25,000-year late Quaternary records of carbonate preservation in the South China Sea

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Abstract

Pelagic sediments recovered from the South China Sea (SCS) exhibit characteristic carbonate preservation fluctuations that are closely linked to major cycles of climatic change. Two deep sea cores (SCS 90-36: 17°59.70'N, 111°29.64'E, water depth 2050 m; and 31-KL: 18°45.4'N, 115°52.4'E, 3360 m) that are located respectively above and below the regional lysocline (~3000 m) were selected for studying the late Quaternary pattern of carbonate preservation in the SCS. Preservation indices measuring changes in the abundances of dissolution-resistant planktonic foraminifers and coccoliths, and in ratios of foraminiferal fragments, benthonic foraminifers, and radiolarians, have revealed increased carbonate preservation during the last glacial period and/or during the last deglaciation (Termination I; ~12,000 k.y.), but stronger carbonate dissolution during the interglacial period of the Holocene. Detailed comparisons of preservation and oxygen isotope records further indicate that the level of carbonate preservation increased steadily from 25 k.y. through Termination I and that preservation reached a maximum during Termination I. From Termination I into the Holocene, however, the level of preservation decreased gradually to a minimum. Both records exhibit noticeable glacial/interglacial variations (1/f on the order of 10⁴ years), while the higher resolution record of 31-KL records a series of higher frequency oscillations (1/f on the order of 10³ years). These data confirm the observation that maximal carbonate preservation tends to occur during the later half of glacial stages as shown previously in many Indian and Pacific deep sea core studies. ©1997 Elsevier Science B.V. All rights reserved.

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1. Introduction

Studies on air trapped in Vostok polar ice cores (Barnola et al., 1987) have shown that atmospheric carbon dioxide concentrations ($p\text{CO}_2$) during the last ice age were significantly lower than during the preindustrial Holocene—about 200 ppm com-

pared to 280 ppm. Attempts to explain atmospheric $p\text{CO}_2$ changes on this time scale focus on ocean chemistry and circulation (Shackleton, 1977; Broecker, 1982a,b) since the ocean acts as a large reactive carbon reservoir that must play an important role in the global carbon cycle.

The preservation of calcium carbonate sediments on the sea floor provides important information about the carbon-carbonate chemistry of the oceans. Over the time scale of orbital variations

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(10^4 – 10^5 years), the change in the saturation state of carbonate in CO_2 -rich deep waters depends primarily on the global mass balance of carbonate and the vigor of deep ocean thermohaline circulation. Previous investigations on deep sea carbonate records from the Pacific (Arrhenius, 1952, 1988; Chuey et al., 1987; Farrell and Prell, 1989, 1991; Archer, 1991; Le and Shackleton, 1992) and the Atlantic (Ruddiman, 1971; Gardner, 1975; Damuth, 1975; Curry and Lohmann, 1986) indicate that the Pacific preservation pattern is out of phase with that of the Atlantic. Good preservation conditions occurred during glacial periods in the Pacific but during interglacial periods in the Atlantic. This asymmetry between the Pacific and Atlantic preservation patterns has been attributed to changes in basin-to-basin fractionation resulting from variations in North Atlantic Deep Water (NADW) production (Berger and Winterer, 1974). Carbonate preservation records of the Indian Ocean (Peterson and Prell, 1985a,b; Naidu et al., 1993; Bassinot et al., 1994) exhibit a Pacific-type pattern with good preservation during glacial periods and poor preservation during interglacial periods. Detailed comparisons of Indian carbonate preservation and oxygen isotope records further indicate that preservation maxima and minima tend to occur during the latter half of glacial and interglacial periods, respectively (Peterson and Prell, 1985b). These phase relationships were later confirmed in central equatorial Pacific studies (Farrell and Prell, 1989, 1991) in which the bathymetric variations in carbonate preservation of the past 800,000 years were reconstructed by using transects of deep sea cores.

The deep sea cores used in the previous studies (Peterson and Prell, 1985b; Farrell and Prell, 1989, 1991) were drilled from open oceans. Sedimentation rates of these cores approximated 2–3 cm/k.y., which were too low to be used to identify rapid climatic changes of the phase lagging patterns which vary on the time scale of only one thousand year. In this study, we used high-resolution deep sea records SCS 90-36 ($17^\circ 59.7' \text{N}$, $111^\circ 29.6' \text{E}$; 2050 m) and 31-KL ($18^\circ 45.4' \text{N}$, $115^\circ 52.4' \text{E}$; 3360 m) from the South China Sea (SCS) to investigate the temporal changes in carbonate preservation during the last glacial/interglacial climatic

change of the past 25,000 yr (Fig. 1). Average sedimentation rates are estimated at 4.0 cm/k.y. for SCS 90-36 and 6.7 cm/k.y. for 31-KL. These high sedimentation rates permit the recovery of short-term climatic signals, especially during the last glacial-to-interglacial transition. Previous studies (Wang, 1992; Bian et al., 1992; Cheng, 1992; Wang and Jian, 1992; Thunell et al., 1992) have documented changes in carbonate accumulation and preservation patterns in the SCS during the last glacial/interglacial period. Little information is available, however, on the relationship between carbonate preservation and climatic changes during this time. In this paper, we present a reconstruction of carbonate preservation patterns in the SCS during the past 25,000 yrs based on multiple sedimentological and micropaleontological indices of carbonate preservation from these two high-resolution records. With reliable age controls provided by AMS ^{14}C dating and oxygen isotope stratigraphic correlation methods, we show that the carbonate preservation level was enhanced during the last glacial and reached a maximum between 13,000 and 11,000 k.y. Carbonate preservation levels decreased suddenly starting about 11 k.y. and have continued to decrease up to the present. This pattern is consistent with that predicted by geochemical models (Berger and Winterer, 1974; Broecker and Peng, 1987; Boyle, 1988).

2. Data and methods

2.1. Stratigraphy and age model

The age models for cores SCS 90-36 and 31-KL were established by the use of AMS ^{14}C dating (Table 1) and oxygen isotope stratigraphic correlation (Imbrie et al., 1984; Prell et al., 1986). The ^{14}C dating control points that were used to obtain the chronology of the cores give consistently older ages with depth (Table 1). These ^{14}C ages correlate well with the age scheme estimated from oxygen isotope curves of planktonic foraminifers (Figs. 2 and 3). The ages of sediment samples between the control points of these age models were obtained by linear interpolation, and such comparisons

Fig. 1. Location map showing positions of South China Sea (SCS) core sites SCS 90-36 and 31-KL used in this study.

allowed estimates of carbonate sedimentation rates to be made (Table 1). The unexpectedly low sedimentation rates prior to 16 k.y. in core SCS 90-36 seem unreliable and may have resulted from erosion or a depositional hiatus such as has been observed near the Xisha Trough (Kudrass et al., 1992).

2.2. Preservation indices

Variations in carbonate preservation reflect primarily the calcium carbonate saturation state of bottom waters and interstitial pore waters. It can be assumed that the state of fragmentation of foraminifers is controlled by the preservation conditions under which the sediments were deposited, so that the percentages of foraminifer fragments increase as the sedimentary environments become more corrosive (Berger, 1967). Studies based on sediment-trap data have shown a linear increase in foraminiferal fragmentation with decreasing carbonate weight (Berger, 1967; Thunell and Honjo, 1981; Thunell et al., 1981). Similarly, data from surface sediment studies reveal increasing fragmentation with decreasing carbonate accumulation rate (Curry and Lohmann, 1986) and increasing carbonate loss (Peterson and Prell, 1985a). The fragmentation index is now the most widely used indicator of carbonate preservation in deep-sea

core studies (Thunell, 1976, 1982; Peterson and Prell, 1985a; Prell, 1984b; Le and Shackleton, 1992; Howard and Prell, 1994). Similar bathymetric trends are also observed for percentage variations in benthonic foraminifers or radiolarians (Peterson and Prell, 1985a), indicating a similar sensitivity to preservation changes for these indices.

The depth level where an abrupt change in dissolution rate gradients separate well preserved from poorly preserved sediments is termed the lysocline, and its depth is determined by a balance of three processes: (1) CaCO_3 productivity of calcareous planktonic organisms (foraminifers, pteropods, and coccoliths), which increases the supply of biogenic CaCO_3 sediments to the sea floor and in turn leads to better preservation; (2) non- CaCO_3 productivity of siliceous planktonic organisms (radiolarians and diatoms), which increases the downward flux of organic carbon and CO_2 and results in increased dissolution in bottom water. In the SCS, some of organic matter inputs into the ocean may be attributed to the supply of terrigenous materials from the Asian Continent; and (3) production and circulation patterns of intermediate and deep waters, the distribution of which, as well as their degree of corrosiveness, change with age. The modern position of the lysocline depth in the SCS has been

Table 1

AMS ^{14}C dating and sedimentation rate of cores SCS 90-36 and 31-KL

Depth (cm)	^{14}C age (k.y.)	Error bar ($\pm$ k.y.)	Sedimentation rate (cm/k.y.)
SCS 90-36			
0	1.3	0.1	
			4.5
27	7.3	0.1	
			6.2
51	11.2	0.1	
			5.7
63	13.3	0.1	
			17.2
87	14.7	0.1	
			0.8(?)
90	15.5	0.1	
			0.8(?)
96	22.9	0.2	
			1.6(?)
105	28.5	0.2	
31-KL			
5	1.8	0.1	
			7.0
35	6.1	0.1	
			4.5
41	7.4	0.1	
			10.9
81	11.1	0.1	
			14.1
133	14.8	0.2	
			19.4
273	22.0	0.2	
			10.7
301	24.6	0.3	

estimated to be 3000 m (Rottman, 1979), which is shallower than the 3400–3800 m reported in the western tropical Pacific (Parker and Berger, 1971; Thompson, 1976; Valencia, 1977; Berger et al., 1982).

We have used a variety of commonly used sedimentological and micropaleontological indicators to measure the relative changes of carbonate preservation in downcore samples from the SCS. The indicators used include: (1) percentage of dissolution-resistant planktonic foraminifer species (RSP%); (2) percentage of whole planktonic foraminifers (WPF%) = $100 \times \text{number of whole planktonic foraminifers} / (\text{number of whole planktonic foraminifers} + \text{number of planktonic foraminifer}$

fragments); (3) percentage of dissolution-susceptible nannoplankton species; (4) ratio of dissolution-susceptible nannoplankton = $\text{number of dissolution-susceptible nannoplankton} / (\text{number of dissolution-susceptible nannoplankton} + \text{number of dissolution-resistant nannoplankton})$; (5) percentage of benthonic foraminifers = $100 \times \text{number of whole planktonic foraminifers} / (\text{number of whole planktonic foraminifers} + \text{number of benthonic foraminifers})$; and (6) percentage of radiolarians = $100 \times \text{number of whole planktonic foraminifers} / (\text{number of whole planktonic foraminifers} + \text{number of radiolarians})$. For core SCS 90-36 indices 1, 2, 3, and 4 were used (Fig. 2; Table 2), and for core 31-KL indices 1, 2, 5, and 6 were used (Fig. 3; Table 3). Since we were able to develop age models for these two cores based on AMS ^{14}C dating and oxygen isotope stratigraphies, we present these preservation index data relative to age rather than core depth. The time series of the preservation indices from these two cores are shown in Figs. 2 and 3, and are arranged in the same direction as preservation–dissolution changes.

The concept of using percentages or ratios of dissolution-susceptible and dissolution-resistant fossil species as preservation indicators was proposed earlier by Berger (1967, 1968). A systematic study of Indian Ocean surface sediment data (Cullen and Prell, 1984) indicated that a summation index of the percentages of fourteen dissolution resistant planktonic foraminifer species (RSP%) could be used to reconstruct the depth of the lysocline. This same formula for calculating RSP% was used in the present study. During the course of sample processing, we counted at least 300 specimens of whole planktonic foraminifers for the abundance analyses.

Changes in the abundance of dissolution-susceptible nannoplankton species and in the ratio of the susceptible vs. resistant nannoplankton species can also be used as sensitive indicators for carbonate preservation. For the nannoplankton analyses, we counted at least 700 specimens for each sample. The taxonomy scheme used here is primarily based on the criteria outlined by Rahman and Roth (1989), Matsuoka and Okada (1989), and Biekart (1989). Explanation for the taxonomic criteria is

Fig. 2. Variations in carbonate preservation indices for core SCS 90-36, compared to downcore AMS ^{14}C dating and $\delta^{18}\text{O}$ stratigraphy (*G. sacculifer*) and plotted against age. A line shows the boundary between $\delta^{18}\text{O}$ stage 1 and stage 2. Abbreviations for the preservation indices are explained in Table 2.

also given in Wei et al. (in press). The following nannoplankton species were defined as susceptible species: *Umbellosphaera tenuis*, *U. irregularis*, *Thoracosphaera* spp., *Syracosphaera pulchra*, *Rhadosphaera longistylis*, *R. clavigera*, *Oolithotus fragilis cavum*, *O. fragilis fragilis*, *Discosphaera* spp., *Cyclolithus annula*. Resistant species are: *Umbilicosphaera sibogae sibogae*, *U. spp.*,

Helicosphaera carteri, *H. hyalina*, *H. kamptneri*, *H. wallichi*, *Ceratolithus cristatus*, *Calcidiscus leptoporus*.

Carbonate contents ($\text{CaCO}_3\%$) in cores SCS 90-36 and 31-KL (Figs. 2 and 3) are relatively low, ranging from approximately 10 to 30%. Our analyses show that the low carbonate contents primarily reflect dilution by terrigenous material,

Table 2

Preservation indices of downcore samples from core SCS 90-36

Core ID	Depth (cm)	Age (k.y.)	RSP ¹ (%)	WPF ² /(WPF + FG ³) (%)	Core ID	Depth (cm)	Age (k.y.)	Nanno – S ⁴ (%)	Nanno – S/(S + R ⁵) (%)
SCS 90-36	0	0.9	25.1	51.5	SCS 90-36	2	1.2	1.4	0.3
SCS 90-36	3	1.5	16.0	59.5	SCS 90-36	6	2.1	2.3	0.3
SCS 90-36	6	2.2	21.2	42.3	SCS 90-36	8	2.8	1.9	0.3
SCS 90-36	9	2.9	17.9	44.4	SCS 90-36	12	3.4	2.5	0.4
SCS 90-36	12	3.5	20.8	58.4	SCS 90-36	14	4.1	3.3	0.4
SCS 90-36	15	4.2	22.5	67.3	SCS 90-36	18	4.8	3.3	0.4
SCS 90-36	18	4.9	16.6	84.5	SCS 90-36	20	5.4	3.4	0.5
SCS 90-36	21	5.5	25.9	58.3	SCS 90-36	24	6.1	1.8	0.3
SCS 90-36	24	6.2	27.1	59.6	SCS 90-36	26	6.8	1.7	0.3
SCS 90-36	27	6.9	29.1	65.6	SCS 90-36	30	7.4	1.2	0.3
SCS 90-36	30	7.4	24.5	71.3	SCS 90-36	32	7.7	1.2	0.4
SCS 90-36	33	7.8	31.9	55.2	SCS 90-36	36	8.2	1.3	0.3
SCS 90-36	36	8.3	34.8	65.9	SCS 90-36	38	8.7	1.8	0.5
SCS 90-36	39	8.8	27.3	78.6	SCS 90-36	42	9.2	2.6	0.3
SCS 90-36	42	9.3	38.9	86.6	SCS 90-36	44	9.7	2.8	0.5
SCS 90-36	45	9.8	32.5	85.6	SCS 90-36	48	10.2	2.8	0.5
SCS 90-36	48	10.3	28.4	73.9	SCS 90-36	54	11.2	2.8	0.7
SCS 90-36	51	10.8	41.9	92.3	SCS 90-36	56	11.7	2.3	0.7
SCS 90-36	54	11.3	36.3	56.2	SCS 90-36	60	12.2	3.8	0.8
SCS 90-36	57	11.8	36.0	88.8	SCS 90-36	62	12.8	5.8	0.6
SCS 90-36	60	12.3	41.8	75.2	SCS 90-36	66	13.0	2.6	0.5
SCS 90-36	63	12.9	48.4	66.8	SCS 90-36	68	13.2	3.7	0.6
SCS 90-36	66	13.0	46.6	72.5	SCS 90-36	72	13.4	2.3	0.5
SCS 90-36	69	13.2	42.2	78.4	SCS 90-36	74	13.6	3.9	0.7
SCS 90-36	72	13.4	47.8	37.4	SCS 90-36	78	13.6	1.9	0.4
SCS 90-36	75	13.6	55.3	76.3	SCS 90-36	80	13.9	3.1	0.5
SCS 90-36	78	13.7	55.2	66.2	SCS 90-36	84	14.1	3.0	0.5
SCS 90-36	81	13.9	59.4	77.9	SCS 90-36	86	14.3	3.0	0.5
SCS 90-36	87	14.3	51.9	72.0	SCS 90-36	90	15.0	1.4	0.4
SCS 90-36	90	15.1	41.1	78.9	SCS 90-36	92	17.6	3.1	0.7
SCS 90-36	93	18.8	42.1	65.5	SCS 90-36	96	21.3	2.0	0.6
SCS 90-36	99	24.2	36.5	62.9	SCS 90-36	98	23.9	2.7	0.6
SCS 90-36	102	26.3	32.2	75.7	SCS 90-36	102	26.0	2.0	0.5

¹RSP = Number of dissolution-resistant planktonic foraminifers (defined based on Cullen and Prell, 1984).²WPF = Number of whole planktonic foraminifers.³FG = Number of planktonic foraminifer fragments.⁴S = Number of dissolution-susceptible nannofossil species.⁵R = Number of dissolution-resistant nannofossil species.

coming from nearby continental margins. Close examination indicates that the changes in carbonate contents as well as in carbonate accumulation rates (not shown) bear no obvious relationship to variations in the other preservation indices. Thus the index of carbonate content was excluded in our further computation of a composite preservation index.

2.3. Composite Preservation Index (CPI)

Most of the indices discussed above are also influenced by environmental variables other than carbonate preservation. For example, the number of fragments may be diminished by winnowing processes that result in sediment transfer to smaller size fractions. Ecological conditions may also cause

Fig. 3. Variations in carbonate preservation indices for core 31-KL, compared to downcore AMS ^{14}C dating and $\delta^{18}\text{O}$ stratigraphy (*G. sacculifer*) and plotted against age. A line shows the boundary between $\delta^{18}\text{O}$ stage 1 and stage 2. Abbreviations for the preservation indices are explained in Table 3.

changes in the percentages of benthonic foraminifers and radiolarians. To extract the variance that is common among these preservation indices, we have used R-mode Principal Component Analyses (PCA) (Thunell, 1976; Peterson and Prell, 1985a,b) on the SCS 90-36 and 31-KL downcore preservation index data (Tables 3 and 4). These

downcore data were interpolated by even time intervals ($\Delta t = 0.75$ k.y. for SCS 90-36 and $\Delta t = 0.30$ k.y. for 31-KL) before running the PCA. The width of these time intervals was chosen to be equivalent to the average temporal sampling of the two cores. The Composite Preservation Index (CPI) was then adopted as the first principal

Table 3

Preservation indices of downcore samples from core 31-KL

Core ID	Depth (cm)	Age (k.y.)	RSP (%)	WPF/(WPF + FG) (%)	WPF/(WPF + BF ¹) (%)	WPF/(WPF + R ²) (%)
31-KL	5	1.4	51.8	17.1	88.4	24.3
31-KL	9	2.0	34.5	30.4	97.2	54.0
31-KL	11	2.2	50.6	11.8	90.6	25.1
31-KL	13	2.5	66.7	7.4	86.8	12.1
31-KL	15	2.8	58.3	9.3	89.6	17.7
31-KL	17	3.1	64.0	12.8	87.7	29.4
31-KL	19	3.4	59.2	14.9	88.4	23.8
31-KL	21	3.7	48.9	7.8	87.0	19.8
31-KL	23	4.0	70.3	9.2	86.0	41.1
31-KL	25	4.2	61.0	15.0	96.2	67.6
31-KL	27	4.5	73.9	20.2	92.6	37.1
31-KL	29	4.8	63.5	22.3	93.7	45.8
31-KL	31	5.1	60.6	24.6	98.1	55.3
31-KL	33	5.4	55.4	22.8	94.4	53.2
31-KL	35	5.7	56.4	21.3	93.6	75.5
31-KL	37	6.1	78.8	30.6	94.5	72.2
31-KL	39	6.6	60.0	27.1	97.5	72.3
31-KL	41	7.0	51.7	29.2	95.0	79.5
31-KL	43	7.2	51.8	36.4	94.0	77.5
31-KL	45	7.4	47.6	39.2	97.3	83.1
31-KL	47	7.6	62.1	23.3	92.0	85.1
31-KL	49	7.7	52.9	23.6	94.4	77.3
31-KL	51	7.9	60.5	27.8	97.5	90.8
31-KL	53	8.0	61.7	37.3	95.4	94.6
31-KL	55	8.3	47.7	33.5	98.6	84.3
31-KL	57	8.5	70.6	27.1	96.9	85.7
31-KL	59	8.7	61.8	38.7	99.4	92.9
31-KL	65	9.2	63.7	38.6	97.6	95.4
31-KL	67	9.4	54.9	37.6	98.2	84.7
31-KL	69	9.6	51.9	38.8	99.2	88.9
31-KL	71	9.8	55.8	53.9	99.2	96.5
31-KL	75	10.1	55.0	55.6	97.0	96.3
31-KL	77	10.3	51.7	66.7	99.4	96.5
31-KL	79	10.5	50.6	67.5	98.8	97.4
31-KL	81	10.7	43.4	36.4	94.7	94.3
31-KL	83	10.8	38.7	54.8	98.7	94.8
31-KL	85	11.0	42.2	69.2	98.6	98.2
31-KL	87	11.1	44.3	63.3	99.7	96.8
31-KL	89	11.3	52.1	79.0	99.3	98.2
31-KL	91	11.4	43.7	52.5	97.6	95.2
31-KL	96	11.7	52.5	70.7	99.0	97.0
31-KL	101	12.1	53.6	65.2	98.9	95.5
31-KL	105	12.4	51.2	66.3	99.0	94.5
31-KL	109	12.7	47.3	60.0	99.2	82.2
31-KL	117	13.2	57.6	70.8	98.5	94.1
31-KL	121	13.5	55.1	65.7	99.4	91.3
31-KL	125	13.8	52.8	55.5	97.3	88.2
31-KL	129	14.1	60.8	70.1	99.0	91.6
31-KL	133	14.4	56.5	75.5	98.8	92.1
31-KL	137	14.6	62.8	69.0	96.9	93.2
31-KL	145	15.0	53.6	84.7	98.7	93.5
31-KL	149	15.2	54.0	73.7	97.6	84.7

Table 3 (continued)

Core ID	Depth (cm)	Age (k.y.)	RSP (%)	WPF/(WPF + FG) (%)	WPF/(WPF + BF ¹) (%)	WPF/(WPF + R ²) (%)
31-KL	153	15.4	52.9	67.0	99.5	90.8
31-KL	157	15.6	61.1	77.9	98.6	92.8
31-KL	161	15.8	54.4	49.2	99.2	85.6
31-KL	165	16.0	59.4	70.2	98.8	92.2
31-KL	169	16.2	55.1	66.0	99.4	81.6
31-KL	173	16.4	59.9	73.7	99.4	93.8
31-KL	177	16.6	56.3	77.6	98.9	92.6
31-KL	181	16.9	55.2	65.4	98.4	86.5
31-KL	185	17.1	51.7	73.1	99.5	90.0
31-KL	189	17.3	51.4	52.9	97.8	59.2
31-KL	197	17.7	51.1	64.9	99.0	82.3
31-KL	201	17.9	49.3	71.7	99.1	89.9
31-KL	205	18.1	55.8	64.2	99.3	94.5
31-KL	209	18.3	58.3	60.5	97.8	71.4
31-KL	213	18.5	64.0	74.4	99.2	88.8
31-KL	217	18.7	55.5	75.4	99.4	85.6
31-KL	233	19.5	58.7	74.2	99.0	88.9
31-KL	241	19.9	69.4	53.0	99.0	84.1
31-KL	249	20.4	58.4	69.7	98.1	76.4
31-KL	253	20.6	68.0	60.1	98.3	85.3
31-KL	261	21.0	66.3	59.9	99.0	82.7
31-KL	269	21.4	67.0	54.0	98.5	68.1
31-KL	273	21.6	58.6	63.0	98.3	90.6
31-KL	281	22.3	60.7	64.8	98.9	92.1
31-KL	289	23.1	67.7	50.2	99.4	60.8
31-KL	293	23.5	57.8	57.6	97.1	79.1
31-KL	301	24.2	62.7	55.1	97.0	85.7
31-KL	313	25.3	71.9	62.7	97.5	81.7

¹BF = Number of benthonic foraminifers.²R = Number of radiolarians.

Table 4

SCS 90-36: Factor scores for the Composite Preservation Index (CPI) and the average percentage of variance for each downcore preservation index explained by the CPI

Preservation indices	Factor scores for CPI	Variance explained (%)
– RSP (%)	–0.54	29.2
WPF/(WPF + FG) (%)	0.63	39.7
Nanno – S (%)	0.80	64.0
Nanno – S/(S + R) (%)	0.88	77.5

Total variance explained = 52.6%.

component, which accounts for about 52% and 76% of the variance in the SCS 90-36 and 31-KL data, respectively (Tables 4 and 5). Well preserved sediments are indicated by more positive CPI

Table 5

31-KL: Factor scores for the Composite Preservation Index (CPI) and the average percentage of variance for each downcore preservation index explained by the CPI

Preservation indices	Factor scores for CPI	Variance explained (%)
– RSP (%)	0.64	41.0
WPF/(WPF + FG) (%)	0.89	79.2
WPF/(WPF + BF) (%)	0.96	92.2
WPF/(WPF + R) (%)	0.96	92.2

Total variance explained = 76.2%.

values, whereas poorly preserved sediments are indicated by more negative values (Fig. 4).

Examination of the factor scores reveal different degrees of correlation between the preservation

Fig. 4. Composite preservation index (CPI) for cores SCS 90-36 and 31-KL. Positive values indicate enhanced preservation and negative values indicate greater dissolution. A line shows the boundary between stage 1 and stage 2 of the $\delta^{18}\text{O}$ stratigraphy.

indices and the CPI (Tables 4 and 5). Squaring the value of each factor score gives the average percentage of variance for each index explained by the CPI. The variance unaccounted for by the CPI indicates that other processes, such as local ecology, sediment transport, and/or winnowing, may affect the carbonate preservation index. In both cores, all indices except for RSP% indicate high values of the factor loadings. In fact, core SCS 90-36 revealed a negative correlation of RSP% with CPI. This relationship indicates that the processes other than preservation/dissolution may affect the variations of RSP%. A recent analysis (Chen, 1994) using Pacific coretop planktonic foraminifer data has already shown a dominant ecological control of thermocline depth which influences the abundance changes of these dissolution-resistant species.

3. Results

Over the past 25 k.y., the CPI curves for the two deep sea cores are quite similar. Both cores show preservation levels that are relatively higher before Termination I (from 25 k.y. through 12 k.y.) compared to those after Termination I. In

addition, we also observe a trend in which the degree of preservation increases steadily from 25 k.y. up to 12 k.y. Carbonate preservation reached maximal levels around Termination I, and then began its steady decline to the present level (Fig. 4). This change in carbonate preservation was also observed in the western equatorial Pacific between 12 and 13 k.y. and has been considered an excellent stratigraphic marker (Berger, 1977; Berger and Killingley, 1977). Close examination indicates that the CPI record from core 31-KL show not only glacial-to-interglacial fluctuations ($1/f$ on the order of 10^4 yr) but also high frequency oscillations ($1/f$ on the order of $\leq 10^3$ yr) (Fig. 4), which suggests there may be multiple sources producing this variability. The presence of these high frequency oscillations is certainly due to the higher sedimentation rates in this core. However, at this stage, we do not rule out the possibility that these oscillations are only random noise signals.

Because several preservation indices were used to extract the CPI, the preservation states of these two cores can not be directly compared. The shallower core SCS 90-36 (2050 m) was drilled well above the modern lysocline (3000 m) so this core is assumed to have better preservation than

31-KL, which was taken from within or below the lysocline (3360 m). When compared by a common index generated for both cores ($WPF\% = (WPF/(WPF + PFF) \times 100)$), we found that prior to Termination I, there was no noticeable preservation difference between these two cores. After Termination I, however, sediments from core 31-KL appear to have experienced stronger dissolution conditions than core SCS 90-36. At this time, the values of $WPF\%$ of core 31-KL sediments decreased from a maximum of 60% to values falling near 20% (Fig. 3). In contrast, the index of core SCS 90-36 sediments exhibited much higher $WPF\%$ values with a maximum value of 90% falling to a minimum of 40%, indicating better preservation (Fig. 2).

The above observation demonstrates a glacial/interglacial preservation fluctuation occurring in SCS sediments. The data indicate that preservation fluctuations in SCS deep sea cores exhibit a Pacific-type pattern with good preservation during glacial periods and poor preservation during interglacial periods. These records show that the preservation fluctuations are not synchronous with the glacial/interglacial change, but appear to be somewhat offset. The period of maximum preservation occurred during the Termination I, a period of rapid deglaciation, rather than during the latter glacial period. Our high resolution records also show that, as indicated by AMS ^{14}C dating and oxygen isotope stratigraphy, this maximum preservation event occurred between 13 and 11 kya. Similar observations of a preservation spike occurring during deglaciation stages have been reported elsewhere, primarily in the tropical Indian and Pacific Oceans (Berger, 1977; Berger and Killingley, 1977; Peterson and Prell, 1985b; Farrell and Prell, 1989, 1991).

4. Discussion

Our results show that over the past 25,000 yr, changes in the level of carbonate preservation occurred in SCS deep sea sediments from about 2000 to 3000 m water depths. These carbonate preservation changes took place during the last glacial-to-interglacial transition period and follow a similar temporal pattern as other preservation

records from the tropical Indian and Pacific Oceans. This indicates that the SCS pattern reflects a regional carbonate response that is probably driven by global changes in deep water circulation, variations in surface water productivity, or a combination of both factors.

The deep water in the SCS is well ventilated through exchange with the western tropical Pacific. The sill that separates the SCS from the western tropical Pacific is estimated to lie at 2500–2600 m, which allows Pacific intermediate and deep waters to enter the SCS (Thunell et al., 1992). A hydrographic study (Gong et al., 1992) of the SCS west of the Luzon Strait also reported similarities of deep water properties between the SCS and western Philippine Sea (WPS), suggesting the presence of inflow through the Luzon Strait from the WPS to the SCS. In this light, the deep sea preservation records from the SCS may also provide a good opportunity to examine variations in Pacific deep water ventilation and chemistry, though the preservation signals in SCS records can be more complicated due to the marginal sea effects of local productivity or organic matter remineralization in the water column.

Deep waters in the Pacific Ocean originate mostly from a mixture of both North Atlantic and Antarctic waters. Though a potential regional source of glacial Pacific deep waters is still under investigation, existing evidence of such a local deep water source is inconclusive (Corliss et al., 1986; Keigwin, 1987; Curry et al., 1988). The processes that control deep water formation in the North Atlantic are very sensitive to climatic changes. In the present-day ocean, NADW forms and accumulates CO_2 oxidized from organic matter descending from surface waters. Through a large-scale transportation system driven by thermohaline circulation (Broecker and Denton, 1989), NADW is transported to the deep basins of the Pacific, and as a consequence the Pacific deep waters are more corrosive so that levels of carbonate preservation are low. During the last glacial stage, however, the production of NADW was diminished relative to that of Antarctic Bottom Water and a general slowing of deep and bottom water circulation occurred. Thus, less CO_2 was transported into deep Pacific waters so that the level of carbonate preservation increased in the glacial deep Pacific.

Our records from the SCS clearly show a Pacific-type of preservation pattern, with good preservation from 25 k.y. through Termination I, and low preservation from the end of Termination I to the present. This would suggest that the SCS preservation pattern is primarily controlled by processes of basin-to-basin fractionation of deep waters as suggested by Berger and Winterer (1974).

Geochemical models (Broecker and Peng, 1987; Boyle, 1988) designed for explaining the glacial-to-interglacial variation of atmospheric $p\text{CO}_2$ have successfully predicted the carbonate preservation spike observed across Termination I in the present SCS study. In the Broecker and Peng (1987) model, the preservation spike in the deep ocean at the time of deglaciation resulted from an adjustment of the ocean carbonate system to nutrient changes. In this model, the rise of atmospheric $p\text{CO}_2$ at the time of the last deglaciation was caused by less efficient surface water productivity and upwelling, which in turn decreased the content of CO_2 and increased the carbonate ion concentration (CO_3^{2-}) in deep waters. The preservation spike recorded in deep sea sediments basically reflects the sudden increase of CO_3^{2-} in deep waters. In response to the sudden increase of CO_3^{2-} , the rate of carbonate removal of the ocean is increased through more rapid burial of carbonate sediments on the sea floor. Restoring the ocean carbonate system to its initial state by this manner took several thousand years (Broecker and Peng, 1987). In the Boyle (1988) model, a mechanism of vertical nutrient transfer was proposed to explain the preservation spike. During the last deglaciation period when the global thermohaline circulation pattern was reorganized, changes in intermediate and deep water production rates caused nutrient and CO_2 to transfer from deep to intermediate waters. This resulted in CO_3^{2-} concentrations in deep waters to increase abruptly, producing a preservation spike in deep sea sediments. As in the model proposed by Broecker and Peng (1987), the carbonate system took several thousand years to restore to its initial state, leading to a gradual increase in atmospheric $p\text{CO}_2$.

These models suggest that carbonate preservation of SCS deep sea sediments is primarily controlled by mechanisms dependent on global ocean processes. The consistency between the temporal

pattern of SCS preservation during the last 25,000 yr and that predicted by global geochemical models (Broecker and Peng, 1987; Boyle, 1988) strongly suggests that the SCS deep waters are well ventilated by inflows from western Pacific deep waters, which is consistent with the observation of regional hydrography.

In addition to the influence of changes in deep water circulation, local productivity variations may also play a significant role in shaping preservation patterns. On examining differences in the preservation levels from cores SCS 90-36 (above modern lysocline) and 31-KL (below modern lysocline), we found that the gradient of dissolution increases after 25 k.y., with the record from core 31-KL exhibiting stronger dissolution than SCS 90-36 after Termination I. The temporal change of preservational gradients within this part of SCS waters (2050–3360 m) may indicate a gradual shoaling of the regional lysocline after 25 k.y. The lysocline shoaling in the SCS may either reflect a decrease in the state of CO_3^{2-} saturation of Pacific deep waters or less deep water exchange between the WPS and SCS. It could also be caused by a decrease of CaCO_3 productivity and/or an increase of organic (siliceous) productivity in SCS surface waters. If productivity changes are indeed one of the important controls in carbonate preservation in the SCS, strong precession-related cycles (23 k.y.) should be found in our records. Recent studies (Huang et al., in press) indicate that the variation of surface ocean productivity in the SCS is closely associated with the intensity of the Asian winter monsoon systems, which are largely regulated by precession cycles as well-documented by many other studies (Prell, 1984a,b; Clemens et al., 1991). However, the present records from cores SCS 90-36 and 31-KL are not long enough to permit spectral identification of precession cycles. Longer records containing more precession cycles and analysis of proxy indices for productivity changes from the sedimentary records will help verify this prediction.

5. Conclusions

A study of variations in carbonate preservation indices over the past 25 k.y. in two SCS cores,

SCS 90-36 and 31-KL, reveal a clear change in carbonate preservation between the last glacial and interglacial periods. A preservation spike that peaked during Termination I, the last deglaciation period (~12,000 k.y.), is also clearly preserved in these records. This overall pattern is similar to that reported in the western Pacific and suggests that similarities in deep water circulation prevailed during the late Quaternary in the SCS and the western Pacific. These data confirm the observation that preservation maxima tend to occur during the latter part of glacial stages, as shown previously in many Indian and Pacific deep sea core studies.

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