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Current understanding of tropical cyclone structure and intensity changes – a review

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With 7 Figures

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Summary

Current understanding of tropical cyclone (TC) structure and intensity changes has been reviewed in this article. Recent studies in this area tend to focus on two issues: (1) *what factors determine the maximum potential intensity (MPI) that a TC can achieve given the thermodynamic state of the atmosphere and the ocean*? and (2) *what factors prevent the TCs from reaching their MPIs*? Although the MPI theories appear mature, recent studies of the so-called superintensity pose a potential challenge. It is notable that the maximum intensities reached by real TCs in all ocean basins are generally lower than those inferred from the theoretical MPI, indicating that internal dynamics and external forcing from environmental flow prohibit the TC intensification most and limit the TC intensity. It remains to be seen whether such factors can be included in improved MPI approaches.

Among many limiting factors, the unfavorable environmental conditions, especially the vertical shear-induced asymmetry in the inner core region and the cooling of sea surface due to the oceanic upwelling under the eyewall region, have been postulated as the primary impediment to a TC reaching its MPI. However, recent studies show that the mesoscale processes, which create asymmetries in the TC core region, play key roles in TC structure and intensity changes. These include the inner and outer spiral rainbands, convectively coupled vortex Rossby waves, eyewall cycles, and embedded mesovortices in TC circulation. It is also through these inner core processes that the external environmental flow affects the TC structure and intensity changes. It is proposed that future research be focused on improving the understanding of how the eyewall processes respond to all external forcing and affect the TC structure and intensity changes. Rapid TC intensity changes (both strengthening and weakening) are believed to involve complex interactions between different scales and to be worthy of future research.

The boundary-layer processes are crucial to TC formation, maintenance, and decaying. Significant progress has been made to deduce the drag coefficient on high wind conditions from the measurements of boundary layer winds in the vicinity of hurricane eyewalls by Global Positioning System (GPS) dropsondes. This breakthrough can lead to reduction of the uncertainties in the calculation of surface fluxes, thus improving TC intensity forecast by numerical weather prediction models.

1. Introduction

Tropical cyclones (TCs) are intense atmospheric vortices that develop over the warm tropical oceans. They are among the most feared and deadly weather systems on Earth. Severe TCs produce destructive winds, high surges, torrential rains and severe floods, usually resulting in serious property damage and loss of life. However, tremendous rainfall brought by a TC can temporarily relieve draught conditions in some cases and thus benefit residents and agriculture in the vicinity. Accurate forecasting of both track and intensity of a TC is critical to mitigation of disasters potentially caused by an approaching TC.

Forecasts of TC motion have improved steadily over the last three decades, mostly due to a combination of better observations, especially the satellite (Soden et al, 2001) and dropsonde (Burpee et al, 1996; Tuleya and Lord, 1997; Aberson and Franklin, 1999) data; improvements in dynamical models (Kurihara et al, 1998); and improved understanding of physical processes and mechanisms that govern the motion of TCs (Wang et al, 1998; Emanuel, 1999). In contrast, there has been relatively little advance in prediction of TC intensity and its change, in spite of the application of sophisticated numerical models and availability of advanced satellite observations (Avila, 1998; McAdie and Lawrence, 2000). The best forecasts of TC intensity in operational use are still statistically based (DeMaria and Kaplan, 1999). Among many reasons (such as inadequate observations over the ocean, inadequate model resolution and physics, and poor initial conditions), deficient understanding of the physical processes governing the intensity change is a fundamental one.

The life cycle of a TC consists of a succession of stages (Fig. 1) starting from the establishment of an environment with the required thermodynamic capacity for development. Initiation of a region of sustained convection, which may stem from a variety of processes, enables the development of a surface mesoscale or synoptic vortex. The system then becomes self-sustaining and capable of intensifying through a series of internal and external processes and complex interactions. At the climax we have an intense vortex that is able to extract enormous energy from the ocean and may undergo intensity oscillation, and then decay in an unfavorable environment, such as moving to a region with cold sea-surface temperature (SST) or strong vertical wind shear, or moving inland.

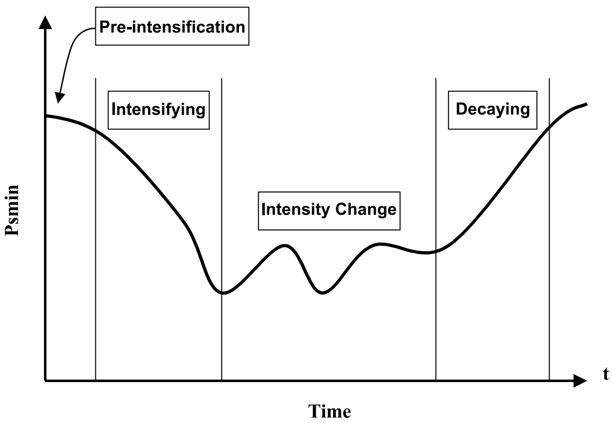

Fig. 1. Schematic diagram showing the lifecycle of a tropical cyclone

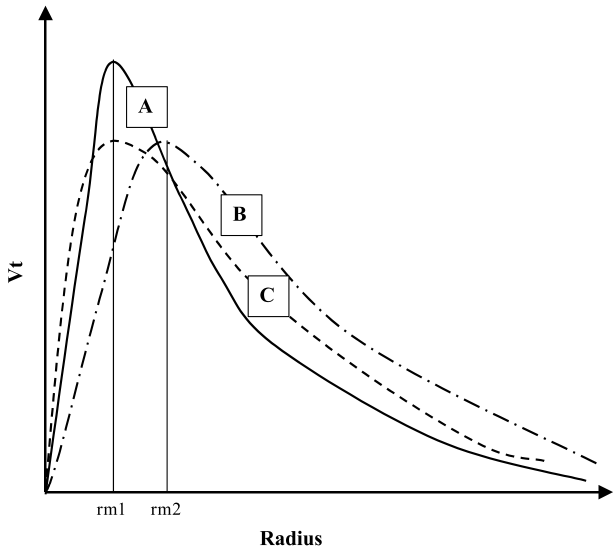

Fig. 2. Schematic diagram showing the variation of radial profiles of typical tropical cyclones

There are multi-scale interactions related to TC development, motion, and structure and intensity changes. While the motion is mostly controlled by the steering flow associated with the large-scale environment, as well as the beta-gyres and the upper-tropospheric negative potential vorticity anomalies (Wu and Emanuel, 1993; Wu and Kurihara, 1996; Wang et al, 1998), the structure and intensity changes are affected at any time by large and complex arrays of physical processes that govern the inner core structure and the interaction between the storm and both the underlying ocean and its atmospheric environment (Wu and Cheng, 1999; Bosart et al, 2000; Hong et al, 2000; Wang, 2002a, b). As a result, understanding and forecasting of TC structure and intensity changes are much more difficult than those of TC track.

It is our attempt in this review article to provide an overview of our current understanding of TC structure and intensity changes and also suggest a number of directions for future studies in this important area. The next section defines the

TC structure and intensity used in this article. This is followed by discussions on the current understanding of the physical processes that are believed to be responsible for TC structure and intensity changes, including the maximum potential intensity (Sect. 3), internal dynamics (Sect. 4), external forcing from environmental flow (Sect. 5), and planetary boundary layer processes (Sect. 6). Several future research directions toward improving the understanding of TC structure and intensity changes are briefly discussed in the last section, along with concluding remarks.

2. Definition of TC structure and intensity

The structure of a TC covers a range of scales from storm scale to mesoscale, including the size of the low-level cyclonic circulation (such as the outermost closed isobar, or the radius of sustained $15\,\mathrm{m\,s^{-1}}$ gale-force winds), the radius of maximum wind (RMW), the eyewall with its symmetric and asymmetric structures, and both the inner and outer spiral rainbands. To facilitate our presentation, we define the TC structure by three components: storm-scale structure, inner-core structure, and spiral rainbands. The storm-scale structure includes the whole cyclonic circulation with sustained tangential winds larger than $15\,\mathrm{m\,s^{-1}}$. The inner-core structure represents the structure within a radius of twice the RMW, including the deep eyewall clouds, the eye of the storm, and convective asymmetries in the eyewall, such as the convectively coupled vortex Rossby waves and spiral rainbands.

Intensity of a TC is usually defined as either the maximum surface wind or the minimum central surface pressure. In order to distinguish the intensity of the TC inner-core from that of the outer circulation, Holland and Merrill (1984) defined strength as the magnitude of the cyclonic circulation in the region outside the inner-core. As shown schematically in Fig. 2, the two tangential wind profiles B and C give two cyclones with similar intensity but different strengths. By comparison, A is greater than B in intensity but with a weaker strength. Intensity and strength changes are rarely correlated. Since the strength change is closely related to the TC size change, we use the maximum surface wind or the minimum central surface pressure as the TC intensity.

The TC structural change is quite complicated, including changes of the RMW and cyclone size; developments of asymmetric structure in the core region, major spiral rainbands, and polygonal and double eyewall structures. These structural parameters can be monitored by advanced observational systems, such as the satellite remote sensors and Doppler radars. However, physical processes responsible for their changes and their effect on intensity change are not well understood. In general, the structure changes in the inner core region are accompanied by or closely correlated to some kind of intensity changes.

Recent studies on the TC intensity tend to focus on two issues. One is concerned with *what factors determine the maximum potential intensity (MPI) that a TC can achieve given the thermodynamic state of the atmosphere and the ocean* (Emanuel, 1986; 1988; 1991; 1995a; Holland, 1997). The statistical analysis of Atlantic TCs by DeMaria and Kaplan (1994) revealed that most storms could only reach 55% of their MPIs and only about 20% reach 80% or more of their MPIs at the time when they are most intense. Therefore, an accompanied issue is *what factors prevent the TCs from reaching their MPIs*. Among many limiting factors, unfavorable environmental conditions, especially the vertical shear-induced asymmetry in the inner core region, have been postulated as the primary impediment to a TC reaching its MPI (Elsberry et al, 1992). We consider this to be a simplistic view and that the structure and intensity changes may be largely controlled by the complex interactions among the internal dynamics, the external forcing from the large-scale environmental flow, and the underlying ocean which may be modified by the cyclone passage.

3. Maximum potential intensity (MPI)

As a step toward understanding the TC intensity and intensity changes, attempts have been made to estimate an upper bound on TC intensity for given atmospheric and oceanic conditions (Kleinschmidt, 1951; Miller, 1958; Malkus and Reihl, 1960; Emanuel, 1986; 1988; 1991; 1995a; 1997; Holland, 1997). A complete review of the MPI theories can be found in Holland (1997) and

Camp and Montgomery (2001). There are two typical MPI theories that are mostly referred in recent years: one proposed by Emanuel (1986; 1988; 1991; 1995a; 1997, hereafter EMPI), and the other proposed by Holland (1997, hereafter HMPI). The methods used to formulate the MPI by Emanuel and Holland are quite different while they lead to similar results, but the MPI in both models is governed by SST and thermodynamic structure of the atmosphere.

There are two versions of the EMPI: EMPI1 (Emanuel, 1986; 1988; 1991) and EMPI2 (Emanuel, 1995a; 1997; Bister and Emanuel, 1998). In a quite similar manner to that proposed by Kleinschmidt (1951), EMPI1 is based on the assumption that the TC behaves like a classic Carnot heat engine in which energy is added at the underlying warm ocean surface and lost in the cool outflow area. An array of assumptions were made to derive a relationship between the central pressure and environmental parameters, including axisymmetric structure, both hydrostatic and gradient balances, slantwise moist-neutral condition, etc. Therefore, the MPI (the minimum central surface pressure) of a TC can be determined entirely by the cyclone radius at which environmental parameters of sea-level pressure, SST, and relative humidity are defined, and the air temperature of the outflow layer in the upper troposphere. Since the available energy for TC in EMPI1 is determined by the maximum entropy difference between the cyclone center and the environment, the EMPI1 is markedly sensitive to the choice of environmental parameters, such as the environmental relative humidity (Holland, 1997; Wu and Lu, 1997; Camp and Montgomery, 2001).

By incorporating eye dynamics, which is closed by assuming a balance between the radial entropy advection and the surface entropy flux together with an assumption of cyclostrophic balance, Emanuel (1995a) modified his original EMPI1. This revised prediction of MPI (EMPI2; still measured by minimum central surface pressure) shows an explicit dependence on the ratio of the exchange coefficient to the drag coefficient at the air sea interface (C_k/C_D) and has a better agreement with the numerical model results (Rotunno and Emanuel, 1987). A by-product of EMPI2 is the MPI defined by the maximum azimuthal surface wind, which is derived based on a balance between the radial entropy advection and the surface entropy flux at the RMW under the eyewall (Emanuel, 1995a; 1997). The EMPI2 was further elaborated (Bister and Emanuel, 1998) by incorporating the effect of dissipative heating in the boundary layer, thus producing somewhat stronger storms, which were later verified in a full physics model by Zhang and Altshuler (1999).

Another well-recognized MPI theory is the thermodynamic model HMPI proposed by Holland (1997), which includes several significant modifications to previous thermodynamic approaches, such as the one discussed by Miller (1958) who related the MPI (minimum central surface pressure) to the maximum achievable temperature anomalies in the TC eye due to subsidence warming by assuming a hydrostatic balance. The HMPI requires an atmospheric sounding, SST, and surface pressure, and includes the oceanic feedback of increasing moist entropy associated with the falling surface pressure over a steady SST. Based on this approach, the surface pressure under the eyewall falls due to the warming from moist-adiabatic ascent in the eyewall convection. This pressure fall leads to an increase in boundary layer equivalent potential temperature (θ_e), allowing for a further warming in the eyewall and the surface pressure drop under the eyewall region. Such a process proceeds until a quasi-steady state is reached. An eye parameterization is then utilized if the net pressure fall under the eyewall is larger than 20 hPa. The θ_e throughout the eye is assumed equal to the surface equivalent potential temperature (with 90% relative humidity) under the eyewall. The temperature anomaly in the eye is obtained by parameterized relative humidity in the eye, which is constrained and adjusted by the cyclone energy budget. The pressure fall at the cyclone center (HMPI) is then obtained by hydrostatic relationship. The HMPI also assumes an axisymmetric structure, similar to the EMPI, and is largely determined by the temperature anomaly in the upper-troposphere, while it is less sensitive to the warming in the mid-lower troposphere.

Both the EMPI and HMPI consider the processes under the eyewall with a passive eye. However, the lower surface pressure in the eye provides a potential source of higher entropy air, which could be entrained into the eyewall, increasing the eyewall entropy and thus the MPI. This possibility was originally proposed by Malkus (1958) and later simulated by Holland (1997) with his MPI model by using the entropy calculated at the cyclone center instead of that under the eyewall. In such an exercise, Holland found a dramatic increase of the MPI (from 883 hPa to 864 hPa based on the January mean sounding at Willis Island). However, Holland found that in this extreme case the frictional dissipation exceeded the buoyant energy in the eyewall and the eyewall θ_e reached an unrealistically high value compared with observations. As discussed by Holland, his calculation was based on a full replacement of the eyewall θ_e, so the effect of entrainment of high eye θ_e might be overestimated. He suggested a plausible scenario that the air rising in the eyewall consists of a combination of high entropy air from inside the eye that has entrained into the eyewall by eddy mixing and mixed with drier, lower entropy air spiraling inward above the cold surface layer that might result from evaporation of sea spray and the convective downdrafts.

This problem is recently revisited by Persing and Montgomery (2003) with the axisymmetric hurricane model of Rotunno and Emanuel (1987). They found that the model MPI increases with the model resolution and is much more intense than the theoretical EMPI at very high resolutions. This model MPI, which they termed as superintensity, is due to the presence of high-entropy air in the low-level eye, which is entrained into the eyewall and represents a source of latent heat for the eyewall convection of the storm, as expected by both Malkus (1958) and Holland (1997). Persing and Montgomery (2003) concluded that the introduction of heat from the eye leads to a modified Carnot cycle that allows for a stronger storm than that originally formulated for EMPI2. In Holland's view, the higher boundary layer θ_e, once distributed aloft, produces a hydrostatic pressure fall and thus higher intensity.

The superintensity concept challenges the MPI theories and may stimulate new development or modifications to the current theories. However, both EMPI and HMPI have proven to be very good at predicting the MPI of real TCs (Tonkin et al, 2000), indicating that this superintensity might be either an artifact of the model used or offset by some other negative feedbacks that have not been included, such as the downdrafts in the inflow boundary layer, and assumptions in the estimation of the surface exchange coefficients. Further research is required to improve our understanding of the MPI and to provide more accurate MPI calculations. Another issue related to the MPI theories is the constraint of the axisymmetric configuration. It is so far not clear whether, how, and to what degree the three-dimensional structure, in particular, the internally generated asymmetric feature in the inner core region of a TC, affects the MPI.

4. Internal dynamics

4.1 Vortex Rossby waves (VRWs)

TCs are localized vortices with elevated cyclonic potential vorticity (PV) concentrated in the inner core region near the RMW with large radial gradients. Any radial perturbation of air parcel would experience restoring force due to the presence of PV gradients, and generate edge PV waves. Because their resemblance to the Rossby waves in the midlatitude large-scale motion, these PV waves are usually termed as vortex Rossby waves (VRWs), i.e., Rossby-type waves in circular vortex. This concept was originally proposed by MacDonald (1968) to explain the spiral rainbands in TCs. However, it was not well recognized until recent years that the Rossby-type waves can exist in the TC core region and may play important roles in TC structure and intensity changes (Guinn and Schubert, 1993; Montgomery and Kallenbach, 1997; Montgomery and Lu, 1997; Kuo, 1999; Reasor et al, 2000).

Building on earlier work, which has identified vortex axisymmetrization as a universal process of smoothly distributed perturbed vortices (Melander et al, 1987; Guinn and Schubert, 1993; Smith and Montgomery, 1995), Montgomery and Kallenbach (1997) developed a theoretical framework on the VRW-guide model of vortex axisymmetrization (see also a later refinement by Brunet and Montgomery, 2002; Montgomery and Brunet, 2002). They showed that axisymmetrization of asymmetric perturbations by the strong shearing tangential flow of the mean vortex is accomplished by outward propagating

VRWs whose restoring mechanism is associated with the radial gradient of storm vorticity. They suggested that the radially outward-propagating VRWs are responsible for initiation of the inner spiral rainbands, and can affect the structure and intensity of the mean vortex by wave-mean flow interaction. Recent studies also show that the interaction of convectively forced VRWs with the mean vortex flow could lead to intensification of the mean vortex (Montgomery and Enagonio, 1998; Enagonio and Montgomery, 2001; Möller and Montgomery, 1999; 2000; Shapiro, 2000).

VRWs in a TC can be forced and driven by, and coupled with, convective asymmetries in the eyewall. Such convective asymmetries in the eyewall can be generated by barotropic or three-dimensional instabilities of the TC vortex (Schubert et al, 1999; Nolan and Montgomery, 2002), the beta effect (Wang and Holland, 1996a, b), or environmental flow and shear (Shapiro, 1983; Wang and Holland, 1996c; Bender, 1997; Frank and Ritchie, 1999). Propagation of these waves along the eyewall can produce an apparent cyclonic rotation of the polygonal eyewall in some TCs (Kuo et al, 1999; Reasor et al, 2000). In an observational study, Reasor et al (2000) examined the low-wavenumber asymmetric structure and evolution of Hurricane Olivia's (1994) inner core. They found that the asymmetry was characterized by a wavenumber two discrete vortex Rossby edge wave, which was suggested to be the result of barotropic instability of the symmetric vorticity structure. Propagation of the wavenumber two VRWs along the eyewall offered a physical explanation to the storm eye rotation. Trailing vorticity bands with radial wavelength of 5–10 km were observed and ascribed to axisymmetrizing VRWs, which contribute to the observed rainbands with similar radial wavelength.

Based on a full physics model, Wang (2001; 2002a, b) identified and studied the VRWs in a simulated TC and found that VRWs with low azimuthal wave numbers dominate the asymmetric structure in the eyewall and play an important role in mixing angular momentum and PV between the eye and the eyewall where both are at a maximum (Fig. 3). These waves have their maximum amplitudes near the RMW, generate a spiral structure in PV fields, tilt outward with height, and have a coherent structure in the mid-lower troposphere where the radial PV gradients are large. A salient feature of these waves is their quasi-balanced nature with strong divergent flow, consistent with the asymmetric balance dynamics of TCs as first proposed by Shapiro and Montgomery (1993) and later clarified by Montgomery and Franklin (1998). Another distinct nature of the VRWs is their coupling with eyewall convection. Eyewall convection is enhanced in the region where low-level inflow is associated with the VRWs, while suppressed in the region with outflow. The VRWs propagate upwind around the eyewall with a period of about twice the period of a parcel advected by the local tangential flow near the RMW of the azimuthal mean cyclone. This is consistent with the corresponding ratio found in Typhoon Herb (1996) by Kuo et al (1999), and hurricane Olivia (1994) by Reasor et al (2000). In addition to the azimuthal propagation, VRWs also propagate outward. Wang (2002b) showed that the outward propagation of the VRWs could initiate inner spiral rainbands as proposed by Montgomery and Kallenbach (1997) and observed in hurricane Olivia (Reasor et al, 2000).

4.2 Spiral rainbands

Spiral rainbands are a unique feature of TCs and may produce severe rainfall outside the eyewall and play an important role in TC structure and intensity changes. Most of the earlier theories of TC rainbands were associated with inertia-gravity waves (Diercks and Anthes, 1976; Kurihara, 1976; Willoughby, 1978) with one exception proposed by MacDonald (1968) who hypothesized that hurricane rainbands are Rossby-type waves. The latter view was clarified theoretically by Montgomery and Kallenbach (1997), who related the VRWs to the axisymmetrization of any asymmetries in the eyewall. The axisymmetrization process is usually accompanied by emanating vorticity filaments that are part of VRWs. Reasor et al (2000) found evidence of a close relationship between the vorticity filaments and inner spiral rainbands in Hurricane Olivia (1998). It is now generally agreed that the inner spiral rainbands are closely related to convectively coupled VRWs (Chen and Yau, 2001; Wang, 2001; 2002b). However, the outer rainbands (outside a radius of about 80 km from the TC center) could

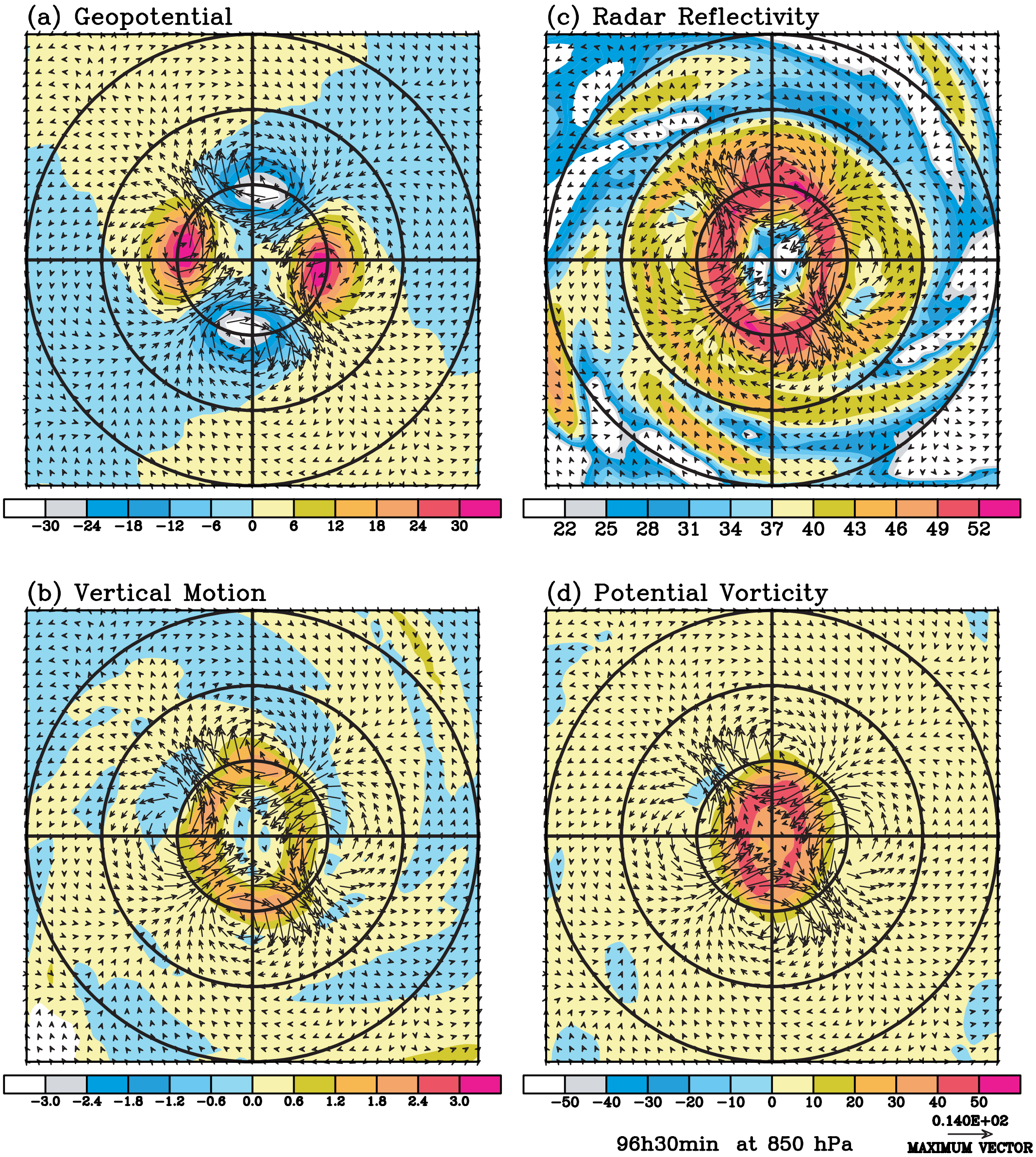

Fig. 3. Shaded in color are (**a**) asymmetric geopotential height ($m^2\,s^{-2}$), (**b**) the total vertical motion ($m\,s^{-1}$), (**c**) estimated radar reflectivity (dBZ), and (**d**) the potential vorticity (PVU), with the asymmetric winds ($m\,s^{-1}$) relative to the moving storm superposed in each panel, at 850 hPa after 168 h of simulation on an *f*-plane (18° N) in a quiescent environment simulated in TCM3 described in Wang (2001; 2002c). The domain shown in each panel is 180 km by 180 km. The circles are placed at radii of 30, 60, and 90 km from the storm center. The figure shows the simulated inner and outer spiral rainbands, elliptic eyewall structure, the convectively coupled vortex Rossby waves in the eyewall, and their relationship with asymmetries in eyewall convection and PV. The maximum wind vector is 24 $m\,s^{-1}$

not be explained by VRWs where the radial PV gradients become too weak. Chow et al (2002) show that fluctuation of the PV distribution in the TC core region can act as a source, generating gravity waves that produce banded structures and the moving spiral rainbands. Such fluctuation or rotation of inner core PV distribution can result from the activity of VRWs in the TC eyewall (Kuo, 1999; Wang, 2002a, b).

In his numerical simulation, Wang (2001; 2002b) showed that the outer spiral rainbands most frequently form and develop between 80 km and 150 km from the TC center. There are several possible explanations for this preference.

One explanation lies in the effect of downdrafts from anvil clouds in the outflow layer (Willoughby et al, 1984). The location of outer spiral rainbands is thus mainly determined by the outflow radial wind speed and the terminal velocity of the ice species (snow and graupel) in the upper part of the eyewall. This may occur at a radius of about 100 km for strong TCs, consistent with the model results. In particular (Wang, 2002c), when the melting of snow and graupel and the evaporation of rain are excluded, the simulated TC is much stronger and the outer spiral rainbands almost disappear, indicating the importance of downdrafts to the generation of outer spiral rainbands and to limiting the TC intensity. Another mechanism is that proposed by Montgomery and Kallenbach (1997), who suggested that the outer spiral rainbands could form near the stagnation radius of the VRWs at which the outward group velocity of the VRWs vanishes. The stagnation radius in the simulated TC is at about 80 km from the TC center (Wang, 2002b), which is also coincident with the active outer spiral rainbands in the simulation. Since both mechanisms coexist, further studies are needed to assess which mechanism is dominant in generating the outer spiral rainbands.

TC rainbands are largely made up of extensive regions of stratiform rain with embedded convection showing various degrees of organization (e.g., Barnes et al, 1983; May, 1996). Many of these rainbands have middle level jets (May et al, 1996; Samsury and Zipser, 1995). May (1996) suggested that these jets could result from PV production by the vertical gradient of heating in the stratiform rain regions. A potential positive role of rainbands in the intensification of storms has been proposed by May and Holland (1999), who argued that PV generated in the stratiform rainbands could be transported into the eyewall region and enhance the core intensity. This view is in contrast to the generally accepted adverse effects due to the generation of downdrafts and the blocking of inflow in the boundary layer (Barnes et al, 1983; Powell, 1990a, b; Wang, 2002c) and is likely secondary in our opinion. Another possible effect may be latent heat released in the rainbands, which may reduce the radial pressure gradients by hydrostatic adjustment, and thus weaken the TC (Bister, 2001).

4.3 Mesoscale vortices

As discussed by Raymond and Jiang (1990), vertical gradient of diabatic heating due to latent heating in the stratiform cloud and evaporational cooling below the cloud base can concentrate cyclonic PV at mid-troposphere. This may result in formation of mid-tropospheric mesoscale vortices (Houze, 1977; Zipser, 1977). Mesoscale vortices generated in such a way are frequently observed within the TC circulation, especially in the outer spiral rainbands (e.g., Holland and Lander, 1993; Simpson et al, 1997). The interaction between these vortices and the TC core can lead to meandering nature of TC tracks with periods of several days and amplitudes of around 100 km (Holland and Lander, 1993). Willoughby (1994) also showed that convective cells in the eyewall could produce trochoidal motion of TCs.

Simpson et al (1997) proposed that in the tropical stratiform clouds when more than one vortex forms, they interact in a complex and stochastic manner with each other and with the monsoon environment, initiating cyclogenesis. This mesoscale TC genesis process was investigated by Montgomery and Enagonio (1998) in the context of TC genesis from convectively forced VRWs in a three-dimensional quasi-geostrophic model. In this theory, the mesoscale vortices can be convectively generated in the eyewall region and then sheared and damped by the axisymmetrization processes, which are followed by intensification of the weak parent storm-scale vortex. However, mesovortices some distance outside the eyewall behave as the outer spiral rainbands, and thus weaken the storm, prohibiting the intensification.

Mesovortices are frequently observed in the eyewall of strong TCs (e.g., Black and Marks, 1991; Willoughby and Black, 1996; Kossin et al, 2002; Knaff et al, 2003). These eyewall mesovortices have horizontal scales smaller than the diameter of the eye yet larger than the individual convective clouds that constitute the eyewall. The mesovortices in the TC eyewall are recently simulated through laboratory experiments by Montgomery et al (2002). The results show many similarities to the observed mesovortices in real TCs. The eyewall mesovortices can mix PV and angular momentum between the eye and the eyewall and thus are believed to contribute to the inner-core structure (Willoughby, 1998; Zhang et al, 2002). However, it is unclear so far whether the eyewall mesovortices play a significant positive or negative role in determining the intensification rate and the MPI of a TC.

The eyewall mesovortices can originate from barotropic shear instability. Schubert et al (1999) studied the barotropic instability of the symmetric circulation of a TC-like vortex and found that the unstable development of the asymmetries in the eyewall could cause the eyewall breakdown accompanied by formation of mesovortices, leading to the appearance of polygonal eyewalls. Kossin and Schubert (2001) found a strong relationship between the vorticity mixing and the rapid pressure fall in TC-like vortices using a nondivergent barotropic model. They initialized the model with thin annular rings of enhanced vorticity embedded in a nearly irrotational flow. This initial annular vorticity ring is highly unstable, and thus breaks down rapidly into a number of mesovortices in their model. Associated with each mesovortex is a local pressure perturbation, or mesolow. In cases where the mesovortices merge to form a monopole, dramatic central pressure falls but the maximum tangential wind decreases. Their findings only explain some dry dynamics since the deepening in central pressure in a real TC is generally accompanied by an increase in maximum tangential wind (Wang, 2002b). In a simple TC model with parameterized convective heating, Wang and Holland (1995) showed that at the time two TC-like vortices merged, the binary system might experience a rapid intensification. This was found to occur only when the diabatic heating was included. Studies with full physics models can provide further insight into the role of embedded mesovortices in TC structure and intensity changes. In addition, some mesovortices in the TC may be related to activities of tornadoes and the extreme damage distribution, in particular, in landfalling TCs (Willoughby and Black, 1996).

4.4 Eyewall processes

The eyewall of a TC is a ring with deep convection near the RMW. The TC gets most of its energy transferred from the ocean surface and released as latent heat in moist convection in the eyewall. The strongest winds and torrential rains in a TC occur mainly in the eyewall. The TC eyewall can be highly asymmetric and is usually characterized by polygonal structure, embedded mesovortices, and VRWs (Lewis and Hawkins, 1982; Black and Marks, 1991; Kuo et al, 1999). Why the asymmetries can develop within the highly rotating core of a TC has been a mystery for about half a century. Lewis and Hawkins (1982) explained the polygonal eyewalls as the internal gravity wave interference patterns due to superposition of differing wavenumbers and periods. In recent years, more evidences show the possible role of VRWs in producing the polygonal eyewall structure (Schubert et al, 1999; Kuo et al, 1999; Kossin and Schubert, 2001; Wang, 2002b). Both Kuo et al (1999) and Wang (2002b) found that the elliptic eyewall could result from the azimuthally propagating wavenumber-two VRWs around the eyewall.

The most remarkable eyewall process related to TC intensity change is the one originally proposed by Willoughby et al (1982), who documented the concentric double eyewalls and eyewall replacement processes. In this concentric eyewall model, as a TC and its primary eyewall intensifies, convection outside the primary eyewall becomes organized into a ring (usually named second eyewall) that encircles the inner eyewall and coincides with local tangential wind maximum. As the second eyewall propagates inward and amplifies, the inner eyewall weakens and is eventually replaced by the outer eyewall. This eyewall replacement (eyewall cycle) is usually accompanied by a weakening and then an intensification of the TC; and therefore is believed to be responsible for large fluctuations in TC intensity.

Another eyewall process is the so-called partial eyewall replacement proposed and studied numerically by Wang (2002b). He showed that strong perturbation from an outer spiral rainband could amplify the VRWs in the eyewall, causing a large distortion of the eyewall and partial eyewall breakdown accompanied by a weakening of the TC. The eyewall can later be recovered from breakdown via axisymmetrization process, resulting in a reintensification of the storm. Therefore the eyewall breakdown/recovery is accompanied by a weakening/intensifying cycle of the TC. This partial eyewall cycle is an asymmetric process, and hence is different from the symmetric double eyewall phenomenon studied by Willoughby et al (1982). The latter is only possible for highly symmetric, intense storms, while the former can happen to both highly symmetric and asymmetric, and both intense and weak, storms.

5. Effects of environmental flow

In addition to the internal dynamics, external forcing from large-scale environmental flow is also a key factor affecting TC structure and intensity changes (Wu and Cheng, 1999). In this section, we will discuss progress in understanding the effects of large-scale environmental flow and its vertical shear, and interaction with upper tropospheric trough on TC structure and intensity changes.

5.1 Uniform environmental flow

A uniform environmental flow can introduce asymmetric surface heat and moisture fluxes and friction to a TC, generating the asymmetric structure in the inner core region (Shapiro, 1983; Kepert, 2000; Kepert and Wang, 2001). In their numerical experiments, Peng et al (1999) found a reduction of the intensity for a TC embedded in a uniform environmental flow. They showed that the intensity of a TC is inversely proportional to the magnitude of the wavenumber-one asymmetry. The relative phasing of asymmetric surface fluxes and moisture convergence in the boundary layer was found to be important. Peng et al (1999) indicated that it was due to an out-of-phase relation between the two asymmetries that the TC intensity was reduced. Dengler and Keyser (1999) found that, in their three-layer model, it was the penetration of stable dry air from mid-levels to the regions of boundary-layer convergence that reduced the intensity of the storm embedded in uniform environmental flows.

The effect of the environmental flow can also be inferred from the relationship between the translation speed of the TC and its intensity. Holland (2002, personal communication) provides an initial assessment, showing that the intense TCs and their translation speed are highly correlated. An example is given in Fig. 4, showing the relationship between translation speed and intensity of TCs in the Australian region during 1958–1998. We see that the very intense TCs can only develop under a narrow range of translation speeds between 3 and 6 m s^{-1} and there is a clear trend towards weaker cyclones as the speed of movement increases. The explanation is that if TCs are moving too slow, oceanic cooling induced by the vertical mixing in the upper ocean caused by turbulence generated by surface wind stress curl under the TC will disrupt the intensification, while if they move too fast the resulting asymmetric structure will also inhibit intensification as shown by Peng et al (1999). Similar analyses for TCs in other basins are required to examine the robustness of this relationship and to isolate storms affected primary by quasi-uniform environmental flow from those affected by vertical wind shear or moving to cold SST-region.

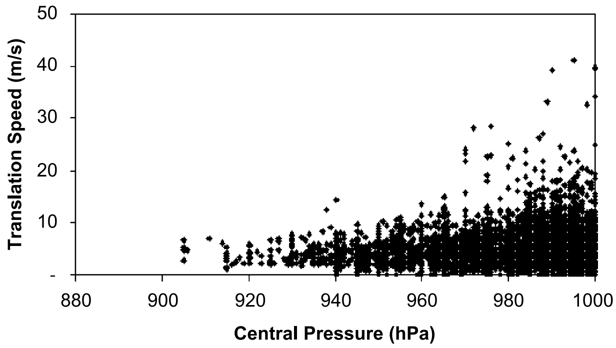

Fig. 4. Relationship between translation speed and intensity of tropical cyclones in the Australian region from 1958 to 1998 (provided by G. J. Holland, 2002)

5.2 Vertical shear of horizontal wind

It has been shown in earlier observational studies (e.g., Gray, 1968; Merrill, 1988) that the vertical shear of horizontal wind has a negative influence on the intensification of TCs. A common explanation of the effect of vertical wind shear is that the heat of condensation released at upper levels is advected in a different direction relative to the low-level cyclonic circulation and therefore the ‘‘ventilation’’ of heat away from the circulation inhibits the development of the storm (Gray, 1968). In addition to the idea of ‘‘ventilation,’’ an astute explanation concerning the effect of vertical wind shear on TC intensity was proposed by DeMaria (1996). A simple two-layer model was used to show that the tilt of the upper and lower level PV produces a mid-level temperature increase near the vortex center. This mid-level warming was hypothesized to reduce the convective activity and thus inhibit storm development. The resistance of a vortex to vertical wind shear is a function of the Rossby penetration depth, which increases with the latitude, horizontal scale, and vortex amplitude (Jones, 1995). In the regression analyses, DeMaria (1996) demonstrated that high-latitude, large, and intense TCs tend to be less sensitive to the effect of vertical shear than low-latitude, small, and weak storms.

The vertical shear-induced convective asymmetry in the inner core region is considered to be negative to TC intensification (Elsberry et al, 1992). Wang and Holland (1996c) and Bender (1997) attributed the development of convective asymmetries in the inner core region of a TC embedded in a vertically sheared environmental flow to the relative flow across the elevated cyclonic relative vorticity core. They showed that both upward motion and convection were enhanced to the downshear left of the TC center but suppressed to the upshear right when facing down shear. This distribution of asymmetry in eyewall convection due to vertical shear has also been found later by Frank and Ritchie (1999; 2001) using more sophisticated high-resolution model and by Corbosiero and Molinari (2002; 2003) and Black et al (2002) from observations (Fig. 5). Frank and Ritchie (2001) attributed the weakening of the TC in their simulation to the outward mixing of high values of PV and equivalent potential temperature by the vertical shear-induced asymmetry in the upper troposphere, resulting in a loss of the warm core at upper levels and weakening of the storm. This mechanism does not need evident vertical tilt of the TC as it was found in a low shear of 5 m s^{-1} in which a mature TC weakened after 3 days. Since observations show that TC can sustain vertical shear as large as 12 m s^{-1} (Zehr, 1992), it is suspected that Frank and Ritchie's simulations might be affected by the lateral boundaries of the fixed finest mesh configuration since when the TC moved toward the lateral boundary outer rainbands might be generated ahead of the storm center due to the mesh-interface discontinuity.

A fundamental frequently asked question is how a TC-like vortex can sustain a coherent vertical structure in vertical shear. Wang and Holland (1996c) found in a numerical study that the cyclonic portion of the TC could remain upright in a moderate vertical shear. The TC core underwent successive downshear tilting during the first 24-h while realigning over a 72-h period. A quasi-steady tilt to the downshear left was found even in the case where the diabatic heating was not considered (also see Wu and Wang, 2001a). They explained this steady-state tilt configuration as a result of complex interactions among the upper-level anticyclonic PV anomalies, the cyclonic portion of the vortex and the external vertical shear. Wu and Wang (2001b) found that the asymmetric diabatic heating due to vertical shear tends to offset the vertical tilt through a fast adjustment between the asymmetric flow and asymmetric diabatic heating. To understand the vertical alignment of a titled

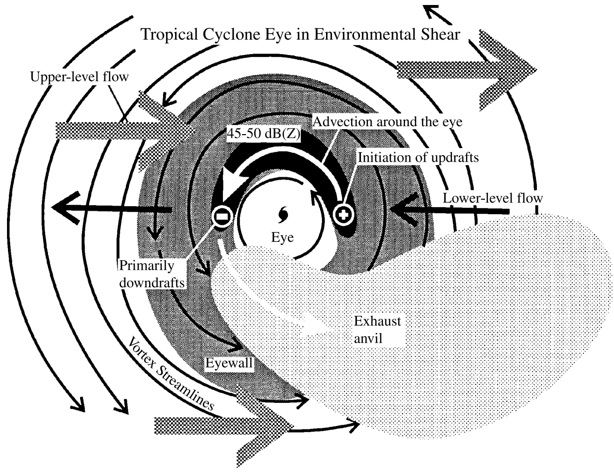

Fig. 5. Schematic illustration of the shear-induced convective asymmetry based upon observations of hurricanes Jimena and Olivia. The low-level environmental flow is indicated by the two solid black arrows. Upper-level flow is indicated by the three stippled arrows. Convective cells form somewhat upwind of the downshear side of the eyewall. They advect around the eye into the semicircle to the left of the shear vector where warm rain processes generate hydrometeors large enough to reflect radar effectively. Precipitation-driven downdrafts begin about 90° to the left of the shear vector. By the time the cells reach the upshear side of the eyewall they have ascended through the 0 °C isotherm and downdrafts predominate below 6 km. As the cells move into the semicircle on the right of the shear vector, most condensate freezes or falls out of the active updrafts. The unloaded updrafts accelerate upward. They detach themselves from the eyewall and approach the tropopause as they rotate through the semicircle to the right of the shear. (From Black et al, 2002)

TC-like vortex, Reasor and Montgomery (2001) developed a new theory, which separates the mean vortex evolution from the evolution of the tilt asymmetry. From this new perspective, the vertically averaged azimuthal mean component of a tilted vortex is defined as the mean, while the departure from this mean as the tilt perturbation. The subsequent evolution of the tilt perturbation was captured by a linear, dry VRW mechanism. They found that the continuum modes in the dynamical system destructively interact with the VRW, leading the decay of the VRW and hence the vortex tilt. Schecter et al (2002) viewed the vertical alignment as a result of the damping of the VRW due to its interaction with the mean vortex circulation. Reasor et al (2003) further show that the VRW damping mechanism provides a direct means of eradicating the tilt of intense TC-like vortices in unidirectional vertical shear, and that intense TC-like vortices are much more resilient to vertical shear than previously believed. Therefore, the VRW damping mechanism intrinsic to the dry adiabatic dynamics of the TC vortex may play a crucial role in maintaining the vertically coherent structure of TCs in moderate vertical shear. However, how the diabatic heating including both the symmetric and asymmetric components affects this dry dynamics is still unknown and serves as an important topic of future study. In addition, since so far idealized simulations have focused on unidirectional shear, it is not clear how a non-unidirectional shear affects the TC structure and intensity.

5.3 Upper-tropospheric trough interaction

Since the inertial stability in a TC is lower in the outflow layer than that in the middle and lower layers, the influence of upper environmental forcing may extend to the inner parts of a TC more easily. Holland and Merrill (1984) showed that the radial-vertical circulation induced by the upper-level momentum forcing has a direct and substantial effect on the core region of a TC. They also proposed that the cooperative interaction between a TC and a passing trough could enhance the outflow jet of a TC, invigorating core region convection and initiating the deepening of the cyclone. Based on some numerical experiments, Challa and Pfeffer (1980), and Pfeffer and Challa (1981) showed that a sufficiently intense and properly organized eddy flux convergence of momentum (EFC) is an essential ingredient for the development of Atlantic tropical disturbances into hurricanes.

DeMaria et al (1993) investigated the relationship between EFC and TC intensity for the named TCs during the 1989–1991 Atlantic hurricane seasons. They showed that about one third of the storms with enhanced EFC intensified just after the period of enhanced EFC. Most of the storms that did not intensify after the enhanced EFC usually experienced increasing vertical shears, moved over cool water, or became extratropical. Their results indicated that although the EFC may be positive but the eddy-related vertical wind shear could be negative to TC intensification.

Molinari and Vollaro (1989; 1990) found that upper-level EFC occurred during the intensifying period of hurricane Elena (1985), indicating that the upper-tropospheric trough with a positive PV anomaly played an important role in the reintensification of Elena. Their balanced vortex solution showed an in-up-out circulation that shifted inward with time, which is consistent with the inward propagation of the maximum outflow. Molinari et al (1995) proposed that the interaction between the outflow from Elena (1985) and an upper-tropospheric trough reduced the penetration depth of the upper westerlies and the length of time during which the vertical wind shear occurred, thus preventing the destruction of Elena by the shear.

Wu and Cheng (1999) found little evidence that the intensification of typhoons Flo and Gene was directly associated with the superposition of a positive PV anomaly as described by Molinari et al (1995). Persing and Montgomery (2002) also showed that the upper tropospheric trough played a minimal role in intensification of hurricane Opal (1995) based on a diagnosis from the GFDL hurricane model output. Hanley et al (2001) examined 121 Atlantic TCs in an attempt to differentiate between troughs which lead to intensification (good trough), and those which lead to decay (bad trough). They found that in the Atlantic basin a larger and stronger upper PV anomaly (bad trough) induces more vertical shear than a small scale PV anomaly (good trough), and has a negative impact on

TC intensity. Therefore, the ''bad trough'' or ''good trough'' for TC intensity is still under debate, and is likely to be case-dependent. In the interaction between the TC and upper trough, the horizontal scale, depth, and amplitude of the PV anomaly associated with the upper trough, and the size and intensity of the TC itself are important for the resultant structure and intensity changes. The processes may involve how the TC core responds to the forcing from the upper PV anomaly.

6. Boundary layer processes

6.1 Surface fluxes

Surface fluxes of sensible and latent heat play a vital role in the development and maintenance of TCs. However, available observations show that the exchange coefficient for sensible and latent heat fluxes is almost independent of wind speed up to $25\,\mathrm{m\,s^{-1}}$ (Liu et al, 1979; Smith, 1988), which corresponds to weak tropical storms, while the drag coefficient strongly depends on wind speed up to $32\,\mathrm{m\,s^{-1}}$ (Smith, 1988). We have known little about the exchange and drag coefficients at high wind conditions. Previous studies showed a strong dependence of the maximum tangential winds of TCs on the surface exchange coefficients for heat and momentum (Malkus and Riehl, 1960; Ooyama, 1969; Rosenthal, 1971). Emanuel (1995a) showed the dependence of the MPI on $(C_k/C_D)^{1/2}$ under the eyewall, where C_k is the exchange coefficient for enthalpy and C_D the drag coefficient for momentum. Emanuel proposed that the ratio of enthalpy to momentum exchange coefficients in real hurricanes must be larger than three fourths; otherwise, the wind speeds would be much lower than observed. This, however, has been challenged by Persing and Montgomery (2003) by using the same numerical model with higher resolutions as used by Emanuel (1995a). Results of Wang (2001; 2002c) also show that realistic TC intensity can be achieved in high-resolution models with the ratio of enthalpy to momentum exchange coefficients less than one half.

The exchange coefficients under TC conditions could be affected significantly by ocean breaking waves and sea spray. Although there have been some efforts contributing to improving the parameterization of these effects, little significant progress has been made so far because the great difficulty in measurements on extremely high wind conditions. Recently based on scaling arguments Emanuel (2003) proposed that in the limit of very high wind speed, the air-sea transition layer becomes self-similar, permitting deductions of air-sea exchange. He hypothesized that exchange coefficients based on the gradient wind speed should become independent of wind speed in the high wind limit. However, it is not clear at what high wind speed the exchange coefficients should become independent of wind speed.

The experimental deductions of the drag coefficient suggest that its wind dependence is reduced at 10-m height wind speed of about $25\,\mathrm{m\,s^{-1}}$, characteristic of tropical storms (Alamaro et al, 2002). This reduction tendency has recently been verified by Powell et al (2003) based on the Global Positioning System dropwindsonde, but with a transition occurring at about $40\,\mathrm{m\,s^{-1}}$ instead of $25\,\mathrm{m\,s^{-1}}$. A possible physical explanation for the transition is the development of a sea foam layer at the air-sea interface. As surface winds increase above $40\,\mathrm{m\,s^{-1}}$, the sea becomes completely covered by a layer of foam and it is difficult to discern individual wave-breaking elements from spray and rain. In such extreme cases, the foam layer impedes the transfer of momentum from the wind to the ocean. As a result increased foam coverage could progressively form a ''slip'' surface at the air-sea interface as winds increase above $40\,\mathrm{m\,s^{-1}}$, leading to a weak increase of friction velocity and a decrease of drag coefficient with wind speed (Fig. 6). This finding is significant in advancing our understanding of the air-sea interaction at high wind regime and can reduce the uncertainties in the calculation of surface fluxes, and thus improving TC intensity forecast, in particular, by numerical weather prediction models. We call for a prompt development of a new parameterization scheme for the drag coefficient based on the new measurements and an evaluation of its effect on TC intensity.

6.2 Ocean mixing

The intensity of TCs is quite sensitive to SST, which mainly determines the surface energy

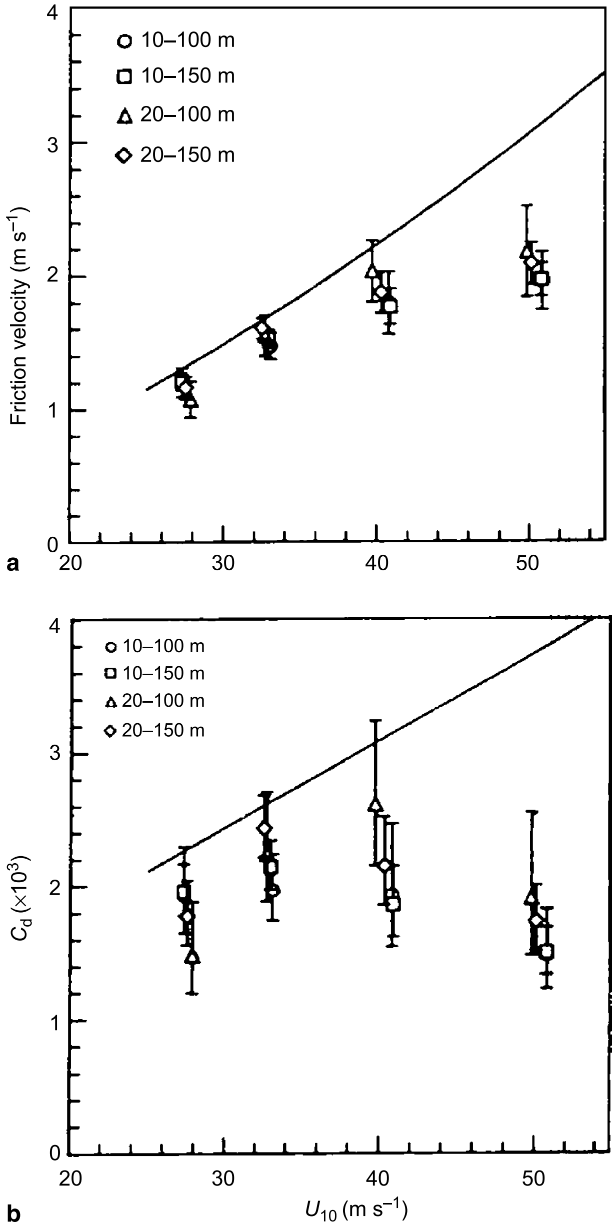

Fig. 6. Surface friction velocity (**a**), and drag coefficient (**b**) as a function of 10-m height wind speed. Vertical bars represent the range of estimates based on 95% confidence limits and symbols at the upper-left of each panel show estimates based on the depth of the surface layer used to fit. The solid curves are corresponding values based on the traditional extrapolation from the relationship given by Large and Pond (1981), reproduced from Powell et al (2003)

transfer from the ocean to the atmosphere. However, strong surface wind stress under a TC can generate strong turbulence in the upper ocean, deepening the ocean mixed layer. This causes significant decreases in SST due to entrainment of cooler water from the thermocline into the mixed layer. The observed SST cold anomalies under TC core vary from 1 °C to 6 °C (Sanford et al, 1987; Shay et al, 1992; Lin et al, 2003). The reduction of SST can result in the reduced enthalpy flux from the ocean to the atmosphere, leading to a decrease in TC intensity. This is a major negative feedback mechanism in limiting the TC intensity. Although the numerical atmospheric models coupled with ocean mixed layer models have been shown to be able to simulate this negative feedback reasonably well (Sutyrin and Khain, 1984; Bender et al, 1993), the simulated TC intensity is sensitive to the vertical structure of the ocean (Bao et al, 2000; Cubukcu et al, 2000; Shay et al, 2000; Chan et al, 2001).

Bender and Ginis (2000) showed overall improvement of TC intensity prediction using a high-resolution coupled model. It is no doubt that future prediction of TC should include the coupling with the ocean as well as ocean waves. In that case, however, the initialization and data assimilation of the ocean will become critical. Although the advanced satellite measurements provide unprecedented datasets, the ocean status and atmospheric variables under the eyewall of a TC, while critical to TC structure and intensity changes, measured by the satellites are largely distorted by the heavy precipitation and thus need correction and improvement.

6.3 Sea spray

Strong winds in severe weather systems, such as TCs, over the open ocean can generate substantial amounts of sea spray into the lowest few meters of the atmosphere by both bursting air bubbles in whitecaps and whipping spume from the tips of waves. Speculation as to the possible role of sea spray in enhancing air-sea fluxes has a long history dating back at least to Ling and Kao (1976). Interest in sea spray evaporation in TCs can be traced back to Anthes (1982), who proposed that evaporation of sea spray would produce a strong cooling in the layer of air within a few meters or so of the ocean's surface, and thus would enhance the sensible heat transfer and strengthen the storm.

Lighthill et al (1994), and Henderson-Sellers et al (1998) suggested that evaporation of sea spray would provide a self-limiting process for

TC intensity. They argued that the energy input per unit mass of air would be reduced if air at the eyewall base is well below the underlying sea surface temperature due to the evaporative cooling by sea spray. On the other hand, Emanuel (1995b) argued that since the evaporation of sea spray does not significantly affect the enthalpy flux at the sea surface and scarcely changes the thermodynamic efficiency of the energy input, it would not limit or affect the MPI of a TC. Recent studies with numerical models show considerable sensitivity of the simulated TC intensity to the details of sea spray parameterization (Bao et al, 2000; Andreas and Emanuel, 2001; Wang et al, 2001). Although the results are diverse to a considerable degree, these modeling results all show a possible increase of TC intensity due to the presence of sea spray, indicating that sea spray at least may not be a limiting factor to TC MPI, as suggested previously by Lighthill et al (1994) and Henderson-Sellers et al (1998). However, we notice that the lack of observations at extreme wind conditions prohibits the accurate praramerization of sea spray. In particular, recent finding by Powell et al (2003) suggests an increasing coverage of the sea by a layer of foam as wind speed increases above 40 m s^{-1}. In that case, the sea spray and rain will coexist and become hard to be distinguished. Therefore, the uncertainty in parameterizing the effect of sea spray will remain, especially under strong TC conditions. This may limit the predictability of TC intensity by numerical models.

6.4 Land surface processes and topographic interaction

After making landfall, a TC usually weakens due to reduction of entropy flux and increase in surface friction (Powell, 1982; 1987; Tuleya et al, 1984; Tuleya, 1994; Powell and Houston, 1998; Farfán and Zehnder, 2001). However, a landfalling TC can produce large amount rainfall and result in accumulated surface water over land. Shen et al (2002) studied the effect of surface water over land on the decay of landfalling hurricanes with the GFDL hurricane model. They found that the local surface cooling near the hurricane core with the largest cooling occurred behind and on the right side of the hurricane center. They showed that a layer of half-meter water can noticeably reduce landfall decay due to large entropy flux from the water surface although the local surface temperature around the hurricane core region is more than 4 °C lower than in its environment. However, the whole coverage of land surface by water under the TC circulation seems impossible and thus the effect discussed by Shen et al (2002) might be overestimated.

Cubukcu et al (2000) found that land–sea contrast could reduce the intensification of the tropical disturbance even when the disturbance is far offshore. They showed that the effect of drier air coming off the landmasses surrounding the Gulf of Mexico can penetrate into the storm core region in a day, reducing the moisture supply and thereby decreasing the intensification rate of the storm. The results from Tuleya (1994), however, showed little remote effect for landfalling TCs. Therefore it remains unclear whether the dry intrusion is a general feature for TCs offshore and for landfalling TCs since the theoretical MPI as discussed in Sect. 3 is determined primarily by the ocean and atmospheric states under the eyewall.

In some geographical region, landfalling TCs experience strong interaction with topography. Understanding the complex interaction is important both scientifically and practically. A typical TC-topographic interaction occurs in Taiwan area where the Central Mountain Range (CMR) runs nearly 400 km north–south over Taiwan Island with about 100 km in east–west width (Wu and Kuo, 1999). Previous studies have shown that while approaching the island of Taiwan, a TC may exhibit dramatic motion, structure and intensity changes, and the enhancement of rainfall in some mountainous areas depending on a number of factors, such as the intensity, size, and translation speed of the TC itself, the landfalling latitude relative to the CMR, and the large-scale weather conditions in the surrounding area (Yeh and Elsberry, 1993a, b; Wu and Kuo, 1999; Wu, 2001; Wu et al, 2002; Lin et al, 2002). Although the complex TC-topographic interaction has been extensively studied both from observations and numerical simulations, detailed studies to improve our understanding of the interaction among the TC eyewall processes, cloud microphysics, and the mesoscale topography

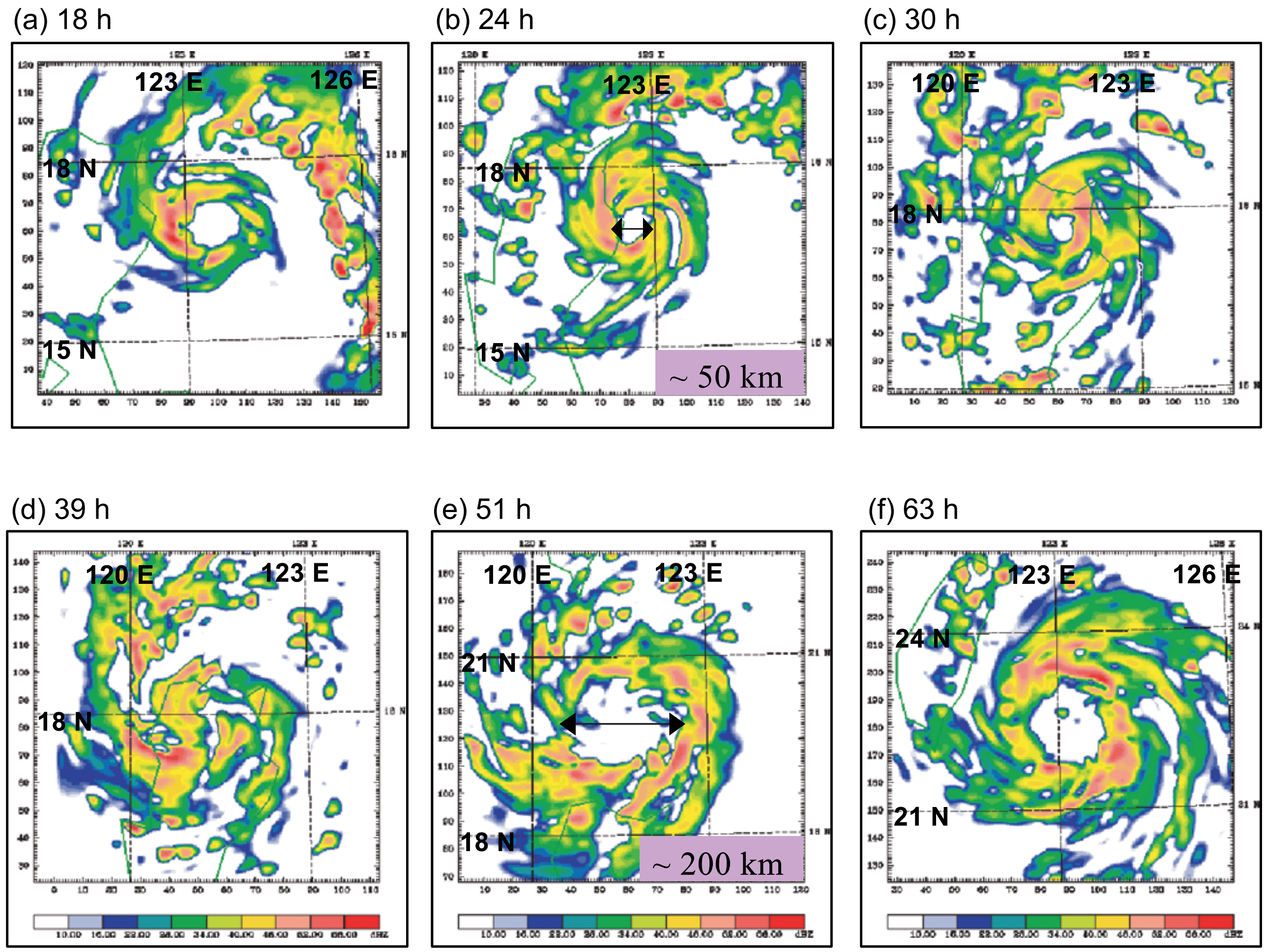

Fig. 7. Model simulated radar reflectivity (dBZ) at 700 hPa for typhoon Zeb (1998) before, during and after its landfall at Luzon: (**a**) 18 h; (**b**) 24 h; (**c**) 30 h; (**d**) 39 h; (**e**) 51 h; and (**f**) 63 h. Showing the evolution of the eyewall of typhoon Zeb, including the eyewall contraction just before landfall (**a**)–(**c**), the following breakdown after landfall (**d**), and then the reformation of the eyewall after it left Luzon and reentered the ocean (**e**), (**f**)

are quite few, even though such studies offer insights into the dynamics of TC-topography interaction and improvements of numerical weather prediction models.

Recently, Wu et al (2003) studied the eyewall evolution of Typhoon Zeb (1998) and found that the eyewall evolution comprised an eyewall contraction just before landfall, a following breakdown after landfall at Luzon, and then a reformation of the eyewall after the storm left Luzon and reentered the ocean. They showed that high-resolution model could reproduce such an eyewall evolution (Fig. 7) but the physical processes responsible for this eyewall evolution are left for follow-up investigation. Since this eyewall structure change was related to intensity change of the storm, it is believed that a detailed study of such special eyewall evolution processes may improve our understanding of TC structure and intensity changes, especially for those TCs making landfall and interacting with mesoscale terrains.

7. Conclusions

We have provided a review of our current understanding of TC structure and intensity changes. We have identified two issues that are mostly concerned by TC research community. One is what factors determine the MPI that a TC can achieve given the thermodynamic state of the atmosphere and the ocean. The other is what factors prevent the TC from reaching their MPIs. It is generally accepted that there exists an upper bound (MPI) on TC intensity given the thermodynamic state of the atmosphere and the ocean, although the theoretical MPI may include some uncertainties and need further improvements. Observational studies have shown that the maximum intensities reached by real TCs in all ocean basins are generally less than that inferred from the theoretical MPI. This indicates the existence of potential limiting factors that prohibit TCs from intensifying and reaching their MPI. These factors include the internal dynamics, external forcing from large-scale environmental flow, and the oceanic upwelling.

Significant progress has been made in the last decade or so in the studies of TC structure and intensity changes, especially those processes in the TC core region. The dynamics of the TC core is believed to be the key to TC structure and intensity changes. The inner core processes include the inner and outer spiral rainbands, convectively coupled vortex Rossby waves, eyewall cycles, and embedded mesovortices in TC circulation. Although studies in recent years have revealed the role of inner core dynamics in TC geneses, structure and intensity changes, most of them have been limited to dry dynamics or idealized simulations with full physics models. Future studies are required to understand the response of the internal dynamics and the TC intensity to external forcing, including the complex interactions that potentially occur between the internal dynamics and the environmental forcing. For example, in the presence of upper-level trough interaction, the so-called ''bad trough'' or ''good trough'' for TC intensity may depend considerably on how the TC core responds to the forcing from the upper-level trough. Such an interaction is also relevant to the structure and intensity changes of TCs experiencing extratropical transitions (Ritchie and Elsberry, 2001). Since the current MPI theories are based on an axisymmetric assumption, questions remain as to whether, how, and to what degree the internal dynamics can affect the TC intensity or MPI.

The boundary layer processes are crucial to TC formation and maintenance or decaying, significant progress has been made recently in the estimates of drag coefficient at high winds under TCs due to the availability of Global Positioning System dropwindsonde, which allows for measuring the profiles of strong winds in the marine boundary layer associated with TCs in the Atlantic basin since 1997. The results revealed that surface momentum flux levels off as wind speeds increase above $40\,\mathrm{m\,s^{-1}}$ with the drag coefficient decreasing as wind speed increasing. This phenomenon is contrary to surface flux parameterizations currently used in a variety of modeling applications, including TC risk assessment and prediction of TC motion, intensity, ocean waves and storm surges. This finding is significant in advancing our understanding of the air-sea interaction at high wind regime and can reduce the uncertainties in the calculation of surface fluxes, and thus improving TC intensity forecast, in particular, by numerical weather prediction models. A prompt development of a new parameterization scheme for surface flux calculation based on the new measurements and an evaluation of its effect on TC intensity are in great request in the TC community.

Important yet less studied so far are the physical processes and mechanisms determining the rapid intensification of a TC. This is both scientifically and practically fundamental since the forecast errors are mostly related to those storms that are experiencing rapid intensity changes (DeMaria and Kaplan, 1999). Such rapid intensity changes (both rapid deepening or weakening) could involve complex interactions between the storm, its environmental flow and the underlying surface (ocean or land). Therefore, this will be a fertile area worth further studies.

Another practical problem is to understand the transition of boundary layer structure of a TC making landfall. Observations indicate complex interactions and transitions for TCs approaching coastlines. Several numerical simulations also revealed many important dynamics of TC transition at landfall. However, the evolution and transition of the internal planetary boundary layer at landfall are not well understood, while understanding these is important to shedding lights to the dynamics of the TC boundary layer at landfall and also to the dynamics of disasters caused by landfalling TCs (Willoughby and Black, 1996). Future studies should include the detailed land-atmosphere interaction and TC-topography interaction during landfalling and their effects on the distribution of precipitation (Elsberry, 2002).

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