Different Influences of Two El Niño Types on Low-Level Atmospheric Circulation over the Subtropical Western North Pacific

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ABSTRACT

This study focuses on different evolutions of the low-level atmospheric circulations between eastern Pacific (EP) El Niño and central Pacific-II (CP-II) El Niño. The western North Pacific anomalous anticyclone (WNPAC) originates from the northern South China Sea for EP El Niño, and moves to the western North Pacific (WNP) afterward. Compared with EP El Niño, the origin of the WNPAC is farther west during CP-II El Niño, with the center over the Indochina Peninsula. Moreover, the WNPAC shows a weaker eastward shift. Such discrepancies are attributed to different evolutions of the cyclonic response over the WNP, which can suppress the convection in the western flank of the anomalous cyclone. The eastward retreat of the anomalous cyclone is significant for EP El Niño, but less evident for CP-II El Niño. These discrepancies are related to zonal evolutions of the increased precipitation over the equatorial Pacific. Following the southward migration of the intertropical convergence zone (ITCZ), the deep-convection region extends eastward along the equator, reinforcing the atmospheric response to the eastern Pacific warming in EP El Niño. For CP-II El Niño, the atmospheric response is insignificant over the eastern Pacific without warming. Moreover, the meridional migration of the ITCZ can modulate zonal variations of the easterly trade wind and specific humidity as well. Due to the combined effects of the climatological background and atmospheric anomalies, the specific humidity-induced and wind-induced moist enthalpy advection contribute to different shifts of the precipitation center.

1. Introduction

As the most dominant interannual ocean–atmosphere phenomenon in the tropical Pacific, El Niño plays a crucial role in East Asian climate (Webster and Yang

1992; Webster et al. 1998; Chang et al. 2000; Zhou et al. 2007). The western North Pacific anomalous anticyclone (WNPAC) is the important “atmospheric bridge” conveying the El Niño signal to the East Asian monsoon system (Zhang and Sumi 2002; Wu et al. 2003; Li et al. 2017; Zhang et al. 2017). The WNPAC was first discovered when Zhang et al. (1996) analyzed the 1986/87

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and 1991/92 El Niño cases, and was further confirmed by composite analysis (Zhang et al. 1999).

The WNPAC is considered as a Rossby wave response to the anomalous cooling over the western North Pacific (WNP) during El Niño (Zhang et al. 1996). It forms in boreal autumn (all the seasons are for the Northern Hemisphere in this paper) and can persist until the following summer (Wang et al. 2000; Wang and Zhang 2002). Therefore, a positive thermodynamic feedback is needed for its maintenance (Wang et al. 2000). Due to the northeasterly anomalies in the eastern flank of the WNPAC, the strengthened northeasterly trade wind reinforces the evaporation and entrainment, leading to cold sea surface temperature (SST) anomalies in the WNP. This anomalous cooling further excites an anticyclonic Rossby wave response to favor the WNPAC. Therefore, the WNPAC associated with negative SST anomalies in the WNP can be regarded as a self-sustained ocean–atmosphere system (Wang et al. 2000, 2003). Additionally, the role of the cold SST anomalies in the central North Pacific is also important and was validated by numerical experiments (He and Wu 2014; Wu et al. 2014).

The Indian Ocean warming also contributes to the maintenance of the WNPAC (Watanabe and Jin 2002; Yang et al. 2007). Following El Niño, a basinwide warming occurs in the tropical Indian Ocean, called the Indian Ocean basin mode (IOBM) (Yang et al. 2007; Du et al. 2009). The IOBM can stimulate easterly anomalies to the east of the heating source in the form of equatorial Kelvin waves. Due to the surface friction and decrease with latitude, the Kelvin wave can induce low-level divergence and suppressed convection over the WNP, causing an anticyclonic response (Wu et al. 2009; Xie et al. 2009). Alternatively, the IOBM can generate an anomalous zonal overturning circulation, and thus the sinking branch over the South China Sea (SCS) and Philippine Sea works for the WNPAC as well (He and Wu 2014). The IOBM persists from developing winter to decaying summer of El Niño, which can maintain the WNPAC to the following summer, acting as a “discharging capacitor” (Wu et al. 2009; Xie et al. 2009). In fact, only in the decaying summer can the IOBM work (Wu et al. 2009). The local cold SST anomalies are another contributor (Wang et al. 2000; Lau et al. 2004); however, the local contribution gets weakened gradually, while the remote contribution from IOBM intensifies from June to August (Wu et al. 2010). It is the absence of the IOBM that weakens the WNPAC in late summer of 2016, even though 2015/16 is the strongest El Niño event since 2000 (Liu et al. 2018).

The WNPAC does not appear in the developing summer of El Niño, even though the spatial pattern of the tropical Pacific SST anomalies is quite similar to that in peak winter. The moist enthalpy advection (MEA) process is able to address this problem (Wu et al. 2017a,b). Based on the moist static energy (MSE) budget, the negative MEA anomalies in winter can suppress the convection over the WNP, and thus generate the WNPAC. In the developing summer, the lack of negative MEA anomalies is not conducive to the formation of the WNPAC (Wu et al. 2017a,b). Owing to the altered Walker circulation, the equatorial central Pacific (ECP) cooling (warming) can stimulate the anticyclonic (cyclonic) Rossby wave response directly (Lau and Nath 2006; Fan et al. 2013; Wang et al. 2013; Xiang et al. 2013), which often appears in the rapid warm/cold transition period (Chen et al. 2016). Additionally, the warm SST anomalies in the Maritime Continent can also generate the WNPAC via the sinking branch of an anomalous Hadley circulation (Sui et al. 2007; Chung et al. 2011).

There is another viewpoint that the WNPAC is not generated locally. An anomalous anticyclone originates from the northern Indian Ocean (NIO) in autumn, then moves eastward and anchors over the WNP in winter (Chou 2004; Chen et al. 2007; Yuan et al. 2012). The eastward displacement is primarily driven by zonal asymmetry of the atmospheric conditions, such as meridional MEA, divergence, vertical motion, and precipitation (Chou 2004; Chen et al. 2007). Wu et al. (2003) considered that the anomalous anticyclone merely originates from the SCS, instead of the NIO, and its eastward development is caused by the evolution of equatorial heating anomalies.

Recently, a new type of El Niño has been proposed. Compared with conventional El Niño, the warming center is concentrated in the ECP, rather than equatorial eastern Pacific (EEP). It is usually defined as date line El Niño (Larkin and Harrison 2005a,b), El Niño Modoki (Ashok et al. 2007), central Pacific (CP) El Niño (Kao and Yu 2009), or warm pool El Niño (Kug et al. 2009). In this study, the name CP El Niño is used, and the conventional El Niño is referred to as eastern Pacific (EP) El Niño for comparison. Due to their distinct warming origins and patterns, the two types of El Niño can induce different atmospheric responses (Weng et al. 2007, 2009; Feng et al. 2010, 2011, 2018; Feng and Li 2011, 2013; Hu et al. 2012; Yu et al. 2012, 2015; Yuan and Yang 2012; Karori et al. 2013; Paek et al. 2015, 2016; Xu et al. 2017). The WNPAC occurs in the developing autumn of EP El Niño. In contrast, an anomalous cyclone exists over the WNP for CP El Niño (Zhang et al. 2011, 2013).

Based on the opposite rainfall variation in southern China in autumn, CP El Niño can be further divided into two subtypes [called El Niño Modoki I and El Niño Modoki II in Wang and Wang (2013)]. Following Chen et al. (2019), the names CP-I and CP-II El Niño are used in this paper to represent El Niño Modoki I and II. This separation is proven to be necessary (Wang et al. 2019; Chen et al. 2019), and these two subtypes exhibit different evolutions of SST anomalies. The warm SST anomalies originate directly from the ECP during CP-I El Niño, and develop locally afterward. In CP-II El Niño, the warm SST anomalies originate from the subtropical northeastern Pacific, then develop southwestward, and finally reach maximum in the ECP. Although the warming center is also in the ECP, the location is farther west compared to the CP-I type (Wang and Wang 2013). Such different SST anomalies can give rise to distinct climate responses (Wang and Wang 2014; Tan et al. 2016). In fact, only during the developing fall of CP-II El Niño does the anomalous cyclone appear. For CP-I El Niño, the response is similar to that in EP El Niño, indicating an anomalous anticyclone (WNPAC), but with a weaker magnitude (Wang and Wang 2013). According to such different anticyclonic/cyclonic responses in the developing autumn, a CP-II El Niño index [i.e., El Niño Modoki II index in Wang et al. (2018)] is defined as the first principal component of multivariate empirical orthogonal function of three variables (EMI index, Niño-4 index, and 850-hPa relative vorticity anomalies over the Philippine Sea) averaged in boreal autumn [September–November (SON)].

This paper attempts to investigate the evolutions and mechanisms of the anomalous anticyclone/cyclone during the developing phase of EP and CP-II El Niño. For the CP-I type, the WNPAC is found to be contributed by the precursory IOBM, and partly canceled by the direct (one step) cyclonic response to the ECP warming in the developing phase, resulting in a weaker intensity, and thus CP-I El Niño is considered as “the least influential type” (Chen et al. 2019). Given this, only EP El Niño and CP-II El Niño are analyzed in the following.

The rest of this paper is categorized as follows. The datasets and methods are introduced in section 2. The different spatial–temporal evolutions of the low-level atmospheric responses are described in section 3, followed by the associated mechanisms of the formation and eastward displacement of the WNPAC in section 4. The eastward shift of the WNPAC is related to the precipitation anomalies over the equatorial Pacific. Therefore, the evolutions and eastward-evolving mechanisms of the precipitation anomalies are further presented in

sections 5 and 6, respectively. Finally, the conclusions and discussion are summarized in section 7.

2. Datasets and methods

a. Datasets

In this study, all the monthly atmospheric variables (such as wind, precipitation, evaporation, heat flux, geopotential height, air temperature, specific humidity, etc.) are provided by the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis of the twentieth century (ERA-20C) spanning from 1900 to 2010 (Poli et al. 2016). ERA-20C has several sets of horizontal resolution, and $1.5^\circ \times 1.5^\circ$ is selected in this study. In the vertical direction, there are 37 vertical levels ranging from 1000 to 1 hPa. The Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al. 2003) is also used. Besides, the subsurface temperature from the German partner of the consortium for Estimating the Circulation and Climate of the Ocean (GECCO), version 2 (Köhl 2015), is applied to discuss possible influence of the WNPAC on the phase transition from El Niño to La Niña in the last section. In this paper, all the datasets are adopted from 1950 to 2010. Before analyses, the linear trend is removed, and the anomalies are calculated as the departures from the 1971–2000 climatology.

b. Methods

To identify the relative contributions of the physical processes to rainfall variation, a moisture budget is performed from the perspective of moisture balance (Neelin 2007). On an interannual time scale, the moisture balance equation can be approximately expressed as (Wu et al. 2017a)

$$P' \approx E' - \langle \bar{\mathbf{u}} \cdot \nabla_h q' \rangle - \langle \mathbf{u}' \cdot \nabla_h \bar{q} \rangle - \left\langle \bar{\omega} \frac{\partial q'}{\partial p} \right\rangle - \left\langle \omega' \frac{\partial \bar{q}}{\partial p} \right\rangle + \text{NL}. \quad (1)$$

For any variable X , X' corresponds to the anomaly, $\bar{X}$ is the climatological mean, and $\langle X \rangle$ is the mass-weighted vertical integral through the atmospheric column, that is, $\langle X \rangle = \int X dp/g$, where g is gravitational acceleration. In this study, the lower limit is 1000 hPa, and the upper limit is 100 hPa. The terms P , E , q , $\mathbf{u}$, and ω represent precipitation, evaporation, specific humidity, horizontal wind velocity, and vertical pressure velocity, respectively, and NL is the nonlinear term. This equation indicates that the rainfall variation can be mainly balanced by the evaporation (the first term on the right side), horizontal moisture advection (the second and

third terms), and vertical moisture advection (the fourth and fifth terms).

In addition to the moisture budget, the MSE budget is also employed from the viewpoint of energy balance (Neelin 2007). On interannual time scale, the MSE equation can be approximately characterized as (Wu et al. 2017a,b)

$$\begin{aligned} \left\langle \omega' \frac{\partial \bar{h}}{\partial p} \right\rangle &\approx -\langle \bar{\mathbf{u}} \cdot \nabla_h (c_p T + L_v q)' \rangle \\ &\quad - \langle \mathbf{u}' \cdot \nabla_h (c_p T + L_v q) \rangle \\ &\quad - \left\langle \bar{\omega} \frac{\partial h'}{\partial p} \right\rangle + F'_{\text{net}} + \text{NL}, \end{aligned} \quad (2)$$

in which $c_p = 1004.64 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat capacity of dry air at constant pressure, T is air temperature, $L_v = 2.5104 \times 10^6 \text{ J kg}^{-1}$ is the latent heat of condensation, and $c_p T + L_v q$ is the moist enthalpy (ME). The MSE is the sum of ME and geopotential, that is, $h = c_p T + L_v q + \phi$. Moreover, F_{net} is the net heat flux (NHF) entering the atmospheric column, and consists of nine components in total. At the top of the atmosphere, there are three components, downward and upward shortwave radiative flux and upward longwave radiative flux. The other six components are at the surface: downward and upward surface shortwave radiative flux, downward and upward surface longwave radiative flux, latent heat flux, and sensible heat flux (Neelin 2007). For all these components, the positive value implies that the atmospheric column gains heat from outside. In other words, the positive direction is downward at the top, while upward at the surface. Equation (2) can be used to diagnose the contributions to vertical motions of the atmosphere.

In this study, the composite analysis is employed to investigate the spatial-temporal variation of each variable. In the research period (1950–2010), there are six EP El Niño events (1951/52, 1965/66, 1972/73, 1976/77, 1982/83, and 1997/98) and six CP-II El Niño events (1968/69, 1979/80, 1991/92, 1992/93, 2004/05, and 2009/10) (Wang and Wang 2013). For the confidence of composite analysis, the Student's t test is performed with 90% confidence level.

3. Different evolutions of the low-level atmospheric circulation

According to Wang et al. (2000), the positive feedback between WNPAC and underlying cold SST anomalies in the WNP forms a self-sustained air–sea coupled process, and thus the WNPAC is relatively steady. In contrast, the WNPAC is replaced by an anomalous cyclone in the developing fall of CP-II El Niño, despite the cold SST

anomalies in the WNP [Fig. 1 in Tan et al. (2016)]. This is different from the aforementioned theory (Wang et al. 2000), implying that the anomalous cyclone over the WNP may be not a steady state. Therefore, it is necessary to investigate the spatial-temporal evolution of the WNPAC.

During EP El Niño, an anomalous cyclone is first observed over the WNP and SCS in August (0) (Fig. 1a); here (0) and (1) indicate the developing and decaying year, respectively. In the Southern Hemisphere, there is a counterpart symmetric with respect to the equator. Such cyclonic pair reflects the typical Gill-type Rossby wave response to diabatic heating over the equatorial Pacific (Gill 1980). The associated positive precipitation anomalies (indicating the diabatic heating anomalies) are over the ECP, with the center around the date line. In September (0) (Fig. 1b), the SCS and WNP are still influenced by the anomalous cyclone. Over the NIO, an anomalous anticyclone appears with two centers near the northern Arabian Sea (NAS) and Bay of Bengal (BOB). This is related to reduced meridional MEA over South Asia (Chou 2004) and sinking branch of the anomalous Walker circulation (Wang and Zhang 2002; Yuan et al. 2012). In response to the eastward shift of the increased precipitation over the ECP, the anomalous cyclone retreats to the east of the Philippines in October (0) (Fig. 1c). The SCS is beyond the influence of the cyclonic anomalies. Instead, an anomalous anticyclone emerges, with the center over the northern SCS. From November (0) to January (1) (Figs. 1d–f), the cyclonic anomalies keep retreating, and even move to the east of 150°E . Accordingly, the anticyclonic center over the SCS moves eastward with strengthened intensity, and finally reaches the Philippine Sea. To its west, the anticyclonic center over the NAS disappears, and the center over the BOB gets weakened locally, showing no zonal movement. Therefore, we are only concerned with the easternmost anticyclonic center in this paper, referred to as WNPAC. The WNPAC originates from the SCS, rather than the NIO, which is consistent with Wu et al. (2003). It should be noted that the waning of the western two centers as well as the strengthening of the eastern center is apt to induce the illusion that the anomalous anticyclone shifts from the NIO to the WNP.

During CP-II El Niño, the warming center is confined in the ECP, rather than the EEP, and thus the cyclonic Gill-type response and positive precipitation anomalies are located farther west in August (0) (Fig. 1g). The atmospheric response is similar in September (0) except over the eastern Indian Ocean, where a weak cyclonic pair appears (Fig. 1h). This is in accordance with Wang and Wang (2014), and may favor the onset of negative Indian Ocean dipole. In October (0) (Fig. 1i),

FIG. 1. Composites 850-hPa streamfunction (contours; $10^6 \text{ m}^2 \text{ s}^{-1}$) and precipitation anomalies (shading; mm day^{-1}) from August (0) to January (1) during (a)–(f) EP and (g)–(l) CP-II El Niño. The zero contours are thickened, and the solid (dashed) contours denote positive (negative) streamfunction anomalies with the interval of $0.5 \times 10^6 \text{ m}^2 \text{ s}^{-1}$ in EP El Niño and $0.4 \times 10^6 \text{ m}^2 \text{ s}^{-1}$ in CP-II El Niño. For precipitation anomalies, only the composites exceeding the 90% confidence level are shaded.

the cyclonic response begins to withdraw eastward, but does not move out of the SCS completely due to its weaker retreat. Correspondingly, an anticyclonic center emerges over the western Indochina Peninsula, which is farther west compared with EP El Niño (Figs. 1c,i). Moreover, the evolution of the cyclonic anomalies is quite different. The cyclonic anomalies retreat eastward significantly during EP El Niño, and thus the Philippine Sea is already under the control of the WNPAC in October (0) and November (0) (Figs. 1c,d). For CP-II El Niño, the Philippine Sea is still governed by the cyclonic anomalies owing to its weaker retreat (Figs. 1i,j). This can explain why an anomalous cyclone (rather than anticyclone) emerges over the WNP in fall of CP-II El

Niño (Wang and Wang 2013). The essence of this anomalous cyclone is the Gill-type response to the ECP warming (Fan et al. 2013). In fact, it can be also found in EP El Niño, but the rapid retreat allows the WNPAC to move to the Philippine Sea early (Figs. 1d–f). For CP-II El Niño, only from December (0) onward does the WNPAC evolve eastward. Nevertheless, the center still resides over the SCS (Xu et al. 2019), due to the weaker retreat of the cyclonic anomalies (Figs. 1k,l). Accordingly, the influence of the WNPAC on the Philippine Sea is much smaller.

Considering the uncertainties of various reanalysis datasets associated with El Niño (Kumar and Hu 2012), the other two datasets [the Twentieth Century Reanalysis

FIG. 2. Vertical moisture advection induced by anomalous descending/ascending motion ($-\langle \omega' \partial_p \bar{q} \rangle$; mm day^{-1}) from August (0) to January (1) during (a)–(f) EP and (g)–(l) CP-II El Niño. The composites exceeding the 90% confidence level are covered by white dots.

(20CR) (Compo et al. 2011) from National Oceanic and Atmospheric Administration (NOAA) Earth System Research Laboratory (ESRL), and the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996)] are also used for validation. The evolutions of 850-hPa streamfunction and precipitation anomalies derived from 20CR and NCEP–NCAR (figures not shown) agree well with Fig. 1, further confirming the robustness of the results, which are independent of datasets.

4. Formation/evolving mechanisms of the WNPAC

Section 3 reveals the nature of the cyclonic anomalies over the WNP in the developing fall of CP-II El Niño. The essence of the anomalous cyclone is actually a Gill-type

response to the ECP warming. This verifies the initial hypothesis that the anomalous cyclone over the WNP is not a steady state indeed. To its west, the anticyclonic center over the Indochina Peninsula moves eastward since December (0) and can reach the Philippine Sea although with a smaller coverage. Therefore, the emphasis in this section is not the anomalous cyclone, but the easternmost anticyclonic center (WNPAC).

Wu et al. (2017a,b) revealed that the negative MEA anomalies play a dominant role in the formation of the WNPAC during El Niño, but these researches did not distinguish El Niño types. On this basis, following the same method (moisture budget and MSE budget), the differences between EP and CP-II El Niño are compared.

During both EP El Niño and CP-II El Niño, the WNPAC is related to the negative precipitation

FIG. 3. MSE budget terms (W m^{-2}) averaged within (a) 5° – 20°N , 130° – 160°E during OND(0) of EP El Niño and (b) 5° – 20°N , 110° – 140°E during OND(0) of CP-II El Niño.

anomalies to its east (Fig. 1); that is, the anomalous cooling associated with decreased rainfall over the SCS and WNP can excite the anticyclonic Rossby wave response to its west. Based on the moisture budget, the decreased precipitation is mostly contributed by the negative vertical moisture advection induced by anomalous descent ($-\langle\omega'\partial_p\bar{q}\rangle$) (Fig. 2), while the other physical processes in Eq. (1) have less contribution, and thus are neglected in this paper. The anomalous descending motion can reduce the MSE export out of the atmospheric column (Back and Bretherton 2009). For quantitative comparison, two boxes are selected. The box is within 5° – 20°N , 130° – 160°E during OND(0) (October–December) of EP El Niño and 5° – 20°N , 110° – 140°E during OND(0) of CP-II El Niño. Restricted by the MSE budget balance in the tropical deep-convection area, the negative anomalies of the vertical wind-related MSE advection [$\langle\omega'\partial_p\bar{h}\rangle$, which is moved to the left-hand side of Eq. (2) for better comparison] are largely associated with the decreased NHF (F'_{net}) and

wind-induced horizontal MEA [$-\langle\mathbf{u}' \cdot \nabla_h(\bar{c}_p T + L_v q)\rangle$] in both types of El Niño (Fig. 3). The MSE-related vertical advection ($-\langle\bar{\omega}\partial h'/\partial p\rangle$) plays a third role. This is in accordance with Wu et al. (2017a), although they did not distinguish El Niño types.

a. Contribution of the NHF

The NHF is one of the major contributors to the anomalous sinking over the SCS and WNP during EP and CP-II El Niño (Fig. 3). Based on further decomposition, the NHF anomalies are primarily contributed by the cloud-related longwave radiative flux (Fig. 4), which coincides well with Fig. 2, especially for the negative anomalies. This is in agreement with Wu et al. (2017a), reflecting an internal positive feedback between cloud forcing and convection (Su and Neelin 2002; Neelin and Su 2005): the suppressed convection reduces the cloudiness, then the decreased cloud-related longwave radiative flux inhibits the warming of the atmosphere to enhance the gross moist stability, and

FIG. 4. As in Fig. 2, but for cloud-related longwave radiative flux anomalies (W m^{-2}).

thus further suppresses the deep convection in turn. Therefore, this is important to the maintenance of the WNPAC.

b. Contribution of the wind-induced MEA

The wind-induced horizontal MEA $[-\langle \mathbf{u}' \cdot \nabla_h (c_p T + L_v q) \rangle]$ is associated with wind anomalies $\mathbf{u}'$, climatological specific humidity $\bar{q}$, and air temperature $\bar{T}$. Due to the weak gradient of air temperature in the tropics, the contribution of air temperature $\bar{T}$ is almost confined to the north of 20°N . Therefore, the anomalous descent over the northern SCS and WNP in October (0) is mostly related to the horizontal advection of latent energy $(-\langle \mathbf{u}' \cdot \nabla_h L_v \bar{q} \rangle)$ (Figs. 5c,i). Since specific humidity is mostly concentrated at low levels, the 850-hPa level is selected for the analysis of specific humidity and wind (Fig. 6). Following the southward migration of the intertropical convergence zone (ITCZ) after summer, the dry air develops

equatorward, leading to negative meridional gradient (wet in the south, dry in the north) of the climatological specific humidity around 25°N in October (0) (Figs. 6c,i). In the western flank of the cyclonic response, the northerly anomalies can transport the dry air southward to reduce the horizontal MEA over the northeastern SCS and Philippine Sea during EP El Niño (Fig. 5c). For CP-II El Niño, the northerly anomalies can even reach southern China (Fig. 6i). Consequently, the negative MEA anomalies are located farther west and are more vigorous (Fig. 5i), due to the strong difference of specific humidity between land and sea. Such negative MEA anomalies can suppress the local convection, and inhibit the precipitation in October (0). This process explains why the WNPAC originates from the northern SCS for EP El Niño (Fig. 1c), while it originates from the Indochina Peninsula for CP-II El Niño (Fig. 1i).

FIG. 5. As in Fig. 2, but for wind-induced horizontal advection of latent energy ($-\langle \mathbf{u}' \cdot \nabla_h L_v \bar{q} \rangle$; W m^{-2}).

After October (0), the cyclonic response and associated northerly anomalies in its western flank withdraw eastward during EP El Niño (Figs. 6d–f). Correspondingly, the northerly-induced negative MEA anomalies retreat from the SCS and move to the Philippine Sea (Figs. 5d–f). Meanwhile, influenced by the southwesterly anomalies, the MEA increases in the northwest of the WNPAC, forming a southeast–northwest dipole associated with the negative MEA anomalies. The positive MEA anomalies also move eastward following the WNPAC. Compared with EP El Niño, the eastward displacement is much weaker during CP-II El Niño. Only from December (0) do the negative MEA and cyclonic anomalies show a slight eastward movement (Figs. 5k,l, 6k,l). Consequently, the northerly-induced MEA has a large contribution to the formation and different eastward evolutions of the WNPAC.

In addition, the WNPAC also exhibits a weak southward movement in both EP and CP-II El Niño (Fig. 1).

In reality, this was first found by Wang et al. (2000), but has not drawn enough attention. In this study, it can be well explained by the northerly-induced MEA. In response to the seasonal migration of the ITCZ, the maximum negative meridional gradient of the specific humidity keeps shifting equatorward after October (0) (Fig. 6), leading to the southward movement of the negative MEA anomalies (Fig. 5) and WNPAC (Fig. 1).

5. Different evolutions of the precipitation anomalies over the equatorial Pacific

As mentioned in section 4, the eastward development of the WNPAC is primarily associated with the northerly-induced MEA in response to the retreat of the cyclonic anomalies. Actually, the retreat is further attributed to eastward shift of the positive precipitation (diabatic heating) anomalies and meridional gradient of the relative vorticity (Wu et al. 2017b). The latter can inhibit

FIG. 6. Composites 850-hPa wind anomalies (vectors; m s^{-1}) from August (0) to January (1) during (a)–(f) EP and (g)–(l) CP-II El Niño. For better comparison, climatological 850-hPa specific humidity (shading; g kg^{-1}) is underlaid in both El Niño types.

westward extension of the Rossby wave when its meridional gradient is negative (Hoskins and Ambrizzi 1993). But it is related to the climatological background, and thus has almost the same effects during two types of El Niño. Therefore, the discrepancies of the cyclonic anomalies are mainly owing to the former process (i.e., different evolutions of the positive precipitation anomalies over the equatorial Pacific).

The Hovmöller diagram of the precipitation anomalies along the equatorial Pacific (averaged between 5°S and 5°N) is presented in Fig. 7. During EP El Niño, the positive rainfall anomalies develop eastward significantly, together with the intensified magnitude (Fig. 7a). The center locates around 160°E in September (0) and moves to 150°W in January (1). To the east of the center, the

eastward extension of the contours is also evident, and the positive rainfall anomalies can reach the EEP. For CP-II El Niño, the zonal displacement is weaker (Fig. 7b). The rainfall center stays around 150°E before October (0). Since November (0), the center begins to move eastward slightly but is always limited to the west of the date line. Moreover, the coverage of the positive rainfall anomalies is much smaller, mostly confined between 150°E and 150°W, which is different from EP El Niño.

Such different evolutions are related to the warming location and changing deep-convection region. Forced by same SST anomalies, the atmospheric response is more vigorous in the deep-convection region than that in descending region, due to nonlinear modulation of the background precipitation (Hong et al. 2008a,b;

FIG. 7. Hovmöller diagram of composited precipitation anomalies (mm day^{-1}) along the equatorial Pacific (averaged between 5°S – 5°N) from August (0) to July (1) during (a) EP and (b) CP-II El Niño. The composites exceeding the 90% confidence level are covered by white dots.

Xiang et al. 2011). During August (0)–October (0), the climatological deep convection is limited to the west of 170°W (Fig. 8a). Accordingly, the maximum precipitation anomalies are around the date line in EP El Niño (Fig. 7a), even if the EEP is warmer (Fig. 8a). As the ITCZ moves equatorward seasonally, the deep-convection region expands to the east (Fig. 8a), and thus the EEP warming begins to work, giving rise to eastward development of the atmospheric response (positive precipitation anomalies) after October (0) (Fig. 7a). For CP-II El Niño, although the deep-convection region also stretches eastward, the warming is confined in the ECP (Fig. 8b). Therefore, the atmospheric response is quite weak over the EEP, which is unfavorable for eastward extension of the positive precipitation anomalies (Fig. 7b). In the next section, the moisture budget and MSE budget are performed again to further diagnose the contributors to different precipitation evolutions along the equatorial Pacific.

6. Formation/evolving mechanisms of the precipitation anomalies over the equatorial Pacific

Based on the moisture budget analysis, the precipitation anomalies over the equatorial Pacific are mainly provided

by vertical moisture advection ($-\langle \omega' \partial_p \bar{q} \rangle$), which is related to the anomalous vertical motion in both EP and CP-II El Niño (Figs. 9a,b). The ascending (descending) anomalies are in favor of MSE export (import) (Back and Bretherton 2009) to decrease (increase) the vertical MSE advection ($-\langle \omega' \partial_p \bar{h} \rangle$) (Figs. 9c,d). This evolving pattern is similar to the precipitation (Fig. 7) and 500-hPa vertical velocity anomalies (Figs. 9a,b), especially for the center in the deep-convection region. Therefore, the rest mainly focuses on the center shift from the perspective of MSE.

For quantitative comparison, four boxes are selected along the equator. For EP (CP-II) El Niño, the western box is within 5°S – 5°N , 150°E – 180° (145° – 165°E), and the eastern box is within 5°S – 5°N , 180° – 150°W (165°E – 175°W). During EP El Niño, the vertical wind-related MSE advection [$\langle \omega' \partial_p \bar{h} \rangle$, which is moved to the left-hand side of Eq. (2) for better comparison] averaged in the western box decreases rapidly from September (0) to November (0) (Fig. 10a). Meanwhile, the eastern-box average increases (Fig. 10b), indicating an evident shift from west to east. In comparison with EP El Niño, the selected eastern box is farther west during CP-II El Niño. Besides, the decrease of the western-box average is much weaker and occurs one month later (Fig. 10c), indicating a weak shift. To keep the MSE budget balance, both the magnitude and evolution of the vertical

FIG. 8. Hovmöller diagram of composited SST anomalies (contours; $^{\circ}\text{C}$) along the equatorial Pacific (averaged between 5°S and 5°N) from August (0) to July (1) during (a) EP and (b) CP-II El Niño, underlaid by climatological 500-hPa vertical velocity (shading; $10^{-2} \text{ Pa s}^{-1}$) in both El Niño types. The zero contours are thickened, and the solid (dashed) contours denote positive (negative) SST anomalies with the interval of 0.3°C . The pink/thickened line is the zero contour of vertical velocity, indicating the boundary of the deep-convection region in this paper.

wind-related MSE advection are mainly associated with the horizontal MEA induced by ME anomalies $[-\langle \bar{\mathbf{u}} \cdot \nabla_h (c_p T + L_v q) \rangle]$, horizontal MEA induced by wind anomalies $[-\langle \mathbf{u}' \cdot \nabla_h (c_p T + L_v q) \rangle]$, and NHF (F'_{net}) anomalies (Fig. 10). The MSE-related vertical advection $(-\langle \bar{\omega} \partial h' / \partial p \rangle)$ is always around 0, which has less contribution.

a. Contribution of the ME-induced MEA

The ME-induced horizontal MEA is mainly contributed by the zonal advection of latent energy (ZALE) related to specific humidity anomalies $(-\langle \bar{u} \partial_x L_v q \rangle)$ (Figs. 11a,b), while the meridional component and air temperature have less contribution. The eastward shift of the positive ZALE center is evident (weak) in EP (CP-II) El Niño (Figs. 11a,b). This is roughly similar to the vertical MSE advection (Figs. 9c,d), indicating that the specific humidity-induced ZALE (MEA) has a large contribution to the vertical motion anomalies over the equatorial Pacific.

Forced by the anomalous warming during EP El Niño, the specific humidity increases over the equatorial central-

eastern Pacific, with the maximum center around 150°W (Fig. 11c). Such anomalous moisture is advected by the climatological easterly trade wind, increasing the ZALE to the west of the date line before early autumn (Fig. 11a). Following the southward migration of the ITCZ, the westerly appears after October (0) (Fig. 11c), reducing the ZALE over the equatorial western Pacific (EWP) (Fig. 11a). Correspondingly, the easterly trade wind begins to retreat (Fig. 11c), leading to eastward shift of the ZALE center (Fig. 11a).

During CP-II El Niño, the positive specific humidity anomalies locate farther west due to the more concentrated warming in the ECP. The center is around 160°E from August (0) to October (0) (Fig. 11d). Accordingly, the significantly positive ZALE anomalies can be only found to the west of 150°E before October (0) (Fig. 11b). In January (1), the center of specific humidity anomalies moves to the date line (Fig. 11d). Therefore, although the climatological easterly wind also retreats eastward after October (0), the significantly positive ZALE anomalies could only exist to the west of 180° , and the eastward shift is less evident (Fig. 11b).

FIG. 9. As in Fig. 7, but for (top) 500-hPa vertical pressure velocity anomalies ($10^{-2} \text{ Pa s}^{-1}$) and (bottom) vertical MSE advection induced by anomalous descending/ascending motion ($-\langle \omega' \partial_p \bar{h} \rangle$; W m^{-2}).

b. Contribution of the wind-induced MEA

Similar to the ME-induced MEA, the wind-induced horizontal MEA is primarily attributed to zonal wind and specific humidity [i.e., the ZALE associated with zonal wind anomalies ($-\langle u' \partial_x L_v \bar{q} \rangle$)] (Figs. 12a,b). In spite of different eastward evolutions as well, this term is only half of the specific humidity-induced ZALE

(see the color bar in Figs. 11a,b and 12a,b). Hence, the wind-induced ZALE plays a minor role in the vertical motion (precipitation) anomalies over the equatorial Pacific.

During EP El Niño, the anomalous westerly center first appears over the EWP, then extends eastward (Fig. 12c). The westerly anomalies can transport the climatological wet air over the EWP to the east,

FIG. 10. MSE budget terms (W m^{-2}) averaged within (a) 5°S – 5°N , 150°E – 180° and (b) 5°S – 5°N , 180° – 150°W for EP El Niño, and within (c) 5°S – 5°N , 145° – 165°E and (d) 5°S – 5°N , 165°E – 175°W for CP-II El Niño.

increasing the ZALE over the ECP (Fig. 12a). Meanwhile, the maximum zonal gradient of the climatological specific humidity moves westward before early autumn, canceling out the eastward development of the westerly anomalies (Fig. 12c). Therefore, the positive ZALE center does not show a zonal shift from August (0) to October (0) (Fig. 12a). In October (0), the anomalous westerly center moves to the vicinity of the date line, in coincidence with the maximum zonal gradient of the specific humidity. After October (0), in response to the equatorward migration of the ITCZ, the maximum zonal gradient of the specific humidity also displaces eastward associated with the westerly center (Fig. 12c). Due to the combined effects, the maximum ZALE moves to the east of the date line in January (1) (Fig. 12a).

In response to the ECP warming during CP-II El Niño, weak easterly anomalies appear over the EEP, which are unfavorable for the eastward development of the westerly anomalies (Fig. 12d), leading to spatial mismatch between the westerly center and maximum zonal gradient of the specific humidity (Fig. 12d). Therefore, the eastward shift of the positive ZALE anomalies is less significant (Fig. 12b). Due to the spatial mismatch, there are actually two centers in February (1), and both of them are located to the west of the date line (Fig. 12b).

c. Contribution of the NHF

Different from the former two processes, the eastward shift of positive NHF anomalies is found in neither EP

El Niño nor CP-II El Niño (Figs. 13a,b). Nevertheless, the contribution is nonnegligible, especially in the ECP during the peak phase (winter). Such evolutions are mainly attributed to the latent heat flux (figure not shown). Besides, the cloud-related longwave radiative flux is the second largest component (Figs. 13c,d).

Although the total NHF anomalies do not show any eastward movement (Figs. 13a,b), the cloud-related longwave component is perfectly consistent with the vertical motion (Figs. 9a,b) and precipitation anomalies (Figs. 7a,b); that is, the eastward shift of the positive anomalies is more (less) significant in EP (CP-II) El Niño (Figs. 13c,d). This evolving consensus further confirms the positive feedback between convection and cloud forcing (section 4a) and its important role in maintaining the vertical motions (Su and Neelin 2002; Neelin and Su 2005; Wu et al. 2017a).

7. Summary and discussion

a. Summary

Forced by different SST anomalies, the atmospheric responses are significantly distinct over the WNP. During the developing autumn of EP and CP-I El Niño, an anomalous anticyclone occurs over the Philippine Sea. In CP-II El Niño, however, an anomalous cyclone, rather than anticyclone, is located near the Philippine Sea (Wang and Wang 2013). In this study, such an anomalous cyclone in the developing fall of CP-II El Niño is essentially a Gill-type response to the ECP warming.

FIG. 11. (top) As in Fig. 7, but for specific humidity-induced ZALE ($-\langle \bar{u} \partial_x L_v q' \rangle$; W m^{-2}). (bottom) As in Fig. 8, but for 850-hPa specific humidity anomalies (shading; g kg^{-1}) and climatological zonal wind (contours; m s^{-1}) from June (0) to July (1). The contour interval is 1.5 m s^{-1} .

The WNPAC originates from the northern SCS during developing phase of EP El Niño. To the west, there are two additional anticyclonic centers. One is located over the NAS and the other is over the BOB. The western two centers do not show any zonal movement; only the easternmost center over the SCS can develop eastward and reach the WNP, and thus is called WNPAC in this paper. This is slightly

different from the previous results that the WNPAC stems from the NIO (Chou 2004; Chen et al. 2007; Yuan et al. 2012). In fact, the weakening of the western two centers as well as the strengthening of the easternmost center is apt to induce the illusion that the anomalous anticyclone shifts from the NIO to WNP. During CP-II El Niño, the origin of the WNPAC is farther west, with the center located in the Indochina

FIG. 12. (top) As in Fig. 7, but for wind-induced ZALE ($-u' \partial_x L_v \bar{q}$; W m^{-2}). (bottom) As in Fig. 8, but for 850-hPa zonal wind anomalies (contours; m s^{-1}) and climatological specific humidity (shading; g kg^{-1}) from June (0) to July (1). The contour interval is 1 m s^{-1} . The thick pink line is the contour of 10 g kg^{-1} , indicating the maximum zonal gradient of specific humidity in this paper.

Peninsula. Compared with EP El Niño, the eastward displacement of the WNPAC is less significant in CP-II El Niño, with a slower speed and shorter path. Such discrepancies directly induce different atmospheric circulations over the WNP in autumn.

The formation and evolution of the WNPAC are primarily associated with the northerly-induced MEA in

the western flank of the cyclonic response. The northerly anomalies can advect dry air (negative MEA anomalies) from high latitude to suppress the convection, and finally generate the WNPAC. Following significant withdrawal of the cyclonic anomalies in EP El Niño, the negative MEA anomalies shift to the east, resulting in significantly eastward movement of the WNPAC. For CP-II El

FIG. 13. As in Fig. 7, but for (top) NHF anomalies (F'_{net} ; W m^{-2}) and (bottom) cloud-related longwave radiative flux anomalies (W m^{-2}).

Niño, the eastward displacement of the WNPAC is weaker due to the slower/shorter retreat of the cyclonic anomalies.

Such discrepancies of the cyclonic response are related to the increased precipitation over the equatorial Pacific. Although the interannual anomalies are concerned in this study, the modulation of the climatological background is quite important. As the ITCZ

moves southward seasonally, the deep-convection region extends eastward along the equator, reinforcing the atmospheric response to the EEP warming in EP El Niño. For CP-II El Niño, the atmospheric response and positive precipitation anomalies are insignificant without the EEP warming, showing weaker shift. On the other hand, the meridional migration of the ITCZ can regulate zonal variations of

FIG. 14. As in Fig. 7, but for SST anomalies ($^{\circ}\text{C}$) from January (0) to December (1).

the easterly trade wind and specific humidity. Jointly affected by the climatological background and atmospheric anomalies, the ME-induced (or specific humidity-induced) and wind-induced MEA contribute to different evolutions of the precipitation center. Besides the aforementioned processes, the cloud-related longwave radiative flux is also important to the maintenance of the precipitation anomalies, reflecting a positive feedback between convection and cloud forcing.

b. Discussion

In general, a La Niña event often follows the decaying of EP El Niño, forming a warm–cold cycle (Fig. 14a). However, no cold events could be found after CP-II El Niño (Fig. 14b) (Kug et al. 2009, 2010). Actually, there is not an absence of negative SST anomalies in the EEP during decaying phase of CP-II El Niño; they are merely insignificant, with less than 90% significance (Fig. 14b). This implies that there may exist the same phase-transition mechanism for CP-II El Niño, but the intensity is too weak to generate significant cooling in the EEP.

The western Pacific oscillator theory highlights the important role of the WNPAC in the phase transition of El Niño (Weisberg and Wang 1997; Wang et al. 1999). In the southern flank of the WNPAC, the easterly

anomalies can initiate a cold/upwelling Kelvin wave to raise the thermocline in the EWP in October (0) (Fig. 15a). The cold Kelvin wave then propagates eastward along the equator. When it reaches the EEP, it cools the sea surface significantly due to the shallower thermocline, providing precondition for the growth of La Niña (Figs. 15a–e). For CP-II El Niño, the WNPAC is still over the Indochina Peninsula and SCS in October (0) (Fig. 1i) and thus cannot influence the EWP temporarily (Fig. 15f). The cold signal emerges in November (0) (Fig. 15g), one month later than that in EP El Niño. After November (0), the cold signal also propagates eastward in the form of equatorial Kelvin wave. However, the strength is weaker, and thus is unable to influence the sea surface (Figs. 15h–j). This powerless signal may be related to the insignificantly eastward shift of the WNPAC. Accordingly, the influence of the easterly anomalies on the EWP is quite small. The details deserve further analysis in the future.

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FIG. 15. Tendencies of subsurface temperature anomalies ($^{\circ}\text{C month}^{-1}$) in a vertical section along the equatorial Pacific (averaged between 5°S – 5°N) from October (0) to May (1) during (a)–(e) EP and (f)–(j) CP-II El Niño. The tendency is calculated as the difference between this month and last month, and the white dots denote the tendency exceeding the 90% confidence level. The green solid line is the climatological 20°C isotherm, indicating the thermocline depth in this paper.

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