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South China Sea surface water evolution over the last 12 Myr: A south-north comparison from Ocean Drilling Program Sites 1143 and 1146

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[1] Planktonic foraminifera (PF) from Ocean Drilling Program (ODP) Sites 1143 and 1146 in the southern and northern South China Sea (SCS), respectively, were quantitatively analyzed in order to reconstruct the sea-surface environment over the last 12 Myr. The observed decrease in deep-dwelling PF species after ~10 Ma at both sites is interpreted to reflect a depression of the upper water thermocline, corresponding to the closure of the Indonesian Seaway around 11–9 Ma. This upper water column structure implies the intensification of equatorial Pacific warm currents and the initial formation of the western Pacific “warm pool” (WPWP) during the early Late Miocene. The consistent pattern of south-north thermocline evolution and the synchronous disappearance of *Globoquadrina dehiscens* (9.8 Ma) at both Sites 1143 and 1146 together imply that the entire SCS was likely under the influence of the newly developed WPWP at ~10 Ma. After ~8 Ma, sea-surface temperatures and thermocline variations evolved differently between the southern and northern SCS. The total deep-dwelling PF fauna at Site 1143 decreased gradually in abundance from 6.6 to 2 Ma, indicating a deepening of the thermocline in the southern SCS. In contrast, deep-dwelling PF species increased in abundance from 3.1 to 2 Ma at Site 1146, reflecting a shoaling of the thermocline in the northern SCS. This south-north contrast reflects two major environmental regimes: (1) the southern SCS, which has mainly been under the influence of the WPWP since the late Late Miocene, and (2) the northern SCS, where effects of the east Asian winter monsoon have prevailed, especially since the Late Pliocene. Estimate of past sea-surface temperatures (SSTs) at Site 1143 suggests a relatively stable and warm environment in the southern SCS since about 2.5 Ma, with an increased influence of warm subsurface waters after the mid-Pleistocene transition (1.2–0.9 Ma). In the northern SCS, however, a gradual decrease in winter SST recorded at Site 1146 over the last 4 Myr records east Asian monsoon evolution, especially the enhancement of the east Asian winter monsoon between 3.1 and 2 Ma. **INDEX TERMS:** 1899 Hydrology: General or miscellaneous; 3030 Marine Geology and Geophysics: Micropaleontology; 4267 Oceanography: General: Paleooceanography; 4572 Oceanography: Physical: Upper ocean processes; 9604 Information Related to Geologic Time: Cenozoic; **KEYWORDS:** late Cenozoic, planktonic foraminifera, sea-surface environment, South China Sea, west Pacific “warm pool,” east Asian winter monsoon

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1. Introduction

[2] The El Niño-Southern Oscillation (ENSO), the spectacular climate phenomenon whose origins lie in the equatorial Pacific, is reflected by a zonal seesaw of the upper ocean thermocline between the eastern and western sectors

of the Pacific. The thermocline deepens in the east but shoals in the west during El Niño events; vice versa during La Niña events [*Le Borgne et al.*, 2002]. The western Pacific “warm pool” (WPWP), with average annual temperatures >28°C and occupying much of the tropical-subtropical region of the western Pacific and eastern Indian oceans, is an important source of water vapor and latent heat for the higher latitudes [*Yan et al.*, 1992]. Since the influences of ENSO and the WPWP on continental climate are likely connected, research has increasingly focused on the character and evolution of equatorial Pacific paleoceanography [*Lukas and Lindstrom*, 1987; *Linsley and Thunell*, 1990; *Thunell et al.*, 1994; *Martinez*, 1994; *Martinez et al.*, 1997, 1998; *Lea et al.*, 2000; *De Deckker et al.*, 2002; *Visser et al.*, 2003]. Whether the WPWP was a stable feature during late Quaternary glacial-interglacial cycles has been a subject of dispute for some years [*Stuifjes et al.*, 1988;

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Thunell *et al.*, 1994; Van der Kaars and Dam, 1995; Pelejero *et al.*, 1999]. While a number of studies indicate that the glacial-interglacial difference for mean annual sea-surface temperatures (SST) in the equatorial Pacific reached 2.8° to 5°C for the Late Pleistocene [Lea *et al.*, 2000; Visser *et al.*, 2003], others suggest that there were no significant SST changes in the WPWP region [Patrick and Thunell, 1997].

[3] Up to now very little has been known about the long-term history of the WPWP, although it has been suggested on the basis of planktonic foraminifer paleogeography that the WPWP likely formed after the closure of the Indonesia Seaway during the Late Miocene [Kennett *et al.*, 1985]. Faunal studies of ODP Sites 847 (0°N, 95°W) and 807 (0°N, 165°E) have revealed significant steps in the development of the modern east-west hydraulic gradient across the equatorial Pacific between 4.5 and 4 Ma, which have been linked to the closure of the Central American Seaway [Chaisson and Ravelo, 2000].

[4] Recent studies of the southern South China Sea (SCS), a region lying within the modern WPWP, have revealed an amplified climate signal during glacial-interglacial cycles as compared to the open ocean [Wang, 1998; Jian *et al.*, 2000]. Thus well-preserved sediments in long cores from the southern SCS may provide a solution to the long-term enigma of WPWP evolution. Sites 1143 and 1146, drilled during ODP Leg 184, were targeted to sample the Neogene paleoceanographic record within and immediately outside the modern WPWP, respectively, and to provide material for the reconstruction of SST gradients between the southern and northern SCS over the last 15 Myr. The successful drilling effort recovered excellent continuous sediment sections with moderate accumulation rates down to the Middle Miocene at Site 1143 and to the Early Miocene at Site 1146 [Wang *et al.*, 2000].

[5] In this study we analyzed the abundance of planktonic foraminifera in sediments from ODP Sites 1143 and 1146, estimated the character of thermocline variations over the last 12 Myr in the SCS, and reconstructed SST changes for the last 4 Myr. By comparing the different modes of thermocline evolution and SST variation between the southern and northern SCS, we assess here the relative influence of both the western Pacific “warm pool” and the east Asian winter monsoon on the evolution of the sea-surface environment in this region since the Late Miocene.

2. Materials and Methods

[6] Sites 1143 and 1146 were drilled and cored in the southern and northern South China Sea, respectively, during ODP Leg 184 (Figure 1). Site 1143 (9°21.72'N, 113°17.11'E; 2772 m water depth), the only site occupied in the southern SCS during the cruise, is located within the region of the modern WPWP [Yan *et al.*, 1992]. A total of 526 samples were collected from Site 1143 at intervals of 30–150 cm, equivalent to a temporal resolution of 10–60 kyr. Site 1146 (19°27.4'N, 116°16.4'E; 2092 m water depth) lies on the northern slope of the SCS, immediately outside the modern WPWP. A total of 796 samples were taken from Site 1146

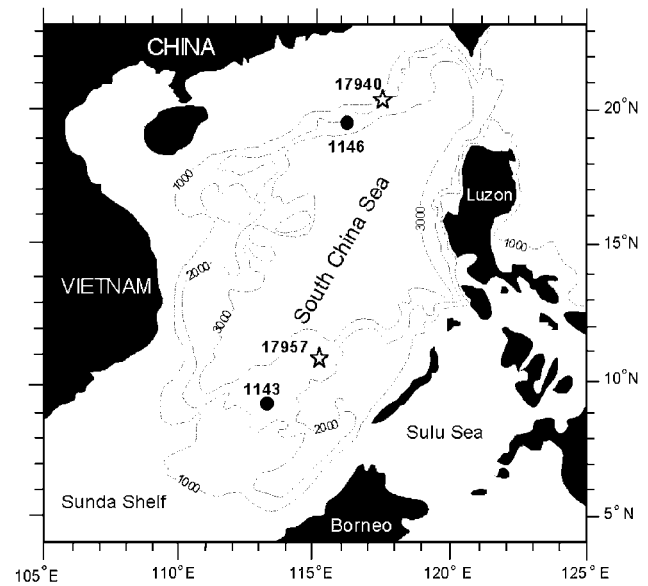


Figure 1. Locations of ODP Sites 1143 and 1146 in the South China Sea drilled during ODP Leg 184 [Wang *et al.*, 2000]. Sites 17957 and 17940 are from cruise the *Sonne 95* [Sarnthein *et al.*, 1994].

at intervals of 10–150 cm, corresponding to a resolution of 1–20 kyr [Wang, 2001; Huang, 2002].

[7] All samples, each of about 10–15 mL size, were oven-dried at 80°C, weighed, and soaked in water for one week. Diluted H₂O₂ was added to help disaggregate more indurated samples from stratigraphically older intervals. Samples were then washed through a 63- μ m sieve, and the >63 μ m fraction was dried, weighed, and stored for foraminiferal studies.

[8] Only planktonic foraminiferal specimens larger than 154 μ m were examined. Quantitative counting proceeded after subdividing the coarse fraction (>154 μ m) into a workable proportion from which enough specimens (>300 pieces on average) could be ensured. The taxonomy employed was based on the work of Thompson [1981], Kennett and Srinivasan [1983], Bolli and Saunders [1985], and Hemleben *et al.* [1988]. The chronology of planktonic foraminiferal datums follows those of Thompson *et al.* [1979], Berggren *et al.* [1995], and Chaisson and Pearson [1997].

[9] Information on planktonic foraminiferal depth habitats comes from previous investigations of their ecology [Bé, 1977; Hemleben *et al.*, 1988; Ravelo and Fairbanks, 1992; Patrick and Thunell, 1997], as well as from paleoecological estimates based on their morphological evolution and oxygen isotope evidence [Kennett *et al.*, 1985; Dowsett and Poore, 1990; Gasperi and Kennett, 1993; Faul *et al.*, 2000; Wang, 2001]. Species that inhabit the mixed layer are often characterized by a spinose wall, while most deep-dwelling species have a nonspinose, pitted to smooth, or postulate test surface [Saito *et al.*, 1981]. The mixed-layer group is represented in the SCS by *Globigerinoides ruber*, *Gs. sacculifer*, *Gs. fistulosus*, *Gs. extremus*, *Gs. obliquus*, *Globigerina rubescens*, *Globigerinita glutinata*, and *Dentoglobigerina altispira*. The deep-dwelling species include

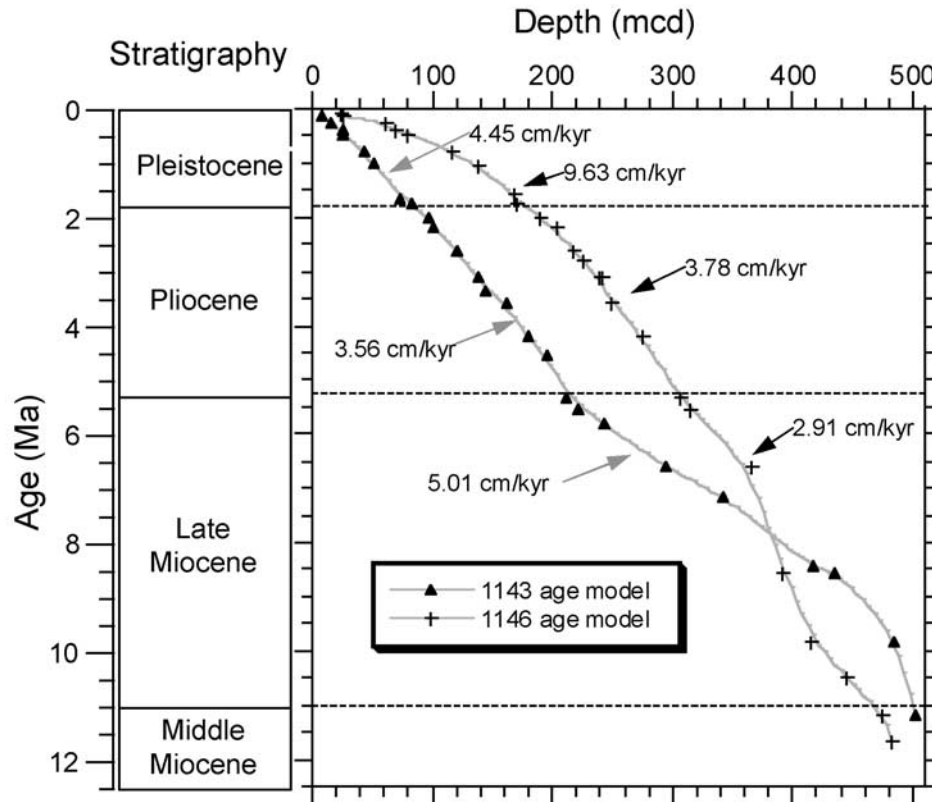


Figure 2. Chronology and stratigraphic subdivisions of Sites 1143 and 1146 based on planktonic foraminiferal and calcareous nannofossil datums (see Table 1). Planktonic foraminiferal datums modified after Li *et al.* [2004], Wang [2001], and Huang [2002]; calcareous nannofossil datums are from Wang *et al.* [2000] and Su [2004].

mainly *Sphaeroidinella dehiscens*, *Candeina nitida*, *Globigerina bulloides*, *Ga. praebulloides*, *Ga. nepenthes*, *Globotrifarina dehiscens*, *Gq. venezuelana*, *Pulleniatina* spp., *Neoglobotrifarina* spp., *Globotrifarina* spp., and *Sphaeroidinellopsis* spp.

[10] At Site 1143, preservation is generally better in upper Miocene and lower Pliocene sediments, a phenomenon probably related to a deeper lysocline in the Pacific before 4.5 Ma [Li *et al.*, 2004]. The effects of dissolution become more frequent and intense during the Late Pliocene-Pleistocene interval, although the dissolution-susceptible (mostly mixed-layer) species remain common. At Site 1146, carbonate dissolution as indicated by PF fragmentation appears to have followed a similar trend as at Site 1143 over the last 12 Myr [Wang, 2001; Huang, 2002]. The observed decrease of mixed-layer species since the Late Pliocene at Site 1146 is interpreted to be mainly due to decreasing temperature under a more prevalent east Asian winter monsoon (see section 4.2). Hence differential preservation appears to have had little effect on our paleoenvironmental interpretations and discussion.

3. Stratigraphy of Sites 1143 and 1146

[11] The stratigraphic subdivision and chronology for Sites 1143 and 1146 are based on a combination of

planktonic foraminiferal and calcareous nannofossil first occurrence (FO) and last occurrence (LO) datums, a microtektite event, and paleomagnetic results (Table 1 and Figure 2). The ~500-m long sedimentary sequences at Sites 1143 and 1146 are comparable in age and represent the last 12 Myr from the southern basin and northern shelf of the South China Sea.

[12] The Pleistocene/Pliocene boundary (LO *Globigerinoides fistulosus*, 1.77 Ma) occurs at 83.4 m (meters of composite depth below seafloor) in Site 1143 and at 170.5 m from Site 1146. The Pliocene/Miocene boundary (5.3 Ma) lies at ~210 m at Site 1143 and 304 m at Site 1146. The Middle/Late Miocene boundary, as defined by the FO of *Globigerina nepenthes* (11.19 Ma), was identified at 502.99 m and 475.21 m at Sites 1143 and 1146, respectively.

[13] As shown in Figure 2, a sedimentation rate of 4.54 cm/kyr is recorded for the Pleistocene section at Site 1143, which is roughly half as much as that at Site 1146 (9.63 cm/kyr). The higher sedimentation rate at Site 1146 is the result of increased terrestrial input owing to the enhanced erosion and uplift of Tibet Plateau [Wang, 2001], and possibly in part owing to frequent glaciations and the increased offshore transfer of sediments from global sea-level fluctuations [Zhang *et al.*, 2001]. During the Pliocene, estimated sedimentation rates between the two sites are similar: 3.56 cm/kyr at Site 1143 and 3.78 cm/kyr

Table 1. Revised Biohorizons, Microtektite Event, and Paleomagnetic Boundary for Sites 1143 and 1146

Events	Age, Ma	Depth, m		Sources ^a
		Site 1143	Site 1146	
FO <i>E. huxleyi</i> Acme ^b	0.09		22.96	A
LO pink <i>Gs. Ruber</i> ^c	0.12	8.12	25.95	B, C
FO <i>E. huxleyi</i> ^b	0.26	14.91	61.66	D, A
FO pink <i>Gs. ruber</i> Acme ^c	0.4	25.03	69.5	B, C
LO <i>P. lacunosa</i> ^b	0.46	26.26	78.31	D, A
Microtektite layer ^d	0.79	42.82	115.85	B, C
LO small <i>Geophyrocapsa</i> Acme ^b	1.01	52.32		D
Base Jaramillo ^c	1.07		137.9	D
LO <i>C. macintyre</i> ^b	1.58		167.41	A
FO medium <i>Geophyrocapsa</i> ^b	1.67	73.845		D
LO <i>Gs. fistulosus</i> ^c	1.77	83.4	170.5	B, C
FO <i>Gr. truncatulinoides</i> ^c	2	96.09	190.3	B, D
LO <i>Gr. multicamerata</i> s. l. ^c	2.17	101.77	203.5	B, C
LO <i>D. surculus</i> ^b	2.63	120.45	217.41	D, A
LO <i>D. tamalis</i> ^b	2.78		225.04	A
LO <i>Dt. altispira</i> ^c	3.09	138.98	239.2	B, C
LO <i>Ss. seminulina</i> ^c	3.12	138.98	241.47	B, D
FO <i>Gr. tosaensis</i> ^c	3.35	144.39		B, D
LO <i>Gr. margaritae</i> ^c	3.58	161.58	248.7	B, E
LO <i>Ga. nepenthes</i> ^c	4.2	179.72	274.89	B, E
LO <i>Ss. Kochi</i> ^c	4.53	195.5		B
FO <i>C. acutus</i> ^b	5.34	210.62	305.68	D, A
LO <i>D. quinqueramus</i> ^b	5.55	220.64	314.65	D, A
FO <i>Gr. tumida</i> ^c	5.82	242.5		D
FO <i>Gs. conglobatus</i> ^c	6.6	294.04	366.15	B, E
FO <i>A. primus</i> ^b	7.39	342.57		D
FO <i>D. berggreni</i> ^b	8.2	417.32		D
FO <i>Gs. extremus</i> ^c	8.58	435.4	391.9	B, E
FO <i>Nq. acostaensis</i> ^c	9.82	484.5	415	B, E
LO <i>Gr. mayeri</i> ^c	10.49		445.12	E
FO <i>Ga. nepenthes</i> ^c	11.19	502.99	475.21	B, E
LO <i>Gr. fohsi</i> s.l. ^c	11.68		482.19	E
FO <i>Gr. fohsi</i> s.l. ^c	13.42		505.69	E

^aSources are as follows: A, Su [2004]; B, Li et al. [2004]; C, Huang [2002]; D, Wang et al. [2000]; E, Wang [2001].

^bCalcareous nannofossil event. FO, first occurrence.

^cPlanktonic foraminiferal event. LO, last occurrence.

^dMicrotektite layer that approximates Brunhes-Matuyama paleomagnetic boundary.

^ePaleomagnetic boundary.

at Site 1146. For the Late Miocene interval, Site 1143 has a sedimentation rate up to 5.01 cm/kyr, while Site 1146 has a lower sedimentation rate of 2.91 cm/kyr. However, large sedimentation rate variations existed for different Late Miocene intervals (e.g., lower parts of both sites in Figure 2). Between ~10 and 6.5 Ma the sedimentation rate at Site 1143 (~6 cm/kyr) was much higher than that at Site 1146 (~2 cm/kyr), likely owing to the presence of abundant turbidite layers [Wang et al., 2000].

4. Results and Discussion

4.1. Planktonic Foraminiferal and Thermocline Evolution in the SCS During the Last 12 Myr

[14] To reconstruct upper water column structure, a combination of faunal and isotopic evidence is often employed because changes in the oceanic environment are recorded by both changing faunal abundances and isotopic signals [Kennett et al., 1985; Ravelo and Fairbanks, 1992; Patrick and Thunell, 1997; Cannariato and Ravelo, 1997]. Many

planktonic foraminifera are depth-related: “shallow-dwelling species” such as *Globigerinoides ruber* and *Globigerinoides sacculifer* live in the mixed layer above the thermocline, while “deep-dwelling species” including *Globorotalia menardii*, *Neogloboquadrina dutertrei*, *N. pachyderma*, and *Pulleniatina obliquiloculata* live in the subsurface water within or below the thermocline. When the thermocline is shallow, the deep-dwelling species are abundant; when the thermocline is deep, the mixed-layer species typically become more frequent [Patrick and Thunell, 1997]. Previous studies have shown that abundance variations of the deep-dwelling species are more significant in tracing paleoenvironmental changes in the tropical and subtropical Pacific [Kennett et al., 1985].

[15] We discuss below the thermocline evolution in the SCS based on the relative abundances of the total deep-dwelling species, supplemented by the down-core percentages of such species groups as *Neogloboquadrina* spp., *Pulleniatina* spp., and *Globoquadrina dehiscentes*. The combination of these data provides a means of identifying which species/group has been more impacted by paleoceanographic changes over the last 12 Myr in the southern and northern SCS.

4.1.1. Decrease of Deep-Dwelling PF Species at ~10 Ma

[16] As shown in Figure 3, a major decrease in the abundance of deep-dwelling PF occurred between 11 and 9 Ma at both sites. The event is centered at a subbottom depth of 476 m in Site 1143 and 440 m in Site 1146, both having an assigned age of ~10 Ma. The decrease in abundance of deep-dwelling PF indicates a deeper thermocline in the South China Sea after ~10 Ma. Lower total abundances of deep-dwelling PF species from 10 to 8.2 Ma (Site 1143) and 8.58 Ma (Site 1146) suggest that a deeper thermocline persisted until 8.2 and 8.58 Ma in the southern and northern SCS, respectively.

[17] In the western equatorial Pacific circulation system the modern deep thermocline of 175–300 m is a hydraulic characteristic of the WPWP owing to the piling up of warm surface water by the easterly trade winds. The thermocline here is 50–100 m deeper than in the eastern Pacific [Levitus and Boyer, 1994]. It has been suggested that the closure of the Indonesia Seaway played an important role in the early formation of the WPWP [Kennett et al., 1985]. The closure of the Indonesia Seaway could have contributed greatly to WPWP evolution owing to the piling up of warm surface water in the western equatorial Pacific and the strengthening of the Equatorial Under Current [Maier-Reimer et al., 1990; Hirst and Godfrey, 1993]. Recent plate tectonic reconstructions for southeast Asia indicate that the Indonesia Seaway was significantly narrowed or partially closed by ~10 Ma (Hall [2000, 2002]; “first closure” stage of Zhou [2002]) (Figure 4). We interpret the synchronous thermocline deepening at ~10 Ma at Sites 1143 and 1146 to likely have been caused by initial formation of the WPWP owing to the closure of the Indonesia Seaway in the early part of the Late Miocene.

[18] The parallel changes in the total deep-dwelling species observed between Sites 1143 and 1146 during the early Late Miocene clearly suggest that warm water existed

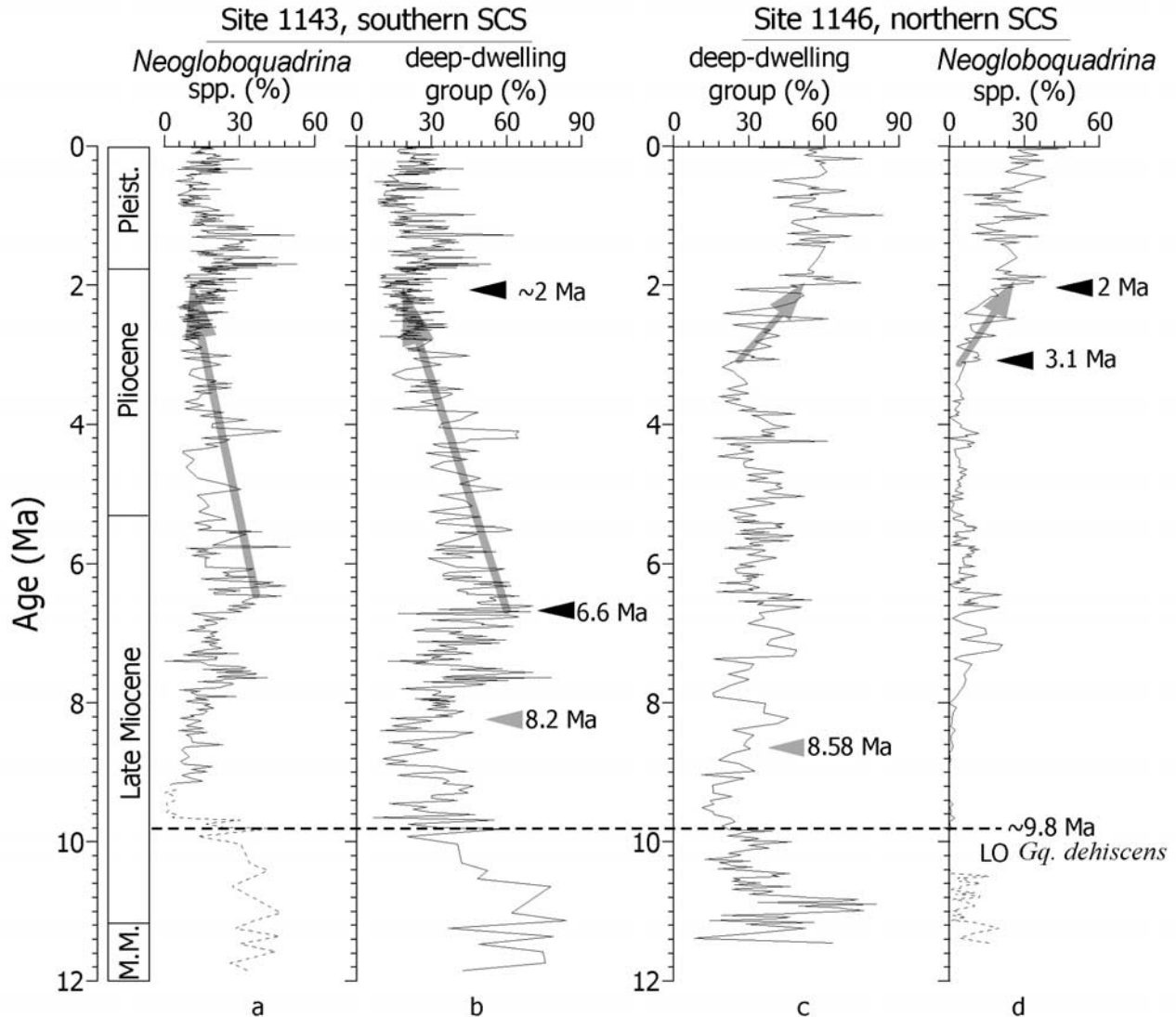


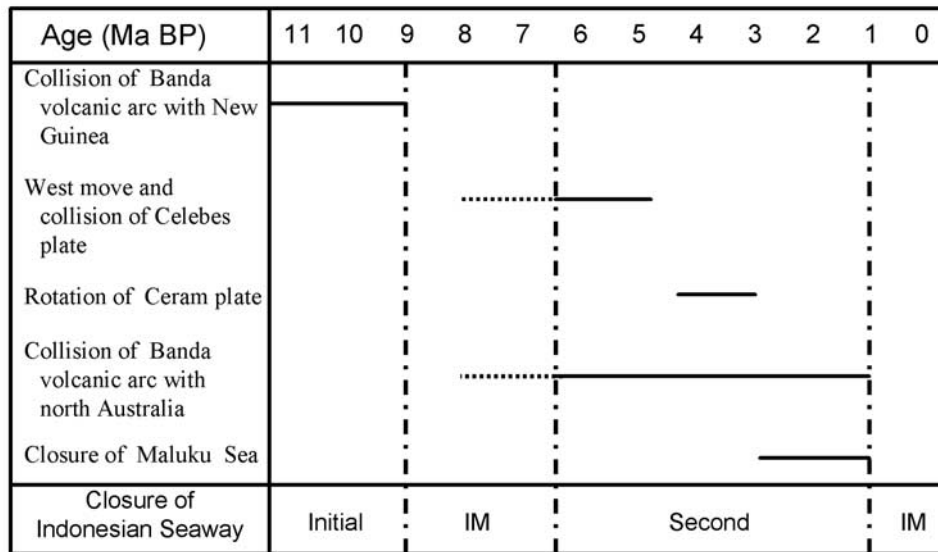
Figure 3. Down-core variations of *Neogloboquadrina* spp. (plus *Gr. siakensis-Gr. mayeri*) and total deep-dwelling planktonic foraminifera in Sites 1143 and 1146. Dashed lines in Figures 3a and 3d indicate *Gr. siakensis-Gr. mayeri* variations. *Neogloboquadrina* spp. (solid line) includes *N. dutertrei*, *N. pachyderma*, *N. humerosa*, and *N. acostaensis*. MM, Middle Miocene. Planktonic foraminifera data from Site 1146 are modified after Wang [2001]. Arrows in Figures 3a and 3b show a major decrease in *Neogloboquadrina* spp. and total deep-dwelling PF in Site 1143, indicating a gradual deepening of the thermocline from 6.6 to 2 Ma in the northern SCS. Arrows in Figures 3c and 3d show increases in the total deep-dwelling PF and *Neogloboquadrina* spp. at Site 1146 that reflect a shoaling thermocline from 3.1 to 2 Ma in the northern SCS.

in the northern as well as the southern SCS by this time. The warm water was likely related to the initial presence of the WPWP. Thus the “warm pool” appears to have extended further north beyond its present boundaries during its initial period of development. As illustrated in Figures 3b and 3c, the abundance patterns of the total deep-dwelling species at Sites 1143 and 1146 imply that a deeper thermocline existed longer in the southern SCS (until ~8.2 Ma) than it did in the northern SCS (until only ~8.58 Ma). The difference in timing of changes in inferred thermocline depth at Sites

1143 and 1146 suggest that the regional influence of the WPWP began to diminish, with the changes recorded first in the northern SCS, and then later in the southern SCS by about 8 Ma.

4.1.2. Disappearance of *Globoquadrina Dehiscens* at ~9.8 Ma

[19] *Globoquadrina dehiscens* was a typical subsurface-favorable species living mainly in warm temperate waters during the Miocene. This species became extinct from the world oceans at 5.6 Ma near the Miocene/Pliocene



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Figure 4. Major steps in the evolution of the Indonesia Seaway during the last 11 Ma [after Zhou, 2002].

boundary [Kennett *et al.*, 1985; Berggren *et al.*, 1995]. In the South China Sea, however, our data show that the disappearance of *Gq.* was several million years earlier than its record elsewhere. The last occurrence of *Gq. dehiscens* is recorded at core depths of 470 m in Site 1143 and 412 m in Site 1146, corresponding to an age of ~9.8 Ma [Wang, 2001; Q. Li *et al.*, manuscript in preparation, 2004] (Figure 3). This event has also been reported from numerous petroleum wells on the northern shelf of the SCS, such as BY 7-1-1 [Qing, 1996a, 1996b]. The LO of *Gq. dehiscens* at these industrial sites is close to the zone N15/N16 boundary, and is accompanied by the first occurrences of *Neogloboquadrina acostaensis* and *Globorotalia merotumida*.

[20] We consider the disappearance of *Gq. dehiscens* after ~10 Ma from both the southern and northern parts of the South China Sea as additional evidence for the initial development of the WPWP during the early Late Miocene. During the formation of the WPWP the pile up of warm water eliminated the temperate waters and dwellers such as *Gq. dehiscens* that characterized the fauna before 10 Ma in the SCS.

4.1.3. Thermocline Deepening After 6.6 Ma in the Southern SCS

[21] A remarkable change in the planktonic foraminiferal fauna at Site 1143 occurred during the Late Miocene-Pliocene between 6.6 and 2 Ma. At approximately 6.6 Ma the deep-dwelling PF species had a relative abundance of more than 60% of the total fauna, reflecting a shallower thermocline at that time. The deep-dwelling PF fauna then decreased gradually to their lowest abundance values of 10% by ~2 Ma (Figure 3b). A similar trend can also be observed in the variations of the *Neogloboquadrina* group (Figure 3a). These decreases imply a trend of thermocline deepening after 6.6 Ma in the southern SCS.

[22] Site 1143 now lies within the modern WPWP, so thermocline variations at this locality should have responded to the warm pool's development. A deepening thermocline after 6.6 Ma may thus imply that the local hydrography in the western Pacific evolved toward modern "warm pool" conditions after ~6.6 Ma.

[23] At ODP Site 807 from the Ontong Java Plateau, today near the center of the WPWP, the contribution of mixed-layer dwelling PF species also increased gradually after 6.4 Ma, implying a major development of the upper water structure of the WPWP [Chaisson and Ravelo, 2000]. Beginning at ~6.5 Ma, another active period of tectonic movement in the western Pacific region led to the complete closure of the Indonesia Seaway by ~1 Ma (Hall [2000, 2002]; or "second closure" stage of Zhou [2002]) (Figure 4). The coincidence of these events implies that the development of the modern WPWP in the latest Miocene was likely in part caused by the Indonesia Seaway closure.

[24] The contrasting evolutionary pattern of *Globorotalia* and *Neogloboquadrina* species in ODP Sites 847 and 806 from the equatorial Pacific has been suggested to mark a significant step in the thickening of the "warm pool" between 4 and 4.5 Ma and in the development of the modern east-west gradient in surface hydrography [Chaisson and Ravelo, 2000]. Similarly, the thermocline gradient between the southern and northern SCS also increased after 3.6 Ma, suggesting final formation of the modern WPWP [Jian *et al.*, 2003a].

[25] Modeling studies have demonstrated that a closed Central American Seaway at about 4.5 Ma could also have resulted in the thermocline deepening in the WPWP [Maier-Reimer *et al.*, 1990; Mikolajewicz *et al.*, 1993; Mikolajewicz and Crowley, 1997; Haug and Tiedemann, 1998]. The evolution of an east-west gradient in surface hydrography in

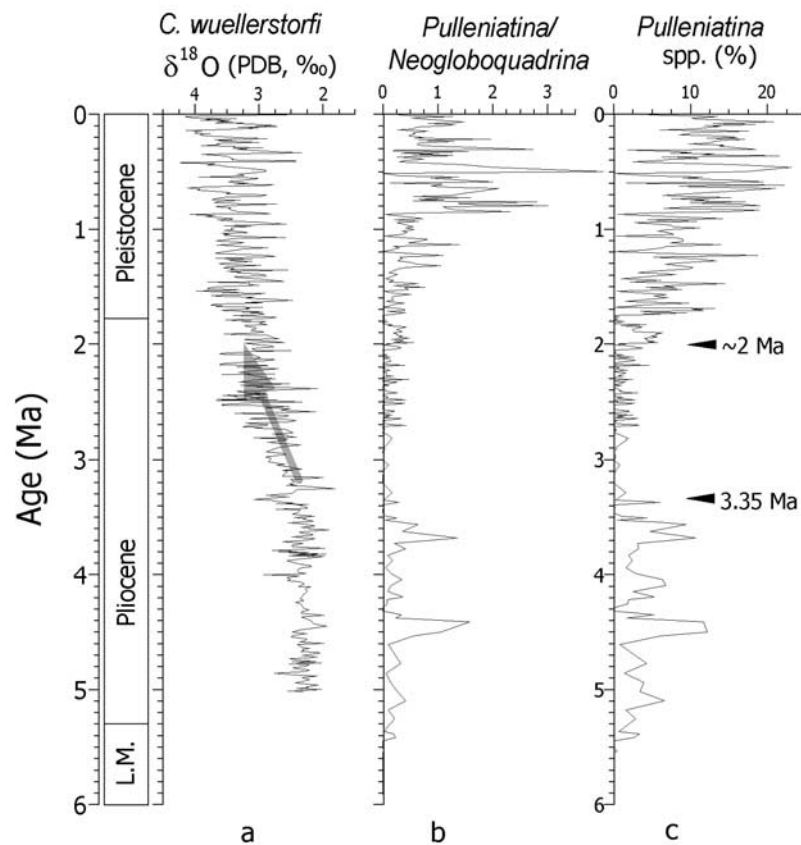


Figure 5. Abundance of *Pulleniatina* group, *Pulleniatina/Neogloboquadrina* ratio, and benthic foraminiferal $\delta^{18}\text{O}$ at Site 1143. The large arrow in Figure 5a indicates the increasing trend in $\delta^{18}\text{O}$ that corresponds to major global cooling and ice sheet expansion. *Pulleniatina* group includes *P. obliquiloculata*, *P. primalis*, *P. precursor*, and *P. spectabilis*. LM, Late Miocene. Benthic foraminiferal oxygen isotope data are from Tian et al. [2002].

the equatorial Pacific and a north-south thermocline gradient in the SCS during the Pliocene support the idea that closure of the Central American Seaway played an important role in the strengthening and enlargement of the WPWP [Chaisson and Ravelo, 2000].

4.1.4. Thermocline Shoaling Since 3.1 Ma in the Northern SCS

[26] At Site 1146 in the northern SCS the total deep-dwelling species and *Neogloboquadrina* spp. did not change significantly until 3.1 Ma (Figures 3c and 3d). *Neogloboquadrina* spp. increased markedly over the 3.1–2 Ma period, from less than 10% to more than 20% of the fauna, and increased further still to ~30% for the Pleistocene. The total deep-dwelling species also increased from 25 to 55% over the same period, reaching the highest value of >83% at ~1 Ma in the Middle Pleistocene. The abundance increases recorded by these taxa after ~3.1 Ma at Site 1146 reflected a sudden shoaling of the thermocline that continued into the Pleistocene in the northern SCS. This pattern contrasts sharply with the record of Site 1143, indicating that different controlling factors have operated on the surface waters of the southern and northern parts of the SCS since the late Late Miocene.

[27] The onset of upper water conditions analogous to the modern northern SCS, as inferred from Site 1146 data,

appears to occur in the middle Pliocene. The modern northern SCS is characterized by the strong mixing of surface water and a shallow thermocline because of the influence of the east Asian winter monsoon. The shoaling thermocline from 3.1 to 2 Ma at Site 1146 suggests an intensified winter monsoon during that period, as also indicated by eolian sediment records from central China [An et al., 2000] and Pliocene-Pleistocene oxygen isotope variations in the northern SCS [Jian et al., 2003b].

4.1.5. Variation of *Pulleniatina* spp. Content and *Pulleniatina/Neogloboquadrina* Ratio Since the Latest Miocene in the Southern SCS

[28] In the modern equatorial Pacific the *Neogloboquadrina* group is considered to be an eastern thermocline dwelling fauna, while *Pulleniatina* spp. represent central and western thermocline dwellers [Chaisson, 1995]. The *Pulleniatina* group first appeared in the SCS at ~5.5 Ma, almost 1 million years after its first occurrence at 6.6 Ma in the open western Pacific and elsewhere [Berggren et al., 1995; Li et al., 2004] (Figure 5c). Its lagged appearance possibly reflects a relatively weak thermocline before ~5.5 Ma in the SCS or its late migration into this marginal sea due to the barrier of the Philippines, or both.

[29] A comparison of the Site 1143 records with the global cooling and ice sheet expansion reflected in benthic

foraminiferal $\delta^{18}\text{O}$ makes it evident that the “three-period” pattern of evolution in the *Pulleniatina* group is correlated with down-core variations in the *Pulleniatina/Neogloboquadrina* ratio (Figure 5). On the basis of the ecology of *Pulleniatina* and *Neogloboquadrina*, the consistent changes between *Pulleniatina* content and *Pulleniatina/Neogloboquadrina* ratio indicate variations in the subsurface waters of the southern SCS. The “three-period” evolution of the *Pulleniatina* group thus would seem to suggest a two-step expansion of the warm subsurface water.

[30] This suggestion is consistent with the observation of Chaisson [1995] that the western Pacific “warm pool” developed in two steps after 5.8 Ma. From about 5.5 Ma to 3.35 Ma and after 2 Ma, extensive warm subsurface water existed in the southern South China Sea. From 3.35 to 2 Ma, low *Pulleniatina* abundance and low *Pulleniatina/Neogloboquadrina* ratios were likely the result of a decrease in warm subsurface waters at the time of ice sheet buildup in the Northern Hemisphere (Figure 5a).

4.2. Sea-Surface Temperature Variations Since the Pliocene

[31] The South China Sea experienced high-amplitude sea-surface temperature changes throughout the glacial-interglacial cycles of the Late Pleistocene. For example, published data from Sonne core 17940 (Figure 1) indicate that winter SST in the northern South China Sea varied by more than 8.6°C from the last glacial maximum to the Holocene, largely as a result of changes in the east Asian winter monsoon [Jian et al., 1996].

[32] We attempted to reconstruct long-term SST variations in the South China Sea by using transfer function PF-12E of Thompson [1981], one of the best methods for SST estimates in this region. For extinct species we used their modern counterparts for the SST calculations [see Anderson, 1997]. The modern and phylogenically related extinct species pairs include the following: *Globigerinoides ruber* (*Gs. extremus*, *Gs. obliquus*), *Gs. sacculifer* (*Gs. fistulosus*), *Globorotalia menardii* (*Gr. limbata*, *Gr. multi-camerata*, *Gr. plesiotumida*, *Gr. miocenica*), *Sphaeroidinella dehiscentis* (*Sphaeroidinellopsis seminulina*), and *Gr. scitula* (*Gr. margaritae*).

4.2.1. Amplitude of SST Variations at Sites 1143 and 1146

[33] Figure 6a shows the down-core variation of estimated SST over the last 2.5 Ma at Site 1143 in the southern SCS. In the period since 0.9 Ma, summer SST fluctuated by 0.9°C, while the winter SST varied by up to 3.5°C. Between 1.2 and 2.5 Ma, estimated summer and winter SST variations were 1.3°C and 4.5°C, respectively. These fluctuations are similar to previously estimated values of 1.3°C and 3.3°C variations for summer and winter SST at nearby core site 17957 (Figure 1) for the last 1.5 Myr [Jian et al., 2000].

[34] For the northern SCS at Site 1146 the summer SST varied by more than 3°C and the winter SST by close to 10°C over the Pleistocene glacial-interglacial cycles (Figure 6d). Both the winter and summer SST decreased gradually toward the low modern values of 18–19°C and 27–28°C, respectively, with large fluctuations across glacial-interglacial transitions since the Late Pliocene. Similar amplitude SST

fluctuations, of >2.5°C and 8.6°C, respectively, for summer and winter, have been also reported from a nearby locality, core site 17940 [Jian et al., 1996]. This comparison between estimated SST at Sites 1143 and 1146 indicates that the southern SCS has maintained relatively warm and stable SST, while larger amplitude SST fluctuations have characterized the northern SCS, especially during the Pleistocene.

4.2.2. Mid-Pleistocene Climate Transition and SST “Increase” at 1.2–0.9 Ma in the Southern SCS

[35] The transfer-function estimated SST shows evidence of a general increase in temperature between about 1.2 and 0.9 Ma at Site 1143 in the southern SCS, though glacial-interglacial fluctuations still remain evident (Figures 6a and 7b). The winter SST has been ~2°C higher on average since 0.9 Ma as compared to before 1.2 Ma.

[36] In transfer function PF-12E, *P. obliquiloculata* contributes 80% to a tropical dissolution-resistant factor [Thompson, 1981]. Given this, the relatively high abundance of *P. obliquiloculata* after 0.9 Ma at Site 1143 (Figure 5c) may have contributed to the higher estimated SST as compared to before 1.2 Ma. In fact, the SST “increase” at Site 1143 between 1.2 and 0.9 Ma actually resembles the relative increase of *Pulleniatina* spp. in the PF fauna. The warming event implied by the transfer function is also characterized by an increase in the *Pulleniatina/Neogloboquadrina* ratio (Figures 6b and 7c). *Pulleniatina* and *Neogloboquadrina* are both deep-dwelling species but indicate different water temperatures [Bé, 1977; Hemleben et al., 1988; Chaisson, 1995]. The observed increase in the *Pulleniatina/Neogloboquadrina* ratio, together with the absence of a significant response from the mixed-layer species from 1.2 to 0.9 Ma (Figure 6c), might thus be attributed to the presence of a warm subsurface water mass rather than to thermocline variations in the southern SCS.

[37] The subsurface environmental changes indicated by these faunas can be correlated with the mid-Pleistocene climate transition (MPT) from predominantly 40- to 100-kyr periodicity reflected in both benthic and planktonic stable oxygen isotopes at Site 1143 (Figures 7a and 7d, after Wang et al. [2001]). We interpret variations in the SST estimates, *Pulleniatina* spp. content, and *Pulleniatina/Neogloboquadrina* ratio at Site 1143 to reflect a situation where the southern SCS was dominated by warm stable surface waters during the Pleistocene, with relatively enhanced warm subsurface waters possibly present after the Mid-Pleistocene climate transition between 1.2 and 0.9 Ma.

4.2.3. SST Decrease Since 3.1 Ma in the Northern SCS

[38] The pattern of SST variation at Site 1146 in the northern SCS is quite different from that in the southern SCS. The transfer-function estimated SST at Site 1146 decreased gradually toward lower modern values since the Late Pliocene, with larger amplitude fluctuations superimposed (Figure 6d). From 3.1 Ma to the Late Pleistocene the estimated decrease in winter SST at Site 1146 was more than 7°C on average.

[39] Under the modern east Asian monsoon system, the northern SCS experiences lower SST in winter when the northeast winds blow across the SCS [Chen et al., 1991]. The evident decrease in SST over the last 3.1 Myr can be

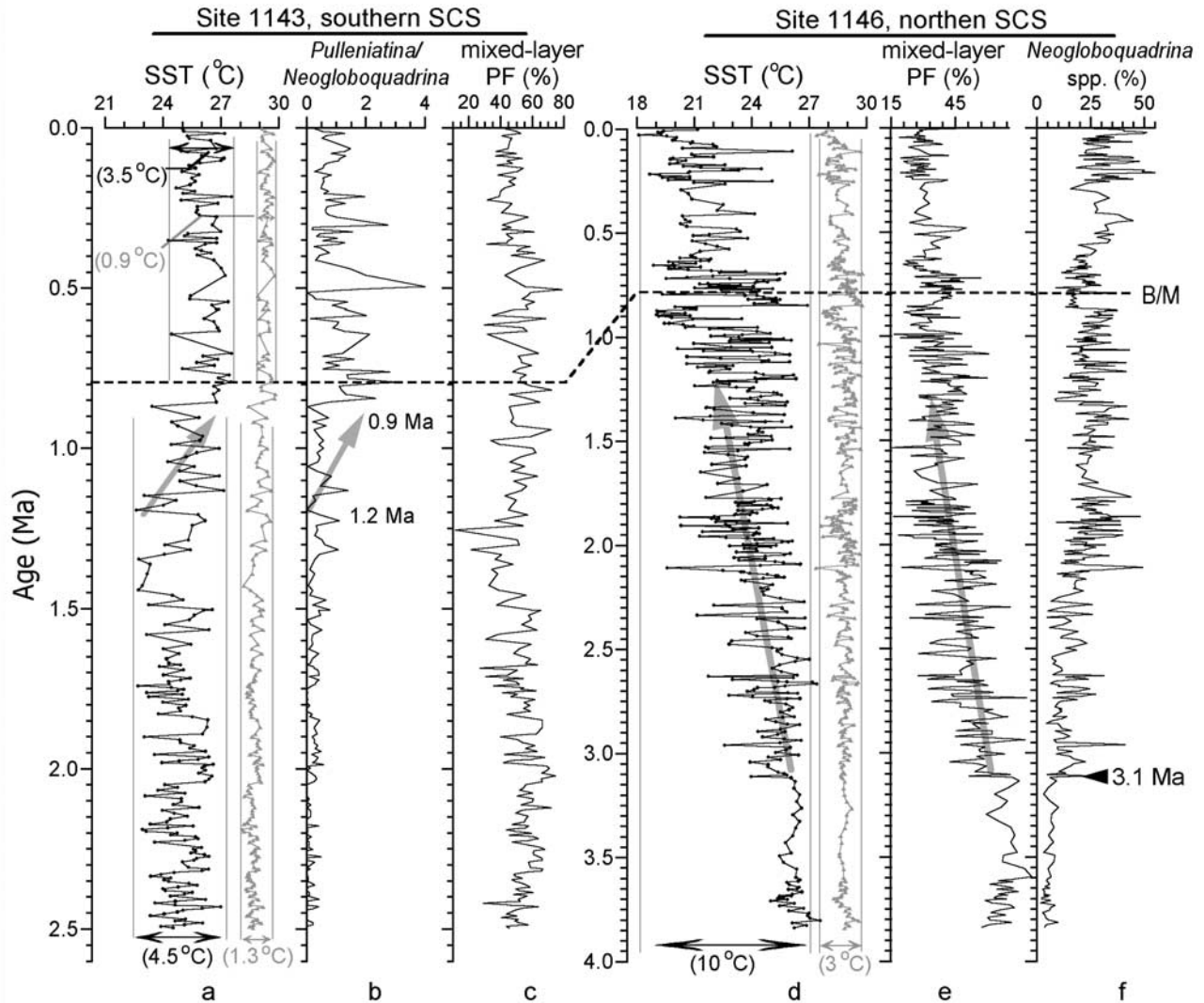


Figure 6. Comparison of transfer-function estimated SST variations between Sites 1143 and 1146 since the Late Pliocene. B/M, Brunhes-Matuyama paleomagnetic boundary (0.78 Ma). Planktonic foraminifera data at Site 1146 are modified from Huang [2002].

correlated with the downcore variations in mixed-layer species at Site 1146 (Figure 6e). Thus we conclude that the SST decrease recorded at Site 1146 since 3.1 Ma reflects an intensified winter monsoon in the northern SCS, consistent with what was discussed above for the shoaling thermocline. On the basis of these two lines of evidence we suggest that the sea-surface environment of the northern SCS has been strongly affected by the evolution of the east Asian winter monsoon since the Pliocene, as also indicated by the Pliocene-Pleistocene $\delta^{18}\text{O}$ results from the northern SCS [Jian *et al.*, 2003b].

4.3. Evolution of the WPWP and the East Asian Winter Monsoon and Their Influence on the SCS Environment

[40] The above reconstruction of South China Sea surface water over the last 12 Myr makes it possible to assess the relative influence of the WPWP and the east Asian monsoon on the environment in this region. The initial formation of

the WPWP at around 10 Ma caused the deepening of the thermocline in the SCS. It is likely that the range of the early WPWP extended further northward than at present, covering the whole SCS including the Site 1146 locality.

[41] Since the late Late Miocene, different factors have controlled the hydrographies of the southern and northern parts of the SCS, as expressed by the different modes of evolution of upper water column structure and sea-surface temperature at the two ODP sites. The evolution of the WPWP affected the variations of the PF fauna and upper water column structure at Site 1143 in the southern SCS, while the development of the east Asian winter monsoon had greater influence on the sea-surface environment at Site 1146, similar to the present conditions in the northern SCS.

[42] From about 6.6 to 2 Ma, gradual warming and deepening of the thermocline occurred in the southern SCS with the development of the WPWP (Figure 3b). In addition, SST have been relatively warm and stable over the

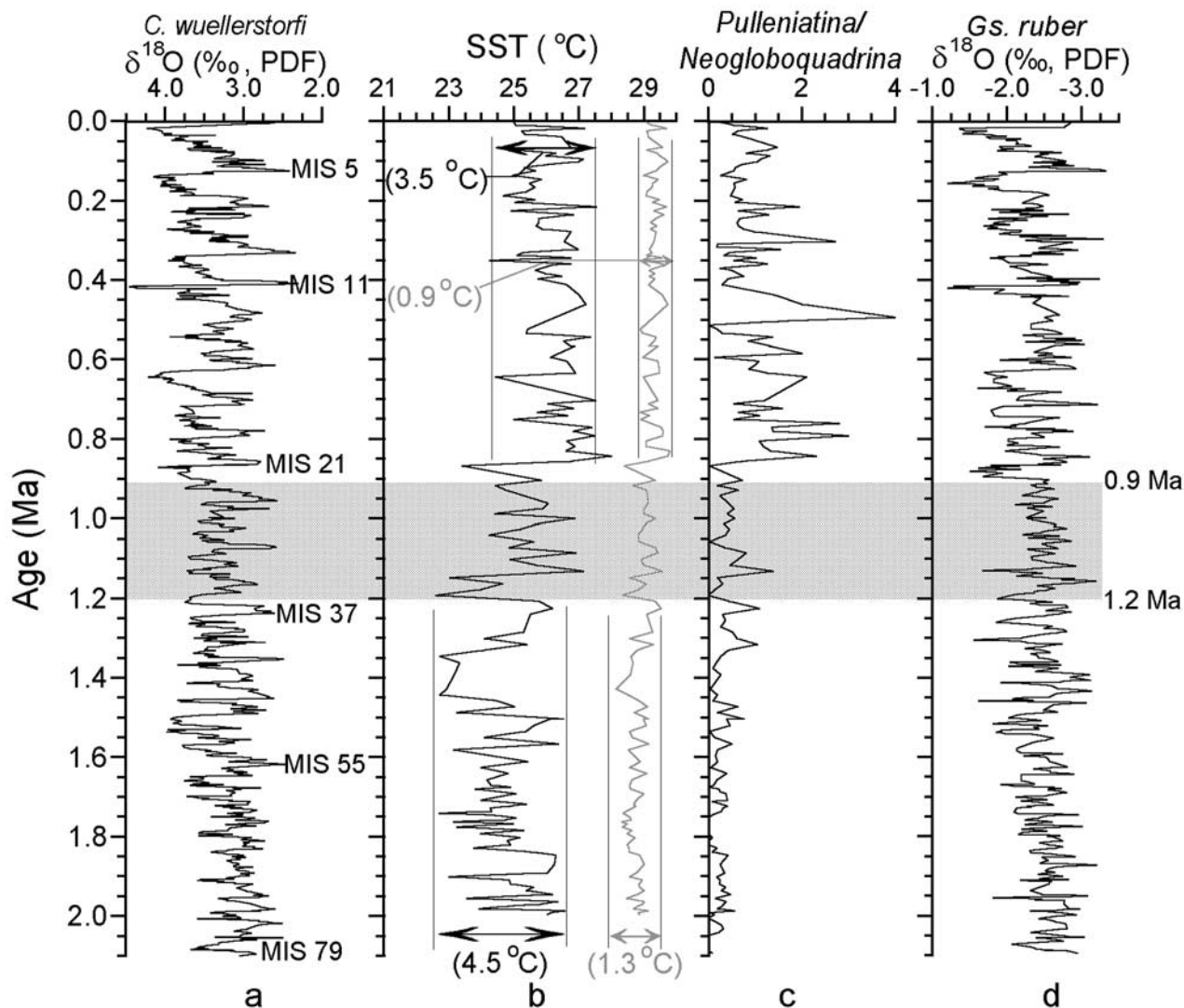


Figure 7. Benthic and planktonic foraminiferal $\delta^{18}\text{O}$, estimated SST, and *Pulleniatina/Neogloboquadrina* ratio for the last 2 Myr at Site 1143. The “mid-Pleistocene transition” (MPT) interval is shaded. Oxygen isotopes are after Wang *et al.* [2001].

last 2.5 Myr in the southern SCS, though an increasing influence of warm subsurface waters is inferred after the mid-Pleistocene transition (1.2–0.9 Ma) at Site 1143 (Figure 6b). In the northern SCS an intensified east Asian winter monsoon since the Late Pliocene resulted in shoaling of the thermocline and a decrease in SST, especially between 3.1 and 2 Ma at Site 1146 (Figures 3c and 6d).

5. Conclusions

[43] On the basis of continuous planktonic foraminiferal records from ODP Sites 1143 and 1146, we reconstructed the evolution of sea-surface environments in the South China Sea over the last 12 Myr. Environmental changes in the region were punctuated by major events occurring at ~ 10 , 6.6–2 Ma, 3.1–2 Ma, and 1.2–0.9 Ma.

[44] A decrease in the relative abundance of deep-dwelling planktonic foraminiferal (PF) species and the disappearance

of *Globoquadrina dehiscens* at around 10 Ma in both ODP sites are interpreted to reflect a deepening of the thermocline in the southern and northern parts of the SCS. This event was likely related to the initial development of the WPWP owing to the tectonic narrowing of the Indonesian Seaway and a pile-up of warm water in the western Pacific. The initial WPWP appears to have extended further northward than at present and to have affected both the southern and northern SCS including the Site 1146 locality. Parallel changes in sea-surface environment between the southern and northern SCS continued until ~ 8.2 Ma.

[45] A gradual decrease in the total deep-dwelling PF fauna at Site 1143 between 6.6 and 2 Ma reveals a further depression of the thermocline in the southern SCS. Toward the end of this time interval an increase in the total deep-dwelling PF species from 3.1 to 2 Ma at Site 1146 indicates a shoaling thermocline in the northern SCS. This contrasting evolution of the upper water column structure between the

southern and northern SCS reflects two different factors controlling the sea-surface environment since the latest Miocene.

[46] The thermocline deepening since 6.6 Ma recorded at Site 1143 is interpreted to mark an important step in the early development of the modern WPWP. Sea-surface temperature estimates at Site 1143 suggest a relatively stable warm environment during the Pleistocene in the southern SCS, while variations in the ratio of *Pulleniatina/Neogloboquadrina* and in the *Pulleniatina* spp. content suggest the enhancement of warm subsurface waters in the southern SCS after the Mid-Pleistocene transition at 1.2–0.9 Ma.

[47] In contrast, the sea-surface environment at Site 1146 in the northern SCS has been strongly affected by the east

Asian winter monsoon since the Late Pliocene. In addition to the shoaling thermocline, the gradual decrease in SST and the large SST fluctuations recorded at Site 1146 imply an intensified east Asian winter monsoon since the Pliocene, especially during the period of 3.1–2 Ma.

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