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Interior hydrography and circulation of the glacial Pacific Ocean

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Abstract

The deep water of the Pacific Ocean is a key component of the global climate system on the time scale of late-Pleistocene glaciation and deglaciation. Despite its importance, the deep Pacific during the last glacial maximum has received relatively little attention compared to the deep Atlantic, in part, because the Pacific poorly preserves carbonate sediments on the sea floor. Here, we review the current state of knowledge of the deep hydrography and circulation of the glacial Pacific by examining available nutrient-proxy data, including some new $\delta^{13}C$ and $\delta^{18}O$ data measured on benthic foraminifera *Planulina wuellerstorfi* from the vicinity of Japan. Available benthic $\delta^{13}C$ and $\delta^{18}O$ and radiocarbon data from the Pacific support the presence of a deep hydrographic boundary at around 2000 m during the Last Glacial Maximum (Paleoceanography 3 (1988) 343; Paleoceanography 7 (1992) 273; Paleoceanography 13(4) (1998) 323). The deep hydrographic divide in the glacial Pacific is similar to what is inferred in the Atlantic (Quaternary Research 18 (1982) 218; Paleoceanography 3 (1988) 317; Paleoceanography 3 (1988) 343; Annual Reviews of Earth Planetary Sciences 20 (1992) 245; Science 259 (1993) 1148), the Indian (Nature 333 (1988) 651; Paleoceanography 13 (1998) 20), and the Southern Ocean (Paleoceanography 11 (1996) 191), suggesting that this is a global phenomenon during the glacial time. The upper water mass has a distinctly enriched $\delta^{13}C$ compared to the deeper water mass, whose possible origins are discussed.

1. Introduction

Deep-water circulation and hydrography have been the foci of many paleoceanographic investigations over the years. Because of its sheer size, the deep ocean has a large capacity to store heat and various dissolved species of gases and salts. In addition, the slow movement of water gives the deep ocean one of the longest memories in the climate system. For these reasons, the deep ocean is believed to have played a central role in the large late-Pleistocene climate oscillations.

The Pacific deep water is volumetrically the most important, which has significant climatic implications, including the global carbon cycle on a glacial–interglacial time scale. On time scales much shorter than that for plate tectonics, the total surface carbon reservoir in the atmosphere–ocean–biosphere system can be assumed to be largely in steady state. Therefore, any significant change in atmospheric CO_2 concentration over glacial–interglacial time scale as recorded in polar ice cores (e.g., Barnola et al., 1987) must involve a redistribution of CO_2 amongst the carbon reservoirs (Broecker, 1982), of which the deep Pacific is by far the largest. Despite its obvious importance, the glacial Pacific deep water has received comparatively little attention than the Atlantic counterpart. One reason for this is that the carbonate compensation depth is shallower and carbonate preservation is poorer in the Pacific than in the Atlantic Ocean. The deficiency of carbonate material hinders paleoceanographic investigations, which have traditionally made use of the deep-sea sediment's carbonate fraction, from which foraminiferal $\delta^{18}O$ chronology are constructed (e.g., Hays et al., 1976; Martinson et al., 1987). Commonly used seawater nutrient proxies, $\delta^{13}C$ and cadmium concentration (Cd/Ca ratio) in benthic foraminiferal calcite tests, are also derived from the carbonate fraction.

For this and other reasons, our state of knowledge of the glacial Pacific deep water is limited. This was apparent, for example, in the work of Duplessy et al. (1988a), who produced a two-dimensional depth–latitude distribution of benthic foraminiferal $\delta^{13}C$ in

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the eastern Atlantic Ocean during the Last Glacial Maximum (LGM) and graphically identified the Glacial North Atlantic Intermediate Water (GNAIW). In the same work, only a $\delta^{13}C$ depth profile for the entire glacial Pacific was generated, since two-dimensional reconstruction was "impossible because of data limitation" (Duplessy et al., 1988a).

Since 1988, there have been few studies that have produced new glacial $\delta^{13}C$ data from the Pacific (Herguera et al., 1992; Keigwin, 1998; Matsumoto and Lynch-Stieglitz, 1999; Matsumoto et al. 2001). Some of these and other works have shown that the water mass above approximately 2000 m is distinct in its nutrient content from waters below it but have yet to produce a consensus on its circulation even on whether deep waters were flowing northward or southward.

In this paper, we review the current state of knowledge of the deep Pacific hydrography and circulation during the LGM focusing on the meaning of deep-water nutrient proxies. As part of the review, we present a vertical profile of new benthic foraminiferal $\delta^{13}C$ measurements from the vicinity of Japan. The new data clearly confirm the deep hydrographic boundary at around 2000 m in the glacial Pacific Ocean.

2. New benthic foraminiferal isotope data

Here, we briefly describe the new $\delta^{18}O$ and $\delta^{13}C$ measured on benthic foraminifera *Planulina (Cibicidoides) wuellerstorfi* from sediment cores raised from the vicinity of Japan between latitudes 28°N and 36°N and from a range in water depth from 740 to 3320 m (Table 1 and Fig. 1). The subtropical western boundary current, Kuroshio Current, overlies most of these core sites. The east coast of Japan where the nutrient-depleted Kuroshio and the nutrient-rich, subpolar western boundary current Oyashio meet is an area of high productivity. The influence of these surface currents reaches to the depth of the main thermocline, which in this region is approximately 1 km.

The utility of benthic foraminiferal $\delta^{13}C$ as a nutrient proxy rests on the general observation that $\delta^{13}C$ of dissolved inorganic carbon is inversely related to nutrient distribution (e.g., Broecker and Peng, 1982) and the assumption that calcite tests of benthic foraminifera faithfully record the ambient seawater $\delta^{13}C$. While the inverse relationship between seawater $\delta^{13}C$ and nutrient content holds true in a broad sense, seawater $\delta^{13}C$ can be affected by temperature dependent isotopic equilibration and net CO_2 transfer across the air–sea interface (Lynch-Stieglitz et al., 1995; Inoue and Sugimura, 1985; Mook et al., 1974; Vogel et al., 1970; Zhang et al., 1995). Also, the assumed relationship between seawater $\delta^{13}C$ and foraminiferal $\delta^{13}C$ has been shown to hold for *P. wuellerstorfi* by a number of core-top studies (Belanger et al., 1981; Graham et al., 1981; Duplessy et al., 1984; McCorkle et al., 1990; McCorkle et al., 1997). However, negative deviations of *P. wuellerstorfi* $\delta^{13}C$ from ambient seawater $\delta^{13}C$ have been pointed out under conditions of high surface productivity (Mackensen et al., 1993) as have positive deviations in relatively shallow depths in some parts of the World Ocean (e.g., McCorkle et al., 1998).

In the examined cores, *P. wuellerstorfi* was very scarce in the Holocene horizons but relatively more abundant in the glacial sections, consistent with the notion of a better carbonate preservation state in the Pacific during glacial than interglacial times (Farrell and Prell, 1989). While most isotope measurements consisted of multiple individuals, about a quarter of all measurements consisted of individual shells. In some cores, planktonic foraminifera *Globorotalia truncatulinoides* and *Globorotalia inflata* were also analyzed for $\delta^{18}O$ in order to obtain stratigraphic control.

Isotope measurements were made at Hokkaido University using a common acid bath coupled to a Finnigan MAT 251 isotope ratio mass spectrometer and at Lamont–Doherty Earth Observatory (LDEO) using a Multiprep carbonate preparation device coupled to a Micromass Optima isotope ratio mass spectrometer. The precision is better than ± 0.02‰ for both $\delta^{18}O$ and $\delta^{13}C$ on the MAT 251 at Hokkaido University, which is calibrated with in-house standards and NBS-20. The Micromass Optima at LDEO is calibrated with NBS-18, NBS-19 and in-house standards and has a precision of

Table 1
Cores

Core	Latitude (°N)	Longitude (°E)	Depth (m)	Reference
KH82-4-14	31.44	129.02	740	Oba (1983)
BO94-20 PN3PC	28.06	127.55	1058	Wahyudi and Minagawa (1997)
V20-133	32.58	140.34	1503	
V28-297	31.59	140.26	2047	
MR97-4 St.3	35.59	141.48	2308	Oba et al. (1999)
KT89-18 P4	32.09	133.54	2700	Oba and Yasuda (1992)
KT92-17 PC14	32.40	138.27	3252	Nishina (1995)
KT92-17 PC16	31.55	138.25	3320	Nishina (1995)

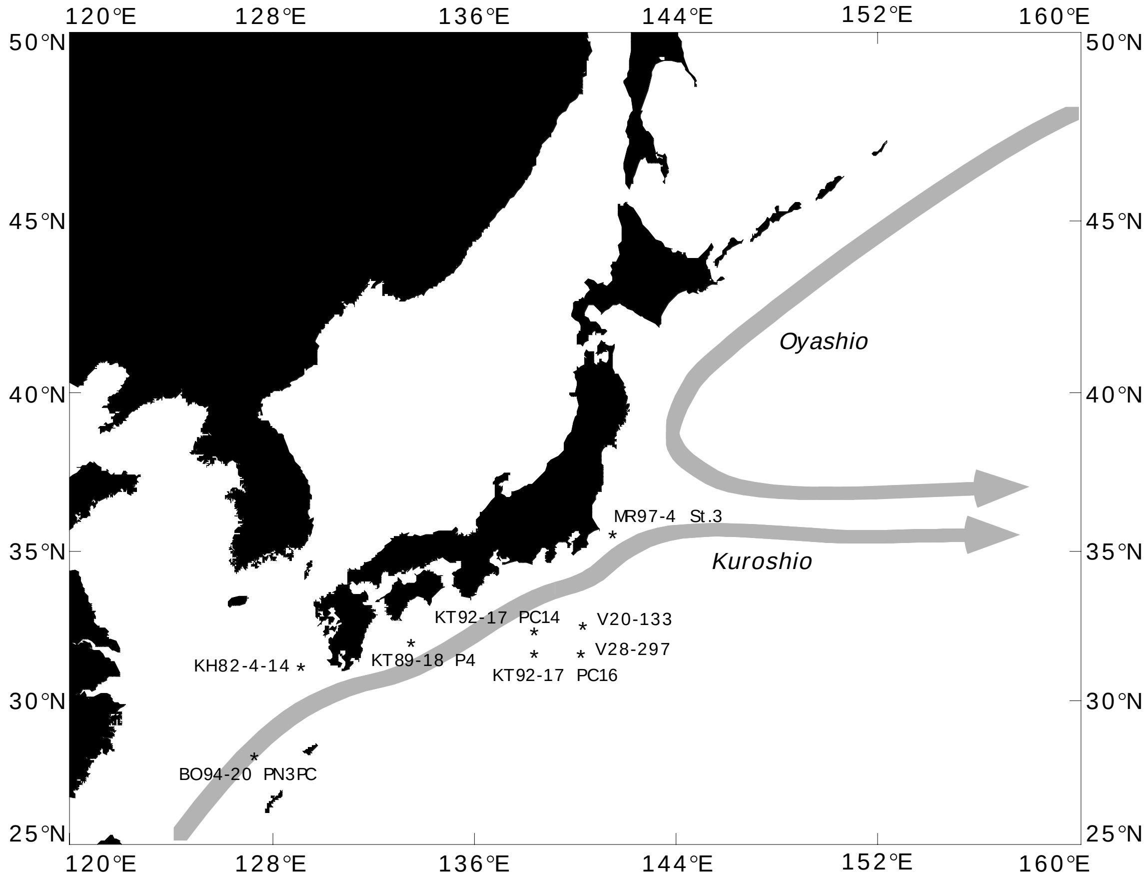

Fig. 1. Core sites and approximate positions of Kuroshio and Oyashio.

± 0.05‰ for $\delta^{13}C$ and ± 0.06‰ for $\delta^{18}O$ for samples larger than 30 μg. For samples <30 μg, the precision for both $\delta^{13}C$ and $\delta^{18}O$ deteriorates by 0.03‰.

2.1. *$\delta^{18}O$ data*

Holocene and glacial sediment horizons are determined on the basis of downcore $\delta^{18}O$ stratigraphies (Fig. 2). Cores KH82-4-14 and V28-297 contained no *P. wuellerstorfi* from their Holocene sediments, but Holocene *P. wuellerstorfi* $\delta^{18}O$ from other cores largely agree with the estimated $\delta^{18}O$ of calcite in equilibrium with ambient seawater (Fig. 3). However, $\delta^{18}O$ of 2.40‰ from PT89-18 P4 (2700 m) is lighter than the estimated value by 0.3‰. This light value is from one measurement and likely does not represent a mean Holocene value (Table 2). It is unlikely that downslope transport of foraminifera from shallower depth is the cause of this light value, since PT89-18 P4 was raised from a topographically flat part of the continental slope (Oba et al., 1999). Also, the Holocene value of 2.69‰ from V20-133 (1503 m) is heavier than the estimated value by 0.2‰, which may have resulted from upward mixing of glacial or deglacial foraminifera, since the core has only 20 cm of Holocene (Fig. 2).

The Holocene $\delta^{18}O$ become increasing light above 1500 m, reflecting higher temperatures (Fig. 3). The glacial $\delta^{18}O$ profile shows a clearer trend with depth, but amplitude is larger and the trend begins at significantly deeper water depth. There is a gradient of more than 1‰ between the maximum glacial $\delta^{18}O$ value of 4.52% at 2700 m and 3.35‰ at 1058 m, just below the influence of Kuroshio. In contrast, the Holocene gradient is about half the glacial gradient.

2.2. *$\delta^{13}C$ data*

Time-slice *P. wuellerstorfi* $\delta^{13}C$ values from each core were determined on the basis of their $\delta^{18}O$ (Table 2). The Holocene $\delta^{13}C$ depth profile generally shows increasingly heavy values with water depth, although the trend is not very tight (Fig. 4a). The scatter in the new data appears comparable in magnitude to the scatter in the far North Pacific data (Keigwin, 1998), but the Ontong Java Plateau data (Herguera et al., 1992) is much tighter. The larger $\delta^{13}C$ data scatter at higher latitudes may be related to increased seasonality in the surface productivity and its effect on benthic foraminiferal $\delta^{13}C$. Despite the scatter, the general depth trend in the new data is consistent with the data from the far

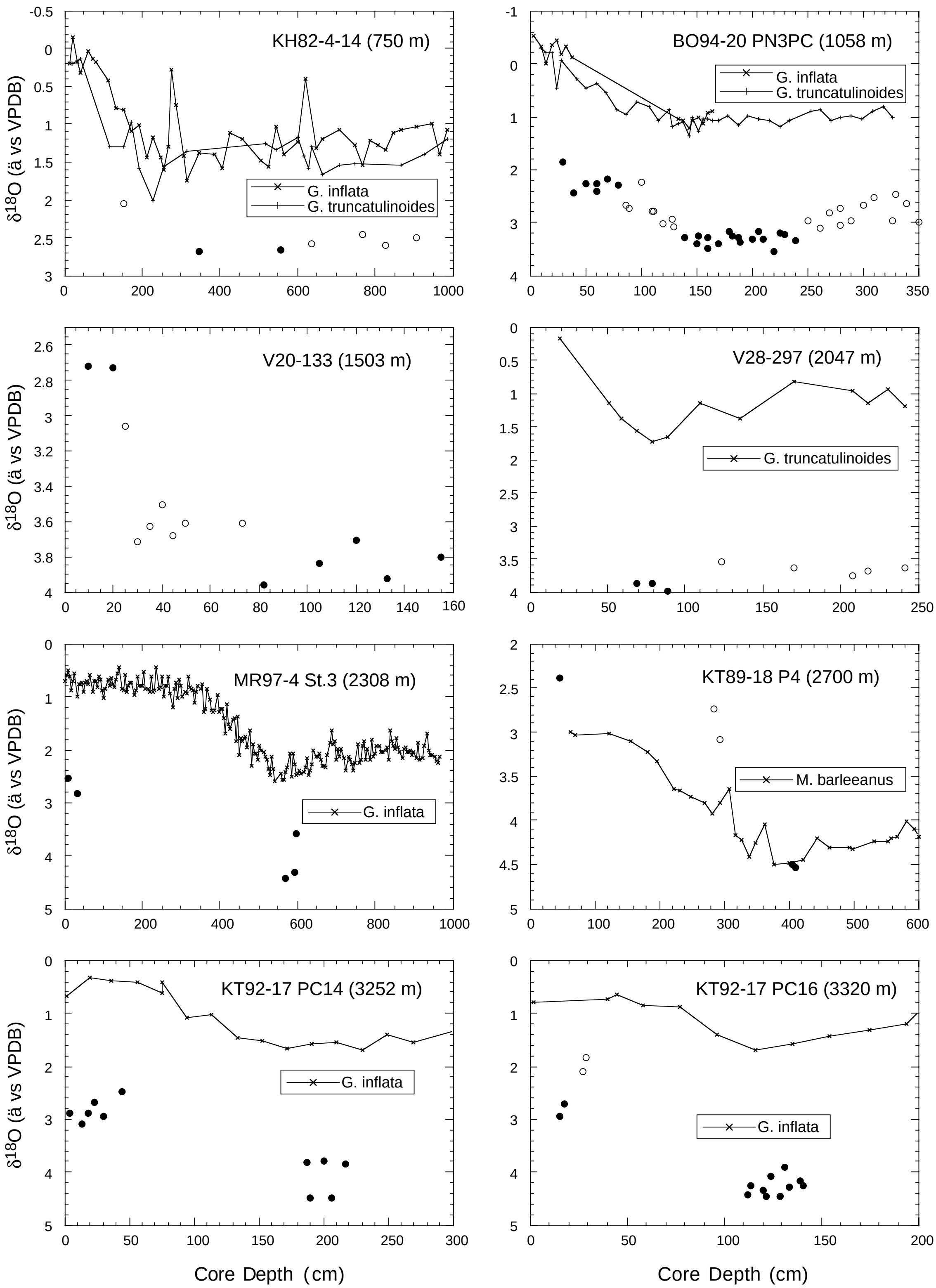

Fig. 2. Downcore oxygen isotope measurements. Circles are *P. wuellerstorfi* $\delta^{18}O$ values; filled circles were used to obtain the time-slice values (Table 2). Part of the *P. wuellerstorfi* data from BO94-20 PN3PC were previously reported by Wahyudi and Minagawa (1997) and Wahyudi (1997). Other data previously generated are *G. inflata* $\delta^{18}O$ from MR97-4 Station 3 (Oba et al., 1999) and *M. barleeanus* $\delta^{18}O$ from KT89-18 P4 (Oba and Yasuda, 1992). All other data were generated by this study.

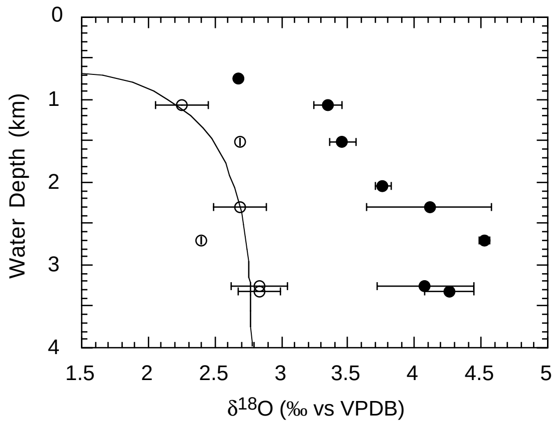

Fig. 3. Vertical profiles of the new *P. wuellerstorfi* $\delta^{18}O$ data from around Japan. Open circles are the Holocene data and filled circles are glacial data (Table 2). Line indicates the estimated calcite $\delta^{18}O$ in equilibrium with ambient seawater using the hydrographic data from GEOSECS station 225 (Fig. 2), the salinity–$\delta^{18}O$ relationship in the northwest Pacific (Keigwin, 1998), and the $\delta^{18}O$ equation from Lynch-Stieglitz et al. (1999).

North Pacific (Keigwin, 1998). The trend is also consistent with the modern distributions of nutrients and seawater (dissolved inorganic carbon) $\delta^{13}C$, which reaches the lowest values in the deep North Pacific centered at approximately 2000 m (Kroopnick, 1985).

The glacial $\delta^{13}C$ profile in contrast shows a clear enrichment with decreasing depth (Fig. 4b). Not plotted on Fig. 4b is an extremely low $\delta^{13}C$ value of $-1.75‰$ (1SD = $\pm 0.20‰$, $n = 3$) from MR97-4 St. 3 (Fig. 2, Table 2). This value is lower than all other values from this study by more than 1.5‰. Core MR97-4 St. 3 is located today near the southern edge of the highly productive Kuroshio and Oyashio mixing zone (Fig. 1). Its anomalously low glacial $\delta^{13}C$ may reflect artifact from increased productivity (Mackensen et al., 1993) due to a migration of the highly productive mixing zone over the core site. Such a migration would be consistent with a postulated glacial southerly shift of the Kuroshio (Chinzei et al., 1987).

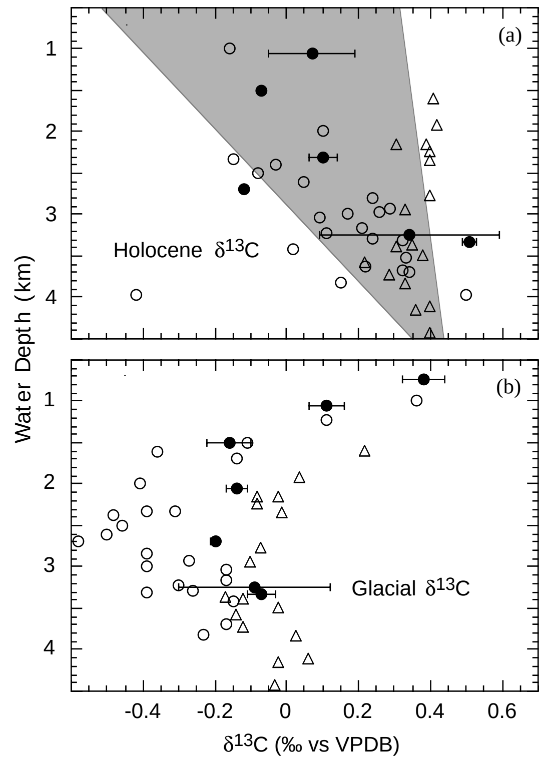

Fig. 4. Vertical $\delta^{13}C$ profiles for (a) Holocene and (b) LGM time slices. Filled circles are *P. wuellerstorfi* data from this study. Open circles are *Cibicidoides* species data from around 50°N (Keigwin, 1998), and open triangles are *P. wuellerstorfi* data from the equator (Herguera et al., 1992). Shading in subplot (a) is bounded by seawater $\delta^{13}C$ measurements from Ontong Java Plateau to the right and Emperor Seamounts to the left (McCorkle and Keigwin, 1994).

3. The deep Pacific during the LGM

3.1. Hydrography

The Duplessy et al. (1988a) one-dimensional vertical profile of glacial benthic foraminiferal $\delta^{13}C$ suggested importantly the presence of a high $\delta^{13}C$ water mass at 700–2600 m. If one takes their vertical profile at face

Table 2
Time slice *P. Wuellerstorfi* stable isotope data

Core	Depth (m)	Holocene			Glacial			Laboratory
		n	$\delta^{18}O$	$\delta^{13}C$	n	$\delta^{18}O$	$\delta^{13}C$	
KH82-4-14	740				2	2.68±0.01	0.38±0.06	LDEO
BO94-20 PN3PC	1058	7	2.25±0.20	0.07±0.12	16	3.35±0.10	0.11±0.05	LDEO
V20-133	1503	2	2.69±0.01	−0.07±0.01	5	3.85±0.10	−0.19±0.06	LDEO
V28-297	2047				3	3.91±0.06	−0.14±0.03	LDEO
MR97-4 St.3	2308	2	2.69±0.20	0.10±0.04	3	4.11±0.47	−1.75±0.20[a]	Hokkaido
KT89-18 P4	2700	1	2.40	−0.12	2	4.52±0.04	−0.20±0.01	Hokkaido
KT92-17 PC14	3252	6	2.83±0.21	0.34±0.25	5	4.08±0.36	−0.09±0.21	Hokkaido
KT92-17 PC16	3320	2	2.83±0.16	0.51±0.02	10	4.26±0.18	−0.07±0.04	Hokkaido

All isotope data are reported in ‰ (vs VPDB). Part of the BO94-20 PN3PC data was previously published (Wahyudi, 1997; Wahyudi and Minagawa, 1997).

[a] Not plotted in Fig. 6 due to its anomalous value; see text.

value (i.e., reflecting nutrient content), the high $\delta^{13}C$ water mass at 700–2600 m would indicate a low nutrient, well-ventilated water mass. This interpretation is in sharp contrast with the modern macronutrient distribution of the Pacific (Fig. 5). Today, the most nutrient-rich water in the World Ocean is found in the North Pacific below the main thermocline and above approximately 2500 m. This distribution is a direct consequence of the modern deep ocean circulation, whose meridional overturning in the Pacific is "nutrient-trapping". In a simplified view, the deep Pacific is ventilated from the south by the densest variety of the lower Circumpolar Deep Water, including the Antarctic Bottom Water, that upwells to mid-depth and returns south as the Pacific Deep Water (Schmitz, 1996). Nutrients accumulate continually throughout this deep-water journey, including the return path. However, because the return flow mixes with the relatively nutrient-depleted Circumpolar Deep Water, the nutrient content of the Pacific Deep Water decreases along the path of its flow. For this reason, nutrients are "trapped" in the mid-depth of the North Pacific. These features, described for nitrate in Fig. 5, are also obvious in the modern seawater $\delta^{13}C$ distribution of Kroopnick (1985).

Notwithstanding its significant departure from the modern oceanography, the basic interpretation of Duplessy et al.'s (1988a) vertical profile of glacial benthic foraminiferal $\delta^{13}C$ as reflecting the nutrient content of the deep Pacific has gained support over the years.

A two-dimensional view of our new and published benthic foraminiferal $\delta^{13}C$ data from the Pacific (Duplessy et al., 1988a; Herguera et al., 1992; Keigwin, 1998; Matsumoto and Lynch-Stieglitz, 1999; Matsumoto et al., 2001) show clearly the change in their distribution between the Holocene and LGM (Fig. 6). The Holocene map shows the lowest $\delta^{13}C$ (i.e., the oldest and most nutrient-rich water mass) in the North Pacific centered around ~2000 m (Fig. 6a), which is consistent with the large-scale distribution of nitrate today (Fig. 5). It also indicates that, despite the extraneous effects of benthic foraminiferal $\delta^{13}C$ as a nutrient proxy, a distribution of benthic foraminiferal $\delta^{13}C$ that spans a sufficiently large area can capture the first-order feature of the corresponding nutrient distribution (Matsumoto and Lynch-Stieglitz, 1999). The glacial $\delta^{13}C$ distribution shows that the lowest $\delta^{13}C$ data have shifted in water depth to 2500–3000 m, compared to ~2000 m during the Holocene (Fig. 6b). Above

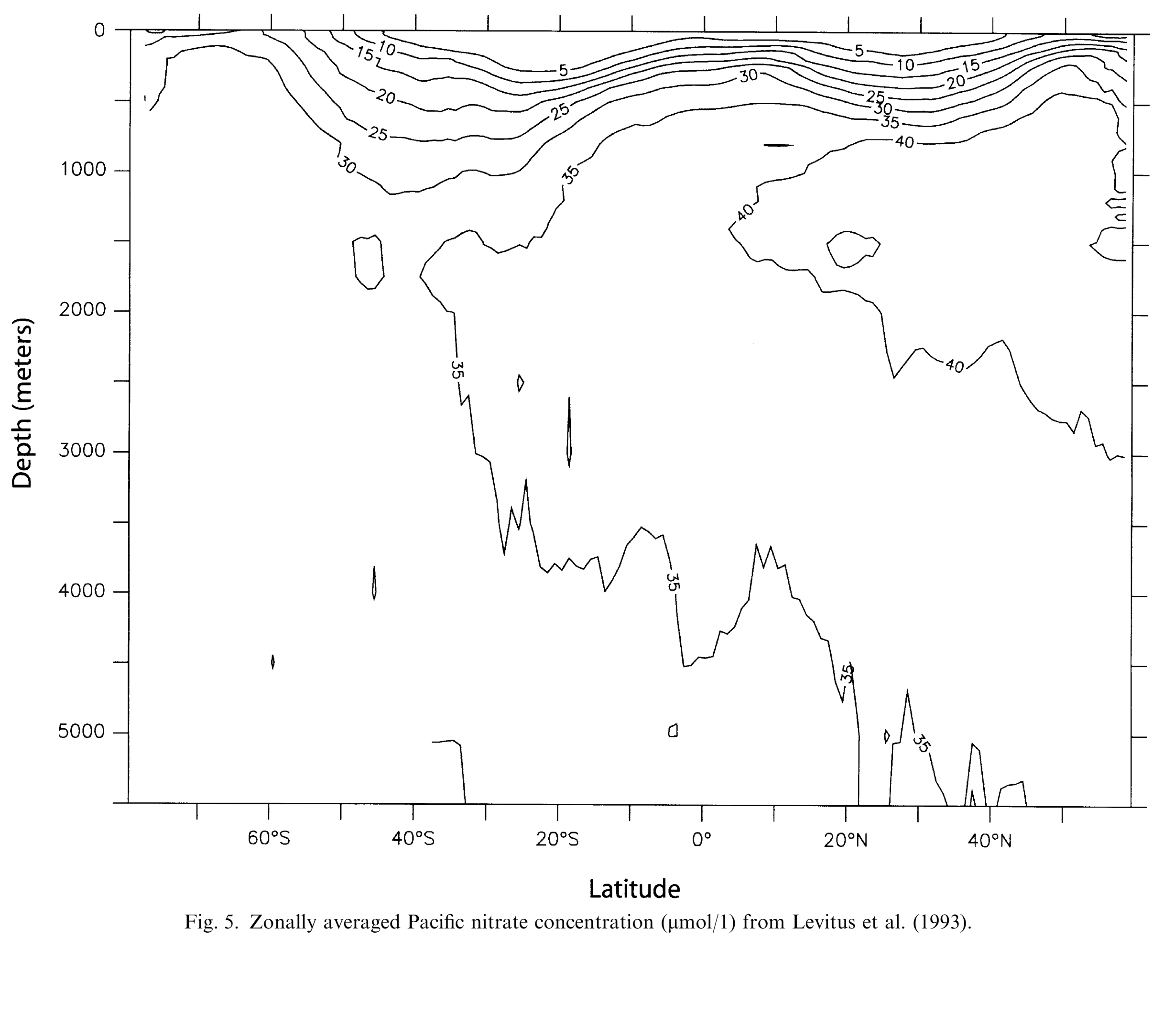

Fig. 5. Zonally averaged Pacific nitrate concentration (μmol/l) from Levitus et al. (1993).

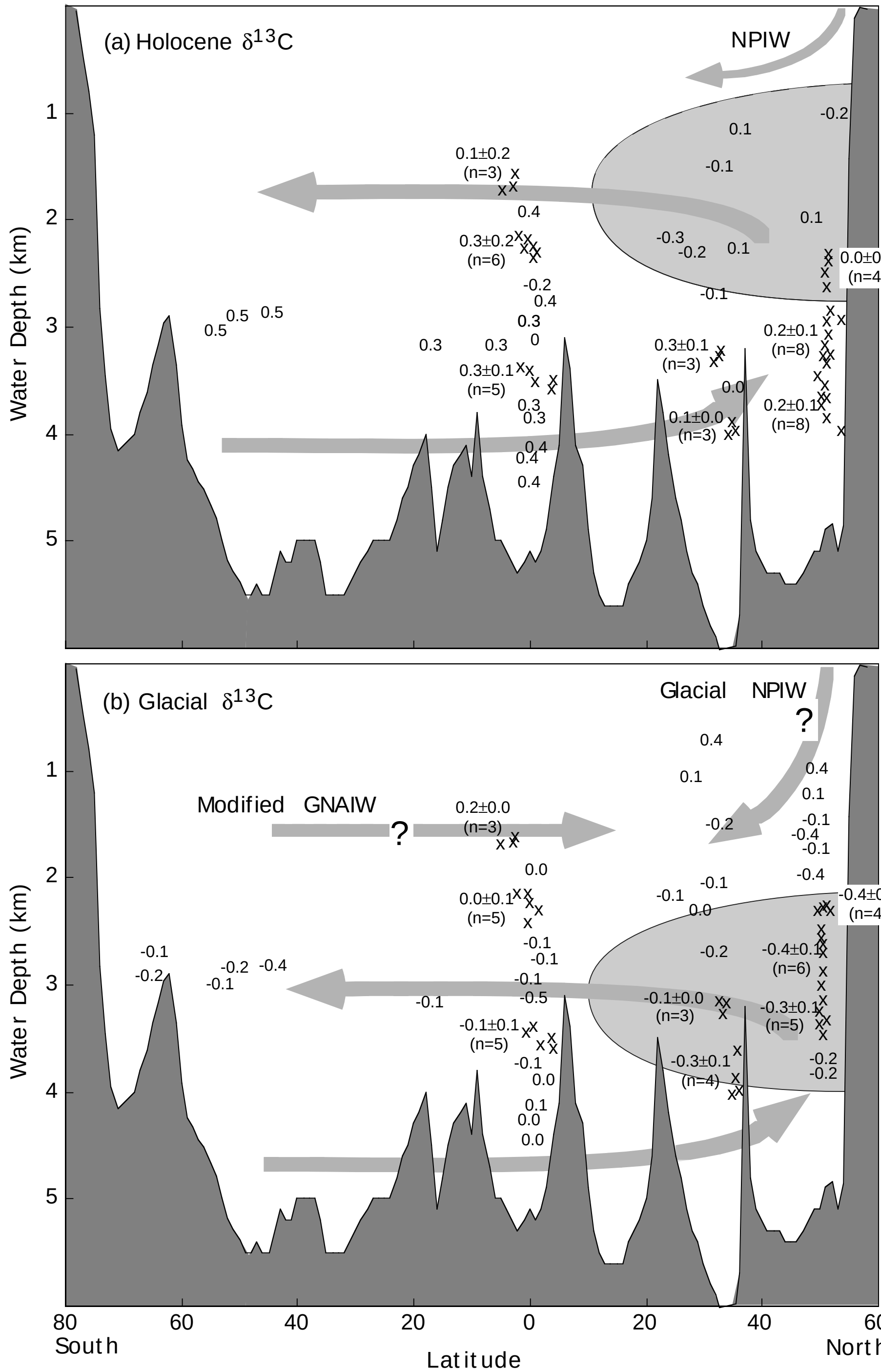

Fig. 6. Meridional distribution of benthic foraminiferal $\delta^{13}C$ in the western Pacific Ocean. Data from Duplessy et al. (1988a), Herguera et al. (1992), Matsumoto and Lynch-Stieglitz (1999), and Matsumoto et al. (2001) were measured on *P. wuellerstorfi*. Measurements from Keigwin (1998) were on *Cibicidoides* species. The lowest $\delta^{13}C$ indicated by shading is found in the deep North Pacific centered around 2000 m water depth during the Holocene and around 3000 m during glaciation. Arrows indicate suggested circulation. The higher $\delta^{13}C$ in the upper 2000 m in the North Pacific reflect a local source of newly ventilated glacial NPIW, influx of modified GNAIW, or both.

2000 m, the glacial $\delta^{13}C$ become increasingly enriched, which may reflect one or some combination of nutrient depletion, air–sea gas exchange signal, or a positive deviation of benthic foraminiferal $\delta^{13}C$ from the ambient seawater $\delta^{13}C$. With the exception of the positive $\delta^{13}C$ deviation, which would not substantiate the interpretation presented thus far, both nutrient depletion and air–sea gas exchange signal would indicate that the water mass above ~2000 m was quite distinct from the underlying water mass.

Two other types of measurement indicate that the deep hydrographic divide in the glacial Pacific, as inferred from benthic foraminiferal $\delta^{13}C$, is real. First, the vertical profiles of benthic foraminiferal $\delta^{18}O$ at the Ontong Java Plateau (Herguera et al., 1992), in the far North Pacific (Keigwin, 1998), and in the vicinity of Japan (this work; Fig. 3) all show more highly depleted $\delta^{18}O$ above 2000 m than in deeper waters during the LGM compared to the Holocene. While it is unclear whether the $\delta^{18}O$ gradient predominantly reflects changes with depth of temperature (Herguera et al., 1992) or salinity (Keigwin, 1998), it is a clear indication of a hydrographic boundary. An important future contribution would be to determine the relative contribution of salinity and temperature to the observed $\delta^{18}O$ gradient by independently estimating the temperature component, for which benthic foraminiferal Mg/Ca measurements appear to hold some promise (Rathburn and DeDekker, 1997; Rosenthal et al., 1997; Toyofuku et al., 2000). The second independent support for a deep hydrographic boundary is provided by apparent ventilation ages, estimated from paired radiocarbon ages of benthic and planktonic foraminifera. Available estimates from the deep Pacific indicate older, relatively poorly ventilated water mass during LGM than today (Broecker et al., 1988; Duplessy et al., 1988b; Shackleton et al., 1988). In contrast, a pair of radiocarbon ages at around 1000m (Duplessy et al., 1988b) indicate a much younger, better ventilated water mass, which is consistent with a high resolution record of bioturbation (indicative of oxygen content) from the Santa Barbara basin (Kennett and Ingram, 1995).

On the other hand, benthic foraminiferal Cd/Ca ratio, another major nutrient proxy commonly used in paleoceanography, does not indicate a deep hydrographic boundary in the glacial Pacific in the same sense as $\delta^{13}C$. In a direct conflict with benthic $\delta^{13}C$ data, available Cd/Ca data indicate that the nutrient levels of the upper waters of the deep glacial Pacific are "consistently higher than nutrient levels of deeper waters" (Boyle, 1992). This discrepancy has two potential causes. The first is that the elevated glacial $\delta^{13}C$ in the upper 2000 m compared to deeper waters is an artifact of positive benthic foraminiferal $\delta^{13}C$ deviation from the ambient seawater $\delta^{13}C$. This feature is apparently not ubiquitous, since such a deviation has been reported only from some parts of the ocean (Ahmad et al., 1995; Slowey and Curry, 1995; McCorkle et al., 1998; Oppo and Horowitz, 2000) and not others (e.g., Lynch-Stieglitz et al., 1994). To our knowledge, it has not been reported from the Pacific basin. That the Holocene benthic foraminiferal $\delta^{13}C$ is able to capture the first-order nutrient distribution of the modern Pacific (Figs. 5 and 6a) suggests that the $\delta^{13}C$ artifact, if it occurs in the glacial Pacific at all, does not mask the large-scale nutrient distribution. The second potential cause is that the glacial Cd/Ca data from the deep Pacific are compromised by selective removal of trace metals from foraminiferal calcite tests accompanying dissolution (McCorkle et al., 1995). This may be more probable a cause, because this feature has been observed in the modern Pacific (McCorkle et al., 1995), where deep waters are close to being undersaturated everywhere (Broecker and Peng, 1982).

While the verdict is not entirely clear, the ensemble of available evidence suggests that this discrepancy between benthic foraminiferal $\delta^{13}C$ and Cd/Ca in the glacial Pacific arises from an artifact of Cd/Ca as a strict nutrient proxy. It is worth noting that a more nutrient depleted upper waters, as indicated in the glacial Pacific by $\delta^{13}C$ data (Fig. 6b), is also inferred in the Atlantic and the Indian Ocean during the LGM from both $\delta^{13}C$ and Cd/Ca (Boyle, 1992). A very important work in the future is to evaluate this discrepancy more rigorously with core-top measurements of benthic foraminiferal Cd/Ca from the modern North Pacific. If the above argument were correct, benthic foraminiferal Cd/Ca would be depleted relative to ambient seawater nutrient content.

3.2. Circulation

There appear to be two candidates for the source of the nutrient-depleted upper water mass in the glacial Pacific: (1) a northern source water mass that is newly ventilated in the North Pacific (Duplessy et al., 1988a; Herguera et al., 1992; Keigwin, 1998) and (2) a southern source water mass, which could have originated in the North Atlantic as GNAIW (Lynch-Stieglitz and Fairbanks, 1994; Lynch-Stieglitz et al., 1996), but may have undergone substantial modification in the Southern Ocean (Oppo and Horowitz, 2000).

On a regional scale, a logical explanation would be a local source of newly ventilated water mass. Nutrient depletion or significant air–sea gas exchange signal implied from enriched $\delta^{13}C$ are both surface processes which, if local, would "require the production of a new water mass to mix, diffuse, or convect surface-influenced waters as deep as 2 km" (Keigwin, 1998). This water mass may have been a stronger and more deeply penetrating version of today's North Pacific Intermediate Water (NPIW) (Duplessy et al., 1988a; Mix et al., 1991; Herguera et al., 1992; Oba and Yasuda, 1992; Keigwin, 1998). Although a water mass that reaches 2000 m should not be called an intermediate water in the sense of modern physical oceanography, here we will refer to it as glacial NPIW (GNPIW) for the lack of a better name. If GNPIW is genetically related, the modern NPIW, the site of GNPIW formation may have been Okhotsk Sea, where NPIW formation today is geographically tied to (Reid, 1965; Reid, 1973; Alfultis and Martin, 1987; Talley, 1991; Freeland et al., 1998;

Wong et al., 1998). The large continental shelf in the Bering Sea also offers a place for sea ice to form in coastal polynas (Sancetta et al., 1985) and thereby form dense waters.

One possible southern source water is GNAIW, which may have reached the Pacific Ocean after being modified to some extent during its transit from the Atlantic via the Southern Ocean (Lynch-Stieglitz and Fairbanks, 1994; Lynch-Stieglitz et al., 1996). This scenario is also consistent with the benthic foraminiferal $\delta^{13}C$ data presented here. While the "high" $\delta^{13}C$ above 2000 m in the glacial North Pacific (Figs. 5 and 6) is indeed much higher than the Holocene, it is actually still lower by more than 1‰ compared to the $\delta^{13}C$ of 1.3‰ in the North Atlantic at similar depths (Matsumoto and Lynch-Stieglitz, 1999). This gradient would be expected if GNAIW reached the Pacific Ocean via the Antarctic Circumpolar Circulation, because during its long transit, GNAIW would continue to accumulate nutrients from remineralization of raining organic matter and thus acquire lower $\delta^{13}C$.

Recently, Oppo and Horowitz (2000) have concluded on the basis of benthic foraminiferal Cd/Ca and $\delta^{13}C$ analysis that GNAIW did not significantly influence the glacial Pacific nutrient content. Instead, they suggest that the upper Pacific was bathed by a "subantarctic water", whose characteristic glacial nutrient content and $\delta^{13}C$ gas exchange signal are defined in the subantarctic Indian Ocean (Lynch-Stieglitz et al., 1996). Oppo and Horowitz (2000) suggest that the Cd/Ca and $\delta^{13}C$ data can be reconciled with strong export of deep water out of the Atlantic basin (Yu et al., 1996) and a GNAIW source (Lynch-Stieglitz et al., 1996), if GNAIW was modified in the Southern Ocean. The distinction between GNAIW and "subantarctic water" on the basis of benthic foraminiferal Cd/Ca and $\delta^{13}C$ then reflects the degree to which GNAIW was modified in the Southern Ocean before being exported out to the Pacific.

A comparison of benthic foraminiferal $\delta^{18}O$, essentially a conservative tracer that reflects seawater temperature and salinity, from the North Atlantic and North Pacific above 2000 m (Fig. 7) is also consistent with GNAIW reaching the Pacific. The average values of the LGM $\delta^{18}O$ data from the two oceans are statistically indistinguishable, implying that GNPIW could indeed be composed largely of GNAIW. This would be analogous to the situation today, where waters in the upper North Pacific (1–2 km depth) today, are dominated by waters which ultimately have their source in the North Atlantic. The quasi-conservative tracer $PO_4*(= PO_4+O_2/175 + 1.95)$ of deep water in the North Pacific is essentially that of Circumpolar Deep Water, which itself is an equal mixture of the North Atlantic Deep Water and Antarctic Bottom Water (Broecker et al., 1998). As during the LGM, the

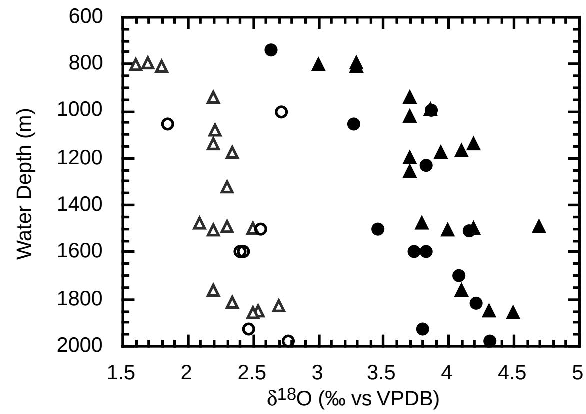

Fig. 7. "Intermediate" depth North Atlantic (triangles) and North Pacific (circles) $\delta^{18}O$ measured on *Cibicidoides* species (including *P. wuellerstorfi*). Open and filled symbols represent, respectively, the Holocene and glacial data. There is no statistically significant difference between the average $\delta^{18}O$ of the Atlantic and Pacific data for the two time slices. Data are from this study and previous studies (Duplessy et al., 1984; Curry et al., 1988; Duplessy et al., 1988a; Herguera et al., 1992; Sarnthein et al., 1994; van Geen et al., 1996; Keigwin, 1998).

Holocene $\delta^{18}O$ average values from the two oceans are also statistically indistinguishable (Fig. 7), reflecting their common heritage. This indicates that benthic foraminiferal $\delta^{18}O$ is insensitive to the degree of temperature and salinity modifications that the North Atlantic Deep Water has experienced during the Holocene prior to entering the Pacific. The failure of the glacial $\delta^{18}O$ data to distinguish GNAIW from GNPIW (Fig. 7) therefore implies, if GNAIW reached the Pacific, that the degree to which GNAIW was modified was no more extreme than how much the North Atlantic Deep Water is modified today during its transformation into Circumpolar Deep Water and then into the Pacific Deep Water.

The paleoceanographic evidence presented thus far cannot distinguish the two possible sources for the $\delta^{13}C$-enriched water mass in the upper 2000 m of the glacial Pacific, either a locally ventilated "intermediate water" from the north or some kind of modified "intermediate water" from the south. It is worth noting that an ocean box model simulating glacial intermediate waters shows that an export of intermediate out of a basin is necessary to simultaneously raise its intermediate water $\delta^{13}C$ and deplete its nutrient content (Sigman, 1997). Otherwise, the model shows a "nutrient trapping" condition, where the intermediate water that is not exported out, ends up upwelling in the low latitudes, bringing to the surface nutrients that it has accumulated prior to upwelling. GNAIW is a good analog of the model behavior when intermediate water is exported out. The high $\delta^{13}C$ of GNPIW compared to the Holocene then also seems to require that it get exported out of the Pacific basin, which would favor the northern source for GNPIW. On the other hand, if air–sea exchange processes at deep-water formation sites were to change in favor of a lower $\delta^{13}C$ relative to nutrient content, by mass balance, there must be higher $\delta^{13}C$ relative to nutrients in the upper portion of the ocean.

However, the 1‰ gradient in benthic foraminiferal $\delta^{13}C$ between GNAIW and GNPIW appears to favor the southern source. Of course, it is very possible that both sources of GNPIW existed, since a water-mass formation in the glacial North Pacific and GNAIW reaching the Pacific do not appear mutually exclusive. This issue cannot be resolved here, but its resolution would make a very important contribution to paleoceanography by filling a major gap in our understanding of the glacial ocean.

The circulation and origin of the very low $\delta^{13}C$ water mass in the glacial Pacific below 2000 m (Fig. 6b) hinges on the discrepancy between benthic foraminiferal $\delta^{13}C$ and Cd/Ca. At face value, glacial Cd/Ca data (and air–sea gas exchange signal of $\delta^{13}C$ (Lynch-Stieglitz and Fairbanks, 1994)) would suggest the formation of a newly ventilated deep water in the North Pacific (Boyle, 1992; Ohkouchi et al., 1994). This is because glacial Cd/Ca in the deep North Pacific is lower (i.e., lower nutrient content) than in the deep eastern equatorial deep Pacific. However, the ensemble of available evidence, including the selective removal of trace metals from foraminiferal calcite tests accompanying dissolution (McCorkle et al., 1995), appears to favor $\delta^{13}C$ as a more faithful recorder of nutrients in the deep Pacific as discussed above.

Assuming then that the large-scale nutrient distribution is correctly depicted by the $\delta^{13}C$ data, the presence of the most nutrient-rich water in the far north indicates that the basic thermohaline circulation as we have today operated in the glacial Pacific (Matsumoto and Lynch-Stieglitz, 1999). The northward penetrating bottom water rises today in the North Pacific to about 2000 m where nutrients accumulate and thereafter returns southward. During the LGM, bottom water from the south probably rose only to 2500–3000 m and that the outflow occurred deeper than today. That the outflow occurred at deeper depth than today (Fig. 6) suggests that the flux of the densest water from the south that filled the glacial Pacific was smaller relative to the flux(es) of the GNPIW. In addition, the clear distinction of the GNPIW from the deeper water mass in terms of $\delta^{13}C$ (Fig. 6), $\delta^{18}O$ (Herguera et al., 1992; Keigwin, 1998; this work), and apparent ventilation ages (Sikes et al., 2000) indicates that the density contrast between the two water masses was relatively large. This in turn suggests that the mode of formation of those waters was quite distinct. While paleoceanographic studies to date are unable to provide clues as to how the densest waters were formed in the Southern Ocean during the LGM, it seems likely that the colder Antarctic continent and its margins contributed to their formation. In comparison, the involvement of Antarctic continent in the formation of the more buoyant GNPIW was probably less.

4. Conclusions

Examination of available nutrient proxy and other data from the Pacific Ocean indicates the presence of two different water masses in the glacial Pacific that were separated at approximately 2000 m water depth. The upper water mass has distinctly depleted $\delta^{18}O$ and enriched $\delta^{13}C$, which may either reflect a water mass formed locally in the North Pacific, a modified GNAIW reaching the Pacific Ocean via the Southern Ocean, or some combination of the two as they are not necessarily mutually exclusive. The origin of this water mass cannot be resolved at this time. Below 2000 m, the deep water was very rich in nutrients, which suggests a more vertically compressed but similar deep circulation as today. The outflow of the lower water mass (analogous to the modern Deep Pacific Water) occurred deeper than today, suggesting that the influx of the southern source water was relatively weak compared to today. The boundary between GNPIW and the nutrient-rich deep water in the glacial North Pacific at about 2000 m is consistent with reported bathyal nutrient front at similar water depth in the Atlantic (Curry and Lohmann, 1982; Curry et al., 1988; Duplessy et al., 1988a; Boyle, 1992; Oppo and Lehman, 1993), the Indian (Kallel et al., 1988; McCorkle et al., 1998), and the Southern Ocean (Lynch-Stieglitz et al., 1996).

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