

$\delta^{13}\text{C}$ depleted oceans before the Termination 2: More nutrient-rich deep-water formation or light-carbon transfer?

Virupaxa K. Banakar

Geological Oceanography Division, National Institute of Oceanography, Dona Paula, Goa-403 004, India
[E-mail: banakar@darya.nio.org]

Received 15 February 2005, revised 11 July 2005

Carbon-isotopes ($\delta^{13}\text{C}$) composition of benthic foraminifera has been extensively used to understand the link between deep-water circulation and climate. Equatorial Indian Ocean $\delta^{13}\text{C}$ records of planktic- and benthic-foraminifera together show an unexplained shift in the long-term mean oceanic- $\delta^{13}\text{C}$ around the penultimate glacial termination (T2: 132 ka). The time-series planktic- and benthic- species $\delta^{13}\text{C}$ records exhibit two distinct mean- $\delta^{13}\text{C}$ levels. The low mean- $\delta^{13}\text{C}$ characterises the pre-T2 period (250 ka – 132 ka), while the post-T2 (~95 ka – Present) period records high mean- $\delta^{13}\text{C}$, generating a one-time shift of ~0.4 % within the last ~250 kyr time-period. This shift is a result of consistently higher- $\delta^{13}\text{C}$ in post-T2 glacial (and interglacial) periods as compared to the pre-T2 glacial (and interglacial) periods, and begins around the T2 (~132 ka), lasts until ~95 ka, and sustained through the T1. The normally observed glacial-interglacial $\delta^{13}\text{C}$ variations of ~0.3 % occur as secondary fluctuations around the long-term primary mean-levels in the Indian Ocean, as well as in other oceans. The T2- $\delta^{13}\text{C}$ shift appears to be an inherent feature of the world oceans although with certain timing offsets. Therefore, it should represent a fundamental change in deep-ocean circulation (nutrient) dynamics. But, the leading hypotheses of circulation-driven oceanic distribution of $\delta^{13}\text{C}$ fail to explain the observed mean- $\delta^{13}\text{C}$ shift. Therefore it is proposed that, in addition to changes in deep-water circulation, the oceans before T2 were characterised by significantly lower- $\delta^{13}\text{C}$ than after. Such low- $\delta^{13}\text{C}$ mean-ocean during the pre-T2 period might have been the result of significantly increased transfer of terrestrial light-carbon to the ocean reservoir due to changes in global wind patterns.

[Key Words: Equatorial Indian Ocean, foraminifera- $\delta^{13}\text{C}$, glacial termination-2, deep-water circulation, role of atmosphere]

Introduction

Deep-water circulation and past climate

The global ocean deep-water circulation is believed to be an important player in Pleistocene glacial-interglacial climate cycles mostly based on foraminiferal $\delta^{13}\text{C}$ distribution and other sedimentary proxies¹⁻¹⁷. In the last two decades of palaeoclimate research, the North Atlantic Deep Water (NADW: nutrient-depleted, $\delta^{13}\text{C}$ -enriched) driven thermohaline circulation has emerged as the major inter-oceanic communicator and climate modulator²⁻⁵. Several previous studies have demonstrated that the NADW formation was greatly weakened during glacial periods^{1,6-9}. However, conflicting observations were reported based on geochemical and radiochemical proxy records, which rather indicated unchanged NADW circulation during the last glacial maximum (LGM)^{10,11}. Broecker *et al.*¹² proposed two modes of deep-water circulation driving the global thermohaline circulation, in which, the NADW in its strong mode dominated during interglacial periods,

whereas, during the glacial periods was replaced by the North Pacific Deep Water (NPDW) dominant mode. Nutrient depleted deep-water formation in both Atlantic and Pacific oceans was suggested to explain the lower deep-water $\delta^{13}\text{C}$ in the Southern Ocean than in the Pacific during the LGM⁶. The chemical stratification of the intermediate depth water in the Southern Ocean during glacial periods and its co-variation with atmospheric $p\text{CO}_2$ was interpreted as a significant modulator of the Pleistocene climate cycle^{18,19}.

Mixing of the presently known two deep waters NADW and Antarctic Bottom Water (AABW) in the circumpolar region generates characteristic deepwater ΣCO_2 - $\delta^{13}\text{C}$ ²⁰ and is preserved in the foraminiferal calcite¹. Therefore, the $\delta^{13}\text{C}$ of benthic foraminiferal calcite has been universally accepted tool to monitor changes in deep-water circulation and nutrient distribution. The deep-water circulation scheme represented by the $\delta^{13}\text{C}$ distribution, and the past climate has been seen as strongly coupled entities in

spite of several conflicting views. The oceanic $\delta^{13}\text{C}$ chemistry may deviate from standard circulation dependent chemistry when the atmospheric system changes significantly because the atmosphere transfers low- $\delta^{13}\text{C}$ characteristic terrestrial carbon to the ocean. Therefore, major reorganization in atmospheric circulation may produce significant changes in oceanic nutrient ($\delta^{13}\text{C}$) dynamics, which is not looked into. Based on the foraminiferal- $\delta^{13}\text{C}$ inventory, this paper demonstrates the possible role of atmosphere in driving the oceanic $\delta^{13}\text{C}$ changes.

Indian Ocean deep-water

As the Indian Ocean does not have its own deep-water formation zone, the Pacific and Atlantic waters get exported to Indian Ocean at depth below 2000 m from the Southern Ocean²¹⁻²⁴. The circumpolar deep-water (CDW) formed in the Southern Ocean enters the Indian Ocean abyss from both western and eastern sides of the basin unlike the single path inflow into the Pacific and Atlantic oceans. Modern CDW has intermediate $\delta^{13}\text{C}$ (~0.4 ‰) of the NADW (~1 ‰) and the Pacific recirculated water (~0 ‰)^{7,20}. The nearly homogenized modern CDW- $\delta^{13}\text{C}$ characteristics are well-reflected in the nearly uniform average Holocene benthic- $\delta^{13}\text{C}$ (~0.5 ‰) in a Cape Basin core ODP1090, which is located in the frontal zone of CDW-NADW¹⁸, as well as in the southwest Pacific core ODP1123, which is located below the deep western boundary currents feeding the Pacific¹⁷. A strong hydro-chemical front at ~10° S latitude that vertically extends down to a depth of ~1 km, separates low-oxygen and high-nutrient waters of the northern monsoon gyre from the high-oxygen and low-nutrient

waters of the southern subtropical gyre, thus preventing penetration of the southern source intermediate water into the northern Indian Ocean²³. The depth distribution of ΣCO_2 - $\delta^{13}\text{C}$ in the equatorial regions of western, central, and eastern Indian Ocean is nearly similar (see ref. 20), indicating horizontally homogenized carbon-isotope chemistry in the Equatorial Indian Ocean (EIO). Such hydrographic set-up is well suited to monitor the changes in $\delta^{13}\text{C}$ character of the deepwater source, i.e., CDW.

This paper presents new evidence for long-sustained presence of low- $\delta^{13}\text{C}$ surface and bottom water in the EIO prior to the T2 period (250 ka to 132 ka) than the period after (~95 ka to Present). As the present record dates back only to 250 ka, it is not known how far in the past this low- $\delta^{13}\text{C}$ event really continues. This study compares the EIO time-series planktic- and benthic- $\delta^{13}\text{C}$ variation with other oceans' records to demonstrate that the global ocean has witnessed a major shift from low- $\delta^{13}\text{C}$ reservoir to a high- $\delta^{13}\text{C}$ reservoir around T2. In addition the paper explores the possible meaning of such change.

Materials and Methods

Intact and clean tests of the upper mixed layer-, thermocline-, and bottom-dwelling foraminifera were picked from a sediment core SK129/CR2 (hereafter SK129), collected from the EIO at 3°N and 76°E (Fig. 1) from a water depth of 3800 m, to reconstruct the time-series water-column $\delta^{13}\text{C}$. Microscopically noticeable dissolution symptoms such as damaged tests and *Globorotalia menardii* keels are not evident suggesting sediment deposition above the carbonate lysocline throughout the time span covered by the

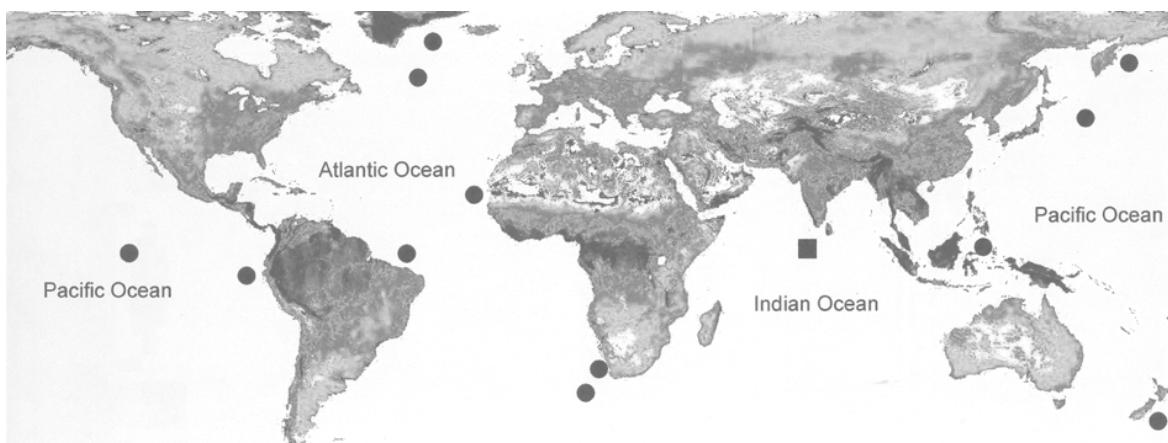


Fig. 1 — Location of SK129/CR2 sediment core used in this study [■]. The approximate locations of published $\delta^{13}\text{C}$ records used for comparing SK129/CR2 records are shown with [●]filled circles (see text and Table 1 for details). [The background map is downloaded from NASA-JPL Web and edited to requirement]

core. Around 60-80 individuals of *Globigerinoides sacculifer* (without terminal sac) and 15-30 individuals of *Uvigerina peregrina* (mixed-layer, thermocline and bottom dwelling species respectively²⁵, from 250-350 μm fractions of the same sub-sections of the core SK129 were crushed open under microscope in presence of ample reverse osmosis treated water and agitated with a fine soft brush. The visible clay lumps, silicate grains, and Mn-oxide particulates were physically removed using a fine brush, and the fine-detritus was sucked using tissue-paper thread, before homogenizing the crushed aliquot. One portion of this aliquot was used for stable isotope measurement and other was preserved for Mg/Ca-palaeothermometric study. The sonicated and further cleaned aliquot was treated with dilute H_2O_2 to remove organic contaminants and measured on Micromass-Sera Mass Spectrometer. The long-term measurements of the in-house standard calibrated with PDB yielded a precision of 0.1 ‰ for $\delta^{18}\text{O}$ and 0.06 ‰ for $\delta^{13}\text{C}$. Strict-pairing could not be maintained for *U. peregrina* due to its highly variable abundance (0-140 tests/fraction), which forced to combine some intervals for obtaining adequate material. As the abundance of *Cibicidoides spp.*, which are epibenthic (epifauna) and considered to be the most reliable for recording deep-water $\delta^{13}\text{C}$ characters was extremely low throughout the core, the *U. peregrina* were analyzed to obtain complete isotope records. Although this species is endobenthic (infauna), has

been successfully utilized by several workers in the past after applying correction of +0.9 ‰ for isotopic-equilibrium offset⁹. The SK129 down-core *N. dutertrei*- $\delta^{18}\text{O}$ was tuned to the SPECMAP²⁶ to derive time-scale. The ~72 ka Youngest Toba Tuff²⁷ and ~120 ka last appearance of *G. ruber* (pink variety)²⁸ have served as useful independent control-points for the oxygen-isotope chronology (Fig. 2). The 5.5 m long SK129 core contains well-defined marine isotope stages and covers the last 250 ka. The last three glacial terminations occur at 242 ka (T3), 132 ka (T2), and 12 ka (T1) in the oxygen-isotope record (Fig. 2).

Results and Discussion

Today the initial- $\delta^{13}\text{C}$ of CDW (~0.5 ‰) progressively depletes to ~0.2 ‰ at the SK129 site due to gradual enrichment of ^{12}C as a result of continuous oxidation of organic matter during its transit to the seafloor²⁰. The Holocene-LGM contrast in $\delta^{13}\text{C}$ of both planktic and benthic species varies between 0.2 and 0.4 ‰ (Fig. 3). As the entire Holocene is represented by only one benthic datapoint generated from combined intervals due to rarity of *Uvigerina* in this interval, the generally observed high- $\delta^{13}\text{C}$ of the late Holocene might have been diluted by early Holocene low- $\delta^{13}\text{C}$. Therefore, the Holocene-LGM $\delta^{13}\text{C}$ change in the EIO can be assumed to be closely comparable to the expected²⁹ mean-ocean change of ~0.3 ‰.

The time-series $\delta^{13}\text{C}$ records of all the three species contain a pronounced positive shift of 0.5 to 0.7 ‰ at the T2 (Fig. 3). This $\delta^{13}\text{C}$ excursion begins at around the T2 and continues until the Marine Isotope Stage (MIS) 5c. Prior to this shift, the mean- $\delta^{13}\text{C}$ of *Uvigerina* sp. is -0.3 ‰, *N. dutertrei* is 0.7 ‰, and the *G. sacculifer* is 1.1 ‰, while after the shift they are 0.1 ‰, 1.1 ‰, and 1.4 ‰ respectively, thus generating distinct pre-T2 low and post-T2 high mean- $\delta^{13}\text{C}$ levels separated by ~0.4 ‰. Further, the low- and high-mean $\delta^{13}\text{C}$ levels contain respective glacial-interglacial $\delta^{13}\text{C}$ changes of ~0.3 ‰ as secondary fluctuation around the mean-levels in the planktic records (Fig. 3).

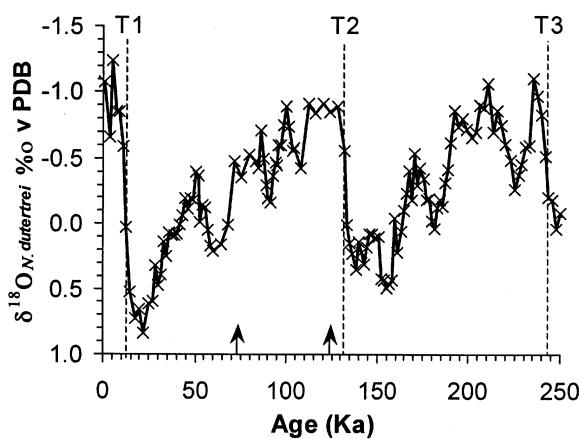


Fig. 2 — The $\delta^{18}\text{O}_{N_{dutertrei}}$ based chronology derived for SK129/CR2. The occurrence of Youngest Toba Ash bed²⁷ and the last appearance datum of the *Globigerinoides. ruber* (pink variety)²⁸ are shown as upward arrowheads on the age axis. Vertical broken lines are the glacial terminations (T) numbered accordingly.

Past-ocean $\delta^{13}\text{C}$ changes

A significantly larger $\delta^{13}\text{C}$ shift at the T2 than at the T1 was first observed in all major ocean sedimentary records by Duplessy & Shackleton¹. This shift was attributed to the diminished production of

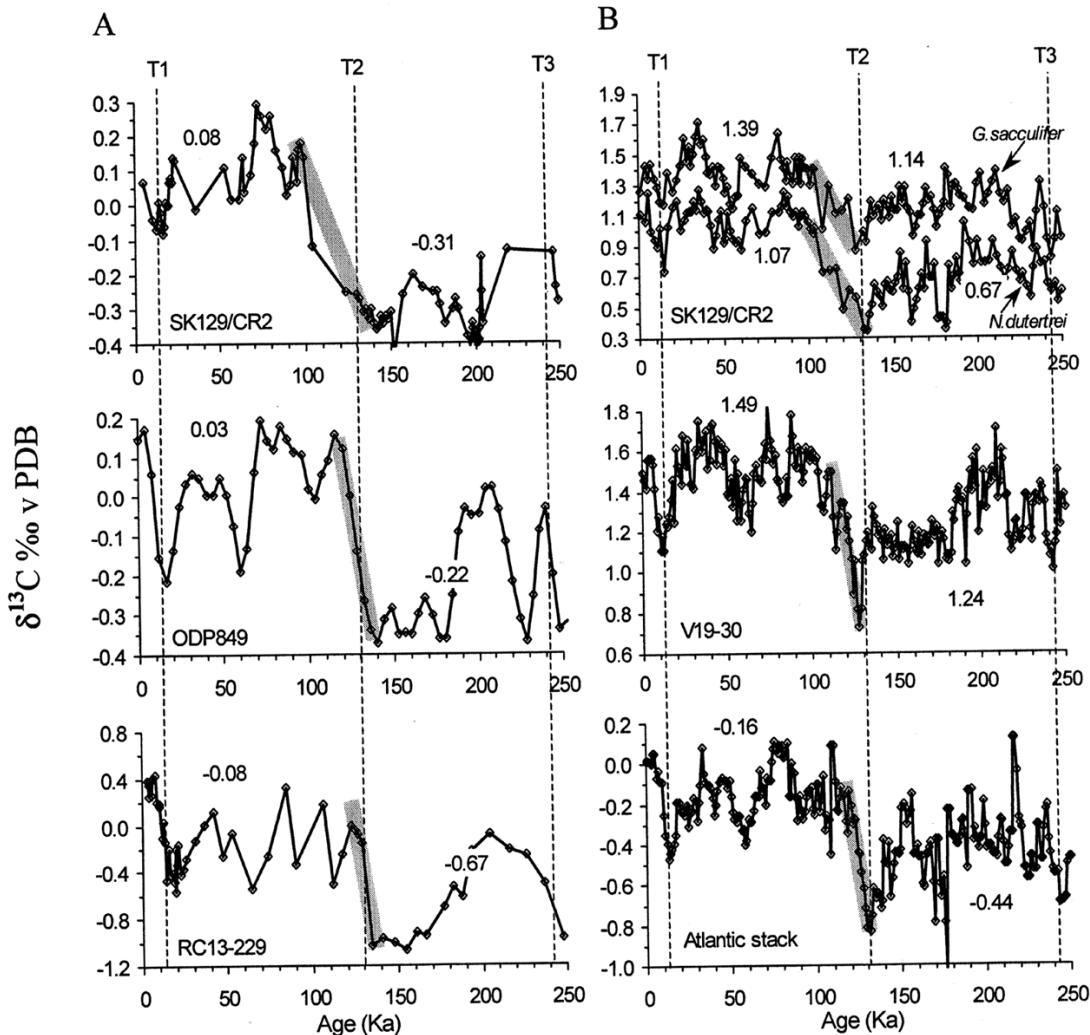


Fig. 3 — Time-series $\delta^{13}\text{C}$ records of the surface water (*Globigerinoides sacculifer*), thermocline water (*Neogloboquadrina dutertrei*) and the bottom water (*Uvigerina*) obtained for the Equatorial Indian Ocean from SK129/CR2 sediment core compared with few key locations of the world oceans (core identity is given in each panel). Left hand side panels (A) are benthic records and right hand side panels (B) are planktic records. The broken vertical lines running through all the panels are the glacial terminations (T) on the SK129/CR2 time-scale and are numbered on the top. The slanting shaded bars indicate the $\delta^{13}\text{C}$ shift around the T2. The pre-T2 and post-T2 mean- $\delta^{13}\text{C}$ are shown with respective values on either side of the T2-shift. The Atlantic planktic stack is the output obtained from several *Globigerinoides sacculifer* records after normalizing with Holocene average³⁷.

NADW during the penultimate glaciation (MIS-6), which resulted the global oceans to be filled by deep-water formed through mixing of Pacific recirculated water and AABW. Similarly the end of the $\delta^{13}\text{C}$ minima events at the terminations recorded in the tropical Pacific planktic species have been interpreted as the resumption of NADW influence on the Southern Ocean³⁰.

Even though the averaging of $\delta^{13}\text{C}$ for pre- and post-T2 periods eliminates the normally observed glacial-interglacial variations, which is not the focus of this study, it is necessary to evaluate the mean- $\delta^{13}\text{C}$

shift in light of previous leading hypotheses proposed based on glacial-interglacial deep-water circulation changes.

The agreement in timing between the planktic- $\delta^{13}\text{C}$ minima in the last glacial Subantarctic and the tropical Indo-Pacific have led Ninnemann & Charles¹⁹ to propose a transfer of low- $\delta^{13}\text{C}$ signal from the former region to the low-latitude surface water through thermocline. First of all such transmission in the Indian Ocean is probably unlikely because of the presence of a strong hydro-chemical front at $\sim 10^\circ\text{S}$ extending to a depth beyond the thermocline²³, assuming that it existed during the pre-T2 period.

Secondly, the EIO benthic record also requires similar amplitude low- $\delta^{13}\text{C}$ deep-water transfer to the EIO abyss. Further, the Subantarctic past-nutrient consumption (surface-deep $\delta^{13}\text{C}$ -gradient) is controversial. The expected changes in $\delta^{13}\text{C}$ of EIO, taking in to account the above contrasting possibilities, are: a) if the nutrient increase in Subantarctic surface has occurred during the pre-T2 due to incomplete consumption, and similar high-latitude to low-latitude transfer of nutrients ($\delta^{13}\text{C}$) also existed, then the pre-T2 carbon-pump in EIO should be stronger, as the nutrient consumption in tropical oceans is generally complete because both light and iron are abundant; b) if the nutrient decrease had taken place in the Subantarctic surface due to complete consumption, then the opposite situation could be expected. Either of such nutrient consumption schemes based on nutrient transfer should have been associated with opposite trends in the $\delta^{13}\text{C}$ distribution between the surface and bottom water of EIO, which is not evident in the present records. Another argument could be the temporal change in vital effects of the planktic species causing the mean $\delta^{13}\text{C}$ shift at T2. There are no obvious reasons why the vital-effects should change so drastically at around the T2; note the deep-water also recorded similar shift.

An extreme depletion of the LGM $\delta^{13}\text{C}$ (up to -1.1‰) in benthic records of the Arabian Sea and Bay of Bengal (BOB) led Kallel *et al.*³¹ to propose an existence of a strong chemical-front in the northern Indian Ocean resulting in differing ventilation of intermediate- and deep-water. But such anomalous depletions are not evident either in the present records (SK129), or in the central-BOB deep-water benthic records³² or in the southeastern Indian Ocean (~30°S) benthic records³³. Interestingly, the extreme $\delta^{13}\text{C}$ depletion in the northern Indian Ocean during the LGM observed by Kallel *et al.*³¹ is evident only in those sediment cores, which were from marginal seas close to the land-mass. Therefore, extremely depleted $\delta^{13}\text{C}$ during LGM observed by them might be due to increased supply of light-carbon from the surrounding land rather than existence of chemical front at intermediate depth.

A major change in the tropical Indian Ocean biological pump or productivity³⁴ causing the pre-T2 low mean- $\delta^{13}\text{C}$ feature may be simply ruled-out, because the mean- $\delta^{13}\text{C}$ shift is unidirectional in both

planktic- and benthic-records. Additionally, the past 270 ka record of the primary production in the EIO does not exhibit any major productivity shift³⁵ around the T2. The other possibilities for $\delta^{13}\text{C}$ variability at a location are changes in deep-water circulation and mean-ocean $\delta^{13}\text{C}$ inventory. The glacial-interglacial mean-ocean $\delta^{13}\text{C}$ change was associated with variations in climate dependent terrestrial carbon reservoir²⁹. A ~0.4‰ lower pre-T2 mean- $\delta^{13}\text{C}$ in the EIO water column than during the post-T2 period, or the additional ~0.4‰ lower $\delta^{13}\text{C}$ during the MIS-6 (penultimate glaciation) than that occurred during the LGM (Fig. 3), would require greatly reduced terrestrial photosynthetic carbon-reservoir during the pre-T2 glacial period. This would necessitate large-scale reduction of tropical rain forests due to freezing during MIS-6, and LGM like climate during MIS-7 (penultimate interglacial). This could be possible if the continental ice cover was significantly larger during the pre-T2 period than after. The pre-T2 period ice-volume as depicted by the past sea level or $\delta^{18}\text{O}$ records³⁶ do not contain such change.

The present $\delta^{13}\text{C}$ shift from pre-T2 low mean to the post-T2 high mean in the EIO planktic and benthic species (Fig. 3) might suggest a fundamental shift in the carbon-isotope inventory of the Indian Ocean, in turn the CDW feeding it. If this shift has occurred in the CDW then it is expected to reflect in other oceans also. In the process of addressing complex glacial-interglacial circulation and climate link, a globally evident similar mean- $\delta^{13}\text{C}$ shift that has occurred during the period covered by last two climate cycles has not been recognized.

$\delta^{13}\text{C}$ -shift around the T2, a global feature

A closer examination of the global long-term $\delta^{13}\text{C}$ records suggest that the unique $\delta^{13}\text{C}$ feature observed in sedimentary records of the EIO is also clearly reflected in planktic records of the tropical Atlantic³⁷, tropical Pacific³⁸, north Pacific³⁹, and Southern Ocean¹⁸, as well as in several deep-and intermediate-water benthic records of almost all the oceans except in the Arctic Sea⁴⁰, but with varying amplitude (Table 1, and e.g., Fig. 3). The relatively lower benthic- $\delta^{13}\text{C}$ during the MIS-6 (and MIS-7) than the LGM (and MIS-5) can clearly be seen in the benthic records of north Pacific [S2^(ref. 39), ODP849^(ref. 41)], north Atlantic [ODP552^(ref. 5)], all the tropical oceans [ODP849^(ref. 42), DSDP677^(ref. 43), ODP659^(ref. 44)], and the Southern Ocean [RC13-229^(ref. 45), ODP1090, ODP1088^(ref. 18), MD962080^(ref. 46)].

Table 1 — The $\delta^{13}\text{C}$ (‰) inventory of key locations in major oceans surrounding the penultimate termination based on records for a period of last 250 ky.

Location/Core	Water depth (m)	Post-T2 mean [@]	Pre-T2 mean [@]	Δ -mean	Excursion at T2	Source reference
<i>Tropical Atlantic</i>						
KNR110 (Planktic) ^S	3436	1.9	1.7	-0.2	1.0	(37) [#]
ODP659 (Benthic)	3081	0.6	0.4	-0.2	0.8	(44)
<i>Tropical Pacific</i>						
V19-30 (Planktic)	3091	1.7	1.4	-0.3	0.8	(38) [#]
ODP849 (Benthic)	3851	0.0	-0.2	-0.2	0.6	(42)
<i>North Pacific</i>						
S2 (Planktic) ^S	3100	1.0	0.7	-0.3	0.6	(39)
(Benthic)		-0.2	-0.4	-0.2	0.6	(39)
<i>Tropical Indian</i>						
SK129 (Planktic) ^S	3800	1.4	1.1	-0.3	0.6	(SK129 present study)
(Planktic) ^T		1.1	0.7	-0.4	0.7	
(Benthic)		0.1	-0.3	-0.4	0.5	
<i>Southern Ocean (Atlantic sector)</i>						
ODP1090 (Planktic) ^S	3702	0.0	-0.5	-0.5	1.5	(18) [#]
(Benthic)		0.0	-0.5	-0.5	1.3	(18) [#]
ODP1088 (Benthic)	2100	0.5	0.2	-0.3	0.8	(18) [#]
<i>Southern Ocean (Pacific sector)</i>						
E33-22 (Planktic) ^S	2500	0.8	0.6	-0.2	0.7	(18) [#]
ODP1123 (Benthic)	3290	0.2	-0.1	-0.3	0.9	(17)

Values were extracted from the published figures, hence may contain small errors.

@ Mean- $\delta^{13}\text{C}$ excludes the excursion period at T2.

S = Surface species,

T = Thermocline species

Numbers in parenthesis are the data sources.

A significant lowering of $\delta^{13}\text{C}$ in planktic and benthic records of the Southern Ocean during MIS-6 as compared to LGM, and lower $\delta^{13}\text{C}$ during MIS-7 than during MIS-5 [see e.g., ODP1090 and ODP1088^[ref. 18]], were not recognized as part of the global signal. The pre-T2 low mean- $\delta^{13}\text{C}$ level in many of the examined records appears to extend beyond the presently considered 250 ka. Therefore, the mean- $\delta^{13}\text{C}$ shift from a low- to a high-level, beginning at around T2 in surface-, thermocline- and bottom-water of EIO (Fig. 3), and the occurrence of similar feature in all the major oceans' planktic- and benthic-records (Table 1, Fig. 3) is remarkable. This feature at the outset suggests that either the world oceans were fed with low- $\delta^{13}\text{C}$ (nutrient enriched) deep-water or there was enhanced supply of terrestrial light-carbon prior to T2.

Assessment of possible causes for mean- $\delta^{13}\text{C}$ shift

In the modern oceans, deep-water driving the thermohaline circulation is nutrient-depleted and is

formed in the North Atlantic, while relatively nutrient-rich deep-water is formed in the Southern Ocean⁹. The largest $\delta^{13}\text{C}$ excursions (>1 ‰; Table 1) occur in tropical Atlantic and Southern Ocean records implicating them as probable candidate sites for generating the presently observed shift from low- to high- $\delta^{13}\text{C}$. However, even the complete shut-down of NADW (accounting for ~0.3 ‰ reduction in the CDW) during the pre-T2 should not have produced lower $\delta^{13}\text{C}$ mean level (~ -0.5 ‰) in the Southern Ocean¹⁸ than the mean-ocean $\delta^{13}\text{C}$ of ~0 ‰ (Table 1), had the world ocean nutrient inventory was similar to the modern. Around 0.5 ‰ lowered mean- $\delta^{13}\text{C}$ in the pre-T2 than in the post-T2, and much larger $\delta^{13}\text{C}$ gradients (>0.8 ‰) between MIS-6 and MIS-5 compared to LGM and Holocene gradients (<0.6 ‰) in the Southern Ocean (Table 1, Fig. 3) require nutrient-enriched (low- $\delta^{13}\text{C}$) deep-water to form somewhere and contribute to the CDW in addition to the shut-down of NADW.

The recirculated Pacific water with the $\delta^{13}\text{C}$ of $\sim 0\text{ ‰}$ combined with certain transfer of nutrients (light carbon) from intermediate depth to deep-water¹⁸ in the Southern Ocean might produce low- $\delta^{13}\text{C}$ CDW. The low- $\delta^{13}\text{C}$ deep-water appears to have formed in the pre-T2 Southern Ocean as evident in the ODP1123^(ref. 17). The reduced ventilation of deep Southern Ocean and low- $\delta^{13}\text{C}$ transfer from intermediate-water to deep-water^{18,19} would have been a valid argument for explaining the pre-T2 low mean- $\delta^{13}\text{C}$ in the EIO, assuming such water flooded it's abyss. In that case, the intermediate depth benthic- $\delta^{13}\text{C}$ record of ODP1088 should have exhibited not only an opposite trend to that of deep-water benthic-records of the same region (ODP1089 and ODP1090), but also higher $\delta^{13}\text{C}$ values for the pre-T2 period than the post-T2, simply because in such transfer, the intermediate-water would loose light-carbon to deep-water. But those Southern Ocean records of $\delta^{13}\text{C}$ generated from the ODP1088, ODP1089, and ODP1090^(ref. 18) exhibit nearly similar unidirectional shift around the T2 with distinctly lower (and higher) mean- $\delta^{13}\text{C}$ during the pre- (and post)-T2 periods. Such $\delta^{13}\text{C}$ patterns clearly oppose the vertical transfer of low- $\delta^{13}\text{C}$ in the Southern Ocean. On the other hand, the pre-T2 benthic- $\delta^{13}\text{C}$ in Pacific sector of the Southern Ocean [see ODP1123^(ref. 17)] is not enough to elevate the pre-T2 CDW- $\delta^{13}\text{C}$ to the mean-Pacific value (0 ‰) in presence of very low- $\delta^{13}\text{C}$ Atlantic sector water, which is required to explain the pre-T2 EIO benthic- $\delta^{13}\text{C}$ of $\sim -0.3\text{ ‰}$.

The local mutation of CDW in the deep EIO by mixing with other high- $\delta^{13}\text{C}$ water is another possibility to produce the observed pre-T2 low mean- $\delta^{13}\text{C}$. Therefore, it is necessary to address the possibility of high- $\delta^{13}\text{C}$ deep-water formation during the pre-T2 period. The potential candidates could be the north Pacific and the northern Arabian Sea. The NPDW contributing directly to EIO is highly unlikely, because first it should alter the CDW in the circumpolar region itself. Further, the planktic- and benthic- $\delta^{13}\text{C}$ evidences from the Shatsky Rise in the north Pacific exhibit unchanged vertical- $\delta^{13}\text{C}$ gradient of $\sim 1\text{ ‰}$ for the last 250 ky period³⁹ ruling out that possibility. The northernmost Arabian Sea in the western Indian Ocean, which presently experiences high-salinity water formation due to excess evaporation and shallow sinking and mixing with Red Sea water^{47,48} presents a remote possibility of high- $\delta^{13}\text{C}$ deep-water formation. However,

for the deep vertical convection to occur there, the estimated sea surface temperature drop⁴⁹ would be $>10^\circ\text{C}$, which is neither feasible in the tropical oceans nor evident in long-term sea surface temperature proxy records⁵⁰.

Thus most of the leading hypotheses do not explain adequately the observed shift in the mean-ocean $\delta^{13}\text{C}$ that began around T2. The lower mean- $\delta^{13}\text{C}$ during the glacial (and interglacial) periods of pre-T2 than during the glacial (and interglacial) periods of post-T2 therefore warrants a proposal for an alternative mechanism.

Proposed hypothesis

As most of the deep-water circulation schemes fail to explain the observed mean-ocean $\delta^{13}\text{C}$ shift, it appears that the pre-T2 global oceans fundamentally were characterized by lower- $\delta^{13}\text{C}$ than after. As the pre-T2 oceans were conditioned with low- $\delta^{13}\text{C}$, and the post-T2 oceans were conditioned with high- $\delta^{13}\text{C}$, the $\delta^{13}\text{C}$ -shift around T2 (MIS-6 to MIS-5 transition) was much larger than around T1 (LGM to Holocene transition).

The source of pre-T2 low- $\delta^{13}\text{C}$ however is an open question, and also the changes in various carbon-reservoir are not possible to quantify yet due to lack

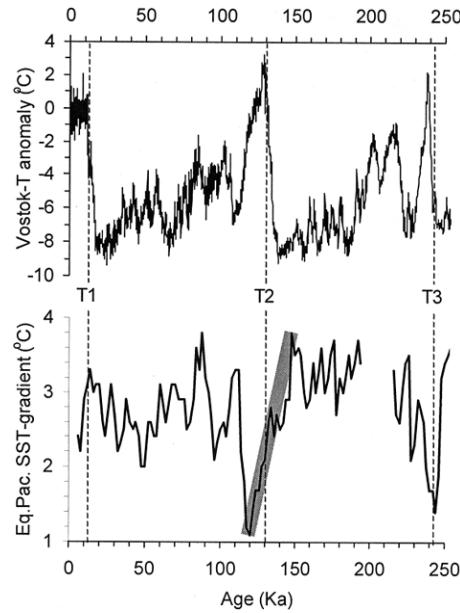


Fig. 4 — Past variation in the Equatorial Pacific surface temperature gradient⁵¹ and Vostok air temperatures⁵³. A major reorganization in the tropical Pacific wind pattern appears to have taken place around T2 as evident in $\sim 5^\circ\text{C}$ shift in the SST gradient within ~ 20 kyr (shown with shaded slanting bar). Note a similar gradient change around T3 but not around T1.

of data. Nevertheless, it is proposed that this fundamental shift in the oceanic carbon-reservoir around the T2 may have been associated with a major change in ocean-atmosphere coupled system significantly reorganizing the global wind patterns. The preliminary evidences in support of this may be traced in: i) the equatorial Pacific sea surface temperature (SST) gradient across the T2⁵¹ (Fig. 4), and ii) the Antarctic atmosphere records^{52,53} (Fig. 4).

A small variation in the equatorial Pacific SST has the potential to induce large-scale changes in atmospheric circulation, which may have the effect on global wind patterns⁵⁴. It appears from the past-SST gradient between west- and east-equatorial Pacific that, a major reorganization has occurred in the Walker Circulation around the T2 (see Fig. 4). This might have affected the supply of terrestrial light-carbon to the oceans from a pre-T2 high input to post-T2 low input, because, most of the terrestrial carbon reservoirs (both organic and inorganic) are significantly lower in $\delta^{13}\text{C}$ compared to mean-ocean value of $\sim 0.0\text{‰}$. This change could have been possible without significantly altering the terrestrial photosynthetic carbon-reservoir, but by a drastic change in global wind intensity and pattern. An initial support for this proposal emerges from Antarctic ice-core record, where the penultimate climate cycle (MIS-7 through MIS-6) was relatively less warmer than the last climate cycle (MIS-5 through LGM), and the MIS-6 glacial period was much longer than the LGM^{52,53} (also see temperature anomaly at terminations in Fig. 4, where air temperature at T2 is warmest). Such climate set-up during the pre-T2 period probably has resulted in general more intense winds than the post-T2 winds.

More intense global winds could have enhanced the terrestrial-carbon input to the oceans during the penultimate climate cycle. Oba & Pedersen⁵⁵ have recognized the importance of eolian carbonate supply to the world ocean during the LGM as one of the key factors that could have lowered ~ 20 ppmv of the atmospheric CO_2 by strengthening the alkalinity-pump. The effect of increased carbonate-dust (alkalinity) on the $\delta^{13}\text{C}$ of seawater- ΣCO_2 would be due to increased burial of biogenic-calcite having high- $\delta^{13}\text{C}$. As a result of simultaneous remineralisation of low- $\delta^{13}\text{C}$ marine organic matter at depth coupled with removal of high- $\delta^{13}\text{C}$ calcite to the sediment, the overall ^{13}C of ΣCO_2 might have depleted. This oceanic ^{13}C depletion combined with

transfer of ^{12}C -enriched dust-carbon, would provide a base for reasonable mechanism to explain the pre-T2 low mean- $\delta^{13}\text{C}$. However, this hypothesis would require quantification of the terrestrial carbon-reservoir and -transfer and calcite burial to account for the lowered mean- $\delta^{13}\text{C}$ of the pre-T2 oceans by $\sim 0.4\text{‰}$ without significantly altering the terrestrial biomass.

From a visible examination of the available data (Figs. 3 and 4), the sequence of events surrounding the T2, based on the oxygen-isotope chronology of SK129 suggests that, the initiation of major decrease in Equatorial Pacific SST-gradient preceded the events of glacial termination and initiation of planktic- $\delta^{13}\text{C}$ excursion by ~ 20 ky, and initiation of Antarctic warming and benthic- $\delta^{13}\text{C}$ excursion by $\sim 5-10$ ky. This approximate sequence of events suggests a role of Tropical Pacific SST in initiating climatically important events surrounding the T2. Interestingly, the past climate rhythm as evident in marine sedimentary records^{26,29}, and Antarctic atmosphere records^{52,53,56}, do not contain one-time large shift comparable to that of the oceanic- $\delta^{13}\text{C}$ shift evident around T2. In that case, to what extent the deep-ocean circulation changes may have played a role in the past climate change needs to be clearly assessed for a better understanding of the deep-water circulation link with climate change.

Acknowledgement

The Department of Ocean Development (Government of India) provided ship time on board *ORV Sagar Kanya* for sampling. I wish to thank, Sir Nick Shackleton, Prof. Nick McCave, and Prof. Harry Elderfield of Cambridge University, UK., and Prof. Raghu Murtugudde of Maryland University, USA for initial useful discussions. The suggestions and comments by anonymous reviewers were very helpful. I thank Prof. David Lea and Dr. Ian Hall for providing their published data. Ms. Anjali Chodankar and Mr. Mike Hall helped in sample processing and isotope measurements respectively. I also thank the St. John's Council, Cambridge University, UK for electing me as one of the seven visiting scholars for 2002, from all professional backgrounds. The isotopic measurements were carried out at Godwin Laboratory, Cambridge University, UK. This is NIO contribution No. 4012.

References

- 1 Duplessy, J-C. & Shackleton, N. J., Response of global deep-water circulation to Earth's climate change 135000-107000 years ago, *Nature*, 316 (1985) 500-507.
- 2 Broecker, W. S., The great ocean conveyor, *Oceanography*, 4 (1991) 79-89.
- 3 Broecker, W. S. & Denton, G. H., The role of ocean-atmosphere reorganizations in glacial cycles, *Geochim. Cosmochim. Acta*, 53 (1989) 2465-2501.
- 4 Charles, C. D. & Fairbanks, R. G., Evidence from Southern Ocean sediments for the effect of North Atlantic deep-water flux on climate, *Nature*, 355 (1992) 416-419.
- 5 Raymo, M. E., Ruddiman, W. F., Shackleton, N. J. & Oppo, D. W., Evolution of $\delta^{13}\text{C}$ gradients over the last 2.5 m.y. *Earth Planet. Sci. Lett.*, 97 (1990) 353-368.
- 6 Curry, W. B., Duplessy, J. C., Labeyrie, L. D., & Shackleton, N. J., Changes in the distribution of $\delta^{13}\text{C}$ of deep-water ΣCO_2 between the last glaciation and the Holocene, *Paleoceanography*, 3 (1988) 317-341.
- 7 Oppo, D. W. & Fairbanks, R. G., Variability in the deep- and intermediate water circulation of the Atlantic Ocean during the past 25000 years: Northern hemisphere modulation of the Southern Oceans, *Earth Planet. Sci. Lett.*, 86 (1987) 1-15.
- 8 Raymo, M. E., Oppo, D. W. & Curry, W. B., A deep-sea carbon isotopic perspective, *Paleoceanography*, 12 (1997) 546-559.
- 9 Shackleton, N. J., Hall, M. A., Line, J. & Shuxi, C., Carbon isotopic data in core V19-30 confirm reduced carbon dioxide concentration in the ice-age atmosphere, *Nature*, 306 (1983) 319-322.
- 10 Boyle, E. A., Cadmium and $\delta^{13}\text{C}$ paleochemical ocean distribution during the stage 2 glacial maximum. *Annual Rev. Earth Planet. Sci.*, 20 (1992) 245-287.
- 11 Yu, E. F., Francois, R. & Bacon, M. P., Similar rates of modern and last glacial ocean thermohaline circulation inferred from radiochemical data, *Nature*, 379 (1996) 689-694.
- 12 Broecker, W. S., Peteet, D. M. & Rind, D., Does the ocean-atmosphere system have more than one stable mode of operation, *Nature*, 315 (1985) 21-26.
- 13 Rutberg, R. L., Hemming, S. R. & Goldstein, S., Reduced north Atlantic deep-water flux to the glacial Southern Ocean inferred from Nd-isotopes, *Nature*, 405 (2000) 935-938.
- 14 Keigwin, L. D., North Pacific deep-water formation during the latest glaciation, *Nature*, 330 (1987) 362-364.
- 15 Sigman, D. M., Jaccard, S. L. & Haug, G. H., Polar ocean stratification in a cold climate, *Nature*, 428 (2004) 59-63.
- 16 Adkins, J. F., McIntyre, K. & Schrag, D. P., The salinity, temperature, $\delta^{18}\text{O}$ of glacial deep ocean, *Science*, 298 (2002) 1769-1772.
- 17 Hall, I. R., McCave, I. N., Shackleton, N. J., Weedon, G. P. & Harris, S. E., Intensified deep-Pacific inflow and ventilation in Pleistocene glacial times, *Nature*, 412 (2001) 809-812.
- 18 Hodell, D. A., Venz, K. A., Charles, C. D. & Ninnemann, U. S., Pleistocene vertical carbon isotope and carbonate gradients in the south Atlantic sector of the Southern Ocean, *Geochem. Geophys. Geosyst.*, 4 (2003) 1-19.
- 19 Ninnemann, U. S. & Charles, C. D., Regional differences in Quaternary Subantarctic nutrient cycling: Link to deep and intermediate water ventilation, *Paleoceanography*, 12 (1997) 560-567.
- 20 Kroopnick, P. M., The distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the world oceans. *Deep Sea Res.*, 32 (1985) 57-84.
- 21 Mantalya, A. W., & Reid, J. L., On the origin of deep and bottom waters of the Indian Ocean. *J. Geophys. Res.* 100C (1995) 2417-2439.
- 22 Ganachaud, A., Wunsch, C., Marotzke, J. & Toole, W., Meridional overturning and large-scale circulation of the Indian Ocean, *J. Geophys. Res.* 105C (2000) 26117-26134.
- 23 Wyrtki, W. K., Physical Oceanography of the Indian Ocean, in: *The biology of the Indian Ocean*, edited by B. Zeitzschel, S. A. Gerlach, (Springer-Verlag, Berlin) 1973, pp. 18-36.
- 24 You, Y., Implication of deep-circulation and ventilation of the Indian Ocean on the renewal mechanism of the North Atlantic Deep Water, *J. Geophys. Res.* 105C (2000) 23895-23926.
- 25 Hemleben, C., Spindler, M. & Anderson, R. O., *Modern planktonic foraminifera*, (Springer-Verlag, NY) 1988, pp. 363
- 26 Imbrie, J., Hays, J. D., Martinson, D. G., McIntyre, A., Mix, A. C., Moley, J. J., Pisias, N. G., Prell, W. L. & Shackleton, N. J., The orbital theory of Pleistocene climate: support from a revised chronology of the marine $\delta^{18}\text{O}$ record. in: *Milankovitch and climate*, edited by A. Berger, N. J. Imbrie, (Plenum-Reidel, London) 1984, pp. 269-305.
- 27 Ninkovich, D., Distribution, age, and chemical composition of western Indonesia, *J. Volcanol. Hydrometeorol. Res.*, 5 (1979) 67-86.
- 28 Thompson, P. R., Be, A. W. H., Duplessy, J-C. & Shackleton, N. J., Disappearance of pink pigmented *G. ruber* at 120000 y B. P. in the Indian and Pacific oceans, *Nature*, 280 (1979) 554-558.
- 29 Shackleton, N. J., Carbon-13 in Uvigerina: Tropical rain forest history and the equatorial Pacific carbonate dissolution cycles. in: *The fate of fossil fuel CO₂ in the oceans*, edited by N. R. Anderson, A. Malahoff (Plenum, London) 1977, pp. 401-427.
- 30 Spero, H. J. & Lea, D. W., The cause of carbon isotope minimum event on glacial terminations, *Science*, 296 (2002) 522-525.
- 31 Kallel, N., Labeyrie, L. D., Leclerc, A. J & Duplessy, J-C., A deep hydrological front between intermediate and deep-water masses in the glacial Indian Ocean, *Nature*, 333 (1988) 651-655.
- 32 Naqvi, W. A., Charles, C. D. & Fairbanks, R. G., Carbon and oxygen isotopic records of benthic foraminifera from northeast Indian Ocean, *Earth. Planet. Sci. Lett.*, 121 (1994) 99-110.
- 33 McCormle, D. C., Heggie, D. T. & Veeh, H. H., Glacial and Holocene stable isotope distribution in the southeast Indian Ocean, *Paleoceanography*, 13 (1998) 20-34.
- 34 Broecker, W. S., Ocean chemistry during glacial times, *Geochim. Cosmochim. Acta*, 46 (1982) 1689-1706.
- 35 Beaufort, L., Lancelot, Y., Camberlin, P., Cayre, O., Vincent, E., Bassinot, F. & Labeyrie, L., Insolation cycles as a major control of Equatorial Indian Ocean primary productivity, *Science*, 278 (1997) 1451-1454.
- 36 Shackleton, N. J., The 100000 year ice-age cycle identified and found to lag temperature CO_2 , and orbital eccentricity, *Science*, 289 (2000) 1897-1902.
- 37 Curry, W. B. & Crowley, T. J., The $\delta^{13}\text{C}$ of equatorial Atlantic surface waters: implication for ice-age $p\text{CO}_2$ levels, *Paleoceanography*, 2 (1987) 489-517.

- 38 Shackleton, N. J. & Pisias, N. G., Atmospheric carbon dioxide, orbital forcing, and climate, *Geophys. Monogr.*, A.G.U. 32 (1985) 303-317.
- 39 Yamane, M., Late Quaternary variations in water mass in the Shatsky Rise area, North Pacific Ocean, *Mar. Micropal.*, 48 (2003) 205-223.
- 40 Flower, B. P., Late Quaternary stable isotopic stratigraphy of Hole 910, Yermak Plateau, Arctic Ocean, in: *Proc. ODP Sci. Results*, 151 (1996) 445-448.
- 41 Keigwin, L. D., Stable isotope stratigraphy and chronology of the upper Quaternary section at Site 883, Detroit Seamount, *Proc. ODP Sci. Results*, 145 (1995) 257-259.
- 42 Mix, A. C., Pisias, N. G., Rugh, W., Wilson, J., Morey, A. & Hagelberg, T. K., Benthic foraminifera stable isotope record from Site 849: Local and global climate change, *Proc. ODP. Sci. Results*, 138 (1995) 371-412.
- 43 Shackleton, N. J. & Hall, M. A., Stable isotope history of the Pleistocene at ODP Site 677, *Proc. ODP Sci. Results*, 111 (1989) 295-320.
- 44 Sarnthein, M. & Tiedemann, R., Towards high-resolution stable isotope stratigraphy of the last 3.4 million years: Sites 658 and 659 off northwest Africa, *Proc. ODP Sci. Results*, 108 (1989) 167-185.
- 45 Oppo, D. W., Fairbanks, R. G., Gordon, A. L. & Shackleton, N. J., Late Pleistocene Southern Ocean $\delta^{13}\text{C}$ variability, *Paleoceanography*, 5 (1990) 43-54.
- 46 Rau, A. J., Rogers, J., Lutjeharms, J. R. E., Giraudeau, J., Lee-Thorp, J. A., Chen, M. T. & Waelbroeck, C., A 450 ky record of hydrological condition on the western Agulhas Bank slope, South Africa, *Mar. Geol.*, 180 (2002) 183-201.
- 47 Kumar, S. P. & Prasad, T. G., Formation and spreading of Arabian Sea high salinity water mass, *J. Geophys. Res.*, 104C (1999) 1455-1464.
- 48 Shenoi, S. S. C., Shetye, S. R., Gouveia, A. D. & Michael, G. S., Salinity extrema in the Arabian Sea. In: *Monsoon biogeochemistry*, edited by V. Ittekot, R. R. Nair, SCOPE-UNEP, Heft 76, (Hamburg Univ.), 1993, pp. 37-50.
- 49 Reichart, G. J., Laurens, L. J. & Zachariasse, W. J., Temporal variability in the northern Arabian Sea oxygen minimum zone during the last 225000 years, *Paleoceanography*, 13 (1998) 607-621.
- 50 Emeis, K. C., Anderson, D. M., Doose, H., Kroon, D. & Schulz-Bull, D., Sea surface temperature and history of monsoon upwelling in the northwest Arabian Sea, *Quat. Res.*, 43 (1995) 355-361.
- 51 Lea, D. W., Pak, D. K. & Spero, H. J., Climate impact of late Quaternary Equatorial Pacific sea surface temperature variations, *Science*, 289 (2000) 1719-1724.
- 52 Epica community members, Eight glacial cycles from an Antarctic ice core, *Nature*, 429 (2004) 623-628.
- 53 Petit, J. R., Jouzel, J., Raynaud, D., Barkov, N. I., Barnola, J. M., Basile, I., Bender, M., Chapellaz, J., Davis, J., Dealygue, M., Delmotte, M., Kotlyakov, V. M., Legrand, M., Lipenkov, Y. V., Lorius, C., Pepin, L., Ritz, C., Saltzman, E. & Steinenard, M., Climate and atmospheric history of the past 420000 y from Vostok ice core, *Nature*, 399 (1999) 429-436.
- 54 Cane, M. A., A role for tropical Pacific, *Science*, 282 (1998) 59-61.
- 55 Oba, T. & Pedersen, T. F., Paleoclimatic significance of eolian carbonate supplied to the Japan Sea during the last glacial maximum, *Paleoceanography*, 14 (1999) 34-41.
- 56 Barnola, J. M., Raynaud, D., Korotkevich, Y. S. & Lorius, C., Vostok ice-core provides 160000 years record of atmospheric CO₂, *Nature*, 329 (1987) 408-414