Oxygen isotope stratigraphy and events in the northern South China Sea during the last 6 million years

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Abstract Based on the stable isotopic analysis of more than 1000 samples of planktonic and benthic foraminifers from ODP Site 1148 in the northern South China Sea (SCS), the oxygen isotope stratigraphy has been applied to the last 3 million years for the first time in the SCS. Furthermore, the paleoceanographic changes in the northern SCS during the last 6 million years have been unraveled. The benthic foraminiferal $\delta^{18}O$ record shows that before ~3.1 Ma the SCS was much more influenced by the warm intermediate water of the Pacific. The remarkable decrease in the deepwater temperature of the SCS during the period of 3.1—2.5 Ma demonstrates the formation of the Northern Hemisphere ice-sheet. However, the several sea surface temperature (SST) reductions during the early and middle Pliocene, reflected by the planktonic foraminiferal $\delta^{18}O$, might be related to the ice-sheet growth in the Antarctic region. Only those stepwise and irreversible SST reductions during the period of ~2.2—0.9 Ma could be related to the formation and growth of the Northern Hemisphere ice-sheet.

Keywords: oxygen isotope stratigraphy, Pliocene-Pleistocene, Northern Hemisphere glaciation, middle Pleistocene revolution, South China Sea.

During the Pliocene-Pleistocene, the Earth changed from a warm and uniform environment into an extreme climate with typical glacial-interglacial cycles. Therefore, the Pliocene-Pleistocene stratum is a natural laboratory for testing many paleoclimatological and paleoceanographic hypotheses[1,2]. Moreover, the astronomically tuned isotope stratigraphy has been applied to the last 6 million years[3,4], providing the possibility for the Pliocene high-resolution research. Consequently, the Pliocene-Pleistocene has become a hot spot of international paleoceanographic studies. Previous studies of the South China Sea (SCS) laid their emphases on the late Quaternary[5,6], while the early Pleistocene[7] and Pliocene[8] were rarely explored. A large number of sediment cores were collected during the ODP Leg 184 to the SCS in 1999, among which ODP Site 1148 of the northern SCS (18° 50′N, 116° 34′E; water depth ~3297 m; base age ~32 Ma) provides a relatively continuous paleo-environmental profile since the early Oligocene[9]. Until now, we have finished more than 2000 samples of planktonic and benthic foraminifers from this site for the stable isotopic analyses. This paper mainly presents the oxygen isotopic changes of the last 6 million years since the latest Miocene, and hence discusses the responses of the SCS to the formation of the Northern Hemisphere ice-sheet and other paleoceanographic events, and their impacts on global climate.

1 Sample preparation and stable isotopic measurements

Based on the shipboard preliminary biostratigraphic and magnetostratigraphic chronology of ODP Leg 184[9], we sampled the composite profile of Site 1148 at a sampling interval of 80 cm for the part of 0—125.8 mcd (meter, composite depth) and of 15 cm for the part of 125.8—201.7 mcd. There are 755 samples in total. About 10 cubic centimeter sediment was taken out from each sample and processed using the standard micropaleontological technique. Planktonic and benthic foraminiferal tests (>0.154 mm) were picked, ultrasonically cleaned in analytical grade ethanol, and then dried in an oven at 60℃. The stable isotopic measurements were carried out at the Laboratory of Marine Geology, Ministry of Education, Tongji University using a Finnigan/MAT 252 mass spectrometer with an automated carbonate device. Calibration to the international PeeDee Belemnite (PDB) scale was through NBS19 and NBS18 standards. The analytical resolution of $\delta^{18}O$ is 0.07 ‰ over the period of analysis.

Because the carbonate dissolution at this site is very strong due to its deep water depth, in many samples there are not enough clean tests of planktonic foraminifer *Globigerinoides ruber* or *Globigerinoides sacculifer* (7—8 specimens) and benthic foraminifers *Cibicidoides wuellerstorfi* (3—5 specimens) for the isotopic analyses. Finally we measured 475 planktonic foraminiferal and 538 benthic foraminiferal samples in this study. Besides *C. wuellerstorfi*, *Uvigerina* sp., *Cibicidoides kullenbergi*, *Cibicidoides* spp. and *Oridorsalis* sp. were also used for the isotopic analysis of benthic foraminifer. There are interspecies offsets for the isotopic values of these species/genera. In this paper, all isotopic values of benthic foraminifer have been corrected into *C. wuellerstorfi* equivalent values following the method of Shackleton et al. (1995)[3]. The corrected values of the same sample from different species/genera were averaged. In the isotopic analysis of planktonic foraminifer, *G. ruber* was used for the upper part (0—160 mcd) at this site while *G. sacculifer* was measured for the lower part. Because the two species are surface-dwelling species[10], no interspecies correction was done for the two species in this study.

2 Stratigraphy

The chronological framework of the last 3 million years at ODP Site 1148 was mainly based on astronomically tuned oxygen isotope stratigraphy. Due to the little influence of local water temperature and salinity, the benthic foraminiferal $\delta^{18}O$ is convenient for comparison between oceans[3,4]. Therefore, on the basis of the shipboard biostratigraphy and magnetostratigraphy[9], we recognized every oxygen isotopic event mainly from the benthic foraminiferal $\delta^{18}O$ curve. The results are shown in fig. 1. There are 126 oxygen marine isotopic stages (MIS) counted above 164 mcd at this site, among which MISs 32—34 were based on the planktonic foraminiferal $\delta^{18}O$

curve. MISs 76—77 could not be distinguished possibly due to a stratigraphic hiatus or very low sedimentation rate. In this study, we followed the isotope timescale of Imbrie et al. (1984) for 0—0.6 Ma[11], Shackleton et al. (1990) for 0.6—2.0 Ma[12], and Shackleton et al. (1995) for 2.0—3.1 Ma[3]. Thus, the average time resolution of samples for the isotopic analyses is about 9 ka during the period of 0—3.1 Ma. The reliability of isotope stratigraphy at this site can further be generally validated by the spectral analysis. Fig. 2 shows that during the period of 0—0.9 Ma, a 100 ka periodicity dominated the $\delta^{18}O$ record while a 41 ka periodicity was more important before it. The transformation of climatic periodicities occurred at the boundary of MIS 22/23 (~0.9 Ma B.P.), that is, the middle Pleistocene Revolution (MPR), being consistent with previous studies[7,13].

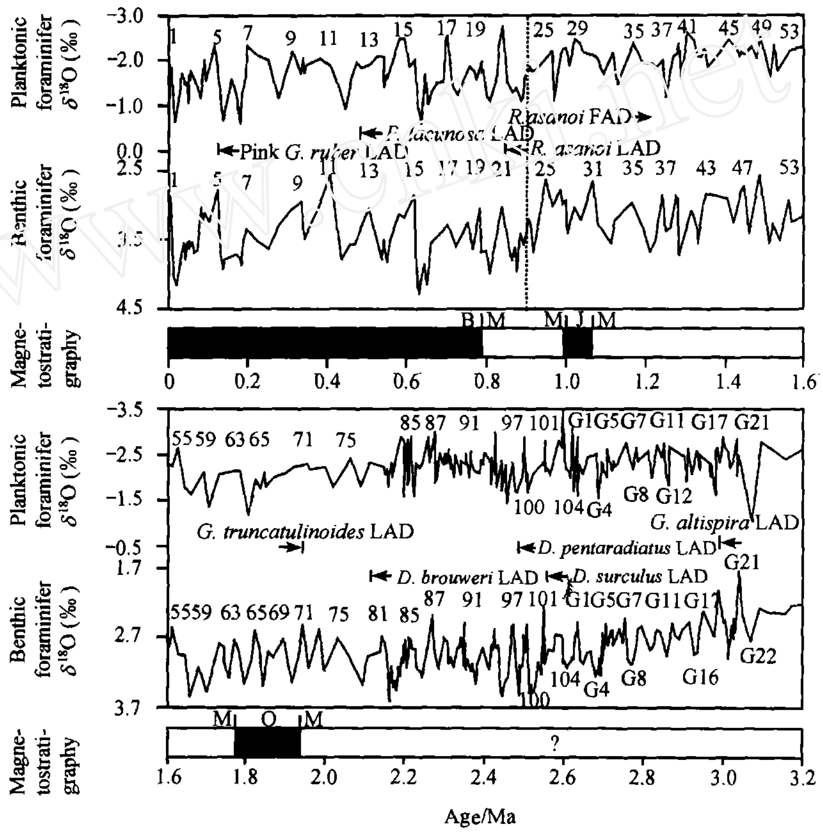

Fig. 1. Oxygen isotope stratigraphy since 3.1 Ma at ODP Site 1148. Biostratigraphic and magnetostratigraphic boundaries are based on the shipboard work of ODP Leg 184[9]. LAD is the last appearance datum, FAD is the first appearance datum. B, M, J and O represent Brunhes, Matuyama, Jaramillo and Olduvai, respectively, while question mark shows that no magnetic reversal was found. Numbers indicate oxygen isotopic stages. Vertical dashed line denotes the MPR.

Before MIS G22, because the average time resolution of planktonic and benthic foraminiferal samples for the isotopic analyses are 25 ka and 16 ka, respectively, it is difficult to recognize MIS. We mainly adopted the ages of biostratigraphic datum[9], and interpolated the age of each sample according to the average sedimentation rate between two datums. Planktonic foraminifer *Globorotalia tumida* firstly appeared at the depth of 198.8 mcd, corresponding to an age of 5.85 Ma.

Therefore, the sediments above 200 mcd belong to the latest Miocene to Holocene. Based on the above-mentioned isotope stratigraphy and biostratigraphy, the changes of sedimentation rate during the last 6 million years at ODP Site 1148 are mainly divided into two sections. The average sedimentation rate during the period of 3.1—6.0 Ma is only 12.6 m/Ma, while it remarkably increased during the period of 0—3.1 Ma with an average of 53.6 m/Ma.

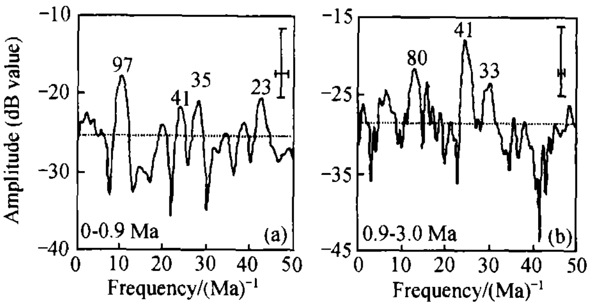

Fig. 2. Results of spectral analysis on the benthic foraminiferal $\delta^{18}O$ of the last 3 Ma at ODP Site 1148. (a), 0—0.9 Ma; (b), 0.9—3.0 Ma. Numbers above peaks indicate period (ka). Horizontal dashed line denotes the average value of the spectrum, considering the peaks above it to be significant. Cross in upper-right corner marks 6-dB bandwidth (horizontal) and 80% confidence interval (vertical).

3 Oxygen isotope and the Northern Hemisphere glaciation

The $\delta^{18}O$ record of benthic foraminifer *C. wuellerstorfi* is extensively used to reflect the change of the polar ice volume and bottom water temperature. The benthic foraminiferal $\delta^{18}O$ curve at ODP Site 1148 distinctly displays stepwise change (fig. 3). The early and middle Pliocene (~3.1—6.0 Ma) and ~0—0.9 Ma after the MPR are two relatively stable stages, with the average $\delta^{18}O$ values of 2.12 ‰ and 3.15 ‰ respectively. Between the two stages, there were two major $\delta^{18}O$ increases: one was the stepwise gradual increase during the period of 3.1—2.5 Ma while the other was the abrupt increase during the MPR. Before ~3.1 Ma, the fluctuation in the deepwater $\delta^{18}O$ of the SCS was small (<0.9 ‰), implying that the deepwater temperature was relatively stable corresponding to the warm global climate during this period[14]. During the period of 3.1—2.5 Ma, that is, from MIS G22 to MIS 100 (fig. 1), the deepwater $\delta^{18}O$ of the SCS increased by about 1.2‰, confirming that the Northern Hemisphere ice-sheet formed during this period[3,4,14]. Particularly, at ~2.7 Ma B.P., the deepwater $\delta^{18}O$ of the SCS decreased below the Holocene value, suggesting that the Northern Hemisphere ice volume had already increased above the modern level at ~2.7 Ma[15]. Because the $\delta^{18}O$ increase between 3.1 and 2.5 Ma is only visible in the benthic record (fig. 3), it is inferred that the deepwater temperature of the SCS decreased during this period. Taking into consideration the ice volume effect, the amplitude of the deepwater temperature decrease should be smaller than 3.8—4.9℃ (temperature change of 1℃ is equivalent to $\delta^{18}O$ change of 0.21‰— 0.27‰[16]). After the formation of the Northern Hemisphere ice-sheet, the average deepwater $\delta^{18}O$ value of the SCS increased by 0.53‰ at the MPR. The $\delta^{18}O$ increase at the MPR was also visible in the planktonic record (fig. 3), it seems that this is a global $\delta^{18}O$ change of sea water, indicating the further growth of the Northern Hemisphere ice volume. In fact, the amplitude of fluctuation in the deepwater $\delta^{18}O$ of the SCS also increased, from <0.9‰ before

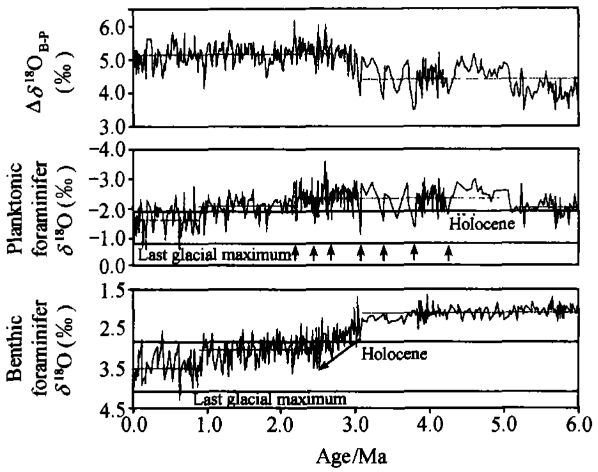

Fig. 3 Benthic and planktonic foraminiferal $\delta^{18}O$ and their difference ($\Delta\delta^{18}O_{B-P}$) of the last 6 Ma at ODP Site 1148. Horizontal solid lines show the values of the Holocene and Last Glacial Maximum, while the dashed lines show the average values of each section. Oblique arrow denotes the formation of the Northern Hemisphere ice-sheet, while small arrows indicate several obvious decreases of the planktonic foraminiferal $\delta^{18}O$ during the Pliocene.

~3.1 Ma to 1.4‰ of 2.5—0.9 Ma and then to 1.8‰ of 0—0.9 Ma, also indicating the growth of the Northern Hemisphere ice volume. The causes of the MPR, particularly the transformation of climatic periodicities from 41 ka to 100 ka during this period (fig. 2), are still vigorously debated in academic circles[13]. However, it is no doubt that the Northern Hemisphere ice volume increased after the MPR.

Planktonic foraminiferal $\delta^{18}O$ displays that there are three major changes in the sea surface temperature (SST) of the SCS since the Pliocene (fig. 3), the increase of 5.2—5.0 Ma and the decreases at ~2.2 Ma B.P. and ~0.9 Ma B.P. During the period from the latest Miocene to the early Pliocene (6.0—5.2 ma), the SST of the SCS was relatively cold, with an average $\delta^{18}O$ value of −1.98‰. From ~5.2 Ma B.P., the planktonic $\delta^{18}O$ distinctly became light, but after ~5.0 Ma it basically fluctuated around −2.32‰. In fact, during the period of ~5.0—2.2 Ma, the planktonic $\delta^{18}O$ became heavy several times (~4.3, 3.8, 3.4, 3.1, 2.7 and 2.5 Ma), implying that even during the warm period of the Pliocene there were brief cooling events in the SST of the northern SCS in about every 400 ka. After each event, the SST and $\delta^{18}O$ returned to the average value. These SST cooling events during the Pliocene, for example the cooling at ~4.3 Ma and ~3.8 Ma, have been already found in the North Pacific[17] and the Southern Ocean[18], attributed to the cryosphere development in the Antarctic region[17]. During the formation of the Northern Hemisphere ice-sheet reflected by the benthic foraminiferal $\delta^{18}O$, there were several very brief decreases in the SST of the SCS, which, however, returned to the level of the Pliocene. Only during the late Pliocene at ~2.2 Ma B.P., the surface water $\delta^{18}O$ of the SCS irreversibly became heavy for the first time and once again during the MPR at ~0.9 Ma B.P., suggesting that the SST of the SCS had ever stepwise decreased. This also indicates the formation and growth of the Northern Hemisphere ice-sheet, to which the response of the surface water, however, lagged behind that of the deepwater in the SCS. Calculating from the difference in the average $\delta^{18}O$ value between the Pliocene (5.0—2.2 Ma; −2.32‰) and the period after the MPR (−1.61‰), the SST decrease should be smaller than 2.6—3.4℃. Wang (1994) analyzed the data of eight DSDP sites and concluded that the SST of the subtropical North Pacific irreversibly stepwise decreased at ~2.2 Ma B.P. and ~1.0 Ma B.P, while that of the tropical Pacific did not change[17]. ODP Site 1148 of the northern SCS is located in the tropical sea area, but its SST change is similar to that of the subtropical Pacific. This might be related to the position of this site close to the Asian continent, also reflecting the particularity of marginal sea.

The difference ($\Delta\delta^{18}O_{B\text{-}P}$) between benthic and planktonic foraminiferal $\delta^{18}O$ at ODP Site 1148 can represent the vertical temperature gradient from the surface to bottom water (fig. 3). The $\Delta\delta^{18}O_{B\text{-}P}$ value before 3.1 Ma (averaging 4.38‰) was clearly lower than that after 2.7 Ma (averaging 5.16‰), indicating that the vertical temperature gradient was smaller during the warm period of the Pliocene. From ~3.1 Ma, the $\Delta\delta^{18}O_{B\text{-}P}$ began to increase and reached the Holocene value at ~2.7 Ma B.P, implying that the Northern Hemisphere ice-sheet formed at this time. Because the temperature decrease of the deepwater was much more remarkable than that of the surface water, as a result, the vertical temperature gradient of the SCS increased.

4 Comparing with the oxygen isotopic records of open oceans

We selected ODP851 of the eastern equatorial Pacific (12° 46′ N, 110° 34′ W; water depth 3760 m)[19] and ODP806B of the western equatorial Pacific (0° 09′ N, 159° 2′ E; water depth 2520 m)[20], to compare the difference of the surface water stable isotopic records between the northern SCS and the Pacific. The isotopic data of *G. sacculifer* were adopted for the two Pacific sites. As shown in fig. 4, the surface water $\delta^{18}O$ values at ODP Site 1148 are clearly more negative compared with those of the equatorial Pacific. Because the modern SST of the northern SCS is lower than that of the equatorial Pacific, the surface water $\delta^{18}O$ values in the SCS should be more positive. Therefore, the more negative values of the surface water $\delta^{18}O$ in the northern SCS are likely related to the lower sea surface salinity caused by the input of river-discharged fresh water. During the Pliocene, there were several cooling events in the equatorial Pacific reflected by the surface water $\delta^{18}O$, for example, at 3.5—3.4 Ma, 3.1—2.7 Ma, 2.6—2.4 Ma and 2.0—1.9 Ma. These events are possibly related to the close of the Panama Isthmus and the onset of the Northern Hemisphere glaciation, or the evolution of the vertical structure of the upper water in this region[19]. Nevertheless, the cooling event at ~2.2 Ma B.P. in the northern SCS was irreversible, different from the SST change in the equatorial Pacific and similar to the characteristics of the subtropical sea area[17].

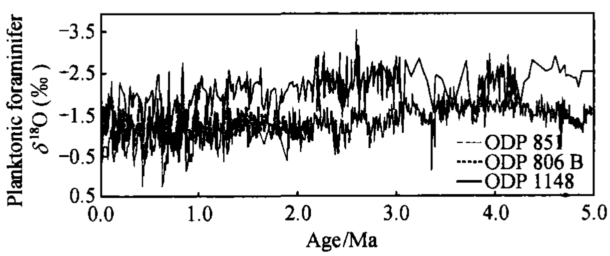

Fig. 4. Comparison of planktonic foraminiferal oxygen isotopes since the Pliocene between the northern SCS and ODP851, ODP 806B of the equatorial Pacific.

In addition, ODP659 of the Atlantic (18° 05′ N, 21° 02′ W; water depth 3070 m)[4], and ODP849 (0° 11′ N, 110° 31′ W; water depth 3851 m)[21] and ODP1014 (32° 50′ N, 119° 58′ E;

water depth 1165 m)[22,23] of the Pacific were selected for comparing the isotopic difference between the deepwater of the northern SCS and the intermediate (ODP1014) and deep water (ODP849 and ODP659) of open oceans. In fig. 5, all isotopic data have been corrected to the *C. wuellerstorfi* equivalent value. During the period of ~3.1—2.5 Ma, the deepwater $\delta^{18}O$ became heavy simultaneously in the northern SCS, the Pacific and North Atlantic (fig. 5), further indicating that the Northern Hemisphere ice-sheet formed during this period. Afterwards, the deepwater

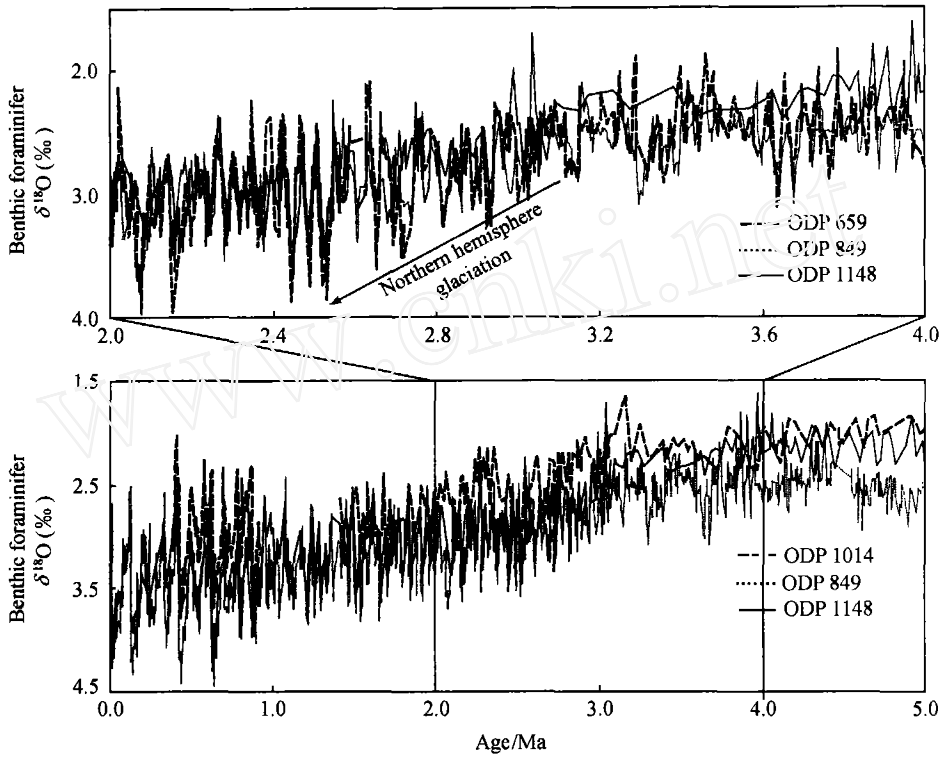

Fig. 5. Comparison of benthic foraminiferal oxygen isotopes since the Pliocene between the northern SCS and ODP849, ODP1014 of the Pacific and ODP659 of the Atlantic. Oblique arrow indicates the formation of the Northern Hemisphere ice-sheet.

$\delta^{18}O$ changes in the northern SCS and the Pacific were similar. During the MPR, the deepwater $\delta^{18}O$ became heavy again and the average $\delta^{18}O$ value changed from ~3.1‰ of 0.9—2.5 Ma into ~3.5‰ of 0—0.9 Ma, but with slightly greater amplitude of the deepwater $\delta^{18}O$ fluctuation in the Pacific. Interestingly, before the formation of the Northern Hemisphere ice-sheet (>3.1 Ma), the deepwater $\delta^{18}O$ of the SCS was generally lighter than that of the Pacific at similar water depth (ODP849; averaging 2.53‰). On the contrary, it was close to the $\delta^{18}O$ of the Pacific intermediate water (ODP1014, averaging 2.05‰). Because the sill depth of the SCS is ~2500 m, the deepwater of the SCS originates only from the Pacific deepwater above 2500 m. Before ~3.1 Ma, the $\delta^{18}O$ value of the SCS deepwater at the depth of ~3300 m was close to that of the Pacific intermediate water, suggesting that at that time the Pacific intermediate water was warmer[23] and had more influence upon the SCS than at present. After the formation of the Northern Hemisphere ice-sheet, the influence of the Pacific deep water upon the SCS increased and hence the deepwater $\delta^{18}O$ of both the SCS and the Pacific became close and heavier than the $\delta^{18}O$ of the Pacific intermediate water (fig. 5).

5 Conclusions

Planktonic and benthic foraminifers from ODP Site 1148 of the northern SCS have been studied in detail for the stable isotopic analyses. On the basis of the shipboard biostratigraphic and magnetostratigraphic work, there are 126 MIS counted above 164 mcd at this site. For the first time the oxygen isotope stratigraphy has been applied to the last three million years in the SCS.

The remarkable decrease of the deep water temperature of the SCS during the period of 3.1—2.5 Ma, reflected by the benthic foraminiferal $\delta^{18}O$, demonstrates that the Northern Hemisphere ice-sheet developed during this period and its volume probably increased above the modern level at ~2.7 Ma B.P. The several SST reductions during the early and middle Pliocene, reflected by the planktonic foraminiferal $\delta^{18}O$, might be related to the ice-sheet growth in the Antarctic region. Only those stepwise and irreversible SST reductions during the period of ~2.2—0.9 Ma could be related to the formation and growth of the Northern Hemisphere ice-sheet.

Before ~3.1 Ma, the $\delta^{18}O$ value of the SCS deepwater at the depth of ~3300 m was close to that of the Pacific intermediate water, suggesting that at that time the Pacific intermediate water was warmer and had a greater influence upon the SCS.

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