Asymmetric Structures in a Simulated Landfalling Hurricane

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ABSTRACT

Highly asymmetric structures in a landfalling hurricane can lead to the formation of heavy rains, wind gusts, and tornados at preferred locations relative to the center of the hurricane. In this study, the development of asymmetric structures in an explicitly simulated idealized hurricane during landfall was investigated. It was found that the boundary layer friction and its associated convection produce a low-level positive potential vorticity (PV) band ahead of the hurricane. The interaction between the PV band and the eyewall PV ring leads to a temporary weakening and reintensifying cycle. Asymmetric structures arise from the near discontinuity of the surface friction and the latent heat flux. The breaking of the eyewall in the rear quadrants is favorable for the intrusion of the low moist entropy air into the core. Consequently, PV increases significantly in the core, in and just above the boundary layer due to the stabilization. After the hurricane makes landfall, the diabatic heating in the eyewall is reduced and cannot generate enough PV to maintain the PV ring in the middle and upper troposphere. The PV ring evolves into a monopolar structure through the nonlinear mixing process.

The Eliassen–Palm (EP) flux and its divergence in the Eulerian mean equations in isentropic coordinates are applied to explore the wave dynamics and wave–mean flow interactions. The vortex Rossby wave–related eddy momentum and heat transports, indicated by the EP flux, vary as a response to the evolution of the PV structure. The wave–mean flow interaction has a significant effect on the tangential wind, which is dominated by the mean circulation, especially the symmetric diabatic heating. Together with the asymmetric diabatic heating, the waves tend to counteract the effect of the mean circulation.

1. Introduction

A hurricane usually is not symmetric. It is made up of a strong symmetric circulation and relatively weaker asymmetric features. The asymmetries may arise from the environmental wind shear (e.g., Jones 1995; DeMaria 1996; Bender 1997; Frank and Ritchie 1999), the interaction with upper-level synoptic systems (e.g., Molinari et al. 1995, 1998; Shi et al. 1997; Bosart et al. 2000; Persing et al. 2002), and the nonuniform surface characteristics (e.g., Bosart et al. 2000; Shay et al. 2000; Hong et al. 2000). Observational studies (e.g., Willoughby et al. 1984; Powell 1987; Powell and Houston 1998; Blackwell 2000) and numerical studies (e.g., Liu et al. 1997; Zhang et al. 1999) have shown that land–sea contrast tends to produce highly asymmetric structures in a landfalling hurricane. These asymmetries may affect not only the intensity of the hurricane (e.g., Montgomery and Kallenbach 1997; Möller and Montgomery 2000; Wang 2002b), but also induce the distribution of precipitation, wind gusts, and even tornadoes at preferred locations. Quantitative precipitation forecasting is currently an important issue in hurricane prediction. Understanding the dynamics and thermodynamics of asymmetric structures formed during hurricane landfall is therefore important for improving our forecasting skills and establishing a better warning system.

In a rapidly rotating hurricane environment, asymmetries often appear in the form of spiral rainbands. Two types of spiral rainbands, namely outer and inner spiral rainbands, are observed. Radar imagery often reveals some intense outer convective spiral bands forming ahead of the center of a hurricane when it approaches land (e.g., Parrish et al. 1982; Burpee and Black 1989; Wakimoto and Black 1994). When interacting with the inner core of a hurricane, such enhanced convective bands may influence the storm to make a track acceleration (Willoughby 1992) and intensity fluctuations (Wang 2002b). Willoughby (1990) showed that the weakening of Atlantic hurricanes just before landfall often results from the formation of a concentric eyewall and the eyewall replacement process. From extensive case studies of 22 concentric eyewall hurricanes, Nong (2000) argued that interactions between a hurricane and its upper-level synoptic environment is neither sufficient, nor necessary for the eyewall replacement process. It is possible that outer spiral rainbands induced
by land–sea contrast may play a certain role in some cases.

As the other main form of asymmetry, inner spiral rainbands connect directly to the eyewall of a hurricane.¹ Usually, inner spiral rainbands have stronger convectons and are closer to the center than outer rainbands; thus, they affect more significantly the symmetric vortex. Observational (Reasor et al. 2000) and numerical studies (e.g., Guinn and Schubert 1993; Chen and Yau 2001; Wang 2001, 2002a,b) show that inner spiral rainbands exhibit characteristics of vortex Rossby waves, which exist on the radial gradient of the potential vorticity (PV) of the storm. In recent years, vortex Rossby wave dynamics as a mechanism governing inner spiral rainbands in a hurricane has been studied extensively.

In the context of dry dynamics, various studies using filtered models (e.g., Montgomery and Kallenbach 1997; Montgomery and Enagyon 1998; Möll er and Montgomery 1999, 2000) have shown that, in a vortex with monopolar PV structure, the axisymmetrization of the initial PV anomalies can intensify the symmetric vortex by redistributing the angular momentum. For a mature hurricane with ring structures in PV, vortex Rossby waves tend to spin down the vortex in the absence of diabatic heating (Chen and Yau 2001) through the PV mixing process (Schubert et al. 1999). On the other hand, full-physics simulations of rapidly deepening hurricanes suggest that vortex Rossby waves substantially intensify the symmetric circulation (Chen and Yau 2001; Chen et al. 2003; Möller and Shapiro 2002). Since the propagation characteristics and dynamical effects of vortex Rossby waves on the symmetric vortex depend strongly on the structure of the symmetric circulation (e.g., the radial gradient of the basic-state PV), we expect that the inner spiral bands may play different roles at different stages (namely, the spinup and spin-down stages) of development of a landfalling hurricane. It is our goal to clarify the behavior of the inner spiral rainbands as well as their interactions with the outer rainbands in a landfalling model hurricane.

In past studies mentioned above, vortex Rossby wave–mean flow interaction was investigated by using the tangential wind tendency equation. An alternative and more powerful tool is the Eliassen–Palm (EP) flux, which has been widely used in large-scale dynamics. A few studies have extended its application to hurricane dynamics (Schubert 1985; Molinari et al. 1995, 1998; Chen et al. 2003). We will apply the EP flux and its divergence to indicate the eddy momentum and heat transport, wave propagation, and wave–mean flow interaction.

In section 2, we briefly describe the model and the design of the numerical experiments. Section 3 presents the simulation results. Discussion and conclusions are presented in section 4.

² The eyewall of a hurricane usually is not symmetric. We shall define the asymmetric component of the eyewall circulation also as inner spiral rainbands.

<table>
<thead>
<tr>
<th>Land use category</th>
<th>Albedo (%)</th>
<th>Moisture availability (%)</th>
<th>Roughness length (cm)</th>
<th>Thermal inertia (10³ J m⁻² K⁻¹ s⁻¹/²)</th>
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<tr>
<td>Wetland</td>
<td>14</td>
<td>50</td>
<td>20</td>
<td>2.51</td>
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2. Model and simulation design

The model employed here is an improved version of the nonhydrostatic Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model version 5 (MM5) with 6-km resolution and 232 × 169 grid points in the horizontal and 24 σ levels in the vertical. The moist processes are simulated explicitly with the Tao and Simpson (1993) microphysics scheme. The Burk and Thompson (1989) planetary boundary layer scheme, which predicts turbulent kinetic energy using the 1.5-order–level-2.5 Mellor–Yamada (1974) formulation, is applied to resolve the boundary layer processes. The model was initialized with a symmetric hurricane at 0800 local time and ran for 24 h. The initial conditions are provided by the simulation of Hurricane Andrew in Liu et al. (1997), but are modified to remove asymmetries within a radius of 390 km from the center of the vortex. A detailed description of the model and the initial conditions can be found in Chen and Yau (2001) and the references therein.

Two simulations are performed. In the control simulation, the surface is assumed to be all water with a constant sea surface temperature (SST) of 28°C. The results of the control simulation have been presented in Chen and Yau (2001) and Chen et al. (2003). The landfall simulation has the same setting as the control run except that one-third of the grid mesh to the west of the domain is prescribed to be over flat land (1-m elevation), with marsh or wetland characteristics. The remaining two thirds are assumed to be over ocean. The land surface temperature, initially the same as the constant SST (28°C), is allowed to evolve with time according to the surface energy budget. Table 1 compares some physical parameters of the land and sea surfaces (cf. Grell et al. 1995).

3. Simulation result

a. Prelandfall rainband

The hurricane generally moves from east to west at an average speed of 7.8 m s⁻¹ in both simulations. Since the steering flow is fairly fast and there is no topography included in the landfall simulation, the track differs in-
The eye of the hurricane makes landfall at 19 h. Figure 1 compares the minimum central sea level pressures and the maximum symmetric component of the surface ($\sigma = 0.995$) tangential winds for the control and the landfall runs. As expected, the landfalling hurricane undergoes two stages: a deepening stage before 15 h, and a rapidly weakening stage after 18 h. In between we see an interesting 2-h temporary weakening and reintensifying cycle. Willoughby et al. (1984) and Willoughby (1990) showed that hurricanes with concentric eyewall and the eyewall replacement would experience a temporary but significant weakening and reintensifying cycle. Wang (2002b) demonstrated that the interaction between an outer spiral band and the eyewall can lead to the breakdown and recovery of the eyewall, which also yields a weakening and reintensifying cycle. In our simulation, we did not observe a concentric eyewall. The small 2-hPa weakening and reintensifying cycle results from an outer PV band interacting with the eyewall PV ring.

Ertel PV maps [see Eq. (1) in Chen and Yau (2001) for the governing equation for Ertel PV] show a quasi-stationary outer rainband forming inland near the shoreline at around 12 h when the hurricane center is 215 km away from the coastline (Fig. 2). The rainband is initiated by friction-induced convergent winds in the boundary layer (Parrish et al. 1982; Burpee and Black 1989) and is characterized by a positive PV band associated with shallow convective clouds extending to a height of 3 km. The positive PV band is generated through two processes, which are both related to boundary layer friction. First, frictional convergence induces convection. The accompanying latent heat release is an important source for PV (Raymond and Jiang 1990; May and Holland 1999; Chen and Yau 2001; Wang 2002a; and others). Second, the boundary layer friction decreases with height, which has a tendency to produce vertical wind shear and the associated horizontally pointing vorticity vector. Positive PV is generated when the horizontal vorticity vector has a component along a preexisting thermal gradient. In our simulation, the land surface is generally colder than the fixed SST when the hurricane is near the land and when it makes landfall because of the lack of shortwave radiation at night and the enhanced evaporative cooling effect. Specifically, the near-surface air is cooled in response to the reduced latent heat transfer from the cold land surface (Tuleya 1994). Therefore, a strong thermal gradient is formed in the boundary layer (cf. Fig. 3c) and low-level PV is generated. Figure 2 illustrates that when the hurricane moves further toward the land, the quasi-stationary PV band develops and interacts with the eyewall PV ring.

Figure 3 depicts the low-level PV, wind, equivalent potential temperature ($\theta_e$), and surface latent heat flux structures valid at 15 h 20 min. The eyewall PV ring [denoted by 20 PV units (PVU) contours] is broken at this time (Fig. 3a), as a consequence of an inner spiral rainband development and the eyewall PV inward mixing (Montgomery and Kallenbach 1997; Schubert et al. 1999; Chen and Yau 2001; Wang 2002b; and others). The mesovortex inside the eyewall PV ring has a local minimum pressure, which makes the low-pressure center slightly off the geometric center (Kossin and Schubert 2001). However, the mesovortex does not decrease the central pressure due to compensation by the cold air import through the spiral PV band (see the thick dashed lines in Fig. 3). The PV band is associated with strong inward radial wind and upward motion. However, the indented contour in wind speed in Fig. 3b indicates that the total wind in the band is weaker than the surrounding environment. The band also provides a narrow channel.
Fig. 2. Hourly evolution of the horizontal structures of PV (PVU, 1 PVU = 10^{-5} kg m^{-1} K m^2 s^{-1}) at the lowest model level ($\sigma = 0.995$), from (a) 12 h to (f) 17 h. The thick lines denote the shore. The coordinates are centered at the hurricane center.

Inland PV trough which cold and dense inland air parcels are advected toward the inner core region. A trajectory analysis shows that some air parcels originating from the inland PV band are advected through the gap in the broken PV ring into the eye, while some other parcels move cyclonically and are elevated upward in the eyewall (graph not shown). The weaker surface wind in the band results in a reduced latent heat transfer from the ocean surface along the PV band (Fig. 3d). Hence, air parcels streaming in do not increase sufficiently their
equivalent potential temperature to enhance the moist static energy in the subcloud layer. Convection in the eyewall is then reduced due to the convective stabilization. This process, related to wind-induced surface heat exchange (WISHE; Emanuel 1989; Emanuel et al. 1994), therefore weakens the vortex.

The broken PV ring is quickly recovered with the help of the axisymmetrization process. The central pressure drops 2 hPa before the eyewall of the vortex makes landfall (cf. Fig. 1). In this experiment, the magnitude of the weakening and re-intensifying cycle is very small (2 hPa) since the boundary layer friction only produces a weak, shallow PV band. If topography is included, the interaction between the PV band and the eyewall PV ring can be stronger. In another experiment when a 1-km-high narrow mountain ridge is added along the shore, we obtain a weakening and re-intensifying cycle with 4-hPa pressure change. The vortex also feels the influence of the land 2 h earlier (graph not shown).

b. Asymmetric structures during landfall

It has long been recognized that the surface latent heat flux from the ocean is the primary energy source for tropical cyclones. The high correlation between the maximum latent heat flux from the ocean and the maximum surface wind indicates a positive feedback process between surface evaporation and the increased low-level wind (cf. Figs. 3b,d). The latent heat flux over the land is one order of magnitude smaller than that over the ocean and this implies that the hurricane will decay rapidly upon making landfall. Figure 1 shows that the
symmetric component of the low-level winds decrease by 19 m s\(^{-1}\) and surface pressure falls 27 hPa in the last 6-h period. Figure 1 also shows that the surface wind weakens 1 h prior to the sea level pressure, which does not start to increase until significant portions of the eyewall have been over land. This suggests that the dynamical adjustment of the vortex to the change of the boundary layer friction is faster than the thermodynamical adjustment to the change of the surface heat source (Yau et al. 1999). The asymmetric distribution of the surface latent heat flux and the friction also implies that highly asymmetric structures will be generated.

Figure 4 depicts the surface structures of the hurricane at 21 h when the vortex center has passed 66 km on-shore. Compared to Fig. 3, the surface wind is much weaker (a maximum value of 80 m s\(^{-1}\) prelandfall versus 51 m s\(^{-1}\) postlandfall). The maximum wind shifts from the right-front quadrant to ahead of the vortex center, in agreement with previous studies (e.g., Shapiro 1983). Both the wind and the surface latent heat flux are nearly discontinuous at the shore (Figs. 4b,d). The maximum surface latent heat flux under the core of the hurricane is 375 W m\(^{-2}\), which is still 5 times less than that over the ocean even though the maximum surface wind is over the land. Tuleya (1994) conducted a series of numerical experiments for the development of tropical cyclones with different surface characteristics (e.g., moisture availability, roughness length, thermal inertia, fixed or varying surface temperature) and concluded that the major reason for the reduced evaporative flux is not the lack of moisture nor the increased roughness, but rather the development of a cool, relatively dry pool of air over cool ground near the vortex core.

The conclusion of Tuleya (1994) is also borne out in our simulation. When the hurricane is not far from the shoreline, it is still being supplied with warm and moist air from the ocean. Figure 4c shows that the low-level equivalent potential temperature field in the near core region has a wavenumber-1 structure. The relatively warm and moist air flow to the north (or to the right of the track) is wrapping cyclonically into the core from the front quadrants. On the other hand, the relatively cold and dry air flow to the south (or left of the track) is wrapping into the core at the rear quadrants. Because
of the downward sensible heat flux over the land surface, air parcels in the cold branch of the flow obtain nearly no net heat from the underlying surface except within a radius of about 30 km from the center or when they cross the sharp gradient of $\theta_e$ in the rear-right quadrant. The horizontal thermal gradient and the low-level wind shear contribute to the generation of the very shallow negative PV in the boundary layer along the coast (Fig. 4a). The formation of the sharp gradient of $\theta_e$, which extends from the surface up to 3 km (graph not shown), indicates the low-level frontogenesis in the rear quadrants, in agreement with the finding of Eliassen (1959) who was the first to notice that the flow in the eyewall is inherently frontogenetic. Emanuel (1995, 1997) further emphasized this idea and demonstrated that his balance-symmetric hurricane model entails strong frontogenesis at the inner edge of the eyewall. The front in the rear-right quadrant is not associated with the eyewall, but is the result of an inner rainband marked by enhanced upward motion (Fig. 5). As shown in Fig. 5a, the low-level eyewall, signified by the strong upward vertical velocity, is broken in the region where cold and dry air intrusion lowers the moist entropy in the eye. The enhanced radar reflectivity (Fig. 5b) suggests that the front also contributes to the formation of asymmetric precipitation patterns of the storm.

The model-accumulated precipitation pattern over the 24-h period, depicted in Fig. 6, is obviously asymmetric. The maximum precipitation often occurs on the right of the track even when the hurricane is over the ocean. Vertical wind shear has been known to be a dominant factor in the formation of precipitation asymmetries in

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**Fig. 5.** As in Fig. 3, but for (a) vertical velocity $w$ at intervals of 0.3 m s$^{-1}$, with regions $\geq 0.9$ m s$^{-1}$ shaded; (b) simulated radar reflectivity contoured at 30, 40, 45, and 50 dBZ, with regions $\geq 45$ dBZ shaded, at $z = 1.2$ km and valid at 21 h.

**Fig. 6.** Accumulated total precipitation contoured at 5, 10, 15, 20, 30, 40, and 50 cm, with regions $\geq 30$ cm shaded for the 24-h integration period. The thick solid line denotes the track and 6-h eye positions are indicated by the hurricane symbols. The shore is denoted by the thick dashed line. The origin of the coordinates is the eye position at 24 h.
tropical cyclones without complex underlying topography. For example, observational studies of Atlantic hurricanes (e.g., Willoughby et al. 1984; Marks et al. 1992; Franklin et al. 1993) and numerical simulations (e.g., Bender 1997; Frank and Ritchie 1999, 2001) have shown that the precipitation in hurricanes away from the land tends to be maximized on the left side of the vertical shear vector (i.e., downshear left).

To investigate the shear–precipitation asymmetry relation, we note that, in addition to the environmental shear at the lateral boundary, there is also the shear generated by the storm that needs to be taken into account. In our simulation, the lateral boundary conditions contain only weak (less than 4.2 m s\(^{-1}\)) between 8- and 1-km altitude and in general northwesterly shear (Fig. 7). The storm-scale vertical wind shear is calculated following Marks et al. (1992) and Rogers et al. (2002) as follows. First, the storm motion is subtracted from the total wind field to obtain the relative wind. Second, we average spatially the relative wind over a 540 km \(\times\) 540 km box centered on the eye and use this profile to compute the vertical wind shear. The evolution of the magnitude and the direction of the storm-scale shear between the 8- and 1-km heights is shown in Fig. 7. Comparing with Fig. 1, we note that the magnitude of the shear is proportional to the maximum wind at the surface. The shear increases from 1.7 to 10.6 m s\(^{-1}\) in the deepening stage and then decreases to 4.9 m s\(^{-1}\) at the end of the simulation. Vertical wind shear is thought to have a generally negative impact on the intensity of hurricanes. The time evolution of the shear may indicate that it acts to limit the storm in reaching its maximum potential intensity. The shear changes from northwesterly before landfall to southwesterly after landfall due to the development of the southerly shear.

By applying the simple shear-precipitation relation (e.g., Frank and Ritchie 2002) to the shear in Fig. 7, we found that the maximum precipitation is located downshear left, just as depicted in Fig. 6 when the hurricane is over the ocean. The same conclusion remains valid when the environmental shear is included because its direction is similar to the storm-scale shear but its magnitude is two times less. It is recognized that the \(\beta\) effect and the boundary layer friction may also produce enhanced precipitation, respectively, in the rear quadrants (Bender 1997) and in the right-front quadrant relative to a moving hurricane (Shapiro 1983). However, their influence on the asymmetric precipitation distribution is likely to be much weaker than the vertical wind shear effect (e.g., Bender 1997). To verify this conjecture, our sensitivity experiment with the same configuration as the current landfall simulation but on a \(f\) plane without boundary layer friction was performed. The storm-scale shear was found to be similar to Fig. 7, except its direction remain northwesterly. The downshear left bias is still present in the accumulated precipitation field.

After landfall, a careful inspection of Fig. 6 shows that the precipitation, in agreement with some observations (e.g., Burpee and Marks 1984), tends to be equally distributed on both the left and right of the shear vector lying along the track. Our results over the ocean are in agreement with those from Frank and Ritchie (1999). In their simulation with easterly zonal flow and westerly shear and a uniform ocean surface, they showed that convection is concentrated in two regions, west and northeast of the center. However, over land, our simulation shows another maximum of convection occurring southwest of the center (not shown). This suggests that the surface characteristics and the associated boundary layer processes need to be considered when discussing the precipitation asymmetries of a hurricane in a sheared environment.

c. Symmetric structure change

To reveal the evolution of a hurricane during landfall, we need to discuss its symmetric structure change not only because the symmetric circulation is much stronger than its asymmetric counterpart, but also because the dynamics of the asymmetries depend on the symmetric structure.

Figure 8 depicts the symmetric wind and thermal structures of the mature hurricane before landfall at 13 h. The storm has a vigorous tangential wind field with a maximum value of 77 m s\(^{-1}\) in the boundary layer. The radius of maximum wind (RMW) tilts slightly inward below 7 km and then turns to slant outward above this level (Fig. 8a). As an important branch of the secondary circulation (Fig. 8b), the strong boundary layer radial inflow supplies angular momentum to the vortex core, which balances both the frictional loss to the surface and the transport to the upper-level atmosphere by the strong updraft in the eyewall. The inflow layer extends from the boundary layer up to 6 km at the 90-km radius and becomes shallower at smaller radii. The radial outflow lies just above the inflow layer, with a relatively strong low-level jet (8.5 m s\(^{-1}\)) in the eyewall at the top of the boundary layer and a strong jet (20 m s\(^{-1}\)) near the tropopause. The upper-level outflow rotates...
anticyclonically beyond the 200-km radius. The vertical velocity in the eyewall has its maximum value of 2.5 m s$^{-1}$ at a radius of 45 km and at a height of 7 km. In a saturated environment, such strong updraft results in strong convection, which releases a huge amount of latent heat. The response of the vortex to this heat source in a balanced Eliassen model (e.g., Ooyama 1969) is to generate a forced secondary circulation [essentially vertical motion as shown by Holland and Merrill (1984)], which in turn redistributes heat and angular momentum. For an intensifying tropical storm, the induced secondary circulation in balanced flows tends to, on the one hand, accelerate the tangential wind (Shapiro and Willoughby 1982; Hack and Schubert 1986; Möller and Shapiro 2002). On the other hand, it tends to cancel the direct effect of heating through adiabatic cooling and results in little net warming in the eyewall (Andrews and McIntyre 1976, 1978; Boyd 1976; Zhang et al. 2002). Consequently, the maximum warming occurs in the eye as a result of the adiabatic warming by the forced subsidence seen in Fig. 8b. The warm core structure shown in Fig. 8c is typical in a tropical cyclone. The strong gradient of $\theta_e$ in the eyewall indicates that the eyewall can be described as an atmospheric front (Eliassen 1959; Emanuel 1995). Note that the boundary layer inflow air increases its $\theta_e$ value as it approaches the center. Liu et al. (1999) estimated the contributions from the surface latent heat flux, sensible heat flux and isothermal expansion to the $\theta_e$ increase of an air parcel moving cyclonically from a radius of 150 km to the eye center and concluded that the surface latent heat flux, accounting for 64% of the total increase, dominates the other two processes. The eyewall convection is maintained or even strengthened by these air parcels with elevated $\theta_e$ values when they enter the updraft in the eyewall (Willoughby 1995; Zhang et al. 2002).

After landfall, the hurricane intensity decays rapidly. Figure 9 depicts the symmetric structures at 23 h (4 h after landfall). Although the maximum tangential wind is still located in the boundary layer, the value has decreased 15 m s$^{-1}$ to 62 m s$^{-1}$ (Fig. 9a). The RMW shrinks (expands) about 10 km below (above) the 4-km height that results in a more tilted eyewall. The boundary layer inflow is augmented in response to the in-
creased friction over land. Just above the boundary layer, however, an outflow layer develops, which offsets the increased mass flux by the inflow that limits the strength of the eyewall updraft. The updraft is significantly reduced, as is the eyewall convection. It has been found that without condensational heating in the eyewall, air converging near the surface will rise to the top of the boundary layer and flow outward immediately instead of at the tropopause (Willoughby 1979; Chen and Yau 2001). Our simulation suggests that this outflow layer develops as long as the eyewall convection is weakened. Meanwhile the upper-level outflow jet is slowed down and the air parcels flow outward through nearly the whole troposphere. The maximum updraft (1 m s\(^{-1}\)) is at the top of the boundary layer. Compared to Fig. 8b, Fig. 9b shows two updraft branches. When the previous eyewall updraft weakens, the second updraft begins to develop at about 80-km radius and extends downward to the boundary layer. A plane view of the vertical velocities shows that the eyewall convection has a highly asymmetric structure (graph not shown). The outer symmetric updraft is actually the zero wavenumber component of the strong vertical motion in the spiral rainbands. Another interesting change in the symmetric secondary circulation is the gentle upward motion in the eye. With this upward motion, the warm and moist air that previously remained in the low-level eye region can be lifted to a higher level. Meanwhile, the near surface region in the eye is refilled with cold and dry air so that the low-level atmosphere in the eye becomes more stable (Fig. 9c). Also note that the weak radial gradient of \(\theta\), near the surface indicates small increases of moist entropy in the inflow air parcels. This is one reason why the convection in the eyewall is enfeebled.

In the context of PV dynamics, the condensational heating in the eyewall is the primary source for the maintenance of the eyewall PV ring depicted in Fig. 8d (May and Holland 1999; Chen and Yau 2001; Wang 2002a). The radial gradient of the basic-state (i.e., axisymmetric component) PV changing sign is a necessary condition for the instability of PV disturbances (Ren 1999; Nolan and Farrell 1999; Schubert et al. 1999). Schubert et al. (1999) illustrated how a PV ring structure evolves to a monopolar distribution by the PV mixing mechanism in a dry barotropic balanced system governed by the 2D vorticity equation. In a 3D full-physics simulation, Chen and Yau (2001) demonstrated that the "bowl-shaped" PV structure can be sustained due to the large generation of PV by the eyewall convection.
while it becomes a midlevel monopolar PV structure without moist processes. As depicted in Fig. 8d and Fig. 9d, when the eyewall convection is weakened, an annular PV ring tends to evolve into a monopolar structure. The ring still exists in the upper levels (above 8 km), however, the PV contours become nearly flat. There are two PV maxima in the eye. The low-level monopolar PV located at 1 km results from the stabilization of the atmosphere in the core, while the upper-level PV maximum located at 6 km forms from the PV mixing process. The PV mixing process is governed by vortex Rossby waves, which redistribute not only the PV, but also the angular momentum.

Along with the PV structure changes, the tangential winds in the eye region increase substantially, while at RMW the winds decrease (Schubert et al. 1999; Montgomery and Kallenbach 1997; Chen et al. 2003; and others). On careful comparison of Figs. 8a and 9a, one can find that the tangential winds are decelerated at the RMW and are accelerated on either side of this radius. However, the total kinetic energy is reduced after landfall.

d. Wave–mean flow interaction

The symmetric primary circulation of a hurricane slowly evolves when heat and momentum sources force the secondary circulation. The heat and momentum sources are often associated with asymmetric motions. The significant changes of the symmetric structures are actually asymmetric in manner, especially during landfall (cf. Figs. 4 and 5).

Asymmetric structures in the near-core region, mainly the inner spiral rainbands, are found to show characteristics of vortex Rossby waves (Reasor et al. 1999; Chen and Yau 2001; Wang 2001, 2002a,b). In recent years, extensive theoretical studies (e.g., Montgomery and Kallenbach 1997; Montgomery and Enagionio 1998; Möller and Montgomery 1999, 2000; Shapiro 2000) have provided fundamental insights into the vortex Rossby wave–mean flow interaction. Results from these theoretical studies and from full-physics numerical studies of Möller and Shapiro (2002) and Chen et al. (2003) have demonstrated how vortex Rossby waves help a hurricane to strengthen during its rapid intensification. In their study, Möller and Shapiro (2002) obtained vortex Rossby waves by the use of asymmetric balance (AB; Shapiro and Montgomery 1993) model and PV inversion. Chen et al. (2003) separated vortex Rossby waves from gravity waves by applying the empirical normal mode method (Brunet 1994). They concluded that the vortex Rossby waves dominate gravity waves in terms of wave activity in the inner-core region. Therefore, the total (balanced and unbalanced) eddy contribution to the momentum change is essentially from the vortex Rossby waves.

To study the wave processes, the atmospheric flow is usually divided into a symmetric basic-state part and an asymmetric disturbance part. The effect of waves or eddies on the symmetric circulation is usually measured in terms of eddy momentum and heat fluxes in a set of Eulerian mean equations. Considering the strong cancellation between the eddy heat flux convergence and adiabatic cooling by the mean flow, Andrews and McIntyre (1976) introduced the so-called transformed Eulerian mean formulation in which the eddy heat and momentum fluxes act together in the form of the divergence of the EP flux to drive changes in the mean circulation. The EP flux and its divergence are good indicators of the wave propagation and wave–mean flow interaction. They have been widely used in large-scale dynamics and have been extended to hurricane studies. Schubert (1985) derived the EP theorem for hurricanes and suggested that the EP flux and its divergence are helpful for better understanding the interactions of symmetric and asymmetric components in actual and model hurricanes. Molinari et al. (1995, 1998) applied such theorem in real hurricanes to investigate the hurricane-rough interactions. The EP flux, as well as the generalized wave activity conservation laws, were further used by Chen et al. (2003) to explore the dynamics of vortex Rossby waves in the inner-core region of a high-resolution explicitly simulated hurricane.

The interaction is generally a two-way process. The mean flow structure can strongly modify the propagation of the disturbances, while the wave disturbances can bring significant mean flow changes through rectified nonlinear effects. It has been verified that the propagation of vortex Rossby waves depends on the radial gradient of the basic-state PV (Montgomery and Kallenbach 1997; Chen and Yau 2001). Figures 8 and 9 show that the basic-state PV undergoes substantial changes throughout landfall. It implies that vortex Rossby waves in the inner-core region will change their propagation properties, the associated energy transports and probably the eddy forcing on the mean flow during landfall. It is interesting to see how vortex Rossby waves experience changes and how they impose different forces on the mean flow.

Following Schubert (1985) and Molinari et al. (1995), we use the forced primitive equations in storm-following cylindrical and isentropic coordinates \((r, \lambda, \theta)\) to derive a set of Eulerian azimuthal mean equations. These equations bear a close formal resemblance to the transformed Eulerian mean equations and share the advantages of that set. Another benefit that we receive by the use of isentropic coordinates, is the direct measurement of the diabatic heating effect on the tangential wind change, which has been shown to be the dominant term in the balanced Eliassen model (Möller and Shapiro 2002). The assumption we need to transform our dataset from the model coordinates to the isentropic coordinates is the hydrostatic balance, which is valid. The deviation from hydrostatic balance in the model data is within 5% even in the eyewall region. The Eulerian azimuthal mean equations are
\[ \begin{align*}
\pi^* + \pi^* \pi^* + \theta^* \pi^* & - \left( \frac{\vec{f} + \nabla}{r} \right) \nabla \\
+ \bar{M}_r - \bar{X}^* &= \frac{\bar{D}}{\bar{\sigma}}, \quad (1a) \\
\n\bar{v}_r + \pi^* \bar{u}^* + \theta^* \pi^* - \bar{Y}^* &= \frac{1}{r} \nabla \cdot \mathbf{F} - \frac{1}{\sigma^*} \left[ (\sigma \dot{\theta} \bar{v}' \bar{u}') \right]_0 \\
- \left( \frac{\sigma^* \bar{v}'}{\bar{\sigma}} \right) &= \left( \frac{\sigma^* \bar{u}'}{\bar{\sigma}} \right), \quad (1b) \\
\bar{\sigma}_r + \frac{1}{r} (r \bar{\sigma} \bar{u}^*), + (\bar{\sigma} \bar{\theta}^*)_0 &= 0, \quad (1c) \\
\bar{M}_\theta &= C_p \left( \frac{\dot{\rho} - \rho}{\bar{\rho}} \right), \quad (1d) \\
\bar{p}_\theta &= -g \bar{\sigma}, \quad (1e)
\end{align*} \]

where

\[ \mathbf{F} = \left[ -\frac{r \langle \sigma \bar{u} \rangle \bar{v}'}{g}, \frac{\bar{P}^*}{g} \bar{M}^*_r \right] \] and

\[ \nabla \cdot \mathbf{F} = \frac{1}{r} \left[ -r \langle \sigma \bar{u} \rangle \bar{v}' \right], + \left( \frac{\bar{P}^*}{g} \bar{M}^*_r \right) \] are the EP flux and its divergence, respectively. The symbol

\[ \bar{D} = \frac{1}{r} [r \pi^* \bar{\sigma} \bar{u}' - r \langle \sigma \bar{u} \rangle \bar{u}'] + [\bar{\theta}^* \bar{\sigma}' \bar{u}' - (\sigma \dot{\theta} \bar{u}')]_0 + \left( \frac{\vec{f} + \nabla}{r} \right) \bar{\sigma}' \bar{v}' + \frac{1}{r} (\sigma \bar{v}') \bar{u}' + \sigma \bar{f}' \bar{v} \\
+ \bar{\sigma} \bar{f}' \bar{v}' - \sigma \bar{M}_r, \] divided by \( \bar{\sigma} \), contains all the eddy terms for the momentum equation in the radial direction. An overbar signifies an azimuthal mean, and a prime denotes the deviation from the mean. Variables \( u, v \) are, respectively, the radial wind and tangential wind relative to the moving hurricane center. We have assumed that the motion of the storm is constant. In (1), \( M = C_p T + \Phi \) is the Montgomery function (\( T \) is the temperature and \( \Phi \) is the geopotential), \( \sigma \) the isentropic density, \( p \), a constant reference pressure, and \( \bar{\eta} = \vec{f} + (r \bar{\sigma})/r \) is the vertical component of the azimuthal mean absolute vorticity, \( \dot{\theta} \) is the diabatic heating rate, and \( \bar{X} \) is the vertical component of the azimuthal mean absolute motion of the storm is constant. In (1), \( \pi \) is the EP flux (star is defined as the Montgomery function (\( \star \) is de®ned as the Montgomery function)).

The vector \((\pi^*, \dot{\theta}^*)\) represents the so-called residual circulation.

In (1a)–(1c), eddy terms appear on the right-hand side while the mean terms are on the left. If we neglect the variation of the Coriolis parameter, which has been shown to be small in the near core region (Molinari et al. 1995), (1) is identical to that of Schubert (1985). Only the mean tangential wind budget will be computed here. The second to sixth terms in the mean tangential wind tendency Eq. (1b) are, respectively, the mean radial vorticity flux (VORM), the mean diabatic heating term (DIABM), the mean friction (FRIC), the wave–mean flow interaction term or the divergence of the EP flux term (EPFD), and the asymmetric diabatic heating term (DIABE). The last two terms in (1b) (respectively, the transient eddy momentum and the eddy radial planetary vorticity flux) usually are very small, and are not presented.

The horizontal component of the EP flux is antiproportional to the eddy momentum transport in the radial direction. Associated with an outward pointing EP flux, eddies transport momentum inward and vice versa. The vertical component of the EP flux is related to the eddy heat transport in the radial direction. Eddies transport heat outward when the EP flux points downward and vice versa (see also Chen et al. 2003). The divergence of the EP flux \( \nabla \cdot \mathbf{F} \) is equivalent to the net azimuthal pressure force (Andrews et al. 1987). When \( \nabla \cdot \mathbf{F} > 0 \), eddies tend to accelerate the mean tangential wind. The nonacceleration theorem states that \( \nabla \cdot \mathbf{F} = 0 \) for steady, conservative, and small-amplitude wave; that is, such waves cannot induce mean flow changes (Charney and Drazin 1961; Andrews and McIntyre 1976; Boyd 1976; Schubert 1985). Hence, the EP flux can be used to indicate the propagation of the wave activities (cf. Chen et al. 2003, and references therein). The divergence of the EP flux is also antiproportional to the radial eddy PV flux (Tung 1986; Molinari et al. 1995).

For the sake of simplicity, previous studies of hurricanes using the Eulerian mean equations in the isentropic coordinates have ignored the lower boundaries by choosing the lowest isentrope above the ground (e.g., Molinari et al. 1995, 1998; Chen et al. 2003). Such an isentrope is usually well above the boundary layer outside the eye in a mature hurricane. For a landfalling hurricane, however, the boundary layer processes are crucial in determining the due course of the evolution of the vortex. The boundary inflow layer is especially important for the momentum budget. The boundary layer where the isentropes intersect the ground can be taken into account by suitably extending the definitions of various quantities (Andrews 1983). Andrews proved that the EP theorem is still held under his extended definitions. Here, we adopt his definitions. Specifically, we extend a very thin layer at the bottom of the model domain, with the lowest isentropic level to be the minimum value of the potential temperature that occur on
the surface (the lowest model level). In this thin layer, the geopotential, the wind and the pressure have the same value as those on the surface. Consequently, there is no mass in this underground thin layer. With these extensions, the model data (state variables and instantaneous forcing) are interpolated from the model terrain-following \( \sigma \) coordinates onto isentropic coordinates with 25 isentropic levels ranging from 292 to 388 K (about 16 km) at intervals of 4 K.

Figure 10 depicts the changes of the EP flux and its
divergence from the prelandfall condition to the postlandfall condition. The thick solid lines included in the figure mark the azimuthal mean surface potential temperatures at the corresponding time. They represent the “mean ground” below which an isentropic level may not be necessarily under the “true ground,” but rather a level where the potential temperature is colder than the mean value (cf. Held and Schneider 1999). The vertical component of the EP flux is generally small. It has been scaled to be easily seen in Fig. 10. The EP fluxes are nearly horizontal. There are mainly three branches of the EP fluxes: the below-ground inward fluxes, the above-ground low-level outward fluxes and the upper-level inward fluxes. Confined to the strong boundary inflow layer, the eddies have inward fluxes that tend to transport eddy momentum outward. The layer of the outward fluxes extends from the mean ground surface to 330 K, which is about the top of the radial inflow layer for radii larger than 90 km. The maximum flux at the 50-km radius coincides with the low-level outflow jet where the eddies transport momentum inward. In a wide range of vertical levels from 330 K to the top of the domain, the eddies exhibit inward EP fluxes and thus the eddy momentum transport is outward, with maximum values in the outflow jet region at the upper level. The EP flux in the vertical direction is very weak except in the eyewall region. At 13 h, the EP fluxes point downward in the eyewall region at the low level and inside the eye, while they point upward in the upper part of the eyewall and in a narrow region just inside the eyewall. In terms of eddy heat transport, the eddies tend to flux heat from the eyewall both outward and inward.

At 23 h, responding to the changes of the low-level radial flow due to friction (cf. Figs. 8b and 9b), the below-ground inward fluxes are significantly intensified, and the region of the above-ground low-level outward fluxes extend horizontally. The vertical components of the EP fluxes are weakened as a result of the reduced convection. In the midlevel of the eye, the EP fluxes are nearly reversed so that the eddies tend to flux heat inward.

The divergence of the EP flux is related to the eddy PV flux. There is a dipole structure in the divergence field. The strong divergence in the low-level eyewall suggests that the eddies can flux the excessive PV generated in the eyewall into the eye and therefore increase the PV inside the eyewall. Note the fact that the divergence band spreading through the troposphere shifts from radii of 20–30 km into the center after landfall, reflecting the changes of the symmetric structure, which determines the behavior of the vortex Rossby waves.

The configuration of the EP flux and its divergence at 13 h is quite similar to that of Chen et al. (2003) who showed the time-average EP flux and its divergence for the leading vortex Rossby wave modes of wavenumber-1 and wavenumber-2 disturbances of a rapidly intensifying hurricane (cf. their Fig. 10). The similarity suggests that the EP flux and its divergence of the total eddies are indeed mostly from the vortex Rossby waves in the inner core region.

The wave–mean flow interaction, which is proportional to the divergence of the EP flux, is shown at 13 h in Fig. 11, along with other mean circulation and eddy terms. Consistent with recent studies by Möller and Shapiro (2002) and Persing et al. (2002), the mean diabatic heating makes the greatest contribution to the acceleration of the tangential wind in an intensifying hurricane (Fig. 11b). At 13 h, the acceleration happens throughout the troposphere with two local maximum values along the RMW: 48 m s$^{-1}$ h$^{-1}$ in the upper troposphere (356 K) and 29 m s$^{-1}$ h$^{-1}$ in the lower level (314 K). The upward advection of the slow wind from the surface by the convective updraft in the boundary layer under the eyewall decelerates the tangential wind near the ground. This deceleration mechanism, as well as those by the symmetric friction, asymmetric diabatic heating and the divergence of the EP flux, compensate the intense acceleration by the mean radial vorticity flux. Besides the two areas of acceleration in the midlevel inside the RMW, the mean vorticity flux tends to decelerate the tangential wind in the radial outflow region. The mean friction (and diffusion) only has an apparent spindown effect in the boundary layer. The total effect of the mean terms is the acceleration at the RMW and deceleration just inside. Compared to the mean circulation terms, the eddies produce significant tendencies on the tangential wind (of order 10 m s$^{-1}$ h$^{-1}$). The wave–mean flow interaction indicated by the weighted divergence of the EP flux redistributes the momentum in the lower troposphere where waves tend to decelerate the wind at the RMW and accelerate it on either side. In the upper troposphere, as shown by previous studies (e.g., Montgomery and Kallenbach 1997; Schubert et al. 1999; and others), waves tend to spin up the wind inside the RMW but spin down the wind at and outside the RMW. Together with the asymmetric diabatic heating, asymmetric structures tend to counteract the symmetric mean circulation.

During the weakening stage, as shown in Fig. 12 at 23 h, the most prominent variations are in the mean diabatic heating term and the mean vorticity flux term. Associated with the weak convection in the eyewall, the diabatic heating-induced acceleration is drastically reduced. However, the mean vorticity flux increases its contribution to the acceleration of the tangential wind in the center of the vortex. The total mean circulation effect becomes spinup in the center, with the maximum value in the midlevel, and spindown along the RMW. Meanwhile, the wave–mean flow interaction tends to decelerate the wind in a wider range in the middle and upper troposphere but accelerate the wind in the lower levels. In total, the wave–mean flow interaction and the asymmetric diabatic heating lead to, still contrary to the mean circulation, the acceleration of the tangential wind at the RMW and deceleration just inside. Hence the
asymmetries prevent an even more rapid weakening of a landfalling hurricane.

4. Discussion and conclusions

Landfalling hurricanes pose a significant threat to the coastal communities. During landfall, a hurricane inner core often exhibits highly asymmetric features that are associated with the distribution of the strongest wind and the heaviest rainfall. Asymmetric structures may also produce intensity fluctuations. Therefore, studies of the asymmetric structures of a hurricane at landfall are of great importance to improving our ability to prepare for landfalling storm events. However, lack of observational data often precludes detailed analysis of the mesoscale structures at landfall. In this study, asymmetric structures formed in a numerically simulated landfalling hurricane are analyzed. The idealized mature hurricane has an axisymmetric initial condition and is simulated for 24 h using MM5. The hurricane deepens rapidly during the first 15 h; it then makes landfall over a flat surface with wetland characteristics and undergoes structural and intensity changes.

Our main new results include the following.

- The hurricane develops storm-scale shear, which affects the distribution of precipitation.
- The symmetric PV structures of the hurricane change dramatically during landfall by cold air intrusion and PV mixing.
- The change in the mean PV structure leads to the variation of the behavior of the inner-core asymmetries.
- The EP flux and the Eulerian mean equations were
applied successfully to study the inner-core dynamics using the results from a high-resolution (6 km) model.

- The effect of diabatic heating and eddy forcing to the mean circulation was evaluated directly. At the lower levels, the contributions to the tangential wind change before landfall were $\sim 30 \text{ m s}^{-1} \text{ h}^{-1}$ by diabatic heating and $\sim 10 \text{ m s}^{-1} \text{ h}^{-1}$ by the eddies.

It is also found that asymmetric structures in the hurricane arising from the boundary layer processes can change the mean circulation significantly. As the hurricane approaches land, a low-level PV band forms along the shore ahead of the system due to frictionally induced shallow convection, the vertical wind shear and the preexisting thermal gradient. When the PV band merges with the eyewall PV ring, a 2-h temporary weakening and reintensifying cycle is observed. Although this kind of cycle is weaker and shorter than the usually observed concentric eyewall and eyewall replacement processes, it resembles such phenomena. If the hurricane encounters complex topography or upper tropospheric systems during landfall, the intensity change could be greater. The interaction of the land–sea boundary layer and a hurricane could be responsible for some eyewall replacement cycles.

The near discontinuity in the surface friction and the latent heat flux prompts the hurricane to generate highly asymmetric structures. At low levels, a wavenumber-1 feature in the equivalent potential temperature field is formed with the warm and moist air to the right of the storm and cold and dry air to the left. Low-level frontogenesis occurs in between these regions in the rear quadrants. The hurricane has the ability to develop...
storm-scale vertical wind shear. The shear evolves with the intensity of the storm and changes direction as the hurricane makes landfall. In agreement with other studies, the accumulated precipitation has a downshear left bias over the ocean. However, the fact that it tends to be uniform when the hurricane is over the land suggests that the boundary layer processes need to be carefully considered when studying the shear-induced asymmetries in precipitation.

The symmetric circulation spins down in an asymmetric manner. The increased boundary layer friction slows down the low-level wind. However the primary reason for the hurricane spindown is the loss of the surface heat flux. The low-level cold branch of inflow air parcels cannot obtain enough moist entropy to maintain the convection, which then reduces the azimuthal mean secondary circulation. An outflow layer just above the boundary inflow develops when the convection becomes weak. The upper-level tangential wind is weakened from the lack of momentum supply by the updraft in the eyewall.

In the context of PV dynamics, the eyewall PV ring is maintained by the large generation of the PV in the eyewall. After landfall, the eyewall PV ring evolves to a monopolar structure. The low-level monopole forms from the stabilization of the low-level atmosphere in the core. The upper monopole results from the nonlinear PV mixing process, which is governed by vortex Rossby waves. In response to the mean PV structure changes, vortex Rossby waves change their behavior.

Using the Eulerian mean equations in storm-following cylindrical and isentropic coordinates with boundary extensions, the EP flux and its divergence describing the inner core asymmetries (mainly vortex Rossby waves) are discussed. Three branches of the eddy momentum fluxes related to the horizontal component of the EP flux are identified. The changes in the eddy heat fluxes associated with the vertical component of the EP flux reflect the thermal structural evolution during landfall. The mean tangential wind changes due to the wave-mean flow interaction, as well as other mean circulation and eddy terms are diagnosed. It is found that the diabatic heating has the greatest contribution to the intensification of the hurricane, while the eddy effect (evaluated as the total effect of the wave-mean flow interaction and the asymmetric diabatic heating) is also significant. The eddies tend to counteract the mean circulation. During the rapid intensification stage, the eddies accelerate the tangential wind inside the RMW and decelerate the wind at the RMW. During the weakening stage, the eddies prevent the fast decay of the hurricane by accelerating the maximum tangential wind.

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