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CONNECTION OF THE SOUTH CHINA SEA SUMMER MONSOON TO MARITIME CONTINENT CONVECTION AND ENSO

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Abstract: The relationship between the intensity of the South China Sea summer monsoon (SCSSM) and the Nino3.4 index and anomalous atmospheric circulation patterns associated with a strong and weak SCSSM are investigated using the NCEP/NCAR reanalysis data, Extended Reconstructed Sea Surface Temperature (ERSST) data and Climate Prediction Center Merged Analysis of Precipitation (CMAP) data. The SCSSM is significantly positively correlated with the Nino3.4 index in the succeeding northern autumn and winter. In the strong minus weak SCSSM composite, a positive East Asia-Pacific teleconnection (EAP) pattern and a negative Europe-Asian-Pacific teleconnection (EUP) pattern appear in the 500 hPa height difference field; low-level cross-equatorial flows are strengthened over the Maritime Continent (MC) region; positive (negative) precipitation anomalies occur in the South China Sea and western north Pacific (MC). A possible mechanism through which SCSSM affects ENSO is proposed. A strong (weak) SCSSM strengthens (weakens) cross-equatorial flows over the MC. The anomalous cross-equatorial flows cool (warm) the SST around the MC through enhanced (reduced) surface latent heat fluxes. The cooling (warming) further leads to suppressed (enhanced) convection over the MC, and causes the anomalous westerly (easterly) in the equatorial western Pacific, which favors the onset of El Niño (La Niña) through modulating the positive air-sea feedback process.

Key words: South China Sea summer monsoon; cross-equatorial flow; Maritime Continent; anomalous westerly: ENSO

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1 INTRODUCTION

The planetary-scale thermal circulation caused by the seasonal variation of incoming solar radiation is the primary driving forcing for the tropical monsoon, whereas the semi-permanent planetary wave due to the seasonal difference between the thermal characters of the earth surface (i.e., the ocean and continent) is another important forcing. The two forcings are in phase in the eastern hemisphere, resulting in the pronounced Asian-Australia monsoon[1]. The strong correlation between the Indian summer monsoon and an ENSO index in the subsequent winter indicates a possible active role of the Indian summer monsoon in modulating the ENSO[2].

During boreal spring, the equatorial zonal wind and zonal SST gradient are weakest. The onset of the Asian summer monsoon may invigorate the weak coupled air-sea system. This may be partially responsible for the spring prediction barrier[3]. The lateral monsoon circulation over the tropical Indian Ocean may affect the air-sea interaction over the equatorial Pacific through a pair of gear operating over the equatorial Indian and Pacific (GIP) Oceans and may even trigger the occurrence of ENSO events (e.g., Wu and Meng[4]).

Lau and Wu[5] investigated the covariability of the Asian summer monsoon and ENSO and found that the

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first mode can account for 47% of total variance. Characterized by a pronounced biennial variability, the rainfall pattern is dominated by negative anomalies over the western Pacific, the MC, the equatorial eastern Indian Ocean, and the southern Bay of Bengal, and positive anomalies over the northern Bay of Bengal, south China, and the western Indian Ocean. The SST field is corresponding to the developing phase of El Niño in the Pacific. A strong Asian winter monsoon may sometimes lead to positive SST anomalies in the equatorial Pacific Ocean[6, 7]. Wang et al.[8] discussed the possible role of the South China Sea (SCS) in the decaying of ENSO. The above results suggest that both the South Asian summer monsoon and the East Asian winter monsoon may trigger the ENSO events.

Climatologically, the onset of the SCS summer monsoon (SCSSM) occurs in the middle of May[9, 10]. This onset time occurs in the spring-to-summer transition period, when both the Walker cells over equatorial Indian and Pacific Ocean are weakest. This might be the reason why the interannual variation of the intensity of SCSSM has a great impact on the atmospheric circulation[11]. So far the possible impact of SCSSM on ENSO is seldom explored. The goal of this study is to investigate the observed relationship between the intensity of SCSSM and the Nino3.4 index, the difference of large-scale circulation between strong and weak SCSSM years, and the possible mechanism through which SCSSM affects the ENSO evolution.

2 DATA AND METHODS

The NCEP/NCAR monthly atmospheric reanalysis dataset with a resolution of $2.5^\circ\times2.5^\circ$[12], monthly ERSST with a resolution of $2^\circ\times2^\circ$[13] and CMAP rainfall dataset with a resolution of $2.5^\circ\times2.5^\circ$[14, 15] are used. The intensity of SCSSM is defined by mean southwesterly wind velocity at 850 hPa averaged over the SCS (10°N– 20°N, 110°E– 120°E) following Wu and Liang[16], and is computed using the NCEP/NCAR monthly reanalysis dataset. A potential problem of the NCEP/NCAR reanalysis is the low quality of the data over Asia prior to 1968, because only very few SLP data were used in the reanalysis before 1968[17]. To avoid this problem, we only use the 1968-2003 reanalysis data in the following analysis. The Nino3.4 index is defined in the same way as that of the NOAA climate prediction center (CPC, http://www.cdc.noaa.gov/ClimateIndices/). For individual ENSO events, readers are referred to the following website: http://www.cpc.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml. The cross-equatorial flow over the MC is defined as the integral of meridional winds at 850 hPa from 80°E to 120°E. The Australian High is defined based on the mean sea level pressure field averaged over (120°E – 150°E, 25°S – 35°S)[18]. Both the correlation analysis and the composite analysis are applied. While the correlation analysis is used to reveal the relationship between the intensity of SCSSM and the Nino3.4 index, the composite analysis is used to reveal the circulation difference between the strong and weak SCSSM.

3 RELATIONSHIP BETWEEN SCSSM INDEX AND NINO3.4 INDEX

Based on the SCS wind index, the mean intensity of SCSSM is about 5m/s, whereas the root square variance is about 1m/s. Figure 1 shows the time series of the SCSSM index. As one can see, there is a pronounced interannual variability. A strong (weak) SCSSM is defined when the index is more (less) than 1 (–1) standard deviation. With this criterion, five strong SCSSM years and six weak SCSSM years are chosen. The five strong summer monsoon years are 1972(2.7, normalized anomaly), 1985(1.5), 1994(1.1), 1997(1.5) and 2002(1.2), and the six weak summer monsoon years are 1980(–1.6), 1983(–1.0), 1988(–1.3), 1989(–1.2), 1996(–1.8) and 1998(–1.6).

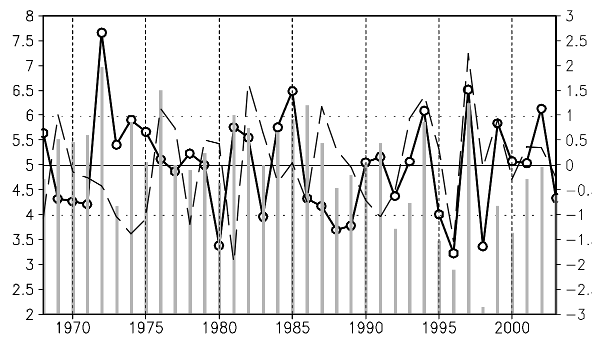


Fig.1 Time series of the SCSSM index (thick, solid line with circles with the unit (m/s) in the left; horizontal, dotted line stands for 1 standard deviation), normalized cross-equatorial flow at 850 hPa over the MC (bar), and the Australia High index (thin, dash line) from 1968 to 2003.

Figure 2 shows the lead-lag correlation coefficient between the SCSSM and Nino3.4 indices. Note that a significant negative correlation between the SCSSM index and the Nino3.4 index occurs from September to March before the onset of SCSSM and a significant positive correlation occurs from October to February after SCSSM. The significant negative correlation before the onset of SCSSM suggests that ENSO may influence the circulation over SCS, which is consistent with the result of Liang and Wu[19] and is explained by Chang et al.[20] with the role of subtropical ridges and

Wang et al.[21] with the Pacific-East Asian teleconnection.

The significant positive correlation between the SCSSM and Nino3.4 indices implies that SCSSM may affect the ENSO evolution. In fact, according to the ENSO events defined by CPC, four out of the five strong SCSSM events (i.e., 1972, 1994, 1997 and 2002) are followed by the development of El Niño episodes, which accounts for 80%. Three out of the six weak SCSSM events (i.e., 1983, 1988 and 1998) are succeeded by La Niña episodes, accounting for only 50%. Such an asymmetry is possibly attributed to the asymmetry of SST anomalies over the eastern Indian Ocean[22, 23] and associated Walker cells over Equatorial Indian and Pacific Ocean. The specific process through which the SCSSM affects the Pacific SST will be investigated in the following sections.

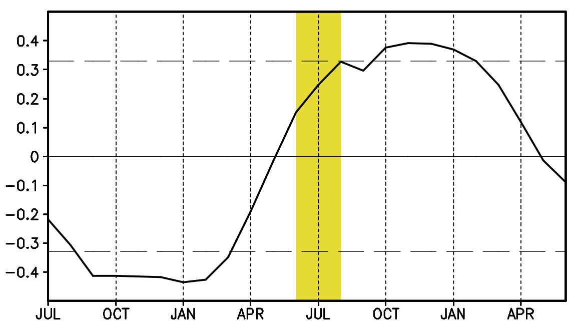


Fig.2 The lead and lag correlation coefficient between the intensity of SCSSM and Nino3.4 index. (the light shade represents for active SCS monsoon duration and the dashed line represents the level of significance at $\alpha = 0.05$.)

In the following sections we first reveal the composite circulation patterns related to strong and weak SCSSM and then explore the possible physical processes through which anomalous SCSSM affects the ENSO.

4 ATMOSPHERIC CIRCULATION ANOMALIES ASSOCIATED WITH ANOMALOUS SCSSM

The composite analysis of 500 hPa height for strong and weak SCSSM illustrates that Europe and Lake Baykal are controlled by a ridge with significant positive anomalies, the Subtropical High over the western Pacific is weaker than normal and extends more northeastward, and the south of China is controlled by negative anomalies for a strong SCSSM (Fig. 3a). For a weak SCSSM, Ural Mountain and Sea of Okhotsk are controlled by a ridge with significant positive anomalies, the Subtropical High over the western Pacific is stronger than normal and extends more southwestward, and the south of China is controlled by positive anomalies (Fig. 3b). Wu et al.[24] noted that the position of the Subtropical High over western Pacific was closely related to geopotential height over the Sea of Okhotsk. When both the geopotential heights over the Sea of Okhotsk and low-latitude western Pacific are high, the Subtropical High over the western Pacific extends more southward. Miao et al.[25] argued that the setup of the blocking over East Asia may delay the northward seasonal march of the Subtropical High over western Pacific. The 500 hPa height difference field (the strong minus weak SCSSM, Fig. 3c) reveals that a large significant negative height anomaly appears from India to 140°E and from 10°N to 30°N. Another large significant negative anomaly appears near the southeast Ural region and the Sea of Okhotsk (Fig. 3). Over East Asia, there exhibits a positive East Asian-Pacific (EAP) teleconnection pattern[26]. A negative Europe-Asia-Pacific teleconnection (EUP) pattern appears in the middle latitudes[27]. The two teleconnection patterns intersect over Japan. The observational evidences above point out that the intensity of SCSSM is related to not only low-latitude circulation, but also mid- and high-latitude circulation.

Figure 4 shows the climatological summer wind at 850 hPa and the wind difference (strong minus weak SCSSM) field. In the climatological map, the cross-equatorial flow is strongest over Somali. The wind is more zonally oriented over the northern Indian Ocean and becomes southwesterly over SCS (Fig. 4a). This is why the SCSSM index is defined based on the southwesterly velocity. The wind becomes more meridionally oriented over East Asia. The difference wind reveals a remarkable cyclonic wind over SCS and western North Pacific, an anticyclonic wind over Japan, and a cyclonic wind over the Sea of Okhotsk. This is consistent with the positive EAP teleconnection pattern in the height field. It is also found that cross-equatorial flows over the MC are greatly enhanced, and the anomalous flow direction is consistent with the climatological one. The strongest cross-equatorial flow occurs west of Sumatra. The cross-equatorial flow over Somali is weakened (Fig. 4b). The results indicate that the strongest response of cross-equatorial flows to the anomalous SCSSM heating appears over the eastern Indian Ocean/west of Sumatra.

There is an argument regarding to whether the anomalous cross-equatorial flow west of Sumatra is a result or a cause of the anomalous SCSSM. Dynamically, both ways are possible. An enhanced northward cross-equatorial wind may strengthen the SCSSM through anomalous moisture transport. An equatorial asymmetric heating distribution, according to Gill[28], may also excite cross-equatorial flows. Just as the intensity of SCSSM may be affected by many different factors, the cross-equatorial flow can also be modulated by either anomalous heating over SCS and western North Pacific or local air-sea interactions over southeastern Indian Ocean[29-31]. Kajikawa et al.[32] noted that the southeasterly wind acceleration over the southeastern Indian Ocean is closely linked to the meridionally asymmetric anomalies of convection between the MC and SCS/Philippine Sea. The argument above suggests that the anomalous cross-equatorial flow west of Sumatra is possibly the result of the anomalous SCSSM. A strong SCSSM is accompanied by the increase of precipitation (and condensational latent heating) in the SCS and tropical western North Pacific (Fig. 5). According to Gill's solution[28], the atmospheric response to a heating north of the equator is characterized by the ascending Rossby wave northwest of it (which corresponds to the anomalous low-level cyclone over the south of China), and northward cross-equatorial flows[28]. Upward motion is co-located with the heating anomalies and a low pressure center. There is a dominant local Hadley cell, with a rising branch in the maximum heating region and a subsidence branch over the equatorial region.

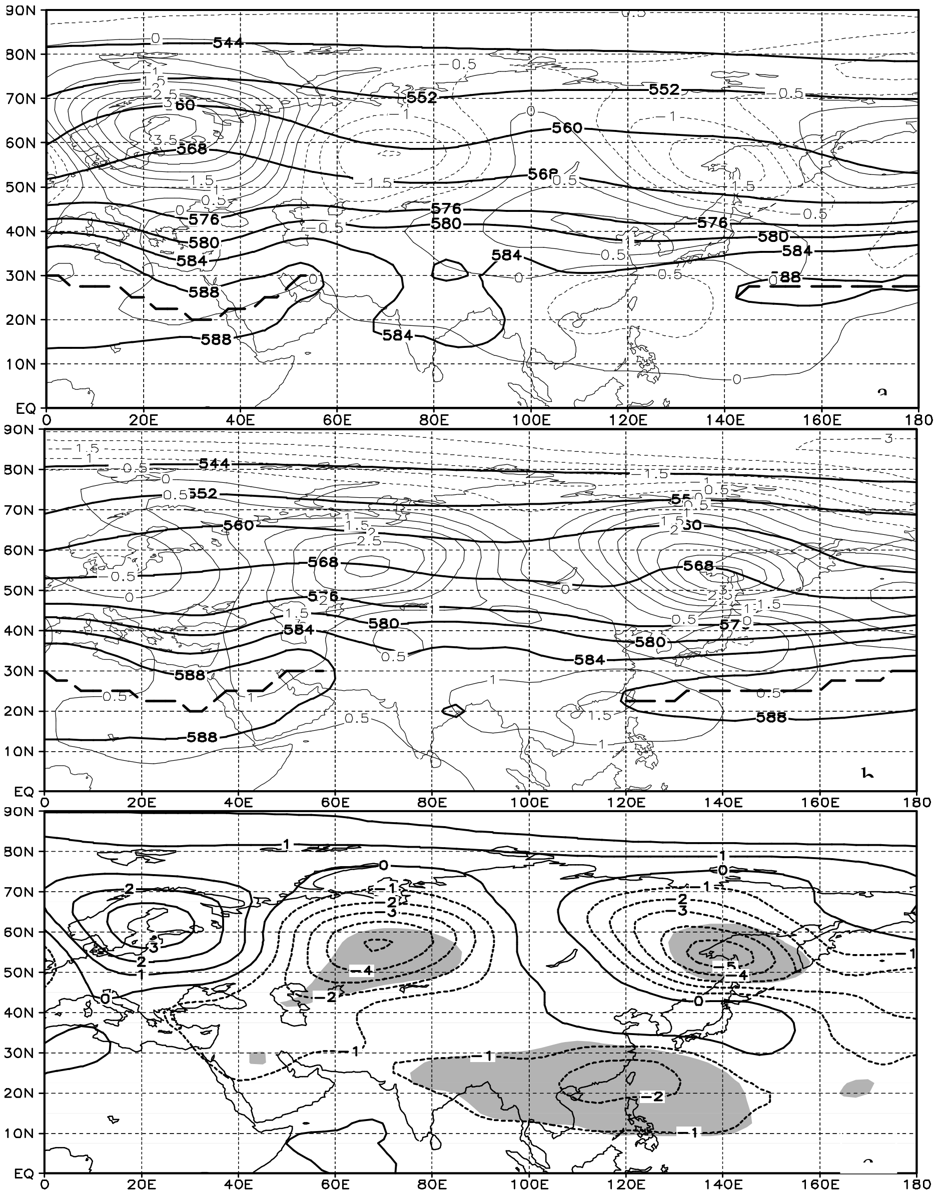


Fig.3 Summertime composite 500 hPa height field and difference for strong (a) and weak (b) SCSSM, and difference (strong minus weak) field (c). Unit: ×10gpm. The regions exceeding the 95% significance level are shaded.

To understand the relative role of the SCSSM versus Australian High in forcing the cross-equatorial flow, we computed the correlation coefficients among the SCSSM index, the cross-equatorial flow over the MC and the Australia High index (as shown in Fig. 1). It turns out that the correlation coefficient between SCSSM and the cross-equatorial flow is 0.59, which exceeds the 0.01 level of significance; but the correlation coefficients of the Australia High with SCSSMI and the cross-equatorial flow are 0.03 and 0.17 respectively, both of which do not exceed the 0.01 level of significance. The observational evidences demonstrate that the anomalous cross-equatorial flow over the MC is primarily induced by the anomalous SCSSM.

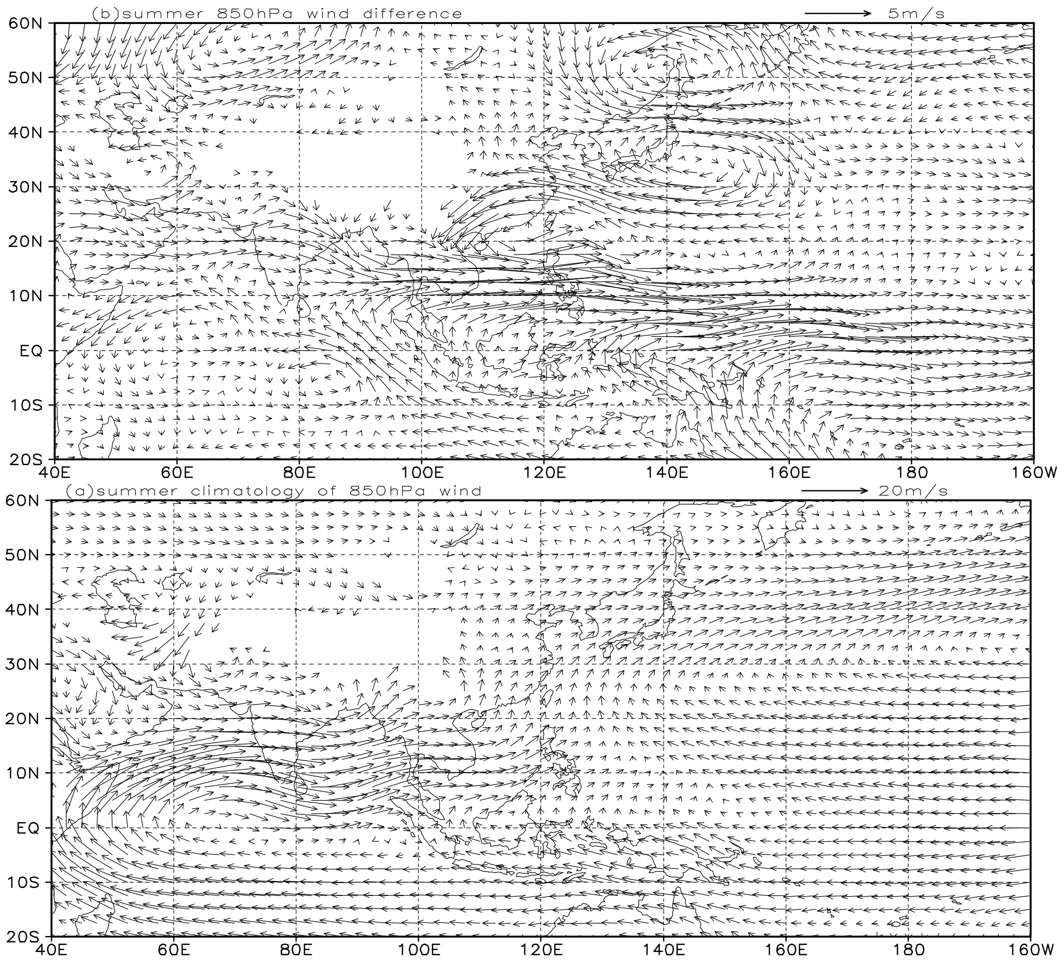


Fig.4 The climatological summer 850 hPa circulation (a) and the composite difference of summer 850 hPa circulation for strong minus weak SCSSM (b).

5 SST AND WIND ANOMALIES INDUCED BY SCSSM

In boreal summer, the climatological low-level mean flow is southeasterly off Sumatra. For a strong SCSSM, the enhanced low-level southeasterly increases the total wind speed and cools the SST off Sumatra through enhanced surface evaporation, oceanic vertical mixing and coastal upwelling[33, 34]. Since the southeast Indian Ocean is a region with intense mean convection, the cold SST anomalies imply the reduction of atmospheric convective heating (Fig. 5). According to Gill's solution[28], the negative heating anomaly may induce a descending Rossby wave response to its southwest, resulting in an anomalous anticyclone that further increases the southeasterly off Sumatra. Through this positive air-sea feedback mechanism, the cold SST anomalies off Sumatra develop[30]. For a weak SCSSM, all processes reverse. Hence, warm SST anomalies off Sumatra develop.

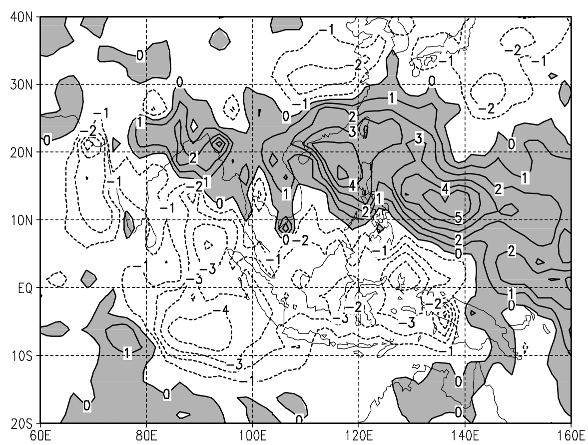
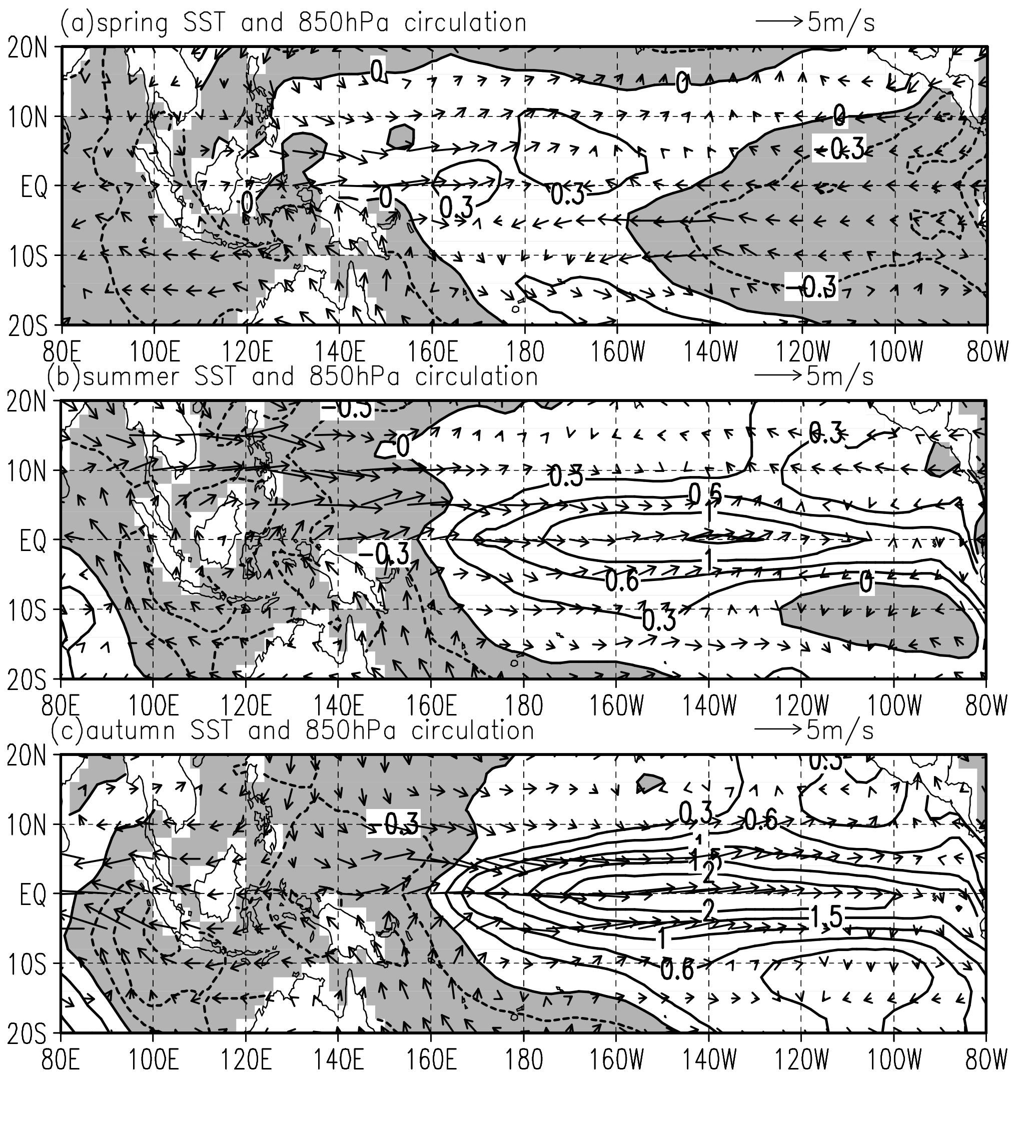


Fig.5 The summer rainfall difference field for the strong minus weak SCSSM composite. (unit: mm/day, positive values are shaded.)

Figure 6 illustrates the evolution of the composite SST and 850 hPa wind difference (a strong minus weak SCSSM) fields from MAM to DJF. During spring, the SST difference field exhibits very weak positive anomalies over 140°E－160°W and 10°S－20°N and weak negative anomalies in other regions. The atmospheric response induced by the SST pattern is also quite weak (Fig. 6a). During boreal summer, the enhanced cross-equatorial wind is co-located with negative SST anomalies over the warm pool and the MC. This large-scale cooling is primarily caused by enhanced surface evaporation associated with the enhanced cross-equatorial flow[34]. The negative SST anomalies over the warm pool and the MC, on one hand, lead to the reduction of atmospheric convective heating over the MC in the subsequent season as the seasonal convective maximum moves southeastward[34, 35]. The cold SST anomalies (particularly those off Sumatra), on the other hand, favor the development of the Indian Ocean dipole and promote the reversed Walker cells in both the equatorial Indian and Pacific Oceans. Both the processes favor the anomalous westerly over the equatorial western central Pacific (Fig. 6b). The anomalous westerly triggers downwelling oceanic Kelvin waves and leads to warm SST anomalies in the eastern equatorial Pacific through the Bjerknes dynamic air-sea interaction[36]. The positive dynamic air-sea feedback processes cause the rapid growth of the SST anomalies in the eastern equatorial Pacific from the boreal summer to winter. For example, the local SST anomaly near 140°W is 1.2°C in summer and exceeds 2.0°C in the subsequent autumn and winter (Fig. 6c & 6d). The negative SST anomalies in the western North Pacific persist from autumn to winter.



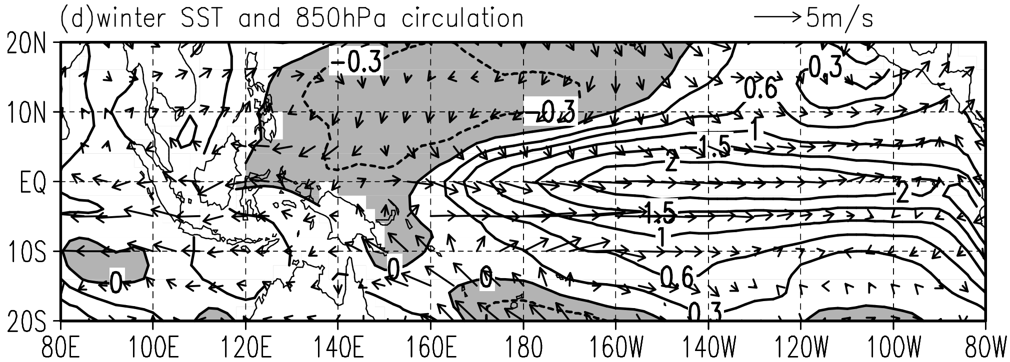


Fig.6 The composite difference (strong minus weak SCSSM) fields of SST and 850 hPa wind field for (a) spring, (b) summer, (c) autumn, and (d) winter. (units: m/s for wind field and °C for SST)

6 THE POSSIBLE MECHANISM OF THE SCSSM IMPACT ON ENSO

The annual cycle of the equatorial atmosphere-ocean system experiences the weakest phase (with the weakest zonal wind and zonal SST gradient) during boreal spring. The onset of SCSSM in late spring infuses a strong forcing into the weak system. Based on the observational analysis above, we argue that the anomalous SCSSM can modulate SST anomalies in the eastern equatorial Pacific in the succeeding autumn and winter. Figure 7 is a schematic diagram that illustrates key physical processes through which SCSSM affects the El Niño evolution. A strong SCSSM can strengthen northward cross-equatorial flows over the MC. The strengthened cross-equatorial flow leads to the large-scale SST cooling around the MC due to enhanced surface evaporation, on one hand, and triggers the positive thermodynamic air-sea feedback process in the southeast Indian Ocean and leads to the cooling of SST off Sumatra, on the other hand. Both the processes lead to the reduction of deep atmospheric convective heating over the MC and reversed anomalous Walker cells over the equatorial Pacific and Indian Oceans. The suppressed convection over the MC induces anomalous westerly over the equatorial western Pacific, which further modulates the coupled air-sea interaction process in the Pacific and affects the El Niño evolution.

The argument above suggests a possible scenario that the anomalous SCSSM may trigger the onset of ENSO. This is consistent with the statistical result in Fig. 2. Figure 2 also illustrates that a preceding El Niño event may weaken SCSSM. The circular argument above implies a two-way interaction process; a strong SCSSM leads to the onset of El Niño, and an El Niño episode after its mature phase may cause the weakening of SCSSM. Thus the two-way interaction between SCSSM and ENSO may cause the quasi-biennial oscillation of SCSSM.

7 CONCLUSION AND DISCUSSION

In this study we investigate the relationship of SCSSM with the Nino3.4 index and the anomalous circulation patterns associated with a strong and weak SCSSM. Based on the observational analysis, a mechanism that connects the anomalous SCSSM, convection over the Maritime Continent and ENSO is proposed. The main results are described as follows.

(1) The anomalies of the SCSSM intensity are related to not only low-latitude circulation, but also mid- and high-latitude circulation. The composite difference of 500 hPa height for the strong minus weak SCSSM exhibits a positive EAP teleconnection pattern along the coast of East Asia and a negative EUP teleconnection pattern in the middle and high latitudes.

(2) The anomalous SCSSM may induce anomalous cross-equatorial flows over the MC. For the strong minus weak SCSSM composite, the cross-equatorial flow is strengthened over the MC in the 850 hPa wind difference field. In the rainfall difference field, positive anomalies appear in SCS and western North Pacific and negative anomalies appear over the MC.

(3) The anomalous SCSSM may affect the ENSO evolution through the modulation of convection over the MC. A strong (weak) SCSSM induces a strong (weak) cross-equatorial flow over the MC, which cools (warms) local SST through anomalous surface latent heat flux and oceanic mixing. The negative (positive) SST anomalies further suppress (enhance) the convection over the MC, which induces anomalous westerlies (easterlies) in the equatorial western Pacific. The equatorial westerly (easterly) anomalies trigger the El Niño (La Niña) onset or modulate the El Niño (La Niña) evolution through coupled air-sea interaction processes.

The relaxation of the equatorial trades and the setup of the anomalous westerly are crucial for triggering the El Niño onset[37-39]. In this study we propose a new westerly-triggering scenario in which an anomalous SCSSM may affect the MC convection and induce an anomalous westerly over the western equatorial Pacific. Thus, in addition to the conventional oceanic dynamical initiation process (through the Rossby wave reflection in the western boundary of the ocean basin), the anomalous SCSSM is another possible cause for the onset of ENSO events.

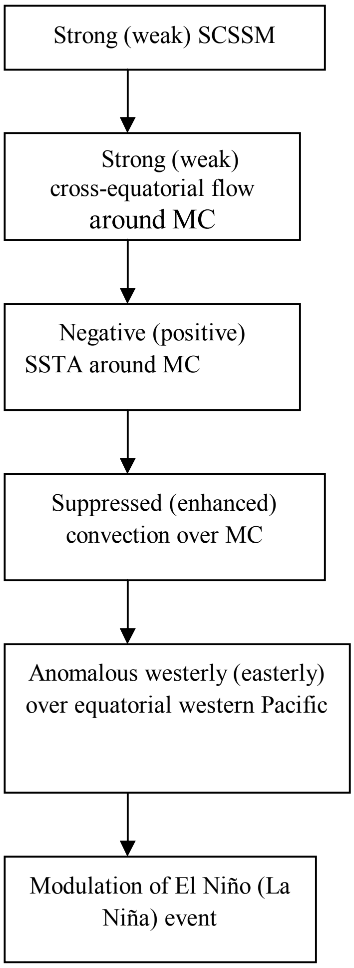


Fig.7 The schematic diagram illustrating the processes by which the connection of SCSSM to the MC convection and ENSO is set up.

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