

WATER MASS CONVERSION IN THE GLACIAL  
SUBARCTIC PACIFIC (54°N, 148°W): PHYSICAL  
CONSTRAINTS AND THE BENTHIC-  
PLANKTONIC STABLE ISOTOPE RECORD

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**Abstract.** Benthic (*Uvigerina* spp., *Cibicidoides* spp., *Gyroidinoides* spp.) and planktonic (*N. pachyderma* sinistral, *G. bulloides*) stable isotope records from three core sites in the central Gulf of Alaska are used to infer mixed-layer and deepwater properties of the late glacial Subarctic Pacific. Glacial-interglacial amplitudes of the planktonic  $\delta^{18}\text{O}$  records are 1.1-1.3‰, less than half the amplitude observed at core sites at similar latitudes in the North Atlantic; these data imply that a strong, negative  $\delta_w$  anomaly existed in the glacial Subarctic mixed layer during the summer, which points to a much stronger low-salinity anomaly than exists today. If true, the upper water column in the North Pacific would have been statically more stable than today, thus suppressing convection even more efficiently. This scenario is

further supported by vertical (i.e., planktic versus benthic)  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  gradients of  $>1\text{‰}$ , which suggest that a thermohaline link between Pacific deep waters and the Subarctic Pacific mixed layer did not exist during the late glacial. Epibenthic  $\delta^{13}\text{C}$  in the Subarctic Pacific is more negative than at tropical-subtropical Pacific sites but similar to that recorded at Southern Ocean sites, suggesting ventilation of the deep central Pacific from mid-latitude sources, e.g., from the Sea of Japan and Sea of Okhotsk. Still, convection to intermediate depths could have occurred in the Subarctic during the winter months when heat loss to the atmosphere, sea ice formation, and wind-driven upwelling of saline deep waters would have been most intense. This would be beyond the grasp of our planktonic records which only document mixed-layer temperature-salinity fields extant during the warmer seasons. Also we do not have benthic isotope records from true intermediate water depths of the Subarctic Pacific.

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## INTRODUCTION

Growing interest in global climatic fluctuations and the need for more precise modeling of past and future climatic change has increased the demand for a better understanding of how the ocean's physical circulation and its chemical cycling have changed during glacial-interglacial times. In the search for possible mechanisms to explain the link between glacial-interglacial variations of deep-ocean ventilation and high-latitude surface-ocean processes, paleoceanographers have so

far concentrated mostly on the North Atlantic [Mix and Fairbanks, 1985; Labeyrie and Duplessy, 1985; Boyle and Keigwin, 1987; Curry et al., 1988; Duplessy et al., 1988] and the Southern Ocean [Labeyrie et al., 1987; Boyle, 1988; Charles and Fairbanks, 1990].

Paleoceanographic studies in the high-latitude North Pacific lag far behind those in the North Atlantic. To date, Keigwin's [1987] benthic isotope record from core RAMA 44 from the Meiji Tongue is the only isotope record available for the Subarctic Pacific north of 50°N (Figure 1). Major obstacles to obtaining further records are the vast turbidite sequences found in the far northeastern Pacific and the scarcity of carbonates due to the relatively shallow carbonate compensation depth (CCD). Both factors make the pursuit of stratigraphic and paleoenvironmental work in the Subarctic Pacific difficult (*e.g.*, McCoy and Sancetta [1985]). However, the North Pacific contains more than 25% of the world ocean volume and, in the absence of local deepwater formation, is a major carbon reservoir [Kroopnick, 1985]. Thus variations through time in the North Pacific ocean system may have profound effects not only in redistributing carbon within the world ocean but also in enhancing variations of the ocean-atmosphere carbon flux and, ultimately, fluctuations of global climate.

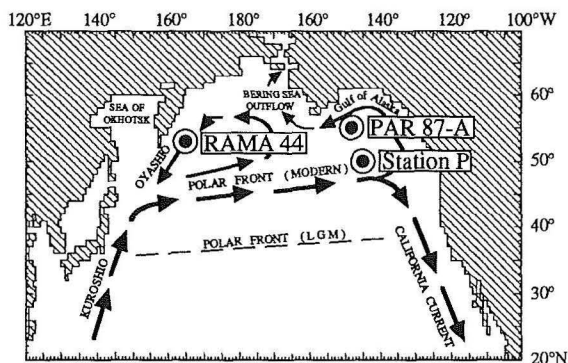


Fig. 1. Position of Parizeau cores (labeled PAR 87-A) in the Subarctic Pacific. Also shown are positions of Ocean Station P (50°N, 145°W), Subarctic core RAMA 44 [Keigwin, 1987], and major current systems.

In this study we present the first stable isotope data for the Subarctic east Pacific which have been obtained from sediment cores from the Patton-Murray Seamount Group in the central Gulf of Alaska (Figure 1). These seamounts comprise an elevated plateau area which rises above the regional turbidite sequences and the CCD. Thus the stratigraphy of the

sediment cores is intact and foraminifera are abundant. The isotope data are used to derive the history of water mass formation and distribution in the Gulf of Alaska region over the past 25,000 years.

## SOURCES AND FLOW PATTERN OF WATER MASSES IN THE NORTH PACIFIC

### Modern

The low salinity of the upper layer waters in the North Pacific inhibits convection. The low salinity is typically limited to the upper 100 m of the water column (Figure 2) and results from only minor advective flow of tropical-subtropical saline surface waters to the north coupled with low evaporation rates [Warren, 1983]. Surface waters are buoyant and deep convection does not occur [Warren, 1983]; even during winter convection does not occur when severe cooling, enhanced wind stress, and low river discharge prevail [Luick et al., 1987].

In the absence of local sources, the Pacific imports its deep waters from the North Atlantic and the Southern Ocean. By the time these waters reach the North Pacific, the oxygen content of the deep waters is extremely low (Figure 2), and concentrations of total dissolved carbon dioxide ( $\Sigma\text{CO}_2$ ) and biologically cycled nutrients are high. Because of these properties, deep waters of the North Pacific have been referred to as the oldest water of the world ocean [Kroopnick, 1985; Broecker et al., 1988]. Underlying bottom waters in the North Pacific are better ventilated (Figure 2), documenting the advection of Antarctic Bottom Water.

While no deepwater convection occurs at present in the Subarctic Pacific, the area is nevertheless a site of limited intermediate water formation. Mid-depth isopycnal surfaces are not directly ventilated by convection of Subarctic surface waters. Instead, intermediate water properties are largely determined by diapycnal diffusion [Tully and Barber, 1960; Reid, 1965]. Recent hydrographic evidence for ventilation of the Sea of Japan and Sea of Okhotsk on the same isopycnal surfaces as found in the mid-depth North Pacific has called attention to these marginal seas as sources for direct ventilation of North Pacific intermediate waters [Ohtani and Nagata, 1990; Riser, 1990; Talley, 1990]. The very narrow seaways connecting the Sea of Japan and Sea of Okhotsk with the open North Pacific, however, strongly limit the water exchange so that the contribution from these marginal seas alone probably cannot explain the properties of North Pacific Intermediate Water. This is true even if the highest possible advective flow out of these regions to the North Pacific is assumed (Stephen Riser, personal communication, 1990).

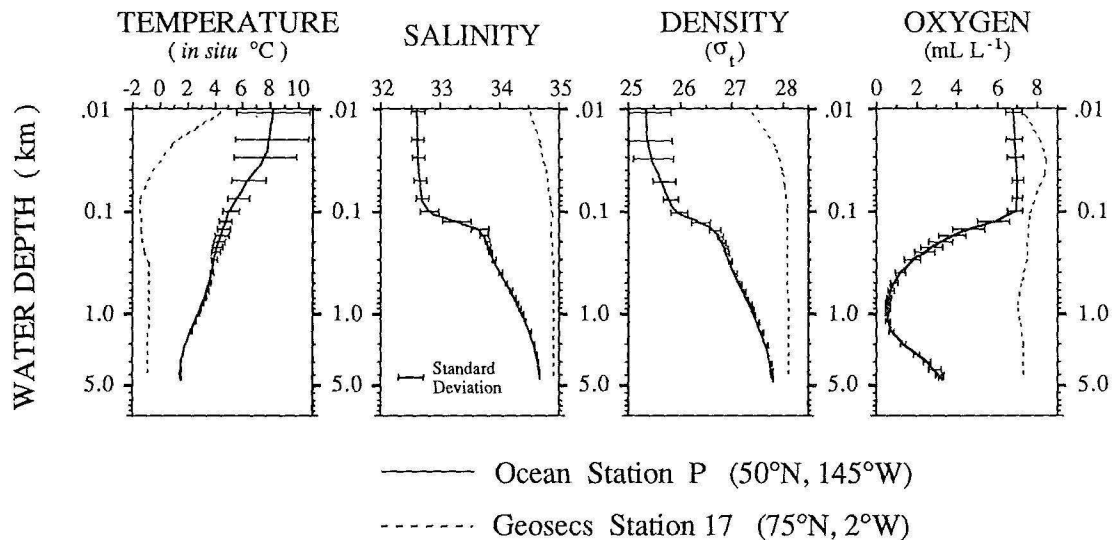


Fig. 2. Vertical profiles of temperature, salinity, density, and oxygen obtained from long-term (1951-1981) observations at Ocean Station P [Tabata and Peart, 1985]. Hydrographic profiles from Geosecs Station 17 in the Norwegian-Greenland Sea are shown for comparison. Note that the pycnocline in the Subarctic Pacific is mainly a function of the sharp halocline whereas in the Norwegian-Greenland seas it is controlled by the strong thermocline.

### Last Glaciation

The surface hydrography of the North Atlantic changed dramatically during glacial-interglacial transitions. The CLIMAP [1981] research group provided convincing evidence for a southward shift of the North Atlantic polar front to a nearly zonal position at 40°N at the last glacial maximum (LGM). As a result, convection cells shifted to the south from their modern positions in the Norwegian-Greenland seas, and thermohaline overturning probably started in the North Atlantic [Boyle and Keigwin, 1987; Duplessy et al. 1988]. CLIMAP [1981] also inferred (based on much less data though) a southward migration of the polar front in the North Pacific, from 50°N at present to ~40°N at the LGM (Figure 1). This shift was associated with a decrease in sea surface temperature (SST) in the Subarctic region north of 50°N, which was presumably largest ( $\Delta$ SST~3°C) in the east [CLIMAP, 1981].

Circumstantial evidence for glacial-interglacial changes in the North Pacific deepwater circulation comes, for instance, from <sup>230</sup>Th/<sup>231</sup>Pa data [Mangini et al., 1982] and seismic reflection profiles [Damuth et al., 1983; Mammerickx, 1985; Mayer et al., 1987] which indicate enhanced sediment erosion in the deep Pacific during glacial times and possibly document increased bottom water flow. Dean et al. [1989] suggest that glacial-age oxidized brown clays north of 50°N reflect a change of redox conditions, presumably indicating increased deepwater ventilation. This

could occur through either enhanced advection of AABW or formation of deepwater locally in the North Pacific. In contrast, Keigwin [1987] was the first to provide a detailed benthic  $\delta^{13}\text{C}$  record from a high-latitude North Pacific sediment core which clearly showed that glacial  $\delta^{13}\text{C}$  was more negative than modern values. These data are corroborated by benthic  $\delta^{13}\text{C}$  profiles throughout the deep Pacific (see Boyle and Keigwin [1987], and data compiled in Curry et al. [1988] and Duplessy et al. [1988]). Together with greater radiocarbon ages of the glacial Pacific deep waters [Shackleton et al., 1988; Broecker et al., 1988], these data suggest that ventilation of the deep Pacific would have actually been decreased during glacial times.

Benthic  $\delta^{13}\text{C}$  patterns at Pacific intermediate water core sites are also inconclusive. Glacial  $\delta^{13}\text{C}$  values, which are the same as or more positive than today, have been obtained from tropical (5°S-1°N) Pacific core sites at depths between 0.7 and 1.7 km [Duplessy et al., 1988] and imply enhanced middepth ventilation during glacials. Benthic  $\delta^{13}\text{C}$  at South China Sea core V35-05 (1950 m [Oppo and Fairbanks, 1987]) and eastern equatorial Pacific site V19-27 (1693 m [Mix et al., 1991]), on the other hand, varies coherently with the deep-ocean benthic  $\delta^{13}\text{C}$  record; that is, benthic  $\delta^{13}\text{C}$  at these sites is more negative during glacial stages. These data imply less extreme increases in  $\delta^{13}\text{C}$  gradients between Pacific intermediate and deep waters than observed, e.g., in the North Atlantic. Thus the data available to



date are inconclusive, and even contradictory, with regard to changes in the intermediate and deepwater circulation of the Pacific Ocean.

The intention of this study is to examine whether convective overturn in the glacial Pacific is physically plausible. This will be done by predicting vertical density profiles needed to start convection and using Warren's surface-layer salinity box model for the North Pacific. The changes of the surface-water temperature-salinity field required by the model will be translated into hypothetical, glacial-interglacial amplitudes of planktonic  $\delta^{18}\text{O}$ . The model output will then be tested using measured planktonic  $\delta^{18}\text{O}$  data. Vertical, i.e., benthic-planktonic isotope gradients will be examined as a further test whether or not a thermohaline link existed between the deep and shallow North Pacific during the last glacial.

#### REGIONAL PHYSIOGRAPHY AND SEDIMENTARY RECORDS

The Patton-Murray Seamount Group is the largest of the elevated plateau areas in the northeastern Pacific, covering >17,000 km<sup>2</sup> above 3.7 km water depth. An extensive seismic survey, which was carried out prior to coring using both single-channel air gun and water gun profiling, revealed a thick (up to 500 ms two-way travel time) sequence of presumed hemipelagic sediments over much of the area. The plateau sequence is acoustically mainly transparent and is, as such, in marked contrast to the acoustic character of the massive turbidite sequences in the adjacent abyssal plain.

The upper sedimentary record of cores PAR 87A-01, PAR 87A-02, and PAR 87A-10 consists of medium to fine silt and silty clay with intercalated layers of ice-rafted debris (IRD). Carbonate contents range from 10-60% by weight at the shallower core site PAR 87A-02 to 5-35% at the deeper sites PAR 87A-01 and PAR 87A-10. Lowest carbonate contents are observed during the last glacial-interglacial climatic transition and are probably the result of massive incursions of IRD. The oxygen isotope stratigraphies show the sedimentary record of the past 25,000 years at all three core sites are complete, with no indication of hiatuses or turbidites.

#### STABLE ISOTOPE ANALYSIS

##### Analytical Methods

We have obtained stable isotope records for the upper sections of cores PAR 87A-01, PAR 87A-02, and PAR 87A-10 (hereafter referred to as cores 01, 02, and 10; Table 1) from the Patton-Murray Seamount Group, which have been sampled at 5 cm intervals. Isotope analyses were performed at the University of British Columbia on a VG Isogas Prism mass spectrometer. All samples were mechanically crushed, sonicated, and roasted in vacuo at 430°C for 30 min prior to analysis. Carbonate dissolution occurred at 90°C in 4 cm<sup>3</sup> orthophosphoric acid using an online VG Isogas Autocarb automated common-acid-bath device. Calibration of the spectrometer to the Pee Dee Belemnite (PDB) carbonate standard was done through National Institute of Standards and Technology "NBS" carbonate standards 19 and 20. External precision (i.e., including gas preparation and spectrometric procedures) was better than 0.04‰ and 0.03‰ respectively for  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  (1 $\sigma$ , 73 replicates of an internal, laboratory standard calcite run during the period of the measurements). Six replicate samples of *G. bulloides* (20 specimens each) from a sediment sample at 50 cm depth in core 10 yielded a 1 $\sigma$  standard deviation of 0.06‰ and 0.07‰ for  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ , respectively. Poorer precision is expected for samples of benthic foraminifera which contained fewer specimens (see below).

##### Selection of Foraminiferal Species and Experimental Strategy

Stable isotope measurements were carried out on planktonic foraminifers *Globigerina bulloides* (250-300  $\mu\text{m}$ ) and *Neogloboquadrina pachyderma* (sinistral) (180-212  $\mu\text{m}$ ), and on benthic foraminifers *Uvigerina* spp. (mostly *U. senticosa*) (250-350  $\mu\text{m}$ ), *Cibicidoides* spp. (*C. wuellerstorfi* and *C. kullenbergi*) (250-350  $\mu\text{m}$ ), and *Gyroidinoides* spp. (>425  $\mu\text{m}$ ). The isotope samples contained 30 planktonic specimens and 1-8 benthic specimens on average. For the benthic records, analysis of three different genera was necessary because *Cibicidoides* spp. and

TABLE 1. Position of Sediment Cores and Average Sedimentation Rates

Core	Latitude	Longitude	Water Depth	Sedimentation Rate
PAR 87A-01	54° 24.9'N	149° 25.9'W	3480 m	7.4 cm kyr <sup>-1</sup>
PAR 87A-02	54° 17.4'N	149° 36.3'W	2920 m	9.7 cm kyr <sup>-1</sup>
PAR 87A-10	54° 21.8'N	148° 28.0'W	3664 m	6.5 cm kyr <sup>-1</sup>



*Uvigerina* spp., which are commonly used in stable isotope paleoceanography, were not continuously available along the cores. The benthic oxygen isotope data are reported on the *Uvigerina* scale of  $\delta^{18}\text{O}$  [Shackleton, 1974]. Values of *Cibicidoides* spp. were converted to the *Uvigerina* scale by adding 0.64‰ [Shackleton and Opdyke, 1973]. No correction was applied to  $\delta^{18}\text{O}$  data from *Gyroidinoides* spp. as they are similar to those obtained from *Uvigerina* spp. (Figure 3a). All isotope data are listed in Tables 2, 3, and 4.

$\delta^{13}\text{C}$  values of endobenthic *Uvigerina* spp. presumably record the  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  of pore waters in the surface sediments [McCorkle et al., 1990]. The  $\delta^{13}\text{C}$  values of *Gyroidinoides* spp. are similar to those of *Uvigerina* spp. (Figure 3b), implying that both genera inhabit similar, endobenthic microhabitats.  $\delta^{13}\text{C}$  from epibenthic *Cibicidoides* spp., on the other hand, is offset from that of the endobenthic genera (Figure 3b). It is assumed to record closely variations in deepwater  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  and is used here as a tracer for water mass variation, i.e., deep-ocean ventilation [see Curry et al., 1988 and references therein].

Epibenthic  $\delta^{13}\text{C}$  fluctuations, however, do not uniquely document changes in relative fluxes of different water masses. They may also be affected by  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  variations of the source waters, i.e., by variations of preformed  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  [e.g., Mix and Fairbanks, 1985]. The  $\delta^{13}\text{C}$  records of planktonic *G.*

*bulloides* and *N. pachyderma* sinistral are used here to constrain potential source water  $\delta^{13}\text{C}$  variations. Even though it is not clear to what extent planktonic  $\delta^{13}\text{C}$  accurately records seawater  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  [see Labeyrie and Duplessy, 1985; Keigwin and Boyle, 1989; Charles and Fairbanks, 1990] such analysis appears important for this study since the North Pacific deep-waters are extremely enriched in nutrients and deficient in  $^{13}\text{C}$ . If deep waters welled up in the glacial Subarctic Pacific and contributed more profoundly to the property field of the surface waters, then the source waters could have had preformed  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  values low enough to confound the traditional interpretation of the benthic  $\delta^{13}\text{C}$  records. This would impede detection of deep water ventilation coming from a North Pacific source.

The planktonic  $\delta^{18}\text{O}$  records are used to infer possible variations of the temperature-salinity (T-S) field of the North Pacific mixed layer. Knowledge of T-S variations is fundamental to estimating whether gravitational instabilities could have occurred which would have fostered convective overturn. Extraction of temperature signals from foraminiferal  $\delta^{18}\text{O}$  records is straightforward, given the thermodynamically well-constrained equilibrium isotope exchange between calcite and seawater [e.g., O'Neil et al. [1969]. Unfortunately, however, our knowledge of variations of the  $\delta^{18}\text{O}_{\text{seawater}}(\delta_w)$ -salinity relationship on glacial-to-interglacial time scales is too incomplete

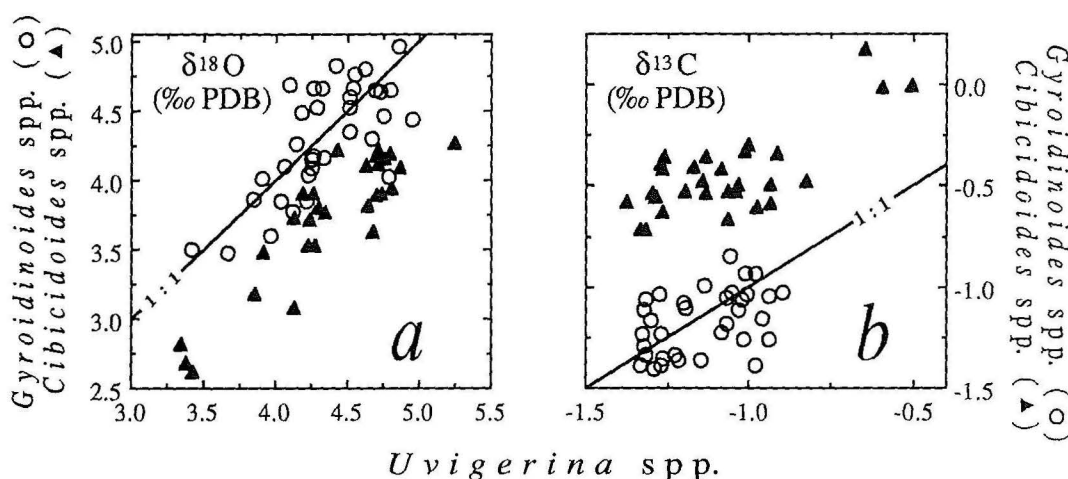


Fig. 3. Comparison of paired stable isotope data obtained from three benthic genera used in this study. (a)  $\delta^{18}\text{O}$  of *Cibicidoides* spp. is consistently offset from *Uvigerina* spp. values by  $0.61 \pm 0.18\text{‰}$  ( $1\sigma$ );  $\delta^{18}\text{O}$  of *Gyroidinoides* spp. is similar to that of *Uvigerina* spp. (t statistics at  $\nu=33$ :  $\Delta\delta^{18}\text{O} = 0.02 \pm 0.28$  ( $1\sigma$ ),  $T = 0.471$ ,  $P = 0.641$ ). (b)  $\delta^{13}\text{C}$  of *Gyroidinoides* spp. is similar to that of *Uvigerina* spp. (t statistics at  $\nu=33$ :  $\Delta\delta^{13}\text{C} = 0.04 \pm 0.18$  ( $1\sigma$ ),  $T = 1.434$ ,  $P = 0.161$ ), suggesting that both genera inhabit similar (endobenthic) microhabitats. The  $\delta^{13}\text{C}$  offset between *Cibicidoides* spp. and *Uvigerina* spp. is not constant, thus supporting the hypothesis that both genera inhabit different, i.e. epibenthic versus endobenthic microhabitats.

TABLE 2. Stable Isotope Data From Core PAR 87A-01

Depth in Core, m	Age <sup>14</sup> C-kyr	<i>G. bulloides</i>		<i>N. pachyderma sin.</i>		<i>Uvigerina spp.</i>		<i>Cibicides spp.</i>		<i>Gyroidinoides spp.</i>	
		$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
0.01	6.01					3.19	-0.35				
0.07	7.01	2.04	0.12	2.28	0.35	3.42	-0.60	2.63	-0.01	3.50	-1.10
0.12	8.00	1.93	0.03	2.28	0.27	4.12	-1.30	3.09	-0.53	3.77	-1.17
0.17	9.00			2.34	0.34	4.22	-1.27	3.54	-0.41	3.85	-1.39
0.21	10.00	2.34	0.05			4.23	-1.29	3.73	-0.55	4.04	-1.40
0.27	10.89	2.50	0.07	2.79	0.06	4.26	-1.31			4.17	-1.33
0.32	11.77	2.18	-0.35	2.79	0.09	4.06	-1.22			4.10	-1.37
0.36	12.35	2.49	-0.45	3.21	0.08	4.33	-1.27	3.78	-0.62	4.16	-1.24
0.41	12.92	3.43	-0.46	3.81	0.04			3.91	-0.56		
0.47	13.50			3.57	-0.04			4.24	-0.48		
0.52	13.88			3.63	-0.11						
0.56	14.23			3.43	-0.19			3.61	-0.47	4.67	-1.07
0.61	14.60	3.32	-0.24			4.79	-1.04	4.20	-0.49		
0.67	14.97	3.26	-0.38	3.46	-0.20			4.16	-0.60		
0.71	15.34	3.45	-0.43	3.58	-0.07	4.72	-0.94	3.92	-0.58	4.64	-1.04
0.77	15.71	3.15	-0.35	3.28	-0.14	4.69	-0.98	3.89	-0.59	4.65	-0.94
0.81	16.07	3.12	-0.47	3.05	-0.12	4.75	-1.14	4.16	-0.53	4.47	-1.00
0.86	16.44	3.10	-0.63	3.45	0.03			4.25	-0.64		
0.92	16.97	2.56	-0.31	3.39	-0.08						
0.96	17.50	2.83	-0.38	3.71	0.05			3.83	-0.61	4.31	-1.05
1.01	18.02			3.62	-0.12	4.66	-1.07	3.64	-0.66	4.30	-1.05
1.06	18.55	2.98	-0.17	3.75	0.00			3.62	-0.62		
1.11	19.08			2.73	0.18	4.32	-1.33			4.66	-1.23
1.17	19.61	2.45	-0.10			4.26	-1.23			4.67	-1.34
1.22	20.13					4.26	-1.35				

Isotope values are in ‰PDB. Age scale has been developed by visual correlation of the oxygen isotope records with those from core PAR87A-10. Age scale is corrected for reservoir effects by subtracting 717 years.

to allow extraction of salinity signals from the  $\delta^{18}\text{O}$  records with similar confidence [e.g., Broecker, 1986; Zahn and Mix, 1991]. We will estimate the glacial T-S flux needed to decrease the buoyancy of the surface waters so as to start convection, and translate the estimates into a hypothetical glacial-interglacial amplitude of  $\delta^{18}\text{O}_{\text{calcite}}$  by using paleotemperature equations and the slope of the modern  $\delta_{\text{w}}$ -salinity relationship. The results will then be compared to the  $\delta^{18}\text{O}$  amplitude observed in the planktonic records at our core sites.

#### CHRONOLOGY

A radiocarbon chronology was adopted for core 10 using five accelerator-mass spectrometer <sup>14</sup>C dates which were obtained from monospecific foraminiferal samples containing 200-400 planktonic specimens. The measurements were carried out by the Radio-Isotope Direct Detection Laboratory at Simon Fraser University (Burnaby, British Columbia) following

procedures described by Nelson et al. [1986] and Vogel et al. [1987]; the results are listed in Table 5. We have subtracted 717 years from the <sup>14</sup>C raw ages so as to accommodate the late Quaternary reservoir age of the Northeast Pacific mixed layer [Southon et al., 1990] (Table 5). Since we have at present no control on potential changes of the reservoir age prior to 11,000 years B.P. [Southon et al., 1990], we use this reservoir correction for the past 25 kyr.

A time scale for core 10 was developed by linear interpolation between the <sup>14</sup>C control points. The core-top age was determined by extrapolation, using mean sedimentation rates calculated from the upper three age control points. The age models for cores 01 and 02 have been developed by visually correlating their isotope records to those of core 10. Sedimentation rates at the three sites are 6-10 cm kyr<sup>-1</sup> (Table 1), yielding a time resolution of 1-2 kyr of the isotope records at sampling intervals of 5 cm.

Recent results from a comparison of uranium-thorium (U-Th) ages and <sup>14</sup>C ages from Barbados

corals suggest that the late glacial  $^{14}\text{C}$  time scale is probably too young by some 3000 years compared to the "true", i.e.  $^{234}\text{U}/^{230}\text{Th}$ - "calendar-year" time scale [Bard et al., 1990]. Even though these results are

still preliminary, we have plotted a "true", i.e.  $^{234}\text{U}/^{230}\text{Th}$ - "calendar-year" time scale along with our isotope records.  $^{14}\text{C}$  ages  $\geq 6,000$  years were converted toward calendar-year ages by using the

TABLE 3. Stable Isotope Data From Core PAR 87A-02

Depth in Core, m	Age, $^{14}\text{C}$ -kyr	<i>G.bulloides</i>		<i>N.pachyderma sin.</i>		<i>Uvigerina</i> spp.		<i>Cibicidoides</i> spp.		<i>Gyroidinoides</i> spp.	
		$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
0.01	1.75					3.41	-0.42				
0.07	3.50			2.53	0.47	3.39	-0.59				
0.12	5.25	2.57	0.22	2.44	0.33					3.39	-0.93
0.17	7.01			2.42	0.29					3.29	-0.96
0.21	10.91	2.56	0.32	2.75	0.25	3.96	-1.06			3.60	-0.85
0.27	11.34	2.53	0.27	2.60	0.26	3.67	-1.02			3.48	-1.07
0.32	11.77	2.38	0.22	2.77	0.18	4.03	-1.08			3.85	-1.22
0.36	11.95	2.77	0.14	2.75	0.18	3.90	-1.02	3.48	-0.33	4.02	-1.26
0.41	12.13	2.75	0.16	2.85	0.01	4.25	-1.26	3.91	-0.35	4.08	-1.36
0.47	12.31	2.93	0.15	3.10	-0.09	4.12	-1.18	3.73	-0.41		
0.52	12.49	2.74	0.15	3.36	0.02	4.79	-0.98			4.02	-1.39
0.56	12.67	2.98	-0.36	3.37	-0.06	4.63	-1.04	3.82	-0.52		
0.61	12.85	3.65	-0.35	3.42	-0.11	4.53	-1.04			4.66	-1.11
0.67	13.03	3.50	-0.18	3.62	0.02	4.51	-1.01			4.60	-0.94
0.71	13.21	2.79	0.06	3.54	-0.09	4.95	-1.05			4.44	-1.03
0.77	13.39	3.69	-0.37							4.86	-1.10
0.81	13.57	3.85	-0.23	3.75	-0.00	4.55	-0.96			4.76	-1.16
0.86	13.75	3.64	-0.13	3.76	-0.14	5.24	-0.92	4.28	-0.33		
0.92	14.11	3.70	-0.15	3.64	-0.04			4.23	-0.42		
0.96	14.47	3.56	-0.16	3.37	-0.07			4.17	-0.43		
1.01	14.82	3.69	-0.18	3.62	-0.08					4.82	-1.25
1.06	15.18	3.90	-0.23	3.43	-0.08	4.85	-1.01	4.10	-0.29	4.96	-1.04
1.11	15.54	3.85	-0.33	3.52	-0.07	4.62	-1.28	4.12	-0.38	4.79	-1.04
1.17	15.75	3.60	-0.25	3.50	-0.04					4.62	-0.87
1.22	15.95	3.18	-0.06	3.57	0.02					4.68	-0.93
1.26	16.07			3.92	-0.07					4.75	-0.93
1.31	16.18			3.71	-0.13			4.24	-0.36	4.83	-0.88
1.36	16.30			3.49	-0.07			4.08	-0.25	4.63	-0.92
1.42	16.41	3.34	-0.15	3.37	0.04			4.32	-0.21	4.64	-0.97
1.47	16.53	3.55	-0.26	3.41	-0.06	4.69	-1.14	4.19	-0.35		
1.51	16.65			3.63	0.08					4.75	-1.02
1.56	16.76	3.42	-0.24	3.38	-0.04			4.46	0.01		
1.61	16.88			3.30	0.05			4.45	-0.04		
1.67	16.99									4.62	-0.75
1.72	17.11									4.58	-0.87
1.76	17.22			3.58	0.24	4.60	-1.25				
1.81	17.34	2.66	0.08	3.50	0.21	4.51	-1.32			4.52	-1.06
1.86	17.45									4.50	-0.86
1.92	17.57			3.46	0.22			4.08	-0.32	4.41	-0.79
1.97	17.68	3.17	0.10	3.53	0.15			3.92	-0.16	4.32	-1.01
2.02	17.80			3.23	0.24	4.51	-1.20			4.35	-1.11

Isotope values are in ‰ PDB. Age scale has been developed by visual correlation of the oxygen isotope records with those from core PAR87A-10. Age scale is corrected for reservoir effects by subtracting 717 years.



TABLE 4. Stable Isotope Data From Core PAR 87A-10

Depth in Core, m	Age, <sup>14</sup> C-kyr	<i>G. bulloides</i>		<i>N. pachyderma</i> sin.		<i>Uvigerina</i> spp.		<i>Cibicides</i> spp.		<i>Gyroidinoides</i> spp.	
		δ <sup>18</sup> O	δ <sup>13</sup> C	δ <sup>18</sup> O	δ <sup>13</sup> C	δ <sup>18</sup> O	δ <sup>13</sup> C	δ <sup>18</sup> O	δ <sup>13</sup> C	δ <sup>18</sup> O	δ <sup>13</sup> C
0.01	1.16					3.31	-0.33				
0.04	1.81			2.20	0.36	3.22	-0.53				
0.10	3.11					3.32	-0.35				
0.11	3.32										
0.14	3.87			2.20	0.25	3.21	-0.48				
0.19	5.06	2.26	0.35	2.34	0.35	3.31	-0.42				
0.21	5.49										
0.23	5.92	2.51	0.15	2.34	0.46	3.33	-0.65	2.83	0.18		
0.28	7.01	2.20	0.40	2.37	0.36	3.37	-0.51	2.69	0.00		
0.30	7.44										
0.32	7.87	2.35	0.31	2.29	0.31	3.54	-0.92				
0.37	8.96	2.62	0.23	2.75	0.23	3.85	-1.15	3.19	-0.47	3.86	-1.36
0.39	9.39										
0.41	9.82					4.14	-1.32			4.26	-1.29
0.46	10.91	2.74	0.12	2.84	0.04	4.26	-1.38	3.54	-0.57		
0.48	11.34										
0.50	11.77	2.49	0.04			4.25	-1.33			4.15	-1.39
0.55	12.07	2.67	0.05	2.50	-0.20	4.31	-1.34	3.86	-0.71		
0.57	12.19										
0.64	12.60	2.97	-0.02	3.43	-0.20	4.18	-1.32	3.91	-0.71	4.49	-1.11
0.65	12.66										
0.67	12.77									4.42	-1.06
0.68	12.83	3.02	-0.48	3.61	-0.05	4.10	-0.90			4.69	-1.03
0.72	13.07	3.46	-0.66	3.73	-0.07					4.68	-1.01
0.74	13.19	3.31	-0.69	3.84	-0.20	4.28	-1.07	3.81	-0.52	4.52	-1.18
0.75	13.24										
0.77	13.36	3.59	-0.40	3.75	-0.17						
0.79	13.48			3.86	-0.04					4.97	-1.05
0.83	13.72	3.54	-0.51	3.57	-0.10						
0.84	13.78										
0.86	13.89			3.55	-0.09						
0.87	13.95	3.60	-0.75					4.36	-0.39		
0.92	14.25	3.43	-0.17	3.51	-0.08			4.17	-0.65		
0.94	14.36										
1.00	14.72										
1.02	14.83	3.48	-0.44	3.51	0.12	4.75	-1.04				
1.06	15.07	3.58	-0.27	3.64	-0.12			4.33	-0.40	4.94	-0.93
1.08	15.19	3.12	-0.09	3.40	-0.05	4.70	-1.09	4.22	-0.41		
1.09	15.25										
1.11	15.36	3.08	-0.35	3.35	0.02	4.80	-1.20	3.95	-0.52	4.65	-1.08
1.14	15.54	3.75	-0.33							4.85	-1.15
1.17	15.72	3.36	-0.51	3.40	-0.14	4.42	-0.94	4.23	-0.49	4.83	-1.26
1.19	15.84										
1.21	15.95	3.14	-0.41	3.16	-0.12	4.70	-0.83	4.12	-0.47		
1.24	16.44							4.32	-0.34	4.75	-1.27
1.37	18.55	2.99	-0.09	3.18	-0.15			3.61	-0.41		
1.38	18.71										
1.40	19.04			3.17	-0.09						
1.47	20.17							3.73	-0.83		

Isotope values are in ‰ PDB. Age scale has been developed using <sup>14</sup>C data listed in Table 2. Age scale is corrected for reservoir effects by subtracting 717 years.

TABLE 5. Radiocarbon Data for Core PAR 87A-10

Depth in Core, cm	$^{14}\text{C}$ Age raw	$^{14}\text{C}$ Age res.corr. <sup>a</sup>	"True" Age <sup>b</sup>	Foraminiferal Species	Number of Specimens
	1000 years B.P.	1000 years B.P.	[1000 yr B.P.]		
23.0	6,640 ± 300	5,923 ± 304	6,924 ± 738	<i>N.pachyderma</i>	250
50.0	12,490 ± 230	11,773 ± 235	13,839 ± 713	<i>G.bulloides</i>	200
121.0	16,670 ± 560	15,953 ± 562	18,725 ± 877	<i>N.pachyderma</i>	275
213.0	31,600 +1500 -2600	30,883 +1501 -2600	36,178 +1645 -2686	<i>N.pachyderma</i>	370
274.0	>35,000	-	-	<i>N.pachyderma</i>	425

<sup>a</sup>For correction a reservoir age of  $717 \pm 47$  years has been applied, which is indicated by data listed in Table 3 of Southon et al. [1990] (their reservoir age estimate of  $1180 \pm 110$  years B.P. at 6360 years B.P. has not been included for computing the average).

<sup>b</sup>Calibration of the corrected  $^{14}\text{C}$  data to the U/Th time scale of Bard et al. [1990] was done using equation 1. The standard error of the "true" age estimate is  $\pm 673$  years. Standard errors increase as a function of the square root of the pooled variances of input data and correction factors.

linear regression equation

$$\text{Age}_{(\text{true})} = 1.169^{\pm 0.03} \times \text{Age}_{(^{14}\text{C})} + 0.076^{\pm 0.45} \quad (1)$$

(n = 18,  $r^2 = 0.99$ )

where  $\text{Age}_{(\text{true})}$  and  $\text{Age}_{(^{14}\text{C})}$  are in thousands of years (superscripts give standard errors of the regression coefficients; standard error of estimate, i.e.,  $\text{Age}_{(\text{true})}$ , is  $\pm 673$  years; Table 5). Equation (1) was determined using the data listed in Table 1 of Bard et al. [1990].

## RESULTS

### Oxygen Isotopes

The  $\delta^{18}\text{O}$  records (Figures 4a-4c) all show the last glacial-interglacial transition, i.e., oxygen isotope Termination I, between 7 and 14 kyr B.P. and display Termination I as a two-step transition with step I<sub>A</sub> lasting from 13.3 kyr to 12.0 kyr B.P. and step I<sub>B</sub> lasting from 9.4 kyr to 7.3 kyr B.P. (these ages have been obtained from core 10 only, since we

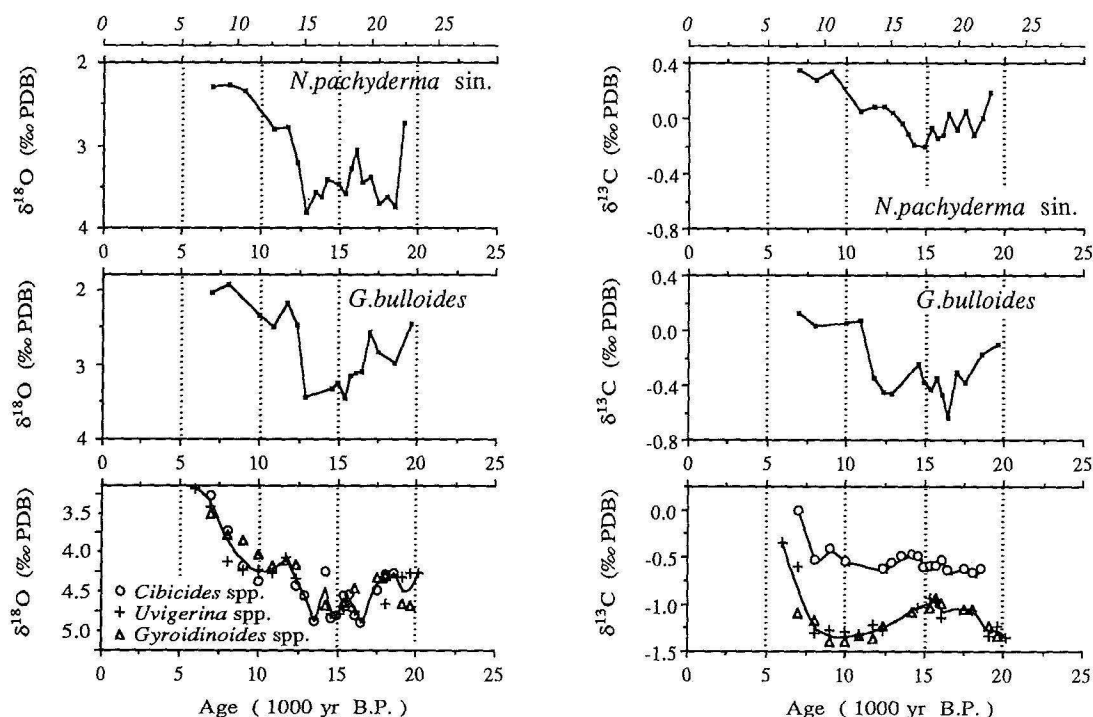


Fig. 4a. Benthic and planktonic isotope records of core PAR 87A-01 (3480 m water depth). Time scale is in (reservoir-corrected)  $^{14}\text{C}$ -kyr. For comparison, a "true" U/Th calendar-year time scale [Bard et al., 1990] is shown at the top. The  $^{14}\text{C}$  scale has been converted into the U/Th scale using equation (1). Vertical, dotted lines are at 5, 10, 15, and 20- $^{14}\text{C}$ -kyr.

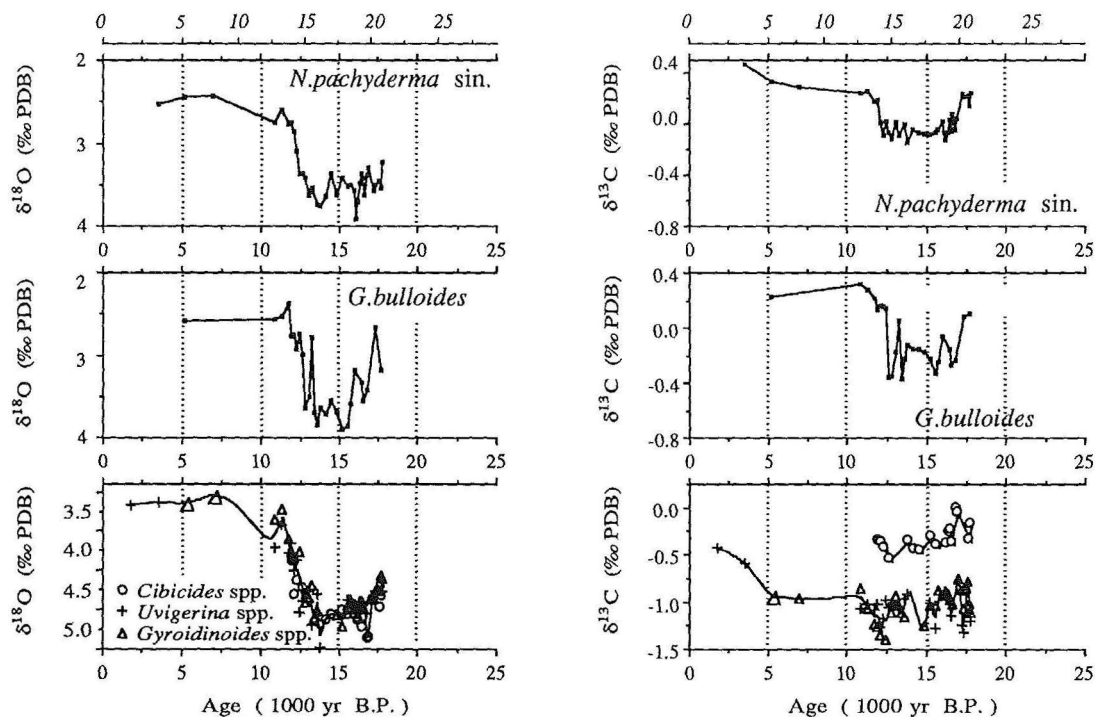


Fig. 4b. Same as Figure 4a except for core PAR 87A-02 (2920 m water depth).

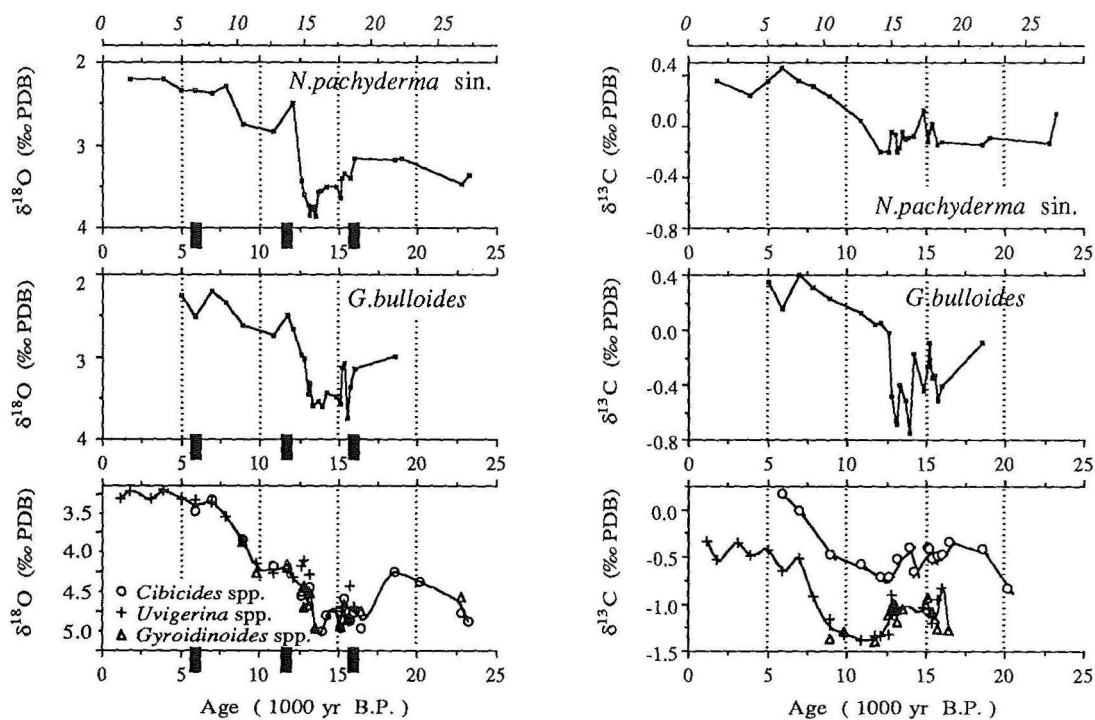


Fig. 4c. Same as Figure 4a except for core PAR 87A-10 (3664 m water depth). Black bars along the time axes of the  $\delta^{18}\text{O}$  records indicate positions of the samples which were radiocarbon dated (see Table 3).



have no independent age control on the isotope records of the two other sediment cores). In the North Atlantic realm, the two-step character of the last deglaciation is due to the Younger Dryas cooling event [e.g., Duplessy et al., 1981] in association with deglacial meltwater events [e.g., Fairbanks, 1989; Broecker, 1990]. For the northwest Pacific, a link to a mid-Termination I climatic reversal has also been suggested [Kallel et al., 1988a], but a major influence of local salinity/meltwater effects cannot be excluded [Keigwin, 1989]. The existence of a mid-Termination I break in our planktonic  $\delta^{18}\text{O}$  records suggests that similar events occurred also in the Subarctic east Pacific. Evidence in support of a climatic reversal in the northeast Pacific region, which would have been coeval with the Younger Dryas event, comes from pollen records from the Glacier Bay area, southeastern Alaska [Engstrom et al., 1990].

Glacial-interglacial  $\delta^{18}\text{O}$  amplitudes in the benthic and planktonic records are, respectively, 1.5‰ and 1.1‰ (Table 6). The planktonic  $\delta^{18}\text{O}$  amplitudes are 1.4–1.5‰ if we take the heaviest, glacial-maximum values just before the onset of the  $\delta^{18}\text{O}$  decrease rather than average values between 13.5 and 17 kyr B.P. (see Figure 4). These amplitudes are considerably lower than those observed at core sites in the Atlantic [e.g., Duplessy et al., 1980; Duplessy, 1982; Shackleton et al., 1983]. Reduced benthic  $\delta^{18}\text{O}$  amplitudes, which have been observed at other Pacific core sites as well, have been assumed to document a much smaller temperature variability in

the deep Pacific compared to the deep Atlantic [Shackleton et al., 1983].

Planktonic  $\delta^{18}\text{O}$  records from high-latitude North Atlantic core sites show glacial-interglacial  $\delta^{18}\text{O}$  shifts of >2‰ [Labeyrie and Duplessy, 1985; Jansen and Veum, 1990] which are well correlated to the dramatic variations of sea surface temperature and salinity at these core sites. Our planktonic  $\delta^{18}\text{O}$  amplitudes are much smaller; they are close to the glacial-interglacial shift of the mean-ocean oxygen isotope composition which is assumed to be about 1.3‰ [Labeyrie et al., 1987; Fairbanks, 1989]. This may point to competing effects of local salinity variations, which appear to have compensated much of the combined isotope signal of the mean-ocean  $\delta^{18}\text{O}$  shift and regional temperature variations (see below).

As Figures 4a–4c also show, the magnitude of the  $\delta^{18}\text{O}$  shifts during terminations I<sub>A</sub> and I<sub>B</sub> is not the same in the planktonic and benthic isotope records at the different core sites. At the deeper sites 01 and 10 (3.5 km and 3.7 km, respectively) the isotope shifts during steps I<sub>A</sub> and I<sub>B</sub> each contribute roughly 50% to the entire deglacial  $\delta^{18}\text{O}$  shift observed in the benthic records. In the planktonic records, about 75% of the total glacial-interglacial amplitude is already accomplished during Termination I<sub>A</sub>. At the shallower site 02 (2.9 km) the amplitude of the benthic  $\delta^{18}\text{O}$  shift during Termination I<sub>A</sub> is about 0.5‰ higher than in the benthic records at the deeper sites and is similar to the  $\delta^{18}\text{O}$  shift observed in the planktonic records at all three sites. We cannot exclude the possibility that this discrepancy between the

TABLE 6. Modern and Late-Glacial Means of  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  at Core Sites From the Patton-Murray Seamount Group

	<i>N. pachyderma</i> (sin.)		<i>G. bulloides</i>		Benthics		
	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}^a$	$\delta^{13}\text{C}^b$
	<i>Modern</i>						
PAR 87A-01 (< 8.5 kyr B.P.)	2.30 <sub>3</sub> <sup>±.04</sup>	0.32 <sub>3</sub> <sup>±.04</sup>	1.98 <sub>2</sub> <sup>±.08</sup>	0.08 <sub>2</sub> <sup>±.07</sup>	3.28 <sub>2</sub> <sup>±.13</sup>	0.01 <sub>1</sub>	-0.35 <sub>1</sub>
PAR 87A-02 (< 8.5 kyr B.P.)	2.46 <sub>3</sub> <sup>±.06</sup>	0.36 <sub>3</sub> <sup>±.10</sup>	2.57 <sub>1</sub>	0.22 <sub>1</sub>	3.37 <sub>4</sub> <sup>±.05</sup>	-	-0.73 <sub>4</sub> <sup>±.26</sup>
PAR 87A-10 (< 8.5 kyr B.P.)	2.29 <sub>6</sub> <sup>±.07</sup>	0.35 <sub>6</sub> <sup>±.07</sup>	2.33 <sub>4</sub> <sup>±.14</sup>	0.30 <sub>4</sub> <sup>±.11</sup>	3.33 <sub>8</sub> <sup>±.11</sup>	0.09 <sub>2</sub> <sup>±.13</sup>	-0.52 <sub>8</sub> <sup>±.19</sup>
Mean <sup>c</sup>	2.34 <sub>12</sub> <sup>±.10</sup>	0.35 <sub>12</sub> <sup>±.07</sup>	2.27 <sub>7</sub> <sup>±.24</sup>	0.23 <sub>7</sub> <sup>±.13</sup>	3.34 <sub>14</sub> <sup>±.09</sup>	0.06 <sub>3</sub> <sup>±.10</sup>	-0.57 <sub>13</sub> <sup>±.23</sup>
	<i>Late-Glacial</i>						
PAR 87A-01 (13.4–16.4 kyr B.P.)	3.43 <sub>8</sub> <sup>±.19</sup>	-0.11 <sub>8</sub> <sup>±.08</sup>	3.23 <sub>6</sub> <sup>±.14</sup>	-0.42 <sub>6</sub> <sup>±.13</sup>	4.77 <sub>8</sub> <sup>±.15</sup>	-0.55 <sub>8</sub> <sup>±.06</sup>	-1.03 <sub>5</sub> <sup>±.05</sup>
PAR 87A-02 (13.4–16.9 kyr B.P.)	3.55 <sub>17</sub> <sup>±.17</sup>	-0.04 <sub>17</sub> <sup>±.06</sup>	3.61 <sub>13</sub> <sup>±.21</sup>	-0.21 <sub>13</sub> <sup>±.08</sup>	4.82 <sub>18</sub> <sup>±.15</sup>	-0.28 <sub>11</sub> <sup>±.15</sup>	-1.01 <sub>14</sub> <sup>±.11</sup>
PAR 87A-10 (13.3–16.4 kyr B.P.)	3.52 <sub>11</sub> <sup>±.19</sup>	-0.07 <sub>11</sub> <sup>±.08</sup>	3.43 <sub>11</sub> <sup>±.22</sup>	-0.39 <sub>11</sub> <sup>±.18</sup>	4.83 <sub>11</sub> <sup>±.11</sup>	-0.46 <sub>8</sub> <sup>±.10</sup>	-1.07 <sub>9</sub> <sup>±.13</sup>
Mean <sup>c</sup>	3.52 <sub>36</sub> <sup>±.18</sup>	-0.07 <sub>36</sub> <sup>±.07</sup>	3.47 <sub>30</sub> <sup>±.25</sup>	-0.32 <sub>30</sub> <sup>±.16</sup>	4.80 <sub>37</sub> <sup>±.14</sup>	-0.41 <sub>27</sub> <sup>±.16</sup>	-1.03 <sub>28</sub> <sup>±.11</sup>
$\Delta$ Glacial-Modern	1.15 <sub>3</sub> <sup>±.07</sup>	-0.42 <sub>3</sub> <sup>±.02</sup>	1.13 <sub>3</sub> <sup>±.11</sup>	-0.54 <sub>3</sub> <sup>±.13</sup>	1.46 <sub>3</sub> <sup>±.03</sup>	-0.56 <sub>3</sub> <sup>±.01</sup>	-0.50 <sub>3</sub> <sup>±.20</sup>

Means of  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  are in ‰ PDB. Values are given in the form  $\bar{X}_N^s$  where  $\bar{X}$  = mean,  $s$  = 1 $\sigma$ ,  $N$  = number of observations.

<sup>a</sup>epibenthic  $\delta^{13}\text{C}$  (*Cibicides* spp.)

<sup>b</sup>endobenthic  $\delta^{13}\text{C}$  (pooled *Uvigerina* spp. and *Gyroidinoides* spp. mean)

<sup>c</sup>weighted arithmetic mean

benthic  $\delta^{18}\text{O}$  amplitudes at different water depths is the result of bioturbational smoothing, in combination with variations of benthic foraminiferal abundances. However, it is also conceivable that the high-amplitude shift of benthic  $\delta^{18}\text{O}$  during Termination I<sub>A</sub> at the shallower core site documents a rapid warming or freshening of the mid-depth North Pacific during the early deglaciation.

### Carbon Isotopes

The pattern of  $\delta^{13}\text{C}$  variation in the planktonic and benthic isotope records is consistent with the global trend in that glacial values are more negative than the Holocene values (Figures 4a-4c). The mean deglacial  $\delta^{13}\text{C}$  shift recorded in the planktonic records is 0.4-0.5‰, similar to the shift observed in the benthic  $\delta^{13}\text{C}$  profiles (Table 6). Both are similar to current estimates of the "global"  $\delta^{13}\text{C}$  shift during the last glacial-interglacial transition. Epibenthic  $\delta^{13}\text{C}$  at the deeper cores 01 and 10 is similar to that recorded at site RAMA 44 in the Northwest Pacific (-0.41‰ PDB; Keigwin [1987]) (Figure 5) and implies a

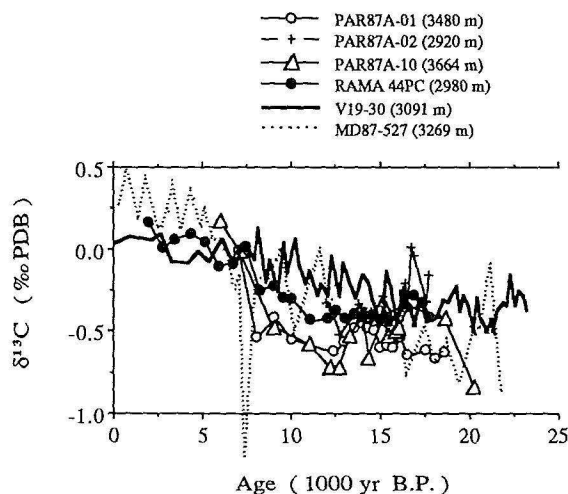


Fig. 5. Comparison of epibenthic  $\delta^{13}\text{C}$  records from our Northeast Pacific sites with those from Northwest Pacific core RAMA-44PC [Keigwin, 1987], equatorial Pacific site V19-30 [Shackleton and Pisias, 1985], and Southern Ocean core MD87-527 [Labeyrie et al., 1987]. At the last glacial maximum  $\delta^{13}\text{C}$  at our deeper core sites appears to be similar to that in the Southern Ocean but more negative than in the equatorial Pacific. If true, this would imply a meridional  $\delta^{13}\text{C}$  distribution in the deep Pacific much different from today. Epibenthic  $\delta^{13}\text{C}$  at the shallower site 02 is more positive at the LGM than at the deeper sites suggesting better ventilation of the upper deep waters and perhaps intermediate waters above.

monotonous distribution of nutrients in the Subarctic Pacific below about 3 km water depth. The values appear to be slightly more negative than values recorded at tropical Pacific site V19-30 (Figure 5), supporting Keigwin's [1987] hypothesis that no local deepwater source existed in the Subarctic Pacific during the last glaciation.

Glacial-maximum  $\delta^{13}\text{C}$  at our shallower site 02 (2920 m) appears to be more positive than at the deeper sites (Table 6 and Figure 5). These data would suggest that the upper deep waters (and possibly the intermediate waters above) were better ventilated. Moreover, it seems that our deepwater  $\delta^{13}\text{C}$  data are similar to those found at Southern Ocean site MD87-527 (Figure 5). A  $\delta^{13}\text{C}$  gradient from site V19-30 towards more negative values in the Southern Ocean and the Subarctic would suggest that the  $\delta^{13}\text{C}$  distribution and thus the circulation pattern in the deep Pacific were much different than today. This meridional  $\delta^{13}\text{C}$  pattern, however, has to be tested by collecting more data especially from the South Pacific.

### DISCUSSION

#### *Modeling Constraints on North Pacific Mixed-Layer Salt Inventories*

Due to the low salinity of the North Pacific mixed layer, the upper water column is so stable that convection cannot occur even if surface-water temperatures (SST) are brought to the freezing point [Warren, 1983] (Figure 6). At the LGM, wintertime SST in the Northeast Pacific appears to have been 3°C lower [CLIMAP, 1981], Pacific deepwater temperatures were around 0°C [Chappell and Shackleton, 1986; Labeyrie et al., 1987]. Assuming a mean-ocean salinity increase of 1, due to the glacial ice volume effect, the North Pacific water column would have still been statically stable (Figure 6). Even if mixed-layer temperatures had approached the freezing point salinities would have to increase by 2 beyond the ice volume effect to ~35.5 so as to reduce the vertical density contrast and to start overturning.

Warren's [1983] near-surface layer box model for the North Pacific predicts that glacial salinities of 35.5 could have been reached if evaporation rates had nearly doubled, with no change in Ekman pumping (Figure 7a). If wind-driven upwelling of saline deep waters had doubled, the required increase in evaporation would have been reduced but not eliminated (Figure 7b). In both cases, the result would have been only weakly sensitive to river runoff. Implications from general circulation models of the last glacial maximum [Kutzbach, 1987; COHMAP, 1988] predict enhanced cyclonic wind stress over the

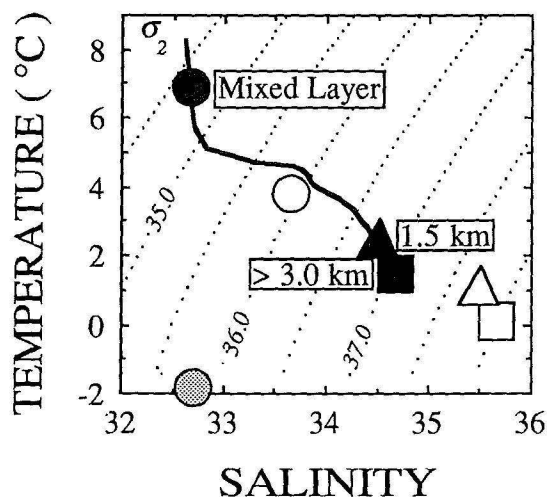


Fig. 6. Temperature-salinity (T-S) diagram showing T-S profile at Subarctic Pacific Station P [Tabata and Peart, 1985]. Density isolines are referred to the  $\sigma_2$  level (approximately 2 km water depth). Mixed-layer, intermediate-water (1.5 km) and deepwater (> 3.0 km) T-S points are means at respectively 0-100 m,  $27.4 < \sigma_\theta < 27.7$ , and  $\sigma_\theta \geq 27.7$ . Even at freezing temperatures ( $\sim -1.9^\circ\text{C}$ , shaded circle) the mixed-layer density is much lower than that of intermediate and deep waters ( $\Delta\sigma_t = 1.0$  and  $1.2$ , respectively) and water mass stratification remains stable. Open symbols are hypothetical glacial-maximum water masses: all salinities have been increased by 1; mixed-layer temperatures have been increased by  $3^\circ\text{C}$  [CLIMAP, 1981], intermediate-water and deepwater temperatures have been lowered by  $1.5^\circ\text{C}$ . Only if mixed-layer temperatures were at  $-0^\circ\text{C}$  and salinities had risen to  $\sim 35.5$  could convection have occurred. See text for discussion.

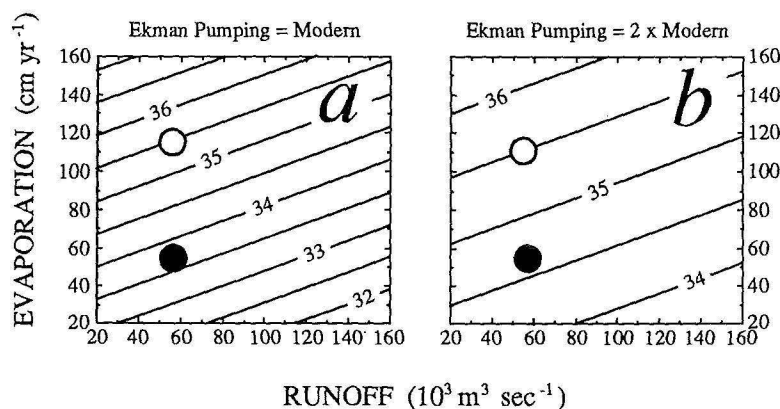


Fig. 7. Modeled mixed-layer salinities for the North Pacific during the last glacial maximum, using Warren's [1983] surface-layer box model. (a) Ekman pumping held at its modern value. (b) Ekman pumping twice the modern. Closed circles show mixed-layer salinity at modern rates of evaporation ( $55 \text{ cm yr}^{-1}$ ) and continental runoff ( $57 \cdot 10^3 \text{ m}^3 \text{ s}^{-1}$ ). Open circles show salinity needed to start convection.

Northeast Pacific in January, due to a stronger Aleutian low, which could indeed have increased midgyre Ekman pumping, thus advecting salt into the mixed layer. The models also indicate little change in oceanic precipitation but slight increases in evaporation, probably due to the combined effects of colder air temperatures, stronger and drier winds, and reduced runoff.

We can conclude here that overturn in the glacial Subarctic Pacific is physically plausible. The question is whether the T-S fields required by the model yield predictions that can be tested with paleoceanographic data. To address this question we examine

$\delta^{18}\text{O}$  data from planktonic foraminifera and compare them to similar data from benthic foraminifera. Vertical, i.e., benthic-planktonic  $\delta^{13}\text{C}$  gradients will be used as an additional constraint regarding convective links between surface and deep waters in the North Pacific.

#### *The Benthic-Planktonic Isotope Record of Changing Water Mass Hydrography*

Peak abundances of *N. pachyderma* sinistral and *G. bulloides* in the Subarctic east Pacific are commonly observed during late spring/early summer



and on occasion during fall and winter [Reynolds-Sautter and Thunell, 1989] (Figures 8a and 8b). Using  $\delta_w$ -salinity data from the North Pacific at 40°-60°N (Figure 9) and seasonal variations of surface temperature and salinity [Tabata and Peart, 1985] (Figure 8c) we have calculated annual mean  $\delta_c$  values of *N. pachyderma* sinistral and *G. bulloides* (Figure 8d). Apparently, measured  $\delta^{18}\text{O}$  from late Holocene planktonic foraminifera of +2.3‰ (PDB) is heavier by >1‰ than predicted surface-water equilibrium- $\delta_c$  (1.1-1.3‰ (PDB), Figure 8d). Planktonic  $\delta^{18}\text{O}$  data from trigger cores taken at the sites of our piston cores match the piston-core-top data presented here.

Thus for cores 02 and 10 we can exclude the possibility that more than a few centimeters of the original sediment surface were lost during the coring process. It is still possible that selective dissolution of Holocene shells in conjunction with an upward bioturbation of glacial shells could have shifted the isotope signal towards more positive glacial values. For now, however, we assume that the heavier than expected planktonic  $\delta^{18}\text{O}$  values document a subsurface growth habitat of both species during the Holocene.

If our assumptions are true, water column equilibrium- $\delta_c$  estimates predict that *N. pachyderma* sinistral

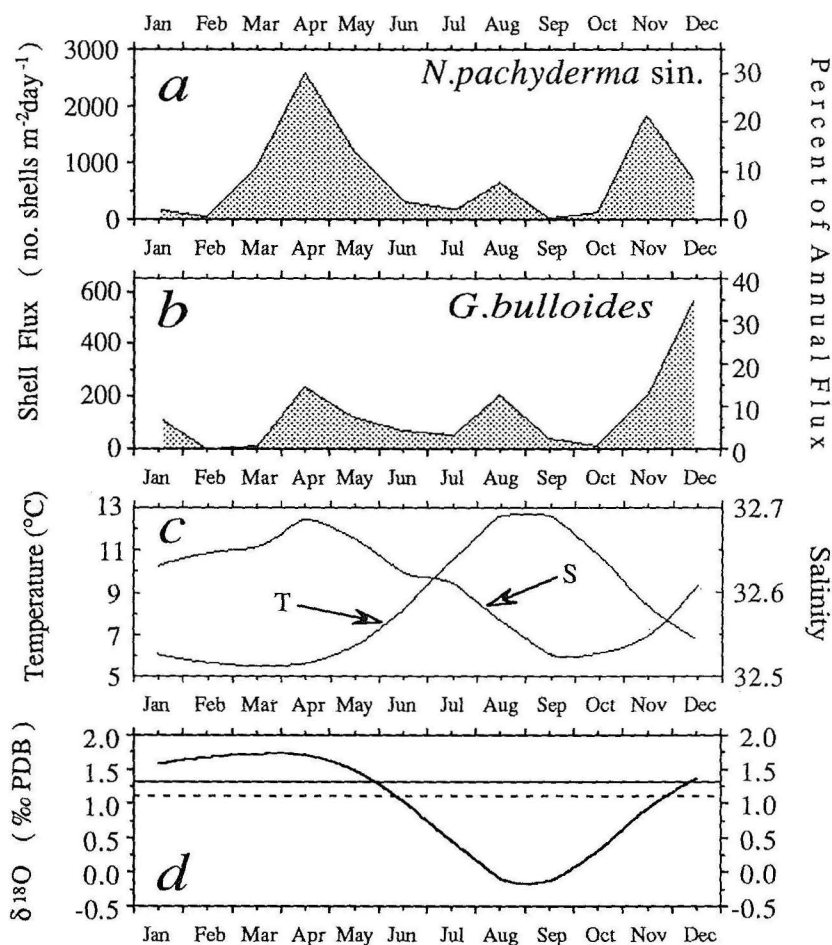


Fig. 8. Seasonal shell flux of (a) *N. pachyderma* sinistral and (b) *G. bulloides* (both from Reynolds-Sautter and Thunell [1989]), anomalies of (c) surface temperature and salinity [Tabata and Peart, 1985], and (d) equilibrium- $\delta_c$  at Ocean Station P. The  $\delta_c$  was predicted using the seasonal salinity variation in Figure 8c and the  $\delta_w$ -salinity relationship shown in Figure 9. Solid and dashed lines in Figure 8d give mean  $\delta_c$  values (weighted to percentage of monthly mean shell fluxes) which *N. pachyderma* sinistral (+1.28‰ PDB) and *G. bulloides* (+1.11‰ PDB) should display if both species calcify their tests in equilibrium with surface waters.

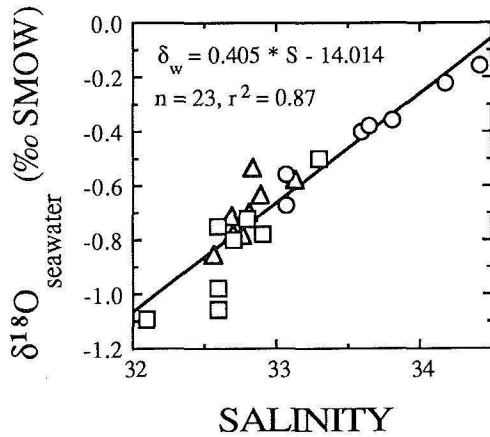


Fig. 9. The  $\delta_w$ -salinity relationship for the North Pacific ( $40^\circ$ - $60^\circ$ N). The diagram shows data from Epstein and Mayeda [1953] (corrected to the SMOW scale of  $\delta^{18}\text{O}$  as recommended by Craig and Gordon [1965]; squares), Craig and Gordon [1965] (Leapfrog Expedition; triangles), and Geosecs station 219 (0-1200 m water depth; Östlund et al. [1987]; circles). The linear regression model applies to an isopycnal range of  $25.4 < \sigma_\theta < 27.5$ . This equation has been used to estimate water column  $\delta_w$  needed to compute the seasonal and vertical equilibrium- $\delta_c$  profiles shown in Figures 8d and 10.

and *G. bulloides* calcify near the base of the mixed layer, at  $\sim 150$  m water depth (Figure 10). In order to destabilize the mixed layer, glacial salinities had to be increased by 2 beyond that induced by the ice volume effect, and glacial temperatures had to be  $\sim 0^\circ\text{C}$  (see Figure 6). This would have resulted in an apparent glacial-interglacial  $\Delta\delta_c$  of  $3.7\text{‰}$  (i.e., mean-ocean  $\Delta\delta_w \approx 1.3\text{‰}$ , temperature effect =  $1.8\text{‰}$ , salinity effect =  $0.6\text{‰}$ ). Yet, not only is the amplitude observed in the planktonic records less than half of that (Table 3), but it is also less than expected from the mean-ocean  $\delta_w$  shift and the CLIMAP [1981] estimate of  $\Delta T = 3^\circ\text{C}$ , which would predict a  $\Delta\delta_c$  of  $2.0\text{‰}$  (assuming the present  $\delta_w$ -salinity relationship applied at the LGM). Therefore our data would imply that a strong, negative- $\delta_w$ /meltwater anomaly existed in the Subarctic Pacific mixed layer during glacial times or that a warming relative to modern SST of  $\sim 1^\circ\text{C}$  occurred, which we consider unlikely. Another possibility could be that the planktonic foraminifera occupied shallower depth habitats during the last glacial where the influence of low-salinity waters was stronger. In either case, the data give no supporting evidence for convection but rather suggest that vertical stratification was similar or even more stable than today. This would be supported by the benthic-plank-

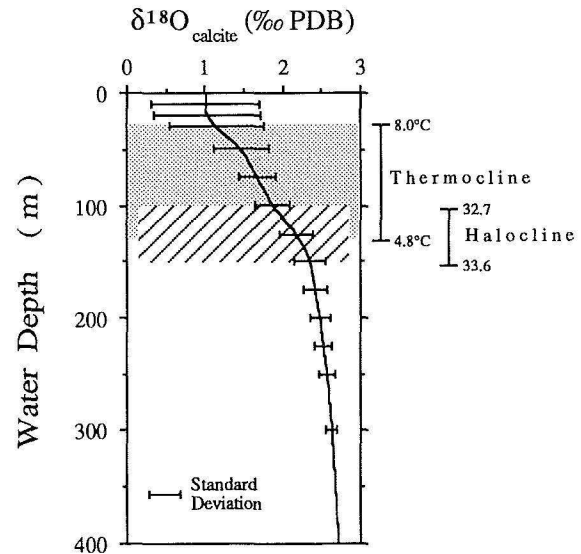


Fig. 10. Water column equilibrium- $\delta_c$  estimates for the upper 400 m at Ocean Station P. The isotope profile was computed using the temperature and salinity profiles in Figure 2 and applying the  $\delta_w$ -salinity relationship shown in Fig. 9. Values of  $+2.3\text{‰}$  (PDB), which are displayed by *N. pachyderma sinistral* and *G. bulloides* in early Holocene ( $< 10$  kyr B.P.) sections of our sediment cores, are found just below the halocline. This implies that both planktonic species calcify their tests near the base of the mixed layer.

tonic  $\delta^{18}\text{O}$  difference of  $1.3\text{‰}$  (see Table 6) which suggests that mixed-layer and deepwater hydrographs were significantly different ( $\Delta T > 5^\circ\text{C}$  or  $\Delta S > 3$ ).

The use of planktonic  $\delta^{13}\text{C}$  as a proxy for seawater- $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  is controversial [see Labeyrie and Duplessy, 1985; Keigwin and Boyle, 1989; Charles and Fairbanks, 1990]. However, using a correction factor of  $0.84\text{‰}$  [Labeyrie and Duplessy, 1985] to convert the modern  $\delta^{13}\text{C}$  values of *N. pachyderma sinistral* to apparent seawater- $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  we arrive at a  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  value of  $+1.2\text{‰}$  (PDB) which is similar to the  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  values found just below the mixed layer at Geosecs Station 219 ( $54^\circ\text{N}$ ,  $177^\circ\text{W}$ ;  $\delta^{13}\text{C}_{\Sigma\text{CO}_2} = 0.8$ - $1.4\text{‰}$  (PDB) at  $25.9 < \sigma_\theta < 26.5$ ; Östlund et al. [1987]). Applying the same correction to the late glacial  $\delta^{13}\text{C}$  values of *N. pachyderma sinistral* implies that  $\delta^{13}\text{C}_{\Sigma\text{CO}_2}$  values of the mixed layer were about  $+0.8\text{‰}$  (PDB). Epibenthic  $\delta^{13}\text{C}$  at our core sites during late glacial is  $-0.5\text{‰}$  (PDB) yielding a benthic-planktonic  $\delta^{13}\text{C}$  difference of  $1.3\text{‰}$ . Thus the benthic-planktonic  $\delta^{13}\text{C}$  gradient does not provide any evidence that a thermohaline link between the Subarctic mixed layer and the deep Pacific existed during glacial times.

## SUMMARY AND IMPLICATIONS

The data presented here suggest that water mass stratification in the glacial North Pacific was the same as or even more stable than today. Strongly reduced amplitudes of the  $\delta^{18}\text{O}$  records obtained from *N. pachyderma* sinistral and *G. bulloides* at three core sites in the central Gulf of Alaska (54°N, 148°W), relative to isotope data in the North Atlantic, suggest that the Subarctic mixed layer was fresher than today, perhaps because of locally more intense meltwater surges from the Cordilleran ice sheet during the summer seasons. Large offsets between benthic and planktonic  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  at the glacial maximum of more than 1‰ imply that the property fields of the Subarctic mixed layer and deep waters were significantly different and further support the contention that a thermohaline link between the deep Pacific and the North Pacific surface ocean did not exist during the LGM. Rather, it seems that epibenthic  $\delta^{13}\text{C}$  at our core sites is more negative than values at tropical-subtropical Pacific core sites but similar to values at Southern Ocean sites (Figure 5); this may point to a ventilation of the deep, central Pacific from mid-latitude sources, perhaps from marginal seas such as the Sea of Okhotsk.

Our data can not exclude the possibility that convection did occur to intermediate depths in the Subarctic Pacific outside the paleoceanographic window which is defined by the growth season of the planktonic foraminifera. Hydrographic evidence from the Weddell Sea and modeling results suggest that daisy-chained, seasonal cycles of melt water discharge, cooling, and sea ice formation precondition polar surface waters towards overturning [Martinson et al., 1981; Lembke et al., 1990]. Strong, stationary low-pressure systems, which are predicted for the glacial Subarctic Pacific during winter [Kutzbach, 1987; COHMAP, 1988], could have fostered the upwelling of saline deep waters thus raising the pycnocline close enough to the surface so that convection could have finally been triggered by cooling of the surface waters. Convective processes which would have occurred during winter time, however, are unlikely to be recorded by our planktonic isotope data because they are merely providing "snapshots" of the mixed-layer T-S field during the warmer seasons, when the plankton were growing and meltwater surges were strongest.

Finally, the planktonic and benthic  $\delta^{18}\text{O}$  records in the Subarctic east Pacific display a Younger Dryas-like break during mid-Termination I. In the light of similar data from the northwest Pacific [Kallel et al., 1988a; Keigwin, 1989] and new pollen records from the Glacier Bay area [Engstrom et al., 1990], which suggest that a climatic reversal coeval with the Younger Dryas event occurred in southeastern

Alaska, the question raises as to whether the Younger Dryas was a hemisphere-wide cooling event.

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