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TROPICAL CYCLONE FORMATIONS IN THE SOUTH CHINA SEA

CHENG-SHANG LEE¹ AND YUNG-LAN LIN^{*1, 2}

¹Department of Atmospheric Sciences, National Taiwan University, Taipei, Taiwan

²Taipei Aeronautic Meteorological Center, Civil Aeronautics Administrator, Taipei, Taiwan

CHIA-MEI HUANG³

³Weather Forecast Center, Central Weather Bureau, Taipei, Taiwan

1. INTRODUCTION

In the past 34 years (1972-2005), there were about one thousand tropical cyclones (TCs) formed in the western North Pacific (WNP). During the same period, 131 TCs formed in the South China Sea (SCS) with an annual average of 3.9. Almost no TC formation occurs in the SCS from January to March, but the number of TC formation cases increases significantly in May and June (mei-yu season) and accounts for 18.3% of the total number of formations in the SCS. Such results indicate that the two months of the mei-yu season are particularly favorable for TC formations in the SCS (Lee et al. 2006).

The mei-yu season in Southeast Asia is from about mid-May to mid-June when a stationary front in the northern SCS often occurs. The mei-yu frontal system is part of the East Asia monsoon system. Chen and Chang (1980) showed that the mei-yu front and midlatitude fronts have similar structures in that they possess a large horizontal temperature gradient and vertical baroclinic structures. The major difference is that the mei-yu front is only weakly baroclinic and there can be strong low-tropospheric vertical wind shear, which is especially true for the frontal systems west of Japan (Ninomiya and Akiyama 1992). Lander (1996) examined TC formations during 1978-1994 associated with the so-called reverse-oriented monsoon trough, which is similar to the frontal orientation in the East Asia mei-yu season. Such mei-yu formations may be related to the subtropical cyclone (STC), which Hebert and Poteat (1975) defined as a TC-like circulation that forms in the baroclinic subtropical environment associated with a front.

Different statistics have been given on TC formation associated with baroclinic systems. Frank and Clark (1977) showed that less than 5% of all TC formations in the Atlantic occurred in close vicinity of a front or east side of a westerly trough. However, Gray (1998) indicated that about 25% of the global TC formations were not in a pure tropical environment. Davis and Bosart (2002) explored the model sensitivities in simulating Diana (in 1984). When the upper-level trough was removed from the initial condition, cyclogenesis did not occur, which indicates the importance of appropriate coupling between the upper and lower levels in transforming the baroclinic system into a tropical depression. In a further study, Davis and Bosart (2003) showed that a rapid decrease in the 200-900-hPa vertical wind shear occurred before the STC became a TC. They also simulated the formation of Hurricane Michael

(in 2000) and found that the vertical wind shear had to decrease rapidly from 30 m s⁻¹ to realize the warm-core structure. Therefore, the reduction of the vertical wind shear is an essential requirement for a baroclinic system to transform to a TC. To study how a baroclinic system which is associated with a mei-yu front can develop to a TC, we composited the synoptic flows of TC formation and non-formation cases (Lee et al., 2006) . However, such analyses only demonstrate the average synoptic pattern and the overall evolution of the environment. Many important issues have yet to be explored, especially the evolution of the meso-scale structure. Therefore, this study, referring to the definition of Lee et al. (2006), attempts to simulate the formation of Typhoon Noguri (2002) and Russ (1994) using MM5 in order to discuss the key mechanism leading to TC formation in the weak baroclinic environment.

Besides the mei-yu season, the percentage of SCS TC formations in November and December is also higher (16%) when compared to that in the WNP. Chang et al. (1979) showed that the northeasterly cold surge off the Asia continent lead to intensify the convective disturbances in the near equatorial region. Some of these disturbances which are either originated from the semi-stationary near-equatorial trough over the coast of north Borneo or from a westward propagating wave in the WNP might intensify to TC intensity (25 kt). The unique topography of the southern SCS including Malay Peninsula and Borneo acts to direct the flow toward the equator. Cold surge airs are dry, but are moistened from the bottom when moving across the open ocean of SCS. Therefore, the lower troposphere is convectively unstable in the southern SCS. The gradient of planetary vorticity together with the blocking and deflection effects due to topography may contribute to TC formation, such as the equatorial typhoon Vamei in 2001 (Chang et al. 2003). The interactions among the synoptic-scale Borneo vortex, northeasterly cold surge and the intraseasonal Madden-Julian oscillation during the boreal winter also contribute to the variability of deep convection in the region (Chang et al. 2005). In this study, the preliminary analyses of TC formations in the SCS during the late season are also presented.

2. SCS TC FORMATION IN THE MEI-YU SEASON

Lee et al. (2006) examined 119 TC formations in the SCS during 1972–2002, and in particular the 20 in May and June. Eleven of these storms are associated with the weak baroclinic environment of a mei-yu front, while the remaining nine are nonfrontal. Comparison of the frontal

^{*} Corresponding author: Yung-Lan Lin, Dept. of Atmos. Sci., NTU, 1, Sec. 4, Roosevelt Rd., 106 Taipei, Taiwan. (ryanlin@nat.as.ntu.edu.tw)

and nonfrontal formation shows that a nonfrontal formation usually occurs at a lower latitude, is more barotropic, develops faster, and possibly intensifies to a stronger TC. Seven of the 11 frontal disturbances are originated over land and have a similar evolution pattem. For comparison purpose, six nonformation cases in the SCS are also identified. These cases which are located near the western end of a mei-yu front have surface closed isobar lasted for over 48 hours but do not develop further. For the nonformation cases, both the northeasterlies north of the front and the monsoonal southwesterlies are weaker in magnitude compared to the formation cases (fig. 1). The nonformation cases also have an average of $2-3 \text{ m s}^{-1}$ larger vertical wind shear than the formation cases (Fig. 2a).

A conceptual model is proposed for the typical frontal-type TC formations in the SCS that consists of three essential steps (Fig.3). First, an incipient low-level disturbance that originates over land moves eastward along the stationary mei-yu front. Second, the low-level circulation center with a relative vorticity maximum moves to the open ocean with the stationary front. Finally, with strengthened northeasterlies cyclonic shear vorticity continues to increase in the SCS, and after detaching from the stationary front, the system becomes a tropical depression.

To examine environmental parameters and to diagnose the large-scale heat and moisture budgets during formation process, typhoon Noguri (2002) and tropical storm Russ (1994) are analyzed using ECMWF advanced dataset. Results show that the incipient disturbance of both TCs are originated from the Hainan Island and are at the southwest end of the mei-yu front. To the north-northeast of the incipient disturbance, there is a high pressure system which moves eastward at later time periods when the incipient disturbance moves eastward and developes. At 850 hPa, the large potential temperature gradient associated with the mei-yu front starts to decrease especially at the western end of the stationary front when the mei-yu front moves eastward slowly. The satellite imageries show that there are active convections at the western end of the stationary front especially to the south of the front (or the northern SCS). The interactions between the mei-yu front and convections in the northern SCS occur during the formation process resulting in the merging of convections and the strengthening of the incipient disturbance. It has to be noted that such process occurs during a 48-hours period before TCFA.

Simulations using MM5 for Noguri and Russ show that there are strong vorticity patches along the mei-yu front during the formation process. The merging of the vorticity patches leads to a stronger vorticity center which later develops to a TC. As an example, Figure 4 shows the model results of Noguri. At 12-hour of model simulation (Fig 4a), there are two cyclonic circulations at 925 hPa. The western one is better-organized but with 3 vorticity maxima. During the following 12-hour period (Fig 4b), these 3 vorticity patches rotate cyclonically and merge together. The two vortices then further rotate cyclonically with respect to each other (Fig 4c). Such evolution pattern is fairly similar to that revealed by the EC data and the satellite images figure not shows except the difference in time. It is also interesting to note that the two strong vorticity patches are related to the strengthening of the northeasternlies and the southwesternlies, individually. The vertical cross sections along the vorticity maxima reveal that the strong vorticity concentrates at low level at the early formation stage. After the mergence of vorticity patches, the low-level vorticity of the disturbance increases quickly with the strong vorticity extends vertically to 300 hPa. The positive temperature anomaly which exists initially at upper level also extend to low levels.

3. SCS TC FORMATION IN THE LATE SEASON

In 1972-2005, there are 22 TC formations over the SCS during the late season (November to January). Eleven of these storms are originated from the southern SCS (called the semi-stationary cases) and the others are associated with westward-propagating waves from the WNP that pass the Philippines (called the westward cases). Base on the best track of JTWC, the tracks of the semi-stationary cases are more diversive when compared to those of the westward cases which are steered by the easterlie south of the subtropical high. Analyzes also show that the semi-stationary cases are often located at the western edge of the subtropical high where the steering flow is not well-defined.

The 925-hPa composites of the westward cases show that at 36 hours before formation (V_{max} = 25 kt), the northeasterlies are strong over a broad area with large positive relative vorticity near the equator (Fig 5b). For the semi-stationary formations at the same period (Fig. 5a), the northeasterlies are also strong over a large area which is more NE-WS oriented and is limited in the SCS. A strong wind axis is located at the western SCS just north of disturbance. At 36 hours after formation (Fig. 5d), the wind speed of northeasterlies associated with the westward cases increases to 15 m s⁻¹ which is larger than that for the semi-stationary cases (Fig. 5c). In consequence, the relative vorticity of the westward cases is larger than that of the semi-stationary cases. It is thus speculated that the strong northeasterlies over the SCS play the key role in the formation of the westward cases.

To identify the key feature leading to the formation of the semi-stationary cases, 18 nonformation cases which have closed surface isobar lasted for more than 48 h but do not further develop are studied. The composite of 925-hPa flow for nonformation cases show that the center of the disturbance is located near the coast of north Borneo. At 36 hours after the first closed isobar is identified, the center of half of the cases are located at Borneo. Thus no further development occurs for these cases although the relative vorticity also reaches a maximum magnitude at this time (Fig. 6b). At the same period, the northeasterlies of the formation cases are stronger in magnitude and located at higher latitudes when compared to the nonformation cases (Fig. 6a-b). It is quite surprisingly to find that the formation cases have stronger vertical wind shear compared to that of the non-formation cases (Fig 2b). However, for the formation case, the vertical wind shear decreases slightly thought the period. For the nonformation cases, although the vertical wind shear is smaller at earlier time periods, it increases continuingly and becomes larger than the formation cases after 66 h of the zero-reference time (Fig. 2b). Because the climatological flow of winter monsoon is that the strong 200-hPa southwestlies are over the low-level northeasterlies in the SCS, the late season formation cases have 2-4 m s⁻¹ larger values of vertical wind shear than that for the typical frontal-type formation cases.

To compare the convection distribution during formation process, the outgoing longwave radiation comparison (OLR) for the formation and nonformation cases are showed in Fig. 8. Three days before the zero-reference time, there is a stationary mode with lower OLR located at the southern SCS extending longitudinally; during the formation period, the convention in Pacific Ocean east of Philippine decreases day after day. In contrast for the nonformation case, the stationary convention at the southern SCS is weaker and spreaded over a smaller domain (Fig 8b).

4. DISCUSSION AND CONCLUSIONS

This study analyzes the circulation pattern and the important process associated with the formation of a TC embedded in a weak baroclinic environment located at the northern SCS during the mei-yu season. Results of composite analyzes reveal that the typical frontal-type TC formations in the SCS that consists of three essential steps. Results from MM5 simulations of Noguri (2002) and Russ (1994) indicate that the strengthening of the northeasterly flow which is associated with an eastward-moving anticyclone helps concentrate the vorticity and creates strong vorticity patches to the northeast of the low center. The merging of strong vorticity patches in the lower troposphere plays an important role during TC formations and the vorticity field become more axis-symmetric near the center at the later stage of formation. Superposition of strong upper level divergence over the area where the strong low-level convergence and vorticity maximum enhances low-level vorticity and help the merging of vorticity patches.

Results show that the percentage of all incipient lows (which have surface closed isobar for at least 48 hrs) during the late season that develop to TC intensity is lower when compared to that associated with the mei-yu front. TC formations in late season generally occur at the southern SCS where strong cyclonic shear vorticity associated with the strong low-level northeasterlies exist. At 200 hPa, the southeasterly prevails over the SCS with large divergence over the southern SCS. Such circulation patterns provide favorable environment for TC formation to occur in the southern SCS. On the other hand, strong vertical wind shear presents over the central SCS where the low-level northeasterlies prevail which is not favorable for further development of any disturbances that might exist. The composite analyses show that the major difference between the formation and the non-formation cases is the northeasterlies to the north of the incipient disturbance. In addition, the non-formation cases are located closer to the Borneo coast when compared to the formation cases thus have higher chance to move inland and experience no further development. Chang et al. (2005) argued that although the northeasterly surge acts to increase the strength of the vortex located at the southern SCS off the Borneo coast, it also causes the vortex to move toward the Borneo landmass. Therefore the weakening of northeasterlies at 36 h after the zero-reference time for the formation cases may be critical for TC formation to occur because it prevents the vortex from moving to the area too close to the Borneo landmass (Fig. 7).

5. ACKNOWLEDGEMENT

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FIG. 1. Composite 925-hPa flow at 36 h for (a) formation, (b) nonformation cases during the mei-yu season. The heavy contours indicate wind speed greater than 6 m s⁻¹ (interval: 2 m s⁻¹) and the shading shows positive relative vorticity (interval: $0.5 \times 10^{-5} \text{ s}^{-1}$).



FIG.3. (Same as in Fig.1) except for frontal-type formations (upper) corresponds to three steps (lower) in the conceptual model of TC formation associated with a mei-yu front.



FIG. 2. 200-850-hPa vertical wind shear (unit: $m s^{-1}$) for the formation (solid) and nonformation (dashed) cases during the (a) mei-yu season and (b) the late season.



FIG. 4. 925-hPa relative vorticity of simulation association with Typhoon Noguri (in 2002) at integration of (a) 12, (b) 24, and (c) 36 hr. Shading shows the positive relative vorticity (interval: $5 \times 10^{-5} \text{ s}^{-1}$).



FIG. 5. (Same as in Fig.1) except at (a), (b) -36 and (c), (d) 36 h for the (a), (c) semi-stationary and (b), (d) westward cases during the late season.



FIG. 6. (Same as in Fig. 1) except at 36 h for (a) formationand (b) nonformation



cases during the late season.

FIG. 7. The difference in 925-hPa northeasterlies at 7.5 ° north of disturbance center between formation and nonformation cases during the late season (unit: m s⁻¹).

FIG. 8. Composite OLR distribution from -3 to 2 day of the zero-reference time for the (a) formation and (b) nonformation cases.