Influences of the Pacific–Japan Teleconnection Pattern on Synoptic-Scale Variability in the Western North Pacific

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ABSTRACT
This study investigates the influences of the Pacific–Japan (PJ) teleconnection pattern on synoptic-scale variability (SSV) in the western North Pacific (WNP). The PJ pattern exhibits salient intraseasonal variations, with a dominant peak at 10–50 days. During positive PJ phases, strengthened SSV is found in the WNP, with a much stronger and better organized synoptic wave train structure. Such a synoptic-scale wave train, however, is greatly weakened during negative PJ phases. Examination of the vertical profiles of the observational data suggests that environmental parameters are generally more (less) favorable for the growth of synoptic disturbances under positive (negative) PJ conditions.

Observational results are further verified with an anomaly atmospheric general circulation model, which reveals faster (slower) growth of the synoptic-scale wave train when the environmental anomalies associated with positive (negative) PJ phases are incorporated into the summer mean state of the model. In addition, sensitivity experiments indicate that thermodynamic parameters of the planetary boundary layer (PBL) play a determining role in controlling the development of synoptic disturbances in the WNP. The increase (decrease) in background PBL moisture during positive (negative) PJ phases enhances (suppresses) perturbation moisture convergence and thus the convective heating associated with SSV, leading to strengthened (weakened) synoptic-scale activity in the WNP. Serving as potential seed disturbances for cyclogenesis, the strengthened (weakened) synoptic-scale activity may also contribute to the enhancement (suppression) in intraseasonal TC frequency during positive (negative) PJ phases.

1. Introduction
Over the tropical western North Pacific (WNP), synoptic-scale variability (SSV), including tropical cyclones (TCs) and tropical depression (TD)-type disturbances, has been shown to be closely related to the boreal summer intraseasonal oscillation (ISO). Gray (1979) found that global TC activity occurs in clusters, with 1–2 weeks of active TC formation followed by a similar period of quiescence. Nakazawa (1988) showed that synoptic-scale disturbances are embedded and enhanced within “super clusters” during the convective phase of the ISO. Subsequent studies have demonstrated that TC genesis and activity in different regions, including the WNP (Liebmann et al. 1994; Kim et al. 2008; Li et al. 2012; Li and Zhou 2013a,b), the eastern Pacific (Maloney and Hartmann 2000), and the Indian Ocean (Liebmann et al. 1994; Bessafi and Wheeler 2006), are significantly modulated by the ISO. In addition, changes in synoptic TC activity as well as TD-type disturbances can be attributed either to modifications of the background mean flow by the ISO (Liebmann et al. 1994; Kim et al. 2008; Zhou and Chan 2005; Li and Zhou 2013a,b) or to wave accumulation in terms of barotropic energy conversions (Maloney and Hartmann 2001; Maloney and Dickinson 2003; Mao and Wu 2010). Li et al. (2012) further

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suggested that the interannually varying background state related to the El Niño–Southern Oscillation (ENSO) can affect the ISO–TC relationship, resulting in enhanced TC modulations during El Niño events. More recently, Li and Zhou (2013a,b) examined and revealed the significant influences of the 10–20-day quasi-biweekly oscillation on TCs in the WNP. In addition to the ISO influences on SSV, a recent series of studies (Zhou and Li 2010; Hsu et al. 2011; Hsu and Li 2011) has also identified a two-way interaction between the ISO and SSV. Zhou and Li (2010) discovered that SSV can exert an upscale feedback to the ISO through nonlinear rectification of the surface latent heat flux, while Hsu et al. (2011) and Hsu and Li (2011) confirmed this by analyzing the barotropic energy conversion and nonlinear rectification of the apparent heat and moisture sources.

Apart from the ISO, there is also an extratropical wave train in boreal summer, which is characterized by anomalous convections over the Philippine Sea and East Asia. Such a wave train is termed the Pacific–Japan (PJ) teleconnection pattern (Nitta 1987, 1989) and has been previously shown to have significant impacts on summer climate variability in East Asia and Japan (Nitta and Hu 1996; Wakabayashi and Kawamura 2004; Kosaka et al. 2011). For example, Nitta and Hu (1996) noticed that coupled summer rainfall and temperature patterns in China are closely connected to a 500-hPa geopotential height anomaly associated with the PJ pattern. Wakabayashi and Kawamura (2004) found that the convective anomalies associated with the PJ pattern give rise to abnormally hot or cool summers in Japan. Kosaka et al. (2011) showed that the PJ pattern mediates the influences of ENSO from the preceding winter and contributes to the mei-yu–baiu precipitation variability in boreal summer. Recent studies focusing on the impacts on synoptic-scale disturbances have suggested that the summertime interannual variability of TC tracks in the WNP (Choi et al. 2010) and TC-related precipitation over Korea (Kim et al. 2012) are significantly modulated by the PJ pattern, while others (Kawamura and Ogasawara 2006; Yamada and Kawamura 2007) have proposed that TCs may also alter the PJ pattern by inducing a stationary Rossby wave train over the WNP. However, when compared with the well-documented relationship between the ISO and SSV, much less attention has been paid to the PJ–SSV relationship, and relevant studies are limited. It is not yet clear how and to what extent different PJ states affect the growth of SSV in the WNP. Therefore, in the present study, we will examine, in depth, the impacts of the PJ pattern on SSV based on both observational analysis and numerical studies.

The rest of the paper is organized as follows: data and methodology are described in section 2. Section 3 identifies the dominant time scale and reveals the basic characteristics of the PJ pattern, while section 4 investigates its impacts on SSV. Section 5 describes the numerical experiments carried out to further verify the results, while section 6 examines the possible influences of the PJ pattern on TCs. Finally, a summary and discussion are given in section 7.

2. Data and methodology

a. Data

Daily atmospheric data including wind, geopotential height, air temperature, omega, specific humidity, and relative humidity for 1979–2010 were taken from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996), while TC datasets were acquired from the Joint Typhoon Warning Center (http://www.usno.navy.mil/NOOC/nmfc-ph/RSS/jtwc/best_tracks/). The present study focuses on boreal summer [June–August (JJA)] when SSV is the strongest in the WNP (Lau and Lau 1990; Li 2006; Tam and Li 2006). Following previous studies (Nitta 1989; Wakabayashi and Kawamura 2004; Choi et al. 2010; Kim et al. 2012), a PJ index, which is defined as the difference in 850-hPa geopotential height anomalies \(Z_{850}\) between a grid point east of Taiwan (22.5°N, 125°E) and a grid point east of Japan (35°N, 155°E), is used to describe the state of the PJ pattern:

\[
\text{PJ index} = \left[ Z_{850}(155\degree E, 35\degree N) - Z_{850}(125\degree E, 22.5\degree N) \right] / 2.
\]

Despite its simplicity, the PJ index has been shown to be applicable and robust in capturing the essential circulation features associated with the PJ teleconnection (Wakabayashi and Kawamura 2004) and has been previously employed by a number of studies for PJ-related diagnosis (Wakabayashi and Kawamura 2004; Choi et al. 2010; Kim et al. 2012). A positive PJ pattern is thus associated with negative height anomalies (enhanced convection) in the subtropics over the Philippines and positive height anomalies (suppressed convection) in the midlatitudes over Japan, whereas a negative PJ pattern is simply the reverse (Nitta 1987, 1989; Wakabayashi and Kawamura 2004). A 10–50-day Lanczos filter is used throughout the study to extract the intraseasonal signals associated with the PJ pattern. This frequency band is chosen based on spectral analysis, which will be discussed in detail in section 3. In addition, a 3–8-day filter is applied to extract signals related to synoptic-scale variability.
b. The anomaly model

An anomaly atmospheric general circulation model (AGCM) (Li 2006) based on the Princeton dynamic core (Held and Suarez 1994) is also employed for this study. This is a multilevel baroclinic model with five evenly distributed sigma levels with an interval of 0.2 and a horizontal resolution of T42. The model consists of primitive equations linearized by a realistic three-dimensional basic state but it retains nonlinearity in the second-order perturbation terms of the prediction equations, such that the evolution of a specified initial perturbation under an idealized or realistic mean basic state can be examined (refer to the appendix for a more detailed description of the model). Basic inputs into the model include the three-dimensional summer mean state of \( u, v, T, \) and \( P_s \), obtained from the long-term average of the NCEP–NCAR reanalysis. Apart from the three-dimensional dynamic parameters, the planetary boundary layer (PBL) thermodynamic factors including the specific humidity \( q_b \) and temperature \( T_b \) are also incorporated in the model to calculate the heating term. To represent the interactive nature of the circulation and convection, an interactive heating scheme is prescribed in the model (Kuo 1974; Wang and Li 1993; Li 2006) in which the perturbation convective heating is proportional to the PBL specific humidity, as well as the perturbation divergence:

\[
Q' = -\alpha \delta q_b d' f(\sigma),
\]

where \( q_b \) is the mean state PBL specific humidity, \( d' \) is the PBL perturbation divergence, and \( \alpha \) denotes the heating coefficient representing the strength of the convection–frictional convergence (CFC) feedback. For current study, \( \alpha \) is set as such that a PBL convergence of \( 10^{-7} \) s\(^{-1}\) corresponds to a heating rate of \( 1^\circ\text{C day}^{-1} \), which is reasonable compared to the observed climatological monthly value over the intertropical convergence zone. In the above equation \( \delta \) is a SST-dependent

![FIG. 1. (left) Power spectrum of the summertime PJ index based on the entire period from 1979 to 2010 and (right) the mean power spectrum averaged over each of the 32 summers. The green dashed line denotes the Markov red noise spectrum, while the red and blue dashed lines represent the 95% and 5% confidence levels, respectively.](image1)

![FIG. 2. Standardized time series of the original (bar chart) and 10–50-day filtered (dotted line) PJ index during (top) 1996 and (bottom) 1997.](image2)
coefficient (Wang and Li 1993), such that $\delta = 1$ when $\text{SST} > 29.5^\circ\text{C}$, $\delta = 0$ when $\text{SST} < 26.5^\circ\text{C}$, and $\delta = (\text{SST} - 26.5^\circ\text{C})/3$ when $26.5 < \text{SST} < 29.5^\circ\text{C}$. Such SST criteria are employed based on the observed convection–SST relationship. As shown by Waliser et al. (1993), deep convection rarely occurs when SST is below $26.5^\circ\text{C}$; for SST between $26.5^\circ\text{C}$ and $29.5^\circ\text{C}$, convective activity increases quasi linearly with SST; and for SST greater than $29.5^\circ\text{C}$, the frequency of occurrence of deep convection tends to level off. A similar SST threshold has also been applied by Sobel et al. (2002) and Huang and Huang (2009) to study the temperature response to ENSO SST. Finally, $f(\sigma)$ represents the heating profile in the vertical direction, with the maximum value being specified in the middle troposphere ($\sigma = 0.5$) and is multiplied by coefficients of $0.2, 0.7, 0.5,$ and $0.1$ at $\sigma = 0.1, 0.3, 0.7,$ and $0.9$, respectively, to imitate the deep convection in tropics (Li 2006; Chen 2012) in the present study.

In this model, Rayleigh friction, with a damping rate of 1 day in the lowest model level ($\sigma = 0.9$), is used to mimic the PBL dissipation, while Newtonian cooling with an $e$-folding time scale of 10 days is applied to the temperature equation at all model levels. Since the present study focuses mainly on the impacts of the PJ pattern on tropical SSV, a strong damping rate of 1 day is applied to the perturbation momentum and temperature equations.

FIG. 3. Composites of 10–50-day filtered 850-hPa geopotential height (contours, m; values over 95% confidence are shaded) and wind (vectors, m s$^{-1}$; only values over 95% confidence are shown) anomalies for positive PJ phases. Day 0 is the day when the filtered PJ index attains its maximum value, while day $n (-n)$ refers to $n$ days after (before) day 0.
over extratropical regions (beyond 40°N and 40°S) at all levels. The effect of atmospheric transients on the mean basic state can be considered negligible in the tropics, which has also been previously justified by Hirota and Takahashi (2012). Wang et al. (2003) used this model to examine the equatorially asymmetric atmospheric response to a symmetric forcing, and Jiang and Li (2005) used it to investigate the initiation of the Madden–Julian oscillation (MJO) in the Indian Ocean. In addition, Li (2006) used the model to study the origin of the summertime synoptic-scale wave train and the associated role of the CFC feedback, while Chen (2012) applied the model to look into the transitions of equatorial mixed Rossby–gravity waves to off-equatorial TD disturbances under different ENSO backgrounds.

3. Dominant time scale and basic characteristics of the PJ pattern

Before diagnosing the influences of the PJ pattern on synoptic-scale variability, the following section examines the basic characteristics of the PJ pattern. To identify the dominant time scale of the PJ pattern, two different approaches of the spectral analysis have been applied (Fig. 1). In the first method the power spectrum is derived from the summertime PJ index based on the entire time series, while in the second method the mean power spectrum is computed by averaging the individual spectrum over each summer. From Fig. 1, it is clear that both spectra reveal a prominent peak at the intraseasonal time scale at 10–50 days. A closer examination of the individual PJ time series confirms this (Fig. 2). The original PJ time series exhibits clear intraseasonal variations, which can be accurately captured by the 10–50-day filtered time series. Indeed, the filtered time series can explain up to 50% of the total variance of the original time series. Therefore, in the rest of this study, we will focus primarily on this frequency band. A positive and negative PJ phase is defined using one standard deviation of the filtered time series as a threshold.

Next we examine the circulation features associated with different PJ phases. Figure 3 illustrates the evolution of the 850-hPa geopotential height and wind anomalies during positive 10–50-day filtered PJ phases. Four days prior to the peak of positive PJ phases, negative geopotential height anomalies and cyclonic circulations

![Fig. 4. Composites of 10–50-day filtered geopotential height (contours, m; values over 95% confidence are shaded) and wind (vectors, m s⁻¹; only values over 95% confidence are shown) anomalies for positive PJ phases at (a) 850 hPa, (b) 500 hPa, and (c) 200 hPa. (d–f) As in (a–c), but for negative 10–50-day filtered PJ phases.](https://example.com/fig4.png)
begin to develop in the subtropics, while positive height anomalies and anticyclonic circulations appear in the midlatitudes. Such a pattern becomes more mature and attains its maximum in day 0, characterizing the peak of positive PJ phases. The alternating circulation anomalies then start to weaken during day 4 and are replaced by an opposite pattern during day 8, with the whole cycle being completed at about 30 days. The circulation features associated with negative PJ phases are similar but with a reversed sign (figure not shown). To illustrate the vertical structure of the PJ pattern, Fig. 4 further shows the associated circulation anomalies at different pressure levels for both positive and negative PJ phases. Consistent with previous studies (Nitta 1987, 1989; Kosaka and Nakamura 2006, 2010), the PJ-related circulation anomalies are approximately barotropic, though there is a bit of a northward tilt in the circulation center in the upper troposphere in the subtropics. Such a poleward tilt with height has also been noted previously by Kosaka and Nakamura (2010) using monthly datasets, who suggested that the interannual PJ pattern behaves as a moist dynamical mode, where anomalous circulations can effectively extract energy from the mean state and, in turn, enhance anomalous convective activities.

4. Influences of the PJ pattern on SSV

Section 3 reveals distinct differences in the circulation features for positive and negative PJ phases. It is thus anticipated that synoptic-scale variability may also vary under different PJ conditions. Figure 5 shows SSV (represented by the variance of the 3–8-day filtered relative vorticity anomalies) in the WNP and the associated deviation from the climatology during positive and negative 10–50-day filtered PJ phases. A noticeable feature here is the strengthened (weakened) SSV during positive (negative) PJ phases. To clearly elucidate the differences in SSV, an empirical orthogonal function (EOF) analysis is further conducted based on the 3–8-day filtered 850-hPa vorticity anomalies with respect to different PJ phases over the WNP ($0^\circ$–$30^\circ$N, 100$^\circ$E–180$^\circ$). Figure 6 shows the horizontal patterns of the first and second EOF modes of SSV during positive and negative 10–50-day filtered PJ phases. For both PJ phases, the dominant SSV pattern is a wave train with alternating circulation anomalies aligned in a northwest–southeast orientation. The wave possesses typical characteristics of a synoptic-scale wave train, with a wavelength of approximately 2500 km and a phase speed of about 4 m s$^{-1}$, which is consistent with previous studies (Lau

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**Fig. 5.** (a) SSV (denoted by the variance in the 3–8-day filtered 850-hPa relative vorticity anomalies, $10^{-12}$ s$^{-2}$) and (b) its deviation from climatology during positive 10–50-day filtered PJ phases; (c),(d) as in (a),(b), but for negative filtered phases.
and Lau 1990; Li and Wang 2005; Li 2006; Tam and Li 2006). The first two EOF modes are in quadrature, with the maximum correlation between the principal component (PC) time series occurring at a lag of 1 day. Together they describe the northwestward-propagating nature of the synoptic wave train in the WNP (Li 2006; Tam and Li 2006). Though the spatial pattern of the SSV looks similar, the amplitude of the SSV differs significantly when the PCs of the EOF mode are regressed onto the atmospheric fields. As shown in Fig. 7, the amplitude of the synoptic wave train in positive PJ phases is much stronger than that in negative PJ phases. The results here further support our findings that SSV tends to be strengthened (weakened) during positive (negative) PJ phases.

Given the significant differences in SSV in positive and negative PJ phases, the next question is what causes such differences. Here we examine the vertical profiles of different dynamic and thermodynamic parameters, including geopotential height, zonal wind, meridional wind, omega, relative vorticity, temperature, relative humidity, and specific humidity under different PJ conditions. As shown in Fig. 8, most parameters during positive PJ phases are generally favorable for the development of synoptic disturbances. During positive PJ phases, there is a sharp increase in PBL moisture, which can in turn strengthen PBL convergence through Ekman pumping and convective heating, contributing positively to the growth of SSV. In addition, the weakened geopotential height, as well as enhanced vorticity and rising motion, in the lower and middle troposphere, further support the strengthening of the synoptic disturbances. Meanwhile, though the westerly shear and the more stable atmosphere in positive PJ phases apparently have an inhibiting effect on the development of the synoptic disturbances (Wang and Xie 1996; Li 2006), such negative impacts seem to be offset by positive impacts, which lead to an overall strengthened SSV during positive PJ phases. The situation during negative PJ phases is similar but reversed. The dominant unfavorable conditions during negative PJ phases, including the weakened PBL moisture, vorticity, and rising motion, seem to override the positive effects contributed by the easterly shear, resulting in an overall weakened SSV in the WNP. To sum up, observational results suggest that environmental factors are generally more (less) favorable for the growth of synoptic disturbances during positive (negative) PJ phases, which result in distinctive differences in SSV in the WNP.

Fig. 6. (a) EOF1 and (b) EOF2 of the 3–8-day filtered 850-hPa vorticity anomalies during positive 10–50-day filtered PJ phases. (c) Correlation coefficients between the time series of PC1 and PC2 at different lags for positive PJ phases. (d)–(f) As in (a)–(c), but for negative filtered phases.
5. Numerical experiments

Finally, in order to verify the observational results, several numerical experiments have been conducted using the anomaly AGCM. The primary focus here is on how the growth of the initial perturbation is affected under different PJ conditions. For the control experiment, the model is linearized using the summer mean state to see how the initial perturbation grows under the realistic summer-mean basic state. In subsequent runs (positive PJ run and negative PJ run hereafter), anomalies associated with positive PJ and negative PJ phases are then added on top of the summer mean state to investigate how the growth of the perturbation is influenced by the different PJ phases.

Figure 9 shows the initial perturbation vorticity field prescribed in the model and the horizontal pattern of the most unstable mode at day 5 for the control run. Note that the final structure of the unstable mode is insensitive to the initial pattern of the perturbation field, which Li (2006) has also tested and confirmed. As shown in Fig. 9, the unstable mode, normalized by the maximum amplitude, depicts a typical synoptic-scale wave train structure with alternating positive and negative circulation anomalies aligned in a northwest–southeast orientation in the WNP. The results are comparable to that of Li (2006), who captured a similar wave train in the presence of the summer mean flow and the CFC feedback using the same model. In other words, the model is capable of reasonably reproducing the essential

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**Fig. 7.** (a)–(c) Lagged regression of 3–8-day filtered 850-hPa vorticity (shading, $10^{-6}$ s$^{-1}$) and wind (vectors, m s$^{-1}$) anomalies against PC1 for positive 10–50-day filtered PJ phases; (d)–(f) as in (a)–(c), but for negative filtered phases. Values in the top left corners denote the lag day. Only values exceeding 95% confidence have been plotted.
features, including the northwest–southeast orientation, the alternating cyclonic and anticyclonic structure, as well as the northwestward-propagating nature of the synoptic-scale wave train in the WNP. Given the same initial perturbation, Li pointed out that the background favorable dynamic and thermodynamic conditions in the WNP lead to stronger CFC feedback, thereby favoring the growth of the perturbation in the region. Specifically, the large warm pool and the greater water vapor content in the WNP favor a larger amount of moist static energy and thus conditionally unstable stratification in the region (Li and Wang 1994). Besides, the background monsoon confluent flow and the easterly shear in the WNP also contribute to local energy accumulation (Kuo et al. 2001) and Rossby wave amplification (Wang and Xie 1996) in the lower troposphere during boreal summer, thereby favoring rapid development of the synoptic wave train in the WNP.

We then see what happens when the PJ background anomalies are added to the summer mean flow. Figure 10 compares the evolution of the maximum perturbation kinetic energy (PKE) for the control, positive PJ, and negative PJ runs. The maximum PKE for all three runs shows a dramatic increase after day 5, indicating the growth of the synoptic perturbation in the presence of convection–frictional convergence. Obviously, the maximum PKE for the positive PJ run after day 5 is much greater than that of the negative PJ and the control run. Indeed, the PKE value on day 6 for the positive PJ run is already double that of the negative PJ run. This indicates that positive PJ phases actually support faster growth of synoptic perturbation. The modeling results here agree well with our observational studies, where strengthened (weakened) SSV is found in positive (negative) PJ phases. We have also checked the horizontal patterns of the unstable mode, though no significant differences for the different runs can be observed (figure not shown). For all three runs, the model reproduces the unstable modes that have similar synoptic wave train structures in the WNP.

In addition, two sensitivity experiments have been carried out to further compare the contributions of dynamic and thermodynamic parameters to wave growth under different PJ conditions. In the first experiment, only the PJ-related near-surface thermodynamic parameters, including $\theta_b$ and $T_b$, are specified and added to the summer mean state, with the three-dimensional dynamic parameters of the summer mean flow ($u$, $v$, $T$, and $P$) remaining unchanged. In the second experiment, the initial condition is changed by specifying and incorporating the PJ-related three-dimensional dynamic parameters, instead of the thermodynamic parameters, onto the summer mean state. By comparison with the results of the control run, the relative importance of the dynamic and the thermodynamic parameters can be revealed. Figure 11 presents the changes in the maximum

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**Fig. 8.** Vertical profiles of 10–50-day filtered anomalies (averaged over 10°–30°N, 120°–150°E) of relative humidity (RH), specific humidity ($q$), omega, geopotential height (GPH), zonal wind ($u$), meridional wind ($v$), temperature (Temp), and relative vorticity (Vort) for positive and negative 10–50-day filtered PJ phases. Solid lines represent anomalies for positive phases and dotted lines denote values for negative phases.
PKE at day 6 under the different aforementioned scenarios. For positive PJ phases (Fig. 11a), we can see that the PJ-related PBL thermodynamic factors contribute positively to the maximum PKE, whereas PJ-related dynamic parameters have a negative effect on the growth of PKE. Although the dynamic and thermodynamic factors behave differently, it is worth noting that the positive contributions of the PBL thermodynamic parameters are actually stronger, which results in a net increase in maximum PKE under positive PJ conditions. The modeling experiments here further extend the results of our observational studies by illustrating the relative importance of the dynamic and thermodynamic fields. Similarly for negative PJ phases, the PJ-related PBL thermodynamic parameters also play a dominant role in controlling the perturbation growth. As shown in Fig. 11b, the negative influences contributed by reduced PBL moisture outweigh the enhancing effects of the dynamic factors and lead to an overall reduction in maximum PKE under negative PJ conditions. The modeling results here reveal that the PJ-related PBL thermodynamic parameters play a determining role in controlling the growth of the synoptic perturbations. An increase (decrease) in background PBL moisture during positive (negative) PJ phases enhances (suppresses) perturbation moisture convergence through Ekman pumping and thus the convective heating associated with SSV, leading to strengthened (weakened) synoptic-scale activity in the WNP.

6. Implications for TC activity in the WNP

Previous sections reveal significant differences in synoptic-scale variability in the WNP from both observational and numerical perspectives. Since synoptic-scale disturbances are important energy sources for cyclogenesis and often treated as seeds or precursors of an individual TC event (Maloney and Hartmann 2001; Fu et al. 2007; Li 2012), it is of interest to further take a look at whether the intraseasonal PJ pattern can modulate TC activity in the WNP.

Figure 12 shows the TC distribution during positive and negative 10–50-day filtered PJ phases, while Table 1 summaries the corresponding TC statistics. The average daily genesis rate (DGR) during positive PJ phases is 15.02%, which is 1.5 times more than that in negative PJ phases (DGR = 9.87%). The major changes in TC activity occur within 10°–30°N, 120°–150°E, which coincide well with the corresponding enhancement or suppression in synoptic-scale activity over the region (Fig. 5). The TC statistics are consistent with our findings. The favorable background conditions and faster growth of
synoptic eddies during positive PJ phases provide favorable environments as well as seed disturbances for TC formation, whereas unfavorable background and weakened synoptic-scale activity during negative PJ phases tend to suppress the overall TC frequency in the WNP. Apart from cyclogenesis, another discernible difference here is the alternations of prevailing TC tracks during different PJ phases. During positive PJ phases, there is a higher tendency for TCs to recurve owing to the eastward retreat of the western North Pacific subtropical high (WNPSH). On the contrary, the westward-extending WNPSH during negative PJ phases tends to reduce the probability of recurring TCs. Using monthly datasets, Choi et al. (2010) found that the interannual PJ pattern can significantly modulate TC activity in the WNP. As an extension, we here further show that the changes in the summer mean state and synoptic-scale activity associated with intraseasonal PJ patterns play an important role in modulating the subseasonal TC activity in the WNP.

7. Summary and discussion

This study systematically investigates the influences of the intraseasonal PJ pattern on synoptic-scale activity in the WNP based on both observational analysis and numerical studies. It is found that the boreal summer PJ teleconnection exhibits obvious intraseasonal variations, with a dominant peak at 10–50 days. Composite and EOF analysis indicate that the intraseasonal PJ pattern has significant influence on synoptic-scale variability in the WNP. During positive PJ phases, strengthened SSV is found in the WNP, with a much stronger and better organized synoptic wave train structure. Such a synoptic-scale wave, however, is greatly weakened during negative PJ phases. As shown by the vertical profiles of the observational data, the environmental parameters associated with positive (negative) PJ phases are generally more (less) favorable for the growth of the synoptic disturbances. The strengthened PBL moisture, together with the weakened geopotential height and enhanced vorticity, apparently outweighs the inhibiting effect of westerly shear and contributes positively to the overall strengthened synoptic-scale disturbances during positive PJ phases. On the other hand, during negative PJ phases, the dominant unfavorable conditions, including weakened PBL moisture, vorticity, and rising motion seem to override the positive effects contributed by easterly shear, leading to an overall weakened SSV in the WNP.
The observational results are further verified with the aid of an anomaly AGCM. By introducing a small perturbation initially and prescribing different background states, we examine how the perturbation evolves under different PJ conditions. The unstable mode captures well the properties of the synoptic-scale wave. Consistent with the observational results, the numerical experiments reveal a much faster (slower) growth of the synoptic-scale wave train when the environmental anomalies associated with positive (negative) PJ phases are incorporated into the model. Sensitivity tests also illustrate the relative importance of PBL thermodynamic parameters as well as three-dimensional dynamic parameters on SSV. It turns out that PBL moisture dominates and plays a decisive role in controlling the growth of synoptic perturbations, while three-dimensional dynamic factors exert secondary reverse effects. The increase (decrease)

<table>
<thead>
<tr>
<th>Days</th>
<th>TC frequency</th>
<th>Daily genesis rate (TC frequency × 100/days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Positive PJ</td>
<td>486</td>
<td>73</td>
</tr>
<tr>
<td>Negative PJ</td>
<td>537</td>
<td>53</td>
</tr>
<tr>
<td>Climatology</td>
<td>2944</td>
<td>371</td>
</tr>
</tbody>
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in background PBL moisture can enhance (suppress) perturbation moisture convergence through Ekman pumping, which can further strengthen (weaken) convective heating and contribute to strengthened (weakened) SSV in the WNP during positive (negative) PJ events. Besides, serving as potential seed disturbances for cyclogenesis, the strengthened (weakened) synoptic-scale activity during positive (negative) PJ phases also results in an enhancement (suppression) in intraseasonal TC frequency in the WNP.

In this study, we focus primarily on the distinctive influences of different PJ patterns on SSV in the WNP. As pointed out by previous studies (Zhou and Li 2010; Hsu et al. 2011; Hsu and Li 2011), SSV may also exert an upscale feedback on the intraseasonal oscillation through nonlinear rectification. A follow-up study is currently underway to investigate how SSV may nonlinearly rectify the intraseasonal PJ pattern. In addition, Kosaka et al. (2011) suggested that the interannual variability of the PJ pattern is closely related to ENSO. SSV, on the other hand, has also been shown to be intimately linked to ENSO in the WNP (Li and Wang 2005; Zhou and Chan 2007; Li and Zhou 2012; Chen 2012; Wang and Wang 2013; etc.). Therefore, it will also be interesting to look into how the interannually varying basic state related to ENSO interacts with both the PJ pattern and SSV in the WNP.

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APPENDIX

A Multilevel Baroclinic Anomaly Model

The primitive equations of the AGCM include momentum, temperature, and continuity equations, which can be expressed in the vertical sigma coordinate as

$$\frac{\partial \mathbf{V}}{\partial t} = -\mathbf{V} \cdot \nabla \phi - RT_{S0} \nabla \ln P_S + G_m,$$

$$\frac{\partial T}{\partial t} = \frac{T_{S0} \kappa}{\sigma} \sigma_D - \kappa T_{S0} \langle D \rangle + G_T,$$

$$\frac{\partial (\ln P_S)}{\partial t} = -\langle D \rangle + G_C,$$

where $T_{S0} = 250 \text{ K}$ is the reference atmospheric temperature, $\sigma = \sigma_D + \sigma_A$ denotes vertical velocity in the sigma $(\sigma = p/p_s)$ coordinate, $\sigma_D = \int_0^\sigma (D - \langle D \rangle) \, d\sigma$, $\sigma_A = \int_0^\sigma (\mathbf{V} - \langle \mathbf{V} \rangle) \cdot \nabla P_S \, d\sigma$, and angle brackets represent a vertical average from the surface to the top of the atmosphere. In addition, $G_m$, $G_T$, and $G_C$ are nonlinear terms representing slow advective adjustment and atmospheric diabatic processes in the forms of

$$G_m = - (\zeta + f) \mathbf{k} \times \mathbf{V} - \sigma \frac{\partial \mathbf{V}}{\partial \sigma} - \nabla E - R(T - T_{S0}) \mathbf{V} \ln P_S,$$

$$G_T = - \mathbf{V} \cdot \nabla T - \sigma \frac{\partial T}{\partial \sigma} + Q - \kappa (T - T_{S0}) \langle D \rangle + \frac{(T - T_{S0}) \kappa}{\sigma} \sigma_D + \frac{T_{S0} \kappa}{\sigma} \sigma_A,$$

$$G_C = - \langle \mathbf{V} \rangle \cdot \nabla \ln P_S.$$

The basic state variables are assumed to follow the above governing equations such that the effect of atmospheric transients on the mean basic state can be considered to be negligible in the tropics. The perturbation governing equations can then be obtained by subtracting the basic state equations from the total equations:

$$\frac{\partial \mathbf{V}'}{\partial t} = -\mathbf{V} \cdot \nabla \phi' - RT_{S0} \nabla \ln P_S' + G_m',$$

$$\frac{\partial T'}{\partial t} = \frac{T_{S0} \kappa}{\sigma} \sigma_D' - \kappa T_{S0} \langle D' \rangle + G_T',$$

$$\frac{\partial (\ln P_S')}{\partial t} = -\langle D' \rangle + G_C',$$

where $G'_x = G_x(\bar{a} + a') - G_x(\bar{a})$; the subscript $x$ corresponds to the subscripts $m$, $T$, and $C$ in the above equations, and $a$ represents model-dependent variables. The perturbation field and the basic-state field are denoted by a prime and an overbar, respectively.

REFERENCES


