The Zonal Oscillation and the Driving Mechanisms of the Extreme Western North Pacific Subtropical High and Its Impacts on East Asian Summer Precipitation

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ABSTRACT

This paper scrutinizes the zonal oscillation of the western North Pacific subtropical high (WNPSH) via diagnosing its two extreme phases, which are defined by the top 10% strongest (positive phase) and the weakest (negative phase) WNPSH index (WNPSHI) days during summers in 1979–2016. Key findings include the following: a tripole pattern consisting of intensified (weakened) precipitation over the Maritime Continent and the East Asian summer monsoon regions, and suppressed (strengthened) precipitation over the western North Pacific summer monsoon region during positive (negative) WNPSH phases; a westward movement of WNPSH-induced precipitation anomalies that subsequently affects eastern China, Japan, and the Korean Peninsula at different time lags; an OLR–vorticity pattern explained by atmospheric responses to thermal sources is suggested to drive the oscillation; and the competitive interaction of local air–sea feedbacks, especially during the positive phase. In addition, moderate-to-strong positive correlations between the WNPSHI and the Niño-3.4 index are found on 1–2-, 2–3-, and 3–6-yr time scales; both exhibit decadal shifts to a higher-frequency mode, suggesting the intensification of both the zonal WNPSH oscillation and the ENSO under the changing climate and their close interdecadal association. A nonlinear quasi-biennial WNPSH–ENSO relationship is identified: the positive (negative) WNPSH phase sometimes occurs during 1) a decaying El Niño (La Niña) in the preceding summer/autumn, and/or 2) a developing La Niña (El Niño) in the current summer/autumn. A full ENSO transition from moderate-to-strong El Niño to La Niña is often seen during the positive phase, offering potential in predicting ENSO events and extreme WNPSH phases and thereby the summer monsoon rainfall in East Asia.

1. Introduction

The western North Pacific subtropical high (WNPSH) (e.g., Xiang et al. 2013; Mao et al. 2010; Park et al. 2010; Sui et al. 2007; Yun et al. 2015) is a subtropical anticyclonic system in the lower and midtroposphere over the western North Pacific (WNP) stemming from the western flank of the summertime North Pacific subtropical high (NPSH) (Lu 2001; Park et al. 2010; Li et al. 2010; Lu and Dong 2001;

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Yun et al. 2015). It is also known as the western Pacific subtropical high (WPSH) (Ren et al. 2013; Wang et al. 2013; Zhou et al. 2009; Wang et al. 2008). The western extension of the WNPSH has been found to have a great influence on the East Asian (EA) summer monsoon and the regional climate (Lee et al. 2013; Xiang et al. 2013; Ren et al. 2013; Lu and Dong 2001; Lu 2001; Lau and Chan 1986; Chang et al. 2000). Recent studies have pointed out that the western extension of the WNPSH modifies the wind circulation patterns over the WNP and brings warm and moist airflow from the South China Sea (SCS) and the Philippine Sea to interact with the relatively cooler air over lands, and eventually leads to enhanced synoptic-scale rainfall in the Korean Peninsula, Japan, eastern China, and the Maritime Continent (MC) (Ren et al. 2013; Wang et al. 2013; Xiang et al. 2013; Mao et al. 2010). However, most of them focused on the westward extension of the WNPSH only, while the opposite phase (i.e., the eastward retreat of WNPSH) and the corresponding driving mechanisms during the two extreme phases have rarely been explored in the literature. This study attempts to offer a comprehensive diagnosis and discussion on the meteorological influences of the two extreme WNPSH phases, later defined as positive (westward extension) and negative (eastward retreat) phases in this article, on the regional climate system that affects the EA and the MC summertime moisture distribution and precipitation pattern.

The zonal WNPSH oscillation is prominent, ranging from subseasonal to interannual time scales (Ren et al. 2013; Park et al. 2010; Mao et al. 2010; Zhou et al. 2009; Wu and Zhou 2008; Sui et al. 2007). Substantial efforts have been made by other researchers to explore various factors on different time scales that together drive the zonal WNPSH oscillation. For instance, Mao et al. (2010) reported that the intraseasonal (20–50 day) oscillation of the summer monsoon over the Yangtze River basin (YRB) was the atmospheric response to the WNPSH with the same time scale of variation. Similarly, Ren et al. (2013) showed strong lagged correlations between the rainfall periods over the YRB and the zonal WNPSH oscillation, but with a shorter subseasonal (10–30 day) time scale. Furthermore, several studies identified the potential relationship between the interannual time scales of the zonal WNPSH oscillation and some large-scale climatic variabilities. For example, Sui et al. (2007) and Wu and Zhou (2008) stated that the 2–3-yr WNPSH oscillation (quasi-biennial cycle¹) could be associated with the anomalous Hadley circulation due to the positive sea surface temperature anomalies (SSTA) over the

MC. The 3–5-yr cycle of the WNPSH oscillation could be the response to the local negative SSTA and anomalous Walker circulation (Sui et al. 2007). Owing to the interannual variation of SSTA forcing over the equatorial Pacific Ocean, the quasi-biennial oscillation of the WNPSH was suggested to have a lead-lag correlation with El Niño–Southern Oscillation (ENSO) (Wu and Zhou 2008; Sui et al. 2007). But later, Wang et al. (2013) and Xiang et al. (2013) showed that strong WNPSH years did not always follow the decay of the El Niño, leaving the relationship between the WNPSH and ENSO remain ambiguous and unresolved. Li et al. (2010) pointed out that the quasi-biennial WNPSH oscillation could have a selective interaction with the ENSO transition, which may reflect that the relationship between the two synoptic variabilities is not forthright to describe and explain. Therefore, the WNPSH–ENSO relationship on different time scales is discussed based on the full spectrum analysis presented in this paper.

Recent studies suggested that the local air–sea feedbacks, such as the wind–evaporation–SST (WES) feedback and the convection–wind–evaporation–SST (CWES) feedback, are likely the key local air–sea feedbacks in the zonal WNPSH oscillation (Wang et al. 2013; Xiang et al. 2013). In addition to the air–sea interaction, the land–sea thermal contrast was also suggested to be one of the key factors to the variation of the intensity and the position of the WNPSH (He et al. 2001). Besides, Wang et al. (2008) suggested the potential role of the Tibetan Plateau warming in strengthening both the East Asia summer monsoon (EASM) and the WNPSH through the Sverdrup vorticity balance and the Rossby wave trains at both the lower and upper troposphere. Although the impacts of the zonal WNPSH oscillation on some regional rainfall cycles have been investigated in the literature (Ren et al. 2013; Mao et al. 2010), more efforts are still needed to have a comprehensive understanding of the interactions among the zonal WNPSH oscillation, atmospheric circulations, and moisture fluxes. Therefore, we strive to construct the causal framework with the goal of providing a deeper comprehension of the characteristics and possible driving factors of the zonal WNPSH oscillation, and its impacts on the EA summer climate and rainfall predictability. The conceptual framework of this work is illustrated in Fig. 1, together with a schematic diagram of the positive WNPSH phase that includes the key processes and their links investigated and discussed in this study.

In line with the conceptual framework, this diagnostic study attempts to do the following:

- 1) Profile the temporal characteristics of the zonal WNPSH oscillation and clarify its association with the ENSO in the time–frequency space.

¹ A quasi-biennial cycle has a mean period of roughly 2 years; see Angell and Korshover (1964) and Baldwin et al. (2001).

FIG. 1. (top) In the conceptual framework of this study, driving mechanisms of the zonal WNPSH oscillation (solid black arrow) and its nonlinear association with the ENSO (dashed black arrow) are discussed. Key processes during extreme WNPSH phases are listed (dashed box), which result in anomalous East Asian JJA precipitation. Linear quantile regression is adopted to further quantify the relationship between extreme WNPSH phases and regional precipitation (gray arrow). (bottom) A schematic diagram of the positive WNPSH phase. The negative WNPSH phase shares a similar but reverse behavior as the positive one.

- 2) Scrutinize its influences on the atmospheric circulations, moisture transports and rainfall; and investigate the dynamic and thermodynamic processes behind.
- 3) Quantify the relationship between the zonal WNPSH oscillation and regional rainfall.

This paper is organized as follows. Data sources and the definition of the WNPSHI are provided in [section 2](#). [Sections 3–5](#) present the results addressing the three

objectives listed above, respectively. At the end of this paper, key findings are discussed and summarized.

2. Data and the WNPSHI

a. Data

The variables investigated in the study such as geopotential height at 850 hPa (Z850), precipitation (PP), horizontal winds at 850 hPa (uv850), vertically integrated

FIG. 2. The JJA climatology of Z850 (shaded; interval: 25 m) and its daily standard deviation (contours; interval: 5 m) and uv850 (vectors) during 1979–2016. The red box (18° – 26° N, 127° – 148° E) indicates the region where the WNPSHI is defined in this study.

water vapor transport (IVT), sea surface temperature (SST), outgoing longwave radiation (OLR), surface solar radiation (SSR), and vorticity at 850 hPa (Vor850) from the ERA-Interim reanalysis dataset (Dee et al. 2011), with a spatial gridded resolution of $1^{\circ} \times 1^{\circ}$ (available at <http://apps.ecmwf.int/datasets/data/interim-full-daily/>). All of the aforementioned variables are examined over the EA region (10° S– 50° N, 60° E– 180°). The anomalies of these variables are obtained by subtracting the pentad-day (5 day) moving average of the calendar day climatology during the boreal summer [June–August (JJA)] in 1979–2016. The entire daily time series from 1 January 1979 to 31 December 2016 is used for the wavelet analysis presented in section 3. The Niño-3.4 index is the averaged SSTA in the region 5° S– 5° N, 190° – 240° E (Bamston et al. 1997).

b. Definition of the WNPSHI

To quantify the zonal WNPSH oscillation, a common practice is to define an area-averaged index (Lu and Dong 2001; Park et al. 2010; Wang et al. 2013; Lee et al. 2013; Yun et al. 2015; Lu 2001). Although Park et al. (2010) mentioned that a single area-averaged index could mask the spatiotemporal variability of the WNPSH, the WNPSHI defined in this study is the Z850 anomalies averaging over a WNPSH-active region identified by the center of the Z850 peak variability, with the aim of well capturing the WNPSH variability. From the JJA climatology (Fig. 2), the NPSH is always located at the northeastern Pacific approximately centered at 35° N, 150° W (Salby 2012). Along its western ridge extending to the WNP, a climatological low-level anticyclonic circulation pattern over the EA is clearly featured, which encourages substantial moisture transports from the oceanic areas to the EA lands, and thus regulates the EASM system (Wang et al. 2013; Xiang et al. 2013; Park et al. 2010; Rodwell and

Hoskins 2001). Moreover, there is a relatively large standard deviation of the JJA climatological pentad moving average Z850 over the WNP (or the northern Philippine Sea), representing the active region of the spatial variability of the WNPSH (Fig. 2, red box). In this study, the daily WNPSHI is thus defined as the pentad-moving average of the Z850 anomalies over the WNPSH-active region of 18° – 26° N, 127° – 148° E. This WNPSH-active region is largely consistent with the regions defined in the previous studies done by the others (Lee et al. 2013; Wang et al. 2013; Sui et al. 2007; Lu 2001). It is important to note that previous studies used the geopotential height either at 850 hPa (Z850) (e.g., Wang et al. 2013; Park et al. 2010; Lee et al. 2013; Lu and Dong 2001; Lu 2001) or at 500 hPa (Z500) (e.g., Sui et al. 2007; Wu and Zhou 2008; Zhou et al. 2009) as an indicator for the WNPSH. Zhou et al. (2009) argued that Z500 had a close connection to the EA climate because the center of the WNPSH at the midlevel troposphere was closer to the WNP, comparing to that at the low-level troposphere. Park et al. (2010) emphasized the close association between the low-level circulation field and the summer monsoon in the East Asia. We use the Z850 to define and measure the WNPSH as it better reflects the low-level thermodynamic processes. Moreover, Z850 is distinctly associated with the water vapor transport in the EA when comparing the JJA climatology maps of the circulations at 850 and 500 hPa, respectively, with the IVT patterns (see Figs. S1a,b in the online supplemental material). Thus, Z850 suits the aim of this study to diagnose the mechanisms of the zonal WNPSH oscillation and its impacts on summertime EA precipitations. Also, the center of the climatological daily standard deviation in the composite map of Z850 (Fig. 2) appears to be more discernible in contrast with that of Z500 (Fig. S1c), making Z850 a better variable in defining the WNPSH-active region and thus the WNPSHI. In addition, considering that the

FIG. 3. (a) The lag-1 Morlet wavelet power spectrum of the WNPSHI from 1979 to 2016. The black contours indicate the variance at the 95% confidence level. The black dashed line shows the cone of influence due to edge effects. (b) The time-averaged global wavelet spectrum. (c) The 3–6- (red), 2–3- (blue), and 1–2-yr (purple) scale-averaged time series of the variance. The solid line segments in (b) and (c) indicate that the scale-averaged variances are at the 95% confidence level, while the dashed line segments are not.

variability of the lower-level WNPSH could be different from the midlevel WNPSH (Li et al. 2010), we compare the WNPSHI measured by Z850 and Z500 anomalies and find that the two indices are significantly cross-correlated, especially from lag -1 to lag 1 (Fig. S2). This implies that the variabilities of the WNPSH at these two levels are nearly concurrent, and the aforementioned variabilities at different levels of the WNPSH do not affect the consistency between the two indices. The definition of WNPSHI assists the interpretation of the WNPSH oscillation and associated circulation patterns: positive (negative) WNPSHI indicates the anomalous anticyclone (cyclone) mainly due to the westward extension (eastward retreat) of the WNPSH, although activities such as tropical cyclones may potentially modulate the index. The WNPSHI enables quantitative study of the relationship between the zonal WNPSH oscillation and other important meteorological variables or indices with a series of statistical analyses presented below.

3. Temporal characteristics of the zonal WNPSH oscillation and its association with the ENSO in time–frequency space

a. Temporal characteristics of the WNPSHI using wavelet analysis

To understand the influence of the WNPSH on the EASM system on different time scales, a lag-1 Morlet

wavelet analysis is employed to profile the variation modes of the WNPSHI (Torrence and Compo 1998). (Wavelet software was provided by Torrence and Compo, available at <http://paos.colorado.edu/research/wavelets/>.) The resulting wavelet power spectrum and global wavelet spectrum (Figs. 3a,b) show that the statistically significant modes of the WNPSHI range from weeks to years, with the consideration of the cone of influence due to edge effects. Although different individual modes of the zonal WNPSH oscillation, ranging from subseasonal to interdecadal time scales, have been explored by other research groups (Ren et al. 2013; Park et al. 2010; Mao et al. 2010; Zhou et al. 2009; Wu and Zhou 2008; Lu 2001), a complete analysis on the whole spectrum of time scales was rarely discussed in the literatures. The wavelet analysis presented here thus assists a comprehensive discussion over the temporal variations of the WNPSH and the detection of any changes over the data period. Our result not only reveals prominent modes ranging from subseasonal to interannual time scales, but also shows a shortening in the dominant time scale over the data period (Figs. 3a,c): a remarkably strong 3–6-yr variability of the WNPSH is found during the 1980s; later, the leading mode shifts to 2–3-yr in the mid-1990s, and to an even shorter time scale (i.e., 1–2 yr) in the late 2000s. These findings generally agree with the results by Sui et al. (2007) and extend to show that the

FIG. 4. (a) The lag-1 Morlet wavelet power spectrum of the Niño-3.4 index from 1979 to 2016. Black contours indicate the variance at the 95% confidence level. The black dashed line indicates the cone of influence due to edge effects. (b) The time-averaged global wavelet spectrum. The Spearman rank correlation of WNPSHI and Niño-3.4 index in (c) 3–6-, (d) 2–3-, and (e) 1–2-yr scale-averaged time series. The solid line segments from (b) to (e) indicate the scale-averaged variances are at the 95% confidence level.

prominent interannual variation has its own decadal shifting trend to an even shorter time scale of 1–2 years based on the updated data period (1979–2016). Sui et al. (2007) argued that the 2–3-yr zonal WNPSH oscillation is the response to the anomalous Hadley circulation on the same time scale, by examining the anomalous vertical velocity and OLR, while the 3–5-yr zonal WNPSH oscillation is associated with the ENSO phenomenon and anomalous Walker circulation across the equatorial Indo-Pacific Ocean. In addition, the significant subseasonal mode of the zonal WNPSH oscillation is found in the wavelet spectrum (Fig. 3b). This might be linked to the summer monsoon cycles in the YRB that are associated with the variation of SSTA over the WNP (Ren et al. 2013). Investigations of extreme WNPSH phases on the subseasonal time scale are presented in section 4. Given our preliminary results and others' previous findings, we speculate that the decadal shifting of the leading mode of the zonal WNPSH oscillation might be associated with low-frequency climatic variabilities, such as the ENSO, and potentially modify the characteristics of the associated atmospheric circulations in the summer monsoon regions in the EA. This speculation leads to our following diagnosis and discussion.

b. Relationship between ENSO and zonal WNPSH oscillation in time–frequency space

To investigate the possible association between the decadal shifting of the WNPSH leading mode and the low-frequency global climatic variability (i.e., the ENSO), we first profile the Niño-3.4 index using the same Morlet wavelet analysis approach. The dominant time scales of the variability range from 1 to 5 years as expected (Figs. 4a,b). Interestingly, the Niño-3.4 index also exhibits a similar shift of its dominant time scales as that of the WNPSHI during 1979–2016 (cf. Figs. 3a and 4a). To quantify this observed association, the Spearman rank correlation is calculated on the 3–6-, 2–3-, and 1–2-yr scale-averaged time series of the WNPSH and the Niño-3.4 indices (Figs. 4c–e), respectively. A strong correlation coefficient at the 95% confidence level is found on the 3–6-yr time scale ($r = 0.88$), with moderate but statistically significant correlations on the 2–3-yr ($r = 0.53$) and 1–2-yr ($r = 0.54$) time scales. The 3–6-yr mode of the ENSO is well associated with the zonal WNPSH oscillation before the mid-1990s (Fig. 4c), while its 2–3- and 1–2-yr modes are more correlated with the zonal WNPSH oscillation after the mid-1990s (Figs. 4d,e), covering remarkable ENSO transitions in 1997–99 and

2009–11. This supports our earlier speculation that the ENSO might be closely associated with the interannual WNPSH oscillation and the decadal shifting of its leading modes. Yun et al. (2015) showed a decadal change in the covariability of the NPSH–WNPSH was associated with a dipole-like SST pattern in the tropical Pacific Ocean [i.e., a warming (cooling) in the WP and a cooling (warming) in the EP]. Such a tropical dipole-like SST pattern is likely the result of the air–sea interaction during La Niña (El Niño) phase. Comparing our findings with those of Yun et al. (2015), the decadal frequency shift of both the WNPSH and ENSO is in phase with the stronger covariability of the NPSH–WNPSH, as well as the stronger signal of tropical dipole-like SST, especially after the mid-to-late 1990s. Moreover, there were significantly weakened westerlies near the subtropical jet over the EA with a distinct increase in precipitation over southeastern China and in number of typhoons passing through the region after the mid-1990s (Kwon et al. 2007, 2005). These suggest that the zonal WNPSH oscillation, ENSO, and the EASM all covariate with each other and are likely intensified under the changing climate on the interdecadal time scale.

Up to this point, the subseasonal to interannual cycles of the zonal WNPSH oscillation are evident based on the Morlet wavelet analysis. We also reveal the decadal shifting of the dominant WNPSH interannual variations from a temporal scale of 3–6 years to 1–2 years over the 38-yr data period. The moderate-to-strong correlations between the WNPSH and Niño-3.4 index at 3–6-, 2–3-, and 1–2-yr cycles as well as the decadal shifting of their leading modes suggest a close coupling interaction between the WNPSH and the ENSO on the interannual and interdecadal time scales. Other potential climate variabilities could also have close associations with the zonal WNPSH oscillation in time–frequency space but are certainly beyond the scope of this study, such as the boreal summer intra-seasonal oscillation (BSISO), the Hadley circulation, the Tibetan Plateau warming, and nonlocal SST forcings in the Maritime Continent (Wang et al. 2018, 2008; Wu and Zhou 2008; Sui et al. 2007; He et al. 2001). A future study is suggested to clarify and integrate all these teleconnections with WNPSH.

In the following sections, dynamic processes associated with the WNPSH oscillation and its link to the EA summer climate on subseasonal time scales are diagnosed. The positive (negative) phase is defined as the days with the top 10% strongest (weakest) WNPSHI during JJA. The top 10% WNPSHI days are therefore those with their WNPSHI ranking from the 1st to the 350th ($0.1 \times 92 \times 38 \approx 350$), given 92 days in JJA and 38 years of the data period (1979–2016). In

addition to a thorough discussion over the key processes in extreme WNPSH phases as illustrated in Fig. 1, findings regarding the nonlinear quasi-biennial association of the extreme WNPSH phases and ENSO transitions are discussed in section 4e. Regions that are strongly influenced by the WNPSH phases are further explored in section 5 in a more quantitative framework.

4. Impacts of the extreme WNPSH phases on EA summer precipitation and their driving mechanisms

a. The tripole pattern of moisture distribution during positive WNPSH phase

Our recent endeavors have shown a strong association among synoptic atmospheric circulation, moisture transport, and anomalous wet conditions at various regions in midlatitudes (Lu and Lall 2017; Najibi et al. 2017; Lu and Hao 2017; Lu et al. 2013) with a diagnosis of the nexus of tropical moisture exports, associated atmospheric dynamics, and teleconnected climate variability. The composites of the anomalies of selected variables (PP, IVT, OLR, Vor850) for the positive WNPSH phase are therefore constructed (Fig. 5). The 25-day evolution of these composites, from 12 days ahead (day –12) to 12 days after (day 12) the onset of the positive WNPSH phase is examined. From here, the top 10% strongest (weakest) WNPSHI days are selected as the onset days for the positive (negative) WNPSH phase. Interestingly, the distribution of the top 10% strongest WNPSHI days is approximately at a ratio of 1:2:4 in June, July, and August, which agrees with the Xiang et al.'s (2013) finding that significant westward extension of the WNPSH frequently occurs in the late summer (i.e., the peak monsoon and typhoon period in EA).

Starting from day –12, an anomalous anticyclonic circulation emerges over the WNP and to the northwest of the suppressed PP [Fig. 5a(1)] and the positive OLR anomalies [Fig. 5a(2)] over the same region. As the anomalous anticyclone is strengthening and propagating westward to the Philippine Sea before the onset of the positive WNPSH phase, anomalous southwesterlies are blowing toward the Korean Peninsula and Japan, while anomalous northeasterlies are observed over the MC [Figs. 5a(1)–d(1)]. When such an anomalous IVT field keeps developing around the onset, warm and moist air is continuously transported from the SCS and the Philippine Sea to the MC and EA land areas (Ren et al. 2013; Lee et al. 2013), favoring cloud formation and resulting in two discernible moisture sinks (i.e., positive PP anomaly) there [Fig. 5e(1)].

Positive WNPSH Phase

FIG. 5. The composites of 1) the PP (shaded) and the IVT (vectors) anomalies and 2) the OLR (shaded) and the Vor850 (contours; interval: 2 s^{-1}) anomalies from (a) 12 days ahead (day -12) to (i) 12 days after (day 12) the top 10% strongest WNPSHI days (i.e., positive WNPSH phase) in 38 summers during 1979–2016 (base period). The solid (dotted) contour denotes positive (negative) values. Only those at the 95% confidence level are plotted, except the statistically significant PP anomalies in composite 1 that are circled with green (+) and brown (−) contours (Student's t test).

These two rain belts are considerably consonant with the widespread pattern of OLR and Vor850 anomalies. As the positive OLR anomaly is fully grown with increasing negative Vor850 anomaly, an enhanced

moisture divergence that prevents convection from developing emerges over the Philippine Sea, resulting in strongly suppressed PP there [Fig. 5e(2)]. As the two moisture sinks always situate to the northwest and

FIG. 5. (Continued)

southwest of the suppressed PP area, a distinct tripole pattern (sink–source–sink) of anomalous moisture distribution is formed during the positive WNPSH phase [Figs. 5c(1)–f(1)]. This tripole pattern demonstrates the prominent influence of the synoptic-scale WNPSH in reallocating atmospheric moisture and modulating regional weather over the WNP, EA, and the MC.

The tripole pattern is attributable to the anomalous wind field during the positive WNPSH phase, as a result of the atmospheric response to the thermal forcing, which is discussed next in section 4b. Regarding the formation of the two moisture sinks in the tripole pattern, the moisture sink to the northwest of the anticyclone (Fig. 5e) is caused by the prevailing anomalous southwesterlies over the EA lands. The dominant low-level southwesterlies favor cloud formation at the intersectional region of the warm and cool air masses stemming from different latitudes (Ren et al. 2013) and/or due to the land–sea thermal contrast (He et al. 2001). On the other hand, the northward displacement of the summertime ITCZ band (Krishnamurti et al. 2013) and the enhanced easterlies over the equatorial western

Pacific [Fig. 5e(1)] together encourage the formation of another moisture sink to the southwest of the anticyclone over the MC. Interestingly, the tripole pattern peaks on day 3 [Fig. 5f(1)], implying a 3-day lagged response of rainfall in the EASM and the MC regions. Mao et al. (2010) also identified a similar lead-lag relationship between the 20–50-day filtered WNPSH onset and intensified 20–50-day Yangtze rainfall, but with a longer leading time of at least 7.5 days. However, in this study it is shown that the lower and upper reaches of the YRB experience significantly anomalous precipitation 3 and 6 days after the WNPSH onset, respectively [Figs. 5f(1),g(1)]. Furthermore, the WNPSH onset does not always lead the YRB rainfall, since regional PP anomalies over the northern YRB had already been observed 3 days before the WNPSH onset [Fig. 5d(1)]. These might be due to the different definitions of the WNPSH and target regions as well as the time scales considered. Nevertheless, the identification of such a lead-lag association of the atmospheric circulation and regional rainfall provides the foundation to develop a predictive model for extremes as exemplified in Lu et al. (2016).

Three monsoonal regions are identified to be directly modified by this tripole pattern of moisture source/sink distribution, including two prominent moisture sinks with enhanced rainfall in the EASM and MC regions, and another strong moisture source with suppressed rainfall in the western North Pacific summer monsoon (WNPSM) region. The EASM and MC regions on average have their positive-WNPSH-phase-associated rainfalls at approximately the 65th–75th percentiles of their daily JJA rainfall in 1979–2016. More details on the relationship between regional rainfall and the WNPSH phases will be presented in section 5. Agreeing with our identified tripole pattern of PP anomalies, Kwon et al. (2005) showed a significantly negative correlation of rainfall anomalies between the EASM and WNPSM based on the data from 1979 to 2004. We argue that this negatively correlated relationship found by Kwon et al. (2005) could be explained as part of the influence of the extreme WNPSH phases. The tripole pattern found in this work further demonstrates the consistency of rainfall anomalies in response to an extreme WNPSH phase over a large domain covering the entire EASM, WNPSM, and MC regions.

From day 6 to day 12 (Figs. 5h,i), the anomalous anticyclone gradually weakens when it approaches the landmass (Hsu and Weng 2001). The PP anomalies in both the EASM and WNPSM regions return to their climatological states, concurrent with the dimming of the anomalous OLR and Vor850, marking the decay of the positive WNPSH phase. Interestingly, enhanced summer rainfall over the MC region persists throughout the entire diagnostic period (i.e., 25 days) during the positive WNPSH phase [Figs. 5a(1)–i(1)], which is not mentioned in the literature to the best of our knowledge. The maximum rainfall over the MC normally occurs in DJF, with its dry season in JJA (Robertson et al. 2011; Chang et al. 2005). The rainfall differences between the DJF and JJA over the MC generally range from 2 to 6 mm day^{−1} (Chang et al. 2005), which is comparable to the PP anomalies in the MC (0.75–2.25 mm day^{−1}) found in this study [e.g., Fig. 5f(1)]. This suggests that the positive WNPSH phase could considerably alter the monsoon climate in the MC region with significantly stronger than usual summer (dry season) precipitation.

Summer rainfall over South Asia and the Indian Ocean basin is also intensified because of the anomalous easterlies from the SCS and the Philippine Sea during the positive WNPSH phase [Figs. 5d(1)–f(1)] (Lee et al. 2013), revealing the teleconnected influence on the Indian and the South Asian summer monsoon climate. Similar piecewise findings were also presented in other studies. For example, Lee et al. (2013) exhibited the potential of strong WNPSH in inducing more precipitation in the EASM region (30°–40°N, 105°–150°E) as

well as in the Indian Ocean monsoon region (5°–15°N, 70°–105°E). Also, Wang et al. (2013) showed that the WNPSH enhances rainfall over Japan, the Korean Peninsula, and the equatorial Pacific, but suppresses rainfall over the WNP. Therefore, the tripole pattern found from the lead–lag composites further confirms and unifies those findings in the previous studies. [e.g., Fig. 5e(1)]. In addition, Mao et al. (2010) and Ren et al. (2013) found that summer rainfall over the YRB was positively correlated with the anomalous southwesterlies during the positive WNPSH phase. However, our result shows that significantly enhanced PP anomalies generally situate to the north of the YRB, while the south of the YRB experiences suppressed monsoons from day 3 to day 9 [Figs. 5f(1)–h(1)], forming an interesting south-to-north anomalous precipitation dipole across the YRB. The close association between the positive WNPSH phase and regional PP anomalies can offer improved predictability of the summer monsoon rains, with clearer space–time features of the moisture transports over EA.

In the next two sections, interplays between the anomalous anticyclonic circulation and the thermal forcing are investigated in order to diagnose the underlying dynamics and feedback mechanisms during the positive WNPSH phase.

b. Dynamics behind the OLR–vorticity pattern and its role in positive WNPSH phase

In addition to the tripole pattern, there is a consistent shift in space between OLR and Vor850 anomalies throughout the diagnostic period. A negative Vor850 anomaly, in particular, is always located to the west/northwest of the positive OLR anomaly since day −9, and becomes even more significant when approaching to the onset day [Figs. 5b(2)–f(2)]. It is therefore plausible that this OLR–vorticity pattern drives the westward propagation of the anomalous anticyclone during the positive WNPSH phase. Similar OLR–vorticity patterns were also documented in previous literature (Yun et al. 2008; Hsu and Weng 2001; Lu and Dong 2001); however, the role of the OLR–vorticity pattern was not well explained. To fill in this gap, we adopt and extend Hoskins and Karoly's (1981) theory of the atmospheric response to the perturbations induced by diabatic heating/cooling:

$$\bar{u}\xi'_x + \beta v' = fw'_z, \quad (1a)$$

$$\beta v' \approx fw'_z, \quad \text{for } L \gg \sqrt{\bar{u}/\beta}, \quad \text{and} \quad (1b)$$

$$f\bar{u}w'_z - f\bar{u}_z v' + w'N^2 = Q. \quad (2)$$

Most of the notations follow Hoskins and Karoly's (1981) work, such that $\bar{u}$ denotes the climatological zonal flow

(background state), v' is the anomalous meridional geostrophic velocity, ξ'_x denotes the zonal derivative of anomalous relative vorticity, w'_z denotes the vertical derivative of anomalous vertical velocity, L denotes the horizontal length scale of the heating, N is the Brunt–Väisälä frequency, and Q is the anomalous diabatic heating term.

To assist the diagnosis of the role of the OLR–vorticity pattern in the westward propagation of the anomalous anticyclone, the vertically integrated apparent heating Q_1 (Yanai et al. 1973; Luo and Yanai 1984) is adopted to investigate the diabatic heating of the troposphere and is computed as

$$\langle Q_1 \rangle = \frac{1}{g} \int_{100 \text{ hPa}}^{1000 \text{ hPa}} c_p \left[\frac{\partial T}{\partial t} + \mathbf{V} \cdot \nabla_h T + \left(\frac{P}{P_0} \right)^\kappa \omega \frac{\partial \theta}{\partial P} \right] dP. \quad (3)$$

Notations in the above equation are conventional, where T , θ , and ω are air temperature, potential temperature, and p velocity, respectively; the angle brackets $\langle \rangle$ denote a vertical integral.

It is suggested that during the early stage of the positive WNPSH phase, suppressed rainfall (i.e., less latent heat released) [Fig. 5a(1)], positive OLR anomaly [Fig. 5a(2)], and negative local SSTA [Fig. 6a(2)] all together induce an anomalous diabatic cooling ($Q = \langle Q_1 \rangle < 0$) [Fig. 6a(1)] over the WNP. Such a diabatic cooling source propagates westward following the anomalous OLR center and with an anomalous high to its west/northwest flank throughout the diagnostic period [Figs. 6a(1)–i(1)], resembling the aforementioned OLR–vorticity pattern.

Based on Hoskins and Karoly's (1981) vorticity equation, the ratio of the first and the second term in Eq. (1a) is roughly $\bar{u}/(\beta L^2)$ by scale analysis, showing that the first term can be ignored once $L \gg \sqrt{\bar{u}/\beta}$, and the vorticity equation then reduces to a simple vorticity balance [Eq. (1b)]. Given $\bar{u} \sim 5 \text{ ms}^{-1}$ over the WNP during JJA and $\beta \approx 2.17 \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$ at $\phi = 18^\circ \text{N}$, the term $\sqrt{\bar{u}/\beta} \approx 480 \text{ km}$. From the composite map on day –12 [Fig. 6a(1)], the diabatic cooling spreads over an large area (140°E – 180° , 10° – 30°N) in the WNP, and the horizontal scale of the cooling is therefore at least $L \sim 2000 \text{ km} \gg \sqrt{\bar{u}/\beta}$. Thus, for such a synoptic-scale thermal forcing, the stretching/squashing effect of the air column must be balanced by the meridional geostrophic motion due to the β -effect [Eq. (1b)]. Considering a diabatic cooling ($Q < 0$) at low latitudes, the third term on the LHS of Eq. (2) dominates (Hoskins and Karoly 1981), indicating the dominant vertical advection in balancing the heating/cooling. As a result, the negative thermal forcing ($Q < 0$) is balanced by the anomalous sinking motion ($w' < 0$) [Eq. (2)], causing the air columns to shrink ($w'_z = \partial w'/\partial z < 0$). This shrinking effect, from the simplified vorticity equation [Eq. (1b)], is balanced by the β -term via

inducing anomalous equatorward geostrophic motion ($v' < 0$) across the negative thermal source. This creates an anomalous high pressure associated with a negative vorticity anomaly ($\xi' < 0$) to the west of the cooling ($Q < 0$) by the geostrophic balance [e.g., Figs. 5d(2), 6d(1)]. As a result, the OLR–vorticity pattern, as an atmospheric response to the tropospheric cooling, plays an important role in driving the western extension of the anomalous anticyclone during positive WNPSH phase. Lu and Dong (2001) also pointed out that the negative SST anomalies off-equator over the western Pacific play a vital role in the westward extension of the WNPSH, which supports the mechanism of the atmospheric response to diabatic cooling of the troposphere discussed above.

In addition, it is noteworthy that the anomalous diabatic cooling during the positive WNPSH mainly centers at $\sim 18^\circ \text{N}$ [Figs. 6a(1)–c(1)], while the above atmospheric response with dominant vertical advection in balancing the diabatic heating/cooling is perfectly satisfied at tropics where $\gamma \equiv f^2 \bar{u}/(\beta N^2 H_Q H) \gg 1$ (Hoskins and Karoly 1981), which measures the ratio between the second and the third term in Eq. (2). Here $H_Q = Q/Q_z$ is the scale height of the heat source, $H = \min(H_Q/H_u)$, where $H_u = \bar{u}/\bar{u}_z$ is the scale height of the background zonal velocity. For a deep heating/cooling in the troposphere at $\sim 18^\circ \text{N}$ during an extreme WNPSH phase, $H_Q \sim 3 \text{ km}$, $H_u \sim 27 \text{ km}$, $\bar{u} \sim 5 \text{ ms}^{-1}$, and $N \sim 1.2 \times 10^{-2} \text{ s}^{-1}$, the corresponding γ value is around 0.36, implying that the meridional advection (i.e., the second term) is not negligible although the vertical advection (i.e., the third term) still dominates in balancing the anomalous heating/cooling in the troposphere. The above scale analysis provides an insight on why the observed OLR–vorticity pattern is not exactly east–west-oriented.

Another possible explanation for the westward propagation of an anticyclone in the Northern Hemisphere was provided by van Leeuwen (2007), based on the difference between the Coriolis forces acting on masses transported at the northern and southern flanks of the vortex. It was stated that, for an anticyclone without any meridional acceleration, the Coriolis force on the eastward-propagating air parcels in the northern vortex is larger than that on the westward-propagating parcels in the southern part. A westward translation of the whole vortex is thus required to compensate the difference (van Leeuwen 2007). Considering the large meridional extension of the synoptic-scale WNPSH [i.e., extending from 10° to 30°N in Figs. 6a(1)–f(1)], the variation of the Coriolis parameter f across the anticyclone might also play a role in the westward propagation of the anomalous anticyclone during the positive WNPSH phase, on top of the atmospheric response to the thermal forcing in the troposphere.

Positive WNPSH Phase

FIG. 6. The composites of 1) the $\langle Q_1 \rangle$ (shaded), Z850 (contours; interval: 5 m) and uv850 (vectors) anomalies and 2) the SST (shaded), SSR (contour; interval: 10 W m^{-2} starting from $\pm 5 \text{ W m}^{-2}$), and uv10m (vectors) anomalies from (a) 12 days ahead (day -12) to (i) 12 days after (day 12) the top 10% strongest WNPSH days (i.e., positive WNPSH phase) in 38 summers during 1979–2016 (base period). The solid (dotted) contours denote positive (negative) values. Only those at the 95% confidence level are plotted (Student's t test).

c. The local air–sea interaction during positive WNPSH phase

The composite analyses of the SSR and SST anomalies [Fig. 6(2)] reveal the role of the local air–sea interaction

in developing and driving the anomalous anticyclone during the positive WNPSH phase. On day -12, a contemporaneous negative SSTA and positive SSR anomaly are observed at the same location of the emerging tropospheric cooling over the WNP [Figs. 6a(1),a(2)],

FIG. 6. (Continued)

implying that this local negative SSTA likely triggers the diabatic cooling in the troposphere that formulates the OLR–vorticity pattern as discussed above (Hoskins and Karoly 1981). To further understand the local air–sea interactions during the positive WNPSH phase, the areal means of the local raw anomalies are computed based on a fixed-size domain (with 6° in latitude and 10° in longitude) following the center of the OLR anomaly (Fig. 7). From the extended diagnostic period from days -21 to 12 , the areal means of both the negative $\langle Q_1 \rangle$ and the apparent moisture sink $\langle Q_2 \rangle$ (Yanai et al. 1973) develop gradually with a persistent negative SSTA in the WNP (Fig. 7a), showing that the WNP cooling very likely triggers the tropospheric diabatic cooling and therefore the westward extension of the WNPSH. The anomalous anticyclone over the WNP was successfully reproduced by Wang et al. (2013) with a negative local SSTA perturbation in a coupling model, although they failed to reproduce the Indian Ocean warming that was argued to be a triggering factor for the positive WNPSH. Nevertheless, findings from this work and also others confirm the crucial role of the local air–sea interaction in the developing positive WNPSH phase, while nonlocal air–

sea interactions are still important but rather secondary. To explain the local SST cooling during the positive WNPSH formation, the convection–wind–evaporation–SST (CWES) feedback mechanism proposed by Xiang et al. (2013) is adopted. According to their illustration of the convection–divergence feedback, the convective precipitation that is suppressed by the SST cooling leads to the intensification of the low-level divergence and eventually contributes to an even more suppressed convection. This SST cooling could be further sustained through the positive CWES feedback as manifested in the wind–evaporation process (Xiang et al. 2013; Wang et al. 2013). In Fig. 7a, the profile of the anomalous $\langle Q_1 \rangle$ is found to be very close to the anomalous moisture sink $\langle Q_2 \rangle$, implying that the anomalous diabatic cooling may primarily be attributed to the suppressed latent heat released to the troposphere, likely due to the anomalous SST cooling that discourages convective precipitation based on the convection–divergence feedback.

Back to the composite maps, under the anomalous surface wind circulation, the negative SSTA in the WNP indeed persists for a few days before the positive

FIG. 7. Composite of the areal mean of raw anomalies for OLR and SST as well as (a) Z850, $\langle Q_1 \rangle$, and $\langle Q_2 \rangle$; (b) SLHF $\uparrow$, u10m, v10m, and UV10m; and (c) SSR, SLHF $\uparrow$, SLHF $\uparrow_{\text{Air}}$, and SLHF $\uparrow_{\text{AirSea}}$ over a fixed-size region with 6° in latitude and 10° in longitude, following the center of the OLR anomaly. The period is from 21 days ahead (day -21) to 12 days after (day 12) during the positive WNPSH phase.

WNPSH onset [Figs. 6a(2)–c(2)]. However, as the positive SSR anomalies intensify throughout the diagnostic period [Figs. 6a(2)–f(2)], they destabilize the anomalous anticyclone by warming up the SST after the positive WNPSH onset. This WNPSH-induced SST warming-up process, as opposite to the positive CWES feedback, can be explained by the negative convection–solar–SST (CSS) feedback: owing to the stable condition in the anticyclone (positive OLR anomaly), the downward surface solar radiation fluxes is enhanced (positive

SSR anomaly), and eventually the sea surface beneath the system is warmed up [Figs. 6d(2)–g(2) and 8]. Such a positive SSTA, induced by the CSS feedback, discourages the low-level divergence and favors the convective cloud formation, which is consistent with the observed weakening of both the negative Vor850 and the positive OLR anomalies [Figs. 5e(2)–i(2)], ultimately terminating the positive WNPSH phase.

Moreover, the local OLR anomaly and SSTA are discovered to be nearly 90° out of phase (Fig. 7a), which

Positive WNPSH phase

FIG. 8. Proposed feedback mechanism explaining the role of the air–sea interaction on the development of the anomalous anticyclone during the positive WNPSH phase. Convection–divergence feedback was illustrated by Xiang et al. (2013). The CWES feedback proposed by Xiang et al. (2013) and Wang et al. (2013) and the CSS feedback proposed in this work are adopted to explain the life cycle of the WNPSH phase.

implies an important local interaction between atmosphere and ocean under the WNPSH system (Wang et al. 2018). The local SST cooling since day -12 is found to be dominant until the OLR anomaly reaches its peak value, suggesting the role of the initial local SST cooling over the WNP in building up the anomalous anticyclone. Both the near-surface northeasterlies (i.e., u_{10m} , v_{10m} , and UV_{10m}) and the upward surface latent heat flux ($SLHF\uparrow$) strengthen from day -12 to -2 (Fig. 7b). This supports our hypothesis that there is a local CWES feedback in which the SST cooling is maintained through wind–evaporation during the developing stage of positive WNPSH phase. This is the first air–sea process between the initial local SSTA and the anomalous anticyclone before the positive WNPSH onset.

When the local OLR anomaly reaches its maximum on around day -2 , both the $SLHF\uparrow$ and the near-surface wind anomalies start to weaken (Fig. 7b), while the local positive SSTA starts to strengthen. In line with the proposed negative CSS feedback, the prevailing SSR anomaly arising from the high pressure anomaly heats up the local SST and encourages convection and thus weakens the anomalous high. These constitute the second and the third air–sea processes explaining how the anomalous high induces SST warming, as well as the way that the ocean feeds back on the atmosphere by weakening the anomalous high, respectively. The CSS feedback proposed in this study agrees with Ren et al.’s (2013) finding that the westward extension of WNPSH tends to warm up the ocean beneath through reducing latent heat flux and increasing incident solar radiation, and eventually acts as a negative feedback to WNPSH. This work attempts to demonstrate the complex local air–sea interactions that the positive CWES (negative CSS) feedback is dominant at the developing (decaying) stage of the positive WNPSH phase.

To verify the proposed hypothesis of the competitive interaction between the CWES and the CSS feedbacks during the positive WNPSH phase as illustrated in Fig. 8, the $SLHF\uparrow$ in the bulk parameterization can be decomposed into three components following the derivation in Wang et al.’s (2018) work:

$$SLHF\uparrow = \rho L_e C_e U \Delta q = \underbrace{\rho L_e C_e U' \overline{\Delta q}}_{(I)} + \underbrace{\rho L_e C_e \overline{U} \Delta q'}_{(II)} + \underbrace{\rho L_e C_e (U' \Delta q')'}_{(III)}, \quad (4)$$

where ρ is the density of air, L_e is the latent heat of vaporization, C_e is the turbulent exchange coefficient for latent heat, U is the near-surface wind speed, and $\Delta q = q_s - q_a$ is the difference in the specific humidity between sea surface and near-surface atmosphere (Liu et al. 1979; Bourras 2006; Yu et al. 2007). The overbar and prime are the Reynolds averaging operators for the climatological and the anomalous components, respectively. Term I is the air component of the $SLHF\uparrow$ (denoted as $SLHF\uparrow_{Air}$) determined by anomalous near-surface winds; term II is the air–sea component (denoted as $SLHF\uparrow_{AirSea}$) driven by the humidity difference at the air–sea interface, and term III is a nonlinear term that is generally negligible compared to the first two terms (Wang et al. 2018). Therefore, $SLHF\uparrow_{Air}$ is adopted to diagnose the CWES feedback as manifested in the wind–evaporation process, while $SLHF\uparrow_{AirSea}$ investigates the collective changes in Δq at the air–sea interface, near-surface temperature, and SST based on the Clausius–Clapeyron equation. Therefore, $SLHF\uparrow_{AirSea}$ is adopted to diagnose the CSS feedback as manifested in the thermal evaporation due to SST warming. From Fig. 7c, profiles of the $SLHF\uparrow_{Air}$ and the $SLHF\uparrow_{AirSea}$ intersect on day -2 , showing that the

Negative WNPSH Phase

FIG. 9. As in Fig. 5, but for the top 10% weakest WNPSHI days (i.e., negative WNPSH phase).

SLHF $\uparrow$ is no longer dominated by the wind–evaporation (of the CWES feedback) but the thermal evaporation (of the CSS feedback) during and after the positive WNPSH onset. This flux decomposition analysis supports our hypothesis of the competition between the CWES and the CSS feedbacks during the positive WNPSH phase.

In brief, given the initial SST cooling in the WNP, the positive CWES feedback likely plays a crucial role in triggering and enhancing the anomalous anticyclone at its early stage (Fig. 7). The negative CSS feedback then dominates the air–sea interaction and is responsible for the decay of the anomalous anticyclone. It might also serve as a sign of the transition to the negative WNPSH

FIG. 9. (Continued)

phase (Ren et al. 2013). In addition, the eastern equatorial Indian Ocean (IO) experiences anomalous SST warming throughout the positive WNPSH phase [Figs. 6a(2)–i(2)], which might be associated with the Indian dipole mode (IDM) (Black et al. 2003) and is likely due to the anomalous easterlies induced by the positive WNPSH that suppresses the Indian summer monsoon (ISM) (Wang et al. 2013; Lee et al. 2013), suggesting the potential teleconnected influence of the WNPSH on the IO climate systems.

In the next section, the negative WNPSH phase will be discussed with the aim of generalizing the zonal WNPSH oscillation, without much repetition of the discussion done for the positive phase.

d. An almost reversed but stronger signal during the negative WNPSH phase

Compared to the positive WNPSH phase, the negative phase (i.e., the eastward retreat of the WNPSH) has nearly reverse but stronger impacts on the EA summer climate. Similar to the diagnostic procedures above, the top 10% weakest WNPSHI days are selected to

investigate the negative WNPSH phase (Fig. 9). During the preonset period (day –12 to day –3), a moisture sink with a stronger than usual PP anomaly develops over the WNP and later propagates westward into the Philippine Sea [Figs. 9a(1)–d(1)], which is concurrent with the movement of the anomalous convective system with negative OLR and positive Vor850 anomalies [Figs. 9a(2)–d(2)]. This anomalous cyclone with cyclonic IVT anomalies enervates the background summer monsoonal winds (Fig. 2) and blocks the moisture transports from the warm and moist areas. Ultimately, anomalous droughts are induced in Japan, the Korean Peninsula, and the MC before the onset [e.g., Fig. 9d(1)]. During the onset of the negative WNPSH phase, a nearly reversed tripole pattern with a strong moisture sink over the WNPSM region and continent-wide anomalous droughts over the EASM and MC regions is found [Fig. 9e(1)]. It is noteworthy that the negative WNPSH phase is associated with a larger extent and bigger magnitude of the droughts than those of the wet conditions triggered by positive WNPSH [cf. Figs. 9e(1) and 5e(1)]. This implies that the eastward retreat of the

Negative WNPSH Phase

FIG. 10. As in Fig. 6, but for the top 10% weakest WNPSHI days (i.e., negative WNPSH phase).

WNPSH, which has seldom been discussed in previous studies, could have an even more pronounced impact on the deficits in summer monsoon rains over the EA and the MC. The synoptic-scale drought persists for about one week since the onset day, which is consistent with the time scale of the strengthened wet condition observed during the positive WNPSH phase [cf. Figs. 9e(1)–h(1)]

and 5e(1)–h(1)]. Teleconnection with the anomalous droughts over India is also noted during and after the negative phase, which suggests that the strongly anomalous cyclone over the WNP extracts moisture from not only local but also remote regions like India and the adjacent seas. The finding shown above reveals that extreme WNPSH phases (either positive or negative)

FIG. 10. (Continued)

could play crucial roles in modulating summer rainfall over the EA, MC, and Indian Ocean basin. Impacts from the negative WNPSH phase (i.e., the eastward retreat) can be even stronger and therefore deserve more attention.

Again, a similar OLR–vorticity pattern (i.e., a positive vorticity anomaly to the west/northwest of the negative OLR anomaly) is found [e.g., Fig. 9d(2)]. Recalling the atmospheric response to the thermal heating (Hoskins and Karoly 1981) as illustrated in section 4c, the anomalous diabatic heating (i.e., positive $\langle Q_1 \rangle$ anomaly) found over the WNP must induce an anomalous low pressure (i.e., positive Vor850/negative Z850 anomaly) to its west by geostrophic balance [Figs. 10a(1)–e(1)]. As the anomalous $\langle Q_1 \rangle$ is slightly larger than $\langle Q_2 \rangle$ before the negative WNPSH onset (Fig. 11a), this indicates that other than the latent heat released, the trapping of radiation by clouds also contributes to the anomalous diabatic heating in the troposphere. This OLR–vorticity pattern is responsible for the westward propagation of the anomalous cyclone during the negative WNPSH phase, explained as follows. Based on the barotropic divergent

vorticity equation, the rate of local change of the relative vorticity ($\partial\xi/\partial t$) is mainly due to the horizontal advection term of absolute vorticity $[-V_H \nabla(\xi + f)]$ (Holland 1983). When the background flow (V_H) is weak, $\partial\xi/\partial t$ is mainly balanced by the horizontal planetary advection term (i.e., $\partial\xi/\partial t \approx -V_H \nabla f = -\beta v$). As the air parcels are moving southward at the western flank of the cyclonic circulation, a region with a positive local change of relative vorticity ($\partial\xi/\partial t \approx -\beta v > 0$) develops to the west of the cyclone [Fig. 9e(1)]. As Holland (1983) showed that a cyclone tends to move in a direction with increasing relative vorticity, the OLR–vorticity pattern formed by the atmospheric response to diabatic heating in the negative WNPSH phase [Figs. 10a(1)–e(1)] creates a region with positive vorticity tendency that drives the westward propagation of the anomalous cyclone. This may also explain the westward propagation of the anomalous anticyclone in the positive WNPSH phase when there is always an anomalous high (i.e., negative vorticity tendency) to its west as if to drive it westward.

Moving forward to the local air–sea interaction during the negative phase, the anomalous convection and the

FIG. 11. As in Fig. 7, but for the negative WNPSH phase.

associated anomalous diabatic heating in the troposphere are possibly induced by an initial SST warming in the WNP found between day -21 and day -16 (Fig. 11a). Although the local SSTa under the WNPSH system is slightly negative since day -15 [Figs. 10a(2)–(d2), 11a], the warm summertime SST over WNP may still maintain the development of the anomalous convection. Later during the negative WNPSH onset, the anomalous near-surface southwesterlies in the negative OLR center become stronger [Figs. 10c(2)–e(2)]. Although it encourages the wind–evaporation process ($\text{SLHF} \uparrow_{\text{Air}}$; Fig. 11b) that cools down the SST, it also improves ventilation at surface such that more water

vapor is available to fuel the convective system, as revealed from the abrupt increase in $\langle Q_2 \rangle$ (Fig. 11a). This process is favorable for cloud formation and pushes the negative OLR anomaly to reach its peak on day -1. In terms of the local air–sea feedbacks, the wind–evaporation ($\text{SLHF} \uparrow_{\text{Air}}$; CWES feedback) and the thermal evaporation ($\text{SLHF} \uparrow_{\text{AirSea}}$; CSS feedback) profiles intersect on day -15 (Fig. 11c). Therefore, competitive interaction between the two air–sea feedbacks ends on day -15, far earlier than that in the positive WNPSH phase. The wind–evaporation process then becomes comparable in magnitude with the thermal evaporation until day -3. The $\text{SLHF} \uparrow_{\text{Air}}$ then

becomes far stronger than the $SLHF \uparrow_{AirSea}$ since day -3 , suggesting that wind–evaporation process is dominant in cooling down the local SST underneath the anomalous convection and terminating the negative WNPSH phase [Figs. 10e(2)–g(2), 11c].

With the understanding of the dynamics and air–sea interactions, a quantitative modeling will be applied in the future research pursuit, which might offer certain predictability of the anomalous summer monsoon rains/droughts over the EASM, WNPSM, and MC regions.

e. Quasi-biennial WNPSH–ENSO relationship during extreme WNPSH phases

As mentioned in section 3, the zonal WNPSH oscillation has a close association with ENSO on interannual and interdecadal time scales. Extending from the discussions over the local air–sea interactions during the extreme WNPSH phases, potential association between the extreme WNPSH phases and the ENSO events as defined by Trenberth (1997) are further explored as follows. The results show that up to 48% of the strongest WNPSHI days are found to occur 9–12 months after short El Niño events [i.e., JJA(−1) and SON(−1)], and at most 44% are 1–3 months ahead of the subsequent persistent La Niña events [i.e., JJA(0) and SON(0)] (Fig. S3a). A similar but reversed pattern is found during the negative WNPSH phase (Fig. S3b). This quasi-biennial WNPSH–ENSO relationship is consistent with the moderate correlation between the WNPSHI and Niño-3.4 index on the 1–2- and 2–3-yr time scales (section 3b). To further understand this quasi-biennial relationship, various types of ENSO transitions from lag -10 to lag 2 are categorized based on the ENSO transition pattern. Note that lag -10 and lag 2 are chosen here because they are the time lags with maximum numbers of lagged El Niño and La Niña events, so the range better captures the lead–lag relationship. The result reveals that up to 31% of the strong WNPSHI days occur under the transition from El Niño to La Niña (denoted as El-Neu-La), while the Neu-La transition accounts for 14% of the strong WNPSHI days (Fig. S3c). However, 18% of the WNPSHI days occur during a neutral event, and the remaining 37% occurs under other types of ENSO transitions. Since a substantial number (55%) of strong WNPSHI days seem to not simply follow the quasi-biennial WNPSH–ENSO relationship, together with the moderate correlation between the two variabilities on the 1–2- and 2–3-yr time scales, the WNPSH–ENSO relationship is thus likely nonlinear and conditional. Referring to the negative WNPSH phase, 28% and 15% of the weak WNPSHI days occur during a developing El Niño (Neu-El) and during a decaying La Niña (La-Neu), respectively. These

considerably resemble the reversed ENSO transition in the positive WNPSH phase (Figs. S3c,d). Again, more than half of the weak WNPSHI days (57%) seem not to simply follow this linear relationship.

In general, the positive (negative) WNPSH phase sometimes occurs during (i) a decaying El Niño (La Niña) in the preceding summer/autumn, and/or (ii) a developing La Niña (El Niño) in the current summer/autumn. A full ENSO transition (i.e., i + ii) is more frequently seen during the positive WNPSH phase than its counterpart, as exemplified by the transition from a moderate-to-strong El Niño year (e.g., 1982/83, 1994/95, 1997/98 and 2009/10) to a La Niña year. This quasi-biennial ENSO–WNPSH relationship resembles the tropospheric biennial oscillation (TBO) (Meehl 1987) and largely supports the interaction of near-annual ENSO transition and the WNPSH behaviors explained by the combination mode (C-mode) dynamics (Stuecker et al. 2013, 2015; Timmermann et al. 2018). Another possible explanation would be the stronger surface wind stress anomalies over the equatorial western Pacific in the positive WNPSH phase [Fig. 6(2)] that could generate equatorial Kelvin waves to stimulate the biennial ENSO cycle, as demonstrated in Kim and Lau's (2001) idealized ENSO–monsoon coupled system. Specifically, Li et al. (2010) argued that the summertime El Niño event was triggered by the weakened WNPSH through anomalous surface westerlies in the tropical western Pacific. While the enhanced Hadley circulation due to the convective heating over the central Pacific Ocean during an El Niño event could strengthen the sinking motion in the WNPSH, Yun et al. (2008, 2010) indicated that the weakened Walker circulation due to the SST warming in both the eastern Pacific and the Indian Ocean during an El Niño event was responsible for the suppressed convection over the Philippine Sea. These findings all suggest that extreme WNPSH phases and ENSO events can be each other's precursor, especially when a strong El Niño is present (Wang et al. 2001). However, it should be recalled that at least half of the extreme WNPSHI days appear not to follow the discussed relationship, suggesting that the occurrence of extreme WNPSHI days is not simply linearly associated with the ENSO transitions on a quasi-biennial time scale, which agrees with and further confirms previous findings (Li et al. 2010; Wang et al. 2013; Xiang et al. 2013). Considering the findings in section 3 that the WNPSHI exhibits significant modes on time scales ranging from subseasonal to interannual and its association with ENSO on interannual and interdecadal time scales, we speculate that the nonlinearity of the WNPSH–ENSO relationship might also be on some longer time scales. This work only demonstrates a

preliminary statistic exploration to address the non-linear quasi-biennial WNPSH–ENSO relationship and the interdecadal association; a complete and in-depth diagnosis is needed to fully understand the WNPSH–ENSO relationship on other crucial time scales, such as the significant 2–3- and 3–6-yr time scales shown in [section 3b](#). Furthermore, the zonal WNPSH oscillation is likely not just associated with the ENSO because of its long-range significant oscillation modes, so a better understanding of its interaction with other potential climatic variabilities such as the boreal summer intraseasonal oscillation over the WNP, the Hadley circulation, Tibetan Plateau warming, and nonlocal SST forcing ([Wang et al. 2018, 2008](#); [Wu and Zhou 2008](#); [Sui et al. 2007](#); [He et al. 2001](#)) may help close up this research gap.

5. Quantitative relationship between the zonal WNPSH oscillation and regional rainfall

Based on the diagnostic analyses of the association between the extreme WNPSH phases (i.e., the top 10% strongest/weakest WNPSHI days) and PP anomalies in the EA, we select several regions to further quantify their relationship with WNPSH phases, including eastern China (EC), the Korean Peninsula (KR), central Japan (CJP), the WNP and the MC ([Fig. S4](#)). The first four originate from the EASM and WNPSM regions defined by [Ding and Chan \(2005\)](#). We first explore the linear association (Spearman rank correlation) between their areal mean JJA daily PP anomalies and the WNPSHI at different time lags. We find that only the precipitation over the WNP region is linearly correlated with the WNPSHI ($r = -0.6$ at lag 0), while the linear correlations are weak in other regions. This is as expected since the WNP region covers the propagation pathway of the anomalous circulation system during its lifetime, the convective precipitation over the WNP is directly induced by the anomalous WNPSH ([Figs. 5 and 9](#)). This preliminary correlation analysis demonstrates that the influence of WNPSH on the EA summer precipitation is not simply linear. A further investigation regarding their relationship is presented next.

As discussed in [section 4a](#), the anomalous rainfall over lands is on average at the 70th percentile of the historical summer rainfall during the positive WNPSH phase, suggesting that the positive phase is associated with moderate-to-strong rainfall over the regions. We therefore adopt the linear quantile regression (LQR) to explore the associations between the normalized PP anomalies and the WNPSHI at different quantile levels from 10% to 90% for the selected regions (i.e., EC, KR, CJP, and MC) in [Fig. S4](#). The LQR specifies the change

of the mean of the dependent variable in the conditional quantile as the independent variable changes ([Koenker 2017, 2005](#); [Hao and Naiman 2007](#); [Koenker and D'Orey 1987](#)). The analysis is conducted using the open-source R package “quantreg” ([Koenker 2017](#)). More details regarding this method and its execution in this study can be found in the [appendix](#). Complete documentation of the method and R package can be found in [Koenker \(2017, 2005\)](#).

To quantitatively describe the lead–lag relationship between the WNPSHI and the regional PP anomalies over the selected regions, different time lags in days (from lag -6 to lag 6) are explored by the LQR analysis. Here lag -6 (lag 6) denotes that the regional PP anomalies are leading (lagging) the WNPSHI by 6 days. Since this method only serves as a supporting analytical tool, the model fitting is provided in the supplemental material ([Figs. S5–S7](#)) for additional reference. An LQR slope is considered significant only when the slope coefficient's 95% confidence interval (CI) ([Figs. S5–S7](#), shaded areas) does not overlap with the 95% CI of the linear regression slope using the entire dataset ([Figs. S5–S7](#), red dashed line). From the LQR analysis, responses of the regional PP anomalies to the variation in the WNPSHI at different time lags are quantified. Significant LQR slopes at the lower and higher quantiles of the WNPSHI are found from lag 2 to lag 5 over the EC, suggesting a dramatic change of rainfall 2–5 days after the onset of extreme WNPSH phases ([Fig. S5](#)). Similar responses of PP anomalies are also found in KR and CJP from lag 0 to lag 1 ([Fig. S6](#)) and from lag -3 to lag -2 ([Fig. S7](#)), respectively. Results from the lead–lag LQR analysis show the movement of WNPSH-induced PP anomalies: CJP (before the WNPSH onset) $\rightarrow$ KR (around the onset) $\rightarrow$ EC (after the onset). Moreover, by reducing the quantile interval, the LQR slopes generally follow an exponential curve at almost all the time lags for the EC, KR, and CJP regions, while the curvatures of the regression line are more remarkable at the significant time lags mentioned above ([Figs. S5–S7](#), blue curve). This lead–lag relationship promises a great potential in predicting anomalous summer precipitation during extreme WNPSH phases.

Different from these extratropical areas, the LQR analysis reports that the varying WNPSHI does not significantly alter the distribution of the rainfall anomalies in the MC region (figure not shown). Since most of the MC islands feature a less profound seasonal cycle of rainfall, and are regulated by different monsoon systems, intraseasonal oscillations and the ENSO events ([Lee 2015](#); [Robertson et al. 2011](#); [Chang et al. 2005](#)), impacts from the extreme WNPSH phases may thus be

diminished. Nevertheless, from the simple boxplot diagram for the MC region, moderate increasing trends of the boxplot quantile values (i.e., 25%, 50%, and 75%) are still found for all the time lags, implying a fair contribution of the zonal WNPSH oscillation to the summer rainfall in the MC region (Fig. S8).

6. Summary

This study begins with the wavelet analysis on the WNPSHI and the Niño-3.4 index, in line with the first objective of providing complete temporal variability profiles and a big picture of the relationship between the motivating phenomenon (i.e., the zonal WNPSH oscillation) and one of the most important global climatic signals (i.e., the ENSO) in the time–frequency space. To investigate the two extreme phases of the zonal WNPSH oscillation, diagnosis on the top 10% strongest (positive phase) and weakest (negative phase) WNPSHI days is conducted and reveals the crucial role of extreme WNPSH phases in influencing the summertime EA climate. Interesting findings include the tripole pattern of the moisture distribution, the OLR–vorticity pattern as a manifestation of the atmospheric responses to the tropospheric heating/cooling, and the competitive interaction between local air–sea feedbacks especially during the positive WNPSH phase. This study aims to have a close-circle analysis on the zonal WNPSH oscillation as illustrated in Fig. 1. Major results of this study are summarized as follows:

- 1) Moderate-to-strong positive correlations between the WNPSHI and Niño-3.4 index are found on the 1–2-, 2–3-, and 3–6-yr time scales. The mid-1990s and the late 2000s are identified to be the two important time points for the decadal shift in the dominant time scale of the zonal WNPSH oscillation and ENSO, from 3–6- to 2–3-yr and finally to 1–2-yr cycles during 1979–2016. Similar decadal changes in the EASM were also documented in the literature, suggesting an intensified zonal WNPSH oscillation and ENSO in the face of global climate change, as well as their close interdecadal association. A quasi-biennial WNPSH–ENSO relationship is identified as follows: the positive (negative) WNPSH phase sometimes occurs during the ENSO transitions of (i) a decaying El Niño (La Niña) in the preceding summer/autumn, and/or (ii) a developing La Niña (El Niño) in the current summer/autumn. A complete ENSO transition from moderate-to-strong El Niño to La Niña is often seen during the positive WNPSH phase, offering potential in predicting ENSO events and extreme WNPSH phases and thereby anomalous summer rainfall over the EASM, WNPSM, and MC regions. However, more than half of the extreme WNPSHI days occur under ENSO transitions that do not follow the quasi-biennial WNPSH–ENSO relationship, implying a nonlinear nature of the relationship and requiring further studies on the full picture of the WNPSH–ENSO relationship.
- 2) A tripole pattern of anomalous precipitation is identified during the positive (negative) WNPSH phases, which reveals intensified (weakened) precipitation over the EASM and MC region but suppressed (strengthened) precipitation over the WNPSM region. Stronger influences of the negative WNPSH phase are noted, suggesting that more attention should be paid to the eastward retreat of the WNPSH, which could substantially suppress the summer monsoon rains in EA land areas and the MC. Under such a tripole pattern during an extreme WNPSH phase, the onset time of the significantly anomalous precipitation varies in different regions based on the lead–lag LQR analysis. It suggests a gradual movement of WNPSH-induced rainfall anomalies starting from central Japan and going to the Korean Peninsula and last eastern China. A fair contribution of the zonal WNPSH oscillation to the summer precipitation extreme in the MC region is observed, although the influences of the complex monsoon systems possibly conceal the contribution to some extent. Meanwhile, the anomalous precipitation over the WNPSM region is found to be highly correlated with the contemporaneous WNPSHI signal. These all provide some predictability of summer precipitation over the EA, WNP, and MC regions induced by extreme WNPSH phases.
- 3) An OLR–vorticity pattern is identified as an anomalous high (low) always to the west/northwest of the OLR center and is believed to drive the westward propagation of the anomalous circulation during the positive (negative) WNPSH phase. The tropospheric diabatic cooling (heating) over the WNP is suggested to be responsible for this distinct pattern and therefore the propagation of the system.
- 4) Local air–sea interaction over the WNP serves as the primary factor in the formation of WNPSH phases. The WNP cooling (warming) triggers the anomalous diabatic cooling (heating) in the troposphere, which encourages early development of positive (negative) WNPSH phase. From the flux decomposition diagnosis, competitive interaction between the CWES and the CSS feedback is prominent in the positive WNPSH phase, with the former dominating during

the developing stage of positive WNPSH phase and the latter dominating in the decaying stage. Competition of feedbacks is not prominent in the negative WNPSH phase, and yet the CWES feedback prevails in its decaying stage.

From the diagnosis of the observable features and the underlying mechanisms of the extreme WNPSH phases, the zonal WNPSH oscillation is undeniably responsible for amplifying the historical summer rainfall extremes in the entire EA and MC regions, closing up the circle for the framework of understanding the role of the zonal WNPSH oscillation in the EA and MC summer climate system and extreme precipitation.

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APPENDIX

Linear Quantile Regression

Comparing with the ordinary least squares regression, the quantile regression is more robust against non-normal errors and outliers in the response measurement (Okada and Samreth 2012). As mentioned in section 5, LQR specifies the change of mean of the dependent variable in the conditional quantile as the independent variable changes (Hao and Naiman 2007; Koenker and D'Orey 1987). The regression process is described as follows:

Suppose $\{y_i; i = 1, \dots, n\}$ denotes the dependent variables, $\{x_i; i = 1, \dots, n\}$ denotes the independent variables, and β_θ is the coefficient in the regression process given the quantile θ ($0 < \theta < 1$). The LQR model can be described as

$$E(y_i|\theta) = x_i\beta_\theta.$$

For a specific quantile θ , the estimator $\widehat{\beta}_\theta$ is given by

$$\widehat{\beta}_\theta = \arg \min_{\beta \in R^k} \left[\sum_{i \in \{i: y_i \geq x_i \beta\}} \theta |y_i - x_i \beta| + \sum_{i \in \{i: y_i < x_i \beta\}} (1 - \theta) |y_i - x_i \beta| \right].$$

The estimators are calculated by the Simplex algorithm described in Koenker and D'Orey (1987).

LQR is thus adopted in this study to explore the associations between the areal mean PP anomalies and the WNPSHI at different quantile levels from 10% to 90%, for the selected regions. When the LQR slope is significant (as defined in section 5), it means the change of the dependent variable (i.e., the areal mean of JJA precipitation anomalies for a region) in response to any unit change of the independent variable (i.e., the WNPSHI) at that specific quantile level of the independent variable must be significantly different from the linearity based on the entire dataset. This reveals the effects of the WNPSHI on the distribution of the precipitation data. Thus, at the significant time lags of the regions discussed in section 5, the LQR slopes of the relationship of "precipitation–WNPSHI" exponentially increase with the quantile levels (blue curve in Figs. S5–S7), indicating more extreme summer rainfall/droughts when the WNPSHI is being extremely anomalous.

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