

## Millennial-scale deep water oscillations: Reflections of the North Atlantic in the deep Pacific from 10 to 60 ka

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**Abstract.** Northeast Pacific benthic foraminiferal  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  reveal repeated millennial-scale events of strong deep-sea ventilation (associated with nutrient depletion and/or high gas exchange) during stadial (cool, high ice volume) episodes from 10 to 60 ka, opposite the pattern in the deep North Atlantic. Two climate mechanisms may explain this pattern. North Pacific surface waters, chilled by atmospheric transmission from a cold North Atlantic and made saltier by reduced freshwater vapor transports, could have ventilated the deep Pacific from above. Alternatively, faster turnover of Pacific bottom and mid-depth waters, driven by Southern Ocean winds, may have compensated for suppressed North Atlantic Deep Water production during stadial intervals. During the Younger Dryas event (~11.6-13.0 cal ka), ventilation of the deep NE Pacific (~2700 m) lagged that in the Santa Barbara Basin (~450 m) by >500 years, suggesting that the NE Pacific was first ventilated at intermediate depth from above and then at greater depth from below. This apparent lag may reflect the adjustment time of global thermohaline circulation.

### 1. Introduction

Evidence for millennial-scale climatic shifts is beginning to emerge on a global basis [e.g., Grimm *et al.*, 1993; Behl and Kennett, 1996; Bard *et al.*, 1997]. Rapid fluctuations appear to be associated with changes in North Atlantic Deep Water (NADW) formation rate [e.g., Oppo and Lehman, 1995], but the behavior of abyssal waters in other ocean basins has remained a mystery.

Here we present isotopic evidence for millennial-scale oscillations in North Pacific Deep Water (NPDW) composition from 10 to 60 ka. The pacing of NPDW change is similar to climatic cycles observed in the North Atlantic. Unlike the Atlantic, however, ventilation in the Pacific generally increases during cold (i.e., stadial) intervals. We offer two possible explanations for this phenomenon, one which involves deep water formation in the North Pacific and the other which calls upon enhanced Pacific mid-depth outflow to compensate deep upwelling near Antarctica. Both are tied to Atlantic thermohaline overturn and therefore represent potential mechanisms for propagating millennial-scale climate events around the globe.

Most of the early evidence for abrupt glacial climate change comes from the North Atlantic region. Oxygen isotope ratios in Greenland ice outline 2-3 kyr warm-cold oscillations, known as Dansgaard-Oeschger temperature cycles [Dansgaard *et al.*, 1993; Grootes *et al.*, 1993]. Similar patterns exist in the temperature-sensitive coiling ratio of the planktonic foraminifer *Neogloboquadrina pachyderma* from North Atlantic sediments [Lehman and Keigwin, 1992; Bond *et al.*, 1993; Oppo and Lehman, 1995]. Layers of ice-rafted detritus in these sediments imply that increased iceberg flux occurred near the end of each stadial period [Bond and Lotti,

1995]. Groups of stadial intervals fall into longer-term (7-10 kyr) cooling cycles, themselves terminated by massive "Heinrich" ice-rafting events [Bond *et al.*, 1993].

North Atlantic Deep Water properties shifted with millennial-scale climatic oscillations. Benthic foraminiferal  $\delta^{13}\text{C}$  from the subpolar North Atlantic suggests that low sea surface temperatures during the last glacial period were generally associated with reduced Atlantic thermohaline circulation [Oppo and Lehman, 1995]. Carbonate records from the midlatitude Atlantic indicate similar shifts in NADW production [Keigwin and Jones, 1994]. Consistent with this pattern, millennial-scale sea surface temperature lows in the equatorial Atlantic (planktonic foraminiferal  $\delta^{18}\text{O}$  highs) match apparent decreases in NADW (benthic foraminiferal  $\delta^{13}\text{C}$  lows) [Curry and Oppo, 1997].

Geologic records from shallow marginal basins of the Northeast Pacific indicate that millennial-scale climate changes extended beyond the North Atlantic region. In the Santa Barbara Basin (which has a sill depth 450 m below modern sea level and contains laminated sediments), well-dated intervals lacking laminations, with high benthic and planktonic foraminiferal  $\delta^{18}\text{O}$  and abundant left-coiling *N. pachyderma*, show that surface water cooling and enhanced basin ventilation occurred near the time of the North Atlantic Younger Dryas cool event [Kennett and Ingram, 1995]. Laminations in the Gulf of California [Keigwin and Jones, 1990] contemporary with those in the Santa Barbara Basin imply a regional switching from oxygen-poor to oxygen-rich conditions at intermediate water depth (<700 m). Behl and Kennett [1996] suggest that laminated intervals in the Santa Barbara Basin correlate with interstadial events in Greenland over the past 60,000 years, hinting at a connection between the processes which control surface temperatures in the North Atlantic and intermediate water composition in the Northeast Pacific. Coupled ocean-atmosphere model results indicate the atmosphere may have transmitted North Atlantic cooling to North Pacific surface and intermediate waters during the Younger Dryas event [Mikolajewicz *et al.*, 1997].

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## 2. Modern Oceanographic Setting

Because of the large salinity-driven density contrast between surface and underlying waters in the modern North Pacific, there is little or no deep or bottom water formed there [Warren, 1983]. Rather, Pacific Bottom Water (PBW) is a blend of ~70% southern (Antarctic) component, ~15% northern (NADW) component, and ~15% intermediate component water [Broecker *et al.*, 1985].

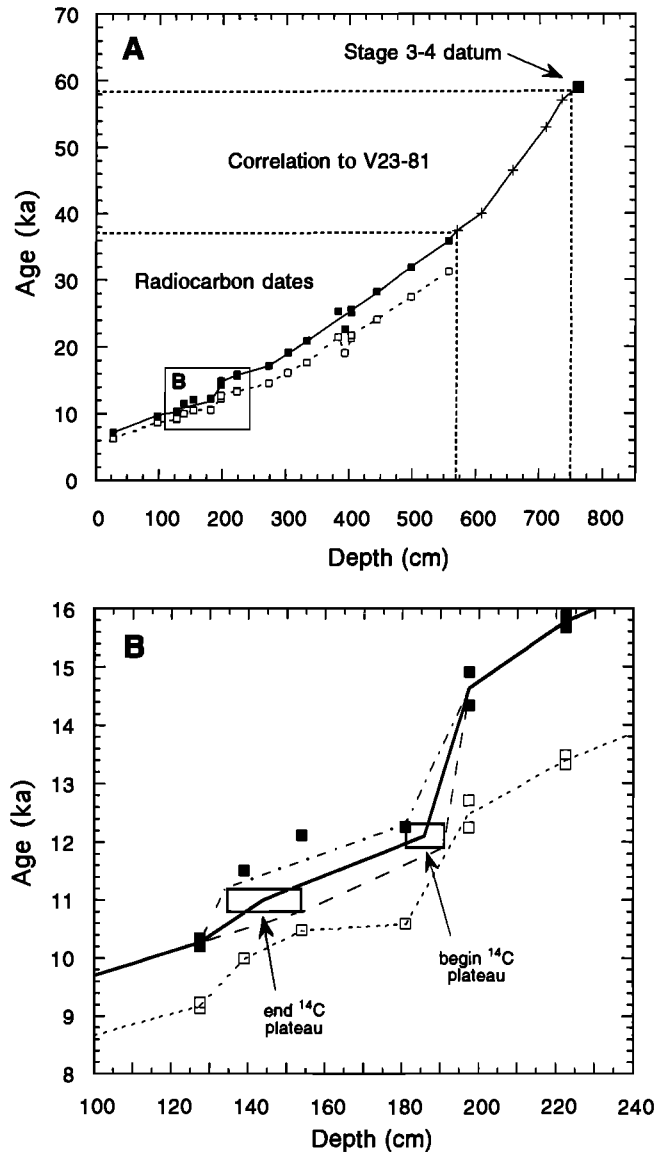
Moving from the southwest to northeast Pacific, PBW becomes warmer and less saline through mixing, and its oxygen content decreases [Mantyla and Reid, 1983]. NPDW is apparently created in the Northeast Pacific Basin through exposure of PBW to geothermal heat and cross-isopycnal mixing with overlying warmer, less saline water [Mantyla, 1975]. NPDW spreads southward as a mid-depth return flow, separated from the northward flowing PBW by a stability maximum near 4 km depth [Mantyla, 1975]. The core of the southward flow is at 2-3 km depth, just below the effective sill depth of Drake Passage. This topographic barrier facilitates southward geostrophic flow to the Southern Ocean, and thus plays a key role in ventilating the ocean's interior [Toggweiler and Samuels, 1995].

## 3. Methods and Materials

The sediment core used for this study, W8709A-13PC (2712 m depth; 42.1°N, 125.8°W), comes from the eastern flank of the Gorda Ridge and is composed predominantly of hemipelagic clay with low (<15%) carbonate and opal content. Today, the site is bathed in NPDW and is below the regional carbonate compensation depth of 2400 m [Lyle *et al.*, 1992]. From 10 to at least 45 ka, however, the carbonate compensation depth (CCD) was deeper than 2700 m [Karlín *et al.*, 1992]. There are five small sandy silt layers (inferred to be distal turbidites) in W8709A-13PC at 206-208, 230-232, 252-256, 322-325, and 355-358 cm.

Stable isotope samples were taken at 3 cm intervals from 99 to 761 cm and 20 cm intervals from 40 to 99 cm and 761 to 868 cm. We separated three species of benthic foraminifera from the >150  $\mu\text{m}$  fraction, *Cibicidoides wuellerstorfi*, *Uvigerina senticosa*, and *Uvigerina peregrina*. On average there were six *C. wuellerstorfi* individuals per analysis. For *Uvigerina* spp. the average was seven. The microfossils were cleaned ultrasonically in alcohol, dried, and then roasted for 1 hour at 400°C under high vacuum. All isotope measurements were made at Oregon State University using a Finnigan/MAT251 mass spectrometer coupled to an Autoprep Systems carbonate reaction device. Reactions occurred under vacuum in ~100% orthophosphoric acid at 90°C.

Stable isotopic ratios are expressed following standard practice as  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  relative to Pee Dee belemnite (PDB). Primary calibration to the PDB standard was through U. S. National Institute of Standards and Technology NBS-20 calcite, assuming  $\delta^{18}\text{O} = -4.14\text{‰}$  and  $\delta^{13}\text{C} = -1.06\text{‰}$  (supported by analyses of NBS-19 calcite). Analytical precision of replicate analyses of NBS-20 made over the same period as the foraminiferal analyses was  $\pm 0.06\text{‰}$  for  $\delta^{18}\text{O}$  and  $\pm 0.02\text{‰}$  for  $\delta^{13}\text{C}$  ( $\pm 1 \sigma$ ,  $n = 14$ ). Precision of analyses of a local standard (a pure finely-ground calcitic marble) was  $\pm 0.05\text{‰}$  for  $\delta^{18}\text{O}$  and  $\pm 0.02\text{‰}$  for  $\delta^{13}\text{C}$  ( $\pm 1 \sigma$ ,  $n = 328$ ).



**Figure 1.** (a) Age model for W8709A-13PC, including reservoir-corrected radiocarbon dates (open squares) and corresponding calendar ages (solid squares). Dates from 36-57 kyr BP are based on correlation of millennial-scale  $\delta^{18}\text{O}$  highs to stadial intervals in North Atlantic Core V23-81. The datum at 59 kyr BP marks the marine isotope stage 3-4 boundary in W8709A-13PC. All dates are listed in Table 1. (b) Age model options for the interval between 130 and 195 cm. Calendar ages in this interval calibrated using standard techniques [e.g., Stuiver and Braziunas, 1993; Bard *et al.*, 1992] are ambiguous, because of the radiocarbon plateau at 10,000  $^{14}\text{C}$  years. We use reservoir-corrected dates in this interval to define the stratigraphic position of the  $^{14}\text{C}$  plateau and then apply known ages for the beginning and end of the plateau to the appropriate depths (see text). The oldest possible age model is defined by the upper left-hand corner of each box, and the youngest possible age model is defined by the lower right-hand corner. The age model used in this paper is an average of these two end-members (solid line).

## 4. Chronostratigraphy

The age model for W8709A-13PC was constructed using 21 calendar-corrected  $^{14}\text{C}$  ages, the marine oxygen isotope stage 3-4 boundary at 59 ka [Imbrie *et al.*, 1984], and 5 ages

assigned by correlation to North Atlantic Core V23-81 (Figure 1a). Radiocarbon control for W8709A-13PC spans the interval from 7 to 36 calendar ka. Dates original to this paper are based on 3.5 - 12.1 mg samples of planktonic foraminifera reacted to carbon dioxide at Oregon State University and analyzed at Lawrence Livermore National Laboratory (Table 1). Accelerator mass spectrometry (AMS) dates from *Gardner et al.* [1997] are based on 5 mg samples of mixed planktonic or benthic foraminiferal species. Those from *Lyle et al.* [1992] are based on 5-10 mg samples of mixed planktonic foraminiferal species.

Reservoir-corrected ages were calculated by subtracting 718 and 2380 years from the raw planktonic and benthic  $^{14}\text{C}$  ages, respectively [*Ortiz et al.*, 1997]. At depths where both benthic and planktonic reservoir-corrected values were determined we used the average.

Radiocarbon dates younger than 10,500 reservoir-corrected years were converted to calendar ages following *Stuvier and Braziunas* [1993]. In this calibration, calendar ages are calculated as:

$$\text{Calendar age (years)} = R_m - (R_t + \Delta R) \quad (1)$$

**Table 1.** Age Model for W8709A-13PC

Depth, cm	Raw $^{14}\text{C}$ Age, ka	Reservoir-Corrected $^{14}\text{C}$ Age, ka	Calendar Age, ka	Inferred Calendar Age, ka	Notes	Data Source
25-30	7.00 ± .23	6.28	7.18	7.18	mixed planktonic species	1
95-100	11.01 ± .36	8.63	9.64	9.64	mixed benthic species	1
125-130	11.51 ± .33	9.14	10.20	10.27	mixed benthic species	1
125-130	9.96 ± .23	9.24	10.34	10.27	mixed planktonic species	1
138-140	10.72 ± .07	10.00	11.50	--	<i>N. pachy</i> and <i>G. bull</i> (3.5 mg)	2
144	--	--	--	11.00	end of $^{14}\text{C}$ plateau	2
153-155	11.20 ± .06	10.48	12.12	--	<i>N. pachy</i> (5.6 mg)	2
180-182	11.31 ± .14	10.59	12.26	--	<i>N. pachy</i> (6.1 mg)	2
186	--	--	--	12.10	beginning of $^{14}\text{C}$ plateau	2
195-200	14.63 ± .39	12.25	14.34	14.63	mixed benthic species	1
195-200	13.43 ± .19	12.71	14.91	14.63	mixed planktonic species	1
220-225	15.86 ± .50	13.48	15.87	15.77	mixed benthic species	1
220-225	14.04 ± .28	13.32	15.67	15.77	mixed planktonic species	1
270-275	15.27 ± .22	14.55	17.18	17.18	mixed planktonic species	3
300-305	18.49 ± .38	16.12	19.08	19.10	mixed benthic species	1
300-305	16.87 ± .27	16.15	19.12	19.10	mixed planktonic species	1
330-335	18.37 ± .27	17.65	20.90	20.90	mixed planktonic species	3
381-383	22.14 ± .14	21.42	25.28	--	<i>G. bull</i> (10.6 mg)	2
393	--	--	--	24.64	average of 381-405 cm	2
390-395	19.82 ± .61	19.10	22.61	--	mixed planktonic species	3
400-405	24.05 ± 1.53	21.67	25.57	--	mixed benthic species	1
400-405	21.96 ± .49	21.24	25.08	--	mixed planktonic species	1
442-444	24.78 ± .14	24.06	28.25	28.25	<i>G. bull</i> (9.8 mg)	2
496-498	28.14 ± .40	27.42	31.91	31.91	<i>G. bull</i> (12.1 mg)	2
556-558	31.97 ± .32	31.25	35.92	35.92	<i>G. bull</i> (11.7 mg)	2
570	--	--	--	37.4	correlate to 452 cm in V23-81	4
608	--	--	--	40.0	correlate to 492 cm in V23-81	4
658	--	--	--	46.5	correlate to 549 cm in V23-81	4
710	--	--	--	53.0	correlate to 616 cm in V23-81	4
735	--	--	--	57.0	correlate to 663 cm in V23-81	4
760	--	--	--	59.0	$\delta^{18}\text{O}$ stage 3/4 boundary	5

*N. pachy* is *Neogloboquadrina pachyderma* and *G. bull* is *Globigerinia bulloides*. Inferred calendar ages comprise the "average" age model in Figure 1. Data sources are (1) *Gardner et al.* [1997]; (2) this study; (3) *Lyle et al.* [1992], (4) *Bond et al.* [1993], and (5) *Imbrie et al.* [1984]. Ages in data source (4) based on  $^{14}\text{C}$  were corrected to calendar ages (see text).

where  $R_m$  is the measured ( $\delta^{13}\text{C}$  corrected) radiocarbon age,  $R_t$  is a time-dependent simulation of the global surface ocean which includes a nominal 400 year reservoir correction, and  $\Delta R$  is an additional regional shift in the reservoir age that is assumed, as a first approximation, to be constant. We used a  $\Delta R$  value of 320 years and 1980 years for the planktonic and benthic raw  $^{14}\text{C}$  ages, respectively.

Reservoir-corrected ages in W8709A-13PC older than 10,500 radiocarbon years were transformed to calendar years using the equation of Bard *et al.* [1992]:

$$\text{Calendar age (years)} = -5.85 \times 10^{-6}(A^2) + (1.39A) - 1807 \quad (2)$$

where  $A$  is the reservoir-corrected radiocarbon age in years. This equation is calibrated from 10,000 to 38,000 radiocarbon years, but corrections older than 22,000 years are not well constrained [Bard *et al.*, 1992]. The calendar ages from 381 to 405 cm were averaged to produce one datum for the age model.

Ages from 36 to 57 ka are based on the correlation of millennial-scale  $\delta^{18}\text{O}$  peaks in W8709A-13PC to stadial intervals in North Atlantic Core V23-81 [Bond *et al.*, 1993]. We converted the reservoir-corrected  $^{14}\text{C}$  age model from V23-81 [Bond *et al.*, 1993] to calendar ages using the Bard *et al.* [1992] equation. Ages for samples between the control points were calculated by linear interpolation.

A zone of nearly constant  $^{14}\text{C}$  dates is apparent in W8709A-13PC near 10,000  $^{14}\text{C}$  years (Figure 1b), which we infer to be the deglacial radiocarbon "plateau" [Stuvier and Braziunas, 1993; Edwards *et al.*, 1993]. Using standard calendar correction formulas in this zone produces ambiguous results because the formulas cannot distinguish between the beginning and end of the plateau. Rather than

use these ages directly, we used them to define the stratigraphic position of constant  $^{14}\text{C}$  in W8709A-13PC. We then applied known calendar ages for the beginning (11.9-12.3 ka) and end (10.8-11.2 ka) of the plateau [Stuvier and Braziunas, 1993; Edwards *et al.*, 1993] to the  $^{14}\text{C}$ -constrained depths for the beginning (181-191 cm) and end (134-154 cm) of the zone of constant  $^{14}\text{C}$ . Possible depth versus time positions for the plateau are outlined with rectangular boxes in Figure 1b. The youngest possible age model is defined by the lower right-hand corner of each box, and the oldest possible age model is defined by the upper left-hand corner. The age model used here (Table 1) is constrained by the center point of these boxes.

## 5. Results

On glacial-interglacial timescales, high values of benthic  $\delta^{18}\text{O}$  correspond to low (more negative) values of  $\delta^{13}\text{C}$  in W8709A-13PC (Figure 2). During marine oxygen isotope stages 2 and 4,  $^{18}\text{O}$  is enriched by as much as 1.5‰ relative to the early Holocene. In Figure 2,  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  of *Uvigerina* spp. is used for the upper 110 cm of the core because of a lack of *C. wuellerstorfi* in this interval.

*C. wuellerstorfi*  $\delta^{13}\text{C}$  is low during stages 2 and 4, with values typically between  $-0.5\text{‰}$  and  $-0.6\text{‰}$ ,  $\sim 0.3\text{‰}$  lower than in stage 1. This  $0.3\text{‰}$  shift is similar to the global average glacial-interglacial change [Curry *et al.*, 1988], and the pattern is consistent with that attributed to fluctuations in terrestrial biomass [Shackleton, 1977]. In stage 3,  $\delta^{13}\text{C}$  averages  $-0.3\text{‰}$  and reaches  $-0.2\text{‰}$  in some intervals, close to modern  $\delta^{13}\text{C}$  values for total dissolved inorganic carbon ( $\Sigma\text{CO}_2$ ) in the deep Northeast Pacific [Kroopnick, 1985].

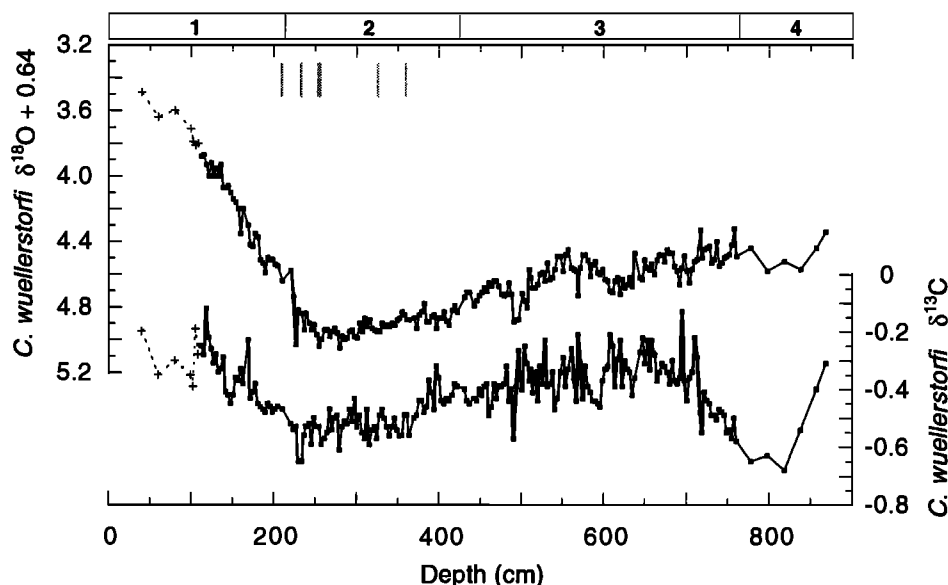


Figure 2. Oxygen and carbon isotopes versus depth in W8709A-13PC. Data at depths  $<110$  cm (+) are from *Uvigerina* spp. (*U. senticosus* and *U. peregrina*). All other data (dots) are from *Cibicoides wuellerstorfi*. *C. wuellerstorfi*  $\delta^{18}\text{O}$  values were adjusted by  $+0.64\text{‰}$  to match the *Uvigerina* scale. *Uvigerina*  $\delta^{13}\text{C}$  values were adjusted by  $+0.70\text{‰}$  (the mean offset in samples where both genera were measured) to match the *Cibicoides* scale. Marine oxygen isotope stages 1-4 are indicated on the upper x axis. On long timescales,  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  are inversely correlated, while on millennial timescales,  $\delta^{13}\text{C}$  maxima generally occur during  $\delta^{18}\text{O}$  maxima. Small sandy silt layers (apparently turbidites) at 206-208, 230-232, 252-256, 322-325, and 355-358 cm are marked with shaded rectangles.

Both  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  display millennial-scale (<10 kyr period) variability during marine oxygen isotope stages 2 and 3. Positive  $\delta^{18}\text{O}$  excursions (values 0.1-0.3‰ higher than adjacent intervals) occur repeatedly, with the most variability concentrated in stage 3. Millennial-scale peaks in  $\delta^{13}\text{C}$  of ~0.1-0.3‰ tend to occur during intervals of high  $\delta^{18}\text{O}$ , opposite the glacial-interglacial pattern, but there are exceptions. For example,  $\delta^{13}\text{C}$  lows near 290, 410, 500, 570, and 615 cm interrupt positive  $\delta^{13}\text{C}$  excursions, producing double  $\delta^{13}\text{C}$  peaks. In general, it appears that  $\delta^{13}\text{C}$  is more variable during  $\delta^{18}\text{O}$  highs than lows.

## 6. Discussion

### 6.1. Timing of Events

From 10 to 60 calendar ka,  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  maxima occurred every 4-6 kyr on average, similar to the pacing of millennial-scale polar front advances in the North Atlantic (Figure 3). Direct comparison of the benthic oxygen isotope record in W8709A-13PC with the percentage of *N. pachyderma* (sinistral) in Core V23-81 suggests that each  $\delta^{18}\text{O}$  high in the NE Pacific has a corresponding sea surface temperature low in the North Atlantic. Maxima in  $\delta^{18}\text{O}$  centered near 17, 25, 32, 41, and 50 ka align with the stadial intervals terminated by Heinrich events 1, 2, 3, 4, and 5, respectively. Other maxima in  $\delta^{18}\text{O}$  near 37, 46, and 57 ka also correspond to high percentages of *N. pachyderma* (sinistral) in Core V23-81. There is a small  $\delta^{18}\text{O}$  maximum (<0.1‰) near the time of the Younger Dryas, interrupting the glacial-interglacial transition.

Some of the Dansgaard-Oeschger temperature cycles apparent in the Greenland Ice Core Project (GRIP) core [Dansgaard *et al.*, 1993] are absent from the NE Pacific  $\delta^{18}\text{O}$  and North Atlantic sea surface temperature records (see Bond *et al.* [1993] for comparison of V23-81 and GRIP). For example, in the time interval between Heinrich events 3 and 4, there are four apparent interstadial events in GRIP, two interstadial events in V23-81, and two  $\delta^{18}\text{O}$  lows in W8709A-13PC. This may be a function of bioturbation in the sedimentary records and the higher sampling resolution for the GRIP core. Between 10 and 60 ka, samples in W8709A-13PC were taken every 250 years on average, about the same as in V23-81 [Bond *et al.*, 1993]. In the GRIP ice core samples were taken every 130 years [Dansgaard *et al.*, 1993].

Positive  $\delta^{13}\text{C}$  excursions in W8709A-13PC generally occur during stadial conditions. Maxima in  $\delta^{13}\text{C}$  at 24.5, 26, 31, 33, 37, 40, 46, 51, and 53 ka each correspond to low sea surface temperatures in the North Atlantic (Figure 3). The correlation between  $\delta^{13}\text{C}$  and low sea surface temperature is less consistent than for  $\delta^{18}\text{O}$ , however. For example, relatively low  $\delta^{13}\text{C}$  at 25 and 32 ka occur during stadial conditions. The processes controlling deep Pacific  $\delta^{13}\text{C}$  must have been particularly sensitive during intervals of extreme cold in the North Atlantic.

Ages for isotopic events younger than 36 ka are well constrained by radiocarbon dating. Ages from 37 to 59 ka were tuned by correlation of  $\delta^{18}\text{O}$  maxima in W8709A-13PC to V23-81 stadial intervals. Therefore, in this interval, the correlation between the two records should be considered

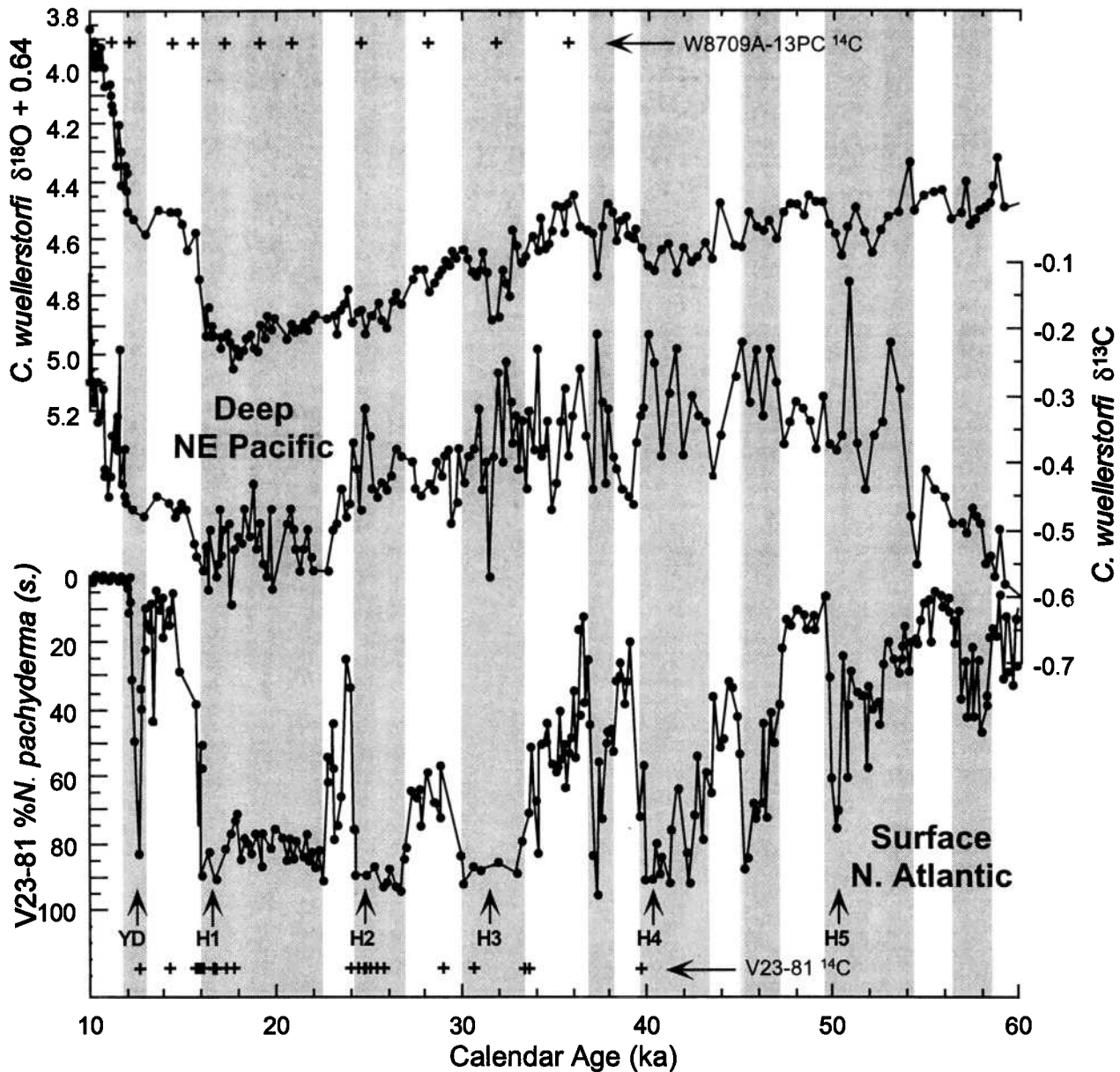
tentative. During cold episodes, however, it is likely that either (1) global ice volume increased or (2) deep waters became cooler. In either case,  $\delta^{18}\text{O}$  would increase, consistent with our method of matching  $\delta^{18}\text{O}$  maxima to stadial intervals. Also, older isotopic events were similar to those in the interval constrained by radiocarbon dating, both in signal amplitude and period, and dates determined by the correlation of  $\delta^{18}\text{O}$  maxima to V23-81 stadial intervals align well with  $^{14}\text{C}$  ages and the marine oxygen isotope stage 3-4 boundary in W8709A-13PC (Figure 1a).

Shifting ventilation patterns could potentially affect reservoir ages in the NE Pacific. Southon *et al.* [1990] calculated a NE Pacific surface water reservoir age in coastal upwelling systems of 1200 years at 6,400  $^{14}\text{C}$  years B.P., nearly 500 years older than the correction used in this paper. Altering the correlation of  $\delta^{18}\text{O}$  maxima to North Atlantic stadial conditions, however, would require reservoir age fluctuations on the order of 2 kyr (Figure 3). Benthic  $^{14}\text{C}$  ages were corrected using a reservoir age of 2380 years. Although this too may have changed through time, five of the six benthic foraminiferal dates in W8709A-13PC were cross-checked with a planktonic date in the same sample, with an average difference of <250 years [Ortiz *et al.*, 1997]. Since ~2 kyr changes in reservoir age are highly unlikely, we infer that millennial-scale isotopic enrichment events in the NE Pacific were essentially contemporary with (i.e., occurred within a few hundred years of) extreme cooling in the North Atlantic.

### 6.2. Oxygen Isotopes

Benthic foraminiferal  $\delta^{18}\text{O}$  is subject to both temperature and ice volume effects. If the ~0.2‰  $\delta^{18}\text{O}$  excursions in W8709A-13PC were entirely due to temperature, this would require deep water temperature oscillations of ~1°C (assuming a thermodynamic relationship of 0.22‰ per °C [Epstein *et al.*, 1953]). Is this reasonable? Deep Pacific temperatures during oxygen isotope stages 2 and 3 were ~2°C lower than at present, on the basis of comparison of benthic foraminiferal  $\delta^{18}\text{O}$  in Core V19-30 from the equatorial Pacific and sea level estimates from coral terraces in New Guinea [Chappell and Shackleton, 1986; Chappell *et al.*, 1996]. Today, potential temperature is ~1.5°C in the vicinity of W8709A-13PC and V19-30 [Broecker *et al.*, 1982]. Similar glacial  $\delta^{18}\text{O}$  values at the two sites [Shackleton *et al.*, 1983] imply that glacial deep water temperature at each location was near -0.5°C. Since the coral terraces which calibrate this estimate formed mostly during sea level high stands (low ice volume and depleted oceanic  $\delta^{18}\text{O}$ ), millennial-scale  $^{18}\text{O}$  enrichments in W8709A-13PC, if due to temperature alone, would suggest cooling to -1.5°C, close to the freezing point of seawater (-1.9°C). Results from a simple box model imply that this would require an unrealistic increase in deep water turnover rates [Mix and Pisias, 1988].

Alternatively, fluctuations in global ice volume could have influenced NE Pacific  $\delta^{18}\text{O}$ . A ~0.2‰  $\delta^{18}\text{O}$  oscillation is equivalent to about 20 m of sea level change [Fairbanks, 1990; Chappell and Shackleton, 1986]. Stage 3 sea level fluctuations of 20-30 m have been observed in coral terraces of New Guinea [Chappell *et al.*, 1996]. Also, several  $\delta^{18}\text{O}$  maxima in W8709A-13PC end near the time of ice-rafter events, when massive quantities of  $^{18}\text{O}$ -depleted ice



**Figure 3.** *C. wuellerstorfi*  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  in W8709A-13PC and the percentage of *Neogloboquadrina pachyderma* (s.) in North Atlantic Core V23-81 plotted versus calendar age. Shaded areas represent cold (stadial) intervals in the North Atlantic. Millennial-scale  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  maxima generally correspond to stadial conditions, and  $^{18}\text{O}$  depletions tend to occur near the time of ice-rafting events. Ages for Heinrich ice-rafting events from Bond *et al.* [1993] are given in calendar years. Calendar-corrected radiocarbon dates for W8709A-13PC (pluses) are marked on the upper age axis, and those for V23-81 (pluses) are marked on the lower age axis. Samples near turbidites in which sand grains composed >50% of the >150  $\mu\text{m}$  particles were excluded from Figure 3. This includes samples at 227, 230, 234, 252, 255, 258, 319, 322, 325, 355, 358, and 361 cm in W8709A-13PC. The isotope values at these depths are shown in Figure 2.

discharged into the North Atlantic (Figure 3). Thus, although temperature may account for a portion of each  $\sim 0.2\%$  oscillation, the timing and magnitude of  $\delta^{18}\text{O}$  shifts in the NE Pacific point to ice volume as a more probable explanation.

### 6.3. Carbon isotopes

Oscillations in benthic  $\delta^{13}\text{C}$  can be caused by changes in terrestrial biomass, organic carbon input from surface waters, air-sea gas exchange, and water mass composition. On

millennial timescales, peaks in  $\delta^{13}\text{C}$  generally occur during positive  $\delta^{18}\text{O}$  excursions, opposite the glacial-interglacial pattern in which low  $\delta^{13}\text{C}$  is associated with high  $\delta^{18}\text{O}$ . Results from the North Atlantic show that benthic  $^{13}\text{C}$  was generally depleted during stadial intervals of the last 60 ka [Oppo and Lehman, 1995], opposite the millennial-scale pattern in the NE Pacific. Given these temporal and regional differences, it seems unlikely that changes in terrestrial biomass caused the short-term  $\delta^{13}\text{C}$  fluctuations in W8709A-13PC.

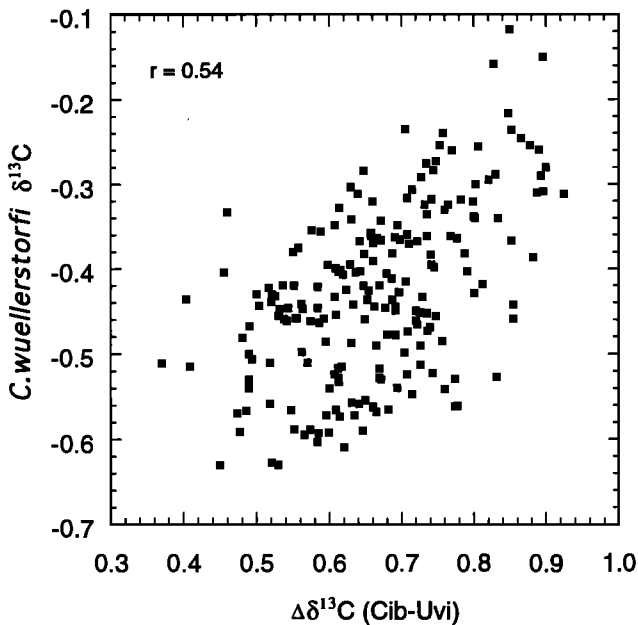


Figure 4.  $\Delta\delta^{13}\text{C}$  (*C. wuellerstorfi*  $\delta^{13}\text{C}$  - *Uvigerina* spp.  $\delta^{13}\text{C}$ ) versus *C. wuellerstorfi*  $\delta^{13}\text{C}$ . During intervals of enhanced organic carbon delivery,  $\Delta\delta^{13}\text{C}$  should increase. If the Mackensen effect drove *C. wuellerstorfi*  $\delta^{13}\text{C}$ ,  $\Delta\delta^{13}\text{C}$  and *C. wuellerstorfi*  $\delta^{13}\text{C}$  would be negatively correlated. Instead, there is a weak positive correlation ( $r = 0.54$ ).

Although *Cibicidoides* spp. are generally considered the most reliable foraminiferal recorders of deep water  $\delta^{13}\text{C}$  [Curry *et al.*, 1988], the  $\delta^{13}\text{C}$  signal may be complicated by organic carbon input to the surface sediments. For example, in the modern Antarctic, *Cibicidoides* spp. are sometimes depleted in  $^{13}\text{C}$  relative to bottom waters, probably as a result of epibenthic decay of  $^{13}\text{C}$ -depleted organic matter [Mackensen *et al.*, 1994]. To determine whether the so-called "Mackensen effect" altered the signal in W8709A-13PC, we plotted  $\delta^{13}\text{C}$  in *C. wuellerstorfi* versus the difference in  $\delta^{13}\text{C}$  between *C. wuellerstorfi* and *Uvigerina* spp. ( $\Delta\delta^{13}\text{C}$ ), measured in the same samples (Figure 4). Since *Uvigerina* spp. have an infaunal habitat [Zahn *et al.*, 1986] and *C. wuellerstorfi* have an epifaunal one [Corliss, 1985], higher organic carbon flux should result in higher  $\Delta\delta^{13}\text{C}$  [Zahn *et al.*, 1986]. Figure 4 shows that  $\delta^{13}\text{C}$  in *C. wuellerstorfi* has a weak positive correlation with  $\Delta\delta^{13}\text{C}$  ( $r = 0.54$ ), opposite the trend expected if the Mackensen effect drove local  $\delta^{13}\text{C}$  variations. Therefore, it appears that pore water effects have not significantly altered  $\delta^{13}\text{C}$  in *C. wuellerstorfi*.

The  $\delta^{13}\text{C}$  of deep water, and hence benthic foraminifera, can also be altered through surface ocean  $\text{CO}_2$  exchange with the atmosphere. At isotopic equilibrium, colder water has higher  $\delta^{13}\text{C}$  and warmer water has lower  $\delta^{13}\text{C}$  [Broecker and Maier-Reimer, 1992]. This thermodynamic effect ( $\delta^{13}\text{C}_{\text{as}}$ , for air-sea exchange) is independent of nutrient-related changes in  $\delta^{13}\text{C}$ . Glacial maximum reconstructions suggest that deep northwest Pacific  $\delta^{13}\text{C}_{\text{as}}$  was up to  $\sim 0.7\text{‰}$  higher than at present [Lynch-Stieglitz and Fairbanks, 1994]. If gas exchange effects were this large in the Pacific, the  $\sim 0.2\text{‰}$  millennial-scale  $\delta^{13}\text{C}$  oscillations in W8709A-13PC could be related to the extent of air-sea equilibration in North Pacific source waters. With  $\delta^{13}\text{C}$  data alone it would be

difficult to distinguish between greater gas exchange and lower nutrients, but both imply greater ventilation during cold events.

The simplest, and we think most likely, explanation for the  $^{13}\text{C}$  enrichments in W8709A-13PC is that NPDW was influenced by a  $^{13}\text{C}$ -enriched water mass during these episodes. Enrichment of  $^{13}\text{C}$  could have occurred through mixing with a "new" deep water source proximal to the NE Pacific with high preformed  $\delta^{13}\text{C}$  ( $> 0.2\text{‰}$ ). A glacial deep water source in the Pacific was proposed by Curry *et al.* [1988] to reconcile higher mean  $\delta^{13}\text{C}$  in the Pacific relative to the Southern Ocean during the last glacial maximum, by Ohkouchi *et al.* [1994] to explain low last glacial maximum Cd/Ca in the deep northwest Pacific, and by Lynch-Stieglitz and Fairbanks [1994] to account for global  $\delta^{13}\text{C}_{\text{as}}$  patterns during the last glaciation.

Fluctuations in residence time of PBW could also account for the  $\sim 0.2\text{‰}$   $\delta^{13}\text{C}$  peaks noted in W8709A-13PC. The longer a water mass resides in the deep sea, the more organic material degrades within it, consuming oxygen and depleting  $^{13}\text{C}$ . In the modern ocean, water mass "aging" accounts for the 0.4‰ depletion of  $\delta^{13}\text{C}$  of PBW relative to southern component waters [Curry *et al.*, 1988]. Increasing the turnover rates of deep and bottom waters in the North Pacific should decrease residence time and thus enrich local  $^{13}\text{C}$ . Deep water changes may have occurred sporadically during each interval of high  $\delta^{18}\text{O}$ , since  $\delta^{13}\text{C}$  is more variable during these times.

#### 6.4. Ventilation From Above

In the modern ocean, deep water forms in the North Atlantic but not in the North Pacific. The primary reason for this contrast is the precipitation-evaporation patterns in the two basins. In the North Pacific, precipitation greatly exceeds evaporation, whereas in the North Atlantic they are nearly balanced [Baumgartner and Reichel, 1975]. Low evaporation in the North Pacific causes a large surface to deep density contrast, limiting the sinking of surface water to intermediate depths [Warren, 1983].

The tendency for intermediate-depth ventilation to occur in the northwest Pacific [Reid, 1965; Talley, 1991, 1993] highlights this region as a candidate for deep water formation during glacial times. A similar potential exists in the Bering Sea, where the presence of chlorofluorocarbons in bottom waters implies that some ( $\sim 1$  Sv) deep water production may occur there today [Warner and Roden, 1995]. The NE Pacific seems an unlikely candidate as an ice age deep water source because the upper water column there was more stable at the last glacial maximum than today [Zahn *et al.*, 1991].

If deep water formed in the North Pacific during stadial intervals, the density contrast between surface and deep waters must have decreased relative to the modern ocean. One way to do this would be to change the precipitation-evaporation balance at the sea surface by altering atmospheric water vapor transport patterns. Today, there is net vapor transport from the Atlantic to the Pacific [Broecker, 1992]. Results from a coupled ocean-atmosphere general circulation model indicate that this pattern may change during intervals of low Atlantic thermohaline overturn. Surface cooling due to a diminished "conveyor" results in less net evaporation

from the northern Atlantic [Manabe and Stouffer, 1988]. Under these conditions, Atlantic to Pacific water vapor transport over Central America by tropical easterlies and over Asia by northern westerlies should decrease, favoring greater Pacific salinity, as modeled by Manabe and Stouffer [1988]. We call this mechanism the “salt seesaw” because of the tendency for salt to accumulate in the North Pacific, lessening water column stability, when NADW production is low.

A strengthened east Asian winter monsoon would increase evaporation rates in the western Pacific near southeast Asia. Input of these waters to the northwest Pacific via the Kuroshio Current would also enhance the likelihood of deep ventilation. Consistent with this idea, data from the Chinese loess plateau imply that the strength of the winter monsoon increased near the time of stadial events [Porter and Zhisheng, 1995]. Although box model results from Warren [1983] indicate that the salinity of the Kuroshio Current would need to increase by more than 1.5‰ relative to today for deep water formation to occur, this effect, in concert with reduced Atlantic to Pacific vapor transport, may have yielded the salinity necessary for deep thermohaline overturn in the North Pacific.

### 6.5. Ventilation From Below

In the South Pacific and South Atlantic, water at depths of 2-3 km spreads south toward Antarctica to contribute to the so-called Common Water, a major component of the world's deep water system [Broecker et al., 1985]. What causes this flow? One idea is that mid-depth waters are drawn southward to compensate northward Ekman transport around Antarctica. Water near the surface cannot flow south in the circumpolar region because of a lack of topographic barriers for geostrophic transport. Deeper water must counterbalance the Ekman process. Presently, NADW appears to be the primary source of this southward flow [Toggweiler and Samuels, 1995].

The demand for southward deep water advection at 2-3 km depth would occur whether or not NADW formed (such as during stadial intervals). At these times the burden of compensating flow may shift to the Pacific, enhancing overturn of its abyssal waters. This would decrease the residence time of NPDW and thus increase its  $\delta^{13}\text{C}$ . In this scenario, reduced buoyancy forcing in the North Atlantic, perhaps as a result of iceberg flux and melting, is automatically compensated by enhanced overturn in the deep Pacific. We call this mechanism the “Antarctic flywheel” because of the tendency for circumpolar winds to draw deep water from the Atlantic or the Pacific as necessary.

Unlike the salt seesaw, the flywheel mechanism does not require deep water formation in the North Pacific, just enhanced turnover of deep and bottom waters. These two mechanisms are not mutually exclusive. Both depend on reduced NADW formation and therefore could occur together. Isolating one of the two mechanisms will require comparison of W8709A-13PC  $\delta^{13}\text{C}$  to high-resolution records from intermediate and deep sites in the Pacific.

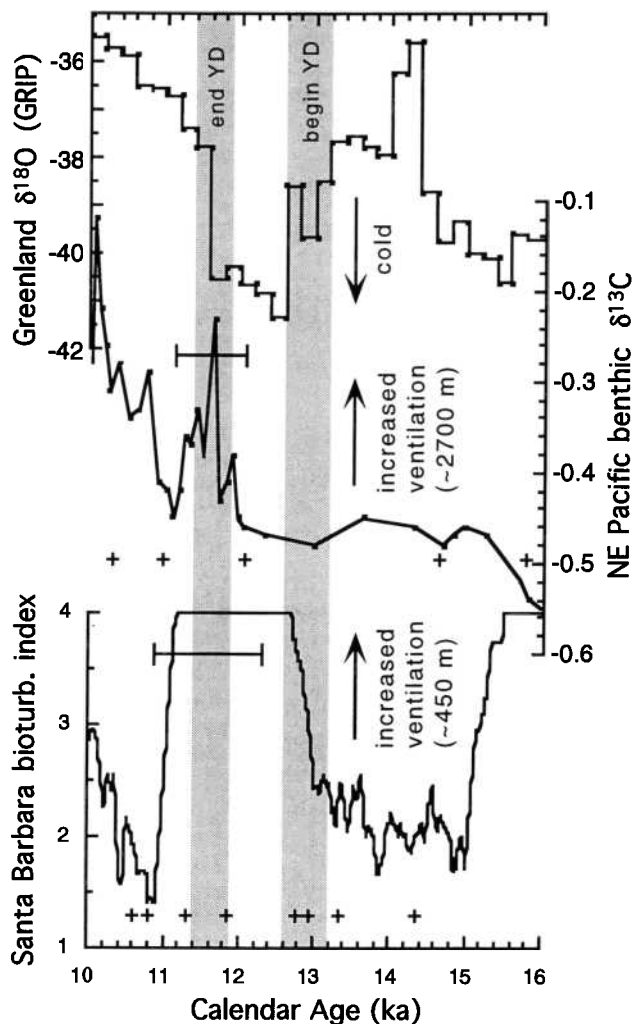


Figure 5. Comparison of W8709A-13PC  $\delta^{13}\text{C}$ , Santa Barbara Basin (Ocean Drilling Program (ODP) 893) bioturbation index [Behl and Kennett, 1996], and Greenland Ice Core Project (GRIP)  $\delta^{18}\text{O}$  [Dansgaard et al., 1993] during the last deglaciation. In the GRIP record the Younger Dryas (YD) begins at  $12,700 \pm 100$  years B.P. and ends at  $11,550 \pm 70$  years B.P. [Johnsen et al., 1992], consistent with the YD age range in the Greenland Ice Sheet Project (GISP) II ice core ( $12,940 \pm 250$  to  $11,640 \pm 250$  years B.P. [Alley et al., 1993]). Radio-carbon control points (pluses) for W8709A-13PC and ODP 893 are indicated next to each time series. The  $\delta^{13}\text{C}$  record reveals an interval of enhanced deep NE Pacific ventilation centered at 11.6 ka, near the end of the YD. This age range for this  $\delta^{13}\text{C}$  peak (11.1-12.1 ka indicated by a horizontal bar), constrained by the age model options in Figure 1b, indicates that deep NE Pacific ventilation lagged the YD initiation and enhanced ventilation in the Santa Barbara Basin by at least 500 years. Age uncertainty for the termination of the YD ventilation event in the Santa Barbara Basin (11.0-12.3 ka [Kennett and Ingram, 1995] shown here by a horizontal bar) precludes a detailed comparison to deep Pacific  $\delta^{13}\text{C}$  in this interval, but it is likely that the ventilation events overlap.

### 6.6. Comparison to Shallower Pacific Records

During the last deglaciation, millennial fluctuations in oxygenation occurred both at 700 m depth in the Gulf of California [Keigwin and Jones, 1990] and at 450 m sill depth of the Santa Barbara Basin [Kennett and Ingram, 1995].



Careful comparison of W8709A-13PC  $\delta^{13}\text{C}$  and the Santa Barbara record during the Younger Dryas event shows that the increase in deep NE Pacific ventilation lagged that of the intermediate-depth NE Pacific by at least 500 years, although the ventilation events overlap near the Younger Dryas termination (Figure 5). This pattern is consistent with a joint salt seesaw and Antarctic flywheel scenario in which changes in NADW production first affect the intermediate-depth NE Pacific via the atmosphere and then the deep NE Pacific via fluctuations in global thermohaline circulation.

Benthic foraminiferal Cd/Ca results from the California margin (core F8-90-G21) imply that ventilation decreased at 1600 m depth near the time of the Younger Dryas [van Geen *et al.*, 1996]. While  $\delta^{13}\text{C}$  in W8709A-13PC does not decrease noticeably in this interval, a direct comparison of the two records may be misleading, due to dating uncertainties from 11 to 16 ka in F8-90-G21 [van Geen *et al.*, 1996]. Also, the deglacial  $^{14}\text{C}$  plateau makes it difficult to compare time series unless the stratigraphic position of the plateau is constrained.

From 15 to 60 ka, nonlaminated intervals representing periods of high oxygen content in the Santa Barbara Basin seem to correlate with cold events in Greenland [Behl and Kennett, 1996]. Since we show that ventilation of the deep NE Pacific generally increased during stadial conditions, ventilation of the intermediate and deep Pacific may have been in phase during this time. Determining if the apparent ventilation lag during the Younger Dryas holds for earlier climatic events will require better age control in both W8709A-13PC and Ocean Drilling Program (ODP) 893 from 15 to 60 ka. Independent of chronology, however, our finding that millennial-scale  $\delta^{13}\text{C}$  maxima generally occur during intervals of high  $\delta^{18}\text{O}$  (ice advance and deep-sea cooling) in W8709A-13PC implies that the deep NE Pacific was better ventilated during stadial than interstadial conditions.

## References:

- Alley, R. B. *et al.*, Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event, *Nature*, 362, 527-529, 1993.
- Bard, E., M. Arnold, and B. Hamelin, Present status of the radiocarbon calibration for the late Pleistocene (abstract), *GEOMAR Rep.*, 15, 52-53, 1992.
- Bard, E., F. Rostek, and C. Sonzogni, Interhemispheric synchrony of the last deglaciation inferred from alkenone palaeothermometry, *Nature*, 385, 707-710, 1997.
- Baumgartner, A., and E. Reichel, *The World Water Balance: Mean Annual Global, Continental and Maritime Precipitation, Evaporation and Run-off*, 179 pp. Elsevier, New York, 1975.
- Behl, R. J., and J. P. Kennett, Brief interstadial events in the Santa Barbara Basin, NE Pacific, during the past 60 kyr, *Nature*, 379, 243-246, 1996.
- Bond, G., and R. Lotti, Iceberg discharges into the North Atlantic on millennial timescales during the last glaciation, *Science*, 267, 1005-1010, 1995.
- Bond, G., W. Broecker, S. Johnsen, J. McManus, L. Labeyrie, J. Jouzel, and G. Bonami, Correlations between climate records from North Atlantic sediment and Greenland Ice, *Nature*, 365, 143-147, 1993.
- Broecker, W. S., The Great Ocean Conveyor, in *AIP Conference Proceedings 247: Global Warming: Physics and Facts*, edited by B. G. Levi *et al.*, pp. 129-161, Am. Inst. of Phys., College Park, Md., 1992.
- Broecker, W. S., and E. Maier-Reimer, The influence of air and sea exchange on the carbon isotope distribution in the sea, *Global Biogeochem. Cycles*, 6, 315-320, 1992.
- Broecker, W. S., D. W. Spencer, and H. Craig (Eds.), *GEOSecs Pacific Expedition*, Vol. 3, *Hydrographic Data 1973-1974*, Nat. Sci. Found., Washington, D. C., 1982.
- Broecker, W. S., T. Takahashi, and T. Takahashi, Sources of flow patterns of deep-ocean waters as deduced from potential temperature, salinity, and initial phosphate concentration, *J. Geophys. Res.*, 90, 6925-6939, 1985.
- Chappell, J., and N. J. Shackleton, Oxygen isotopes and sea level, *Nature*, 324, 137-140, 1986.
- Chappell, J., A. Omura, T. Esat, M. McCulloch, J. Pandolfi, Y. Ota, and B. Pillans, Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records, *Earth Planet. Sci. Lett.*, 141, 227-236, 1996.
- Corliss, B. H., Microhabitats of benthic foraminifera within deep-sea sediments, *Nature*, 314, 435-438, 1985.
- Curry, W. B., and D. W. Oppo, Synchronous, high-frequency oscillations in tropical sea surface temperatures and North Atlantic Deep Water production during the last glacial cycle, *Paleoceanography*, 12, 1-14, 1997.
- Curry, W. B., J. C. Duplessy, L. D. Labeyrie, and N. J. Shackleton, Changes in the distribution of  $\delta^{13}\text{C}$  of deep water  $\Sigma\text{CO}_2$  between the last glaciation and the Holocene, *Paleoceanography*, 3, 317-341, 1988.
- Dansgaard, W. *et al.*, Evidence for general instability of climate from a 250-kyr ice-core record, *Nature*, 364, 218-220, 1993.
- Edwards, R. L., J. W. Beck, G. S. Burr, D. J. Donahue, J. M. A. Chappell, A. L. Bloom, E. R. M. Druffel, F. W. Taylor, A large drop in atmospheric  $^{14}\text{C}/^{12}\text{C}$  and reduced melting in the Younger Dryas, documented with  $^{230}\text{Th}$  ages of corals, *Science*, 260, 962-968, 1993.

## 7. Conclusions

Stable isotopic variations recorded by *C. wuellerstorfi* in a core from the deep northeast Pacific outline repeated millennial-scale deep water oscillations during the last glacial period. Intervals of high  $\delta^{18}\text{O}$  appear to correlate to North Atlantic stadial events, and are probably partially controlled by ice volume, implying  $\delta^{18}\text{O}$  has potential as a high-resolution stratigraphic index. Enrichment of  $^{13}\text{C}$  during  $\delta^{18}\text{O}$  maxima suggests that intervals of strong ventilation in the deep Pacific occurred during ice advances, opposite the pattern in the deep North Atlantic. Thus, during the last glacial period, there appears to be an alternating relationship between deep ventilation in the Pacific and the Atlantic on millennial timescales.

This relationship could be a function of two separate mechanisms, both of which are internal to Earth's climate system. The first involves salt buildup and cooling in the North Pacific, decreased water column stability, and ventilation of the deep Pacific from above when the North Atlantic cools. We call this mechanism the salt seesaw. Alternatively, ventilation may have occurred from below, because of increased Pacific outflow compensating wind-driven northward advection in the Southern Ocean. We call this mechanism the Antarctic flywheel. While data from intermediate depths in the Pacific (e.g., Santa Barbara Basin) are consistent with the salt seesaw scenario, the apparent lag (>500 years) of deep water ventilation behind intermediate water oxygenation during the Younger Dryas event suggests that the two mechanisms may have acted together.

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- Epstein, S., R. Buchsbaum, H. A. Lowenstam, and H. C. Urey, Revised carbonate water isotopic temperature scale, *Geol. Soc. Am. Bull.*, 64, 1315-1326, 1953.
- Fairbanks, R. G., The age and origin of the "Younger Dryas Climate Event" in Greenland ice cores, *Paleoceanography*, 5, 937-948, 1990.
- Gardner, J. V., W. E. Dean, and P. Dartnell, Biogenic sedimentation beneath the California Current system for the past 30 kyr and its paleoceanographic significance, *Paleoceanography*, 12, 207-225, 1997.
- Grimm, E. C., G. L. Jacobson Jr., W. A. Watts, B. C. S. Hansen, and K. A. Maasch, A 50,000-year record of climate oscillations from Florida and its temporal correlation with Heinrich events, *Science*, 261, 198-200, 1993.
- Grootes, P. M., M. Stuvier, J. W. White, W. C. Johnsen, and J. Jouzel, Comparison of oxygen isotope records from the GISP II and GRIP Greenland ice cores, *Nature*, 366, 552-554, 1993.
- Imbrie, J., J. D. Hays, D. G. Martinson, A. McIntyre, A. C. Mix, J. J. Morley, N. G. Pisias, W. L. Prell, and N. J. Shackleton, The orbital theory of Pleistocene climate: Support from a revised chronology of the marine record, in *Milankovitch and Climate, Part 1*, edited by A. L. Berger et al., pp. 269-305, D. Reidel, Norwell, Mass., 1984.
- Ingram, B. L., and J. P. Kennett, Radiocarbon chronology and planktonic-benthic foraminiferal  $^{14}\text{C}$  age differences in Santa Barbara Basin sediments, Hole 893A, edited by J. P. Kennett, J. G. Baldauf, and M. Lyle, *Proc. Ocean Drill. Program Sci. Results*, Vol. 146, 19-27, 1995.
- Johnsen, S. J., H. B. Clausen, W. Dansgaard, K. Fuhrer, N. Gundestrup, C. U. Hammer, P. Iversen, J. Jouzel, B. Stauffer, and J. P. Steffensen, Irregular glacial interstadials recorded in a new Greenland ice core, *Nature*, 359, 311-313, 1992.
- Karlin, R., M. Lyle, and R. Zahn, Carbonate variations in the northeast Pacific during the late Quaternary, *Paleoceanography*, 7, 43-61, 1992.
- Keigwin, L. D., and G. A. Jones, Deglacial climatic oscillations in the Gulf of California, *Paleoceanography*, 5, 1009-1023, 1990.
- Keigwin, L. D., and G. A. Jones, Western North Atlantic evidence for millennial-scale changes in ocean circulation and climate, *J. Geophys. Res.*, 99, 12,397-12,410, 1994.
- Kennett, J. P., and B. L. Ingram, A 20,000-year record of ocean circulation and climate change from the Santa Barbara basin, *Nature*, 377, 510-514, 1995.
- Kroopnick, P. M., The distribution of  $^{13}\text{C}$  of  $\text{CO}_2$  in the world oceans, *Deep Sea Res.*, 32, 57-84, 1985.
- Lehman, S. J., and L. D. Keigwin, Sudden changes in North Atlantic circulation during the last deglaciation, *Nature*, 356, 757-762, 1992.
- Lyle, M., R. Zahn, F. Prah, J. Dymond, R. Collier, N. Pisias, and E. Suess, Paleoproductivity and carbon burial across the California Current: The Multitracers Transect, 42°N, *Paleoceanography*, 7, 251-272, 1992.
- Lynch-Stieglitz, J., and R. G. Fairbanks, A conservative tracer for glacial ocean circulation from carbon isotope and palaeo-nutrient measurements in benthic foraminifera, *Nature*, 369, 308-310, 1994.
- Mackensen, A., H. Grobe, H. W. Hubberten, and G. Kuhn, Benthic foraminiferal assemblages and the  $\delta^{13}\text{C}$ -signal in the Atlantic sector of the Southern Ocean: Glacial-to-interglacial contrasts, in *Carbon Cycling in the Glacial Ocean: Constraints on the Ocean's Role in Global Change, NATO ASI Series*, Vol. 1, 17, edited by R. Zahn et al., pp. 105-144, Springer-Verlag, New York, 1994.
- Manabe, S., and R. J. Stouffer, Two stable equilibria of a coupled ocean-atmosphere model, *J. Clim.*, 1, 841-866, 1988.
- Mantyla, A. W., On the potential temperature in the abyssal Pacific Ocean, *J. Mar. Res.*, 33, 341-354, 1975.
- Mantyla, A. W., and J. L. Reid, Abyssal characteristics of the World Ocean waters, *Deep Sea Res.*, 30, 805-833, 1983.
- Mikolajewicz, U., T. J. Crowley, A. Schiller, and R. Voss, Modeling teleconnections between the North Atlantic and North Pacific during the Younger Dryas, *Nature*, 387, 384-387, 1997.
- Mix, A. C., and N. G. Pisias, Oxygen isotope analyses and deep-sea temperature changes: Implications for rates of oceanic mixing, *Nature*, 331, 249-251, 1988.
- Ohkouchi, N., H. Kawahata, M. Murayama, M. Okada, T. Nakamura, and A. Taira, Was deep water formed in the North Pacific during the late Quaternary? Cadmium evidence from the northwest Pacific, *Earth Planet. Sci. Lett.*, 124, 185-194, 1994.
- Oppo, D. W., and S. J. Lehman, Suborbital timescale variability of North Atlantic Deep Water during the past 200,000 years, *Paleoceanography*, 10, 901-910, 1995.
- Ortiz, J. D., A. C. Mix, S. Hostetler, and M. Kashgarian, The California Current of the last glacial maximum: Reconstruction at 42°N based on multiple proxies, *Paleoceanography*, 12, 191-205, 1997.
- Porter, S. C., and A. Zhiseng, Correlation between climate events in the North Atlantic and China during the last glaciation, *Nature*, 375, 305-308, 1995.
- Reid, J. L., Jr., *Intermediate waters of the Pacific Ocean*, 85 pp., Johns Hopkins Univ. Press, Baltimore, Md., 1965.
- Shackleton, N. J., Carbon-13 in *Uvigerina*: Tropical rainforest history and the equatorial Pacific carbonate dissolution cycles, in *The Fate of Fossil Fuel  $\text{CO}_2$  in the Oceans*, edited by N. R. Andersen and A. Malahoff, pp. 401-427, Plenum Press, New York, 1977.
- Shackleton, N. J., J. Imbrie, and M. A. Hall, Oxygen and carbon isotope record of East Pacific core V19-30: Implications for the formation of deep water in the late Pleistocene North Atlantic, *Earth Planet. Sci. Lett.*, 65, 233-244, 1983.
- Southon, J. R., D. E. Nelson, and J. S. Vogel, A record of past ocean-atmosphere radiocarbon differences from the northeast Pacific, *Paleoceanography*, 5, 197-206, 1990.
- Stuvier, M., and T. F. Braziunas, Modeling atmospheric  $^{14}\text{C}$  influences and  $^{14}\text{C}$  ages of marine samples to 10,000 BC, *Radiocarbon*, 35, 137-189, 1993.
- Talley, L. D., An Okhotsk Sea water anomaly: Implications for ventilation in the North Pacific, *Deep Sea Res.*, 38, 171-190, 1991.
- Talley, L. D., Distribution and formation of the North Pacific intermediate water, *J. Phys. Oceanogr.*, 23, 517-537, 1993.
- Toggweiler, J. R., and B. Samuels, Effect of Drake Passage on the global thermohaline circulation, *Deep Sea Res.*, 42, 477-500, 1995.
- van Geen, A., R. G. Fairbanks, P. Dartnell, M. McGann, J. V. Gardner, and M. Kashgarian, Ventilation changes in the northeast Pacific during the last deglaciation, *Paleoceanography*, 11, 519-528, 1996.
- Warner, M. J., and G. I. Roden, Chlorofluorocarbon evidence for recent ventilation of the deep Bering Sea, *Nature*, 373, 409-412, 1995.
- Warren, B. A., Why is no deep water formed in the North Pacific?, *J. Mar. Res.*, 41, 327-347, 1983.
- Zahn, R., K. Winn, and M. Sarnthein, Benthic foraminiferal  $\delta^{13}\text{C}$  and accumulation rates of organic carbon: *Uvigerina peregrina* group and *Cibicidoides wuellerstorfi*, *Paleoceanography*, 1, 27-42, 1986.
- Zahn, R., T. F. Pedersen, B. D. Bornhold, and A. C. Mix, Water mass conversion in the glacial subarctic Pacific (54°N, 148°W): Physical constraints and the benthic-planktonic stable isotope record, *Paleoceanography*, 6, 543-560, 1991.

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