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Contrasting ocean changes between the subpolar and polar North Atlantic during the past 135 ka

Henning A. Bauch,[1] Evguenia S. Kandiano,[2] and Jan P. Helmke[3]

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[1] Variations in the poleward-directed Atlantic heat transfer was investigated over the past 135 ka with special emphasis on the last and present interglacial climate development (Eemian and Holocene). Both interglacials exhibited very similar climatic oscillations during each preceding glacial terminations (deglacial TI and TII). Like TI, also TII has pronounced cold–warm–cold changes akin to events such as H1, Bølling/Allerød, and the Younger Dryas. But unlike TI, the cold events in TII were associated with intermittent southerly invasions of an Atlantic faunal component which underscores quite a different water mass evolution in the Nordic Seas. Within the Eemian interglaciation proper, peak warming intervals were antiphased between the Nordic Seas and North Atlantic. Moreover, inferred temperatures for the Nordic Seas were generally colder in the Eemian than in the Holocene, and vice versa for the North Atlantic. A reduced intensity of Atlantic Ocean heat transfer to the Arctic therefore characterized the Eemian, requiring a reassessment of the actual role of the ocean–atmosphere system behind interglacial, but also, glacial climate changes. **Citation:** Bauch, H. A., E. S. Kandiano, and J. P. Helmke (2012), Contrasting ocean changes between the subpolar and polar North Atlantic during the past 135 ka, *Geophys. Res. Lett.*, *39*, L11604, doi:10.1029/2012GL051800.

1. Introduction

[2] The Atlantic Meridional Overturning Circulation (AMOC) is an important part of the climate system transfering an enormous amount of heat via ocean currents and atmospheric circulation into the high-northern latitudes (Figure 1). For example, the North Atlantic warms Western Europe by several °C on average, making it much more habitable than it would be otherwise. In paleoceanographical and paleoclimatological studies, it has often been concluded that a disruption of that AMOC-system through high-latitude surface ocean freshening would dramatically change the climate of this region [*Rennermalm et al.*, 2007]. While this paradigm seems to hold for the glacial periods [*Bond et al.*, 1993; *Oppo and Lehman*, 1995], rapid and dramatic shifts in climate are far more muted during peak warm periods such as the Holocene [*North Greenland Ice Core Project Members*, 2004].

[3] Because of the increasing anthropogenic influence on the climate system, model projections predict temperatures to rise on a global scale, but the polar amplification hypothesis calls for even more warming in the Arctic, causing seasonally ice-free conditions [*Wang and Overland*, 2009]. Insight into oceanographic conditions in the Arctic and subarctic during past warm periods can provide a necessary test of these climate predictions. The last interglaciation (the Eemian or marine isotope stage 5e), centered around 125 ka, is such a period because model experiments often suggest higher-than-Holocene temperatures over most of Europe and the Arctic [*Kaspar et al.*, 2005; *Otto-Bliesner et al.*, 2006]. Temperatures over Greenland and the Arctic Ocean were apparently up to 5°C higher in the Eemian leaving a widely reduced Arctic sea-ice cover [*CAPE Last Interglacial Project Members*, 2006]. However, crucial aspects such as the involvement of the AMOC remains poorly known, as is the timing when exactly during the last interglacial cycle did the peak warmth occur in the Arctic.

2. Material, Methods, and Strategy

[4] To groundtruth the hypothesis of extreme Arctic-subarctic warmth during the Eemian interglacial cycle, we used two sediments cores from sites that monitor the flow of warm North Atlantic Drift water. Core M23414 is from the eastern edge of the Subpolar Gyre within the North Atlantic and core PS1243 from the eastern Polar Gyre near the Arctic Front of the southern Nordic seas (Figure 1). Both cores have glacial-interglacial records that extend back in time to about 500 ka with high quality chronology and proxy records of paleoceanographic changes in their respective region [*Bauch and Erlenkeuser*, 2003; *Helmke et al.*, 2002; *Kandiano and Bauch*, 2003].

[5] A suite of well-proven proxy methods were used to evaluate the paleoceanographic conditions over the last 135 ka, i.e., since the penultimate glacial maximum (PGM, Marine Oxygen Isotope Stage (MIS) 6), during Termination II and the peak interglacial MIS5e, and through the onset and the maximum of the last glacial period (LGM or MIS2), Termination I, and finally into the present warm period, the Holocene (MIS1). The proxies include quantification of iceberg-rafted detritus (IRD), and warm (mainly subpolar) vs. cold (polar) water planktic foraminiferal species used as specific water mass indicators and as a basis to estimate past surface ocean temperatures via the modern analog method (MAT) [*Prell*, 1985]). Oxygen isotope data of planktic foraminiferal species together with epibenthic oxygen and

[1]Akademie der Wisssenschaften und der Literatur Mainz, Helmholtz-Zentrum für Ozeanforschung, Kiel, Germany.
[2]Helmholtz-Zentrum für Ozeanforschung, Kiel, Germany.
[3]Institute for Advanced Sustainability Studies, Potsdam, Germany.

Corresponding author: H. A. Bauch, Akademie der Wisssenschaften und der Literatur, Mainz, Helmholtz-Zentrum für Ozeanforschung, Wischhofstrasse 1-3, D-24148 Kiel, Germany. (hbauch@geomar.de)

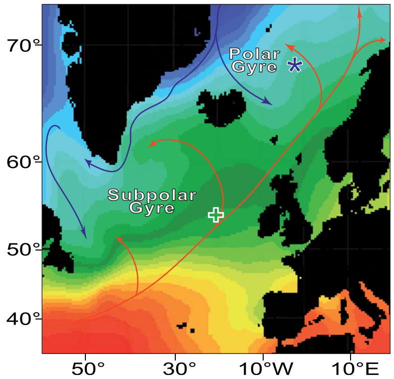

Figure 1. Overview of summer sea surface temperature in the North Atlantic (NCEP; Aug. 2010) showing the southern transfer of oceanic heat (in changeable colors) towards the polar Nordic Seas alongside with the general surface circulation pattern. Cross and asterisk mark the two studied core sites (M23414: 53°32N/20°17W, 2196 m water depth; PS1243: 69°22N/6°32W, 2710 m water depth).

carbon isotopes are used to monitor the temporal changes in surface and bottom water properties, and to guide a SPECMAP-based δ^{18}O stratigraphy.

[6] Both cores have sound age frameworks published elsewhere based on radiocarbon dates for the upper parts including the Holocene and the last glacial and stable isotope chronology for sediments deposited prior to ~30 ka; To reduce bias but to keep full insight into our interpretative basis we deliberately refrain from (re)producing an age model. Since the two sediment cores show good data comparability, we rather emphasize the unequivocal trends and events.

[7] With various proxy data we try to address a number of issues that are critical to the understanding of the glacial-interglacial ocean-climate interaction over the North Atlantic region. First, when in the past 135 ka did a tight meridional surface ocean connection occur between the subpolar and polar North Atlantic? Second, what were the conditions like during the terminal phases of the last two glacial periods and how did ocean changes progress during the subsequent interglacial climate ameliorations? Finally, to better evaluate our present climatic status in the context of a natural versus man-influenced climate change, we allude to the Holocene (i.e., the "pre-industrial" part of it) and how it compares to the previous interglaciation in terms of ocean heat transfer towards the Arctic.

3. Marine Isotope Stage 5 and the Holocene in the North Atlantic

[8] Proxy data from core M23414 (Figure 2) demonstrate major oceanic changes since the PGM (~135 ka). The peak of MIS5e is a prominent feature in all δ^{18}O records, and in the faunal and sedimentological data. The upper age boundary of MIS5e is bounded by the isotopic event 5.4 (~111 ka). The glacial inception after MIS5e occurred in accord with an increase in δ^{18}O, decreasing SSTs, and recurrence of IRD before 110 ka. The remainder of the last glaciation was punctuated by a series of well-known major and minor ice-rafting events. These events not only cooled the surface ocean due to meltwater release, but they also had a profound impact on the deep-water circulation [*Hodell et al.*, 2010; *Oppo and Lehman*, 1995; *Seidov and Maslin*, 1999] as verified in benthic δ^{13}C (Figure 2). Two important marker horizons are the iceberg surge events H11 and H1 each of which occurred during early deglaciation after the PGM and the LGM, respectively. By comparison, H11 is the more prominent of the two because of its association with a strong depletion in bottom water δ^{13}C. In M23414, H11 also has a precursor event with decreased δ^{13}C as well as increased abundance of IRD and the polar water foraminiferal species *Neogloboquadrina pachyderma* sinistral (*NP*s). Within the deglacial progression after H11, there is some further variability recognizable in the highly-resolved sediment reflectance data (410 cm core depth).

[9] Due to several intervening warm Greenland Interstadials (GI), the recognition of the classical marine δ^{18}O-based substructure of MIS 5 (MIS 5a–5e) is difficult to ascertain in Greenland ice. For instance, it is by no means straightforward to decide if just GI 21 or also parts of GIs 19–20 correlate to MIS5a (Figure 2). Similar difficulties arise when judging both the planktic and benthic δ^{18}O records. A clearer picture of the MIS5 structure, however, emerges when evaluating the δ^{18}O and faunal SST records against the IRD and sediment reflectance: (1) MIS5 was warmest in MIS5e, and second warmest in MIS5a (two events), whereas in between SSTs were substantially colder; (2) The warm peak of MIS5e is divided, the earlier part right after the deglaciation of MIS6 being the warmest. In fact, SSTs within MIS5e were warmer (by ~2–3° C) than at any time in the Holocene at this location [*Bauch and Kandiano*, 2007]. This is corroborated by the planktic δ^{18}O, in which both species show considerably lighter values in MIS5e than in MIS1 (~1‰ in *NPs*; ~0.2‰ in *GB*). Considering that *NPs* likely reflects a sporadic invasion of polar waters at the subsurface, perhaps due to enhanced westerly winds, the 1‰ difference in δ^{18}O of *NPs* would imply that this species was bathed (i.e., grew) in relatively warm polar waters during MIS5e. That the heavy Holocene δ^{18}O values were brought about by more saline waters (i.e., caused by evaporation from a warmer water) can be ruled out because of the low SSTs.

4. Marine Isotope Stage 5 and the Holocene in the Nordic Seas

[10] In core PS1243 from the Nordic Seas, the deglaciation of the PGM and the LGM is marked by pronounced variability in all proxies. Supported by radiocarbon dates, the upper 80 cm comprise the time since 30 ka [*Bauch et al.*, 2001]. During the late glacial, the Bølling/Allerød (B/A) warming features a 40% decrease of *NP*s, largely due to the subpolar species *Turborotalita quinqueloba* (*TQ*) which is the main interglacial counterpart of *NP*s in this region. The peak sample of the B/A was recently radiocarbon-dated (KIA 42610: 11,720 yrs ± 55 BP), and yielded a calendar age of 13,157 yr when using a 400 yr marine reservoir age [*Fairbanks et al.*, 2005]. The Younger Dryas (YD) cold event that followed is characterized again by a return to glacial-like, polar conditions. During the first half of the Holocene subpolar abundance of nearly 80% indicates warmest

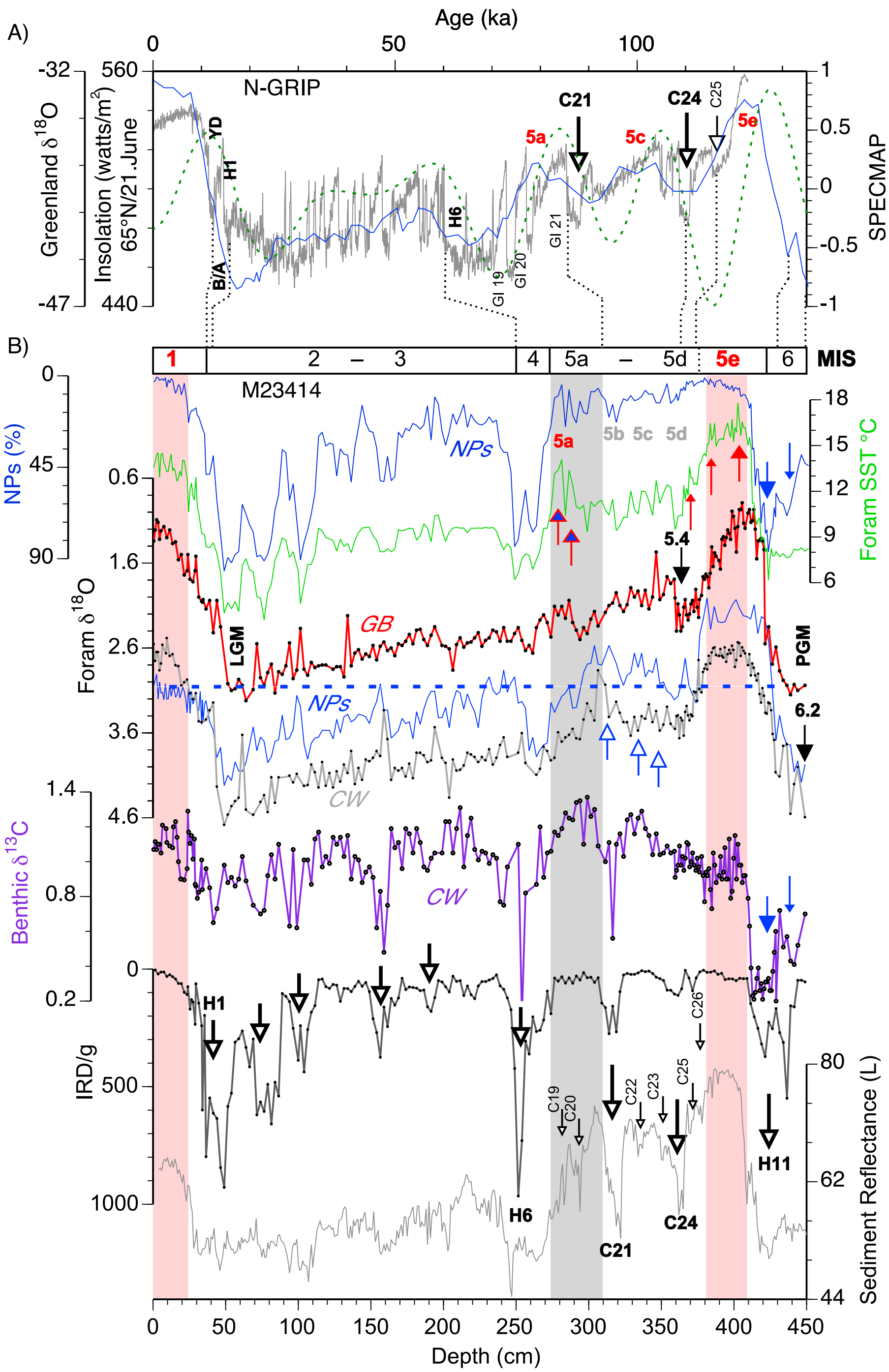

Figure 2. (a) Comparison of Greenland ice core data in grey [*North Greenland Ice Core Project Members*, 2004], SPECMAP chronology [*Martinson et al.*, 1987] in blue and with major isotope stages for MIS5e, and northern hemisphere insolation in green [*Laskar et al.*, 2004] for the last 135 ka with downcore proxy records and correlation scheme of core M23414 (Figure 2b). (b) Top to bottom: stratigraphic subdivision of M23414 showing oxygen isotope stage (MIS) boundaries and substages as well as the two conspicuous SPECMAP date events 5.4 and 6.2 (110.79 and 135.1 ka) which frame MIS5e; relative abundance of the polar foraminifer *Neogloboquadrina pachyderma* sin. (*NPs*); SST-estimates based on planktic foraminiferal assemblages; $\delta^{18}O$ records of planktic *Globigerina bulloides* (*GB*), *NPs*, and benthic *Cibicidoides wuellerstorfi* (*CW*); bottom water $\delta^{13}C$ of *CW*; iceberg-rafted debris (IRD) per gram sediment and >250 μm; sediment reflectance data (L); black-white arrows denote major and minor ice-rafting events (H, C) as described from other marine sediment records [*Chapman and Shackleton*, 1999]; blue, red-blue and red arrows denote some noteworthy cold-warm events between MIS6 and 4; LGM – Last glacial maximum; PGM – Penultimate Glacial Maximum; YD – Younger Dryas; B/A – Bølling/Allerød; GI – Greenland Interstadial.

conditions, before polar water masses again became increasingly prevalent at site PS1243 until the recent time.

[11] By comparison to the Holocene, the peak of MIS5e shows abundances of *NP*s that never fall short of 50%, implying relatively colder conditions. In contrast to that, but in accordance with M23414, the IRD-free interval of MIS5e has consistently lower planktic δ^{18}O values (by ~0.3‰). Despite some notable intra-interglacial variability – this has been described from the region before [*Fronval et al.*, 1998] – the upper part of MIS5e seems strongest influenced by warm Atlantic conditions because of few finds of temperate foraminiferal species (*Globorotalia truncatulinoides* and *G. scitula*). This idea gains further support from a peak abundance of the small-sized subpolar species *Globogerinita uvula* (*GU*) in this interval. Foraminiferal-based paleo-SST estimates relate to the species distribution in surface sediments within the size fraction >150 μm [*Pflaumann et al.*, 2003]. Being a small and trochospirally-shaped species, *GU* has no record in that data base at all. Although *GU*s' presence in the Holocene remains negligible at site PS1243, its main occurrence in MIS5e just prior to the glacial inception confirms other inferences of a late Eemian warming in the Nordic Seas [*Bauch and Erlenkeuser*, 2008; *Van Nieuwenhove et al.*, 2008; *Rasmussen et al.*, 2003] as well as a two-part interglacial peak [*Bauch et al.*, 2011].

[12] At site PS1243, TII and TI appear to follow a similar climatic pattern in showing comparable sequences of cold-warm events. Event H11 during early TII is, as in the North Atlantic, associated with a massive depletion in bottom water δ^{13}C. A second depletion, which followed upon a time of warming that had 40% of subpolar foraminifers (B/A-II) is thus akin in timing and character to a YD-style cold event which we call YD-II.

[13] Despite the overall similarity between TI and TII, there remains a significant difference. This is to do with the sudden but very widespread appearance of the rare, low-latitude foraminifer *Beella megastoma* (*BM*) in the Nordic Seas. As described previously [*Bauch et al.*, 2000], *BM* occurs during cold conditions, and in PS1243 together with the two major depletions in bottom water δ^{18}O. By analogy, rather low benthic δ^{18}O values are found already during the LGM (not in the PGM) but here in association with small-sized TQ providing good evidence for the advection of Atlantic waters into the central Nordic Seas at this time [*Bauch et al.*, 2001; *Kandiano and Bauch*, 2002]. The first major meltwater spike (time-coeval with H1) that followed upon the LGM has notably no record of *BM* – for various reasons the part just below the VA was repeatedly sampled, and each one yielded a single specimen of *BM.*

[14] Between the glacial inception after MIS5e and the LGM, the highly-resolved record of small-sized *TQ* shows a minor double-spike in MIS5a. This corroborates our SST-estimates from the North Atlantic of a weakened, but functioning, AMOC at this time. Prior to MIS5a, there are three notable spikes in the planktic δ^{18}O record. As no warming can be linked to them, these events appear to be meltwater-induced, and might correspond to C21-C23 in M23414. In addition, our record of small-sized *TQ* indicates a cold-warm variability before isotope event 5.4, which seems reflected also in the benthic δ^{18}O. It is intriguing that the associated IRD recurrence at this time (possibly C25) occurred at an abundance level of ~75% in *NP*s (>125 μm). Within the uppermost samples *NP*s again co-exists with IRD at a similar level.

5. Discussion

[15] The low SSTs in Nordic Seas during MIS5e reflected in consistently lowered subpolar foraminiferal abundance clearly contradict any notion of an anomalously warm subpolar North Atlantic. Furthermore, it casts doubt on the interpretations of Eemian records of small-sized *TQ* from North of Greenland [*Nørgaard-Pedersen et al.*, 2007]. Those finds, if stratigraphically correct, may be a local phenomenon like others in the modern Arctic [*Bauch*, 1999; *Volkmann*, 2000]. Hence, they can neither constitute evidence for enhanced AMOC nor for a supposedly, warmer and more ice-free Arctic as suggested [*CAPE Last Interglacial Project Members*, 2006]. Although small-sized *TQ* have a likely higher advectional potential than larger specimens [*Kandiano and Bauch*, 2002], we confirm their significance as to record AMOC intensity at the polar latitudes during times colder than peak interglaciations, such as in MIS5a.

[16] Lower-than-Holocene δ^{18}O and SSTs in the Nordic seas during MIS5e would imply lower salinities at the calcification depth of *NP*s, the latter being always below the mixed layer [*Simstich et al.*, 2003]. By contrast, although δ^{18}O of *GB* was also slightly lower in parts of MIS5e in the North Atlantic, the huge $\Delta\ \delta^{18}$O of 1.9‰ between *NP*s and *GB* indicates much more enhanced thermal stratification for the Holocene [*Mulitza et al.*, 1997; *Simstich*, 1999] and/or major variations in the seasonal calcification depth behavior between these two species [*Chapman*, 2010; *Jonkers et al.*, 2010]. The heavy Holocene values of *NP*s (~3.1 ‰; as in PS1243), heavier even when compared with bottom water δ^{18}O, however, indicate that *NP*s lived and calcified in polar waters of the Labrador Sea which later became advected to site M23414. The reason for the major spatio-temporal differences between the Eemian and Holocene climate at each site and between the two regions may have its origin in the different sizes and melting history of northern ice sheets in MIS6 and MIS2 which affected the surface ocean properties in the Nordic Seas in particular thereby changing both prevalent atmospheric circulation and meridional water mass transfer [*Bauch et al.*, 2011]. Despite some step-like features in various TII-records [*Sarnthein and Tiedemann*, 1990; *Siddall et al.*, 2006; *Cheng et al.*, 2009], a widespread YD-style cooling is apparently missing from TII [*Carlson*, 2008]. Our recognition of a TI-style deglaciation in the central Nordic Seas (with both YD-II and B/A-II) thus emphasizes the sensitivity of the polar region to temperature changes. The apparent lack of pronounced climatic perturbations elsewhere [*Carlson*, 2008] is masked, perhaps, by the rapidity of environmental changes and its specific timing relative to the global rises in sea-level and temperature .

[17] Although TI started at ~20 ka, air temperatures over Greenland increased in parallel with northern summer insolation (Figure 2), i.e., before the LGM [*Alley et al.*, 2002]. At face values, the increased appearance of small-sized *TQ* in the Nordic Seas around 25 ka implies initiation of modern AMOC already in early MIS2 (Figure 3). An intensified heat transfer from the lower latitudes should have led to enhanced moisture supply via the atmosphere and hence ice growth in the cold Polar North which then resulted in the LGM. New data now indicate that bottom overflow water from

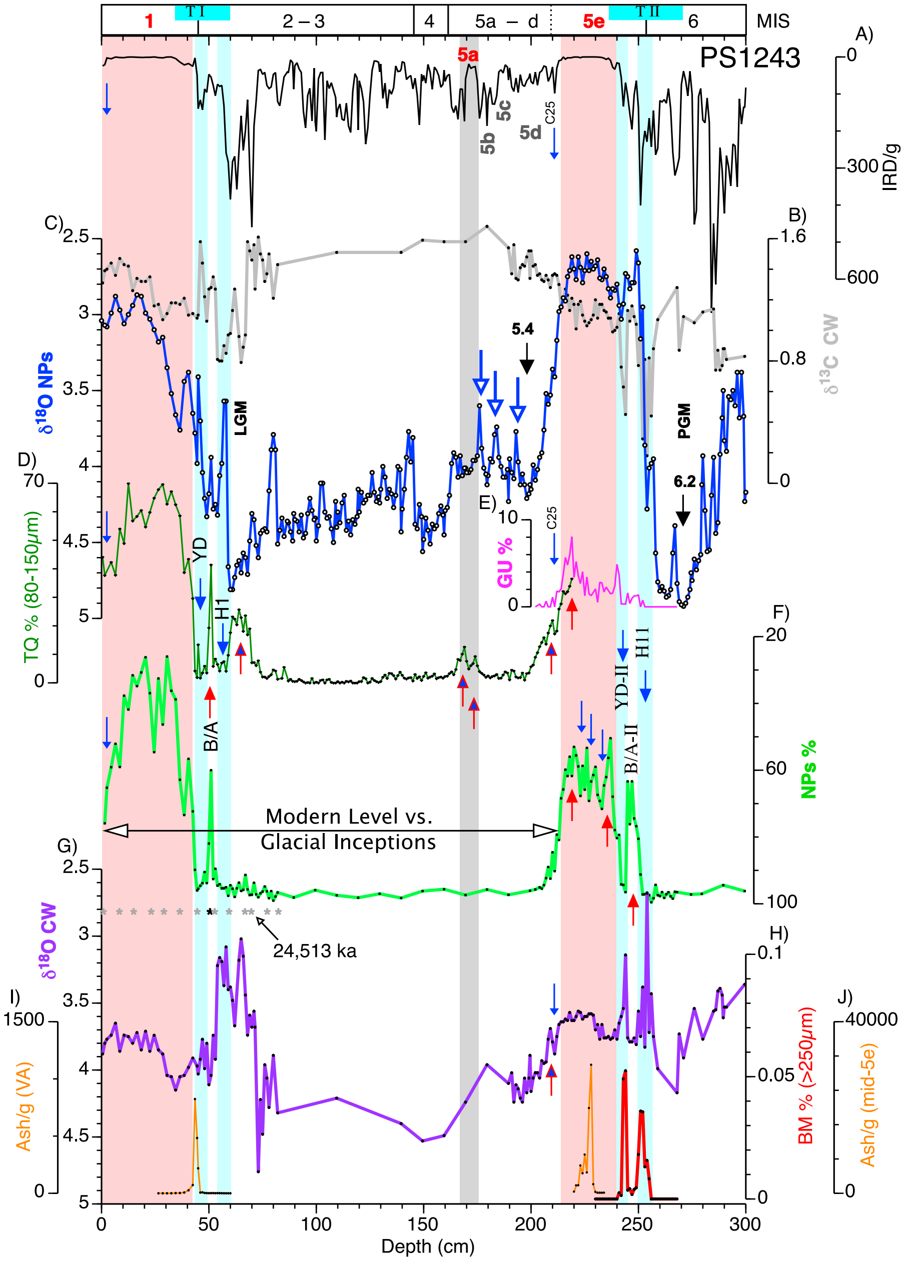

Figure 3. Proxy records and general stratigraphic overview from site PS1243 in the central Nordic Seas since MIS 6. (a) Number of iceberg-rafted debris per gram sediment and >250 μm; (b) bottom water δ^{13}C of the species *C. wuellerstorfi* (*CW*); (c) planktic δ^{18}O record of polar species *N. pachyderma* sin. (*NPs*); (d) relative abundance of the small-sized, subpolar, planktic foraminifer *Turborotalita quinqueloba* (TQ); (e) relative abundance of the subpolar, planktic foraminifer *Globogerinita uvula* (*GU*); (f) relative abundance of larger-sized TQ; (g) Bottom water δ^{18}O of the species *C. wuellerstorfi* (*CW*); (i) relative abundance of the large-sized, subtropical planktic foraminifer *Beella megastoma* (*BM*); (j) rhyolithic grains of the Vedde Ash (VA) in the YDand the mid-5e rhyolithic tephra for stratigraphic support [*Rasmussen et al.*, 2003]. Glacial terminations (TI and TII) and cold-warm events within are marked in light blue. Please note and compare with Figure 2 the 2-spiked MIS5a warming and the three melt-water spikes between MIS5d to b (blue-white arrows); asterisks denote positions of AMS C^{14} dates.

the Nordic Seas started to increase from 24 ka on [*Hodell et al.*, 2010] thus supporting our *TQ*-results and other interpretations of a seasonally ice-free central Nordic Seas with enhanced water mass overturn during the LGM [*Weinelt et al.*, 1996].

[18] Although no small-size *TQ*-record is available to interpret AMOC changes in late MIS6, comparison of the benthic and planktic δ^{18}O in PS1243 would indicate quite different water mass conditions and structures between the PGM and the LGM as well as each following termination. In particular the low planktic δ^{18}O values in B/A-II seem more influenced by higher SSTs than those in B/A, and that the YD-II occurred comparatively late during TII. Altogether this may be the combined effect of the 'unusual' water mass changes, as indicated by our *BM*-record, and a relatively higher sea-level at this time [*Thomas et al.*, 2009]. The occurrence of *Beella* in the Nordic Seas during TII remains enigmatic. In terms of its modern biogeography and in spite of

its general rareness, the fact that this subtropical, deep-dwelling species [*Hemleben et al.*, 1989] co-occurred with strong depletions in benthic $\delta^{13}C/\delta^{18}O$ during deglacial cold events calls for a very specific type of meridional water mass transfer and surface ocean structure. Inflow of relatively warm but also salty Atlantic water, probably at the subsurface below a thick overlying halocline, may have been partly responsible for the strong depletions in the benthic $\delta^{18}O$ in TII [*Bauch et al.*, 2000].

6. Conclusions

[19] A comparison of paleorecords from sites in the polar Nordic Seas and the subpolar North Atlantic (since ~135 ka), which includes the present and last interglacial climate cycle, suggests a warmer surface ocean in the North during the Holocene than in Eemian times. The reconstructed colder last interglacial Nordic Seas - and a brief but much colder warming later during MIS5 (5a) - contradicts modelling efforts on sea-ice and other reconstructions that would opt for the Eemian as past analogue of an Atlantic-influenced polar amplification of present-day global warming in the Arctic. Instead, our proxies reveal some major discrepancies in the polar ocean during glacial terminations I and II. While there is a good agreement in alternating cold-warm events, TII in particular experienced intrusions of Atlantic waters at the surbsurface which also left an imprint on the bottom water $\delta^{18}O$ in the Nordic Seas during times of massive deglacial meltwater discharges. Because deglacial processes during TII lasted well into the Eemian at the high latitudes, the influence of Atlantic ocean heat was effectively reduced at the surface due to enhanced freshening and seasonal sea ice. The recognition of a late warm peak in the Eemian rather than early as in the Holocene is thus expression of a profoundly different postglacial reorganization of the ocean–atmosphere system over the polar North Atlantic.

[20] **Acknowledgments.** The Editor thanks two anonymous reviewers for assisting in the evaluation of this paper.

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