Millennial-scale dynamics of the winter cold tongue in the southern South China Sea over the past 26 ka and the East Asian winter monsoon

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A B S T R A C T

Millennial-scale variations of the East Asian winter monsoon (EAWM) remain elusive due to sparse and controversial reconstructions. By compiling a variety of alkenone-based sea surface temperature (SST) estimates, we find that the west–east SST gradient in the southern South China Sea (SCS) well documents the temporal dynamics of the winter “cold tongue” off the southern Vietnam and by inference, variations in the EAWM intensity over the past 26 ka. Our results reveal that the winter “cold tongue” SSTs were significantly colder during Heinrich event 1 and the Younger Dryas event, resulting in an increased west–east SST gradient in the southern SCS due to a strengthened EAWM. Within dating uncertainties, an intensified EAWM during cold stadials was coeval with the shutdown or a reduction in strength of the Atlantic meridional overturning circulation (AMOC), exhibiting a strong linkage between the AMOC and the EAWM system. The west–east SST gradient also indicates an enhanced EAWM during the early Holocene, which may be induced by postglacial ice-sheet dynamics and a strong seasonal contrast in solar insolation. Our findings suggest that the EAWM was probably modulated by a complex interplay between the AMOC, solar insolation and ice-sheet dynamics on sub-orbital time scales.

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Introduction

Abrupt climate change during the last glaciation and the Holocene has been observed at both high and low latitudes. Trigger mechanisms for abrupt climate change and processes propagating these signals globally are thought to reside either in the ocean or in the atmosphere. Hypotheses attribute such changes to reorganizations of the ocean thermohaline circulation and/or to changes in tropical atmosphere–ocean dynamics (Broecker, 2003). One attempt to test such hypotheses is unraveling the relationship between the Northern and Southern Hemisphere high-latitude climate and the Asian monsoon climate (Rohling et al., 2009). Within dating uncertainties, the high latitude North Atlantic and the Asian monsoon climate systems exhibit a very similar millennial- and centennial-scale variability over the last glacial and interglacial periods, as recorded in Greenland ice cores (Johnsen et al., 1992; Dansgaard et al., 1993; North Greenland Ice Core Project members, 2004) and cave deposits from China (Wang Y. et al., 2001, 2005, 2008; Yuan et al., 2005), Yemen (Burns et al., 2003) and Oman (Fleitmann et al., 2003), as well as in North Atlantic (Bond et al., 1993, 2001) and Indian Ocean (Schulz et al., 1998; Altabet et al., 2002; Gupta et al., 2003) marine sediments.

Which process (or processes) is (are) responsible for the linkage between these two climate systems? Modeling experiments carried out by Barnett et al. (1988) show a strong influence of late-lasting snow cover on Eurasia to weaken the following summer monsoon. This suggests that a long-lasting winter snow cover on Eurasia during glacial periods might have weakened the Asian summer monsoon. Denton et al. (2005) further suggested that changes in wintertime climate were probably the common linkage between Greenland–European temperatures, the Intertropical Convergence Zone (ITCZ) and Asian monsoons. According to Denton et al. (2005), the expansion of sea ice across the North Atlantic during cold stadials, in particular during Heinrich events, resulted in much colder winters than normal. The cold wintertime anomalies in the North Atlantic realm might have amplified and propagated a signal of abrupt climate change throughout the Northern Hemisphere and into the low latitudes. Through atmospheric reorganizations and teleconnections, these cold anomalies most likely weakened the Asian summer monsoon but enhanced the winter monsoon. This hypothesis provides a promising way to explain the stadial and interstadial switches revealed in monsoon proxy records.

Testing such a hypothesis requires reliable reconstructions of both the summer and winter monsoon. The development and evolution of the East Asian summer monsoon (EASM) on millennial and centennial time scales has been well-reconstructed from cave deposits of southern and southwestern China for the past 360 ka (Cheng et al., 2009). However,
the East Asian winter monsoon (EAWM) has left nearly no signature in the hydrological cycles and thus has been proven difficult to reconstruct. Studies on loess–paleosol sequences revealed that the EAWM was unstable on millennial time scales (Porter and An, 1995; An and Porter, 1997; Chen et al., 1997; Liu and Ding, 1998; Xiao et al., 1999; An, 2000; Yu et al., 2006). Strengthened episodes of the EAWM may be coeval with cold stadials or Heinrich events in high northern latitudes (Porter and An, 1995; An and Porter, 1997; Yu et al., 2006). However, paleomonsoon records derived from the loess sequences are considered to be unreliable on sub-orbital time scales because of disconformities and dating issues (Stevens et al., 2007). In the Sulu Sea, periods of increased primary productivity over the last glaciation as well as during Heinrich events were interpreted to be caused by intensified EAWM winds (de Garidel-Thoron et al., 2001).

Recently, high-resolution sediment records from Lake Huguang Maar of southern China revealed that the EAWM strengthened during Heinrich event 1 and the Younger Dryas (YD) periods but weakened during the warm Bølling–Allerød (B–A) interval (Yancheva et al., 2007; Wang L. et al., 2008). However, the interpretation of the titanium content of the Lake Huguang Maar sediments in terms of changes in the EAWM intensity (Yancheva et al., 2007) was challenged by the recent study of Zhou et al. (2009). Zhou et al. (2009) have shown that instead of being transported by winds, as suggested by Yancheva et al. (2007), the titanium most likely come from the catchment area of Lake Huguang Maar. Consequently, variations in the titanium content are most likely controlled by the hydrology in the lake catchment area rather than by the strength of the EAWM winds. Therefore, the millennial-scale variability of the EAWM remains elusive.

Oceanographic setting

Surface water circulation in the SCS is predominantly driven by the East Asian monsoon winds. The southwest EASM appears in May and reaches its maximum strength during July and August, while the northeast EAWM can last from October through March (Shaw and Chao, 1994). As demonstrated by modern observations and simulations, the southwest EASM drives a basin-scale anticyclonic circulation in the SCS while the northeast EAWM drives a cyclonic circulation, both being associated with a strong western boundary current along the Vietnamese coast (Shaw and Chao, 1994; Chu et al., 1999; Chu and Wang, 2003; Liu et al., 2004; Fig. 1). In summer, the boundary current moves northward and splits into two branches between 11°N and 14°N, with one branch going across the central basin towards the northeast and the other continuously flowing northward along the Vietnamese coast (Chu et al., 1999; Wang et al., 2006; Fig. 1a). The winter boundary current flows to the south and reaches the southern tip of Vietnam between October and March (Chu et al., 1999; Fig. 1b). The more intense EAWM wind produces a much stronger boundary current in winter than in summer (Chu et al., 1999).

Modern SST distribution patterns in the SCS are largely controlled by the seasonal reversal of the monsoon-driven surface water circulation. During the EASM season, SSTs are quite uniform over the entire basin with temperatures around 28.5°C except for the summer upwelling area off the middle Vietnamese coast (Fig. 2a). The winter boundary current transports cold surface waters from the northern SCS to the south along the Vietnamese coast, resulting in a distinct “cold tongue” area in the southern SCS (Liu et al., 2004; Fig. 2b). Thus, the winter SST distribution exhibits a strong north–south temperature gradient over the entire SCS basin and an evident west–east temperature gradient in the southern SCS (Fig. 2b). The development of the “cold tongue” is greatly determined by the wind-driven southward cold surface water advection, which fully develops in November–February and then dissipates as the northeast EAWM declines (Liu et al., 2004).

Figure 1. Mean surface circulation over the SCS derived from the Princeton Ocean Model (after Chu et al., 1999): (a) boreal summer; (b) boreal winter. Dash lines denote the direction and track of the summer and winter boundary currents, respectively.
Monthly mean time series of the “cold tongue” SST and the southern SCS west–east SST gradient are tightly correlated with the surface vector wind speed as shown for the time period 1995–2005 (Kalnay et al., 1996), with a correlation coefficient R of 0.70 and −0.62, respectively (Fig. 2c). In general, SSTs vary between 28.5 and 30.0°C during summer (May–August) but decrease to 25.5–27.0°C during winter (December–February) when the “cold tongue” appears. The west–east SST gradient is fairly small (0–0.8°C) in summer but increases up to 1.3–2.2°C during the EAWM season. Colder SSTs and a higher west–east SST gradient are always associated with more intense EAWM winds (Fig. 2c). Stronger EAWM winds transport more cold waters from the northern SCS to pile up off the southern Vietnamese coast, which are then diverted to the southeastern SCS through mixing and diffusing (Liu et al., 2004). The outspread of the cold upper ocean waters expands the “cold tongue” area and increases the west–east SST gradient. It is noteworthy here that the “cold tongue” SST variability is also modulated by El Niño events. For instance, during the strong El Niño years of 1997/1998 and 2002/2003, the anomalous anticyclonic circulation over the Indo-western Pacific weakened the northeast EAWM over the SCS, leading to a weakened boundary current and a warm winter SST in the “cold tongue” area (Fig. 2c; Liu et al., 2004).

We further compare the winter west–east SST gradient with winter surface vector wind speed during the time period 1948–2009 (Fig. 2d; Kalnay et al., 1996). These two data sets also reveal a marked correlation on interannual time scale, with a correlation coefficient R of −0.52. Over the past 60 yr, the EAWM wind speed over the SCS increased from −6.8 m/s to −7.8 m/s, and the west–east SST gradient increased from 1.33°C to 1.53°C. The stronger EAWM years are basically associated with a greater west–east SST gradient in the southern SCS, and vice versa. Therefore, we propose that the temporal variability of this winter “cold tongue” and the west–east SST gradient in the southern SCS can be used to record past changes of the EAWM intensity.

**Strategy and proxy variable used for reconstructing “cold tongue” dynamics**

Accurate and reliable SST estimates are needed in order to investigate the dynamics of the winter “cold tongue” in the southern SCS over the past 26 ka. Planktonic foraminifera transfer functions have been widely used to estimate past seasonal SST changes in the SCS (e.g., Chen et al., 2005). However, planktonic foraminifera transfer functions have been proven to be inapplicable to estimate glacial SSTs in the SCS (e.g., Chen et al., 2005). Planktonic foraminifera transfer functions have been proven to be inapplicable to estimate glacial SSTs in the SCS (e.g., Chen et al., 2005). The application of the regional calibration equation allows an SST estimate with an accuracy of ±0.5°C (Pelejero and Grimalt, 1997). In addition, SST determinations become less accurate when SST index values get higher because of difficulties in the quantification of C37:3 alkenone. Therefore, in this study we only use measurements with SST index values lower than 0.97 for core MD01-2390 (Fig. 3a).

Except for core MD01-2390, SST–SST records from cores located within the modern “cold tongue” area (MD97-2151, 18252, MD01-2390; Figs. 2b and 3b) and outside the “cold tongue” area (18287, 17961, 17964; Figs. 2b and 3a) show very similar changes both in amplitude and temporal pattern over the past 26 ka, respectively. We thereby group these SST records in order to derive an SST stack for the southwestern and the southeastern SCS, respectively (Fig. 3c). Stacks could retain their common features but suppress stochastic noise in each record. The following steps were performed to produce SST stacks: (1) SST records with a time resolution better than 0.4 ka (Table 1) are smoothed at a scale of 0.4 ka. (2) The smoothed and non-smoothed SST records were interpolated to a new evenly spaced time series with a resolution of 0.4 ka. (3) In each group, the stack was obtained by taking average means of coring sites 18252, MD97-2151, 18252, MD01-2390; Figs. 2ba and 3a) show very similar changes both in amplitude and temporal pattern over the past 26 ka, respectively. This suggests that the EAWM season with predominant in the cold months (winter and spring) when the EAWM prevails (Chen et al., 2007). This suggests that the EAWM season with predominant in the cold months (winter and spring) when the EAWM prevails (Chen et al., 2007). This suggests that the EAWM season with predominant in the cold months (winter and spring) when the EAWM prevails (Chen et al., 2007). This suggests that the EAWM season with predominant in the cold months (winter and spring) when the EAWM prevails (Chen et al., 2007).
some common effects both on the western and eastern SST variations in the southern SCS, such as SST and sea-level changes over glacial-interglacial cycles, the intrusion and shut-off of the Indo-Pacific warm water into the SCS through the Sunda Shelf, freshwater input and its impacts on sea surface salinity and possibly further on alkenones records. Therefore, several non-monsoon effects on SST variations could be removed from SST gradient records. As the SST series and stack estimates have an accuracy of ±0.5°C, the accuracy for SST gradient estimates should be ±1.0°C. Although an uncertainty of ±1.0°C is relatively large compared to the west-east SST gradient variations over the past 26 ka (Fig. 3d), we consider using multi-SST sequence and the stack method could still yield meaningful results. The west-MDO1-2390 SST gradient shows a much higher estimate of 2.6–4.6°C between 18.9 and 9.1 cal ka BP (Fig. 3d), which is far beyond the modern winter SST gradient of ~1.0°C (November–February). Moreover, the west-east and west–MD01-2390 SST gradient records show consistent temporal patterns within the time period of 18.9–9.1 cal ka BP, which also confirms the reliability of our estimates.

In order to get better alignments of the millennial-scale events observed in different SST records, we shifted the original age models of SST records. We took the defined time intervals for Heinrich event 1 (17.5–14.7 cal ka BP), the B–A (14.7–12.8 cal ka BP) and the YD event (12.8–11.7 cal ka BP) in stalagmite δ18O records (Wang et al., 2001) as our tuning target. All radiocarbon-based age models (Table 1) were adjusted within radiocarbon dating uncertainties in each core. Two non-radiocarbon-based age models (core MD97-2151 and MD01-2392) were also tuned to fit those of other SST records. Afterward we followed the above methods to obtain new SST series, SST stacks and SST gradient records (dashed lines in Figs. 3c and d), which are basically consistent with original results both in temporal patterns and amplitudes. This confirms that the SST gradient results are not greatly biased by age-model uncertainties.

### SST gradient variations and implications for the “cold tongue” and EAWM development

As shown in Figures 3a and b, the winter $\Delta^2$SST estimates show an LGM (Last Glacial Maximum, 23–19 cal ka BP, Mix et al., 2001) to Late Holocene difference of 5.0–6.0°C. Although planktonic foraminifera transfer functions are inapplicable in the glacial southern SCS (Steinke et al., 2008a), they work well in the northern SCS. The 5.0–6.0°C warming estimate here is comparable with the winter LGM–Late Holocene SST contrast of 4.0–7.0°C as inferred from planktonic foraminifera transfer-function estimates in the northern SCS (Huang et al., 1997; Chen and Huang, 1998; Chen et al., 1999; Wang et al., 1999; Wang, 1999). The SST estimates of core MD01-2390 show 1.0–1.5°C higher temperatures than other southeastern records between 18.9 and 9.1 cal ka BP. We infer that winter surface waters over this core location were less influenced by “cold tongue” water than other cores during this time period because its modern position is the furthest away from the “cold tongue” area (Fig. 2b).

In the southeastern SST records (except for core MD01-2390), the last deglacial warming was characterized by an abrupt increase in temperature that corresponds to the onset of the B–A interval. The warming trend was then punctuated by a subtle cooling episode between 12.8 and 11.5 cal ka BP, coeval with the YD event in the North Atlantic region (Fig. 3a). In the southwestern and core MD01-2390 SST records, the deglacial warming was punctuated by two cooling episodes between 17.5 and 14.8 cal ka BP and between 12.6 and 10.8 cal ka BP (Fig. 3b). These two cooling events are coeval with Heinrich event 1 and the YD event, respectively. In addition, a cooling of 1.5–2.0°C between 25 and 23 ka was also observed in core 17964 and MD97-2151 that is approximately coeval with Heinrich event 2. However, the occurrence of Heinrich event 2 in SST records is not verified by any radiocarbon ages.

As shown in Figures 3c and d, a large west-east SST gradient of 1.6–2.1°C is observed during the time period 25–23 ka, which might correspond to Heinrich event 2. The winter southwestern SCS was relatively warm between 22.0 and 18.0 ka, leading to a small SST gradient of 0.5–1.2°C. This gradient is similar to or even lower than the modern winter SST gradient of ~1.0°C. Explicit discrepancies between these two SST records occur during the last deglaciation, resulting in a large SST gradient of 1.5–2.3°C during Heinrich event 1 and the YD event, and an intermediate SST gradient of 1.0–1.7°C during the B–A interval. The west–east SST gradient shows a gradual decrease from 2.2°C to 0.4°C between 11.0 and 4.0 cal ka BP. Thereafter, it increases towards the modern value of ~1.0°C. The west–MD01-2390 SST gradient shows a similar temporal pattern compared to the west–east one between 18.9 and 9.1 cal ka BP. However, the amplitudes of changes are much higher. The west–MD01-2390 SST gradient is around 3.5–4.6°C during Heinrich event 1, 3.0–3.3°C during the B–A 3.0–4.0°C during the YD event and a decreasing trend from 4.0°C to 3.0°C in the early Holocene.

Higher SST gradients during Heinrich event 1 and the YD event indicate a strengthened winter “cold tongue” that is most likely caused by an enhanced EAWM. The intermediate SST gradients during the B–A interval and the early Holocene also imply a stronger EAWM intensity relative to the late Holocene (Fig. 3d).

### Controls on the EAWM intensity over the past 26 ka

#### The EAWM during the LGM

Interestingly, the small SST gradient in the southern SCS between 22.0 and 18.0 ka might indicate a weaker EAWM during the LGM that is comparable to the late Holocene condition. This result is in conflict with previous findings (see below). Previous modeling studies suggested that the Siberian High and the Northern Hemisphere westerlies have been strengthened and/or displaced southward under full glacial conditions (Kutzbach et al., 1993), resulting in a strengthened EAWM during the LGM. Numerous terrestrial and marine records also suggest an intensified EAWM during the LGM.

# Table 1

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Water depth (m)</th>
<th>Modern winter mean SST (November–February; °C)</th>
<th>SST sequences time resolution (ka)</th>
<th>References</th>
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</thead>
<tbody>
<tr>
<td>MD97-2151</td>
<td>09°14’N</td>
<td>109°23’E</td>
<td>1273</td>
<td>26.43</td>
<td>0.26</td>
<td>Kienast et al. (2001)</td>
</tr>
<tr>
<td>MD01-2392</td>
<td>09°51’3’N</td>
<td>110°12’E</td>
<td>1966</td>
<td>26.84</td>
<td>0.72</td>
<td>Xie et al. (2007)</td>
</tr>
<tr>
<td>18287c</td>
<td>05°39’N</td>
<td>110°39’E</td>
<td>598</td>
<td>27.57</td>
<td>0.19</td>
<td>Kienast et al. (2001), Steinke et al. (2001)</td>
</tr>
<tr>
<td>17961</td>
<td>08°30’N</td>
<td>112°19’E</td>
<td>1908</td>
<td>27.55</td>
<td>0.64</td>
<td>Pelejero et al. (1999a,b)</td>
</tr>
<tr>
<td>17964b</td>
<td>06°09’N</td>
<td>112°12’E</td>
<td>1556</td>
<td>27.77</td>
<td>0.46</td>
<td>Pelejero et al. (1999a,b)</td>
</tr>
<tr>
<td>MD01-2390b</td>
<td>06°38’12”</td>
<td>113°24’56”</td>
<td>1545</td>
<td>27.84</td>
<td>0.22</td>
<td>Steinke et al. (2010)</td>
</tr>
</tbody>
</table>

**Notes:**

- **a** Mean winter SST data are from World Ocean Atlas, 2005 (Locarnini et al., 2006).
- **b** Sediment records with AMS 14C dating.
- **c** The age model of core MD97-2151 is derived from Lee et al. (1999).
and a predominant ice volume control on the EAWM variability on orbital time scales (Fig. 3f; e.g., Chen and Huang, 1998; Liu and Ding, 1998; de Garidel-Thoron et al., 2001; Tian et al., 2006). Here we address several possibilities to reconcile this conflict.

First, it should be noted that the published LGM SST records (Table 1) are not fixed by any radiocarbon data. An accurate calculation of the LGM west–east SST gradient requires well-dated LGM SST estimates from both the southwestern and southeastern SCS in future studies. A second possibility is related to the glacial sea-level lowering and its possible effect on the SST distribution patterns in the SCS. During the LGM when the eustatic sea level was around 120 m below the present level (Fig. 3g), the SCS lost almost half of its size (Fig. 4). Due to the exposure of the Sunda Shelf, the winter southward

![Figure 4](image_url)
western boundary current could not enter into the Java Sea but flowed back to the southeastern SCS along the glacial coastline, forming a counterclockwise surface circulation (Fig. 4; Wang et al., 1995). The glacial sea-level low stand and the establishment of a counterclockwise surface circulation may have favored a uniform SST distribution pattern in the southern SCS. However, it should be noted that the Sunda Shelf was completely flooded and the southern straits were opened only after 8.0 cal ka BP (Fig. 3g; Geyh et al., 1979; Hanebuth et al., 2000). Strong SST gradients during the last deglaciation indicate lowering sea level and glacial surface circulation pattern could not be a dominant control on the west–east SST gradient changes, although they might tend to reduce SST gradients.

A third possibility is that the EAWM intensity during the LGM was probably not that strong as proposed by earlier studies. Although the state of the AMOC during the LGM is still a matter of debate, it seems that the strength of the AMOC was only slightly weaker (Fig. 3e; McManus et al., 2004; Gherardi et al., 2005; Lynch-Stieglitz et al., 2007). This may have favored a northward shift of the ITCZ (Chiang and Bitz, 2005), leading to a longer duration of the EASM over southern Asia. Furthermore, a comparison between the Hulu Cave and polar ice-core records revealed that the Asian monsoon was predominantly controlled by the Southern Hemisphere climate variability during the LGM through changes in the cross-equatorial air airflow (Rohling et al., 2009). The inflow of moisture from the Southern Hemisphere resulted in a relatively stronger EASM precipitation in eastern China during the LGM, which is comparable to that during the late Holocene (Wang et al., 2001). Accordingly, the southern SCS was probably dominated by the air inflow from the southern ocean during the LGM. The prevailing southwest monsoon wind with a longer duration could have suppressed the formation of the “cold tongue,” leading to a smaller west–east SST gradient in the southern SCS. However, this final most attractive LGM scenario requires additional west–east SST gradient reconstructions covering the LGM interval and evidence from other marine and terrestrial studies.

The EAWM during the last deglaciation

The west–east SST gradient over the last 26 ka reveals a strengthened EAWM during Heinrich event 1 and the YD event (Fig. 3d). Our results are in good agreement with previous studies from the SCS (see below). An increased north–south SST gradient over the SCS due to a stronger EAWM was observed during Heinrich event 1 and the YD event (Tian et al., 2010). The strengthened EAWM also led to a stronger mixing of the upper ocean water column in the northern (Chen et al., 1999) and southern SCS (Steinke et al., 2010) during these two deglacial stadial periods. In addition, the abundance of the planktonic foraminifera Neogloboquadrina pachyderma (dextral), a proxy of the intrusion of cold waters from the open Pacific through the Luzon Strait forced by the EAWM, significantly increased in the SCS during Heinrich event 1 (Pflaumman and Jian, 1999; Huang et al., 2002; Steinke et al., 2008a; Xiang et al., 2009). This also indicates a strengthened EAWM at that time.

The reconstructed SST gradient in the southern SCS (Fig. 3d) greatly resembles the 231Pa/239Th records from the North Atlantic Ocean, a kinematic proxy for the intensity of the Atlantic meridional overturning circulation (AMOC, Fig. 3e; McManus et al., 2004; Gherardi et al., 2005). Although the 231Pa/239Th record of core GGC5 from the Bermuda Rise is questioned to be influenced by particle composition and flux (Lippold et al., 2009), a parallel particle-flux analyses on core GGC5 does not support such an argument (McManus et al., 2004). The similarity of the SST gradient records and the 231Pa/239Th records indicates that the intensified EAWM during cold stadges is most likely related to changes in the state of the AMOC. The distinct increase of the SST gradient during Heinrich event 1 and the YD event correspond to a shutdown and a slowdown of the AMOC, respectively. In contrast, the relatively smaller SST gradient during the B–A warming interval is coeval with a rapid resumption of the AMOC.

How did changes in the AMOC affect the EAWM system? During the shutdown or reduction in the strength of the AMOC, the northern high latitudes experienced a marked cooling while the low latitudes maintained warm or experienced only a subtle cooling (Rahmstorf, 2002). Expanded winter sea ice coverage in the subarctic North Atlantic prevented heat release from the ocean and thus further intensified winter cooling in the Northern Hemisphere high latitude (Denton et al., 2005). The severe winter cooling could enhance cold–air subsidence in the northern high latitudes and strengthen the inner–land surface high. This combined a strengthened meridional temperature gradient favored the transport of cold streams southward across the subtropical and tropical areas, resulting in strengthened EAWM winds. In addition, the southward retreat of the ITCZ during Heinrich events and the YD event (Wang et al., 2001; Griffiths et al., 2009) most likely has resulted in a longer duration of the EAWM over southern Asia.

The EAWM during the early Holocene

The larger west–east SST gradient between 11 and 7.2 cal ka BP suggests a stronger EAWM during the early Holocene compared to the mid-late Holocene (Fig. 3d). This finding is in relatively good agreement with diatom flux data from Lake Huguang Maar in southeastern China that also reveal a stronger EAWM between 10 and 8.5 cal ka BP and between 7.0 and 6.0 cal ka BP (Wang L. et al., 2008). Recent published studies based on loess and lacustrine sediments also indicate an arid environment and a strong EAWM over northern China and central Asia before 8.0 cal ka BP (Fig. 3e; Stevens et al., 2007; Chen et al., 2008; Mason et al., 2009). Stalagmite δ18O records from southern China imply stronger summer monsoon precipitation during the early Holocene (e.g., Yuan et al., 2005; Wang et al., 2005). Based on these findings, it seems that the variability of the EAWM and the EASM are not anti-correlated on a millennial time scale during the early Holocene.

We propose that two factors may have caused the prolonged stronger EAWM after the last deglaciation. These are most likely related to the presence of residual ice sheets (Fig. 3g) and the slow warming of the Northern Hemisphere high latitudes (Kaplan and Wolfe, 2006). The delayed response of the ice sheets to the solar insolation may have enhanced the Siberian High during the early Holocene winter and thus strengthened the EAWM. The orbital configuration as follows may have also played an important role: the boreal summer was at the perihelion while the boreal winter was at the aphelion during the early Holocene, which might have amplified seasonal contrast in insolation. This likely enhanced the thermal contrast between the ocean and the Eurasian continent, which favored the establishment of both a strong summer and winter monsoon (Wang L. et al., 2008).

Conclusion

Based on historic data, we find that the winter west–east SST gradient variability in the southern SCS is highly correlated with the EAWM intensity. We further present a west–east SST gradient record over the past 26 ka based on several U37C––SST records from the southern SCS. Stronger SST gradients during Heinrich event 1 and the YD event indicate an enhanced EAWM. The stronger EAWM during these deglacial stadial periods coincides with the shutdown or reduction in strength of the AMOC. An intermediate SST gradient during the B–A warming corresponds to a resumption of the AMOC intensity. The west–east SST gradient also indicates a stronger EAWM during the early Holocene, which is most likely due to residual ice sheets in the Northern Hemisphere and an enhanced seasonal contrast in solar insolation during the early Holocene. Our new reconstructions suggest that the EAWM system exhibits a complex response to various forcing mechanisms, such as the AMOC, solar insolation and ice volume, operating over different time periods and scales. Our study suggests that the winter west–east SST gradient in the southern SCS is a promising proxy for
EAWM variations. Longer alkenone records from the southern SCS which cover the LGM and marine oxygen isotope stage 3 should be obtained in order to investigate millennial-scale EAWM variations further back in time.

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