Atmospheric Dynamic and Thermodynamic Processes Driving the Western North Pacific Anomalous Anticyclone during El Niño. Part I: Maintenance Mechanisms

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(Manuscript received 6 July 2016, in final form 1 June 2017)

ABSTRACT

The western North Pacific anomalous anticyclone (WNPAC) is an important low-level circulation system that connects El Niño and the East Asian monsoon. In this study, the mechanisms responsible for the formation and maintenance of the WNPAC are explored. Part I of this study focuses on the WNPAC maintenance mechanisms during El Niño mature winter and the following spring. Moisture and moist static energy analyses indicated that the WNPAC is maintained by both the remote forcing from the equatorial central-eastern Pacific via the atmospheric bridge and the local air–sea interactions. Three pacemaker experiments by a coupled global climate model FGOALS-s2, with upper-700-m ocean temperature in the equatorial central-eastern Pacific restored to the observational anomalies plus model climatology, suggest that about 60% (70%) intensity of the WNPAC during the winter (spring) is contributed by the remote forcing from the equatorial central-eastern Pacific. The key remote forcing mechanism responsible for the maintenance of the WNPAC is revealed. In response to El Niño–related positive precipitation anomalies over the equatorial central-eastern Pacific, twin Rossby wave cyclonic anomalies are induced to the west. The northern branch of the twin cyclonic anomalies advects dry and low moist enthalpy air into the tropical western North Pacific, which suppresses local convection. The suppressed convection further drives the WNPAC.

1. Introduction

The western North Pacific anomalous anticyclone (WNPAC), which is also referred to as the Philippine Sea anomalous anticyclone or anomalous northwest Pacific anticyclone, is a core research object in studies of the relationship between El Niño–Southern Oscillation (ENSO) and the East Asian–western North Pacific (EA–WNP) monsoon [see reviews by Wang and Li (2004), Li and Wang (2005), and Zhou et al. (2014a)]. The WNPAC not only influences the EA–WNP winter (Zhang et al. 1996, 1999; Wang et al. 2000; Lau and Nath 2000) and summer (Chang et al. 2000a,b; Yang et al. 2007; Li et al. 2008; Xie et al. 2009; Wu et al. 2009, 2010a; Wang et al. 2013) monsoon, but also contributes to the decay of El Niño (Weisberg and Wang 1997a,b; Wang et al. 1999; Wang et al. 2001; Kug et al. 2006; Li et al. 2007; Ohba and Ueda 2009; Wu et al. 2010b; Okumura et al. 2011; Chen et al. 2016).

The life cycle of the WNPAC is tightly linked with both the phase of El Niño and the annual cycle of the tropical climate. It forms during El Niño–developing fall, fully establishes during El Niño mature winter, maintains during the following spring and summer, and decays after that (e.g., Wang et al. 2000; Wang and Zhang 2002; Wang et al. 2003). At present, it is widely accepted that the maintenance mechanisms responsible for the WNPAC during El Niño mature winter and the following spring are distinct from those for the WNPAC during El Niño decaying summer (Yang et al. 2007; Xie et al. 2009; Wu et al. 2010b; Okumura et al. 2011; Chen et al. 2016).
et al. 2009; Wu et al. 2009, 2010b; Wang et al. 2013). However, at present, the WNPAC maintenance mechanisms during winter and spring remain inconclusive.

The temporal evolutions of El Niño–related sea surface temperature anomalies (SSTAs) and the corresponding precipitation and low-level circulation anomalies, which were obtained through regressions against the December (year 0)–February (year 1) [D(0)JF(1)] mean Niño-3.4 index (area-averaged SSTAs over 5°S–5°N, 120°–170°W), are shown in Fig. 1. We used years 0 and 1 to represent El Niño developing and decaying years, respectively. As noted in previous studies (e.g., Rasmusson and Carpenter 1982), typical El Niño events are fully established during boreal summer (Fig. 1b), with warm SSTAs dominating the equatorial central-eastern Pacific (CEP). The warm SSTAs further strengthen in fall, reach a maximum in winter, and then gradually weaken in spring (Figs. 1b,d,f,h). The cold SSTAs in the western North Pacific (WNP) establish in June–August (year 0) [JJA(0)] and are maintained in the following three seasons (Figs. 1b,d,f,h).

From JJA(0) to D(0)JF(1), although the cold SSTAs in the tropical WNP generally tend to intensify, local precipitation and circulation anomalies experience remarkable changes in their spatial patterns. In JJA(0), the tropical WNP is dominated by positive precipitation and cyclonic anomalies (Fig. 1a). In September–November (year 0) [SON(0)], the positive precipitation anomalies and the center of the cyclonic anomalies shift eastward and negative precipitation anomalies are seen over the east of the Philippine Sea (Fig. 1c). In D(0)JF(1), the negative precipitation anomalies dominate the entire tropical WNP and the WNPAC fully establishes (Fig. 1e). The negative precipitation anomalies and WNPAC are maintained during the following spring (Fig. 1g).

Wang et al. (2000, 2003) proposed that the persistence of the WNPAC relies on the local wind–evaporation–SST
feedback. The northeasterly anomalies to the southeastern flank of the WNPAC strengthen the climatological northeasterly trade wind and thus enhance evaporation and cool the SST in situ. Lau and Nath (2003) conducted numerical experiments to support this mechanism. On the contrary, Stuecker et al. (2015) proposed that the life cycle of the WNPAC does not rely on the local air–sea interactions in the WNP, but instead on interactions between El Niño and the annual cycle. They designed idealized SST boundary conditions to drive an atmospheric general circulation model (AGCM), in which the SSTAs were only specified in the equatorial CEP. The model reproduced the major life cycle of the WNPAC. However, the detailed dynamic processes causing the formation and maintenance of the WNPAC were not studied in Stuecker et al. (2015).

As the spatial pattern of the SSTAs from JJA(0) to MAM(1) does not change greatly, it is conceivable that the tight seasonal phase locking of the WNPAC is caused by the change in the background climatological state. However, what seasonal change in the climatological state plays a fundamental role and through what dynamic or thermodynamic processes the background state modulates the precipitation and circulation anomalies remain unknown.

This study consists of two parts. In this paper (Part I), we aim to answer the following question: what are the mechanisms responsible for the persistence of the WNPAC during El Niño mature winter and the following spring?

The remaining sections of this paper are arranged as follows. Section 2 introduces the observational and reanalysis datasets, analysis methods, and numerical models used in the study. Section 3 explores the WNPAC maintenance mechanisms during El Niño mature winter. Section 4 investigates whether the mechanisms proposed in section 3 work in the following spring. Section 5 discusses some uncertain issues and summarizes the major conclusions.

2. Datasets, methods, and models

a. Datasets

The datasets used in this study were 1) monthly precipitation data from the Global Precipitation Climatology Project (GPCP; Adler et al. 2003); 2) monthly SST data from the Met Office Hadley Centre Sea Ice and SST dataset (HadISST; Rayner et al. 2003); and 3) 6-hourly wind, geopotential height, temperature, specific humidity, and surface heat fluxes from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011). The horizontal resolutions of the GPCP, HadISST, and ERA-Interim datasets were $2.5^\circ \times 2.5^\circ$, $1^\circ \times 1^\circ$, and $1.5^\circ \times 1.5^\circ$, respectively. All the datasets cover the period 1979–2012.

b. Methods

To understand the mechanisms responsible for the precipitation anomalies, a moisture equation was diagnosed. The vertically integrated anomalous moisture equation is written as

\[
\partial_t (\hat{c}_p T + L_v q) + \langle \mathbf{u} \cdot \nabla_h (\hat{c}_p T + L_v q) \rangle - \langle \omega \hat{q} \rangle - \langle \omega' \hat{p} \rangle + F'_{\text{net}},
\]

where the primes denote the monthly anomaly with the climatological annual cycle removed; angle brackets denote the mass integral through an entire atmospheric column; \( q \) denotes specific humidity; \( \mathbf{u} \) and \( \omega \) denote horizontal wind and vertical pressure velocity, respectively; and \( E \) and \( P \) represent evaporation and precipitation, respectively. All the terms are derived from the 6-h ERA-Interim dataset.

On the interannual time scale, the time tendency term [first term of Eq. (1)] is far smaller than the other terms and is thus negligible. The advection terms can be further decomposed into two linear advection terms, a nonlinear advection term and a nonlinear transient term. Therefore, Eq. (1) becomes

\[
P' = E' - \langle \mathbf{u} \cdot \nabla_h q' \rangle - \langle \mathbf{u}' \cdot \nabla_h q' \rangle - \langle \omega \hat{q} \rangle - \langle \omega' \hat{p} \rangle + NL,
\]

in which NL represents the sum of all the nonlinear terms, and the overbars denote the monthly mean. As will be shown below, the precipitation anomalies over the tropical WNP are primarily dominated by the anomalous vertical motion. A further question is, what causes the anomalous vertical motion? This cannot be answered by Eq. (2). Here we try to address this issue from the perspective of energy balance.

In the tropics, vertical motion is constrained by the moist static energy (MSE) budget (Neelin and Held 1987). We diagnosed the MSE equation for the study region. Following Neelin (2007), the MSE equation can be written as

\[
\partial_t \langle c_p T + L_v q \rangle' + \langle \mathbf{u} \cdot \nabla_h (c_p T + L_v q) \rangle' + \langle \omega \hat{h} \rangle' = F'_{\text{net}},
\]

where MSE is represented by \( h = c_p T + L_v q + \phi \); \( \phi \) denotes the geopotential; \( T \) denotes air temperature; \( c_p \) and \( L_v \) denote the specific heat at constant pressure and the latent heat of vaporization, respectively; and \( c_p T + L_v q \) is moist enthalpy. The net MSE flux coming
into the atmospheric column from the surface and the top of atmosphere is denoted by $F_{\text{net}}$, which is equal to

$$F_{\text{net}} = 5^{\text{S}_{\text{yt}}^2} 5^{\text{S}_{\text{yt}}^2} 5^{\text{R}_{\text{yt}}^1} 5^{\text{R}_{\text{yt}}^1} 5^{\text{LH}^2} 5^{\text{SH}}.$$

(4)

The first three terms in parentheses are the radiative fluxes at the top of atmosphere, including the downward and upward shortwave radiative fluxes ($S_{\text{yt}}^1$ and $S_{\text{yt}}^1$, respectively) and the upward longwave radiative flux $R_{\text{yt}}^1$, which are positive when heating the atmosphere. The last six terms in parentheses are the surface fluxes, including the downward and upward shortwave radiative fluxes ($S_{\text{ys}}^1$ and $S_{\text{ys}}^1$, respectively), the downward and upward longwave radiative fluxes ($R_{\text{ys}}^1$ and $R_{\text{ys}}^1$, respectively), and the latent heat (LH) and sensible heat (SH) fluxes, which are positive when heating the ocean.

Similar to the moisture equation, the MSE equation can be simplified to

$$h_v \partial_p \tilde{h} \approx F'_{\text{net}} - \langle \mathbf{u} \cdot \nabla_h (c_p T + L_e q) \rangle - \langle \mathbf{u}' \cdot \nabla_h (c_p T + L_e q) \rangle - \langle \mathbf{v} \cdot \nabla_h \mathbf{p} \rangle + \mathbf{h'} + \mathbf{NL}.$$

(5)

In the tropical WNP with active deep convection, the vertical profile of $\omega'$ shows a top-heavy bow structure with a maximum at about 450 hPa (Fig. 2), suggesting that it is dominated by the deep mode of the tropical vertical motion (Back and Bretherton 2009). Meanwhile, the vertical profile of climatological MSE $\tilde{h}$ shows a bottom-heavy bow structure with a maximum at 750–600 hPa (Fig. 2). A previous study noted that the ascending motion of the vertical profile of the deep mode tends to export the MSE out of the atmospheric column in the tropics, causing positive gross moisture stability of the column (Back and Bretherton 2009). If the physical processes on the right-hand side of Eq. (5) tend to reduce (increase) the MSE in the column, descending (ascending) anomalies should be generated to decrease (increase) the MSE exported out of the column and thus keep the MSE budget balance.

c. Models

A state-of-the-art coupled global climate model (CGCM) FGOALS-s2 was used in this study. The model was developed by the State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG) at the Institute of Atmospheric Physics (IAP), Chinese Academy of Sciences (CAS) (Bao et al. 2013; Zhou et al. 2014b). Its four components—atmosphere, land, ocean and sea ice—are coupled together by the CCSM3.0 coupler, version 6.0, developed in the National Center for Atmospheric Research (NCAR) (Collins et al. 2006). The atmospheric component is Spectral Atmosphere Model of IAP LASG, version 2 (SAMIL2), with the horizontal resolution of about 2.81 $\lambda$ longitude $\times$ 1.66 $\lambda$ latitude and 26 levels in the vertical direction (Bao et al. 2013). The ocean component is LASG IAP Climate System Ocean Model, version 2 (LICOM2), with a horizontal resolution of about 1 $\lambda$ $\times$ 1 $\lambda$ in the extratropical zone and 0.5 $\lambda$ $\times$ 0.5 $\lambda$ in the tropics and 30 levels in the vertical direction (Liu et al. 2012). Evaluation on the performances of CMIP5 models shows that the FGOALS-s2 is one of the best models in simulating the WNPAC, with amplitude of the simulated WNPAC comparable with that in the observation (Wu and Zhou 2016).

3. WNPAC during El Niño mature winter

a. Observational analysis

In the observations, the WNPAC is tightly linked to the local negative precipitation anomalies (Fig. 1e). In terms of the Gill model, tropical circulation anomalies can be understood as passive responses to specified diabatic heating anomalies (Gill 1980). Hence, we focused on the processes causing the negative precipitation anomalies over the tropical WNP.

Moisture budget analysis indicated that the negative precipitation anomalies over the tropical WNP are
caused by negative anomalous advection of the climatological vertical moisture by descending anomalies \( \langle -\omega h \rangle \), while the evaporation and horizontal moisture advection anomalies have small contributions (Fig. 3a). In the tropics, the vertical motions are constrained by the MSE budget balance. The descending anomalies tend to decrease energy exported out of the atmospheric column in deep convection regions; that is, anomalous vertical advections of the climatological MSE \( \langle -\omega h \rangle \) are positive [Fig. 3b; the advection term is moved to the left side in Eq. (5) and is thus multiplied by \(-1\)]. MSE budget analysis indicated that this term is primarily balanced by negative net energy flux anomalies \( F_{\text{net}} \), the horizontal advection of climatological enthalpy by anomalous wind \( \langle -u \cdot \nabla_h (c_p T + L_e q) \rangle \), and anomalous vertical MSE advection by climatological vertical motion \( \langle -\omega h \rangle \) (Fig. 3b). These three terms represent different physical processes, which were individually analyzed as follows.

1) NET MSE FLUX

In terms of the MSE budget analysis, the negative net MSE flux \( F_{\text{net}} \) had the largest contribution to the anomalous descending motion over the tropical WNP (Fig. 3b). We show below that the negative \( F_{\text{net}} \) is generated by a local atmospheric internal positive feedback and thus cannot be the mechanistic root of the negative precipitation anomalies over the tropical WNP and the WNPAC.

As shown in Eq. (4), \( F_{\text{net}} \) includes nine components. Considering that the atmosphere is largely transparent to solar radiation, the related terms in Eq. (2) were merged together to represent the net solar radiative flux. The net longwave radiative flux anomalies were separated into clear-sky and cloud-related components. Hence, Eq. (4) was simplified to

\[
F_{\text{net}} = R_{\text{cloud}} + R_{\text{clear}} + S_{\text{net}} + LH + SH.
\] (6)

Figure 4 shows the spatial patterns of \( F'_{\text{net}} \), \( R'_{\text{cloud}} \), and \( LH' \) during El Niño mature winter. It can be seen that \( F'_{\text{net}} \) over the tropical WNP is dominated by the longwave cloud radiative forcing anomalies (more than 60%), while the contributions from the net solar radiative, clear-sky longwave radiative, and sensible and latent heat flux anomalies are only about 14.9%, 12.9%, 11.8% (not shown), and 0.9%, respectively. The spatial pattern of \( R'_{\text{cloud}} \) highly resembles that of the precipitation anomalies over the tropical WNP (Figs. 4b and 1c). This is because \( R'_{\text{cloud}} \) over the tropical ocean is primarily generated by an internal positive feedback between convection and cloud radiative forcing in the
tropical atmosphere (Su and Neelin 2002; Neelin and Su 2005; Bretherton and Sobel 2002). The suppressed deep convection over the tropical WNP corresponds to a decrease in deep convective cloud and associated cirrostratus and cirrocumulus. Hence, the warming effect of the longwave cloud radiative forcing weakens. The net cooling effect for the atmospheric column further strengthens the gross moist stability of the atmospheric column and thus suppresses the active ascending motion and associated deep convolutions over the tropical WNP.

The above analysis indicates that in the tropical WNP, the negative net MSE flux anomalies are primarily contributed by the weakened longwave cloud radiative forcing anomalies induced by the internal positive feedback with the suppressed convection, while the latent heat flux anomalies have no contribution. The situation is distinct from the equatorial CEP, where El Niño–related warm SSTAs drive the atmosphere by increasing the upward latent heat flux, although internal positive feedback due to longwave cloud radiative forcing is still very important (Su and Neelin 2002; also seen in Figs. 4b,c).

2) **Horizontal Advection of Climatological Moist Enthalpy by Anomalous Wind**

The \( \langle -\mathbf{u} \cdot \nabla_h (c_p T + L_o q) \rangle \) term is the second-largest term on the right-hand side of Eq. (5) (Fig. 3b). The moist enthalpy \( c_p T + L_o q \) is a function of both temperature and specific humidity. Analysis indicated that the \( \langle -\mathbf{u} \cdot \nabla_h (c_p T + L_o q) \rangle \) term is primarily contributed by moisture advection in the tropical WNP (>84%, figure not shown); therefore, we used \( \langle -\mathbf{u} \cdot \nabla_h (L_o q) \rangle \) to represent \( \langle -\mathbf{u} \cdot \nabla_h (c_p T + L_o q) \rangle \). As we know, most moisture is confined in the planetary boundary layer. Figure 5 shows climatological 925-hPa specific humidity and anomalous wind during El Niño mature winter. The spatial pattern of climatological specific humidity resembles the climatological wintertime intertropical convergence zone (ITCZ; Fig. 5a). It has a strong negative meridional gradient in the tropical WNP. The anomalous northerly wind component advects dry (low moist enthalpy) air into the tropical WNP (Fig. 5b), which weakens upward motion and thus precipitation there.

The anomalous northerly wind component is part of both the cyclonic anomalies generated by the positive precipitation anomalies over the equatorial CEP and the WNPAC itself (Fig. 5a), which indicates that negative \( \langle -\mathbf{u} \cdot \nabla_h (c_p T + L_o q) \rangle \) is associated with both remote forcing from the CEP and a local positive feedback. This positive feedback is referred to as wind–moist enthalpy advection–convection feedback hereafter. The northeasterly anomalies to the southeastern flank of the WNPAC suppress precipitation over the tropical WNP through advection of off-equatorial dry (low moist enthalpy) air. The suppressed convection further feeds the WNPAC to its west in terms of the Gill model.

A similar mechanism associated with the \( \langle -\mathbf{u} \cdot \nabla_h (c_p T + L_o q) \rangle \) term has been used to explain the formation of the descending anomalies to the north of El Niño–related positive precipitation anomalies (Su and Neelin 2002) and the descending branch of the anomalous Walker circulation associated with El Niño (Ham et al. 2007).

3) **Vertical Advection of Anomalous MSE by Climatological Vertical Motion**

Compared with the horizontal moist enthalpy advection, the \( \langle \nabla \cdot h \rangle \) term plays a secondary role (Fig. 3b). In the tropics, this term is related to the change in the gross moist stability caused by the change in the vertical structure of the MSE \( h \). Figure 6 shows the vertical profiles of the D(0)JF(1)-mean area-averaged \( h' \) over the tropical WNP, together with its subcomponents: dry static energy \( s' = c_p T' + \phi' \) and latent energy \( L_o q' \). The vertical profile of \( h' \) exhibits a bottom-heavy bow structure with a minimum at about 775 hPa. In the deep convection region, the vertical structure of \( h' \) indicates an increase in the gross moist
stability \((\overline{-\omega\partial h'/\partial p}) < 0\), which tends to suppress local convection.

The vertical structure of \(h'\) is dominated by the \(L_q q'\) component, while the contribution from the anomalous dry static energy \(s'\) is relatively small (Fig. 6). The moisture content changes in tropical atmospheric columns are associated with the following two processes: 1) In the boundary layer, \(q'\) is primarily modulated by underlying SSTAs. The spatial pattern of the 925-hPa specific humidity anomalies resembles that of the SSTAs, with their pattern correlation reaching 0.81 (Figs. 7c,d). 2) In the lower free troposphere, \(q'\) is constrained by the local anomalous vertical motions. The pattern correlation between 750-hPa specific humidity and 750-hPa \(\omega\) for the entire tropics reaches -0.7 (Figs. 7a,b). This process is associated with another atmospheric internal positive feedback, known as the moisture–convection feedback [see review by Emanuel (2007)]. For the tropical WNP, large-scale descending motion anomalies associated with the negative precipitation anomalies cause decreases in the midlevel moisture, which in turn suppresses local convection through increasing gross moist stability. For the entire atmospheric column, the contribution of the SSTA-related negative \(q'\) in the boundary layer to the \((\overline{-\omega\partial h'/\partial p})\) term is largely hidden by that of the negative \(q'\) in the lower free troposphere. The impacts of underlying cold SSTAs are further explored through numerical experiments in section 3b.

The above analyses for individual terms on the right-hand side of Eq. (5) suggest that the negative precipitation anomalies over the tropical WNP are driven by local forcing of the cold SSTAs and remote forcing from the equatorial CEP. The remote forcing is the negative anomalous enthalpy advection by teleconnected northerly anomalies. For the local forcing, the cold SSTAs in the tropical WNP tend to suppress local convection through decreasing overlying moisture and thus increasing atmospheric gross moist stability. The negative precipitation anomalies are further amplified by the three positive feedbacks; that is, the convection–cloud radiative forcing, wind–moist enthalpy advection–convection, and moisture–convection feedbacks.
b. Numerical experiments

Observational analysis indicated that both the remote forcing from the equatorial CEP and local forcing are favorable for the negative precipitation anomalies over the tropical WNP and thus the WNPAC. However, it is impossible to separate the local and remote forcing cleanly in the observational analysis. Hence, we conducted a series of numerical experiments using the CGCM FGOALS-s2 to resolve this issue.

The three sets of designed experiments are listed in Table 1. The first experiment was a pacemaker-coupled simulation. The upper-700-m ocean temperature in the equatorial CEP was restored to the observational anomalies plus model climatology (black dashed triangles in Figs. 8b,d,f), while in the other ocean areas the oceanic and atmospheric components of the model were freely coupled (hereafter resCEP run). In the second experiment, the upper-700-m ocean temperature in the equatorial CEP was also restored to the observational anomalies plus model climatology, while in the other ocean areas, the ocean temperature was restored to the model climatology (hereafter resCEP_clmGLB run). The third experiment was similar to the resCEP_clmGLB run, but only ocean temperatures in the tropical WNP (green dashed quadrilateral in Fig. 8f) are restored to the model climatology. The atmospheric and oceanic components are freely coupled in other ocean areas, as in the resCEP run. This experiment is referred to as resCEP_clmWP run.

In the resCEP clmGLB run, only the direct remote forcing from the equatorial CEP works. In the resCEP_clmWNP run, the direct remote forcing of El Niño still worked, while the air–sea interactions in the tropical WNP were explicitly suppressed. Hence, through the comparisons among the three runs, the relative contributions of the remote forcing and local air–sea interactions to the maintenance of the WNPAC can be separated.

The WNPAC and the negative precipitation anomalies over the tropical WNP were realistically reproduced by the resCEP run (Fig. 8a). Furthermore, the underlying cold SSTAs were also well reproduced, suggesting that the pacemaker experiment can capture the air–sea interactions in the tropical WNP (Fig. 8b). The MSE budget analysis for the resCEP run indicated that the modeled suppressed convection over the tropical WNP (5°–15°N, 125°–165°E; red dashed box in Fig. 8a) is associated with three terms: $F_{\text{net}}$, $\langle -\mathbf{u}' \cdot \nabla h(c_p T + L_vq) \rangle$, and $\langle -\hat{\mathbf{m}}_h \hat{\mathbf{a}} \hat{\mathbf{p}} \rangle$, as in the observational analysis above (Table 2). These results suggested that the physical processes responsible for the maintenance of the observed WNPAC can be reproduced by the model, which gave us confidence to further explore the relative contributions of the local air–sea interactions and remote forcings and their related mechanisms. Compared with the observational analysis, there was one major discrepancy in the MSE budget for the resCEP run: the magnitude of $F_{\text{net}}'$ was much smaller than the $\langle -\mathbf{u}' \cdot \nabla h(c_p T + L_vq) \rangle$ term. This is because the longwave cloud radiative forcing anomalies induced by the internal positive feedback with the suppressed convection are weaker than those in the observation (figure not shown).

The resCEP_clmGLB run represents pure impacts of remote forcing from the equatorial CEP through an atmospheric bridge. The resCEP_clmWNP run reproduced

### Table 1. Experiments using the FGOALS-s2 model.

<table>
<thead>
<tr>
<th>Expt name</th>
<th>Description of the expt design</th>
<th>Integration time</th>
</tr>
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<tbody>
<tr>
<td>resCEP</td>
<td>Pacemaker-coupled run; the upper-700-m ocean temperature in the equatorial CEP (black dashed triangle in Fig. 8b) is restored to the observational anomalies plus the model climatology. Atmospheric and oceanic components are freely coupled in other ocean areas. Refer to the appendix for how to restore the ocean temperature.</td>
<td>1979–2012, three members</td>
</tr>
<tr>
<td>resCEP_clmGLB</td>
<td>The upper-700-m ocean temperature anomalies in the equatorial CEP (black dashed triangle in Fig. 8d) are restored to the observational anomalies. However, in other ocean areas, the upper-700-m ocean temperature is restored to the model climatology. Refer to the appendix for how to restore the ocean temperature.</td>
<td>1979–2012, three members</td>
</tr>
<tr>
<td>resCEP_clmWNP</td>
<td>As in the resCEP_clmGLB, but only the upper-700-m ocean temperature in the tropical WNP (green dashed quadrilateral in Fig. 8f) is restored to the model climatology. The atmospheric and oceanic components are freely coupled in other ocean areas.</td>
<td>1979–2012, three members</td>
</tr>
</tbody>
</table>
the WNPAC (Fig. 8c). However, the intensity of the WNPAC in it was weaker than that in the resCEP run (Figs. 8a,c). The magnitude of area-averaged 925-hPa streamfunction anomalies over 8°–25°N, 105°–165°E in the resCEP_clmGLB run reached about 60% of that in the resCEP run (Figs. 8a,c).

On the other hand, the differences between the resCEP and resCEP_clmWNP runs represent the impacts of the cold SSTAs in the tropical WNP (Fig. 8g). The intensity of the WNPAC in Fig. 8g reaches about 43% of that in the resCEP run (Figs. 8a,c). The differences between (b) and (f). Red dashed box in (a) is used for the MSE budget analysis. The location of the red dashed box is 5°–15°N, 125°–165°E. Black dashed triangles in (b),(d),(f) denote the ocean areas, in which modeled upper-700-m ocean temperatures are restored to the observational anomalies plus the model climatology. Green dashed quadrilateral in (f) denotes the ocean areas in which upper-700-m ocean temperatures are restored to the model climatology.

Table 2. MSE budgets (W m⁻²) over the tropical WNP (5°–15°N, 125°–165°E, red dashed box in Fig. 8a) during El Niño mature winter for the resCEP and resCEP_clmGLB runs.

<table>
<thead>
<tr>
<th></th>
<th>(\langle u'v' \rangle)</th>
<th>(F_{\text{net}})</th>
<th>(-\langle \mathbf{u} \cdot \mathbf{V}_L + L_v q \rangle)</th>
<th>(-\langle \mathbf{u}' \cdot \mathbf{V}_L (T + L_v q) \rangle)</th>
<th>(-\langle \mathbf{m} \cdot f \rho' \rangle)</th>
</tr>
</thead>
<tbody>
<tr>
<td>resCEP</td>
<td>-8.0</td>
<td>-2.3</td>
<td>2.4</td>
<td>-16.4</td>
<td>-3.4</td>
</tr>
<tr>
<td>resCEP_clmGLB</td>
<td>-5.5</td>
<td>5.2</td>
<td>0.4</td>
<td>-11.0</td>
<td>-2.4</td>
</tr>
</tbody>
</table>

Above results indicated that both the remote forcing from the equatorial CEP and the local air–sea interactions contribute to the maintenance of the WNPAC (Fig. 8c). However, the intensity of the WNPAC in it was weaker than that in the resCEP run (Figs. 8a,c). The magnitude of area-averaged 925-hPa streamfunction anomalies over 0°–25°N, 105°–165°E in the resCEP_clmGLB run reached about 60% of that in the resCEP run (Figs. 8a,c).
WNPAC, with scale ratio of 6:4. Further comparisons among these experiments reveal some more interesting results.

First, the $h^2 u_0/C_1 = h(cpT_1L_0q)$ term is the dominant forcing term in the right-hand side of Eq. (5) for both the resCEP and resCEP_clmGLB runs (Table 2), suggesting that the advection of the mean moist enthalpy by anomalous wind plays a central role in the maintenance of the negative precipitation anomalies over the tropical WNP and the WNPAC.

Second, cold SSTAs can contribute to the WNPAC through modulating the local latent heat flux. During El Niño mature winter, the WNP is dominated by the northeasterly trade wind. The northeasterly wind anomalies to the southeastern flank of the WNPAC strengthen the trade wind and thus tend to enhance the evaporation (Wang et al. 2000). However, this effect is greatly offset by the underlying cold SSTAs (cold SSTAs tend to suppress the evaporation). As a result, the latent heat flux anomalies are very weak in both the observation (Fig. 4c) and resCEP run (Fig. 9a). However, if the cold SSTAs were not generated, the evaporation would be greatly intensified by the enhanced wind speed. The corresponding positive latent heat flux anomalies tend to weaken the negative precipitation anomalies and WNPAC in terms of the Eq. (5). This phenomenon can be seen in the resCEP_clmWNP run (Fig. 9b). Because of the strong positive latent heat flux anomalies, even the $F_{net}$ term becomes positive in the resCEP_clmWNP run, opposite to that in the observation and the resCEP run (not shown).

4. WNPAC during El Niño decaying spring

During El Niño decaying spring, the negative precipitation anomalies over the tropical WNP and the WNPAC are still maintained, although their intensities weaken. Do the forcing mechanisms for El Niño mature winter work in the spring? We also conducted budget analyses of moisture and MSE for the spring (Fig. 10). The results were generally consistent with those for the preceding winter. The negative precipitation anomalies over the tropical WNP were caused by a negative $\langle -\omega \cdot \nabla h(c_pT + L_0q) \rangle$ term associated with the anomalous descending motion. The reduced MSE energy export associated with the negative $h^2 u_0/C_1 = h(cpT_1L_0q)$ and $h^2 v_0/h_0$ terms, with their relative contributions similar to those for the preceding winter. As demonstrated above, $F_{net}$ is mostly generated by the atmospheric internal positive feedback between deep convection and longwave cloud radiative forcing. The detailed physical processes associated with the negative $h^2 u_0/C_1 = h(cpT_1L_0q)$ and $h^2 v_0/h_0$ were investigated as follows. Figure 11a shows the 925-hPa wind anomalies and the climatological specific humidity during El Niño decaying spring. It was calculated that the negative meridional gradient of the climatological 925-hPa $q$ over the tropical WNP during boreal spring was very close to that during winter (Figs. 5a and 11a). The northeasterly anomalies to the western flank of the cyclonic anomalies (Fig. 11a) therefore can advect low moist enthalpy air into the tropical WNP as in the preceding winter. The $\langle -u \cdot \nabla h(c_pT + L_0q) \rangle$ term averaged over $1^\circ$–$14^\circ$N, $125^\circ$–$160^\circ$E was $-4.1 \text{ W m}^{-2}$ (Fig. 11b), weaker than $-5.3 \text{ W m}^{-2}$ in the preceding winter (Fig. 5b).
The results indicated that the forcing mechanism of moist enthalpy advection also works during El Niño decaying spring.

The vertical structures of the area-averaged anomalous MSE and its $s'$ and $L_vq'$ subcomponents over the tropical WNP in MAM(1) resembles that in the preceding D(0)JF(1) (figure not shown). This suggests that both the surface cold SSTAs and atmospheric internal feedback between the anomalous downward motions and negative specific humidity anomalies in the lower free troposphere contribute to the negative $\neg \mathbf{\overline{\omega}b'H}'/\partial p$.

As in D(0)JF(1), the magnitude of $\neg \mathbf{\overline{\omega}b'H}'/\partial p$ is far smaller than that of $\neg(\mathbf{u}' \cdot \nabla_h (c_p T + L_v q))$, indicating that the negative precipitation anomalies over the tropical WNP and associated WNPAC are also primarily maintained by the negative enthalpy advection mechanism in MAM(1).

To further investigate the relative importance of the remote forcing from the equatorial CEP and the local air–sea interaction in the maintenance of the WNPAC in MAM(1), we compared the WNPAC in MAM(1) simulated by the resCEP, resCEP_clmGLB, and resCEP_clmWNP runs (Fig. 12). It was calculated that the magnitude of area-averaged 925-hPa streamfunction anomalies over $0^\circ$–$25^\circ$N, $105^\circ$–$165^\circ$E in the resCEP_clmGLB run reached about 73% of that in the resCEP run (Figs. 12a,c), indicating that the remote forcing from the equatorial CEP is a dominant factor for maintaining the WNPAC in MAM(1). On the other hand, the differences between the resCEP and resCEP_clmWNP runs represent the impacts of the local air–sea interactions. It is estimated that the relative contribution of the local air–sea interactions to the WNPAC in MAM(1) is about 42% (Fig. 12e). The sum of the contributions of the CEP and the WNP is greater than 1, suggesting that SSTAs in other ocean areas, such as the Indian Ocean, may have some negative contributions to the WNPAC.

5. Discussion and conclusions

a. Discussion

ENSO tends to reach its mature phase during boreal winter and decay in the following spring, with the Niño-3.4 index decreasing by about 45% (Figs. 1f,h). However, idealized numerical experiments indicated that the relative contribution of the remote forcing from the equatorial CEP to the WNPAC in El Niño decaying spring is larger than that in El Niño mature winter. It is speculated that the increased relative contribution is associated with stronger responses of low-level circulation anomalies to El Niño–related positive precipitation anomalies over the equatorial CEP. Though the positive precipitation anomalies over the equatorial CEP in El Niño decaying spring are weaker than that in the preceding winter (Figs. 1e,g), the intensity of the northern branch of the twin Rossby wave–like cyclonic anomalies over the central Pacific in MAM(1) is close to that in D(0)JF(1) (Figs. 1e,g). Because the magnitude of the meridional gradient of the climatological specific humidity in MAM is only slightly smaller than that in DJF, the magnitude of the $\neg(\mathbf{u}' \cdot \nabla_h (c_p T + L_v q))$ term in MAM(1) reaches about 80% of that in D(0)JF(1) (Figs. 3 and 10).

The major bias of the resCEP run is that the magnitude of the $\neg(\mathbf{u}' \cdot \nabla_h (c_p T + L_v q))$ ($F_{\text{net}}$) term is larger (smaller) than that in the reanalysis. The bias of the $\neg(\mathbf{u}' \cdot \nabla_h (c_p T + L_v q))$ term is partly caused by larger modeled meridional gradient of climatological specific humidity. It is estimated that the 925-hPa $\sigma, q$ averaged over $1^\circ$–$14^\circ$N, $125^\circ$–$160^\circ$E in the resCEP run is about 10% larger than that in the reanalysis. On the other hand, the wind anomalies term $\mathbf{u}'$ is associated with both the remote forcing from the CEP and WNPAC itself. The WNPAC simulated by the resCEP is stronger than that in the observation. The root cause of the bias in the $\neg(\mathbf{u}' \cdot \nabla_h (c_p T + L_v q))$ remains unknown because of the complicated feedbacks involved.

The overestimation of the $F_{\text{net}}$ term is primarily caused by the underestimate of the longwave cloud radiative forcing anomalies. The magnitude of the $R'_{\text{cloud}}$ in the resCEP is only 50% of that derived from ERA-Interim. It is a long-standing puzzle that models have large biases in reproducing the cloud radiative forcing
(e.g., Potter and Cess 2004). However, it is worth stressing that the $R_0$ cloud is associated with internal feedbacks in local atmospheric columns and thus should not be able to alter the relative contributions of remote and local forcing. The uncertainties in the study associated with the model biases deserve further experiments using additional models.

b. Conclusions

The WNPAC plays a central role in linking the EA–WNP monsoon and El Niño [e.g., reviews by Wang and Li (2004) and Li and Wang (2005)]. However, the WNPAC maintenance mechanisms during El Niño mature winter and the following spring are still controversial (Li et al. 2016). In this study, we tried to resolve this issue through moisture and MSE budget analyses and idealized numerical experiments. The main conclusions are summarized as follows.

In terms of the Gill model, the WNPAC is a Rossby wave–like response to the negative precipitation anomalies over the tropical WNP. The moisture and MSE budget analyses indicated that the negative precipitation anomalies are stimulated by the combined effects of 1) the negative moist enthalpy advection anomalies of the northerly component to the western flank of the northern branch of the twin cyclonic anomalies, which are driven by the enhanced convection over the equatorial CEP (a schematic for the mechanism is shown in Fig. 13) and 2) increased gross moist stability of the dry anomalies in the planetary boundary layer driven by the underlying cold SSTAs.

Then the negative precipitation anomalies and associated WNPAC are further amplified by three positive feedbacks. The first is the convection–cloud radiative forcing feedback; that is, the suppressed convection over the tropical WNP causes a decrease in deep convective cloud and associated cirrostratus and cirrocumulus in situ, which in turn further suppresses the convection through weakening the longwave cloud radiative warming effect. The second is the wind–moist enthalpy advection–convection feedback; that is, the northeast anomalies to the southeastern flank of the WNPAC advect low moist enthalpy air into the tropical WNP and thus suppress convection over there, which further intensifies the WNPAC. The last is the moisture–convection feedback; that is, the downward motion anomalies associated with the suppressed convection reduce the moisture in the lower free troposphere, which increases the gross moist stability and thus further suppresses local convection.

The observational analysis indicated that both the cold SSTAs in the tropical WNP, which are generated by the
local air–sea interactions, and remote forcing from the warm SSTAs in the equatorial CEP contribute to the formation of the WNPAC. To separate their relative contributions, we conducted three idealized pacemaker experiments, with the ocean temperature in the equatorial CEP restored to the observational anomaly plus model climatology. The only difference among the three experiments is that the resCEP run has free air–sea interactions outside the equatorial CEP, while the resCEP_clmGLB (resCEP_clmWNP) run restores ocean temperature outside the equatorial CEP (in the tropical WNP) to the model climatology. The intensity of the D(0)JF(1)- [MAM(1)-] mean WNPAC in the resCEP_clmGLB run reaches about 60% (70%) of that in the resCEP run, indicating that the contribution of the remote forcing is larger than that of the local air–sea interactions. Furthermore, the MSE budget analyses for the numerical experiments confirmed the findings from the observational analysis that the most important remote forcing mechanism is the negative anomalous moist enthalpy advection by teleconnected northerly anomalies.

In Part II of this study (Wu et al. 2017), we explore the formation mechanisms of the WNPAC during the late fall of the El Niño developing phase. A key issue is why the WNPAC forms in the late fall instead of the preceding El Niño–developing summer and early fall, under the condition that the spatial distributions of the SSTAs in the tropical Pacific have not changed significantly since the summer (Fig. 1).

Acknowledgments. This work is jointly supported by the National Natural Science Foundation of China (Grants 41661144009, 41675089, and 41330423), China National Key R&D Program (2017YFA0603802 and 2015CB453201), NSF AGS-1565653, and Jiangsu Collaborative Innovation Center for Climate Change. This is SOEST publication number 10213, IPRC publication number 1282 and ESMC publication number 177.

APPENDIX

Restoring Method used in the resCEP, resCEP_clmGLB, and resCEP_clmWNP Runs

The three sensitivity numerical experiments were conducted based on an ocean data assimilation system. The assimilation system was constructed on the CGCM FGOALS-s2.

The used assimilation scheme is referred to as Ensemble Optimal Interpolation–Incremental Analysis Update (EnOI-IAU) scheme. The assimilated observational records are derived from the EN4.1.1 dataset offered by the Hadley Centre, which collected all available global oceanic temperature and salinity profiles (Good et al. 2013). In the study, an anomaly-field assimilation approach was used to avoid the fact that the model drifts away from its preferred climatology.

The width of the assimilation cycle is 1 month. The EnOI-IAU scheme includes three major steps. The first step is “forecast,” which generates the first guess of the assimilation cycle. The second step is “EnOI,” which calculates analysis increment through combining the first guess and constructed observational data in the window (Oke et al. 2002). Here the constructed observational data are the sums of the model climatology and observational anomalies. The third step is “IAU,” which incorporates the analysis increments in the upper 700 m to the model as small constant forcing terms of the prognostic equations in each integration step (Bloom 1996). The outputs of the IAU were used in the study.

For the resCEP run, the assimilation domain was confined to the equatorial CEP (black dashed triangle in Fig. 8b). Strictly speaking, only analysis increments in the domain were added to the oceanic prognostic equations during the IAU step. The model was integrated freely outside the domain. In this way, the ocean temperature anomalies in the equatorial CEP are restored to the observational anomalies, while in other ocean areas, the ocean temperature anomalies are modulated by free air–sea interactions.

For the resCEP_clmGLB run, the assimilation domain was nearly global ocean (70°N–70°S). In the EnOI step, we constructed observational data through adding observational anomalies in the equatorial CEP (black dashed triangle in Fig. 8d) to the global model.
climatology. In this way, the ocean temperature anomalies in the equatorial CEP are restored to the observational anomalies, while in other ocean areas, the enhancement of the ocean temperature anomalies are suppressed.

For the resCEP_clmWNP run, the assimilation domains were both the equatorial CEP (black dashed triangle in Fig. 8f) and tropical WNP (green dashed quadrilateral in Fig. 8f). The constructed data for the assimilation are the observational anomalies plus model climatology in the CEP and model climatology in the WNP, respectively. Outside the tropical WNP and CEP, the atmospheric and oceanic components are coupled freely.

REFERENCES


