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Palaeoceanographic change in the northeastern South China Sea during the last 15000 years

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ABSTRACT: The sedimentary succession of piston core RC26-16, dated by ^{14}C accelerator mass spectrometry, provides a nearly continuous palaeoceanographic record of the northeastern South China Sea for the last 15 000 yr. Planktic foraminiferal assemblages indicate that winter sea-surface temperatures (SSTs) rose from 18°C to about 24°C from the last glacial to the Holocene. A short-lived cooling of 1°C in winter temperature centred at about 11 000 ^{14}C yr ago may reflect the Younger Dryas cooling event in this area. Summer SSTs have remained between 27°C and 29°C throughout the record. The temperature difference between summer and winter was about ca. 9°C during the last glacial, much higher than the Holocene value of ca. 5°C. During the late Holocene a short-lived cooling event occurred at about 4000 ^{14}C yr ago. Oxygen and carbon isotopic gradients between surface (0–50 m) and subsurface (50–100 m) waters were smaller during the last glacial than those in the Holocene. The fluctuation in the isotopic gradients are caused most likely by changes in upwelling intensity. Smaller gradients indicate stronger upwelling during the glacial winter monsoon. The fauna-derived estimates of nutrient content of the surface waters indicate that the upwelling induced higher fertility and biological productivity during the glacial. The winter monsoon became weaker during the Holocene. The carbonate compensation depth and foraminiferal lysocline were shallower during the Holocene, except for a short-lived deepening at about 5000 ^{14}C yr ago. A preservation peak of planktic foraminifera and calcium carbonate occurred between 13400 and 12 000 ^{14}C yr ago, synchronous to the global preservation event of Termination I.

KEYWORDS: palaeoceanography; South China Sea; deglaciation; Holocene.

Introduction

Quaternary glacial–interglacial oscillations result in changes in sea-surface temperature, sea-level, water-mass circulation, fertility and ice-volume. The South China Sea, a semi-closed marginal ocean, may amplify such glacial–interglacial variations because of its sensitivity to the rise and fall of sea-level. It has been documented that the extensive development of glaciers in polar areas during the Last Glacial Maximum (LGM) resulted in a sea-level drop of approximately 120 m (Fairbanks, 1989). Most straits that border the South China Sea (SCS) became too shallow to allow exchange of waters between the South China Sea and the open ocean, except for the Luzon Strait. The Luzon Strait, located between Taiwan and Luzon Island (Fig. 1), with a sill depth of 2500 m, was the only passage to the West Philippine Sea during the last glacial. The exposure of continental shelves, especially in the southern part, caused a one-fifth reduction in the surface area of the South China Sea (P. Wang, 1990). The lower sea-level might also have caused the Kuroshio to shift offshore, east of its current position (Ujiie *et al.*, 1991; Ahagon *et al.*, 1993). In the northwestern Pacific, the Kuroshio Front advanced southward during the last glacial (Moore *et al.*, 1980; Thompson, 1981). The Kuroshio Front then moved northward at about 13 000 to 10 000 ^{14}C yr ago (Chinzei *et al.*, 1987), synchronous with the post-glacial warming of the North Atlantic Ocean (Duplessy *et al.*, 1981).

At present, the sea-surface temperature and water circulation of the South China Sea are generally governed by the East Asia monsoon climate (Wyrtki, 1961; Tchernia, 1980; Shaw and Chou, 1994). Warm southwesterly monsoon winds prevail during the summer and the warm Indian Ocean surface waters flow eastward into the southern SCS over the Sunda Shelf. The SSTs are uniformly high (28–29°C) all over the SCS. During the winter, cold northeasterly winds together with southward flow of the cold China Coastal Currents result in low SSTs, especially in the northern part of the SCS, and cause a strong north–south temperature gradient

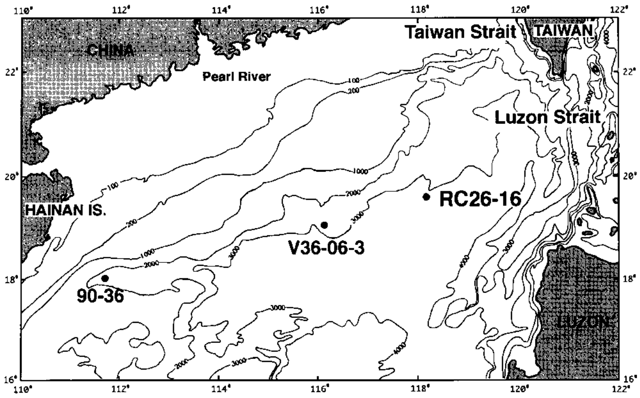

Figure 1 Locations of the studied piston core RC26-16 and two other cores (90-36 and V36-06-3) mentioned in the text.

(21 versus 27°C). The drop of sea-level by 120 m during the last glacial would expose the Taiwan Strait (<70 m in depth) and changed the position and route of the China Coastal Current. A sea-level rise of more than 60 m during the last deglaciation would reopen the Taiwan Strait as a major passage through which the resumed China Coastal Current flowed into the northern part of the SCS.

Besides sea-level change, the South China Sea also has been sensitive to changes of the monsoon system. Several studies based upon proxies in the loess–palaeosol sequences in central China have shown that the winter monsoons were strengthened during the glacial periods (An *et al.*, 1990, 1991; Maher *et al.*, 1994). Correspondingly, the SCS has also witnessed dramatic changes in hydrography, productivity and sedimentation during the last few glacial cycles (P. Wang, 1990; L. Wang and P. Wang, 1990; Thunell *et al.*, 1992; Schonfeld and Kudrass, 1993; Miao *et al.*, 1994; P. Wang *et al.*, 1995; Miao and Thunell, 1996).

Although the late Quaternary palaeoceanographic evolution of the northern SCS has been the subject of many studies (e.g., Feng *et al.*, 1988; L. Wang and P. Wang, 1990; Winn *et al.*, 1992; Schonfeld and Kudrass, 1993), only in recent years has a high-resolution time-scale been developed for the last-glacial–Holocene based on AMS ^{14}C dates (e.g., Chen *et al.*, 1997; Huang *et al.*, 1997). As a complement to our recent studies of the northwestern parts of the SCS (Chen *et al.*, 1997; Huang *et al.*, 1997, in press; Wei *et al.*, in press), this study attempts to provide a high-resolution palaeoceanographic record for the northeastern part of the SCS by examining the last-glacial–Holocene sequence of the core RC26-16 (19°53.23′N, 118°01.97′E at 2912 m water depth, Fig. 1). Because of its proximity to the main area of upwelling at the northwestern offshore of Philippines induced by the winter northeasterly monsoon (Pohlmann, 1987; State Oceanic Administration, 1988; Gong *et al.*, 1992; Chao *et al.*, 1996; Shaw *et al.*, 1996; Wiesner *et al.*, 1996), this core also serves to monitor changes in palaeomonsoonal upwelling.

Materials and methods

Piston core RC26-16 was cored by the Research Vessel *Conrad* from the Lamont–Doherty Geological Observatory in 1985. The sequence consists of hemipelagic silt and clay. From this 265-cm-long core a total of 66 samples was taken.

Mixed planktic foraminifers of four species, *Pulleniatina obliquiloculata* (Parker and Jones), *Neogloboquadrina dutertrei* (d'Orbigny) *Globorotalia mendardii* (Parker, Jones and Brady) and *G. inflata* (d'Orbigny) were picked from eight samples, and then cleaned and converted to CO_2 gas for ^{14}C accelerator mass spectrometry dating at the Institute of Nuclear and Geological Sciences, New Zealand. We adopted Bard (1988) to correct for the reservoir effect by deducting 400 yr from the measured marine ^{14}C ages. The corrected ages were further converted to calendar ages using the revised calib 3.0 Program (Stuiver and Reimer, 1993). The results are listed in Table 1. From extrapolation, the core-top is dated to be ca. 1300 ^{14}C yr ago, whereas the bottom is ca. 15 080. The temporal resolution for the upper part (1300 to 12 000 ^{14}C yr ago) is about 250–330 yr per sample on average, whereas the lower part is about 125 yr. The resolution is the highest so far among all the results published for this area.

Carbon and oxygen stable isotopic values were obtained from *Globigerinoides sacculifer* (Brady) (from 350 to 500 μm size fraction) and *Ga. menardii* (500–700 μm). The restriction to a certain size range for each taxon excludes the possible vital effects associated with planktic foraminiferal ontogeny (Berger *et al.*, 1978; Oppo and Fairbanks, 1989; Ravelo and Fairbanks, 1992; Kroon and Darling, 1995). The isotopic signal of *Gs. sacculifer* is considered to reflect the surface water (0–50 m) condition, whereas those of *Ga. menardii* reflect waters below the thermocline (Vergnaud Grazzini *et al.*, 1995; Mulitza *et al.*, 1997). The specimens were reacted with 100% H_3PO_4 at 60°C to yield CO_2 gas. The purified gas was analysed using a VG SIRA 10 mass spectrometer at the Earth Science Institute, Academia Sinica, Taipei, Taiwan. The standard deviations (1σ) of reference samples during the isotopic analysis were within 0.06‰ and 0.07‰ for $\delta^{18}O$ and $\delta^{13}C$ values, respectively.

Relative abundances of planktic Foraminfera were obtained by counting at least 250 specimens from the >150 μm fraction. We adopted the taxonomic scheme of Saito *et al.* (1981). Winter sea-surface temperatures (SSTs) were estimated using Berger's species-temperature-preference equation (Berger and Gardner, 1975), and summer and winter SSTs were also calculated using Thompson's (1981) western Pacific transfer function *FP* – 12*E*.

The species-temperature-preference equation is simply a weighted average of the optimum temperatures of all the foraminifers in the assemblage:

$$T_{est} = \Sigma\,(p_i t_i)/\Sigma p_i$$

Table 1 Ages derived from accelerator mass spectrometer ^{14}C measurement of planktic Foraminifera for eight samples in Core RC26-16

Depth (cm)	Species[a]	Measured ^{14}C age (yr BP)	Corrected ^{14}C age (yr BP)	Calendar year (yr BP)
10	P + M + D	2366 ± 71	1966	1888
34	P + M + D	4310 ± 83	3910	4372
58	P	6320 ± 110	5920	6742
90	P	8470 ± 84	8070	8983
110	P	9703 ± 103	9303	10238
150	P	12080 ± 120	11680	13617
206	M + D + I	13890 ± 130	13490	16149
250	P + M + D	15140 ± 130	14740	17636

[a]P, *Pulleniatina obliquiloculata;* M, *Globorotalia menardii;* D, *Neogloboquadrina dutertrei;* I, *Globorotalia inflata.*

where p_i is the percentage of species in the assemblage, and t_i is the 'optimum' temperature of the species. The optimum (mean) winter SSTs of planktic Foraminifera were compiled from Coulbourn *et al.* (1980) and Bé (1977) (Table 2). According to Berger and Gardner (1975), the standard error of the estimated modern temperatures in the South Pacific is 2.2°C. The standard error for the transfer function is 1.48°C and 2.48°C for summer and winter temperature, respectively (Thompson, 1981).

The concentration of inorganic phosphate–phosphorus in surface water was estimated by using the species preference equation similar to the palaeotemperature estimation (Berger and Gardner, 1975). The optimum phosphate–phosphorus concentrations are also listed in Table 2.

Carbonate and total organic carbon (TOC) contents in the bulk sediments were determined by using the LECO WR-112 carbon analyser. The pooled standard deviation for the total carbon is 0.020% (Chang *et al.*, 1991).

Results and discussion

Age model and sedimentation rate

Ages of samples in this study are reported as corrected ^{14}C age and calendar age (Table 1). However, to facilitate comparison with results published previously, all the events are reported in corrected ^{14}C age. The calibrated calendar age was used only for calculating sedimentation rates and accumulation rates. As shown in Fig. 2, the ^{14}C age and calendar age models yield somewhat different sedimentation rates. In general, the true sedimentation rates (based upon calendar age) are lower than the apparent sedimentation rates (based upon ^{14}C age) (Fig. 2). The sedimentation rates during the glacial interval (prior to 13 300 ^{14}C yr ago are 22.1 to 29.5 cm kyr^{-1} (corresponding to the apparent sedimentation rates of 30.9–35.2 cm kyr^{-1}), which are about two times, or even higher, than those during the deglaciation stage (11.8 to 15.9 cm kyr^{-1}) and the Holocene (9.7–14.3 cm kyr^{-1}).

Table 2 Mean winter sea-surface temperature (SST) and concentration of inorganic phosphate–phosphorus for modern planktic Foraminifera species used for species-preference equations in this study. Data were mainly from Coulbourn *et al.* (1980) and augmented from Bé (1977)

Species	Mean winter sea-surface temperature (°C)	[PO_4–P] (μg at l^{-1})
Globigerinella aequilateralis[a]	23.5	0.218
Globigerina bulloides	—	0.95
G. calida	22.6	0.34
G. rubescens	24.2	0.27
G. falconensis	18.2	0.53
G. quinqueloba[a]	10.5	0.90
Globigerinoides ruber	23.8	0.27
G. sacculifer	24.4	0.33
G. conglobatus	21.3	0.28
G. tenellus	23.8	0.27
Orbulina universa	17.9	0.55
Globigerinita glutinata	20.7	0.51
Pulleniatina obliquiloculata	25.1	0.38
Globorotalia menardii[a]	23.1	0.91
G. tumida	24.9	0.43
G. truncatulinoides	18.4	0.28
G. inflata	17.4	0.28
G. scitula[a]	22.7	0.43
Neogloboquadrina dutertrei	17.7	0.59
N. pachyderma (dextral coiling)	13.4	0.80
N. hexagona	22.6	0.41
Globoquadrina conglomerata	26.0	0.41
Sphaeroidinella dehiscens	22.6	0.39

[a]Data from Bé (1977).

The sediments of the core studied consist mainly of clay and silt, with subordinate biogenic calcareous carbonates. According to Chen (1978), the clay minerals in the bottom sediments of the area studied consist mainly of illite, smectite, chlorite and kaolinite, indicative of the provenance characters of coastal China, Taiwan and the Philippine archipelagoes. In view of the fact that the widespread laterites in southern China contain high amounts of kaolinite (50%–70%) (State Oceanic Administration, 1988), whereas the clay fraction of the Plio - Pleistocene sediments of southern Taiwan consists mainly of illite (50–65% on average in different sections) (Wang, 1970; Lin and Hsueh, 1979), we consider that the high content of illite (35–62%) and low kaolinite (ca. 15%) in the down-core sediments in the northeastern SCS (Feng *et al.*, 1990) reflects the fact that a major part of the sediments was originated from Southern Taiwan.

Recent studies of sediment traps deployed in the northern SCS also confirms that the highest sedimentary flux occurred annually during November to January, when the catchment area of the Pearl River was in the dry season (Jennerjahn *et al.*, 1992). On the other hand, the surface currents in Taiwan Strait and in Luzon Strait were driven by the northeast monsoon to flow southward into the South China Sea (Shaw and Chou, 1994). The subsurface contour currents flowing along the northern slope also might have contributed to lateral transportation (Schonfeld and Kudrass, 1993). Both currents would transport the eroded sediments from southern Taiwan to the northern SCS. The significant changes in sedimentation rate through the last-glacial–Holocene cycle are related primarily to sea-level change. This controlled the partition of fluviatile terrigenous materials for deposition on shelf, slope and abyssal plain (Schonfeld and Kudrass, 1993), and also contributed to variations in production of biogenic carbonate in surface water and corrosiveness of the bottom water.

Based upon compilation of sedimentation rates obtained from 13 cores, Schonfeld and Kudrass (1993) concluded that the highest sedimentation rates were recorded during Termination I (defined as 18 300–9800 yr ago by them), and also suggested that the rapid rise of sea-level during deglaciation may have enhanced erosion on land and continental shelves and induced along-slope mass transportation to the lower continental slope of the SCS. Our results are consistent with their data except that the highest sedimentation rates are associated with the last glacial and the first half of the deglaciation period. We suggest that during the last glacial, the low sea-level caused wide exposure of continental shelf and stronger erosion of rivers on the continent and on Taiwan Island, as well as a more rigorous ocean circulation, which strengthened the western boundary/contour currents. These two factors, in combination, contributed to the observed high rates of sedimentation. Once the sea-level rose to –60 m relative to the

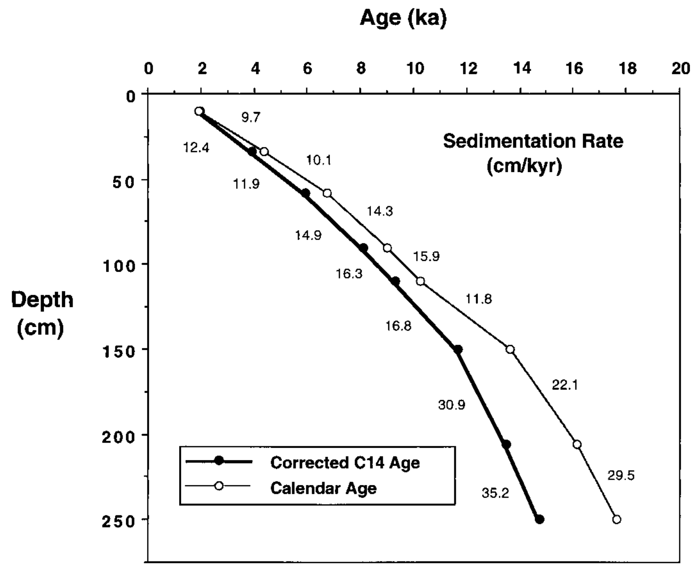

Figure 2 Depth–age plot showing two curves of sedimentation rates derived from corrected ^{14}C age and calibrated calendar year shown in Table 1, respectively.

present sea-level at about 10 000 ^{14}C yr ago (Fairbanks, 1989), the Taiwan Strait was submerged and became a buffer zone to the deposition of terrigenous sediments. Sedimentation rates in the northern SCS would have decreased in the last half of deglaciation and during the post-glacial.

Oxygen isotope and palaeotemperature profiles

The $\delta^{18}O$ values obtained from the two planktic foraminiferal species show a deglaciation trend, from a maximum value at about 14 300 ^{14}C yr ago to a stable Holocene value at about 9000 ^{14}C yr ago (Fig. 3). The surface-dwelling species, *Gs. sacculifer*, shows a higher variability in $\delta^{18}O$ values compared with the subsurface species *Ga. menardii*. *Globorotalia menardii* displays a similar $\delta^{18}O$ trned to that of *Gs. sacculifer*, except for a local positive excursion of 0.5‰ at 12 500 ^{14}C yr ago (Fig. 3). The excursion shown by *Ga. menardii* was also recorded in Core 90-36 located in the northwest of the South China Sea (Huang *et al.*, in press). The significance of this excursion is not clear. The gradients between these two species have become smaller from the glacial to the Holocene, suggesting a smaller vertical temperature gradient and stronger mixing during the last glacial compared with the post-glacial.

The difference between the maximum glacial $\delta^{18}O$ value (–0.06‰) and the Holocene value (–2.03‰) is about 1.97‰ for *Gs. sacculifer*. Such a difference is larger than the values of 1.0~1.2‰ reported from the open ocean of the equatorial western Pacific (Broecker, 1986; Thunell *et al.*, 1994). By subtracting the ice-volume effect of 1.2‰ (Fairbanks, 1989) from the last-glacial–Holocene $\delta^{18}O$ difference (ca. 2.0‰), the average temperature difference between the last glacial

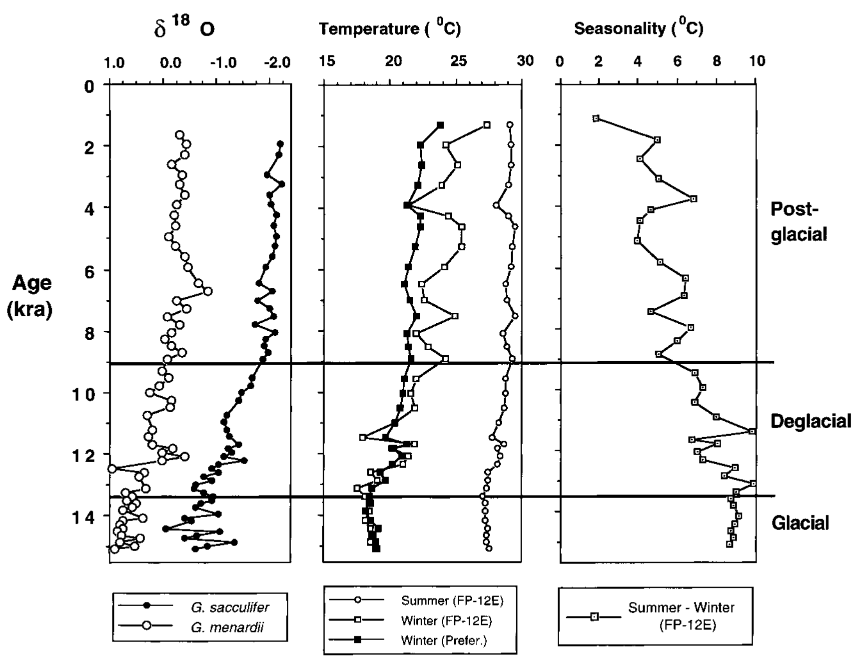

Figure 3 Time-series of planktic $\delta^{18}O$, palaeotemperatures and seasonality. The palaeotemperatures were estimated using Thompson's (1981) transfer function $FP - 12E$ and the species-temperature-preference equation (Berger and Gardner, 1975).

and post-glacial for the surface water is estimated to be 3°C (ignoring salinity effects).

Winter sea-surface palaeotemperatures derived from the species preference equation show a warming trend, from 18°C during the glacial to about 22°C in the Holocene (Fig. 3). The brief cold episode centred at 11 000 ^{14}C yr ago may represent the cooling effect of the Younger Dryas event (discussed later). The transfer function *FP* – 12*E* yields a similar winter-temperature trend, whereas the summer SSTs rose from the glacial temperature of 27°C to an interglacial value of 29°C. The estimated winter and summer temperatures of the core-top (1300 ^{14}C yr ago) are 24.3/22.5°C and 29.1°C, respectively. These values are consistent with the present mean winter (23.5°C) and summer (29°C) temperatures (fig. 2-6 of Feng *et al.*, 1988), suggesting that the palaeotemperature estimation is reasonable. The seasonal difference in SSTs (summer–winter) was about 9°C during the last glacial and decreased to about 5°C during the Holocene (Fig. 3). The 9°C difference in glacial seasonality is much larger than that recorded from the southern part of the SCS of about 3.3–7.3°C (Miao *et al.*, 1994; P. Wang *et al.*, 1996). Similar results were also shown in a nearby core V36-06-3 by L. Wang *et al.* (1994). The high seasonality in temperature of the northeastern SCS during the last glacial is attributed to the unusually low winter temperature at this site. This unusually low winter temperature, in turn, is considered as an indication of intensified upwelling caused by a stronger winter monsoon (discussed later).

Thunell *et al.* (1994) used the modern analogue technique (MAT) to reconstruct palaeo-SSTs over the low latitudes from 20°N to 20°S of the western Pacific and concluded that the glacial–interglacial difference in annual mean SST is less than 2°C. Our oxygen isotopic data suggest that the difference in annual mean temperature (3°C) is in the same range, but the winter temperature shows a dramatic change between the glacial and post-glacial in the northern part of the SCS at the same latitude. Such a large temperature change might reflect the fact that the winter northeasterly monsoon was much intensified during the last glacial. The intensified winter monsoon might have caused the South China Sea to be a 'cold pool' in juxtaposition with the much larger 'constant warm' Western Pacific Warm Pool to the east of the Philippines (Thunell *et al.*, 1994). The temperature gradient between the South China Sea and the western Philippine Sea must have been greater than today during the last glacial and thus a different regime would have influenced heat–water transport in this area. This situation may shed some light on explaining the discrepancy in palaeotemperature estimates between the terrestrial records (e.g, Webster and Streten, 1978; Rind and Peteet, 1985; Stuijts *et al.*, 1988) and marine proxies of foraminifera in the equatorial western Pacific (CLIMAP Project Members, 1976).

Carbon isotope and fertility profiles

In general, the subsurface-dwelling planktic species *Ga. menardii* exhibits the typical tropical pattern of carbon isotope variation characterised by similar δ^{13}C values during the last glacial and the late Holocene and a broad deglacial to early Holocene minimum (Fig. 4) (Oppo and Fairbanks, 1989). Oppo and Fairbanks (1989) argued that the observed deglacial δ^{13}C minimum in tropical planktic Foraminifera cannot be attributed simply to mean ocean δ^{13}C changes, but is a reflection of the isotopic composition of the underlying intermediate water and/or indicative of upwelling. *Globigerinoides sacculifer*, however, exhibits an atypical pattern by showing lighter δ^{13}C values. The depletion of ^{13}C in *Gs. sacculifer* during the last glacial makes the δ^{13}C values very similar to those of *Ga. menardii*. This similarity in δ^{13}C values between the surface (0–50 m) and subsurface dwellers (50–100 m for *Ga. menardii*, Schweitzer and Lohmann, 1991) indicates that at least the top 100 m of surface waters were well mixed during the last glacial. Following Sarnthien *et al.* (1990) who suggested that in upwelling areas the δ^{13}C values of the surface water will decrease due to the upwelling of the nutrient-enriched subsurface water, we further interpret the low δ^{13}C values in glacial *Gs. sacculifer* to be an indication of upwelling. At present, winter-monsoon-induced upwelling does occur in the areas off the western side of the Luzon island (Pohlmann, 1987; State Oceanic Administration, 1988; Gong *et al.*, 1992; Shaw *et al.*, 1996; Wiesner *et al.*, 1996), and these two planktic foraminiferal species show peak fluxes during the winter and spring upwelling seasons (Anonymous, 1992). The proximity of RC26-16 to this upwelling area leads us to infer that the upwelling during the last glacial was stronger and caused by strengthened winter monsoonal winds.

Other lines of circumstantial evidence also suggest that the winter monsoon and upwelling were stronger during the last glacial. The nutrient level indicated by the PO_4–P concentration was higher during the period prior to 13 000 ^{14}C yr ago (Fig. 4). The high fertility in the surface waters is likely to yield higher biological productivity. Weight percentage of total organic carbon in sediments, as well as accumulation rates of total organic carbon, are all relatively higher during the last glacial and the first half of the deglacial stage than in the post-glacial (Fig. 4). The alkenones concentration ([C_{37}]) also shows the highest values during the same period (Chen and Sheu, pers. comm.; Fig. 5). This is in agreement with the observation made by Winn *et al.* (1993) that the ice-age organic carbon in the northern SCS shows characteristic marine δ^{13}C values of >-20‰ (Winn *et al.*, 1992). The compositional change in planktic foraminiferal assemblage also indicates the existence of high primary productivity during the same period. As an indicator of upwelling and high biological production in this area (Chen and Kuo, 1986), *Neogloboquadrina dutertrei* dominated the assemblages (Fig. 5). The absolute abundance of *N. dutertrei*, measured as number of specimens per gram of sediment, also shows high values during the intervals prior to 12 500 ^{14}C yr ago (Fig. 5).

We propose that the high concentration and high mass accumulation rates of organic carbon, the high concentration of alkenones, and the dominance of *N. dutertrei*, consistently suggest that the last glacial and the first half of the deglaciation interval witnessed high surface-water biological productivity, which, in turn, is caused by upwelling or intensified mixing of the surface and subsurface waters, as indicated by the decreased gradients in δ^{13}C and in δ^{18}O between *Ga. menardii* and *Gs. sacculifer*.

Carbonate dissolution

At present the calcite carbonate compensation depth at about 3000 m in the northeastern SCS is the shallowest in the entire basin (Zheng *et al.*, 1993). The depth of the CCD became deeper by 200 m in the northern SCS during the last glacial (Zheng *et al.*, 1993). The water depth (2912 m) of our studied site is very close to the present CCD and slightly above its uppermost level during the late Quaternary.

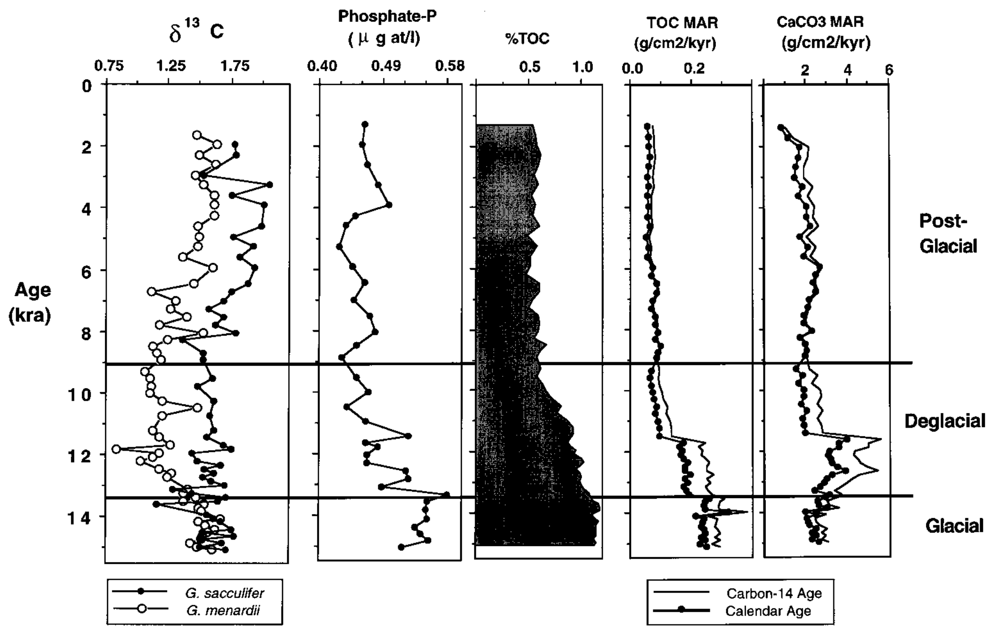

Figure 4 Time-series of planktic $\delta^{13}C$, estimated concentration of phosphate–phosphorus, content of total carbon (TOC) in weight percentage and mass accumulation rates of TOC and calcium carbonate.

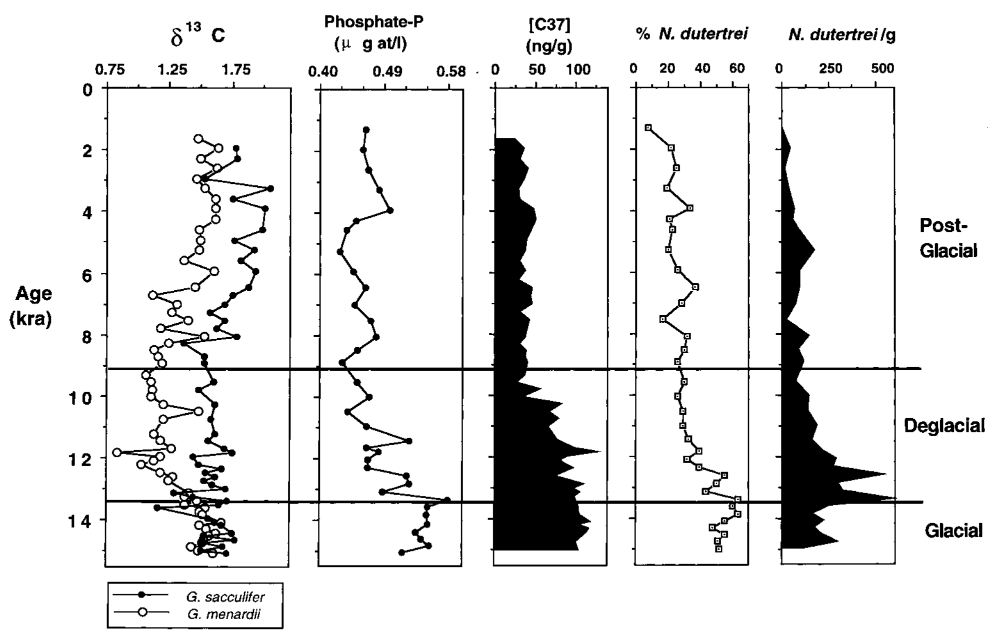

Figure 5 Proxies of palaeo-productivity: estimated concentration of phosphate–phosphorus, alkenones concentration ([C37]) in bulk sediments, relative and absolute abundance of *Neogloboquadrina dutertrei*, in comparison with $\delta^{13}C$ of two planktic Foraminifera species. The alkenone concentration data were provided by M.-J. Chen and D. D.-L. Sheu (pers. commun.).

Hence, the carbonate preservation in this core serves as a good monitor for the rise and fall of the CCD and the corrosiveness of deep waters of the northeastern SCS.

Weight percentage of calcium carbonate fluctuates throughout the record, ranging from the core-top minimum of 6% to a maximum of about 20% in the middle Holocene and in the deglaciation intervals. The % $CaCO_3$ cycles show a pattern transitional between the 'Pacific' and 'Atlantic' carbonate stratigraphies (Fig. 6). The variation of %$CaCO_3$ at this site appears to be a result of the interplay between three factors: (i) changes in biogenic carbonate production (e.g., sinking flux of Foraminifera and nannoplankton); (ii) carbonate dissolution intensity of bottom-water; and (iii) dilution effect of terrigenous sediment input. We calculated two dissolution indices to monitor the bottom-water dissolution intensity: (i) absolute abundance of whole planktic Foraminifera tests per gram sediments and (ii) %fragments defined as (number of planktic foraminiferal fragments/(number of planktic foraminiferal fragments + whole Foraminifera test)). The first index shows that the preservation of planktic Foraminifera is relatively better during 13 400–12 000 ^{14}C yr ago and at about 5000 ^{14}C yr ago than during other intervals. The second index suggests that the dissolution intensity of the bottom waters is generally low during the glacial and deglaciation periods, but it increases progressively after 8000 ^{14}C yr ago except for a brief drop centred at 5000 ^{14}C yr ago. The low %$CaCO_3$ in the last glacial is therefore not due to dissolution, rather it is a result of dilution caused by increased terrigenous influx of sediments. The high content of $CaCO_3$ and the highest mass accumulation during the deglaciation period is due to low corrosiveness of bottom waters and high productivity in surface waters.

The preservation/dissolution patterns are in good agreement with the preservation pattern of planktic Foraminifera (Huang *et al.*, in press) and calcareous nanofossils (Wei *et al.*, in press) in the northwestern site Core 90-36. The mass accumulation rate of carbonate reached a peak during the first half of the deglacial stage (13 000–11 500 ^{14}C yr ago), coinciding with the global carbonate preservation peak of Termination I as first defined by Berger (1977).

Broecker and Peng (1987) proposed that surface water productivity and upwelling became less efficient during the Termination, and hence there was a decrease in the total CO_2 content of surface water and an increase in carbonate (CO_3^{2-}) concentration in deep waters. The preservation spike recorded in deep-sea sediments during the Termination reflects basically the sudden increase of CO_3^{2-} in deep waters. Alternatively, Boyle (1988) suggested that during the onset of the last deglaciation, differential changes in the production rate of intermediate and deep waters caused CO_2 and nutrient to transfer from deep to intermediate waters. This results in an increase in CO_3^{2-} concentration in deep waters.

With a sill depth of 2500 m, the Luzon Strait allows only exchange of the intermediate waters of the western Phillipine Sea (WPS) with the intermediate and deep waters of the SCS, and the bottom waters of the WPS (>2500 m) do not enter into the SCS (Gong *et al.*, 1992; Chao *et al.*, 1996). The preservation peak at Termination I in the core studied appears to reflect an increase of CO_3^{2-} in the deep waters of the SCS as well as in the intermediate waters of the WPS. Our results, therefore, are in agreement with the model of Broecker and Peng (1987) but not of Boyle (1988).

The minor preservation peak at about 5000 ^{14}C yr ago (Fig. 6) was also observed in Core 90-36 (Huang *et al.*, in press; Wei *et al.*, in press), but it may be only a local phenomenon. Its significance awaits further analysis.

Deglaciation structure

The structure of deglaciation in the South China Sea has been a controversial issue. Broecker *et al.* (1988) argued that sedimentary characteristics of the southern SCS show an abrupt change at the close of the last glacial period, but that there was no sign of the Younger Dryas. In contrast, preliminary marine records from the SCS along with loess records from China led P. Wang (1990) to recognise a Younger Dryas cooling in this area. High-resolution oxygen isotopic records supported by a number of AMS ^{14}C dates from the Sulu Sea provided further support to the idea of a Younger Dryas cooling (Linsley and Thunell, 1990; Kudrass *et al.*, 1991). More recently, L. Wang *et al.* (1994) suggested that the last deglaciation in the northern SCS as recorded in core V36-06-3 was more or less a gradual process punctuated by two to three short-term climatic oscillations. P. Wang *et al.* (1995) reiterated the view that the palaeotemperature record of core V36-06-3 from the northern SCS shows an abrupt cooling event corresponding to the Younger Dryas. Due to the low resolution and lack of AMS ^{14}C dating of the record, however, their conclusions were somewhat tentative. On the basis of a more extensive record covering the past 150 kyr, Linsley (1996) argued that the planktic $\delta^{18}O$ variation in the Sulu Sea reflects primarily global sea-level and associated salinity change, and therefore the observed Younger Dryas-like pattern is merely a result of a teleconnection caused by advection along the sea surface (or through the atmosphere) of ^{18}O depleted meltwater from higher latitudes. In other words, in the Sulu Sea records, the ligher $\delta^{18}O$ values that bracket the 'Younger Dryas' result from meltwater pulses at 12 ka and 9.5 ka (Fairbanks, 1989), and there is no obvious cooling associated with the Younger Dryas *per se*. Thunell and Miao (1996) also reached the same conclusion for the southeastern SCS based upon palaeo-SST reconstruction.

The current study, with its high resolution and sufficient AMS ^{14}C dating points, can further elucidate the nature of the deglaciation pattern in the northeast SCS. Our results indicate that, starting from 13 700 ^{14}C yr BP the planktic foraminiferal $\delta^{18}O$ values show clear depleting trends until about 9000 ^{14}C yr BP, namely, from –0.5‰ to –1.80‰ for *G. sacculifer*, a change (1.3‰) that is almost equal to the glacial–interglacial difference of 1.25‰ (Fairbanks, 1989). Other proxies, such as winter SSTs, temperature seasonality (difference between summer and winter temperatures), nutri-

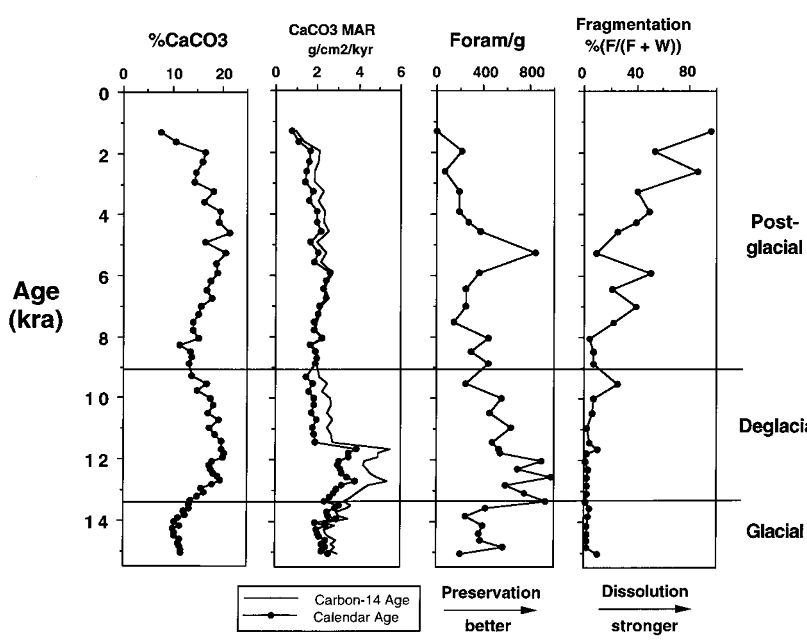

Figure 6 Various indices of carbonate preservation.

ent level (indicated by concentrations of phosphate–phosphorus and alkenone), together show a transitional phase from the glacial to post-glacial state (Figs 3–5).

The deglaciation can be subdivided further into three intervals. The first, from 13 700 to 11 700 ^{14}C yr BP, is characterised by continuous shift to lighter values of δ^{18}O in the two planktic foraminiferal species by about 0.9‰ (Fig. 3), and an increase in δ^{13}C gradient between them (Fig. 4). Global sea-level during that interval rose from –110 m to –70 m relative to the present, which corresponds to a change of 0.4‰ in sea water δ^{18}O (Fairbanks, 1989). The difference in δ^{18}O (0.9‰ – 0.4‰ = 0.5‰) suggests that there was an attendant warming of surface water by about 2°C within this 2 kyr, in agreement with the palaeo-SST estimates derived from planktic foraminiferal assemblages (Fig. 3). This interval is also marked by relatively high accumulation rates of $CaCO_3$ and organic carbon (Fig. 4). The high accumulation is attributed mainly to high sedimentation rates and not necessarily to high biogenic productivity, because %C_{org}, nutrient level and productivity (shown by ketones concentration and %*N. dutertrei*) had already declined (Figs 4 and 5).

The first stage was ended by the return of δ^{18}O to heavier values, to a local maximum of –1.25‰, at about 11 000 ^{14}C yr BP. This period was short-lived, however, and lasted from 11 500 to 10 500 ^{14}C yr BP. The increase in planktic foraminiferal δ^{18}O is only about 0.25‰, which corresponds to 1°C drop in temperature, assuming that the salinity remained unchanged. Foraminiferal-fauna-derived winter SSTs show a cooling by 1°C (preference equation) or 4°C (*FP* – 12*E*) (Fig. 3). Alkenones $U^{k'}_{37}$ palaeothermotry suggests also a moderate cooling of 0.9°C (from 24.3 to 23.4°C; Chen and Sheu, pers. comm., 1997). This Younger Dryas-like cooling event at the site studied is less pronounced when compared with the Sulu Sea, where the SST was estimated to drop by 3°C (Kudrass *et al.*, 1991).

The SSTs and planktic δ^{18}O returned to the general deglaciation trend after 10 500 ^{14}C yr BP and reached post-glacial values at about 9000 ^{14}C yr BP (Fig. 3). The sea-level at this time rose from –60 m to about –40 m relative to the present level, and the Taiwan Strait would have been submerged entirely. Nutrient levels continued to drop and the primary productivity indices show a declining trend in the third interval of the deglaciation sequence (Figs 4 and 5).

In summary, our record suggests that the deglaciation was not a rapid event in the northeastern SCS. Instead its structure is similar to the pattern of the 'two-step deglaciation model' with little or no ice disintegration in between (model 2 of Ruddiman, 1987; or the two meltwater-pulses model of Fairbanks, 1989). The Younger Dryas event, although pronounced in high latitudes and in the amphi-Atlantic area, appears to have had only a moderate impact on the northeastern SCS. It does not register strongly in either δ^{18}O or palaeo-SSTs. In contrast, the good preservation of Foraminifera and high accumulation of carbonate during 13 000 to 11 500 ^{14}C yr BP coincides with the 'global carbonate preservation peak of Termination I' recognised in the open oceans.

Conclusions

High-resolution planktic foraminiferal oxygen and carbon isotope stratigraphy, along with foraminiferal-fauna-derived palaeotemperature estimates and other proxies in piston Core RC26-16 document dramatic palaeoceanographic change over the past 15 000 yr. During the last glacial (prior to 13 700 ^{14}C yr BP), the winter sea-surface temperatures in the northeast South China Sea were about 18°C, which is 6°C lower than the Holocene temperature of 24°C. The summer SSTs showed less change and rose from 27°C to 29°C from the last glacial to post-glacial. Seasonal temperature difference between summer and winter during the last glacial was about 9°C, which is much larger than the present value of 5°C. The relatively colder glacial winter temperatures and larger seasonality compared with that in the open ocean in the western tropical Pacific indicates a strong amplification of the glacial effect in the northern South China Sea. Deglaciation began at about 13 700 ^{14}C yr ago, followed by a moderate cooling by 1°C between 11 500 and 10 500 ^{14}C yr ago. This Younger Dryas-like cooling event is less pronounced when compared with those found in the Sulu Sea (Kudrass *et al.*, 1991). The deglaciation trend resumed at about 10 500 ^{14}C and finally reached the post-glacial state at about 9000 ^{14}C yr ago. Since then, palaeotemperature and hydrographic conditions have remained more or less constant, except for a short-lived cooling at about 4000 ^{14}C yr ago.

Multiple proxies suggest consistently that the East Asia winter monsoons were strengthened during the last glacial, causing stronger upwelling in the northeastern South China Sea off Luzon. The combination of high surface-water productivity and high influx of terrigenous sediments resulted in very high sedimentation rates and accumulation rates during the last glacial and the first half of the deglaciation period. The carbonate compensation depth (CCD) and the foraminiferal lysocline were deeper during the last glacial than the present. The preservation peak of carbonate at 13 000 to 11 500 ^{14}C yr ago is synchronous with the global preservation peak of Termination I. The carbonate dissolution intensity of the bottom waters increased steadily during the Holocene, except for a brief drop at about 5000 ^{14}C yr ago.

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