Respective and Combined Impacts of North Indian Ocean and Tropical North Atlantic SST Anomalies on the Subseasonal Evolution of Anomalous Western North Pacific Anticyclones

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ABSTRACT: The subseasonal variation of the anomalous western North Pacific anticyclone (WNPAC) has important implications for East Asian summer monsoon variability. How the WNPAC evolves on the subseasonal time scale under the different configurations of tropical North Atlantic (TNA) and North Indian Ocean (NIO) SST warming is elucidated in this study. The WNPAC forced by individual TNA SST warming shows an obvious subseasonal variation with a stepwise northward movement. In contrast, the WNPAC forced by individual NIO SST warming shows a weak subseasonal variation, being nearly stabilized at around 20°N from June to August and thereby causing long-lasting and intense positive mei-yu–baiu–changma rainfall anomalies. The physical mechanism for the different subseasonal variation of WNPAC is further investigated. The TNA SST warming generates a WNPAC via a Rossby wave–induced divergence/convergence chain response. In this process, the TNA SST warming–induced suppressed convection over the western Pacific moves northward with the northward movement of climatological intertropical convergence zone and summer monsoon region, which generates a northward shift of the WNPAC. However, the NIO SST warming produces a WNPAC via a Kelvin wave–induced suppressed convection over the western Pacific Ocean. This suppressed convection is stabilized at around 20°N because of the Kelvin wave activity scope being limited within 20°N, which finally produces a nearly stationary WNPAC from June to August. In addition, under the simultaneous occurrence of the TNA and NIO SST warming, the subseasonal variation of WNPAC bears a resemblance to that for the individual NIO SST warming condition, where the TNA SST warming fails to exert its impact.

KEYWORDS: Atmosphere–ocean interaction; Anticyclones; El Niño; Monsoons; Summer/warm season

1. Introduction

The anomalous western North Pacific anticyclone (WNPAC) often appears during El Niño mature winter, decaying in spring and summer (e.g., Wang et al. 1999; Wang et al. 2000; Wu et al. 2003; Feng et al. 2011; Li et al. 2018). Especially, during El Niño decaying summer, this WNPAC is the key for El Niño to affect East Asian summer monsoon (e.g., Weng et al. 1999; Huang and Wu 1989; Feng et al. 2014; Chen et al. 2018). The WNPAC is able to cause severe climate disasters around East Asian countries in summer via modulating the western Pacific Ocean subtropical high (Chang et al. 2000; Feng et al. 2011). In general, a WNPAC on the one hand hampers the western Pacific subtropical high to shift northward, and on the other hand causes western Pacific subtropical high to extend westward into the Chinese mainland, causing abundant moisture to be transported into central China and then giving rise to severe flooding in East Asia (Chang et al. 2000). Most previous studies focused on investigating the seasonal mean (i.e., June–August mean) WNPAC characteristics and their impact on East Asian summer monsoon (e.g., Chen et al. 2016; Feng et al. 2020; Zhou et al. 2019; Xie et al. 2010). Such a way is actually not conducive to understanding East Asian summer monsoon variability more clearly, because the East Asian summer monsoon has a strong subseasonal variability, with stepwise northward shifts of the western Pacific subtropical high and the related monsoon rainfall belt (Ding and Chan 2005; Zhu et al. 2019). For example, in the decaying summer of 2015/16 El Niño, the Yangtze River Valley experienced its worst floods since 1998. However, there are no observed seasonal mean rainfall anomalies in China in summer of 2016. Thus, the seasonal mean anomaly in 2016 cannot reflect the actual flooding situation (Wang et al. 2017). Therefore, the subseasonal variation of the WNPAC has a crucial role in understanding and predicting the subseasonal variation of East Asian summer monsoon.

During El Niño decaying summer, the WNPAC has a large variation in its intensity and location that is due to the differences in El Niño characteristics (Wang et al. 2017; Feng et al. 2011, 2014; Chen et al. 2016; Jiang et al. 2019). El Niño with warming in the central Pacific (called central Pacific El Niño or El Niño Modoki or warm-pool El Niño) is accompanied by a large domain of WNPAC, while El Niño with warming in the eastern Pacific (called eastern Pacific El Niño, conventional El Niño, or cold-tongue El Niño) stimulates a WNPAC with a smaller domain concentrating on the low latitudes (Feng et al. 2011). Feng et al. (2020) further reported that the central Pacific El Niño, which decays quickly into a La Niña in the post–El Niño summer, stimulates a weak and small WNPAC. In contrast, the central Pacific El Niño, which decays slowly and is accompanied by tropical North Atlantic Ocean sea surface temperature (SST) warming, generates a strong and large WNPAC. Wang et al. (2017) classified El Niño in terms of its intensity and found that the strong

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El Niño often induces a robust WNPAC from El Niño mature winter to decaying spring, while the weak El Niño often induces a weak and insignificant WNPAC from winter to decaying spring. However, in the decaying summer, strong and weak El Niños stimulate a similar WNPAC in its intensity and domain (Wang et al. 2017). Moreover, El Niños with different decaying speeds also induce different WNPAC features (Zhou et al. 2019; Feng et al. 2014; Jiang et al. 2019). Therefore, it can be seen that the relationship between El Niño and WNPAC is complicated, especially in El Niño decaying summer when El Niño SST anomalies in Pacific decay to a normal condition. The El Niño–WNPAC relationship needs further and in-depth investigation.

The WNPAC generation is a Rossby wave response to the suppressed convection over the western Pacific (Wang et al. 2000; Xie et al. 2009; Ham et al. 2013). This suppressed convection in El Niño mature winter and decaying spring can be induced by the local air–sea interaction over the western North Pacific via the wind–evaporation feedback (Wang et al. 2000), where the WNPAC is coupled with the cold SST anomalies over the western North Pacific. On the eastern side of the WNPAC, the anomalous northeasterly winds are consistent with the climatological winds. Thus, the total wind speeds are increased, and evaporation is enhanced. The SST in the western North Pacific is cooled. Such cooling in turn yields a WNPAC by inducing Rossby waves. In addition, the suppressed convection over the western Pacific is also induced by the moist enthalpy advection/Rossby wave modulation mechanism (Wu et al. 2017a,b). In response to the El Niño positive SST anomalies, the northern branch of the twin Rossby wave cyclonic anomalies advects dry and low moist enthalpy air into the tropical western North Pacific, which suppresses local convection and further drives the WNPAC. Therefore, based on the above review of WNPAC maintenance mechanisms, it can be seen that the climatological mean state plays an important role. In El Niño decaying summer when the climatological mean state is completely changed, the WNPAC maintenance mechanisms, namely “local air–sea interaction” and “moist enthalpy advection,” do not work. Although El Niño SST anomalies decay into a normal state in El Niño decaying summer, the remote oceans over the north Indian Ocean (NIO) and tropical North Atlantic (TNA) show significant anomalous SST warming (Alexander et al. 2002; Obha and Ueda 2005; 2009), which plays a key role in maintaining the WNPAC (Yang et al. 2007; Xie et al. 2009; Rong et al. 2010; Ham et al. 2013; Feng and Chen 2021). The NIO SST warming generates a WNPAC via the Kelvin wave response, which is described as the “capacitor effect” by Xie et al. (2009). In addition, the TNA SST warming generates a WNPAC dominantly via the Rossby wave response. It is reported that the TNA SST warming causes suppressed convection over the central-eastern Pacific via exciting a Rossby wave, which is expected to directly generate a WNPAC (Ham et al. 2013). However, based on a numerical simulation, Feng and Chen (2021) suggested that the central-eastern Pacific suppressed convection is unable to directly stimulate a strong and significant WNPAC owing to the fact that it is far from the western North Pacific. They proposed that the TNA SST warming induces a WNPAC via a complicated Rossby wave–induced divergence/convergence chain process. The Rossby wave–induced central-eastern Pacific suppressed convection can induce anomalous low-level convergence and positive rainfall anomalies over the tropical western-central Pacific due to the mass balance law, which further generates anomalous low-level divergence and suppressed convection over the western North Pacific due to the same mass balance law as well. Thus, a strong and significant WNPAC is eventually produced (Feng and Chen 2021).

Previous studies demonstrated that both TNA and NIO SST warming are able to induce a WNPAC. Feng and Chen (2021) found that the TNA SST warming-induced WNPAC in summer often extends northward, while the NIO SST warming-induced WNPAC concentrates in the low latitudes. However, their study focused on the seasonal mean time scale. How WNPAC evolves in the subseasonal time scale under the individual and combined forcing of TNA and NIO SST warming is still unknown. This issue is important to predict the subseasonal variation of the WNPAC and East Asian summer monsoon. Nevertheless, this issue has not attracted much attention at present. Therefore, this study aims at investigating the respective and combined impacts of TNA and NIO SST warming on the subseasonal evolution of WNPAC and related physical mechanisms.

The structure of the rest of this paper is arranged as follows. Section 2 gives the introduction of the data and methods that are used in this study. The separate influence of TNA and NIO SST warming on the subseasonal evolution of the WNPAC and the related physical mechanisms are elucidated in section 3. Subsequently, the combined influence of TNA and NIO SST warming on the subseasonal evolution of the WNPAC is investigated in section 4. The related physical mechanisms that are responsible for this combined impact are given as well. Section 5 gives the summary of this study.

2. Data, methods, and model introduction

The monthly reanalysis data from the National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP–NCAR) are used in this study. The horizontal resolution is 2.5° latitude by 2.5° longitude (Kalnay et al. 1996). Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) rainfall data with the horizontal resolution of 1° latitude by 1° longitude are also employed in this study. The SST data are obtained from The Hadley Centre Global Sea Ice and Sea Surface Temperature dataset (HadISST), which has a 1° latitude by 1° longitude horizontal resolution and covers the period from 1870 to present (Rayner et al. 2003). In this study, the period from 1979 to 2020 is used, and the linear trend of all data is removed to avoid the impact of global warming.

To reflect the NIO SST interannual variability, a NIO index is defined by averaging SST anomalies over the region 0°–20°N, 40°–100°E. In the same way, a TNA index is defined by averaging SST anomalies over the region 0°–20°N, 80°–20°W. These two indices have a similar standard deviation (0.3). To
understand the respective impacts of the NIO SST warming and TNA SST warming on the subseasonal evolution of WNPAC, a partial correlation/regression method is used in this study (Ashok et al. 2007). In addition, since the tropical central-eastern Pacific (CEP) SST cooling can also generate a WNPAC via inducing a Rossby wave response. Here, to remove the interference of the CEP SST cooling on the impact of TNA and NIO SST warming on WNPAC, a CEP SST index is defined by averaging SST anomalies over the region 10°S–10°N, 160°E–130°W.

When we perform the partial regression/correlation analysis using the following equation:

$$r_{12,34} = \frac{r_{12,4} - r_{13,4}r_{23,4}}{\sqrt{(1 - r_{13,4}^2)(1 - r_{23,4}^2)}}. \quad (1)$$

the CEP SST influence is also removed in this study. In Eq. (1), the term $r_{ij}$ represents the linear correlation coefficient between $A_i$ and $A_j$; $r_{12,34}$ is the partial coefficient between two variables $A_1$ and $A_2$, after removing the influence of the variables $A_3$ and $A_4$ (Ashok et al. 2007).

To illustrate the WNPAC subseasonal variation under the different SST forcing, an atmospheric general circulation model (AGCM) is used to perform a series of sensitivity experiments. Here, a spectral atmospheric model (SAMIL) is used in this study, which is developed at the Institute of Atmospheric Physics (IAP)/State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG). The dynamical framework of SAMIL is a hybrid-coordinate system with 26 vertical layers, rhomboidally truncated at wavenumber 42 in the horizontal (R42), which contains a nominal Gaussian grid resolution of 2.81° latitude by 1.66° latitude. For details of this model refer to the studies of Bao et al. (2013) and Wang et al. (2013). The experiments designed in this study are introduced in section 3.

3. Separate influence of TNA and NIO SST warming on WNPAC

a. Subseasonal WNPAC evolution differences

To see the separate influence of TNA and NIO SST warming on the subseasonal WNPAC evolution, the partial regression of 850-hPa wind and streamfunction against the TNA SST warming condition only reflects the WNPAC features in July and August (Figs. 1b,c). However, the southward location of WNPAC for the NIO SST warming can roughly reflect the WNPAC features in June, July, and August due to its weak subseasonal variation.

The WNPAC location plays a key role in determining summer drought and flooding regions in East Asia (Fig. 2). The TNA SST-warming-induced WNPAC leads to positive rainfall in southern China in June due to the transportation of water vapor by WNPAC-related southerly winds (Fig. 2a). With the strengthening and northward movement of WNPAC in July, the WNPAC transports abundant water vapor to the central China and northern China and causes maximum 0.6 mm day$^{-1}$ positive rainfall anomalies there (Fig. 2b). In August, the WNPAC extends farther north, and thus plentiful water vapor is transported to central and northern China (Fig. 2c). The resultant positive rainfall expands to a much larger domain than that in July (Fig. 2c). In contrast, since the NIO SST warming-driven WNPAC has long been stable in the low latitudes, a large amount of water vapor is continuously transported to the meiyu–baiu–changma region in June and July (Figs. 2d–e), leading to long-lasting and intense positive rainfall anomalies over the meiyu–baiu–changma region (Figs. 2d–e). In such a situation, severe flooding is prone to occur in these regions. In August, with slightly northward extension of WNPAC, the positive rainfall anomalies move slightly northward as well (Fig. 2f).

In addition, to further show the different subseasonal variation of the WNPAC location, partial regression of zonally averaged vorticity anomalies against TNA index and NIO index is shown in Fig. 3. It can be seen that for the TNA SST warming condition, the anomalous negative vorticity, which reflects the WNPAC, initially develops in the low latitudes in June and moves northward in the following July and August (Fig. 3a). Comparably, for the NIO SST warming condition, the anomalous negative vorticity is always located in the low latitudes from June to August, indicating a roughly stationary WNPAC (Fig. 3b).

b. The possible physical mechanism

According to the above results, the WNPAC has a strong subseasonal variation for the TNA SST warming condition with an obvious northward movement from June to August.
FIG. 1. Partial regression of 850-hPa wind anomalies (vectors) and streamfunction anomalies (shading; $10^6$ m$^2$ s$^{-1}$) against the summer monthly (a)–(c) TNA and (d)–(f) NIO indices in (top) June, (middle) July, and (bottom) August. The absolute value of wind vectors showing here is greater than 0.2 m s$^{-1}$. Letter A denotes the WNPAC. The location of A is determined by the maximum streamfunction anomalies within the domain of WNPAC (100°–140°E). The stippled regions indicate streamfunction anomalies exceeding the 90% confidence level according to the Student’s $t$ test.
whereas the WNPAC has a weak subseasonal variation for the NIO SST warming condition with a weak position shift. Thus, the physical mechanism that is responsible for such different subseasonal variation of the WNPAC position is explored. The SST anomaly and its generated anomalous vertical atmospheric circulation are shown in Fig. 4. For the TNA SST warming condition, the most obvious SST anomalies lie in the TNA region, where positive SST anomalies with a maximum value of 0.4°C persist through all summer months and slightly weaken from June to August (Figs. 4a–c). The warm SST anomalies produce anomalous upper-level divergence over the TNA and anomalous convergence over the central-eastern Pacific from June to August (Figs. 4a–c). In contrast, for the NIO SST warming condition, obvious positive SST anomalies cover the Indian Ocean from June to August with a maximum value of 0.4°C, which is comparable.
to that for the TNA SST warming condition (Figs. 4d–f). Although over the southeastern Pacific there are positive SST anomalies, they cannot force the atmospheric circulation anomalies due to the low climatological SST over the southeastern Pacific. As a response to the NIO SST warming, the anomalous upper-level divergence over the Indian Ocean is observed, and the related anomalous convergence is seen over the western Pacific and Maritime Continent from June to August (Figs. 4d–f). The anomalous vertical zonal circulation shows distinct characteristics between the TNA SST warming condition and NIO SST warming condition. How these vertical zonal circulation and their related rainfall anomalies induce the different subseasonal variation of WNPAC position is further explored.

**Fig. 3.** Partial regression of zonally averaged 850-hPa vorticity anomalies (100°–150°E; 10^{-6} \text{s}^{-1}) against the (a) TNA and (b) NIO indices from May to August. The stippled regions indicate vorticity anomalies exceeding 90% confidence level according to the Student’s t test.

**Fig. 4.** As in Fig. 1, but of SST anomalies (shading; °C) and 200-hPa velocity potential (contours). The contour interval is 1.5 × 10^5 \text{m}^2 \text{s}^{-1}.
It is reported that the TNA SST warming generates a WNPAC via a Rossby wave–induced divergence/convergence chain response over the Pacific (Feng and Chen 2021; Fig. 11a, described in more detail in section 5). Based on numerical simulation results, Feng and Chen (2021) suggested that the TNA SST warming-induced Rossby wave leads to central-eastern Pacific suppressed convection, which is far away from the western North Pacific (WNP) region and thereby cannot directly produce a strong and obvious WNPAC (Feng and Chen 2021, their Fig. 11). However, the central-eastern Pacific suppressed convection can produce low-level convergence and enhanced convection on its southwest side (i.e., the tropical western-central Pacific in June) due to the mass balance law (Fig. 11a). The WNP convection is suppressed, which thus produces a strong and robust WNPAC (Feng and Chen 2021). Here, we diagnose this physical mechanism on the subseasonal time scale and make clear what controls the northward movement of the WNPAC from June to August for the TNA SST warming condition.

Figure 5 shows the partial regression of the 850-hPa streamfunction anomalies and rainfall anomalies against summer monthly TNA index in (top) June, (middle) July, and (bottom) August. In (d)–(f), the black contours indicate the climatological rainfall with the value of 5 mm day$^{-1}$, which roughly stands for the ITCZ and climatological summer monsoon rainfall regions. Absolute wind values greater than 0.2 m s$^{-1}$ are plotted in (a)–(c). The stippled regions in (d)–(f) denote rainfall anomalies exceeding 90% confidence level according to the Student’s t test. The red- and blue-outlined rectangles in (d)–(f) represent the key suppressed and enhanced convection regions, respectively, in the process of WNPAC northward movement.

On the other hand, the Rossby wave–induced westerly winds gradually weaken over the eastern Pacific (Fig. 5a). Thus, the anomalous low-level divergence is seen over the central-eastern Pacific (Fig. 4a). Correspondingly the rainfall anomalies over the central-eastern Pacific, where the intertropical convergence zone (ITCZ) is located, is suppressed in June (Fig. 5d). Based on an atmospheric general circulation model, Feng and Chen (2021) demonstrated that the central-eastern Pacific negative rainfall anomaly would induce low-level convergent winds and positive rainfall anomalies on its southwest side (i.e., the tropical western-central Pacific in June) owing to the mass balance law, which can be clearly seen in Fig. 5d. Such positive rainfall anomalies are expected to further yield low-level divergent winds and suppressed convection over the WNP due to the same mass balance law according to the study of Feng and Chen (2021). However, their results are obtained based on the summer mean time scale. On the subseasonal time scale in our study, owing to the southward location of ITCZ and the climatological summer monsoon region in June, the WNP suppressed convection is insignificant and extremely weak (Fig. 5d). Hence, no obvious WNPAC is induced in June (Figs. 1a and 5a). This result further confirms that the suppressed convection that this generates over the central-eastern Pacific cannot effectively produce a significant WNPAC without the WNP suppressed convection.

In July, a pair of cyclones as a Rossby wave response to the TNA SST warming moves westward into the eastern Pacific (Fig. 5b). Its generated suppressed convection over the central-eastern Pacific moves northward due to the northward shift of
the ITCZ in July (Fig. 5e). More important, in July the suppressed convection over the WNP is successfully induced by the positive rainfall anomalies over the tropical western-central Pacific because of the northward expansion of the climatological summer monsoon region (Fig. 5e). Therefore, a significant WNPAC is induced by the WNP suppressed convection (Figs. 1b and 5b). In August, with farther northward movement of the ITCZ and climatological summer monsoon region, the Rossby wave–induced suppressed convection over the central-eastern Pacific shifts farther northward, which leads to positive rainfall anomalies over the tropical western-central Pacific (Fig. 5f). This positive rainfall anomaly over the tropical central-western Pacific induces and pushes the suppressed convection over the WNP more poleward due to the continuous northward movement of the climatological summer monsoon region (Fig. 5f). Correspondingly, the generated WNPAC is located more northward in August (Figs. 1c and 5c). In terms of the above analysis, the northward movement of the ITCZ and climatological summer monsoon region plays a crucial role in the northward shift of the WNPAC from June to August for the TNA SST warming condition. In other words, the subseasonal evolution of climatological ITCZ and summer monsoon north border determines the location of the suppressed WNP convection and eventually controls the WNPAC subseasonal position.

In contrast, the NIO SST warming generates a WNPAC via the Kelvin wave–induced divergence response (Yang et al. 2007; Xie et al. 2009; Wu et al. 2009). Xie et al. (2009) reported that the NIO SST warming excites anomalous easterly winds as a Kelvin wave propagating into the western Pacific. The anomalous easterly winds reach their peak at the equator and weaken toward the north and south, forming anticyclonic shear and further leading to divergence off the equator over the WNP as a result of the Ekman pumping process (Wu et al. 2009). Thus, the suppressed WNP convection generates a strong WNPAC (Xie et al. 2009). Here, we diagnose this physical mechanism in the subseasonal time scale and make clear what controls the weak subseasonal variation of the WNPAC position from June to August for the NIO SST warming condition.

Figure 6 shows the partial regression of the tropospheric temperature anomalies, 850-hPa wind anomalies, and rainfall anomalies against summer monthly NIO index. To clearly see the NIO SST warming-induced Kelvin wave, the tropospheric temperature anomalies are given in Fig. 6a. In June, the tropospheric temperature anomalies show a wedge shape over the eastern Indian Ocean and western Pacific, indicating a Kelvin wave intruding into the western Pacific (Fig. 6a). This anomalous tropospheric temperature distribution is similar to that in the study of Xie et al. (2009), who demonstrated that this wedge shape of tropospheric temperature is a Kelvin wave response to NIO SST warming via observed and numerical simulated analysis. This Kelvin wave is able to suppress convection over the western Pacific. This occurs because the Kelvin wave–related easterly winds over the western Pacific reach their peaks...
at the equator and decrease toward north and south (Fig. 6a), leading to anticyclonic shear and causing low-level divergence over the western Pacific (Fig. 4d; Wu et al. 2009). Thus, the suppressed convection over the western Pacific is induced (Fig. 6d). Since the domain of the Kelvin wave activity is limited to south of 20°N, the suppressed convection over the western Pacific induced by Kelvin wave is dominantly concentrated south of 20°N in June (Fig. 6d); thus, a WN PAC being located southward is produced in such a condition (Figs. 1d, 6a). In July, a similar wedge shape of tropospheric temperature anomalies is observed, indicating a Kelvin wave propagating into the western Pacific (Fig. 6b). The Kelvin wave–induced suppressed convection over the western Pacific is mainly limited south of 20°N as well, although the ITCZ and climatological summer monsoon domain expand northward in July (Fig. 6e). Hence, the generated WN PAC in July covers a similar domain as that in June, maintaining a stationary state (Figs. 1c and 6b). In addition, the suppressed convection over the western Pacific is weaker in July than in June. It is possibly a result of the modulation of intraseasonal oscillation processes, which is beyond the scope of this study. In August, the NIO SST warming generates a similar Kelvin wave that intrudes into the western Pacific (Fig. 6c). At this time, the domain of the ITCZ and climatological summer monsoon extends farther northward than that in July. However, the suppressed convection over the western Pacific caused by the Kelvin wave is still concentrated south of 20°N in August, which induces a WN PAC being located in a similar region. As a result, the NIO SST warming-induced WN PAC shows a weak subseasonal position variation. The limited domain of the Kelvin wave activity is the main cause of this nearly stationary WN PAC from June to August.

To further verify the results obtained from the observational data, a series of numerical experiments is conducted here, including one control experiment and two sensitivity experiments forced by different SST boundary conditions (Table 1). We integrate each experiment for 20 years. The first 5 years of integration are removed to avoid the initial influence. The remaining 15 years of integration are averaged to remove the internal variability. The anomalous atmospheric responses to the prescribed SST anomalies are calculated by the difference between the sensitivity experiment and control experiment. Figure 7 shows 850-hPa wind responses to the TNA warming (TNA_run minus CLIM_run) and the NIO warming (NIO_run minus CLIM_run), respectively. In response to the TNA warming forcing, there exists an obvious

### Table 1. Experiment details.

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<th>Expt name</th>
<th>Introduction of imposed SST boundary</th>
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<td>CLIM_run</td>
<td>Monthly climatological SST</td>
</tr>
<tr>
<td>TNA_run</td>
<td>SST anomalies obtained from the partial regression analysis in the TNA (0°–25°N, 80°–10°W) from January to December are imposed on monthly climatological SST</td>
</tr>
<tr>
<td>NIO_run</td>
<td>As in TNA_run but with SST anomalies in the NIO (10°S–25°N, 40°–100°E)</td>
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Fig. 7. The 850-hPa wind responses (vectors) to the SST anomalies (shading; °C) over the (a)–(c) TNA and (d)–(f) NIO in (top) June, (middle) July, and (bottom) August. Absolute wind values greater than 0.5 m s⁻¹ are plotted. Letter A denotes the WN PAC. The location of A is determined by the maximum streamfunction anomalies within the domain of WN PAC (100°–140°E).
WNPAC from June to August (Figs. 7a–c). The simulated WNPAC is slightly stronger than the observed one in June (Fig. 7a), whereas the simulated WNPAC is slightly weaker in August (Fig. 7c). Although there exists a bias in the WNPAC intensity in the numerical simulation, the northward movement of the WNPAC from June to August is well simulated (Figs. 7a–c), which bears a resemblance to the observational result. In response to the NIO SST warming, a nearly stationary WNPAC is simulated from June to August due to the limited domain of Kelvin wave (Figs. 7d–f).

4. Combined influence of NIO and TNA SST warming on WNPAC

a. Subseasonal WNPAC evolution

The previous section focuses on the separate impact of the NIO and TNA SST warming on the subseasonal evolution of the WNPAC in summer. Then a question is raised how the summer WNPAC evolves in the subseasonal time scale with the positive SST anomalies simultaneously appearing over the TNA and NIO. Thus, the combined impact of the TNA and NIO SST warming on the subseasonal evolution of the WNPAC is elucidated in this section. Here, the composite method is used based on the TNA and NIO index. A criterion of both of the summer-mean TNA and NIO indices being greater than 0.1 is defined to choose the simultaneous TNA and NIO SST warming events. Accordingly, five events (1987, 1998, 2010, 2016, and 2020) were eventually chosen.

Figure 8 shows the composite rainfall anomalies and 850-hPa wind anomalies in June, July, and August for the simultaneous TNA and NIO SST warming events. In June, a WNPAC is observed, which transports abundant water vapor to the southern China and the mei-yu–baiu–changma region where positive rainfall anomalies are seen with a maximum magnitude of 1.6 mm day$^{-1}$ (Fig. 8a). In July, the WNPAC persists and intensifies (Fig. 8b). However, its location is nearly the same as in June, leading to the water vapor continuously being transported to the mei-yu–baiu–changma region. Significant positive rainfall anomalies are observed in July with a maximum magnitude of 4 mm day$^{-1}$ (Fig. 8b). In such a situation, the mei-yu–baiu–changma region is prone to experience long-period and intense rainfall anomalies, leading to severe flooding disaster. In August, WNPAC and its related positive rainfall anomaly have a weakly northward movement (Fig. 8a). As a result, it can be seen that the position of WNPAC has a weak variation from June to August with the combination forcing of TNA and NIO SST warming. Thus, in comparing Fig. 8 and Fig. 1, one can find that the combined impacts of the TNA and NIO SST warming on the subseasonal evolution of the WNPAC bear a resemblance to the individual impact of NIO SST warming.

b. The possible physical mechanism

In this section we aim to understand why the subseasonal evolution of the WNPAC forced by the combination of TNA and NIO SST warming is similar to that forced by the individual NIO SST warming. The distribution of SST anomalies is first diagnosed in Fig. 9, where significant positive SST anomalies are seen over the TNA and NIO from June to August (Fig. 9). In addition, over the South China Sea (SCS) there are obvious positive SST anomalies as well.
Since anomalous descending motion and negative rainfall anomalies occupy the SCS (Figs. 9 and 10), the positive SCS SST anomalies are forced by the atmospheric process. The SST anomalies over the Pacific are weak and insignificant. Hence, the TNA and NIO SST warming can be considered as the main forcing of the anomalous atmospheric circulation anomalies. The anomalous vertical zonal circulation shows ascending motion over the NIO and TNA and descending motion over the whole Pacific from June to August (Fig. 9). Note that the center of descending motion is dominantly located in the western Pacific (Fig. 9), which is similar to that for the individual NIO SST warming condition.

Based on the aforementioned results, both the TNA and NIO positive SST anomalies stimulate a WNPAC by suppressing the convection over the western Pacific. The TNA SST warming-induced suppressed convection over the western Pacific moves gradually northward from June to August with the northward expansion of the ITCZ and climatological summer monsoon region, while the NIO SST warming-induced suppressed convection always concentrates south of 20°N from June to August owing to the limited domain of the Kelvin wave activity. How the western Pacific suppressed convection evolves under the combined impact of TNA and NIO SST anomalies is further diagnosed in Fig. 10. The suppressed convection over the western Pacific is seen in June and is kept south of 20°N (Fig. 10a). Although the western Pacific suppressed convection is intensified in July, the main body of this suppressed convection is still confined south of 20°N (Fig. 10b). In August when the climatological ITCZ and summer monsoon region continue to expand north to about 30°N, the western Pacific suppressed convection is still located south of 20°N (Fig. 10c). Therefore, the subseasonal variation of the western Pacific suppressed convection bears a large resemblance to that forced by separate NIO SST warming (Figs. 6d–f and 10a–c). The resultant WNPAC shows a weak subseasonal position variation.

A question is raised as to why the NIO SST warming and not the TNA SST warming acts as a dominant role in forcing the subseasonal evolution of WNPAC. In other words, the role of TNA SST warming in inducing a northward movement of western Pacific suppressed convection does not work under the combined impact of NIO and TNA SST warming. Based on the aforementioned analysis, the positive rainfall anomalies over the tropical western-central Pacific, indirectly forced by the TNA SST warming, play a crucial role in generating the suppressed convection over the western North Pacific via the mass balance law. However, under the combined forcing of TNA and NIO SST warming, it can be seen that the positive rainfall anomalies over the tropical western-central Pacific disappear, and instead negative rainfall anomalies occupy this region (Figs. 10a–c). It is because the NIO SST warming-induced Kelvin wave propagates to the western-central Pacific, thereby directly inducing low-level divergence and negative rainfall anomalies there. In addition, the TNA SST warming-induced suppressed convection over the central-eastern Pacific, which has a gradually northward shift from June to August with the northward expansion of the ITCZ, can be observed in Figs. 10a–c. However, this suppressed convection fails to produce positive rainfall anomalies over the tropical central-western Pacific due to the intervention of the NIO SST warming. Thus, the role of TNA SST warming
in inducing a northward movement of WNPAC from June to August is not effective.

5. Summary and outlook

It is known that the TNA SST warming and NIO SST warming play a crucial role in producing a subseasonal variation of WNPAC in summer, which acts as a key atmospheric circulation determining the East Asian summer monsoon drought and flooding regions. In this study, the respective impacts of TNA SST warming and NIO SST warming on the subseasonal evolution of WNPAC are discussed, which is conducive to understand the differences in the influence of TNA and NIO SST anomalies on the WNPAC. As a response to the TNA SST warming forcing, the generated WNPAC is extremely weak in June, becomes strong and moves northward in July, and continues to shift northward in August. Thus, the most obvious feature of the WNPAC subseasonal variation is the northward movement from June to August. The physical mechanism that is responsible for this northward movement is further investigated. The TNA SST warming first induces a Rossby wave with a pair of cyclones over the western TNA and eastern Pacific, which leads to suppressed convection over the central-eastern Pacific owing to the gradual weakening of the Rossby wave–related westerly winds. The central-eastern Pacific suppressed convection moves northward with the northward expansion of the ITCZ from June to August. Subsequently, the suppressed convection over the central-eastern Pacific leads to low-level convergence and positive rainfall anomalies over the tropical western-central Pacific, which finally produces low-level divergence and suppressed convection over the WNPAC. With the northward movement of ITCZ and summer monsoon region from June to August, the WNPAC suppressed convection shifts northward as well, which eventually yields a subseasonal northward movement of the WNPAC (Fig. 11a).

In contrast, as a response to the NIO SST warming, the WNPAC shows a completely different subseasonal evolution. The WNPAC maintains a nearly stationary state, which is located in the low latitudes from June to August. Thus, the WNPAC shows a weak subseasonal variation. Such WNPAC subseasonal evolution leads to a long duration and intense positive rainfall anomalies over the mei-yu-baiu–changma region. The NIO SST warming first stimulates a Kelvin wave propagating into the western Pacific, which causes the convection over the western Pacific to be suppressed. Since the Kelvin wave activity domain is limited south of 20°N, the suppressed convection over the western Pacific from June to August is always confined to south of 20°N. Therefore, the resultant WNPAC is located in a similar area, showing a weak subseasonal variation. In such a situation, the scope of the suppressed convection is not affected by northward movement of the ITCZ and summer monsoon region, which is different from that for the TNA SST warming condition (Fig. 11b).

In some cases in which the TNA and NIO SST warming appear simultaneously, the subseasonal variation of the WNPAC is weak, showing a nearly stationary state from June to August. The subseasonal evolution of the WNPAC bears a strong resemblance to that for the separate NIO SST warming condition. It is because the suppressed convection over the western Pacific is always located south of 20°N from June to August, which induces a weak subseasonal WNPAC variation. The forcing of the TNA SST warming does not work because the TNA SST warming-induced positive rainfall anomalies over the tropical western-central Pacific, which plays an important role in producing and pushing the western Pacific suppressed convection northward, are replaced by negative rainfall anomalies. These negative rainfall anomalies are forced by the NIO SST warming-induced Kelvin wave. Therefore, the TNA SST warming fails to force a northward movement of WNPAC under the interference of NIO SST warming. These results are obtained via composite analyses based on five samples. There is a caution that we should keep in mind for the composite method. Owing to the limited samples, these results are inevitably affected by internal atmospheric variability.

From the results of this study, it can be seen that the impact of TNA and NIO SST warming on the WNPAC is completely different on the subseasonal time scale. This result has an important implication for predicting the monthly variability of East Asian summer monsoon based on the WNPAC. This study
dominantly discusses the impacts of different configurations of TNA and NIO SST warming on the WPAC. In fact, in addition to the TNA and NIO SST anomalies, the SST anomalies over the Pacific can also modulate the seasonal variation of WPAC (Wang 2019). Therefore, considering three ocean SST anomalies together is helpful to deeply understand the WPAC activity in the seasonal time scale, which will be explored in future study.

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Data availability statement. The reanalysis data were obtained from NCEP-NCAR (https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.html). The SST data were obtained online (https://www.metoffice.gov.uk/hadobs/hadisst/data/download.html). The GPCP data were obtained online (https://psl.noaa.gov/data/gridded/data.cmap.html). The outputs of the model experiments in this study are available from the corresponding author upon request.

REFERENCES


