Article ID: 1006-8775(2017) 01-0068-13

INFLUENCE OF THE INTERANNUAL VARIATION OF CROSS-EQUATORIAL FLOW ON TROPICAL CYCLOGENESIS OVER THE WESTERN NORTH PACIFIC

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Abstract: The influence of the interannual variation of cross-equatorial flow (CEF) on tropical cyclogenesis over the western North Pacific (WNP) is examined in this paper by using the tropical cyclone (TC) best track data from the Joint Typhoon Warning Center and the JRA-25 reanalysis dataset. The results showed that the number of TCs forming to the east of 140°E over the southeastern part of the western North Pacific (WNP) is in highly positive correlation with the variation of the CEF near 125° E and 150° E, i.e., the number of tropical cyclogeneses increases when the cross-equatorial flows are strong. Composite analyses showed that during the years of strong CEF, the variations of OLR, vertical wind shear between 200-850 hPa, 850 hPa relative vorticity and 200 hPa divergence are favorable for tropical cyclogenesis to the east of 140°E over the tropical WNP, and vice versa. Moreover, it is also discussed from the view of barotropic energy conversion that during the years of strong CEF, an eastward-extended monsoon trough leads to the rapid growth of eddy kinetic energy over the eastern part of WNP, which is favorable for tropical cyclogenesis; but during the years of weak CEF, the monsoon trough is located westward in the western part of the WNP, consistent with the growth area of eddy kinetic energy. As a result, there are fewer TC geneses over the eastern part of WNP. Besides, the abrupt strengthening of a close-by CEF 2-4 days before tropical cyclogenesis may be the one of its triggers.

Key words: tropical cyclogenesis; statistical analysis; cross-equatorial flow; western North Pacific; barotropic energy conversion

CLC number: P444 **Document code:** A

doi: 10.16555/j.1006-8775.2017.01.007

1 INTRODUCTION

Cross-equatorial flow (CEF) is a major pathway for momentum and moisture exchange between the Northern and Southern Hemispheres. In the boreal summer, there are three notable channels of southerly CEF over the western North Pacific (WNP). These channels are located at approximately 105° E, 125° E, and 150°E near the equator[1-2], among which the former two are more significant because of the stronger southerlies[3]. Previous studies have indicated that the variation of the CEF is closely related to the East Asian monsoon, El Niño-Southern Oscillation, and tropical cyclone (TC) activities[3-6].

The relationship between the CEF and the formation frequency of TCs over the WNP has been discussed in various studies. On interannual time-scales, a positive correlation exists between the intensity of the CEF and the frequency of TCs in the main development region during the typhoon season in boreal summer. Wang and Leftwich[7] stated that the cross-equatorial currents are stronger during the active periods for TC formation, and weaker during inactive periods. Using statistical analyses, Li et al.[8] found that TC activities are positively correlated with the intensity of the northward flows near 45° E during June to September. Furthermore, the abnormally weak CEFs between 90°E to 180° may have been the main reason why TC activity was rare. The short-term change in CEF also plays a significant role in tropical cyclogenesis. Li et al.[7] pointed out that TC formation can be induced by the rapid intensification of the CEFs near the genesis longitude, based on an examination of 14 events in 1998. Xiao[9] studied TC activities and corresponding variations of the CEFs in 1982. They found that the southerlies turned northeastward at 130°E and 150-160°

Received 2014-07-17; **Revised** 2016-02-03; **Accepted** 2017-02-15

Foundation item: Special Scientific Research Project for Public Interest (GYHY201006021); National Basic Research Program of China (2010CB950403); Special Scientific Research Project for Public Interest (GYHY201005019-2)

Biography: FENG Tao, Ph. D., primarily undertaking research on typhoon climatology and air-sea interaction.

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E, and quickly strengthened after crossing the equator. Therefore, they affected the tropical cyclogenesis during that year. Liu et al. [10] studied the relationship of the corresponding variation between the CEF and the monsoon trough, or the Intertropical Convergence Zone (ITCZ), in the Northern Hemisphere. Because the monsoon trough is the most favorable regime for tropical cyclogenesis [11], the accompanying variations between the CEF and the monsoon trough are critical for modulating the frequency and position of tropical cyclogenesis.

The above-mentioned studies show a clear relationship between the intensification of the CEF and TC genesis over the WNP. However, the interannual variation of the CEF and its impact on the frequency and position of tropical cyclogenesis over the WNP has not been investigated. In addition, few studies have mentioned the relevant mechanisms. Therefore, this work mainly focuses on the relationship between the interannual variation of the CEF and the frequency of tropical cyclogenesis over the WNP. Furthermore, barotropic processes are responsible for the main growth of kinetic energy for synoptic perturbations during the formation period through the convergence and the shear of the mean flow [12-14]. Thus, this paper will also discuss the barotropic energy conversions related to the intensity of the CEF, and their impacts on tropical cyclogenesis over the WNP. Finally, we will discuss the role of the rapid intensification of the CEF on TC formations.

2 DATA

The present study focuses on the typhoon season (June-October) of the WNP during 1979-2010. Japanese 25-year Reanalysis (JRA-25) data with 1.25° × 1.25° horizontal resolution, obtained from the Japan Meteorological Agency, is used in this study. Because of the computational efficiency, the six-hourly data are averaged to daily values. The best track data from the Joint Typhoon Warning Center (JTWC) is also used to provide TC information, such as, genesis time and position. The best track data includes information of tropical depressions (TD) before the intensity reaches the tropical storm (TS) category. However, only those TCs that have reached a TS category (maximum sustained surface wind reaches or exceeds 35 knots, about 18 m/s) are selected. Furthermore, the genesis time is defined as the warning time when the maximum sustained surface wind of a TC reaches 35 knots. According to these definitions, a total of 579 TCs over the WNP are included during the study period.

The outgoing longwave radiation (OLR) data from the National Oceanic and Atmospheric Administration (NOAA) is used in addition. OLR is observed by measuring the longwave radiation emitting from land surface to space, which usually represents convective activity over the WNP. Lower values of the OLR represent stronger convective activity.

3 CORRELATION BETWEEN THE FREQUENCY OF TROPICAL CYCLOGENESIS AND THE INTENSITY OF THE CROSS-EQUATORIAL FLOW

3.1 *Cross-equatorial flow index*

A CEF index needs to be defined to examine the interannual variation of the CEF. The vertical distribution of the meridional flow shows the wind maximum is found at approximately 925 hPa near the equator (figure not shown). There are three dominant channels of CEF over the WNP, and three indices (VI1, VI2, and VI3) are defined to represent the intensity of the CEF. These indices are calculated by averaging the meridional wind speed at 925 hPa within [102–112°E, 5°S–5°N], [123–133°E, 5°S–5°N], and [145–155°E, 5°S–5°N], respectively, during June to October in each year, as expressed by Eq.1:

$$VI_m=\frac{1}{K}\frac{1}{J}\frac{1}{I}\sum_{k=1}^{K}\sum_{j=1}^{J}\sum_{i=1}^{I}V_{i,j,k}\begin{pmatrix} i=1,2,\cdots,I \\ j=1,2,\cdots,J \\ k=1,2,\cdots,K \end{pmatrix} \quad (1)$$

where, the subscript m = 1, 2, and 3 represents the three specific domains, as shown in Fig.1; k is a specific day from June to October of each year; (i,j) represents a fixed grid point inside the domain m; and $V_{i,j,k}$ is the daily averaged meridional wind at the grid point (i,j) at 925 hPa pressure level inside the domain m on day k.

Due to their zonal distribution, the three CEFs impact on the frequency of tropical cyclogenesis over the entire WNP in different ways. As Fig.1 shows, the WNP is divided into five areas marked D1-D5. D1 denotes the South China Sea [105–120°E, 5–25°N]; D2 the western part of the tropical WNP [120–140°E, 5–25°N]; D3 the central part of the tropical WNP [140–160°E, 5–25°N]; D4 the eastern part of the tropical WNP [160°E–180°, 5–25°N]; and D5 the northern part of the tropical WNP [120°E–180°, 25–35°N]. Among them, the central points of the D1, D2, and D3 regions correspond to the longitude of the three primary CEFs. In the 22-year period, 96, 218, and 177 TC formations are counted in these three regions, respectively. There is no significant channel of CEF to the south of the D4 region. In addition, 47 TCs formed in D5, which is the high-latitude area. Although the TCs that formed in D5 are usually associated with the synoptic systems of the westerlies at middle latitudes[15], the correlation between the tropical CEF and TC formations in the D5 area is discussed as well. Three TCs formed outside these five areas during the 22-year dataset. In total, 576 TCs are examined in this study.

3.2 *Correlation between the frequency of tropical cyclogenesis and the CEF indices*

The first issue to address is whether a reliable relationship exists between the frequency of tropical cyclogenesis and the intensity of CEFs over the WNP. The present study mainly focuses on the interannual variability of the CEF and the associated tropical

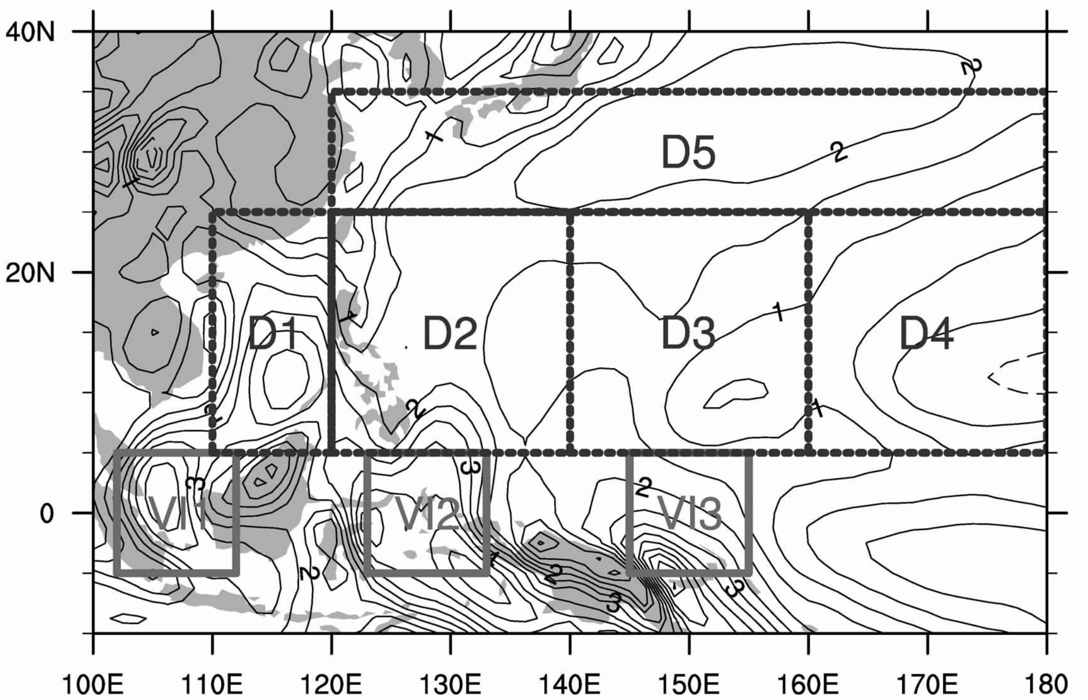

Figure 1. Horizontal distribution of 925 hPa meridional wind (contour) over the WNP during typhoon season (June-October) averaged for 1979-2010. The solid and dashed lines denote southerlies and northerlies, respectively. Heavy solid boxes denote the main channels of the cross-equatorial flows. Heavy dashed boxes indicate the main development regions for tropical cyclones.

cyclogenesis. As a result, the long-term trends of VI1, VI2, and VI3 are removed first. Next, cross correlations between the annual frequency of TC formation and the normalized CEF indices are calculated. As the WNP-column in Table 1 shows, correlations between TC frequency and the CEF indices are 0.2–0.3 for the entire WNP, and they did not exceed the 95% significance level. Dividing the WNP into five regions, the correlations are still not high in D1 and D2. However, the TC frequencies in the D3, D4, and D5 regions are significantly correlated with the CEFs, as shown in Table 1. A total of 261 TCs form over the D3, D4, and D5 regions, which account for 45.3% of all samples. Among them, the correlations between VI2 and D3, and VI3 and D4 are 0.48 and 0.63, respectively, and both exceed the 99% significance level. This indicates that the TC frequencies in D3 and D4 regions are strongly related to the low-level meridional wind over [102–112°E, 5°S–5°N] and [145–155°E, 5°S–5°N], respectively. The correlations between VI1 and D4, VI2 and D4, VI2 and D5, and VI3 and D5 are 0.41, 0.41, -0.37, and -0.45, respectively, which are all above the 95% significance level. In other words, the TC frequency in D4 is closely related to the low-level meridional wind in [102–112°E, 5°S–5°N] and [123–133°E, 5°S–5°N], and the TC activity in D5 region is clearly related to the low-level meridional wind in [123–133°E, 5°S–5°N] and [145–155°E, 5°S–5°N]. Significant negative correlations are found between the TC activity in the D5 area and VI1, VI2. This negative correlation suggests that TC formation is active in the D5 area when the CEF is abnormally weak. However, when the CEF is abnormally strong, TC formation is inhibited in this area. Therefore, TC activities over the central-eastern and northern parts of the WNP appear to be correlated with the intensity of CEF. When the CEF is abnormally strong, TCs form preferentially over the central-eastern part of the WNP, rather than over the northern part of WNP. Conversely, when the CEF is abnormally weak, the TC frequency over the northern part of the WNP is higher, compared with that over the central-eastern part of the WNP.

Table 1. The correlations between the CEF indices and the frequency of TC activities over the WNP and during typhoon seasons. The regions D1-D5 are shown in Fig.1 and the “WNP” column denotes the sum of these five regions.

	WNP	D1	D2	D3	D4	D5
VI1	0.22	0.02	−0.08	0.32	**0.41***	−0.23
VI2	0.20	−0.14	−0.04	**0.48****	**0.41***	**−0.37***
VI3	0.27	0.07	−0.04	0.33	**0.63****	**−0.45***

Note: *: above the 95% significance level; **: above 99% significance level.

High correlations (0.88, 0.62, and 0.69) indicate that there are remarkable correlations between VI1, VI2, and VI3, respectively. As a result, the three CEFs correspond well with each other on the interannual time-scale. The TC activities are less affected by the value of VI1 south of the D1 and D2 regions. Similarly, TC activities in the D3 and D5 areas are less affected by the value of VI1. The combined results based on the intensity of VI2 are similar to those based on VI3. Therefore, the following results are based on the interannual variation of VI2 to investigate the influence of the CEF on the TC activities.

Strong (weak) CEF years are selected as the standard deviation of VI2 greater than 1 (less than -1) after removing the linear trend. Six representative years of strong CEF (hereafter, strong years) are included, which are 1982, 1987, 1993, 1997, 2002, and 2006. Four representative years of weak CEF (hereafter, weak years) are included, which are 1985, 1988, 1989, and 1998. During strong years, 8.3 TCs form on average, in the D3 and D4 areas, and only 0.5 TC in D5. During weak years, 5.0 TC form, on average, in the D3 area, and 3.0 over D5.

4 INFLUENCE OF THE INTENSITY OF CROSS-EQUATORIAL FLOW ON THE LARGE-SCALE ENVIRONMENT

4.1 *Influence on low-level circulation*

To examine the influence of CEF intensity on the large-scale circulations over the WNP, wind fields at 850 hPa are composed for strong and weak years. As Fig.2a shows, during the strong years, strengthened meridional flows occur near 125° E and 150° E on the equator. Affected by the Coriolis force, the southerlies shifted northeastward after crossing the equator, thereby intensifying the low-level westerlies at lower latitudes. The monsoon trough extends close to 160°E because of the strengthened westerlies. As a result, tropical cyclogenesis are found in the entire tropical WNP, even in regions near the dateline. In contrast, during the weak years (Fig.2b), zonal wind is dominant near the equator due to the weaker southerlies from the Southern Hemisphere. In this case, the monsoon trough extends only to 135°E. As a result, TC formations are concentrated in the western part of the WNP at a relatively high latitude. Liu et al.[10] suggested that enhanced CEF plays a significant role in maintaining a strong ITCZ in the Northern Hemisphere. Affected by the strong ITCZ, the wavelength of the mixed-Rossby gravity wave shrinks. Thus, the waves transform into TD-type waves because of the zonal convergence. Conversely, the kinetic energy of the mean flow converts to eddy kinetic energy through the shear and convergence of the zonal mean flow, making the large-scale dynamical conditions favorable for tropical cyclogenesis. Based on the thermodynamics, the presence of sufficient water vapor favors the development of deep convection in the convergence zone in the middle-lower troposphere. This intensifies the vertical motion and boundary convergence near the surface layer, and promotes TC formation inside the convergence zone. As a result, the frequency of TCs is closely linked to the intensity of CEF over the WNP. When the CEF enhances, the monsoon trough strengthens and extends eastward, leading to more TC formation, and vice versa[18].

4.2 *Influence of CEF on low-level relative vorticity, upper-level divergence, convective activity, and vertical wind shear over the WNP*

In addition to the wind field, other atmospheric factors also play important roles in tropical cyclogenesis, including, convection activity, vertical wind shear, low-level relative vorticity, upper-level divergence, and sea surface temperature (SST). In boreal summer, SST is always warm enough for TC formation because it is always above 28.5℃ in the tropical WNP. SST will not therefore be discussed in this paper. Instead, the composite distributions of relative vorticity at 850 hPa (Fig.3a-3c), divergence at 200 hPa (Fig.3d-3f), vertical wind shear between 200 and 850 hPa (Fig.4a-4c), and OLR (Fig.4d-4f) are examined for June-October during strong years (1982, 1987, 1993, 1997, 2002, and 2006) and weak years (1985, 1988, 1989, and 1998).

4.2.1 LOW-LEVEL RELATIVE VORTICITY

The cyclonic relative vorticity associated with the monsoon trough in the lower troposphere over the WNP are favorable to tropical cyclogenesis. Low-level relative vorticity is significantly affected by the CEF because of the corresponding variation between the monsoon trough and the CEF. As Fig.3a shows, during the strong years, a large area with a positive relative vorticity extends from the Philippines to central WNP, among which four positive centers are found near 125°E, 140°E, 155°E, and 170 °E. These areas spread as an elongated belt. The maximum value of relative vorticity reaches 6×10^{-6}-10×10^{-6} s^{-1}. However, as Fig.3b shows, two positive centers appear near 125° E and 155° E. The maximum relative vorticity reaches only 4×10^{-6} s^{-1} during weak years.

Comparing the results between strong and weak years (Fig.3c), the relative vorticity at 850 hPa is approximately 2×10^{-6}-6×10^{-6} s^{-1} greater during the strong than the weak years. Anomalous centers are found near 140°E, 150°E, and 170°E, and the difference reaches up to 8×10^{-6} s^{-1} at the western-most center. However, to the north of 20° N, the variation of low-level relative vorticity is opposite to that in the lower latitudes. That is, the low-level relative vorticity decreases at higher latitudes during strong CEF years (Fig.3c).

4.2.2 UPPER-LEVEL DIVERGENCE

Upper-level divergence is one of the major factors influencing tropical cyclogenesis. Strong upward motion is triggered due to the co-effects of upper-level divergence and lower-level convergence, which favors TC formation. The variation of upper-level divergence correlates well with the intensity of CEFs over the WNP. During the weak years (Fig.3d), three divergent centers are found over the tropical WNP, which are situated at 135° E, 155° E, and 170° E. The maximum

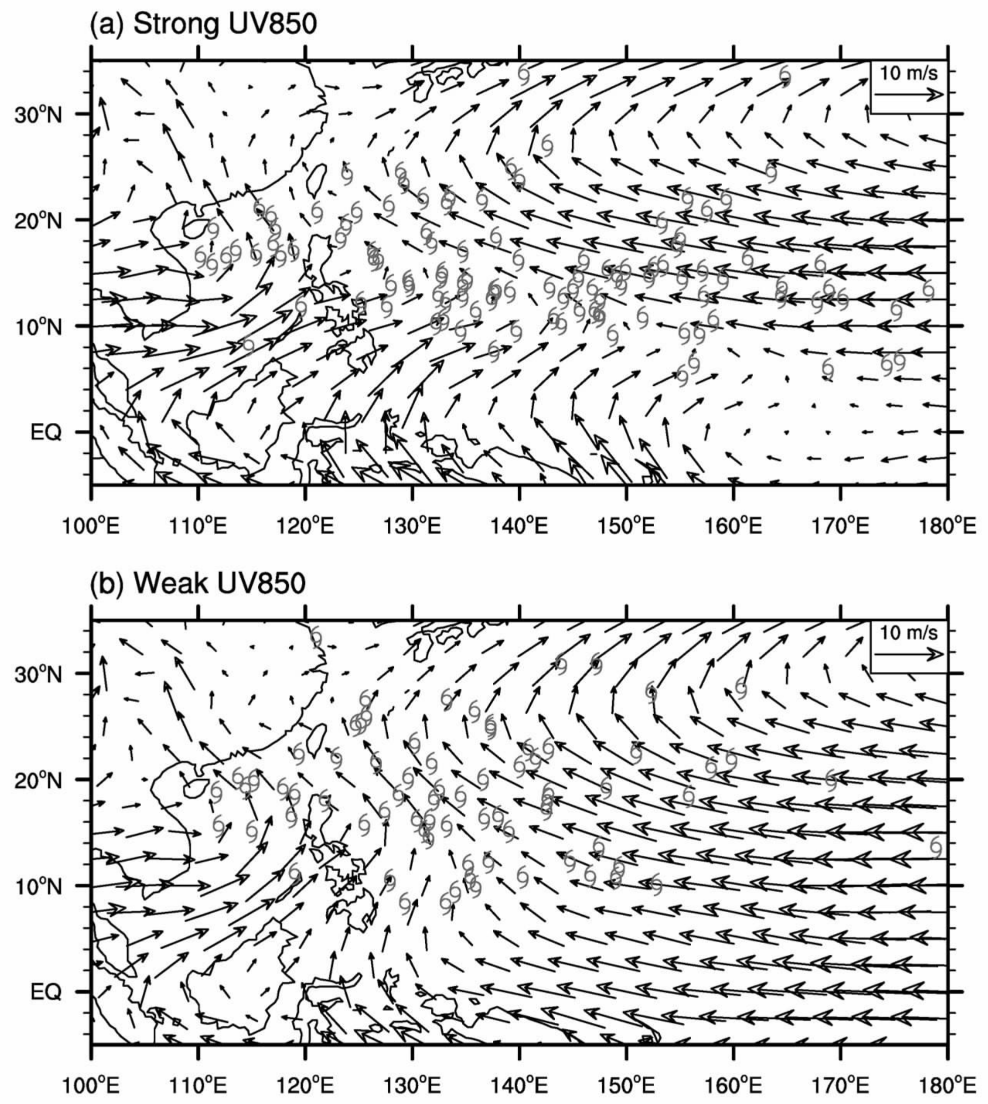

Figure 2. Compositions of horizontal wind at 850 hPa from June to October during strong CEF years (a) and weak CEF years (b). Typhoon symbols denote genesis positions.

values of divergence are all above 12×10^{-6} s^{-1} in these three areas. As shown in Fig.3a and 3d, these three strong divergent areas are consistent with low-level cyclonic vorticity. This configuration of the high- and low-level atmosphere is dynamically favorable to the intensification of large-scale vertical upward motion. In contrast, the three divergent areas in Fig. 3e also exist over the tropical WNP, but their values decrease to approximate 8×10^{-6} s^{-1}. The divergent center at 135°E shifts slightly northward with respect to that during the strong years, but no significant meridional shift is observed for the other two.

Divergence at 200 hPa is significantly greater during the strong years than that during the weak years (Fig.3f). Compared with the weak years, three divergent anomalous centers are found near 135°E, 150°E, and 170°E, and the divergences are 2×10^{-6}-6×10^{-6} s^{-1} greater in the strong years. However, divided by 15°N, the out-of-phase relationship between the divergences in the lower and higher latitudes suggests that the divergence in the higher latitudes decreases when the CEF strengthens.

4.2.3 VERTICAL WIND SHEAR BETWEEN 200-850 hPa

Vertical wind shear (VWS) between 200–850 hPa is a critical factor controlling tropical cyclogenesis. Because convective activities are essentially inhibited when large VWS exists, the upper-level warm core, which is necessary for TC formation, is difficult to establish due to the weakened cyclonic circulation. Therefore, TCs usually form in the region with small VWS over the WNP. The vertical shear of horizontal wind discussed here is defined as:

$$VWS=\sqrt{(U_{200}-U_{850})^2+(V_{200}-V_{850})^2} \tag{2}$$

where U, V are the zonal and meridional components of the horizontal wind, respectively, at 850 hPa or 200 hPa levels. When VWS decreases rapidly during several days, it satisfies the requirements for tropical cyclogenesis. However, for longer time-scales, the frequency of TC formation is statistically higher when the time-averaged VWS is relatively small. In strong

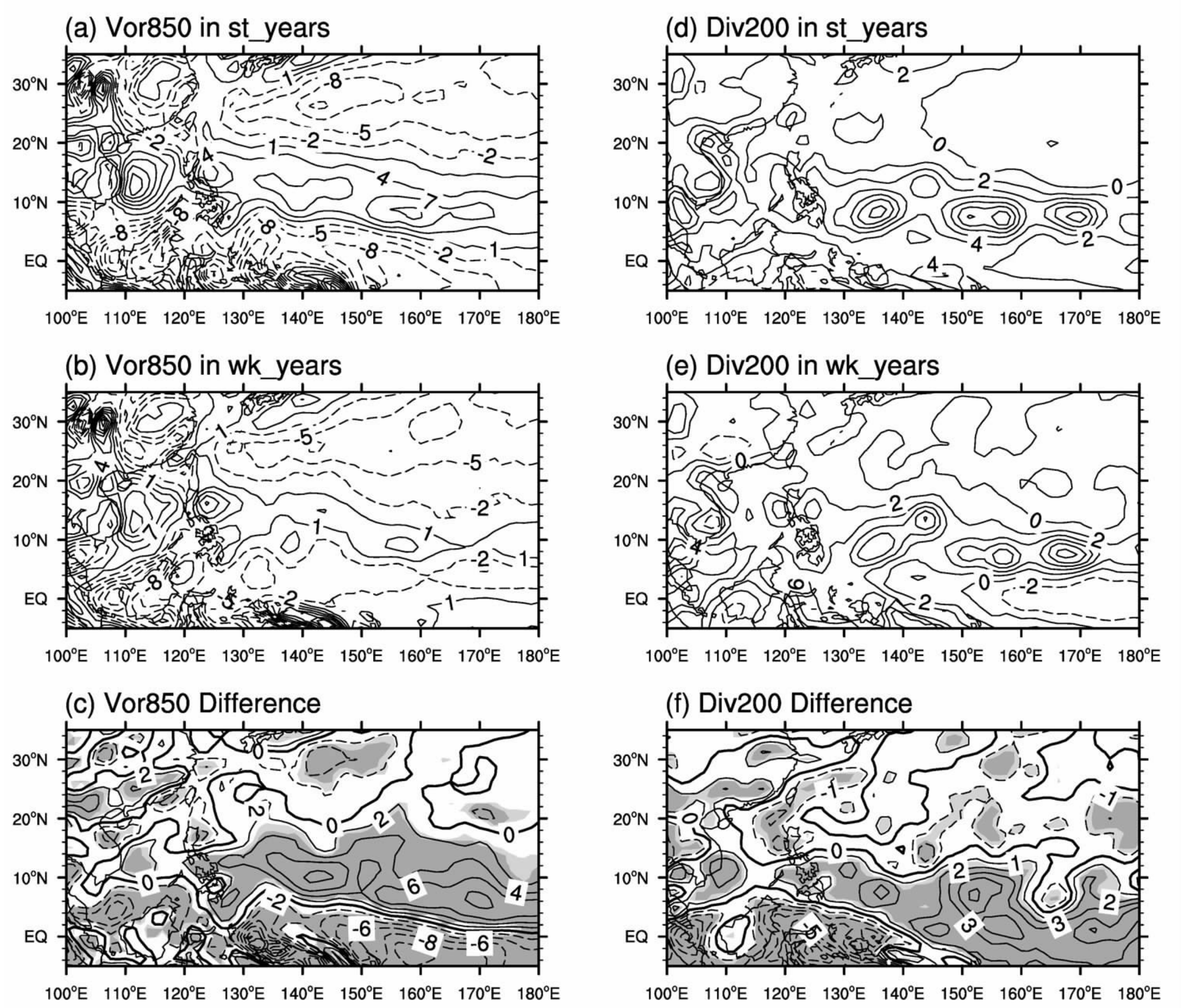

Figure 3. Composition of relative vorticity (units: 10^{-6} s^{-1}) at 850 hPa over the WNP from June to October during strong CEF years (a), weak CEF years (b), and the difference between strong and weak CEF years (c). The shaded area in (c) is above the 95% level of significance. (d-f) are the same as (a-c) except for divergence at 200 hPa (units: 10^{-6} s^{-1}).

CEF years (Fig.4a), the areas where the VWS is less than 12 m/s extend from 120° E to the dateline in west-east orientation with a belt-shaped distribution. In comparison, during weak years (Fig.4b), the contour of 12 m/s is located near 140° E, and the regions with small VWS can be found east of 140°E only.

The latitude belt between 12°N and 20°N has small VWS over the WNP and is a favorable region for TC formation. As Table 1 shows, the TC frequencies inside D1 and D2 areas are not well correlated with the CEF. Identically, the differences in VWS are relatively small over the South China Sea and the Philippines (Fig.4c). However, VWS is significantly weakened in D3 and, especially, D4 during the strong years (Fig.4c). TCs usually form in these regions because of the lower VWS. This is consistent with the statistical results (Table 1) in section 3. During the years with strong CEF, the reduction in the VWS could be attributed to the simultaneous variations of the intensity of CEF and the position of the western Pacific subtropical high over the eastern part of the WNP[10]. When the CEFs intensified, the trade wind easterlies and the subtropical high moved northward, resulting in weaker easterlies in the upper troposphere. Therefore, the VWS decreased over the southeastern part of the WNP. In the central WNP, the increasing VWS is caused by the intensification of the low-level westerlies over the lower latitudes.

4.2.4 OUTGOING LONGWAVE RADIATION

OLR is a measure of the longwave radiation emitting from the land surface or the cloud tops. It is useful for analyzing the convective activities over the ocean, where lower values of OLR indicate stronger convection. As Fig.4d shows, during years with strong CEF, the value of OLR is mostly less than 220 W/m^2 over the tropical WNP south of 18° N. Vigorous convection exists over the entire tropical WNP. The areas with enhanced convective activities extend southeastward to the Philippines, and there is a low center near where the OLR value is less than 210 W/m^2. However, over the ocean north of 18° N, the OLR values are mostly greater than 240 W/m^2, which represents suppressed convection in these regions.

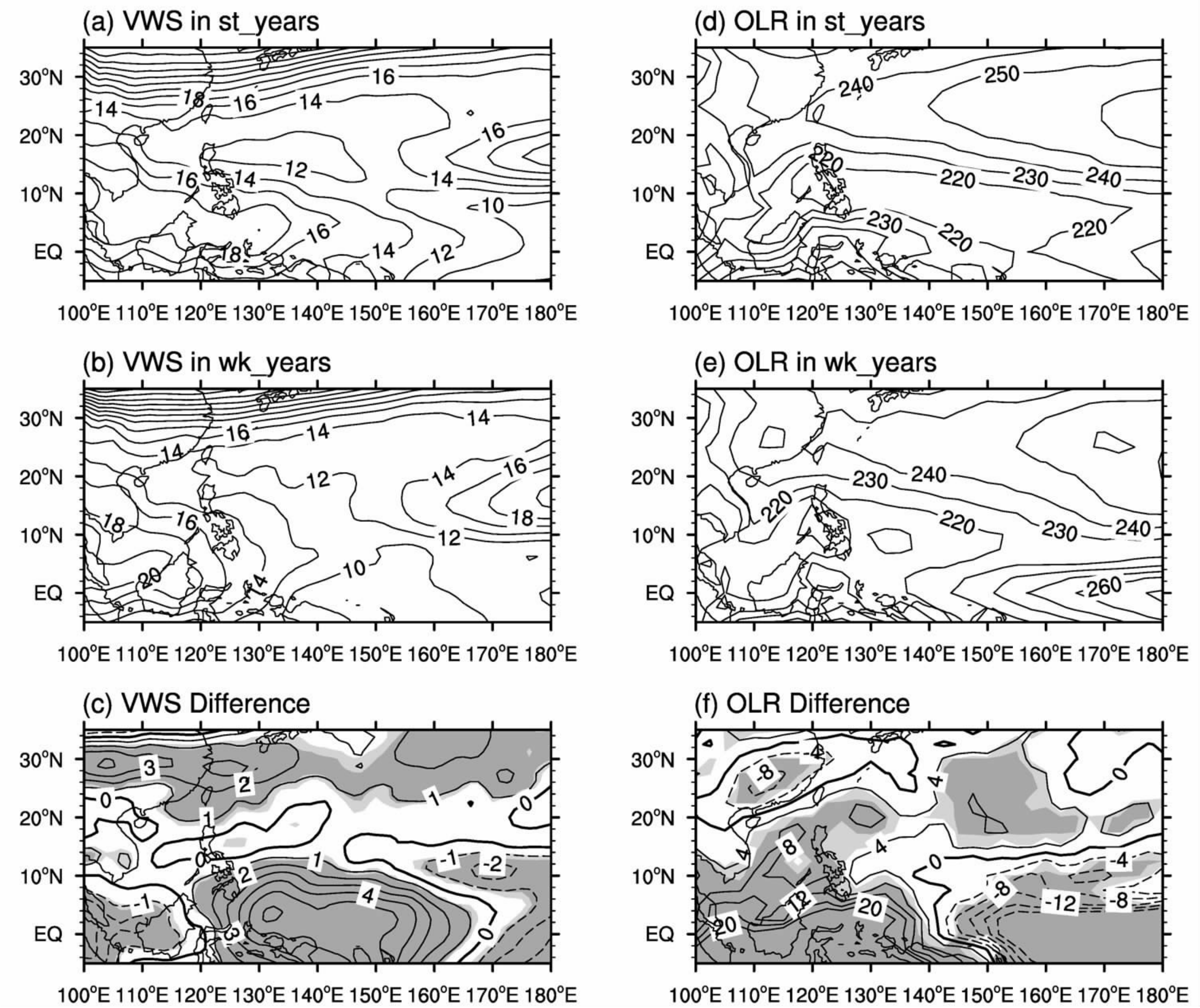

Figure 4. Composition of 200-850 hPa vertical wind shear (units: m/s) over the WNP from June to October during strong CEF years (a), weak CEF years (b), and the differences between strong and weak CEF years (c). The shaded area in (c) is above the 95% significance level. (d-f) are the same as (a-c) except for OLR (units: W/m^2).

However, as indicated in Fig. 4e, during weak years the active convection region is still located near the Philippines and the nearby OLR is less than 220 W/m^2. The regions with high OLR values, as well as those with suppressed convection, are found in the eastern and northern parts of the WNP.

There are significant differences between the OLR distributions during the strong and weak years. As Fig. 4f shows, during strong years, the differences are negative over the tropical ocean east of 140°E, which indicates active convection. In this case, the eastern part of the tropical WNP is favorable for tropical cyclogenesis. Conversely, the differences are positive to the north of 15°N, indicating that suppressed convections occur over the northern part of the tropical WNP during these years. The opposite values of OLR over the southern and northern parts of the tropical WNP may have caused the TC frequencies to vary out of phase between the lower and higher latitudes.

During years of strong CEF, the convective activity, VWS, low-level relative vorticity, and upper-level divergence are significantly different from those in the weak years during the typhoon season from June to October. During the strong years, the regions with a favorable environment for tropical cyclogenesis, including active convection, small VWS, large low-level relative vorticity and upper-level divergence, are located at the southeastern part of the tropical WNP. Conversely, during the weak years, the regions with these favorable factors are located at the northwestern part of the tropical WNP. As a result, TCs usually form in the D3 and D4 regions over the WNP during the strong years, while TC activity over D5 are suppressed. The frequency of tropical cyclogenesis and the regions of TC formation are both affected by CEF through the variation of the monsoon trough. When the CEFs are strong, the monsoon trough extends southeastward, and therefore TCs tend to form over the southeastern part of the WNP. However, during weak CEF years, the monsoon trough shifts northwestward, and therefore

TCs usually form in the northwestern part of the WNP at a relatively high latitude. Furthermore, at higher latitudes from 25°N to 35°N, during strong CEF years, the 850 hPa relative vorticity, and the 200 hPa divergence are abnormally small, and VWS and OLR are abnormally large. These are unfavorable conditions for tropical cyclogenesis and as a result the TC frequency in D5 is negatively correlated with the intensity of the CEF.

5 BAROTROPIC ENERGY CONVERSION

The above results show that the low-level convergence and cyclonic shear of the mean flow are significantly strengthened during strong CEF years because of the strong southerlies near the equator over the WNP. The kinetic energy of the mean flow is converted to eddy kinetic energy through the horizontal shear and convergence of the mean flow, known as barotropic energy conversion. Thus, the kinetic energy of the mean flow is a major source for the development of synoptic-scale disturbances [14,16]. TCs are most likely to form when a continuously intensifying disturbance moves into favorable background environment (as described in Section 4). Based on the work of Chen[16], the 3-8-day filtered wind fields are regarded as the perturbation components, and the large-scale flow with a frequency less than 10 days is regarded as the basic component. The effects of the intraseasonal oscillation in the 10-90-day periods and long-term mean circulation with a frequency less than 90 days are both considered in the calculations and discussion. The simplified equation of barotropic energy conversion is:

$$\frac{\partial K'_{baro}}{\partial t}=-\overline{u'v'}\frac{\partial}{\partial y}\bar{u}-\overline{u'v'}\frac{\partial}{\partial x}\bar{v}-\overline{u'^2}\frac{\partial}{\partial x}\bar{u}-\overline{v'^2}\frac{\partial}{\partial y}\bar{v} \tag{3}$$

where (U', V') represent the zonal and meridional comp onents of eddy flow, ($\bar{U}$, $\bar{V}$) represent the zonal and meridional components of the basic flow, and the eddy kinetic energy is $K'=(\overline{u'^2}+\overline{v'^2})/2$. The left side of Eq. (3) is the growth rate of the eddy kinetic energy and the eddy kinetic energy tendency. The four terms on the right side represent the energy growth caused by the meridional shear of the zonal wind, the zonal shear of the meridional wind, the zonal convergence of the zonal wind, and the meridional convergence of the meridional flow, respectively.

5.1 *Growth of the eddy kinetic energy*

The composite distribution of the eddy kinetic energy tendency and horizontal wind at 850 hPa over the WNP during June to October in strong- and weak-CEF years is shown in Fig.5. The eddy kinetic energy tendency is distributed along the axis of the monsoon trough during the entire typhoon season despite the intensity of CEF (Fig.5a and 5b). During strong CEF years (Fig.5a), the main growth area of the eddy kinetic energy is located inside [110–160°E, 5–15°N], together with the eastward spread of the monsoon trough. The growth rate of the eddy kinetic energy is greater than 1×10^{-5} m²/s³ in these regions, where the maximum value of the growth rate reaches above 3×10^{-5} m²/s³ near the shear region [130–140°E, 10–20°N]. In this case, when a cyclonic perturbation moves into the area, large amounts of kinetic energy are provided for the development of cyclonic perturbations by the large-scale mean flow through barotropic energy conversion. This usually leads to tropical cyclogenesis in the D3 and D4 regions. In contrast, during the weak years (Fig.5b), the monsoon trough shifts westward. As a result, the main growth areas are found in the low-latitude region between 100° E and 140° E, where the growth rate of the eddy kinetic energy is above 1×10^{-5} m²/s³. Although the coverage of the growth region is smaller in the weak years than that in the strong years, the growth rate over the Philippines and the South China Sea reaches up to 5×10^{-5} m²/s³. That is to say, during the weak CEF years, developing perturbations are usually located over the western part of the WNP, which frequently leads to TC formation in the west. However, less kinetic energy converted to eddies from the mean flow. This results in suppressed TC activity over the eastern part of the WNP.

5.2 *Four terms of barotropic energy conversion*

To analyze the influence of the different mean flows on the growth of the eddy kinetic energy, each term on the right hand side in Eq. (3) is calculated by averaging over the genesis period and a 10°×10° box around the genesis position. The averaged growth rates of the eddy kinetic energy during tropical cyclogenesis in strong and weak years are also calculated. As Table 2 shows, the major contributors to energy growth are $-\overline{u'v'}\partial\bar{u}/\partial y$ and $\overline{u'^2}\partial\bar{u}/\partial x$, irrespective of a strong or weak year. The other two terms, $\overline{v'^2}\partial\bar{v}/\partial y$ and $-\overline{u'v'}\partial\bar{v}/\partial x$, are always negative. During strong years, the growth rate of eddy kinetic energy is 3.6×10^{-5} m²/s³ on average during the formation period. The growth rate is 3.3×10^{-5} m²/s³ from the meridional shear of the zonal wind ($-\overline{u'v'}\partial\bar{u}/\partial y$), and about 1.6×10^{-5} m²/s³ from the zonal convergence of zonal wind ($\overline{u'^2}\partial\bar{u}/\partial x$). This indicates that the effect of the meridional shear of zonal wind is greater than the effect of the zonal convergence of the zonal wind on eddy kinetic energy growth during strong CEF years.

However, in weak CEF years, the growth rates of the eddy kinetic energy are 1.7×10^{-5} m²/s³ and 1.3×10^{-5} m²/s³ from the meridional shear and zonal convergence of zonal wind, respectively. Because of the strong CEFs, the monsoon trough intensifies and shifts eastward. Therefore, the growth rate of the eddy kinetic energy provided by the mean flow is about three times greater than that during weak years. The rapid growth of the eddy kinetic energy leads to the intensification of the low-level synoptic perturbation. This results in TC formation over the eastern part of the WNP.

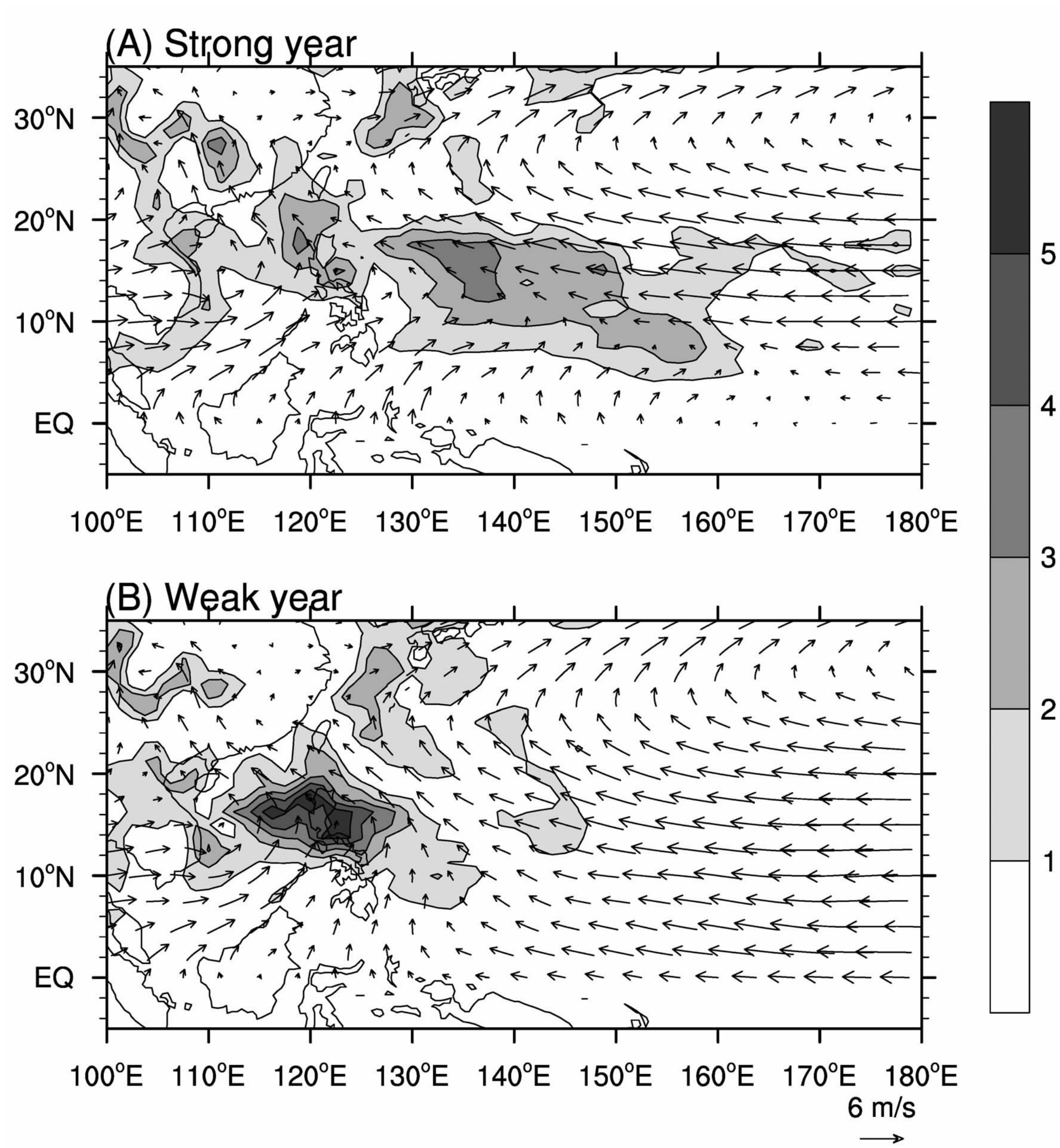

Figure 5. Eddy kinetic energy tendency (units: 10^{-5} m^2s^{-3}) at 850 hPa over the WNP from June to October during strong CEF years (a), and weak CEF years (b). Only contours with a value greater than 1 are plotted. The contour interval is 1.

Table 2. Growth rates of eddy kinetic energy at 850 hPa through each term of Eq.(3) averaged by a box centered at genesis position during genesis period. Unit: 10^{-5} m^2 s^{-3}.

	$-\overline{u'v'}\partial\bar{u}/\partial y$	$\overline{v'^2}\partial\bar{v}/\partial y$	$-\overline{u'^2}\partial\bar{u}/\partial x$	$-\overline{u'v'}\partial\bar{v}/\partial x$	$\partial K'_{baro}/\partial t$
Strong years	3.3	−1.1	1.6	−0.2	3.6
Weak years	1.7	−1.6	1.3	−0.2	1.2

6 DYNAMICAL EFFECT OF THE SHORT–TERM VARIATION OF CROSS–EQUATORIAL FLOW ON TROPICAL CYCLOGENESIS OVER THE WNP

The above analysis focused on the impact of the interannual variation of the CEFs on the frequency and genesis region of TCs over the WNP. However, the short-term variations of the CEFs also have significant dynamical effects on tropical cyclogenesis. Therefore, the horizontal wind at 850 hPa is first interpolated onto a horizontal grid at 1.25° resolution with the center at the genesis position during each genesis period. The zonal flow is then averaged from 2.5° to 10.0° to the south of the genesis latitude. The equatorial meridional flows are also averaged at 125° E and 150° E to represent the variation of the CEF during the genesis period in strong and weak years, respectively (Figs.6–7).

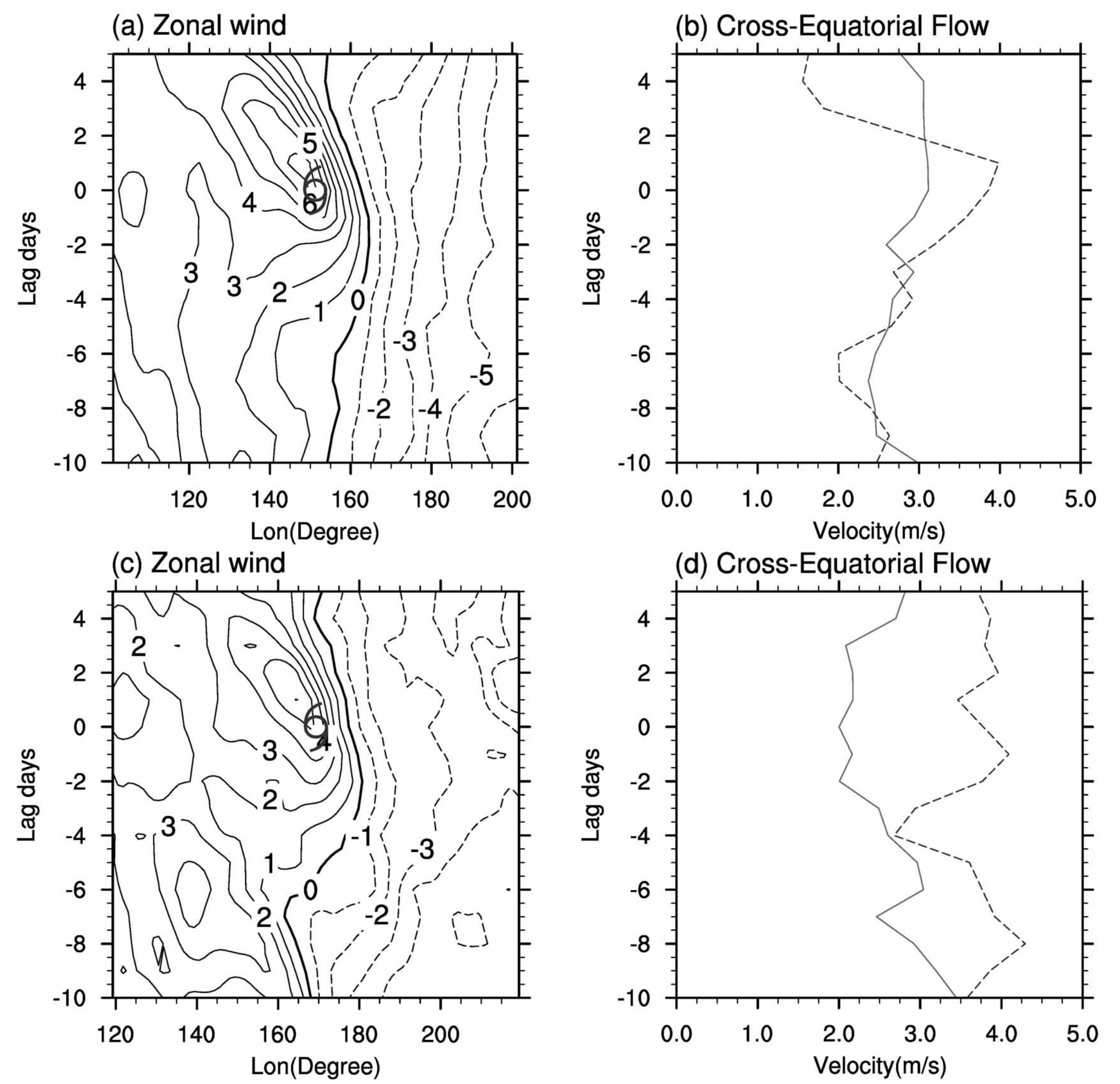

Figure 6. The "−" symbol on the vertical axis represents the time before genesis. Panels (c) and (d) are the same as (a) and (b) except for the TCs that formed in the D4 region.

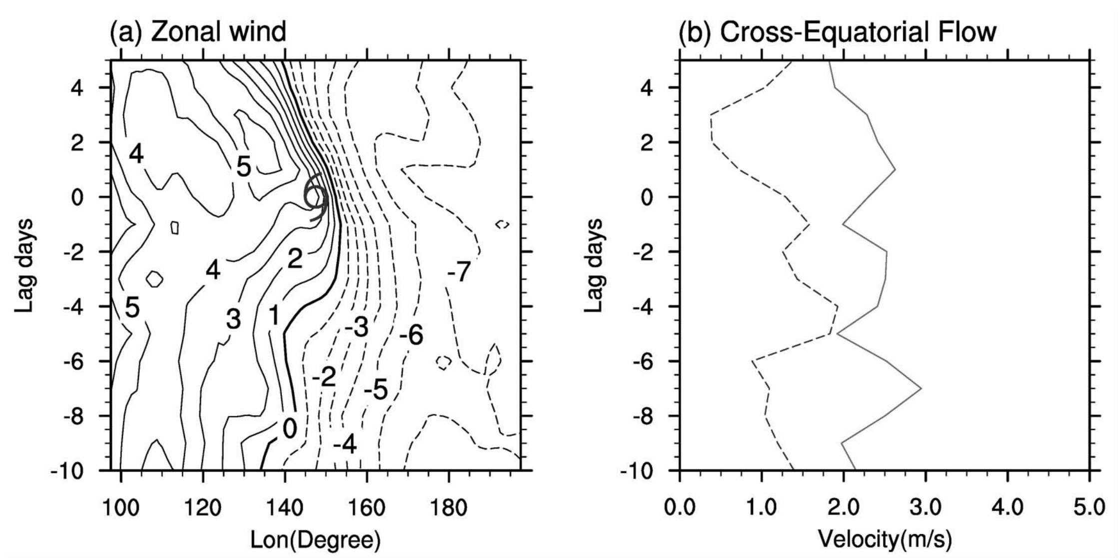

Figure 7. As Fig.6a and 6b but for weak CEF years.

6.1 *Strong years*

During the strong CEF years, the average position of the tropical cyclogenesis in the D3 region is located at [151.2° E, 13.6° N]. Fig.6a shows the temporal variation of the 850 hPa zonal wind. At six days before genesis (-6 day), a weak zonal flow is situated around the genesis area, and the eastern end of the monsoon trough is located at about 2,000 km west of the genesis position. On the same day, the meridional winds over the two CEF channels begin to intensify at the same time (Fig.6b). The monsoon trough begins to shift eastward. At four days before genesis (−4 day), the strengthening of the southerlies slows down and remains at about 2.8 m/s. From 4 to 2 days before genesis, the monsoon trough continues to shift eastward because of the strong southerlies across the equator, although no further strengthening of the southerlies is observed[4]. At about 2 days before genesis, the eastern end of the monsoon trough reaches approximately 1,000 km west of the genesis longitude (Fig. 6a). Simultaneously, the CEF near 150°E rapidly strengthens from 2.6 m/s to 4 m/s, leading to the intensification of the cyclonic vorticity near the eastern end of the monsoon trough. At this stage, the monsoon trough persists and strengthens with the increase of the low-level relative vorticity. Hence, when a TC finally forms, the zonal westerlies shift slightly westward together with the intensifying and northwest moving mature TC. During the entire period of tropical cyclogenesis, the boundary between the westerlies and easterlies is maintained at approximately 157°E regardless of the strength of the westerlies, and the −2 m/s contour is about 1,500 km to the east of the TC.

Fig.6c−6d shows the composite results for the TCs that formed in D4 during strong CEF years. At seven days before TC formation, the boundary of the monsoonal westerlies and trade wind easterlies is located at approximately 164°E. There is a low center at about 2,500 km west of the boundary, which split the monsoonal southwesterlies from the Indian Ocean and the CEF over the WNP. A significant weakening trend is also found for the meridional wind from the Southern Hemisphere during several days (Fig.6d). At 4 days before TC genesis, the main part of the monsoon trough is located near 145°E, and the southerlies at 150°E intensify from 2.6 m/s to 4 m/s, resulting in the westerlies strengthening from 1.0 m/s to 3.0 m/s inside 145°E to 165°E. At the same time, the northeast trade winds are weak in the low troposphere over the WNP, whereas the westerlies at the low latitudes persist. At two days before genesis, the main part of the monsoon trough is still located over the ocean west of 140°E. The westerlies within [150−170°E, 10°N] strengthens with a slight eastward shift about 1,000 km east of the genesis position due to the strong CEF at 150°E. Along with the strengthening of the southerlies, the TCs usually form at the eastern end of the monsoon trough. In summary, during strong CEF years, D4 is the favorable region for tropical cyclogenesis because of the strong CEF at 150° E, which is essential for the maintenance of the monsoon confluence region. The confluence region shifts eastward due to the weak easterlies and the strong westerlies, which provides favorable dynamical conditions for tropical cyclogenesis.

6.2 *Weak years*

During the weak CEF years, TCs usually form in the D3 region. As a result, only the TCs formed in D3 are discussed in this section. At 10 days before genesis (Fig.7a), the main region of development is dominated by zonal easterlies. Fig.7a indicates that a period of west-east fluctuation existed at approximately 4 days for both of the two CEFs. The westerlies move slowly eastward until genesis. As shown by Fig.7b, at 4 days before genesis, the main part of the monsoon trough is located near 135°E. The CEFs at 125°E and 150°E are relatively weak, and the southwest winds only reach about 2,000 km west of the genesis position. At about 2 days before genesis, the meridional winds of the two cross-equatorial channels strengthen one after another, and merge into the westerlies south of the monsoon trough. Therefore, the monsoon trough intensifies and extends eastward, reaching the genesis longitude. However, because of the strong easterlies, the monsoon trough with strong zonal convergence is sustained near 145°E without penetrating any further eastward. As a result, the TCs usually form within this convergence region between the monsoonal westerlies and the trade easterlies.

7 SUMMARY

This study examined the impact of the interannual variation of the CEF on the frequency and location of tropical cyclogenesis over the tropical WNP during typhoon season (June to October) from 1979 to 2010 by using best track data from JTWC, OLR data from NOAA, and the wind fields of the JRA-25 reanalysis data from the Japan Meteorological Agency. Correlation analysis reveals that the frequency of tropical cyclogenesis over the entire WNP is not significantly correlated with the intensity of CEF. However, TC activities to the east of 140°E are well correlated with the southerlies near 125°E and 150°E. This indicates that TC formations are enhanced over the eastern part of the WNP during strong CEF years.

Composite analysis was applied to explain the correlation between TC activity and the CEFs. We discussed the large-scale atmospheric factors responsible for tropical cyclogenesis, including low-level horizontal wind, convective activity, 200−850 hPa vertical wind shear, 850 hPa relative vorticity, and 200 hPa divergence. The results show that when the CEFs are strong, the monsoon trough shifts southeastward, and the above-mentioned factors are favorable for tropical cyclogenesis over the oceans east of 140°E. This leads to further TC formation in the eastern part of the WNP. The variability of these factors are opposite north of 25° N. This indicates that tropical cyclogenesis often occurs at higher latitudes when the CEFs are relatively weaker. In summary, when the CEFs are strong, the TC development area shifts southeastward; and when the CEFs are weak, TCs tend to form in the high-latitude regions.

The role of the CEFs on tropical cyclogenesis was also discussed based on barotropic energy conversion processes. The results show that during strong CEF years the horizontal convergence and shear of the mean flow are located in the eastern part of the WNP. As a result, the kinetic energy of the mean flow is transferred to eddy kinetic energy in the eastern part of the WNP. This leads to the growth of tropical synoptic perturbations, and further induces TC formation over the central and eastern parts of the WNP. However, during weak CEF years, the growth area of the eddy kinetic energy is located in the western part of WNP, and therefore TCs usually form in the west. The growth rate of the eddy kinetic energy is three times larger during the strong years than that during weak years due to the stronger meridional shear and stronger zonal convergence of the zonal wind. Tropical perturbations are therefore more likely to develop into TCs in the strong years.

We also investigated the impact of the rapid strengthening of the CEFs on tropical cyclogenesis. In both strong and weak years, the tropical cyclogenesis is related to the sudden intensification of nearby CEFs. Formation often occur about 2–4 days after the sudden intensification. During the strong years, the signals of sudden intensification are more evident than those in the weak years. The sudden intensification of the CEFs is accompanied by the eastward shift of the monsoon trough. Thus, more TC development occurs inside the monsoon trough during strong years. However, during the weak years, the variation of the southerlies appears as a 4-day period without significant eastward extension of the monsoon trough. During these years, TCs tend to form in the confluent regions, which are located at the eastern end of the monsoon trough.

This study discussed the influence of the CEFs over the western Pacific on tropical cyclogenesis over the southeastern and northern parts of the tropical WNP. The climatology of the monsoon trough is located over the WNP west of 140° E, and some of the local atmosphere related to tropical cyclogenesis is less affected by CEFs. The TC formation over these regions is therefore not well correlated with the intensity of the CEFs. The relationship between tropical cyclogenesis, CEFs, and the monsoon trough, and the essential factors controlling the interannual variation of TC activities, are important issues that need to be investigated in the future. Furthermore, Southern Hemisphere atmospheric signals are becoming more important to the prediction of TC activities over the WNP[19-20]. As a primary channel for the exchange of momentum and moisture between the Northern and Southern Hemispheres, the CEF forms a link for an inter-hemispheric teleconnection, which is another important issue.

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Citation: FENG Tao, SHEN Xin-yong, HUANG Rong-hui et al. Influence of the interannual variation of cross-equatorial flow on tropical cyclogenesis over the western North Pacific [J]. J Trop Meteorol, 2017, 23(1): 68-80.