Available online at www.sciencedirect.com
ScienceDirect
PALAEO
Palaeogeography, Palaeoclimatology, Palaeoecology 246 (2007) 440–453
www.elsevier.com/locate/palaeo

Sea surface temperature reconstruction for the middle Okinawa Trough during the last glacial–interglacial cycle using C_{37} unsaturated alkenones

Houyun Zhou [a,c,d,*], Tiegang Li [b], Guodong Jia [c], Zhaoyu Zhu [a], Baoquan Chi [a,e], Qiyuan Cao [b], Rongtao Sun [b], Ping'an Peng [c]

[a] *Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, Guangzhou, 510640, China*
[b] *Institute of Oceanology, Chinese Academy of Sciences, Qingdao, 266071, China*
[c] *State Key Laboratory of Organic Geochemistry, Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, Guangzhou 510640, China*
[d] *State Key Laboratory of Loess and Quaternary Geology, Institute of Earth Environment, Chinese Academy of Sciences, Xi'an, 710075, China*
[e] *The Graduate School of the Chinese Academy of Sciences, Beijing, 100039, China*

Received 29 June 2005; received in revised form 14 October 2006; accepted 18 October 2006

Abstract

C_{37} unsaturated alkenones were analyzed on a core retrieved from the middle Okinawa Trough. The calculated $U^{K'}_{37}$ displays a trend generally parallel with those of the oxygen isotopic compositions of two planktonic foraminiferal species, *Neogloboquadrina dutertrei* and *Globigerinoides sacculifer*, suggesting that in this region, SST has varied in phase with global ice volume change since the last glacial–interglacial cycle. The $U^{K'}_{37}$-derived SST ranged from ca. 24.0 to 27.5 °C, with the highest value 27.5 °C occurring in marine isotope stage 5 and the lowest ~24.0 °C in marine isotope stage 2. This trend is consistent with the continental records from the East Asian monsoon domain and the marine records from the Equatorial Pacific. The deglacial increase of the $U^{K'}_{37}$-derived SST is ~2.4 °C from the Last Glacial Maximum to the Holocene.

Keywords: $U^{K'}_{37}$ index; SST; Paleoceanography; Kuroshio Current; Glacial/interglacial variations

1. Introduction

The Okinawa Trough is one of the back-arc basins of the western Pacific Ocean. It was formed between ca. 2 and 0.5 Ma BP (Lee et al., 1980; Kobayashi, 1985; Letouzey and Kimura, 1986). It extends from north–northeast to south–southwest, joining the East China Sea (ECS) on its western side and facing the Ryukyu Island Arc to the east (Fig. 1). The trough is an important site for paleoclimatic, paleohydrological and paleoceanographic studies. Firstly, it is influenced by both the open ocean (the west Pacific Ocean) and the coastal water from the ECS. Enormous nutrients and suspended sediments from the Changjiang River are supplied to the trough, where they interact with the water mass from the open ocean. The sediments in the trough therefore bear information about both terrestrial and marine processes.

* Corresponding author. Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, Guangzhou, 510640, China. Fax: +86 20 85290130.
E-mail address: hyzhou@gig.ac.cn (H. Zhou).

0031-0182/$ - see front matter © 2007 Elsevier B.V. All rights reserved.
doi:10.1016/j.palaeo.2006.10.011

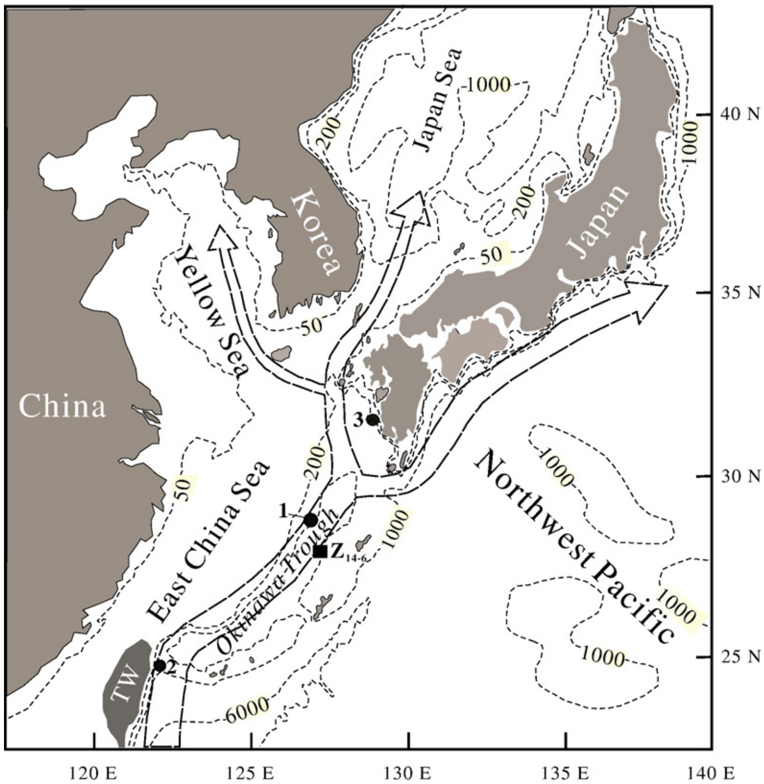

Fig. 1. Location of the Okinawa Trough and core $Z_{14\text{-}6}$. The dashed arrows indicate the Kuroshio Current and its branches. Dashed lines with numbers are bathymetric contours. Solid circles are cores studied previously in the Okinawa Trough. 1: DGKS9603 (Li et al., 2001); 2: 1202B (Zhao et al., 2005); 3: MD982195 (Ijiri et al., 2005).

Secondly, the trough can provide continuous paleoclimatic and paleoceanographic records from the last glacial–interglacial cycle, especially from the Last Glacial Maximum (LGM), when sea level in the west Pacific marginal seas was more than 100 m lower relative to present (Lambeck and Chappell, 2001). During the LGM, most of the ECS was exposed, while the Okinawa Trough was still below sea level and kept receiving sediments from the continental Chinese land mass and from the Ryukyu Island Arc. This makes the trough an ideal site for investigating the interaction between continental and marine processes during the last glacial–interglacial cycle. Thirdly, the trough is presently bathed by the western boundary current of the northwest Pacific, the Kuroshio Current (Fig. 1). The current originates from the North Equatorial Current in the western Pacific, carries warm and saline water, crosses between the southern end of the Ryukyu Island Arc and northeast of Taiwan and flows along the Okinawa Trough towards the north until it crosses the arc again at the Tokara Strait (Jian et al., 2000; Li et al., 2001). The Kuroshio Current exerts a significant influence on the sea surface temperature (SST) and salinity distribution in this region. The Okinawa Trough is occupied by the open-sea water mass with high temperature and deep thermocline (Jian et al., 2000). The summer, winter and annual SSTs in the ECS range between 26–29, 9–23 and 16–26 °C, respectively (Fig. 2). At the site of core $Z_{14\text{-}6}$ (27°07′N, 127°27′E; see Fig. 1 for its location), the summer, winter and annual SSTs are 28.5, 22 and ca. 25 °C, respectively. All these SSTs seem to display relatively higher values along the track of the Kuroshio Current, indicating the influence of this warm current. Furthermore, the main axis of the Kuroshio Current is in the area of highest heat and vapor supply to the atmosphere in the western Pacific, strongly influencing the East Asian climate, upper-ocean thermal structure and distribution of marine sediments in this region (Li, 1993). Change in the pathway and intensity of the Kuroshio Current during the Quaternary, especially during the last glacial–interglacial cycle, has attracted much attention (Jian et al., 1996; Li et al., 1997; Shieh et al., 1997; Jian et al., 1998; Sawada and Handa, 1998; Liu et al., 1999; Xu and Oda, 1999; Ujiié and Ujiié, 1999; Jian et al., 2000; Li et al., 2001; Wang et al., 2001a; Ujiié et al., 2001; Liu et al., 2003).

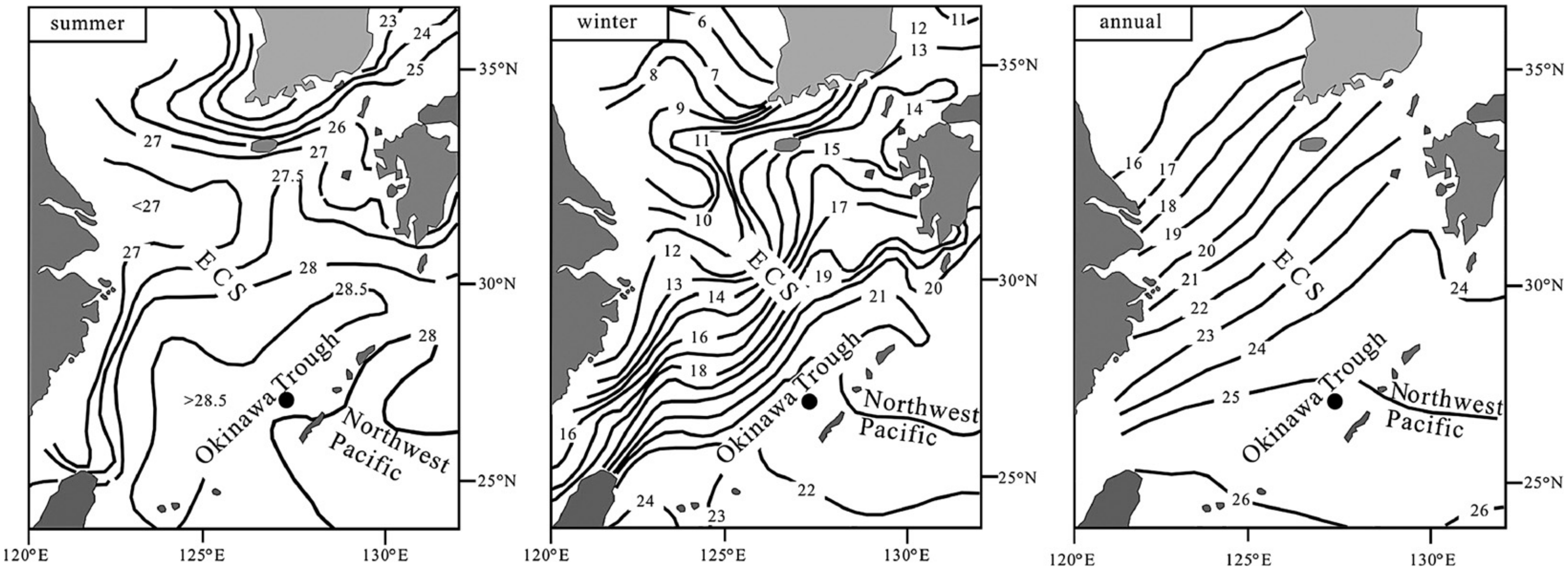

Fig. 2. Modern summer, winter and annual surface temperature distributions over the East China Sea. The solid circles represent core $Z_{14\text{-}6}$. The SST data were downloaded from the web site http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NODC/.WOA98/.ANNUAL/.analyzed/.temperature/.

Up to now, however, many paleoclimatic and paleoceanographic investigations in the Okinawa Trough have been based on assemblages and/or stable isotopes (such as the stable isotopic compositions of oxygen and carbon, $\delta^{18}O$ and $\delta^{13}C$) of foraminifera. Much less attention has been paid to organic matter, especially to biomarkers in sediments (Ijiri et al., 2005; Zhao et al., 2005). In fact, more than 10 years ago Zheng and Shi (1991) already detected long-chain alkenones in surface sediments from the Okinawa Trough. The investigations on sinking particles and sediments in the northwestern North Pacific off Japan and on two cultured marine coccolithophorids, *Emiliania huxleyi* and *Gephyrocapsa oceanica* (Sawada et al., 1996, 1998), indicated that the alkenone-based temperature mirrored SST or ambient air temperature. Thus biomarkers such as long-chain alkenones are important proxies for paleoclimate and paleoceanography. The SSTs reconstructed by Zhao et al. (2005) and Ijiri et al. (2005) in the Okinawa Trough covered only a small part of the Late Pleistocene. Here we report a longer SST record reconstructed for the middle Okinawa Trough using long-chain alkenones, and covering the last glacial–interglacial cycle.

2. Materials and methods

Core $Z_{14\text{-}6}$ was retrieved on the east side of the middle Okinawa Trough from a water depth of 739 m (Fig. 1). The core is 8.96 m long and composed mainly of silty clay and clay. It was first dated by Yan et al. (1990) back to marine isotope stage MIS 6 based on oxygen isotopic stratigraphy and the disappearance of *Globigerinoides rubber* pink (Thompson et al., 1979). Here some minor revisions are made according to the SPECMAP stacked $\delta^{18}O$ chronology (Fig. 3; Martinson et al., 1987). All the age control points are listed in Table 1. The age model was established by assuming constant sedimentation rates between two contiguous dates (Fig. 4).

In recent years, C_{37} unsaturated long-chain alkenones have been increasingly used to reconstruct SST. In the marine environment, the long-chain C_{37} alkenones are produced only by Haptophyceae algae in surface water. The variation in the number of the double bonds of long-chain alkenones depends largely on the ambient temperature (Marlowe et al., 1984). It was found that during sinking from sea surface to bottom, and during the early stage of diagenesis after deposition, temperature

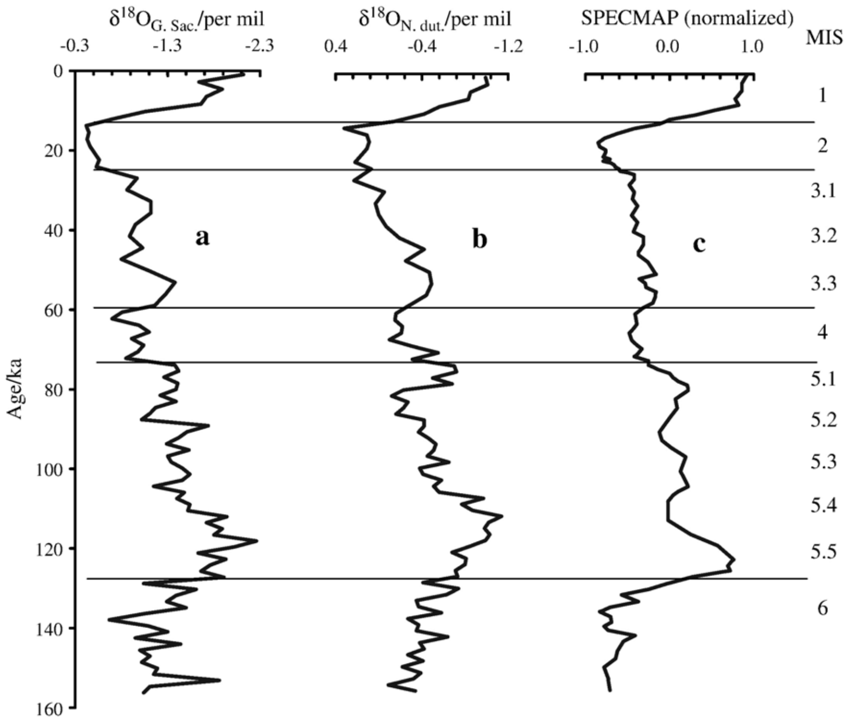

Fig. 3. Oxygen isotope stratigraphy of core $Z_{14\text{-}6}$. The $\delta^{18}O$ records of two species of planktonic foraminifera, *N. dutertrei* and *G. sacculifer* ($\delta^{18}O_{N.dut}$ and $\delta^{18}O_{G.sac}$) (Yan et al., 1990) were compared with the standard SPECMAP curve (Martinson et al., 1987).

Table 1
Age control points of core $Z_{14\text{-}6}$ based on oxygen isotopic stratigraphy and microfaunal event

MIS boundary	Depth below core top/cm[a]	Age/ka BP[b]	Notes
	0(0)	0(0)	
1–2	64(65)	10(12.05)	
2–3	170(135)	32(24.11)	
3–4	265(255)	56(58.96)	
4–5	340(345)	75(73.91)	
5–6	700(700)	128(128)	*Globigerinoides rubber* pink disappeared

a, b: depths and ages are from Yan et al. (1990), and numbers in brackets are our own estimations based on the oxygen isotopes of *G. sacculifer* and *N. dutertrei*. Although there are some minor differences, the two sets of values are generally in good agreement.

information contained in long-chain alkenones would not be qualitatively altered (Prahl et al., 1989; McCaffery et al., 1990; Conte et al., 1992; Madureira et al., 1995; Sawada et al., 1998). This makes it possible to reconstruct SST variation during the late Quaternary using the long-chain alkenones contained in marine sediments. According to the finding of Marlowe et al. (1984), Brassell et al. (1986a,b) first defined the temperature-dependent unsaturation index of C_{37} alkenones, $U_{37}^{K'}$, as follow,

$$U_{37}^{K'} = (C_{37:2} - C_{37:4})/(C_{37:2} + C_{37:3} + C_{37:4}) \quad (1)$$

where $C_{37:x}$ denotes the concentration of alkenones consisting of 37 carbon atoms with x the number of double bonds. Because of the low value of $C_{37:4}$ encountered in marine sediments, Prahl and Wakeham (1987) suggested a simplified expression of Eq. (1),

$$U_{37}^{K'} = C_{37:2}/(C_{37:2} + C_{37:3}) \quad (2)$$

In 1986, Brassell et al. (1986a) found that in a sediment core from the Atlantic Ocean, $U_{37}^{K'}$ was parallel to the $\delta^{18}O$ of foraminifera. Later, culture of *E. huxleyi* indicated that for the temperature range between 8 and 25 °C, there was a linear positive correlation between $U_{37}^{K'}$ and ambient temperature (Prahl and Wakeham, 1987; Prahl et al., 1988). Thereafter this relationship was confirmed in various studies encompassing water column particulate organic matter and coretop sediments (Freeman and Wakeham, 1992; Conte and Eglinton, 1993; Prahl et al., 1993; Sikes and Volkman, 1993). More recent work carried out by Pelejero and Grimalt (1997) in the South China Sea and a revision of the $U_{37}^{K'}$-temperature calibration (Pelejero and Calvo, 2003) indicated that this relationship applied at higher temperature of up to 25–29 °C. Although Bentaleb et al. (2002) suggested that when SST was above 26.4 °C, $U_{37}^{K'}$ reached a constant value of unity (1.0) and could not be used for accurate SST reconstruction, Pelejero and Calvo (2003) argued that the authors probably erred in their arbitrary assignment of unity to $U_{37}^{K'}$ when the $C_{37:3}$ alkenone was not detected. Cultures of *G. oceanica* have provided different results, some suggesting a different $U_{37}^{K'}$-temperature relationship (Volkman et al., 1995), while some others indicating a similar relationship to what is observed in the culture of *E. huxleyi* (Conte et al., 1998). For the sedimentary record, however, a thorough comparative study indicated that the $U_{37}^{K'}$-SST relationship would not be affected by changes in Haptophyta species population (Müller et al., 1997).

Sub-samples for C_{37} alkenone analysis were taken from Core $Z_{14\text{-}6}$ with an average interval of 5 cm. In total, 113 sub-samples were obtained. According to the age model (Fig. 4), the average sample spacing is ~1 ka. The procedure used for unsaturated alkenone analysis refers to Villanueva et al. (1997). First, sub-samples were dried and manually ground for homogeneity. Then they were extracted three times with dichloromethane. The extracts were hydrolyzed with 6% potassium hydroxide in methanol to eliminate interference from alkenoates and wax esters. The hexane extracts from the alkaline solution were then fractionated by silica column chromatography to separate hydrocarbons and ketones from the polar fraction. After derivatization with bis (trimethylsilyl)trifluoroacetamide (BSTFA) the extracts were analyzed by gas chromatography (HP5890) with a flame ionization detector. All the analyses were conducted at the State Key Laboratory of Organic Geochemistry in Guangzhou Institute of Geochemistry.

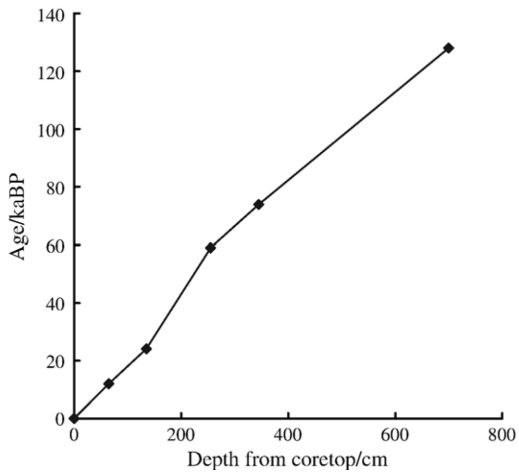

Fig. 4. Age versus depth plot for core $Z_{14\text{-}6}$ based on age control points listed in Table 1.

In all the sub-samples both the $C_{37:2}$ and $C_{37:3}$ alkenones were detected. $U_{37}^{K'}$ was calculated using Eq. (2) and SST was reconstructed using the following $U_{37}^{K'}$-SST relationship established by Müller et al. (1998),

$$U_{37}^{K'} = 0.033*SST + 0.044 \quad (3)$$

where SST represents seawater temperature at a depth between 0–30 m.

3. Results and discussions

The $U_{37}^{K'}$ variations of core $Z_{14\text{-}6}$ for the last glacial–interglacial cycle is shown in Fig. 5. Also shown in Fig. 5 are the $\delta^{18}O$ records of two planktonic foraminiferal species, *Neogloboquadrina dutertrei* and *Globigerinoides sacculifer* ($\delta^{18}O_{N.dut}$ and $\delta^{18}O_{G.sac}$) (Fig. 5a, b; Yan et al., 1990), and the relative abundances of *G. sacculifer*, *N. dutertrei* and *Pulleniatina obliquiloculata* (Fig. 5c–e; Yan and Thompson, 1991). *G. sacculifer* lives in surface water (Fairbanks et al., 1982). *N. dutertrei* and *P. obliquiloculata* are thermocline-dwellers (Ravelo et al., 1990). *P. obliquiloculata* is very abundant in the surface sediments of the ECS beneath the main axis of the Kuroshio Current (Wang et al., 1985) and thus is usually regarded as an indicator of the Kuroshio Current (Jian et al., 1996; Li et al., 1997; Jian et al., 1998; Ujiié and Ujiié, 1999; Wang et al., 2001a).

3.1. Variations of $U_{37}^{K'}$ and comparison with other proxies

The $U_{37}^{K'}$ of core $Z_{14\text{-}6}$ varied between 0.83 and 0.95 during the last glacial–interglacial cycle (Fig. 5f). The highest value (i.e. 0.95) occurred during the last interglacial stage MIS 5. Thereafter, the $U_{37}^{K'}$ fluctuated with an overall decreasing trend until the LGM (MIS 2) when the lowest value of 0.83 occurred. After the LGM, it increased to ca. 0.91. The $U_{37}^{K'}$ shows a trend generally parallel to the $\delta^{18}O_{G.sac}$ and $\delta^{18}O_{N.dut}$ records of the same core (Fig. 5a, b). This is verified by its covariations with the two $\delta^{18}O$ records (Fig. 6) and is consistent with results from Brassell et al. (1986a). $\delta^{18}O_{G.sac}$ displays a higher coherence with $U_{37}^{K'}$ than $\delta^{18}O_{N.dut}$ does (Fig. 6). This is because *N. dutertrei* lives in deeper water than *G. sacculifer* (Fairbanks et al., 1982; Andreasen and Ravelo, 1997) and $U_{37}^{K'}$ reflects annually averaged temperatures of the surface 0–30 m water column (Pelejero and Grimalt, 1997).

The abundances of *N. dutertrei*, *P. obliquiloculata* and *G. sacculifer* display different trends during the last glacial–interglacial cycle (Fig. 5c–e; Yan and Thompson, 1991). *P. obliquiloculata* is generally parallel to the $U_{37}^{K'}$ record, with the highest abundance (average ca. 10%) occurring in MIS 5.5 and the lowest in MIS 2 (~1%) (Fig. 5c, f). *G. sacculifer* varied mostly between 3–9%. The lowest abundance occurred in MIS 3 and thereafter increased significantly (Fig. 5d). Its general trend is different from those of $U_{37}^{K'}$ and *P. obliquiloculata*. The difference between *P. obliquiloculata* and *G. sacculifer* may be due to their different habitats and maybe they respond differently to climatic and hydrological changes (Jian et al., 2000). The former is a thermocline-dweller while the later is a mixed layer-dweller (Fairbanks et al., 1982; Ravelo et al., 1990). However, the other thermocline-dweller *N. dutertrei* (Ravelo et al., 1990) also displays a different overall trend compared with *P. obliquiloculata* (Fig. 5c, e). This again may be due to their different habitats. Although both *P. obliquiloculata* and *N. dutertrei* are thermocline-dwellers (Ravelo et al., 1990) and are regarded as representative species of the Kuroshio Current, the former has a much narrower temperature tolerance range (18–30 °C) than the later (10–30 °C) (Ijiri et al., 2005).

In addition to temperature, salinity is another factor possibly influencing the unsaturation index $U_{37}^{K'}$ (Blanz et al., 2005). However, this effect becomes notable only when the salinity decreases significantly and the percentage of $C_{37:4}$ shows a remarkable increase (5–10%, see Blanz et al., 2005 and references therein). No $C_{37:4}$ was identified in core $Z_{14\text{-}6}$. Thus we suggest that at the site of core $Z_{14\text{-}6}$, salinity variation since the LGM should have limited effect on the $U_{37}^{K'}$ record.

3.2. $U_{37}^{K'}$-derived SST

The SST reconstructed with $U_{37}^{K'}$ using Eq. (3) is presented in Fig. 7a. It ranges from ca. 24.0 to 27.5 °C, with the highest value (27.5 °C) occurring in MIS 5.5 and the lowest value (ca. 24.0 °C) in MIS 2 (the LGM). This trend is consistent with the marine record from the Equatorial Pacific (Lea et al., 2000) and the continental records from the East Asian monsoon domain (Thompson et al., 1997; Liu and Ding, 1998). For example, the loess–paleosol series in the Loess Plateau (Liu and Ding, 1998) and the ice core from Guliya Ice Cap on the Qinghai-Tibet Plateau (Thompson et al., 1997) indicate that the climate over the East Asia was warmest in MIS 5 and coldest in MIS 2 (the LGM). The $U_{37}^{K'}$-derived SST displays a strong coherence with the $\delta^{18}O_{G.sac}$ and $\delta^{18}O_{N.dut}$ records (Figs. 5a, b; 7a; Yan et al., 1990). This suggests that in the study area, SST has changed in phase with global ice volume change during the last glacial–interglacial cycle. The average SSTs are 25.2,

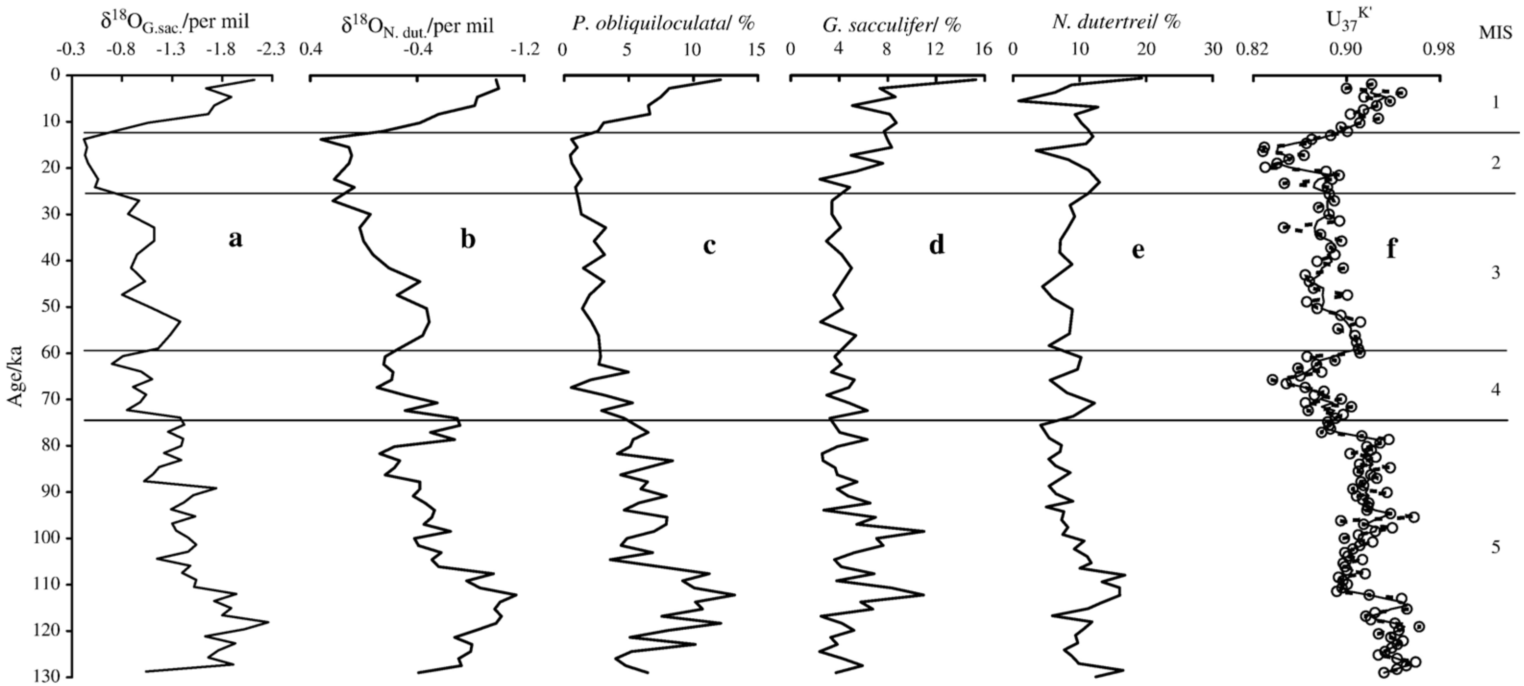

Fig. 5. The calculated $U_{37}^{K'}$ from core $Z_{14\text{-}6}$ since the last interglacial (f) and comparison with the $\delta^{18}O_{N.dut}$ (a) and $\delta^{18}O_{G.sac}$ (b) records, the relative abundances of *P. obliquiloculata* (c), *G. sacculifer* (d) and *N. dutertrei* (e). In (f), the thin line represents a three points running average.

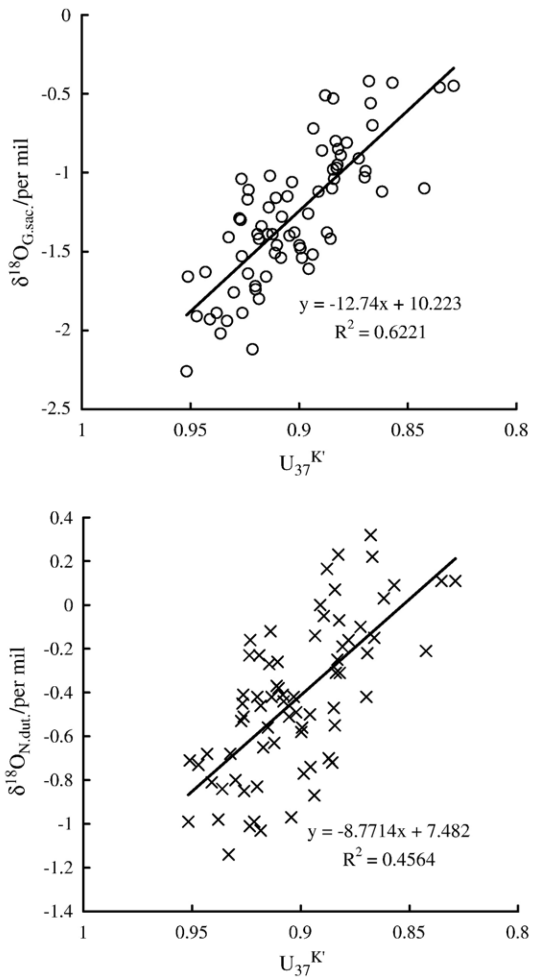

Fig. 6. Correlations between $U_{37}^{K'}$ and the (a) $\delta^{18}O_{N.dut}$ and (b) $\delta^{18}O_{G.sac}$ records.

25.5 and 26.4 °C for MIS 4, 3 and 1, respectively. The $U_{37}^{K'}$-derived SST was not obtained for the latest 1.5 ka because the coretop sediment was not available. The average SST for MIS 1 (26.4 °C) is higher than the modern SST at the site of core $Z_{14\text{-}6}$ (Fig. 2), but falls between the $U_{37}^{K'}$-derived SSTs for MIS 1 reconstructed from core MD982195 to the north (Ijiri et al., 2005) and core 1202B to the south (Zhao et al., 2005) (see Fig. 1 for the locations of the three cores), respectively. It is also consistent with the seasonal SSTs reconstructed in this region using the foraminiferal transfer function (Jian et al., 2000). Therefore, it should be acceptable when taking into consideration of the error in U-SST calibration (Villanueva and Grimalt, 1996; Villanueva et al., 1997; Bentaleb et al., 2002) and the seasonal bias of $U_{37}^{K'}$-derived SST (Sawada et al., 1998).

Based on the foraminiferer assemblages of core $Z_{14\text{-}6}$, Yan and Thompson (1991) reconstructed the annual and seasonal SSTs using the transfer function FP-12E (Thompson, 1981) (Fig. 7b). These SSTs are generally parallel to the $U_{37}^{K'}$-derived SST, but have different ranges. Since the last interglacial stage, the annual SST fluctuated between 18 °C and 24 °C. The winter SST varied between 13 °C and 21 °C while the summer SST had a much narrower range of 26 °C–28.6 °C. The $U_{37}^{K'}$-derived SST falls between the winter and summer SSTs. However, it is apparently higher than the annual SST. This may be related to: (1) errors in SST estimation. The errors in SST estimation using the transfer function FP-12E is ±1.46 °C for summer and ±2.48 °C for winter (Thompson, 1981), and the error in $U_{37}^{K'}$-derived SST estimation is ±0.5 °C or more (Villanueva and Grimalt, 1996; Villanueva et al., 1997; Bentaleb et al., 2002). These errors are big and make the difference insignificant between the $U_{37}^{K'}$-derived SST and the annual SST obtained by Yan and Thompson (1991); (2) bias of the $U_{37}^{K'}$-derived SST because of the seasonal variation of alkenone production (Sawada et al., 1998). Investigation on trapped sediments in the northwestern North Pacific off Japan revealed that a pronounced maximum in alkyl alkenoate fluxes occurs in late spring to summer (Sawada et al., 1998). In the ECS, the higher flux of *E. huxleyi* occurs in spring to summer (Tanaka, 2003) in the euphotic zone, which has a depth of 60–65 m (Gong et al., 2000). This may lead to an overestimation of the annual mean SST using the $U_{37}^{K'}$-SST calibration; (3) biases of the SST reconstructed using the transfer function FP-12E because of the dissolution of foraminifera (Thunell et al., 1994). However, the water depth of core $Z_{14\text{-}6}$ is 739 m and is smaller than the carbonate compensation depth in the northwestern Pacific and its marginal seas (Kato et al., 1983; Yamamoto et al., 2003). Thus the effect of dissolution should be limited. The annual SST for MIS 1 obtained by Yan and Thompson (1991) (Fig. 7b) is ca. 1 °C lower than the modern annual SST at the site of core $Z_{14\text{-}6}$ (Fig. 2).

3.3. SST variations since the Last Glacial Maximum

The $U_{37}^{K'}$-derived SST for core $Z_{14\text{-}6}$ shows an abrupt and fast increase since 15–16 ka BP (Fig. 7a), which is in accordance with the timing of the last deglaciation recorded by the Arctic ice core (Mayewski et al., 1994), the loess–paleosol series (Xiao et al., 1995) and the stalagmite from Hulu Cave, East China (Wang et al., 2001b). This is also consistent with the $U_{37}^{K'}$-derived SST for core MD982195 (Ijiri et al., 2005) in the study area. Based on the foraminiferer assemblages of core DGKS9603, Li et al. (2001) suggested that the core site had been increasingly influenced by the Kuroshio

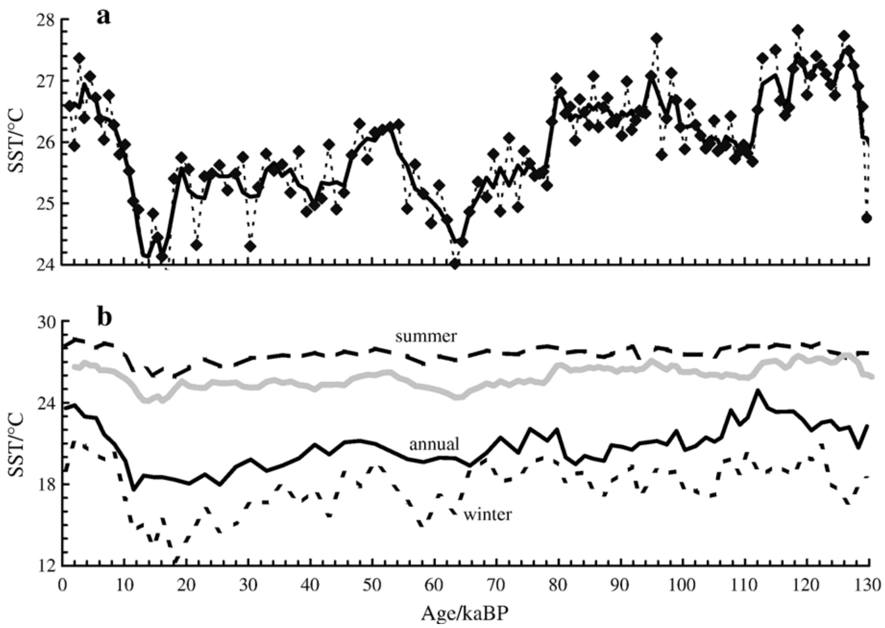

Fig. 7. (a) The $U^{K'}_{37}$-derived SST from core $Z_{14\text{-}6}$ since the last interglacial and comparison with (b) the summer, winter and annual SSTs reconstructed by Yan and Thompson (1991) using the foraminiferal transfer function FP-12E (Thompson, 1981). In (a), a three points running average is represented by a thin line. It is also shown in (b) as a thick grey line.

Current since ca. 16 cal ka BP. The $U^{K'}_{37}$-derived SST for core 1202B, however, indicate a significant increase since ca.13 ka BP (Zhao et al., 2005), about 2 ka later than indicated by other records. Zhao et al. (2005) suggested that this might be related with the uncertainties of their age model. Therefore, we suggest that in the study area, the last deglaciation started at ~15 ka BP, in phase with the global ice volume change, but later than the last deglacial warming of the Equatorial Pacific (Visser et al., 2003). A latitudinal shift in the deglaciation timing was also observed in the eastern Pacific (Lea et al., 2000; Herbert et al., 2001).

Most recently, Schaefer et al. (2006) found that cosmogenic ^{10}Be exposure dates indicated near-synchronous interhemispheric termination of the LGM between 19 and 17 ka BP in mid-latitudes, which is consistent with the onset of temperature and atmospheric CO_2 increases in Antarctic ice cores (Blunier and Brook, 2001; Monnin et al., 2001), but unambiguously earlier than the dramatic warming event recorded by Greenland ice core at the onset of the Bölling, which occurred at 14.7 ka (Severinghaus and Brook, 1999). They interpreted this as a reflection of a global summer temperature warming. However, in the Northern Hemisphere, destabilization of ice sheets lead to a near-shutdown of the meridional overturning circulation and, in turn, to a substantial spread of North Atlantic sea winter ice and hypercold winters between about 17.5 and 14.7 ka BP. This explanation seems not consistent with the speleothem δ^{18}O records from East Asia, which indicated that the East Asian summer monsoon was weakest during the 17.4–14.7 ka BP period and strengthened abruptly thereafter (Wang et al., 2001b; Yuan et al., 2004). Maybe there is a different mechanism in the Asian monsoon domain to end the LGM and start the last deglaciation. Further study is needed to check whether this is true.

Climatic and environmental change since the LGM is important because it displays the latest turnover of the climate and environment on the earth (Jouzel et al., 1987; Martinson et al., 1987; Thompson et al., 1997; Liu and Ding, 1998). Previous SST reconstruction by the CLIMAP Project Members (1976) indicated that SST has increased by about 2 °C over most of the middle latitude Pacific Ocean since the LGM; while for the subtropical region, SST in the LGM was similar to modern SST. With simultaneous application of a variety of transfer function techniques, Kucera et al. (2005) reconstructed most recently the SSTs of the Atlantic and Pacific Oceans using assemblages of planktonic foraminifera. They found that at some sites close to core $Z_{14\text{-}6}$, the LGM SSTs were more than 1 °C higher relative to present (their Fig. 16 (b), 17 (a) and (b)). However, this is not supported by most other SST reconstructions in the region using transfer function or $U^{K'}_{37}$-SST calibration (Yan and Thompson, 1991; Ijiri et al., 2005; Zhao et al., 2005; this work), and is not consistent with the weakening of the Kuroshio Current and cooling in its source region in the LGM (Sawada and Handa, 1998). This is probably due to the paucity of calibration samples from subtropical sediments (Kucera et al., 2005).

In core $Z_{14\text{-}6}$, the $U_{37}^{K'}$-derived SST increased ca. 2.4 °C from the LGM to the Holocene (Fig. 7a). The SSTs reconstructed by Yan and Thompson (1991) using transfer function FP-12E (Fig. 7b) display different deglacial increases. The annual SST increased ∼5 °C from the LGM to the Holocene, the summer and winter SSTs 1.6 °C and 6–8 °C, respectively. The $U_{37}^{K'}$-derived SST shows a deglacial increase of ∼2.4 °C (Fig. 7a), falling between 1.6 and (6–8) °C and apparently less than ∼5 °C. However, the $U_{37}^{K'}$-derived SST and the annual SST obtained by Yan and Thompson (1991) (Fig. 7a, b) are not conflicting considering their uncertainties and biases (Thompson, 1981; Villanueva and Grimalt, 1996; Villanueva et al., 1997; Sawada et al., 1998; Bentaleb et al., 2002). The last deglacial increase in SST indicated by the $U_{37}^{K'}$ proxy of core $Z_{14\text{-}6}$ is higher than what was suggested by the CLIMAP Project Members (1976). This may be caused partly by inherent difference between SST reconstructions using $U_{37}^{K'}$-SST calibration and transfer function FP-12E, and also by the difference in hydrography. It is found that the SST increase from the LGM to the Holocene (1) is relatively small for open ocean compared with marginal sea, and (2) is usually greater for closed marginal sea than for semi-closed marginal sea. For example, SST increase from the LGM to the Holocene is 1.3 °C and 1.8 °C, respectively, for the East Equatorial Pacific Ocean and the Equatorial Atlantic Ocean (Prahl et al., 1989; Sikes and Keigwin, 1994), while for the South China Sea, a closed marginal sea, the rise is 2.8 °C (Pelejero et al., 1999). This value is greater than its counterpart in the Okinawa Trough in the ECS, a semi-closed marginal sea (∼2.4 °C, this work).

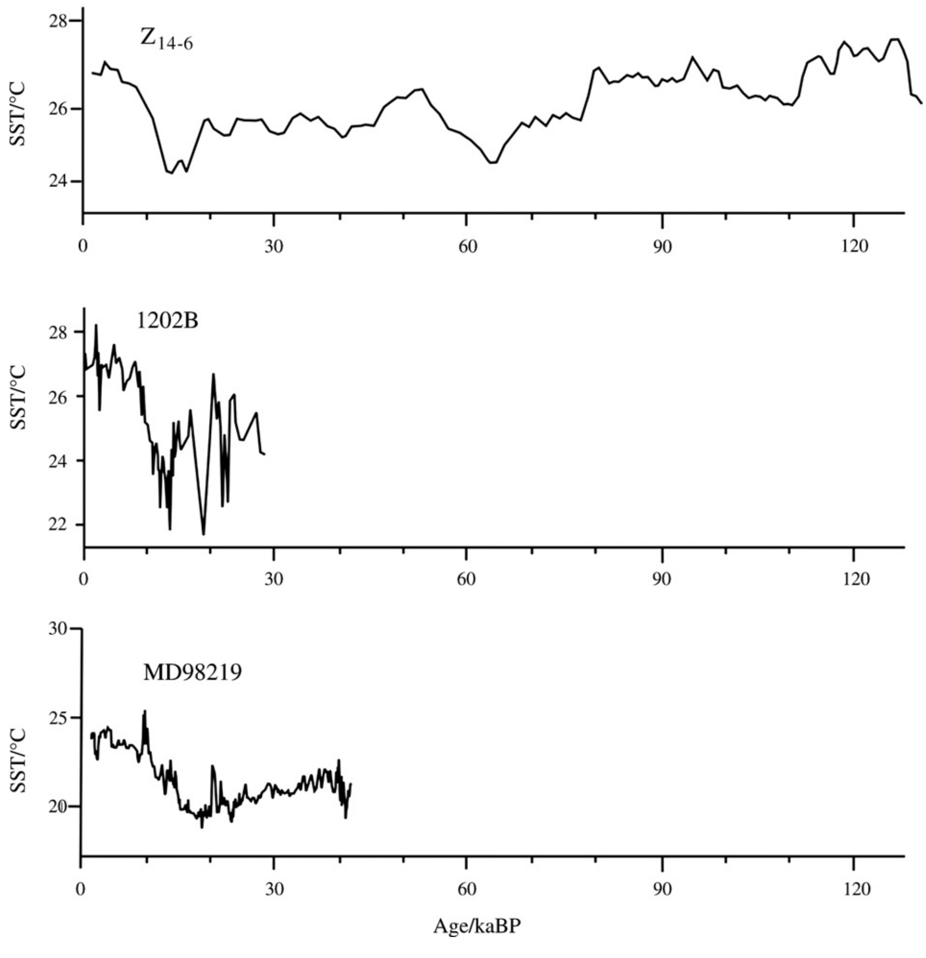

Fig. 8. Comparison of the $U_{37}^{K'}$-derived SST from core $Z_{14\text{-}6}$ with those from cores 1202B (Zhao et al., 2005) and MD982195 (Ijiri et al., 2005). See Fig. 1 for the location of the three cores.

The hydrographical effect on SST variation is similar to that on the $\delta^{18}O$ variation of planktonic foraminifera (Wang, 1999; Li et al., 2001).

In 2005, two $U_{37}^{K'}$-derived SST records were established in the south and north of the Okinawa Trough, respectively, both showing significantly greater deglacial SST increase of ~5 °C (Fig. 8b, c; Ijiri et al., 2005; Zhao et al., 2005). Core 1202B was drilled offshore of NE Taiwan in the south of the Okinawa Tough (see Fig. 1 for its location). The greater deglacial SST increase recorded by this core is related to the enhanced glacial cooling at this site, which in turn may be related with its proximity to Taiwan, the variability of the Kuroshio Current, sea level variations, and coastline shift (Zhao et al., 2005). The larger deglacial SST increase indicated by core MD982195 in the north of the Okinawa Trough (Ijiri et al., 2005) may be related to (1) its proximity to the land mass of Japan, (2) higher sensitivity of SST in the north of the Okinawa Trough compared to variations of the Kuroshio Current, sea level and coastline, and (3) greater temperature change at higher latitude relative to lower latitude.

As shown in Fig. 9, the $U_{37}^{K'}$-derived SST increased since the last deglaciation until ca. 10 ka BP when it reached ca. 26.4 °C. Then it kept stable for 2.5 ka. From ca. 7.4 ka BP, it increased again. However, this increase is relatively small (~0.6 °C). At ca. 4.6 ka BP, the $U_{37}^{K'}$-derived SST showed a slight decrease. Its post-glacial evolution is consistent with the paleoclimatic and paleohydrological records for other cores in this region. Core E017 retrieved in the southern Okinawa Trough indicated that at 10.1–9.2 cal. ka BP, the influence of the Kuroshio Current increased significantly (Xiang et al., 2003). Xu and Oda (1999) suggested that two cores in the Okinawa Trough, KH82-4-14 and RN80-PC3, showed that conditions there have been controlled by the modern open-sea water since 10.5–8.5 ka BP. Some cores in northwestern Pacific Ocean off Japan showed that the influence of the warm Kuroshio Current peaked at around 7–6 ka BP (Chinzei et al., 1987). Sawada and Handa (1998) also reported a remarkable strengthening of the Kuroshio Current at ~7.1 cal. Ka BP in the northwestern Pacific off central Japan. In the study on core OR281-12P in the south of the Okinawa Trough, Shieh et al. (1997) indicated that at ca. 7.5 ka BP, the $\delta^{18}O_{N.dut}$ became much lighter and might reflect the influence of warm water mass after that time. The investigation carried out by Jian et al. (2000) on cores 255 and B-3GC in the south and north of the Okinawa Trough, respectively, also revealed that at ~7.5–7.3 ka BP, the winter SST and the abundance of the indicator of the Kuroshio Current, *P. obliquiloculata*, showed remarkable increases. During the period from 5.3–4 to 3–2 ka BP, the abundance of *P. obliquiloculata* in many cores in the ECS decreased abruptly and significantly (Jian et al., 1996; Li et al., 1997; Jian et al., 1998; Ujiié and Ujiié, 1999; Jian et al., 2000; Li et al., 2001; Wang et al., 2001a). This is called the *P. obliquiloculata* minimum (PM) event. However, the decrease of the $U_{37}^{K'}$-derived SST at ca. 4.6 ka BP (Fig. 7a) doesn't match that of *P. obliquiloculata* abundance in magnitude (Jian et al., 1996; Li et al., 1997; Jian et al., 1998; Ujiié and Ujiié, 1999; Jian et al., 2000; Li et al., 2001; Wang et al., 2001a).

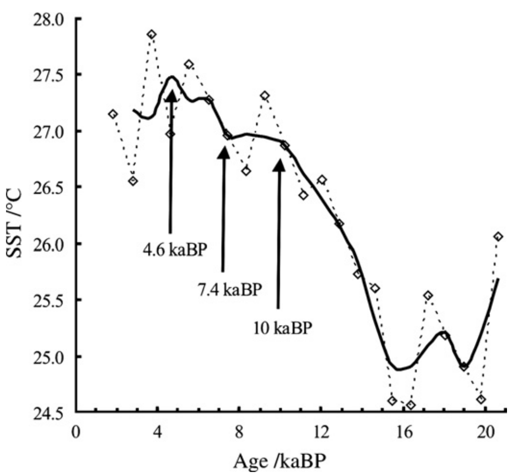

Fig. 9. Enlargement of Fig. 7a for the period since 20 ka BP. It can be seen that SST displays an abrupt and fast increase from ca. 15–16 ka BP until ca. 10 ka BP when it became relatively stable. Since ca. 7.4 ka BP SST increased again. Decrease of SST from ca. 4.6 ka BP can also be observed.

In the west Pacific, some important events are not yet thoroughly addressed. These events include the Younger Dryas (YD) cooling event that happened in the last deglacial period (Chinzei et al., 1987; Thunell and Miao, 1996; Shieh et al., 1997; Xu and Oda, 1999; Wang, 1999; Ujiié et al., 2001; Li et al., 2001; Ijiri et al., 2005) and the PM event that occurred in the late Holocene (Jian et al., 1996; Li et al., 1997; Ujiié and Ujiié, 1999). In core $Z_{14\text{-}6}$, the YD event was not recorded by the $U_{37}^{K'}$-derived SST, nor by the SSTs reconstructed using transfer function (Fig. 7; Yan and Thompson, 1991). No reversal was observed in the deglacial decrease in the $\delta^{18}O_{N.dut}$ and $\delta^{18}O_{G.sac}$ records (Fig. 3a, b; Yan et al., 1990). Likewise, the PM event is not clear in the SST and $\delta^{18}O$ records (Figs. 3, 7). Especially, the *P. obliquiloculata* abundance doesn't show a decrease (Fig. 5c) in the PM or YD event. However, due to the low resolution of the records from core $Z_{14\text{-}6}$, it is not appropriate to conclude definitely that the YD and PM events are not recorded in the middle Okinawa Trough.

4. Conclusions

(1) The $U_{37}^{K'}$ of core $Z_{14\text{-}6}$ fluctuated between 0.83 and 0.95 during the last glacial–interglacial cycle, with the highest value (i.e. 0.95) occurring in MIS 5 and the lowest (i.e. 0.83) in the LGM (MIS 2). It is generally parallel to the $\delta^{18}O$ records of *N. dutertrei* and *G. sacculifer* of the same core, indicating that in this region, the SST varied in phase with global ice volume change.

(2) The $U_{37}^{K'}$-derived SST recorded by core $Z_{14\text{-}6}$ ranges from ca. 24.0 to 27.5 °C, with the highest SST (27.5 °C) occurring in MIS 5.5 and the lowest (~24.0 °C) in MIS 2. This trend is consistent with the continental records from the East Asian monsoon domain and the marine record from the Equatorial Pacific.

(3) The onset of the last deglaciation in the middle Okinawa Trough is ~15 ka BP, in phase with global ice volume change but later than the last deglacial warming in the Equatorial Pacific.

(4) The $U_{37}^{K'}$-derived SST for core $Z_{14\text{-}6}$ displays an increase of ~2.4 °C from the LGM to the Holocene. This increase is smaller than the deglacial SST increases recorded by cores MD982195 and 1202B in the north and south of the Okinawa Trough, respectively. It is greater than what is observed in open-sea sediments, but is smaller than what is obtained in closed marginal seas, suggesting the effect of hydrography on SST variation.

(5) The Younger Dryas and *P. obliquiloculata* minimum events are not recognized in core $Z_{14\text{-}6}$, but it is not appropriate to conclude definitely that the YD and PM events are not recorded in the middle Okinawa Trough due to the low resolution of the records for core $Z_{14\text{-}6}$.

Acknowledgements

We sincerely thank C. Pelejero and an anonymous reviewer for their thorough reviews and useful comments on earlier versions of this paper, and the editor in chief (TC) for his comments and editing of this paper. We are grateful to Dr. Chang Feng-ming and Dr. Zhao Jing-tao for their help in sampling. Help from Ms. Yu Chi-ling while analyzing unsaturated alkenones at the State Key Laboratory of Organic Geochemistry is appreciated. Thanks go to Prof. Shi Ji-yang, Dr. Peng Xian-zhi and Dr. Hu Jian-fang for their helpful suggestions for this work. This paper is financially supported by the CAS Pilot Project (No. KZCX3-SW-220, KZCX3-SW-152), the State Key Laboratory of Organic Geochemistry (No. OGL-200308), the State Key Laboratory of Loess and Quaternary Geology (No. SKLLQG0408) and the Key Laboratory of Marginal Sea Geology, CAS (No. GIGCX-04-01).

References

Andreasen, D.J., Ravelo, A.C., 1997. Tropical Pacific Ocean thermocline depth reconstructions for the Last Glacial Maximum. Paleoceanography 12, 395–413.

Bentaleb, I., Fontugne, M., Beaufort, L., 2002. Long-chain alkenones and U_{37}^{k} (variability along a south–north transect in the western Pacific Ocean. Global and Planetary Change 34, 173–183.

Blanz, T., Emeis, K.-C., Siegel, H., 2005. Controls on alkenone unsaturation ratios along the salinity gradient between the open ocean and the Baltic Sea. Geochimica Cosmochimica et Acta 69, 3589–3600.

Blunier, T., Brook, E.J., 2001. Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. Science 291, 109–112.

Brassell, S.C., Eglimton, G., Marloweetal, I.T., 1986a. Molecular stratigraphy: a new tool for climatic assessment. Nature 320, 129–133.

Brassell, S.C., et al., 1986b. Palaeoclimatic signals recognized by chemometric treatment of molecular stratigraphic data. Organic Geochemistry 10, 649–660.

Chinzei, K., Fujioka, K., Kitazato, H., Koizumi, I., Oba, T., Oda, M., Okada, H., Sakai, T., Tanlmura, R., 1987. Postglacial environmental change of the Pacific Ocean off the coasts of central Japan. Marine Micropaleontology 11, 273–291.

CLIMAP Project Members, 1976. The surface of the ice-age earth. Science 191, 1131–1137.

Conte, M.H., Eglinton, G., 1993. Alkenone and alkenoate distributions within the euphotic zone of eastern North Atlantic: correlation with production temperature. Deep-Sea Research 40, 1935–1961.

Conte, M.H., Eglinton, G., Madureira, L.A., 1992. Long-chain alkenones and alkyl alkenoates as paleotemperature indicators: their production, flux and early diagenesis in the eastern North Atlantic. In: Eckardt, C.B., Larter, S.R., et al. (Eds.), Advances in Organic Geochemistry 1991. Organic Geochemistry, vol. 19, pp. 287–298.

Conte, M.H., Thompson, A., Lesley, D., Harris, R.P., 1998. Genetic and physiological influences on the alkenone/alkenoate versus growth temperature relationship in *Emiliania huxleyi* and *Gephyrocapsa oceanica*. Geochimica Cosmochimica et Acta 62, 51–68.

Fairbanks, R.G., Sverdlove, M., Free, R., Wiebe, P., Bé, W., 1982. Vertical distribution and isotopic fractionation of living planktonic foraminifera from the Panama Basin. Nature 298, 841–844.

Freeman, K.H., Wakeham, S.G., 1992. Variations in the distributions and isotopic compositions of alkenones in Black Sea particles and sediments. Organic Geochemistry 19, 277–285.

Gong, G.-C., Shiah, F.-K., Liu, K.-K., Wen, Y.-H., Liang, M.-H., 2000. Spatial and temporal variation of chlorophyll a, primary productivity and chemical hydrography in the southern East China Sea. Continental Shelf Research 20, 411–436.

Herbert, T., Schuffert, J., Andreasen, D., Heusser, L., Lyle, M., Mix, A., Ravelo, A.C., Stott, L.D., Herguera, J.C., 2001. Collapse of the California Current during glacial maxima linked to climate change on land. Science 293, 71–76.

Ijiri, A., Wang, L., Oba, T., Kawahata, H., Huang, C.-Y., Huang, C.-Y., 2005. Paleoenvironmental changes in the northern area of the East China Sea during the past 42,000 years. Palaeogeography, Palaeoclimatology, Palaeoecology 219, 239–261.

Jian, Z., Li, B., Pflaumann, U., Wang, P., 1996. Late Holocene cooling event in the western Pacific. Science in China, Series D: Earth Sciences 39 (5), 543–550.

Jian, Z., Saito, Y., Wang, P., Li, B., Chen, R., 1998. Shifts of the Kuroshio axis over the last 20,000 years. Chinese Science Bulletin 43, 1053–1056.

Jian, Z., Wang, P., Saito, Y., Wang, J., Pflaumann, U., Oba, T., Cheng, X., 2000. Holocene variability of the Kuroshio Current in the Okinawa Trough, northwestern Pacific Ocean. Earth and Planetary Science Letters 184 (1), 305–319.

Jouzel, J., Lorius, C., Petit, J.R., et al., 1987. Vostok ice core: a continuous isotope temperature record over the last climatic cycle (160,000 years). Nature 329, 403–408.

Kato, T., Endo, M., Kato, M., 1983. Vertical distribution of various elements in sediment cores from the Japan Sea. Marine Geology 53, 277–290.

Kobayashi, K., 1985. Sea of Japan and Okinawa Trough. In: Nairn, A.E.M., Stehli, F.G., Uyeda, S. (Eds.), The Ocean Basins and Margins. The Pacific Ocean, vol. 7A, pp. 419–459.

Kucera, M., Weinelt, M., Kiefer, T., et al., 2005. Reconstruction of sea-surface temperatures from assemblages of planktonic foraminifera: multi-technique approach based on geographically constrained calibration data sets and its application to glacial Atlantic and Pacific Oceans. Quaternary Science Reviews 24, 951–998.

Lambeck, K., Chappell, J., 2001. Sea level change through the last glacial cycle. Science 292, 679–686.

Lea, D.W., Pak, D.K., Spero, H.J., 2000. Climate impact of late Quaternary Equatorial Pacific sea surface temperature variations. Science 289, 1719–1724.

Lee, C.-S., Shor Jr., G.G., Bibee, L.D., Lu, R.S., Hilde, T.W.C., 1980. Okinawa Trough: origin of a back-arc basin. Marine Geology 35, 219–241.

Letouzey, J., Kimura, M., 1986. The Okinawa Trough — genesis of a back-arc basin developing along a continental-margin. Tectonophysics 125 (1–3), 209–230.

Li, K., 1993. Climate of the China Seas and Northwestern Pacific. China Ocean Press, Beijing. (in Chinese).

Li, B., Jian, Z., Wang, P., 1997. *Pulleniatina obliquiloculata* as a paleoceanographic indicator in the southern Okinawa Trough during the last 20,000 years. Marine Micropaleontology 32, 59–69.

Li, T.G., Liu, Z.X., Hall, M.A., Berne, S., Saito, Y., Cang, S.X., Cheng, Z.B., 2001. Heinrich event imprints in the Okinawa Trough: evidence from oxygen isotope and planktonic foraminifera. Palaeogeography, Palaeoclimatology, Palaeoecology 176 (1), 133–146.

Liu, T.S., Ding, Z.L., 1998. Chinese loess and the paleomonsoon. Annual Review of Earth and Planetary Sciences 26, 111–145.

Liu, Z.X., Saito, Y., Li, T.G., Berne, S., Cheng, Z., Li, P.Y., Li, Z., Guichard, F., Floch, G., 1999. Millennial-scale paleoceanography in Okinawa Trough during late Quaternary period. Chinese Science Bulletin 44 (18), 1705–1709.

Liu, Y.G., Fu, Y.X., Du, D.W., Meng, X.W., Liang, R.C., Li, T.G., Wu, S.Y., 2003. Paleoceanographic records in the sedimentary cores from the middle Okinawa Trough. Chinese Science Bulletin 48, 74–81 (supp.).

Madureira, L.A.S., Conte, M.H., Eglinton, G., 1995. Early diagenesis of lipid biomarker compounds in North Atlantic sediments. Paleoceanography 10, 627–642.

Marlowe, I.T., Green, J.C., Neal, A.C., Brassell, S.C., Eglinton, G., Course, P.A., 1984. Long-chain (n-C_{37}–C_{39}) alkenones in the Prymnesiophyceae. Distribution of alkenones and other lipids and their taxonomic significance. British Phycological Journal 19, 203–216.

Martinson, D.G., Pisias, N.G., Hays, J.D., Imbrie, J., Moore Jr., T.C., Shackletion, N.J., 1987. Age dating and the orbital theory of the ice ages: development of a high-resolution 0 to 300,000-year chronostratigraphy. Quaternary Research 27, 1–29.

Mayewski, P.A., Meeker, L.D., Whitlow, S.I., Twickler, M.S., Morrison, M.C., Bloomfield, P., Bond, G.C., Alley, R.B., Gow, A.J., Grootes, P.M., Meese, D.A., Ram, M., Taylor, K.C., Wumkes, M.A., 1994. Changes in atmospheric circulation and ocean ice cover over the North Atlantic during the last 41,000 years. Science 263, 1747–1751.

McCafferey, M.A., Farrington, J.W., Repeta, D.J., 1990. The organic geochemistry of Peru margin surface sediments, I. A comparison of the C_{37} alkenone and historical El Nino records. Geochimica Cosmochimica et Acta 54, 1671–1682.

Monnin, E., Indermühle, A., Dällenbach, A., Flückiger, J., Stauffer, B., Stocker, T.F., Raynaud, D., Barnola, J.-M., 2001. Atmospheric CO2 concentrations over the last glacial termination. Science 291, 112–114.

Müller, P.J., Cepek, M., Ruhland, G., Schneider, R., 1997. Sediments from the Walvis Ridge: implications for the alkenone paleotemperature method. Palaeogeography, Palaeoclimatology, Palaeoecology 135, 71–96.

Müller, P.J., Kirst, G., Ruhland, G., von Storch, I., Rosell-Mele, A., 1998. Calibration of the alkenone paleotemperature index $U_{37}^{K'}$ based on core-tops from the eastern South Atlantic and the global ocean (60° N–60° S). Geochimica Cosmochimica et Acta 62, 1757–1772.

Pelejero, C., Calvo, E., 2003. The upper end of the U37K′ temperature calibration revisited. Geochemica Geophysics Geosystem 4, 1014, doi:10.1029/2002GC000431.

Pelejero, C., Grimalt, J.O., 1997. The correlation between the U37K′ index and sea surface temperatures in the warm boundary: the South China Sea. Geochimica Cosmochimica et Acta 61, 4789–4797.

Pelejero, C., Grimalt, J.O., Sarnthein, M., Kienast, M., Wang, L.J., 1999. High-resolution U_{37}^{K} temperature reconstructions in the South China Sea over the past 220 kyr. Paleoceanography 14, 224–231.

Prahl, F.G., Wakeham, S.G., 1987. Calibration of unsaturation patterns in long-chain ketone composition for paleotemperature assessment. Nature 330, 367–369.

Prahl, F.G., Muelhansen, L.A., Zahnle, D.A., 1988. Further evaluation of long-chain alkenones as indicators of paleoceanographic conditions. Geochimica Cosmochimica et Acta 52, 2303–2310.

Prahl, F.G., Muehlhausen, L.A., Lyle, M., 1989. An organic geochemical assessment of oceanographic conditions at MANOP Site C over the past 26,000 years. Paleoceanography 4, 495–510.

Prahl, F.G., Collier, R.B., Dymond, J., Lyle, M., Sparrow, M.A., 1993. A biomarker perspective on prymnesiophyte productivity in the northeast Pacific Ocean. Deep-Sea Research 40, 2061–2076.

Ravelo, A.C., Fairbanks, R.G., Philander, S.G.H., 1990. Reconstructing tropical Atlantic hydrography using planktonic foraminifera and an ocean model. Paleoceanography 5, 409–431.

Sawada, K., Handa, N., 1998. Variability of the path of the Kuroshio ocean current over the past 25,000 years. Nature 392, 592–595.

Sawada, K., Handa, N., Shiraiwa, Y., Danbara, A., Montani, S., 1996. Long-chain alkenones and alkyl alkenoates in the coastal and

pelagic sediments of the northwest North Pacific with special reference to the reconstruction of *Emiliania huxleyi* and *Gephyrocapsa oceanica* ratios. Organic Geochemistry 24, 751–764.
Sawada, K., Handa, N., Nakatsuka, T., 1998. Production and transport of long-chain alkenones and alkyl alkenoates in a sea water column in the northwestern Pacific off central Japan. Marine Chemistry 59, 219–234.
Schaefer, J.M., Denton, George H., Barrell, David J.A., Ivy-Ochs, Susan, Kubik, Peter W., Andersen, Bjorn G., Phillips, Fred M., Lowell, Thomas V., Schluchter, Christian, 2006. Near-synchronous interhemispheric termination of the Last Glacial Maximum in mid-latitudes. Science 312, 1510–1513.
Severinghaus, J.P., Brook, E.J., 1999. Abrupt climate change at the end of the last glacial period inferred from trapped air in Polar Ice. Science 286, 930–934.
Shieh, Y.-T., Wang, C.-H., Chen, M.-P., Yung, Y.-L., 1997. The Last Glacial Maximum to Holocene environment changes in the southern Okinawa Trough. Journal of Asian Earth Sciences 15, 3–8.
Sikes, E.L., Keigwin, L.D., 1994. Equatorial Atlantic sea surface temperature for the last 30 kyr: a composition of Uk-37, $\delta^{18}O$ and foraminiferal assemblate temperature estimates. Paleoceanography 9, 31–45.
Sikes, E.L., Volkman, J.K., 1993. Calibration of alkenone unsaturation ratios ($U^{k}{}_{37}$) for paleotemperature estimation in cold polar waters. Geochimica et Cosmochimica Acta 57, 1883–1889.
Tanaka, Y., 2003. Coccolith flux and species assemblages at the shelf edge and in the Okinawa Trough of the East China Sea. Deep-Sea Research. Part 2. Topical Studies in Oceanography 50, 503–511.
Thompson, P.R., 1981. Planktonic foraminifera in the western North Pacific during the past 150,000 year, comparison of modern fossil assemblages. Palaeogeography, Palaeoclimatology, Palaeoecology 35, 241–279.
Thompson, P.R., Be, A.W.H., Duplessy, Shacklton, N.J., 1979. Disappearance of pink-pigmented *Globigerinoides rubber* at 120,000 years BP in Indian and Pacific oceans. Nature 280, 554–558.
Thompson, L.G., Yao, T.D., Davis, M.E., Henderson, K.A., Mosley-Thompson, E., Lin, P.-N., Beer, J., Synal, H.-A., Cole-Dai, J., Bolzan, J.F., 1997. Tropical climate instability: the last glacial cycle from a Qinghai-Tibetan ice core. Science 276, 1821–1825.
Thunell, R.C., Miao, Q., 1996. Sea surface temperature of the western Equatorial Pacific Ocean during the Younger Dryas. Quaternary Research 46, 72–77.
Thunell, R.C., Anderson, D., Gellar, D., Miao, Q., 1994. Sea surface temperature estimates for the tropical western Pacific during the last glaciation and their implications for the Pacific warm pool. Quaternary Research 41, 255–264.
Ujiié, H., Ujiié, Y., 1999. Late Quaternary course changes of the Kuroshio Current in the Ryukyu Arc region, northwestern Pacific Ocean. Marine Micropaleontology 37, 23–40.
Ujiié, H., Hatakeyama, Y., Gu, X.X., Yamamoto, S., Ishiwatari, R., Maeda, L., 2001. Upward decrease of organic C/N ratios in the Okinawa Trough cores: proxy for tracing the post-glacial retreat of the continental shore line. Palaeogeography, Palaeoclimatology, Palaeoecology 165, 129–140.
Villanueva, J., Grimalt, J.O., 1996. Pitfalls in the chromatographic determination of the alkenone U^{k}_{37} index for paleotemperature estimation. Journal of Chromatography, A 723, 285–291.
Villanueva, J., Pelejero, C., Grimalt, J., 1997. Clean-up procedures for the unbiased estimation of C_{37}–C_{39} alkenones sea surface temperatures and terrigenous *n*-alkane inputs in paleoceanography. Journal of Chromatography, A 757, 145–151.
Visser, K., Thunell, R., Stott, L., 2003. Magnitude and timing of temperature change in the Indo-Pacific warm pool during deglaciation. Nature 321, 152–155.
Volkman, J.K., Barrett, S.M., Blackburn, S.I., Sikes, E.L., 1995. Alkenones in *Geophyrocapsa oceanica*: implications for studies of paleoclimate. Geochimica Cosmochimica et Acta 59, 513–520.
Wang, P.X., 1999. Response of Western Pacific marginal seas to glacial cycles: paleoceanographic and sedimentological features. Marine Geology 156, 5–39.
Wang P., Zhang J., Min Q., 1985. Distribution of foraminifera in surface sediments of the East China Sea, in: Wang P. (Eds.), Marine Micropaleontology of China, Beijing: China Ocean Press and Springer-Verlag, 34–69.
Wang, J.L., Saito, Y., Oba, T., 2001a. High-resolution records of thermocline in the Okinawa Trough since about 10,000 aBP. Science in China, Series D: Earth Sciences 44, 193–200.
Wang, Y.J., Cheng, H., Edwards, R.L., An, Z.S., Wu, J.Y., Shen, C.-C., Dorale, J.A., 2001b. A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. Science 294, 2345–2347.
Xiao, J., Porter, S.C., An, Z., Kumai, H., Yoshikawa, S., 1995. Grain size of quartz as an indicator of winter monsoon strength on the Loess Plateau of Central China during the last 130,000 years. Quaternary Research 43, 22–29.
Xiang, R., Li, T.G., Yang, Z.S., Li, A.C., Jiang, F.Q., Yan, J., Cao, Q.Y., 2003. Geological records of marine environmental changes in the southern Okinawa Trough. Chinese Science Bulletin 48, 194–199.
Xu, X., Oda, M., 1999. Surface-water evolution of the eastern East China Sea during the last 36,000 years. Marine Geology 156, 285–304.
Yamamoto, K., Kurata, Y., Takayanagi, Y., Nishimura, A., Mimura, K., 2003. Latitudinal change of normal paraffin composition in the northwest Pacific sediments. Marine Geology 196, 157–170.
Yan, J., Thompson, P.R., 1991. Paleoceanographic evolution in the Okinawa Trough during the late Pleistocene. Oceanologia et Limnologia Sinica 22, 264–271 (in Chinese with English Abstract).
Yan, J., Cang, S.X., Qin, Y.S., 1990. The oxygen isotopic stratigraphy of core $Z_{14\text{-}6}$ in Okinawa Trough. Oceanologia et Limnologia Sinica 21, 442–448 (in Chinese with English Abstract).
Yuan, D.-X., Cheng, H., Edwards, R.L., et al., 2004. Timing, duration, and transitions of the last interglacial asian monsoon. Science 304, 575–578.
Zhao, M.X., Huang, C.-Y., Wei, K.-Y., 2005. A 28,000 year $U^{K'}_{37}$ sea-surface temperature record of ODP Site 1202B, the southern Okinawa Trough. TAO 16, 45–56.
Zheng, S.L., Shi, J.Y., 1991. Detection of long-chain unsaturated alkenones in surface sediments at South China Sea and Okinawa Trough. Acta Sedimentologica Sinica 9, 90–96 (supp. In Chinese with English Abstract).