Relative Contributions of the Indian Ocean and Local SST Anomalies to the Maintenance of the Western North Pacific Anomalous Anticyclone during the El Niño Decaying Summer*

BO WU
LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences, and Graduate University of Chinese Academy of Sciences, Beijing, China

TIM LI
IPRC, and Department of Meteorology, University of Hawaii at Manoa, Honolulu, Hawaii

TIANJUN ZHOU
LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

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ABSTRACT

To investigate the relative role of the cold SST anomaly (SSTA) in the western North Pacific (WNP) or Indian Ocean basin mode (IOBM) in maintaining an anomalous anticyclone over the western North Pacific (WNPAC) during the El Niño decaying summer, a suite of numerical experiments is performed using an atmospheric general circulation model, ECHAM4. In sensitive experiments, the El Niño composite SSTA is specified in either the WNP or the tropical Indian Ocean, while the climatological SST is specified elsewhere. The results indicate that the WNPAC is maintained by the combined effects of the local forcing of the negative SSTA in the WNP and the remote forcing from the IOBM. The former (latter) contribution gradually weakens (enhances) from June to August. The negative SSTA in the WNP is crucial for the maintenance of the WNPAC in early summer. However, because of a negative air–sea feedback, the negative SSTA gradually decays, as does the local forcing effect. Enhanced local convection associated with the IOBM stimulates atmospheric Kelvin waves over the equatorial western Pacific. The impact of the Kelvin waves on the WNP circulation depends on the formation of the climatological WNP monsoon trough, which does not fully establish until late summer. Therefore, the IOBM plays a crucial role in late summer via the Kelvin wave induced anticyclonic shear and boundary layer divergence.

1. Introduction

The most pronounced low-level circulation anomalies over the tropical Pacific during El Niño decaying summer is an anomalous anticyclone over the western North Pacific (WNPAC; see review papers by Li and Wang 2005; Lau and Wang 2006). The WNPAC, persisting from the El Niño mature winter to the subsequent summer, plays a crucial role in the El Niño–East Asian summer monsoon teleconnection (Zhang et al. 1996; Wang et al. 2000; Chang et al. 2000a,b; Lau and Nath 2000; Wang and Zhang 2002; Wu et al. 2003; Li et al. 2007; Zhou et al. 2009a).

During the El Niño mature winter and the following spring, the WNPAC would enhance the precipitation over southeastern China through anomalous moisture transport (Zhang and Sumi 2002). During the El Niño decaying summer, the WNPAC would enhance the meiyu–baiu precipitation by modulating the western Pacific subtropical high (Chang et al. 2000a). In addition, the WNPAC forced by negative precipitation anomalies over the Philippine Sea is a part of meridional wave train propagating northward from the Philippine Sea, which

The maintenance of the WNPAC from the El Niño mature winter to the subsequent spring is attributed to a positive thermodynamic air–sea feedback (Wang et al. 2003). A cold sea surface temperature anomaly (SSTA) in the western North Pacific (WNP) suppresses local convection (Su et al. 2000), which stimulates a low-level atmospheric anticyclone anomaly to its west. The north-easterly anomalies to the eastern flank of the WNPAC enhance the mean trade wind and cool the in situ SST through enhanced upward surface latent heat flux (Wang et al. 2000).

With the onset of the WNP summer monsoon, the mean southwesterly monsoon circulation replaces the northeasterly trade wind. As a result, a negative evaporation–wind–SST feedback appears in the region. The negative air–sea feedback weakens the negative SSTA in the WNP during the El Niño decaying summer (Chou et al. 2009). In contrast to the weakening of the local negative SSTA, the WNPAC is strengthened from the preceding spring to the summer (Wu et al. 2009b).

A possible cause of strengthening of the WNPAC during the El Niño decaying summer is attributed to the remote SSTA forcing from the tropical Indian Ocean (TIO; see Yang et al. 2007; Li et al. 2008; Wu et al. 2009b; Xie et al. 2009). An Indian Ocean basin mode (IOBM) establishes after the ENSO mature winter. The basinwide warming in the TIO is possibly caused by anomalous surface heat flux associated with the descending branch of the anomalous Walker circulation (Klein et al. 1999; Venzke et al. 2000; Lau and Nath 2003) or the change of tropical tropospheric temperature (Chiang and Sobel 2002). In addition, ocean dynamic processes also play a role (Li et al. 2002, 2003). The IOBM may strengthen the WNPAC in boreal summer through the following two processes: First, through enhanced convection over the TIO, the IOBM may stimulate a Kelvin wave–type easterly response in the western Pacific. The anticyclonic shear vorticity associated with the easterly anomalies may weaken the WNP summer monsoon through Ekman pumping divergences (Wu et al. 2009b; Xie et al. 2009). Second, the IOBM may increase the surface moisture and thus enhance the low-level moisture transport and convection over the Maritime Continent, the latter of which may further induce the subsidence over the WNP through an anomalous Hadley circulation (Chang and Li 2000; Li et al. 2001; Sui et al. 2007; Wu and Zhou 2008; Wu et al. 2009b).

Both the local forcing of the negative SSTA in the WNP and the remote SSTA forcing from the TIO may impact the WNPAC during the El Niño decaying summer. However, their relative contributions are still not clear. The clarification of this issue may help us to identify better predictors for climate prediction. Through just an observational analysis, it is difficult to indentify the relative contributions of the local and remote forcing. In the study, our strategy is to examine the response of an atmospheric general circulation model (AGCM) to specified regional SSTA forcing.

The rest of the paper is organized as follow. A description of the model and data is provided in section 2. Section 3 reveals the essential characteristics of the atmospheric circulation, precipitation, and surface flux anomalies during the El Niño decaying summer through observational diagnosis. The results of numerical experiments are analyzed in section 4, in which we focus on the relative contribution of the TIO and WNP in maintaining the WNPAC. Section 5 summarizes our findings.

2. Model and data description

The AGCM used in the study is ECHAM version 4.6 (hereafter ECHAM4), which was developed by the Max Planck Institute for Meteorology (MPI; Roeckner et al. 1996). The model was run at a horizontal resolution of spectral triangular 42 (T42), roughly equivalent to 2.8 latitude \( \times \) 2.8 longitude, with 19 vertical levels in a hybrid sigma–pressure coordinate system extending from surface to 10 hPa. The mass flux scheme of Tiedtke (1989) is used to represent the deep, shallow, and middle convection. The original moisture convergence closure has been replaced by a convective available potential energy (CAPE)-based closure scheme (Nordeng 1994). Previous evaluation shows that ECHAM4 is one of the best AGCMs in the simulation of the Asian–Australian monsoon (Zhou et al. 2009a), and it thereby has been used in many modeling studies of monsoon variability (Fu et al. 2008; Li et al. 2008; Zhou et al. 2009b).

The SST data constructed for phase II of the Atmospheric Model Intercomparison Project (AMIP II) are used as the model lower boundary condition and for the observation analysis (Hurrell et al. 2008). The dataset is a merged product based on the Hadley Centre sea ice and SST dataset version 1 (Rayner et al. 2003) and version 2 of the National Oceanic and Atmospheric Administration (NOAA) optimum interpolation SST analysis (Reynolds and Smith 1994).

Other observational datasets used in the study include 1) the observational precipitation field derived from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) for the period of 1979–2007 (Xie and Arkin 1997); 2) National Centers for Environment Prediction (NCEP)/Department of Energy (DOE) AMIP II Reanalysis (Kanamitsu et al. 2002) for the period of
1979–2007; 3) objectively analyzed air–sea fluxes (OAFlux) for global oceans (Yu et al. 2008), with latent heat flux from 1982 to 2004 and shortwave radiative flux from 1984 to 2004. To focus on the interannual time scale, variations longer than 8 yr are filtered out from the original datasets with a Lanczos filter (Duchon 1979).

In the paper, all of the observational analysis and numerical experiments are based on five strong El Niño events (i.e., the 1982, 1991, 1994, 1997, and 2004 events) selected from the period of 1979–2008. In the period, the rest of two strong events, the 1986 and 2002 events, are excluded for following reasons. The 1986 event persists for 2 yr, which is much longer than the normal El Niño life cycle. During its decaying summer (the 1988 summer), a cold SSTA, rather than a warm SSTA, covers large area of the TIO, so that the in situ precipitation is suppressed, rather than enhanced as conventional El Niño events (figure not shown). For the 2002 events, a significant Indian Ocean dipole pattern, rather than an IOBM, is seen in the TIO (figure not shown) during its decaying summer. To enhance the Indian Ocean and local SSTA forcing signals, we exclude the two “unconventional” El Niño events.

3. Observational analysis

The composite SSTA evolution patterns during the El Niño decaying summer are shown in Fig. 1. The TIO is controlled by the IOBM throughout the summer. The largest warm SSTA appears in the southeastern Indian Ocean and the tropical western Indian Ocean. The amplitude of the warm SSTA in the TIO slightly weakens from June to August. Over the WNP, a significant negative SSTA occurs in June, and it gradually attenuates from west to east and weakens with time (Fig. 1).

Corresponding to the IOBM during the El Niño decaying summer, convection over the TIO is enhanced (Figs. 2a–c). The maximum precipitation anomalies are located in the northern and southeastern Indian Ocean. In contrast, negative precipitation anomalies persistently control the WNP throughout the summer. The negative precipitation center slowly moves eastward from June to July. As a Rossby wave response to the negative heating, a large-scale anomalous anticyclone (WNPAC) extending from 110° to 170°E maintains over the WNP throughout the summer (Figs. 2d–f). Although the WNPAC persists throughout the summer, its impact on the East Asian summer monsoon is the most significant in June. A positive anomalous rainband extends from central China to southern Japan (Fig. 2a), indicating that the mei-yu–baiu rainband is significantly enhanced. The temporal dependence of the WNPAC impact is associated with the northward movement of the East Asian climatological rainfall band (Tao and Chen 1987).

Fig. 1. Observed monthly mean SST anomalies (°C) from June to August for the El Niño composite. The solid lines represent the 5% significant level.

The WNPAC and associated negative precipitation anomalies have a strong feedback to the underlying SST (Chou et al. 2009; Wu et al. 2009a). Figure 3 shows the June–August mean downward shortwave radiative and downward latent heat flux anomalies for the El Niño composite. They indicate that the decay of the negative SSTA in the WNP is caused by the combined effects of increased downward shortwave radiative and downward latent heat flux anomalies. The increased shortwave radiation is closely associated with the negative in situ precipitation anomalies for their analogous spatial patterns throughout the summer (Figs. 3a,b). The increase of the downward latent heat flux is caused by the easterly anomalies in the southern flank of the WNPAC, which decrease the southwesterly wind speed of the WNP monsoon. Because the monsoon southwesterly flow is established from west to east, the positive downward latent
heat flux anomalies in the Philippine Sea gradually intensify and expand eastward (Figs. 3d–f). It is worthy to note that even in the presence of the negative thermodynamic air–sea feedback, the amplitude of the negative SSTA in the WNP is still quite large in early summer. It is expected that such a strong SSTA may have a significant impact on the local circulation.

4. Numerical experiments

a. Experiment design

To reveal the relative contribution from the TIO and WNP SSTA forcing, we design idealized numerical experiments in which we separate the anomalous SST forcing from the TIO and WNP. Figure 4 shows the model geographic locations for the TIO and WNP domains.

Table 1 lists four sets of designed numerical experiments. The first is a control run, in which ECAHM4 was integrated for 20 yr, forced by monthly climatological SST. The climatological SST is based on the period of 1979–2004. This experiment is referred to as CTRL run. To investigate the relative contributions of the remote TIO forcing and the local WNP forcing to the WNPAC during the El Niño decaying summer, three sets of sensitivity experiments are designed, in which the composite SSTA shown in Fig. 1 was added to the climatological SST in the global ocean (hereafter the GB run), the TIO only (hereafter the TIO run), or the WNP only (hereafter the WNP run). Note that in the TIO (WNP) run the SST climatology was prescribed in the regions outside of the TIO (WNP). Each experiment was integrated from April to August, with 20 ensemble members. The ensemble mean results of June–August are analyzed.
The difference between the GB and CTRL runs provides an evaluation on the performance of the model in reproducing the WNPAC. The difference between the TIO and CTRL runs examines the sole contribution of the remote IOBM forcing to the WNPAC, and the difference between the WNP and CTRL runs examines the sole contribution of the local WNP negative SSTA forcing. A Student’s $t$ test is used to examine the statistical significance of model response.

b. Model climatology

Because of the close relationship between the WNP circulation anomaly and the climatological circulation (Chou et al. 2009; Wu et al. 2009b), it is necessary to evaluate the model’s performance in the climatology simulation. The simulations of June–August precipitation and 850-hPa wind fields derived from the CTRL run are shown in Fig. 5, along with the observed counterparts. In boreal summer, the mean precipitation centers are located in the southeastern Indian Ocean, the Arabian Sea, the Bay of Bengal, the South China Sea, and the WNP, respectively. Correspondingly, strong low-level winds blow from the Southern Hemisphere to the Northern Hemisphere and from the Arabian Sea toward the Philippine Sea. In June, the westerlies are confined to the west of the South China Sea, and then they gradually extend eastward as the summer progresses. The rapid eastward expansion of the westerlies in late summer is accompanied by an enhanced convection associated with the onset of the WNP monsoon. Over the East Asia, the most remarkable feature is a mei-yu–baiu rainband extending from central China to south of Japan in June.

The pattern correlations between the simulated and observed climatological precipitation fields in the region
The correlation coefficients for June, July, and August reach 0.65, 0.68, and 0.69, respectively. ECHAM4 reproduces the major features of the Asian summer monsoon (Figs. 5d–f), such as the precipitation centers over the Arabian Sea and the Bay of Bengal, the Somalia cross-equatorial jet, and westerlies blowing from the Arabian Sea to the Bay of Bengal. The model also reasonably reproduces the WNP monsoon trough and its buildup and eastward advance from June to August. The main deficiency is that the simulated monsoon trough shifts northward by about 5° latitude and extends eastward excessively. The simulated WNP rainband is stronger than the observation. The model also fails to simulate the mei-yu–baiu rainband. The discrepancies of the simulated WNP monsoon trough and mei-yu–baiu rainband have been common problems for many AGCMs (Lau et al. 1996; Kang et al. 2002).

c. Response of the WNPAC to remote and local SSTA forcing

The precipitation and 850-hPa wind anomalies simulated by the GB run are shown in Fig. 6. The GB run realistically reproduces low-level circulation anomalies in the WNP from June to August. Some key characteristics of the stimulated WNPAC, such as zonal extension, pronounced easterly anomalies in its southern flank, and eastward migration of the anticyclone center from June to August, resemble the observation. The associated negative precipitation anomalies in the WNP are also reasonably simulated. The major difference between the GB run and the observation is that the simulated WNPAC and negative precipitation anomalies are slightly stronger than those in the observation. Compared with the observation, the negative precipitation anomalies shift northward slightly, and unrealistic positive precipitation anomalies appear in the equatorial WNP. The discrepancies may be associated with the northward shift of the climatological rainband over the WNP.

In the GB run, the TIO is dominated by positive precipitation anomalies over the tropical southeastern and northwestern Indian Ocean. The amplitude of the positive precipitation anomalies gradually intensifies from June to August, consistent with the observation. Negative precipitation anomalies and biased atmospheric circulation anomalies appear in some part of the TIO in June and August. However, they are statistically insignificant. The net effect of the anomalous heating over the TIO is enhanced convection as anomalous low-level winds to its east converge into the positive heating region.

Although the simulated WNPAC resembles the observation, the simulated mei-yu–baiu rainfall anomalies are much weaker than those observed (Fig. 6a). The discrepancy is associated with the model climatology, that is, the model fails to reproduce the climatological mei-yu–baiu rainband (Fig. 5).

Corresponding to the reasonable precipitation and low-level wind anomalies over the WNP, the model reproduces the major characteristics of the surface flux anomalies (Fig. 7). The positive downward shortwave radiative flux anomaly caused by the negative precipitation anomaly is dominant throughout the summer, although its meridional extent is narrower and its strength is greater than that observed (Figs. 7a–c). Meanwhile, the model reproduces the eastward extension of the positive downward latent heat flux anomalies associated with the gradually establishment of the WNP summer monsoon trough (Figs. 7d–f).

<table>
<thead>
<tr>
<th>Experiment</th>
<th>SST forcing field</th>
<th>Integration</th>
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<tbody>
<tr>
<td>Control run (CTRL)</td>
<td>Global climatological SST</td>
<td>20 yr</td>
</tr>
<tr>
<td>Global SSTA forcing (GB)</td>
<td>Add the composite SSTA to climatological SST in the global ocean</td>
<td>20 realizations</td>
</tr>
<tr>
<td>Tropical Indian Ocean forcing (TIO)</td>
<td>Add the composite SSTA to climatological SST in the tropical Indian Ocean only</td>
<td>20 realizations</td>
</tr>
<tr>
<td>Western North Pacific forcing (WNP)</td>
<td>Add the composite SSTA to climatological SST in the western North Pacific only</td>
<td>20 realizations</td>
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The reasonable responses of the GB run in reproducing the WNPAC give us confidence to further investigate the relative roles of the remote IOBM forcing and the local WNP forcing through comparing the results of the WNP and TIO runs with that of the GB run.

The precipitation and 850-hPa wind anomalies simulated by the WNP run are shown in Fig. 8. The WNPAC and local negative precipitation anomalies are reproduced throughout the summer. In June, the patterns of the WNPAC and negative precipitation anomalies in the WNP run resemble those in the GB run (Figs. 6a,d), with weaker amplitudes. In July, the WNPAC and the local negative precipitation anomalies further weaken. In August, a positive precipitation anomaly appears over the Philippine Sea in response to the underlying warm SSTA. This causes the eastward shift of the WNPAC by more than 10° longitude, relative to that in the GB run. For all 3 months, easterly anomalies in the southern flank of the WNPAC simulated by the GB run extend to the west of the South China Sea (cf. Fig. 6), whereas the easterly anomalies in the WNP run are confined to the east of the Philippine Sea. The difference between the WNP and GB runs is mainly attributed to the lack of the remote forcing from the TIO.

The precipitation and 850-hPa wind anomalies in the TIO run are shown in Fig. 9. The WNPAC is reproduced throughout the summer. Compared with the WNP run, the WNPAC and the associated negative precipitation anomalies in the TIO run are clearly shifted westward and trapped in the South China Sea and Philippine Sea, particular in June and July. The intensity of the WNPAC gradually intensifies from June to August. In June, easterly anomalies appear in the tropical Pacific, but the anticyclonic vorticity is weak and confined to the South China Sea. In July and August, the WNPAC significantly strengthens and expands eastward. The eastward expansion of the WNPAC is accompanied by the establishment of the WNP monsoon trough (Fig. 5).

The mechanism through which the IOBM impacts the WNP circulation anomaly during the El Niño decaying summer was discussed in Wu et al. (2009b). Enhanced convective heating anomalies in the TIO stimulate Kelvin
The above idealized numerical experiments demonstrate that both the remote IOBM forcing and the local WNP SSTA forcing contribute to the maintenance of the WNPAC during the El Niño decaying summer. To reveal the relative contribution of the remote and local forcing, we calculate the area-averaged vorticity anomaly over the region of 10°–35°N, 120°–160°E from June to August, simulated by the WNP and TIO runs. The results are shown in the Fig. 10. While the seasonally averaged vorticity intensities are comparable in the two runs, the subseasonal evolutions are different. The local SSTA forcing leads to the strongest WNPAC response in June, and the response weakens in July and even turns to a cyclone anomaly in August. The remote TIO forcing, on the other hand, is relatively weak in early summer and strengthens from June to August. While the wave–type easterly anomalies in the western Pacific. The anticyclonic shear of the easterly anomalies causes a boundary layer divergence in the off-equatorial WNP through an Ekman pumping processes. The extent to which the boundary layer divergence affects the local precipitation anomaly depends on its influences on the mean state in the WNP. In June, the WNP monsoon trough has not fully established. As a result, the divergence has a limited impact on the precipitation and wind anomalies over the WNP. In contrast, in July and August, the WNP monsoon trough fully establishes. The Kelvin wave–induced boundary layer divergence suppresses the mean convection, leading to a significant negative heating anomaly. The negative heating anomaly further stimulates an atmospheric Rossby wave response, and thus a low-level anomalous anticyclone in the WNP.

**Fig. 6.** Difference between the (left) precipitation (mm day⁻¹) and (right) 850-hPa wind (m s⁻¹) between the GB and CTRL runs. The contour values at left panel are ±1, ±2.5, and ±4. Solid (dashed) lines denote positive (negative) values. The shading represents the 5% significant level for the (left) precipitation and (right) zonal wind.
weakening of the local response is primarily attributed to the weakening of the local SSTA because of the negative thermodynamic air–sea feedback, the strengthening of the remote response is caused by the evolution of the mean flow, even though the IOBM weakens slightly from early to late summer.

5. Conclusions and discussion

a. Conclusions

Some recent studies (e.g., Xie et al. 2009) stressed the effect of the remote forcing from the tropical Indian Ocean in maintaining the WNPAC during the El Niño decaying summer, by arguing that the seasonal mean SSTA is negligibly small in the WNP resulting from the negative in situ evaporation–wind–SST feedback. Here, with the aid of a set of numerical experiments, we demonstrate that both the negative SSTA in the WNP and the IOBM contribute to the maintenance of the WNPAC. When solely forced by the SSTA in the western North Pacific (WNP run) or the tropical Indian Ocean (TIO run), the WNPAC is somewhat reproduced, but its amplitude is weaker than that forced by the global SSTA (GB run).

The WNPAC that is simulated by the WNP and TIO runs shows distinctive evolution characteristics. In the WNP run, the WNPAC is forced by the local SSTA. A significant negative SSTA appears in the WNP prior to the El Niño decaying summer because of a positive thermodynamic air–sea feedback (Wang et al. 2003). As the summer comes, the reversal of the mean wind leads to a negative air–sea feedback. Despite of the negative feedback, the cold SSTA in the WNP is still quite significant in June and July (Fig. 1) and impacts the local in situ circulation anomaly. The decay of the local SSTA forcing leads to the weakening of the simulated WNPAC from June to August.

Different from the WNP run, the evolution of the WNPAC in the TIO run is associated with the change of
the background mean circulation. Whereas the amplitude of the IOBM weakens slightly from June to August, the simulated WNPAC significantly intensifies and expands eastward. The intensification and expansion are consistent with the establishment of the climatological WNP monsoon trough. As a Kelvin wave response to positive heating anomalies in the TIO, easterly anomalies in the western Pacific cause an anticyclonic shear, and thus a boundary layer divergence over the WNP through Ekman pumping (Wu et al. 2009b). The boundary layer divergence leads to a greater negative precipitation anomaly response in late summer (July and August), when the monsoon trough is fully established. This explains why the remote Indian Ocean forcing effect is strengthened from June to August even though the IOBM weakens.

In summary, the WNPAC during the El Niño decaying summer is attributed to both the remote and local SSTA forcing. In early summer, the WNPAC is primarily supported by the local negative SSTA, whereas in late summer, the IOBM plays a more important role in maintaining the WNPAC.

b. Discussion

Although the results above are robust based on the ECHAM4 numerical experiments, the quantitative estimation of relative contributions of the local and remote SSTA effects on the WNPAC might be model dependent. It is worth noting that the strengths of the WNP cold SSTA and the IOBM differ in each El Niño event. The contributions shown in the ideal experiments may be model dependent, especially under the condition of biased model climatology. Also, in reality there are two-way interactions between the WNP and TIO SST/circulation anomalies; as a result, it is difficult to assess the pure effect of the TIO SSTA on the WNP circulation anomaly. Although both the IOBM and the WNP cold SSTA contribute to the WNPAC, from a prediction point of

FIG. 8. Same as Fig. 6, but for the difference between the WNP and CTRL runs.
view, the latter may be a better predictor for the East Asian summer monsoon. This is because the WNPAC exerts a dominant impact on the East Asian summer monsoon in June, when the climatological mei-yu–baiu rainband is the strongest. Therefore, the WNP negative SSTA, which dominates the WNPAC in June, may deserve more attention for the seasonal predication of the East Asian summer monsoon.

As noted in the introduction, the PJ pattern is primarily excited by the convection anomalies over the Philippine Sea. Because the convection anomalies result from both the local SSTA and the TIO remote forcing during the El Niño decaying summer, it is interesting to investigate their relative roles. Figure 11 shows the 200-hPa vorticity anomalies in the observation and the GB, TIO, and WNP simulations. It is clear that the GB run can simulate the well-organized meridional wave train propagating from the Philippine Sea, which is analogous to that in the observation. However, the vorticity anomalies in the TIO and WNP runs are weaker and loosely organized, suggesting that the formation and maintenance of the PJ pattern may rely on both the remote and local forcing during the El Niño decaying summer.

FIG. 9. Same as Fig. 6, but for the difference between TIO and CTRL runs.

FIG. 10. Temporal evolutions of area-averaged vorticity anomalies ($10^{-6} \text{ m}^2 \text{s}^{-1}$) over the region of $10^\circ$–$35^\circ$N, $120^\circ$–$160^\circ$E for the WNP (solid line) and TIO (dashed lines) runs.
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Fig. 11. (a) Observed composite pattern of June–August mean 200-hPa vorticity anomalies during the El Niño decaying summer. (b) The difference of June–August mean 200-hPa vorticity between (b) the GB, (c) the TIO run, and (d) the WNP run and the CTRL runs.


