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FORMATION AND STRUCTURAL EVOLUTION OF INITIAL DISTURBANCE OF TYPHOON FUNG-WONG (2008)

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Abstract: The formation of a tropical cyclone is the result of a process in which an initial disturbance evolves into a warm-core low-pressure system; however, the origin of the initial disturbance and the features of the initial fields are overlooked in most existing theories. In this study, based on FY-2C brightness temperature data and the Japan reanalysis dataset, the origin and evolution of the tropical disturbance that became Typhoon Fung-Wong (2008) were examined. The results demonstrated that the initial disturbance emerged within a saddle-type field with large vertical tropospheric wind shear. The vertical wind shear decreased with the adjustment of the upper circulation; moreover, accompanied by convection over the warm section around the upper cold vortex, it provided favorable thermal and dynamic conditions for the development of a tropical vortex. During its development, the zone of associated positive relative vorticity strengthened and descended from the mid-troposphere to lower levels. This rapid strengthening of lower-level vorticity was due to increasing convergence related to the intensification of the pressure gradient southwest of the subtropical high. This indicated that the upper cold vortex and West Pacific subtropical high played very important roles in this case.

Key words: structural evolution; diagnostic analysis; tropical cyclone Fung-Wong; vertical wind shear; subtropical high

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1 INTRODUCTION

Tropical cyclones (TCs) often threaten many coastal regions of the world. China, which is located on the western fringe of the Northwest Pacific Ocean, is one such area that is seriously affected by TCs. The generation of a TC is a complex process that involves the interaction of multi-scale weather systems. Research into TCs suffered originally because of the lack of observational data over ocean regions. However, the rapid development of satellite remote sensing has improved data acquisition; moreover, developments in computer technology have enhanced data processing capabilities, model performance, and computing power. Based on these improvements, observational composite analyses and numerical experiments are now used commonly by researchers to investigate TC generation (Zhi and Zhang[1]).

In previous studies, scientists usually focused on large-scale environmental parameters. For example, Gray[2] summarized six primary thermodynamic and dynamic conditions of TC generation: (1) ocean thermal energy (i.e., upper-layer (to a depth of 60 m) ocean temperature >26℃),(2) conditional instability through a deep atmospheric layer, (3) high relative humidity in the lower and middle troposphere, (4) the Coriolis effect, (5) large values of low-level relative vorticity, and (6) weak vertical shear of the horizontal winds. These six favorable factors have been generally accepted and are applied because of their existence during TC genesis, except for a very few cases (Chang et al.[3]; Cowan et al. [4]).

The generation of a TC is a process in which the initial disturbance is transformed into a system with a warm-core structure. The classic physical mechanisms for TC generation are conditional instability of the second kind (CISK) (Charney and Eliassen[5]) and wind-induced surface heat exchange (WISHE) (Emanuel[6]). To some extent, the CISK and WISHE theories are able to explain the process of intensification following the formation of a warm-core vortex; however, the origin of the initial disturbance and the features of the initial fields have not yet been described in detail (Zhang and Guo [7]). Therefore, while considering possible mechanisms for TC generation, researchers have proposed various hypotheses: southern hemisphere cold surges (Li[8]; Lee[9]; Xu and Wu[10]), terrain effects (Chang et al[3]; Farfan and Zehnder [11]), wind-driven surges (Lee et al.[12]), merging of mesoscale vortices (Simpson et al.[13]; Ritchie

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et al.[14]), reciprocity of a mid-level cyclonic vortex and lower-layer monsoonal trough[13], and convection onset[14]. Currently, investigations of TC triggering mechanisms have been limited to case studies; moreover, they are not understood fully for differing sea areas and environmental backgrounds[7].

In this study, considering the example of the Typhoon Fung-Wong, which formed in the Northwest Pacific in 2008, the TC-generation process, including how the initial disturbance was triggered and transformed into a tropical depression (TD) with a warm core, was investigated using FY-2C brightness temperature satellite data and the Japan reanalysis dataset. The features of TC's structural evolution during generation were examined because the majority of domestic research on the TC structure has usually concentrated on its development stage (Jiang et al.[15]; Shu et al.[16]; Chen et al.[17]).

2 FUNG-WONG OVERVIEW

According to the monitoring data of the Japan Meteorological Agency (JMA), Fung-Wong was classified as a TD over the Northwest Pacific Ocean, east of Taiwan Island, at 0800 BJT on July 24, 2008. It moved steadily to the west and northwest, was upgraded to a tropical storm at 1400 BJT on July 25, and then developed into a severe tropical storm the following morning. At 0800 BJT on July 27, Fung-Wong reached typhoon status with a minimum pressure of 960 hPa and a maximum sustained wind speed of 40 m/s near the surface center (Fig.1). Note that Fung-Wong was the strongest typhoon to make landfall in China in 2008, and it brought severe gales and torrential rain over a large area that persisted for a long-time.

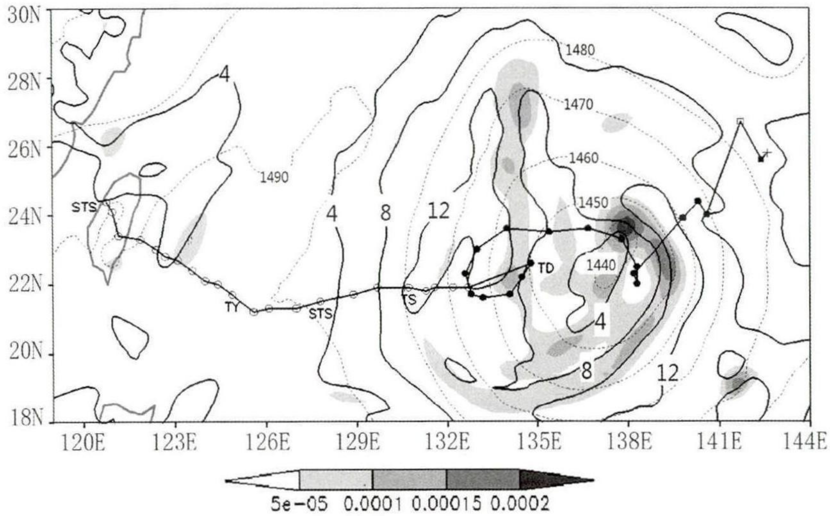


Figure 1. Storm track of typhoon Fung-Wong. Cross: tropical vortex center at 1400 BJT July 19, 2008; solid box: vortex center at 2000 BJT July 19, 2008; hollow box: vortex center at 0200 BJT July 20, 2008; solid circle: vortex center from 0800 BJT on July 20 to 0800 BJT on July 24, 2008 (the definition of vortex center can be seen in Table 1); hollow circle: best track and intensity data from the JMA following the TD formation. Overlain with relative vorticity (shading), wind speed (solid line, m/s), and 850-hPa geopotential height (dashed line, gpm) at 2000 BJT on July 21, 2008.

3 ORIGIN OF TROPICAL CYCLONE

To examine the formation process of the tropical cyclone, the preceding satellite images and streamline patterns were analyzed. Fig.2 shows that there was a saddle-type field that had persisted for many days before the formation of the low-pressure disturbance (the box in Fig.2a marks the disturbed region 24°–29° N, 142.5° E). At 2000 BJT on July 19, 2008, a N–S mesoscale convergence line (marked by a blue curve in Fig.2a) evolved within this region, with positive relative vorticity appearing at the sea level and in the lower-tropospheric layer (not shown).

Under favorable conditions, a tropical disturbance will undergo a continuous self-development process when its structure presents a closed cyclonic circulation near the surface layer (Briegel and Frank[18]). In the case of Fung-Wong, there appeared to be a closed sea-level streamline at 0200 BJT July 20, 2008 (Fig.2b), and a closed isobar at 0800 BJT on July 20, 2008 (closed solid line in Fig.2c). The locations of the low-pressure center coincided with the circulation center, and the convective cloud system in the southwestern periphery of the disturbance became further developed. Zehr's [19] statistics for the Northwest Pacific revealed that intense development of convective cloud as well as the appearance of a corresponding mesoscale vortex are conditions necessary for a TD to form.

In the development process of the tropical vortex, the temperature at its center remained cold compared with its surroundings, as shown at 950 hPa at 0800 BJT July 21, 2008 (Fig.3a). It was not until 2000 BJT on Ju-

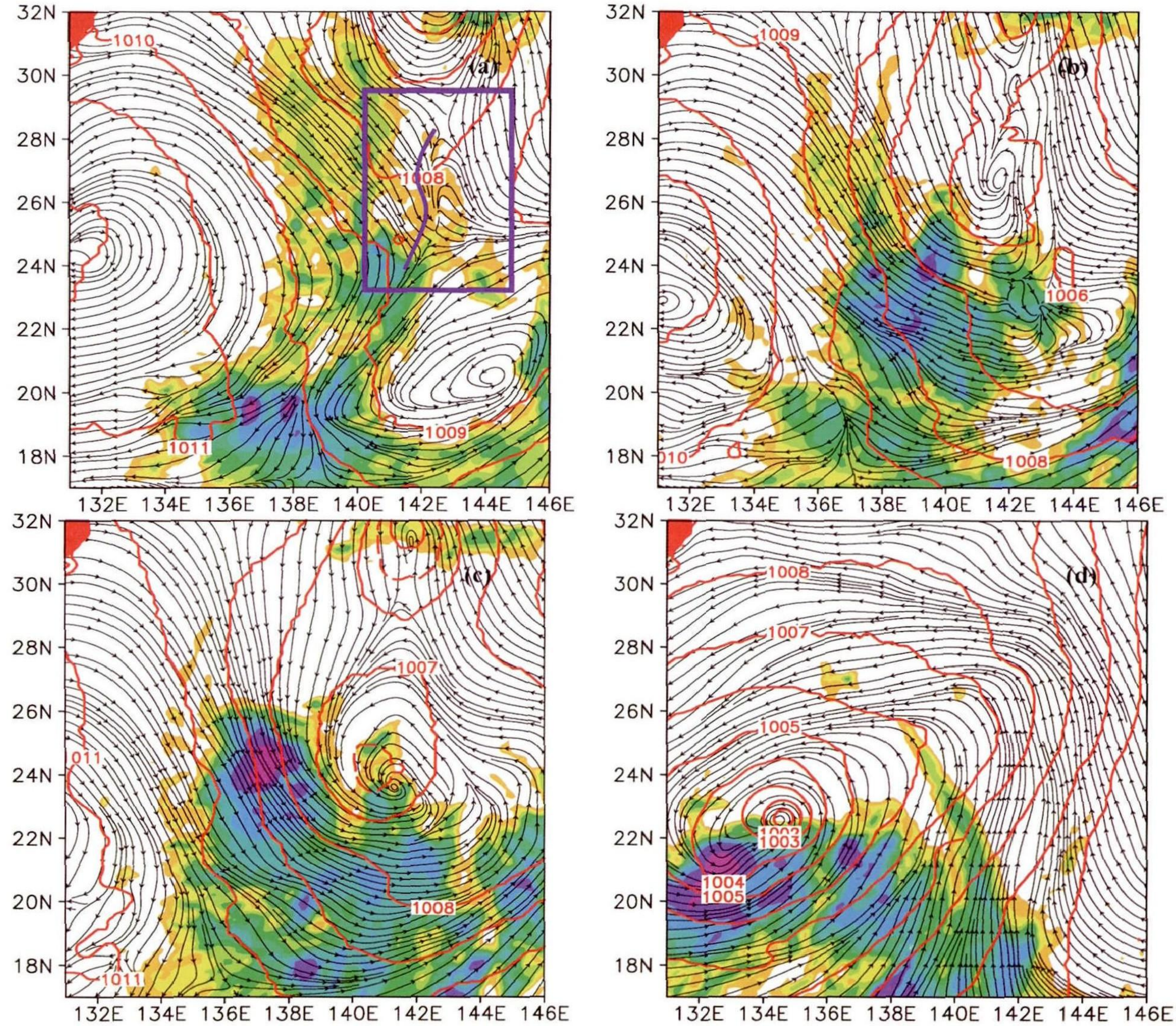


Figure 2. Temperature of brightness blackbody (℃, shaded), sea-level streamline analysis, and pressure field (1-hPa-interval contours, of which the dotted isoline represents pressure value of 1006.5 hPa): (a) 2000 BJT July 19; (b) 0200 BJT July 20; (c) 0800 BJT July 20; and (d) 0800 BJT July 24, 2008.

ly 21, 2008, that the vortex transformed into a warm-core system in an environment with sufficient vapor transportation. By that time, the sea-level pressure near the vortex center had deepened, the radius of maximum wind speed had decreased (Fig.3b), and positive relative vorticity had become distributed closely around the vortex center. All of these indicate that the cyclonic circulation of the tropical vortex had developed quite well by 2000 BJT on July 21, 2008.

In the following days, the vortex continued to develop steadily. By 0800 BJT July 24, 2008, the sea-level pressure near the vortex center had fallen to 1002 hPa and the wind speed had reached the intensity of a tropical depression; thus, the system was designated as Fung-Wong (Fig.2d). At this time, according to the brightness temperature analysis (Fig.2d), severe convective clouds developed to the south of the cyclonic circulation within an area of nearly 2°×2° latitude and longitude. The process of TC genesis (Table 1) demonstrated that there were two very important moments: the formation of a sea-level mesoscale convergence line, i.e., the initial disturbance at 2000 BJT July 19, 2008, and the intensification and transformation of the vortex into a lower-layer warm-core system at 2000 BJT on July 21, 2008.

To identify the location of the system before its designation as Fung-Wong, it was defined as follows. Using the appearance of closed sea-level isobars at around 0800 BJT July 20, 2008, as the demarcation point, the location of the vortex center after this time was based on the minimum sea-level pressure (solid circles in Fig.1) while its location before this time was defined as the center of circulation (shown in Table 1, and marked by other symbols in Fig.1).

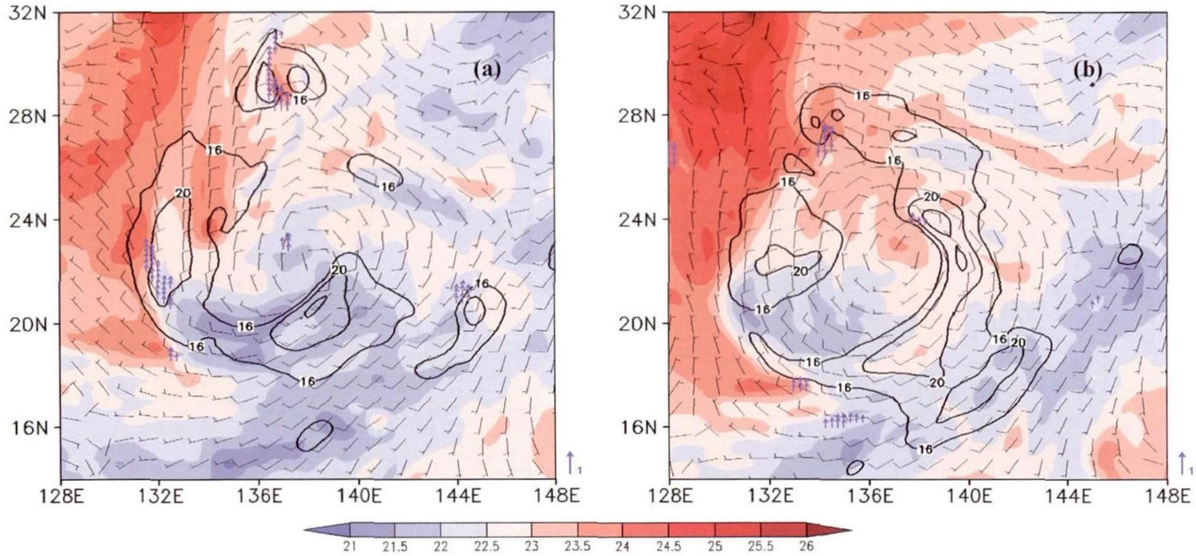


Figure 3. 950-hPa wind field (each full flag represents 4 m/s), temperature (℃, shaded), water vapor flux (g/(cm·hPa·s), contours), and vertical velocity <-0.3 Pa/s (the arrow in the bottom-right corner represents -1 Pa/s): (a) 0800 BJT July 21; (b) 2000 BJT July 21, 2008.

Table 1. Formation process of Typhoon Fung-Wong.

Development stage	Time	Vortex center definition
Saddle-type field	1400BJT19	Sea-level saddle-type field center
Mesoscale convergence line	2000BJT19	Sea-level vorticity center
Closed streamline	0200BJT20	Sea-level stream field vortex center
Closed isobar	0800BJT20	Sea-level pressure center
Vortex development	2000BJT21	Sea-level pressure center
Tropical depression designated	0800BJT24	Sea-level pressure center

4 GENERATION OF INITIAL DISTURBANCE

4.1 *Large-scale environment*

As shown in Figs.4a-c, an upper cold vortex existed to the southeast of the area in which the TC was generated at 2000 BJT on July 19, 2008. This type of vortex is always concentrated near 23°N in the tropical upper troposphere of the Northwest Pacific in summer. It persisted for a long time before the occurrence of the initial disturbance and its early growth stage (the cold vortex was generated at 0200 BJT July 18, 2008, and it moved southwest; figure omitted). From the zonal cross section of the temperature anomaly through the center of the cold vortex (Fig.5), a warm core can be seen above 200 hPa (warmest at 150 hPa), whereas a cold core is present below 200 hPa down to the mid-troposphere. The meridional wind component shows a strong northerly component to the west of the vortex center and a strong southerly component to its east between 100 and 300 hPa. Under the upper cold vortex, there is a cyclonic circulation with closed sea-level streamlines with surrounding convective clouds (Fig.2a).

A low-pressure area to the north of the initial disturbance in the lower troposphere can be seen at 850 hPa (Fig.4a) but not at 500 hPa (Fig.4b). At 200 hPa (Fig.4c), where the TD would form, there is a zone of strong northeasterly winds between the upper cold vortex to the southeast and the subtropical high to the northwest. The strong upper-layer outflow, corresponding to the northerly component of the upper cold vortex in Fig.5, superposed over the lower-layer pressure trough. This led to the enhancement of upper-layer divergence and upward vertical movement within the lower and middle levels, both of which are propitious to development of surface low pressure.

Figure 2 shows that the initial disturbance occurred within a saddle-type field, where two cyclonic vortices were positioned to the north and south, which corresponded to the shallow low pressure described in Fig.4a and the lower circulation of the upper cold vortex, respectively. This saddle-type field had persisted for a long time before 1400 BJT July 19, 2008 (Fig.6), with gentle but erratic winds and stable pressure. A low-level quasi-static region of wind combined with an upper-layer zone of strong winds can lead to considerable vertical wind shear, which is beneficial to the development of lower-layer positive vorticity. As the subtropical high strengthened and moved slightly eastward by 2000 BJT

July 19, 2008 (Fig.2b), the two high pressure systems to the west and east of the saddle clearly squeezed toward each other, which resulted in denser streamlines and a stronger wind between them. Therefore, the area in which positive vorticity increased most obviously is the region where the wind gradient was the strongest (Fig. 6); moreover, at this time, the mesoscale convergence line appeared in the region with the positive vorticity tendency.

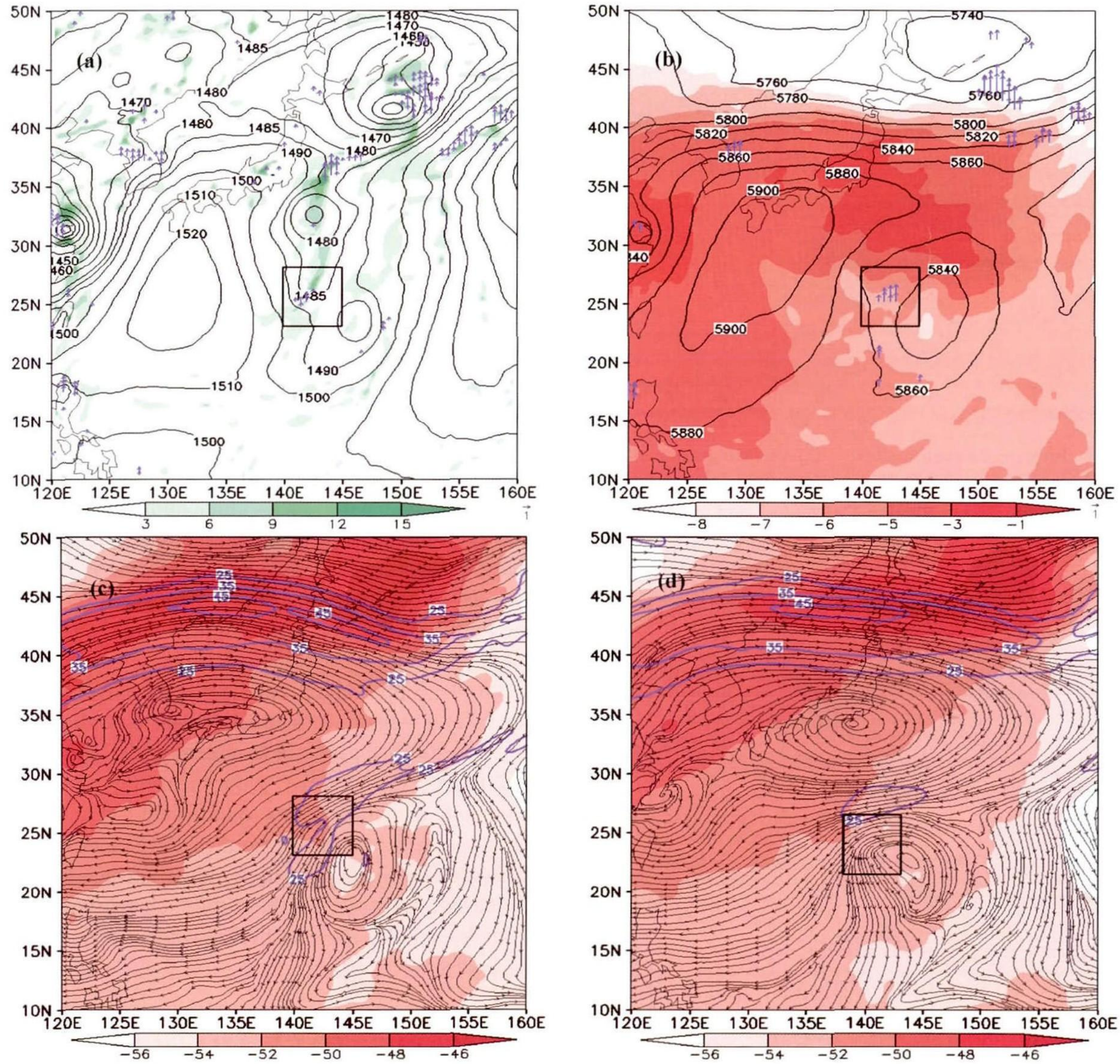


Figure 4. (a) 850-hPa and (b) 500-hPa potential height (gpm, contours), relative vorticity (10^{-5} /s, shading), and vertical velocity (Pa/s, the arrow in bottom-right corner represents −1 Pa/s) at 2000 BJT on July 19, 2008. 200-hPa streamline analysis, temperature (℃, shading), and horizontal wind speed (m/s, contours) at (c) 2000 BJT July 19 and (d) 0800 BJT July 20, 2008. The box marks the location of the disturbance.

4.2 *Vertical wind shear*

Before a tropical disturbance can occur, the thermodynamic environmental conditions concerning the sea surface temperature and deep moist layer must be met; therefore, vertical wind shear is another important condition necessary for the generation and development of this TC.

The average vertical wind shear around the vortex center in 5×5 grids was calculated. Fig.7 shows that the vertical wind shear reached speeds of 16 m/s or more before 0800 BJT July 20, 2008.

The previous sections have described how the strong vertical wind shear resulted from the strong upper northeasterly outflow. This not only enhanced the upper-layer divergence but also facilitated the generation of a rising movement in the downshear direction, which was vital for the initial disturbance (Rotunno and Klemp [20]). However, as an initial disturbance develops into a TD, strong vertical wind shear becomes an adverse condition. Weak shear indicates that latent heat released by the convection of the initial disturbance is concentrated within a limited space and generates the warm core. This lowers the pressure of the initial disturbance and strengthens the low-level cyclonic circulation, and the cyclone can develop. In addition, weak vertical wind shear is beneficial to vertical coupling between the upper and lower layers, which is one of the critical factors in TC genesis (Gray[21]).

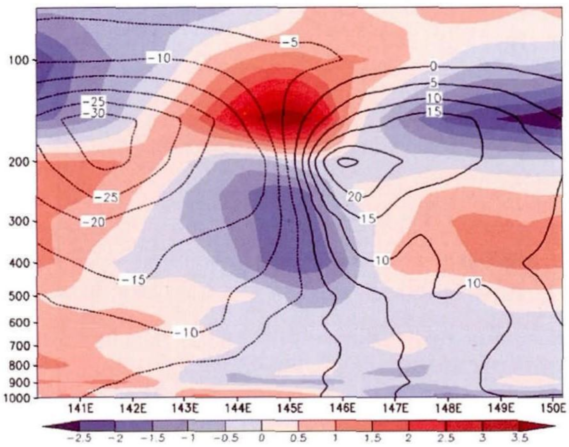


Figure 5. Temperature anomalies (℃) vertical profile of the upper cold vortex at 2000 BJT July 19, 2008, and meridional wind component (m/s); positive for south wind, negative for north wind.

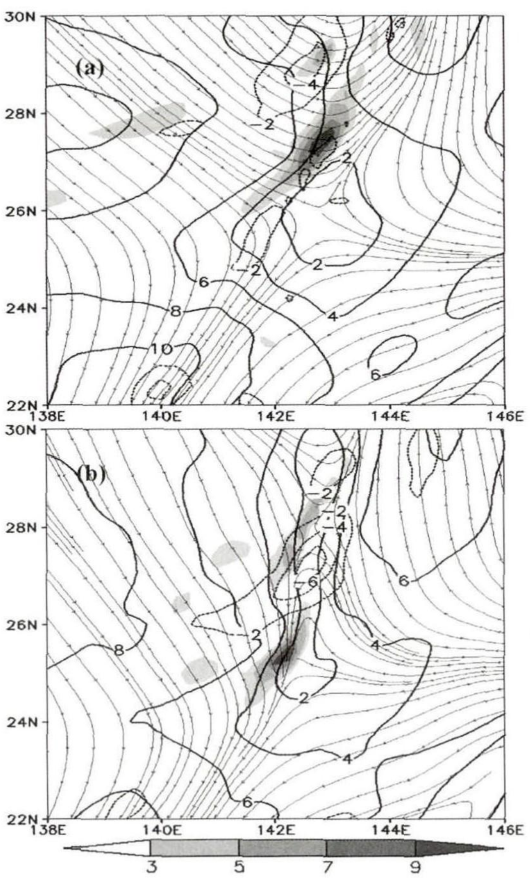


Figure 6. (a) 950-hPa streamlines, wind speed (m/s, solid lines), divergence (10^{-5}/s, dashed lines), and total vorticity tendency ($10^{-9}/s^2$, shading) 1400 BJT July 19, and (b) 950-hPa streamlines, wind speed (m/s, solid lines), divergence (10^{-5}/s, dotted lines), and vertical vorticity tendency ($10^{-9}/s^2$, shading) 2000 BJT July 19, 2008.

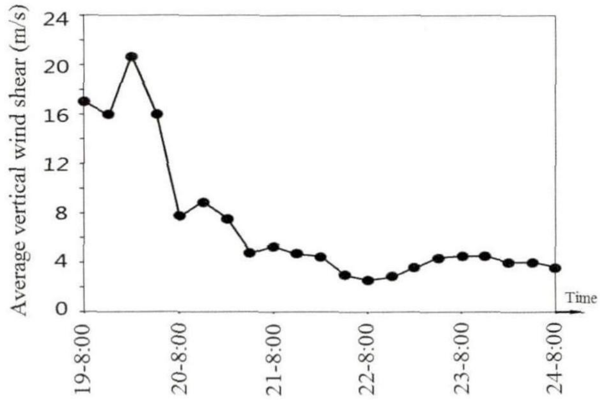


Figure 7. Change of average vertical wind shear (m/s) over the sea.

Statistics show the critical vertical wind shear for the generation of TCs in the western Pacific is approximately 8–10 m/s (Li[8]; Zehr[19]; Chen et al.[22]). When the initial disturbance of Fung-Wong occurred, the associated vertical wind shear decreased immediately, falling to the critical value by 0800 BJT July 20, 2008, and far below that by the following morning. As variation of vertical wind shear might indicate the evolution of a TC, changes in the upper circulation have to be investigated further to explore the reasons behind the changes in vertical wind shear. Controlled by the gradient between the upper subtropical high and the cold vortex, a strong vertical wind shear was generated between the upper-layer zone of strong winds and the lower-layer region of calmer winds before 0800 BJT July 20, 2008 (Fig.4c). Subsequently, the upper circulation transformed as the subtropical high moved eastward and the cold vortex diminished (Fig.4d). Thus, the upper-level zone of northeasterly winds extended eastward and the in situ upper-level northeasterly winds decreased. Because of the subtle change of the lower-layer wind, the vertical wind shear decreased markedly.

5 MECHANISM OF STRUCTURAL EVOLUTION DURING TD GENERATION

5.1 *Structure evolution*

To study the structural characteristics of the disturbance as it developed into a TD, the temporal evolution of some physical quantities over the center of the disturbance was analyzed. The definition of the center was based on Table 1, and the analysis time extended from 2000 BJT July 19, to 0800 BJT July 25, 2008. The temperature anomaly was defined as the difference between the mean temperatures of 1°×1° and 10°×10° areas centered over the disturbance.

Figure 8a shows strong convergence below 700 hPa (maximum at 700 hPa) and strong divergence at 500 hPa (maximum at 200 hPa) from 1400 to 2000 BJT July 19, 2008. The two principal reasons for this were the strong upper outflow and the change in lower-level circulation because of the squeezing of the high-pressure areas on either side of the saddle-type field. The vertical distribution of pseudo-equivalent potential temperature demonstrates that strong ascending motion provided a channel of high potential energy, linking the upper and lower layers of the troposphere, which was beneficial for the initial perturbation. Both convergence and divergence weakened from 2000 BJT July 19 to 1400 BJT July 21, 2008, compared to the former stage. During this period, there was convergence in the low-level troposphere and divergence area in the upper troposphere with a maximum at 300 hPa, which was relative to strong effluent current from the northwest of an upper cold vortex and the south of a subtropical high. In the mid-troposphere, convergence was weak and the pseudo-equivalent potential temperature was low, showing that stratification was not stable within the low-level troposphere. During this time, the low-level positive relative vorticity center dropped from 700–600 to 800 hPa (Fig.8b). Concurrently, the upper cold vortex began to weaken, although evidence of its circulation remained at higher levels (Fig.4d). The associated relative vorticity appeared at 300 hPa (Fig.8b) as it approached the tropical vortex. The development of low-level relative vorticity corresponded to the warming of the air and the positive temperature anomaly center that was located below the relative vorticity center.

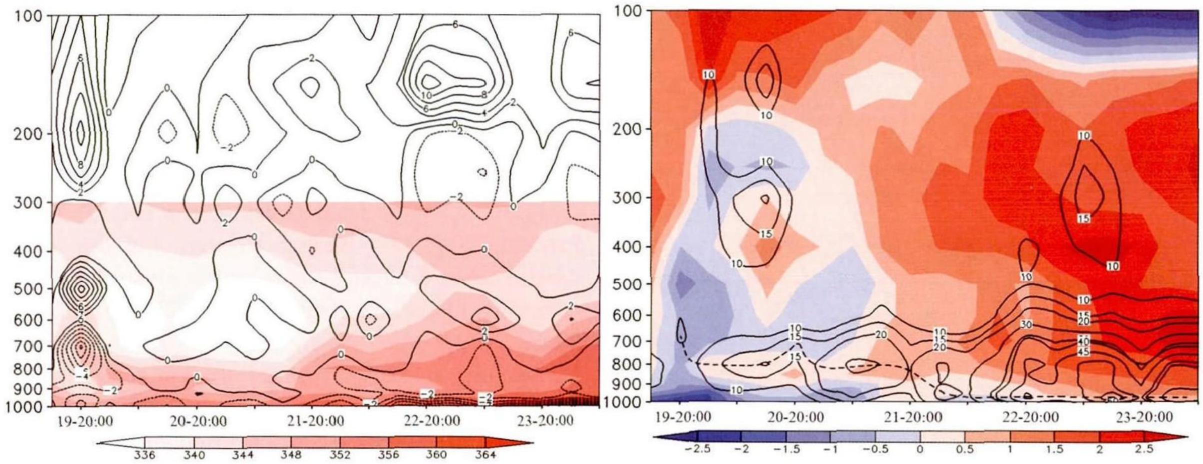


Figure 8. Height-time cross sections: (a) pseudo-equivalent potential temperature (K, shading) and divergence (10^{-5}/s, contours) and (b) temperature anomaly (℃, shading) and relative vorticity (10^{-5}/s, contours) through the disturbance center.

Low-level convergence strengthened after 1400 BJT July 21, 2008, and it extended rapidly to 750 hPa. Concurrently, the zone of high pseudo-equivalent potential temperature in the lower-level troposphere was lifted with a steeper gradient. Correspondingly, divergence in the upper layer strengthened to some extent, but it remained weak by comparison. The low-level positive relative vorticity center dropped again to 975 hPa by 0800 BJT July 22, 2008. During the development of this TC, it is evident that the positive relative vorticity center was located initially in the mid-troposphere and subsequently descended. The vorticity center was present near 975 hPa, i.e., almost at the bottom of the troposphere, for one or two days before the TC was named (Fig.8b). Moreover, the temperature over the tropical vortex center continued to increase until the low-level cold core eventually became a warm core, as seen in Fig.3 (Fig.8b).

During the period from 2000 BJT July 22, 2008, until the time the TC was named (0800 BJT July 24, 2008), this system developed continuously, as evidenced by the increase in the zone of high pseudo-equivalent potential temperature and the strengthening of the low-level convergence. The convergence zone rose to 700 hPa with the maximum in the boundary layer reaching -8×10^{-5}/s, and divergence above 200 hPa increased concomitantly. During this time, the associated relative vorticity increased continuously, although its center remained at 975 hPa. Over time, the position of the main warm core in the upper layer descended, but later than that of the positive relative vorticity center. During the final two days before the tropical disturbance was designated, it continued to develop steadily.

5.2 *Mechanism for structure evolution*

As mentioned above, before 0800 BJT July 21, 2008, the movement of the subtropical high played an important role in both the generation of the initial tropical perturbation and the decrease of vertical wind shear. Between 1400 BJT July 21 and 0800 BJT July 22, 2008, low-level convergence clearly strengthened (Fig. 8), relative vorticity increased and its position descended, and the temperature of the core changed from cold to warm (Fig.3). These and other related conditions portended the development of the tropical perturbation.

Figure 9 shows the wind and relative vorticity tendency at 950 hPa at 2000 BJT July 21, 2008. This moment is significant in the descent of the relative vorticity center (Fig.8). Eq.(1) shows the relative vorticity equation. The distribution of the total vorticity tendency($\frac{\partial\zeta}{\partial t}$)

around the low-pressure area (Fig.9) shows three centers located to the north, east, and southwest, which are coincident with the centers of the divergence contribution term ($-\left(f+\zeta\right)\cdot\left(\dfrac{\partial u}{\partial x}+\dfrac{\partial v}{\partial y}\right)$). Because the divergence term contributes significantly to the first two centers, it can be considered the most important with regard to the increase of relative vorticity.

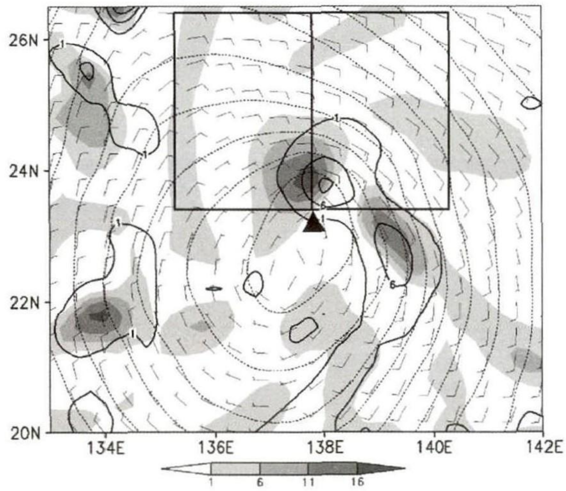
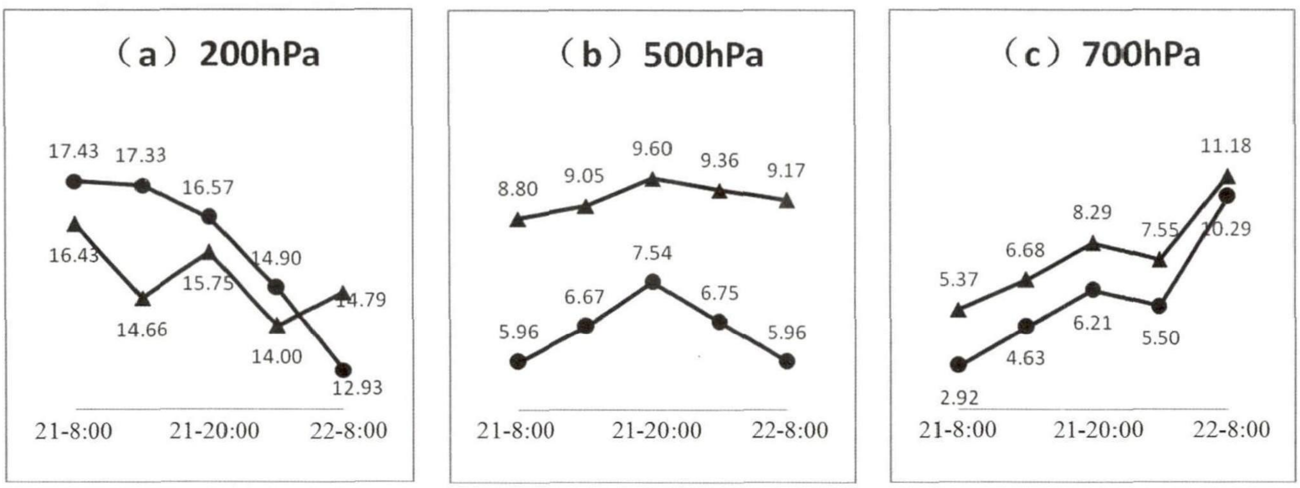


Figure 9. 950-hPa wind and potential height field (center pressure, 475 hPa; dotted contours, interval 5 hPa), total vorticity tendency ($10^{-9}/s^2$, shading), and convergence-divergence ($10^{9}/s^2$, solid contours) 2000 BJT July 21, 2008. ▲ represents the low-pressure center.

$$\frac{\partial\zeta}{\partial t}=-\vec{V}\cdot\nabla\zeta-\beta v-\omega\frac{\partial\zeta}{\partial p}+\left(\frac{\partial\omega}{\partial y}\frac{\partial u}{\partial p}-\frac{\partial\omega}{\partial x}\frac{\partial v}{\partial p}\right)-\left(f+\zeta\right)\cdot\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right) \quad (1)$$

Depending on the divergence equation, there are many factors that can lead to the change of divergence (Lv et al.[23]). The above analysis indicates that 1) the generation of divergence is related to the change of the isobaric surface slope when vertical wind shear is weak, 2) convergence is obvious in the low-levels, and 3) lower-layer relatively vorticity and clear ascending motion associated with the disturbance are not apparent. The relative vorticity strengthened most to the north and east of the low-pressure center, where the isobars are closer together; therefore, the evolution of the pressure gradient in the rectangular window, which was shown in Fig. 9, was analyzed further. Fig.10 shows the pressure gradient and wind speed at various levels from 0800 BJT July 21 to 0800 BJT July 22, 2008. During this time, the pressure gradient at 200 hPa decreased, whereas it changed little at 500 hPa. In the low-level troposphere, the increase in the pressure gradient was more obvious at lower heights. Moreover, the pressure gradient increased most rapidly during the six hours prior to 2000 BJT July 21, 2008. From the change of pressure gradient at each level, it can be seen that the development of low-level tropospheric vorticity is related to the increase of the pressure gradient. The window-averaged wind showed similar characteristics: the upper wind decreased; mid-level wind strengthened before 2000 BJT July 21, 2008, but weakened later; and the low-level wind increased throughout, with the most rapid increase occurring in the six hours prior to 2000 BJT July 21, 2008.

Figure 3a shows that the vortex center was relatively colder than its surroundings at 950 hPa at 0800 BJT July 21, 2008. Ascent occurred mainly in the surrounding warm air transporting a great deal of water vapor and increasing the relative vorticity. The development of organized convection over the perturbation helped the vertical movement of water vapor and the release of latent heat of convection. Because of this convergence of water vapor, the entire troposphere experienced warming and increased humidity. Through the release of the latent heat of convection, thermal energy accumulated around the vortex and the relative vorticity continued to develop. The vortex finally translated into a circulation with a warm core at 2000 BJT July 21, 2008 (Fig.3b). When considering the cause of the warm core development, mid-level relative vorticity center de-



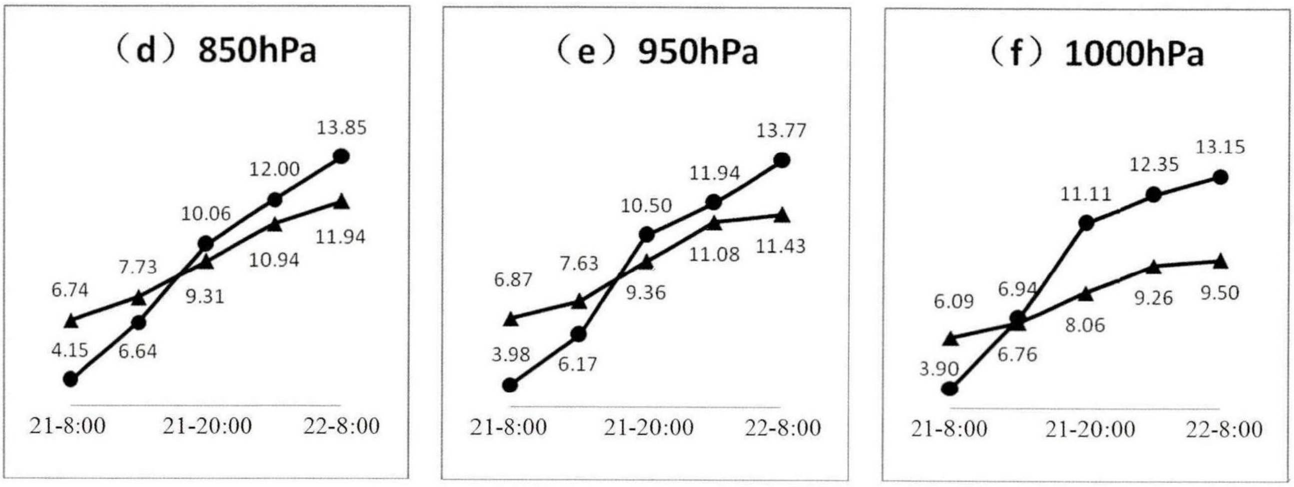


Figure 10. Meridional pressure gradient (Pa/latitude) and window-averaged wind speed (m/s).

scent, and rapid development of the system, the reason could be linked to the increase of the pressure gradient around the low-level vortex. This led to the increase of relative vorticity in the area with compact isobars, which lowered the pressure of cyclone and strengthened its rotation.

6 CONCLUSIONS

Taking Typhoon Fung-Wong (2008) as an example, this study analyzed the processes of its generation and structural evolution based on FY-2C brightness temperature data and the Japan Reanalysis dataset:

The initial disturbance occurred within a persistent saddle-type field. This became distorted and it produced a mesoscale convergence line with concentrated positive vorticity because of the eastward movement of the low-level subtropical high. In the upper troposphere, as a strong northeasterly wind between the northwest of the subtropical high and the southeast of the upper cold vortex became a strong outflow current, high-level divergence was strengthened and strong vertical wind shear was generated. These characteristics were beneficial to the development of the initial disturbance by providing conditions suitable for vertical motion. The mid-troposphere above the initial disturbance had relatively low temperature and high humidity. This provided a channel of high potential energy between the high- and low-level troposphere, which allowed the mesoscale convergence line to develop into a tropical disturbance with closed isobars within 24 h.

Following the generation of the initial disturbance, as the subtropical high moved eastward and the upper cold vortex diminished, vertical wind shear above the tropical disturbance weakened and ultimately reached the critical value that presented beneficial environmental conditions for the continued development of the disturbance into a tropical low-pressure system. Furthermore, both the warm air rising around the cold vortex and the development of convective cloud to the northwest provided considerable potential heat that was advantageous to the reduction of low-level pressure.

When the tropical disturbance developed into a tropical low-pressure system, the associated positive relative vorticity moved its center from the mid-troposphere to lower levels and it strengthened. The analysis of the vorticity budget showed that the increase of low-level vorticity was related to the increase of convergence, which resulted from the increase of the pressure gradient in response to the westward movement and strengthening of the northeastern subtropical high.

There are three important conclusions that can be drawn from the findings of this study. First, during the generation of the TC Fung-Wong, the movement of the subtropical high was vital for the formation of the initial disturbance and for its further development via its influence on the strength of the vertical wind shear. Second, the upper cold vortex was important because convection in the surrounding warm air provided suitable thermal conditions for the development of the initial vortex and the change in its strength affected the vertical wind shear. Third, following the occurrence of the tropical disturbance and its subsequent development into the tropical low-pressure system, the time at which the vertical shear reached the critical condition was much earlier than when the TD was named. Therefore, it could be implied that the alteration of the vertical shear could be considered an indicator for early forecasting of the development of typhoons.

In the case of the generation of Fung-Wong, note that the surface disturbance began to develop rapidly following the disappearance of the upper cold vortex; thus, their relationship and the mechanism of influence should be the subjects of further research.

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