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Late Quaternary dinoflagellate cyst stratigraphy at the Eurasian continental margin, Arctic Ocean: indications for Atlantic water inflow in the past 150,000 years

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Abstract

Four sediment cores located at the Eurasian continental margin underlying the Atlantic layer have been studied for their dinoflagellate cyst content. Concentrations of distinct dinoflagellate cyst taxa display fluctuations in the late Quaternary, which are linked to changes in the inflow of relatively warm Atlantic surface and near-surface waters, resulting in increased local production of cysts in certain time intervals. Based on the assumption that marked changes in strength of inflow occurred synchronously at the Eurasian continental margin, concentration maxima can be used to correlate sediment cores. A dinoflagellate cyst record from the northern Barents Sea continental margin has been related to the stable oxygen isotope and paleomagnetic records to provide direct chronological information. The combination of these methods permits definition of stratigraphic sections equivalent to oxygen isotope stages in carbonate-poor sequences from the Eurasian continental margin. Previous age models of sediment cores are revised, based on dinoflagellate cyst abundance peaks and species distribution, but a firm chronostratigraphy of sedimentary sequences at the eastern Laptev Sea continental margin cannot be established because of the weak signal at the sites furthest from Fram Strait.

In the past 150,000 years, the influence of Atlantic (sub-) surface waters generally decreased from west to east along the Eurasian continental margin, in particular during the glacials. Pronounced concentration maxima of cosmopolitan and temperate–subpolar dinoflagellate cysts indicate the inflow of Atlantic waters and seasonally increased production of cysts in the Holocene and Eemian. The Holocene is well-marked at the entire Eurasian continental margin but it is more difficult to assess the extent of (sub-) surface water inflow during the Eemian, which may have only reached the western Laptev Sea continental margin.

Keywords: Eastern Arctic Ocean; Late Quaternary; stratigraphy; stable oxygen isotopes; magnetostratigraphy; paleomagnetic excursions; dinoflagellate cysts

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1. Introduction

The late Quaternary stratigraphy of sedimentary sequences in the Eastern Arctic Ocean is frequently a matter of intense discussions because of a general lack of well-constrained stable oxygen isotope records. Local freshwater supply from melting continental ice sheets and river discharge may overprint the global signature of the planktic stable isotope signal in the high northern latitudes (e.g., Köhler and Spielhagen, 1990; Voelker et al., 1998; Dokken and Jansen, 1999; Knies et al., 1999; Poore et al., 1999; Spielhagen et al., submitted for publication). Additionally, the low biogenic carbonate production, which is reflected in a 4- to 8-fold decrease of planktic foraminifer abundance in the upper 500 m of the water column from the Fram Strait to the Laptev Sea continental margin (Volkmann, 2000a), results in low carbonate accumulation rates in sediments. Furthermore, dissolution strongly distorts this low productivity signal leading to significant gaps in the planktic and benthic carbonate records (e.g., Köhler and Spielhagen, 1990; Hebbeln and Wefer, 1997; Knies et al., 1999; Volkmann, 2000b; Wollenburg et al., 2001; Vogt et al., 2001-this volume).

In the Eastern Arctic Ocean, carbonate-bearing sequences, which permit establishment of a relatively continuous stable oxygen isotope stratigraphy over the last two glacial/interglacial cycles, are usually confined to the Fram Strait and the Yermak Plateau region (e.g., Markussen et al., 1985; Morris, 1988; Köhler and Spielhagen, 1990; Hebbeln and Wefer, 1997; Knies et al., 1999; Vogt et al., 2001-this volume). Along the Eurasian continental margin and in the adjacent deep sea, biogenic carbonate is rare and stable oxygen isotope stratigraphies have been established only for Weichselian to Holocene sequences (Stein et al., 1994; Nørgaard-Pedersen et al., 1998; Hald et al., 1999).

Various approaches have been therefore applied to establish a stratigraphic framework for late Quaternary Eastern Arctic Ocean sediments. Since the discovery of short-term paleomagnetic excursions in Weichselian sediments from the Yermak Plateau (Løvlie et al., 1986), extensive magnetostratigraphic investigations revealed the presence of up to six excursions in the past 130,000 years (Bleil and Gard, 1989; Nowaczyk, 1991; Nowaczyk and Baumann, 1992; Nowaczyk et al., 1994; Frederichs, 1995; Schneider et al., 1996; Nowaczyk and Knies, 2000; Knies et al., 2000). The magnetostratigraphy has been tied to a coccolith biostratigraphy (Bleil and Gard, 1989; Nowaczyk and Baumann, 1992), which was calibrated vs. stable oxygen isotope records in

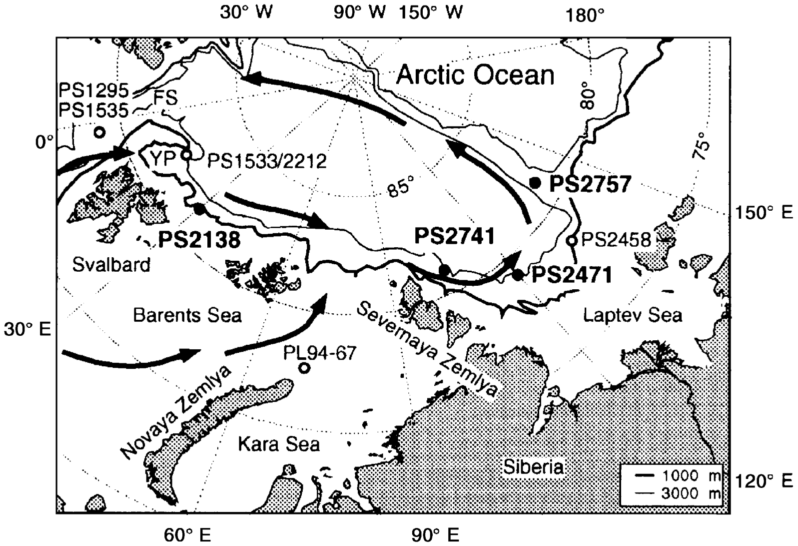

Fig. 1. Location of sediment cores at the Eurasian continental margin. The new cores used for dinocyst analysis are marked in bold. The arrows indicate schematically the path of the Atlantic layer in the Eastern Arctic Ocean (after Rudels et al., 1994). (YP: Yermak Plateau; FS: Fram Strait.)

Table 1
Sediment cores used in this study

Expedition	Year	Station	Gear	Latitude, N	Longitude, E	Water depth (m)	Length (cm)	Reference
ARK III/3	1985	PS1295-4	GKG	77°59.5′	02°25.2′	3112	42.5	Gersonde, 1986
ARK IV/3	1987	PS1533-3	SL	82°01.9′	15°10.7′	2030	485	Thiede, 1988
ARK IV/3	1987	PS1535-8	KAL	78°44.8′	01°52.8′	2557	496	Thiede, 1988
ARK VIII/2	1991	PS2138-1	SL	81°32.1′	30°25.6′	995	629	Rachor, 1992
ARK VIII/3	1991	PS2212-3	KAL	82°04.2′	15°43.0′	2550	770	Fütterer, 1992
ARK IX/4	1993	PS2458-4	KAL	78°10.0′	133°23.9′	983	983	Fütterer, 1994
ARK IX/4	1993	PS2471-4	SL	79°08.7′	119°47.6′	3047	416	Fütterer, 1994
ARK XI/1	1995	PS2741-1	KAL	81°06.0′	105°22.0′	2400	637	Rachor, 1997
ARK XI/1	1995	PS2757-8	KAL	81°09.6′	140°12.0′	1230	840	Rachor, 1997

GKG: Large Box Corer; SL: Gravity Corer; KAL: Kasten Corer.

the Nordic Seas based on species abundance and composition of assemblages (Gard, 1988; Gard and Backman, 1990). Although coccoliths provide the stratigraphy for numerous sediment cores in the Nordic Seas and the Eastern Arctic Ocean (e.g., Gard, 1988, 1993; Baumann, 1990; Nowaczyk and Baumann, 1992; Nürnberg et al., 1995), their occurrence is confined to periods with enhanced inflow of relatively warm surface waters in the last 500,000 years (Gard and Backman, 1990). Low carbonate contents limit the applicability of this coccolith stratigraphy in sediment cores from the eastern Eurasian continental margin (Müller, 1999; Knies et al., 2000). Recently, carbonate-poor sediments in the Eastern Arctic Ocean have been dated with U/Th and Be radioisotopes, offering a unique chronostratigraphic tool (Eisenhauer et al., 1990, 1994; Spielhagen et al., 1997; Aldahan et al., 1997; Strobl, 1998; Strobl et al., 1999). In principal, multidisciplinary approaches revealed the best stratigraphic results although the resolution is always less than that of sedimentary sequences containing a continuous stable oxygen isotope stratigraphy (cf. Knies et al., 2000).

Dinoflagellate cysts (= dinocysts), which have walls composed of organic compounds, are a potentially valuable stratigraphic tool in the Arctic Ocean because of a relatively large species number and a

Table 2
Sources of stratigraphic data available for cores listed in Table 1

Station	^{18}O, ^{13}C	AMS	Magnetic susceptibility	Magneto-stratigraphy	^{230}Th	^{10}Be	Coccoliths	Benthic foraminifers	Dinoflagellate cysts
PS1295-4	1	1	n.d.	n.d.	n.d.	n.d.	19	n.d.	19, 23
PS1533-3	2	2	8	8	9, 15	9, 15	8	n.d.	n.d.
PS1535-8	2, 3	2	8	8	n.d.	n.d.	8, 20	n.d.	n.d.
PS2138-1	4	4, this study	9	9	16	16	n.d.	22	this study
PS2212-3	5	n.d.	10	10	n.d.	n.d.	9	22	this study
PS2458-4	6	6	11	n.d.	n.d.	n.d.	n.d.	n.d.	this study
PS2471-4	n.d.	n.d.	11, 12	n.d.	17, 18	17, 18	21	n.d.	this study
PS2741-1	n.d.	7	7, 13	7	n.d.	n.d.	n.d.	n.d.	this study
PS2757-8	n.d.	n.d.	13, 14	14	17	17	n.d.	n.d.	this study

n.d.: No data.

Sources: (1) Jones and Keigwin, 1988; (2) Köhler, 1992; (3) Köhler and Spielhagen, 1990; (4) Knies and Stein, 1998; Knies et al., 1999; (5) Vogt et al., in press; (6) Spielhagen et al., submitted for publication; (7) Knies et al., 2000; (8) Nowaczyk and Baumann, 1992; (9) Nowaczyk and Knies, 2000 (10) Nowaczyk et al., 1994; (11) Fütterer, 1994; (12) Nürnberg et al., 1995; Stein et al., 1999 (13) Rachor, 1997 (14) Nowaczyk et al., unpublished; (15) Eisenhauer et al., 1994; (16) Schulz, 1997; (17) Strobl, 1998; (18) Strobl et al., 1999; (19) Baumann and Matthiessen, 1992; (20) Baumann, 1990; (21) Nürnberg et al., 1995; Andruleit, unpublished; (22) Wollenburg et al., 2001; (23) Matthiessen and Baumann, 1997.

good preservation potential (e.g., Mudie, 1992; Kunz-Pirrung, 1998; Rochon et al., 1999). Previously, they have been studied mostly in Pliocene and Pleistocene sediments from deep-water areas of the Arctic Ocean where they may occur in intervals, which are barren of calcareous and siliceous microfossils (Aksu and Mudie, 1985; Mudie, 1985; Matthiessen and Brenner, 1996). The main purpose of this paper is to report on new studies of the chronostratigraphic occurrence of dinocysts in late Quaternary sediments from the Eurasian continental margin. Sediment cores were selected for the longest stratigraphic sequences and for location under the core of the so-called Atlantic layer, which flows as a boundary current from the Yermak Plateau along the Eurasian continental margin into the eastern Arctic Ocean (e.g., Rudels et al., 1994; Schauer et al., 1997). We also compare the dinocyst ecostratigraphy of a core from the Yermak Plateau slope with the stable oxygen isotope stratigraphy, supported by AMS ^{14}C dates and magnetostratigraphy in order to calibrate abundance acmes with the marine isotope stages (MIS), so that the ecostratigraphy can be used for rough dating of carbonate-poor sequences. Furthermore, we use all the available stratigraphic data to revise the chronostratigraphic framework of cores from the Laptev Sea continental margin.

2. Material and methods

2.1. Material

The sediment cores used were recovered during several expeditions of RV "Polarstern" (Fig. 1; Table 1). Initial shipboard descriptions and detailed on-

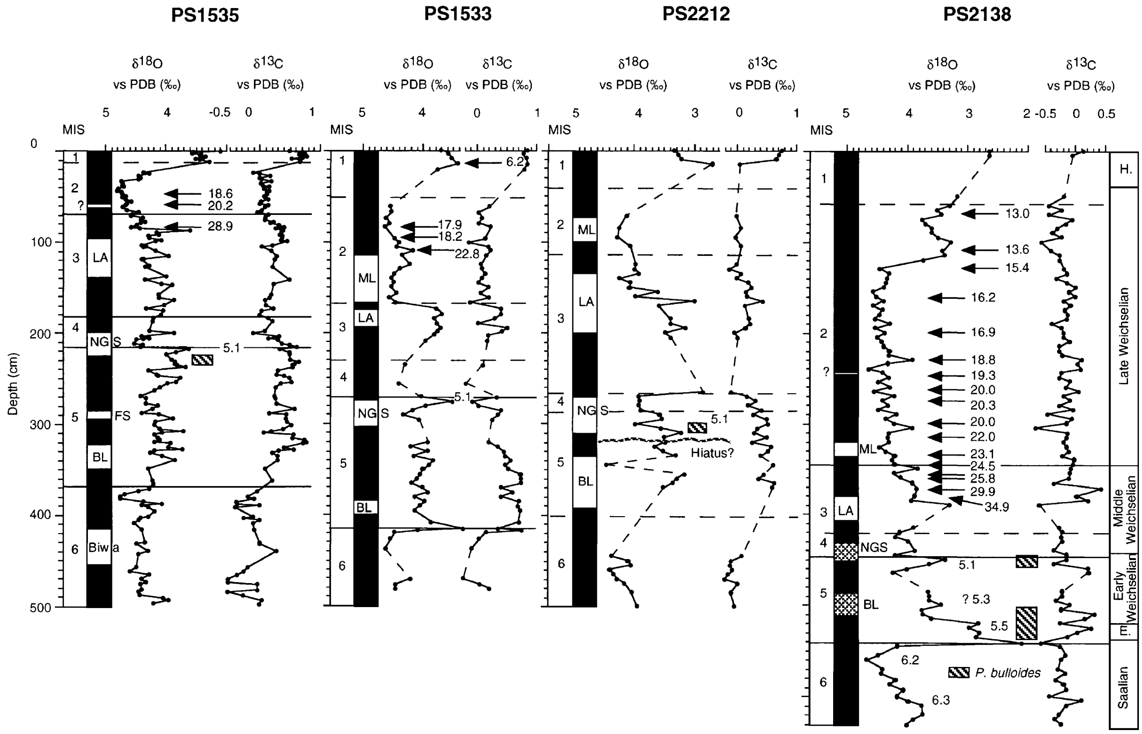

Fig. 2. Relationship between magnetostratigraphy and stable oxygen and carbon isotope stratigraphy in selected sediment cores from Fram Strait and Yermak Plateau. For sources of data, see Table 2. Only the late Saalian to Holocene section of core PS2212 is shown. The AMS ^{14}C ages are in ka years BP. The Norwegian–Greenland Sea and Blake excursions in core PS2138 are inferred from the correlation of magnetic susceptibility record to core PS2212 (Nowaczyk and Knies, 2000). The correlation of isotope stages to terrestrial chronology is from Mangerud (1989). (ML: Mono Lake; LA: Laschamp; NGS: Norwegian–Greenland Sea; FS: Fram Strait; BL: Blake; H: Holocene; E: Eemian.)

shore sedimentological studies including X-ray radiograph analyses showed that the sediment cores are relatively undisturbed, except for core PS2471 which contained numerous turbidites (see cruise reports, Table 1; Vogt et al., in press; Müller, 1999; Knies et al., 1999, 2000). The last deglaciation is

Table 3
AMS ^{14}C dates of sediment cores used in this study

Core number	Depth (cm)	^{14}C Age (BP)	Reservoir corrected (years BP)	Carbon source	Laboratory	Age used (ka BP)	Source
PS1533-3	15.5	6605 ± 80	6165	*N. pachyderma* sin.	ETH	6.2	1
PS1533-3	84.5	18,310 ± 180	17,870	*N. pachyderma* sin.	ETH	17.9	1
PS1533-3	92.5	18,600 ± 130	18,160	*N. pachyderma* sin.	ETH	18.2	1
PS1533-3	110.5	23,230 ± 200	22,790	*N. pachyderma* sin.	ETH	22.8	1
PS1535-8	55.5	19,050 ± 170	18,610	*N. pachyderma* sin.	ETH	18.6	1
PS1535-8	61.5	20,620 ± 190	20,180	*N. pachyderma* sin.	ETH	20.2	1
PS1535-8	83.5	29,340 ± 410	28,900	*N. pachyderma* sin.	ETH	28.9	1
PS2138-1	50	13,460 ± 110	13,020	bivalve shells	KIA1282	not used[a]	2
PS2138-1	65	13,430 ± 70	12,990	*N. pachyderma* sin.	KIA9872	13.0	3
PS2138-1	80	13,040 + 140/ − 130	12,600	bivalve shells	KIA363	not used[a]	2
PS2138-1	110	14,030 ± 80	13,590	mixed forams	KIA4765	13.6	2
PS2138-1	130	15,850 ± 130	15,410	mixed forams	KIA1283	15.4	2
PS2138-1	160	16,670 ± 210	16,230	*N. pachyderma* sin.	KIA364	16.2	2
PS2138-1	200	17,320 ± 130	16,880	*N. pachyderma* sin.	KIA2745	16.9	2
PS2138-1	230	19,230 ± 140	18,790	*N. pachyderma* sin.	KIA9873	18.8	3
PS2138-1	244	19,710 ± 130	19,270	*N. pachyderma* sin.	KIA9874	19.3	3
PS2138-1	260	20,420 ± 130	19,980	*N. pachyderma* sin.	KIA9875	20.0	3
PS2138-1	275	20,700 ± 130	20,260	*N. pachyderma* sin.	KIA9876	20.3	3
PS2138-1	300	20,480 + 330/ − 320	20,040	*N. pachyderma* sin.	KIA365	20.0	2
PS2138-1	316	22,470 ± 160	22,030	*N. pachyderma* sin.	KIA9877	22.0	3
PS2138-1	331	23,540 ± 240	23,100	*N. pachyderma* sin.	KIA2744	23.1	2
PS2138-1	345	24,910 + 200/ − 190	24,470	*N. pachyderma* sin.	KIA10372	24.5	3
PS2138-1	360	26,240 + 280/ − 270	25,800	bivalve shells	KIA4766	25.8	2
PS2138-1	366	26,260 ± 230	25,820	*N. pachyderma* sin.	KIA10373	25.8	3
PS2138-1	372	30,290 + 360/ − 350	29,850	*N. pachyderma* sin.	KIA10374	29.9	3
PS2138-1	375	> 45,190		*N. pachyderma* sin.	KIA10375	not used[b]	3
PS2138-1	380	35,340 + 1570/ − 1310	34,900	mixed forams	KIA1284	34.9	2
PS2138-1	385	30,240 + 380/ − 370		*N. pachyderma* sin.	KIA10376	not used[c]	3
PS2458-4	399	10,090 ± 65	9650	mixed bivalves	AAR3084	9.7	4
PS2458-4	436	10,050 ± 170	9619	mixed bivalves	AAR2418	9.6	4
PS2741-1	160	12,040 ± 70	11,600	mixed forams	KIA4764	11.6	5
PS2741-1	200	38,160 ± 3500	37,720	*N. pachyderma* sin.	KIA110	37.7	5
PS2757-7	0–1	1000 ± 30	560	planktic forams	KIA1469	0.6	6
PS2757-7	4–5	1610 ± 30	1170	planktic forams	KIA1470	1.2	6

The ^{14}C dates are δ^{13}C-normalized and corrected for reservoir effects equal to 440 years (cf. Knies et al., 1999). (*N. pachyderma* sin. = *Neogloboquadrina pachyderma* sin.).

Sources: (1) Köhler, 1992; (2) Knies and Stein, 1998; Knies et al., 1999; Nowaczyk and Knies, 2000; (3) Nowaczyk et al., unpublished; (4) Spielhagen et al., submitted for publication; (5) Knies et al., 2000; (6) Spielhagen, unpublished.

[a] The bivalve ages are not used because they may be reworked.

[b] Infinite age.

[c] Unreliable age because carbon content of sample was too low.

particularly well-dated in cores PS1295 and PS2458 (Fig. 1; Jones and Keigwin, 1988; Spielhagen et al., submitted for publication), which were used to evaluate the synchroneity of dinocyst events in the Holocene and late Weichselian. Additionally, core PS2212 was used to more precisely define the time-stratigraphic position of the Blake paleomagnetic excursion.

2.2. Stratigraphy

Various methods have been previously applied in order to establish a chronostratigraphy of the sediment cores (Table 2 and references therein). Sediment cores from Fram Strait and Yermak Plateau can be dated by stable oxygen isotope stratigraphy and AMS ^{14}C dating (Fig. 2; Table 3). In contrast, there is insufficient carbonate in other cores for a meaningful stable oxygen isotope stratigraphy. Also, note that *Neogloboquadrina pachyderma* sin. was picked from the size fraction >63 μm from core PS2138 instead of 125–250 μm as in the other cores (Knies et al., 1999). The ages of the stage boundaries are taken from Martinson et al. (1987).

The AMS ^{14}C dates must be used with care because of the unknown reservoir effect in the Eurasian shelf seas and adjacent Arctic Ocean. Usually, a reservoir correction of 400–500 years is used both for foraminifers and bivalves (e.g., Knies et al., 1999; Bauch et al., 1999b). Recent studies of the

Table 4
Dinocyst taxa in the sediment cores

	Ecology	Taxa used for stratigraphy	“Atlantic” water taxa	Plankton
Algidasphaeridium ? *minutum* var. *minutum*	polar–subpolar			
Algidasphaeridium ? *minutum* var. *cezare*	polar			
Ataxiodinium choane	temperate		×	×
Bitectatodinium tepikiense	temperate	*B. tepikiense*	×	×
Brigantedinium spp.	cosmopolitan[a]			
Brigantedinium simplex	polar–temperate			×
Impagidinium spp.[b]	temperate–tropical	*Impagidinium* spp.[b]	×	
Impagidinium pallidum	polar–subpolar	*I. pallidum*		
Impagidinium patulum	temperate	*Impagidinium* spp.	×	
Impagidinium sphaericum	subpolar–temperate	*Impagidinium* spp.	×	
Lingulodinium machaerophorum	temperate–tropical		×	×
Nematosphaeropsis labyrinthus	cosmopolitan	*N. labyrinthus*	×	×
Operculodinium centrocarpum sensu Wall and Dale	cosmopolitan	*O. centrocarpum*	×	×
Cyst of *Pentapharsodinium dalei*	subpolar–temperate	*P. dalei*	×	
Cyst of *Polykrikos schwartzii*	temperate			
Cyst of *Polykrikos* sp. sensu Kunz-Pirrung, 1998	polar			
Spiniferites spp.	cosmopolitan	*Spiniferites* spp.	×	
Spiniferites elongatus	subpolar–temperate	*S. elongatus* s.l.	×	×
Spiniferites frigidus	subpolar	*S. elongatus* s.l.	×	
Spiniferites cf. *elongatus* sensu Harland and Sharp	subpolar–temperate	*S. elongatus* s.l.	×	
Spiniferites lazus	temperate	*Spiniferites* spp.	×	
Spiniferites membranaceus	temperate	*Spiniferites* spp.	×	×
Spiniferites hyperacanthus	temperate	*Spiniferites* spp.	×	×
Spiniferites mirabilis	temperate	*Spiniferites* spp.	×	×
Spiniferites ramosus	cosmopolitan	*Spiniferites* spp.	×	×
Selenopemphix quanta	temperate–tropical			×
Species et genus indet.				

The biogeography of taxa is mainly from Harland (1983) and Rochon et al. (1999). The occurrence of the respective vegetative stages in the plankton of Eurasian shelf seas and adjacent Arctic Ocean is noted (from Okolodkov, 1998).

[a] In general, species of *Brigantedinium* spp. are cosmopolitan, but if characterized by *B. cariacoense* and *B. simplex*, this category indicates boreal to polar conditions.

[b] All species of *Impagidinum* without *I. pallidum*.

radiocarbon content of pre-bomb molluscs showed that a variable reservoir correction must be applied depending on the species used for dating (Forman and Polyak, 1997). Therefore, we used AMS ^{14}C dates only to support age models.

The maximum of the benthic foraminifer *Pullenia bulloides*, which is a characteristic species on the upper continental slope (500–1000-m water depth) in the Yermak Plateau area (Wollenburg and Mackensen, 1998), is a biostratigraphic marker in the Nordic Seas and Eastern Arctic Ocean for MIS 5.1, but it has peak abundances in cores from water depths shallower than 1500 m in MIS 5.1 and MIS 5.3–5.5 (e.g., Haake and Pflaumann, 1989; Struck, 1992; Wollenburg et al., 2001).

The magnetostratigraphy was correlated to stable oxygen isotope records in cores from Yermak Plateau to provide independent age control for carbonate-poor sequences from the eastern Arctic Ocean (Fig. 2). It must be noted that variable sampling intervals and temporarily and spatially variable sedimentation rates may influence the resolution of paleomagnetic excursions. The stratigraphy of the sediment cores is expressed in terms of stable oxygen isotope stages and of the standard Quaternary chronostratigraphy (cf. Mangerud, 1989).

2.3. Dinocyst analysis

The new sediment cores were sampled for dinocysts at 3- to 20-cm intervals, with an average of 10 cm. Additionally, the data of core PS1295-4 were taken from Matthiessen and Baumann (1997). Freeze-dried samples were prepared with standard palynological preparation methods (e.g., Matthiessen, 1995; Rochon et al., 1999). The processing method uses cold HCl and HF to remove carbonates and silicates, respectively. After each acid treatment, residues were washed through a 6-μm mesh to remove the bulk of the clay and fine silt fractions. The final residue was mounted with glycerine jelly on microscopic slides, covered with a cover slip, and sealed with paraffin wax.

At least one microscopic slide from each sample has been scanned and dinocysts were counted. All taxa found and their ecological affinity are listed in alphabetic order in Table 4 (for details of taxonomy, see Rochon et al., 1999). Concentrations were calculated according to the marker grain method of Stockmarr (1971). A known number of acetylosed *Lycopodium clavatum* spores were added to weighted dry sediment. Spore tablets of the batch no. 124961 (distributed by Laboratory of Quaternary Biology at the Department of Quaternary Geology in Lund, Sweden) were used with an average number of spores of 12,542 ± 2081.

3. Stratigraphic framework of sediment cores

The sediment cores PS2138, PS1533, PS1535 and PS2212 from the Yermak Plateau and Fram Strait region provide the chronostratigraphic reference sections to date sediments in the Eastern Arctic Ocean (Fig. 2). The interpreted paleomagnetic excursions are consistent time markers with respect to stable isotope stratigraphy, and their time-stratigraphic position is in general agreement with a compilation of age ranges on a global scale (Bleil and von Dobeneck, 1999). The core PS2138 is taken as primary reference section because it has a relatively continuous stable oxygen and carbon isotope record in contrast to other cores, supported by 21 AMS ^{14}C, which can be linked to the standard oxygen isotope stages (Knies et al., 1999; Fig. 2; Table 3). Note that the base of MIS 1 in core PS2138 has been adjusted because of new AMS ^{14}C dates. In particular, the last interglacial (MIS 5.5) in PS2138 is clearly reflected by a plateau of light oxygen isotope values after a distinct meltwater event in Termination II. The magnetostratigraphy of core PS2138 is not as clear as that in other cores from the Yermak Plateau. Only the Mono Lake and Laschamp excursions could be unequivocally identified, whereas the Blake and Norwegian–Greenland Sea excursions had to be transferred by correlating the magnetic susceptibility records of PS2138 and PS2212. Nowaczyk and Knies (2000) assume that early diagenetic processes erased these short-term excursions in the records.

The stratigraphic position of the Mono Lake (ML) excursion can be easily assessed because AMS ^{14}C dates bracket the excursion in core PS2138 (Fig. 2). It is associated with the interval of heaviest stable oxygen isotope values in the last glacial/interglacial cycle, suggesting that it occurred in MIS 2. However, the few AMS ^{14}C dates and the continuously decreasing stable oxygen isotope values make the accurate definition of the stage 2/3 boundary difficult. Thus, it was previously placed at the termination of the Mono Lake excursion in cores PS1533 and PS2212 (Nowaczyk et al., 1994). We tentatively place the base of MIS 2 in all cores before the Mono Lake excursion marking the late/middle Weichselian boundary.

A very short excursion has been tentatively identified in core PS2138 after the Mono Lake excursion (Nowaczyk and Knies, 2000). This may correspond to an excursion in core PS1535 between ca. 18,600 and 20,200 years BP, which was previously assigned to the Mono Lake excursion (Fig. 2; Nowaczyk and Baumann, 1992).

The Laschamp (LA) excursion occurs in an interval with variable, but relatively light stable oxygen isotope values indicating an MIS 3 age (Fig. 2). The termination has an age of ca. 34,900 years BP in core PS2138 and is older than ca. 29,000 years BP in core PS1535 (Fig. 2). Both the Mono Lake and Laschamp excursions have a relatively short duration of ca. 1500 to 2000 years (Voelker et al., 1998; Kissel et al., 1999; Laj et al., 2000; Nowaczyk and Knies, 2000). The variable thickness in the cores from Yermak Plateau and Fram Strait suggests variable sedimentation rates from a few centimeters to more than 10 cm, as already indicated by AMS ^{14}C dates and stable oxygen isotope stratigraphy (e.g., Knies et al., 2000; Nowaczyk and Knies, 2000).

The Norwegian–Greenland Sea (NGS) excursion is associated with a pronounced shift to heavy stable oxygen and carbon isotope values at the MIS 4/5 stage boundary (Fig. 2). The light stable oxygen isotope peak of MIS 5.1 and the occurrence of *Pullenia bulloides* mark the older part of the excursion (Köhler and Spielhagen, 1990; Wollenburg et al., 2001). It is confined to MIS 5 in core PS1533, probably because of a short hiatus spanning the younger part of the excursion (Nowaczyk and Baumann, 1992). The Norwegian–Greenland Sea excursion covers approximately the interval from the termination of MIS 4 to MIS 5.1/5.2.

The Fram Strait (FS) excursion is a characteristic feature of mid-MIS 5 only in two parallel cores from site PS1535. Low sedimentation rates were discussed as possible cause of its absence in core PS1533 in the corresponding time interval by Nowaczyk and Baumann (1992). For core PS2212, between the Norwegian–Greenland Sea and Blake excursions, a hiatus was inferred from correlation of magnetic susceptibility records and paleomagnetic data with core PS1533 (Nowaczyk et al., 1994). The age of the possible hiatus can be calculated from the length of the corresponding stratigraphic interval in core PS2138 that is probably missing in core PS2212, based on the correlation of magnetic susceptibility records (Nowaczyk and Knies, 2000). The hiatus must be located between MIS 5.1/5.2 and possibly MIS 5.2/5.3 (ca. 85,000–95,000 years BP instead of ca. 85,000–115,000 years BP in Nowaczyk et al., 1994). Thus, low sedimentation rates could also explain the relatively short interval of normal polarity between the excursions.

The Blake excursion (BL) occurred during early MIS 5, close to the MIS 5/6 stage boundary (Fig. 2). This boundary has been previously placed inconsistently in cores PS1533 and PS2138 (cf. Nowaczyk et al., 1994; Knies et al., 2000) and is now set at the distinct meltwater peak in Termination II (Fig. 2).

4. Dinocyst ecostratigraphy in the Eastern Arctic Ocean

4.1. Distribution of dinocysts in relation to the environment

The number of dinocyst taxa (> 27 species) is moderately high in the sediment cores from the Eurasian continental margin (Table 4). The absence of first and last occurrence datums means that a phylogenetic biostratigraphic zonation cannot be established, and a dinocyst ecostratigraphy must be supplemented by independent dating methods in order to establish age models. The occurrence of taxa and the composition of assemblages in our cores are therefore correlated to the available stable oxygen isotope stratigraphy and magnetostratigraphy. A similar approach has been already used by Gard and Backman (1990), who defined a coccolith biozonation for the Nordic Seas and Eastern Arctic Ocean. We do not define a new zonation based on dinocysts because they are only abundant in relatively short stratigraphic intervals, but we delineate a distinctive succession of abundance acmes for the eastern Arctic region.

The basis of this study is derived from the known influence of ecological conditions, particularly temperature, on the modern abundance and distribution of dinocysts (e.g., de Vernal et al., 1997; Rochon et al., 1999). Dinocysts are the resting stages of planktic dinoflagellates that are common in the plankton of the Arctic Ocean (Okolodkov and Dodge, 1996). The vegetative stages of some dinocysts recorded in the sediment cores have been observed in the plankton of the Eurasian shelf seas (Okolodkov, 1998; Table 4), suggesting that these species belong to the local plankton.

The taxa found in our cores can roughly be divided into two ecological groups according to their environmental preferences. A small number of taxa are confined to the polar to subpolar regions (Table 4). Most taxa, in particular the cosmopolitan species *Operculodinium centrocarpum* and *Nematosphaeropsis labyrinthus,* are more abundant in temperate environments of the northern latitudes and may be common in polar environments that are influenced by Atlantic waters (Harland, 1983; Kunz-Pirrung, 1998; Rochon et al., 1999; Table 4). The cold-water taxa reflect the local environmental conditions, which may have changed independently between the different sites because of variable ecological conditions such as light, nutrients and food supply. Therefore, changes in the abundance patterns of these taxa may have occurred asynchronously at different locations. In contrast, most other taxa are of primary ecostratigraphic significance because they may only migrate with the Atlantic waters from the Norwegian Sea into the eastern Arctic Ocean (e.g., Okolodkov and Dodge, 1996), leading to the establishment of an autochthonous population in the shelf seas and adjacent Arctic Ocean during periods of optimum conditions. This does not imply that concentrations of dinocysts are necessarily a function of the intensity of influx because suitable local conditions are also required to cause an enhanced production of dinocysts.

The number of taxa was further reduced to get taxonomically relatively stable taxa (Table 4) and to reduce the effect of variable preservation (e.g., Dale, 1976; Zonneveld et al., 1997). Complete remineralization of primarily produced organic carbon is a common feature in sediments at the Laptev Sea continental margin (Boetius and Damm, 1998). Autofluorescence of taxa is a first indicator of potential preservability character because species that have a low preservation potential such as *Brigantedinium*, *Algidasphaeridium*, *Polykrikos* and *Selenopemphix* do not fluoresce (Table 4; e.g., Dale, 1976; Matthiessen, 1995; Zonneveld et al., 1997).

The concentrations of total Atlantic water taxa and selected taxa marked in Table 4 were calculated. A number of species which are occasionally present in some samples are not shown. The concentrations of the cold-water species *Impagidinium pallidum* are figured in this study because of a characteristic stratigraphic distribution pattern. Today, the concentrations of total Atlantic water taxa are strongly decreasing from more than 10,000 to 1000 cysts/g of dry sediment in the Fram Strait in the west, to less than 1000 at the Laptev Sea continental margin in the east (Mudie, 1992; Matthiessen, 1995; Kunz-Pirrung, 1998), probably reflecting the diminishing influence of warmer sea-surface conditions and warmer water inflow. Concentrations are both an indicator of optimum living conditions of dinoflagellates and sedimentary conditions. Thus, cyst concentration is depending upon sediment accumulation rates and grain size distribution (e.g., Dale, 1976) and we interpret change only when exceeding an order of magnitude.

The presence and abundance of the selected Atlantic water dinocysts at an Arctic Ocean location can be related either to local production or transport with currents. Transport of cysts with occasional inflow of relatively warmer Atlantic waters from the Norwegian Sea is more likely when concentrations are extremely low. High concentrations (more than 1000) may be caused by a local production of dinocysts when ecological conditions were favourable. During ecologically optimum conditions when surface waters were warmer and inflow was stronger, the sea–ice margin may have retreated in summer from the shelf seas to the continental margin. These extended open water conditions associated with elevated nutrient concentrations may have then triggered an increase of abundances of these Atlantic water dinoflagellates. We assume that production might be related to a few years of optimum conditions and it does not imply permanently open water conditions.

The transit time of Atlantic waters from the Yermak Plateau to the Laptev Sea continental margin is less than a decade (Frank et al., 1998). Since bathymetric obstacles are absent on the continental slope, major changes in Atlantic water intrusions must have occurred simultaneously at all sites at the Eurasian continental margin on a geological time-scale. Therefore, significant changes in abundance pattern of individual species as well as peaks in concentrations of Atlantic water dinocysts are quasi-synchronous, and have the potential for use in rough (millenial-scale) correlation of sediment cores at the Eurasian continental margin.

4.2. Dinocysts in sediment cores from the Eurasian Continental margin

The concentrations of the Atlantic water dinocyst taxa display distinct fluctuations in all sediment cores showing peak concentrations in few stratigraphic levels (Figs. 3–7 and 9). Most samples are characterized by concentrations less than 100 cysts/g dry sediment.

The taxa display distinct recurrent distribution patterns in the cores. *Operculodinium centrocarpum, Nematosphaeropsis labyrinthus, Spiniferites elongatus* s.l. and other species of *Spiniferites* occur throughout the cores. Fluctuations in concentrations are mainly caused by *O. centrocarpum*, although *N. labyrinthus* and *S. elongatus* s.l. are more abundant than *O. centrocarpum* in some intervals, in particular, when concentrations are relatively low.

Bitectatodinium tepikiense is continuously present only below the mid-Holocene and older parts of the cores. *Pentapharsodinium dalei* and *Impagidinium* spp. are apparently confined to restricted stratigraphic intervals. The warm-temperate to subpolar species of *Impagidinium* have a rather narrow range, being present mainly in the concentration maxima, while *P. dalei* is regularly present in MIS 1 and 5. Concentrations of some taxa show distinct gradients along the Eurasian continental margin. *Operculodinium centrocarpum* and *P. dalei* are present with variable concentrations in all cores, while the other taxa generally decrease in concentrations from core PS2138 to PS2757. Thus, *Nematosphaeropsis labyrinthus, B. tepikiense*, *I. pallidum*, *Spiniferites elongatus* s.l. and other species of *Spiniferites* occur with low concentrations in core PS2757; *Impagidinium* spp. even being absent from the core.

5. Calibration of the dinocyst record PS2138 vs. stable oxygen isotope and magnetostratigraphy

Sediment core PS2138 has been selected as reference section to calibrate the dinocyst record vs. stable oxygen isotope stratigraphy and magnetostratigraphy (Fig. 3). The interval covering the Blake excursion has been additionally analysed in core PS2212 because the stratigraphic position is not well-constrained with respect to the last interglacial and the stage 5/6 boundary (Fig. 4).

Distinct peaks of the dinocyst concentrations ($>$ 1000 cysts/g) do not correspond with isotope stage boundaries but they are confined to intervals with light stable oxygen isotope values in MIS 1 and 5 (Fig. 3). Concentrations in MIS 2–4 and MIS 6 are at least one order of magnitude lower or samples are virtually barren. Small peaks of concentrations occur only at the isotope stage boundary 3/4, in MIS 5.1 and 6.3. Concentrations in the Mono Lake and Laschamp excursions are generally low.

The peak concentrations are apparently associated with the warmest climatic conditions in the past 150,000 years (Fig. 3). When comparing the abundance patterns of core PS2138 with that of the well-dated core PS1295 from Fram Strait, the base of the Holocene, which is marked by a significant increase of *Operculodinium centrocarpum* and *Impagidinium pallidum*, may be tentatively placed at 30-cm core depth (Figs. 3 and 5). The late glacial to early Holocene in both cores is characterized by a concentration maximum of *Nematosphaeropsis labyrinthus* and *Spiniferites elongatus* s.l.

The Eemian is equally characterized by high concentrations and a dominance of *Operculodinium centrocarpum* and may be set at ca. 510- to 535-cm core depth (Fig. 3). This is in good agreement with a maximum of benthic foraminifers from 512- to 544-cm core depth (*Epistominella pusilla*, *Discorbinella berthelothi*, *Pullenia* spp.), indicating relatively warm Atlantic water inflow (Wollenburg et al., 2001). The Eemian differs from the Holocene in having low abundances of *Impagidinium pallidum* and significant amounts of *Bitectatodinium tepikiense*.

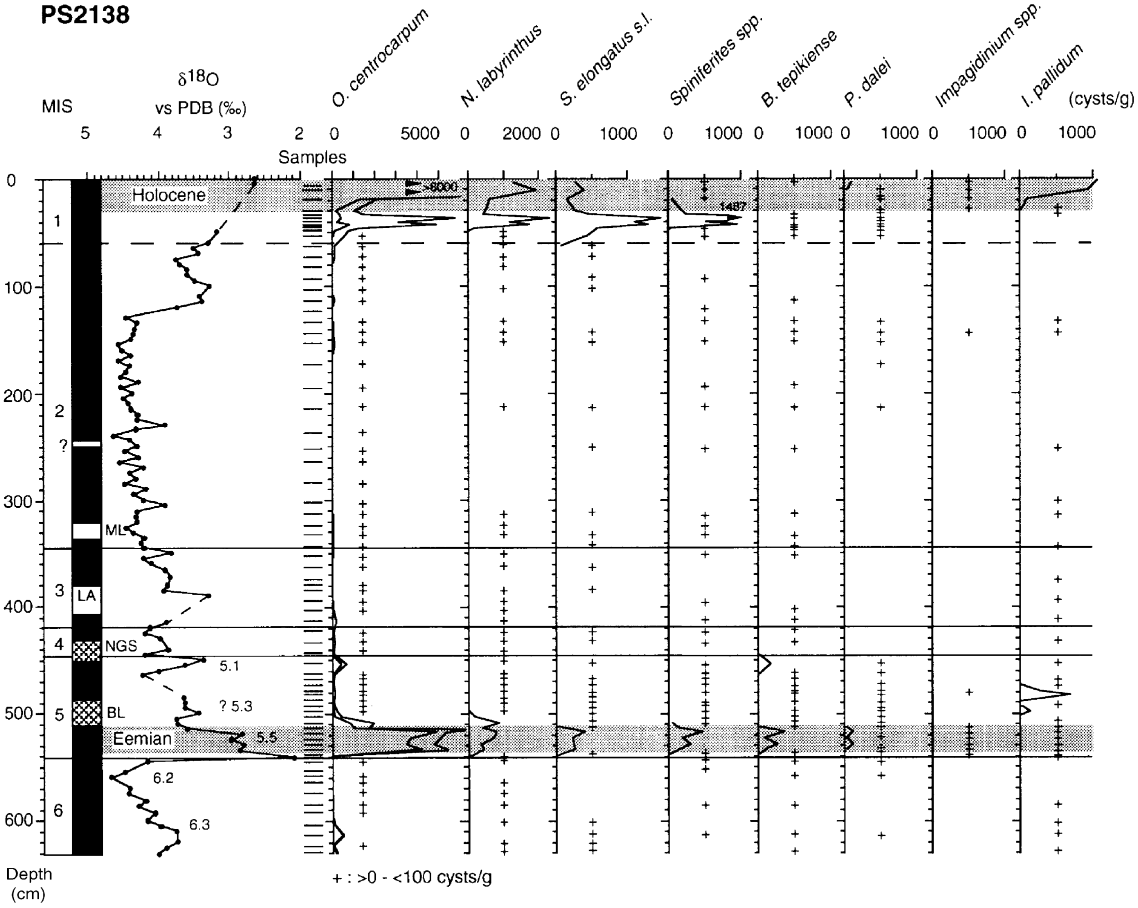

Fig. 3. Chronostratigraphy and dinocyst record of sediment core PS2138. The concentrations of total Atlantic water dinocysts (right line) and of *Operculodinium centrocarpum* (left line) are shown in one column. The reference age model is from Knies et al. (2000) and Nowaczyk and Knies (2000).

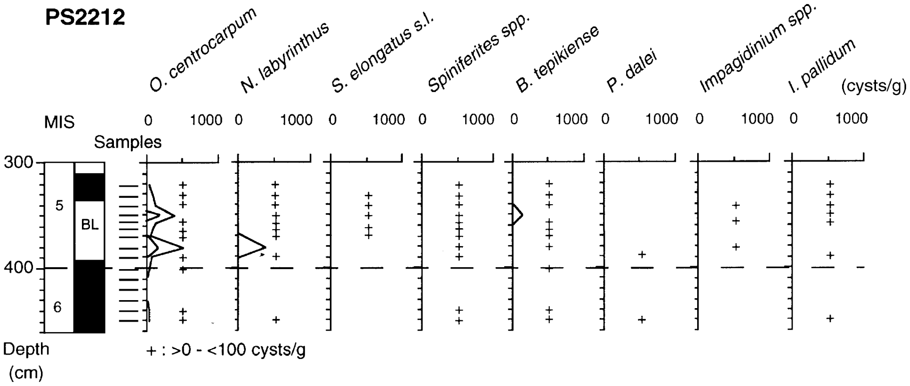

Fig. 4. Dinocyst record across the Blake excursion in sediment core PS2212. The concentrations of total Atlantic water dinocysts (right line) and of *Operculodinium centrocarpum* (left line) are shown in one column. The reference age model is from Nowaczyk et al. (1994) and Vogt et al. (in press).

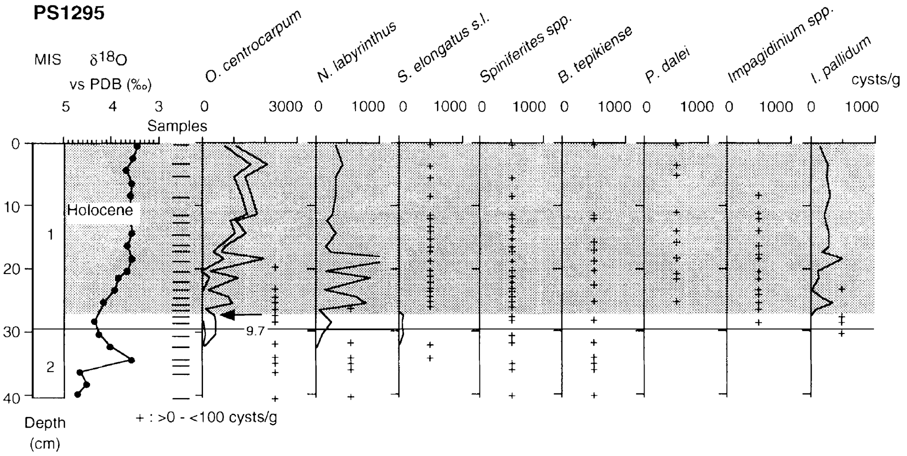

Fig. 5. Dinocyst record of sediment core PS1295. The concentrations of total Atlantic water dinocysts (right line) and of *Operculodinium centrocarpum* (left line) are shown in one column. The AMS ^{14}C ages are in ka BP. The age model is from Jones and Keigwin (1988).

The comparison of the dinocyst records of cores PS2212 and PS2138 confirms the proposed transfer of the Blake excursion to core PS2138 by means of the magnetic susceptibility records (Nowaczyk and Knies, 2000). The interval, which comprises the Blake excursion, is characterized in both cores by low concentrations and similar abundances of *Operculodinium centrocarpum* and *Nematosphaeropsis labyrinthus* together with some *B. tepikiense*, indicative of MIS 5.4–5.1 (compare Figs. 3 and 4). Therefore, the inferred Blake excursion is in core PS2138 younger than the Eemian (Fig. 2; cf. Nowaczyk and Knies, 2000). This is in agreement with previous studies from the Mediterranean Sea by Tucholka et al. (1987), who placed the Blake excursion at the MIS 5.5/5.4 boundary. In contrast, the abundance peak of the coccolith *Calcidiscus leptoporus* at ca. 351- to 360-cm core depth in core PS2212, which approximately corresponds with the Eemian (Gard, 1988; Nowaczyk et al., 1994), suggests an MIS 5.5 age for the Blake excursion but correlates with dinocyst assemblages younger than MIS 5.5 (Fig. 4). The variable coccolith records with a general absence in early MIS 5 in the cores from Fram Strait and Yermak Plateau may be explained with dissolution, indicated by low carbonate contents at Termination II (Hebbeln and Wefer, 1997; Bauch et al., 1999a; Vogt et al., in press).

Operculodinium centrocarpum dominates the assemblages in the other small concentration maxima (MIS 3/4 boundary, 5.1, 6.3) as well. *Bitectatodinium tepikiense* shows peak abundances in MIS 5.1 and 5.5. The occurrence of *Pentapharsodinium dalei* and all species of *Impagidinium* is mainly associated with MIS 1 and 5. *Impagidinium* spp. are apparently confined to the interglacials, while *I. pallidum* is most abundant in MIS 1 and 5 (Fig. 3).

6. Chronostratigraphy of sediment cores at the Laptev Sea continental margin

In this section, we discuss the dinocyst stratigraphy of sediment cores at the Laptev Sea continental margin, and refine previous age models based on all available stratigraphic evidence. The chronostratigraphy of sediment core PS2138 forms the basis for the stratigraphic assignments in the sediment cores PS2741, PS2471 and PS2757.

6.1. Core PS2741

The age model is based on magnetostratigraphy, which is supported by two AMS ^{14}C dates, suggesting an age equivalent to MIS 6 for the base of the core, and a pronounced IRD layer, which probably marks the mid-Weichselian deglaciation in the eastern Arctic Ocean associated with MIS 3.31 (Fig. 6; Knies et al., 2000). This age model is slightly revised because of new evidence from dinocysts (Fig.

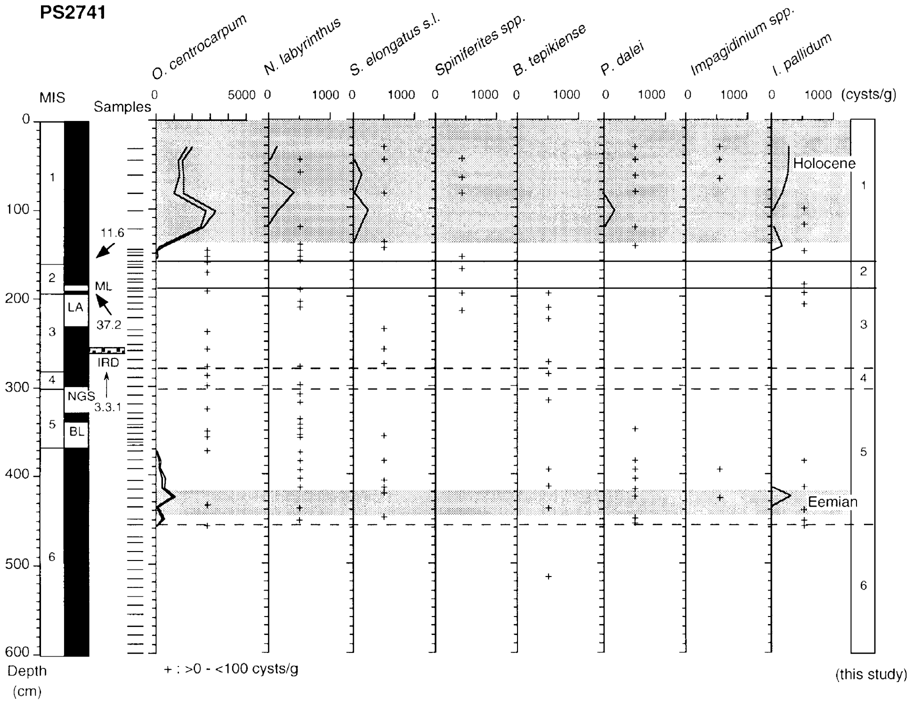

Fig. 6. Age model and dinocyst record of sediment core PS2741. The concentrations of total Atlantic water dinocysts (right line) and of *Operculodinium centrocarpum* (left line) are shown in one column. The AMS ^{14}C ages are in ka BP. The reference age model is from Knies et al. (2000).

6). The base of the Holocene (ca. 135-cm core depth) is marked by a significant increase of total dinocyst concentrations. The Holocene is characterized by high concentrations of *Operculodinium centrocarpum* and *Impagidinium pallidum* (compare with core PS2138). Dinocyst concentrations are low in MIS 2 to mid-MIS 5, in particular during the paleomagnetic excursions, and do not provide additional stratigraphic information. The assemblages in the concentration maximum below the Blake excursion at ca. 420- to 440-cm core depth are comparable to assemblages from the Eemian of core PS2138, although concentrations of most taxa are lower (Fig. 3). Therefore, we now place the MIS5/6 stage boundary, in contrast to Knies et al. (2000), below the Blake excursion at the base of the concentration maximum of dinocysts at ca. 455-cm core depth.

6.2. Core PS2471

Various age models have been proposed for core PS2471 based on coccoliths, ^{230}Th$_{ex}$ and ^{10}Be radioisotopes and sedimentological data suggesting ages for the base ranging from MIS 3–6 (Fig. 7; Nürnberg et al., 1995; Strobl, 1998; Müller, 1999; Stein et al., 1999; Strobl et al., 1999). Stratigraphic assignments are further complicated by intervals containing distal turbidite deposits (Fütterer, 1994; Nürnberg et al., 1995; Müller, 1999). For stratigraphic purposes, only the data from the undisturbed intervals are considered.

Nürnberg et al. (1995) suggested that, based on the occurrence of the coccolith *Gephyrocapsa* sp., the upper 50 cm correspond to MIS 1 and ca. 250-cm core depth to MIS 5. This is supported by the ^{10}Be stratigraphy, which suggests an MIS 1 age for the uppermost 20 cm and an MIS 5 age for the interval from 225 to 260 cm (Strobl et al., 1999; Fig. 7). However, Stein et al. (1999) questioned the MIS 5 age because Baumann (1990) reported *Gephyrocapsa* sp. also from MIS 3 sediments in the Nansen Basin. Andruleit (1998, personal communication) studied coccoliths from selected intervals but could

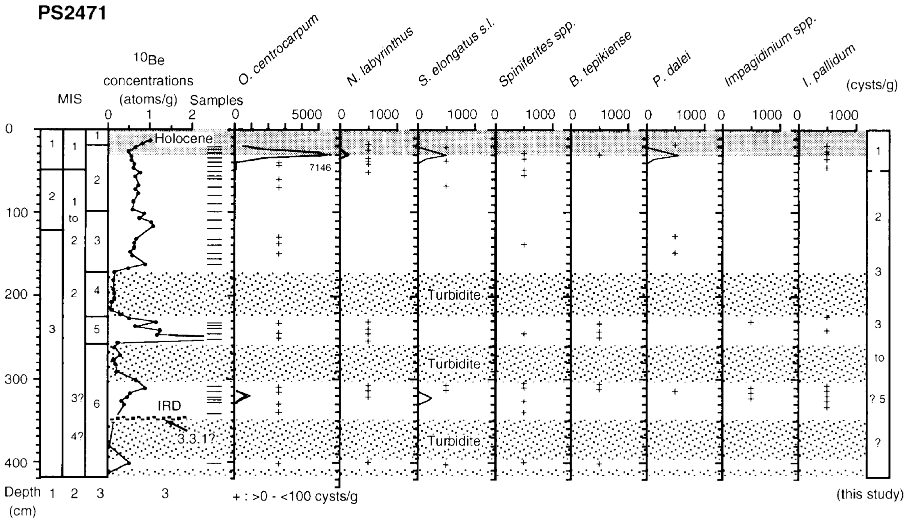

Fig. 7. Age models, ^{10}Be record and dinocyst record of sediment core PS2471. The concentrations of total Atlantic water dinocysts and of *Operculodinium centrocarpum* (left line) are shown in one column. The age models are from: (1) Müller (1999); (2) Stein et al. (1999); (3) Strobl et al. (1999). The ^{10}Be data are from Strobl et al. (1999).

not confirm the coccolith stratigraphy of Nürnberg et al. (1995).

Based on sedimentological and geochemical grounds, Stein et al. (1999) confirmed the age assignment for the upper 50 cm of core PS2471 by Nürnberg et al. (1995), but placed the upper turbidite tentatively into MIS 2 or 4, and the lower two into MIS 4. They argued that the turbidites might have been triggered during low sea level stands in the last glacial. Müller (1999) arrived at an MIS 3 age for all turbidites because she assumed that peaks of the clay mineral smectite in the intervals between the turbidites can only be supplied during periods of relatively high sea level from the source area, Permian flood basalts, in the Putoran Mountains in the hinterland of the western Laptev Sea.

The ^{230}Th$_{ex}$ and ^{10}Be radioisotope stratigraphy indicating an MIS 6 age for the base of the core must be questioned because variable cosmogenic production of ^{10}Be and changes in sediment supply and granulometric composition influence the accumulation of ^{10}Be, making the interpretation in terms of isotope stages difficult (e.g., Eisenhauer et al., 1994; Aldahan et al., 1997; Strobl et al., 1999). Distinct flux peaks of ^{10}Be in the GRIP ice core and marine sediments coincide with the last glacial and are caused by low geomagnetic field intensities during paleomagnetic excursions, such as the Mono Lake and Laschamp (e.g., McHargue et al., 1995; Aldahan et al., 1997; Frank et al., 1997; Yiou et al., 1997; Strobl, 1998). Thus, broad ^{10}Be maxima in cores PS1533, PS2138 and PS2471 in MIS 3 have been attributed to the "Raisbeck Peak" correlating with the Laschamp excursion (Fig. 8; Strobl, 1998). In core PS2471, the low ^{230}Th$_{ex}$ and ^{10}Be concentrations perfectly match the intervals with turbidites (Figs. 7 and 8). These contain high amounts of reworked (probably Mesozoic) palynomorphs (Matthiessen, unpublished data), suggesting that dilution with these old sediments reduced fluxes. If we exclude these turbidite intervals, the ^{10}Be record appears more even.

The chronostratigraphy of core PS2471 can be somewhat improved when considering all stratigraphic data. The composition of Holocene assemblages is comparable with those of core PS2741 but concentrations of *Impagidinium pallidum* are lower. The increase of dinocyst concentrations allows the definition of the base of the Holocene at approximately 35-cm core depth (Fig. 7). This is in agreement with the occurrence of coccoliths in the upper 25 cm of the core, in particular *Coccolithus pelagicus*, indicating a Holocene age (Nowaczyk and Baumann, 1992), and the increase of ^{10}Be concentrations

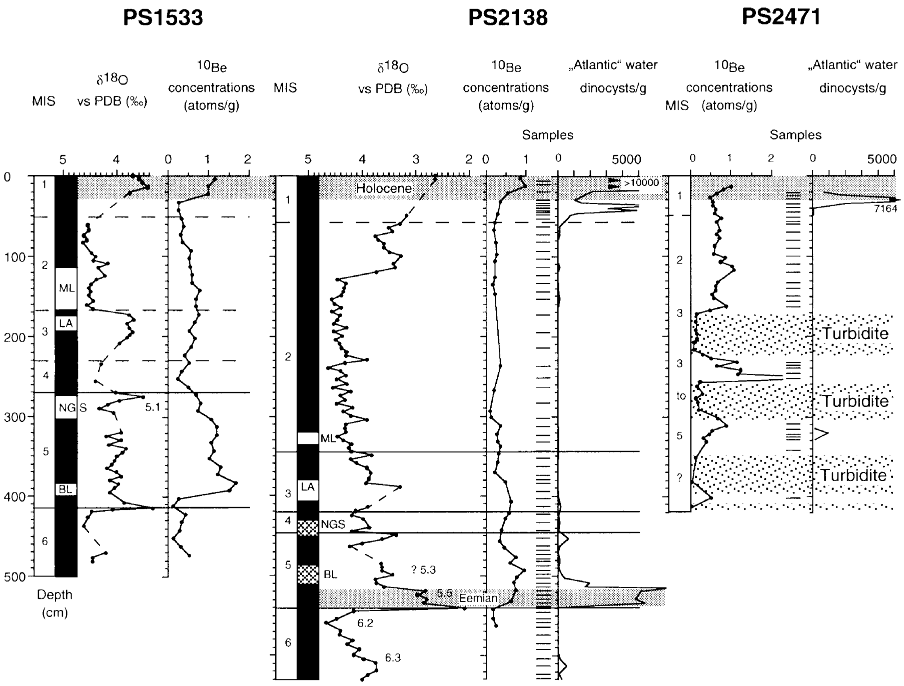

Fig. 8. Relationship between stable oxygen isotope and ^{10}Be stratigraphy and concentrations of total Atlantic water dinocysts in cores PS2138, PS1533 and PS2471. (ML: Mono Lake; LA: Laschamp; NGS: Norwegian–Greenland Sea; BL: Blake).

(Nürnberg et al., 1995; Andruleit, 1998, personal communication; Strobl et al., 1999). The base of MIS 1 is tentatively placed below the lowermost occurrence of *I. pallidum* at ca. 50-cm core depth (compare with core PS2741). The interval from MIS 1 to the top of the first turbidite can be assigned to MIS 2/3 because concentrations of dinocysts are as low as in the corresponding intervals of cores PS2138 and PS2741. Although concentrations are low in the interval between the first and second turbidite, assemblages resemble those from the base of MIS 3 to mid-MIS 5 in cores PS2138 and PS2741. The peak concentrations of *Operculodinium centrocarpum* and *Spiniferites elongatus* s.l. and presence of *Pentapharsodinium dalei* and *Impagidinium* spp. below the base of the second turbidite may indicate the Eemian. However, ^{10}Be concentrations are as low as in the supposedly MIS 2/3 sediments and do not support an MIS 5 age. The dinocyst record might also be related to the warmer climate phase in middle MIS 3 (Nørgaard-Pedersen et al., 1998), which might not be reflected in cores PS2138 and PS2741 because of lower sedimentation rates and lower sample resolution. In this case, the IRD layer at 341-cm core depth may correspond to the IRD layer in core PS2741 assumed to have an MIS 3.31 age (Fig. 6). Therefore, the base of the core might be assigned either to MIS 6 or 4. An independent age control, e.g., by means of luminescence dating, might help to resolve this issue. The stratigraphy might be further complicated by hiatuses that could occur at the base of the turbidites.

6.3. Core PS2757

A chronostratigraphy has been proposed, based on ^{230}Th$_{ex}$ and ^{10}Be stratigraphy, suggesting an MIS 9 age for the base of the analysed interval at 550-cm core depth of the 840-cm-long core (Strobl, 1998). Preliminary results of magnetostratigraphic studies have been used to assign an MIS 6 age for the base of the core (e.g., Nowaczyk in Müller, 1999) but the paleomagnetic data cannot easily be correlated with established magnetostratigraphic records from the eastern Arctic Ocean, and are therefore not used in this study. Stein et al. (in press) also proposed an MIS 6 age for the base by correlating organic geochemical parameters and magnetic susceptibility records of cores PS2741 and PS2757.

Concentration maxima of *Operculodinium centrocarpum* and *Pentapharsodinium dalei* mark the Holocene, which has a thickness of about 50 cm (Fig. 9). The MIS 1/2 boundary is tentatively placed below the lowermost occurrence of *Impagidinium pallidum* (see core PS2471) at ca. 60-cm core depth, indicating relatively high sedimentation rates of ca. 5 cm/ka. This is supported by two AMS ^{14}C dates in the box core from the same station, suggesting a sedimentation rate of ca. 7 cm/ka for near-surface sediments (Table 3; Spielhagen, unpublished data). Further downcore, dinocysts are almost absent. Extrapolating a sedimentation rate of 5–7 cm/ka downcore, the base of core PS2757 should be younger than 200,000 years. Thus, the slight dinocyst peak at 220-cm core depth might have an age of 55,000–30,000 years BP.

These relatively high sedimentation rates are in conflict with the ^{10}Be stratigraphy for the upper 550 cm (Strobl, 1998), suggesting average sedimentation rates less than 1 cm/ka. If we assume high sedimentation rates, the primary radioisotope concentrations may be strongly diluted by old terrigenous sediments leading to low concentrations of ^{10}Be and ^{230}Th$_{ex}$. However, core PS2757 should be completely analysed for ^{10}Be and ^{230}Th$_{ex}$ before a chronostratigraphy is established based on all stratigraphic data.

7. Paleoenvironmental implications

The advection of Atlantic waters provides the northern high latitudes with heat and moisture that is important for ice sheet growth and decay, and is responsible for seasonal open water conditions in the Fram Strait and Yermak Plateau region (e.g., Hebbeln et al., 1994; Dokken and Hald, 1996). Sediment core PS2138 is located in a crucial position underlying the Fram Strait branch of the Atlantic water (Fig. 1). Distinct fluctuations of dinocysts concentrations suggest a variable inflow of Atlantic waters at the northern Barents Sea (Fig. 3), as has been previously suggested by Knies et al. (1999, 2000). The most pronounced inflow events occurred in the Eemian, in the late glacial and the Holocene, leading to seasonally ice-free surface waters and warm climate conditions, while minor events may have occurred in MIS 6.3, 5.1 and at the stage 3/4 boundary. The Eemian dinocyst assemblages differ from those of the Holo-

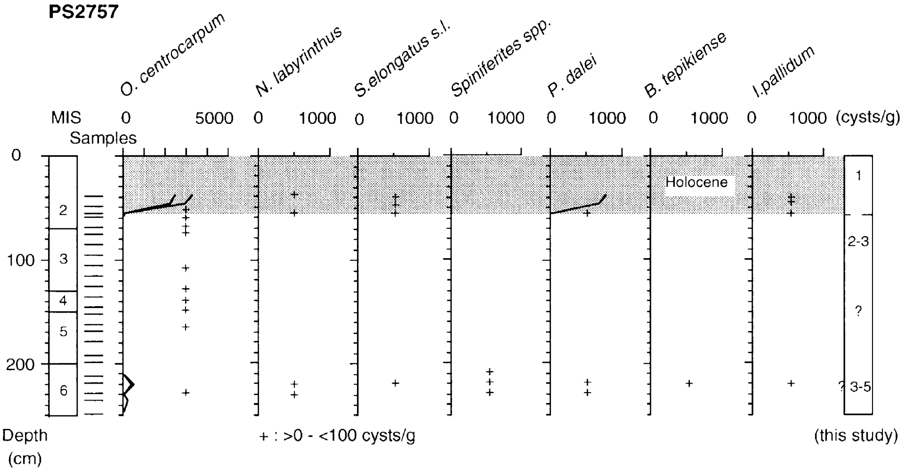

Fig. 9. Age models and dinocyst record of sediment core PS2757. The concentrations of total Atlantic water dinocysts (right line) and of *Operculodinium centrocarpum* (left line) are shown in one column. Only the upper 250 cm of the 810-cm long core are shown. Further 38 samples from the lower part are barren, except for single occurrences of *Nematosphaeropsis labyrinthus* at 645- and 76.5-cm core depth. The reference age model is from Strobl (1998).

cene, suggesting somewhat different sea-surface conditions. *Bitectatodinium tepikiense* is a common species not only in the Eemian (Fig. 3), but also in Pliocene and early Pleistocene sediments of the Yermak Plateau (Matthiessen and Brenner, 1996). A detailed paleoenvironmental interpretation of the Eemian dinocyst assemblages will be published elsewhere.

In contrast, previous studies suggested that Atlantic water inflow was moderate and well below Holocene level in the Eemian (Köhler and Spielhagen, 1990; Hebbeln et al., 1998; Bauch et al., 1999a). However, the occurrence of distinct benthic foraminifers in core PS2138, reflecting a stronger Atlantic water inflow in the Eemian (Wollenburg et al., 2001), supports our interpretation. Benthic foraminifers and molluscs from coastal sections in western Spitsbergen indicate that the Eemian was warmer than today but probably cooler than during the Holocene climate optimum (Mangerud et al., 1998) or at least as warm as MIS 5.1 (Bergsten et al., 1998).

These contrasting interpretations might have resulted from the variable environmental conditions in the Fram Strait and Yermak Plateau area. On the one hand, the reconstruction of a strongly diminished inflow of warmer waters in the Eemian was based on carbonate records from cores located in the central and western Fram Strait (Hebbeln et al., 1998; Bauch et al., 1999a), which may not reflect the conditions in the main path of the Westspitsbergen Current carrying the Atlantic water northward. On the other hand, increased carbonate production along Westspitsbergen in the Eemian might have been altered by enhanced dissolution, leading to lower carbonate contents in the interglacials than glacials (e.g., core PS2138, Knies et al., 1999). Wollenburg et al. (2001) explicitly stated that carbonate dissolution affected benthic foraminiferal faunas during the interglacial periods of enhanced productivity and Atlantic water advection in cores PS2138 and PS2212, supporting the view of Knies et al. (1999) that the almost complete dissolution of biogenic carbonate is linked to the strongest advection of Atlantic waters in MIS 1 and 5.

In the past 150,000 years, two major intrusions occurred along the Eurasian continental margin in the Eemian and Holocene (Figs. 3, 6, 7 and 8). The Holocene conditions were established at all sites from Fram Strait to the Laptev Sea continental margin, whereas the equivocal chronostratigraphies of cores PS2471 and PS2757 do not allow to trace the Eemian inflow beyond Severnaya Zemlya. The onset of interglacial conditions may have occurred almost synchronously along the Eurasian continental margin, based on the available temporal resolution of records (Fig. 10). The increase of dinocyst concentrations is dated in sediment core PS1295 from Fram Strait at ca. 9700 years BP and in core PS2458 from the eastern Laptev Sea continental margin at ca. 9700 years BP, supported by the AMS ^{14}C dates of cores PS2138 and PS2741 at the base of MIS 1. This agrees with a rise of biogenic production in core PL94-67 from the St. Anna Trough (Fig. 1) at 9500 years BP, which was related to the onset of interglacial conditions (Hald et al., 1999).

The strongest inflow of Atlantic waters is primarily related to periods of high sea level and absence or low extent of ice sheets on the Barents Sea/Spitsbergen shelves, when Atlantic water could intrude both around Spitsbergen and across the Barents/Kara Sea. This is clearly demonstrated by the retreat of the Barents Sea Ice sheet during the last deglaciation. After a first major drawdown of the marine-based part in the Barents Sea from ca. 15,200 years BP, major parts of the Barents Sea and the adjacent northern Kara Sea were deglaciated by ca. 13,000 years BP and finally the shallow banks and islands were ice-free by ca. 10,000 years BP (Polyak et al.,

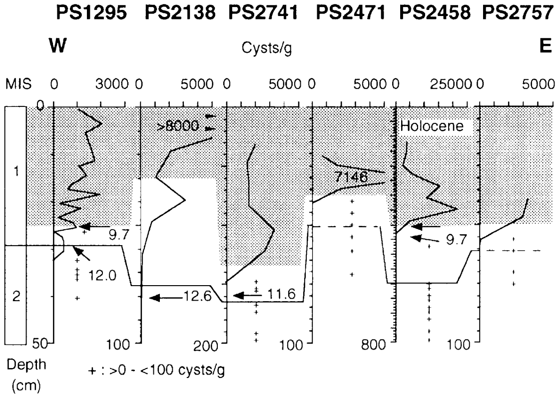

Fig. 10. Concentrations of total Atlantic water dinocysts during the last deglaciation at the Eurasian continental margin. Note the different scales.

1997; Landvik et al., 1998; Hald et al., 1999). Thus, warmer waters probably could not flow through the Barents Sea before 13,000 years BP and the modern circulation system was not established before the onset of the Holocene.

8. Conclusions

Concentrations of cosmopolitan and temperate to subpolar dinocysts display distinct fluctuations in sediment cores from the Eurasian continental margin, and they seem to indicate responses to variations in strength of Atlantic water inflow and/or changes in surface and near-surface water mass conditions. The dinocyst taxa recorded in the Arctic Ocean sediments have long stratigraphic ranges, and thus, changes in dinocyst assemblage composition reflect variable ecological conditions. Therefore, the ecostratigraphy must be supplemented by stable oxygen isotope and paleomagnetic data as well as ^{10}Be and ^{230}Th radioisotopes to establish age models of sediment cores.

Distinct changes in concentrations of Atlantic water dinocyst occurred approximately synchronously during the transition to the Holocene (i.e., within millenia) along the Eurasian continental margin. Pronounced peaks in total concentrations are confined to the interglacials and mainly reflect increases in cosmopolitan, opportunistic species. The lower–middle Holocene is recorded in all sediment cores at the Eurasian continental margin, whereas the extent of climate warming from Atlantic inflow during the Eemian cannot be fully assessed because of contradictory stratigraphic information and variable sedimentation rates of cores from the eastern Laptev Sea continental margin. The dinocyst assemblages, however, show that warmer surface and/or subsurface waters penetrated at least to Severnaya Semlya during the Eemian.

Age models for the Eastern Arctic Ocean cores may be considerably improved if the different stratigraphic methods are calibrated in a standard reference section. In particular, detailed ^{230}Th and ^{10}Be records may provide valuable chronostratigraphic information but they must be calibrated vs. both high-resolution stable oxygen isotope and magnetostratigraphic records, such as those from core PS2138, to improve their stratigraphic applicability. Furthermore, dating methods such as optical stimulated luminescence (OSL) may give additional absolute age information beyond the range of radiocarbon dating and in sediment cores barren of biogenic carbonate. Despite still existing contradictory age models at the Laptev Sea continental margin, a combination of all available physical, micropaleontological and sedimentological methods will lead to more precise age models for carbonate-poor sequences in the eastern Arctic Ocean.

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