

Dinoflagellate cyst evidence for warm interglacial conditions at the northern Barents Sea margin during marine oxygen isotope stage 5

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ABSTRACT: Dinoflagellate cyst analysis has been conducted on a sediment core from the northern Barents Sea margin (eastern Arctic Ocean) to reconstruct sea-surface conditions during marine oxygen isotope stage 5. Cyst concentrations and composition of assemblages display a distinct variability reflecting deglacial warming and the onset of relatively warm interglacial conditions in early stage 5. Dinoflagellate cyst assemblages of the last interglacial, which corresponds broadly to substage 5e, are comparable with those from recent sediments of the area, but have lower abundances of cold water taxa. Such assemblages suggest that sea-surface temperatures of the last interglacial were therefore at least as warm as today or even warmer. After a distinct cooling, relatively stable cold conditions with minor periods of surface water warming prevailed from mid-substages 5d to 5b. Temperatures might have been similar in substages 5a and 5e, but the different compositions of assemblages indicate a stronger stratification of surface waters, probably owing to a larger meltwater supply, in substage 5a. These results are compared with previous studies on marine carbonate and calcareous microfossil records from the Fram Strait and Yermak Plateau area, which show that the general view of a cold last interglacial in the eastern Arctic Ocean is untenable.

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Introduction

The palaeoenvironmental evolution of the eastern Arctic Ocean during late Quaternary glacial–interglacial cycles is known only on a rather broad scale. This is the result mainly of low chronostratigraphical resolution of sedimentary records and limits in the applicability of traditionally used proxies, such as calcareous microfossils and stable isotopes. Despite these restrictions, recent studies provide evidence for a distinct response of the eastern Arctic Ocean to climate-induced environmental change, such as the Atlantic water inflow and surface water productivity (e.g. Hebbeln and Wefer, 1997; Nørgaard-Pedersen *et al.*, 1998; Knies *et al.*, 2000; Wollenburg *et al.*, 2001). Moreover, the glaciations in northern Eurasia led to changes in source areas and transport pathways of terrigenous sediments that are reflected in distinct lithologies in sediment cores (e.g. Flower, 1997; Spielhagen

et al., 1997; Knies *et al.*, 2000; Jakobsson *et al.*, 2000; Vogt *et al.*, in press).

Although numerous interglacials have been identified in marine sequences from the eastern Arctic Ocean, most studies have focused on the last deglaciation and marine (oxygen) isotope stage (MIS) 1 (e.g. Stein *et al.*, 1994; Polyak *et al.*, 1997; Hald *et al.*, 1999; Duplessy *et al.*, 2001). The last marine interglacial (MIS 5e), which is correlated to the Eemian in terrestrial sequences, is of particular importance because palaeoclimate conditions in the Nordic Seas and adjacent land areas were for the last time in the past 150 000 yr comparable to the Holocene (e.g. Kellogg *et al.*, 1978; Mangerud *et al.*, 1979, 1998; Henrich, 1998). Thus, palaeoclimate variability of the last interglacial may provide a good analogue for future climate evolution of the present interglacial. Previously, contradictory palaeoenvironmental scenarios have been proposed for the adjacent eastern Arctic Ocean during the last interglacial (Gard and Backman, 1990; Köhler and Spielhagen, 1990; Hebbeln and Wefer, 1997; Bauch *et al.*, 1999; Knies *et al.*, 1999; Wollenburg *et al.*, 2001). These reconstructions suggest either relatively cold surface water conditions, probably even colder than today, and strongly reduced advection of Atlantic waters, or a relatively strong Atlantic water inflow and distinct interglacial conditions.

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In contrast, palynological data demonstrate that the last interglacial in northern Eurasia was warmer than today or even warmer than during the Holocene climate optimum (e.g. Velichko *et al.*, 1991; Lozhkin and Anderson, 1995; Saarnisto *et al.*, 1999). Shallow-water microfossils from raised marine sediments support this interpretation, indicating that temperatures of circum-Arctic coastal waters were similar to today or even warmer during the last interglacial (e.g. Miller *et al.*, 1989; Brigham-Grette and Hopkins, 1995; Funder *et al.*, 1998; Mangerud *et al.*, 1998; Kelly *et al.*, 1999).

Recently, benthic foraminiferal and dinoflagellate cyst (= dinocyst) data on sediment core PS2138 revealed evidence of a pronounced marine interglacial at the northern Barents Sea margin in MIS 5 (Wollenburg *et al.*, 2001; Matthiessen *et al.*, in press). The core site underlies the Svalbard branch of the West Spitsbergen Current transporting relatively warm Atlantic waters into the eastern Arctic Ocean (e.g. Manley *et al.*, 1992; Fig. 1). The area is ice-free or located in the marginal ice zone for some time during the summer season (e.g. Romanov, 1995; Fig. 1). We present here, for the first time, the detailed dinocyst record of MIS 5 and discuss the palaeoenvironmental development of the northern Barents Sea margin from ca. 130 000 to 60 000 yr BP. Dinocysts are a valuable tool for reconstructing sea-surface parameters such as temperature and salinity in high northern latitudes (Rochon *et al.*, 1999; de Vernal *et al.*, this issue) and thus may provide new insights into sea-surface conditions during the penultimate deglaciation and the last interglacial.

Material and methods

Lithology and sediment composition

Sediment core PS2138-1 was taken at the northern Barents Sea margin (81°32.1' N, 30°35.6' E, 995 m water depth) during the expedition ARK VIII/2 of RV *Polarstern* in 1991 (Rachor, 1992).

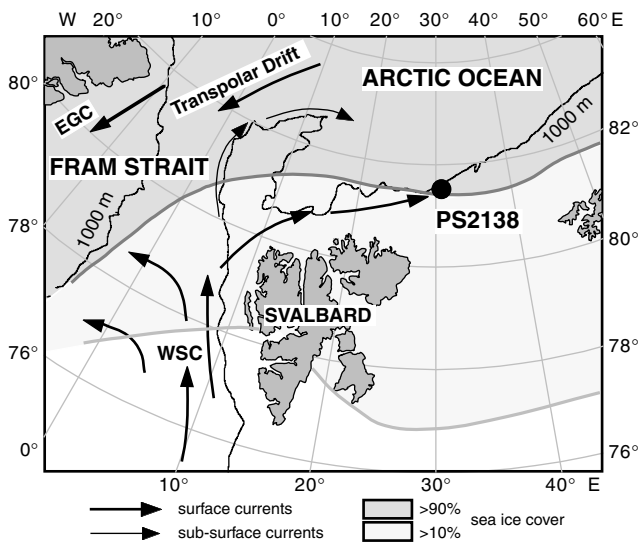


Figure 1 Location of core PS2138 at the northern Barents Sea margin. Surface and subsurface currents are shown schematically (Manley *et al.*, 1992). The mean concentration of sea-ice (>10% and >90%, respectively) is shown for August to September, based on data from 1972–1981 (Romanov, 1995). (EGC: East Greenland Current; WSC: West Spitsbergen Current)

The sediments generally consist of fine-grained bioturbated, carbonate-poor glaciomarine muds with few intercalations of laminated and coarse-grained layers (Figs 2 and 3; Knies *et al.*, 1999).

The sedimentological and geochemical methods applied are described in detail elsewhere (Knies *et al.*, 1999, 2000). Ice-rafted debris (IRD) was counted on X-radiographs as number of grains >2 mm 10 cm⁻². Opal and total organic carbon (TOC) contents show low variability, whereas carbonate values range from a few per cent to ca. 20% (Fig. 3). The TOC predominately consists of terrestrial organic matter, as indicated by low hydrogen indices and light $\delta^{13}\text{C}$ values of bulk TOC (Knies *et al.*, 2000). A relatively constant amount of carbonate consists of detrital dolomite (Knies *et al.*, 1999).

Stratigraphy

The chronostratigraphy of the core is based on $\delta^{18}\text{O}$ stratigraphy of *Neogloboquadrina pachyderma sin.* (>63 μm), supported by 21 AMS ¹⁴C ages, magnetostratigraphy and benthic stable isotopes—all revealing unequivocally the presence of the last two glacial–interglacial cycles (Fig. 2 and Table 1; Knies *et al.*, 1999, 2000; Nowaczyk and Knies, 2000; Wollenburg *et al.*, 2001; Matthiessen *et al.*, in press). Benthic stable isotopes were measured on the foraminifers *Melonis zaandami* and *Cassidulina teretis* which have an intermediate to deep infaunal and shallow infaunal habitats, respectively (Wollenburg and Mackensen,

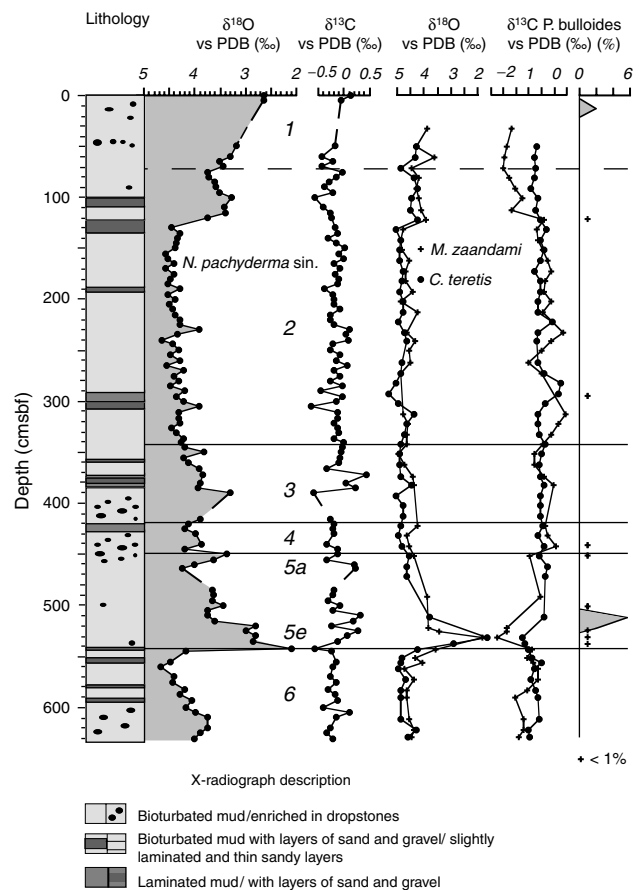


Figure 2 Lithology, and oxygen and carbon isotope stratigraphy of core PS2138 (Knies *et al.*, 1999; Wollenburg *et al.*, 2001). The distribution of the stratigraphic marker *P. bulloides* is shown

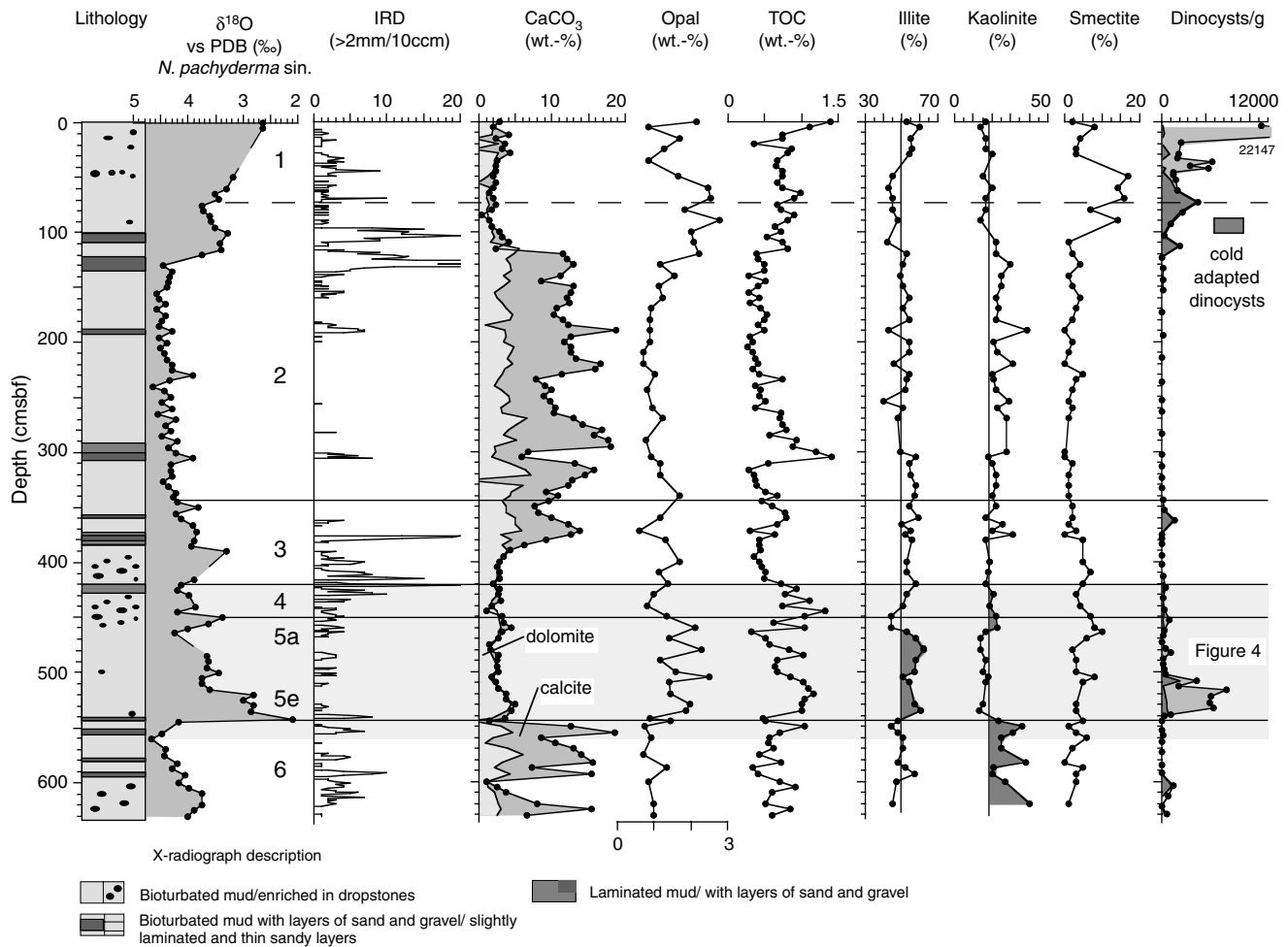


Figure 3 Stable oxygen isotope stratigraphy, lithology, sedimentological parameters, clay mineralogy (Knies *et al.*, 1999, 2000) and total dinocyst concentration of core PS2138. Concentration of cold-adapted dinocysts (*Islandinium minutum* s.l., *Brigantedinium* spp.) that dominate polar assemblages (Rochon *et al.*, 1999) also is illustrated

1998; Wollenburg *et al.*, 2001). Marine isotope stages were derived from AMS ^{14}C ages and isotope events using the orbitally tuned time-scale of Martinson *et al.* (1987). MIS 5 was subdivided into substages according to Shackleton (1969).

Dinocyst analysis

Dinocysts have been analysed on average at intervals of 10 cm, but at 3 to 5 cm in early MIS 5. The freeze-dried samples were processed with standard palynological preparation methods. After each treatment with hydrochloric and hydrofluoric acids, fine silt and clay particles were removed by wet sieving with a 6 μm mesh. The final residues were mounted in glycerin jelly on a microscope slide, covered with a glass slip and permanently sealed with paraffin wax. Variable numbers of dinocysts were counted, but in most samples from MIS 4 and 5 more than 90 to 100 cysts were enumerated. Nomenclature and taxonomy of dinoflagellate cysts (Table 2) basically follows those of Williams *et al.* (1998), Rochon *et al.* (1999) and Head *et al.* (this issue). *Islandinium minutum* s.l., which formerly was known as *Algidasphaeridium? minutum* s.l., comprises *I. minutum* and the related species *I.? cezare* and *Echinidinium karaense* (Head *et al.*, this issue). *Brigantedinium* spp. includes all spherical brown protoperidinioid cysts without processes because orientation of specimens did not usually permit

identification to species level. *Spiniferites* spp. includes all *Spiniferites* cysts except *Spiniferites elongatus* s.l. (*S. elongatus*, *S. frigidus* and intergrades). Concentrations were estimated by adding tablets with a known amount of *Lycopodium clavatum* spores to weighed samples (Stockmarr, 1971). Accumulation rates of dinoflagellate cysts in MIS 5 were calculated according to van Andel *et al.* (1975; for details of method see Knies *et al.*, 1999).

All data are stored in the PANGAEA information system (<http://www.pangaea.de>) at the Alfred Wegener Institute for Polar and Marine Research, Bremerhaven.

Chronostratigraphy of marine oxygen isotope stage 5 in core PS2138

The MIS 6–5 transition is clearly marked by a distinct shift in the stable oxygen and carbon isotope records from heavy to light values between 545 and 535 cm core depth (Fig. 2; Knies *et al.*, 1999; 2000; Wollenburg *et al.*, 2001). A single light planktonic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ value at 542 cm core depth is associated with layers enriched in coarse-grained sediments indicating the retreat of the Svalbard–Barents Sea ice-sheet (SBIS) from the shelf edge and the release of deglacial meltwater (Figs 2 and 3; Knies *et al.*, 2000). Therefore, the MIS 5–6 boundary is difficult

Table 1 The AMS ^{14}C datings and isotope events in core PS2138. *Neogloboquadrina pachyderma* sin. (*N. pachy.* sin.) and mixed foraminifers were picked from the $>63\ \mu\text{m}$ fraction. The AMS ^{14}C ages were $\delta^{13}\text{C}$ -normalised and corrected for a reservoir effect of 440 yr (Mangerud and Gulliksen, 1975). The radiocarbon ages were calibrated to calendar ages using the program CALIB 4.3 for ages younger than 20 700 ^{14}C yr (Stuiver and Reimer, 1993) and a second-order polynomial for older ages (Bard *et al.*, 1998)

Depth (cm)	^{14}C age (BP)/ isotope events	Reservoir corrected (yr BP)	Calendar age (BP)	Carbon source	Laboratory number
50	13 460 \pm 110	13 020	*	Bivalve shells	KIA1282
60	MIS 2.0	12 400	13 443		
65	13 430 \pm 70	12 990	15 524	<i>N. pachy.</i> sin.	KIA9872
80	13 040+140/-130	12 600	*	Bivalve shells	KIA363
110	14 030 \pm 80	13 590	16 205	Mixed forams	KIA4765
130	15 850 \pm 130	15 410	18 299	Mixed forams	KIA1283
160	16 670 \pm 210	16 230	19 243	<i>N. pachy.</i> sin.	KIA364
200	17 320 \pm 130	16 880	19 991	<i>N. pachy.</i> sin.	KIA2745
230	19 230 \pm 140	18 790	22 189	<i>N. pachy.</i> sin.	KIA9873
244	19 710 \pm 130	19 270	22 741	<i>N. pachy.</i> sin.	KIA9874
260	20 420 \pm 130	19 980	23 558	<i>N. pachy.</i> sin.	KIA9875
275	20 700 \pm 130	20 260	23 880	<i>N. pachy.</i> sin.	KIA9876
300	20 480+330/-320	20 040	23 627	<i>N. pachy.</i> sin.	KIA365
316	22 470 \pm 160	22 030	25 943	<i>N. pachy.</i> sin.	KIA9877
331	23 540 \pm 240	23 100	27 177	<i>N. pachy.</i> sin.	KIA2744
345	MIS 3.0	24 000	28 200		
345	24 910+200/-190	24 470	28 748	<i>N. pachy.</i> sin.	KIA10372
360	26 240+280/-270	25 800	30 261	Bivalve shells	KIA4766
366	26 260 \pm 230	25 820	30 284	<i>N. pachy.</i> sin.	KIA10373
372	30 290+360/-350	29 850	34 805	<i>N. pachy.</i> sin.	KIA10374
375	$>45\ 190$		**	<i>N. pachy.</i> sin.	KIA10375
380	35 340+1570/-1310	34 900	40 333	Mixed forams	KIA1284
385	30 240+380/-370		***	<i>N. pachy.</i> sin.	KIA10376
420	MIS 4.0		59 000		
450	MIS 5.0		74 000		
542	MIS 6.0		130 000		

* Bivalve ages are not used because the shell fragments may be reworked.

** Non-finite age.

*** Unreliable age because carbon content of sample was too low for accurate measurement.

to define from the planktonic isotope record alone. We tentatively place it at the light peak in the planktonic record that is also the mid-point of the benthic oxygen isotope shift.

The shape of the planktonic oxygen isotope curve of MIS 5 is comparable to those from a number of cores from Fram Strait (Lloyd *et al.*, 1996; Hebbeln and Wefer, 1997) although a major gap occurs in PS2138 in upper MIS 5 owing to the absence of planktonic foraminifers (Fig. 2). Both planktonic and benthic isotope records indicate the presence of substage MIS 5e. A plateau of relatively light planktonic $\delta^{18}\text{O}$ values from ca. 520 to 532 cm core depth is associated with relatively heavy $\delta^{13}\text{C}$ values. The benthic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data reflect interglacial conditions from ca. 538 to 522 cm core depth (Wollenburg *et al.*, 2001).

Substages 5a to 5d cannot be distinguished unequivocally owing to the low resolution of the isotope records. The light planktonic $\delta^{18}\text{O}$ value at 450 cm core depth is interpreted as isotope event 5.1 (Knies *et al.*, 1999). Stage 4 is characterised by heavy $\delta^{18}\text{O}$ and light $\delta^{13}\text{C}$ values. The MIS 4–5 boundary is placed at the mid-point of the oxygen isotope shift at ca. 448 cm core depth. The resolution of the benthic isotope curve is too low to allow a definition of the MIS 4–5 boundary (Fig. 2).

The age model based on stable isotope stratigraphy is supported by the occurrence of the benthic foraminifer *Pullenia*

bulloides (Fig. 2), a characteristic infaunal species in 500 to 1000 m water depth at the northern Barents Sea margin (Wollenburg and Mackensen, 1998). *Pullenia bulloides* is present in few stratigraphic intervals but distinct abundance maxima occur only in upper MIS 1 and above MIS 5e (Wollenburg *et al.*, 2001).

This distribution pattern is similar to those in well-dated sediment cores from shallow water depths ($<1500\ \text{m}$) in the Nordic Seas (Haake and Pflaumann, 1989; Struck, 1997). In those cores, *P. bulloides* occurs mainly in MIS 1 and 5 and abundance maxima in the last 150 000 yr are confined both to Holocene and MIS 5a, but some cores also have a second maximum in MIS 5e (Struck, 1997). In contrast, in sediment cores from deeper waters in the Nordic Seas *P. bulloides* is confined to the interval from the MIS 4–5 boundary to MIS 5d and has only a single abundance maximum in MIS 5a (Haake and Pflaumann, 1989; Struck, 1997). Based on the studies of Streeter *et al.* (1982) and Haake and Pflaumann (1989), the maximum of *P. bulloides* is used in the Nordic Seas and Fram Strait as a stratigraphical indicator of MIS 5a (e.g. Köhler and Spielhagen, 1990; Fronval and Jansen, 1997; Hebbeln and Wefer, 1997; Bauch *et al.*, 1999). Lloyd *et al.* (1996) noted a somewhat different stratigraphical occurrence in cores from the western Svalbard margin, where the maximum abundance is located below the light stable oxygen isotope peak of MIS 5a (see also Köhler and Spielhagen, 1990; Bauch *et al.*, 1999). This distribution agrees well with

Table 2 Dinocyst taxa in core PS2138

<i>Ataxiodinium choane</i> Reid 1974
<i>Bitectatodinium tepikiense</i> Wilson 1973
<i>Brigantedinium</i> spp. Reid 1977 ex Lentin & Williams 1993
<i>Echinidinium karaense</i> Head <i>et al.</i> this issue
<i>Impagidinium</i> spp. ^a Stover & Evitt 1978
<i>Impagidinium pallidum</i> Bujak 1984
<i>Impagidinium patulum</i> (Wall 1967) Stover & Evitt 1978
<i>Impagidinium sphaericum</i> (Wall 1967) Lentin & Williams 1981
<i>Islandinium minutum</i> (Harland & Reid in Harland <i>et al.</i> 1980) Head <i>et al.</i> this issue
<i>Islandinium? cezare</i> (de Vernal <i>et al.</i> , 1989 ex de Vernal in Rochon <i>et al.</i> , 1999) Head <i>et al.</i> this issue
<i>Lingulodinium machaerophorum</i> (Deflandre & Cookson 1955) Wall 1967
<i>Nematosphaeropsis labyrinthus</i> (Ostenfeld 1903) Reid 1974
<i>Operculodinium centrocarpum</i> sensu Wall and Dale 1966
Cyst of <i>Pentapharsodinium dalei</i> Indelicato & Loeblich III 1986
Cyst of <i>Polykrikos schwartzii</i> Bütschli 1873
<i>Spiniferites</i> spp. Mantell 1850
<i>Spiniferites elongatus</i> Reid 1974 <i>sensu lato</i>
<i>Spiniferites lazus</i> Reid 1974
<i>Spiniferites hyperacanthus</i> (Deflandre & Cookson 1955) Cookson & Eisenack 1974
<i>Spiniferites mirabilis</i> (Rossignol 1964) Sarjeant 1970
<i>Spiniferites ramosus</i> (Ehrenberg 1938) Mantell 1854 <i>sensu lato</i>
Species <i>et</i> Genus indet.

^a All species of *Impagidinium* excluding *I. pallidum*

the maximum abundance of *P. bulloides* in core PS2138 in mid-MIS 5 sediments (Fig. 2), thus supporting our age model.

Based on the age-fix points of events 5.0 and 6.0 (Table 1) an average sedimentation rate of 1.6 cm ka⁻¹ is calculated for MIS 5. The selected sample intervals thus represent an average resolution of 3000 to 5000 yr in MIS 5 and of ca. 3000 yr in MIS 5e.

Dinocyst record of marine oxygen isotope stage 5

Although sedimentation rates were low and bioturbation may have smoothed the dinocyst record, concentrations and composition of assemblages reflect distinct changes in environmental conditions from late MIS 6 to MIS 4 (Fig. 4). These fluctuations occurred against a background of a permanent meridional circulation system, which experienced some variability in the intensity of Atlantic water inflow during the last two glacial–interglacial cycles (Hebbeln and Wefer, 1997; Knies *et al.*, 1999). In the following section we discuss the distribution of selected dinocyst taxa that usually make up more than 85 to 90% of the assemblages in core PS2138, and tentatively define the assemblage zones.

Assemblage zone I (542–535 core depth)

After the termination of the meltwater event, cyst concentrations increase by an order of magnitude allowing us to count a statistically significant number of specimens (Fig. 4). The dominant taxa in zone I, *Islandinium minutum* s.l. and *Brigantedinium* spp., characterise polar assemblages from the high northern latitudes (e.g. Mudie, 1992; de Vernal *et al.*, this issue). *Islandinium minutum* is a polar to subpolar taxon that is common to abundant in environments with low summer sea-surface temperatures (<5 °C) and an extensive seasonal sea-ice cover (Dale, 1996; de Vernal *et al.*, this issue). *Islandinium? cezare* and *Echinidinium karaense*, which are included in *A. minutum* s.l., have comparable ecological preferences (Head *et al.*, this issue).

Brigantedinium spp. is a cosmopolitan taxon that is particularly abundant in regions with enhanced surface water productivity, such as upwelling areas and polynyas (Wall *et al.*, 1977; Dale, 1996; de Vernal *et al.*, this issue). Species of this taxon are cysts of heterotrophic dinoflagellates, which feed on a variety of food sources, such as diatoms (e.g. Jacobsen and Anderson, 1986). Therefore, the availability of food may be a more important ecological factor for *Brigantedinium* than physical parameters (e.g. Dale, 1996).

The only other taxa that are important numerically in assemblage zone I are the cosmopolitan neritic

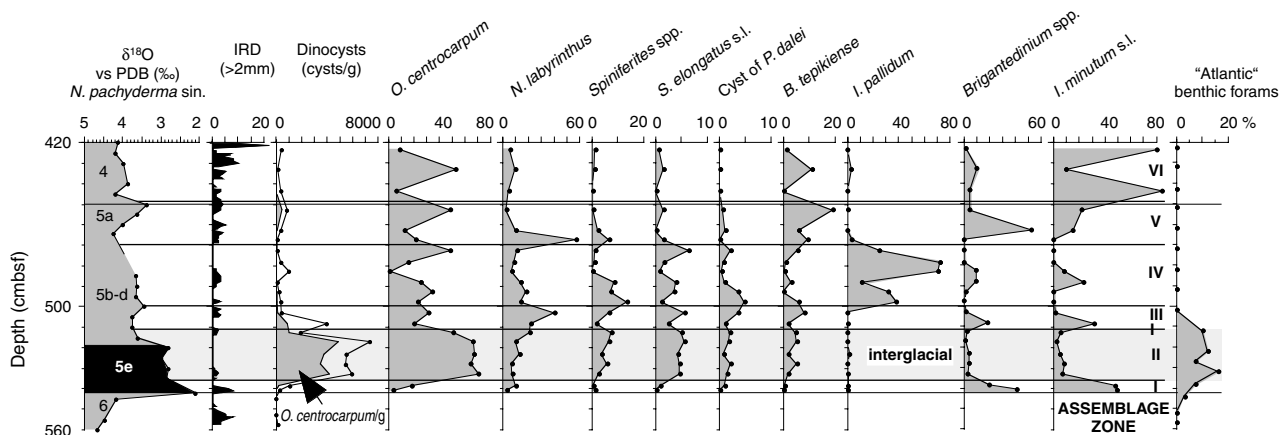


Figure 4 Isotope stratigraphy, total dinocyst concentrations and percentages of selected dinocyst taxa in core PS2138 during MIS 4 to 6. The distributions of 'Atlantic' benthic foraminifers (*Epistominella pusilla*, *Pullenia bulloides*, *Discorbinella bertheloti*) and ice-rafted debris also are shown (Knies *et al.*, 1999; Wollenburg *et al.*, 2001)

to oceanic species *Operculodinium centrocarpum* and *Nematosphaeropsis labyrinthus* (Wall *et al.*, 1977; Dale, 1996). These species tolerate a wide range of temperature conditions and have broad abundance maxima at summer temperatures of ca. 2–16 °C (Wall *et al.*, 1977; Dale, 1996; Rochon *et al.*, 1999; de Vernal *et al.*, this issue), but they also occur in polar environments with seasonal sea-ice cover. The ecological preferences of both species differ in that *O. centrocarpum* tolerates salinities as low as 3 (psu) = practical salinity unit (Dale, 1996), whereas *N. labyrinthus* requires normal marine salinities (>20 psu, de Vernal *et al.*, this issue).

The assemblages of zone I indicate cold polar conditions and an extensive seasonal sea-ice cover. Open-water conditions may have prevailed in a short summer season, with an increased surface water production in the marginal ice-zone.

Assemblage zone II (535–510 cm core depth)

Assemblage zone II corresponds roughly with the plateau of light $\delta^{18}\text{O}$ values indicative of MIS 5e. The assemblages are marked by the dominance of *O. centrocarpum*, a high number of other temperate to subpolar taxa, and a pronounced maximum of cyst concentrations. This zone is bracketed by peak abundances of *I. minutum* s.l. and *Brigantedinium* spp. A number of other taxa are present with low, but slightly higher abundances than in assemblage zone I (Fig. 4).

High percentages of *O. centrocarpum*, which are associated with high cyst concentrations, characterise the pathway of relatively warm Atlantic waters in the northern North Atlantic and the adjacent Norwegian Sea (Matthiessen, 1995; Rochon *et al.*, 1999). Dale (1996) noted a proportional seaward increase of this species in recent sediments at the transition from neritic to oceanic environments, which corresponds to the location of core PS2138. Thus, the peak abundance of *O. centrocarpum* represents an interglacial signal at the northern Barents Sea margin.

The distinct increase of taxa such as *Spiniferites* spp. (mainly *S. ramosus* s.l.), *Spiniferites elongatus* s.l., cyst of *Pentapharsodinium dalei* and *Bitectatodinium tepikiense*, is related to the onset of interglacial conditions (Fig. 4). Cyst of *P. dalei* and *Spiniferites elongatus* s.l. occur today in sediments from polar to subpolar environments with winter sea-ice cover but they prefer temperatures >2 °C (Dale, 1996; Rochon *et al.*, 1999; de Vernal *et al.*, this issue). *Bitectatodinium tepikiense* is considered to be a characteristic species of the transitional zone between temperate and subpolar environments (Dale, 1996). It is abundant in regions with a pronounced seasonality of surface water conditions, having cold winters with sea-ice cover and relatively warm surface waters in summer (>10 °C, Rochon *et al.*, 1999).

A number of rare taxa are present almost exclusively in assemblage zone II, such as *Ataxiodinium choane*, *Impagidinium* spp., *I. sphaericum*, *I. patulum*, *Spiniferites lazus*, *S. hyperacanthus*, *S. mirabilis*, and *Tectatodinium pellitum*. They also are rare in recent sediments from the Norwegian Sea and the Eastern Arctic Ocean (Matthiessen, 1991; Mudie, 1992).

The assemblages of zone II mark relatively warm interglacial conditions at the northern Barents Sea margin. The common occurrence of *I. minutum* s.l. suggests a strong seasonality of surface water conditions, with relatively cold winters with seasonal sea-ice cover and a substantial warming in summer.

Assemblage zone III (510–500 cm core depth)

A sharp decline of dinocyst concentrations occurs at the transition to zone III in substage 5d. Assemblages in this zone have a variable composition. Dominance of *O. centrocarpum* is replaced by an abundance maximum of *N. labyrinthus*.

Islandinium minutum s.l. and *Brigantedinium* spp. dominate in the lowermost sample of assemblage zone III, suggesting polar conditions with an extended duration of sea-ice cover, comparable to those in zone I. Simultaneously, most other taxa show a minimum in abundance. The increase of *N. labyrinthus*, *S. elongatus* s.l., cyst of *P. dalei* and *B. tepikiense* indicates a return to warmer conditions in summer and a reduced seasonal sea-ice cover at the top of zone III.

Assemblage zone IV (500–470 cm core depth)

A pronounced change in assemblage composition occurred in mid-MIS 5. The acme of *Impagidinium pallidum* and a minimum of *B. tepikiense* are bracketed by peak abundances of *N. labyrinthus*.

Percentages of *I. pallidum* in the lower part of assemblage zone IV are similar to those in recent sediments from the central Greenland and Iceland seas (Matthiessen, 1995), although a modern analogue does not exist for the extreme predominance of *I. pallidum* (>70%) at the top of the zone. This oceanic subpolar to polar species is abundant at low temperatures (<5 °C) and occurs in regions with seasonal sea-ice cover (Matthiessen, 1995; Dale, 1996; de Vernal *et al.*, this issue). In the middle part of zone IV, *I. pallidum* is replaced by higher percentages of *I. minutum* s.l. and *Brigantedinium* spp. and a minimum of other taxa, indicating even colder conditions and a more extensive sea-ice cover. Mid-MIS 5 (approximately substage 5d to 5b) is thus characterised by relatively stable cold polar conditions and an extensive seasonal sea-ice cover.

Assemblage zone V (470–450 cm core depth)

This zone is characterised by a significant increase of *B. tepikiense* co-occurring with abundant *O. centrocarpum*. Percentages of other taxa (*Spiniferites* spp., *S. elongatus* s.l. and cyst of *P. dalei*) are low except for isolated peaks of *N. labyrinthus* and *Brigantedinium* spp. (Fig. 4).

A strong decline of *I. pallidum* suggests a warming of surface waters, which commenced with peak abundances of *O. centrocarpum* and *N. labyrinthus* at the transition to zone V. The single peak of *Brigantedinium* spp., which is associated with common *I. minutum* s.l., indicates a return to cooler conditions and may reflect an enhanced availability of food owing to higher surface water production in the marginal ice-zone.

Sea-surface temperatures of upper zone V may have been similar to those of zone II. The assemblage compositions, however, differ considerably because of lower cyst concentrations, co-occurrence of *B. tepikiense* and *O. centrocarpum*, and common occurrence of *I. minutum* s.l. in zone V. This indicates a surface water warming that probably was linked to a more pronounced stratification owing to substantial freshwater supply in summer during zone V.

Assemblage zone VI (450–420 cm core depth)

This zone is characterised by marked shifts between the assemblages dominated either by *I. minutum* s.l. or by *O. centrocarpum* and *B. tepikiense*. Percentages of other taxa remain low and cyst of *P. dalei* disappears at the base of the zone.

The predominance of *I. minutum* s.l. reflects cold sea-surface conditions with an extensive seasonal sea-ice cover at the top and base of the zone. A sea-surface warming in mid-zone VI is indicated by *O. centrocarpum* and *B. tepikiense*, but the absence of cysts of *P. dalei* suggests somewhat colder conditions than in MIS 5a.

Discussion

The onset of the penultimate deglaciation is characterised by a pronounced change in sediment source areas and hydrographic conditions at the northern Barents margin. The disintegration of the Svalbard–Barents Sea ice sheet (SBIS) during Termination II is reflected in a peak supply of IRD and a subsequent meltwater event in the Yermak Plateau area (Figs 3 and 4; Pagels, 1991; Nowaczyk *et al.*, 1994; Knies *et al.*, 1999, 2000, in press; Vogt *et al.*, in press). The change in clay mineral composition indicates a shift in the source area of fine-grained sediments from the Barents Sea, where kaolinite-rich Mesozoic bedrock probably was eroded by glacial activity during MIS 6, to local illite-rich input from the Svalbard archipelago (Fig. 3; Knies *et al.*, 2000; Vogt *et al.*, in press).

The distinct decrease in calcite contents at the MIS 6–5 boundary is caused by an intensified syn- and post-depositional dissolution of calcareous planktonic and benthic microfossils, which reflects an enhanced advection of Atlantic waters and an increased surface-water production at the northern Barents Sea margin (for details see Knies *et al.*, 1999; Wollenburg *et al.*, 2001). The organic-walled dinocyst assemblages are not affected by degradation of organic matter associated with carbonate dissolution. Notably, the presence in most samples of non-fluorescent pigmented cysts of the genera *Brigantedinium* and *Islandinium*, which are more susceptible to dissolution than other taxa (e.g. Dale, 1976; Zonneveld *et al.*, 1997), suggests a negligible effect of organic matter degradation.

Marine palaeoenvironment during the last interglacial

The increase in dinocyst concentrations and species adapted to relatively warm waters lags behind the shift in isotope records in early MIS 5. This lag indicates that ocean climate ameliorated after a cold period associated with meltwater discharge. The distinct supply of cold fresh water may have suppressed surface water warming until the SBIS largely disappeared. Full interglacial conditions may have been established during the interval free of IRD from 530 to 514 cm core depth (Fig. 4; cf. Bauch *et al.*, 1996). The recurrence of IRD, associated with a distinct decrease in dinocyst concentrations and a decline of *O. centrocarpum* marks the termination of the last interglacial (Fig. 4). This event probably reflects the advance of the SBIS corresponding to the initiation of glaciation in northern Svalbard (cf. Knies *et al.*, in press).

A similar offset between an increase of subpolar to temperate planktonic microfossils and a shift to light $\delta^{18}\text{O}$ values has been noted in a number of records from the adjacent Nordic Seas, thus suggesting that early MIS 5 was influenced by a substantial meltwater discharge and cold regional conditions (e.g. Kellogg *et al.*, 1978; Haake and Pflaumann, 1989; Gard and Backman, 1990; Fronval and Jansen, 1997; Bauch *et al.*, 2000). Similarly, the pollen record from Fjøsanger (western Norway) reflects colder conditions at the onset of the Eemian, which is succeeded by a pronounced warming culminating in a distinct thermal optimum in the late part of the last interglacial (Mangerud *et al.*, 1981). This optimum occurred simultaneously with a warming of coastal waters indicated by molluscs, benthic foraminifers and dinocysts (Mangerud *et al.*, 1981; Dale, 1996).

Although MIS 5e generally correlates with the Eemian (e.g. Mangerud *et al.*, 1981, 1998), a detailed comparison of pollen and isotope data from a sediment core off Portugal has revealed a more complex response of proxies to climate change (Sánchez-Goni *et al.*, 1999). The peak warming on land and in surface waters was shown to correspond to a plateau of light oxygen isotope values in MIS 5e and a substantial part of MIS 5d. The clear offset between a change to heavy $\delta^{18}\text{O}$ values, and a decrease in dinocyst concentrations and decline of *O. centrocarpum* in core PS2138 suggests a similar situation (Fig. 4). However, an independent correlation of the PS2138 record to terrestrial interglacial sequences determined by pollen analysis is not possible.

The palaeoenvironmental conditions during the last interglacial were stable within the resolution of the dinocyst record (ca. 3000 yr). A possible environmental variability on submillennial time-scales cannot be resolved by the PS2138 record, but has been shown in palaeoclimate records of MIS 5e from the Nordic Seas and southeastern Siberia (e.g. Fronval and Jansen, 1997; Karabanov *et al.*, 2000).

The peak abundance of benthic foraminifers *Epistominella pusilla*, *Pullenia bulloides* and *Discorbinella bertheloti*, which are considered to be 'Atlantic species' (Wollenburg *et al.*, 2001) supports our conclusion that full interglacial conditions were established in early MIS 5 (Fig. 4). Based on foraminiferal assemblages, Wollenburg *et al.* (2001) estimated considerably higher palaeoproductivities than today for the same interval, consistent with maximum accumulation rates of dinocysts, especially *O. centrocarpum* (Fig. 5). The foraminiferal assemblages include high amounts of the 'phytodetritus' species *Epistominella pusilla*, *Islandiella helenae* and *I. norcrossi* (for discussion see Wollenburg *et al.*, 2001), thus suggesting a higher flux of phytoplankton to the sea floor. Accordingly, peak accumulation rates of dinocysts may result from increased production of cysts in surface waters, indicating optimal environmental conditions for dinoflagellates during the interglacial.

A comparison of the present and last marine interglacials is hampered by the low resolution of late MIS 1 in core PS2138, which is either compressed or was lost during coring, and the absence of any age-fix points during the last 12 000 yr (Fig. 2; Table 1). Recent dinocyst assemblages on the continental slope near the core site are comparable to those of the last interglacial, but differ in lower abundances of *O. centrocarpum* and higher abundances of *Brigantedinium* spp. and *I. pallidum*, indicating somewhat cooler conditions (Mudie, 1992). Therefore, sea-surface temperatures during the last interglacial may have been even warmer than today.

This interpretation agrees with mollusc and benthic foraminiferal data from Eemian coastal deposits of Svalbard, which indicate conditions warmer than present, but possibly cooler than during the Holocene climatic optimum (Miller *et al.*, 1989; Mangerud *et al.*, 1998; Bergsten *et al.*, 1998).

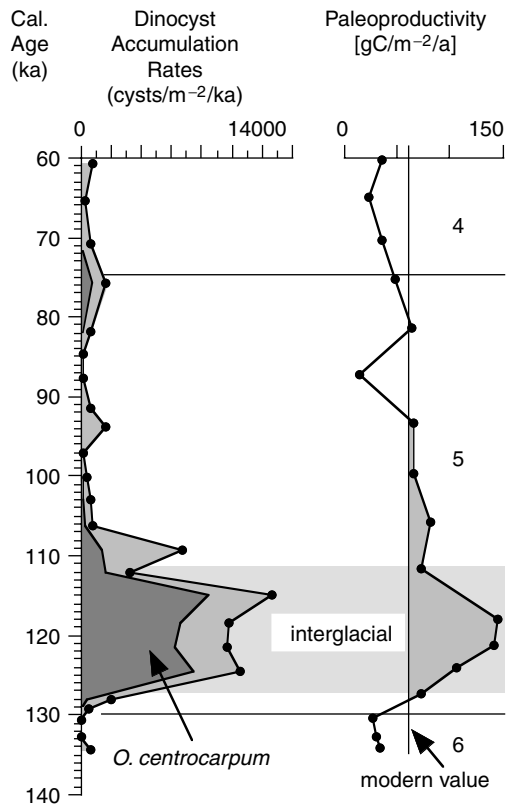


Figure 5 Dinocyst accumulation rates and paleoproductivity estimated from benthic foraminiferal assemblages (Wollenburg *et al.*, 2001) in MIS 4 to 6

The last interglacial may have been a period of warmer conditions than today and of reduced sea-ice cover in the entire circum-arctic shelf regions (Hamilton and Brigham-Grette, 1991; Brigham-Grette and Hopkins, 1995; Funder *et al.*, 1998, submitted; Kelly *et al.*, 1999).

A stronger advection of Atlantic water into the Barents Sea and further east along the Siberian coast occurred during the last interglacial (e.g. Velichko *et al.*, 1991; Mangerud *et al.*, 1998, 1999; Larsen *et al.*, 1999). At the northern Barents Sea margin, the advection of Atlantic waters was slightly variable during the time interval from MIS 6 to 4 (Pagels, 1991; Knies *et al.*, 1999). Nevertheless, an increased input of marine organic matter during MIS 5e and 5a suggests an advection stronger than today and seasonally open water conditions (Knies *et al.*, 2000).

Evidence of a relatively warm last marine interglacial at the northern Barents Sea margin contradicts the data on carbonate and planktonic foraminifer and coccolith abundances from Fram Strait, which suggest a less intense Atlantic water inflow than in the Holocene (Gard and Backman, 1990; Köhler and Spielhagen, 1990; Hebbeln and Wefer, 1997; Bauch *et al.*, 1999). In particular, lower carbonate and coccolith accumulation rates in MIS 5e than in MIS 5a and in the Holocene indicate a lower production and colder surface waters (Hebbeln and Wefer, 1997).

These cores from the central and western Fram Strait, however, may not record the conditions in the main pathway of the West Spitsbergen Current along the Svalbard continental margin. Alternatively, low biogenic carbonate contents may be explained by abiotic factors such as variable dissolution and/or dilution with terrigenous sediments (e.g. Köhler and Spielhagen, 1990; Pagels, 1991; Köhler, 1992; Hebbeln and Wefer, 1997; Knies *et al.*, 1999; Vogt *et al.*, in press; Wollenburg *et al.*, 2001). The inverse relationship of carbonate

contents in core PS2138 with interglacial conditions has been attributed to intensive dissolution during MIS 5 to 3 and MIS 1 (for discussion see Knies *et al.*, 1999; Wollenburg *et al.*, 2001). Thus, the low carbonate contents in MIS 5 consists almost entirely of dolomite (Fig. 3). Carbonate dissolution may have been particularly strong during interglacials, because sea-ice formation on shelves may lead to dense, CO₂ rich bottom waters descending into the basins (e.g. Pagels, 1991; Steinsund and Hald, 1994; Knies *et al.*, 1999; Volkman, 2000). Moreover, an increased supply of organic matter, resulting from enhanced production in the photic zone, may lead to higher CO₂ release in bottom waters by a bacterial degradation of organic matter (Huber *et al.*, 2000; Wollenburg *et al.*, 2001). Therefore, accumulation rates of calcareous microfossils should be used as palaeoproductivity indicators with considerable care (Wollenburg *et al.*, 2001). Thus, in core PS2138 carbonate contents are low throughout MIS 5, whereas dinocyst assemblages show distinct fluctuations. We want to note that despite the signature of total carbonate compositions in the Fram Strait, the occurrence of certain coccolith assemblages indicates that the Atlantic water may have reached the Yermak Plateau in MIS 1, 5a and 5e (Gard and Backmann, 1990; Nowaczyk *et al.*, 1994).

Palaeoenvironmental conditions during middle MIS 5

Previous studies revealed relatively low surface water temperatures, a weak subsurface advection of Atlantic water, and a perennial sea-ice cover in the Fram Strait during mid-MIS 5, corresponding approximately to substages 5d–b (Gard and Backman, 1990; Köhler and Spielhagen, 1990; Hebbeln and Wefer, 1997). This interpretation is supported by abundant cold water dinocysts in zones III and IV, which indicate relatively stable cold conditions and a long duration of sea-ice cover after the termination of the last marine interglacial. Variable abundances of *O. centrocarpum*, *N. labyrinthus* and other taxa reflect fluctuations in the advection of relative warm (sub-) surface water and temporarily somewhat warmer conditions and reduced sea-ice cover when abundances of cold water taxa *A. minutum* s.l. and *I. pallidum* were low, e.g. in upper zone III. This climatic regime may have provided moisture for a progressive build-up of the SBIS that is documented by the onset of IRD input above zone II (Fig. 4).

The coldest period of MIS 5 was probably during upper zone IV in MIS 5b/c when assemblages were extremely dominated by *I. pallidum*, accompanied by *I. minutum* s.l. and *Brigantedinium* spp. This cold phase is terminated by peak abundances of *O. centrocarpum* and *N. labyrinthus*, suggesting a surface warming and enhanced Atlantic water inflow (cf. Knies *et al.*, 2000; Wollenburg *et al.*, 2001).

Palaeoenvironmental conditions in late MIS 5–4

The Fram Strait and eastern Arctic Ocean were characterised by moderate Atlantic water advection and seasonally ice-free conditions with low iceberg occurrences in substage 5a (Gard and Backman, 1990; Köhler and Spielhagen, 1990; Pagels, 1991; Hebbeln and Wefer, 1997; Nørgaard-Pedersen *et al.*, 1998; Vogt *et al.*, in press). Sejrup and Larsen (1991) and Bergsten *et al.* (1998) showed that sea-surface temperatures in Fram Strait may have been comparable in MIS 5e, 5a and late MIS 1. The inflow of Atlantic water may have been much

smaller in MIS 5a as compared with MIS 5e (Köhler and Spielhagen, 1990; Wollenburg *et al.*, 2001). In MIS 4, a dense sea-ice cover may have prevailed in the central Arctic Ocean (Nørgaard-Pedersen *et al.*, 1998).

The dinocyst assemblages generally conform with this interpretation, although the resolution in late MIS 5 and MIS 4 is too low to allow a detailed reconstruction. The strong decline in a number of warm adapted species, and the generally inverse relationship of *I. minutum* to *O. centrocarpum* and *B. tepikiense* in upper zone V and zone VI indicate variable surface water conditions in MIS 5a–4. Peak abundances of *O. centrocarpum* and *B. tepikiense* suggest some surface water warming, in particular during substage 5a. The temperature difference with the last interglacial may not have been very high, but the different composition of assemblages indicates a warming of surface water owing to the stronger stratification that was caused by meltwater supply. The continuous input of IRD in MIS 5a and light $\delta^{18}\text{O}$ and heavy $\delta^{13}\text{C}$ values in core PS2138 suggest a seasonal discharge of meltwater at the Barents Sea margin (Figs 3 and 4). A similar event in cores from Fram Strait at the MIS 4–5 boundary was attributed to a strong freshwater discharge from icebergs and/or adjacent land masses (Köhler and Spielhagen, 1990; Lloyd *et al.* 1996). The coeval increase of kaolinite and smectite contents in association with input of IRD at the MIS 5–4 transition reflect the successive advance of the SBIS to the shelf edge (Fig. 3; Knies *et al.*, 2000).

Conclusions

Dinocyst analysis of a sediment core from the northern Barents Sea continental margin (eastern Arctic Ocean) revealed clear evidence of interglacial conditions during early MIS 5. The last marine interglacial is marked by a distinct peak of dinocyst concentrations and assemblages dominated by *O. centrocarpum*. Simultaneously, the peak abundance of 'Atlantic water' benthic foraminifers indicates a strong advection of water masses from the Norwegian Sea (Wollenburg *et al.*, 2001). The composition of dinocyst assemblages suggests that sea-surface temperatures were similar to present or even warmer.

Dinocyst evidence for a relatively warm last interglacial contrasts with results of previous studies in the Fram Strait–Yermak Plateau area. Based on carbonate and calcareous microfossil records, the last marine interglacial in that area was interpreted as a period of colder climatic conditions than today or during the Holocene climatic optimum. Recent studies, however, showed that dissolution may have affected the carbonate records considerably, as suggested for example by the near absence of calcite in core PS2138 during the entire MIS 5 and 4. Moreover, the primary carbonate production may have been diluted episodically by a strong supply of terrigenous sediments.

Dinocyst assemblages indicate palaeoenvironmental conditions during MIS 5d to 4 that are comparable with previous reconstructions based on carbonate and calcareous microfossil records. A relatively stable cold period with minor intervals of relatively warm surface water prevailed in mid-MIS 5, approximately spanning substages 5d to 5b. A short-term warming occurred in substage 5a, when temperatures may have been similar to the last marine interglacial of the early MIS 5. The different composition of assemblages, however, suggests a stronger stratification of surface waters, probably owing to a larger meltwater supply.

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