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North Pacific response to millennial-scale changes in ocean circulation over the last 60 kyr

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Abstract. Foraminifera-based sea surface temperature estimates in the northwest Pacific (Ocean Drilling Program Site 883; 51°N) varied by 4°C on millennial timescales over the last 60,000 calendar (cal) years, with the most prominent amplitudes during marine isotope stage 3. Age control is based on benthic $\delta^{18}O$ records, ^{14}C ages, and on geomagnetic intensity variations at Site 883 tuned to the North Atlantic paleointensity stack since 75 ka (NAPIS-75), in turn, tied to the Greenland Ice Sheet Project 2 (GISP2) ice core chronology. On the basis of this tuning, northwest Pacific warm phases parallel Greenland cold stadials and viceversa. This contrasts with atmospheric heat transfer, expected to produce quasi-coeval signals across the Northern Hemisphere. The antiphasing may instead stem from variations in global thermohaline circulation. At its North Pacific terminus in the subarctic gyre, any slowdown or shutoff of North Atlantic Deep Water formation during Dansgaard-Oeschger stadials led to a turnoff or reduction of the upwelling of cold Pacific deepwater and thus, to striking, instantaneous, short-term warmings of surface water.

1. Introduction

The thermohaline circulation (THC) of the Atlantic and the global ocean largely follows three different basic modes: (1) the modern, Holocene mode, (2) the peak glacial mode, which is similar to mode 1, but differs by strongly reduced North Atlantic Deep Water (NADW) formation, and (3) the Heinrich meltwater mode [*Sarnthein* et al., 1994, 2001]. The latter mode entirely differs from modes 1 and 2 in showing a reversed Atlantic THC pattern. *Van Kreveld* et al. [2000] recently demonstrated that the short-term reversals from THC modes 1 and/or 2 to mode 3 along with the abrupt transitions to Dansgaard-Oeschger (D-O) stadials may have been initiated by cyclic glacier surges and meltwater injections from the east Greenland ice sheet into the East Greenland Current, which is crucial in controlling the density of surface water in the major centers of NADW formation in the northwest Atlantic [*Dickson*, 1997]. This finding indeed comes close to the concepts suggested by various ice and ocean models [*MacAyeal*, 1993; *Paillard and Labeyrie*, 1994; *Rahmstorf*, 1995].

On the other side of the global ocean circulation system the northwest Pacific represents a sort of "far end" of the global oceanic deepwater flow, which starts from the North Atlantic and is "recycled" in the Weddell Sea [*Gordon*, 1986, 1991; *Broecker*, 1991]. This terminal position is inferred from the maximum modern ^{14}C ages >1000 years [*Ostlund and Stuiver*, 1980; *Stuiver and Braziunas*, 1993] and high nutrient contents found in surface water near the North Pacific margin [*Reid*, 1962, 1965; *Craig* et al., 1981], coupled to extremely high (>2150 years) ^{14}C ages and nutrient contents of deep water [*Stuiver* et al., 1983; *Broecker and Peng*, 1982; *Broecker* et al., 1988].

The outlined unique terminal setting of the North Pacific is still under debate [*Toggweiler and Samuels*, 1993]. However, any real terminal setting should necessarily lead to some regional-scale upwelling of North Pacific Deep Water (NPDW) along the Northern Pacific margin [*Roemmich and McCallister*, 1989; *Gordon*, 1991; *Schmitz*, 1995], in particular, in the northwesternmost Pacific. Here NPDW upwells along the center of the subarctic cyclonic gyre. This upwelling is clearly reflected by an outstanding enrichment in phosphate-phosphorus in the more coolish surface water, up to a level only characteristic of modern circum-Antarctic waters [*Reid*, 1962, 1965; *Dodimead et al.*, 1963; *Craig et al.*, 1981]. It is a major objective of this study to trace how this upwelling cell may have varied over glacial-to-interglacial times and how these variations were possibly linked to past changes in global THC.

The present study is based on sediment cores from Ocean Drilling Program (ODP) Site 883 (51°N, 168°E; 2385 m water depth) which was drilled at the Detroit Seamount right below the center of the northwest Pacific upwelling cell (Figure 1; *Rea et al.*, [1993]). Here Kotilainen and Shackleton [1995] first detected millennial-scale fluctuations in gamma ray attenuation porosity evaluator (GRAPE) density, which they ascribed to a differential input of ice-rafted debris (IRD) over marine isotope stages (MIS) 5-2 stadial and interstadial

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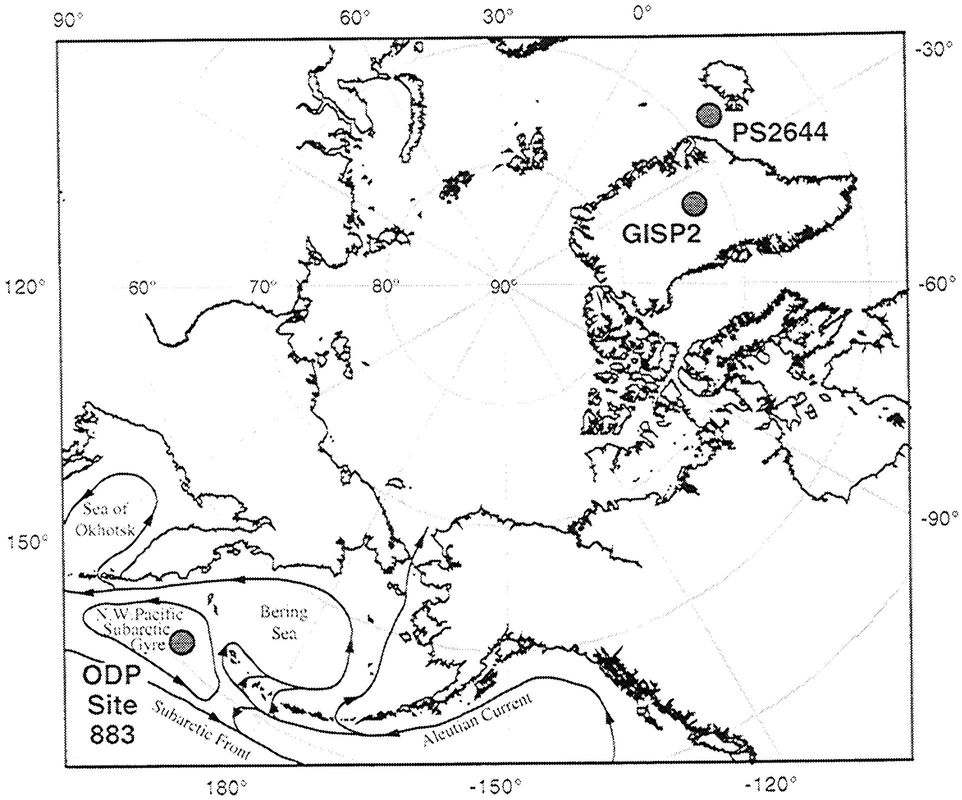

Figure 1. Location of ODP Site 883, marine sediment core PS2644, and Greenland ice core GISP2. Arrows indicate surface currents of the North Pacific subarctic gyre [arrows modified from *Dodimead et al.*, 1963].

events, when tentatively correlated to the Greenland temperature record [*Dansgaard et al.*, 1993]. The variations in IRD supply were attributed to a joint atmospheric forcing of Northern Hemisphere climate, which also was simulated in the coupled models of *Schiller et al.* [1997] and *Mikolajewicz et al.* [1997]. However, no sufficently detailed basis for a rigorous correlation of millennial-scale events from the North Pacific to the North Atlantic was available.

Starting from an initial, low-resolution benthic record of *Keigwin* [1995], our new isotopic, geochemical, and faunal data may provide far more detailed evidence of ocean variability. The new data set establishes a first multicentennial to bimillennial resolution record of the response of northwest Pacific deepwater (T. Kiefer et al., manuscript in preparation, 2001) and surface water properties and, especially, of sea surface temperatures (SST) to the long- and short-term variations in Atlantic and global THC over the last 60 kyr, i.e., to the D-O cycles of MIS 3-1. Moreover, a detailed stratigraphy is established for MIS 4 back to 75 ka. Interoceanic tuning of geomagnetic records is tested as a novel basis for high-quality precision in interhemispheric age correlation. The emerging seesaw variability of THC changes in the northwest Pacific and North Atlantic enables us to approach a further intriguing aspect in this complex but coupled ocean-atmosphere system, the feedback of changes in North Pacific upwelling and global THC on SST. These SST variations may finally have contributed to enigmatic D-O-scale oscillations of northwest Pacific climate.

2. Methods

Stable oxygen isotopes were measured on (frequently duplicate) samples of planktonic *Neogloboquadrina pachyderma* sinistral coiling (12-46 specimens) in the 150-250 μm size fraction, cracked and ultrasonically cleansed in ethanol. The samples were processed in the Carbo Kiel automated carbonate preparation device, linked on-line to a Finnigan MAT 251 mass spectrometer. The isotope values were calibrated to the Peedee Belemnite (PDB) scale using the NBS20 isotope standard. Solnhofen Plattenkalk and NBS19 served as secondary standards. External analytical error is less than 0.08‰ for $\delta^{18}O$.

Two accelerator mass spectrometry (AMS) ^{14}C ages of *Morley et al.* [1995] were supplemented by a further 16 ^{14}C datings (Table 1) of *N. pachyderma* (s.) and 2 datings of mixed benthic foraminifera, measured by accelerator mass spectrometry at the Leibniz Laboratory of Kiel University [*Nadeau et al.*, 1997]. Most datings were obtained from species abundance maxima (Figure 2). The dating at 0.81 meters below the seafloor (mbsf) [*Morley et al.*,1995], which was obtained from an abundance minimum, was ignored because it is probably biased toward a too young value by downcore bioturbational mixing (Figure 2). Carbon 14 ages were calculated as conventional ^{14}C years [*Stuiver & Polach*, 1977] and were reduced by 800 years for the local ocean reservoir effect [*Southon* et al., 1990; *Ingram and Kennett*, 1995]. However, 800 years may be a conservative estimate for the reservoir effect. Moreover, it may have strongly varied over time [*Southon* et al., 1990]

Table 1. AMS ^{14}C Ages Measured From ODP Site 883[a].

Laboratory Number	Depth, mbsf	Material Dated	Weight, mg	Raw ^{14}C Age, years	Reservoir Age, years	Reservoir-Corrected ^{14}C Age, years	$+1\sigma$ SE, years	-1σ SE years	Calendar Age, ka
				Accelerator Mass Spectrometry Dates					
KIA8200	0.03	*N. pachyderma* s.	6.0	5515	800	4715	35	35	5.5
KIA8201	0.145	*N. pachyderma* s.	9.5	7785	800	6985	40	40	7.8
KIA8202	0.24	*N. pachyderma* s.	8.3	9115	800	8315	45	45	9.3
KIA8203	0.44	*N. pachyderma* s.	7.8	10,610	800	9810	50	50	11.2
KIA8748	0.44	mixed benthics	2.4	11,890	2100	9790	80	80	
KIA8204	0.51	*N. pachyderma* s.	7.3	12,790	800	11,990	60	60	14.0
KIA8749	0.51	mixed benthics	4.3	13,420	2100	11,320	90	90	
KIA8750	0.645	*N. pachyderma* s.	10.1	13,490	800	12,690	70	70	15.3
LLNL	0.81	*N. pachyderma* s.		13,630	800	12,830	80	80	
KIA7161	1.07	*N. pachyderma* s.	9.4	17,090	800	16,290	90	90	19.4
LLNL	1.41	*N. pachyderma* s.		20,330	800	19,530	160	160	23.2
KIA7162	1.44	*N. pachyderma* s.	3.9	20,610	800	19,810	160	150	23.5
KIA8419	1.77	*N. pachyderma* s.	5.2	23,130	800	22,330	150	150	~25.5
KIA9356	1.78	*N. pachyderma* s.	6.6	24,040	800	23,240	170	170	~26.5
KIA8420	2.27	*N. pachyderma* s.	10.2	32,360	800	31,560	430	410	37.5[b]
KIA7163	2.44	*N. pachyderma* s.	6.0	39,230	800	38,430	960	860	39.6[b]
KIA9671	2.53	*N. pachyderma* s.	10.0	36,980	800	36,180	710	650	40.7[b]
KIA9672	2.71	*N. pachyderma* s.	10.0	39,220	800	38,420	400	380	44.5[b]
KIA9673	2.74	*N. pachyderma* s.	9.8	35,780	800	34,980	270	260	45.1[b]
				Apparent background ages					
KIA9355	7.49	*N. pachyderma* s.	9.9	60,250	—	—	2000	1600	
KIA9901	7.49	*N. pachyderma* s.	10.0	57,390	—	—	900	810	

[a]Samples with KIA numbers were measured at the Leibniz Laboratory in Kiel; LLNL, Lawrence Livermore National Laboratory [*Morley et al.*, 1995]. Two site-specific apparent background ages were determined from a sample from termination II.
[b]Derived from correlation of geomagnetic paleointensity records (Figure 2).

Planktonic foraminifera species were counted according to *Pflaumann et al.* [1996] in the >150 µm fraction, using sample splits of 250-1135 specimens, however, in some cases only 18-186 specimens were available (marked by larger error bars on curve e in Figure 3). Lithic particles >150 µm were also counted and ascribed to a pyroclastic or ice-rafted origin. Sea surface temperatures (SST) were estimated from the downcore foraminifera assemblages with the SIMMAX modern analog technique [*Pflaumann et al.*, 1996]. Tentatively, we used an enlarged reference database of 977 core top assemblages from the Atlantic (86°N to 57°S; U. Pflaumann, unpublished data, 2000), because no appropriate sediment surface data set is available from the high-latitude Pacific.

Concentrations of total organic carbon (TOC) were measured by gas chromatography with a CHN-O elemental analyzer (Carlo Erba Instruments) on 10 mg of freeze-dried and crushed sediment. For TOC analysis, inorganic carbon was removed with 17% H_3PO_4. TOC values are reproduced within ±0.02 wt % (1 sigma, $n = 103$).

3. Age Control

3.1. General Stratigraphic Framework

The interhemispheric comparison (Figure 1) of paleoceanographic events on millennial timescales, proposed in this paper, requires a precision in stratigraphic correlation hitherto unprecedented. On the basis of the general framework of benthic $\delta^{18}O$ stratigraphy the top 5 m of sediment at ODP Site 883 were deposited during MIS 1-5.1 (Figure 2). The stage 5-4 boundary was defined at 4.70 mbsf, and the stage 3-2 boundary was defined near 1.80 mbsf. The terminations of glacial stages 4 and 2 start near 3.6 and 0.8 mbsf, respectively. At 0.5 mbsf, termination Ib [*Duplessy et al.*, 1981] is reflected by an abrupt $\delta^{18}O$ decrease by 0.6‰, with a Bølling-Younger-Dryas-style $\delta^{18}O$ plateau below and a $\delta^{18}O$ "shoulder" on top, characteristic of the Preboreal at the onset of the Holocene $\delta^{18}O$ plateau.

By and large, this stratigraphy is supported by the planktonic $\delta^{18}O$ record and ^{14}C dating (Figure 2 and Table 1). However, it is difficult to strictly define the actual core depth of several $\delta^{18}O$ stratigraphic events such as the stage 3-2 boundary, which shifts from 1.8 to 1.95 mbsf on the basis of both the planktic $\delta^{18}O$ curve and ^{14}C dating. Major problems of any high-resolution $\delta^{18}O$ stratigraphy in the North Pacific are as follows: (1) The global planktic $\delta^{18}O$ signal may be biased by significant regional SST and salinity effects, induced by variations in upwelling intensity, precipitation, and the low-salinity Aleutian surface current near the North Pacific margin [*Dodimead et al.*, 1963]. (2) The North Pacific $\delta^{18}O$ signal may lag the Atlantic $\delta^{18}O$ signal by up to 1500 years [*Duplessy et*

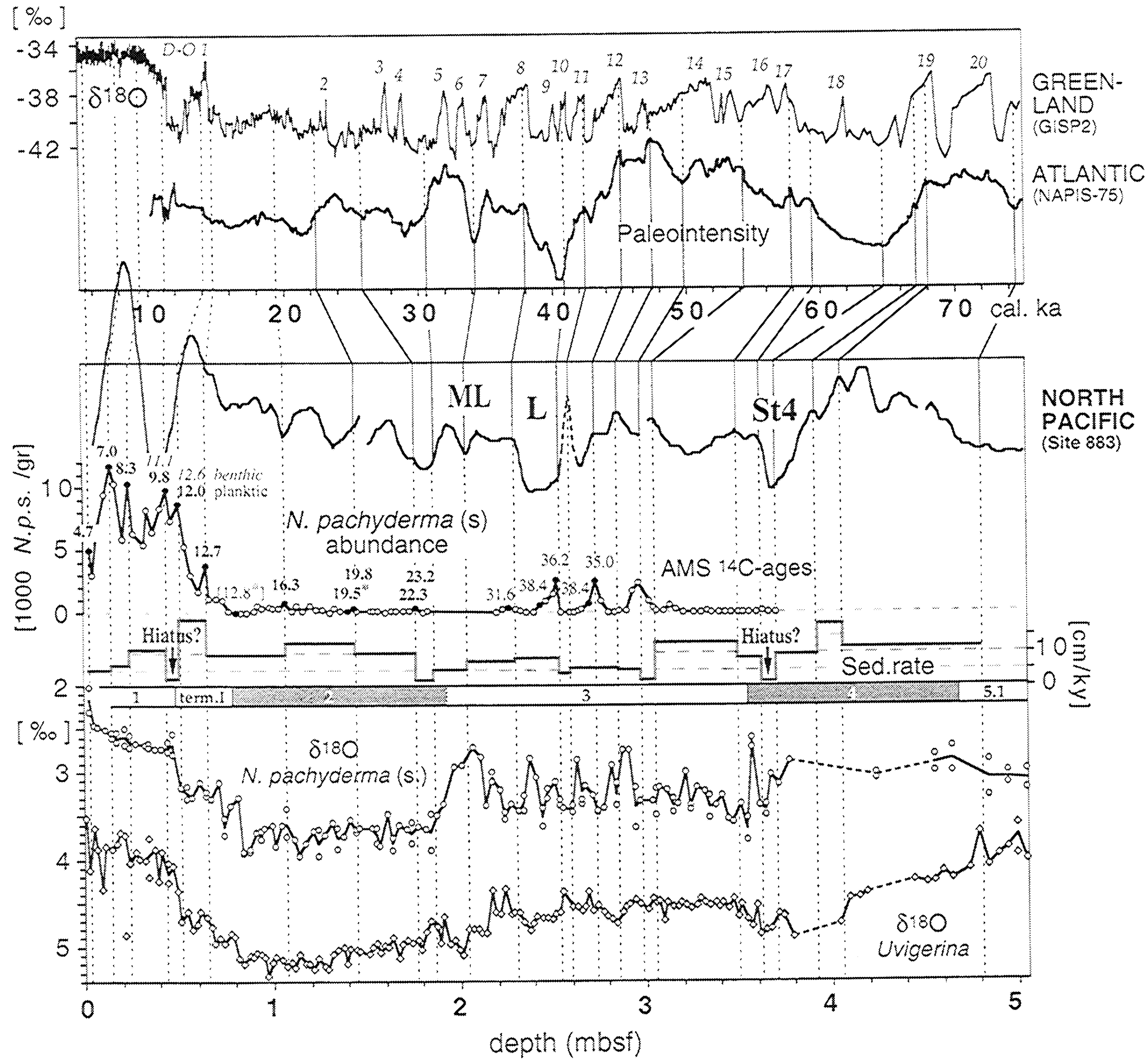

Figure 2. Age control at ODP Site 883, based on ^{14}C ages in the upper 1.80 m and on geomagnetic paleointensity farther downcore. Here, mbsf, meters below seafloor. Paleointensity records are natural remanent magnetization normalized to anhysteretic remanent magnetization (NRM/ARM). Vertical solid lines mark control points in the Site 883 paleointensity record [*Roberts et al.*, 1997] used for tuning to the North Atlantic paleointensity stack NAPIS-75 [*Laj et al.*, 2000]. Prominent paleointensity minima: ML, Mono Lake; L, Laschamp; St4, "late stage 4." Carbon 14 ages are corrected for a -800 year reservoir effect. Ages with asterisks are from *Morley et al.* [1995]. Absolute abundance of the dated species *N. pachyderma* (sinistral) is given. Solid circles represent dated samples. Marine isotope stages (MIS) 1-5.1 and termination I are indicated on the horizontal bar, as derived from the $\delta^{18}O$ curves of *Uvigerina* (data partly from *Keigwin* [1995]) and *N. pachyderma* (s.).

al., 1991], equal to the length of a D-O cycle. (3) Likewise, the ^{14}C reservoir effect in North Pacific surface water is generally high, but little known [*Southon et al.*, 1990; *Stuiver and Braziunas*, 1993]. It may even exceed the time span of a D-O cycle. Furthermore, we expect that the ^{14}C reservoir effect has strongly varied with time on the basis of differential intensities of upwelling of relatively old deep water and differential habitat depths of *N. pachyderma* (s.) [*Simstich*, 1999].

3.2. Carbon 14 Age Control Within Stages 1-2

A detailed age model of the upper 1.80-m-long section at Site 883 was obtained from a series of consistent ^{14}C ages free of age reversals back to 23.2 ^{14}C ka. Farther downcore, the ^{14}C ages appear controversial. Carbon 14 ages <20 ka were converted into calendar ages using the INTCAL98 calibration data [*Stuiver et al.*, 1998]. Two ages >20 ka were converted by adding 3000-3500 years [*Winn et al.*, 1991], which was supported by a best fit correlation of a minor geomagnetic paleointensity maximum found in both the North Atlantic paleointensity stack since 75 ka (NAPIS-75) and the Site 883 records (Table 1 and Figure 2).

The age range of 16.3-23.2 ^{14}C ka is consistent with MIS 2, deduced from the planktonic and benthic $\delta^{18}O$ records. Near 0.5 mbsf the prominent planktic $\delta^{18}O$ decrease by 0.6‰ of an initially presumed Preboreal age is bracketed within 7 cm by planktonic dates of 12.0 ^{14}C ka at the base and 9.8 ^{14}C ka on top of it (corrected for a ^{14}C reservoir age of 800 years). As compared to ^{14}C ages of the Preboreal elsewhere (9.8-9.1 ^{14}C ka; *Bard et al.* [1987], *Winn et al.* [1991], *Sarnthein et al.* [1994], and *Hughen et al.* [1998]) the ^{14}C age difference of 2200 years is large and probably exceeds the reach of extreme variations in the local ^{14}C reservoir effect [*Sikes et al.*, 2000]: The ages may indicate a Younger Dryas section at Site 883 condensed to <7 cm. More likely is a >1.5-kyr-long hiatus, in line with a striking sediment unconformity at 0.47-0.48 mbsf (core photo given by *Rea et al.* [1993]). On the basis of the planktonic ^{14}C age of 9.8 ka at 0.44 mbsf the pronounced negative planktonic and benthic δ^{18}O values at 0.47 mbsf reflect deposition immediately after the Younger Dryas (YD) stadial. The benthic ^{14}C reservoir age then exceeded the planktonic reservoir age by 1300 years, as today, when the Preboreal and Holocene started near 9.8 ^{14}C ka. An enigmatic abrupt planktonic δ^{18}O decrease at the core top occurs near 5 ^{14}C ka.

3.3. Geomagnetic marker horizons

Beyond the basic δ^{18}O stratigraphy. Site 883 shows a series of enigmatic short-term planktonic δ^{18}O minima of Δ0.6-0.9‰ which may reflect a response to the millennial-scale D-O cycles during MIS 3 (Figures 2 and 3). Their understanding requires a semimillennial precision in interhemispheric stratigraphic correlation. In particular, there is a clear need to identify quasi-coeval events in sediments of the North Pacific and North Atlantic oceans on the basis of signals, the timing of which is unbiased by the "aging" of oceanic water masses.

To cope with this problem, we employed characteristic geomagnetic intensity variations in the sediment record of Site 883 [*Roberts et al.*, 1997] as a new technique for tying the stratigraphy of MIS 3-5.1 to that of the stacked North Atlantic paleointensity record NAPIS-75 [*Laj et al.*, 2000] (Figure 2). This record, in turn, is closely tied (with a precision of 80-160 years) to the annual-layer-counted the Greenland Ice Sheet Project 2 (GISP2) ice core chronology [*Meese et al.*, 1994; *van Kreveld et al.*, 2000], on the basis of the planktic δ^{18}O record of semicentennial resolution and a dense series of ^{14}C ages in core PS2644 (*Voelker et al.* [1998] and Figure 2), which contains an individual record of the NAPIS-75 stack from the Icelandic Sea. Correlating the paleointensity records of ODP Site 883 and NAPIS-75 thus allows for a quasi-direct comparison of North Pacific and Greenland climate changes with millennial-scale precision and better. A paleointensity minimum associated with the Laschamp geomagnetic excursion [*Bonhommet and Babkine*, 1967] at 2.38-2.58 mbsf and a second minimum in the upper MIS 4, at 3.68-3.74 mbsf, are used as prime control points for our global age correlation.

By and large, the "Laschamp" paleointensity minimum at Site 883 is supported by ^{14}C ages of 38.4 and 36.2 ka and an age of 38.4 ka a little below, at 2.71 mbsf (Figure 2). These dates come close to the ^{14}C ages near the Laschamp event in the North Atlantic [*Voelker et al.*, 1998]. However, farther downcore, at 2.74 mbsf, where the geomagnetic intensity is high, an extremely young age of 35.0 ^{14}C ka demonstrates the uncertainty in ^{14}C chronology in MIS 3 at ODP Site 883. The partly inverse succession of planktonic ^{14}C ages across and below the Laschamp interval may result from a combination of (1) markedly increased ^{14}C production during the Laschamp event, (2) an extremely and short-term variable ^{14}C reservoir effect in the center of deepwater upwelling in the North Pacific subarctic gyre [*Reid*, 1962, 1965], and (3) strong bioturbational sediment mixing below this upwelling zone because of high carbon and nutrient fluxes (*Trauth et al.* [1997] and curve h of Figure 3).

Proceeding from the Laschamp and MIS 4 paleointensity minima as prime stratigraphic marker horizons, we "tuned" the complete series of minor positive and negative magnetic intensity features at Site 883 (between 1.50 and 5.0 mbsf) to the NAPIS-75 record of the last 75 kyr (Figure 4a). In particular, we correlated the small intensity minimum at 2.05 mbsf to the short minimum associated with the Mono Lake excursion and the intensity maxima/minima at 1.95, 2.6, 2.86, 2,98, 3.06, 3.5, 3.6, 3.91, 4.07, and 4.8 mbsf to a series of corresponding maxima/minima in NAPIS-75 (Figure 2). In this interhemispheric correlation the potential impact of differential magnetization "lock-in" depths [*deMenocal et al.*, 1990] was not quantitatively assessed, either in the NAPIS-75 or in the Site 883 geomagnetic records.

The short-term geomagnetic maximum at 2.6 mbsf, right below the Laschamp minimum, may be due to the presence of a major ash layer and in this case should be ignored as an outlier. Alternatively, as precisely the same spike also occurs in the synthetic field intensity record calculated by *Laj et al.* [2000] from the Greenland Ice Core Project (GRIP) ^{36}Cl record on the GISP2 timescale, the maximum may reflect a real geomagnetic excursion and thus strongly support our correlation.

In the Holocene, over termination I, and in MIS 2 the geomagnetic record at Site 883 was not suitable for age tuning, because it probably was overprinted by major changes in sediment composition. Moreover, the respective geomagnetic records from North Atlantic sediments are not well defined [*Laj et al.*, 2000].

3.4. Check Test of Geomagnetic Correlation

On the basis of our geomagnetic "tuning" and combined ^{14}C and benthic δ^{18}O stratigraphy the short-lasting planktonic δ^{18}O minima generally match the cold D-O stadials during ~60-10 kyr ago (Figures 3 and 4a). This holds true where age control is well-constrained such as prior to D-O interstadials 17-15, 13-11, 9-7, and 2-1. Accordingly, the sampling resolution reaches approximately 500 years and better. Only where the record is more poorly resolved do the δ^{18}O minima match D-O interstadials 7 and 6 at 1.9 - 2.1 mbsf. No signals are resolved at interstadials 15-14 and 11-10 (Figures 2, 3, and 4a), where carbonate dissolution is high or sedimentation rates are very low (Figure 2 and curve g of Figure 3).

The outlined antiphase pattern between the short-term planktonic δ^{18}O variability in the northwest Pacific and temperatures on Greenland is unexpected in terms of ocean-atmosphere general circulation model (O-A-GCM) simulations [*Mikolajewicz et al.*, 1997]. To test the robustness of this controversial correlation model, we employed an inverse approach, plotting the North Atlantic and North Pacific geomagnetic records as a result of tuning the negative δ^{18}O excursions at Site 883 to the D-O interstadials 5-20 in GISP2. A first version of this test (Figure 4b) leads, as compared with Figure 4a, to a significantly increased, almost persistent incongruity of the geomagnetic records with most of their

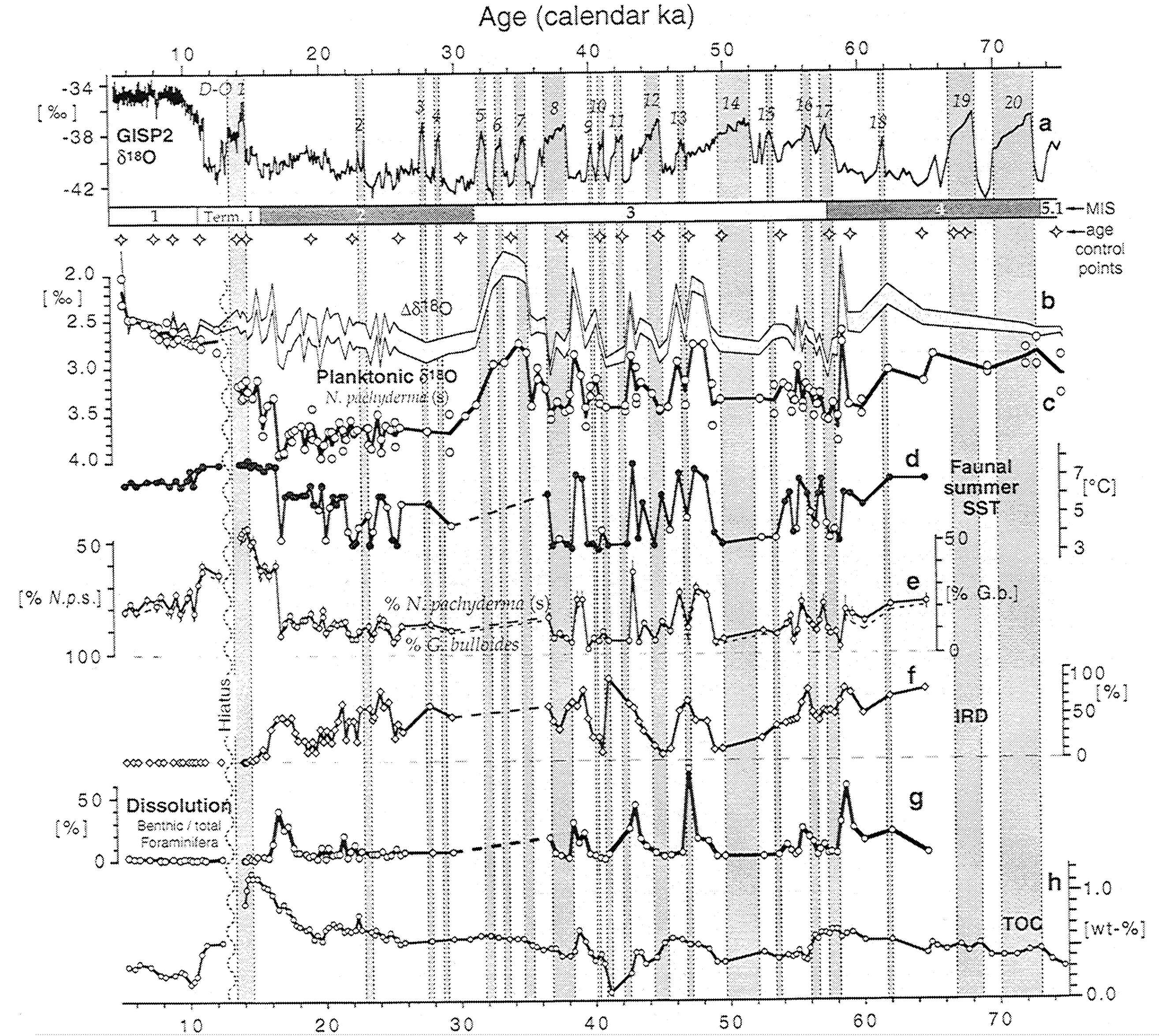

Figure 3. Time series of the last 75 kyr obtained from ODP Site 883 (curves b-h) compared to the Greenland GISP2 $\delta^{18}O$ ice record (curve a). Curves b and c show the planktonic $\delta^{18}O$ values of *N. pachyderma* (s.): $\delta^{18}O$ data which result from a reduction by the benthic $\delta^{18}O$ record of *Uvigerina* (Figure 2) and original data (thick line and open circles), respectively. The benthic $\delta^{18}O$ record representing the approximate variations in $\delta^{18}O$ "ice effect" was normalized to maximum and minimum estimates of the Last Glacial Maximum-Holocene shift of 1.3‰ (*Labeyrie et al.* [1987] and *Fairbanks* [1989]; upper thin curve] and 1.0‰ (*Schrag et al.* [1996]; lower thin curve]. Curve d shows sea surface temperature (SST) for summer, estimated from planktonic foraminiferal assemblages, using the SIMMAX transfer function of *Pflaumann et al.* [1996]. Open circles indicate major statistical uncertainty of >1°C due to there being less than 550 counted foraminifera specimens in a faunal assemblage dominated by a single species. Curve e shows percentages of *N. pachyderma* (s.) and *G. bulloides* (dashed line) >150 µm in percent of total planktonic foraminifera. Error bars in curve e are after *van der Plas and Tobi* [1965]. Curve f shows ice-rafted debris (IRD) >150 µm in percent of ash-free basis. Curve g shows percent benthic foraminifera over total foraminifera as an index of calcite dissolution [*Thunell*, 1976]. Curve h shows weight percent of total organic carbon (TOC).

short- and long-term excursions (r^2 = 0.34 versus r^2 = 0.58 in Figure 4a for the interval 30-60 ka). This incongruity of the geomagnetic records is reduced (r^2 = 0.49) when tuning the planktonic $\delta^{18}O$ minima at Site 883 to the next-younger D-O interstadials (Figure 4c). However, it is necessary in this case to vary the sedimentation rates at Site 883 especially strongly on short millennial timescales.

On the basis of this test we conclude that the immediate correlation of geomagnetic intensity records is adequate for demonstrating a pattern, where the distinct short-term planktonic $\delta^{18}O$ minima in the northwesternmost Pacific tend to correlate with D-O stadials of MIS 4-1 and positive $\delta^{18}O$ excursions match the D-O interstadials. In harmony with this trend the difference values between the planktonic and benthic

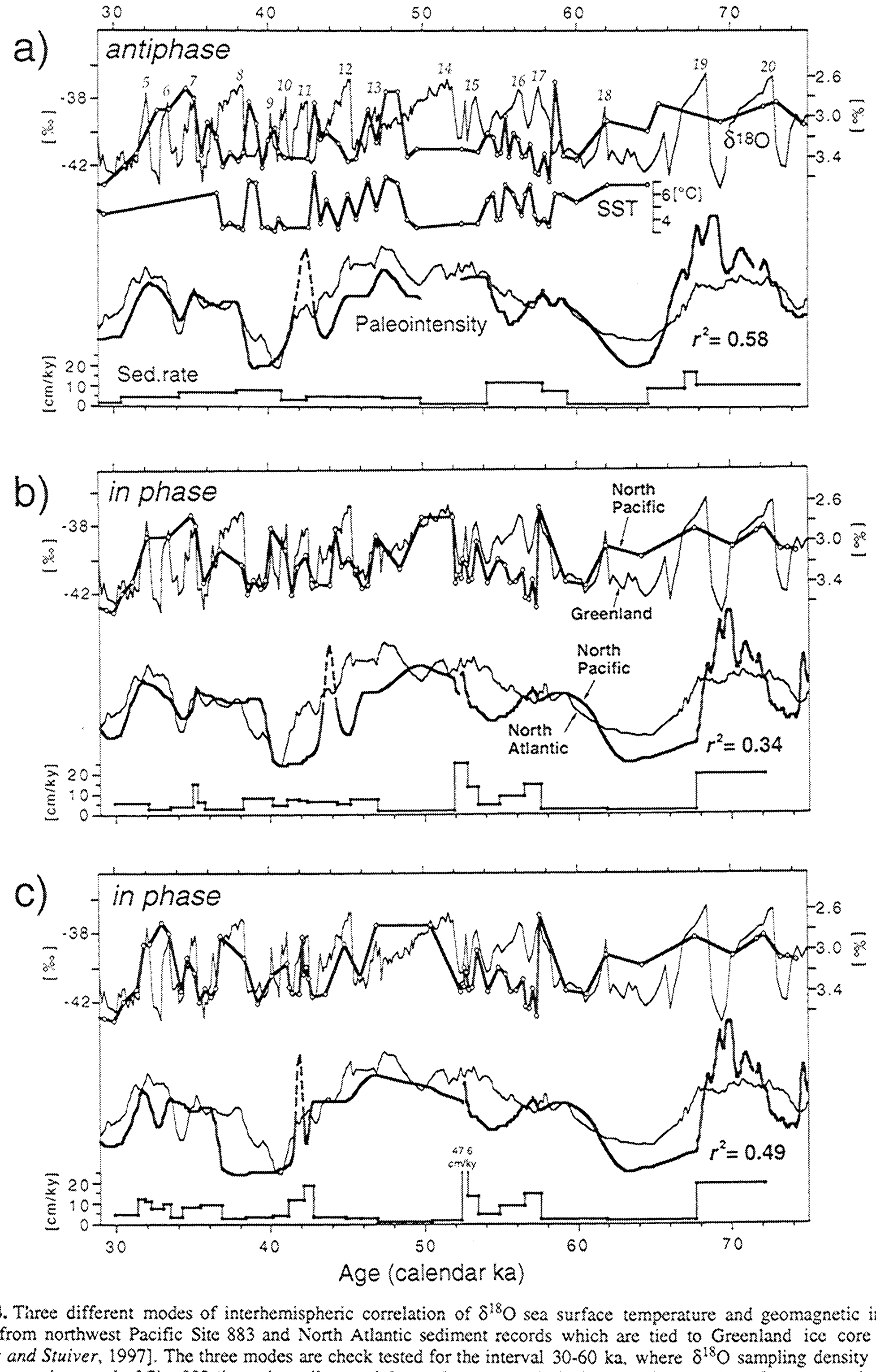

Figure 4. Three different modes of interhemispheric correlation of $\delta^{18}O$ sea surface temperature and geomagnetic intensity records from northwest Pacific Site 883 and North Atlantic sediment records which are tied to Greenland ice core GISP2 [*Grootes and Stuiver*, 1997]. The three modes are check tested for the interval 30-60 ka, where $\delta^{18}O$ sampling density is high. (a) Geomagnetic record of Site 883 (intensity spike at ash layer shown as a dashed curve; interruption of geomagnetic record linked to core break) tuned to the NAPIS-75 stacked record [*Laj et al.*, 2000] from the North Atlantic. (b and c) Planktonic $\delta^{18}O$ minima at Site 883 tuned to maximum temperatures at GISP2 in two different modes. Different sedimentation rate curves resulting from different correlation modes are shown at the bottom of each plot. Modes shown in Figures 4b and 4c produce substantial deviations of the geomagnetic intensity records (with r^2 calculated for 30-60 ka) and are discarded.

$\delta^{18}O$ records (the latter by and large representing the global "ice effect") do not decrease, but slightly increase from the last glacial to the Holocene "warm" interglacial (curve b of Figure 3).

4. Variations in Surface Water Properties

The millennial-scale amplitudes of up to 0.9‰ in the planktonic $\delta^{18}O$ record of MIS 2 and 3 (curves b and c of Figure 3) reflect changes in temperature and/or salinity of the surface water rather than in global ice volume, as documented by the comparatively moderate millennial-scale variance in the benthic $\delta^{18}O$ record (<0.25‰; Figure 2). Different from the short-term changes in MIS 3, most of the 1.4‰ decrease in planktonic $\delta^{18}O$ from the last glacial to the Holocene results from the reduction in global ice volume (1.0-1.3 ‰; *Schrag et al.* [1996], *Labeyrie et al.* [1987], and *Fairbanks* [1989]). The remaining $\delta^{18}O$ decrease of 0.1-0.4‰ reflects minor local changes in the surface water over the deglacial, such as either an insignificant warming by 0.5°-1.5°C and/or a salinity decrease by 0.2-0.8 units. In summary, the major changes in northwest Pacific surface water conditions over the last 75 kyr occurred on millennial rather than on glacial-to-interglacial timescales (curve b of Figure 3). Note that these scales correspond to the scale of short-term changes in the mode of Atlantic THC, which were more pronounced than changes from glacial-to-interglacial times [*Sarnthein et al.*, 1994, 2001; *Keigwin and Boyle*, 1999]. This analogy suggests a direct link between the variations in North Pacific surface water and North Atlantic THC.

During MIS 3 any short-term $\delta^{18}O$ decrease of up to 0.9‰ on D-O timescales may reflect a SST increase by 2.5°-4°C [*Shackleton*, 1974] and/or a salinity decrease of ~2 units. To unravel this ambiguity in oxygen isotope evidence, we used planktonic-foraminifera-based SST estimates and, moreover, the concentration of ice-rafted debris (IRD) as a tracer for melting icebergs (Figure 3).

Distinct warmings of the northwest Pacific surface water from 3.5° to 7°C during summer (curve d of Figure 3) are coeval with negative shifts in planktonic $\delta^{18}O$ during MIS 3, consistent with the short-term minima in the abundance of *N. pachyderma* (s.) (85-70%), well below the general background of 95-88% (curve e of Figure 3). The SST changes closely match the outlined 2.5°-4°C SST variation inferred from the planktonic $\delta^{18}O$ shifts, always in timing and partly in amplitude (curves c and d of Figure 3 and Figure 4a). The same applies, in principle, to the last deglacial. Here, faunal summer SST is 6°C at 18 ka, which falls to 3.5°C near 17 ka, and reaches an extended plateau of 8°C from 16 to 14 and from 12 to 11 calendar ka (cal ka). In the Holocene SST remains almost constant near 7°C from 10 to 5 cal ka. The modest net warming of 1°C from the Last Glacial Maximum to the Holocene is equivalent to the planktonic 0.2-0.5‰ $\delta^{18}O$ decrease, corrected for the global ice effect (curve b of Figure 3).

Within the range of the lysocline, which today reaches up to 1700 m water depth in the northern Pacific [*Craig et al.*, 1981], differential dissolution of foraminiferal tests may be a candidate for triggering short-term $\delta^{18}O$ and faunal changes at Site 883 at 2385 m water depth. In MIS 3 and 2, maxima in dissolution (curve g of Figure 3) match minima in $\delta^{18}O$ and *N. pachyderma* (s.), which equates to warm SST. This is in contrast to the expected enrichment of heavy oxygen isotopes [*Bonneau et al.*, 1980] and "cold" *N. pachyderma* (s.) [*Berger*, 1970] as a result of increased dissolution. Hence the short-term warmings over stage 3 are not dissolution artifacts. On the other hand, the maxima in IRD may be partly enlarged by the impact of maximum dissolution of planktonic foraminifers.

With regard to potential meltwater injections, only a few $\delta^{18}O$ minima in MIS 3 and 2 are in phase with IRD abundance maxima such as prior to D-O interstadials 17, 8, and 2. Most IRD maxima occur several centuries prior or subsequent to the $\delta^{18}O$ minima (curves c and f of Figure 3). Accordingly, only 3 out of 10 $\delta^{18}O$ minima may be linked to potential iceberg melt in the North Pacific subpolar gyre along with D-O climate oscillations. Vice versa, many IRD maxima match the local cold intervals or the transitions between cold and warm. Thus meltwater injections either from Kamchatka or from Alaska via the Aleutian current appear insignificant for explaining the $\delta^{18}O$ minima, each of which coincide with high SST values.

In summary, the close match of the faunal and isotopic SST records, also by the order of magnitude, supports the following interpretations: (1) SST fluctuated by 2.5°-4.0°C on millennial timescales during MIS 3-2, with the warm phases in the northwest Pacific most probably correlated to the cold stadials in Greenland and the North Atlantic. (2) Subsequent to the Last Glacial Maximum, during termination Ia, the SST first increased by (net) 2°C to an extended maximum of 8°C over the Bølling-Allerød and Preboreal. Later, during the early Holocene, SST at Site 883 slightly dropped and exceeded the LGM values by only 1°C.

5. Potential Upwelling Control of Northwest Pacific SST Variations

An atmospheric transfer of the Dansgaard-Oeschger temperature cycles found in Greenland will result in quasi-coeval signals across the high-latitude Northern Hemisphere, as simulated with a coupled O-A-GCM by *Mikolajewicz et al.* [1997]. Therefore the antiphase temperature trend of North Pacific surface water relative to the North Atlantic (Figures 3 and 4a) is unexpected. Only the ocean circulation is capable of forcing the observed west-east interhemispheric seesaw effects such as modeled by *Huang et al.* [2000], by analogy to the seesaw reported for the relationship between northern and southern ice sheets and the northern and southern Atlantic [*Blunier et al.*, 1998; *Ninnemann et al.*, 1999].

In the Northwest Pacific most effective temperature changes may originate from the intensity of regional deepwater upwelling. Here a tongue of both nutrient-enriched deep water and strongly upward deflected isotherms indeed indicates the modern upwelling of upper deep water south of the Aleutian island arc (*Reid* [1965], *Craig et al.* [1981], and *Ganachaud and Wunsch* [2000]: 1.5 Sv). This upwelled water probably constitutes the terminus of the last derivative of North Atlantic Deep Water (NADW) which is much diluted by Antarctic Bottom Water and Central Pacific Water in the Southern Ocean [*Schiller et al.*, 1997] and finally advances into the North Pacific as part of the global THC system [*Broecker*, 1991; *Gordon*, 1986, 1991]. In harmony with *Huang et al.* [2000] we propose that any slowdown or shutoff of NADW formation results in a reduction or turnoff of the upwelling of cold Pacific Deep Water in the northwest Pacific subarctic gyre. This process has led, in turn, to a distinct and instantaneous deepening of the thermocline and thus to a transient warming of surface water during the D-O stadials over the last 60 kyr. During D-O interstadials the THC mode in the North Atlantic was reestablished similar to today, and accordingly, the reinvigorated upwelling of cold deep waters again reduced North Pacific SST by 2.5°-4.0°C.

An interhemispheric thermohaline control of subarctic Pacific SST on a glacial-to-interglacial timescale was first suggested by *Haug* [1996] on the basis of high alkenone-derived SST values during the Last Glacial Maximum. Nevertheless, almost no variability in upwelling intensity is shown in our TOC record, exept for the time prior to and during D-O 1. The TOC minimum at 41 ka is tied to an ash layer (curve h of Figure 3). We relate this lack of variability to a "burn-down" effect which depletes the enriched carbon in thin sediment layers during high-productivity times by subsequent oxidation, when the upwelling productivity and carbon fluxes again are low. On long timescales, however, the coupling between any increase in upwelling-induced opal production in the northwest Pacific and enhanced deepwater formation in the Atlantic was particularly obvious over the time interval 5.3-2.7 Ma [*Haug et al.*, 1999].

Paleoceanographic evidence for the outlined coupling between Atlantic and Pacific thermohaline circulation also comes from variations in the intermediate-water flow into northeast Pacific Santa Barbara Basin [*Kennett and Ingram*, 1995; *Behl and Kennett*, 1996]. Here, low benthic-planktonic ^{14}C age differences in nonlaminated sediments are coeval with the D-O stadials of the past 60 kyr. The low age differences probably reflect a cessation or reduction in the incursion of "old" (oxygen- and ^{14}C -depleted) intermediate water from the southern Pacific into the California margin basins during times when deepwater formation in the North Atlantic had diminished. In harmony with this finding the O-GCM of *Seidov and Haupt* [1997] simulates a reversed deepwater flow in the northwest Pacific in response to a Heinrich event like meltwater disturbance of the North Atlantic.

Further support comes from benthic stable isotope records on glacial-to-interglacial timescales [*Duplessy et al.*, 1988; *Keigwin*, 1998]. These records provide clear evidence of a reversed THC in the upper northwest Pacific, resulting in intermediate-water formation down to 2000 m, when the NADW formation was reduced.

Also on millennial timescales a first high-resolution benthic $\delta^{13}C$ record from the midlatitude northeast Pacific [*Lund and Mix*, 1998] shows increased values along with major North Atlantic stadials, suggesting a pattern of deepwater-to-intermediate-water ventilation opposed to that of the Atlantic. Unfortunately, the benthic $\delta^{13}C$ record of the upper deep water at Site 883 (2385 m) suffers from very small specimen numbers of epibenthic foraminifer species, the signals of which are subject to bioturbational mixing (T. Kiefer et al., manuscript in preparation, 2001).

Haug et al. [1999] postulate an overall glacial increase in stratification of the upper water column in the subarctic North Pacific, largely persistent over the last 2.7 Myr. This model is based on a dramatic drop in diatom productivity together with an increase in nutrient utilization, coincident with the first arrival of ice-rafted debris. The postulated robust halocline during the Pleistocene should have largely impeded the upwelling of deep water to the sea surface and, accordingly, any deep-water-derived cooling of the surface water. However, the new record of SST variations at Site 883 does not support this concept.

6. Conclusions

1. At the northwest Pacific ODP Site 883, abundant relative minima in planktonic $\delta^{18}O$, representing isotopic shifts up to 0.9‰ in marine isotope stage 3, are clearly established as SST maxima. The amplitudes of coeval changes in both the $\delta^{18}O$ and faunal composition of planktonic foraminifera correspond to SST changes of 2.5°-4°C.

2. A first interhemispheric correlation of these northwest Pacific SST and $\delta^{18}O$ fluctuations to the North Atlantic and Greenland on millennial to submillennial timescales is established using geomagnetic paleointensity changes as a stratigraphic tool which is independent of seawater age and thus provides a new quality of dating precision.

3. On the basis of this interhemispheric correlation the SST maxima in the North Pacific are most likely coeval with cold Dansgaard-Oeschger stadials in the North Atlantic. Accordingly, the SST variations in the northern high latitudes of the two oceans are in antiphase, implying an interhemispheric climatic seesaw mechanism.

4. The robustness of this controversial correlation model was tested by tuning the positive SST excursions at Site 883 so that it is in phase to the warm D-O interstadials in the GISP2 ice core. This approach increased the incongruity of the geomagnetic records and thus was discarded.

5. Upwelling of cold Pacific Deep Water at the North Pacific terminus of the global THC [*Gordon*, 1991] is considered as the possible instantaneous forcing mechanism that controls the local thermocline depth and short-term temperature oscillations which are strictly opposed to those in the North Atlantic.

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