Remote Tropical Western Indian Ocean Forcing on Changes in June Precipitation in South China and the Indochina Peninsula

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ABSTRACT

In this study, remote influence originating from the tropical western Indian Ocean on June precipitation in South China and the Indochina Peninsula is documented. Based on numerical simulation and statistical analysis, it is noted that the warm anomaly in the tropical western Indian Ocean can induce a weaker-than-normal Walker circulation across the tropical Indian Ocean and western Pacific Ocean. This further leads to a northeast–southwest-oriented western North Pacific subtropical high and a weaker-than-normal monsoon trough in the South China Sea. In addition, the weak monsoon trough is concurrent with an anomalous rising motion in South China and a sinking motion in the Indochina Peninsula. This enhances precipitation in South China and suppresses precipitation in the Indochina Peninsula on an interannual time scale. On the other hand, the warming trend in the tropical western Indian Ocean also supports the long-term trends of precipitation in the two regions.

1. Introduction

According to early studies (Tao and Chen 1987; Wang and LinHo 2002; Ding 2004), the rain belt moves stepwise northeastward from the southeast Bay of Bengal to northeast China in association with the seasonal march of the East Asian summer monsoon. In the first stage, a sharp increase in precipitation and a change in zonal wind direction in the South China Sea in mid-May indicate the onset of the South China Sea summer monsoon (Wang et al. 2004; Zhou et al. 2005; Zhou and Chan 2005, 2007; Li et al. 2014). Subsequently, the rain belt propagates in two directions. One extends northward to the subtropical western North Pacific Ocean concurrent with the advance of the subtropical high in early to mid-June, forming the meiyu rainfall in China and the Baiu rainfall in Japan (Chang et al. 2000; Zhang et al. 2002; Yuan et al. 2008; Gu et al. 2009; Wang et al. 2012). Another extends northwestward across the Indochina Peninsula to the Indian subcontinent.

Under global warming, it is expected that the intensity of precipitation will increase along with the higher moisture-holding capacity of the atmosphere and changes in vertical motion (Wentz et al. 2007; Chou
et al. 2012). It is also noted that the influence of warming on precipitation varies notably from region to region (Neelin et al. 2006). In China, an increasing trend in precipitation is found in the northwest and southeast, and a decreasing trend is observed in the northeast and southwest (Zhao et al. 2010; Wang et al. 2015, 2018; Leung and Zhou 2018; Leung et al. 2018). The increase in precipitation in southeast China may lead to a higher frequency and intensity of floods (Li et al. 2015; Qiu et al. 2017; Hu et al. 2019). Since extreme rainfall events are controlled by anomalous atmospheric circulation at synoptic and intraseasonal time scales (Li and Zhou 2015; Li et al. 2014), this study investigates trends in the monthly precipitation in South China and the Indochina Peninsula instead of seasonal precipitation, which alleviates intraseasonal variation in precipitation. In addition, the underlying cause of the trends will be investigated.

According to the results of previous studies, sea surface temperature (SST) variation in the different oceans can induce anomalous atmospheric circulation and influence the climate of South China. Past studies show that the El Niño–Southern Oscillation is closely related to variation in the East Asian winter monsoon, by inducing anomalous anticyclonic flow around the Philippine Sea (Wang et al. 2000; Li et al. 2011; X. Li and Zhou 2012; Yuan et al. 2012; Zheng et al. 2013; Leung and Zhou 2016; Leung et al. 2017; Chen et al. 2018; Gu et al. 2018; Li et al. 2019; Wang et al. 2018, 2019; Leung et al. 2019). Its impact may persist to the following summer and alter the summer precipitation in southeast China, the Indochina Peninsula, and even the lower to middle reaches of the Yangtze River region (Chen et al. 2000; Huang et al. 2004; Wu and Zhou 2008; Xie et al. 2009; Chowdary et al. 2011; Xie et al. 2016; Zhang et al. 2019). In addition, a recent study by Li et al. (2018) also illustrates the remote influences from different ocean basins, including the North Atlantic and equatorial Indian Ocean, on early summer precipitation in South China. Numerical simulation by Chen et al. (2018) showed that cold SST in the tropical western Indian Ocean and the tropical North Atlantic could induce an anomalous anticyclone in the western North Pacific, causing stronger-than-normal precipitation in South China in August 2016. According to He and Zhou (2015), the intensity of the western North Pacific subtropical high is controlled by the SST gradient between the tropical Indian Ocean and the tropical western Pacific. This enhances the intensity and frequency of summer precipitation events in South China (Fu et al. 2016). Therefore, the precipitation variation in South China is also linked to SST variation in other oceans apart from the tropical Pacific Ocean.

On the other hand, a large portion of anomalous heating accompanied by anthropogenic warming is stored in the oceans, leading to sharp warming there (Pierce et al. 2006; Lee et al. 2015; Volkov et al. 2017). Model simulations indicate that stronger warming in the tropical eastern Pacific than in the surrounding areas could facilitate a higher frequency of extreme El Niño–Southern Oscillation events in the future (Cai et al. 2014). Hence, this study will further examine the contributions of SST variation to the changes in monthly precipitation in South China and the Indochina Peninsula on both interannual time scale and long-term trend. This paper is organized as follows: section 2 illustrates the data and methods applied in this study; section 3 shows the remote impact of tropical western Pacific SST on June precipitation in South China and the Indochina Peninsula based on statistical analysis and numerical simulation; and section 4 discusses and summarizes the results of this study.

2. Data and method

To investigate long-term changes in precipitation in South China, CPC Global Unified Gauge-Based Precipitation data provided by the NOAA/OAR/ESRL PSD (https://www.esrl.noaa.gov/psd/) are employed in this study (Chen et al. 2008). In addition, ERA-Interim reanalysis data (Dee et al. 2011) are utilized to investigate atmospheric variation in conjunction with changes in precipitation. Different variables are used in this study, including geopotential \( h \), horizontal wind \( \mathbf{U} \), and vertical velocity \( \omega \). Horizontal divergence \( \text{Div} \) and wind speed \( \text{WS} \) are derived from horizontal wind. For SST, ocean reanalysis data from NOAA Extended Reconstructed Sea Surface Temperature, version 5 (ERSSTv5; Huang et al. 2014), are used. The study period covers 40 years, from 1979 to 2018.

In this study, the simplified atmospheric general circulation model SPEEDY (Molteni 2003; Kucharski et al. 2006, 2013) from the International Centre for Theoretical Physics is applied to investigate the forcing of SST on South China precipitation. This model is widely used to quantify the remote atmospheric impacts exerted by anomalous SST in different oceans (Sun et al. 2017; King et al. 2018). In this study, horizontal and vertical resolutions are T30 and 8 levels, respectively. The model is run with the climatological mean SST and an activated land model. A total of 51 initial conditions are obtained by choosing the atmospheric state with different spinup times from 10 to 60 years. In this study, the Student’s \( t \) test is employed to examine the linear trend of one sample and the difference between two samples.
3. Results

The spatial pattern of annual precipitation in South China and the Indochina Peninsula is presented in Fig. 1a. A large amount of precipitation is noted around the Pearl Delta region and Hainan Island. The amount of precipitation generally decreases northward, except for a center of strong precipitation located south of the middle and lower reaches of the Yangtze River. South China is defined as the region from 20° to 27.5°N and from 107.5° to 120°E (red box in Fig. 1a). The Indochina Peninsula is defined as the region from 10° to 22.5°N and from 97.5° to 107.5°E. The monthly precipitation in these two regions is illustrated in Fig. 1b. Maximum and minimum monthly precipitation in South China occur in June and December, respectively. In the Indochina Peninsula, monthly precipitation reaches its peak value in August. June precipitation accounts for 16.5% of annual rainfall and 40.9% of summer (June–August) rainfall in South China, and 13.7% of annual rainfall and 29.1% of summer rainfall in the Indochina Peninsula.

Additionally, the linear trend in monthly precipitation from 1979 to 2018 is shown in Fig. 1c. Significant increases in precipitation in South China are noted in June, November, and December, while a significant decrease is found in February. In the Indochina Peninsula (Fig. 1d), precipitation decreases significantly in June and increases in July. It is noted that the monthly precipitation trends in South China and the Indochina Peninsula are the opposite in June. The remainder of this study focuses on the underlying cause of the trend in June precipitation in these two regions.

a. Opposite precipitation trends in South China and the Indochina Peninsula in June

The climatological mean of precipitation in June is shown in Fig. 2a. Strong precipitation occurs in South China concurrent with the southwesterly East Asian summer monsoon flow, which transfers moisture from the South China Sea and the Bay of Bengal to South China. In Fig. 2b, the linear trend of precipitation in June is illustrated. Increasing and decreasing trends of precipitation are noted in South China and the Indochina Peninsula. It should be noted that the westerly flow between the two regions is also enhanced significantly. This raises the question of whether the long-term changes in precipitation in South China and the Indochina Peninsula are coupled with the anomalous westerly flow in the lower troposphere between the two regions.

The temporal variations in precipitation in South China and the Indochina Peninsula (hereinafter SC and ICP, respectively), along with the westerly wind at
850 hPa between the two regions (20°–22.5°N and 100°–115°E), are illustrated in Fig. 3. Correlations among the precipitation, westerly wind, and meridional gradient of geopotential between the two regions are presented in Table 1. Significant positive or negative correlations are found between the precipitation in SC or ICP, respectively, and the westerly wind, as shown in Figs. 3a and 3b, respectively. In addition, the westerly wind is highly correlated with the meridional gradient of geopotential at 850 hPa, implying a strong geostrophic balance between wind and geopotential gradient. The westerly wind gets stronger along with a steeper meridional gradient of geopotential. Note that the correlation of precipitation between the two regions is insignificant. Nonetheless, the differences in precipitation and westerly wind between the two regions are highly correlated (Fig. 3c). This means that the relative intensity of precipitation in the two regions is related to variation in the westerly wind between the two regions.

To investigate the relation of the meridional gradient of geopotential and the precipitation in the two regions, cases of strong and weak meridional gradient of geopotential are defined as the standardized gradients $< -1$ and $> 1$, respectively. According to this definition, six strong meridional gradient cases are identified, including 1993, 1997, 1998, 2005, 2010, and 2015; and six weak meridional gradient cases are identified, including 1979, 1980, 1992, 1994, 2004, and 2013. June precipitation in SC and ICP for strong and weak meridional gradient cases is presented in Fig. 3d. For strong meridional gradient cases, precipitation in SC is substantially more than that in ICP. However, the difference in precipitation between the two regions is notably weaker for weak meridional gradient cases. This is associated with a decrease in SC precipitation and an increase in ICP precipitation. This supports the view that the relative intensity of precipitation in SC and ICP is related to the westerly wind between the two regions.

The differences in precipitation between strong and weak meridional gradient cases are portrayed in Fig. 4. Significantly more and less precipitation is observed in SC and the ICP, respectively, when the meridional gradient of geopotential is strong. The pattern is similar to that of the linear trend of precipitation in Fig. 2b. This confirms that a strengthened meridional gradient of geopotential and westerly wind is associated with more precipitation in SC and less precipitation in the ICP.

### b. Contributions of atmospheric circulation in Southeast Asia to variations in SC and the ICP precipitation

In this section, the mechanism controlling the changes in SC and ICP precipitation is investigated on interannual time scale. Composites of geopotential at 850 hPa for the weak and strong meridional gradient of geopotential are delineated in Figs. 5a and 5b. A comparison of the two composites shows that a notable weak or strong meridional gradient of geopotential in SC is respectively concurrent with a strong or weak monsoon trough in the South China Sea. This is also demonstrated...
by a significant difference in the southeastern portion of the monsoon trough (Fig. 5c). Hereinafter, weak meridional gradient cases are regarded as strong monsoon trough cases and strong meridional gradient cases are regarded as weak monsoon trough cases.

For the strong monsoon trough (SMT) illustrated in Fig. 5a, the distances between contour lines increase from the west to the center of the monsoon trough and decrease from the center to the east of the trough. According to geostrophic balance, a change in distance between contours implies a change in wind speed. Therefore, a strong monsoon trough may induce deceleration and convergence west of the trough and acceleration and divergence to the east. This leads to anomalous rising and sinking motion to the west and east of the trough, respectively. For the weak monsoon trough (WMT) shown in Fig. 5b, on the other hand, the change in distance between contours across the trough is relatively weak. Consequently, its forcing on the divergence in the lower troposphere and vertical motion in the midtroposphere is weaker.

It is also noted that the orientation of the western North Pacific subtropical high is different in SMT and WMT (Figs. 5a, b). In this study, the axis of the western North Pacific subtropical high is defined as the local maximum of geopotential along each longitude in the region (5°–35°N and 90°–150°E). The axis is east–west oriented for SMT (Fig. 5a) and northeast–southwest oriented for WMT (Fig. 5b). This shows that the intensity of the monsoon trough in the South China Sea is suppressed when the western North Pacific subtropical high is northeast–southwest-oriented and is enhanced when the western North Pacific subtropical high is east–west-oriented. As shown in Fig. 5c, the geopotential in the lower troposphere in the South China Sea and Philippine Sea is significantly higher in June in WMT.

The climatological means of vertical motion and low-level divergence are illustrated in Figs. 6a and 6b. Rising motion prevails in the regions surrounding SC (Fig. 6a). Strong rising motion manifests southeast of the Tibetan Plateau, in the South China Sea, and in the subtropical western North Pacific. A relatively weak rising motion is

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**TABLE 1.** Correlation coefficients among precipitation in South China and the Indochina Peninsula and westerly wind and meridional gradient of geopotential between the two regions. Correlations exceeding the 0.05 significance level of a t-test are shown in boldface type.

<table>
<thead>
<tr>
<th></th>
<th>Westerly</th>
<th>ΔΦ/Δy</th>
<th>SC – ICP precipitation</th>
<th>ICP precipitation</th>
</tr>
</thead>
<tbody>
<tr>
<td>SC precipitation</td>
<td>0.56</td>
<td>−0.53</td>
<td>0.89</td>
<td>−0.15</td>
</tr>
<tr>
<td>ICP precipitation</td>
<td>−0.65</td>
<td>0.63</td>
<td>−0.59</td>
<td></td>
</tr>
<tr>
<td>SC – ICP precipitation</td>
<td>0.76</td>
<td>−0.72</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>ΔΦ/Δy</td>
<td>−0.97</td>
<td>—</td>
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noted in SC in association with divergence in the lower troposphere (Fig. 6b). This low-level divergence is related to the strong rising motion surrounding SC. In short, the low-level divergence plays a role in suppressing the rising motion in SC, according to the continuity equation. The climatological mean of horizontal wind speed at 850 hPa is displayed in Fig. 6c. Wind speed is relatively strong west of the monsoon trough. The wind speed decelerates across the trough and accelerates downstream of the trough. This change in wind speed is possibly related to the weakening horizontal gradient of geopotential at 850 hPa as it approaches the monsoon trough from the west. The gradient strengthens downstream of the trough. This is consistent with the result in Fig. 3, implying a strong geostrophic balance in this region.

In contrast with the climatological mean, the difference in vertical motion between SMT and WMT, shown in Fig. 6d, demonstrates a negative center in SC and a positive center in the ICP and South China Sea. The difference in vertical motion is consistent with the difference in low-level divergence (Fig. 6e). Therefore, the intensity of the monsoon trough can modulate the location of the divergence center in SC. The divergence center displaces southwestward or northeastward when the monsoon trough is respectively weak or strong. As presented in Fig. 6f, wind speed between SC and the ICP is enhanced significantly in June in WMT. The change in wind speed indicates weakened deceleration to the west of the monsoon and acceleration to the east.

c. Remote impact of the SST anomaly in the tropical western Indian Ocean on SC and ICP precipitation

As shown above, precipitation in SC and ICP is modulated by the horizontal orientation of the western North Pacific subtropical high and the intensity of the monsoon trough in the South China Sea on interannual time scale. To investigate the underlying causes of the changes in these two systems, the possible influence of SST anomalies is examined. The difference in SST between SMT and WMT is displayed in Fig. 7a. Significantly warmer SST is observed in the tropical western Indian Ocean (TWIO) concurrent with the weak monsoon trough in the South China Sea.

The regression of the geopotential at 850 hPa onto the SST in TWIO (from 10°S to 10°N and from 50° to 70°E) is delineated in Fig. 7c. A belt of positive values extends from southwest of the Tibetan Plateau to the “Maritime Continent.” The positive values in the Maritime Continent represent a weaker-than-normal monsoon trough in the South China Sea that is similar to Fig. 5c. This means that a weak monsoon trough is concomitant with anomalously warm SST in TWIO. Furthermore, the warm SST in TWIO is associated with a positive anomaly of geopotential in the upper troposphere in the tropical Indian Ocean (Fig. 7e). In addition, the calculation of the regression is repeated with detrended geopotential and SST (Figs. 7d,f). It is noted that the positive values in the regressions remain significant, even though the trends are removed. Hence, the coupling between the SST in TWIO and the intensity of the monsoon trough in the South China Sea exists on an interannual time scale.

In addition, changes in vertical motion in the midtroposphere and divergence in the upper troposphere in response to anomalous warm SST in TWIO are also investigated (Fig. 8). For the climatological mean, strong upper-level divergence and rising motion are located in the tropical western Pacific and eastern Indian Oceans (Figs. 8a,b). On the other hand, weak upper-level convergence and sinking are located in the tropical western Indian Ocean. This forms the Walker circulation across the tropical Indian Ocean and western Pacific, with strong easterlies in the upper troposphere. Regressions of the upper-level divergence and vertical motion in the midtroposphere onto the SST in TWIO are shown in Figs. 8c and 8d. It is noted that the Walker circulation is weakened concurrent with the anomalous warm SST in TWIO.
To clarify the causal relation between the SST in TWIO and the intensity of the monsoon trough, numerical simulation is employed. In each sensitivity simulation, a positive SST anomaly of 0.75 K is added to TWIO (from 10°S to 10°N and from 50° to 70°E) in June. The value of the SST anomaly is selected according to the difference between SMT and WMT in TWIO (Fig. 7a). The differences in geopotential, horizontal wind, vertical motion, and precipitation between 51 pairs of sensitivity and control simulations are calculated. The differences in geopotential at 850 and 300 hPa between the control and sensitivity simulations are
FIG. 6. Climatological means of (a) vertical motion (Pa s\(^{-1}\)) at 500 hPa, (b) divergence (10\(^{-6}\) s\(^{-1}\)) at 850 hPa, and (c) wind speed (m s\(^{-1}\)) at 850 hPa in June. (d)–(f) Similar to (a)–(c), but for the composite differences between WMT and SMT. In (d)–(f), only differences at are significant at the 0.05 significance level of a t test or better are plotted.
elucidated in Fig. 9. It is noted that warmer SST in TWIO can induce a positive belt of geopotential at 850 hPa extending from southwest of the Tibetan Plateau to the Maritime Continent (Fig. 9a). This weakens the monsoon trough in the South China Sea. As shown in Fig. 9b, an anomalous positive geopotential occurs in the upper troposphere in the tropical Indian Ocean due to the anomalous warm SST in TWIO. An anomalous westerly at 300 hPa is also noted along the tropical Indian Ocean. These changes in the atmospheric circulation are consistent with observations (Figs. 7, 8), indicating the impact of SST in TWIO on the Walker circulation across the tropical Indian Ocean and western Pacific and the monsoon trough in the South China Sea. Furthermore, the differences in precipitation and vertical motion between the control and sensitivity simulations are also examined (Fig. 10). The anomalous warm SST in TWIO also induces positive precipitation anomaly in SC and negative precipitation anomaly in ICP (Fig. 10a), in conjunction with anomalous vertical
motion (Fig. 10b). Even though the differences in precipitation and vertical motion are notably weaker in the simulation than in the reanalysis data, the results are generally consistent. Therefore, the warmer TWIO supports more precipitation in SC and less precipitation in ICP.

The above results show the remote influence of TWIO SST on the precipitation in the two regions on interannual time scale. Based on their interannual linkage, the significant warming in TWIO in June (Fig. 7b) is a possible cause of the increasing precipitation in SC and the decreasing precipitation in ICP, by weakening the monsoon trough in the South China Sea.

4. Summary and discussion

In this study, the remote influence of sea surface temperature in the tropical western Indian Ocean on the precipitation in SC and the ICP is investigated and summarized. For an anomalously warm TWIO, the sinking branch of the Walker circulation in TWIO is suppressed. This weakens the Walker circulation across the tropical Indian Ocean and western Pacific and further induces anomalous sinking motion in the South China Sea (Fig. 11a). Subsequently, the intensity of the monsoon trough in the South China Sea is suppressed. For a weak monsoon trough, the westerly wind accelerates as it approaches the monsoon trough from the west, because of the stronger-than-normal horizontal gradient of geopotential. Similarly, it decelerates on the eastern side of the monsoon trough. The change in wind speed across the monsoon trough contributes to anomalous divergence and sinking motion to the west of the trough and anomalous convergence and rising motion to the east. Consequently, the rain belt in SC and ICP is displaced northeastward, leading to stronger
precipitation in SC and weaker precipitation in ICP. This anomalous atmospheric circulation and precipitation are reversed with cold SST in TWIO (Fig. 11b). Based on this dynamic linkage, the warming in the tropical western Pacific is supporting the increasing and decreasing trends of precipitation in SC and ICP, respectively.

The monsoon trough in the South China Sea is considered an important climate system controlling the pathway and frequency of tropical cyclones; R. C. Y. Li and Zhou (2012) noted that El Niño–Southern Oscillation could increase the number of supertyphoons by strengthening the monsoon trough. In addition, it is also noted that the intensity of the monsoon trough is related to the onset time of the South China Sea summer monsoon (Wang and Chen 2018).

The early onset of the South China Sea monsoon is concurrent with an intensive and northward-shifted monsoon trough, which enhances the number of typhoons making landfall on the Southeast China coast. These studies show the influence of the trough on
postsummer precipitation in South China. This study, on the other hand, investigates the role of the monsoon trough in presummer precipitation in South China, supplementing the forcing of the monsoon trough on the summer precipitation in this region.

Furthermore, opposite precipitation trends occur in SC and ICP in June, but not in other months during boreal summer (Figs. 1c,d). Note that there are strong variations in the trends of monthly precipitation in the two regions. This situation implies that the out-of-phase relation between the two regions occurs on an intraseasonal time scale instead of on a seasonal time scale. The out-of-phase relation being confined to June is possibly related to the location of the western North Pacific subtropical high, as the subtropical high demonstrates intraseasonal northward stepwise displacements in mid-June and late July, respectively (Chang et al. 2000). However, further studies are needed to clarify this intraseasonal out-of-phase relation in the two regions.

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