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# Asynchronous variation in the holocene asian monsoon recorded by marine sediments and its implication

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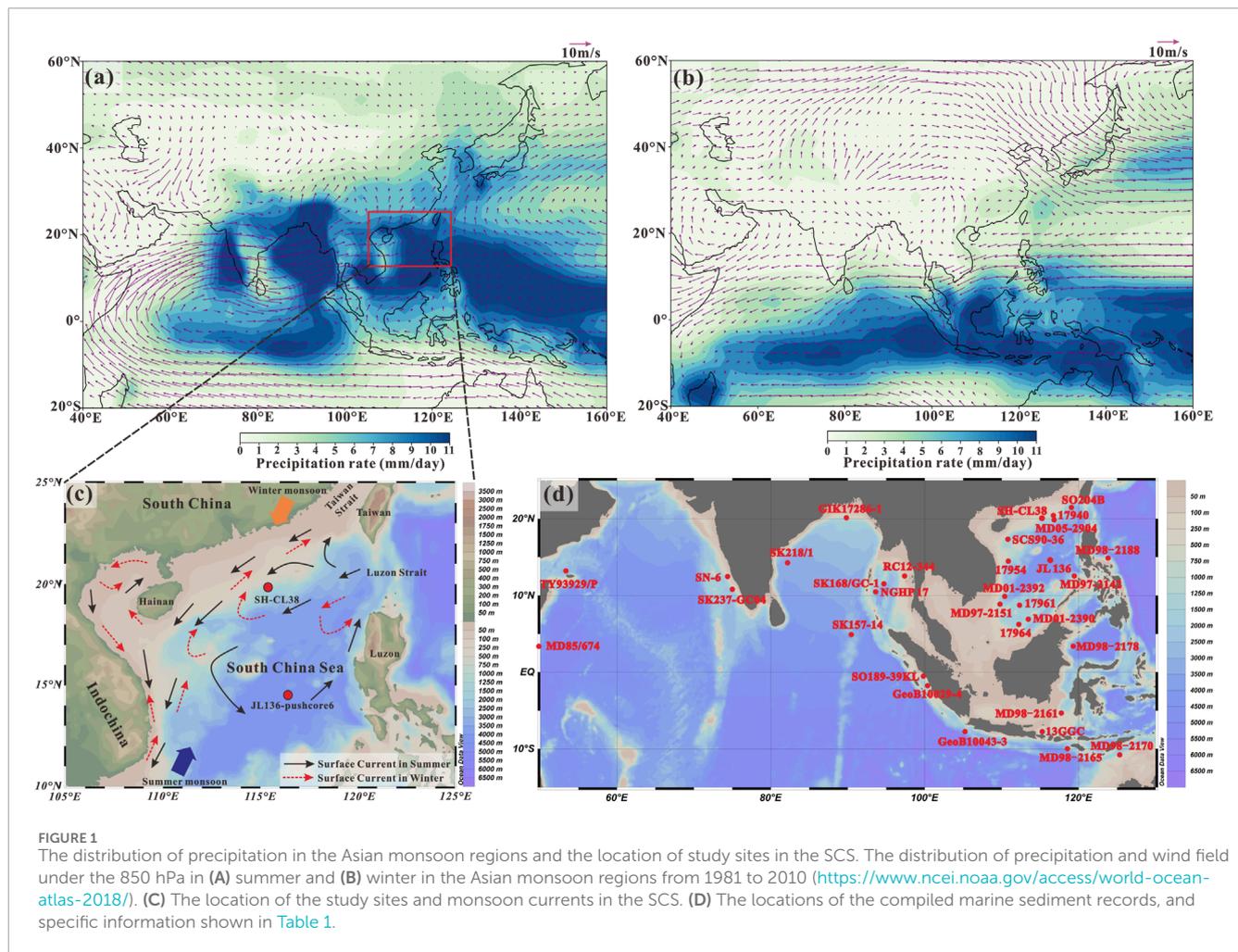
The long-term evolution of climate during the Holocene remains controversial, as proxy and model data, and multiple proxies, show diverging temperature trends between the different reconstructions. Here, we compile sea surface temperature (SST) from multiple marine sediment records in the South China Sea (SCS) and Indo-Pacific over the Holocene, which reveal a phase difference in the precession band of different marine sediment records. Peak identification was performed on the data from each site, and the sea surface temperature anomalies (SSTA) was simply divided into two modes, the Early Holocene (EH-peak) and Middle Holocene (MH-peak), based on the timing of the first maximum peak, using 9 ka as the boundary. The phase difference between the two modes is ~3 ka in the precession band. We suggest that the phase difference corresponds to the shifts in the mean latitudinal position of the Intertropical Convergence Zone (ITCZ) driven by the Northern Hemisphere Summer Insolation (NHSI). Two modes indicate the warming of the SSTA during the late Holocene, which may be attributed to rising  $p\text{CO}_2$ , a strengthening El Niño, and a weakening of the East Asian winter monsoon (EAWM). Furthermore, we observe a partial overlap between the site distribution of the MH-peak and the modern monsoon precipitation domains, which may indicate the shift in the mean latitudinal position of the ITCZ and the dynamics of the monsoon precipitation domains.

## KEYWORDS

holocene, asian monsoon, ITCZ, NHSI, monsoon precipitation domains

## 1 Introduction

The Asian Summer Monsoon (ASM) is a vast system of overturning atmospheric circulation and plays a significant role in the global hydrologic and energy cycles, which consists of two subsystems: the East Asian Summer Monsoon (EASM) and the Indian Summer Monsoon (ISM) (An et al., 2000; Cheng et al., 2021). The ocean has a large heating capacity compared to the continent (Wu et al., 2012). In summer, the Asian continents warm faster than the adjacent oceans, producing an atmospheric



pressure gradient between the ocean and the continent. As a result, warm air carries moisture towards the continent (Figure 1A). In winter, the continents cool faster than the ocean so that the cold and dry air originating from the Siberian high blows from the continent to the ocean (Figure 1B). The ASM system can influence the life and production activities of 60% of the global population (Zhong et al., 2014). Therefore, it is essential to understand the evolution of the monsoon over the geological history.

The Holocene is the major period of human survival and development, whose monsoon climate evolution is crucial to predict the future climate changes. In this way, the research of Holocene monsoon is important. However, the long-term evolution of climate during the Holocene remains controversial, because of often divergent conclusions between multiple proxies and model data, for example, the different temperature trends (Affolter et al., 2019; Baker et al., 2017; Marsicek et al., 2018; Stott et al., 2004). suggests a decreasing trend in temperature over the last 10 ka based on the sediment core records of the western Pacific. In contrast, Marsicek et al. (2018) indicates a warming trend in Holocene temperature based on simulated reconstruction. The main controversies focus on (1) Climate models and pollen-inferred temperature reconstructions showing evidence for lower temperatures during the early Holocene and continuous warming

until the late Holocene in the Northern Hemisphere (Liu et al., 2014b; Marsicek et al., 2018), which challenge previous assumptions of an early to middle Holocene Thermal Maximum (10–6 ka) (Dykoski et al., 2005; Rashid et al., 2007; Selvaraj et al., 2007). (2) The significant phase differences (6–8 ka) in the precession band of Asian monsoon records, observed in a wide range of records varying from marine sediment cores to speleothem  $\delta^{18}O$  records over the Asian continent (Cheng et al., 2021; Clemens et al., 2008). The Asian summer monsoon (ASM) records from the Arabian Sea and other oceans surrounding the Asian continent show a significant lag of  $8 \pm 1$  ka to Northern Hemisphere summer insolation (NHSI) (Clemens et al., 2008; 2010), whereas the Asian cave  $\delta^{18}O$  records lag the NHSI by only  $\sim 2$  ka (Cheng et al., 2016; Cheng et al., 2021). The potential explanations for the asynchronous evolution of the Holocene climate records include underrepresentation of proxies in earlier temperature reconstructions (Marsicek et al., 2018), regional cooling effects related to the remnant ice sheets in the early Holocene (Baker et al., 2017), incomplete forcing or insufficiently sensitive feedbacks of climate model simulations (Liu et al., 2018; Zhang et al., 2018), and seasonal biases between marine records (Liu et al., 2014b; Marsicek et al., 2018). However, the ultimate causes of the asynchronous temperature variations in the Holocene conundrum remain ambiguous.

In order to further understand the asynchronous changes in Holocene temperatures and to provide new insights into solving Holocene conundrum, we estimated sea surface temperature (SST) in the northern and southern South China Sea (SCS) based on the Mg/Ca ratios of planktonic foraminifera shells. Additionally, we compiled more than 30 marine sediment records in the SCS and Indo-Pacific. We compare the compiled results with other Asian monsoon records, expecting to further understand the spatio-temporal distribution characteristics of asynchronous evolution of marine sediment records in the Asian monsoon regions during the Holocene and its implications.

## 2 Materials and methods

Two sediment cores, SH-CL38 (19.9°N, 115.3°E, water depth 1,288 m) and JL136-pushcore6 (14°58.69'N, 116°32.58'E, water depth 2,500 m), were recovered from the northern and central SCS, respectively, during the HYSH201805 cruise and “Jiaolong” 136th dive expedition (Figure 1C). The upper 100 cm of SH-CL38 sediment was described and sampled at 2 cm intervals. However, we only get 31 sediment samples because some samples have been lost in the previous studies, including 0–4 cm, 6–8 cm, 12–14 cm, 22–24 cm, 32–34 cm, 36–38 cm, 42–44 cm, 50–62 cm, 66–70 cm, 72–74 cm, 82–84 cm, 90–92 cm, 96–98 cm. But this does not affect our explanation in this study. The sediments of SH-CL38 primarily consist of silt and clay, with low sand proportion. The core JL136-pushcore6 was described and sampled at 1 cm intervals, resulting in 29 sediment samples. The sediments of the core JL136-pushcore6 mainly consist of gray silty clay with abundant foraminifera. We have further compiled 31 marine sediment records from the Asian monsoon regions to trace the monsoon evolution (Figure 1D; Table 1).

### 2.1 Age models

Accelerator mass spectrometry (Beta Lab, United States)  $^{14}\text{C}$  dates were obtained on mixed planktonic species (*G. ruber* and *G. sacculifer*), and calibrated to calendar age using CALIB8.2 software (Heaton et al., 2020). Based on the Marine Reservoir Correction Database (<http://intcal.qub.ac.uk/calib/>), the regional reservoir correction ages ( $\Delta \pm R$ ) obtained from the five nearby sites were  $5 \pm 52$  years (JL136) and  $-94 \pm 59$  years (SH-CL38), respectively (Yang et al., 2023). The AMS  $^{14}\text{C}$  calibrated results with a  $2\sigma$  error range can offer precise age control for sediment cores. The chronology was based on the results from Bayesian modeling employing the Oxcal program. Subsequently, the Poisson-process deposition model was applied to obtain the age-depth model (Table 2) (Bronk, 2009; Ramsey and Lee, 2013). Temporal intervals between the samples of cores JL136-pushcore6 and SH-CL38 is  $\sim 350$  and  $\sim 280$  years, respectively. The results indicate that the age range of cores JL136-pushcore6 and SH-CL38 is 2.1–12.8 ka, and 1.5–14.4, respectively. In this study, we focus exclusively on Holocene sediment samples for testing and analysis (Figure 2).

### 2.2 Mg/ca ratios

The relationship between calcification temperatures and Mg/Ca ratios in shells (or tests) of foraminifera has been quantified from core-top, sediment trap, and culture studies, resulting in paleotemperature calibration equations (Dekens et al., 2002; Anand et al., 2003; Tierney et al., 2019). This relationship has long been used in paleoclimate reconstructions (Brown and Elderfield, 1996; Nürnberg et al., 1996; Rosenthal et al., 1997; Steinke et al., 2006; Yin et al., 2021; Yang et al., 2023). Typically, SST reconstructions generated from Mg/Ca ratios are obtained from monospecific aggregates of 5–30 shells by the solution method using ICP-AES or ICP-OES (Marchitto, 2006; Rongstad et al., 2017; Groeneveld et al., 2019). In recent years, with the development of testing techniques, Laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) is becoming a widespread technique for analyzing elemental ratios in foraminifera calcite (Eggins et al., 2003; Sadekov et al., 2009; Sadekov et al., 2010; Fehrenbacher et al., 2015). Inter-method comparisons studies comparing average Mg/Ca ratios obtained from LA-ICPMS on discrete chambers and ICP-AES or ICP-OES analysis of complete shells have found that population mean values are consistent (Groeneveld et al., 2008; Sadekov et al., 2009; Jochum et al., 2019; Rustic et al., 2021). Thus, the analysis of Mg/Ca ratios from individual foraminifera using LA-ICP-MS has opened new dimensions in paleoclimate reconstructions, allowing for paleoceanographic reconstruction of oceanic variability (Ford et al., 2015; White et al., 2018; Rustic et al., 2020).

A suitable number (not less than 6) tests of *G. ruber* (s.s.) with sizes ranging from 250 to 350  $\mu\text{m}$  were extracted from each sample of core SH-CL38. In order to avoid the influence of debris materials on the test results, a gentle ultrasonic pre-treatment of the samples with reagent-grade methanol solution was performed with reference to the method of Vetter et al. (2013). The cleaned samples were oven-dried at 40°C. This analysis was carried out at the Key Lab of Submarine Geosciences and Prospecting Techniques, Ocean University of China. Each sample was fixed on a glass slide using carbon tape to facilitate the measurement of laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS). Six erosion points were selected for each foraminifera, with 2 points in each chamber (Figure 2). A high-resolution LA-ICP-MS technique using a pulsed ArF excimer laser ( $k = 193$  nm) coupled to a NexlOn 2000 inductively coupled plasma mass spectrometer (ICP-MS) was employed to generate depth profiles across test walls. Prior to the analysis of the different spot sizes and laser intensities, multiple measurements of the foraminiferal test were conducted to obtain the optimal signal. The optimal laser beam spot diameter was fixed at 32  $\mu\text{m}$ , the energy density at 4 J/cm<sup>2</sup>, and the erosion frequency at 5 Hz (laser shots per second). The total design time of a single point was 105 s, of which the air-carrying blank was 25 s, the erosion time was 40 s, and the washing time was 40 s.  $^{43}\text{Ca}$  was used as an internal standard since all the measurements in this study were obtained from foraminiferal calcite.  $^{44}\text{Ca}$  was used as monitor to ensure consistency, while  $^{27}\text{Al}$  and  $^{55}\text{Mn}$  were used to evaluate the surface contamination and prevent penetrating shells (which produces a diagnostic signal at  $^{27}\text{Al}$ ). In this way, any ratios of Al/Ca >0.10 mmol/mol and Mn/Ca >0.25 mmol/mol were excluded from the average dataset because the contamination always results in higher Al/Ca and Mn/Ca ratios. The NIST SRM

TABLE 1 The information from the marine sediment cores and Asian cave records was used in this study.

Site/Cave	Latitude (N)	Longitude (E)	Water depth (m)	References
SCS9036	17°35.8'	111°17.4'	2050	Huang et al. (1997)
GIK17964	06°09.5'	112°12.8'	1,556	Steinke et al. (2008)
GIK17961	08°30.4'	112°19.9'	1968	Wang et al. (1999)
GIK18252	09°14'	109°23'	1,273	Kienast et al. (2001)
MD01-2,390	06°38.1'	113°24.6'	1,545	Steinke et al. (2008)
MD01-2,392	09°51.1'	110°12.6'	1966	Huang J et al. (2011)
MD97-2,142	12°24.7'	119°16.7'	1,557	Chen et al. (2003)
MD97-2,151	08°43.7'	109°52.2'	1,598	Huang et al. (2002)
MD98-2,178	03°37.2'	118°54'	1984	Fan et al. (2013)
MD98-2,188	14°49.2'	123°29.4'	730	Lin et al. (2006)
SK157-14	05°11'	90°05'	3,306	Raza et al. (2017)
SK168/GC01	11°42.4'	94°29.6'	2064	Gebregiorgis et al. (2016)
SK237/GC04	10°58.6'	74°59.4'	1,245	Saraswat et al. (2013)
GeoB10029-4	-01°58.6'	100°46.1'	964	Mohtadi et al. (2010)
SO189-39 KL	-00°47'	99°54'	517	Mohtadi et al. (2014)
NGHP 17	10°45.2'	93°06.7'	1,356	Gebregiorgis et al. (2018)
MD85674	03°06.1'	50°15.6'	4,875	Bard et al. (1997)
TY93929/P	13°02'	53°05'	2,490	Rostek et al. (1997)
GIK17286-1	19°44.6'	89°52.7'	1,428	Lauterbach et al. (2020)
GIK17938	19°47.2'	117°32.3'	2,840	Chen et al. (1999)
GIK17940	20°07'	117°23'	1728	Huang et al. (2009)
MD05-2,904	19°27.3'	116°15.2'	2066	He et al. (2008)
MD98-2,161	-05°12.6'	117°28.8'	1,185	Fan et al. (2013)
MD98-2,165	-09°39'	118°20.4'	2,100	Levi et al. (2007)
MD98-2,170	-10°21.3'	125°23.3'	832	Stott et al. (2007)
SO50-31 KL	18°45.3'	115°52.4'	3,360	Zhao et al. (2006)
SO204B	21°13'	118°05'	1822	Yang et al. (2019b)
GeoB10043-3	-07°18.6'	105°03.5'	2,171	Setiawan et al. (2015)
13 GG C	-07°14.4'	115°12'	594	Linsley et al. (2010)
SN06	12°29.2'	74°07.8'	589	Tiwari et al. (2015)
RC12-344	12°27.6'	96°02.4'	2,140	Rashid et al. (2007)
SK218/1	14°02'	82°00'	3,307	Govil and Divakar (2011)

(Continued on the following page)

TABLE 1 (Continued) The information from the marine sediment cores and Asian cave records was used in this study.

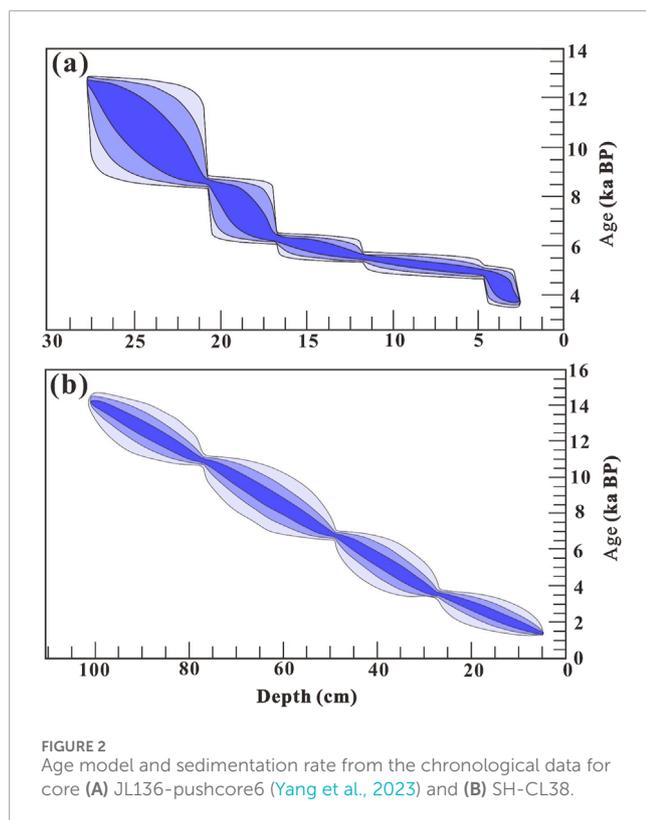
Site/Cave	Latitude (N)	Longitude (E)	Water depth (m)	References
GHE27L	19°51'	115°20.4'	1,533	Yang et al. (2019b)
SH-CL38	19°54'	115°18'	1,288	This study
JL136-pushcore6	14°58.69'	116°32.58'	2,500	Yang et al. (2023)
Jiuxian	38°24'	113°48'	--	Cai et al. (2010)
Sanbao	31°00'	110°36'	--	Cheng et al. (2016)
Heshang	30°27'	110°25'	--	Hu et al. (2008)
Heifeng	29°40'	106°17'	--	Yang et al. (2019a)
Dongge	25°17'	108°5'	--	Dykoski et al. (2005)
Lianhua	38°24'	113°48'	--	Dong et al. (2015)
Mawmluh	25°5'	91°2'	--	Dutt et al. (2015)
Bukit Assam	4°17'	114°55'	--	Carolin et al. (2013)
Tangga	-0°19'	100°43'	--	Wurtzel et al. (2018)
Klang	8°18'	98°42'	--	Chawchai et al. (2021)
Gempa Bumi	-5°18'	120°42'	--	Krause et al. (2019)

TABLE 2 AMS-<sup>14</sup>C ages and calendar ages of core SH-CL38 and JL136-pushcore6 (Yang et al., 2023).

Sample ID	Sample depth (cm)	Materials	AMS <sup>14</sup> C age (yr BP)	Calendar age (cal. yr BP)	2 sigma age (cal. yr BP)
SH-CL38	4–6	<i>G. ruber</i> (s.s.)	2070 ± 25	1,478	1,272–1,684
	26–28	<i>G. ruber</i> (s.s.)	3,790 ± 30	3,595	3,358–3,832
	48–50	<i>G. ruber</i> (s.s.)	6,555 ± 40	6,856	6,581–7,132
	76–78	<i>G. ruber</i> (s.s.)	10,070 ± 30	10,948	10,695–11,202
	100–102	<i>G. ruber</i> (s.s.)	12,745 ± 45	14,357	13,922–14,792
JL136-pushcore6	3–4	<i>G. ruber</i> (s.s.) + <i>G. sacculifer</i>	4,105 ± 30	3,982	3,707–4,258
	5–6	<i>G. ruber</i> (s.s.)	5,045 ± 25	5,149	4,879–5,420
	12–13	<i>G. ruber</i> (s.s.)	5,560 ± 25	5,758	5,543–5,973
	17–18	<i>G. ruber</i> (s.s.)	6,260 ± 30	6,512	6,280–6,744
	21–22	<i>G. ruber</i> (s.s.)	8,430 ± 30	8,808	8,549–9,067
	27–28	<i>G. ruber</i> (s.s.)	11,550 ± 40	12,877	12,656–13,098

610 and 614 soda-lime glass standard were employed to quantify the internal variability within each group and assess the relative differences among groups. In this study, two measurements on per chamber were conducted to calculate the average Mg/Ca ratio for

each foraminifera. The mean Mg/Ca variability in the chamber was about ±0.16 mmol/mol. Finally, the mean Mg/Ca variability from all measurements (at least 30 measurements in one sample) was determined of ±0.37 mmol/mol.



For core JL136-pushcore6, Mg/Ca ratios were analyzed using Arian 720ES ICP-OES in the Key Laboratory of Ocean and Marginal Sea Geology, SCS Institute of Oceanology Chinese Academy of Science. Specific pre-treatment and test parameters are described in Yang et al. (2023).

We converted Mg/Ca ratios to SST using a calibration,  $\text{Mg/Ca (mmol/mol)} = 0.38\text{exp}[0.09\text{SST (}^\circ\text{C)}]$  (Dekens et al., 2002). Testing of Mg/Ca ratios and reconstructed SST can be found in Supplementary Table S1.

## 2.3 Compilation of marine sediment records

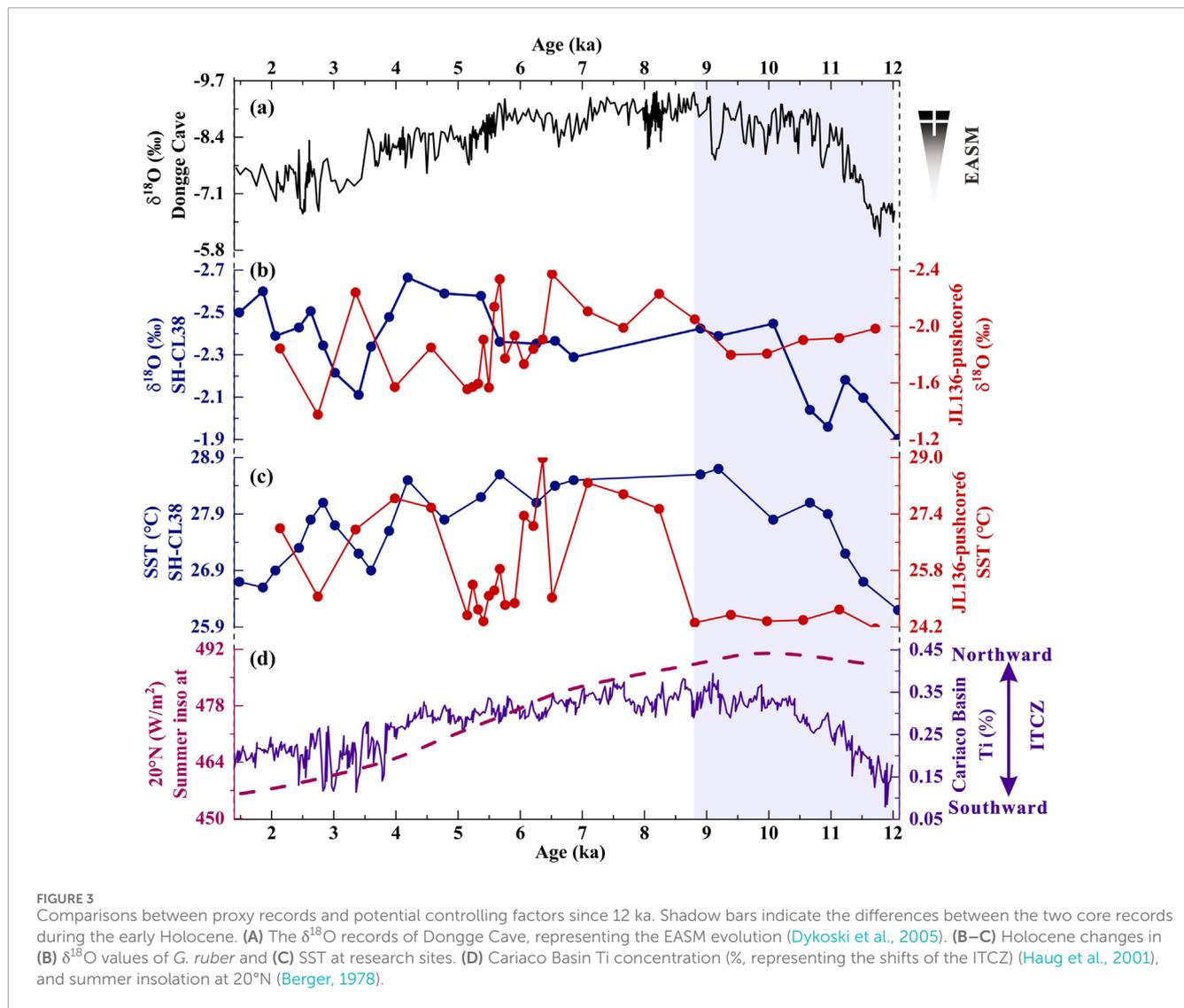
There are a lot of studies on the evolution of the Monsoon evolution (Affolter et al., 2019; Böll et al., 2015; Lo et al., 2013; Raza et al., 2017; Yang et al., 2023; Yin et al., 2021). However, the reconstructed SST values in the studies can be affected by possible inter-laboratory and inter-calibration biases and different cleaning protocols (Dang et al., 2020), which result in the difficulties to reach a consensus. To minimize this biases and differences, we calculated the SST anomalies relative to the average value of each temperature record over 6 to 10 ka, which is denoted as the sea surface temperature anomalies (SSTA) (Supplementary Figure S1). Given the differences between the resolution of samples from different sites, we have selected to bin the SSTA into 500-year non-overlapping bins (Supplementary Figure S2) (Linsley et al., 2010). This composite reconstruction is based on binning of the original data and not interpolation, and the treatment of the data is referenced from Linsley et al. (2010).

## 3 Results and discussion

### 3.1 Asynchronous evolution of the SCS sediment records

In the northern SCS, the results show that the SST range from 26.2°C to 28.7°C at the site SH-CL38 over the last 12 ka, with a rapid increase from 26.2°C to 27.9°C during 12–11 ka. Then, from 11 to 5.4 ka, higher SST values were observed, with an average of 28.3°C. After 5.4 ka, there was a gradual decline in SST, with a mean value of 27.5°C (Figure 3). The general trend shows similarity with the reconstructed SST using alkenone, planktonic foraminiferal  $\delta^{18}\text{O}$  values, and Mg/Ca ratios at 17,940 (Wang et al., 1999), SCS90-36 (Huang et al., 1997), ODP1144 (Wei et al., 2007), and KNG5 (Huang E et al., 2011) sites in the northern SCS. In addition, these records have a similar trend to the stalagmite  $\delta^{18}\text{O}$  values of Dongge Cave, with a rapid warming in the early Holocene, followed by a warmer and wetter middle Holocene, and gradual cooling in the Late Holocene (Cheng et al., 2019; Dong et al., 2010). In contrast, in the central SCS, the SST ranges from 24.2°C to 29.0°C at the site of JL136-pushcore6. The SST (24.2–24.7°C) was low from 11.7 to 8.8 ka, followed by rapid increase from 24.3°C to 29.0°C during 8.8–6.1 ka and a gradual decrease during the late Holocene (after ~6.1 ka). This SST trend indicates that the sediment record in the central SCS did not significantly warm during the early Holocene and rapidly warmed in the middle Holocene. until the mid-Holocene. The warming observed in the sediment record of the central SCS lags that of the northern SCS by ~3 ka (Figure 3).

The evolution trend of the northern SCS sediment records is consistent with the stalagmites  $\delta^{18}\text{O}$  values of Dongge Cave (Cheng et al., 2019; Dong et al., 2010; Dykoski et al., 2005) (Figure 3A). The stalagmites  $\delta^{18}\text{O}$  values were interpreted as a comprehensive representation of the entire water vapor transport process, reflecting the intensity of the EASM across the continent (Cheng et al., 2019; Cheng et al., 2021; Ruan et al., 2019). Thus, the northern SCS sediment records reflect EASM variability (Oppo and Sun, 2005; Yin et al., 2021; Yokoyama et al., 2011). The EASM intensity variations inferred by the stalagmites  $\delta^{18}\text{O}$  values are consistent with NHSI changes, revealing that the EASM is mainly driven by insolation forcing on orbital timescale (Cheng et al., 2019; Wang et al., 2001). The NHSI changes along with the movement of the mean latitude of the Intertropical Convergence Zone (ITCZ) (Cheng et al., 2019; Dong et al., 2018; Fleitmann et al., 2007). During the high NHSI period, the overall northward shifts of the ITCZ is accompanied by a weakened Northern Hemisphere Hadley cell and a strengthened Southern Hemisphere cell, increasing cross-equatorial airflows (Chiang and Friedman, 2012; Du et al., 2021). This results in an enhanced sea-land temperature and pressure gradient, ultimately leading to a strengthening of the EASM and an increase in monsoon precipitation, whereas the opposite effect is observed during the low NHSI (Cheng et al., 2021; Du et al., 2021). Therefore, the rapid warming recorded in the northern SCS during the early Holocene was influenced by an enhanced EASM with high NHSI forcing (Chiang et al., 2015; Kong et al., 2017). The simulation results of the precipitation differences between high NHSI minus low NHSI show that the southerly monsoon over East Asia enhanced at high NHSI along with the enhancement of the western Pacific Subtropical High (Bosmans et al., 2018). This has resulted in the movement of the main precipitation belts deeper into the continent and a decrease

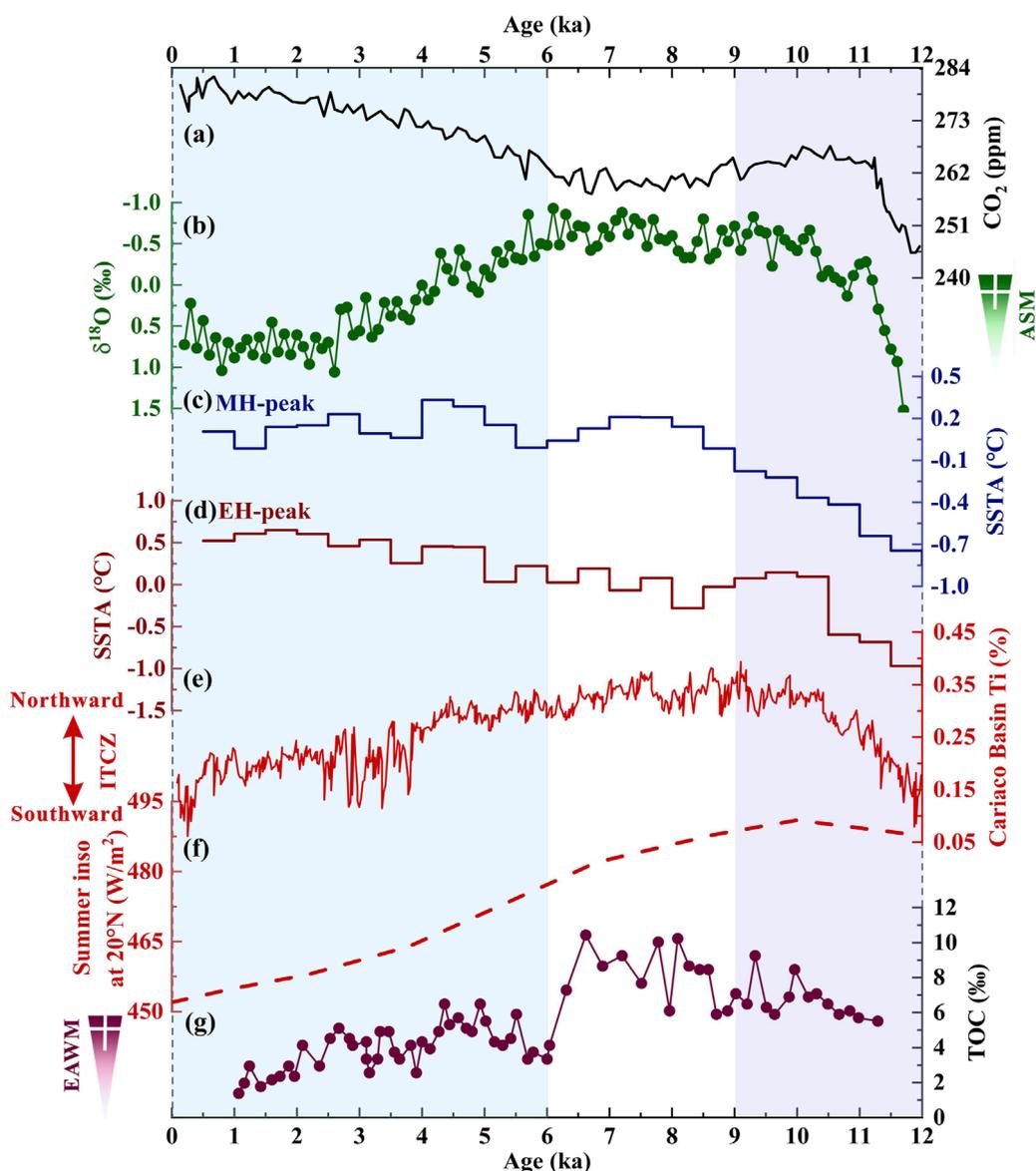


in oceanic precipitation around the southeastern-south of the Asian continent (Bosmans et al., 2018; Huang et al., 2020; Lee et al., 2019; Wang et al., 2016). The northern SCS continuously warmed, with an increasing monsoon rainfall during the early Holocene (Cheng et al., 2019; Dong et al., 2018; Huang et al., 2016). Geographically, the central SCS is closer to the equator, but further away from Asian continents compared to the northern SCS. The asynchronous SST evolution of the Early Holocene in the northern and central SCS records may be a response to the northward shift of the ITCZ forced by the high NHSI (Bosmans et al., 2018; Kong et al., 2017; Zhang et al., 2021). During the middle Holocene, warm and humid Holocene optimums were recorded by cores from both northern and central SCS (An et al., 2000; Zhou et al., 2016). The warm and humid Middle Holocene optimum is also recorded by other proxies in the Asian monsoon region, such as lake sediments (Jia et al., 2015; Rao et al., 2016), Asian cave stalagmite  $\delta^{18}\text{O}$  records (Dong et al., 2018), and loess-paleosol records (Jia et al., 2021; Wang et al., 2014), which may be attributed to the variation of the NHSI. As the NHSI gradually decreases, the ITCZ begins to retreat southwards. During this period, the tropical monsoon trough and the main precipitation belts shift southward, leading to a decrease in total rainfall at the northern edge of the Asian monsoon domain, while the

overall climate in the SCS region is warm and humid (Anwar et al., 2018; Fleitmann et al., 2007; Jia et al., 2021; Zhou et al., 2016).

### 3.2 The different evolution modes of holocene climate

We present a compilation of more than 30 marine sediment records from the Asian monsoon regions to further understand the asynchronous variation of the Asian monsoon records in the precession band and its implications (Table 1). We identified peaks in the data from each station and simply divided the SSTA into two modes, based on the timing of the first maximum peak (Supplementary Figure S3). Because the first rapid warming occurred mainly in the Early and Middle Holocene, we use 9 ka as the boundary and divide it into two modes, the Early Holocene peak (EH-peak) and the Middle Holocene peak (MH-peak). To more clearly describe the patterns of the temperature variation, we calculated its mean SSTA values (Figure 4). The EH-peak mode shows that the mean SSTA continuously warms by  $\sim 0.9^\circ\text{C}$  since  $\sim 12$  to 10.5 ka, followed by a stable period during 10.5–5 ka, with warms by  $\sim 0.3^\circ\text{C}$  after  $\sim 5$  ka. The



**FIGURE 4** Comparisons between proxy records and potential controlling factors during the Holocene. Purple bars indicate differences between the two modes during the early Holocene. Blue bars indicate changes in the two modes during the Late Holocene. **(A)** Holocene changes in CO<sub>2</sub> concentration (Lüthi et al., 2008). **(B)** The anomalous variation of stalagmite δ<sup>18</sup>O records for several caves in the Asian monsoon regions, representing the ASM evolution (c–d) 500 years non-overlapping binned average SSTA in **(C)** MH-peak mode and **(D)** EH-peak mode. **(E)** Cariaco Basin Ti concentration (%), representing the shifts of the ITCZ (Haug et al., 2001). **(F)** Summer insolation at 20°N (Berger, 1978). **(G)** Total organic carbon (TOC) concentration from the Huguangyan Maar Lake, Guangdong province, regarded as an indicator of the EAWM with high/low values indicating strong/weak EAWM (Jia et al., 2015).

MH-peak mode shows that the first peak of mean SSTA occurs at ~8 ka, which lags the EH-peak model by ~3 ka in the precession band. Then the mean SSTA kept stable after 8 ka (Figure 4).

### 3.3 Potential influence factors on different modes

The long-term evolution of climate during the Holocene remains controversial, because we can get the different temperature trends based on the different reconstruction methods (Affolter et al., 2019).

Stott et al. (2004) found a decrease in SST of 0.5°C in the western tropical Pacific Ocean over the past 10 ka using three sediment cores. Linsley et al. (2010) found a cooling trend during the late Holocene in the western tropical Pacific warm pool by eight sediment cores. Marcott et al. (2013) proposed that the climate kept warming during the early-middle Holocene (11.3–5 ka), with a gradually decreasing temperature in the late Holocene from 73 globally distributed records. However, there are also contrary opinions. The SST reconstructions in the eastern equatorial Pacific and SCS show a gradual warming from the early to late Holocene, and the Mg/Ca-SST of the Northwestern Atlantic also shows a relatively stable or

slightly warming during the Holocene (Leduc et al., 2010). Liu et al. (2014b) suggest that the SST warms from the early Holocene and did not cool until the Little Ice Age (1,300–1,850 AD) based on three climate models, which simulated the global and regional mean temperatures over the Holocene. In addition, the results also show warming from the early Holocene up to ~2 ka by reconstruction of Northern Hemisphere temperatures from sub-fossil pollen in North America and Europe (Marsicek et al., 2018). Our EH-peak and MH-peak modes show an increase trend in SSTA after ~6 ka, which supports the argument of Late Holocene warming (Figure 4) (Leduc et al., 2010; Liu et al., 2014b; Marsicek et al., 2018).

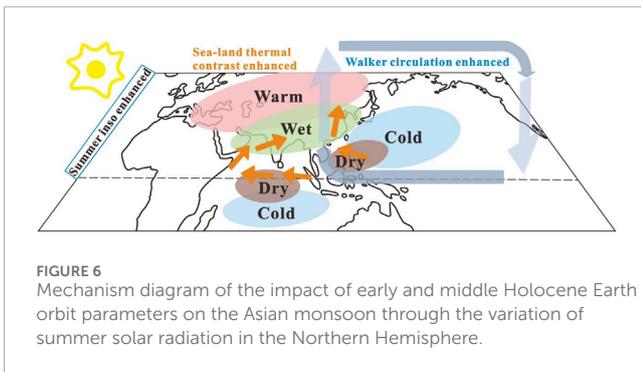
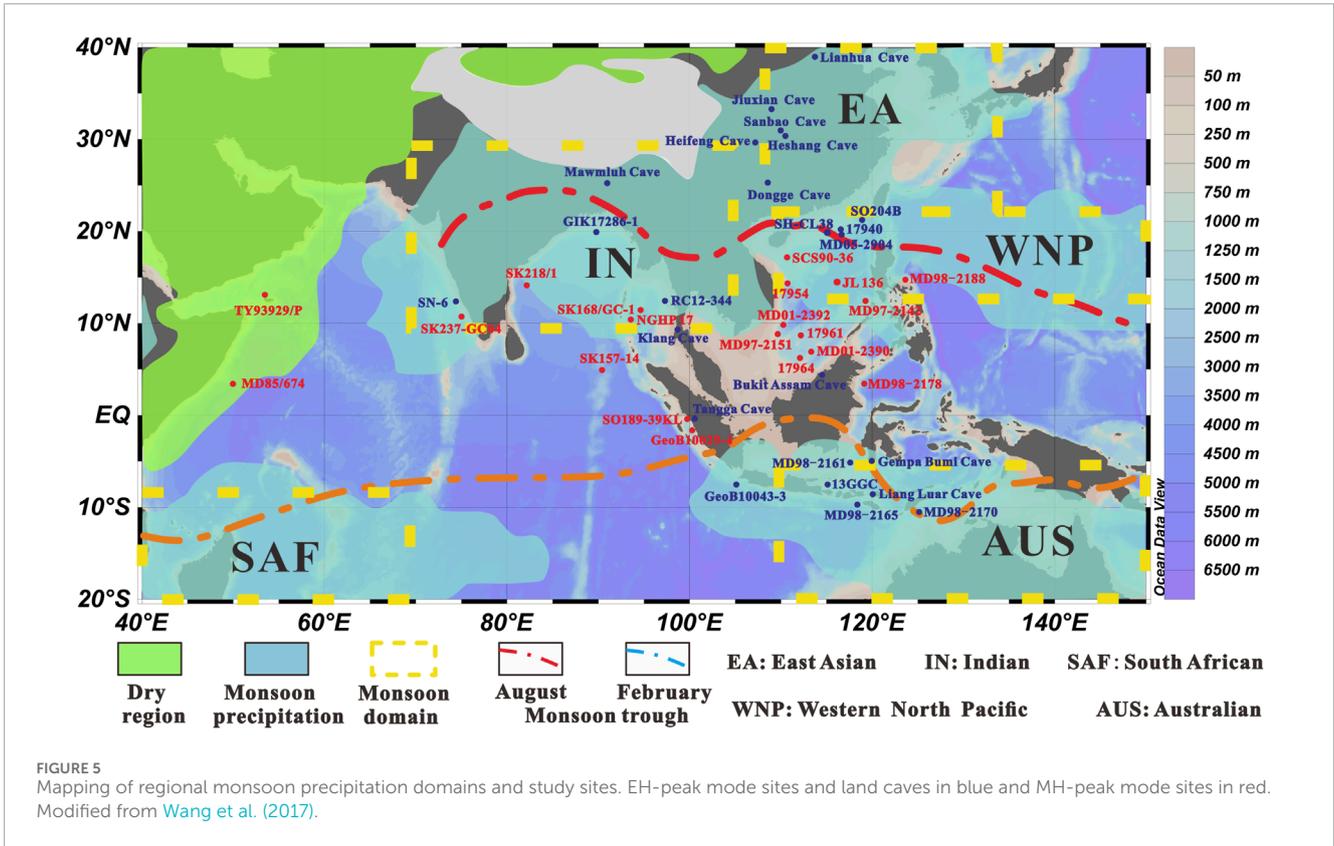
It has been proposed that the monsoon climate is primarily driven by the solar radiation intensity, which can be mainly reflected by NHSI, caused by the Earth's orbital parameters on orbital timescale since the Quaternary (Imbrie et al., 1992). The NHSI reached a Holocene maximum at ~11.5 ka and then decreased, which may controlled the overall trend of the ASM (Laskar et al., 2004). The EH-peak mode is mainly controlled by the NHSI, with a rapidly increase of mean SSTA, which reached the first peak in the early Holocene, and kept relatively stable in the middle Holocene (Figure 4). The MH-peak mode shows that the mean SSTA varied lags the NHSI by ~3 ka. It gradually increased during the early Holocene and reached its first peak in the middle Holocene. The solar radiation variations can cause seasonal migration of the ITCZ (Dong et al., 2010). During the early Holocene, the intensified summer solar radiation in the Northern Hemisphere contributed to an increase in the interhemispheric thermal gradient, which was accompanied by a rapid northward shift of the ITCZ (Chiang and Friedman, 2012). This shift facilitated the transport of more warm water vapor from the Indian Ocean to the Asian Monsoon region, thereby enhancing the EASM and leading to increased monsoon precipitation and warmer temperatures (Chawchai et al., 2021; Fleitmann et al., 2007; Wu et al., 2020). During the middle and late Holocene, the decreasing northern hemisphere summer solar radiation intensity weakens, can lead to the southward shift of the ITCZ and a gradually decreasing ASM intensity (Dong et al., 2010; Fleitmann et al., 2003; Fleitmann et al., 2007).

Both mentioned modes indicate that the mean SSTA warmed during the late Holocene, contrary to the weakening summer monsoon due to the reduction of northern hemisphere summer insolation. This indicates that other factors may influence the SSTA (Huang J et al., 2011; Laskar et al., 2004). Previous studies suggest that SST variations in the Late Holocene were significantly influenced by the East Asian winter monsoon (EAWM) (Kong et al., 2014; Yang et al., 2019b). The EAWM induces strong stirring of seawater, resulting in a net loss of heat from the sea surface to the air, which would lead to a decrease in SST (Chen et al., 1999; Yang et al., 2019b). A weakened EAWM may weaken the mixing of the upper water layer, resulting in a warming of the SST (Yang et al., 2019b). Jia et al. (2015) revealed a weakening of the EAWM from the early to late Holocene, with the most significant transition at ~6 ka, based on the biogeochemical records from Huguang Maar Lake. Kaboth-Bahr et al. (2021b) also revealed a weakening of the winter monsoon from ~5 ka until ~2 ka by stacking marine and terrestrial proxy records. The weakened EAWM was influenced by the enhanced El Niño activity in the late Holocene (Dong et al., 2021). Enhanced El Niño activity suppresses the EAWM through southerly anomalies generated by anticyclonic circulation anomalies

in the western Pacific Ocean (Dong et al., 2021; Huang et al., 2004). The weakened EAWM was accompanied by southeasterly anomalies, which transport warm and humid air from the tropical oceans northward, resulting in increased monsoon humidity and a gradual warming of temperatures (Ma et al., 2018). So, the combination of strengthened El Niño and weakened EAWM causes increased monsoon rainfall and temperatures in southeastern and SCS. In addition, CO<sub>2</sub> as a greenhouse gas has a significant influence on climate change (Wen et al., 2016). The atmospheric CO<sub>2</sub> contents continuously rose since ~6 ka, resulting in the infrared radiation difficulty escape from the atmosphere and lack of heat exchange, leading to a thermal insulation effect and atmospheric warming (Hu et al., 2000; Wen et al., 2016). Observation and model results suggest a negative correlation between atmospheric CO<sub>2</sub> contents and the Asian winter monsoon intensity under global warming conditions (Hori and Ueda, 2006; Kaboth-Bahr et al., 2021b; Li et al., 2019). Therefore, the warming of the mean SSTA during the Late Holocene may have been influenced by the combination of increased CO<sub>2</sub> contents, a strengthened El Niño, and a weakened EAWM.

### 3.4 Implications of the holocene climate mode for the asian monsoon regions

The EH-peak and MH-peak modes have a regular distribution based on the compilation of the sediment records in the different study sites (Figure 5). The EH-peak mode mainly occurred in the oceans surrounding the Asian continent, while the MH-peak mode mainly occurs in the oceans away from the Asian continent (Figure 5). Dong et al. (2010) suggested that shifts in the mean position of the ITCZ control the monsoon intensity throughout the entire low-latitude region of Asia on the orbital timescales based on the Asian cave δ<sup>18</sup>O records. The shift in the mean latitudinal position of the ITCZ are forced by the intensity of the NHSI (Fleitmann et al., 2003; Fleitmann et al., 2007), which reached the Holocene maximum with weak winter insolation intensity during the early Holocene (Laskar et al., 2004). The northern hemisphere is located at perihelion and aphelion in summer and winter, respectively, which can lead to further seasonal contrast in insolation (Wang et al., 2008). This process, in turn, can enhance the sea-land thermal contrast, and eventually increase summer and winter monsoon strength (Wang et al., 2008). In addition, the intensity of summer insolation in the Northern Hemisphere was much stronger than in the Southern Hemisphere during the Early Holocene, which strengthened the cross-equatorial airflows, resulting in a further northward shift in the mean position of the ITCZ (Chawchai et al., 2021; Fleitmann et al., 2007). This transports large amounts of warm water vapor from the Indian Ocean into the Asian monsoon region, resulting in increased monsoon precipitation and temperatures in the interior Asian continent and along the southern coasts (Chawchai et al., 2021; Fleitmann et al., 2007; Wu et al., 2020). Meanwhile, the oceans away from the Asian continent can only accept the relatively less influence of warm and humid air. During the middle Holocene, the NHSI weakened and winter insolation intensity gradually increased, which resulted in a weakening sea-land thermal contrast, and a gradually southward shifting in the mean position of the ITCZ (Zhang et al., 2021). Under this condition, the oceans away from the Asian continent should be strongly influenced by the warm and humid air, which can be observed from the EH-peak and



MH-peak modes (Figure 5). Therefore, the EH-peak and MH-peak modes should be caused by the northern hemisphere solar radiation variability, and the phase differences in the precession band may reflect different aspects of the monsoon dynamics, which is also proposed by Cheng et al. (2019) and Dong et al. (2010).

According to the modern definition of monsoon climate, precipitation becomes an important limited factor in the delineation of the monsoon domains (Wang et al., 2017). The Asian monsoon regions mainly contain the Indian, the western North Pacific, and the East Asian monsoon precipitation domains (Yim et al., 2014). According to the distribution of the modern monsoon precipitation domains, the EH-peak sites are located within the monsoon precipitation domains, while the MH-peak sites are mainly outside the precipitation domains, with parts of the sites at the edge of the precipitation domains (Figure 5) (Yim et al., 2014). These distributions are mainly related to the seasonal

migration of the ITCZ (Fleitmann et al., 2007; Wang et al., 2017), accompanied by the migration of the latitudinal zone of wind convergence and precipitation, which determines the onset, duration, and termination of the rainy season in the tropics and subtropics (Fleitmann et al., 2007). Kaboth-Bahr et al. (2021a) inferred that the mean position of the ITCZ has shifted southward or northward 10°–15° since the 20 ka from its current position based on the variations of terrigenous input fluxes into the northern SCS. The shift in the mean position of the ITCZ are accompanied by monsoon rainfall variability (Fleitmann et al., 2007). Liu et al. (2014a) modeled the precipitation difference between middle Holocene (6 ka) and late Holocene (1 ka) and proposed that the precipitation was strengthened in the Asian continent but weakened in the oceans away from the Asian continent compared to the present, revealing that the mean position of the ITCZ is more northerly than present in the middle Holocene. During the early Holocene, the NHSI experienced a Holocene maximum, with strong insolation seasonal contrasts and enhanced sea-land thermal contrasts, which led to a further amplified the northward shift in the mean position of the ITCZ, and the monsoon precipitation domains was more northerly than in the present (Fleitmann et al., 2007; Wang et al., 2008). In addition, the westward shift of the tropical Pacific Walker circulation also further strengthens the Asian monsoon, resulting in a warmer and wetter climate on the Asian continent, while oceans around the Asian continental are also affected (Figure 6) (Tian et al., 2018; Wang et al., 2020). The distributional characteristics of the EH-peak and MH-peak mode sites coincide with the northward shift in the mean position of the ITCZ during the early Holocene. The partial MH-peak sites

located in the modern monsoon precipitation domains indicate that the mean position of the ITCZ was further north compared to the present  $\sim 24^\circ\text{N}$  during the early Holocene (Kaboth-Bahr et al., 2021a). During this period, the Asian continent and the southern coastal areas are located within the monsoon precipitation domain, where the monsoon precipitation and temperature increase. In contrast, the ocean areas far from the continents had relatively cooler temperatures and less monsoon precipitation. In the Middle and Late Holocene, as the NHTSI decreased, the ITCZ gradually shifted southward, accompanied by a southward shift in the monsoon precipitation domain (Chawchai et al., 2021; Fleitmann et al., 2007). This resulted in increased monsoon precipitation and temperature at some oceanic areas far from the continent. The partial overlap between the site distribution of the MH-peak and the modern monsoon precipitation domains seems to corroborate the latitudinal shift of the ITCZ. Therefore, we suggest that the difference between the two modes, may indicate a change in the mean latitudinal position of the ITCZ and the dynamics of the monsoon precipitation watershed.

## 4 Conclusion

We analyze the core sediment records from the northern and central SCS and find that there are significant phase differences not only between the marine sediment records and the Asian cave  $\delta^{18}\text{O}$  records in the precession band but also between the marine sediment records. We re-analyze SST data from marine sediment records in the Asian monsoon regions and establish two modes, EH-peak and MH-peak. The results show that the mean SSTA peak of the MH-peak mode lags the EH-peak mode by  $\sim 3$  ka in the precession band. On the Holocene long-term trend, the two climate modes respond to the northern hemisphere solar radiation variability, and the phase differences in the precession band may reflect different aspects of the monsoon dynamics. Two modes indicate warming of the SSTA during the late Holocene, which may be attributed to rising  $p\text{CO}_2$ , a strengthening El Niño, and a weakening of the EAWM. In addition, the distribution of the two Holocene modes sites responds to the distribution of monsoon precipitation domains and is implication the dynamics of the monsoon precipitation domains.

## Data availability statement

All relevant data is contained within the article: The original contributions presented in the study are included in the article/Supplementary Material.

## Author contributions

JY Writing–original draft, Writing–review and editing. GZ: Writing–review and editing. YnZ: Writing–review and editing. ZZ:

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Formal Analysis, Methodology, Writing–review and editing. SL: Formal Analysis, Methodology, Writing–review and editing. HW: Formal Analysis, Methodology, Writing–review and editing. XG: Formal Analysis, Methodology, Writing–review and editing. GC: Formal Analysis, Methodology, Writing–review and editing. LY: Formal Analysis, Methodology, Writing–review and editing. YrZ: Formal Analysis, Methodology, Writing–review and editing. SL: Writing–review and editing.

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## Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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## Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2024.1493790/full#supplementary-material>

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