Zonal Oscillation of Western Pacific Subtropical High and Subseasonal SST Variations during Yangtze Persistent Heavy Rainfall Events

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(Manuscript received 12 December 2012, in final form 11 June 2013)

ABSTRACT

This study examines the relationship between the zonal oscillation of the western Pacific subtropical high (WPSH) and underneath sea surface temperature (SST) variation on a subseasonal time scale, associated with the persistent heavy rainfall (PHR) events over the middle and lower reaches of the Yangtze River valley (MLYRV) in China. A total of 76 PHR events and 45 break events in the summers of 1979–2011 are first identified over the MLYRV and divided into early and late summer groups. During the PHR events over the MLYRV for both groups, the WPSH stretches more westward, accompanied by the positive anomalies of the 500-hPa geopotential height field over East Asia and its coastal region south of 30°N and the subseasonal warmer SSTs beneath the WPSH western edge. The time-lagged composites suggest that the WPSH western edge exhibits westward-then-eastward migration on a subseasonal time scale for the PHR events. The zonal changes of the WPSH and anomalous circulation and SST anomaly (SSTA) signals for break events is almost the mirror image of that for the PHR events for the early summer group. Accompanied by the WPSH westward extension, the increased incident solar radiation and decreased latent heat flux over the coastal region of East Asia contribute to the positive SSTAs beneath the western part of the WPSH. The positive SSTAs construct a convective instability that provides an adverse condition for maintaining the anticyclonic anomalies in the mid–lower levels. The persistent SST warming is also favorable to the transition of low-level circulation from anticyclonic to cyclonic anomalies over the coastal region. As a result, the WPSH withdraws eastward after the peak of the rainfall events over the MLYRV.

1. Introduction

The middle and lower reaches of the Yangtze River valley (MLYRV) is located over eastern China, with heavy population and rapid growth in economy. This region is influenced by the East Asian summer monsoon (Tao and Chen 1987; Ding and Chan 2005; Jin et al. 2013); therefore, the variation of flood/drought in this region has been the focus in many studies (Shen and Lau 1995; Mao and Xu 2006; Yang et al. 2005; Sun and Yang 2005; Huang et al. 2007; Zhu et al. 2007; Yang and Zhu 2008; Ye and Lu 2012; Jiang et al. 2013; Greatbatch et al. 2013). The persistent moderate-to-heavy rainfall events cause summer flood/drought disasters over the MLYRV (Tang et al. 2006; Bao 2007). Analyses of the long-term station dataset show that persistent heavy rainfall (PHR) events over the MLYRV are abundant in summer. The PHR features strong intensity (about 15–50 mm day\(^{-1}\) on average), long duration (3–7 days), and a large area of rainfall band (Tang et al. 2006; Bao 2007). Between two PHR events there exists a clear break, characterized by a continuous dry period with little precipitation. Thus, the precipitation over the MLYRV exhibits the feature of subseasonal variation and the PHR occupies the active phase of the rainfall oscillation (Mao et al. 2010; Yang et al. 2010).

The western Pacific subtropical high (WPSH) plays an important role in the MLYRV rainfall variations (Huang 1963; Wang et al. 2000b; Liu and Wu 2004; Sui et al. 2007; Tao and Wei 2009; Mao et al. 2010). The WPSH is usually represented by either the extent of the 5880 geopotential meter (gpm) isoline at 500 hPa or the specific isolines at 850 hPa. The location and intensity of the WPSH usually determine the zone where the warm and humid air from the oceanic region meets the cold air from mid- to high latitudes. Thus, the WPSH strongly influences the weather and climate in the East Asian monsoon region (Tao and Chen 1987; Wu and Wang 2001; Wu et al. 2003; Ding and Chan 2005). The
WPSPH exhibits a pronounced seasonal cycle. It reaches the northernmost position in summer with the western edge at 500 hPa near the coastline of East Asia. The WPSPH displays distinct northward propagations in summer, cooperated with the seasonal marching of the East Asian summer monsoon rainfall. For example, one of the robust northward jumps of the WPSPH occurs in mid-July, when the mei-yu (commonly known as the plum rain) finishes over the MLYRV (Tao and Chen 1987; Lu 2001; Wu et al. 2003; Ding and Chan 2005; Yang et al. 2013). Besides seasonal variation, the WPSPH exhibits both zonal and meridional changes on interannual to decadal time scales that have noticeable connection with the interannual to decadal variations of the East Asian summer rainfall (Lu 2001; Lu and Dong 2001; Wu et al. 2003; Sui et al. 2007; Grotjahn and Oman 2007; Zhou et al. 2009).

On a subseasonal time scale, the western edge of the WPSPH exhibits migrations first westward then eastward, corresponding to persistence and break sequences of the MLYRV rainfall (Mao and Xu 2006; Bao 2007; Wang et al. 2000b; Mao et al. 2010). Mao et al. (2010) found that significant westward extension of the WPSPH leads to an extremely positive anomaly of Yangtze rainfall, and an abrupt eastward retreat links to the extreme negative rainfall anomaly. Yang et al. (2010) identified two dominant modes of subseasonal variation (biweekly and 21–30 days) for the MLYRV summer rainfall. Both modes are closely related to the zonal shift of the WPSPH. The WPSPH impacts features of the subseasonal rainfall by controlling the supply of lower-level moisture from ocean to land (Wang et al. 2000b; Liu et al. 2008). When the WPSPH extends more westward, an anomalous anticyclone in the lower level occupies East Asia and its coastal waters. Thus, the above-normal warm and moist southwest monsoon flow, originated from the South China Sea (SCS) and the western Pacific Ocean, lands on East China. The warm and moist airflow cooperates with the cold airflow from the midlatitudes, resulting in anomalous convergence and ascending motion over southeastern China: hence, providing necessary dynamical and thermal conditions for the persistent rainfall (Tao and Chen 1987; Zhou and Yu 2005; Mao et al. 2010).

The oceanic anomalies in the Indo–Pacific are highly correlated with the interannual to decadal variations of the WPSPH (Lu and Dong 2001; Sui et al. 2007; Xie et al. 2009; Zhou et al. 2009; Li et al. 2010; Wu et al. 2010). Based on analyses of observations and GCM simulations, Lu (2001) and Lu and Dong (2001) suggested that the interannual zonal variation of the WPSPH at its western edge is closely linked with the intensity of atmospheric convection over the warm pool and the SST anomaly (SSTA) off equator over the western Pacific. Cooler SSTs in the western Pacific lead to suppressed local convection and anomalous northward anticyclonic circulation in the lower level. These result in the westward extension of the WPSPH. Sui et al. (2007) demonstrated that the WPSPH exhibits significant oscillations at 2–3 and 3–5 years, which are attributed to the atmospheric response to the remote forcing of the SST in the Maritime Continent and equatorial central-eastern Pacific, respectively. Wang et al. (2000a) presented that the WPSPH variation is also a key factor influencing the ENSO–East Asian teleconnection. Li et al. (2010) presented a selective interaction between ENSO and the North Pacific subtropical high in which the WPSPH is a key. Wu et al. (2010) suggested that the anomalous anticyclone over the western North Pacific is maintained by the combined effects of the local SSTA forcing and the remote forcing from the Indian Ocean during the summer with El Niño decaying. The above-normal Indian Ocean SST warms the tropical troposphere through the adjustment of deep convection and triggers an anticyclonic pattern over the northwestern Pacific via emanation of baroclinic Kelvin waves, thus enhancing the WPSPH (Xie et al. 2009). Most of the above cited studies on the relationship between the WPSPH and SST as well as the connection with East Asian climate focus on variations on the interannual to decadal time scale. The relationship between the WPSPH and SST on a subseasonal time scale, accompanied by rainfall variations over East China, is poorly understood. Particularly, little attention has been paid to the link of the subseasonal changes of the SST with the zonal movement of the WPSPH. The subseasonal SST fluctuations are detectable over the Indian monsoon region (Vecchi and Harrison 2002; Krishnamurthy and Kirtman 2009), the SCS–western Pacific (Wu et al. 2008; Li et al. 2009; Wu 2010; Roxy and Tanimoto 2012; Roxy et al. 2013), and the Kuroshio Extension (Wang et al. 2012) regions in summer. Results from both observations and simulation indicate that the subseasonal variation of the SST and the related air–sea interaction are nonnegligible factors in monsoon variability; thus, they are crucial for the model to reasonably reproduce the low-frequency oscillation of monsoon precipitation (Vecchi and Harrison 2002; Lin et al. 2008; Krishnamurthy and Kirtman 2009; Wu et al. 2008; Fu and Wang 2004; Fu et al. 2008; Duan et al. 2008; Wu 2010; Roxy and Tanimoto 2012; Roxy et al. 2013). The first purpose of this study is to explore the subseasonal signals of the SST linked with the zonal movement of the WPSPH, accompanied by the PHR events over the MLYRV.

The second goal is to document the role of the subseasonal SST changes on the zonal movement of the WPSPH. Previous studies showed that the low-frequency
zonal oscillation of the WPSH can be attributed to the thermodynamic processes, such as the circulation–convection interaction (Hsu and Weng 2001), propagation of subseasonal oscillation in the lower and upper troposphere (Tao and Wei 2009; Yang et al. 2010), or the possible upscale feedback from synoptic-scale variability to the intraseasonal oscillation (Zhou and Li 2010; Hsu and Li 2011). These studies explained well the westward propagation of the WPSH over the western North Pacific. Accompanied by subseasonal rainfall oscillation over the MLYRV in summer, the WPSH first migrates westward, then retreats eastward after reaching the westernmost spot (Wang et al. 2000b; Mao and Xu 2006; Mao et al. 2010). The abrupt eastward movement during the second phase of the WPSH zonal oscillation has received less attention, compared with the westward stretch of the WPSH. We speculate that the subseasonal SST variations and the related air–sea interaction may play a role in the eastward retreat of the WPSH. On one hand, the local wind–evaporation and cloud–radiation processes linked with the WPSH zonal movement could cause the subseasonal variation of the SST. On the other hand, the changed SSTs could feed back on the atmosphere circulation and rainfall via a process such as modifying the lower-level instability (Wu 2010; Roxy and Tanimoto 2012; Wang et al. 2012).

To summarize, the issues to be addressed in this study include 1) the local WPSH–SST relationship on a subseasonal time scale over the western North Pacific accompanied by the PHR events over the MLYRV, and 2) the possible role of subseasonal SST variation and the related air–sea interaction in the WPSH zonal oscillations. The paper is organized as follows: data sources and analysis techniques are described in the next section. Section 3 presents the subseasonal WPSH variations accompanied by the PHR events over the MLYRV. Section 4 investigates the local WPSH–SST relationship and air–sea interaction processes and the possible influence of subseasonal SST changes on the WPSH zonal shift. Finally, conclusions and discussion are provided in section 5.

2. Data and analysis methods

a. Data

The datasets used in this study include 1) daily atmospheric reanalysis for the period 1979–2011 from the National Centers for Environmental Prediction–U.S. Department of Energy (NCEP–DOE) Atmospheric Model Intercomparison Project (AMIP)-II Reanalysis (Kanamitsu et al. 2002). Variables include surface latent and sensible heat fluxes, surface net shortwave radiation, surface air temperature and humidity at 2-m height, and surface winds at 10-m height. These surface variables are on T62 Gaussian grids. In the dataset upward surface heat fluxes are denoted as positive. The tropospheric elements are on a regular $2.5^\circ \times 2.5^\circ$ latitude–longitude grid. 2) A quality controlled dataset of daily precipitation at 756 gauge stations provided by the Climate Data Center, China Meteorological Administration (CMA) is used. These data are widely used to examine the subseasonal features of rainfall in East China (Tang et al. 2006; Mao and Xu 2006; Liu et al. 2008; Mao et al. 2010; Yuan et al. 2010). Among those stations, 53 are located in the MLYRV after removing the stations that have more than 3 days of continuous missing data during the summer months [June–August (JJA)] of 1979–2011 (see Fig. 1c for station’s locations). 3) A daily gridded precipitation dataset is used from the Asian Precipitation—Highly Resolved Observational Data Integration Towards Evaluation of Water Resources (APHRODITE; http://www.chikyu.ac.jp/precip/) project, conducted by the Research Institute for Humanity and Nature (RIHN) and the Meteorological Research Institute of the Japan Meteorological Agency (MRI/JMA) since 2006. The dataset is created primarily with data obtained from a rain gauge observation network and is intended to accurately represent both the mean and extreme values. We used the APHRO_V1101 dataset that covers the region of $15.0^\circ \text{S}–55.0^\circ \text{N}, 60.0^\circ–150.0^\circ \text{E}$ at a resolution of $0.5^\circ \times 0.5^\circ$ for 1979–2007. Further details of the dataset are described in Xie et al. (2007) and Yatagai et al. (2009, 2012). 4) In addition, daily SST is used from version 2 of the National Oceanic and Atmospheric Administration (NOAA) optimal interpolation (OI) SST (OISST) dataset based on Advanced Very High Resolution Radiometer (AVHRR) satellite data. The new OISST dataset is available daily on a $0.25^\circ \times 0.25^\circ$ grid for the period of 1982–2011 and is an improvement over the older weekly OISST dataset (Reynolds et al. 2002). 5) Finally, daily mean outgoing longwave radiation (OLR) data on a $2.5^\circ \times 2.5^\circ$ grid from NOAA is used (Liebmann and Smith 1996).

b. Analysis method

The methodology adopted by Krishnamurthy and Shukla (2000) is used to investigate the subseasonal oscillation signals. It begins with removing very high-frequency fluctuations in the daily data by applying a 5-day running mean then deducing the daily climatology. After removing the interannual signal linked with ENSO by subtracting seasonal anomaly, the subseasonal daily data are obtained and used for the analysis.

The criteria for a PHR event are identified by intensity, duration, and extent region of rainfall. 1) For intensity, when the rainfall $p \geq 25$ (or 50) mm day$^{-1}$ is
found at a particular gauge station, it is called a single-station PHR event. Since CMA categorizes daily precipitation of 50 ≤ p < 100 mm as heavier rain, 25 ≤ p < 50 mm as heavy rain, and 10 ≤ p < 25 mm as moderate rain, the intensity criterion of 25 (50) mm day−1 is considered moderate-to-heavy rainfall for an individual station. 2) For duration, a single-station PHR event over the MLYRV generally has a duration of ≳3 days (Tang et al. 2006; Bao 2007). 3) For extent, a region is defined where a PHR event occurs covering multiple stations. The intensity criterion can be lower than 25 (50) mm day−1 for an area-averaged PHR event.

Figure 1 shows the distribution of the total number for single-station PHR events over eastern China during the summer of 1979–2011 with the intensity criterion set as 30 and 15 mm day−1, respectively, both with a duration of ≳3 days. The climatological summer (JJA) daily rainfall is shown in Fig. 1a as a background. The climatological rainfall amount increases from northern to southern China. To the south of 35°N, there are three regions with rainfall amount ≥7 mm day−1: the MLYRV, the Sichuan basin, and southern China. Abundant single-station PHR events occur over the MLYRV and the region south of 25°N. The ratio of single-station PHR events to the total summer precipitation is about 25% over the MLYRV for the intensity criterion of 15 mm day−1. The red dots in Fig. 1c show the locations of the 53 rain gauge stations used in this study. To identify the PHR events over the MLYRV, the subseasonal rainfall anomalies for the summers of
1979–2011, averaged for 53 stations, were first obtained. The time series were then normalized to unit standard deviation and referred to as the subseasonal rainfall anomaly index (RI). Figure 2 displays time series of the daily rainfall and RI over the MLYRV for the summers of 1991 and 2009. A PHR event (also referred to as active events in this study) is identified when the RI exceeds its one standard deviation in at least three consecutive days. Additionally, a break event is identified when the RI is below minus one standard deviation in at least three successive days, in contrast with the PHR events. During the summers of 1979–2011, 76 PHR events and 45 break events are identified with the above criteria, with the durations ranging from 3 to 6 days for most of the events and longer than 10 days for a few extreme ones.

The time-lagged composition is used in this study. The simultaneous composite (lag = 0) of the variables with respect to rainfall events is constructed by averaging the variables for all the days of JJA 1979–2011 that satisfy the definition for the PHR events over the MLYRV. The lag = −n (lag = n) composite indicates that the variables lead (lag) the PHR events by n days and are obtained by shifting backward (forward) the number of leading (lag) days. In comparison, the simultaneous composition of break events is also carried out in this study in a similar way.

3. Zonal oscillation of the WPSH related to PHR and break events

Figures 3a and 3b show the simultaneous composites of actual daily rainfall anomaly over eastern China from

CMA for 76 PHR (active) and 45 break events, including the 588 and 149–151 geopotential dekameter (gpdam) contour lines from the 500- and 850-hPa geopotential height fields (Z500 and Z850), respectively. For active events accompanied by above-normal precipitation over the MLYRV, the western edge of the WPSH, represented by the western part of the 588 gpdam contour line (hereafter 588-line), extends anomalously westward reaching 115°E. In contrast, for break events the 588-line just slightly crosses 130°E, with anomalous drought over the MLYRV. The 149 and 151 gpdam contour lines (hereafter 149–151-line) at 850 hPa show similar differences between two-type events as the 588-line but with a slightly weaker zonal span. Figures 3c and 3d show that the APHRODITE dataset is nearly identical to the CMA data over eastern China. Moreover, the APHRODITE data shows the above-normal precipitation over southern Japan accompanied by PHR over MLYRV and drought condition over southern Japan and the southern Korean Peninsula with break events. The composite of subseasonal OLR anomalies for active events (Fig. 3c) shows that convection is enhanced over the East Asian summer monsoon region and is suppressed over the off-equatorial warm pool. The convection anomaly for break events (Fig. 3d) is almost the mirror image of that for active ones.

The simultaneous composites of subseasonal Z500 and 850-hPa wind anomalies are plotted in Fig. 4, overlaid with the 588, 585, and 582 gpdam contour lines. Associated with the PHR over the MLYRV (Fig. 4a), the subseasonal Z500 anomalies exhibit a north–south
dipole structure with negative (positive) values north (south) of the 585 gpdam contour. The amplitudes of both negative and positive lobes are significantly large over East Asia and its coastal waters. The positive Z500 anomalies and the anomalous anticyclone at 850 hPa beneath the western edge of 588-line indicate the westward extension of the WPSH. The anomalous cyclone at 850 hPa over the north lobe brings the cold air mass from the mid- and high latitudes. This southward flow is confluent with the anomalous northward moisture transport related to the westward extension of the WPSH. Therefore, the convergence over the MLYRV is strengthened, resulting in a condition favorable for enhanced precipitation over the Yangtze River valley. The patterns of
anomalous subseasonal Z500 and low-level wind for break events (Fig. 4b) are in sharp contrast to those for active events. The negative Z500 anomalies and the anomalous cyclone at 850 hPa west of the 588-line suggest a more eastward position of WPSH, leading to the reduced moisture supply to the MLYRV.

It is well known that climatologically, a northward jump of the WPSH occurs over the western North Pacific in mid-July (Tao and Chen 1987; Lu 2001; Ding and Chan 2005; Yang et al. 2013). The variability of summer rainfall over the MLYRV shows different characteristics before and after mid-July associate with remarkable differences in the state of mean circulation (Ding and Chan 2005; Wang et al. 2009). Thus, we further divide the events into two groups [early summer (1 June–15 July) and late summer (16 July–31 August)]. We get 47 (38) active (break) events for the early summer group and 29 (7) active (break) events for the late summer group. Figure 5 plots the lagged composites of the 588-line at Z500 and the 149–151-line at Z850 for the early summer group for the two-type events. The lags show range from −12 to 12 days with an interval of 3 days. The zonal oscillation of the WPSH at Z500 is the dominant feature associated with the PHR evolution over the MLYRV (Figs. 5a,b). Prior to the continuous rainfall over the MLYRV, the west point of the 588-line is located near 126°E at lag = −12 days; it stretches westward slightly at lag = −9 days. The western edge passes southern Taiwan with an accelerating speed when the lag sets smaller and finally reaches the westernmost location at lag = 0 days. After that, the western part of the WPSH begins to withdraw eastward and is over the western Pacific at lag = 6 days. Associated with the PHR evolution over the MLYRV, the WPSH at Z850 is also anchored by the zonal movement (first westward then eastward) denoted by both the 149 and 151 contour lines.

![Simultaneous composite of subseasonal fields of wind anomaly at 850 hPa (m s⁻¹, vectors) and Z500 anomaly (gpm, shaded and black contours) superimposed on the 588, 585, and 582 gpdam contour lines (red contours). The dots (vectors) in figures denote the areas where the simultaneous correlation coefficients between the RI and Z500 (850-hPa wind field) anomaly are significant at the 0.05 confidence level.](http://journals.ametsoc.org/jcli/article-pdf/26/22/8929/4012762/jcli-d-12-00861_1.pdf)
The lagged composites of the 588-line (149–151-line) for break events shown in Figs. 5c,d (Figs. 5g,h) also present zonal oscillation but with the zonal migration first eastward then westward and with a smaller span. Thus, the zonal changes of the WPSH at the 500- and 850-hPa levels exhibit mirror images between active and break events for the early summer group.

For the late summer group, the seven break events during 1979–2011 produce random results for composites (figures not shown). Therefore, the following analyses are devoted to the 29 active events. The lagged composite of the 588-line and the 149–151-line are shown in Fig. 6. Compared with the active events in the early summer group, the WPSH at 500 hPa moves
northward slightly, while the WPSH main body is still located south of 30°N during the Yangtze PHR period. The 151 contour line at 850 hPa for the late summer group retreats eastward and is far away from the coastline. In general, the 588 and 149 contour lines shown in Fig. 6 still demonstrate the features of zonal shift of first westward then eastward movement.

To further illustrate the predominant signal shown in Figs. 5 and 6 for active events, Fig. 7 displays Hovmöller plots of the subseasonal anomalies of Z500, Z850, and OLR for active events averaged over 110°–140°E for the two groups and the whole summer. All panels in Fig. 7 show that south of 30°N, positive anomalies of Z500 and Z850 are associated with the suppressed convection.
prior and during the Yangtze PHR period, followed by negative Z500 and Z850 anomalies and intensified local convection. Thus, the zonal oscillation of the WPSH at 500 and 850 hPa is dominant during the whole summer for the active events.

4. The subseasonal variation of SSTA in coastal waters and the WPSH zonal oscillation

The local subseasonal signal of the SSTA cooperates with the PHR and the WPSH variations. Figure 8a displays the simultaneous composite of the SSTA in the western Pacific for active events during the whole summers of 1982–2011. It shows the robust subseasonal SSTA signal in the northwest Pacific. Accompanied by the PHR over the MLYRV, the SSTAs are characterized by a north–south dipole structure with negative (positive) SSTA north (south) of the northern edge of the WPSH, bearing some resemblance to the Z500 anomalies shown in Fig. 4a. Figure 8b shows the contrast of break events during early summer. The SSTA for break events is almost the mirror image of that for the PHR, except that the positive SSTAs are everywhere north of the 588-line as in Fig. 8b, while the negative
SSTAs are scattered north of the 588-line in Fig. 8a. The SSTAs also exist south of the 588-line but with relatively weaker amplitude.

Noticeably, two regions of positive SSTAs are found beneath the 588-line in Fig. 8a, over the western and eastern part of WPSH, respectively. The SSTAs beneath the eastern region are located far away from the western edge of the WPSH. Thus, in the following, we investigate the relationship between the zonal variations of the WPSH and coastal SSTAs for active events. The characteristic patterns for break events are almost the mirror image of that for active ones.

### a. Evolution of subseasonal SSTA and Z500 anomaly

To examine the evolution of the subseasonal SST variations over the coastal water and the WPSH zonal oscillation related to the Yangtze rainfall events, we make Hovmöller plots of subseasonal anomalies of SST and Z500 averaged over 120°–140°E for active events of the whole summer (Fig. 9a). In the following, we focus on the region south of 30°N where it is beneath the western part of the 588-line in Fig. 8a. The positive Z500 anomaly south of 30°N occurs incipiently around lag = −10 days, corresponding to the anomalous westward extension of the WPSH. Its amplitude increases gradually on the approach of the Yangtze rainfall events and is at its largest when continuous heavy rainfall begins over the MLYRV. The positive Z500 anomaly becomes weaker after lag = 0, indicating the beginning of the eastward retreat of the WPSH. The Z500 anomaly switches to the negative at lag = 5 days, coherent with the end of the PHR event over the MLYRV. It reaches the minimum at about lag = 10 days, corresponding to the end of the WPSH retreat. The subseasonal evolution of local SSTA appears with a phase difference with respect to the Z500 anomalies. Prior to the westward extension of the WPSH, negative SSTAs cover the East Asian coastal waters south of 30°N. Several days later, when the WPSH is on its midway of westward stretch, the positive SSTA begins to develop. The 0.2°–0.3°C positive anomalies maintain for about 5 days, accompanied by the eastward withdrawal of WPSH. After that, the positive SSTA decreases gradually, accompanied by the WPSH retreating back to the western North Pacific. The relationship between the Z500 anomaly and the coastal SST north of 30°N undergoes the opposite process to that south of 30°N. Strong lower-level cyclonic anomalies cover the coastal waters north of 30°N, accompanied by the PHR events over the MLYRV that may contribute to the development of local SST cooling.

The phase difference between subseasonal SSTA and the WPSH zonal movement as well as the OLR anomalies south of 30°N is further demonstrated in Fig. 9b. Over the East Asian coastal water region, the positive Z500 anomaly and suppressed convection are followed by an increase of SST, implying that the anomalous westward extension of the WPSH leads positive SSTA by several days. After lag = 0, the negative Z500 anomaly and strengthened local convection are accompanied by small but persistent positive SSTA. The above relationship between subseasonal WPSH and SST can be clearly seen by the profiles of lag correlation between

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**Fig. 9.** (a) Hovmöller plots of subseasonal anomalies of SST (K, shaded) and Z500 (gpm, contours) for active events over 120°–140°E. (b) Composites of subseasonal anomalies of Z500 (gpm, purple line), OLR (W m⁻², green line), and SST (K, black line) for active events averaged over 15°–25°N, 120°–140°E. The scales on the left side in (b) are from left to right for Z500 (and OLR) and SST, respectively. Negative (positive) lags on the abscissa (day) indicate SST, Z500, and OLR anomalies leading (lagging) RI. Lag = 0 corresponds to simultaneous composites of subseasonal anomalies.
them (Fig. 10). A positive correlation is obtained when the Z500 and Z850 anomalies lead the SSTAs, indicating that the atmosphere is driving the SST. In addition, a negative correlation is produced when the Z500 and Z850 anomalies lag the SSTAs, indicating that the SST also has an impact on the atmosphere.

b. Role of anomalous westward extension of WPSH on SST changes

The anomalous westward extension of the WPSH has a crucial role in the generation of positive SSTAs shown in Fig. 9, via radiation and heat fluxes at the air–sea interface. Meanwhile, physical processes in the upper ocean may take partial responsibility. According to previous studies, the former plays a dominant role on a subseasonal time scale over the monsoon region (Wu 2010; Roxy and Tanimoto 2012). Therefore, we focus on examining the radiation and heat fluxes. Figure 11 displays the composites of the net solar radiation and the OLR, latent and sensible heat fluxes at the air–sea interface, 10-m wind speed, and 2-m specific humidity, averaged over 15°–25°N, 120°–140°E for all the 76 active cases. The OLR and net solar radiation anomalies increase gradually to above normal in association with the westward extension of the 588-line from lag = −52 days to lag = 50, suggesting suppressed convective activity and more of the incident solar energy. Meanwhile, the latent heat flux switches from small positive (15 W m⁻² at lag = −11 days) to relatively large negative (−17 W m⁻² at lag = −2 days) resulting from the decreased lower-level wind speed during the westward extension of the WPSH. This means that the ocean releases less latent heat to the atmosphere. The changes of sensible heat flux are negligible compared with that of latent heat flux. Thus, the anomalies of both the solar radiation and latent heat release accompanied by the westward extension of the WPSH contribute to the increase of local SST. After the Yangtze rainfall events, the incident solar radiation over the coastal region becomes weaker and the latent heat release goes back to positive anomaly; hence, the positive SSTAs become indistinct.

c. Possible role of SSTA on the WPSH eastward retreat

The above analyses show that the subseasonal increase of SST in the coastal waters results from local atmospheric forcing (i.e., the anomalously westward extension of the WPSH and the related air–sea interaction). The detectable positive SSTAs persist more than 10 days. Therefore, the SSTA may feed back on the atmospheric circulation in the mid–lower levels and
show impact on the WPSH. Previous investigations suggested that the SSTA can play a role in the atmospheric variability in several ways. For example, the SSTA can modify the convective instability via changing the air temperature and humidity in the lower level (Wu 2010; Roxy and Tanimoto 2012; Wang et al. 2012). The SSTA may also influence the lower-level wind field via a Rossby wave type response to the anomalous heating (Hsu and Weng 2001; Wu 2010). In the following, we attempt to analyze the above two processes.

The vertical difference in atmospheric equivalent potential temperature $\theta_e$ between 1000 and 850 hPa $\Delta \theta_e = \theta_{e1000} - \theta_{e850}$ is used to estimate the atmospheric convective instability (Roxy and Tanimoto 2007, 2012). The atmosphere is in a convectively unstable condition when this difference is positive, thus being favorable to the local convection-related activities. Figure 12 exhibits composites of subseasonal anomalies of $\theta_{e1000}$ and $\Delta \theta_e$, averaged over 120°–140°E, 15°–25°N for 76 active events. The $\Delta \theta_e$ anomaly is negative prior to the day of the peak SST, indicating a convectively stable condition with suppressed convective activity. After lag = 2 days, the atmosphere switches to an unstable state indicated by positive anomalies of $\Delta \theta_e$. Unstable conditions are at their maximum at lag = 8 and then gradually evolves to a stable one. Such an evolution of $\Delta \theta_e$ anomalies also corresponds to that of OLR anomalies shown in Fig. 11a. The evolution of the $\Delta \theta_e$ anomaly bears a strong resemblance to that of $\theta_e_{1000}$. Therefore, the convective unstable situation is triggered from the lower level. The larger $\theta_e_{1000}$ anomaly after lag = 0 is a result of the extra supply of moisture to the lower atmosphere related to the positive anomalies of latent heat release after the increase of SST (see Figs. 11b,c). This process results in a moistened and destabilized atmosphere in the lower level that provides an unfavorable condition for maintaining anticyclonic anomalies in the mid–lower levels.

The SSTA can induce anomalous low-level cyclonic/anticyclonic circulation in a way shown in the theoretical study by Gill (1980) in which cyclonic flow is obtained that forms on the western margins of the heating zone. Prior to the examining the process, we first show in Fig. 13 the Hovmöller plot of subseasonal anomalies of 850-hPa relative vorticity $\zeta_{850}$ and wind fields for 76 active events. Figure 13 shows that the anomalous anticyclone and positive occupy the 10°–30°N band before lag = 2 days and then rapidly change to the anomalous cyclone and positive $\zeta_{850}$ after lag = 4 days. These evolutions are consistent with the zonal variations of the WPSH for active events.

We demonstrate from the following results that the persistently positive SSTA in the coastal waters in the PHR events has a strong link with the rapid transition of the anomalous $\zeta_{850}$ shown in Fig. 13. To investigate the relation between the transition of anomalous $\zeta_{850}$ and the SSTA, we composite the tendency of subseasonal anomalous $\zeta_{850}$ and wind fields at 850 hPa for 76 active events. The tendency is obtained by their differences at 2–6 days. The temporal evolution shown in Fig. 14 illustrates the evolution of these tendencies overlapped on subseasonal SSTA and the 588-line over East Asian coastal waters. At lag = 0, the positive tendency of the relative vorticity anomaly $\Delta \zeta_{850}$ is detectable over the warm SSTA. From lag = 2–6 days, the large positive $\Delta \zeta_{850}$ is located slightly at the northwest flank of the warm SSTA, just beneath the western edge of the WPSH. A weak negative one is located to the south of the positive $\Delta \zeta_{850}$ region. The tendency of wind field anomaly $\Delta \mathbf{V}_{850}$ is connected with $\Delta \zeta_{850}$; namely, the positive $\Delta \zeta_{850}$ corresponding to the cyclonic $\Delta \mathbf{V}_{850}$. The temporal evolution shown in Fig. 14 evidently reflects a close conjunction of positive SSTA with the transition of anomalous $\zeta_{850}$ and circulation. The positive SSTA may induce the anomalous cyclonic
circulation in a way suggested by the theoretical study of Gill (1980), which can diminish the preexistent anomalous anticyclonic circulation. As a result, the WPSH moves eastward accompanied by a negative to positive transition of anomalous $\zeta_{850}$ (as well as lower-level wind fields) shown in Fig. 13. This result implies a negative feedback of the SSTA to the WPSH, as suggested by the lag correlation shown in Fig. 10. After lag = 8 days, the positive SSTA fades away and the feedback effect becomes weak.

The subseasonal SSTAs also occur in the midlatitude North Pacific shown in Fig. 8. The anomalous signals

Fig. 14. Composites of the 588-line, subseasonal SSTA (K, colors), and the tendencies of subseasonal anomalous $\zeta_{850}$ ($10^{-6}$ m s$^{-1}$, black contours) and 850-hPa wind fields (m s$^{-1}$, vectors) for active events over East Asian coastal waters. (a)–(f) Lag = −2, 0, 2, 4, 6, and 8 days, respectively. The lag day for SSTA and the 588-line is overlaid in each panel. The lag day for tendency is obtained by 2 days later minus the current.
span the northern edge of the 588-line and show the opposite sign south and north of the northern edge of the 588-line. The subseasonal SSTA in this region appears certain in connection with the northern edge line of the WPSH but has a weaker relationship with the zonal movement of the western edge of the WPSH (figures not shown).

5. Conclusions and discussion

a. Conclusions

This study addresses the local relationship between the WPSH and SST over the western North Pacific on a subseasonal time scale, associated with the PHR events over the MLYRV. The lead–lag composite analyses are carried out to characterize the influence of the westward extension of the WPSH’s western edge on the evolution of the SSTA, and the possible role played by the positive SSTA in the eastward retreat of the WPSH. The datasets include atmospheric reanalysis products, observed daily SST, daily OLR, and a station-based daily rainfall dataset.

Based on station daily rainfall data in China during summers of 1979–2011, a total of 76 PHR (active) events are identified, using the criteria that the normalized subseasonal daily rainfall anomalies over the MLYRV exceed one standard deviation in at least 3 consecutive days. In parallel, 45 break events that mean a continuous dry period, with little precipitation, are chosen with the opposite intensity criterion. The simultaneous composites of subseasonal Z500 anomalies and SSTAs over the western North Pacific corresponding to the PHR events demonstrate dipolar structures with positive (negative) values south (north) of the WPSH’s northern edge. Accompanied by the westward extension of the WPSH, positive Z500 anomalies and the anomalous anticyclone at 850 hPa are located over East Asia and its coastal region south of 30°N. Meanwhile, the warmer waters are beneath the WPSH western edge. During break events, the WPSH is located more eastward with opposite anomalous circulation and SSTA signal compared with the active events.

We further divide the events into the early summer and the late summer groups. Compared with active events in the early summer group, the WPSH at 500 hPa for active events in the late summer group moves slightly northward; however, the WPSH main body is still located south of 30°N. The lagged composites of the 588-line at Z500 and 149–151-line at Z850 for active events suggest that the western edge of the WPSH exhibits zonal migration on a subseasonal time scale. In addition, the zonal changes of the WPSH for active and break events in the early summer group exhibit mirror images with respective to each other.

The above subseasonal evolution of the WPSH has two distinct phases. The first stage is the westward propagation of its western edge and anomalous low-level anticyclonic circulation, prevailing the coastal region prior and during the PHR events over the MLYRV. This stage ends when the WPSH western edge reaches the coastlines of southeastern China at the peak of PHR events. The second stage is the eastward withdrawal of the WPSH and the transition of the lower-level circulation from anticyclonic to cyclonic anomalies. The subseasonal variations of the coastal SST display phase difference with respect to the WPSH zonal shift. Before the WPSH stretching westward, negative SSTAs cover the East Asian coastal waters south of 30°N. An increase of SST occurs several days after the westward stretch of the WPSH. The positive SSTAs persist for several days accompanied by the eastward withdrawal of the WPSH, and then they decrease gradually.

The westward extension of the WPSH has a crucial role in the generation of positive SSTAs via the above-normal incident solar radiation and the below-normal latent heat release at the air–sea interface. The positive SSTAs last about 10 days and act to impact the local circulation in the mid–lower levels. It destabilizes the lower atmosphere via extra moisture supply to the lower atmosphere. This forms an unfavorable condition for maintaining the local anticyclonic anomaly. The persistent positive SSTAs also have a close conjunction with the transition of lower-level circulation from the anomalous anticyclone to the cyclone over the coastal region, just beneath the western edge of the WPSH. As a result, the preexisted anomalous anticyclonic circulation is diminished and the WPSH contracts eastward. These results suggest that the positive SSTAs forced by the westward extension of the WPSH have a negative feedback on the WPSH and may contribute to the eastward retreat of the WPSH. Therefore, the local air–sea interaction in the western North Pacific on a subseasonal time scale may be one of the causes for the zonal oscillation of the WPSH on a subseasonal time scale. The subseasonal fluctuations of the WPSH and SST for break events are almost mirror images of fluctuations of the active events. The negative SSTAs are induced in a way almost opposite to the generation of the positive SSTA for active events.

b. Discussion

It is known that a better understanding of the regional or global low-frequency oscillation in the atmosphere would be helpful to the improvement of seamless weather/climate prediction (Brunet et al. 2010; Shapiro...
et al. 2010). Variation with the period ranging from 10 to 30 days is one of the dominant modes of low-frequency oscillation for the summer monsoon rainfall over the MLYRV and is significantly modulated by atmospheric subseasonal oscillations (Yang et al. 2010). The robust lead–lag relation between the WPSH zonal movement and the SSTAs as revealed by this study suggests the connection of SST with the atmospheric oscillation over the East Asian coastal region on a subseasonal time scale. The induced increase of SST may play a certain role in the WPSH zonal movement, especially during the WPSH retreat phase. Such local air–sea interactions demonstrate an air–sea coupled system on a subseasonal time scale. Our study suggests that the subseasonal SSTAs can feed back to the local atmospheric oscillation. The maximum positive SSTA leads the maximum negative Z500 anomaly by about at least 5 days, as shown in Figs. 9 and 10. This result might be considered to be one of the potential premonitory signals of the eastward retreat of the WPSH and the ending of the PHR over the MLYRV. Further research is needed to perform a suitable prediction scheme to test the possibility. The WPSH–SST relationship may also exist over the northern edge of the WPSH accompanied by the Yangtze River PHR events, as shown in Fig. 8. This region is near the midlatitude North Pacific and is far from the western edge of the WPSH. Thus, the WPSH–SST relationship could be distinct from that over the coastal waters. Further studies are needed to explore the related processes.

Acknowledgments. This work was jointly supported by the 973 Program (Grant 2012CB417203) and the Special Fund for the Meteorological Scientific Research of Public Sector (Grants GYHY200806004 and GYHY201106017). The authors thank the anonymous reviewers for their valuable comments and suggestions and Prof. Youyu Lu for helpful discussion. The NCEP–DOE AMIP-II Reanalysis data is provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado (from their website at http://www.esrl.noaa.gov/psd/).

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