

Carbon and oxygen isotopic records of benthic foraminifera from the Northeast Indian Ocean: implications on glacial–interglacial atmospheric CO₂ changes

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Abstract

We present here continuous records of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in benthic foraminifera, extending well into the last ice age, in two piston cores from the Andaman Sea (sill depth ~ 1.3 km) and the Bay of Bengal (3 km). These show that, contrary to the previous reports, the glacial to interglacial shift in $\delta^{13}\text{C}$ at mid-depths in the Northeast Indian Ocean was indistinguishable from the mean oceanic $\delta^{13}\text{C}$ change, negating a more vigorous renewal of intermediate waters globally during the glacial time. The corresponding $\delta^{13}\text{C}$ shift in deep waters is estimated to be about 50% larger than that reported previously. Jointly with some recent data from the Pacific, our results indicate a modest glacial–Holocene shift in the intermediate to deep water chemical gradients in the Indo-Pacific as a whole, implying that it was perhaps not the dominant mechanism for the glacial–interglacial atmospheric CO₂ variations. Also, in conflict with previous work, our measurements suggest significant cooling of both the intermediate and deep waters during the glacial time. The high-resolution records from the Andaman Sea help reconstruct paleoenvironmental changes at intermediate depths during the last deglaciation. Rapid increases in $\delta^{13}\text{C}$ occurring in two stages during the early deglaciation appear to have been caused by the fluctuations in the North Atlantic Deep Water production. A negative excursion in $\delta^{13}\text{C}$ during the mid-deglaciation is ascribed to enhanced nutrient regeneration at mid-depths associated with the greatly intensified summer monsoon around the Pleistocene–Holocene boundary.

1. Introduction

Analyses of fossil benthic foraminifera in sediment cores from the Northeast Indian Ocean (NEIO) for stable isotopes of oxygen and carbon

have been restricted previously to selected sections of the cores [1,2]. The results have been interpreted to propose that a discontinuity in the vertical hydrochemical structure (at ~ 2 km depth) could have existed in the region during the last glacial maximum (LGM). Waters below the boundary were significantly colder and/or saltier (high $\delta^{18}\text{O}$), those above were about the same as today. In addition, the deep waters were nutrient enriched (low $\delta^{13}\text{C}$), while the upper waters were

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nutrient-depleted (high $\delta^{13}\text{C}$). Some of the cores examined for the stable isotopes [1] have also been subjected to cadmium analysis in foraminifera shells [3]. The Cd data appear to be consistent with the $\delta^{13}\text{C}$ results, in that the Cd content above 2.5 km during the LGM has been found to be about 40% lower than the modern values.

A transfer of nutrients from the intermediate to deep waters during the LGM has been shown by Boyle [4,5] to be a potentially important mechanism for the glacial decrease in atmospheric CO_2 content as observed in polar ice cores [6]. Isotopic evidence for such a shift in the vertical chemical structure is unequivocal for the Atlantic Ocean [7–11]. Outside the Atlantic, the most pronounced glacial to interglacial change in the intermediate to deep water chemical gradients is widely believed to have occurred in the NEIO, and so the aforementioned results from the NEIO form an important pillar on which Boyle's hypothesis rests [4,5]. However, independent verification of these results has not been sought so far from time series analyses utilizing the species whose shells reliably record the isotopic composition of the ambient water. We present here, for the first time, two such records (one representing the changes in intermediate and one in deep waters) which not only provide a more reliable measure of the glacial to interglacial reorganization of the vertical chemical structure in the NEIO, but also help reconstruct high-resolution paleoenvironmental changes at intermediate depths during the deglaciation.

2. Hydrography of coring sites

The two cores (RC12-339 and RC12-344) used in this study were raised from the Bay of Bengal and the Andaman Sea, respectively (Fig. 1). While RC12-339 was collected from the northern end of the Ninetyeast Ridge (water depth 3010 m), RC12-344 was recovered from the continental slope off Burma (water depth 2140 m).

The NEIO comes under the influence of the seasonally changing monsoon gyre [12]. Excessive runoff from several major rivers results in low

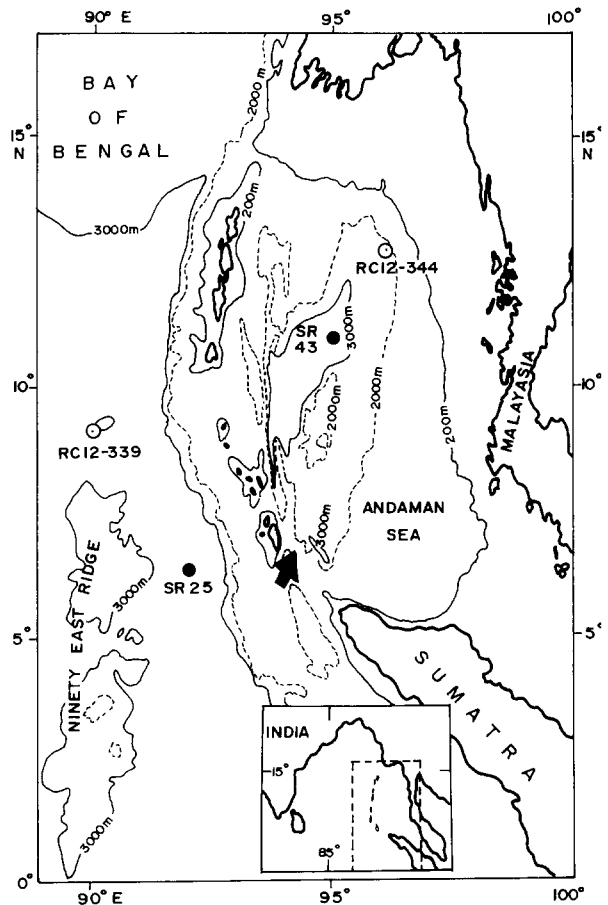


Fig. 1. Locations of the cores (circles) and U.S.N.S. *Serrano* stations (dots). Only 200 m, 2000 m and 3000 m depth contours are shown for brevity. The arrow shows where the Bay of Bengal waters spill over the sill (depth ~ 1.3 km) beneath the Great Passage into the Andaman Basin.

surface salinities [13,14]. However, in spite of the immense freshwater discharge, concentrations of nutrients in the Bay of Bengal surface waters are extremely low even near the mouths of the major rivers [14,15]. At Sandheads, the mouth of the Hooghly river, for example, the concentrations of nitrate, phosphate and silicate are typically < 0.1 , ~ 0.5 and $\sim 2 \mu\text{M}$, respectively [15]. It appears that the rivers flowing into the Bay of Bengal are characterized by modest dissolved nutrient levels and the limited quantities of nutrients brought by them may be largely removed within the estuaries [16]. This, in conjunction with the strong stratification, leads to low primary production [17].

Unlike the Arabian Sea, the thermohaline structure of subsurface waters in the Bay of Bengal is quite smooth. A rather broad salinity maximum occurs at the 300–600 m depth, after which salinity decreases slowly and monotonously with depth (Fig. 2). The intermediate-depth water composition in the NEIO is greatly influenced by circulation in the equatorial Indian Ocean. Distributions of properties along the horizontal surfaces show steep north–south gradients across the equatorial belt and gentle east–west gradients, with the isolines running parallel to the latitude [13]. Also, the equatorial water, formed through mixing of various water masses originating in the northern and southern hemispheres and those advecting from the Pacific Ocean, is nearly homogenous vertically within the depth range 100–1000 m [18]. Thus, the tracer distributions strongly suggest that the flow in the equatorial region in the upper kilometre or so is mostly zonal. This seems to result in very little trans-equatorial exchange of intermediate waters, except along the western boundary [18,19]. Consequently, the waters originating in the southern hemisphere (Antarctic Intermediate Water (AAIW) and Subantarctic Mode Water (SMW)) and in the Indonesian seas do not flow directly into the NEIO. Instead, the subsurface NEIO (to a depth of ~ 1 km) is mostly filled by the North Indian High-salinity Intermediate Water

(NIHSIW) [12], brought from the Arabian Sea by the Monsoon Current [20].

The NIHSIW itself may be formed through mixing of: (1) the intermediate water (a mixture of SMW, AAIW and the through-flow from the Pacific) that crosses the equator in the western Indian Ocean; (2) the high-salinity outflows from the Persian Gulf and the Red Sea; and (3) the water upwelling from below [19]. The salinity maximum observed in the NEIO represents the core of the NIHSIW. This water mass is also characterized by a pronounced oxygen minimum (Fig. 2) because it is derived from the Arabian Sea. Within the NEIO itself, the oxygen consumption at depth is not very large, due to the modest biological productivity of surface waters, coupled with the rapid sinking of particulate organic matter through the water column [15].

The deep waters in the Bay of Bengal are of circumpolar origin and are supplied by two western boundary currents flowing along the Central Indian and the Ninetyeast ridges [21]. The bottom water upwells as it moves northward, eventually to return at shallower depths. Significantly, the rate of deep water upwelling may be several times higher in the Indian Ocean relative to the Pacific or the Atlantic [21]. As a result of this, any changes in the composition of bottom waters could readily affect the water composition at shallower depths.

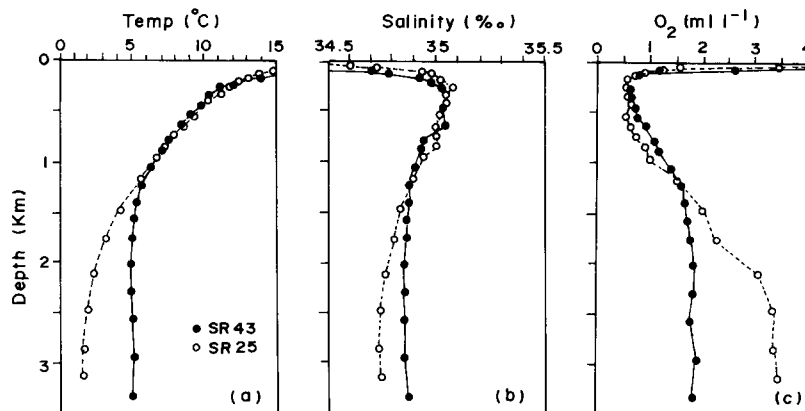


Fig. 2. Vertical profiles of (a) temperature, (b) salinity and (c) dissolved oxygen at two representative stations located in the Bay of Bengal (SR 25) and the Andaman Sea (SR 43). Near-surface data have not been included.

The Andaman Sea is a marginal basin, lying to the east of the Andaman-Nicobar Ridge, with a maximum water depth of ~ 4.4 km. Deep water

exchange with the Bay of Bengal is constrained by relatively shallow sills, the deepest of which is located beneath the Great Passage in the south.

Table 1
Stable isotope data (percentage versus PDB) from the Cores RC12-344 and RC12-339

RC12-344				RC12-339			
Depth (cm)	<i>Cibicidoides pachyderma</i>		<i>Planulina wuellerstorfi</i>		Depth (cm)	<i>C. kullenbergi</i> + <i>P. wuellerstorfi</i>	
	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$		$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
25	1.958	0.193	1.801	0.306	7	2.447	0.318
35	1.894	0.184			18	3.009	0.291
45	2.047	0.176			28	3.210	0.059
55	2.097	0.192	1.950	0.202	38	3.011	0.098
65			1.768	0.184	46	3.986	-0.362
75			1.885	0.232	54	3.880	-0.108
85			1.923	0.284	58	3.947	-0.200
105			2.185	0.109	64	3.897	-0.207
115			2.365	0.073	69	3.735	-0.104
127			2.214	0.053	73	3.876	-0.267
135	2.233	-0.019	2.364	-0.068	88	3.786	-0.106
145			2.317	-0.086	92	3.823	-0.050
155	2.254	-0.004	2.228	0.095	95	3.806	-0.147
165	2.210	0.067			100	3.604	-0.023
175	2.406	-0.029	2.394	0.052			
185	2.551	-0.146	2.608	-0.071			
194	2.719	-0.190	2.632	-0.101			
206	2.909	-0.185					
215	2.728	-0.142					
225	2.312	0.024					
234	2.699	0.049					
246	2.799	0.002					
254	2.839	0.019					
265	3.027	0.096					
275	3.040	0.008					
285	2.927	-0.277					
294	3.371	-0.115					
306	3.376	-0.142					
315	3.492	-0.209					
325	3.463	-0.263					
335	3.409	-0.233					
345	3.433	-0.202					
355	3.434	-0.249					
366	3.349	-0.257					
375	3.454	-0.258					
385	3.318	-0.201					
395	3.418	-0.211					
405	3.274	-0.245					
415	3.326	-0.241					
425	3.374	-0.248					
435	3.326	-0.310					
445	3.120	-0.134					
455	3.371	-0.182					
465	3.202	-0.171					
476	3.276	-0.143					
490	3.276	-0.125					

The sill depth at the Passage is not accurately recorded. In order to infer it from the hydrographic data, we compared vertical profiles of temperature, salinity and dissolved oxygen from inside and outside the silled basin (Fig. 2). The two stations selected for this purpose (Fig. 1) were occupied by U.S.N.S. *Serrano* during December, 1961. At depths < 1.2 km, the distributions of all the tracers in the Andaman Basin are very similar to those observed in the nearby Bay of Bengal. However, a clear divergence of the profiles occurs at greater depths. As is the case with other similar basins, the hydrographical characteristics of deep waters within the Andaman Sea are remarkably uniform. The observed temperature ($\sim 5^{\circ}\text{C}$) of the deep waters at Station SR 43 corresponds to a depth of ~ 1.3 km at Station SR 25 (Fig. 2). This should be the effective sill depth at the Great Passage. This depth falls within the range where the conservative properties, such as potential temperature, salinity, nitrate tracer (NO) and phosphate tracer

(PO), exhibit linear correlations [15], indicating that the water spilling over the Great Passage sill may be regarded as a mixture of two water types: the overlying NIHSIW and the underlying bottom water, which probably has a significant North Atlantic Deep Water (NADW) component [21].

The oxygen content of the Andaman Sea deep waters is strikingly uniform and indistinguishable from the concentration estimated outside the basin at the sill depth (Fig. 2) (the apparent oxygen utilization (AOU) at depths > 1.3 km at Station SR 43 ($239 \pm 2 \mu\text{M}$) is the same as the AOU estimated at 1.3 km at Station SR 25). This indicates that the renewal of deep waters in the Andaman Basin is rapid, with minimal chemical changes occurring during the residence of water within the basin itself.

The North Indian Ocean is known to have experienced pronounced changes in hydrography, near-surface circulation and biological productivity, associated with the glacial–interglacial cycles. The available evidence suggests that, as com-

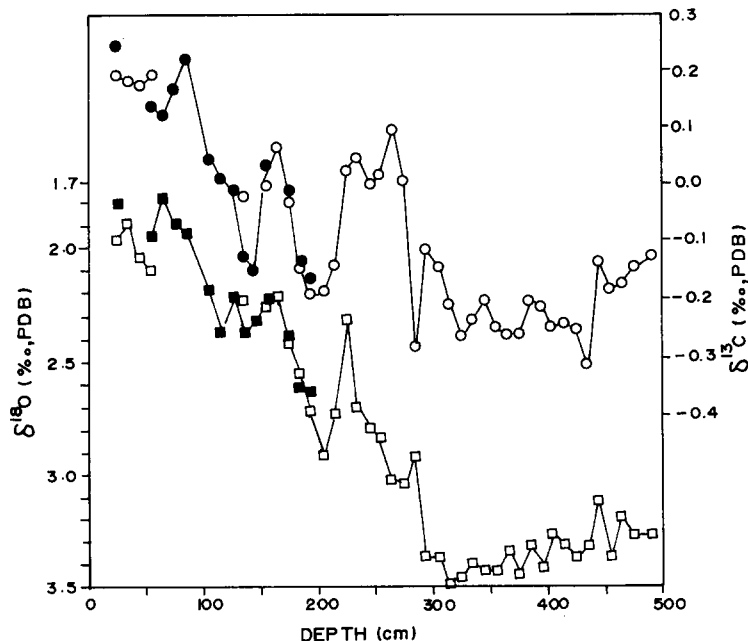


Fig. 3. Downcore records of $\delta^{18}\text{O}$ (squares, lower curve) and $\delta^{13}\text{C}$ (circles, upper curve) in benthic foraminifera (*Planulina wuellerstorfi* (solid symbols) and *Cibicidoides pachyderma* (open symbols)) in RC12-344. The $\delta^{13}\text{C}$ values of *P. wuellerstorfi* have been brought to the *C. pachyderma* scale by subtracting 0.06‰ , the mean offset between the two species (see text).

pared to today, the productivity was lower during the LGM and higher around 9 kyr BP [22]. However, in view of the observed negligible chemical modification of waters within the deep Andaman Basin at present, it would be reasonable to assume that any changes in the ventilation time or surface productivity in the recent geological past would not have led to significant differences in the composition of the Andaman Sea deep waters from that of the inflowing waters. If so, then the Andaman Sea is perhaps the best site for investigating the variations in intermediate water composition in the North Indian Ocean.

3. Methods

Measurements of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ were made at the Lamont-Doherty Earth Observatory (L-DEO) following standard procedures: hand-picked specimens (200–400 μg) from the coarse fraction (> 150 μm) were cleaned ultrasonically in distilled water, roasted under vacuum at 375°C and analyzed using a Finnigan-MAT 251 mass spectrom-

eter equipped with a Carousel-48 automatic carbonate preparation system. Calibration to the PDB standard was achieved through NBS-19 and NBS-20 standards. Analytical precision was better than 0.06‰ for $\delta^{18}\text{O}$ and 0.03‰ for $\delta^{13}\text{C}$.

Isotopic analyses on RC12-344 were performed using *Planulina wuellerstorfi* and *Cibicides pachyderma*. Of the two species, *C. pachyderma* exhibited greater downcore abundance. The results presented here are, therefore, based mostly on the analysis of this species. Isotopic data for *P. wuellerstorfi* were generated only from the upper portion of the core where *C. pachyderma* was less abundant. Results of 7 pairs of analysis from some levels where the two species co-existed (Table 1) suggest little differences in the isotopic composition ($0.03 \pm 0.11\text{‰}$ for $\delta^{18}\text{O}$ and $0.06 \pm 0.06\text{‰}$ for $\delta^{13}\text{C}$, with *P. wuellerstorfi* giving lighter and heavier values, respectively). No single species was present in sufficient quantities in the RC12-339 samples. Therefore, the isotope record for this core was obtained using mixed specimens of *P. wuellerstorfi* and *C. kullenbergi*. Previous investigators [e.g., 23] found that

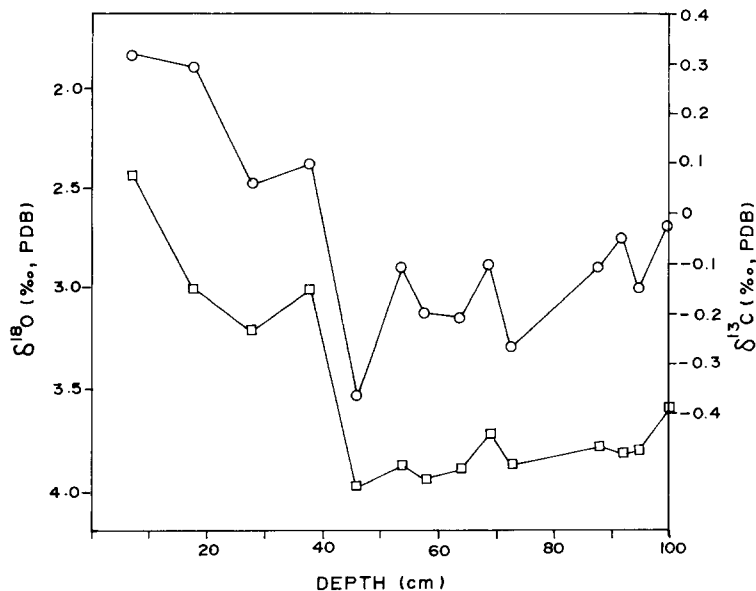


Fig. 4. Downcore records of $\delta^{18}\text{O}$ (squares, lower curve) and $\delta^{13}\text{C}$ (circles, upper curve) in benthic foraminifera (*P. wuellerstorfi* and *C. kullenbergi*) in RC12-339.

these species also do not exhibit a significant offset in either $\delta^{18}\text{O}$ or $\delta^{13}\text{C}$.

4. Results and discussion

The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values measured in the two cores are listed in Table 1. Graphical presentation of these results is made in Figs. 3 and 4 for RC12-344 and RC12-339, respectively. Unfortunately, the relatively low sedimentation rates in RC12-339 do not allow a detailed reconstruction of the changes in deep waters during the last glacial–interglacial transition. In contrast, RC12-344 provides a high-resolution record of the variations in intermediate water composition during and since the LGM. Therefore our attention is focused here mostly on the Andaman Sea core. Results of the analyses of oxygen isotopes in the planktonic foraminifer *Globigerinoides ruber*, carbon isotopes in organic matter, and total organic carbon in sediments from this core have been published previously [24]. Although this core comes from the continental slope, in a region which is greatly influenced by runoff from the rivers Irrawaddy and Salween, the rather smooth paleoenvironmental records obtained earlier indicate that the core did not suffer from slumping.

4.1. Comparison of Recent foraminiferal $\delta^{13}\text{C}$ with GEOSECS water-column data

The high sedimentation rate in RC12-344 affords reliable estimation of the isotopic composition of modern benthic foraminifera even in the absence of the real core-top samples, due to subsampling by previous workers. The $\delta^{13}\text{C}$ values of *C. pachyderma* picked from the upper Holocene section of this core (25–55 cm) are very close to each other ($0.19 \pm 0.01\text{‰}$, $n = 4$). These are also comparable with the data for *P. wuellerstorfi* within the sub-bottom depth interval 25–85 cm ($0.24 \pm 0.06\text{‰}$, $n = 5$), with the two species showing about the same mean offset as deduced from the 7 paired analyses. The $\delta^{13}\text{C}$ of total dissolved carbon dioxide (TCO_2) at the sill depth and $\sim 10^\circ\text{N}$ latitude in the Bay of Bengal is

$\sim 0\text{‰}$ vs PDB [25]. Thus, the $\delta^{13}\text{C}$ of Recent *P. wuellerstorfi* appears to be significantly greater than that of TCO_2 in the ambient water. Data from RC 12-339 also lead to a similar conclusion: the $\delta^{13}\text{C}$ at 7 cm (0.32‰) is heavier than the corresponding water column value by $\sim 0.2\text{‰}$. It may be noted that our value at 7 cm in this core is similar to that measured at the core-top (0.23‰) by Duplessy et al. [26].

The above results appear to go against the widespread perception that the carbon isotopic composition of *P. wuellerstorfi* is very nearly the same as that of TCO_2 in the bottom water in which it grows [23,26–29]. Some divergence between the published water column and core-top *P. wuellerstorfi* $\delta^{13}\text{C}$ data is not very uncommon, however, as manifested by the significant variability ($\delta = \pm 0.21\text{‰}$) in the observed departure of foraminiferal $\delta^{13}\text{C}$ from the computed equilibrium values [23]. A re-examination of the available data (mostly based on the analysis of *P. wuellerstorfi*) reveals that the core-top $\delta^{13}\text{C}$ is consistently higher than the corresponding water column values in most cores raised from the northern Indian Ocean, except the zones of intense upwelling [26]. In MD76-125 ($8^\circ 35'\text{N}$, $75^\circ 20'\text{E}$, water depth 1878 m), for example, the *Cibicidoides* $\delta^{13}\text{C}$ values average around 0.23‰ for the upper Holocene [29]. This, and several other intermediate-depth (1426–1954 m) cores [26] collected from the eastern Arabian Sea between latitudes $8^\circ 35'$ and 15°N give an average core-top $\delta^{13}\text{C}$ value of $0.18 \pm 0.09\text{‰}$. This shows an enrichment of $\sim 0.2\text{‰}$ with reference to the GEOSECS water column data [25], in conformity with our results. A similar offset of the core-top $\delta^{13}\text{C}$ data from the Exmouth Plateau in the Southeast Indian Ocean from the nearest GEOSECS water column data has also been observed [Dr. D.C. McCorkle, pers. commun., 1993]. Thus, there is enough information that points to a systematic disagreement between the carbon isotopic compositions of the Recent *P. wuellerstorfi* and the ambient deep waters in the Indian Ocean. However, it is not clear to what extent this divergence owes to the uncertainties associated with the water column ^{13}C data. Recently, Herguera et al. [30] found a large (0.45‰) offset

in the core-top $\delta^{13}\text{C}$ of *P. wuellerstorfi* on the Ontong Java Plateau from the water-column $\delta^{13}\text{C}$ estimated during the GEOSECS at stations located about 20° to the east. However, subsequent measurements of $\delta^{13}\text{C}$ in bottom water on the Ontong Java Plateau itself have led to values that are about 0.4‰ more positive than the GEOSECS data, indicating that *P. wuellerstorfi* faithfully records bottom water $\delta^{13}\text{C}$ at this site [31]. Might the consistently heavier $\delta^{13}\text{C}$ in the Holocene *P. wuellerstorfi* relative to the published bottom water values similarly arise in the Indian Ocean as well, even though the GEOSECS Indian Ocean $\delta^{13}\text{C}$ data set is generally considered to be cleaner than those from the other oceans? In any event, the use of foraminiferal $\delta^{13}\text{C}$ data for reconstructing paleochemical changes involves the assumption that this 'vital effect' remained constant through the time interval of interest to us.

4.2. Glacial to interglacial shift in benthic $\delta^{13}\text{C}$

Radiocarbon dates were not obtained on RC12-344, due to its low carbonate content, especially during the glacial time, and the identification of LGM has been made based on the oxygen isotope curve. An appreciable decrease in $\delta^{18}\text{O}$ begins around 300 cm in this core. Comparisons with two recent accelerator mass spectrometer (AMS) radiocarbon-dated isotope records from the Southern Ocean [32] and the Northwest Pacific Ocean [33] enable the assignment of an approximate age of 13.5 kyr BP to this level. This would imply an average sedimentation rate of ~ 22 cm/kyr since the beginning of the deglaciation. Extrapolating this rate to the glacial period, one would expect the LGM to occur at about 400 cm. However, the core location is such that the sedimentation rate might have fluctuated greatly in response to both the sea level rise as well as variations in riverine inputs. The glacial sedimentation rates are expected to be lower due to the decreased riverine inputs (resulting from a weaker summer monsoon) and lower productivity (as indicated by the much lower carbonate content). Therefore, the LGM level could actually be shallower than indicated above. Based on the plank-

tonic $\delta^{18}\text{O}$ record, Fontugne and Duplessy [24], assumed the LGM to occur at 290 cm in this core. However, their analysis was restricted to a depth of 340 cm. The benthic record clearly shows that the 290 cm level corresponds to the early deglaciation. However, it is important to note that the above uncertainty in the exact position of the LGM does not affect the major conclusions reached by us, as both the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ remain quite constant within the sub-bottom depth interval of 325–425 cm. The isotopic data within this depth range have been averaged to represent the conditions prevailing during the peak glaciation.

The most significant result obtained by us is that the magnitude of the glacial ($-0.24 \pm 0.02\text{‰}$, $n = 11$) to Holocene ($0.19 \pm 0.01\text{‰}$, $n = 4$) shift in benthic $\delta^{13}\text{C}$ (0.43‰) in the Andaman Sea is virtually indistinguishable from the oceanic mean (0.46‰) [29]. However, since the latter value itself is based on the analysis of deep ocean cores, which generally show a more positive glacial to Holocene shift as compared to the intermediate-depth cores [e.g., 34–36, this work], the inclusion of upper ocean data would produce a global mean somewhat less than that reported by Curry et al. [29]. In other words, the upper Indian Ocean would have actually experienced a $\delta^{13}\text{C}$ shift of *higher* amplitude relative to the global mean. This is at variance with the results of Kallel et al. [1], who combined their results of isotopic measurements on the glacial foraminifera with the GEOSECS water column data [25] and deduced little changes in $\delta^{13}\text{C}$ in the upper Indian Ocean. These results suggested that the dissolved carbon in intermediate waters of the NEIO was isotopically much heavier during the LGM as compared to today after accounting for the global shift in $\delta^{13}\text{C}$. A comparison of the two data sets reveals that the glacial $\delta^{13}\text{C}$ values in RC 12-344 are not very different from those measured by Kallel et al. [1] (lowest -0.11‰) in the nearby Andaman Sea core MD77-169 ($10^\circ 13' \text{N}$, $95^\circ 03' \text{E}$, water depth 1600 m) allowing for the slight offset (0.06‰) between *P. wuellerstorfi* and *C. pachyderma*. Therefore, the different estimates of the glacial to Holocene shift in $\delta^{13}\text{C}$ are mostly due to our observation that the core-top $\delta^{13}\text{C}$ in '*Cibicidoides*' is not the same as

$\delta^{13}\text{C}$ of TCO_2 in the bottom waters as assumed by Kallel et al. [1] **.

The results of Kallel et al. [1] have often been cited as evidence for a more vigorous intermediate circulation globally during the LGM [5,30,35–38]. However, this may be questioned on the following grounds. A better ventilation of the intermediate-depth Atlantic during the LGM was probably caused by the formation of intermediate waters at the expense of deep waters in the North Atlantic [7,34,39]. Intermediate waters could also have formed in the North Pacific during glacial times [29,34,36]. In the North Indian Ocean, on the other hand, it is very unlikely that the local production of intermediate waters was any more intense than it is today. This is because the outflow from the Red Sea was greatly reduced [40], while that from the Persian Gulf must have been non-existent, due to a much lower sea level. Moreover, the area of the continental shelves (where the subsurface waters might be formed) was greatly reduced, and the surface temperatures were not very different from those observed now [41]. It has been postulated that an intensified flow of intermediate waters from the Pacific through the Indonesian Archipelago could bring about the improved renewal [1]. However, this flow should actually have decreased during the LGM, in response to the lower sea level and to the dampening of the southwest monsoon [42]. Therefore, a more energetic renewal of intermediate waters in the North Indian Ocean could have occurred only through enhanced advection from the southern hemisphere. However, the evidence available also precludes this possibility. Recent measurements of the opal accumulation rates [43], $^{231}\text{Pa}/^{230}\text{Th}$ ratio [44] and $\delta^{15}\text{N}$ [37] in the Southern Ocean sediments suggest that during the LGM the zone of maximum productivity migrated north of the present position of the

Antarctic Polar Front. The higher productivity probably led to lower nutrient concentrations within the subantarctic zone [37]. Hence, a stepped-up supply of intermediate waters from the southern hemisphere should have amplified the nutrient depletion at mid-depths in the North Indian Ocean during the LGM. This is irreconcilable with the ^{13}C evidence.

The RC12-339 $\delta^{13}\text{C}$ record (Fig. 4) shows the LGM ($\delta^{18}\text{O}$ maximum, 46 cm) to Holocene (7 cm) shift in $\delta^{13}\text{C}$ to be 0.68‰. This is significantly larger than the mean oceanic shift [29]. Combining the RC12-344 and RC12-339 records (considering the slight offset between the $\delta^{13}\text{C}$ values of *P. wuellerstorfi* and *C. pachyderma*), we estimate that the intermediate to deep $\delta^{13}\text{C}$ gradient increased from $-0.18‰$ (LGM) to $0.07‰$ (Holocene), a shift of $+0.25‰$. This is similar to the shift observed in the Pacific [35,36], but is only about one quarter of the corresponding change in the Atlantic [7–10].

In contrast to our results, Kallel et al. [1] measured extremely light $\delta^{13}\text{C}$ values (lowest $-1.15‰$) in the glacial sediments of the deep Bay of Bengal. Referenced against the estimated core-top values of $0.2\text{--}0.3‰$ these yield a glacial–Holocene change in the benthics of $\sim 1.4‰$, implying that the bottom waters in the NEIO could have turned anoxic during the LGM. A pronounced enrichment of ^{238}U observed recently in the glacial section of a core from the Arabian Sea has also been interpreted as suggesting anoxia in bottom waters [38]. Our results are incompatible with these observations, although we acknowledge that the Bay of Bengal core examined by us was raised from a somewhat greater depth. While it may be argued that the relatively low sedimentation rate in RC12-339 might have led to an underestimation of the magnitude of the glacial–Holocene change in $\delta^{13}\text{C}$, we point out that benthic $\delta^{13}\text{C}$ values as low as $-1.08‰$ at the LGM have been observed in MD77–182 ($16^\circ 10' \text{N}$, 91°E , water depth 2455 m), which has comparable sedimentation rates [1]. However, it may also be noted that the $\delta^{18}\text{O}$ at its maximum in RC12-339 ($\sim 4.0‰$) is substantially lower than the glacial values ($\sim 4.3‰$) measured by Kallel et al. [1] in several cores

** It has been brought to our notice, while the present paper was in review, that the analysis of *P. wuellerstorfi* from the Holocene portions of the intermediate-depth cores examined by Kallel et al. [1] has now been made. The results also show an offset of 0.2–0.3‰ between the foraminiferal and water column $\delta^{13}\text{C}$ values (S.M. Ahmad and L. Labeyrie, manuscript in preparation).

deeper than 2 km. A possible offset, due to unknown reasons, between the $\delta^{18}\text{O}$ values of the two data sets might account for most of this difference. This is because the glacial $\delta^{18}\text{O}$ values in RC12-344 are also lighter by $\sim 0.2\text{‰}$ compared to those from MD77-169 in spite of the lower sedimentation rate at the latter site [1]. Moreover, the observed large amplitude of the glacial to Holocene shift in $\delta^{18}\text{O}$ in RC12-339 (see below) makes it unlikely that we have failed to sample the true glacial level in this core. While another core from the Bay of Bengal with a better sedimentation rate was not available to us for resolving the above discrepancy, we have also analyzed *P. wuellerstorfi* from the core-top and LGM (82 cm) levels in V29-29 ($5^{\circ}07'\text{N}$, $77^{\circ}35'\text{E}$, water depth 2673 m). The LGM (-0.31‰) to Holocene (0.27‰) shift observed in this core was 0.58‰ , not very different from the shift seen in RC12-339.

The similarly modest change in the vertical chemical structure in the Pacific and the Indian Oceans (the combined volume of which far exceeds that of the Atlantic) implies that the mechanism proposed by Boyle [4,5] would probably be inadequate to account for the bulk of the changes in atmospheric CO_2 content associated with the glacial–interglacial cycles. Boyle [3] arrived at the same conclusion based on his Cd data from the Pacific Ocean.

4.3. Glacial to interglacial shift in benthic $\delta^{18}\text{O}$

The glacial to interglacial $\delta^{18}\text{O}$ shifts in benthic foraminifera substantially exceed the ice volume effect (1.2‰) [45] (by 0.26‰ in RC12-344 and by 0.34‰ in RC12-339), requiring a cooling by 1.1°C and 1.5°C , respectively, assuming no additional salinity change. The estimated deep water cooling is in accordance with, while the intermediate-depth change is at variance with, the previous results [1]. It has been postulated [2] that the higher amplitude of the glacial–interglacial $\delta^{18}\text{O}$ shift in deep sea cores in the North Indian Ocean as compared to the equatorial Indian Ocean might reflect a local source of glacial deep waters in the former region. However, as discussed earlier, it is difficult to reconcile the

local source with a lower sea level and negligible change in sea surface temperatures. It seems more likely that the anomaly is an artifact of low sedimentation rates in the equatorial Indian Ocean. It may be added here that the inferred cooling of the glacial deep waters in the NEIO is of the same magnitude as that deduced for the Pacific deep waters [36].

4.4. The mid-depth benthic record of deglaciation

The high sedimentation rates in RC12-344 enable a reconstruction of the changes at intermediate depths over the last deglaciation (Fig. 3). In the absence of radiocarbon ages, the $\delta^{18}\text{O}$ curve affords the fixation of a reasonable time frame for the interpretation of two major features observed in the $\delta^{13}\text{C}$ record.

First, during the early deglaciation, the $\delta^{13}\text{C}$ curve is characterized by rapid increases in $\delta^{13}\text{C}$, interrupted by a brief but abrupt decrease. This is qualitatively very similar to the variations in $\delta^{13}\text{C}$ recorded in the Southern Ocean, believed to reflect the fluctuations in the production of NADW [32]. The initial increase and subsequent decrease in $\delta^{13}\text{C}$ during the early stages of deglaciation are considered to represent a ‘false start’ in the NADW production before the vigorous production of NADW resumed [32]. That the changes in the NADW production appear to be recorded by the sediments in the Andaman Sea is not surprising. This is because, as stated earlier, the bottom water (with a significant NADW component) is one of the end-members regulating the composition of the waters that fill the marginal basin below the sill depth.

Second, the $\delta^{13}\text{C}$ record shows a pronounced negative excursion to near-glacial values during the mid-deglaciation, with the $\delta^{13}\text{C}$ minimum coinciding roughly with the $\delta^{18}\text{O}$ maximum corresponding to the Younger Dryas. This must be related to a regional event, since a pronounced decrease in $\delta^{13}\text{C}$ associated with the Younger Dryas has not been recorded in the Southern Ocean [32], presumably because the NADW continued to reach the Antarctic at that time, albeit at shallower depths [46]. The negative $\delta^{13}\text{C}$ excursion could arise from an intensification of the

southwest monsoon at about the same time [47], which might have led to an enhanced nutrient regeneration at intermediate depths. Support for this explanation comes from a similar spike in $\delta^{13}\text{C}$ in a high-resolution record [29] from the core MD76-135 located in the northwestern Indian Ocean ($14^{\circ}27'$, $50^{\circ}31'E$, water depth 1895 m). As this site is very close to the upwelling centre off southeast Arabia, the spike (which also coincides with the Younger Dryas $\delta^{18}\text{O}$ maximum) is much more prominent there ($\delta^{13}\text{C} = -0.85\text{‰}$). Thus, the $\delta^{13}\text{C}$ spike observed in RC 12-344 does not appear to be a local signal reflecting a more intense oxidation of organic matter within the Andaman Sea itself. Instead, it may be a regional feature produced by an intensification of the mid-depth oxygen minimum in response to the generally elevated primary production in the North Indian Ocean. The subsequent, less developed, features might also be related to monsoonal changes; however, their significance cannot be evaluated due to the lack of a reasonable age control in that portion of the core.

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