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Interannual and interdecadal variability of the summertime western North Pacific subtropical high

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[1] The western North Pacific Subtropical High (WNPSH) in summer exhibits significant 2–3 years and 3–5 years oscillations with interdecadal variability. The 2–3-year oscillation is most evident after 1990. It is accompanied by anomalous meridional overturning circulation characterized by warm SST anomalies (SSTA) and ascending motion in the maritime continent and anomalous descending motion near the Philippine Sea, and by evolving warm to cold SSTA in the central-eastern Pacific from the preceding winter to the summer. The 3–5-year oscillation is most pronounced during the 1980s. It is accompanied by anomalous descending motion over the maritime continent and warm SSTA in the central-eastern equatorial Pacific that persists from the preceding winter to the summer; the complementary cooling and descending motion in the western Pacific are related to anomalous east-west circulation associated with ENSO. **Citation:** Sui, C.-H., P.-H. Chung, and T. Li (2007), Interannual and interdecadal variability of the summertime western North Pacific subtropical high, *Geophys. Res. Lett.*, *34*, L11701, doi:10.1029/2006GL029204.

1. Introduction

[2] Southeast China and Taiwan are located in the East Asian monsoon region at the western side of the western North Pacific Subtropical High (WNPSH). The WNPSH has a marked seasonal change with the peak phase in summer when it reaches the northernmost position. The maintenance of the summertime WNPSH is attributed to the monsoonal diabatic heating [*Ting*, 1994; *Hoskins*, 1996; *Rodwell and Hoskins*, 2001; *Chen et al.*, 2001], land-sea heating contrast [*Wu and Liu*, 2003; *Liu et al.*, 2004; *Miyasaka and Nakamura*, 2005], and air-sea interaction [*Seager et al.*, 2003].

[3] In addition to seasonal change, the western edge of WNPSH exhibits significant interannual changes in its zonal and meridional extent. The position and intensity of a seasonal rain front that forms between the monsoonal flow from the tropics and drier, cooler air from the north [*Tao and Chen*, 1987] may vary with the interannual changes of WNPSH. Such changes affect the weather and climate in regions located along its periphery. For example, *Chang et al.* [2000a, 2000b] found that stronger and westward-extending WNPSH blocks the pre-Mei-yu fronts from moving southward, causing heavier rainfall along Yangtze River Valley and droughts in southeast China in May and June. The anomalous summertime anticyclone is associated with equatorial eastern Pacific warm SST anomalies (SSTA) in previous winter. *Lu* [2001] found a relationship between the zonal extent of the subtropical high and the intensity of atmospheric convection over the warm pool. Stronger (weaker) convection over the warm pool is associated with more eastward (westward) extension and clear (unclear) seasonal northward migration of the subtropical high. *Lu and Dong* [2001] further used a numerical model to show that the suppressed convection caused by cold SSTA in the warm pool results in anomalous anticyclonic circulation in the lower troposphere over the subtropical western North Pacific (WNP), and westward extension of the subtropical high.

[4] In this study, we point out that the interannual variation of WNPSH has two distinctive oscillations on quasi-biennial (2–3-yr) and lower-frequency (3–5-yr) time-scales, and physical processes that determine the two scales are different. The two dominant oscillations also reveal interdecadal changes, which has important implications for the monsoon prediction.

2. Data and WNPSH Index

[5] The primary data used in this study are monthly variables derived from the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) Reanalysis 1 [*Kalnay et al.*, 1996] on a 2.5° × 2.5° grid from 1958 to 2005, and the monthly Extended Reconstructed Sea Surface Temperature version 2 (ERSST.v2) for 1958–2005 developed on a 2° × 2° grid by *Smith and Reynolds* [2004].

[6] The interannual change of the summer WNPSH is measured by the geopotential height at 500 hPa (Z_{500}). Figure 1a shows the standard deviation ($\tilde{\sigma}_Z$) of summer (JJA) mean Z_{500} during the period of 1958–2005. It was calculated from 2–5-year band-pass filtered data, and has been normalized by its zonal mean value. The figure shows that the largest variability occurs in the WNP region (east of Taiwan), western flank of the climatological mean subtropical high as depicted by the contour of Z_{500} = 5870 m. We calculate the area mean value of Z_{500} in JJA within the large $\tilde{\sigma}_Z$ region (120–140E. 10–30N) and use its anomaly from the climatological mean as an index for studying the interannual variability of WNPSH.

[7] A power-spectral analysis of the 48-year WNPSH index (Figure 1b) indicates two dominant peaks at 2.5 years and 3.6 years that pass the 95% significant test. Through a Fourier series decomposition of the index series, we con-

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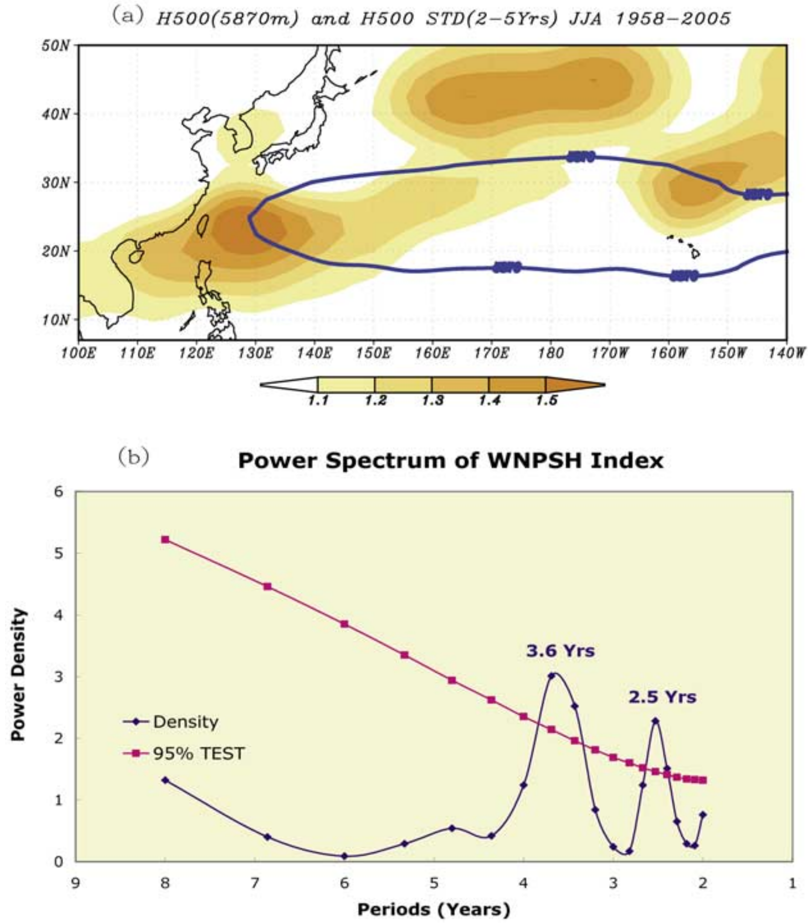

Figure 1. (a) Contour represents the climatological value of 500 hPa geopotential height (Z_{500}) at 5870 m and shading denotes the standard deviation of Z_{500} calculated by 2–5 year band-pass filtered data ($\tilde{\sigma}_Z$) normalized by its zonal mean value. (b) Power spectrum of time series of the summertime WNPSH index.

struct two band-pass time series by summing Fourier components within the 2–3-yr and 3–5-yr period ranges. Hereafter, the two time series are referred to as the 2–3-yr and 3–5-yr WNPSH indices, respectively. Positive indices represent the westward-extending phase of the subtropical high, while negative indices represent the opposite.

3. Characteristics of Large-Scale Patterns

[8] The structures of atmospheric circulation associated with the 2–3-yr and 3–5-yr oscillations of WNPSH are investigated by examining the horizontal patterns of 500 hPa geopotential height (Z_{500}), 500 hPa vertical pressure velocity (ω_{500}), 500 hPa wind ($\mathbf{V}_{500}$) and outgoing longwave radiation (OLR) in the concurrent summer. The results are shown in Figure 2. For the 2–3-yr oscillation, the distribution of anomalous ω_{500} and OLR indicates that the intensified WNPSH tends to accompany the most significant anomalous ascent and enhanced convection over the maritime continent (100–120E, 0–10S), and anomalous descent with suppressed convection resides over the Philippine Sea, while relatively less significant anomalous subsidence and suppressed convection appear in the equatorial eastern Pacific. A zonal band of anomalous ascent with enhanced convection is also observed to the north of WNPSH. The anomalous ω_{500} and OLR in the western Pacific resemble a negative phase of the Pacific-Japan (PJ) pattern [*Nitta*, 1987], which may be triggered and maintained by negative heating in WNP through converting energy from the zonally asymmetric summertime mean flow [*Kosaka and Nakamura*, 2006]. The anomalous subtropical high (Z_{500}) appears slightly north of the subsidence center, accompanied by anomalous anticyclonic circulation ($\mathbf{V}_{500}$), with significant anomalous easterlies extending from northern Australia to Indian subcontinent and anomalous westerlies extending from the Yangtze River to the North Pacific.

[9] Different from the 2–3-yr oscillation, the anomalous ω_{500} and OLR show a distinctive feature in the 3–5-yr oscillation, this is, anomalous subsidence and suppressed convection appear in the maritime continent, while ascending motion and enhanced convection occur in the equatorial eastern Pacific. Over the WNP region, descending motion and suppressed convection anomalies are located southeast of the anomalous anticyclone and subtropical high center. The feature above is consistent to that found by *Chang et al.* [2000a, 2000b].

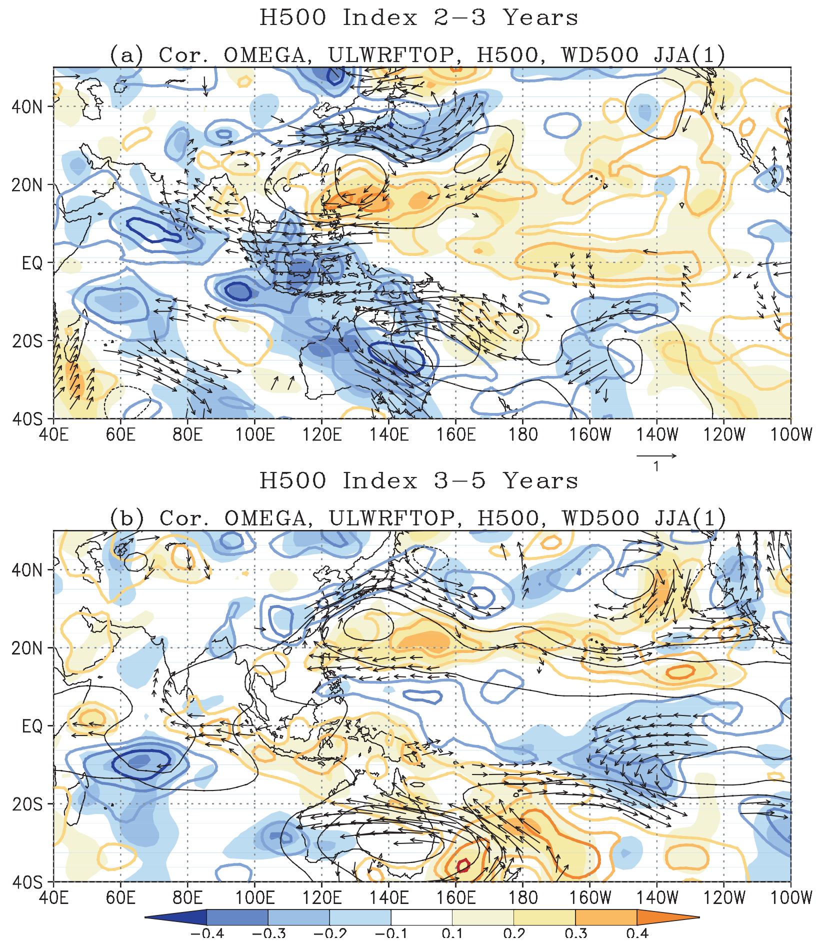

Figure 2. (a) Map of correlation coefficients between the 2–3-year oscillation of the WNPSH index and vertical p-velocity (shading, unit: Pa/s), geopotential height (black contour, unit: m), significant wind fields (vector, unit: m/s) at 500 hPa and outgoing longwave radiation (color contour, unit: W/m^2) in boreal summer. (b) The same as Figure 2a but with the 3–5-yr filtered WNPSH index. The positive (negative) correlation of the vertical p-velocity and OLR corresponds to anomalous descent (ascent) and suppressed (enhanced) convection in the presence of the intensified WNPSH. The black contours are equal or above (below) 0.2 (−0.2) with a contour interval of 0.1, and the color contours are equal or above (below) 0.1 (−0.1) with a contour interval of 0.1. The correlation coefficient at 90% and 95% significant level is 0.2 and 0.26 in Figure 2a, 0.3 and 0.4 in Figure 2b.

[10] Although the subtropical high anomalies on both timescales are associated with the local subsidence, the cause of the anomalous descending motion on these scales might be different. For the 2–3-yr oscillation, it is obviously attributed to anomalous local Hadley circulation in association with the rising motion in the maritime continent. For the 3–5-yr oscillation, it is not attributed to local Hadley circulation but due to local SSTA forcing as shown below.

[11] The evolution of SSTA associated with the 2–3-yr and 3–5-yr oscillations of WNPSH was examined. For the 2–3-yr oscillation, the SSTA evolution resembles the observed tropospheric biennial oscillation (TBO) in the Indo-Pacific warm ocean region as explored using a seasonal-sequence EOF analysis by *Li et al.* [2006, Figure 6]. Over the Pacific, warm SSTA appear in the equatorial eastern Pacific in the preceding winter, and decay through spring and transition into cold SSTA in summer. Over the Indian Ocean, warm SSTA appear first in the tropical Indian Ocean in the preceding winter, and extend slowly eastward as season progresses. By summer, warm SSTA cover the maritime continent and northern Australia. Thus, it is the warm SSTA that lead to the enhanced convection and anomalous ascending motion over the maritime continent in summer.

[12] For the 3–5-yr oscillation, warm SSTA in the equatorial eastern Pacific and cold SSTA in the WNP persist from the preceding winter to summer, while SSTA in the SCS and along the East Asian coast remain warm (Figure 3). This typical tri-pole SSTA pattern was frequently observed during the decaying phase of El Nino episodes [e.g., *Wang et al.*, 2000]. Over the tropical Indian Ocean, warm SSTA also persist from northern winter to summer.

[13] In association with the local cold SSTA, the anomalous 850 hPa anticyclone and 500 hPa descending motion persist in the WNP from the previous winter to the concurrent summer, even though its shape and position change to a certain extent. Along the equator, the reversed anomalous Walker cells appear in both the Pacific and Indian Oceans from the preceding winter to summer, as revealed from velocity potential fields (figure not shown).

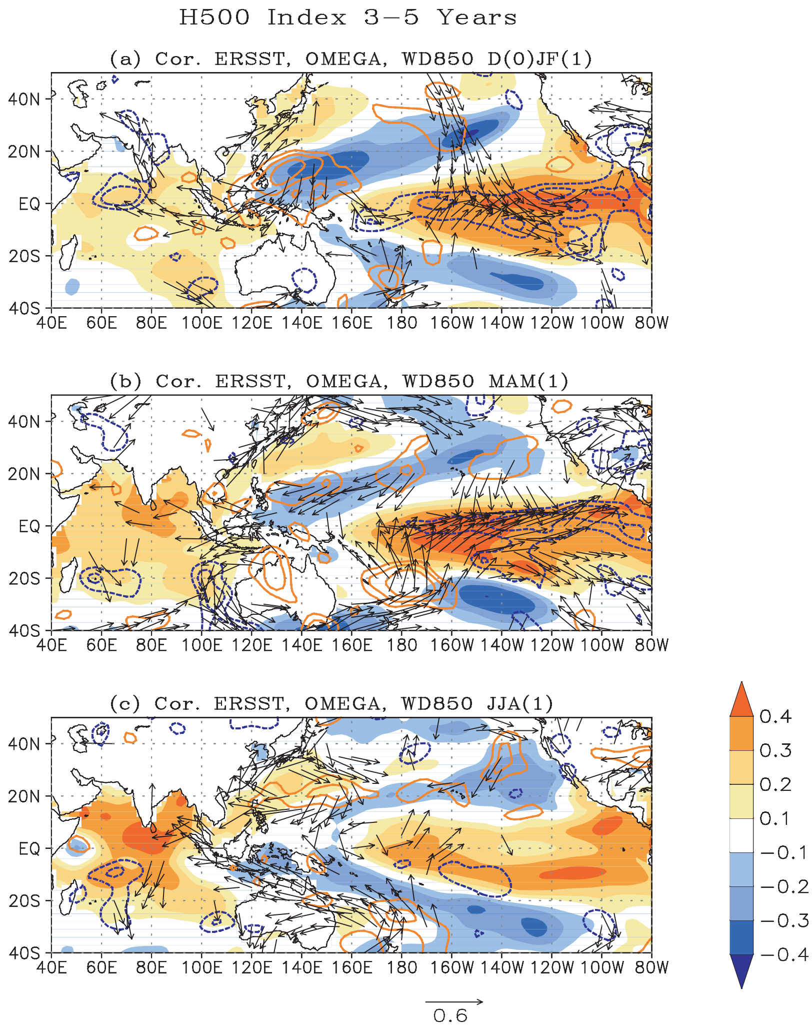

Figure 3. Map of lagged correlation coefficients between the 3–5-yr filtered WNPSH index and global SSTA (shading, unit: °C), vertical p-velocity (color contour, unit: Pa/s) at 500 hPa and significant wind fields (vector, unit: m/s) at 850 hPa in (a) the preceding winter, (b) spring and (c) the concurrent summer. The positive (negative) correlation of the vertical p-velocity corresponds to anomalous descent (ascent) in the presence of the intensified WNPSH. The color contours are equal or above (below) 0.2 (−0.2) with an interval of 0.1, and red (blue) contours denote positive (negative) correlations. The correlation coefficient at 90% and 95% significant level is 0.3 and 0.4.

[14] The lagged correlation analysis in Figure 3 suggests that a strong WNPSH associated with the 3–5-yr oscillation occurs most likely in the El Nino decaying summer. The anomalous WNP anticyclone is maintained for three seasons from the previous winter to summer through a positive air-sea feedback [*Wang et al.*, 2000, 2003] mechanism in which an anomalous anticyclone, initiated by cold SSTA as a Rossby wave response to a negative heat source, may further reinforce the cold SSTA through wind-evaporation-SST feedback under prevailing climatological northeasterly trade winds.

[15] To sum up, the observational analysis suggests that the biennial variability of WNPSH is primarily caused by convection and SST anomalies over the maritime continent while the lower-frequency variability is attributed to the local SSTA in the WNP. Previous studies suggested that the tropical Indian Ocean SSTA may play a role in the 3–5-yr oscillation. For example, *Lee et al.* [2005] showed that convective activities over the Bay of Bengal are responsible for the suppressed convection over the western Pacific and the development of WNPSH in summer. In our analysis, a close look at Figure 3c shows that the warmest SSTA reside over the Bay of Bengal and in the northern Indian Ocean but enhanced convective activities occur south of the equator. Thus the warm SSTA in the northern Indian Ocean do not represent active heating, rather a passive response to local cloud radiative forcing and decrease in surface evaporation due to anomalous low-level easterlies against the mean monsoon westerlies.

4. Modulation by Interdecadal Oscillation

[16] To analyze the decadal modulation of the WNPSH interannual variability, we perform a wavelet analysis of the detrended WNPSH index and the time series of SSTA in the Nino3.4 region (preceding winter) and the maritime continent region (concurrent summer). The analysis result is

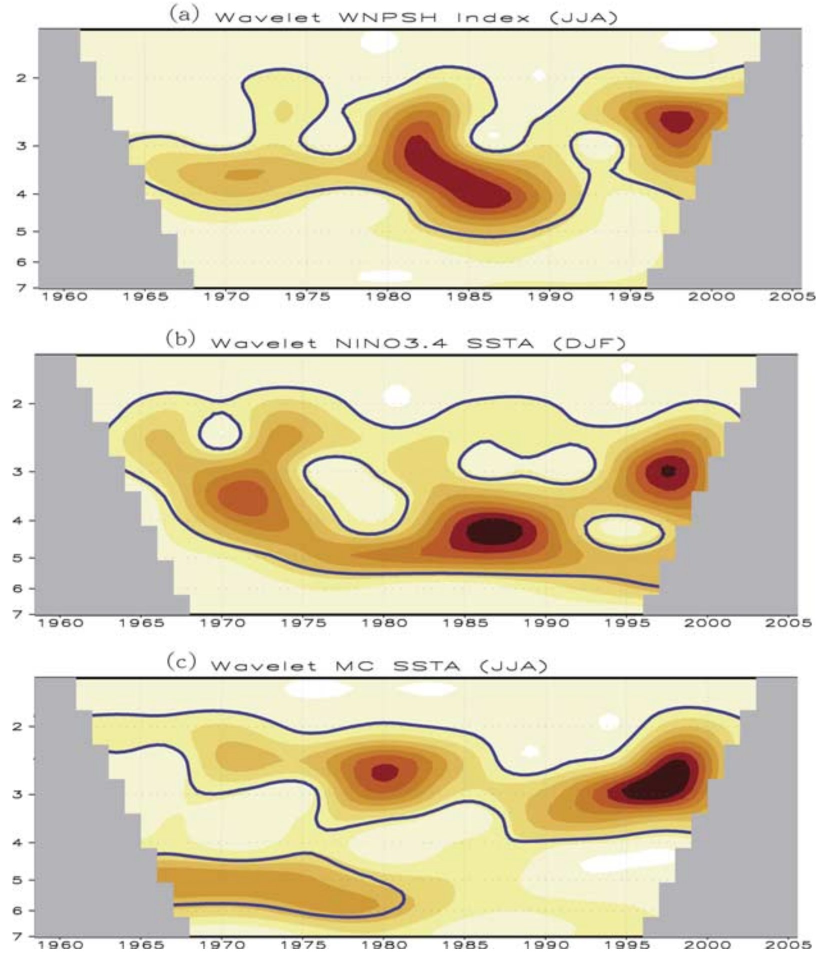

Figure 4. Wavelets power spectrum of (a) JJA WNPSH index with a linear trend removed, (b) DJF Nino3.4 SSTA, and (c) maritime continent JJA SSTA with a linear trend removed. Shading shows the power, and contour represents the 95% significance level (the edge effect blocked by grey areas).

shown in Figure 4. The WNPSH index possesses two dominant periods at 2–3 years and 3–5 years. The Nino 3.4 SSTA has a primary period at 3–5 years (with stronger amplitude) and a secondary period at 2–3 years (with weaker amplitude). On the other hand, the maritime continent SSTA exhibit a dominant period at 2–3 years only.

[17] Figure 4 also shows that the 2–3-yr oscillation of the WNPSH is strongest after 1990. This corresponds well with the 2–3-yr variability of maritime continent SSTA and some portion of Nino3.4 SSTA. Furthermore, the 3–5-yr oscillation of the WNPSH is most evident during the 1980s. This coincides well with the 3–5-yr oscillation of Nino3.4 SSTA during the same period.

[18] Our wavelet analysis is consistent with results by *Kwon et al.* [2005], who recognized a decadal change of the East Asia-WNP summer monsoon from an ENSO-related oscillation in 1979–1993 to a monsoon-dominated oscillation in 1994–2004. The circulation features in the two studies, however, are different, possibly due to the different reference fields used.

5. Conclusion and Discussion

[19] Interannual variability of WNPSH in boreal summer exhibits two distinct periodicity, 2–3 years and 3–5 years. Our analysis reveals that different circulation features are associated with the two oscillations. For the 2–3-yr oscillation, an enhanced WNPSH is associated with warm SSTA and enhanced convection over the maritime continent that connects the local ascending motion to the descending motion in the subtropical WNP through an anomalous Hadley circulation. Meanwhile it is associated with simultaneous cold SSTA in the central to eastern equatorial Pacific. For the 3–5-yr oscillation, an enhanced WNPSH is associated with persistent local cold SSTA and anomalous descending motion in the subtropical WNP from the preceding winter to the summer. The descending branch is well connected to ascending branches over the eastern equatorial Pacific and Indian Ocean through anomalous Walker cells.

[20] The current study found a lower-frequency modulation of 2–3-yr and 3–5-yr oscillations of WNPSH, i.e. pronounced 2–3 yr oscillations occur after 1990 while strong 3–5-yr oscillations happen during the 1980s. In addition, we noticed a marked (interdecadal) warm trend of SST over the maritime continent and South China Sea that is particularly evident after 1986. We speculate that the 2–3-yr oscillation of WNPSH might be related to the monsoon - warm ocean interaction associated with the TBO [*Li et al.*, 2006], while the 3–5-yr oscillation of WNPSH is related to ENSO-East Asia teleconnection [*Wang et al.*,

2000]. Since ENSO has both 2–3-yr (minor) and 3–7-yr (major) spectral peaks, the 2–3-yr oscillation of WNPSH may also be related to the biennial ENSO. The possible causes for the biennial ENSO may include the monsoon-ocean feedback and inter-basin teleconnection between the tropical Pacific and Indian Ocean [*Li et al.*, 2006]. The low-frequency trend of SST in the maritime continent is speculated to have a positive influence on the monsoon-ocean interaction to give rise to dominant 2–3-yr oscillation of WNPSH in 1990s. In addition to the tropical heating, mid-latitude processes may also play a role in the interannual variability of WNPSH [e.g., *Enomoto*, 2004].

[21] The opposite correlation between the WNPSH intensity and convective activity over the maritime continent on the 2–3-yr and 3–5-yr timescales poses an interesting question as how the PJ pattern is related to equatorial heating. We plan to explore this issue and understand the relative roles of air-sea interactions in different ocean basins (such as Pacific cold tongue, warm pool and Indian Ocean) in causing the quasi-biennial and lower-frequency variability of WNPSH through idealized, partial-coupling coupled ocean-atmosphere model experiments. We regard our finding to be of interest of a wider audience, and a step further toward a better understanding of interannual variability of the Asian monsoon climate.

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