Gulliksen, S., and Løvlie, R. (1988). The last deglaciation (20,000 to 11,000 B.P.) on Andøya, northern Norway. *Boreas* 17, 41–77.

Weinelt, M. S., Sarnthein, M., Vogelsang, E., and Erlenkeuser, H. (1991). Early decay of the Barents Shelf Ice Sheet—Spread of stable isotope signals across the eastern Norwegian Sea. *Norsk Geologisk Tidsskrift* 71, 137–140.

Weinelt, M. S. (1993). Veränderungen der Oberflächenzirkulation im Europäischen Nordmeer während der letzten 60.000 Jahre—Hinweise aus stabilen Isotopen. *Berichte Sonderforschungsbereich 313* 41, 106.

Wolf, T. C. W. (1991). Paläo-ozeanographisch-klimatische Entwicklung des nördlichen Nordatlantiks seit dem späten Neogen (ODP Legs 105 und 104, DSDP Leg 81). *GEOMAR Report* 5, 92.