Interannual relationships between the tropical sea surface temperature and summertime subtropical anticyclone over the western North Pacific

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The interannual variability of the Western North Pacific Subtropical High (WNPSH) in boreal summer is investigated with the use of the NCEP/NCAR Reanalysis Data. The most significant change of the 500 hPa geopotential height field appears at the western edge of the WNPSH, with dominant 2–3 year and 3–5 year power spectrum peaks. The 2–3 year oscillation of the WNPSH and associated circulation and sea surface temperature (SST) patterns possess a coherent eastward propagating feature, with a warm SST anomaly (SSTA) and anomalous ascending motion migrating from the tropical Indian Ocean in the preceding autumn to the maritime continent in the concurrent summer of a strong WNPSH. A strong WNPSH is characterized by anomalous anticyclonic circulation and maximum subsidence in the western North Pacific (WNP). The anomalous WNPSH circulation has an equivalent barotropic vertical structure and resides in the sinking branch of the local Hadley circulation, triggered by enhanced convection over the maritime continent. A heat budget analysis reveals that the WNPSH is maintained by radiative cooling. The 3–5 year oscillation of the WNPSH exhibits a quasi-stationary feature, with a warm SST anomaly (anomalous ascending motion) located in the equatorial central eastern Pacific and Indian Ocean and a cold SST anomaly (anomalous descending motion) located in the western Pacific. The anomaly pattern persists from the preceding winter to the concurrent summer of a high WNPSH. The greatest descent is located to the southeast of the anomalous anticyclone center, where a baroclinic vertical structure is identified. The zonal phase difference and the baroclinic vertical structure suggest that the anomalous anticyclone on this timescale is a Rossby wave response to a negative latent heating associated with the persistent local cold SSTA. ECHAM4 model experiments further confirm that the 2–3 year mode is driven by the SSTA forcing over the maritime continent, while the 3–5 year mode is driven by the local SSTA in the WNP.


1. Introduction

The subtropical high over the North Pacific is a dominant component of atmospheric general circulation that experiences a distinct seasonal cycle with the maximum strength and northernmost position occurring in the boreal summer. The evolution of the subtropical high over the western North Pacific (hereafter WNPSH) in summer is associated with the onset and withdrawal of the Asian summer monsoon. The zonal shift of the western part of the WNPSH dominates the position of large-scale frontal zones in East Asia [Tao and Chen, 1987]. A southwestward extension of the WNPSH would result in more water vapor convergence and excessive rainfall along the Yangtze River valley [Chang et al., 2000a, 2000b; Zhou and Yu, 2005]. The precipitation characters in Japan and Korea were also greatly influenced by the movement of the WNPSH through changing the transport of water vapor into the Baiu or Changma frontal area [e.g., Tomita et al., 2004; Lee et al., 2005].

Traditionally, subtropical highs are regarded as the descending branch of the Hadley circulation maintained by radiative cooling. This idealized, zonally symmetric Hadley circulation model, however, would lead to a weaker (stronger) subtropical high in the summer (winter) hemisphere, which is contrary to observations [Hoskins, 1996]. Moreover, the observed summertime subtropical highs are...
in the form of high-pressure cells rather than a zonally uniform high-pressure belt. The dynamics of the summertime subtropical high is better understood in the context of planetary waves, as suggested by Ting [1994], who found that summertime subtropical stationary wave patterns are maintained by the global diabatic heating. [4] Hoskins [1996] and Rodwell and Hoskins [2001] applied the monsoon-desert mechanism proposed by Rodwell and Hoskins [1996] to argue that the remote diabatic heating over the North American monsoon region could induce a Rossby wave pattern to the west and contribute to the lower tropospheric subtropical anticyclone in the North Pacific. Rodwell and Hoskins [2001] further examined the dynamics of the summertime subtropical high in a global nonlinear model and concluded that the combined effect of the topography and the monsoonal heating is of primary importance for the generation of the surface subtropical high and subsidence aloft over the eastern Pacific. This differs from Chen et al. [2001], who argued that summertime subtropical highs can be maintained by downstream energy propagation of stationary Rossby waves forced by convection heating over the Asian monsoon region in a linear quasigeostrophic model. Their experiments, however, could not produce the distinct meridional dependence of the observed vertical structures of the summertime planetary wave, possibly due to the lack of other heating sources such as sensible heating over the western portion of the continent and radiation cooling off the west coast of the continent. Liu et al. [2004] performed atmospheric GCM experiments to show the importance of the global surface sensible heating and longwave radiative cooling in the formation of the subtropical anticyclones. Miyasaka and Nakamura [2005] demonstrated that the summertime subtropical high can be accurately reproduced by the near-surface thermal contrast between a cooler ocean and a hotter land to the east. The role of local air-sea interactions in generating the subtropical anticyclone was investigated numerically by Seager et al. [2003]. All studies above suggest the important role of the land-ocean-atmosphere feedback in maintaining the subtropical high. [5] In addition to the annual cycle, the subtropical high in the boreal summer also exhibits remarkable interannual variations. Convective heating over the western Pacific warm pool is regarded as an important forcing that affects the extent of the subtropical high in the boreal summer. According to Nitta [1987], convective activity in the tropical western Pacific during summer can influence the Northern Hemispheric circulation; that is, Rossby waves generated by the anomalous tropical heat source can propagate into higher latitudes. This leads to a north–south dipole between sub tropics and middle latitude East Asia, known as the Pacific-Japan (PJ) Pattern. Lu [2001] found that stronger convection over the warm pool is associated with more eastward and northward migration of the subtropical high. Lu and Dong [2001] performed numerical model experiments and showed that suppressed convection caused by a cold SST in the warm pool results in anomalous anticyclonic circulation in the lower troposphere over the subtropical western North Pacific (WNP), providing a favorable condition for the westward extent of the subtropical high. [6] In addition to the warm pool heating, the El Niño-Southern Oscillation (ENSO) may also influence the East Asian climate through the modulation of low-level winds over the WNP [e.g., Wang et al., 2000; Chang et al., 2000a, 2000b; Lau and Nath, 2000; Kawamura et al., 2001; Lau and Wu, 2001; Chou, 2004]. An anomalous low-level anticyclone dominates the WNP during the mature phase of an El Niño, whereas an anomalous cyclone occurs during the La Nina mature winter. The anomalous anticyclone persists through the following spring and summer, leading to a weak WNP monsoon in the ENSO decaying summer [e.g., Wang et al., 2000, 2003; Chou et al., 2003]. The anticyclone in early summer favors the northward movement of the Mei-yu/Baiu front and thus enhances (weakens) the East Asia (WNP) summer monsoon rainfall [Chang et al., 2000a, 2000b]. [7] The wintertime anomalous low-level anticyclone in WNP may be regarded as a Rossby wave response to the suppressed convection over the western Pacific, which is possibly caused both by the remote warm SSTA over the eastern Pacific [Wang et al., 2000; Chou, 2004; Chen et al., 2007] and the local cold SSTA [Lau and Nath, 2000; Wang et al., 2000]. The maintenance of the WNP anticyclone from the El Nino mature winter to its decaying early summer is attributed to a season-dependent thermodynamic air-sea feedback [Wang et al., 2000, 2003; Kawamura et al., 2001; Chou, 2004; Li and Wang, 2005]. During the El Nino decaying summer, the SST in the eastern Pacific has diminished, yet the low-level anticyclone still exists over the WNP. It is likely that both the cold SST in the WNP in late spring/early summer [Sui et al., 2007; Wu et al., 2010] and the basin-wide warming in the IO in summer [Yang et al., 2007; Xie et al., 2009; Wu et al., 2009a, 2009b] are responsible for maintaining the WNPSH during the summer. [8] The Asian summer monsoon shows a strong quasi-biennial variability [e.g., Shen and Lau, 1995; Barnett, 1991; Chang and Li, 2000; Li et al., 2001; Li and Zhang, 2002], which is termed as the tropospheric biennial oscillation (TBO) in the Asian-Australian monsoon region [Meehl, 1994, 1997]. The tropical Pacific and southeast IO (SEIO) SST associated with the Asian summer monsoon also exhibits a biennial characteristic [Chang et al., 2000a; Wang et al., 2001]. A hybrid coupled GCM simulation shows that the TBO may exist even in the absence of ENSO [Li et al., 2006]. On the other hand, ENSO may be linked to the monsoon TBO through the following three branches of teleconnection: (1) WNP monsoon–SEIO SST teleconnection or the Philippine-Sumatra pattern [Li et al., 2002; Li et al., 2006; Wu et al., 2009b]; (2) IO convection–Central Eastern Pacific (CEP) SST teleconnection [Meehl, 1987; Chang and Li, 2000; Yu et al., 2002; Watanabe and Jin, 2002; Li et al., 2003]; and (3) Pacific-East Asia teleconnection [Wang et al., 2000, 2003]. [9] Because the WNPSH is regarded as one of the important components of the Asian summer monsoon system, the objective of this study is to conduct an analysis of the interannual evolution of the WNPSH and investigate possible mechanisms that link to the variability. The rest of the paper is organized as follows. Section 2 describes the data used in this study. In section 3 a WNPSH index is defined and two distinctive spectrum peaks on quasi-biennial
(2–3 year) and lower-frequency (3–5 year) periods are identified. A regression analysis is subsequently performed to obtain the large-scale circulation, SST, outgoing longwave radiation, and heat and moisture budgets associated with the two oscillation periods. By synthesizing all features, possible mechanisms responsible for the two oscillations of the WNPSH are proposed. The idealized experiments of the ECHAM4 AGCM are also examined in section 4. In section 5 the seasonal lead–lag patterns of regressed SST and circulation fields associated with the 2–3 year and 3–5 year oscillations of the WNPSH are discussed. Section 6 comprises a summary and discussions.

2. Data and Model Description

2.1. Reanalysis Data

[10] The primary data set used in this study is the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP/NCAR) Reanalysis [Kalnay et al., 1996] that includes wind, geopotential height, stream function, vertical p velocity, sea level pressure, velocity potential, and outgoing longwave radiation flux on a 2.5 x 2.5 grid for 48 years (1958–2005). This is the longest reanalysis product available. Another set of data used in the study is the NOAA Extend Reconstructed Sea Surface Temperature version 2 (ERSST v2) developed on a 2° x 2° grid for the 48 years (1958–2005). The extended reconstruction of historical SST was using improved statistical methods from 1854 to the present. A full description of the ERSST v2 is available in the work of Smith and Reynolds [2004].

[11] Further, this study used the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA), which also contains the above variables at the same resolution from 1957 to 2002. A full description of the ERA is available in the work of Gibson et al. [1996, 1997]. The analysis results from this study, based on NCEP/NCAR data, were compared against those from ERA and the major features were found to be essentially the same. Most of the results discussed in this study are based on NCEP/NCAR reanalysis, except the following.

[12] In order to discuss the dynamic features of the climate oscillations in this study, the heat and moisture budgets were computed, the so-called apparent heat source (Q1) and apparent moisture sink (Q2) [e.g., Yanai et al., 1973], using ERA. This decision is made due to the fact that the overall simulation of the diabatic heating (and therefore the moisture) field over the monsoon region in ERA is quite close to reality [Annamalai et al., 1999]. In addition, a comparison of the heat and moisture budgets based on ERA, NCEP/NCAR R-1 Reanalysis, and NCEP/DOE AMIP-II (R-2) Reanalysis [Kanamitsu et al., 2002] was completed. The results show that budgets computed based on both ERA and R-2 reanalysis data are quite similar to the observed rainfall variability in the tropical and East Asian monsoon regions. However, the R-2 reanalysis data only cover the satellite period from 1979 to present. Therefore the Q1 and Q2 budgets were computed at each vertical level (23 pressure levels) using the 6-hourly ERA data from 1958 to 2001.

[13] According to Yanai et al. [1973], the large-scale heat and moisture budget residuals, Q1 and Q2, respectively, are defined and computed by

\[ Q_1 \equiv c_p \left( \frac{P}{p_0} \right) ^ \lambda \left( \frac{\partial q}{\partial t} + \beta \cdot \nabla q + \omega \frac{\partial q}{\partial p} \right) \]  

and

\[ Q_2 \equiv -L \left( \frac{\partial q}{\partial t} + \beta \cdot \nabla q + \omega \frac{\partial q}{\partial p} \right) \]  

where \( \theta \) is the potential temperature, \( q \) is the water vapor mixing ratio, \( \beta \) is the horizontal velocity, \( \omega \) is the vertical p velocity, \( L \) is the latent heat of vaporization, \( k = R/c_p \) with \( R \) the gas constant and \( c_p \) the specific heat capacity at constant pressure of dry air, and \( p_0 = 1000 \) hPa.

[14] In equations (1) and (2), \( \beta \), \( \theta \), and \( q \) denote large-scale fields that are mathematically defined as Reynolds averaged quantities (normally with an overbar, being neglected here) and are customarily assumed resolvable by reanalysis data. So \( Q_1 \) and \( Q_2 \) are computed by 6-hourly data at each grid point from ERA products. The 6-hourly estimates of \( Q_1 \) and \( Q_2 \) are averaged into monthly mean values and multiplied by \( c_p^{-1} \) to be expressed in terms of equivalent warming/cooling rate (K/day).

[15] The \( Q_1 \) and \( Q_2 \) can be interpreted by:

\[ Q_1 = Q_h + L(\Theta - \Theta_0) - \nabla \cdot \mathbf{s} - \frac{\partial q}{\partial t} \frac{\omega}{\partial p} \]  

and

\[ Q_2 = L(\Theta - \Theta_0) + L \nabla \cdot \mathbf{q} + L \frac{\omega}{\partial p} \]  

where \( Q_h \) is radiative heating rate, \( c \) and \( e \) are condensation and evaporation per unit mass of air respectively, \( s \) is the dry static energy, and the prime denotes deviation from the Reynolds average. The primed variables denote subgrid scale fluctuations that are assumed to be well separated from the resolvable large-scale variable. Equation (3) represents the total effect of radiative heating, latent heat released by net condensation \( L(\Theta - \Theta_0) \), and the horizontal and vertical convergence of sensible heat fluxes due to subgrid-scale eddies such as cumulus convection and turbulence. Equation (4), on the other hand, represents the total effect of net condensation and divergence of eddy moisture fluxes due to subgrid-scale eddies. The contributions of the eddy horizontal transport terms \( \nabla \cdot \mathbf{s} \) in equation (3) and \( \nabla \cdot \mathbf{q} \) in equation (4) are generally small and may be negligible compared to the horizontal transports by the large-scale motion [Yanai and Johnson, 1993].

2.2. Atmospheric General Circulation Model

[16] The Atmospheric General Circulation Model (AGCM) used in the study is ECHAM version 4.6 (hereafter ECHAM4), which was developed by the Max Planck Institute for Meteorology [Roeckner et al., 1996]. The model was run at a horizontal resolution of spectral triangular 42 (T42), with 2.8° grid and 19 vertical levels in a hybrid sigma pressure coordinate system. The climatological monthly SST constructed
by phase II of the Atmospheric Model Intercomparison Project (AMIP II) was used as the model low boundary forcing. The ECHAM4 model is one of the Top-3 AMIP II models in simulating the interannual variability modes of the Asian-Australian monsoon [Zhou et al., 2009a]. It shows excellent performance in simulating the El Nino-related tropospheric temperature anomalies [Zhou and Zhang, 2011]. Thus the model is used in the study for investigating the interannual variations of the summertime circulation in WNP.

3. Characteristics Associated With the Anomalous Subtropical High

3.1. WNPSH Index

[17] In this study the interannual variation of the summer subtropical high over the WNP is measured using the 500 hPa geopotential height field (Z500). Although the summer Pacific subtropical high can be detected in the lower troposphere, the WNPSH is most pronounced in the middle troposphere. The center of the subtropical high in the lower troposphere is located in the eastern Pacific, while the center of the subtropical high in the middle troposphere resides over the western North Pacific [Liu et al., 2001; Liu and Wu, 2004]. For this reason, the variations of the WNPSH in this study are measured by the Z500 variability, following many previous studies [e.g., Tao and Chen, 1987; Zhou and Li, 2002]. Figure 1 shows the standard deviation (\(\sigma_2\)) of the summer (JJA) mean Z500 during the period of 1958–2005. It is calculated from 2 to 7 year band-pass filtered data and has been normalized by its zonal mean value to illustrate the zonally asymmetric character. Figure 1 shows that the largest variability occurs in the northwest Pacific region, southwest of the climatological mean subtropical high as depicted by the contour line of Z500 = 5870 m. A WNPSH index is defined as the area mean value of Z500 anomalies in JJA in the region of (120°–140°E, 10°–30°N). This index is used to measure the strength of the interannual variability of the WNPSH. For comparison, two additional indices are defined based on 850 hPa stream function (\(\psi_{850}\)) and geopotential height anomalies (Z850) in JJA averaged over the region of (120°–150°E, 15°–25°N), where the strongest interannual variation of \(\psi_{850}\) and Z850 occurs (hereafter termed as \(\psi_{850}\) and Z850 indices). The \(\psi_{850}\) and Z850 fields were used by Lee et al. [2005] and Lu [2001] to study the subtropical high variability over the WNP.

[18] The power spectrums of the Z500, \(\psi_{850}\), and Z850 indices are examined. The Z500 index shows two dominant periods at 2–3 years and 3–5 years, both of which pass the 95% confidence level (Figure 2a). The \(\psi_{850}\) and Z850 index also presents two similar peaks although with the 3–5 year band (2–3 year band) being above (slightly below or at) the 95% confidence level (figures not shown). Additional power spectrum analyses with 44-year ERA40 Z500 data (figure not shown) show a similar result. Thus in the following, the Z500 index will be used to represent the WNPSH quasi-biennial (2–3 year) and lower-frequency (3–5 year) variability.

[19] Through a Fourier series decomposition of the WNPSH index series, two band-pass filtered time series were constructed by summing Fourier components within the 2–3 year and 3–5 year periods (Figure 2b). Hereafter, the two time series are referred to as the 2–3 year and 3–5 year WNPSH indices. Positive indices represent the strong and westward extending phase of the subtropical high, while negative indices represent the opposite. In the following analysis, the circulation and SST fields are regressed against the two WNPSH indices to investigate spatial and temporal structures associated with the 2–3 year and 3–5 year oscillations of the WNPSH.

3.2. Circulation Patterns Associated With the 2–3 Year and 3–5 Year Oscillations

[20] The structures related to 2–3 year and 3–5 year oscillations of the WNPSH were investigated by examining the horizontal patterns of regressed fields of summer mean
ERSST, outgoing longwave radiation (OLR), $Z_{500}$, 500 hPa vertical $p$ velocity ($\psi_{500}$), 200 hPa and 850 hPa stream function ($\psi_{200}$ and $\psi_{850}$), 200 hPa and 850 hPa velocity potential ($\chi_{200}$ and $\chi_{850}$), and wind fields at 850 hPa and 500 hPa ($V_{850}$ and $V_{500}$). Since the derived results of regression can be applied to both the positive phase (represented by a strengthened and westward WNPSH) and the negative phase (represented by a weakened WNPSH), the interpretation and discussion of physical mechanisms afterward will be merely based on the features of the positive phase.

[21] The regressed $Z_{500}$, $V_{500}$, $\psi_{850}$, and $\psi_{200}$ are shown in Figure 3. The WNPSH high anomaly of the 2–3 year oscillation is associated with a meridional standing wave pattern with a zonally elongated positive $Z_{500}$ anomaly and anticyclonic circulation in the WNP, a negative $Z_{500}$ anomaly over the northern East Asia, and a positive anomaly over Eurasia (Figure 3b). The horizontal distribution of $\psi_{850}$ in Figure 3c shows a pair of anticyclonic circulations symmetric about 10 N. One is located in the WNP (overlapping with a positive $Z_{500}$ anomaly center) and the other is located south of the Equator. The circulation pattern of $\psi_{200}$ (Figure 3a) shows that an anticyclonic circulation extends from southern to southeastern China and the WNP.

[22] In the 3–5 year oscillation, an anticyclone and a positive $Z_{500}$ anomaly center (Figure 3e) are located near Taiwan when the WNPSH extends westward. The positive $Z_{500}$ anomaly extends further to the tropical central eastern Pacific than that in 2–3 year oscillation. The positive anomaly in $Z_{500}$ and $\psi_{850}$ fields (Figures 3e and 3f) acts as the source of a Rossby wave train propagating from the WNP to North America. The distribution of $\psi_{200}$ (Figure 3d) exhibits a cyclonic circulation pattern over China. In the WNP, $\psi_{200}$ is in general out of phase with $\psi_{850}$.

[23] Figure 4 shows the regressed SST patterns for the 2–3 year and 3–5 year modes. The SST field has been normalized at each grid point based on the climatological standard deviation to capture the obvious SST variations. The SST distribution associated with the westward extending WNPSH in the 2–3 year oscillation (Figure 4a) features a warm SST anomaly extending from the northern IO southeastward to South China Sea, maritime continent and northeastern Australia, and a cold SST in the central eastern equatorial and northeastern and southeastern Pacific. For the 3–5 year oscillation (Figure 4b), however, a warm SST appears in the tropical IO and the central eastern equatorial Pacific. The SST in the central eastern Pacific is surrounded by cold SST anomalies to its north and south and in the maritime continent.

[24] In the 2–3 year oscillation, the large-scale divergent circulation in upper and lower levels are shown by regressed $\chi_{200}$ and $\chi_{850}$ fields in Figures 5a and 5b. A broad overturning circulation occurs in the summer of a strong WNPSH over the western Pacific, with the low–level convergent (divergent) flow toward Java and the eastern IO (out of the tropical central eastern Pacific region) (Figure 5b). The regressed OLR field in Figure 5c shows enhanced convection over the maritime continent and suppressed convection over the Philippine Sea and the equatorial eastern Pacific, consistent with the divergent wind in Figures 5a and 5b. A zonal band of enhanced convection is also observed to the north of the WNPSH. The anomalous OLR pattern in the western Pacific resembles a negative phase of the PJ pattern [Nitta, 1987].

[25] For the 3–5 year oscillation, the divergent circulation pattern differs from that of the 2–3 year oscillation. The regressed $\chi_{200}$ and $\chi_{850}$ fields (shown in Figures 5d and 5e) reveal two anomalous Walker cells in the tropics, with two upper-tropospheric divergence centers located in the tropical central-eastern Pacific and IO, respectively, and a convergence center located in the region extending from southern...
Philippines to the maritime continent and eastern Australia (Figure 5d). The regressed OLR field (Figure 5f) shows the regions of suppressed convection in the maritime continent and subtropical North Pacific and the regions of enhanced convection in the equatorial eastern Pacific and western tropical IO, consistent with the features of the divergent flows shown in the $c_{200}$ and $c_{850}$ fields (Figures 5d and 5e).

The meridional overturning circulation of the 2–3 year oscillation over the western Pacific is further revealed by the regressed zonally averaged vertical $p$ velocity ($\omega$) at each pressure level and latitude–vertical divergent wind profile ($v$, $\omega$) along 110°E–140°E (Figure 6a). Consistent with the regressed OLR and $\chi$ fields, the flow is upward over the maritime continent (0°–10°S) and East Asia near Japan (30°–40°N), with the former being stronger than the latter. The southern branch of upward motion causes upper-tropospheric divergent flows toward the Philippine Sea (at 10°–20°N) where the downward motion is pronounced. The low-level southward divergent flow to the south of the WNPSH may bring the moisture into the equatorial region and enhance convection in the maritime continent. The enhanced convection further reinforces the WNPSH subsidence via a positive feedback process.

For the 3–5 year oscillation, the regressed vertical overturning circulation along 110°E–150°E (Figure 6b) shows two descending branches at 0°–10°S and 15°–25°N, coinciding with the two suppressed convective centers in the OLR field. There is no direct linkage between the WNP and equatorial regions (Figure 5f). It can be seen from $\chi$ analysis in Figures 5d and 5e that the equatorial branch of the descending motion within the western Pacific is a result of the anomalous large-scale Walker circulation. In the $\chi$ field, however, the convergent flow over the WNP is not as strong as that in the maritime continent/eastern Australia (Figure 5d). The result suggests that unlike the 2–3 year oscillation, the subsidence in the WNP is not directly

**Figure 3.** Regressed summer-mean fields of stream function at (a, d) 200 hPa and (c, f) 850 hPa and (b, e) geopotential height with significant wind fields at 500 hPa for the 2–3 year (Figures 3a, 3c, and 3e) and 3–5 year (Figures 3b, 3d, and 3f) oscillations of WNPSH. Shaded (vectors) areas pass the 90% confidence level.
attributed to the equatorial forcing in the 3–5 year oscillation. Rather, it may be forced through a Rossby wave response to anomalous heating associated with the local cold SSTA in Figure 4b [Sui et al., 2007; Wu et al., 2010].

3.3. Vertical Structures of the WNPSH Anomalies

[28] The analysis above indicates that the large-scale circulations associated with an enhanced summertime WNPSH in the 2–3 year and 3–5 year oscillations are distinctly different. The 2–3 year oscillation is accompanied by an anomalous local Hadley circulation that connects the maritime continent and the WNP, while the 3–5 year oscillation is characterized by the anomalous Walker circulation and a cold SSTA to the southeast of the anomalous anticyclone center. To reveal the specific processes that affect WNPSH, the horizontal and vertical structures of various dynamic and thermodynamic variables associated with the two oscillations are further compared.

[29] First, the relative zonal position between the anomalous anticyclone and descending motion centers during a strong WNPSH phase are examined by depicting the regressed $\bar{v}\_{850}$ (contours) and $\bar{\omega}\_{500}$ (blue shades) fields in Figure 7. Only the stronger (at the 95% significant level) portion of the regressed positive $\bar{\omega}\_{500}$ field is shown in Figure 7. While the subsidence center (suppressed convection in Figure 5c) in the 2–3 year oscillation is zonally in phase with the anticyclone center, the maximum subsidence (suppressed convection in Figure 5f) in the 3–5 year oscillation is located to the southeast of the anomalous anticyclone center. Such a phase difference suggests that the anomalous WNPSH in the former is possibly a direct response to the local Hadley circulation anomaly, whereas in the latter it is a baroclinic Rossby wave response to a negative heating anomaly induced by the local cold SSTA (Figure 4b).

[30] To reveal the vertical structure of the anomalous WNPSH, the domains of the strongest downward motion and anomalous subtropical high center on the 2–3 year and 3–5 year timescales respectively were chosen. The vertical profiles of the domain-averaged regressed fields of vertical $p$ velocity, relative vorticity, $Q_1$ and $Q_2$ are computed at each pressure level.

[31] Figure 8a shows the regressed vertical $p$ velocity profiles over the area of strongest downward motion in the 2–3 year and 3–5 year oscillations. The strongest subsidence is located in the upper troposphere (300 hPa) in the 2–3 year oscillation but in the middle troposphere (400–500 hPa) in the 3–5 year oscillation. The magnitude of the subsidence in the 3–5 year oscillation is weaker than that in the 2–3 year oscillation. Figure 8b shows the vertical profiles of the regressed relative vorticity fields over the anomalous anticyclone center associated with the 2–3 year and 3–5 year oscillations.
oscillations. While it exhibits a clear equivalent barotropic structure in the 2–3 year oscillation, the vorticity has a baroclinic structure in the 3–5 year oscillation, changing from negative anomalies in the lower troposphere to a positive anomaly in the upper troposphere (above 250 hPa). This result is consistent with circulation patterns over the WNP area shown in Figures 3a, 3c, 3d, and 3f.

The vertical distributions of regressed $Q_1$ and $Q_2$ anomalies averaged over the strongest descent region are shown in Figures 8c and 8d. Note that the amplitude of the negative $Q_1$ anomaly in the middle-upper troposphere is greater in the 2–3 year oscillation than in the 3–5 year oscillation (Figure 8c), while the amplitude of the negative $Q_2$ anomaly in the lower troposphere is greater in the 3–5 year oscillation than in the 2–3 year oscillation (Figure 8d). This implies that enhanced longwave radiative cooling due to the reduction of the vertically integrated moisture caused by the descent-induced dry advection plays an important role in the establishment of the anomalous WNPSH on the 2–3 year timescale. On the other hand, a peak of anomalous $Q_2$ at 850 hPa in the 3–5 year oscillation implies strong negative condensation heating at the top of the boundary layer. The negative condensation heating is induced by the local cold SSTA over the region of (150°–170°E, 15°–20°N) that is to the southeast of the anomalous anticyclone and is responsible for a baroclinic vertical structure.

In both the oscillations the warm SSTA under the anomalous anticyclone may be regarded as a passive response to the atmospheric forcing [Wang et al., 2005; Wu and Kirtman, 2005; Zhou et al., 2008]. The anomalous anticyclone leads to the suppression of convection and thus the increase of downward solar radiation into the ocean surface. Meanwhile, less evaporation due to the weakening wind speed may also increase SST.

The distinctive horizontal and vertical structures above suggest that the physical mechanisms responsible

Figure 5. Regressed summer-mean velocity potential and divergent wind at (a, d) 200 hPa and (b, e) 850 hPa and (c, f) upward longwave radiation flux (similar to outgoing longwave radiation) for the 2–3 year (Figures 5a, 5c, and 5e) and 3–5 year (Figures 5b, 5d, and 5f) oscillations of WNPSH. The red (blue) contours denote positive (negative) velocity potential anomalies or enhanced (suppressed) convection. Shaded regions pass the 90% confidence level.
for the anomalous WNPSH may differ in the 2–3 year and 3–5 year oscillations. According to the Gill [1980] model, a Rossby wave response is located to the west of a heating source. This corresponds well to the 3–5 year oscillation case, in which there is a phase shift between the anomalous anticyclone and the anomalous heating (or vertical motion). Further, the Rossby wave response to the anomalous heating is baroclinic. These features are accurately observed in the 3–5 year oscillation. Therefore the diagnosis result in this study suggests that the variation of the WNPSH on the 3–5 year timescale is primarily caused by an atmospheric Rossby wave response to a negative heating associated with a cold SSTA to the southeast of the anticyclone center. In contrast, in the 2–3 year oscillation, the anomalous WNPSH center is zonally colocated with the vertical motion, and its vertical structure is approximately barotropic. This implies that the Rossby wave response is unlikely and the downward motion is a major driving mechanism. That is, the strengthening of the WNPSH on the 2–3 year timescale is primarily attributed to the radiative cooling due to the descent-induced dry advection; the less clouds and water vapor in the atmosphere, the more longwave radiation is emitted into space. The change of local vertical motion and circulation in the WNPSH is induced by the remote forcing via anomalous local Hadley circulation associated with enhanced convective heating in the maritime continent.

[35] The regressed results above are consistent with the summertime subtropical LOSECOD diabatic heating pattern identified by Wu and Liu [2003] and Wu et al. [2009], who showed that the dominant diabatic heating sources over the northwestern Pacific region are convective heating and longwave radiation cooling. The corresponding diagnoses of Figures 5, 7, and 8 indeed demonstrate that in the 2–3 year oscillation, the longwave radiation cooling associated with the anomalous descent by the local Hadley circulation prevails over the negative convective heating; while in the 3–5 year mode, the negative convective heating due to the cold SSTA prevails over the longwave radiation cooling. The negative convective heating will induce an atmospheric Rossby wave response to the west as the anomalous WNPSH. As a consequence, the subtropical anticyclone anomaly of the 2–3 year oscillation possesses a barotropic structure, whereas in the 3–5 year oscillation possesses a baroclinic structure as presented in Figure 8b. This is

Figure 6. Regressed vertical velocity (contours) and divergent wind fields (a) averaged over 110°E–140°E for 2–3 year and (b) averaged over 110°–150°E for 3–5 year oscillations of WNPSH. Shaded regions pass the 90% confidence level.
because a convective heating in the subtropics can generate a Sverdrup balance stream function pattern with anticyclone (cyclone) to the east and cyclone (anticyclone) to the west of the heating (negative heating) in the lower troposphere and the opposite in the upper troposphere, as discussed by Wu and Liu [2003], Liu et al. [2004], and Wu et al. [2009]. Therefore the distinct difference between the two oscillation modes should be attributed to the different dominant diabatic heating over the northwest Pacific.

4. Model Experiment Results

To test the hypothesis that the WNPSH variability on the 3–5 year timescale is primarily forced by local SSTA in the WNP while the WNPSH variability on the 2–3 year timescale is attributed to the forcing over the maritime continent, the following control and sensitivity experiments have been developed. In the control run (CTRL), ECHAM4 was integrated for 20 years, forced by monthly climatological SST (based on the period of 1955–2000). In the WNP run, the model was forced by the climatological SST plus a uniform SSTA of −1°C over a WNP domain (150°E–160°W, 10°–25°N) (blue box in Figure 9a). In the MC run, the same simulation strategy was used as in the WNP run except with a +1°C SSTA specified in the maritime continent region (90°–150°E, 20°S–5°N) (red box in Figure 9b). The amplitude of this specified SSTA in the WNP and MC regions is 2–3 times larger than the composite SSTA field associated with the observed high WNPSH indices, in order to obtain a robust response. The WNP and MC runs each consists of seven ensemble members which have initial conditions starting from 27 April 0000 UTC to 30 April 0000 UTC, separating by 12 h apart. All experiments were integrated from May to August. The atmospheric response to the local and remote SSTA forcing is estimated based on the difference between the ensemble mean of the WNP/MC run and the CTRL run (i.e., WNP-CTRL and MC-CTRL).

Figure 7. Regressed summer-mean stream function at 850 hPa (contour) and positive $p$ velocity (only most significant descending region is indicated here by shading) at 500 hPa for (a) the 2–3 year and (b) 3–5 year oscillations of WNPSH.
The amplitude of the simulated precipitation and anticyclonic circulation is too strong, due to the greater SSTA specified. In the observation, the SSTA in the WNP decreases gradually from June to August [Wu et al., 2010].

The strong positive SSTA forcing in the MC induces a positive precipitation anomaly over the northern part of the SSTA domain and a negative precipitation anomaly in the northern Philippines along 15°N (Figure 9b). An anomalous anticyclone is simulated near Taiwan and is located to the north of the negative precipitation anomaly center, similar to the observed anticyclone–vertical motion phase relation. In the tropical western Pacific, anomalous easterly flows in the southern part of the anticyclone extend to the north IO and converge toward the MC region. The simulated patterns in the MC run are quite similar to those observed shown in Figure 3c, 5c, and 7a but with less zonal extent and a slightly northward shift due to the neglect of tropical central Pacific cold SSTA forcing in Figure 4a.

It is seen from Figures 9a and 9b that the spatial phase patterns between the anomalous precipitation center
and the anticyclone center in both the WNP and MC runs are similar to those observed. In addition, the vertical profiles of the vertical velocity and relative vorticity fields averaged over the maximum precipitation and anticyclonic vorticity center in both the WNP and MC runs were evaluated. In Figure 9c the anomalous descending motion in the MC run is larger than that in the WNP run. The strongest descending motion is located in the middle troposphere around 400–600 hPa in the MC run but at 500–700 hPa in the WNP run. In the lower troposphere below 850 hPa, the amplitude of the descent in the WNP and MC run is slightly reversed as in the observation of NCEP/NCAR R-1 (Figure 8a). In addition to the p velocity, the vertical distributions of relative vorticity over the anticyclone center in the WNP and MC run also shows a distinct difference, similar to the observation. The anomalous anticyclone in the MC run (Figure 9d) is associated with the negative vorticity throughout the troposphere that resembles the observed barotropic structure in the 2–3 year oscillation (Figure 8b).

The horizontal and vertical structures of the anomalous circulation in the idealized numerical experiments accurately capture the observed characteristics in the 2–3 year and 3–5 year oscillations. Numerical experiments in the study demonstrate the role of SSTA forcing in the WNP and MC regions in generating the anomalous anticyclone associated with the westward extension of the WPNSH in the two oscillations. The baroclinic vertical structure in the WNP run confirms that a Rossby wave response to the local cold SSTA is a key process involved in the 3–5 year oscillation. Since the cold SSTA in the WNP weakens from early to late summer when El Nino is decaying in the 3–5 year oscillation, the SSTA forcing in the tropical IO may also contribute to the westward movement of the subtropical high [Xie et al., 2009; Wu et al., 2010]. The numerical experiment of Zhou et al. [2009b] has also shown that a warmer tropical IO may contribute to the westward extension WPNSH in the interdecadal time scale. In an additional idealized experiment with a +1°C SSTA specified in the tropical IO basin (20°S–20°N, figure not shown), an anomalous low-level anticyclonic circulation is simulated in the WNP, with a maximum center located slightly southwest over the South China Sea. Since the simulated anticyclone differs from the observed WPNSH in its location (a southwestward shift from the observed center) and vertical structure (the simulated baroclinic structure is much weaker), the warm SSTA in the tropical IO may not be the major forcing in causing the westward shift of the WPNSH in the 3–5 year oscillation. In the MC run, the simulated horizontal pattern and the vertical structure resemble the observed in the 2–3 year oscillation. This confirms the role of the remote warm SSTA forcing in the MC in inducing the anomalous WPNSH through an anomalous local Hadley circulation as discussed in section 3.

5. Temporal Evolution Features

[41] To better understand the interannual variability of the WPNSH, the seasonal evolution of anomalous SST and circulation fields of $\omega_{500h}$, $\vec{u}_{500h}$, and $f_{500h}$ associated with the WPNSH 2–3 year and 3–5 year oscillations are examined. More than one full cycle (from MAM(0) to JJA(2)) and nearly one full cycle (from MAM(0) to JJA(3)) of regressed fields are presented for 2–3 year and 3–5 year oscillations. Seasons of MAM(1), JJA(1), SON(1) denote the concurrent spring, summer and autumn of the westward WPNSH, whereas the numbers 0, 2, and 3 in parentheses denote the lead and lag years relative to the current year (1).

5.1. SSTA and Vertical Velocity

[42] The evolution of regressed SSTA for the 2–3 year oscillation (Figure 10a) shows the following features. Starting from the preceding spring MAM(0), cold SSTA anomalies dominated the northern IO, MC, and northern Australia, while a weak warm SSTA appeared in the equatorial central Pacific. The warm SSTA in the central Pacific moved eastward and intensified from JJA(0) to SON(0), weakened afterward through MAM(1), and became negative by the current JJA(1). After JJA(1), the cold SSTA strengthened until D(1)JF(2) and then decayed from MAM(2) to JJA(2) when the cold SSTA turned into a weak warm SSTA. Accompanying the above evolution, a weak warm SSTA in the central IO in the previous summer JJA(0) grew through SON(0). By D(0)JF(1), the warm SSTA had spread widely to the northern IO, MC, and northern Australia. The broad-scale warm SSTA further extended southeastward from D(0)JF(1) to MAM(1) and eastward from MAM(1) to SON(1). By the following summer JJA(2), the SSTA had evolved to a pattern similar to that in MAM(0). The above evolution feature resembles the quasi-biennial oscillation of the tropical SSTA signal found in previous studies [e.g., Barnett, 1991; Li et al., 2006].

[43] Contrary to the 2–3 year oscillation, the evolution of the 3–5 year oscillation shows a standing oscillation, with the SSTA persisting from MAM(0) to SON(1) and reversing its sign from D(1)JF(2) to JJA(3) (Figure 10b). A warm SSTA developed in the central eastern Pacific from MAM(0), reached its warmest phase in D(0)JF(1), and then weakened in the following three seasons to D(1)JF(2) when a weak cold SSTA grew. The cold SSTA became stronger from MAM(2) to D(2)JF(3), weakened in MAM(3), and continued to last for two seasons until SON(3) (figure not shown). Accompanying the warm SSTA in the equatorial Pacific, the surrounding cold SSTA to the northwest and southwest also persisted from JJA(0) to SON(1). In D(1)JF(2), the areas surrounding the equatorial Pacific were replaced by a warm SSTA that can persist to JJA(3). In addition, warm SSTA anomalies developed in the IO from D(0)JF(1) strengthened and extended to the WNP and marginal seas along the East Asia coast through MAM(1) to SON(1). In the following seasons, weak cold SSTA anomalies appeared in the IO from MAM(2), extended to the north and south from JJA(2) to D(2)JF(3), and reached the strongest intensity over the whole ocean basin in JJA(3). The evolution of the regressed SSTA
Figure 9. Anomalous precipitation (shaded, only areas passing the 90% confidence level plotted) and 850 hPa stream function in (a) the WNP run and (b) the MC run, and vertical profiles of (c) anomalous vertical $\rho$ velocity over the strongest descending region and (d) anomalous relative vorticity over the anticyclone center in the WNP (blue line) and the MC (red line) runs. The thick blue and red box in Figures 9a and 9b shows the region where uniform (either −1°C or +1°C) SSTA is specified.
patterns described above resembles a typical El Nino evolution with a long life cycle [e.g., Wang et al., 2000, 2003]. The 3–5 year oscillation of the WNPSH may be modulated by the remote low frequency El Nino SSTA forcing [Wu and Zhou, 2008] via local air-sea coupling in WNP [Wang et al., 2000; Sui et al., 2007].

[44] The time-longitude section of the regressed SST and \( \omega_{500} \) fields along the Equator from MAM(0) to JJA(2) (to JJA(3) in 3–5 year oscillation) is shown in Figure 11. For the 2–3 year oscillations (Figures 11a and 11c), there is an obvious eastward propagation of the warm SSTA and upward motion from the IO to the western Pacific during the period from JJA(0) to JJA(2). The warm SSTA propagated to the tropical western Pacific in D(0)JF(1) where the upward motion developed two seasons later in JJA(1). In the central eastern Pacific, warm SST (upward motion) existed from MAM(0) to MAM(1) and was followed by a cold SST (downward motion) from JJA(1) to JJA(2) with a slight eastward propagation. The center of the upward motion within (170°W–150°W) is located to the west of the warm SSTA in the region (150°W–120°W) over the tropical central eastern Pacific. However, over the tropical western Pacific, the two fields do not show a significant phase difference.

[45] The regressed fields for the 3–5 year oscillation (Figures 11b and 11d) exhibit a standing band pattern characterized by a warm (cold) SSTA and upward (downward) motion in the IO, a cold (warm) SSTA and downward (upward) motion in the western Pacific, and a warm (cold) SSTA with upward (downward) motion in the central eastern Pacific from MAM(0) to D(1)JF(2) [from MAM(2) to SON(3)]. Although warm (cold) SST anomalies are generally accompanied by upward (downward) motion in the ocean basins, this feature is not so clear in the IO where a significant SST is only associated with weak vertical motion. In the tropical western Pacific, the strongest descending (ascending) motion lagged the strongest cold (warm) SST by 1 month, developing in JJA(0) (MAM(2)), reaching the strongest intensity in D(0)JF(1) [D(2)JF(3)], and then persisting to JJA(1) (JJA(3)).

[46] The spatial and temporal evolutions of the SST in the 2–3 year and 3–5 year oscillations are consistent with the study by Barnett [1991], who found that the near-equatorial characteristics of the quasi-biennial SST/SLP mode are those of a quasi-progressive wave while the 3–7 year mode is closer to a standing wave.

5.2. Stream Function and Wind

[47] The examination of temporal evolution of the regressed \( \chi_{200} \) (figures not shown) and \( \omega_{500} \) (Figure 11c) from JJA(0) to SON(1) indicates that in the 2–3 year oscillation, a large-scale upper-level divergence (ascending) was located in the eastern Pacific and northern IO from JJA(0) to MAM(1), while a large scale upper-level convergence (descending) appeared over the western Pacific. The pattern was nearly out of phase in JJA(1), with a dominant upper-level divergence near Java and a large-scale convergence over the tropical central eastern Pacific (Figure 5a). It should be noted that the \( \chi_{200} \) pattern has no significant signal in the transition season MAM(1). This implies that the signal from the preceding boreal summer JJA(0) to winter D(0)JF(1) may have difficulty to pass through the spring season MAM(1) on the 2–3 year oscillation. This differs from the 3–5 year oscillation, in which the large-scale upper-layer convergent flow over the tropical western Pacific persisted from the preceding summer JJA(0) to the current summer JJA(1).

[48] The evolutions of the regressed \( v_{850} \) and \( V_{850} \) fields (Figure 12) are, in general, consistent with the \( \chi_{200} \) field. In the 2–3 year oscillation (Figure 12a), a strong anomalous anticyclone existed over southern Asia in SON(0) and moved to the Philippine Sea in D(0)JF(1). The anomalous anticyclone weakened and disappeared in MAM(1) but reemerged over the WNP in JJA(1), again indicating a spring barrier for the Philippine Sea anticyclone to pass through the boreal spring on this timescale. The local Hadley circulation in JJA(1), therefore, is essential for the development of the summer anticyclonic circulation over the WNP on this timescale. The evolution is also consistent with the vertical velocity fields that descending motion within the tropical western Pacific weakens in MAM(1), where the ascending motion becomes stronger and statistically significant in JJA (1) within the green box in Figures 11a and 11c.

[49] In the 3–5 year oscillation (Figure 12b), however, the anomalous anticyclone over the Philippine Sea and low-level northeasterly wind over WNP persisted from D(0)JF(1) to JJA(1), consistent with the \( \chi_{200} \) SSTA and vertical velocity pattern in Figure 10b and Figures 11b and 11d. A positive thermodynamic air-sea feedback can explain how the anomalous anticyclone is maintained from the preceding winter to the current summer [Wang et al., 2000, 2003; Sui et al., 2007]. Here the anomalous anticyclone, initiated by a cold SSTA as a Rossby wave response to a negative heat source in the boreal winter, may further reinforce the cold SSTA through the wind-evaporation-SST feedback under prevailing climatological northeasterly trade winds to the spring. Although the seasonal change of prevailing winds leads to a negative air-sea feedback and thus a weakening of the local cold SSTA in the boreal summer, the negative SSTA may still have a significant impact on the WNP monsoon, particularly in early summer. Thus this local cold SSTA may have a large impact on the summertime WNPSH variability.

6. Conclusion and Discussion

[50] The western edge of the summertime Pacific subtropical high in the Northern Hemisphere (WNPSH) exhibits a significant zonal shift on the interannual timescale. The largest variability is found in the western North Pacific. This variability is represented in this study by an index defined by geopotential height anomalies at 500 hPa averaged in the region where the largest WNPSH variability occurs. A power spectrum analysis of the WNPSH index indicates two dominant peaks at 2–3 year and 3–5 year periods. The oscillation characteristics of the WNPSH at the two periods are investigated through a regression analysis of atmospheric and oceanic variables against the two WNPSH indices at the 2–3 year and 3–5 year periods. The idealized numerical experiments with the ECHAM4 model are also carried out to examine the dominated SST and associated mechanisms for the WNPSH variability on the two timescales. Since the location and strength of the subtropical anticyclone greatly change from June to August, the conclusion obtained from
Figure 10. (a) Evolutions of the SSTA field from the preceding spring (MAM(0)) to the summer of the following year (JJA(2)) regressed against the WNPSH index for the 2–3 year oscillation. (b) Evolutions of the SSTA field from the preceding spring (MAM(0)) to the summer of the year after next year (JJA(3)) regressed against the WNPSH index for the 3–5 year oscillation.
The 2–3 year oscillation of the WNPSH and the associated SST and circulation possess a coherent eastward propagating feature. The primary signals include the southeastward spread of a warm SSTA and ascending motion from the Indian Ocean in the preceding autumn, SON(0), to the maritime continent in the current summer JJA(1) of the high WNPSH index. In the central and eastern Pacific, the SSTA changes its sign from positive in the preceding summer through winter to negative in JJA(1), analogous to the La Nina development condition. In the current summer, the WNPSH is characterized by an anomalous anticyclonic circulation and descending motion north of the Philippines. The descent is a part of the anomalous local Hadley circulation, with enhanced convection and a warm SSTA appearing in the maritime continent. A heat budget analysis reveals that the descent and anticyclone are maintained primarily by radiative cooling, and the corresponding vertical structure is equivalently barotropic. The anticyclonic circulation in the WNP is accompanied by an anomalous cyclonic circulation to its north. As a result, the overall structure appears like a meridionally oriented PJ type pattern. The role of the remote SSTA forcing in the maritime continent in affecting the WNPSH variability is verified by the idealized ECHAM4 model experiment.

The 3–5 year oscillation of the WNPSH exhibits a standing wave feature, that the warm SSTA in the equatorial central eastern Pacific, its surrounding cold SSTA to the northwest and southwest, and a warm SSTA in the Indian Ocean and along the East Asian marginal seas, persisting from the preceding winter to the current summer of a high WNPSH index. Along with the evolution above, the ascending motion over the tropical eastern central Pacific and Indian Ocean and the descending motion over the western Pacific also persist from the preceding autumn to the current summer. These features resemble the tripole SSTA pattern and associated overturning Walker circulation.

Figure 11. (a, b) Longitude-time section of regressed SSTA and (c, d) 500 hPa vertical p velocity anomaly fields averaged between 15°S and 15°N from MAM(0) to JJA(2) for 2–3 year oscillation (Figures 11a and 11c) and from MAM(0) to JJA(3) for 3–5 year oscillation (Figures 11b and 11d). Contours in warm (cold) colors indicate warm (cold) SSTA or downward (upward) motion. Shaded regions pass the 90% confidence level. Green boxes denote the current summer of a strengthened WNPSH.
pattern frequently observed during the decaying phase of El Nino episodes [e.g., Wang et al., 2000, 2003]. For the 3–5 year oscillation, the maximum descent of the WNP is slightly east of the anomalous anticyclone center. The greatest descent appears in the lower troposphere, accompanied by a negative latent heating and a cold SSTA to the southeast. The corresponding vertical structure of relative vorticity in the descent region bears a baroclinic nature, and the zonal phase difference between the anomalous anticyclone center and the descent implies a Rossby wave response to a negative heating anomaly in association with the local cold SSTA. The local SSTA forcing mechanism is confirmed by the idealized ECHAM4 model experiment.

[53] Our findings of the 2–3 year and 3–5 year oscillations of the WNPSH are consistent with previous studies of Barnett [1991] and Wang et al. [2001] who found that there are quasi-biennial and low-frequency variabilities in the tropical SST/SLP fields and in Asian monsoon circulation fields. Our findings are also consistent with the observational fact that ENSO has both the quasi-biennial and lower-frequency spectral peaks. The consistency above indicates that the variability of the WNPSH connected with the Asian Monsoon system might be closely linked to the tropical SSTA forcing in the interannual time scale. The possible mechanisms that connect the subtropical high and tropical SST include the monsoon–warm ocean interaction associated with the TBO [Li et al., 2006], ENSO–East Asia teleconnection [Wang et al., 2000; Sui et al., 2007], and a season-dependent Indian Ocean impact on the WNP monsoon [Wu et al., 2009b].

[54] In addition to the tropical forcing, midlatitude processes may also play a role in the interannual variability of

Figure 12. Regressed 850 hPa stream function and wind fields from the preceding summer (JJA(0)) to the current autumn (SON(1)) for (a) the 2–3 year and (b) 3–5 year oscillations of WNPSH. Shadings and vectors denote regions passing the 90% confidence level.
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Acknowledgments

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Chou, C. (2004), Establishment of the low frentics in the WNPSH [e.g., Enomoto, 2004]. Since the present study focus on the relationship between the WNPSH and tropical forcing, the influence of the midlatitude processes have not been discussed here. OEHAM4 model simulation confirms that the 2–3 year mode is driven by the WNPSH over the maritime continent, while the 3–5 year mode is driven by the local SST in the WNPSH. Despite the model results in support of the hypothesized physical mechanisms in the 2–3 year and 3–5 year oscillations, the role of the air-sea interaction in the WNPSH region, where the atmospheric feedback to the SSTA is regarded as an important process, have not been illustrated. Further observational analyses and coupled GCM modeling studies are needed to investigate the relative importance of these factors and their causality in causing the interannual variability of the WNPSH from dynamic and thermodynamic discussions.


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