How can anomalous western North Pacific Subtropical High intensify in late summer?

Baoqiang Xiang,[1] Bin Wang,[1,2] Weidong Yu,[3] and Shibin Xu[1,4]

Received 22 February 2013; revised 26 March 2013; accepted 28 March 2013; published 28 May 2013.

[1] The western North Pacific (WNP) Subtropical High (WNPSH) is a controlling system for East Asian Summer monsoon and tropical storm activities, whereas what maintains the anomalous summertime WNPSH has been a long-standing riddle. Here we demonstrate that the local convection-wind-evaporation-SST (CWES) feedback relying on both mean flows and mean precipitation is key in maintaining the WNPSH, while the remote forcing from the development of the El Niño/Southern Oscillation is secondary. Strikingly, the majority of strong WNPSH cases exhibit anomalous intensification in late summer (August), which is dominantly determined by the seasonal march of the mean state. That is, enhanced mean precipitation associated with strong WNP monsoon trough in late summer makes atmospheric response much more sensitive to local SST forcing than early summer. **Citation:** Xiang, B., B. Wang, W. Yu, and S. Xu (2013), How can anomalous western North Pacific Subtropical High intensify in late summer?, *Geophys. Res. Lett.*, *40*, 2349–2354, doi:10.1002/grl.50431.

1. Introduction

[2] The western North Pacific (WNP) subtropical high (WNPSH) is a primary circulation system of the East Asian-WNP summer monsoon. Investigation of the variations of WNPSH is of utmost importance for understanding the variation and prediction of the East Asian Summer monsoon (EASM) rainfall and tropical storm activities in the WNP [e.g., *Wang et al.*, 2000; *Chang et al.*, 2000; *Ding* 2007; *Zhou et al.*, 2009; *Wang et al.*, 2013].

[3] Extensive studies have been focused on the anomalous WNPSH cases associated with El Niño, and two mechanisms have been proposed to be critical in maintaining the anomalous WNPSH. The first is a local positive thermodynamic feedback between convectively coupled Rossby waves and the underlying SST cooling in the WNP with the aid of background mean flows [*Wang et al.*, 2000]. However, during the ensuing summer of El Niño, exactly how this anomalous WNPSH maintains itself against dissipation remains a fundamental but highly debated issue along with the rapid decay of the Pacific SST anomaly (SSTA). Some recent studies put forward the second mechanism by highlighting the importance of the anomalous tropical Indian Ocean (IO) warming, which may intensify the WNPSH through either the eastward propagation of Kelvin waves or an anomalous Hadley circulation [e.g., *Sui et al.*, 2007; *Xie et al.*, 2009; *Huang et al.*, 2010; *Chowdary et al.*, 2010, 2011]. Numerical experiments have been conducted to support this argument by using the linear baroclinic model (LBM) [*Xie et al.*, 2009], atmospheric general circulation model (AGCM) [*Huang et al.*, 2010], and coupled general circulation model (CGCM) [*Chowdary et al.*, 2010, 2011]. *Wu et al.* [2010] further analyzed the above two mechanisms and argued that the remote forcing effect from the IO warming plays the dominant role in late summer, and the contribution from local air-sea interaction gradually weakens from June to August because the eastward expansion of monsoon trough and the reversal of mean winds result in weak SST cooling in the WNP.

[4] Intriguingly, *Wang et al.* [2013] found that about half of the strong WNPSH cases in the last three decades do not concur with decaying El Niño or the IO warming. They identified two modes predominantly controlling the anomalous WNPSH. The first is a positive local atmosphere-ocean coupled mode between the WNPSH and the Indo-Pacific warm pool oceans, and the second is an external forced mode associated with El Niño/Southern Oscillation (ENSO) development. In the rest of this paper, we tend to examine the contrasting behaviors of these two modes in June and August to address the following three important and relevant issues: (1) What causes the anomalous intensification of WNPSH in late summer (August)? (2) Does the local air-sea feedback become weakened in late summer? (3) How do we understand the role of the IO in maintaining the anomalous WNPSH?

2. Data and Methodology

[5] Several datasets are used in this study, including (1) monthly mean SST from NOAA Extended Reconstructed SST (ERSST, v3b) [*Smith et al.*, 2008], (2) monthly mean precipitation from Global Precipitation Climatology Project (GPCP, v2.2) datasets [*Adler et al.*, 2003], and (3) monthly mean 1000 hPa wind and geopotential height (H850) from National Centers for Environmental Prediction—Department of Energy (NCEP—DOE) Reanalysis 2 products [*Kanamitsu et al.*, 2002]. In this study, the period 1979–2010 is chosen for precipitation and 1979–2011 for other data due to data availability. Summer (July–August, JJA) anomalies are calculated by the deviation of JJA mean from the long-term climatology.

[1]International Pacific Research Center, University of Hawaii at Manoa, Honolulu, Hawaii, USA.
[2]Department of Meteorology, University of Hawaii at Manoa, Honolulu, Hawaii, USA.
[3]Center for Ocean and Climate Research, First Institute of Oceanography, State Oceanic Administration, Qingdao, China.
[4]Physical Oceanography Laboratory, Ocean University of China, Qingdao, China.

Corresponding author: B. Xiang, International Pacific Research Center, University of Hawaii at Manoa, Honolulu, HI 96822, USA. (baoqiang@hawaii.edu)

0094-8276/13/10.1002/grl.50431

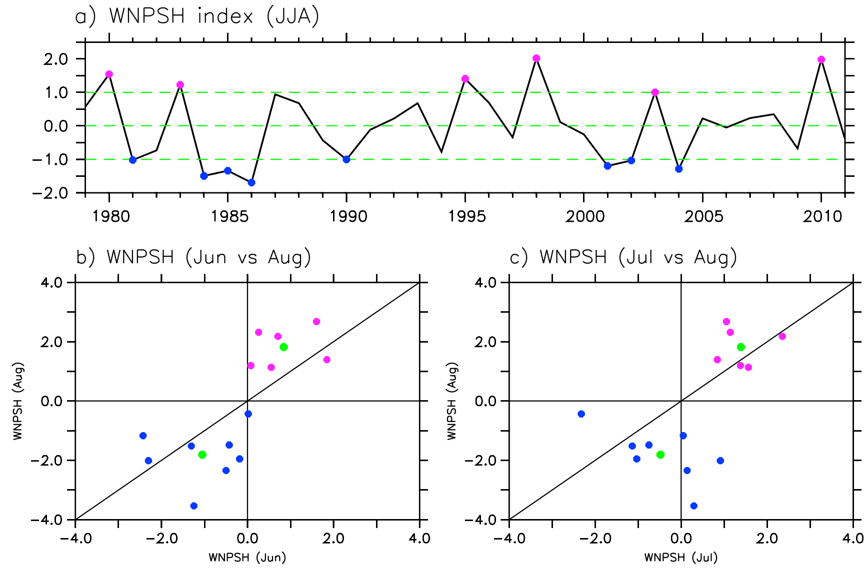

Figure 1. (a) The normalized boreal summer (JJA) WNPSH index defined by the 850 hPa geopotential height (H850) over the domain (15°N–25°N, 115°E–150°E). Six anomalous strong WNPSH and eight anomalous weak WNPSH cases are selected based on the criteria when its absolute values are greater than one standard deviation. The scatter diagram of the normalized WNPSH index between (b) June and August and (c) July and August for the six anomalous strong and eight anomalous weak WNPSH cases as shown in Figure 1a. The green dots denote the corresponding ensemble mean of these positive and negative events.

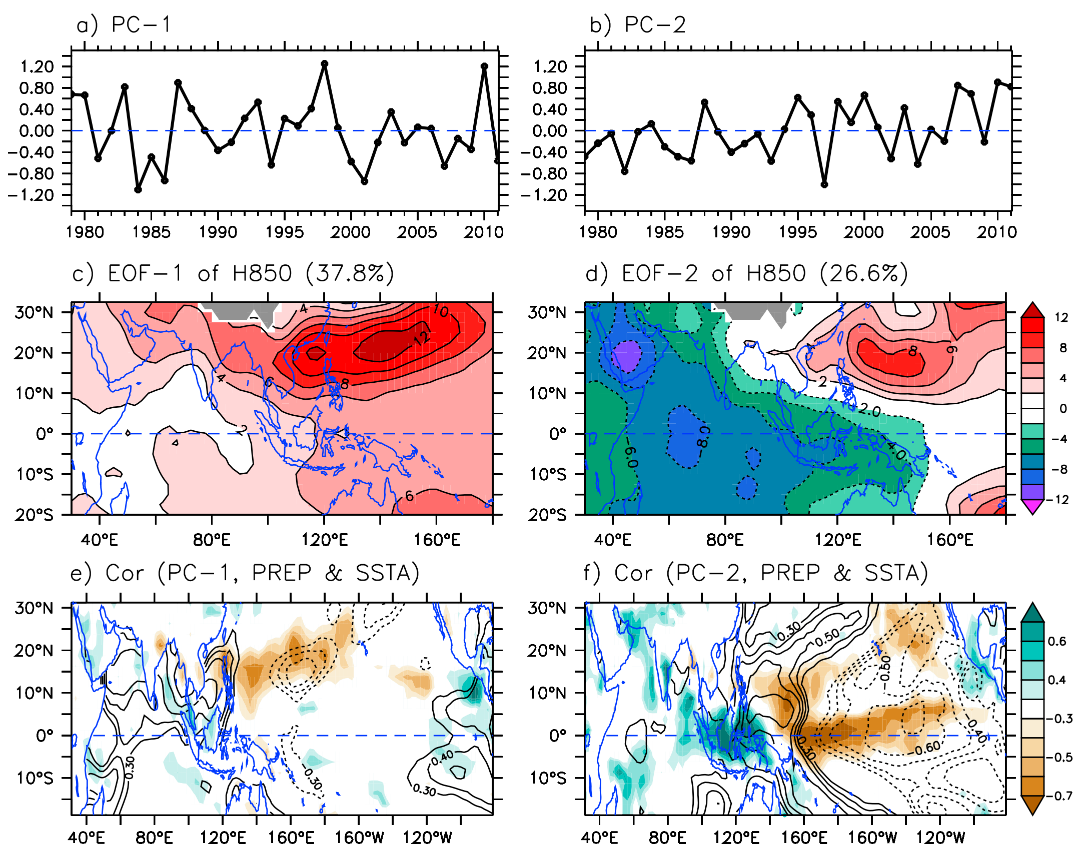

Figure 2. (a and b) The time series of the first two leading EOF modes of H850 in the Asian-Australian monsoon domain during boreal summer (JJA). (c and d) The corresponding spatial patterns of these two modes. (e and f) The simultaneous correlation of these two modes with precipitation (shading) and SSTA (contours). Only the values with confidence level above 91% are shown for the correlation coefficients.

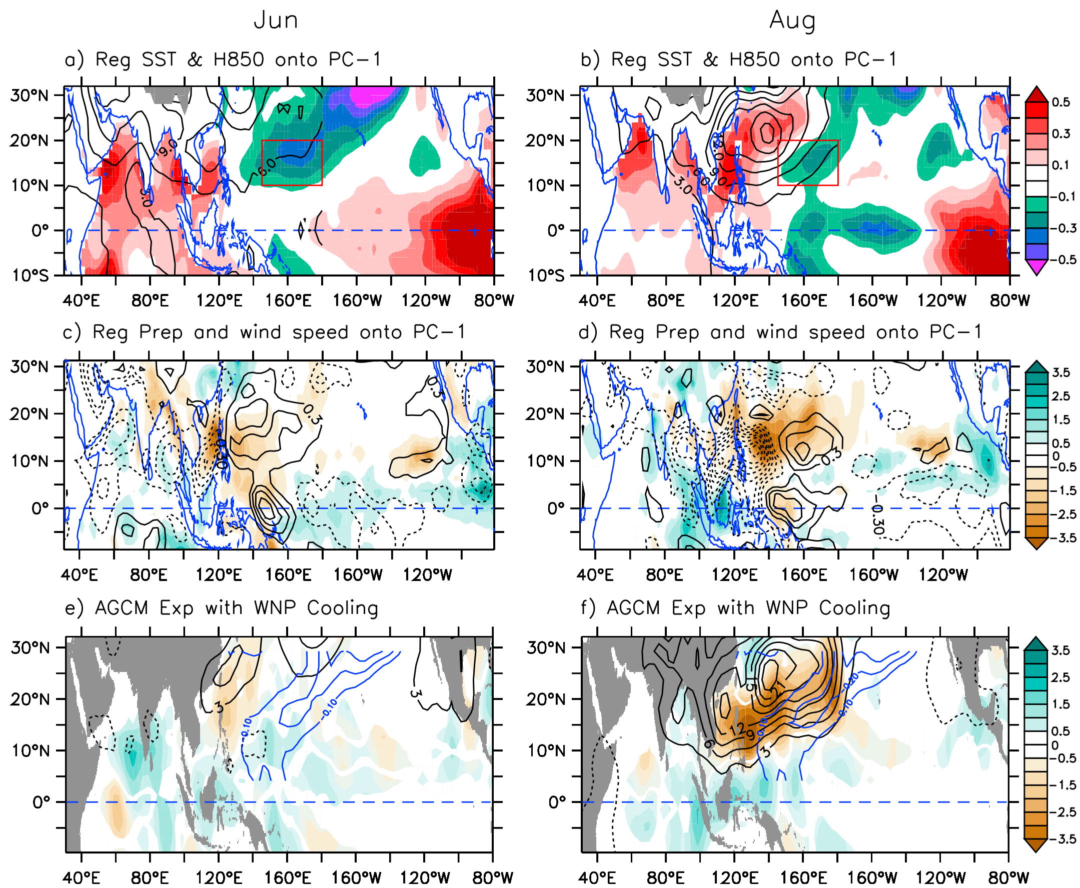

Figure 3. The regressed SSTA (shading) and H850 anomaly (contours) onto PC-1 in (a) June and in (b) August. The red boxes in Figures 3a and 3b denote the regions that are critical in maintaining the WNPSH with strong local feedback. (c and d) The charts shown are the same but for the regressed precipitation (shading) and 1000 hPa wind (contours) speed anomalies. Simulated precipitation (shading in mm/day) and H850 (black contours in m) with prescribed cold SSTA (blue contours in °C) in (e) June and in (f) August by using the ECHAM model. The prescribed SST cooling forcing in Figures 3e and 3f is based on the regressed SSTA (JJA) onto PC-1 over the region (5°N–30°N, 130°E–130°W).

[6] One AGCM ECHAM (v4.6, T42 resolution) is used here [*Roeckner et al*., 1996]. Three experiments are carried out. One is a control run forced with observed climatological SST and sea ice, and the other two sensitivity experiments are performed with imposed SSTA in the WNP and equatorial central Pacific (CP) during JJA. Each experiment has 20 ensembles.

3. Results

3.1. Observed Intensification of the Anomalous WNPSH in Late Summer

[7] Following *Wang et al*. [2013], here we use the normalized H850 averaged over (15°N–25°N, 115°E–150°E) as the WNPSH index. As shown in Figure 1a, six strong positive cases (1980, 1983, 1995, 1998, 2003, and 2010) and eight strong negative cases (1981, 1984, 1985, 1986, 1990, 2001, 2002, and 2004) are selected according to the criteria that the anomaly deviates from its mean by more than one standard deviation. Although about half of the moderate and strong positive WNPSH cases do not concur with decaying El Niño [*Wang et al*. 2013], most of the strong WNPSH cases (except 1980) occur during summer following peak phases of El Niño (Figure 1a). It is clear that the anomalous WNPSH typically has larger amplitude in August than in June and July (Figures 1b and 1c), indicating that the majority of strong WNPSH cases actually intensify in late summer. The amplitudes for the six composite positive WNPSH cases are 0.84, 1.39, and 1.82 for June, July, and August, respectively. The counterparts for the eight composite negative cases are −1.05, −0.48, and −1.81. This finding raises a critical question: How can the WNPSH anomaly amplify more vigorously in August than in June and July?

3.2. Regulation of Seasonal March of Mean State on the WNPSH

[8] To answer the above question, we made an empirical orthogonal function (EOF) analysis of JJA mean H850 in the Asian-Australian monsoon domain (20°S–32.5°N, 30°E–180°E) (Figure 2) that is similar to *Wang et al.* [2013]. The principle component of the first EOF mode (PC-1) is highly correlated with the WNPSH index, with $r = 0.86$ (Figure 2a), and PC-2 is also significantly correlated with the WNPSH index, with $r = 0.42$ (Figure 2b). Based on a multivariate regression, the WNPSH index can be well reconstructed ($1.442 \times$ PC-1 $+ 0.847 \times$ PC-2) with a high correlation coefficient ($r = 0.96$). This faithful reconstruction indicates that understanding the evolution of these two modes may help elucidate the causes for the anomalous intensification of WNPSH in late summer.

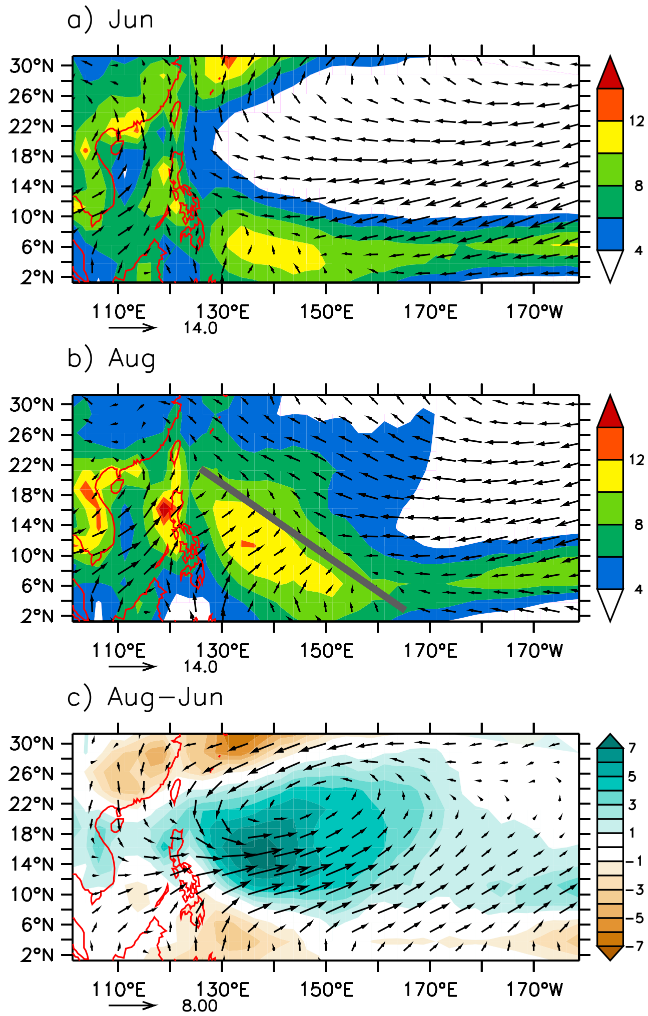

Figure 4. Mean precipitation (shading in mm/day) and 1000 hPa wind (vectors in m) in (a) June and in (b) August and (c) their difference. The gray line in Figure 4b roughly represents the WNP monsoon trough.

3.2.1. The Local Convection-Wind-Evaporation-SST Coupled Mode

[9] The leading EOF (EOF-1) mode of JJA mean H850 is characterized by a southwest-northeast oriented anomaly in the WNP, signifying an enhanced WNPSH (Figure 2c) [*Wang et al.*, 2013]. Some of the positive cases are related to El Niño decaying summer (1983, 1993, 1998, and 2010), but some are not (1979, 1980, and 1987). We present, in Figure 3, the contrasting features between June and August with respect to PC-1. In June, the regressed H850 depicts an anomalous high extending from the WNP to the Asian continent regions (Figure 3a). In August, the maximum of regressed H850 shifts to the WNP paired with weak H850 anomaly over the Asian continent (Figure 3b). The regressed SSTA features a tilted SST cooling band to the southeast flank of the anomalous WNPSH, with its amplitude decreasing from June to August (Figure 3a versus Figure 3b). In August, a contemporaneous SST warming is observed just underlying the prominent anticyclonic anomaly in the WNP (Figure 3b), accompanied by below-normal precipitation and reduced surface wind speed (Figure 3d). This supports an assertion that the SST warming in the WNP is mainly a consequence of atmospheric forcing due to more incoming solar radiation and less upward latent heat flux. Similarly, the resultant weakened surface wind speed is evident in the northern IO (Figure 3c), which is responsible for the in situ northern IO warming (Figure 3a) via reduced upward latent heat flux [*Du et al.*, 2009].

[10] Given the weak precipitation change in the northern IO, the WNP SST cooling is expected to be of central importance in triggering the WNPSH from an atmospheric point of view. We have quantified the ratio between the WNPSH index and the WNP SST cooling averaged over the domain (10°N–20°N, 145°E–180°E, red box in Figure 3), which is −30.6 for June versus −133.7 for August. It implies that the atmosphere is more sensitive to the SSTA forcing in August. How should one understand this? The cold SSTA in the WNP stimulates westward emanations of descending Rossby waves. The resultant suppressed precipitation tends to produce anomalous low-level divergence, which then feeds back to cause less convective precipitation, completing a convection-boundary divergence feedback loop. Note that the WNP mean precipitation is remarkably enhanced in August (Figure 4c), so that the atmosphere becomes more sensitive to the SST forcing as demonstrated by an AGCM model [*Xiang et al.*, 2011]. Given the same magnitude of cold SSTA in the WNP, the ECHAM model reproduces much larger precipitation and H850 anomalies that bear close resemblance to the observed counterparts (Figures 3e and 3f), confirming the importance of background mean precipitation in regulating the atmospheric response to SSTA forcing.

[11] *Wang et al.* [2013] emphasized that the WNPSH and its underlying dipolar SSTA in the WNP and IO are a highly coupled system for the EOF-1 mode, which can sustain itself through the local air-sea feedback, i.e., wind-evaporation-SST (WES) feedback that relies on background mean flows [*Xie and Philander*, 1994]. By contrast, here we suggest that the local feedback is not only relying on the background mean flows but also on the mean precipitation. The mean precipitation can influence the WES feedback via altering the anomalous convection, so we refer to this feedback as the convection-WES (CWES) feedback. The CWES feedback loop can be understood from an initial SST cooling in the WNP, which induces suppressed convection as well as an anomalous anticyclone, and then driving anomalous easterly winds to the east flank of the anticyclone and cooling the initial SST. Note that in the CWES mechanism the convection comes in play, and this is an important difference between the east Pacific WES and the warm pool CWES.

[12] It should also be recognized that the prerequisite for the existence of this EOF-1 mode is the presence of strong enough cold SSTA in early summer, considering the intense damping effect from more incoming solar radiation [*Wu et al.*, 2010]. This can account for the fact that the WNPSH can sustain itself to the ensuing late summer only for some extreme El Niño events [*Wang et al.*, 2000], for which the cold SSTA in the WNP is usually strong in late spring and early summer. Of course, we do not deny that the initial SST cooling may originate from the mid-latitudes in early summer, but the local CWES feedback holds key to a reduction in the dissipation rate of the cold SSTA to sustain the WNPSH.

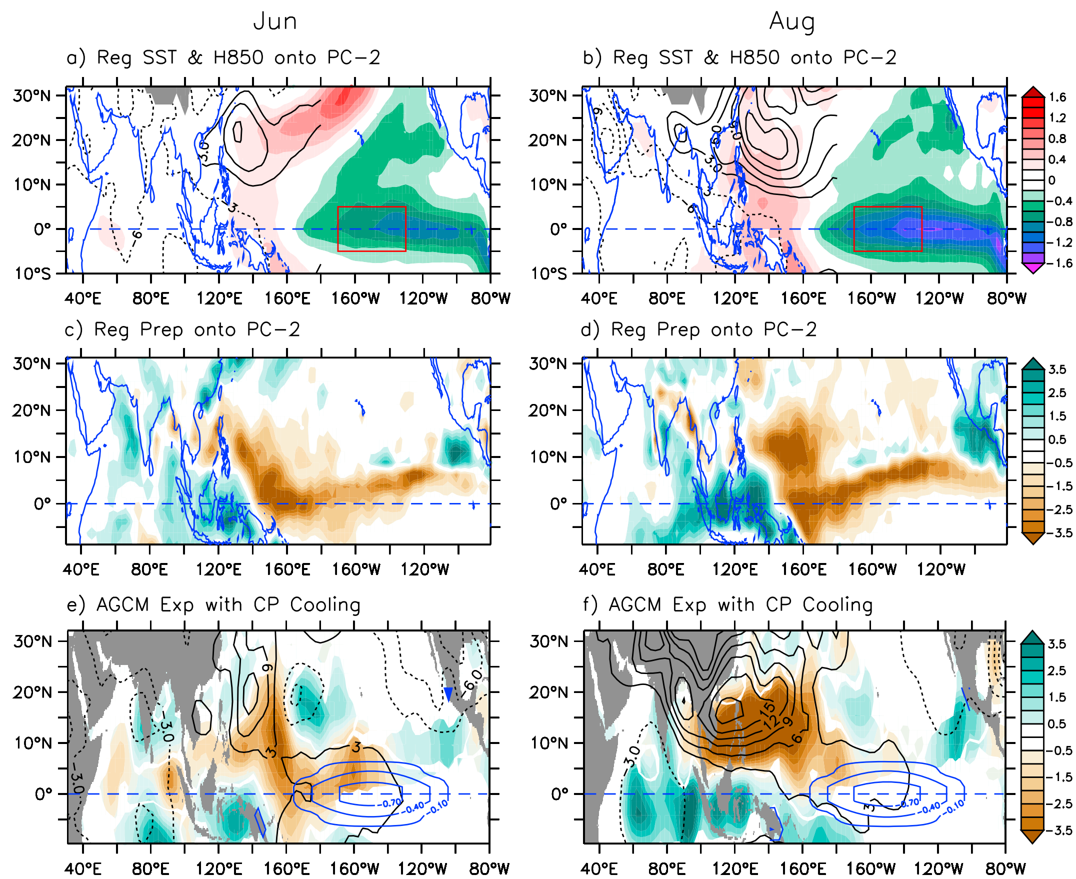

Figure 5. Same as Figure 3 but for the EOF-2 mode. Note that Figures 5c and 5d do not have the regressed wind speed. The red boxes in Figures 5a and 5b denote the key region in driving the anomalous WNPSH.

3.2.2. The Forced Mode by ENSO Development

[13] The EOF-2 mode of JJA H850 features a contrasting pattern between the WNP and the tropical IO/Maritime content (Figure 2d) [*Wang et al*., 2013]. The regressed H850 averaged over the WNP, where the WNPSH index is defined, is much larger in August than in June (12.94 versus 6.43), suggestive of an intensification of WNPSH. Note that the most significant correlation with the simultaneous SSTA is over the equatorial CP (Figure 2f), reminiscent of an ENSO developing mode [*Wang et al*., 2013]. However, the CP SSTA change alone can explain only about 30% of the intensity change of WNPSH from June to August, with the regressed SSTA over the equatorial CP (5°S–5°N, 170°W–130°W, red box in Figure 5) decreasing from −0.75 (June) to −0.98 (August). Again, the seasonal mean state change largely contributes to the intensified WNPSH during the late summer associated with the EOF-2 mode. The atmospheric model experiments with prescribed SST cooling over the equatorial CP also well support the argument that intense WNPSH can be induced with more mean precipitation in the WNP (Figure 5f versus Figure 5e).

[14] A significant trend can be seen for the PC-2 (Figure 2b), indicating that EOF-2-related WNPSH is becoming stronger on a decadal time scale. This is arguably due to the decadal mean state change characterized by a grand La Niña-like mean state change in recent decades [*McPhaden et al*., 2011]. Another point worthy of noting is that some La Niña developing years coincide with El Niño decaying years (such as 1998 and 2010), so one should be cautious when doing composite analyses.

4. Summary and Discussion

[15] In this study, it is demonstrated that the majority of the extreme WNPSH events intensify in late summer. The seasonal march of mean state is argued to be essential in intensifying the anomalous WNPSH through two mechanisms: the local CWES feedback and remote ENSO forcing. Simply, the WNP monsoon trough in the late summer makes the atmosphere more sensitive to SST perturbation than the early summer.

[16] Is the local air-sea feedback weakened in late summer? As mentioned before, the local feedback (i.e., CWES feedback) is strongly dependent on both the background mean winds and precipitation. With the eastward expansion of monsoon trough in August, the mean easterly winds in the WNP weaken compared with those in June (Figure 4). Nevertheless, the SST cooling region well collocates with the region with prominently enhanced wind speed (Figure 3) and background mean easterly winds (Figure 4b). Additionally, the intensified mean precipitation in the WNP makes the atmosphere more sensitive to SST forcing, suggesting that the local air-sea interaction is still active and even stronger in August with the involvement of convection-boundary divergence feedback. However, *Wu et al*. [2010] concluded that the local air-sea interaction is less important in late summer. The discrepancy likely comes from the fact that *Wu et al*. [2010] used both the SST cooling and warming in the WNP as a forcing in AGCM experiments, which may pose problems since

the WNP SST warming largely represents a result of atmospheric forcing.

[17] Many AGCM or simple model experiments have been made to show the importance of the IO warming on WNPSH, but it one should be cautious in taking the northern IO warming as a forcing based on the following pieces of evidence. The northern IO warming does not produce robust enhanced precipitation, and the H850 shows a positive anomaly in the northern IO (Figure 3). The AGCM experiments with SST as a forcing, however, will produce enhanced precipitation which is at odds with observations. It has been suggested that the conventional notion of taking ocean SST as a forcing in the heavily precipitating monsoon region is a fundamental flaw that leads to simulation and seasonal forecast error [*Wang et al*., 2005].

[18] The above argument does not mean that the IO is not important for the WNPSH. The coupled model experiments by *Chowdary et al*. [2010, 2011] have clearly shown the importance of *the IO air-sea interaction* in maintaining the WNPSH. If there were no air-sea interaction in the IO, the northern IO would have suppressed precipitation due to the influence of the westward extension of the WNPSH ridge [*Chowdary et al*., 2012], and this resultant suppressed rainfall in the northern IO (anomalous atmospheric cooling) will weaken the WNPSH. By contrast, with the air-sea interaction in the IO, the northern IO warming produces an "invisible" local rainfall to offset the suppressed rainfall related to the WNPSH, resulting in a rainfall-neutral condition in the northern IO. Therefore, as a coupled system, the northern IO warming can be intensified by the WNPSH forcing, which in turn feeds back to enhance the WNPSH. We conclude that the air-sea interaction and the resultant SST warming in the northern IO are important for the WNPSH.

[19] **Acknowledgments.** The authors are grateful to two anonymous reviewers for their helpful comments. This work has been supported by the APEC Climate Center. The authors acknowledge partial support from International Pacific Research Center which is sponsored by the JAMSTEC, NASA (NNX07AG53G) and NOAA (NA09OAR4320075). This is SOEST contribution number 8912 and IPRC contribution number 972.

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