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ABSTRACT
The effects of multiple mesoscale convective systems (MCSs) on the formation of Typhoon Ketsana (2003) are analyzed in this study. Numerical simulations using the Weather Research and Forecasting (WRF) model with assimilation of Quick Scatterometer (QuikSCAT) and Special Sensor Microwave Imager (SSM/I) oceanic winds and total precipitable water are performed. The WRF model simulates well the large-scale features, the convective episodes associated with the MCSs and their periods of development, and the formation time and location of Ketsana. With the successive occurrence of MCSs, midlevel average relative vorticity is strengthened through generation of mesoscale convective vortices (MCVs) mainly via the vertical stretching mechanism. Scale separation shows that the activity of the vortical hot tower (VHT)-type mesoγ-scale vortices correlated well with the development of the MCSs. These VHTs have large values of positive relative vorticity induced by intense low-level convergence, and thus play an important role in the low-level vortex enhancement with aggregation of VHTs as one of the possible mechanisms.

Four sensitivity experiments are performed to analyze the possible different roles of the MCSs during the formation of Ketsana by modifying the vertical relative humidity profile in each MCS and consequently the strength of convection within. The results show that the development of an MCS depends substantially on that of the prior ones through remoistening of the midtroposphere, and thus leading to different scenarios of system intensification during the tropical cyclone (TC) formation. The earlier MCSs are responsible for the first stage vortex enhancement, and depending on the location can affect quite largely the simulated formation location. The extreme convection within the last MCS before formation largely determines the formation time.

1. Introduction
The formation of tropical cyclones (TCs) has long been a major area of research, and has resulted in theories such as the convective instability of the second kind (CISK; Charney and Eliassen 1964) and wind-induced surface heat exchange (WISHE; Emanuel 1986). Whereas these theories focus on the intensification process of TCs after the basic kinematic structure and, in some cases, the warm-core structure is already established, the formation processes from weak or unorganized disturbances (DB) to a tropical depression (TD) are believed to be different and more stochastic in nature. Although the large-scale environmental conditions necessary for TC formation are well known (Gray 1998), the mesoscale and convective-scale mechanisms for DBs to develop into a TD are not well understood (Tory and Frank 2010).

In the western North Pacific, about 70% of TC formations occur in the monsoon trough (MT) (Lander 1994; McBride 1995; Chen et al. 2004). The MT is often associated with a favorable background environment with high relative humidity (RH), cyclonic relative vorticity, convective instability, strong low-level convergence,
and broad scale of persistent cloud clusters. Therefore, variability in the MT substantially impacts the spatial and temporal variation in TC formation as well as frequency of formation (Chang et al. 1996; Chen et al. 2004; Cheung 2004). Occurrence of mesoscale convective systems (MCSs) at multiple times was identified during the 48 h prior to TC formation in about 70% of all TCs that formed in the monsoon-related synoptic patterns such as monsoon confluence and monsoon shear (see definitions in Ritchie and Holland 1999; Lee et al. 2008). In particular, it was found in these studies that the coexistence of MCSs during TC formation is common in the monsoon-related synoptic patterns.

A midlevel mesoscale convective vortex (MCV) is often generated in MCSs with severe convection (Bartels and Maddox 1991). Once formed, the MCV often continues to exist after its initiating parent MCS has weakened or even dissipated. Recent observational and modeling studies show growing evidence of the important roles of long-lived MCVs during TC formations (Bosart and Sanders 1981; Bartels and Maddox 1991; Harr and Elsberry 1996; Harr et al. 1996; Ritchie and Holland 1997; Simpson et al. 1997; Cheung and Elsberry 2006; Kieu and Zhang 2008). Therefore, whereas the large-scale monsoon trough provides a favorable climatological and synoptic environment, the occurrence of MCSs may be an indispensable and crucial factor for TC formations.

The physical mechanisms through which MCSs influence TC formations have been debated in the past decade. There are two main available theories. One of these theories, the so-called top–down theory, is based on the classic MCS structure with an MCV in the stratiform rain region that usually develops in an environment with substantial low-level vertical wind shear (Chen and Frank 1993). The MCV frequently has a slight extension downward to the surface by mechanisms such as vertical advection (Rogers and Fritsch 2001), and is anticipated to be the potential focal point for TC formation (Harr et al. 1996; Simpson et al. 1997; Ritchie and Holland 1997; Bister and Emanuel 1997). Such a downward extension of MCVs has been observed in U.S. Great Plains MCSs (Maddox 1980; Bartels and Maddox 1991; Fritsch and Maddox 1981a,b; Menard and Fritsch 1989; Miller and Fritsch 1991) as well as simulated MCSs (Zhang and Fritsch 1986, 1987; Chen and Frank 1993).

The bottom–up theory is somewhat based on the observations of Zehr (1992) that low-level vortex intensification sometimes follows bursts of intense deep convection. Montgomery et al. (2006) suggested that this deep convective, low-level vortex enhancement is taking place within the MCSs well before the system-scale vortex becomes self-sustainable. Montgomery et al. further suggested that vortical hot towers (VHTs) with spatial scales of 10–20 km play an important role during the process (see also Hendricks and Montgomery 2006; Tory et al. 2006a,b). In addition, based on the observational and modeling evidence of VHTs, Reasor et al. (2005), Sippel et al. (2006), and Fang and Zhang (2010) indicated that the bottom–up process is likely a realistic pathway to the enhancement of the surface vortex during TC formation.

Both of the theories suggest that MCSs are of great importance in TC formation processes. However, what is the exact role of multiple MCSs in the low-level vortex enhancement? Recently, the enhancement of the low-level vortex and the rapid occurrence of VHTs during MCSs development have been detected by using the latest remote sensing data (Lee et al. 2008), but whether the process is bottom–up or top–down or a mixture of the two could not be revealed by the coarse-resolution satellite data. Based on the work of Lee et al. (2008), this study utilizes high-resolution numerical simulations to further investigate 1) how the multiple MCSs contributed to the organization and growth of the low-level cyclonic, vertical relative vorticity during the process when the DB developed into Ketsana (2003) within a reverse-oriented monsoon trough [southwest to northeast versus the climatological orientation the other way round; see Lander (1994)], and 2) the respective roles of individual MCSs during the evolution of the pre-Ketsana DB.

The next section describes the data and model used in this study. Section 3 provides a synopsis of Typhoon Ketsana (2003) and the evolution of MCS activity prior to Ketsana’s formation. Section 4 shows model verification against available observations. The three-dimensional structure of the involved MCSs and contribution of them to the low-level vortex enhancement during Ketsana’s formation are presented in section 5. In section 6, several sensitivity experiments are conducted to further examine the role of individual MCSs during the transformation from the DB to TD, and the response of the DB’s development to the respective intensities of the MCSs. A summary and discussion are given in the final section.

2. Data sources and model description

The gridded reanalysis data with 2.5° latitude–longitude resolution from the National Centers for Environmental Prediction (NCEP) is used to analyze the large-scale environment associated with Ketsana’s formation. Hourly images from the Geostationary Operational Environmental Satellite-9 (GOES-9) infrared channel-1 (IR1) with wavelength 10.3–11.3 μm and 5-km resolution are used to monitor deep convective clouds and MCSs. The images are from a data archive in Kochi University in Japan that covers the area 20°S–70°N, 70°–160°E. The
cloud-top height is inferred from the infrared brightness temperature (TB). Areas with TB lower than the thresholds of $-32^\circ$, $-60^\circ$, and $-75^\circ$C, which have been applied for MCS activity analyses in Lee et al. (2008), are considered to be convection in cloud clusters, deep convection in MCSs, and extreme convection in vertical hot towers, respectively. Following Lee et al. MCSs in observations and numerical simulations are identified by deep convection with TB $> 214$ K ($-56.15^\circ$C), an area larger than $4 \times 10^4$ km$^2$, and eccentricity larger than 0.5.

In this study, the nonhydrostatic Advanced Research Weather Research and Forecasting (WRF) model version 3.1.1 developed by the National Center for Atmospheric Research (Skamarock et al. 2008) is used to simulate the processes leading to the formation of Ketsana. The model domains are triply nested through two-way nesting with horizontal resolutions of 27/9/3 km, and mesh sizes of 198 $\times$ 154 (D1), 295 $\times$ 232 (D2), 436 $\times$ 346 (D3), respectively (Fig. 4a). The model has 35 vertical levels with model top at 50 hPa. The modified version of the Kain and Fritsch cumulus parameterization scheme (Kain 2004) is used in the 27- and 9-km resolution domains while only explicit moisture calculation is used in the 3-km resolution domain. Other physics packages include the WRF single-moment six-class microphysics scheme with graupel (WSM6; Hong et al. 2004) and the Yonsei University planetary boundary layer (YSU-PBL) scheme (Noh et al. 2003).

The WRF model is initialized at 0000 UTC 16 October 2003 and 84-h integration is carried out until 1200 UTC 19 October 2003. The model initial and lateral boundary conditions are taken from the 1° latitude–longitude-resolution NCEP Global Forecast System final gridded analyses with the outermost lateral boundaries updated every 6 h. The NCEP daily sea surface temperature (SST, resolution 0.5° latitude–longitude) field is used to update the oceanic boundaries. In the original simulation without data assimilation, the location of TC formation is to the south of that reported in the Joint Typhoon Warning Center (JTWC) best track, and the distribution of MCSs does not exactly match the distribution observed in satellite images. To improve the atmospheric conditions as represented in WRF, the National Aeronautics and Space Administration’s Quick Scatterometer (QuikSCAT) oceanic winds available at 0600 UTC 16 and 0600 UTC 17 October 2003 and the Defense Meteorological Satellite Program’s Special Sensor Microwave Imager (SSM/I) oceanic surface wind speed and total precipitable water (TPW) available at 1200 UTC 17 October 2003 are assimilated using the WRF-Var 3.1.1 variational data assimilation system (Skamarock et al. 2008) with a time window of 6 h. The 25-km horizontal resolution, twice daily swath data of QuikSCAT and SSM/I are extracted from the Remote Sensing System (RSS) data archive. This simulation with data assimilation is treated as the control experiment. The 54-h period from 1200 UTC 17 to 1800 UTC 19 October 2003 during the development from DB to TD associated with Ketsana is the focus of the analysis.

3. Synopsis of Typhoon Ketsana

According to the information from JTWC, Ketsana was first detected as a DB approximately 700 nautical miles (1296 km) east of Luzon Island at 0600 UTC 15 October 2003. The TC formation alert (TCFA) was issued at 1300 UTC 18 October 2003 when the system developed to the east of the Philippines (Fig. 1). In this study, the formation time is taken as 1200 UTC 18 October 2003 when the intensity reached 25 kt ($\sim 12.86$ m s$^{-1}$) for the first time. The development of Ketsana was noted as being initially slow, the TD developed into a tropical storm around 0000 UTC 19 October 2003 and then achieved typhoon intensity around 1200 UTC 20 October 2003. The movement of the TC was initially very slow as it was south of the weak subtropical ridge, which did not provide a strong steering flow. For this reason, geographically fixed grids are used in all the WRF model simulations.

Typhoon Ketsana formed in a reverse-oriented MT (Fig. 1) that provided a favorable large-scale environment with high humidity and abundant low-level cyclonic vorticity for TC formation (not shown). There were two low-level jets at each side of the monsoon trough; the
westerlies to the south were particularly strong. During mid October when Ketsana developed, the integrated amplitude of the first two real-time multivariate Madden–Julian oscillation (MJO) indices (Wheeler and Hendon 2004) both exceed 1 and the phase of MJO was 5 in the western Pacific (not shown). Therefore, the MJO was in the active/wet phase and was considered favorable for TC formation (Liebmann and Hendon 1990; Maloney and Hartmann 2000; Kim et al. 2008).

The IR1 satellite images during 17–18 October 2003 show that the persistent cloud clusters are active in the monsoon trough area (Fig. 2). During the 48 h prior to Ketsana’s formation, five MCSs are observed. The first MCS (MCS1) developed around 1300 UTC 16 October 2003 northwest of the low-level circulation center and then MCS2 developed at about 0400 UTC 17 October 2003 to the north. Afterward, two of them (the third and fourth MCSs, hereafter referred to as MCS3 and MCS4, respectively) developed almost simultaneously near 1500 UTC 17 October 2003 and lasted until 0400 UTC 18 October 2003. Later, the fifth MCS (hereafter referred to as MCS5) developed at 0600 UTC 18 October 2003,
which was 6 h prior to Ketsana’s formation. The time series of the percentage area of convection in the region where the DB developed show abundant deep and extreme convection during MCS1 as early as 1200 UTC 16 October 2003 and during development of MCS3 and MCS4 (Fig. 3). After a short while of quietness, deep and extreme convection developed again associated with MCS5, and the convection further amplified right after the formation of Ketsana. As mentioned in the introduction, the coexistence of MCSs during TC formation is common in the monsoon-related synoptic patterns of the western North Pacific. Therefore, the convective episodes during Ketsana’s formation are highly representative in the basin. Since the first two MCSs are relatively weaker and shorter in lifetime, the present study focuses on discussing the possible different roles of the last three MCSs in the TC formation.

4. Model verification

a. Large-scale circulation

Comparison between the WRF simulation in domain 1 and the corresponding NCEP reanalysis data shows that the large-scale circulation has been simulated well (Figs. 4a,b), which provides a credible background environment for the successful mesoscale simulations in the nested fine-mesh domains. With assimilation of the SSM/I and QuikSCAT data, the low-level cross-equatorial flow between 110° and 140°E, and the location and intensity of the reverse-oriented MT are consistent with those in the NCEP reanalyses. In particular, the strong southwesterlies south of the MT are essential for reproducing the orientation and strength of the MT, and enhancing the low-level convergence in the MT that is crucial for generating MCSs. At the upper levels, the location and evolution of the South Asia high, which was northeast of the position where Ketsana developed, also resemble the same in the NCEP reanalysis (Figs. 4c,d). This provides an appropriate outflow channel for the pre-Ketsana DB development.

b. The intensity of Ketsana

During the period of 1200 UTC 17–18 October 2003, the simulated sea level pressure pattern is similar with that in the NCEP reanalysis (not shown). At 1200 UTC 18 October 2003, the domain-1 simulated center of the sea level pressure where Ketsana formed is 14.6°N, 129.8°E, which is close to the JTWC best-track position of 14.7°N, 130.3°E. In addition, the low-pressure centers in the higher-resolution simulations in the inner domains are more to the west of the observed formation location, which is consistent with the fact that MCS5 developed west of the system center before TD intensity was attained.

Ketsana’s formation is not a gradual process of evolution of the preexisting DB into TD, but roughly a two-stage development. There is a period with slow change in the minimum sea level pressure before 0000 UTC 18 October 2003, followed by a period of rapid transformation when the sea level pressure drops at a higher rate through the formation and early intensification of the system (Fig. 5a). Since the reported intensity of Typhoon Ketsana in the best track starts from formation time at 1200 UTC 18 October 2003, the simulated intensity before that cannot be validated directly. However, it can be seen that the simulated maximum surface wind increases rapidly to 15 m s\(^{-1}\) (or TD intensity) 6 h before formation. After formation, the WRF simulated intensity is slightly higher than that in the best track during the intensification to tropical storm but the overall trend until the end of simulation (1800 UTC 19 October 2003) agrees well with observations. Therefore, the formation time of 1200 UTC 18 October 2003 is also applied when analyzing the WRF simulations in the following.

The area average (within 13°–17°N, 128°–132°E) observed TB from satellite images shows a minimum at around 2100 UTC 17 October 2003 that is associated with MCS3 and MCS4, and then there is another major decrease before formation associated with MCS5 (Fig. 5b). Since the depth of convection and thus TB is not usually closely reproduced in numerical models, the simulated radar reflectivity within the same area that measures the convection activity in the model is examined instead for
comparison. The time series of simulated radar reflectivity has a local maximum at the same time of occurrence of MCS3 and MCS4. It then increases rapidly 6 h before Ketsana’s formation, which is due to convective bursts within MCS5 and is consistent with the observed variation of the area-average TB.

c. Evolution of the MCSs

The simulated MCS activities are examined via the simulated radar reflectivity (Fig. 6). MCS3 starts to develop at 1200 UTC 17 October 2003 east of the formation position (Fig. 6, upper panels), and it dissipates at about 0400 UTC 18 October 2003 (not shown). Comparison with the observed TB shows that MCS3 occurs a little earlier in the model, but otherwise its evolution is similar to that observed. MCS4 occurs at 1700 UTC 17 October 2003 in the model that is two hours later than observations (Fig. 6, upper left panel). Although the simulated intensity of MCS4 is slightly weaker and the lifetime shorter than that observed, it moves northwestward and then dissipates gradually as revealed in satellite images. MCS5 starts to develop near the location of the decaying MCS3 at 0400 UTC 18 October 2003 and then strengthens rapidly to lead to the formation of Ketsana (Fig. 6, lower left panel). Both the location of major convection and time of development associated with the simulated MCS5 resemble those implied from satellite images. The spiral-shaped cloud bands at 1200 UTC 18 October 2003 indicate that the system-scale low-level vorticity has been strengthened by the formation time (Fig. 6, lower right panel). Moreover, there are some small-scale, extreme convective systems with large

![Fig. 4. (a),(b) The 850- and (c),(d) 200-hPa geopotential height (unit: m) and wind vector at 1200 UTC 17 and 1200 UTC 18 Oct 2003, respectively; 850-hPa RH larger than 80% is also shown (shaded). Here, (a),(c) are from the 27-km resolution WRF simulation and (b),(d) NCEP reanalysis. (a) The WRF model domains (D1, D2, and D3) with horizontal resolutions of 27, 9, and 3 km are indicated.](image-url)
values of radar reflectivity embedded in the simulated MCSs such as that within MCS5 near 14.6°N, 129.8°E. These convective-scale systems may be candidates for VHTs, the characteristics of which will be further explored in section 6.

To summarize, the WRF model can reproduce well the large-scale environment, the major convective episodes associated with the MCSs, and the formation time and temporal intensification profile of Ketsana. Thus, the model simulations from the 9- and 3-km resolution inner grids are used to further examine the detailed structures of the MCSs.

5. The structure of the MCSs and Ketsana’s formation

a. The three-dimensional structures of the MCSs

The temperature anomaly is calculated between 14° and 16°N with respect to the zonal average in the 9-km WRF simulation. It is found that each MCS from MCS3 to MCS5 has a warm-core structure that concentrates between 400 and 200 hPa (Figs. 7a,b). The largest temperature anomaly is about 2.5 K in MCS5, which just precedes formation of Ketsana. The cold anomalies underneath the warm cores suggest that the heating profiles consist of stratiform type as depicted in Houze (1989). The Houze study showed that a deep convective heating profile within MCSs has maximum heating at a height of about 4–6 km, while stratiform heating profile has maximum heating at about 8 km but cooling below 4 km. On the other hand, examination of the upper-level divergence field reveals the major centers of convection within the MCSs (Figs. 7c,d). For example, at 1800 UTC 17 October 2003 the center at about 14°N corresponds to MCS3 and the two others at 15°–16°N to MCS4 (cf. Fig. 6a). Later at 0900 UTC 18 October 2003 the strong center of divergence is atop MCS5 just south of 15°N (compared with Figs. 6c,d). Comparison between the convection centers and temperature anomalies indicates that the warm cores, especially those for MCS4 and MCS5, do not coincide with the convection centers but displaced about 100 km. This relationship is consistent with the mechanism of MCS development that latent heat release to create the warm anomaly is in the region of stratiform rainfall, which is often located behind the center of major convection with the MCS with respect to the low-level vertical wind shear direction.

The warm anomaly concentrates at the upper levels as in Figs. 7a,b, and slight cold anomalies are found at the lower levels to the west, which may be attributable to the evaporative cooling of rainfall that decreases the temperature. It is not clear at this stage how the individual mesoscale warm-core centers associated with MCS3 to MCS5 lead to the temperature anomaly structure when Ketsana forms. Generally speaking the warm core of a TC is the result of a mesoscale dynamical adjustment to the heating in deep convection that is embedded in MCSs (Hack and Schubert 1986; Nolan et al. 2007). These two studies showed that the dynamic response of an incipient vortex to a diabatic heat source, especially in terms of the horizontal scale, is strongly dependent on the vertical distribution of the heat source. Thus, the warm and cold anomalies in each MCS here may play a role in leading to the final thermodynamic structure of the vortex during formation. Nonetheless, it is reasonable to hypothesize that at least for MCS5, which develops close to the system center of formation, its associated diabatic heating is essential for transformation to the system-scale warm-core structure.

Figure 8a depicts the time–height diagram of the area-average (13°–17°N, 128°–132°E) RH anomaly as well as the average RH itself. It can be seen that the
occurrences of the MCSs increase the midlevel (400–550 hPa) RH by as much as 18%, resulting in average RH of over 70% at this layer. At a lower layer of 600 hPa, RH is over 80%, and the time series in Fig. 8b shows that the specific humidity increases rapidly before formation. These moisture conditions are consistent with the idealized simulation results in Nolan (2007), which were that high RH over most of the troposphere is necessary for midlevel vortex development.

Examination of the temporal variation of the area-average (13°–17°N, 128°–132°E) relative vorticity in the simulation indicates that starting from about 1800 UTC 17 October 2003 the surface to midlevel relative vorticity is already quite high (over $5 \times 10^{-5}$ s$^{-1}$; Fig. 9a). The system relative vorticity then increases gradually as a column until formation time and continues to increase afterward during early intensification. Similar evolution of relative vorticity is found in the idealized simulations in Nolan (2007), with either a surface or midlevel vortex as initial condition. The area-average divergence in Fig. 9a shows strong near-surface convergences that commence at around 1500 UTC 17 and 0600 UTC 18 October 2003, which correspond to the development of MCS3 and MCS4, and MCS5, respectively. The causal relationship between convergence and MCSs development, however, is uncertain and needs further investigation. It is suspected that the low- and midlevel relative vorticity that is associated with the formation of Ketsana may be generated from different mechanisms of various spatial scales. One candidate is the transition of the midlevel MCV within the MCSs toward the surface on the time scale of hours. The time series of the area-average potential vorticity (PV) depicts stronger midlevel PV after the development of MCS3 and MCS4.
Moreover, the midlevel PV extends downward to about 850 hPa after the development of MCS3 and MCS4 early on 18 October 2003 and again just before formation when MCS5 develops. These downward extensions are suggestive of vertical stretching for PV generation within the MCSs.

In the WRF simulation, strong midtropospheric updrafts and PV are identified within the MCSs. Figure 10 shows that there are dominant mesoscale PV cores associated with the MCSs, first for the earlier MCS3 and MCS4 at the periphery of the disturbances and then with MCS5 near the center of circulation. From the movement of these PV cores, merging of the MCVs is not apparent but axisymmetrization should occur as the storm-scale cyclonic circulation gets stronger during the evolution.

**b. Mid- and low-level vortex enhancement**

To quantify the above argument on midlevel MCVs development within the MCSs, budget analysis based on circulation formulation (Weisman and Davis 1998; Trapp and Weisman 2003; Davis and Galarneau 2009) is performed, which is according to the following tendency equation:

\[
\frac{\delta C}{\delta t} = -\eta \delta A - \int \eta \mathbf{V} \cdot \hat{n} \, dl + \int \mu \left( \mathbf{k} \times \frac{\delta \mathbf{V}}{\delta p} \right) \cdot \hat{n} \, dl + \int \mathbf{(k} \times \mathbf{F}) \cdot \hat{n} \, dl,
\]

where \( C \) is the circulation, \( \eta \) is the absolute vorticity, \( \delta \) the divergence, \( A \) the area of the box under consideration,
v (V) the horizontal (3D) wind vector, n the unit vector normal (perpendicular) to the box, ω the vertical velocity in pressure coordinate, k the vertical unit vector, and F the frictional force. The line integral is performed along a close area, which is a box within 13°–17°N, 128°–132°E herein. In the equation, overbars define the average value around the perimeter of the box, primes denote a perturbation from this average, and tildes indicate the average over the area of the box. The terms on the right-hand side of the equation represent stretching, eddy flux, tilting, and frictional effects, respectively. The advantage of this circulation formulation over the traditional form is that one can examine the net mesoscale circulation within a region in which complex convection and vorticity dynamics are taking place. For example, an updraft that tilts horizontal vorticity into the vertical will produce equal yet opposite signed vertical vorticity anomalies except the updraft is along the edge of the box, and either the positive or negative (but not both) vertical vorticity anomaly is within the box. This kind of process can be shown explicitly when a line integral along the peripheral of the box is performed. Moreover, net import or export of vorticity through horizontal advection is taken into account under the circulation budget framework. The

Fig. 8. (a) Time–pressure diagram of the area-averaged (13°–17°N, 128°–132°E) relative humidity anomaly (contour, unit: %) and area-averaged relative humidity (shaded). (b) Time series of the specific humidity anomaly (g kg⁻¹) at 600 and 925 hPa. Specific humidity anomaly is equal to the deviation from the average within the area 10°–20°N, 124°–136°E.
readers are referred to Weisman and Davis (1998), Trapp and Weisman (2003), and Davis and Galarneau (2009) for derivation of the budget equation and more detailed discussion on interpretation of the budget items.

The simulated total tendency of relative vorticity shows a low- to midlevel maximum right after 1800 UTC 17 October 2003, which is associated with MCS3 and MCS4 (Fig. 11a). There is another low-level increase at about 0000 UTC 18 October 2003, but tendency is small afterward. Relative vorticity increases again near the time of development of MCS5 between 0600 and 1200 UTC 18 October 2003 as a column from lower to upper troposphere, which occurs just before the formation of Ketsana.

The stretching term of circulation budget (Fig. 11b) depends on the mean divergence within the box of calculation, and thus its variation is similar to the divergence field in Fig. 9a. The low-level stretching is the strongest throughout the times of examination according to the divergence field. When MCS3 and MCS4 develop, strong stretching effect extends to mid- to upper-mid levels. The
tendency of stronger stretching at the mid levels is also seen during the development of MCS5 before and after the formation time.

The eddy flux term in the circulation budget equation is essentially horizontal advection by the perturbation wind components at the edge of the area for computation (Davis and Galarneau 2009). This eddy flux term shows large values throughout the troposphere during the development of MCS3 and MCS4 at about 1800 UTC 17 October 2003, with the maximum at midlevel (Fig. 11c). Note that the major convective cores associated with MCS3 and MCS4 are a distance away from the system center and near the boundary of the box for calculating budget. Thus, this maximum in eddy flux is closely related to the perturbations in wind field as response to the convection and associated heating of the two MCSs. Later on, convection associated with MCS5 develops before Ketsana’s formation. However, the eddy flux term only has an upper-level maximum. This is likely because MCS5 develops near the system vortex center, and the largest perturbations in wind field are induced by the upper-level divergence associated with deep convection.

Last, the tilting term in the circulation budget equation actually includes vertical advection as in the traditional form of vorticity budget. However, the pattern of variability for this term in circulation form is similar to that in the traditional budget calculation (not shown), which implies that vertical advection does not contribute

FIG. 10. The 500-hPa simulated potential vorticity (shaded, unit: PVU) and wind vectors at (a) 1800 UTC 17 Oct 2003, and (b) 0000, (c) 0600, and (d) 1200 UTC 18 Oct 2003.
much to the budget. The tilting term shows several maximum values at the upper levels with their timing correlated with the development of MCS3 and MCS5 (Fig. 11d). In contrast, the values are small and at some times negative at the lower levels. Overall, the summation of the three budget terms of stretching, eddy flux, and tilting well accounts for the total relative vorticity tendency in Fig. 11a, and thus any frictional effect is negligible. Low-level to midlevel stretching is largely responsible for the increase in relative vorticity during the development of the MCSs. During the several hours before Ketsana’s formation, upper-level relative vorticity also increases, which is attributable to both the eddy flux and tilting mechanisms.

6. Further analysis of the roles of MCSs in Ketsana’s formation

a. General consideration

The above analyses show that the multiple MCSs prior to Ketsana’s formation are closely related to the process of strengthening the relative vorticity in both the middle and low levels. A natural question to arise is how critical are the successive occurrences of MCSs to the TC formation, and do they play different roles during the formation? To explore clues to this question, several sensitivity experiments in WRF are performed that change the intensities of the MCSs individually. The methodology is to modify the vertical distribution of humidity in the region of an MCS by assimilating reduced midlevel RH in that MCS region by using the WRF variational data assimilation system. The first basis of this method is the well-known condition that midlevel abundance of humidity is favorable for TC formation (e.g., McBride 1995; Briegel and Frank 1997; Gray 1998), and indeed large RH anomalies at 500–700 hPa are identified in the coarse grid of WRF control simulation especially during the MCSs development (Fig. 8). From the mesoscale perspective, more moisture content allows larger precipitation and prevents evaporative cooling. Subsequently downdrafts are decreased and low-level net positive latent heating is possible, which can then drive low-level convergence that creates the vortex. On the other hand, some downdrafts associated with heavy precipitation may be favorable for triggering new MCS development given the appropriate vertical wind shear environment. Four sensitivity

![Figure 11](image-url)
experiments are designed with details as described in the following.

b. Design of the four sensitivity experiments

Base on satellite images, the MCS3, MCS4, and MCS5 regions are identified as (12°–15°N, 127°–130°E), (15°–18°N, 130°–133°E), and (12°–17°N, 127°–130°E), respectively. In the control experiment, the midlevel RH in MCS4 is much smaller than that in MCS3 at 1200 UTC 17 October 2003 before their initialization (Fig. 12a), and this is the reason why the simulated activity of MCS4 is weaker and its lifetime shorter than that observed. In the sensitivity experiment 1 (EXP1; Table 1), the area-averaged RH in the MCS3 region between 200 and 700 hPa is assimilated into the MCS4 region within the 6-h assimilation window (similarly for later experiments) centered at 1200 UTC 17 October 2003 in order to strengthen the intensity of MCS4. After the assimilation, the midlevel RH within MCS4 increases substantially (Fig. 12b).

In experiment 2 (EXP2), MCS4 is weakened and retains only 60% of the 200–700-hPa relative humidity in the control experiment and then is assimilated at 1200 UTC 17 October 2003. Likewise, the intensities of MCS3 and MCS5 are weakened in sensitivity experiments 3 and 4 (EXP3 and EXP4), respectively, by adjusting the 200–700-hPa RH at 1200 UTC 17 October 2003 for MCS3 and 0600 UTC 18 October 2003 for MCS5 to 60% of the original values (Fig. 12). Other environmental variables in the model are not changed directly in the sensitivity experiments (but will be indirectly through the changes in RH).

c. Results

When the midlevel RH is adjusted in the sensitivity experiments, the development of the directly associated MCS is modified substantially. For example, the simulated MCS4 in EXP1 is stronger and its lifetime longer than that observed (Fig. 13a) but the convection in the region of MCS4 in EXP2 is almost shut down (Fig. 13c). A similar effect applies to MCS3 in EXP3 (Fig. 13e), and in EXP5 MCS5 does not occur until 1200 UTC 18 October 2003, which is about 6 h later than in the control experiment (Figs. 13g,h).

It is interesting that the results from these four sensitivity experiments show that the subsequent MCSs developments depend greatly on the earlier ones. The strengthened (weakened) MCS4 in EXP1 (EXP2) leads to more (less) severe convection later in the region of MCS5 (Figs. 13b,d). On the other hand, when MCS3 has less convection in EXP3, there is also less inner-core convection associated with MCS5 in the following day (Figs. 13e,f). These results reflect the earlier comment on the interrelationship among MCS developments through the modification of precipitation, latent heat release, and the mesoscale circulation.

To verify the aforementioned relationships between the MCSs development, forward trajectory analysis is performed with respect to the control simulation for air parcels originating from midlevels (500 or 700 hPa) within MCS3 or MCS4 during 1200 UTC 17–18 October 2003. The trajectories of these air parcels follow the cyclonic circulation for some time. While some of them stay at mid levels, some of them get nearer the inner core where MCS5 develops and then follow the updraft associated with convection of that last MCS (Fig. 14). That is, some of the moistened air parcels by the earlier MCS3 and MCS4 did affect the moisture content at the region where MCS5 develops, and through this mechanism influence the strength of the last inner-core updraft before the formation of Ketsana.

The different scenarios of MCS development in the sensitivity experiments affect the pace of system vortex development associated with the formation of Ketsana.
(Fig. 15). In EXP1, the minimum sea level pressure reduces faster and the lower-level relative vorticity is enhanced more quickly than in the control experiment during the development of MCS4. At 0600 UTC 18 October 2003 the system relative vorticity is actually similar to that in the control experiment, but stronger convection in the subsequent MCS5 in EXP1 leads to faster system intensification and an earlier formation time of 0900 UTC 18 October 2003 (versus 1200 UTC 18 October 2003 in the control; Table 1). Moreover, the TC formation location (15.1°N, 130.8°E) is nearer to the MCS4 location and a little northeast of that in the control experiment. On the contrary, system intensification rates are all retarded to different degrees when one of the MCSs is weakened. In EXP2 and EXP3, the early intensification of the DB is much slower than that in the control experiment. Therefore, when MCS5 develops at about 0600 UTC 18 October 2003 it needs a much longer time for the DB to intensify to TD. The estimated formation times, at which the vortices intensify to the same intensity as that at the control-run formation time, in these two experiments are 2100 UTC 18 and 0000 UTC 19 October 2003, respectively. In EXP4, the development of the two earlier MCSs is exactly the same as in the control. When the convection associated with MCS5 is much reduced in EXP4 at 0600 UTC 18 October 2003, the system intensification almost stops completely until inner-core convection finally develops after 1200 UTC 18 October 2003 to increase the DB’s low-level vorticity and deepen the central sea level pressure. However, this system rehabilitation is quite effective, and the estimated TC formation time for this sensitivity experiment is 1800 UTC 18 October 2003, which is only 6 h later than in the control.

These sensitivity experiments indicate that the last MCS5 determines the formation time largely because its convection is near the inner core, and thus the strength of that convection affects the vorticity within the system center. Nonetheless, the earlier convection associated with MCS3 and MCS4 also influences the pace of storm development through their impacts on midtropospheric relative vorticity generation. The available relative vorticity within the MCVs is then subject to possible mechanisms of merging and downward transfer, and enhances the system-scale relative vorticity. Moreover, it will be shown in the following that VHT activity is related to that of the MCS, and that the earlier convection affects the available low-level convergence and vorticity as well.

Note that similar sensitivity experiments were performed in Nolan (2007). In that study, several experiments that also utilized the WRF model and with different RH vertical distributions (environmental versus MCS-specific as applied in this paper) were performed. A surface vortex develops quickly in a near-saturated (90% RH) troposphere, and that development in a dry environment (5% RH) above 7 km is only delayed slightly compared with their control simulation. However, if the environment is dry above 4 km, the upper-level saturation and subsequently the midlevel vortex development occur much later. As a result, TC formation is delayed by about two days, indicating the strong dependence on midlevel moisture.

d. Scale separation

To further explore the mechanism for the generation of near-surface relative vorticity during the TC formation—in particular the potential contribution from small-scale convective bursts in VHT-like systems—scale separation is performed for the 3-km resolution simulation. In further detail, the spatial filter based on the two-dimension spectral decomposition of Lin and Zhang (2008) is applied to the simulated relative vorticity field every 10 min, which separates the mesoscale processes into three scales: the meso-α scales larger than 250 km, the meso-β scales between 50 and 250 km, and the meso-γ scales less than 50 km. The vortices with meso-γ scales are considered to reflect the activity of the VHTs, while those with meso-α and meso-β scales reflect the vorticity associated with the developing DB as well as possible MCVs.

It is found that the activity of the VHTs is closely related to the MCSs development (Fig. 16a). The time series of the percentage area with VHTs show a peak at 1800 UTC 17 October 2003 when the MCS3 and MCS4 are mature. After a quiet period, the activity starts to rise again for the 12 h before formation with a lot of convective bursts in the region of the MCS5 (not shown). The summed relative vorticity of the meso-γ vortices with
FIG. 13. As in Fig. 6 but for the following sensitivity experiments: EXP1 at (a) 1800 UTC 17 and (b) 1200 UTC 18 Oct 2003; EXP2 at (c) 1800 UTC 17 and (d) 1200 UTC 18 Oct 2003; EXP3 at (e) 1800 UTC 17 and (f) 1200 UTC 18 Oct 2003; and EXP4 at (g) 0800 and (h) 1200 UTC 18 Oct 2003. The black dot is the best-track location of Ketsana’s formation.
larger than $5 \times 10^{-4} \, \text{s}^{-1}$ is then examined. (There are also meso-$\gamma$-scale downdrafts but their effects are not considered here.) These convective bursts generate strong relative vorticity at the low levels between 800 and 900 hPa (Fig. 16b), and apparently there are also two periods of substantial contribution from these small systems that correspond to the early MCS development and the last MCS5. In other words, the vertical stretching effect within these VHTs is mainly confined to the near-surface levels (Montgomery et al. 2006), which contributes to the low-level vortex enhancement as seen in the unfiltered relative vorticity field in Fig. 9a. How these convective-scale systems (versus the midtropospheric MCVs) lead to the surface vortex necessary for formation of Ketsana will be one of the major focuses in Part II of this study.

7. Conclusions

a. Summary

The effects of multiple MCSs on the formation of Ketsana (2003) in a reversed-oriented monsoon trough are analyzed in this study. During the 48 h prior to Ketsana’s formation at 1200 UTC 18 October 2003, five MCSs were observed in satellite images, of which two (MCS3 and MCS4) coexisted for about 12 h, and the fifth MCS (MCS5) was associated with the convective burst before formation. Thus, the period with the MCSs developing near a prominent DB is exactly during the developing stage that the DB transformed to the named TC.

Numerical simulations using the WRF 3.1.1 model with the finest grid resolution of 3 km are performed on the formation of Ketsana. By assimilating QuikSCAT oceanic winds as well as SSM/I oceanic surface wind speed and TPW, the WRF model simulates well the large-scale features such as the orientation of the monsoon trough, the convective episodes associated with the MCSs and their periods of development, and the formation time and location of Ketsana. The simulation in the highest-resolution grid of WRF also reproduces the rate of early intensification very well. The roles that the last three MCSs play in this TC formation are the focus of this part of the study.

Analysis of the relative vorticity budget via circulation budget analysis shows that the vorticity tendency mainly contributes to the vertical stretching term during the development of MCS3 and MCS4, but eddy flux and tilting mechanism also contribute largely during the few hours before formation. With the successive occurrence of MCSs, midlevel MCVs are generated. But the downward transfer of midlevel vorticity associated with the MCVs does not appear to be significant.

![Fig. 14. Forward trajectory analysis of an air parcel starting from 15.0°N, 129.5°E during 1200 UTC 17–18 Oct 2003 in the WRF control simulation. The horizontal coordinates are grid points in the model, while height is in km.](image1)

![Fig. 15. Times series of (a) the minimum sea level pressure (unit: hPa) and (b) the area-averaged (12.5°–17.5°N, 127.5°–132.5°E) relative vorticity (unit: $10^{-5} \, \text{s}^{-1}$) at model level $\sigma = 0.845$ in the 9-km resolution domain for the control experiment (CTL) and the four sensitivity experiments (EXP1, EXP2, EXP3, EXP4).](image2)
during the TC formation period. Although there is no explicit merging of MCVs identified in the simulations, the warm-core structure in each of the MCSs, especially just prior to formation, facilitates the thermodynamic structure of the system vortex associated with the DB.

Four sensitivity experiments are performed to analyze the possible different roles of the MCSs during the formation of Ketsana. In particular, the vertical RH profile in each of MCSs is modified through a data assimilation technique to change the strength of convection within the MCS. The results from these additional experiments
and forward trajectory analysis of the control simulation show that the development of an MCS depends substantially on that of the prior ones, thus leading to different scenarios of system intensification during the TC formation. The MCS4 that develops northeast of the system center is crucial in the vortex enhancement prior to the occurrence of the last MCS in the formation process. This MCS also affects quite largely the simulated formation location. Conversely, MCS3 develops west of the system center in close vicinity to the later MCS5. Therefore, the degree of convection within MCS3 affects that in MCS5, which is near the core of the system center. The sensitivity experiment with a weakened MCS5 has a suspended period of intensification, which indicates that the extreme convection within this MCS is determining the final route to the TC formation, and consequently the formation time as well.

On the other hand, scale separation shows that the activity of the VHT-type meso-γ-scale vortices is closely related to the development of the MCSs. These VHTs have large values of positive relative vorticity induced by intense low-level convergence and vertical stretching, and thus play an important role in the low-level vortex enhancement. Meanwhile, the accumulation and axisymmetrization of the strong positive vorticity of the VHTs (not shown in this study) are possible influences upon the generation of meso-β-scale vortices that eventually develop the TC core. Therefore, the “bottom-up” processes are still considered essential during the transformation from the DB into the TD of this typhoon case.

b. Discussion

Whereas it is identified that the MCS processes result in the generation of MCVs and enhance the mid- to low-level cyclonic vorticity and the meso-β-scale convergence forces development of VHTs that represent abundance of cyclonic vorticity in the near-surface level, it is not conclusive in this part of the study on how the final system vortex in the TC formation is attributed quantitatively to the MCVs and the VHTs, respectively. More detailed analysis will be carried out on how the midlevel vorticity and the vorticity at the low levels coming from the convective-scale systems interact with each other to create the column-wise cyclonic vortex. Aggregation of VHTs is indeed observed in the WRF simulations for Ketsana (but not very clearly in term of upscale cascade); however, this may not be the only route to enable a long-lasting vorticity anomaly as pointed out in Fang and Zhang (2010) study, in which bottom–up processes seems to be crucial for the formation.

It is worth mentioning that the dipolar structure of vorticity anomalies associated with VHTs, as discussed in Montgomery et al. (2006), is also found in this study. That is, the very strong positive vorticity anomalies of the VHTs are always accompanied by strong negative anomalies in their neighborhood, and this phenomenon extends to the meso-β vortices as well. The question is then: What is effect of the negative vorticity anomalies on TC formation processes, and do they have a negative influence on the low-level vortex enhancement? For example, studies such as Trapp and Weisman (2003) analyzed the impact of the convergence of planetary vorticity (i.e., vertically stretched in downdrafts), which can significantly inhibit the longevity and intensity of negative relative vorticity anomalies in time scales of a few hours. This is also consistent with the recent study of Fang and Zhang (2011) that both the negative and positive vorticity anomalies, which are manifestations of vorticity tilting, are accumulated into a large single-sign vorticity region through the larger-scale convergent circulation that is driven by latent heating. However, the negative vorticity anomalies are weaker and shorter-lived compared with the positive ones, and are thus absorbed. Issues such as these will be the focus in Part II of the study, which will also diagnose the thermodynamic aspects of these small-scale systems during Ketsana’s formation.

There are often multiple MCSs occurring near a prominent tropical disturbance that is embedded in favorable background environment, but finally the disturbance dissipates. In the earlier sensitivity experiments on MCSs development, the inner-core convection within the last MCS before formation seems to be quite critical in determining whether the system is a formation or non-formation. However, a collection of more TC cases is necessary to establish a consensus on this trigger mechanism of TC formation, and whether there is a consensus on the position of the triggering convection for various synoptic patterns. Moreover, further investigation is also needed on the relation between the inner-core convection as identified in this study and the long-lived updraft near the center of the system that creates the surface vortex necessary for TC formation, as identified in Nolan (2007). If more sensitivity experiments such as those in this study are performed for a collection of TC cases, better ideas can be obtained on the predictability of TC formation time in the western North Pacific.

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