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Location of the preferred region for tropical cyclogenesis in strong monsoon trough pattern over the western North Pacific

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Abstract

We examined the processes of tropical cyclogenesis in strong monsoon trough pattern over the western North Pacific (WNP) using reanalysis data and numerical experiments. Composite analysis showed that more tropical cyclones are likely to form in the central WNP (130°E–165°E) and that fewer tropical cyclones appear in the western (120°E–130°E) and eastern (165°E–180°E) WNP when monsoon trough extends southeastward. Numerical experiments with the same weak artificial vortices inserted into eight different regions of the monsoon trough showed that weak tropical disturbances tend to develop more rapidly in the central WNP near 140°E–160°E, particularly near 150°E–155°E when the monsoon trough extends eastward, whereas weak tropical disturbances tend to develop more slowly in the eastern WNP near 165°E–170°E and do not form in the western WNP near 120°E–137.5°E. Our modeling results are consistent with the observational analyses. The failure of tropical cyclogenesis in the western WNP is due to the decrease of the moisture and heat (including the sensible and latent heat) from the underlying ocean, whereas large vertical wind shear and dry conditions in the upper level of the vortex reduce the gradient of intensification of tropical disturbances in the eastern WNP when the vortices have a similar initial intensity.

Key words: monsoon trough; tropical cyclogenesis; idealized model simulation
1. Introduction

Tropical cyclones are one of the most destructive atmospheric systems worldwide and often give rise to great socioeconomic costs if they make landfall near highly populated coastal regions. Disasters associated with tropical cyclone activity have become more serious in recent decades (Mendelsohn et al. 2012) and predictions of the location of cyclogenesis and the subsequent tracks of tropical cyclones are of profound socioeconomic significance. Tropical cyclones are most active in the western North Pacific (WNP) basin where 30% of tropical cyclones form (Neumann 1993). A monsoon trough occurs at lower levels over the WNP during the tropical cyclone season and generally consists of a meridional shear line and an area of zonal convergence (Li 2012). The development of about 70% of tropical cyclones is related to this monsoon trough (Ritchie and Holland 1999; Molinari and Vollaro 2013; Yoshida and Ishikawa 2013; Feng et al. 2014), which is a region favorable for the development of synoptic-scale disturbances (Shapiro 1977; Zehnder 1991).

Previous observational studies have suggested that the interannual variation of tropical cyclogenesis is closely related to the El Niño–Southern Oscillation (e.g., Chen et al. 1998; Chia and Ropelewski 2002; Wang and Chan 2002; Harr and Chan 2005; Camargo et al. 2007; Wu et al. 2012; Cao and Wu 2018b; Wu et al. 2019). It is generally agreed that tropical cyclones occur more frequently over the southeastern WNP and less frequently over the northwestern WNP during El Niño summers (Chia and Ropelewski 2002; Wang and Chan 2002). Wu et al. (2012) reported that more tropical cyclones form east of 160°E during strong monsoon trough years than during weak monsoon trough years. Their study indicated that the interannual variation in the location of tropical cyclogenesis over the WNP is consistent with the interannual variation in the east–west extension of the monsoon trough. However, the region with
the largest interannual variation in tropical cyclogenesis (in the eastern part) does not correspond to the region of formation of most tropical cyclones (in the middle part) over the WNP when the monsoon trough is strong. As shown by Wu et al. (2012, Fig. 3a in their paper), more tropical cyclones appear near 150°E than east of 160°E. The objective of our study is therefore to investigate the location of the preferred region of tropical cyclogenesis in strong monsoon trough pattern over the WNP.

Cao et al. (2014) used an idealized simulation to show that monsoon trough anomalies on the interannual timescale strongly affect tropical cyclogenesis through both dynamic (cyclonic circulation) and thermodynamic (humidity) factors. The initial large-scale fields in the work of Cao et al. (2014) included only the interannual anomalies of the monsoon trough against a resting background. Cao and Wu (2018a) added climatological mean fields into the initial large-scale fields during the strong and weak monsoon trough years to investigate how the tropical disturbance is triggered under seasonal mean background fields with no pre-existing tropical disturbance. It means that only the composite large-scale fields during the strong and weak monsoon trough years are specified including climatological mean fields, and no pre-existing synoptic-scale tropical disturbances exist at the initial time. The numerical simulations suggested that a tropical disturbance tends to occur near 150°E–160°E when a strong monsoon trough extends eastward, whereas a tropical disturbance tends to form near 120°E–130°E when a weak monsoon trough retreats westward (Cao and Wu 2018a). As an extension of the work of Cao and Wu (2018a), an artificial vortex is inserted into different regions of the strong monsoon trough pattern to determine the preferred region for the tropical cyclogenesis using an idealized numerical model. Compared with the work of Cao and Wu (2018a), the purpose of this study is to examine the effects of large-scale flows on the transition
from the pre-existing tropical disturbances into tropical cyclone over the WNP.

This paper is organized as follows. Section 2 examines the large-scale wind patterns during strong monsoon trough years and the preferred region for the tropical cyclogenesis over the WNP from reanalysis data. Section 3 introduces the design of our numerical experiments and investigates the evolution of tropical cyclogenesis and its sensitivity to the position of the vortex in numerical experiments. Contrasts in the tropical cyclogenesis in two sensitivity experiments are highlighted in Section 4 and a summary and discussion are presented in Section 5.

2. Observational analyses

The large-scale wind fields of strong monsoon trough pattern were the same as those in Cao et al. (2016) and Cao and Wu (2018a). The large-scale fields related to strong monsoon troughs were obtained from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis datasets starting in 1948 (Kalnay et al. 1996) and based on the zonal variation of the intensity of the monsoon trough. The intensity of the monsoon trough was defined by the seasonal mean relative vorticity averaged between 5°N and 20°N from July to November during the time period 1979–2007 (Wu et al. 2012). To extract the dominant interannual variation in the intensity of monsoon trough, an empirical orthogonal function (EOF) method was performed for the time and longitude datasets of latitude averaged relative vorticity between 5°N and 20°N (Lorenz 1956). The EOF technique is a rather common statistical tool for exploratory data analysis and
dynamical mode reduction, which has a fundamental importance in atmosphere, ocean, and climate science (Hannachi et al. 2007; Monahan et al. 2009). The purpose of EOF method is to transform a spatial-temporal dataset to spatial patterns and temporal projections of these patterns. The former is the EOFs and the latter is the temporal coefficients of the EOF patterns called as principal components (PC). Note that the spatial patterns and temporal coefficients produced by EOF analysis are orthogonal. Based on the time coefficients of the leading empirical orthogonal function mode, we identified seven strong monsoon trough years (1982, 1986, 1990, 1991, 1997, 2002 and 2004) and seven weak monsoon trough years (1984, 1988, 1995, 1996, 1998, 1999, 2007), which are approximately 50% of the 29-year period. The absolute PC value larger than 4.5 is used to find the strong and weak monsoon trough years (Fig. 2a in Wu et al. (2012)). The composite method was then applied to obtain the large-scale wind fields in the strong monsoon trough years. A detailed description of this method was reported by Wu et al. (2012).

Information about tropical cyclogenesis over the WNP was obtained from the National Climate Data Center’s International Best Track Archive for Climate Stewardship v3r10 (Knapp et al. 2010). The tropical cyclone best track data include the longitude and latitude of the center of the tropical cyclone and the maximum wind speed at six-hour intervals. We defined the genesis of a tropical cyclone as the first record of a maximum sustained wind speed >25 knots (~12.9 m s\(^{-1}\)). The present analysis is focused on the region (120°E–180°E, 0°–30°N).
Figure 1 shows the lower-level wind fields and tropical cyclone genesis during the strong and weak monsoon trough years. Note that the monsoon trough is characterized by the monsoon westerly winds to its south and the trade easterly winds to its north denoted by the black dashed line. In strong monsoon trough years, the monsoon trough expands southeastward to near 170°E (Fig. 1a), similar to previously reported observations (Wu et al. 2012; Cao and Wu 2018a), whereas in weak monsoon trough years, the monsoon trough retreats northwestward to 140°E. If we divided the WNP into northwestern, northeastern, southwestern and southeastern parts according to the latitude of 15°N and the longitude of 150°E, more TCs are generated in southeastern part of the WNP during the strong monsoon trough years compared to the weak monsoon trough years, which are consistent with the previous studies (Wang and Chan 2002; Wu et al. 2012). In this study, a finer partition is applied to investigate quantitatively the preferred region for tropical cyclogenesis in the strong monsoon trough pattern over the WNP.

We divided the whole WNP into six sub-regions at intervals of 10°E. The frequencies of tropical cyclogenesis in the regions 130°E–140°E, 140°E–150°E, 150°E–160°E and 160°E–170°E during the strong monsoon trough years were 28, 22, 28 and 23, respectively, and the frequencies of tropical cyclogenesis in the regions 120°E–130°E and 170°E–180°E were 12 and 6, respectively (Fig. 1a). To examine the sensitivity of results to the division of the WNP domain, we divided the WNP into five sub-regions with an uneven range as shown in Table 1. The frequency of tropical cyclogenesis in the regions 135°E–145°E, 145°E–155°E and 155°E–165°E were 22,
29 and 26, respectively, whereas the frequency of tropical cyclogenesis in the regions 120°E–135°E and 165°E–180°E was 28 and 14, respectively. If the TC numbers in the sub-region are normalized by averaging area as shown in Table 1, no remarkable difference is found between two types of the domain divisions. Based on the two different divisions, more tropical cyclones appeared in the central WNP near 130°E–165°E and fewer tropical cyclones appeared over the western (near 120°E–130°E) and eastern (near 165°E–180°E) WNP in strong monsoon trough years. On the other hand, during the weak monsoon trough years, more tropical cyclones appeared in the western and central WNP and fewer tropical cyclones appeared over the eastern WNP (Fig. 1b). In this study, we only examine quantitatively the preferred region of tropical cyclogenesis in strong monsoon trough pattern over the WNP.

3. Numerical simulation

3.1. Design of numerical experiments

We used the non-hydrostatic Advanced Research Weather Research and Forecasting (WRF-ARW; Skamarock et al. 2008) model version 3.3.1 to simulate the early stage of tropical cyclogenesis under the background of strong monsoon trough pattern. A single computational domain with a grid spacing of 30 km centered at 15°N was used to weaken the potential mesh interference on the structure of the monsoon trough (Li et al. 2006). The model domain included 391 × 391 grid points on a beta plane with 31 vertical levels from the surface to 50 hPa. The large-scale fields at the
initial time included zonal wind, meridional wind, sea-level pressure, surface pressure, temperature, relative humidity, geopotential height and sea surface temperature (SST) from the NCEP reanalysis dataset. The initial SST data (Fig. 2) was fixed through model integration. The numerical model was integrated for five days. Other model physics consist of the Kain–Fritsch convective scheme (Kain and Fritsch 1993), an explicit microphysics scheme (Lin et al. 1983), the Yonsei University (YSU) planetary boundary layer (PBL) scheme, thermal diffusion land-surface scheme, and Monin–Obukhov surface-layer scheme (Hong et al. 2006). In the simulation, a periodical lateral boundary condition is used. Those model physics were the same as those reported by Cao et al. (2016) and Cao and Wu (2018a).

Eight idealized experiments were performed to determine the preferred region of tropical cyclogenesis over the WNP in strong monsoon trough pattern. Figure 2 shows the composited wind fields at the lower level, SST, the genesis potential index (GPI), and vertical shear of zonal wind between 200 hPa and 850 hPa during strong monsoon trough years. The GPI was developed by Emanuel and Nolan (2004) and is given as:

$$GPI = |10^5 \text{abv}|^2 \left( \frac{RH}{50} \right)^3 \left( \frac{PI}{70} \right)^3 (1 + 0.1V_{\text{shear}})^{-2}$$

where $\text{abv}$ is the absolute vorticity at 850 hPa, $RH$ is the relative humidity at 700 hPa, $PI$ is the maximum potential intensity and $V_{\text{shear}}$ is the magnitude of the vertical wind shear between 200 and 850 hPa defined by $\sqrt{(u_{200}-u_{850})^2 + (v_{200}-v_{850})^2}$. The maximum SST is located north of the equator and east of 160°E in strong monsoon trough years (Fig. 2a). The distribution of the GPI is similar to that of the
monsoon trough and is tilted along a northwest–southeast direction with the
maximum located at (15°N, 145°E) (Fig. 2a). Meantime, the initial vortices are
located in the region of relatively weak vertical shear of zonal wind (between -4 m/s
and 0, Fig. 2b). In order to examine the relative contribution of individual terms to the
total GPI changes, we take the logarithm of GPI equation above and the sum of the
four terms in the right side of equation is identical to the total GPI change (figure not
shown). The composite results in strong monsoon trough years show that the absolute
vorticity in the lower level has the largest contribution to GPI change and the
contribution from the relative humidity is secondary. The vertical wind shear and
potential intensity have a small contribution.

The positions of weak artificial vortices were chosen based on the distribution of
the GPI. In the eight idealized experiments, the weak artificial vortices were inserted
into regions centered at (17.5°N, 130°E), (15°N, 137.5°E), (15°N, 145°E), (12.5°N,
150°E), (12.5°N, 155°E), (10°N, 160°E), (10°N, 165°E) and (7.5°N, 170°E) (Fig. 2).
The eight experiments are referred to as C130E, C137.5E, C145E, C150E, C155E,
C160E, C165E and C170E, respectively.

The composite fields of the atmospheric variables and SST were specified as the
background in each numerical simulation and the only difference was the position of
the artificial vortex. Note that the background fields are fixed and we only change the
location of the initial disturbance vortex at the initial time. Thus, the background
condition change is only because of the relative position of the initial disturbance
vortex with respect to the background field. The radial profile of the tangential wind for the initial artificial vortex is given by:

\[ V(r) = \frac{V_m}{r_m} \left[ \exp \left( 1 - \frac{r}{r_m} \right) - \frac{|r - r_m|}{R_0 - r_m} \exp \left( 1 - \frac{R_0}{r_m} \right) \right], r > r_m, \]

\[ V(r) = V_m \frac{r}{r_m^2} \exp \left( 1 - \frac{r}{r_m} \right), r \leq r_m, \]

where \( r \) is the radius measured from the vortex center, \( r_m \) is the radius of the maximum tangential wind (150 km), \( V_m \) is the maximum tangential wind (8 m s\(^{-1}\) at the radius of 150 km) and \( R_0 \) is 1000 km, where the vortex winds disappear. The wind speed gradually decreased upward and vanished at 100 hPa. The mass and thermodynamic fields associated with the artificial vortex were obtained based on a nonlinear balance equation in which the hydrostatic and gradient wind balances were satisfied for the initial vortex (Wang 1995).

3.2. Simulated evolution of tropical cyclogenesis

Figure 3 shows the time evolution of the intensity of the vortex represented by the minimum sea-level pressure (MSLP) and the maximum azimuthal mean wind (MAMW) of the vortex at 10-m height in the eight experiments from \( t = 48 \) h to \( t = 120 \) h, including the early stage of tropical cyclogenesis. Note that the TC center is defined by the MSLP to calculate the azimuthal mean wind. In the first 48-hour integration, the weak vortices in the eight experiments experience an initial gestation and adjustment period with a rather slow development (figure not shown). If the development speed of vortices is defined as a gradient of the MAMW between \( t = 48 \)
h when the vortices have a similar initial intensity and \( t = 72 \text{ h} \), it is seen that the vortices experienced more rapid development in experiments C145E, C150E, C155E, C160E, particularly in C150E and C155E, whereas the vortices developed relatively slowly in the C165E and C170E runs after \( t = 48 \text{ h} \) (Fig. 3b). The weak vortices did not develop into tropical cyclones in the C130E and C137.5E runs. The time of tropical cyclogensis was defined as when the MAMW reached 12.9 m s\(^{-1}\) (dashed line in Fig. 3b). Based on this definition, tropical cyclogenesis occurred at \( t = 81 \text{ h} \) in C145E, \( t = 75 \text{ h} \) in C150E and C155E, \( t = 84 \text{ h} \) in C160E, \( t = 96 \text{ h} \) in C165E and \( t = 108 \text{ h} \) in C170E. This simulation result indicates that weak vortices in the central WNP were more likely to develop into tropical cyclones, whereas those in the western and eastern WNP developed more slowly or unsuccessfully. Cao and Wu (2018a) showed that a tropical disturbance is easily triggered over the WNP near 150°E–160°E when the monsoon trough expands eastward, assuming that there is no existing tropical disturbance. Thus, the region 150°E–160°E is the most favorable for tropical cyclogenesis when the monsoon trough expands eastward, irrespective of whether there is a tropical disturbance over the WNP.

We then investigated why the weak vortices in runs C130E and C137.5E did not develop into tropical cyclones using C130E as an example. Figure 4 shows the simulated 850 hPa wind fields in run C130E from \( t = 48 \text{ h} \) to \( t = 66 \text{ h} \) at intervals of 6 h. At \( t = 48 \text{ h} \), the vortex had moved northwestward to (20°N, 125°E) from its initial position of (17.5°N, 130°E) as a result of the steering easterly winds. At this time, the maximum wind speed was close to 12 m s\(^{-1}\) (Fig. 4a). In the following 12 hours, the
vortex gradually moved closer to Taiwan and the Philippines. The winds in the southwest quadrant of the vortex gradually weakened as the vortex moved close to the northern Philippines. The closed vortex disappeared at $t = 66$ h and only the southerly winds remained to the east of Taiwan (Fig. 4d). The evolution of the vortex in C137.5E showed similar features to C130E, but with a slower decaying time (figure not shown). Figure 5 further shows the evolution of mid-level relative humidity and surface heat flux (including sensible and latent heat fluxes) in C130E. When the vortex moves northwestward, the mid-level humidity and surface heat flux both become smaller, particularly at $t = 66$ h, which is unfavorable for the development of convection. In addition, with the vortex moving northwestward, the SST gradually decreases as shown in Fig. 2a. This is consistent with the decrease in surface heat flux.

In addition, we performed another sensitivity experiment similar to C130E run, except that the topography height is equal to zero in the region of $5^\circ$N–$25.5^\circ$N and $119.5^\circ$E–$127^\circ$E including Taiwan Island and partly Philippines Island. The simulation result showed that the vortex still cannot develop into a strong TC (figure not shown). This indicates that the failure of tropical cyclogenesis in C130E is not much closely associated with the frictional effects of the land surface. Thus, it is inferred that the failure of tropical cyclogenesis in runs C130E and C137.5E is associated with the decrease of the moisture and sensible and latent heat from the underlying ocean. Note that other possible reasons for the failure of tropical cyclogenesis in the western WNP may exist, for instance, the influences from inefficiency of the energy conversion.
from the background fields, which will be further investigated in the future work.

3.3. Sensitivity to the position of the vortex

Five additional experiments were performed to investigate the sensitivity to the initial positions of the weak vortices to enhance the significance of the model results. The artificial vortices were inserted into regions centered at (15°N, 130°E), (10°N, 140°E), (10°N, 150°E), (8°N, 160°E) and (6°N, 170°E) (Fig. 6a), referred to as runs C130E15N, C140E10N, C150E10N, C160E8N and C170E6N, respectively. The choice of position of the vortices was based on the distribution of the relative vorticity. The positions of the vortices in these five experiments were further south than those in the preceding eight experiments because the regions with large values of relative vorticity were located further south than those of the GPI (Figs. 2a and 6a).

Figure 6b shows the MAMW of the vortex in the additional five experiments from \( t = 48 \) h to \( t = 120 \) h. The vortices developed more rapidly in experiments C140E10N, C150E10N and C160E8N, particularly in C150E10N and C160E8N, whereas the vortex developed at a slower speed in C170E6N. The vortex in run C130E15N did not develop into a tropical cyclone (Fig. 6b). Tropical cyclogenesis occurred at \( t = 81 \) h in both C150E10N and C160E8N and at \( t = 90 \) h in C140E10N. The simulation results reconfirm that the weak vortices in the central WNP are more likely to develop into tropical cyclones, whereas those in the western and eastern WNP develop more slowly.
It is noteworthy that the vortex in regions with a larger GPI value did not necessarily develop faster than those in regions with a smaller GPI. For example, although the GPI around the vortex in runs C165E and C170E was larger than that in runs C150E10N and C160E8N, the time of tropical cyclogenesis in runs C165E and C170E (96 h and 108 h) was much later than that in runs C150E10N and C160E8N (81 h and 81 h). Therefore, the empirical GPI index may be a good tool in a climatological viewpoint, but it does not fully represent the probability of tropical cyclogenesis over the WNP in a short period. It is necessary to design an objectively empirical index to estimate quantitatively the influence of large-scale environmental components on tropical cyclogenesis over the WNP on the synoptic time scale, which has been discussed in DeMaria et al. (2001) and Schumacher et al. (2009).

4. Mechanisms behind runs C150E and C170E

Apart from runs C130E and C137.5E in which no tropical cyclone appeared, there were six experiments in which the weak vortices successfully developed into tropical cyclones. These six experiments were divided into two groups. One group included runs C145E, C150E, C155E and C160E in which the tropical cyclones had a relatively rapid speed of genesis. The other group included runs C165E and C170E in which the tropical cyclones had a slower speed of genesis. The following analysis uses only the results from the runs with the fastest (C150E) and slowest (C170E) speeds of tropical cyclogenesis to illustrate the maximum contrast.
Figures 7 and 8 show the simulated wind fields at 850 hPa from $t = 48$ h to $t = 66$ h in runs C150E and C170E, respectively. As shown in Fig. 3b, the intensity of the vortex in C150E and C170E is almost the same at $t = 48$ h. At this time, the vortex moves from the initial position of (15°N, 150°E) to (16°N, 143°E) in run C150E (Fig. 7a). The vortex gradually intensifies during the following 12 hours and maintains a closed cyclonic structure as it moves westward (Figs. 7b, c). At $t = 66$ h, the vortex develops into a strong tropical depression with a nearly symmetrical wind structure located at (15°N, 140°E), with an MSLP of 1001 hPa and a MAMW of 11.8 m s$^{-1}$ at 10-m height (Fig. 7d). In the process of tropical cyclogenesis, the large-scale control is mainly characterized of the monsoon shear between the easterlies in the north and the westerlies in the south in C150E. From Fig. 2, the initial large-scale environment around the vortex in C150E is better than that in C170E (denoted by GPI).

By contrast, the vortex in run C170E shows a slower genesis. From $t = 48$ h to $t = 66$ h, the maximum wind of the vortex only increased from 7 to 8.4 m s$^{-1}$ (Fig. 8a, d). The maximum wind speed was located in the northern or northeastern part of the vortex due to the superposition of easterly winds. In these periods, the structure of the wind in run C150E changed from asymmetrical to symmetrical (Fig. 7), whereas the structure of the wind in run C170E remained asymmetrical (Fig. 8). Thus, the vortex in run C170E only developed into a weak tropical depression with an MSLP of 1004 hPa and an MAMW of 8.4 m s$^{-1}$ at $t = 66$ h (Fig. 8d). Another remarkable difference between Fig. 7 and Fig. 8 is the intensity of northerly wind component in the trade easterly region. In C170E, the large-scale control is also associated with the monsoon
shear pattern generated by the easterlies in the north, because the easterlies to the north are farther apart from westerlies to the south (Fig. 8).

In addition to the large-scale lower-level wind conditions, we further studied the vertical profile of environmental conditions in runs C150E and C170E to elaborate the particular way in which the vortex in the eastern WNP leads to the slower tropical cyclogenesis. Figure 9 shows the vertical profiles of the area-averaged zonal wind from \( t = 48 \) h to \( t = 66 \) h in runs C150E and C170E. At \( t = 48 \) h, the differences in the zonal wind at upper and lower levels were fairly small in both C150E and C170E (Figs. 9a, e), with a magnitude of 1 m s\(^{-1}\). The differences in the zonal wind in the upper and lower levels changed little in run C150E over the following 18 hours (Figs. 9b–d). The area-averaged winds were easterly winds in both the upper and lower levels, resulting in a small vertical wind shear in run C150E. However, the difference in the zonal wind in the upper and lower levels in run C170E increased over time (Figs. 9f–h). The area-averaged winds were always easterly at lower levels, whereas at upper levels they changed from easterly to westerly, leading to an increase in the vertical wind shear in run C170E.

In general, a weak vertical wind shear favored tropical cyclogenesis. A large vertical wind shear can lead to a slower intensification of the vortex into a strong tropical cyclone by “venting” the accumulated condensational heat away from the tropical disturbance core region, holding back the decrease in surface pressure, the establishment of a warm core in the upper level and the intensification of cumulus convection (Gray 1979; DeMaria 1996; Frank and Ritchie 2001; Chia and Ropelewski
2002; Ge et al. 2013), which will be detailedly discussed in the following analysis. In addition, the vertical wind shear is also calculated more accurately based on the method from previous studies (Hanley et al. 2001). First, the zonal wind in different levels is interpolated into an azimuth polar grid centered on the vortex center. Then, these winds are averaged over a radius of 450 km from vortex center to remove symmetric vortex so that the resulting winds stand for the environmental flows. The results are almost similar to that in Fig. 9 (figure not shown).

Figure 10 shows the vertical height–longitude cross-section of the temperature anomalies and meridional winds in runs C150E and C170E from \( t = 48 \) h to \( t = 66 \) h. As a result of the fairly small vertical wind shear, the condensational heating induced by deep convection led to a upper level warm core in run C150E (Figs. 10a–d). The winds showed a slightly westward tilted vertical structure at \( t = 48 \) h (Fig. 10a). At \( t = 66 \) h, the wind showed a prominent vertical extension without tilting (Fig. 10d), which favored the accumulation of condensational heating and a decrease in surface pressure (Fig. 3a). In run C170E, the wind tilted more clearly with altitude from \( t = 48 \) h to \( t = 66 \) h as the vertical wind shear increased (Figs. 10e–h). Under the influence of westerly shear, the lower level vortex shifted westward and the upper level vortex shifted eastward, delaying the establishment of the deep moist layer in the core region (Ge et al. 2013). At the same time, the warm core in the upper level became asymmetrical and tilted eastward with height in run C170E (Fig. 10h). This structure of the warm core and the winds did not favor a rapid decrease in surface pressure and intensification of the vortex. Although run C170E is consistent with previous studies...
in that vertical wind shear has a detrimental effect on tropical cyclogenesis (Gray 1979; Frank and Ritchie 2001), the vortex still developed into a tropical cyclone eventually. This may occur because a forced secondary circulation could overcome the tilting induced by shear and restore the vertical alignment, allowing tropical cyclogenesis (Zhang and Kieu 2006; Ge et al. 2013).

Why does the vertical wind shear increase in run C170E? Figure 11 shows the wind fields at 300 hPa in runs C150E and C170E during the same periods. In run C150E, the anticyclonic circulation wraps around the cyclonic circulation of the upper level vortex (Figs. 11a–d). The distinct southerly winds are located to the left of the vortex and the zonal wind is fairly small. The maximum wind is observed in the north or northeast quadrant of the vortex, which leads to area-averaged easterly winds in the upper level in run C150E (Figs. 9a–d). By contrast, in run C170E the westerly winds penetrate into the cyclonic vortex though anticyclonic circulation wraps also around the cyclonic circulation of the upper level vortex (Figs. 11e–h), resulting in the area-averaged westerly wind in the upper level shown in Figs. 9e–h.

In order to better examine the effect of background upper-level wind fields on tropical cyclone genesis, a spatial filtering technique is applied to separate the meso-scale vortex from the environmental circulation (Fig. 12). Wavelength larger than 600 km is considered as the background large-scale circulation and the wavelength smaller than 600 km represents the tropical cyclone-scale vortex and smaller-scale disturbance. The obtained background upper-level wind fields in runs C150E and C170E in Fig. 12 are almost similar to Fig. 11. Thus, the difference in
vertical wind shear in runs C150E and C170E is closely associated with the vertical structure of the vortex. The deep and strong vortex in C150E can lead to a larger inertial stability and thus resist the influence from the large-scale wind fields. The deep vortex in C150E in the early stage of tropical cyclogenesis may be mainly ascribed to the better initial large-scale environment including strong relative vorticity and great humidity around the vortex as shown in Fig. 2, which is favorable for the vortex to intensify in the core (Gray 1968; Ritchie and Holland 1999; Fang and Zhang 2010). In C170E, the weak vortex is easily affected by the large-scale wind fields outside of the vortex.

We next examined the evolution with time of the physical variables around the core region in runs C150E and C170E. Figure 13 shows the vertical height–time cross-sections of the relative vorticity, divergence, relative humidity and diabatic heating averaged in the center of the vortex from $t = 48$ h to $t = 120$ h in run C150E. The green dashed lines indicate the times of tropical cyclogenesis. The vortex intensifies from $t = 48$ h to $t = 75$ h, the early stage of tropical cyclogenesis (Fig. 13), with a relatively rapid development of the relative vorticity, which increases by $4 \times 10^{-5}$ s$^{-1}$ (Fig. 13a). The maximum vorticity always appears in the lower level of the vortex during the early genesis stage, accompanied by upper level divergence and lower level convergence (Figs.13a, b). A distinct feature is that relative humidity and diabatic heating have a close relationship prior to tropical cyclogenesis. The maximum relative humidity (>80% at 600 hPa) occurs before tropical cyclogenesis at $t = 54$ h (Fig. 13c). The maximum diabatic heating appears at $t = 57$ h at 350 hPa. We
therefore conclude that moist air in the middle level of the vortex is a precondition for the outbreak of deep convection. Previous studies have also suggested that a high mid-level humidity favors the growth of tropical disturbances (Nolan 2007; Ge et al. 2013; Cao et al. 2016).

In contrast, the vortex was intensified slowly in run C170E. The maximum vorticity first appeared at low levels from \( t = 48 \) h to \( t = 78 \) h (Fig. 14a). A mid-level maximum vorticity followed the low level spin-up from \( t = 78 \) h to \( t = 96 \) h (Fig. 14a). After \( t = 96 \) h, the maximum vorticity appeared again in the lower level. During the 30 hours from \( t = 48 \) h to \( t = 78 \) h, the vortex developed slowly, with the relative vorticity increasing by about \( 2 \times 10^{-5} \) s\(^{-1} \) (Fig. 14a). In sharp contrast, the relative vorticity of the vortex intensified more rapidly during the next 30 hours from \( t = 78 \) h to \( t = 108 \) h, with an increase of about \( 5 \times 10^{-5} \) s\(^{-1} \) (Fig. 14a). After \( t = 78 \) h in run C170E, a weak convergence appeared at 300 hPa (Fig. 14b), accompanied by a low relative humidity in the upper level (Fig. 14c). This upper level convergence was unfavorable for the development of vertical motion. Maximum diabatic heating occurred at \( t = 78 \) h at 450 hPa, followed by a decrease in diabatic heating corresponding to the low humidity in the upper level (Figs. 14c, d).

To conclude, the slow development of tropical cyclogenesis in run C170E was therefore mainly associated with a large vertical wind shear and the penetration of dry air into the upper levels. The large vertical wind shear reflects the effect of large-scale environmental conditions on the development of vortex. Those two processes do not favor the accumulation of diabatic heating and enhancement of the upper level warm
core, which impedes the rapid decrease in surface pressure and intensification of the lower level vortex.

5. Summary and discussion

In this study, we documented the preferred region for tropical cyclogenesis in strong monsoon trough pattern and investigated the physical mechanisms behind this preference and whether tropical cyclogenesis can be simulated in a numerical model. Our composite analysis from observational data of the relationship between the monsoon trough and tropical cyclogenesis shows that more tropical cyclones are likely to appear in the central WNP near 130°E–165°E during the strong monsoon trough years and fewer tropical cyclones appear over the western (near 120°E–130°E) and eastern (near 165°E–180°E) WNP.

The numerical simulation showed that weak vortices developed more rapidly in experiments C145E, C150E, C155E and C160E, and that the vortices in runs C165E and C170E developed more slowly. The weak vortex in runs C130E and C137.5E did not develop into a tropical cyclone. These results suggest that the preferred region for the tropical cyclogenesis is located in the central part of the strong monsoon trough pattern, which is consistent with the composite analysis of the observational data. The failure of tropical cyclogenesis in the western WNP is mainly associated with the decrease of the moisture and heat (including sensible and latent heat) from the underlying ocean. The large vertical wind shear and the penetration of dry air into the
upper level of the vortex are the main factors to the slower genesis of vortices in the 
eastern WNP. The sensitivity of the modeling results to the positions of vortices was 
also investigated and it was confirmed that the preferred region for tropical 
cyclogenesis was the central part of the strong monsoon trough pattern over the WNP. 

Previous studies have suggested that tropical cyclogenesis occurs more 
frequently in the southeastern WNP when the monsoon trough is strong and extends 
eastward during El Niño summers, whereas more tropical cyclones appear in the 
northwestern part of the WNP when the monsoon trough is weak and retreats 
westward during La Niña summers (Wang and Chan 2002; Li 2012; Wu et al. 2012). 
Our study suggests that the preferred region for tropical cyclogenesis is in the center 
of the strong monsoon trough. In El Niño years, the preferred for the tropical 
cyclogenesis moves eastward as the monsoon trough is displaced eastward. Our 
results therefore support the eastward shift of the region of tropical cyclogenesis 
during El Niño years (Cao et al. 2018). However, this does not mean that the largest 
increase in the frequency of tropical cyclones occurs in the southeastern WNP, though 
tropical cyclogenesis is increased in this region. 

We have mainly focused on the processes of tropical cyclogenesis in strong 
monsoon trough pattern. The sensitivities of our simulation results to the model 
physics are yet unknown. Therefore, a series of sensitivity experiments will be carried 
out to investigate the robustness of the numerical simulation results. We will also 
examine multi-scale interactions in the tropical cyclogenesis process when a weak 
vortex is inserted into the WNP through a higher resolution numerical model.
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Table and Figure captions:

Table 1. The list of sub-regions and corresponding tropical cyclogenesis number.

Figure 1. The 850-hPa wind fields (m s\(^{-1}\)) and tropical cyclogenesis during (a) strong and (b) weak monsoon trough years from the reanalysis data. The dashed lines divide the WNP into six sub-regions with equal ranges. The red numbers denote the frequency of tropical cyclogenesis in the corresponding sub-regions. The black dashed line defined by zonal wind equal to zero in (a) indicates the location of monsoon trough.

Figure 2. Composite (a) 850 hPa wind fields (vector, m s\(^{-1}\)), sea surface temperatures (contours, °C), and genesis potential index (shaded) from the reanalysis data as the initial background fields and low boundary condition and (b) vertical shear of zonal wind between 200 hPa and 850 hPa during July–November in strong monsoon trough years (1982, 1986, 1990, 1991, 1997, 2002, and 2004). The black symbols indicate the different positions of the weak artificial vortices at the initial time.

Figure 3. Evolution over time of (a) the minimum sea-level pressure (hPa) and (b) the maximum azimuthal mean wind speed (m s\(^{-1}\)) at 10 m in the eight experiments. The abscissa represents time (hours) and the ordinate corresponds to the intensity. The dashed line in part (b) denotes the time of tropical cyclogenesis when the maximum azimuthal mean wind speed exceeded 12.9 m s\(^{-1}\).

Figure 4. Wind fields (vectors, m s\(^{-1}\)) and total wind speed (shading; m s\(^{-1}\)) at 850 hPa in run C130E from \(t = 48\) h to \(t = 66\) h at intervals of 6 h.
Figure 5. Wind fields (vectors, m s$^{-1}$), relative humidity (shading, %) at 700 hPa and surface heat flux (contour, W m$^{-2}$) in run C130E from $t = 48$ h to $t = 66$ h at intervals of 6 h.

Figure 6. (a) Composite 850 hPa wind fields (vectors, m s$^{-1}$), GPI (shading) and relative vorticity (contours; 10$^{-5}$ s$^{-1}$) from the reanalysis dataset from July to November in strong monsoon trough years. The black symbols indicate the different positions of the weak artificial vortices at the initial time. (b) Evolution with time of the maximum azimuthal mean wind speed (m s$^{-1}$) at 10 m in the additional five experiments. The abscissa represents time (hours) and the ordinate corresponds to the intensity. The dashed line in part (b) denotes the time of tropical cyclogenesis when the maximum azimuthal mean wind speed exceeds 12.9 m s$^{-1}$.

Figure 7. Wind fields (vectors, m s$^{-1}$) and total wind speed (shading; m s$^{-1}$) at 850 hPa in run C150E from $t = 48$ h to $t = 66$ h at intervals of 6 h.

Figure 8. Same as Fig. 7 but in run C170E.

Figure 9. Vertical profiles of area-averaged (600 km × 600 km) zonal wind from $t = 48$ h to $t = 66$ h at intervals of 6 h in runs (a–d) C150E and (e–h) C170E.

Figure 10. Vertical height–longitude cross-sections of meridional wind (contours; m s$^{-1}$) and temperature anomalies (shading; °C) in runs (a–d) C150E and (e–h) C170E from $t = 48$ h to $t = 66$ h at intervals of 6 h.

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× 900 km around the vortex center.

Figure 12. Same as Fig. 11 but with wavelength larger than 600 km.

Figure 13. Vertical height–time evolution of (a) area-averaged (420 km × 420 km) relative vorticity (10⁻⁵ s⁻¹), (b) divergence (10⁻⁵ s⁻¹), (c) relative humidity (%) and (d) diabatic heating (10⁻⁴ K s⁻¹) in run C150E from t = 48 h to t = 120 h. The green dashed lines denote the time of tropical cyclogenesis.

Figure 14. Same as Fig. 13 but in run C170E.
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