

## REVIEW

### Clouds in Tropical Cyclones

ROBERT A. HOUZE JR.

*Department of Atmospheric Sciences, University of Washington, Seattle, Washington*

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#### ABSTRACT

Clouds within the inner regions of tropical cyclones are unlike those anywhere else in the atmosphere. Convective clouds contributing to cyclogenesis have rotational and deep intense updrafts but tend to have relatively weak downdrafts. Within the eyes of mature tropical cyclones, stratus clouds top a boundary layer capped by subsidence. An outward-sloping eyewall cloud is controlled by adjustment of the vortex toward gradient-wind balance, which is maintained by a slantwise current transporting boundary layer air upward in a nearly conditionally symmetric neutral state. This balance is intermittently upset by buoyancy arising from high-moist-static-energy air entering the base of the eyewall because of the radial influx of low-level air from the far environment, supergradient wind in the eyewall zone, and/or small-scale intense subvortices. The latter contain strong, erect updrafts. Graupel particles and large raindrops produced in the eyewall fall out relatively quickly while ice splinters left aloft surround the eyewall, and aggregates are advected radially outward and azimuthally up to 1.5 times around the cyclone before melting and falling as stratiform precipitation. Electrification of the eyewall cloud is controlled by its outward-sloping circulation. Outside the eyewall, a quasi-stationary principal rainband contains convective cells with overturning updrafts and two types of downdrafts, including a deep downdraft on the band's inner edge. Transient secondary rainbands exhibit propagation characteristics of vortex Rossby waves. Rainbands can coalesce into a secondary eyewall separated from the primary eyewall by a moat that takes on the structure of an eye. Distant rainbands, outside the region dominated by vortex dynamics, consist of cumulonimbus clouds similar to non-tropical storm convection.

#### 1. Introduction

It is tempting to think of clouds in the inner regions of tropical cyclones as cumulonimbus that just happen to be located in a spinning vortex. However, this view is oversimplified, as the clouds in a tropical cyclone are intricately connected with the dynamics of the cyclone itself. Perhaps up to now it has not been important to know the detailed inner workings of tropical cyclone clouds. However, as high-resolution full-physics models become more widely used, forecasting the probabilities of extreme weather and heavy precipitation at specific times and locations will become an increasingly feasible goal for landfalling cyclones. In addition, prediction of sudden and rapid changes in storm intensity remains perhaps the greatest challenge in tropical cyclone preparation and

warning. Rapid intensity changes often occur in conjunction with rapid reorganization of the cyclone's mesoscale cloud and precipitation structures. Realistic simulation of the clouds and precipitation, therefore, is a particularly important aspect of the models' ability to make these forecasts accurately and for the right physical and dynamical reasons.

Accurate representation of tropical cyclone clouds in numerical models is also important from a climate dynamics standpoint. Tropical cyclones have been at the center of discussions of global warming effects. Although the issue remains unresolved, tropical cyclones may increase in number and/or intensity as the earth warms. Since tropical cyclones are major producers of both cloud cover and precipitation in the tropics and subtropics, the correct prediction of tropical cyclone behavior in global climate models will depend ultimately on the accuracy with which tropical cyclone clouds are represented.

For these reasons, it seems appropriate to synthesize and organize the available information on the diverse cloud processes within tropical cyclones. This review

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*Corresponding author address:* Robert A. Houze Jr., Dept. of Atmospheric Sciences, University of Washington, Box 351640, 610 ATG, Seattle, WA 98195.  
E-mail: houze@atmos.washington.edu

presents a dynamical and physical description of the main types of clouds in a tropical cyclone with an emphasis on how they relate to the dynamics of the cyclone itself. The paper will proceed by first examining the cloud structures associated with the formation of a tropical cyclone. Then the review will consider in detail how the mean symmetric structure of the vortex affects cloud structure. Following that will be a discussion of how cloud structures are affected and controlled by storm asymmetries, including those brought on by the storm's translation, the large-scale environmental wind shear, the dynamics of rainbands (which may evolve into a new eyewall) and embedded convection. The review will extend to an even smaller scale by examining the microphysics and electrification of clouds within tropical cyclones.

## 2. Definitions, climatology, and the synoptic-scale contexts of tropical cyclones

According to the *Glossary of Meteorology* (Glickman 2000), a tropical cyclone is any low pressure system having a closed circulation and originating over a tropical ocean. It is further categorized according to its peak wind speed: a *tropical depression* has winds of  $<17 \text{ m s}^{-1}$ , a *tropical storm* has a peak wind of  $18\text{--}32 \text{ m s}^{-1}$ , and a *severe tropical cyclone* has a peak wind of  $33 \text{ m s}^{-1}$  or more.<sup>1</sup> In this article, we will use the term *tropical cyclone* to refer primarily to the stronger tropical storms and severe tropical cyclones. Severe tropical cyclones have local names. They are called *hurricanes* in the Atlantic and eastern North Pacific Oceans, *typhoons* in the western North Pacific Ocean, and *cyclones* in the South Pacific and Indian Oceans. They are all the same phenomenon.

Gray (1968) identified most of the environmental factors favoring tropical cyclones. They originate over oceans, as their primary energy source is the latent heat of water vapor in the atmospheric boundary layer. They nearly always form over regions where the sea surface temperature (SST) exceeds  $26.5^\circ\text{C}$  (Fig. 1a). Once tropical cyclones form, they tend to be advected by the large-scale wind. Figure 1b shows how the tracks are generally westward at lower latitudes, where easterlies dominate the large-scale flow. The storms often "recurve" toward the east, when they move poleward into the midlatitude westerlies. When they encounter land or colder water, they die out or (more rarely) transition into extratropical cyclones (Jones et al. 2003).

<sup>1</sup> In the United States the wind speed is considered to be a 1-min average at a standard anemometer level of 10 m. Elsewhere it is a 10-min average. Of course, in a severe tropical cyclone it is not practical to measure the wind by standard methods, and the wind near the surface of the earth must be determined from satellite data or from aircraft flights through the storms.

Although tropical cyclones form at low latitudes, generally between  $5^\circ$  and  $20^\circ$  (Fig. 1a), they rarely form within  $\pm 5^\circ$  of the equator because the Coriolis force is too weak for low-level convergence to be able to generate relative vorticity.<sup>2</sup> Gray (1968) found that genesis regions not only are located off the equator, where the Coriolis force is nonzero, but also have higher-than-average relative vorticity of the surface wind field and temperature stratification that is at least moderately conditionally unstable, and weak vertical shear of the horizontal wind. The strong background positive vorticity helps to trap energy released in convection so that it contributes via generation of potential vorticity to strengthening of the positive vorticity. Weak shear of the background flow allows the cyclone to develop vertical coherence. The climatological presence of strong shear over the South Atlantic, plus the lack of synoptic precursor disturbances, accounts for the extremely rare occurrence of tropical cyclones in that region. Besides the factors highlighted by Gray (1968), the relative humidity of the environment must not be too low (DeMaria et al. 2001). Otherwise, clouds cannot fully develop, and without clouds to transport moist static energy upward from the boundary layer in contact with the ocean, there can be no tropical cyclone.

The large-scale preconditions for tropical cyclogenesis noted by Gray (1968) and DeMaria et al. (2001) may be set up by a variety synoptic-scale processes. Bracken and Bosart (2000), Davis and Bosart (2001), and McTaggart-Cowan et al. (2008) have shown that about half of Atlantic tropical cyclones occur when African easterly waves (Burpee 1972; Reed et al. 1977, 1988; Thorncroft and Hodges 2001) move westward off the African continent, and a tropical cyclone spins up in the trough of the wave, provided the wave trough does not ingest dry midlevel air from the Saharan region (Dunion and Velden 2004). Other Atlantic tropical cyclones develop when midlatitude synoptic-scale dynamics operating over a warm ocean surface set up conditions of large-scale shear and vorticity that lead to tropical cyclogenesis. According to Bracken and Bosart (2000), Davis and Bosart (2001), and McTaggart-Cowan et al. (2008), these midlatitude conditions involve midlatitude troughs and/or fronts extending into low latitudes.

<sup>2</sup> Though extremely rare, tropical cyclone formation can occur near the equator. Typhoon Vamei (2001) developed near Singapore,  $1.5^\circ$  north of the equator, and Cyclone Agni (2004) formed only a few kilometers off the equator. In such cases, the rotation must be derived from some preexisting source of relative vorticity. Analyzing Vamei, Chang et al. (2003) found that the storm drew on the vorticity of two interacting cyclonic disturbances that had moved into the region. The authors estimated that the probability of a similar equatorial development is 1 in about every 100–400 yr.

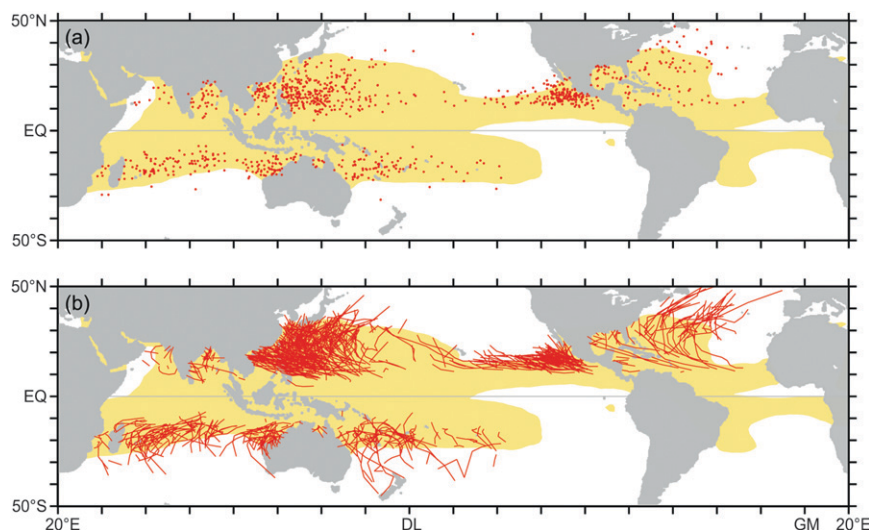


FIG. 1. Locations and tracks of tropical cyclones for 1970–89 relative to the global surface temperature analysis of Legates and Willmott (1990): (a) locations of hurricanes on the first day with hurricane-force winds ( $>32 \text{ m s}^{-1}$ ), and (b) tracks of hurricanes. The shaded oceanic regions are where SST exceeds  $26.5^\circ\text{C}$  in summertime, represented by August in the Northern Hemisphere and February in the Southern Hemisphere. (Provided through the courtesy of T. Mitchell, University of Washington.)

The ultimate formation of a tropical cyclone in a favorable synoptic-scale environment entails the concentration of perturbation kinetic energy into a low pressure system with a horizontal scale of a few hundred kilometers. The concentration of energy on this scale may be partly a *downscale* process, whereby a developing storm extracts energy from larger scales of motion. One idea suggested by Webster and Chang (1997) is that when a synoptic-scale wave in the easterlies enters a region of large-scale negative stretching deformation ( $\partial\bar{u}/\partial x < 0$ , where  $\bar{u}$  is the large-scale mean zonal wind component) it will become more energetic, shrink in scale, and be confined to the lower troposphere. However, tropical cyclone development also certainly involves strong *upscale* feedback in which convective clouds connect the intensifying cyclonic circulation with the boundary layer and focus and intensify vorticity locally. Dunkerton et al. (2008) have suggested a mechanism whereby the parent synoptic disturbance contains a subsynoptic-scale subregion within which smaller-scale phenomena can flourish and efficiently accomplish the upscale feedback. They call this the “marsupial theory,” where the favored subregion for development is analogous to the pouch of a mother kangaroo. Their theory applies specifically to the case in which the parent synoptic system (i.e., the kangaroo) is a tropical easterly wave. In the wave’s critical layer (where the speeds of the wave and mean flow are equal), a “pouch” region develops where parcel trajectories form closed cyclonic loops. Clouds forming in this closed-off

region are protected from deleterious large-scale environmental influences such as the intrusion of dry air. Upscale cyclogenesis processes connected with clouds can therefore flourish in this synoptically protected zone, where they are confined and combine to strengthen the cyclonic circulation. If this protected region (or “sweet spot”) of the wave is located in a region of warm SST and minimal large-scale vertical shear, and preexisting low-level positive relative vorticity, convective clouds and precipitation in this region will be especially conducive to concentrating low-level convergence and positive vorticity and strengthening the low. While this marsupial theory of cyclogenesis has been developed in the context of tropical easterly waves, it may be that the nontropical precursor synoptic-scale features identified by McTaggart-Cowan et al. (2008) also form protected zones and/or regions of negative stretching deformation in ways that have not yet been investigated.

### 3. Clouds involved in tropical cyclogenesis

#### a. Idealization of the clouds in an intensifying depression

As noted in the previous section, tropical cyclogenesis is partly an upscale process, with convective-scale dynamics locally adding energy and vorticity to a preexisting cyclonic disturbance of the large-scale flow, in a region synoptically predisposed to development. Convective clouds are central in the upscale positive feedback

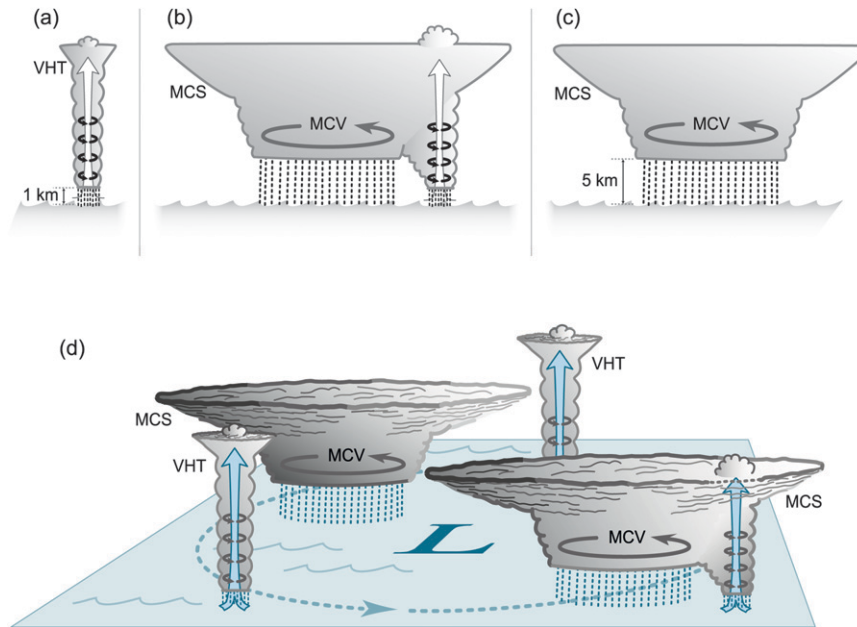


FIG. 2. MCS life cycle within a developing tropical cyclone. (a) The MCS begins as a set of one or more isolated deep VHTs. The vorticity of the low-level environment is stretched by convergence in the lower portions of the updraft and advected upward. (b) The convective-scale cells are transient components of a larger and longer-lived MCS, and as individual vortical hot towers of the MCS die off, they form a precipitating stratiform cloud out of the weakly buoyant upper portions of the old deep cells while new towers form, so that at its mature stage of development the MCS has both convective and stratiform components. The stratiform region accumulates the positive vorticity of the remaining portions of the convective cells to form a midlevel MCV. (c) In the late stages of the MCS life cycle, new cell development ceases while the MCV vorticity remains in the precipitating stratiform cloud region for some hours. (d) An idealized distribution of VHTs and MCSs in various stages of development. The MCS life cycle is adapted from Houze (1982). (Figure is from Houze et al. 2009.)

to tropical cyclogenesis. Steranka et al. (1986) noted from satellite data that prolonged “convective bursts” occurred “in the near region surrounding the depression centers before the maximum winds initially increased.” These bursts of cirrus outflow from convective clouds stand out in satellite imagery and suggest that a very special type of cumulonimbus cloud is involved in tropical cyclogenesis.

Figure 2 illustrates conceptually the variety of convective entities that may exist in the region where a convective burst presages intensification of a preexisting vortex to tropical storm status. Some of these convective entities are individual towers, while others take the form of a mesoscale convective system (MCS), which contains both deep convective cells and stratiform clouds and precipitation (see Houze 2004 for a review of MCSs). The convective entities in Fig. 2d are depicted within a low-level preexisting cyclonic circulation. From Gray’s (1968) climatology we know that cyclogenesis occurs in an environment rich in vorticity at low levels, and the cloud-scale processes depicted

in Fig. 2 can only operate effectively to enhance cyclogenesis in a preexisting region of low pressure and maximum vorticity such as indicated by the L and the dashed cyclonic streamline in the figure. As discussed in section 2, the work of Dunkerton et al. (2008) suggests further that cyclogenesis may be especially effective in a closed cyclonic circulation protected from the larger-scale environment.

Figures 2a–c give a simplified depiction of the life cycle of the cloud-scale MCS as it would occur within an assumed larger-scale environment rich in vorticity at lower levels. The general characteristics of the schematic life cycle are an adaptation of the generic MCS life cycle described by Houze (1982), but with an emphasis on the way vorticity might develop within the MCS. The MCS begins as one or more isolated deep convective towers (Fig. 2a). The vorticity of the low-level environment is stretched by convergence at the base of the buoyant convective updrafts and advected upward. The updrafts thus become upward extending centers of high positive vorticity, which Hendricks et al. (2004) called vortical hot



towers (VHTs; see Fig. 2), as a special case of the “hot tower” terminology introduced by Riehl and Malkus (1958) to describe deep tropical convection in general. As an individual convective tower dies off, it weakens and becomes part of a precipitating stratiform cloud (Houze 1997). New towers form adjacent to the stratiform region, so that at its mature stage of development the MCS has both convective and stratiform components (Fig. 2b). The base of the stratiform cloud deck is in the midtroposphere, as it is mostly composed of the material of the upper portions of convective cells. The vertical profile of heating in the stratiform region leads to the development of a mesoscale convective vortex (MCV) in midlevels in association with the stratiform cloud deck (see Houze 2004). In the case depicted in Fig. 2, the stratiform region is composed of the remnants of convective cells, which themselves have considerable positive vorticity. These vortical hot tower remnants add more vorticity to the MCV. During the middle life cycle state in Fig. 2b, the MCS contains vorticity structures both in the form of convective-scale VHTs and in the form of the wider stratiform-region MCV. In the late stages of the MCS life cycle, the hot towers cease forming, but the stratiform cloud region containing MCV vorticity remains for some hours (Fig. 2c).

### *b. Example of a vortical hot tower*

An example of vortical hot tower has been documented with airborne Doppler radar by Houze et al. (2009) in an MCS located in the depression that became Hurricane Ophelia (2005). The basic structure of the updraft is illustrated in Fig. 3. Doppler radar showed a deep, wide, intense convective cell of a type that has been previously thought to occur in intensifying tropical depressions (Zipser and Gautier 1978) but had not been documented in detail until high-resolution airborne Doppler radar data were collected in the pre-Ophelia depression. The observed updraft of the cell was 10 km wide and 17 km deep and had updrafts of  $10\text{--}20\text{ m s}^{-1}$  throughout its mid- to upper levels (Figs. 3a,b). It contained a cyclonic vorticity maximum (shown at 8-km altitude in Fig. 3c). This vorticity maximum, centered on the updraft, confirms that the cell was a vortical hot tower. The massive convective updraft was maintained by strong positive buoyancy (virtual potential temperature perturbation  $>5^{\circ}\text{C}$  at 10 km; Fig. 3a), probably aided by latent heat of freezing at higher altitudes. The convective updraft was fed by a layer of strong inflow several kilometers deep. Wind-induced turbulence, just above the ocean surface, enriched the equivalent potential temperature of the boundary layer of the inflow air, thus creating an unstable layer with little convective

inhibition. This air was raised to its level of free convection when it encountered the denser air located at low levels in the rainy core of the convection. Although evaporative cooling and precipitation drag occurred in the rain shower of the cell (see the negative temperature perturbation below 2 km in Fig. 3a), the negative buoyancy was insufficient to produce a strong downdraft or gust front outflow to force the updraft. Instead the updraft air was forced initially upward when the ambient low-level wind of the developing depression encountered the density gradient at the edge of the rainshower. This type of updraft formation without the aid of a gravity current outflow from a downdraft is not common, but its possibility of occurrence was pointed out in numerical simulations by Crook and Moncrieff (1988). The lack of a divergent spreading cool-air pool made this intense convection an extremely effective feedback mechanism for cyclogenesis in the pre-Ophelia depression. The production of low-level mass convergence, and associated stretching and upward advection of vorticity in the lower troposphere of the region where cyclogenesis was occurring, was not offset by strong downdraft divergence, and air of low equivalent potential temperature was not transported into the boundary layer to offset the gain of moist static energy from the sea surface.

The dominance of updraft motion and relative lack of downdraft motions in the convective burst of convection in the intensifying depression was also noted by Zipser and Gautier (1978) in their analysis of flight-level data. Nolan (2007) conducted idealized model simulations of convection in a tropical cyclogenesis environment and found that updraft mass transport dominated as the depression approached tropical storm stage. Zipser and Gautier (1978) further noted that the mesoscale convergence feeding the massive convective updraft present in the MCS they investigated was sufficient to account for the intensification of the synoptic-scale depression vortex in which the MCS was located. A sharp increase of upward mass flux with height occurred throughout the lower portion of the updraft in the pre-Ophelia convection (see the isolines of the vertical derivative of the vertical mass flux in Fig. 3b). The concomitant stretching of vorticity and generation of potential vorticity<sup>3</sup> in the lower portion of the updraft, and the strong vertical motions of the updraft advecting the concentrated vorticity vertically, created a deep convective-scale vorticity

<sup>3</sup> The generation of potential vorticity is proportional to the vertical profile of heating (Houze 1993, p. 460, or many other references). In the present case the latent heating is directly proportional to the vertical mass flux. A positive vertical derivative of the mass flux is thus proportional to the potential vorticity generation.

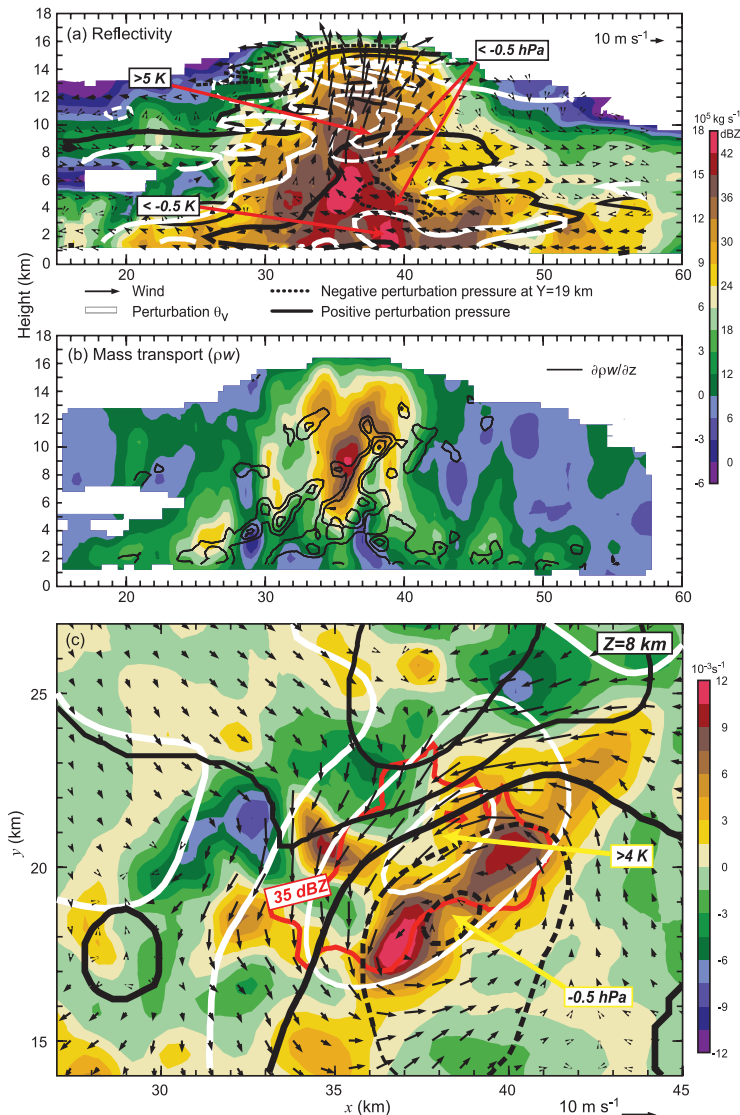


FIG. 3. Analysis of airborne dual-Doppler radar data collected in the cyclogenesis phase of the tropical depression that evolved into Hurricane Ophelia (2005). Data were collected between 2108 and 2123 UTC 6 Sep 2005. Cartesian domain is relative to National Hurricane Center best-track position of Ophelia at 0000 UTC 7 Sep 2005. The vertical cross sections in (a) and (b) are taken along  $y = 19 \text{ km}$  in (c). (a) Reflectivity (color; dBZ) with wind vectors, and overlaid with the buoyancy field in units of virtual potential temperature perturbation ( $\theta'_v$ ; white lines, with 2.5-K contours) and perturbation pressure field (black lines, with 0.5-hPa contours). Thicker lines indicate zero contours, solid lines are positive perturbations, and dashed lines are negative perturbations. Velocity vectors are in the  $x$ - $z$  plane. (b) Mass transport (color;  $10^5 \text{ kg s}^{-1}$ ) and areas of positive vertical mass transport gradient ( $2 \times 10^5 \text{ kg s}^{-1} \text{ m}^{-1}$ , with contours starting at  $1 \times 10^5 \text{ kg s}^{-1} \text{ m}^{-1}$ ). (c) Horizontal pattern at the 8-km level of fields derived from the airborne dual-Doppler data. Relative vertical vorticity (color;  $10^{-3} \text{ s}^{-1}$ ) is shown with perturbation wind vectors, and overlaid with the buoyancy field in units of virtual potential temperature perturbation ( $\theta'_v$ ; white lines, with 1-K contours) and perturbation pressure (black lines, with 0.5-hPa contours). The buoyancy was derived from the wind field and its change with time during the period of observations. Thicker lines indicate zero contours, solid lines are positive perturbations, and dashed lines show negative perturbations. The red contour surrounds the region in which reflectivity was  $>35 \text{ dBZ}$ . (Figure is from Houze et al. 2009.)

perturbation manifested as a convective-scale cyclonic vortex—that is, a vortical hot tower. This vortex is evident in Fig. 3c.

The vorticity field shown in Fig. 3c shows both negative and positive vorticity centers near the center of the updraft. This structure is consistent with the updraft tilting the environmental horizontal vorticity into the vertical to produce a vortex couplet in midlevels, as described by Rotunno (1981) and Montgomery et al. (2006). For simplicity, this tilting-produced couplet is not shown in Fig. 2a since the positive member of the tilting-produced couplet dominates, probably because it combines with the positive vorticity of the stretched and upward-advected boundary layer vorticity to produce deep updrafts that are centers of cyclonic rotation. Moreover, numerical experiments suggest that the negative vorticity anomalies tend to be expelled from the developing tropical cyclone vortex while the positive anomalies are retained (Montgomery and Enagonio 1998). Thus, the negative anomalies are not relevant to the developing cyclone as positive feedback elements.

#### *c. Ensemble of clouds in a developing storm*

The ensemble of clouds in the intensifying depression shown in idealized form in Fig. 2d is based on real examples. Simpson et al. (1997) and Ritchie et al. (2003) showed satellite imagery of MCSs rotating around the center of the developing Southern Hemisphere Tropical Storm Oliver (1993). Sippel et al. (2006) described a similar set of MCSs seen in both satellite imagery and coastal radar observations as the newly formed Tropical Storm Allison (2001) made landfall in Texas. The MCSs in Allison were distributed around the developing cyclone, and the coastal radar showed positive vorticity anomalies in the active convective elements of the MCSs, probably similar to the vortical hot tower illustrated in Fig. 3.

Convective-ensemble behavior similar to Allison and Oliver occurred during the genesis stage of Hurricane Ophelia (2005). Figure 4 shows infrared satellite and coastal radar imagery just prior to the pre-Ophelia tropical depression intensifying to tropical storm intensity. A day before the depression reached tropical storm intensity (Figs. 4a–d), strong convection was prevalent over a wide area, but the clouds and precipitation exhibited no tropical storm–like structure. At the time of Figs. 4c,d, the convection seen on radar had grouped into three MCSs ~200 km in horizontal dimension, each having both active convective cells and areas of stratiform precipitation. The vortical hot tower cell shown in Fig. 3 was located in the MCS located northeast of the Melbourne, Florida, coastal radar. Thus, the pre-Ophelia depression had a cloud population containing intense rotational convective

cells and MCSs scattered about in the low-pressure area as depicted schematically in Fig. 2d.

By the time of Figs. 4e,f, the overall area covered by deep convection in the pre-Ophelia depression had decreased and became focused on a single very intense MCS. In the next several hours, this particular mesoscale precipitation area radically changed its shape and took on the structure of a tropical storm (Figs. 4g,h). Nolan's (2007) idealized model simulations suggest that the long period of moistening by prior convection (e.g., Figs. 4a–d) may facilitate conversion from MCS to tropical storm structure. The radar echo exhibited an incipient eyewall and a well-defined principal rainband (as defined by Willoughby et al. 1984b; Willoughby 1988) extending from south to east to north of the storm and having a convective structure upwind and a stratiform structure downwind, as described in papers on mature hurricane rainbands (Atlas et al. 1963; Barnes et al. 1983; Hense and Houze 2008). Details of principal rainbands will be discussed in section 8c.

#### *d. Cloud feedback in cyclogenesis*

Figure 2d depicts an idealized scenario in which a population of clouds occurs in a region where large-scale conditions have dictated the presence of a preexisting low-level weak cyclonic circulation. The cloud population within this region consists of a combination of isolated deep convective cells with cyclonic vorticity maxima in the updrafts (as in Fig. 2a), one or more mature MCSs with both cyclonically rotating convective cells and MCVs (as in Fig. 2b), and older MCSs with residual MCVs (as in Fig. 2c). Each of the clouds contains significant vorticity perturbations in the form of convective-scale vortical hot towers and/or stratiform-region MCVs. These in-cloud vorticity perturbations can feed back positively to the larger-scale cyclonic vorticity. The deep convection generates potential vorticity. Collectively, the clouds accumulate and distribute vorticity derived from the boundary layer and low levels of the free environment throughout a deep layer. By these means they help to organize the preexisting weaker synoptic-scale cyclonic circulation into a tropical storm. The background positive vorticity helps to reduce the Rossby radius (Houze 1993, p. 53) within which the effect of the clouds is combined.

Tropical cyclogenesis is not just a question of strengthening the preexisting vortex. The vortex must reorganize to become a tropical cyclone. An important question is how a cloud population such as that depicted conceptually in Fig. 2d, or shown by example in Fig. 4, helps convert a preexisting benign synoptic-scale cyclonic circulation to a structure having the specific configuration of a tropical cyclone. One of the distinctive features of a tropical

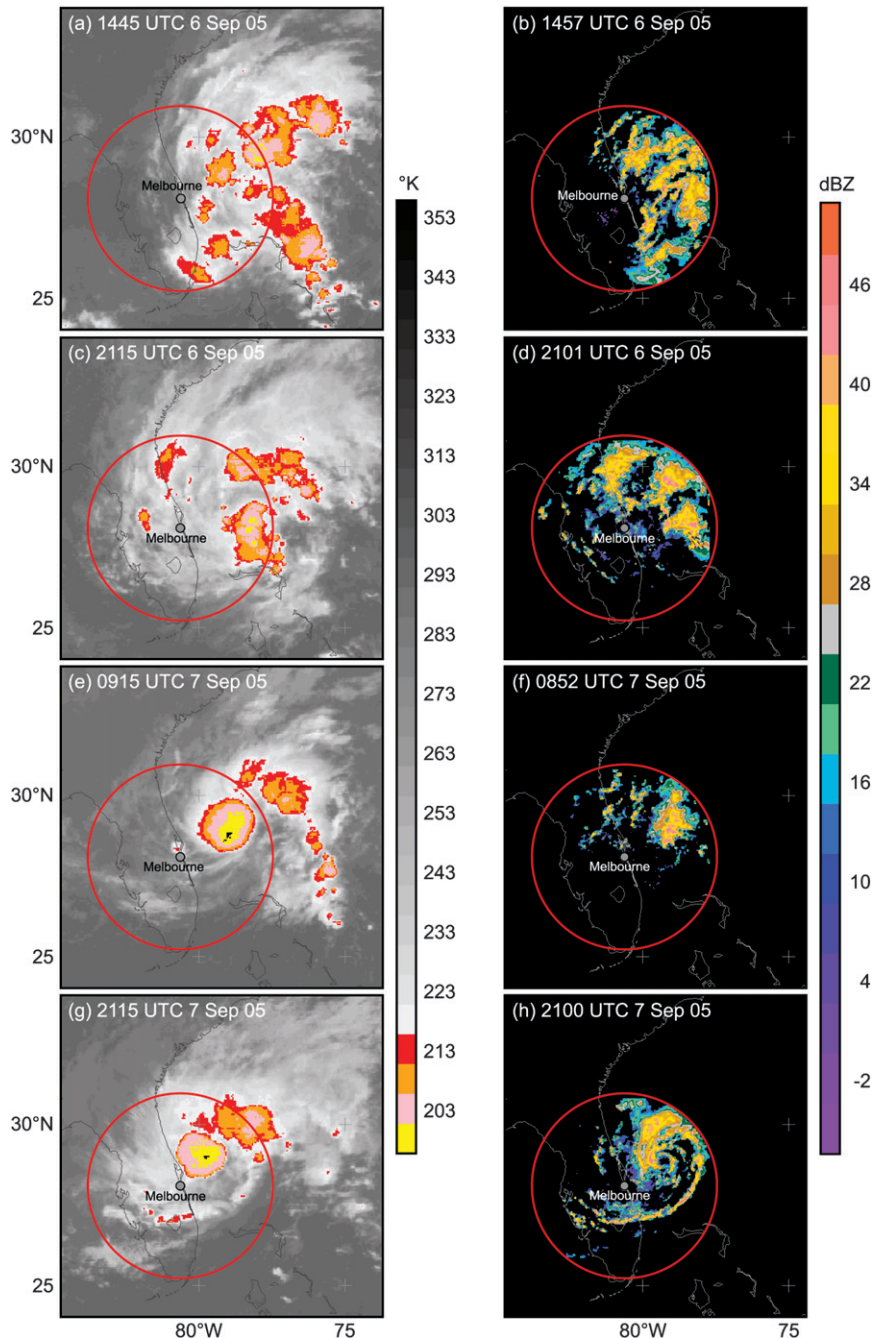


FIG. 4. Satellite and radar overview of the genesis of Ophelia. The circle shows maximum radar range. Satellite imagery shows infrared temperature from *Geostationary Operational Environmental Satellite (GOES)-12*. Radar data are from the Melbourne Weather Surveillance Radar-1988 Doppler (WSR-88D). (Figure is from Houze et al. 2009.)

cyclone is a narrow annulus of maximum wind at some distance (usually between 10 and 100 km) from the storm center. This zone is referred to as the radius of maximum wind (RMW), and it is kept in near-thermal wind balance by a strong secondary circulation, which in turn produces

an *eyewall* cloud. Outside the eyewall precipitation occurs in mesoscale spiral *rainbands*. These cloud and precipitation features of the mature tropical cyclone will be discussed in detail in subsequent sections of this paper. As a cyclonic disturbance reaches tropical storm strength, it



also develops an RMW and incipient eye, eyewall, and rainbands.

The foregoing question then reduces more specifically to how upscale feedback by the cloud population depicted in Fig. 2d helps to convert the preexisting cyclonic circulation to a cyclonic disturbance exhibiting an RMW, eye, eyewall, and rainbands. One idea about how this conversion occurs is that the convective and mesoscale vorticity perturbations contained within the members of the cloud population, as depicted in Fig. 2d, are eventually sheared apart by the radial gradient of the vortex wind and mixed into the mean flow around the low in a process of *axisymmetrization*.<sup>4</sup> Hendricks et al. (2004) and Montgomery et al. (2006) present evidence from a model that the axisymmetrization process distributes subsynoptic-scale vorticity perturbations, such as those contained in the vortical hot towers and MCVs, into a ring of high vorticity at a specific distance from the storm center, giving the enhanced low pressure system an RMW and thus a structure like that of a tropical cyclone.

While axisymmetrization must play a role, a remaining curious fact is that the tropical cyclone center has been observed to occur definitively within a *particular member* of the cloud population rather than as a collective smearing of the effects of the whole population around the preexisting low (Fig. 4h). Bister and Emanuel (1997) have suggested that cooling due to melting and evaporation of precipitation below the base of the stratiform cloud is involved in the extension of the middle level MCV downward. However, their idea requires a particularly strong stratiform-region downdraft, and observations of MCSs in intensifying tropical depressions indicate that these MCSs have relatively weak downdrafts (Zipser and Gautier 1978; Houze et al. 2009). Another way for one MCS to become a storm center was demonstrated in a numerical model of a midlatitude MCS by Chen and Frank (1993). In their simulation, the MCV strengthened and built downward as a result of humidification in the region of the MCV; the humidification lowered the buoyancy frequency and thus led to a reduction in the Rossby radius. This occurred in the absence of a lasting downdraft. Finally, there may be a stochastic aspect to the process. As the larger de-

pression gathers strength from all the ongoing convective activity and gradually takes on tropical storm structure in its wind field, the probability increases that one of the MCSs with rotational convective cells and/or MCV will occur in the ideal spot (presumably the exact center of the depression) where the MCS's cloud structure interacts with the vortex center to metamorphose into a structure that has an incipient eye, eyewall, and spiral rainbands.

## 4. Overview of the mature tropical cyclone

### a. Visible clouds

In satellite pictures, the clouds in a mature tropical cyclone are dominated by upper-level cirrus and cirrostratus spiraling anticyclonically outward. In the visible image of Hurricane Katrina (2005) in Fig. 5a, the rough tops of convective clouds penetrate through the smooth cirrus outflow. The most striking feature is the open *eye* of the storm located in the middle of the spiraling cloud pattern. A zoomed-in satellite view of the eye in Fig. 5b shows that the cloud top surrounding the eye slopes downward toward the ocean surface. Seen from an aircraft flying inside the eye (Fig. 5c), the cloud surface bounding the eye region and sloping at about a 45° angle gives an observer on the plane the impression of being inside a giant circular sports stadium with the grandstand banking upward and outward. This huge sloping cloud bank, called the *eyewall*, is highlighted by the sunlight on the east side of the eye in all three panels of Fig. 5. The ocean surface is not visible in Figs. 5b,c. Rather it is obscured by low stratus or stratocumulus clouds. This low cloud cover is common in the eye of a strong tropical cyclone. The dynamics of clouds in the eye and eyewall regions will be discussed further in sections 5–7.

### b. Three-dimensional wind field

The visible cloud pattern seen in Fig. 5 is determined by the wind and thermodynamic structure of the cyclone. An example of the low-level wind field in a tropical cyclone (Gloria 1985) is shown in Figs. 6a,b. The winds were constructed from rawinsonde, dropsonde, and aircraft Doppler radar measurements. In Fig. 6a, the wind data have been filtered to remove wavelengths less than about 150 km near the center of the storm and 440 km in the outer portions of the figure. This view emphasizes the large-scale flow pattern in which the storm is embedded. In Fig. 6b, the analysis retains wavelengths down to about 16 km in the center of the figure and down to about 44 km in the outer part of the

<sup>4</sup> Axisymmetrization terminology arose in the fluid dynamical field of vortex dynamics. It refers to the tendency for a smoothly distributed vortex when perturbed asymmetrically to relax to a more circular configuration through a combination of wave and mixing processes. The idea was first studied by Melander et al. (1987) and was subsequently linked with the azimuthal and radial propagation of Rossby-like waves within the vortex and their ensuing wave–mean-flow interaction by Montgomery and Kallenbach (1997). This terminology has become common in the tropical cyclone literature.

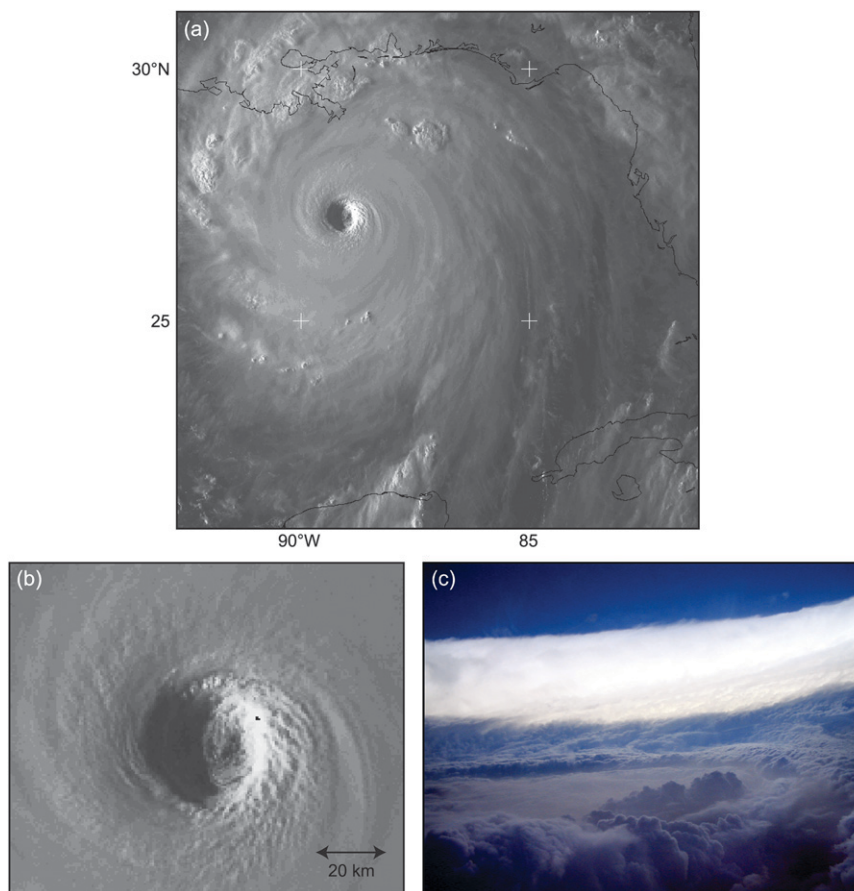


FIG. 5. (a) Visible wavelength satellite view of Hurricane Katrina at 2230 UTC 28 Aug 2005. (b) Zoomed-in satellite view of the eye. (c) View of the eye at 2334 UTC from an aircraft flying inside the eye region (photograph by F. Roux). In all three views, note the late-afternoon sunlight reflecting off the eastern portion of the eyewall.

figure. In this higher-resolution analysis, the tropical cyclone vortex itself is highlighted. The streamlines show boundary layer air spiraling inward toward the center of the storm. The shaded isotachs (Fig. 6b) show the roughly annular zone corresponding to RMW at  $\sim 20$  km ( $0.18^\circ$  latitude) from the storm center. The air parcels spiraling inward at 900 hPa tend to increase their angular momentum, and the RMW occurs where the radial inflow rate abruptly slows down. Radial convergence, increased tangential wind speed, and sudden upward turning of the air current occur at this location and produce the eyewall cloud. Inside the RMW is the eye, where the wind speeds drop off almost immediately to nearly zero, and the vertical air motion is downward, producing the eye of the storm by suppressing clouds, except for the low-level stratus capping the mixed layer (Fig. 5). The dynamics of the eye are discussed further in section 6.

The 200-hPa wind pattern (Figs. 6c,d) illustrates that although the tropical cyclone's outflow is strong, it is

generally asymmetric. In this case, the outflow is concentrated in a channel northeast of the storm center. The depth of the cyclone core is also apparent; even at this high altitude, the flow is cyclonic near the storm center but changes to anticyclonic  $\sim 100$  km from the storm center. Details of the cyclonic flow near the center of the storm are illustrated by the high-resolution analysis in Fig. 6d.

Vertical cross sections of the broadscale mean radial  $u$  and tangential  $v$  components of the wind of a tropical cyclone are shown in Figs. 7a,b, which are composites of data collected in many storms. Most evident in the radial wind field (Fig. 7a) is the increase in the inward directed component at low levels as one approaches the storm center. Strong radial outflow is evident at the top of the storm, at about the 200-hPa level. An important feature missed by these old, low-resolution composite observations is the occurrence of a shallow radial outflow layer at the top of the boundary layer, which is related to the fact that the strong winds near the center of the vortex become

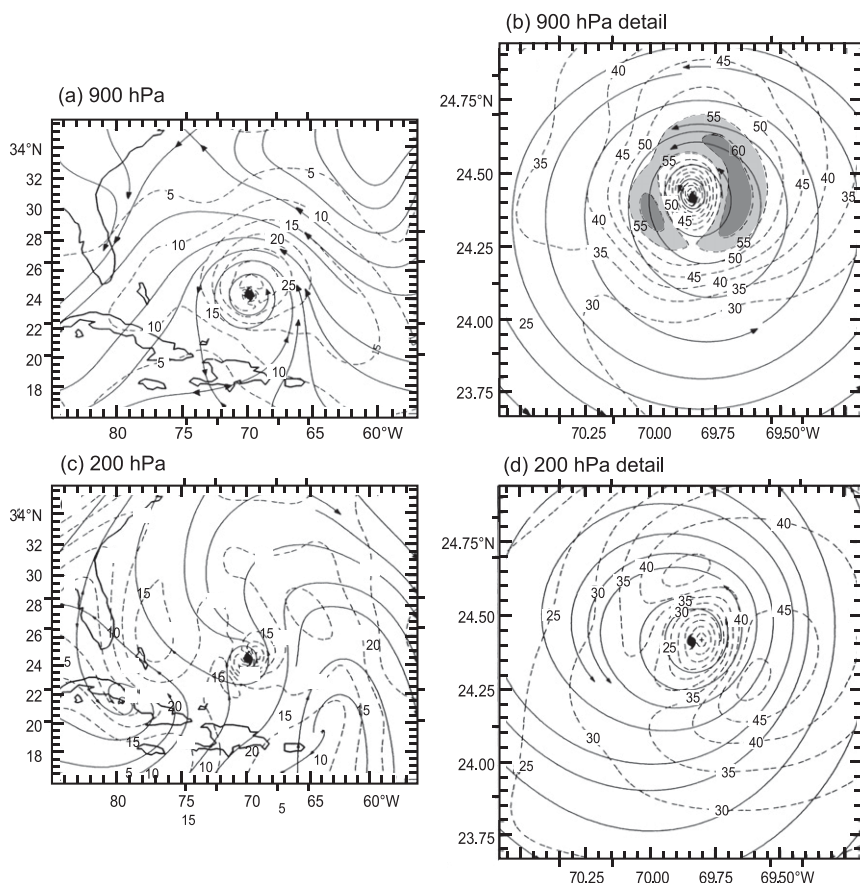


FIG. 6. Wind field associated with Hurricane Gloria (1985). (a) Large-scale flow at 900 hPa. Tick marks indicate boundaries of the three nested rectangular domains defined for the analysis; in the inner domain, wavelengths of  $<\sim 150$  km have been filtered out. In the intermediate and outer domains, wavelengths of  $<\sim 275$  and  $440$  km have been removed. (b) High-resolution wind analysis at 900 hPa, in which wavelengths of  $<\sim 16$ ,  $28$ , and  $44$  km have been filtered out in the three successively larger domains, whose boundaries are indicated by tick marks. Solid lines with arrows are streamlines. Dashed lines are isotachs ( $\text{m s}^{-1}$ ). Winds of  $>55 \text{ m s}^{-1}$  are light gray and  $>60 \text{ m s}^{-1}$  are dark gray. (c) Large-scale flow at 200 hPa. Tick marks indicate boundaries of three nested rectangular domains defined for the analysis; in the inner domain, wavelengths of  $<\sim 150$  km have been filtered out. In the intermediate and outer domains, wavelengths of  $<\sim 275$  and  $440$  km have been removed. (d) High-resolution wind analysis at 200 hPa, in which wavelengths of  $<\sim 16$ ,  $28$ , and  $44$  km have been filtered out in the three domains whose boundaries are indicated by tick marks. Solid lines with arrows are streamlines. Dashed lines are isotachs ( $\text{m s}^{-1}$ ). [Provided through the courtesy of J. Franklin, U. S. National Oceanic and Atmospheric Administration. See also Franklin et al. (1993).]

supergradient (Smith et al. 2008; this will be discussed in section 6). The strongest tangential wind field (Fig. 7b), defining the RMW, is at very low levels. The average dropsonde profile obtained on aircraft penetrations of tropical cyclones is characterized by a broad maximum centered  $\sim 500$  m above the surface (Franklin et al. 2003). The strong convergence in the eyewall at low levels and the strong outflow aloft must be balanced by strong upward motion. The broadscale pattern of vertical motion

in a tropical cyclone is shown in Fig. 7c. The overall cloud and precipitation amounts are determined by this vertical mass transport, which on average is upward over the region located within  $400$  km of the storm center. However, this large-scale mean vertical motion pattern does not have the spatial resolution to provide much insight into cloud structures, nor does it show the downward motion in the eye. To see these details, special aircraft observations and radar instrumentation are required.





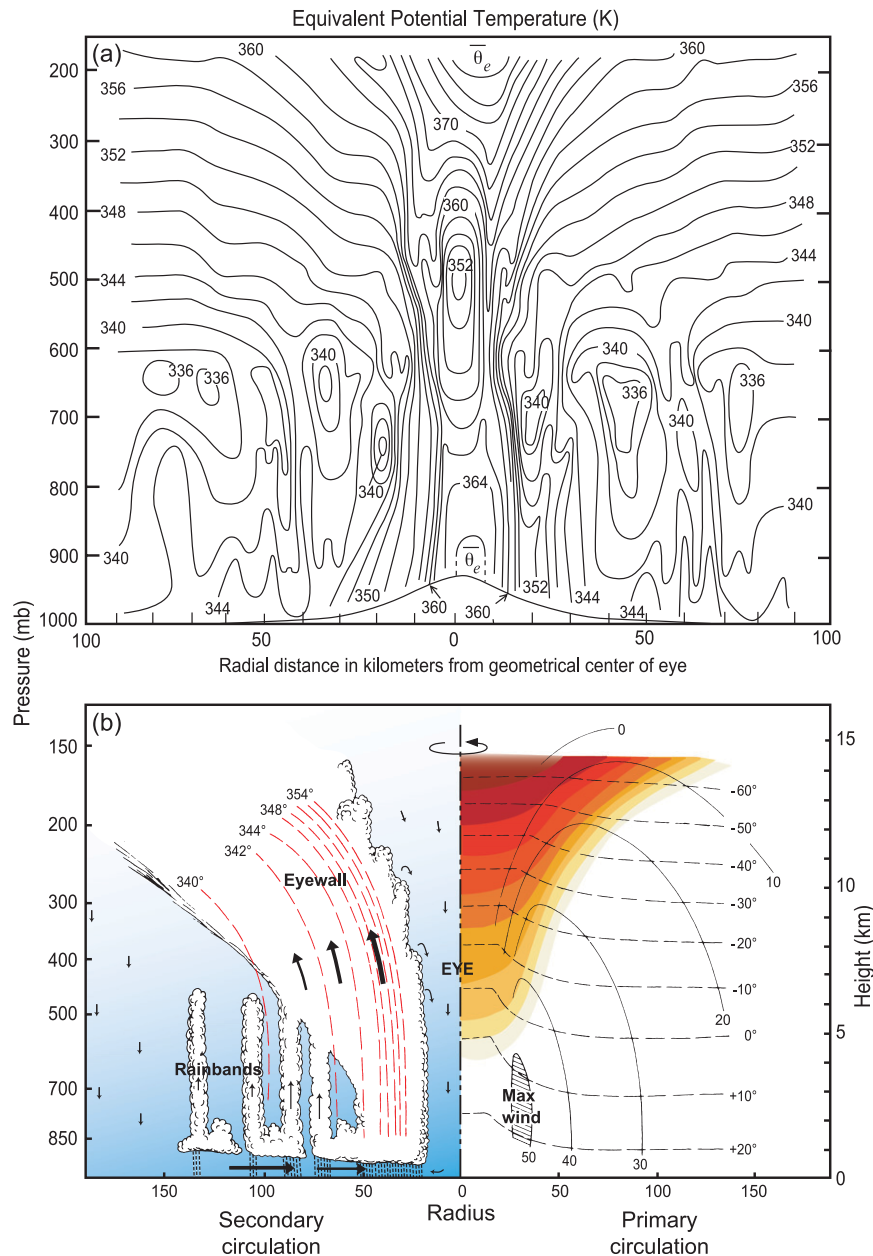


FIG. 8. (a) Equivalent potential temperature  $\theta_e$  in Hurricane Inez (1966). [From Hawkins and Imbembo (1976).] (b) Radial cross section through an idealized, axially symmetric hurricane. On the left the radial and vertical mass fluxes are indicated by arrows and equivalent potential temperature (K) is indicated by dashed lines. On the right tangential velocity ( $\text{m s}^{-1}$ ) is indicated by solid lines and temperature ( $^{\circ}\text{C}$ ) is indicated by the dashed lines. [Adapted from Wallace and Hobbs (2006). Reprinted with permission from Elsevier.]

significantly affects the overall intensity of the tropical cyclone. To fully examine the eye and eyewall dynamics, sections 5 and 6 will discuss how the basic or mean structure of the eye and eyewall cloud can be thought of in terms of the slantwise conditionally symmetrically neutral component of motion in a thermal wind-balanced tropical cyclone vortex. Section 7 will then examine the

buoyancy driven vertical drafts superimposed on the eyewall cloud.

## 5. The eye

As noted in section 4a and illustrated by Fig. 5, the eye is a region of unique and strong dynamics that involves

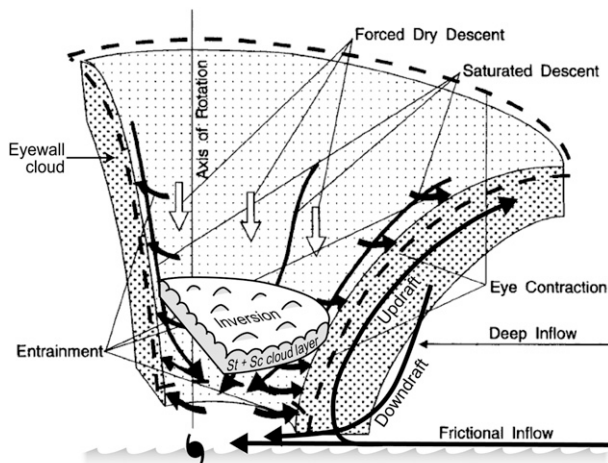


FIG. 9. Schematic illustration of the secondary flow in the eye and eyewall of a hurricane. Dashed lines show an earlier location of the contracting eyewall. [Adapted from Willoughby (1998).]

two distinct and different types of clouds, a layer of low-topped stratus and/or stratocumulus clouds extending horizontally over the eye region and an outward-sloping eyewall cloud. These two types of clouds are depicted schematically in Fig. 9 in relation to the wind field in the tropical cyclone. Both are related to the dynamics of the tropical cyclone vortex. The depiction in Fig. 9 is from a thorough examination of observations in the eye region of hurricanes by Willoughby (1998).

The conditions in the eye region are ideal for the formation of a cloud-topped mixed layer (Lilly 1968; Stevens et al. 2003; Bretherton et al. 2004) because subsidence in the eye produces a stable layer at the top of a vigorous mixed layer over a warm ocean surface. The top of the stratiform cloud with a few cumuliform turrets appears below the aircraft in Fig. 5c at the top of the mixed layer cloud, near the upper extent of the subsidence-induced stable layer.

The sloping eyewall cloud highlighted by sunlight in Fig. 5 is primarily a manifestation of the *secondary circulation*, which, as shown in Fig. 9, is a radial-vertical overturning transverse to the cyclone's *primary circulation*, which refers to the general cyclonic rotation of the wind around the eye. In a balanced tropical cyclone, the secondary circulation consisting of the radial  $u$  and vertical  $w$  components of the wind keeps the primary circulation (i.e., the  $v$  component of the wind) in gradient-wind equilibrium. The low-level radial inflow branch of the secondary circulation gains latent heat energy via upward turbulent flux of sensible and (primarily) latent heat (i.e., moist enthalpy) as it flows over the ocean surface toward the eyewall. Latent heat released in the vertical component of the eyewall circulation provides the energy that

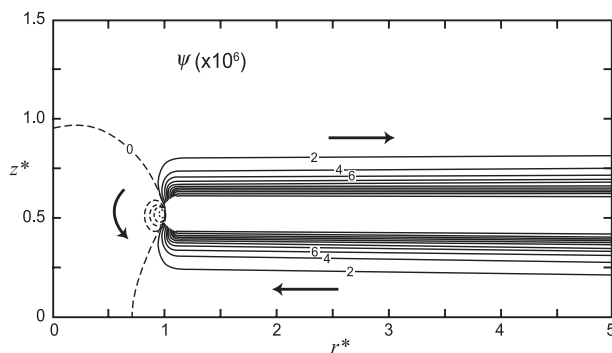


FIG. 10. Mass-flux streamfunction  $\psi$  of the response of an idealized barotropic vortex to a source of heat at nondimensional radius  $r^* = 1$  and height  $z^* = 0.5$ . Contour interval is  $2 \times 10^{-6}$ ; each contour unit ( $1 \times 10^{-6}$ ) is equivalent to  $0.825 \text{ kg m}^{-1} \text{ s}^{-1}$ . (From Shapiro and Willoughby 1982.)

powers the vertical circulation and maintains the tropical cyclone strength and structure.

The secondary circulation is characterized both by rising motion in the eyewall zone and compensating subsidence, occurring partly concentrated in the eye region and partly more diffusely in the far field radially well outside the eyewall. Figure 10 shows the secondary circulation computed by considering point sources of heat and momentum placed near the RMW in an idealized tropical cyclone-like vortex (Shapiro and Willoughby 1982). From this basic result, it is evident that the concentrated subsidence in the eye region is a consequence of the heating in the eyewall cloud. The subsidence in the eye, conceptualized and labeled as “forced dry descent” in Fig. 9, keeps the eye cloud free except for the mixed-layer stratus/stratocumulus at the top of the boundary layer. Pendergrass and Willoughby (2009) and Willoughby (2009) have shown that variations in hypothesized heating patterns associated with the eyewall and different environmental conditions lead to a variety of variations. The latter paper indicates that transient bursts of heating in the eyewall, for example from embedded buoyant convective towers, might produce significant alterations in the storm structure. The subsidence computed for the eye region is accompanied by divergence at low levels within the eye region. Figure 9 indicates that the eyewall is fed in part by the low-level radial outflow from the eye region as well as by radial inflow from outside the eyewall. Important implications of this ingestion by the eyewall of air from the boundary layer of the eye region are discussed in section 7.

Shapiro and Willoughby's (1982) idealized tropical cyclone circulation shown in Fig. 10 is not in a steady state because the RMW contracts over time. Since the RMW is collocated with the eyewall, this behavior is

referred to as *eyewall contraction*. The nonlinear response to the eyewall heat and momentum source has the maximum rate of increase of tangential wind velocity located slightly radially inward of the RMW. This result implies that the RMW (Fig. 8b) of a storm in thermal wind balance should decrease in time, and observations and numerical models confirm that many eyewalls contract (Willoughby et al. 1982; Willoughby 1990, 1998; Houze et al. 2006, 2007). The dashed lines in Fig. 9 indicate an earlier position of the inner edge of the eyewall cloud. As the eyewall contracts, the winds in the eyewall zone increase according to conservation of angular momentum and other factors, and the central pressure of the storm decreases as the pressure gradient associated with the new location and strength of the maximum wind continually adjusts toward gradient-wind balance.

The downward air motion in the eye region and divergence of mass away from the eye at low levels seen in Fig. 9 can be thought of partly as a compensation for the gradual contraction of the eyewall boundary over time. Also contributing to the subsidence in the eye is turbulent mixing of angular momentum at the interface of the rapidly rotating eyewall and more quiescent eye region. This mixing is suggested by the entrainment arrows at the inner edge of the eyewall cloud in Fig. 9. To maintain thermal wind balance in the eye region the increase of tangential wind velocity and relative change in the vertical shear of the horizontal wind in the eye needs to be matched by a stronger horizontal temperature gradient, which can only be provided by adiabatic subsidence warming in the center of the vortex (Emanuel 1997).

Figure 9 includes two types of downdrafts based on close-up observations of tropical cyclones during aircraft penetrations. On the outer side of the eyewall, negatively buoyant precipitation driven downdrafts (see further discussion in section 7d) occur at lower levels. Below cloud base, they merge with the frictional inflow beneath the eyewall and become part of the boundary layer air spreading into the eye region. At higher levels along the inner edge of the eyewall, a thin layer of moist air cascades down the eyewall (Willoughby 1998). Mixing of the dry air in the eye and moist air in the eyewall cloud occurs in this thin layer. Evaporation of cloud particles cools the air and produces a cascade of negatively buoyant air down the side of the eyewall cloud. This cascade is visible as a mist of particles mixed into the dry air adjacent to the eyewall cloud (Fig. 11, from Willoughby 1998). This downdraft on the inner side of the eyewall has also been well documented by aircraft measurements of the vertical velocity (Jorgensen 1984, see also footnote in section 7c).



FIG. 11. Photograph inside the eye of Hurricane Olivia at 2136 UTC 25 Sep 1993, showing the downward cascade of moist air along the inner edge of the eyewall. The moist boundary layer is most clearly visible along the border between the eyewall and the sky that runs from the left center to the upper right. The eyewall slopes away from the camera so that the line of sight is tangent to it along the border. This geometry renders the cascade more visible because it provides a long optical path in the moist air. The mist visible against the sky derives from cloud and precipitation particles mixed a kilometer or two into the eye. [From Willoughby (1998). Photograph by J. Franklin.]

## 6. Dynamics of the mean eyewall cloud

### a. Sloping angular momentum surfaces

The eyewall surrounding the eye of the storm is like no other cloud. Although it is often described as a “convective” cloud, the eyewall does not simply follow the dynamics of ordinary cumulus and cumulonimbus, which arise from purely vertical accelerations releasing buoyant (conditional or potential) instability. Rather, as noted in section 4c, the eyewall cloud is shaped to great measure by slantwise convective motions, where the combination of vertical and horizontal accelerations (responding to both buoyancy and inertial forces) maintain a state of near-conditional symmetric neutrality. In this section, we summarize the dynamics of the slantwise component of the eyewall circulation. Section 7 will describe important buoyant convective perturbations of this structure.

The essential dynamics of the slantwise circulation of the sloping eyewall cloud can be inferred from Emanuel (1986) and Smith et al. (2008), who provide an interconnected theoretical linkage of air-sea interaction at the surface, atmospheric boundary layer structure, the secondary circulation of the balanced vortex, and the moist conditionally symmetrically neutral upflow in the eyewall. Emanuel’s (1986) approach is often parameterized in terms of an approximate Carnot cycle (section 3 of his paper), which gives insight into tropical cyclone intensity. However, the focus of the present review is cloud dynamics, not storm intensity. To understand the air motions

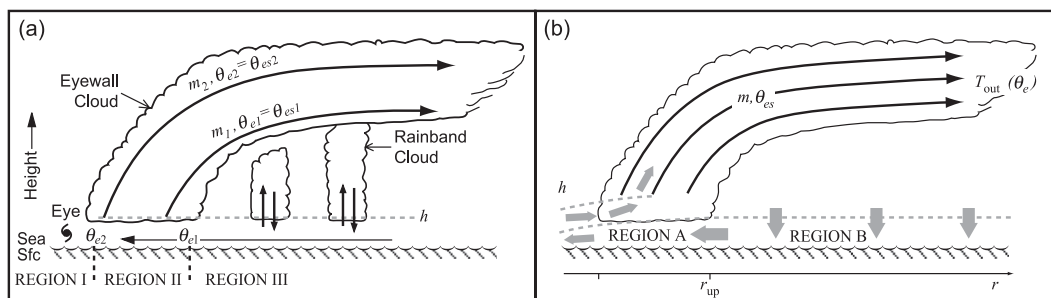


FIG. 12. (a) Idealized air-sea interaction model of a tropical cyclone adapted from Emanuel (1986). Arrows indicate direction of airflow at selected locations. The dashed line represents the top of the boundary layer at height  $h$ . (b) Modification by Smith et al. (2008) of Emanuel's conceptual model. Air subsides into the boundary layer for  $r > r_{up}$  and ascends out of the boundary layer for  $r < r_{up}$ . The frictionally induced net inward force in the outer region produces a radially inward jet at  $r = r_{up}$ . The subsequent evolution of this jet depends on the bulk radial pressure gradient that can be sustained by the mass distribution at the top of the boundary layer. The jet eventually generates supergradient tangential winds whereupon the radial flow rapidly decelerates and turns upward and outward. When the outflow has adjusted to the radial pressure gradient that is sustained by the mass field, the flow turns upward into the eyewall clouds. [Panel (b) is reprinted with permission from the Royal Meteorological Society.]

giving rise to the eyewall cloud, the following discussion makes use of the more explicit version of Emanuel's approach rather than its Carnot parameterization.<sup>5</sup>

The degree to which an  $m$  surface in the eyewall cloud of a balanced tropical cyclone slopes outward can be obtained from the basic equations for a two-dimensional, axisymmetric vortex that is in hydrostatic and gradient-wind balance (see appendix A). For such a vortex, conservation of  $m$  requires that the slope of a particular  $m$  surface be

$$\left. \frac{\partial r}{\partial p} \right|_m = - \frac{\partial m / \partial p}{\partial m / \partial r}, \quad (2)$$

where  $r$  is radius from the storm center and  $p$  is pressure. Thus, the slope of an  $m$  surface is determined by the ratio of the vertical to the horizontal shear of  $m$ . Note also that the right-hand side of (2) is the ratio of the shear of the swirling wind to the local vertical vorticity. In sections 6b–d, it will be shown that with substitution from the thermal-wind equation for a balanced vortex [(A3)], (2) implies that the slope of the  $m$  surfaces with height must be outward, as depicted in Fig. 8b.

Equation (2) provides a basis for obtaining mathematical expressions that describe the dynamics of the eyewall cloud. An expression for  $m(r, p)$  is substituted into (2) to determine the geometry of the streamlines. With the help of the basic equations of a symmetric vortex, one can obtain the magnitudes of the wind components, and the thermodynamic properties of the cir-

culation of an idealized tropical cyclone vortex containing the eyewall cloud. The solution is based on the observation that the tropical cyclone circulation in the eyewall region tends to be conditionally symmetrically neutral [ $\partial \theta_{es} / \partial z = 0$  on a surface of constant  $m$ , where  $z$  is height and  $\theta_{es}$  is the saturation equivalent potential temperature defined by (B1)], but not especially strongly conditionally unstable ( $\partial \theta_{es} / \partial z$  usually only slightly negative in the vertical direction). Consistent with these observations, the eyewall cloud is assumed to be undergoing slantwise, nearly neutral vertical motions above the boundary layer. The ascending air is assumed further to be emanating from a boundary layer whose value of  $\theta_e$  has been determined by upward flux of sensible and latent heat from the ocean surface. For an idealized circularly symmetric tropical cyclone vortex, this approach yields both the primary and secondary circulation by connecting the eyewall dynamics directly to the air-sea interaction that occurs in the boundary layer over the warm ocean.

### b. Boundary layer assumptions and implications

Emanuel (1986) divides the boundary layer of the idealized two-dimensional axisymmetric vortex into the regions I, II, and III shown in Fig. 12a. All three regions are assumed to be well mixed by turbulence and to be of constant depth  $h$ . The production of turbulent kinetic energy in the boundary layer by shear is substantial because of the strong low-level winds circulating around the center of the storm, while the generation of turbulence by buoyancy is strongly promoted by the warm ocean surface. The turbulence produces strong upward eddy flux of latent heat (and to a lesser degree sensible

<sup>5</sup> For a more expanded summary of Emanuel's (1986) theory, see chapter 10 of Houze (1993) and the recent paper by Smith et al. (2008).



heat) everywhere in the lower part of the boundary layer. The flux at the top of the boundary layer, however, is assumed to vary significantly from one region of the storm to the next. In region III, rainbands (to be discussed in section 8) are expected to be important contributors to the turbulent flux at the top of the boundary layer, with convective updrafts and downdrafts transporting high and low  $\theta_e$  out of the boundary layer, respectively. Region II is the region of the eyewall cloud, and the flux at the top of the boundary layer there is assumed to be entirely positive and dominated by the mean upward flux associated with the secondary circulation of the tropical cyclone vortex. Turbulent flux at the top of the boundary layer here is considered to be very small relative to this mean upward flux. Region I is the eye of the storm (section 5).

Emanuel's (1986) simplified treatment of the basics of eyewall dynamics (Fig. 12a) does not consider any interaction between the eyewall and region I. However, in real storms the air diverging out of the eye region at low levels (Fig. 9) can and does feed extremely high- $\theta_e$  air into the base of the eyewall cloud. The eyewall cloud vertical motions are intensified by the entrainment of high- $\theta_e$  air from the boundary layer of the eye region, and local convective cells of buoyant updraft motion can form within the eyewall in locations where the entrainment particularly strongly enhances the buoyancy. To take the entrainment of high- $\theta_e$  air from the eye into account, Smith et al. (2008) replaced Emanuel's (1986) boundary layer model with one that allows for air from the eye region to enter the base of the eyewall cloud (Fig. 12b). Implications for enhancing buoyancy within the eyewall in this manner will be discussed further in sections 6e and 7a–c.

A subtle but significant feature of Emanuel's (1986) boundary layer treatment is that the wind at the top of the boundary layer is assumed to be in gradient balance. Smith et al. (2008) employed a boundary layer model that allows for frictionally induced cross-isobaric flow in the boundary layer. In their model, the frictionally enhanced radial inflow in the boundary layer at the boundary of regions A and B (Fig. 12b) is much stronger than the radial inflow in Emanuel's version. One result is that in region A the tangential wind (into the plane of the cross section) becomes supergradient (as a consequence of the Coriolis force acting on the radial inflow). Outward-directed Coriolis and centrifugal forces acting on the supergradient tangential wind in turn slow the boundary layer inflow. Convergence in region A, resulting in part from this effect, feeds the eyewall cloud with higher- $\theta_e$  air than would be supplied from directly beneath the eyewall as in Emanuel's (1986) model. But the rising supergradient flow turns radially outward in a layer atop the inflow layer as it seeks gradient balance. This low-level

radial flow reversal allows air to flow out of the eye region into the base of the eyewall cloud.

### c. *Connecting the balanced vortex with a simplified boundary layer*

Although Emanuel's (1986) model underestimates the  $\theta_e$  of the air entering the base of the eyewall (cf. Figs. 12a,b), the overall effect of this underestimate is slight on the mean structure of the tropical cyclone. Bryan and Rotunno (2009) show that the parcels of air entering from the eye make the eye only about 0.3 K warmer on average and compose only about 8% of the air rising in the eyewall. The individual parcels entering from the eye do have increased in buoyancy and, as will be discussed in later sections, may account for locally intense updrafts embedded within the eyewall. But since the mean structure of the eyewall cloud is largely accounted for by Emanuel's approach, it remains useful and instructive to utilize his approach to illustrate the basic sloping structure of the streamlines (i.e., the  $m$  and  $\theta_e$  surfaces) in the eyewall cloud. This slantwise circulation remains the primary determinant of the sloping nature of the eyewall cloud, and following this simplified approach we can obtain a view of how the overturning of air required to balance the tropical cyclone vortex takes on the structure made visible by the outward-sloping eyewall cloud (Figs. 5c and 11).

The secondary circulation consistent with the boundary layer regions II and III in Fig. 12a is obtained by considering the equations for a vortex in both hydrostatic and gradient-wind balance (appendix A). The secondary circulation is indicated by mapping the isolines of angular momentum  $m$  in the  $r$ - $p$  plane; since  $m$  is conserved, the  $m$  surfaces are also the trajectories of the mean air motion. Following Emanuel (1986), we obtain the surfaces of angular momentum  $m(r, p)$  implied by the conservation of  $m$  by substituting in (2) and integrating upward, using the value of  $m$  at the top of the boundary layer as a boundary condition and assuming that the mature vortex has adjusted to a state of conditional symmetric neutrality everywhere above the boundary layer.

Since the region above the boundary layer is assumed to have adjusted to conditional symmetric neutrality, the saturation equivalent potential temperature  $\theta_{es}$  (appendix B) is uniform along surfaces of constant  $m$  above height  $z = h$ . The value of  $\theta_{es}$  on the  $m$  surface is assumed to be equal to the value of the boundary layer  $\theta_e$  at the point where the  $m$  surface intersects the top of the boundary layer; that is,

$$\ln \theta_{es} = \ln \theta_e(h), \text{ on an } m \text{ surface.} \quad (3)$$

This assumption assures that a parcel that becomes saturated in the well-mixed boundary layer of the tropical

cyclone will experience neutral buoyancy when displaced along an angular momentum surface extending above the top of the boundary layer—that is, the condition of conditional symmetric neutrality will be maintained in the vortex above the boundary layer, particularly in the eyewall cloud. Emanuel's (1986) assumed relationship of  $m$  and  $\theta_{es}$  above the boundary layer to  $\theta_e$  in the boundary layer is illustrated by Fig. 12a.

Since under conditionally symmetrically neutral conditions there is a 1-to-1 correspondence between thermodynamic variable  $\theta_{es}$  and the angular momentum  $m$ , thermodynamic relations can be employed to relate the radial gradient of  $m$  to thermodynamic variables. The result (see Houze 1993, 418–419, for the derivation) is

$$\frac{\partial \alpha}{\partial r} \bigg|_p = \frac{\partial m}{\partial r} \bigg|_p \frac{\partial T}{\partial p} \bigg|_m \frac{d}{dm} c_p \ln \theta_{es}, \quad (4)$$

where  $\alpha$  is specific volume,  $T$  is temperature, and  $c_p$  is the specific heat at constant pressure. With  $\partial m / \partial r$  given by (4) and  $\partial m / \partial p$  given by the thermal-wind equation in (A6), the slope of the  $m$  surface (2) becomes

$$\frac{\partial r}{\partial p} \bigg|_m = \frac{\partial T}{\partial p} \bigg|_m \frac{r^3}{2m} \frac{d}{dm} c_p \ln \theta_{es}. \quad (5)$$

Integrating (5) along an  $m$  surface from some arbitrary radius  $r$  to  $r = \infty$  leads to

$$\frac{1}{r^2} = \frac{1}{m} \left[ \frac{d}{dm} c_p \ln \theta_e(h) \right] [T_o - T_m(p)], \quad (6)$$

where  $T_m(p)$  is the temperature on the  $m$  surface at pressure  $p$  and  $T_o$  is the outflow temperature on the  $m$  surface (i.e., the temperature at  $r = \infty$ ), and we have made use of (3).

Since it is assumed that the tropical cyclone has adjusted to a state of conditional symmetric neutrality,  $T_m(p)$  in (6) is given by the temperature along the saturation moist adiabat corresponding to  $\theta_e(h)$ . Equations (3) and (6) then allow construction of the fields of  $m$  and  $\theta_{es}$  throughout the region above the boundary layer, provided  $T_o$  and the radial distributions of  $p$ ,  $m$ , and  $\theta_e$  at the top of the boundary layer ( $z = h$ ) are known. To make the calculations tractable, the outflow temperature  $T_o$  is assumed to be a constant, with a value representative of the upper levels of the storm. The rationale for this simplifying assumption is that the streamlines emanating from the eyewall cloud all exit the storm at upper levels (see Emanuel 1986 for further discussion). The radial distributions of  $p$ ,  $m$ , and  $\theta_e$  at  $z = h$  are determined, following Emanuel (1986), by assuming that the temperature at the top of the boundary layer is a constant  $T_B$ .

Then (3), (6), and the gradient-wind equation applied at the top of the boundary layer [the primary assumption removed and revised by Smith et al. (2008)] lead to

$$\begin{aligned} \frac{T_o - T_B}{T_B} \ln \frac{\theta_e(r)}{\theta_{ea}} &= \frac{R_d}{c_p} \ln \frac{p(r)}{p_a} + \frac{R_d}{c_p} \frac{r}{2} \frac{\partial \ln p(r)}{\partial r} \\ &+ \frac{f^2}{4c_p T_B} (r^2 - r_a^2), \quad z = h, \end{aligned} \quad (7)$$

where the subscript  $a$  indicates evaluation at a radius far distant from the storm center and  $R_d$  is the gas constant for dry air. This equation gives one relation between  $\theta_e(r)$  and  $p(r)$  at the top of the boundary layer. If  $\theta_e(r)$  at the top of the boundary layer  $z = h$  can be determined independently, then  $p(r)$  at  $h$  will be determined by (7), and  $m(r)$  at  $h$  will follow from the gradient wind equation [(A3)]. Emanuel (1986) obtains  $\theta_e(r)$  at  $h$  by modeling the effect of boundary layer fluxes via the bulk aerodynamic formula (Roll 1965, p. 251; Stull 1988, p. 262) together with (6) and the assumption that the fluxes at the top of the boundary layer [ $\tau_A(h)$ ] are negligible to obtain

$$\begin{aligned} \ln \theta_e &= \ln \theta_{es}(\text{SST}) \\ &- \frac{C_D}{C_\theta} [c_p(T_B - T_o)]^{-1} \left( v^2 + \frac{frv}{2} \right), \quad z = h, \end{aligned} \quad (8)$$

where  $C_\theta$  and  $C_D$  are mixing coefficients for moist static energy and momentum. For details of the derivation, see Emanuel (1986) or Houze (1993, 416–422). When this expression is substituted into (7) for  $\ln \theta_e$  and  $v$  is expressed in terms of radial pressure gradient by means of the gradient wind equation in (A3), (7) becomes a differential equation for  $p(r)$  at  $z = h$ . Thus, (7) and (8) form a set of simultaneous equations for  $\theta_e$  and  $p$  at the top of the boundary layer. Again, it should be noted that the solutions will slightly underestimate the value of  $\theta_e$  in the eyewall since Emanuel's (1986) model does not allow interaction with the eye of the storm (Bryan and Rotunno 2009).

#### d. Emanuel's solutions for the $m$ and $\theta_{es}$ surfaces in the eyewall cloud

The aim of this paper is to identify the characteristics of clouds of tropical cyclones that distinguish them from other types of convective clouds. Emanuel's (1986) approach to determining the pattern of  $m$  and  $\theta_e$  surfaces is valuable in this regard because it shows that the basic slantwise nature of the eyewall is an unavoidable feature of a strong vortex that is simultaneously in thermal wind balance, conditionally symmetrically neutral, and drawing energy from the sea surface. While buoyant convective motions may be intermittently and importantly

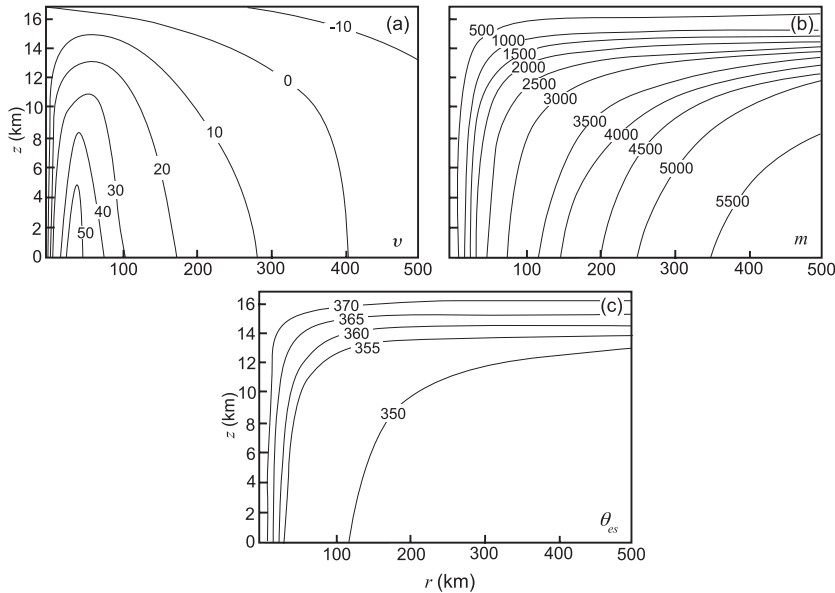


FIG. 13. Structure of a tropical cyclone as calculated with the air-sea interaction model of Emanuel (1986). (a) Gradient wind  $v$  ( $\text{m s}^{-1}$ ) in the air-sea interaction model of a tropical cyclone. (b) Absolute angular momentum  $m$  ( $10^3 \text{ m}^3 \text{ s}^{-1}$ ) in the air-sea interaction model of a tropical cyclone. (c) Saturation equivalent potential temperature  $\theta_{es}$  (K) in the air-sea interaction model of a tropical cyclone. Values of specified variables are  $\text{SST} = 300 \text{ K}$ ,  $T_B = 295 \text{ K}$ ,  $T_o = 206 \text{ K}$ , ambient low-level relative humidity is 80%, latitude is  $28^\circ$ ,  $p_a = 1015 \text{ hPa}$ ,  $r_a = 400 \text{ km}$ , and  $C_\theta = C_D$ . (From Emanuel 1986.)

superimposed on it, the eyewall cloud owes its fundamental slantwise convective nature to the vortex itself. Equations (3), (6), (7), and (A3) are the heart of Emanuel's approach, and they yield the pattern of  $m$  and  $\theta_e$  surfaces, provided that the SST, upper-level outflow temperature  $T_o$ , ambient surface relative humidity, and several other quantities ( $f$ ,  $p_a$ ,  $r_a$ ,  $C_D/C_\theta$ , and  $T_B$ ) are specified. The assumption about relative humidity is the most questionable. The relative humidity is directly proportional to the ratio  $\theta_e/\theta_{es}(\text{SST})$  implied by (8). Emanuel (1986) assumes a fixed value of this ratio in region III. The assumption about relative humidity only works because of the above discussed unrealistic assumption of gradient-wind balance at the top of the boundary layer (Smith et al. 2008). Furthermore, he assumes a constant relative humidity of 80%, whereas buoy data analyzed by Cione et al. (2000) show that the mean relative humidity varies significantly as a function of distance from the center of a tropical cyclone, increasing from  $\sim 85\%$  at  $\sim 200 \text{ km}$  from the storm center to  $\sim 95\%$  within  $\sim 50 \text{ km}$  of the storm center. Despite these serious limitations regarding the quantitatively realistic nature of the eyewall cloud, Emanuel's (1986) assumptions nevertheless lead to an uncluttered view of the basic factors controlling the prevailing slantwise nature of vertical motions in the eyewall cloud. For further details of the solution strategy for the

simplified model, see Emanuel (1986) or chapter 10 of Houze (1993). For a more quantitatively accurate treatment see Smith et al. (2008).

An example of the results of this type of calculation is shown in Fig. 13. The location of maximum winds near the center of the storm (Fig. 13a) is consistent with Fig. 7b, although the winds are stronger in the calculated case than in the composite section, where averaging from many cases has smeared the basic pattern. The modification of the Emanuel model by Smith et al. (2008) indicates that the maximum winds could be even stronger if the interaction with the eye region (Fig. 12b) were taken into account. The  $m$  and  $\theta_{es}$  surfaces in Figs. 13b,c may be regarded as the streamlines of the radial cross-sectional flow above the boundary layer, and near the center of the storm they depict the streamlines of flow within the eyewall cloud. Although the interaction with the eye region is ignored, the  $m$  and  $\theta_{es}$  surfaces of this stripped down model of the symmetric vortex provide a useful view of the essential features that produce the eyewall cloud of a tropical cyclone—namely, the overturning required to keep the storm in thermal-wind balance when the boundary layer is well mixed and the circulation above the boundary layer is conditionally symmetrically neutral. Near the center of the storm, where the  $m$  and  $\theta_{es}$  surfaces depict the upward and outward air motion within the

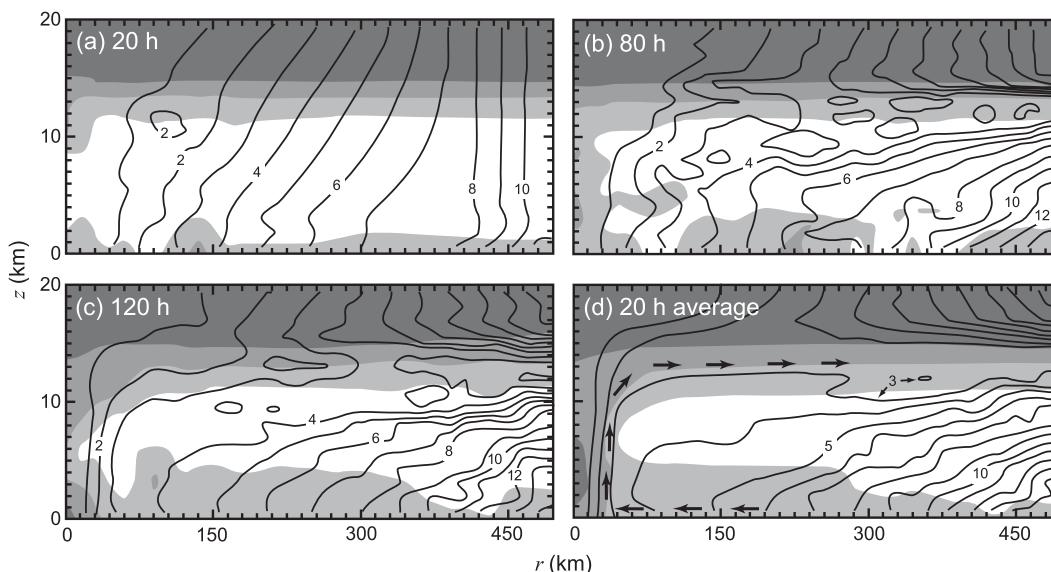


FIG. 14. Results of a two-dimensional nonhydrostatic numerical-model simulation of a tropical cyclone. (a)–(c) Saturation equivalent potential temperature (shaded) and absolute angular momentum (nondimensional units) for three times during the simulation. Closed momentum contours have a value of 3. Shading indicates equivalent potential temperature values of 345–350 K (horizontal lines), 350–360 K (medium shading), and >360 K (heavy shading). (d) Average over a 20-h period during which the storm was approximately steady. Shading indicates equivalent potential temperature values of 345–350 K (light shading), 350–360 K (medium shading), and >360 K (heavy shading). Absolute angular momentum is shown by contours labeled in nondimensional units. Also indicated is the airflow through the updraft. (From Rotunno and Emanuel 1987.)

eyewall cloud, the surfaces are packed closely together. This enhanced gradient of  $m$  and  $\theta_{es}$  in the eyewall cloud is the result of a form of frontogenesis, where radial frictional inflow in the boundary layer advects moist enthalpy inward, thus increasing the gradient of  $\theta_{es}$  in the eyewall region (Emanuel 1997).

#### *e. Temporal development and stability of the mean two-dimensional eyewall cloud*

The theoretical picture of a two-dimensional steady tropical cyclone just discussed is, of course, an idealization. In nature, the cyclone and its eyewall cloud are in a perpetual state of evolution. The idealized structure appears to be a state toward which the mature storm tends to adjust. Rotunno and Emanuel (1987) used a two-dimensional axisymmetric nonhydrostatic model to show this adjustment process (Fig. 14). To be consistent with the idealized case described above, the environment was assumed to be initially conditionally neutral but unstable in the conditionally symmetric sense. This assumption represents an actual tropical cyclone environment fairly well, except that in reality the environment usually has some degree of conditional instability. Figures 14a–c show that the fields of  $m$  and  $\theta_{es}$  gradually become more parallel as time passes. Figure 14d is a 20-h average of the  $m$  and  $\theta_{es}$  fields after the storm reaches a quasi-steady state.

Having reached this equilibrium state the streamlines in the eyewall region are generally parallel to the  $m$  and  $\theta_{es}$  surfaces, indicating that the eyewall region of the model has the form of a conditionally symmetrically neutral circulation connected to a boundary layer in contact with the warm sea surface, as in the idealized model calculations illustrated in Figs. 13.

Examination of Figs. 14a–c further shows that the two-dimensional tropical cyclone circulation tended to create conditional instability where none initially existed. As the storm developed, air in the boundary layer flowing inward increased its value of  $\theta_{es}$ , and conditional instability ( $\partial\theta_{es}/\partial z < 0$ ) developed near the center of the storm in the lower troposphere. Vertical convective motions took place in response to this instability and, in the process, transported high  $m$  upward. This transport produced local maxima of  $m$ , and hence local areas of inertial instability, which was released in horizontal accelerations (as suggested by Willoughby et al. 1984a). The combination of vertical convection in response to conditional instability and horizontal motion in response to local inertial instability created a final mixture in the eyewall region which was neutral to combined vertical and horizontal acceleration. Thus, in the process of achieving its conditionally symmetrically neutral final state, transient buoyant convective cells were superimposed on the developing



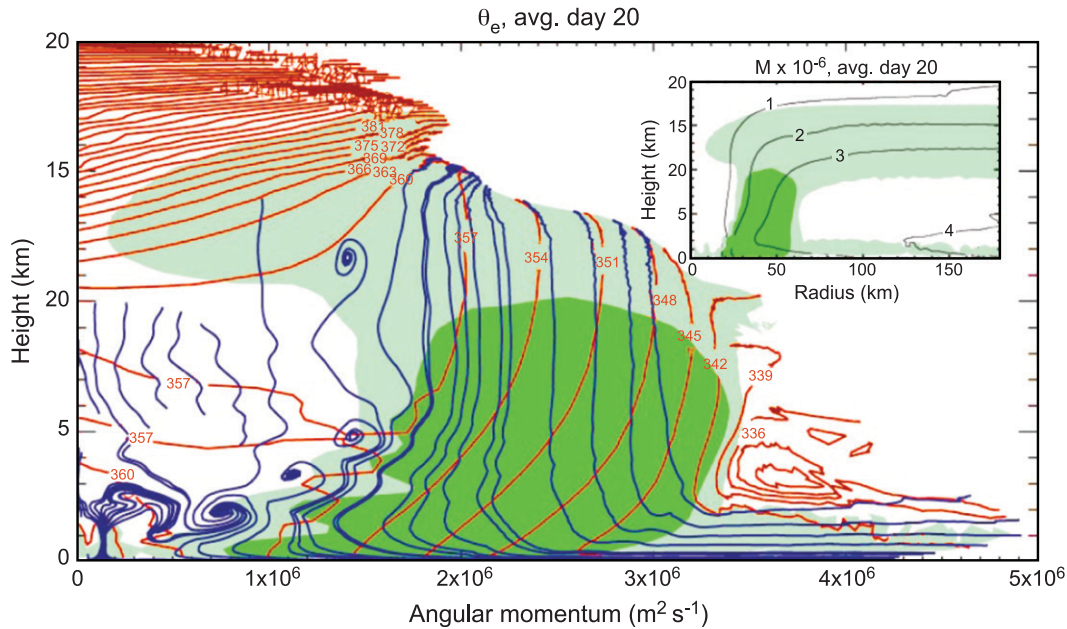


FIG. 15. Results of a two-dimensional simulation of a tropical cyclone. Equivalent potential temperature  $\theta_e$  (K; red, 1-day average) is displayed as a function of absolute angular momentum and height. Overlaid are two-dimensional streamlines of the daily averaged secondary circulation (blue). Light-green shading denotes liquid water mixing ratio  $q_l > 0.3 \text{ g kg}^{-1}$ . Dark-green shading denotes  $q_l > 1.0 \text{ g kg}^{-1}$ . The inset figure repeats the shading for  $q_l$ , displayed as a function of physical radius. The contours of absolute angular momentum (black; inset) are provided for physical reference for the main figure. (From Persing and Montgomery 2003.)

slantwise circulation. These buoyant cells were apparently part of the mixing process that led to the slantwise conditionally symmetrically neutral air motion in the eyewall.

Even after a mature eyewall forms, a tropical cyclone tends to create zones of conditional instability in the eyewall. The boundary layer frictional inflow continues to import air of high  $\theta_e$  from distant radii so that the process illustrated by Figs. 14b–d continues to create instability in the eyewall region. In a further investigation using Rotunno and Emanuel's (1987) two-dimensional model in a high-resolution mode, Persing and Montgomery (2003) found that overturning eddies occur in the radial-vertical plane and draw low-level boundary layer air from the eye into the eyewall (Fig. 15). This behavior is especially powerful since the boundary layer air in the eye tends to have the highest values of  $\theta_e$  in the tropical cyclone (Fig. 8a). This process is probably the two-dimensional model's way of accomplishing the ingestion of high- $\theta_e$  air from the eye into the eyewall, which Smith et al. (2008) show is a necessary outcome of the development of supergeostrophic wind at the top of the boundary layer (Fig. 12b). This entrainment of high- $\theta_e$  air from the eye strengthens the eyewall circulation and creates buoyancy that can be released in strong vertical convection altering the structure of the otherwise moist symmetrically neutral eyewall cloud. Convection, however,

is typically a three-dimensional process, and to consider convection superimposed on the eyewall it is essential to remove the constraint of two-dimensionality. The next section considers convection and other asymmetrical processes without this constraint.

## 7. Substructure and asymmetry of the eyewall cloud

### a. Conditional instability within the eyewall cloud

While a circular outward-sloping eyewall cloud consisting of purely slantwise conditionally symmetric neutral motions (section 6) is a useful idealization that explains a great deal about the typical structure of an eyewall cloud, the eyewall cloud does not exist in reality without containing superimposed cells of buoyant convective updrafts arising from the local release of conditional buoyant instability. The air masses in which tropical cyclones occur are not exactly slantwise neutral; they usually exhibit some degree of conditional instability. In addition, conditionally unstable air processed by the eyewall cloud is not undisturbed large-scale environmental air but rather air whose thermodynamic structure has been modified by the tropical cyclone circulation. The processes by which this modified air enters the eyewall are typically associated with deviations from

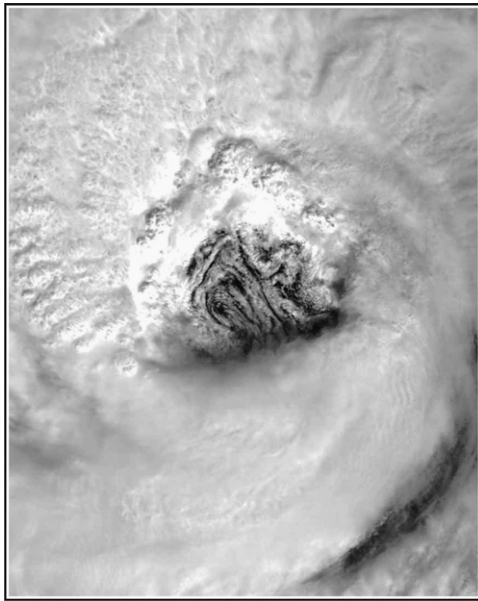


FIG. 16. Moderate-Resolution Imaging Spectroradiometer (MODIS) satellite image of the eye of Hurricane Erin at 1515 UTC 11 Sep 2001. Eyewall has a wavenumber-3 asymmetry, and the stratocumulus in the eye is distorted in a highly convoluted pattern. (From Kossin et al. 2002.)

the idealized symmetrically neutral model of the hurricane dynamics. The two-dimensional modeling discussed in section 6e shows that the vertical overturning of the tropical cyclone in the  $r$ - $z$  plane tends to create conditional instability within the eyewall cloud even if none initially existed in the environment. Observations (summarized in Fig. 9) verify that the eyewall is fed from its base by both frictional inflow from outer radii and by low-level outflow from the eye. The circular shape of the eyewall cloud is often distorted. Figure 16 shows an example where the eyewall exhibits a wavenumber-3 or -4 irregularity, and the undercast of stratus across the eye is highly distorted. These asymmetries are thought to be due at least in part to barotropic instability of the sheared flow in the RMW (Schubert et al. 1999; Kossin and Schubert 2001; Kossin et al. 2002). Vortices associated with eyewall asymmetries constitute a mechanism by which high  $\theta_e$  of the air in the boundary layer may be entrained at low levels from the eye into the eyewall.

If the  $\theta_e$  of the air flowing up into the base of the eyewall cloud as a result of any of these processes is sufficiently high, the eyewall cloud will develop pockets of conditionally unstable air, and small-scale buoyant updrafts will form in those locations. In a realistic three-dimensional simulation of Hurricane Bob (1991), Braun (2002) found that buoyant elements superimposed on the eyewall accounted for over 30% of the vertical mass flux in the eyewall, but that the buoyancy was most often

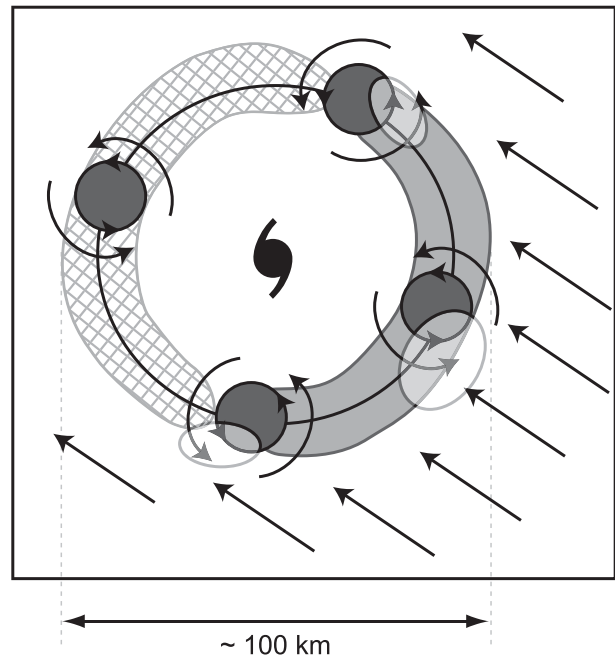


FIG. 17. Schematic diagram summarizing the interaction between eyewall mesovortices and the low-level inflow associated with the environmental wind shear in a Northern Hemispheric tropical cyclone. The elongated semicircular areas indicate where shear effects favor upward (light shading) and downward (cross hatching) motion. The low-level relative flow is indicated by the straight arrows, and the mesovortices and their local cyclonic circulations are indicated by darkly shaded circles and curved arrows. The semitransparent, lightly shaded ovals represent areas of enhanced low-level convergence and upward motion. The direction of the large-scale shear vector (not shown) is from northwest to southeast. (From Braun et al. 2006.)

manifested along outward-sloping rather than purely vertical paths because of being superimposed on the strong slantwise component of the eyewall circulation. These buoyant updrafts are highly three-dimensional and the constraint of two-dimensional modeling must be removed to fully simulate their behavior.

#### *b. Eyewall vorticity maxima and strong updrafts*

Braun et al. (2006) simulated Hurricane Bonnie (1998) with a three-dimensional full-physics model and found that strong vertically extensive updrafts located along the eyewall were associated with small-scale vortices in the horizontal wind field (Fig. 17). Marks et al. (2008) call such subvortices eyewall vorticity maxima (EVMs). The EVMs on the eyewall in the Bonnie simulation were  $\sim 20$  km, small enough in scale for four of these subvortices to form on the eyewall, which was very large in diameter, the eye of Bonnie being  $\sim 100$  km. In the first airborne dual-Doppler radar study of a tropical cyclone,

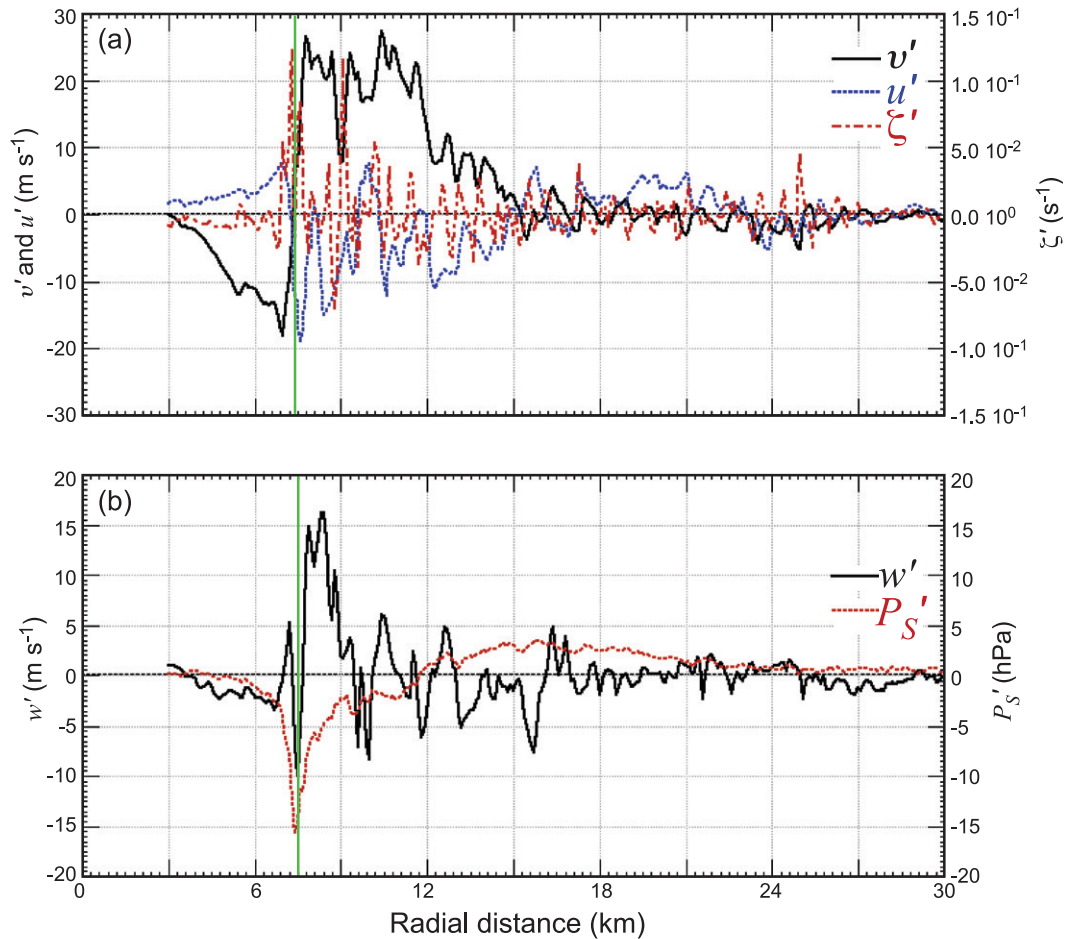


FIG. 18. Data obtained on a flight penetration of an eyewall vorticity maximum (subvortex) in Hurricane Hugo, 15 Sep 1989. Plots show 1-s time series of perturbation components of (a) tangential wind  $v$ , radial wind  $u$ , and relative vorticity  $\zeta$  and (b) vertical velocity  $w$  and surface pressure  $P_s$  with respect to radial distance from the center of Hurricane Hugo, 15 Sep 1989. The perturbations are the residual after the mean vortex profiles have been removed from the time series. The green vertical line denotes the center of the eyewall vorticity maximum. (From Marks et al. 2008.)

Marks and Houze (1984) found an EVM on the developing eyewall of Hurricane Debby (1982) that was  $\sim 5$ – $10$  km in horizontal scale. The primary vortex was  $\sim 30$  km in diameter. Marks et al. (2008) describe a harrowing flight directly across an intense EVM located on the inner side of the eyewall of Hurricane Hugo (1989). Figure 18 contains a record of the meteorological variables on this flight track directly across the subvortex. The mean vortex structure has been removed, so these data show only the perturbation values associated with the EVM. The tangential wind perturbation record in Fig. 18a shows that this subvortex was 6–7 km in overall horizontal extent although the windshift across the center from  $-18$  to  $+25 \text{ m s}^{-1}$  occurred over a distance of only  $\sim 2$  km. The primary vortex of Hugo was  $\sim 25$ – $30$  km in diameter.

Braun et al. (2006) suggest that EVMs form along the eyewall in response to instability of the vortex ring as-

sociated with the RMW, similar to the way “suction vortices” form within the ring of maximum vorticity of a tornado vortex (Fujita 1981; Houze 1993, 298–303). Tornadic suction vortices are often hundreds of meters in dimension and are subvortices of a primary tornado that is hundreds–thousands of meters in diameter. Collectively, the model simulation of Bonnie, the observations of EVM in Debby and Hugo, and the breakdown of a tornado vortex into suction vortices suggest that EVMs are fundamental products of vortices and may come in a range of sizes related the width of the parent vortex.

The EVM described by Marks et al. (2008) contained a very strong updraft, reaching a peak velocity of  $\sim 17 \text{ m s}^{-1}$  (Fig. 18b). This strong updraft is consistent with the Bonnie simulation, in which Braun et al. (2006) found that each EVM coincided with a deep vertically erect updraft. The association of the strong updrafts

with the EVMs in Bonnie and Hugo is not coincidental. The small-scale vorticity maxima entrain extremely high- $\theta_e$  air from the eye region into the base of the eyewall cloud. This entrainment of air from the eye is a primary mechanism for imparting buoyancy to the otherwise slantwise convective eyewall. The intense buoyant updraft associated with an EVM is a direct response to this local enhancement of buoyancy.

The model simulation of Braun et al. (2006) indicates that the convective-scale updrafts associated with the small-scale vortices remain vertically erect and coherent as they are advected around the eyewall. This erect structure of the rotating convective updrafts is somewhat surprising since we might expect the vertical shear of the horizontal wind in the tropical cyclone vortex to produce updraft cells that are distorted in the vertical. Braun et al. (2006) suggest that the small-scale rotating updrafts in the EVMs are inertially resistant to the surrounding airflow so that their central cores are protected and resistant to the shear of the mean tangential wind of the tropical cyclone vortex. The EVM observed by Marks and Houze (1984) in Debby was not vertically erect but rather was tilted in the direction of the shear. It may have been a weaker example of an EVM and therefore unable to withstand the shear in the tangential flow of the main vortex.

Since the EVMs derive their rotation from the vorticity on the inside edge of the zone of maximum wind of a fully formed tropical cyclone, these rotational updrafts are different from, and should not be confused with, the vortical hot towers discussed in section 3a, which occur prior to the formation of a tropical cyclone and derive their vorticity from the vorticity-rich boundary layer and/or tilting of the ambient shear. It is worth noting that the damage that could occur from the locally enhanced winds in an EVM in a landfalling storm is likely to be worse than what would be expected from the mean vortex winds.

### c. Statistics of updrafts and downdrafts in eyewall clouds

We have seen that the EVMs contain strong updrafts associated with entrainment of high- $\theta_e$  air from the eye region. The vertical velocities in those drafts are on the upper extreme of updraft velocities in hurricane eyewall clouds. The general statistics of vertical air motions in hurricane eyewalls are known fairly well since research and reconnaissance aircraft have been flying through tropical cyclones and measuring vertical air motions for over 50 years (Sheets 2003; Aberson et al. 2006). Up to the mid-1980s, aircraft measurements of vertical air motions were obtained only in situ, along the flight tracks of aircraft, and spatial distributions of vertical velocity could be determined only by composites of data

from different flight tracks (e.g., Fig. 7c). One of the best sets of flight-track vertical velocities were obtained in the extremely intense Hurricane Allen (1980). Figure 19a from Jorgensen (1984) shows aircraft-observed vertical velocities in the eyewall of Allen determined by mass continuity from the horizontal wind data along the flight track.<sup>6</sup> Figure 19b shows vertical velocities indicated by the aircraft's inertial navigation system. Both types of measurements indicate peak magnitudes of  $w$  of just over  $6 \text{ m s}^{-1}$ . Figure 20 from Jorgensen et al. (1985) shows that the vertical velocities seen in Allen are typical of those seen at flight level in other tropical cyclones, although this larger sample shows slightly larger peak updrafts  $\sim 7\text{--}10 \text{ m s}^{-1}$ .

Airborne Doppler radar measures the velocity of precipitation particles along the beam of the radar. For data collected with the beam pointed vertically, correction for particle fall speed leads to estimation of the vertical air motion. Black et al. (1996) analyzed vertically pointing Doppler radar collected by aircraft on 185 radial flight legs into and out of the eye regions of hurricanes. Their results are shown in Fig. 21 in the form of a two-dimensional contoured-frequency-by-altitude diagram (CFAD; Yuter and Houze 1995). Note that the broadening of the distributions with height is an artifact resulting from a sampling bias whereby the weaker drafts are difficult to detect at greater distance from the aircraft, which was most frequently flying at altitudes of  $\sim 1.5\text{--}3 \text{ km}$ . In general agreement with the flight-level data in Figs. 19 and 20, the peak updraft values are  $\sim 8 \text{ m s}^{-1}$ . More than 70% of the Doppler-derived vertical velocities were in the range of  $\pm 2 \text{ m s}^{-1}$ . *Drafts* (both up and down) were defined by Black et al. (1996) as zones having  $|w| > 1.5 \text{ m s}^{-1}$  continuously along the flight track and containing a maximum of  $w > 3 \text{ m s}^{-1}$ . Updrafts defined this way were nearly all  $< 3 \text{ km}$  in horizontal dimension; only 5% were wider than  $6 \text{ km}$ . More extreme updrafts such as in the EVM updraft encountered on the Hugo flight (Fig. 18) do not appear in statistics of the type shown in Figs. 19, 20, and 21 because they are rare and because they would normally not be penetrated since the aircraft normally avoids portions of eyewalls that appear dangerous either visually or on radar. The statistics in Fig. 21 show that the magnitudes of updraft velocity tend to increase with height, the largest values being above the 10-km level. It has been suggested that stronger vertical velocities should occur at higher levels because their buoyancy is boosted by the release of latent heat of fusion (Lord et al. 1984; Zipser 2003).

<sup>6</sup> Note that the downdrafts in Fig. 19 confirm the downdrafts on the inner side of the eyewall in Fig. 9 and in the photograph in Fig. 11.



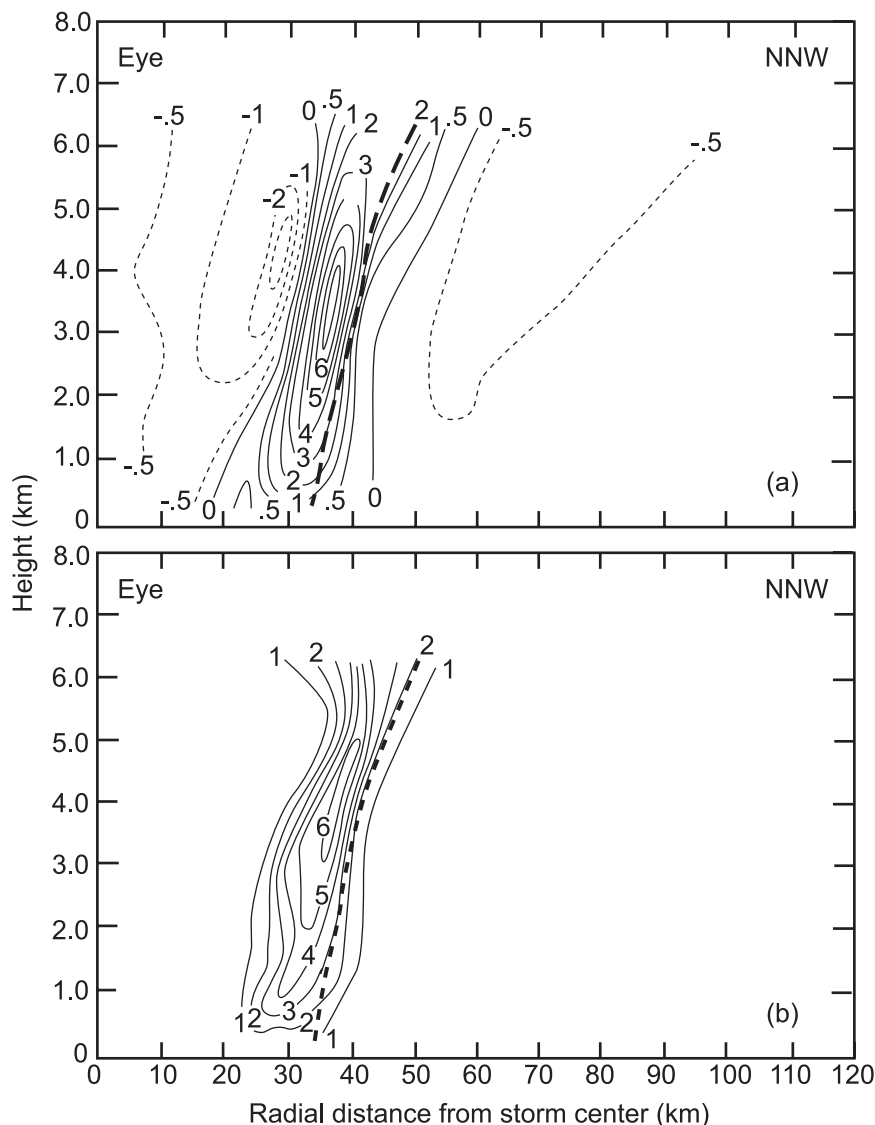


FIG. 19. Vertical air velocity ( $\text{m s}^{-1}$ ) measured aboard aircraft flying through the eyewall of Hurricane Allen (1980). (a) Vertical velocity diagnosed from horizontal wind measurements by integrating the continuity equation. (b) Vertical velocity measured directly by the inertial navigation equipment of the aircraft. (From Jorgensen 1984.)

The airborne Doppler radar data of Black et al. (1996) provide information on the horizontal as well as the vertical structure of the upward motion zone in the eyewall. The statistics exhibit a coherent radially outward-sloping structure in two-dimensional autocorrelations relative to the updraft maxima in the eyewall for updrafts at 2.5- and 7.5-km flight altitudes (Figs. 22a,b). A difficulty in interpreting aircraft-based updraft data in tropical cyclones, whether obtained in situ or by radar, is that no determination can be made observationally between slantwise neutral vertical motions (see section 6) and embedded vertically accelerating buoyant updrafts. The Hugo penetration (Fig. 18) and Bonnie simulation

(Fig. 17) demonstrate that locally intense updrafts that are primarily buoyancy phenomena can be superimposed on the eyewall. In numerical simulations of both Bonnie and Hurricane Bob (1991), Braun (2002, 2006) found that updrafts exhibiting buoyancy relative to the mean vortex circulation in which they were embedded accounted for over half<sup>7</sup> of the condensation in the eyewall. If this result is representative of tropical cyclones in general, it implies

<sup>7</sup> This fraction is probably an overestimate, because Braun (2002, 2006) did not attempt to remove the component of the updraft's condensation that was due to the background slantwise motion. The fraction is nevertheless probably very significant.

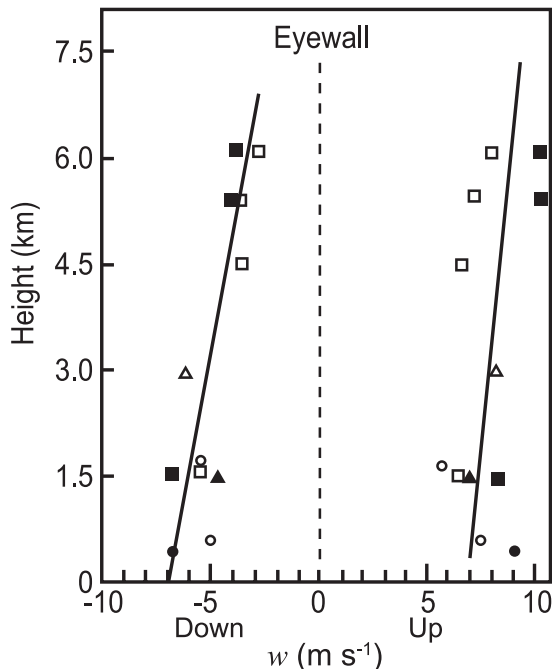


FIG. 20. Peak updraft and downdraft speeds measured aboard research aircraft flying through the eyewalls of tropical cyclones. Each style of symbol indicates a particular research flight. The value shown by a given symbol is the maximum draft speed observed on that flight at that altitude. (From Jorgensen et al. 1985.)

that while the eyewall's mean structure (including the sloping stadium-like structure exemplified in Fig. 5 and evident in Figs. 22a,b) is determined by the adjustment of the mean vortex motion toward an idealized slantwise symmetrically neutral state (see section 6), the dynamics of the eyewall cannot be accurately interpreted without considering the superimposed intense buoyant updrafts.

#### d. Downdrafts in the eyewall

As indicated in Fig. 9, convective-scale downdrafts occur in the zone of heavy rain within the eyewall. The CFAD in Fig. 21 attaches a magnitude to these downdrafts. The downdrafts are observed at all altitudes, but they are more frequent at lower levels than at higher altitudes. Overall they are outnumbered by updrafts by about a factor of 2. Figure 23 from Black et al. (1996) shows that the downdraft mass flux in the eyewall is about 20%–30% of the upward mass flux at any given altitude. Despite their frequent occurrence, neither the causes of the eyewall downdrafts nor their dynamical significance to the overall tropical cyclone circulation have been fully determined. Figures 22c,d show that unlike the upward vertical velocities in the eyewall, these downdrafts show no sloping coherent structure; the

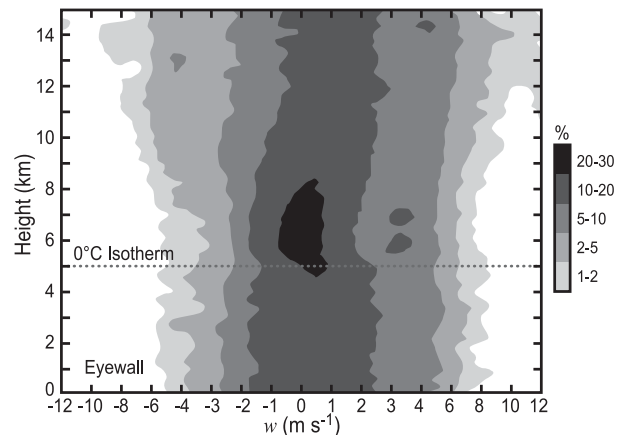


FIG. 21. Statistics of vertical air motions measured by vertically pointing airborne Doppler radar in hurricanes analyzed in the form of contoured-frequency-by-altitude diagrams of vertical velocity  $w$  for the eyewall. (Adapted from Black et al. 1996.)

downdrafts are essentially uncorrelated in the radius–height plane. It seems likely that, as is the case in convection not associated with tropical cyclones (e.g., Yuter and Houze 1995), the downdrafts at upper levels are forced responses to updraft motions while those at lower levels are precipitation driven.

#### e. Eyewall asymmetry owing to storm motion and shear

The clouds and precipitation of the eyewall of a tropical cyclone are practically always distributed nonuniformly around the storm. This asymmetry arises for two reasons: (i) the movement of the storm through the surrounding atmosphere and (ii) the wind shear of the large-scale environment. The first reason is because the eyewall is closely associated with the cyclone's strong circulation, which is inertially stable and resistant to the surrounding airflow. Consequently, the movement of the tropical cyclone through its environment creates convergence on the side of the storm toward which the eyewall is moving (Shapiro 1983). Thus, the mean vertical velocity pattern in the eyewall of a moving tropical cyclone is inherently asymmetric. Of interest is that there is a feedback between the wavenumber-1 asymmetry and the storm track; instabilities in the asymmetry are related to a trochoidal motion sometimes exhibited by tropical cyclone tracks (Nolan et al. 2001). The second reason for the wavenumber-1 asymmetry is vertical wind shear in the environment. Although shear is unfavorable to the development and intensification of tropical cyclones (see section 2), these storms usually exist within environments having some degree of shear. The storm-relative

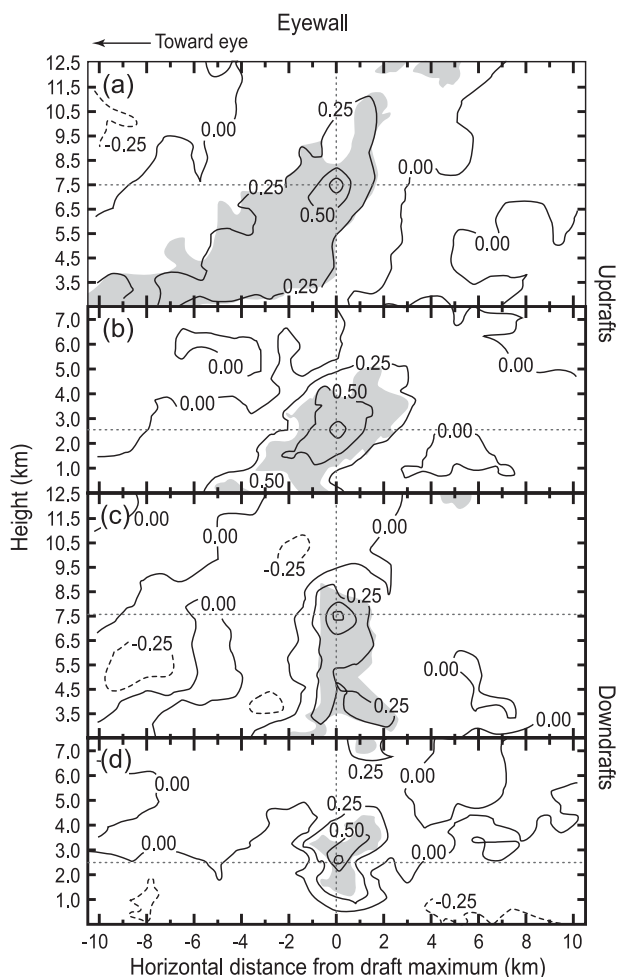


FIG. 22. Statistics of vertical air motions measured by vertically pointing airborne Doppler radar in hurricanes analyzed as two-dimensional autocorrelations relative to the vertical velocity draft maxima in the eyewall region for updrafts at (a) 7.5- and (b) 2.5-km altitude and downdrafts at (c) 7.5- and (d) 2.5-km altitude. The horizontal axis is oriented so that the storm center (eye) is to the left of the figure. The vertical and horizontal scales are approximately equal to minimize vertical exaggeration. The shaded areas indicate correlations that are significant at the 95% level. The intersection of the dashed lines is the location of the draft maximum. (From Black et al. 1996.)

environmental flow is a function of height, and as such it redistributes cloud and precipitation particles around the tropical cyclone. These two factors acting together lead to a variety of wavenumber-1 asymmetries of the eyewall structure and intensity.

Analyzing data from the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (Kummerow et al. 1998) along with large-scale wind databases for tropical cyclones over six ocean basins for a 3-yr period, Chen et al. (2006) have determined the expected rainfall patterns for different combinations of storm motion and

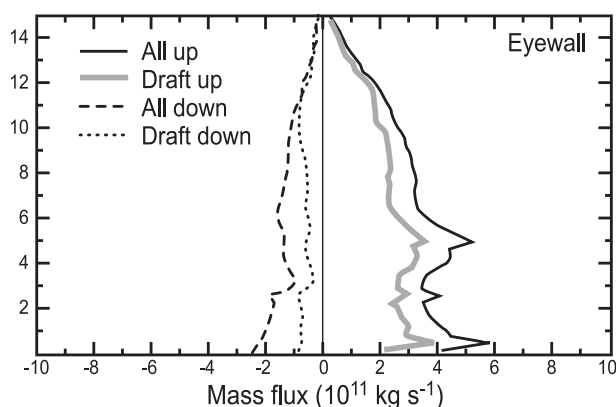


FIG. 23. Statistics of vertical air motions measured by vertically pointing airborne Doppler radar in hurricanes expressed as the total upward (solid) and downward (dashed) mass flux in the eyewall. Data have been normalized by the amount of missing data as a function of altitude for each of the four regions. Results are shown for the total mass flux by all of the vertical motions and only by the drafts (see text for definition). (From Black et al. 1996.)

wind shear vectors. They find that in general, the tropical cyclone rainfall asymmetry has larger amplitudes with respect to vertical wind shear than to storm motion. Regardless of the strength of the environment shear, the maximum in eyewall rainfall is on the downshear left side of the storm. When the shear is strong, this asymmetry is felt at all radial distances from the storm center. When the shear is weak, the shear-related asymmetry is confined to the radii nearer the storm center (including the eyewall), while at long distance from the center, the maximum rainfall is on the downshear side of the storm, as expected in the case of no shear. Rogers et al. (2003) have shown how these asymmetries of rainfall in the eyewall combine with the storm motion to produce a variety of surface rainfall patterns.

Analyzing airborne Doppler radar data and in situ flight track data collected in Hurricanes Jimena (1991) and Olivia (1994), Black et al. (2002) inferred the fate of cloud and precipitation particles produced in an updraft forced by convergence (due to storm motion) on the forward side of a tropical cyclone embedded in a sheared environment (Fig. 24). For illustration, they assumed the shear to consist of upper-level westerly flow and lower-level easterlies. Cloud and precipitation particles generated in the updraft on the downshear side (east) of the storm are advected cyclonically by the primary vortex circulation. Precipitation particles produced in the updraft formed on the east side of the eyewall are advected cyclonically around the north side of the storm, where they produce a strong 45–50 dBZ radar echo. This echo corresponds to the maximum rainfall found by Chen et al. (2006) to occur typically on the downshear left side of the

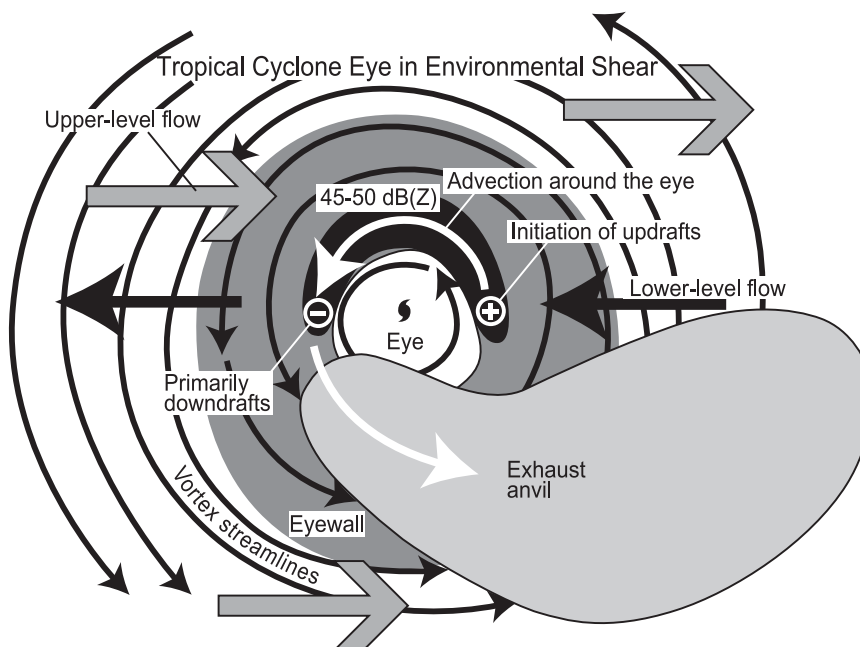


FIG. 24. Schematic illustration of the shear-induced asymmetry of the inner-core region of a tropical cyclone based upon aircraft observations. The low-level environmental flow is indicated by the two solid black arrows. Upper-level flow is indicated by the three shaded arrows. Convective cells form somewhat upwind of the downshear side of the eyewall. They are advected 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  $90^\circ$  to the left of the shear vector. By the time the cells reach the upshear side of the eyewall they have ascended through the  $0^\circ\text{C}$  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 buoyant air in the unloaded updrafts accelerates upward and approaches the tropopause as they rotate through the semicircle to the right of the shear. (From Black et al. 2002.)

storm. Since translation-induced vertical motion does not favor updraft formation on the west side of the eyewall (i.e., upshear), the radar echo is weaker on the south side of the eyewall. Figure 24 further shows how the upper-level cloud produced by the active convection is blown off in an asymmetric pattern. Ice particles are advected outward and cyclonically around to the south side of the storm and exit in a massive cirriform plume on the southeast side of the storm.

#### *f. Cloud microphysical processes in the eyewall and inner-core region*

In the upward branch of the secondary circulation of the tropical cyclone, precipitation particles are generated rapidly to produce the ring of heavy rainfall defining the eyewall. The basic elements of precipitation particle growth and fallout are illustrated by Fig. 25, which is based on Doppler radar observations in Hurricane Alicia (1983) (Marks and Houze 1987). Many of

the precipitation particles in the eyewall are generated by the “warm rain” process—that is, drops are condensed, grow rapidly by drop coalescence below the  $0^\circ\text{C}$  level, and fall out before they have a chance to become ice particles. However, much of the rain generation in the eyewall and surrounding inner-core region of the cyclone involves the ice phase. Just above the  $0^\circ\text{C}$  level, ice particles occur and grow by riming to produce graupel. The heavier graupel particles fall out rapidly (fall speeds of several meters per second), melt to form raindrops, and fall out in the eyewall precipitation region (see trajectory 0–1’–2’ in Fig. 25). However, many of the ice particles formed in the eyewall cloud are ice-particle aggregates (bunches of ice particles stuck together), which fall more slowly ( $\sim 1\text{ m s}^{-1}$ ; see chapter 3 of Houze 1993). The aggregates and other slowly falling ice particles are advected outward by the radial wind component (Fig. 25) and swept great distances around the storm by the strong tangential winds of the cyclone (trajectory 0–1–2–3–4 in Fig. 26). As a result of the ice



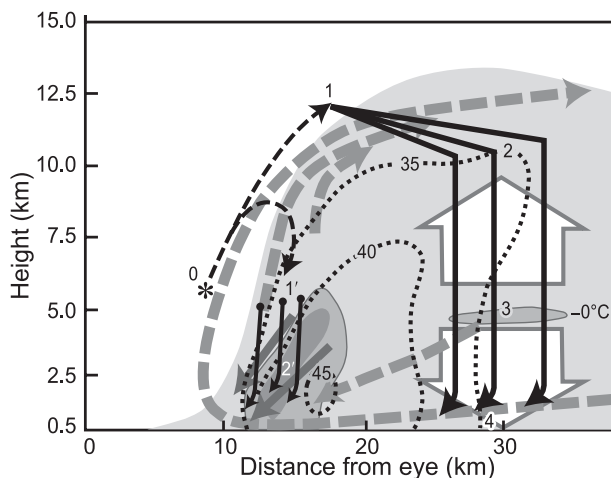


FIG. 25. Schematic of the radius–height circulation of the inner-core region of Hurricane Alicia (1983), as observed by airborne dual-Doppler radar. Shading depicts the reflectivity field, with contours at 5, 30, and 35 dBZ. The primary (tangential) circulation ( $v$ ;  $\text{m s}^{-1}$ ) is depicted by dotted lines, and the secondary circulation is shown by the wide dashed streamlines. The convective downdrafts are denoted by the thick solid arrows, and mesoscale up- and downdrafts are shown by the broad arrows. Hydrometeor trajectories emanate from the asterisk and are denoted by thin dashed and solid lines labeled with numbers corresponding to points along the horizontal projections shown in Fig. 26. (From Marks and Houze 1987.)

particles swirling outward in this way, the eyewall seeds the clouds throughout the inner-core region of the tropical cyclone.<sup>8</sup> When the aggregates finally pass through the melting layer, usually within a rainband located some distance from the eyewall, they produce a bright band just below the  $0^{\circ}\text{C}$  level in the radar reflectivity, which is a signature of stratiform precipitation. This structure is indicated at radii of  $>20$  km in Fig. 25.

Figure 27 is a more detailed schematic of the microphysical processes in the mixed-phase region of the eyewall cloud (enclosed by the dashed circle). This diagram summarizes 20 years of research based on 230 aircraft missions collecting cloud microphysical observations in tropical cyclones (Black and Hallett 1986, 1999). It is consistent with Fig. 25 in that below a mixed-phase region of the cloud (a shallow layer zone between  $\sim 0^{\circ}$  and  $-5^{\circ}\text{C}$  containing both liquid drops and ice

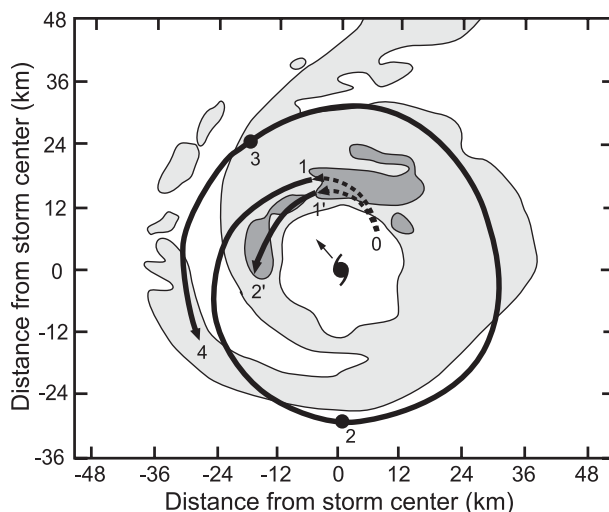


FIG. 26. Horizontal projections of the paths of precipitation particle trajectories superimposed on the radar echo pattern of Hurricane Alicia (1983). The echo contours are for 20 and 35 dBZ. The numbers show how the trajectories correspond to the vertical cross section in Fig. 25. (From Marks and Houze 1987.)

particles), drops grow vigorously by coalescence, and many rain out before they can be lifted into the mixed-phase region. Above the mixed-phase region, the cloud is glaciated, apparently from the highly probable collisions of any newly condensed supercooled drops with the numerous ice particles already present at upper levels in the mature eyewall. In the mixed-phase layer of the eyewall cloud, supercooled liquid water content is generally low ( $<0.5 \text{ g m}^{-3}$ ) and found primarily where the updraft intensity is  $>5 \text{ m s}^{-1}$ , although not even all updrafts of this strength contain supercooled drops (Black and Hallett 1986). This liquid water is likely condensed just below the  $0^{\circ}\text{C}$  level, lofted by the updraft above this level and exists only for a short period of time before coming into contact with preexisting ice particles. The inset of Fig. 27 magnifies the region of the eyewall cloud where supercooled drops are found. This zone can be divided into three subregions: the inner edge adjacent to the eye region, an interior zone, and an outer edge of the eyewall cloud. Supercooled drops are found primarily on the inner edge of the eyewall cloud, where the newly generated drops are unlikely to have yet collided with preexisting ice particles since they occur near the top of the outward-sloping eyewall cloud, where no ice particles are entering the layer from above.

The middle portion of the inset in Fig. 27 illustrates how ice particles from higher levels come into contact with the supercooled droplets in the mixed-phase region. The drops freeze by contact nucleation and grow further by riming as they accrete the continuously produced supercooled cloud droplets. Graupel particles result

<sup>8</sup> This natural seeding of the inner-core region of the storm with ice particles generated in the eyewall cloud is one of the reasons that early attempts to seed tropical cyclone clouds artificially by introduction of ice-nucleating agents were futile. Any supercooled drops appearing in the inner-core region are quickly scavenged by the particles emanating from the eyewall cloud. See Willoughby et al. (1985) and Sheets (2003) for historical accounts of attempts to reduce cyclone intensity via cloud seeding.

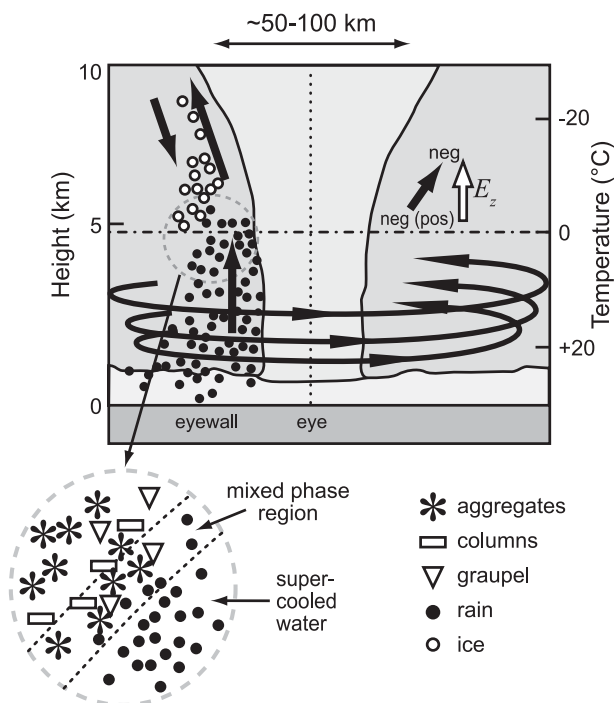


FIG. 27. Schematic depicting precipitation processes, cloud distribution, and preferred regions for charge separation in a tropical cyclone. Ocean surface temperatures are near  $28^{\circ}\text{C}$ , giving cloud-base temperatures near  $22^{\circ}\text{C}$ . In mature tropical cyclones, typical vertical velocities near  $0^{\circ}\text{C}$  level are  $5\text{--}8\text{ m s}^{-1}$ , with convection extending to well beyond the  $-40^{\circ}\text{C}$  isotherm, the temperature for homogeneous ice nucleation. The inset shows the region in which charge separation occurs in the horizontal gradient of vertical shear, limited in altitude by the lack of supercooled water above  $-5^{\circ}\text{C}$ . White dots indicate where small ice particles, frozen drops, and graupel exist, and black drops indicate rain. Air motion arrows show axial and radial flow leading to redistribution of hydrometeors. Tangential (primary) circulations are represented by circular streamlines. Straight arrows show approximate radial and vertical air motions; lengths are not to scale. The observed distribution of electric field polarity is shown. (From Black and Hallett 1999.)

from this riming and continue to grow by collection of supercooled droplets. Riming at  $\sim 0^{\circ}$  and  $-5^{\circ}\text{C}$  produces large numbers of secondary ice particles by the process identified by Hallett and Mossop (1974; see summary in chapter 3 of Houze 1993). High concentrations of ice particles (up to hundreds per liter) are produced by this process (Black and Hallett 1986). These secondary particles combined with the snow particles falling into the region from the eyewall outflow above, make the probability of the supercooled drops being scavenged by ice particles extremely high. The crystal habits of the tiny ice particles produced by the secondary ice-particle production mechanism and growing further by vapor deposition are usually difficult to discern from instruments aboard aircraft. Where identifiable, the pre-

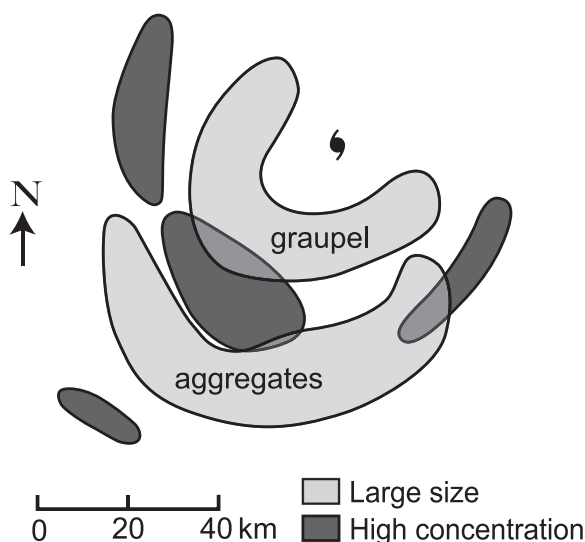


FIG. 28. Mesoscale horizontal distribution of ice particles in Hurricane Norbert (1984), as observed by an aircraft flying within the storm at an altitude of 6 km. The areas of high ice concentration had particle counts of  $\sim 100\text{ L}^{-1}$ . The areas of large size were where the median diameter was  $>1.05\text{ mm}$ . Graupel was thought to be the predominant large ice particle type in the eyewall region. Aggregates were thought to be the primary large particle type in the region outside the eyewall zone. (Adapted from Houze et al. 1992.)

dominant crystal habits are columns, consistent with the temperature regime of  $\sim 0^{\circ}$  and  $-5^{\circ}\text{C}$  (Table 6.1 of Wallace and Hobbs 2006). As indicated in Fig. 27, these columns are part of a population that includes supercooled droplets, graupel and other ice particles in the interior of the updraft region of the eyewall cloud.

The inset in Fig. 27 indicates that the outer edge of the eyewall cloud is dominated by completely glaciated cloud, with high concentrations of ice particles (up to  $200\text{ L}^{-1}$ ; see Black and Hallett 1986). Of these particles, many are small and likely produced by the secondary ice-particle production process. However, large aggregate ice particles also appear in this region. Figure 28 from Houze et al. (1992) shows the pattern of ice particles sampled extensively by aircraft flying at the 6-km level in the eyewall and inner-core regions of Hurricane Norbert (1984). This overall distribution of ice particles is consistent with the processes indicated in Fig. 27. High concentrations of small ice particles ( $0.05\text{--}0.5\text{ mm}$  in dimension) likely resulting from secondary ice-particle production were located just outside the eyewall. On either side of this zone of numerous small ice particles were regions where larger ice particles ( $>1.05\text{ mm}$  in dimension) were concentrated. The larger ice particles in the eyewall region were graupel particles, which because of their higher fall velocities remained in the eyewall region as they fell out. The larger ice particles

found some 50 km outside the eyewall were aggregates, and in trajectories similar to 0–1–2–3–4 in Fig. 26, they were advected radially outward, circulated around the storm, and fell out as stratiform precipitation well outside the eyewall region.

From observations such as those described here, it is evident that the cloud microphysics in the tropical cyclone (and associated cloud electrification to be described below) are complex and linked to the details of the cloud dynamics. The extent to which these processes are represented accurately in numerical models is only just beginning to be explored. Rogers et al. (2007) have compared observational and model statistics of gross vertical velocity and radar reflectivity in tropical cyclones and found rough agreement. However, much remains to be determined about specific particle growth modes, mixing ratios of different hydrometeor species, nucleation and breakup processes, and particle trajectories in relation to both the larger-scale air motions in the cyclone and the mesoscale and convective air motions within specific cloud and precipitation features. Until models represent these details, precise forecasting of rainfall in landfalling hurricanes will remain a major challenge.

#### g. Electrification of the eyewall cloud

Cloud electrification is described in basic references such as Williams (1988) and Wallace and Hobbs (2006). When a smaller ice particle collides with a larger graupel particle the two particles take on opposite electric charges. Air motions in the cloud advect the smaller particles away from the falling graupel to a region of the cloud where they accumulate and give that part of the cloud an overall electric charge opposite in sign to the net charge in the region of the cloud where the graupel is concentrated. The sign of the charge transfer depends on the temperature and liquid water content of the region of the cloud where the ice-particle collisions occur. In the eyewall cloud of a tropical cyclone, graupel occurs generally in the  $\sim 0^\circ$  and  $-5^\circ\text{C}$  temperature layer (Fig. 27, inset), and the liquid water contents there are nearly always low ( $<0.5\text{ g m}^{-3}$ ). Under these conditions, the graupel takes on positive charge, and the small colliding ice particles become negatively charged. The air motions in the eyewall tend to carry the small ice particles upward and outward, away from the region of graupel particles, leading to a negatively charged region of cloud at the  $-10^\circ$  to  $-15^\circ\text{C}$  level, slightly radially outward of the region where the ice-particle collisions occurred. Black and Hallett (1999) have determined, as indicated on the right-hand side of Fig. 27, that this sloping region of negative charge results in an upward component of the electric field vector ( $E_z$ ) within the eyewall cloud. Component  $E_z$  points upward from

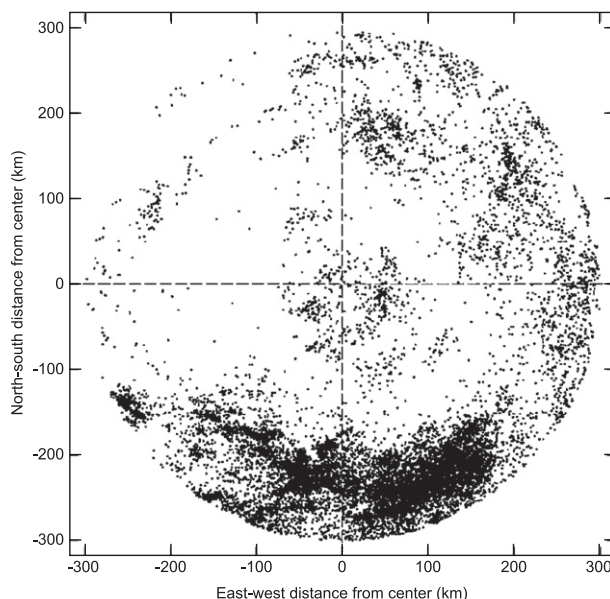


FIG. 29. Plot of the locations of all negative lightning flashes registered at the ground or ocean surface in nine tropical cyclones. Flashes are located with respect to the center of each cyclone. (From Molinari et al. 1999.)

a lower region of less negative to the upper zone of more negative charge that has accumulated from the upward and outward advection of small ice particles. When  $E_z$  becomes sufficiently large, a lightning discharge may result, as a negative strike at the ocean or ground surface. The (pos) in the figure indicates that on occasions when a portion of the eyewall cloud becomes buoyantly unstable, with larger vertical velocities and larger liquid water content ( $>0.5\text{ g m}^{-3}$ ), the small ice particles may become positively charged when they bounce off graupel particles. In that case positive charge is advected upward and outward, the sign of  $E_z$  is reversed, and positive surface lightning strikes may result at the ocean or ground surface. However, positive strikes are rare in tropical cyclones since the liquid water contents are predominantly low in the eyewall cloud, where, as noted above, any newly generated supercooled drops are quickly glaciated because of their highly probable contact with the abundantly present ice particles. An exception might be an extreme updraft of the type documented by Marks et al. (2008), but this has not been documented.

Figure 29 shows the distribution of negative surface lightning flashes observed in nine tropical cyclones (Molinari et al. 1999). These data indicate that a maximum of flashes occurs in the eyewall region, within 100 km of the storm center. A secondary maximum of flash occurrence is at a distance of  $>200\text{ km}$  from the storm center. As will be discussed in section 8b, the clouds in this region of a tropical cyclone appear to be less

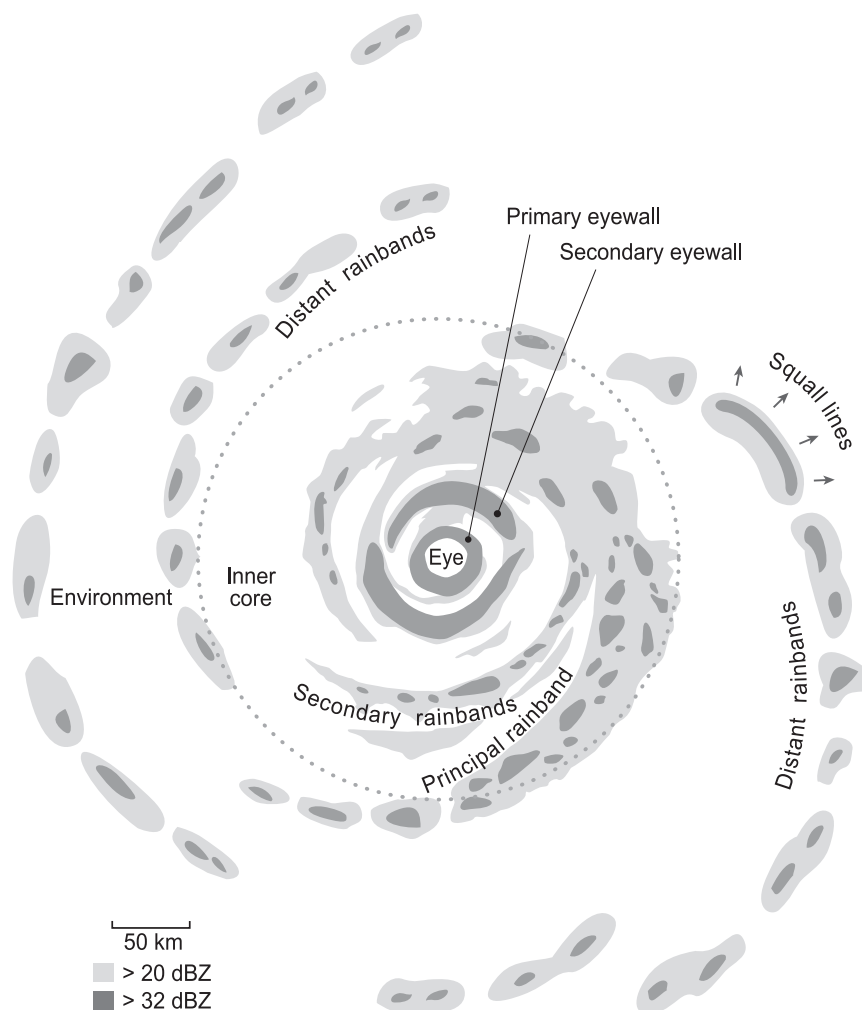


FIG. 30. Schematic illustration of radar reflectivity in a Northern Hemisphere tropical cyclone with a double eyewall. (Adapted liberally from Willoughby 1988.)

strongly influenced by the dynamics of the inner-core vortex, and this lightning is probably produced by processes similar to those observed in thunderstorms not associated with tropical cyclones (Williams 1988; Houze 1993; Wallace and Hobbs 2006) rather than by the eyewall electrification mechanism represented in Fig. 27.

## 8. The region beyond the eyewall: Rainbands and eyewall replacement

### a. The eyewall–rainband complex—An overview

In a tropical cyclone, beneath the overriding cirrus canopy issuing from the primary eyewall, the rainfall outside the eyewall is primarily associated with a complex of *rainbands*, which have a spiral geometry as opposed to the quasi-circular geometry of the eyewall. Figure 30 shows an idealized but typical array of rainbands and

eyewalls in a tropical cyclone. For purpose of discussion, Fig. 30 includes a circle delineating an approximate boundary between the outer *environment* of the tropical cyclone and an *inner core*, which is dynamically controlled by the cyclonic vortex circulation. Two eyewalls are shown because often an older *primary eyewall* will be replaced by a surrounding *secondary eyewall*. Eyewall replacement will be discussed further in section 8e. The rainbands are of three basic types. *Distant rainbands* are located far from the storm center and are composed of buoyant convection aligned with confluence lines in the large-scale low-level wind field spiraling into the hurricane vortex. The distant rainbands are radially far enough from the eye of the storm that the vertical structure of the convection within them is relatively unconstrained by the dynamics of the inner-core vortex of the cyclone. Figure 31 shows how the convective available potential energy (CAPE) increases with increasing radial distance from the



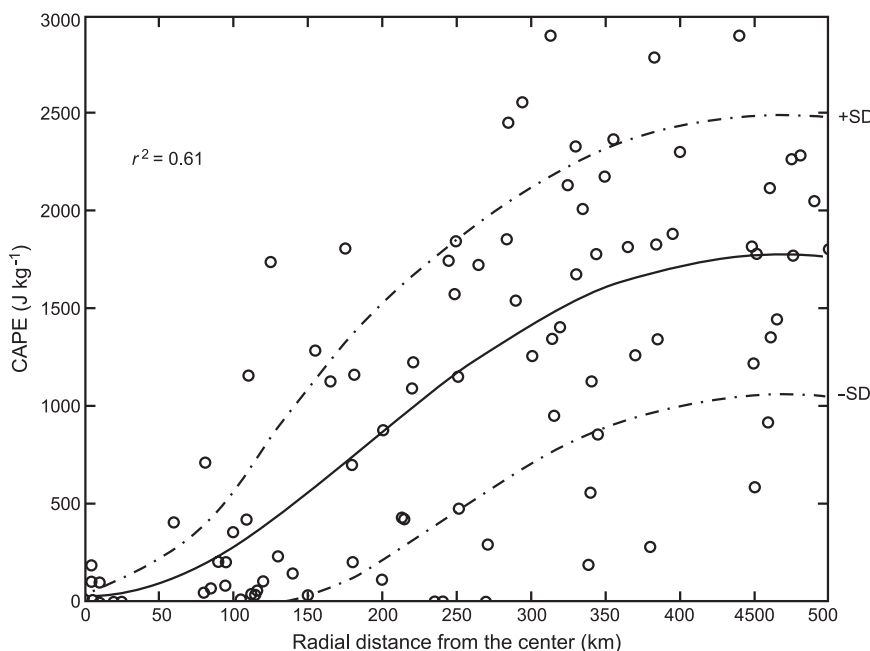


FIG. 31. The radial distribution of CAPE by dropsonde observations in six tropical cyclones as a function of radial distance from the storm center. The solid line is a polynomial curve fit to the data. Plus or minus 1 standard deviation from the range-dependent mean is shown by the dashed lines. Open circles represent the values of individual dropsondes. (From Bogner et al. 2000.)

center of the storm. The distant rainbands occur mainly where the CAPE is highest and most like that of the large-scale environment of the cyclone. The *principal rainband* tends to be more or less stationary relative to the storm. The proximate cause of the principal rainband has not been firmly established. Willoughby et al. (1984b) and Willoughby (1988) argue that the quasi-stationary rainband is a result of some sort of interaction between the wind field of the vortex and that of the large-scale surroundings of the storm. But this has not been verified. In some cases, the upwind end of the principal rainband lies outside the inner-core region of the storm, and the vertical structure of convective elements at the upwind end of the principal rainband is similar to those in the distant rainbands, operating as convection whose vertical structure is relatively unaffected by the vortex dynamics of the tropical cyclone. However, much of the principal rainband's length lies within the inner core of the storm, and the vertical structure of the convective elements along much of the length of the principal rainband is constrained by the dynamics of the inner-core vortex, and these convective cells take on a unique structure, which will be discussed in section 8c. At its downwind end, the principal rainband is more stratiform and becomes nearly tangent to the eyewall. *Secondary rainbands* are located within the inner-core region of the storm and tend to be smaller and more transient than the principal rainband. Their behavior

is consistent with Rossby wave normal modes of the primary vortex, propagating rapidly both radially and azimuthally. The azimuthal component of their motions is cyclonic but somewhat slower than the cyclonic tangential wind component. Often the secondary rainbands connect tangentially to or merge with a principal rainband or an eyewall. Secondary rainbands are discussed further in section 8d.

The different structures of eyewalls, inner-core rainbands, and distant rainbands in the stationary rainband complex are brought out statistically in Fig. 32. The three figure panels contain reflectivity data from the TRMM satellite's precipitation radar (Kummerow et al. 1998) for four hurricanes that reached an intensity of category 4 or 5<sup>9</sup> in the Caribbean Sea/Gulf of Mexico region in 2005 (more information about the intensity scale is also available online at [www.nhc.noaa.gov/aboutsshs.shtml](http://www.nhc.noaa.gov/aboutsshs.shtml)). The radar reflectivity data have been compiled into CFADs stratified according to whether data were obtained in eyewall echoes, in the inner-core region excluding the eyewall, or in the distant environment region, that is, in distant rainbands or in the

<sup>9</sup> Tropical cyclone intensity is commonly expressed on the Saffir–Simpson scale, where category 4 means a maximum wind of 59–69 m s<sup>−1</sup> and category 5 refers to a maximum wind of >69 m s<sup>−1</sup> (Saffir 2003).

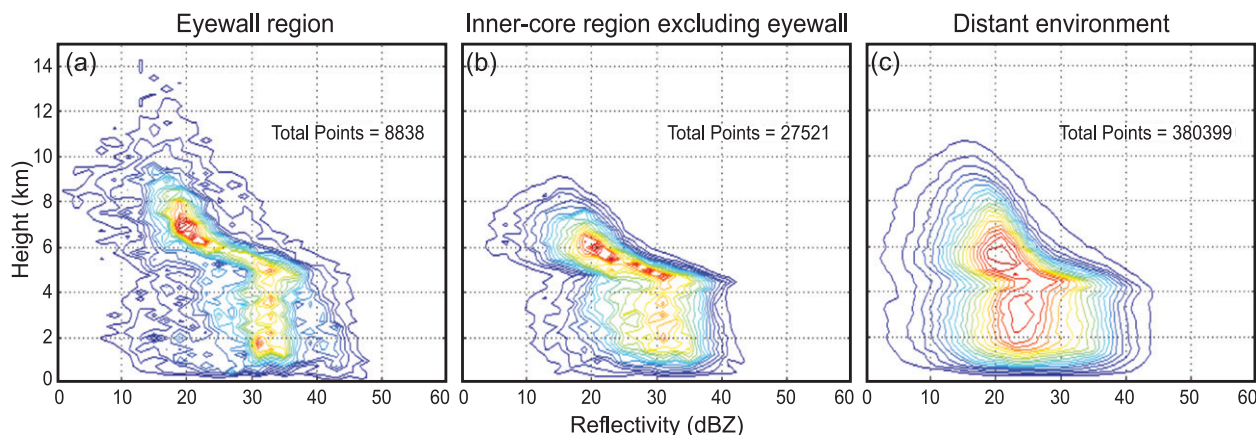


FIG. 32. CFADs of radar reflectivity observed in four intense hurricanes over the Gulf of Mexico (Dennis, Emily, Katrina, and Rita). Contours represent the frequency of occurrence relative to the maximum absolute frequency. Altitudes are geopotential height relative to the surface of the earth. The ordinate of the CFAD is altitude (in 250-m increments, or bins) and the abscissa is reflectivity (in 1-dBZ bins). Data are stratified according to whether they were obtained (a) in the eyewall echoes, (b) in rainbands in the inner-core region outside the eyewall, or (c) in the environment of the storm (i.e., in distant rainbands or the upwind ends of principal rainbands). (Provided through the courtesy of D. Hence.)

extreme upwind ends of principal rainbands. The results in Fig. 32 are highly consistent with the airborne radar data statistics of Black et al. (1996). The CFADs in the three panels of Fig. 32 all contain a mix of convective properties (high reflectivity values in outliers) and stratiform characteristics (hint of brightband at the 4–5-km level). However, the proportions of these characteristics are different in each panel.

The CFAD for the eyewall echoes (Fig. 32a) exhibits a distribution that is highly concentrated (peaked) around its modal value. The statistical uniformity of the radar echo in the eyewall region is guaranteed by the primary circulation of the tropical cyclone. The RMW is located within the eyewall cloud (Fig. 8b), and the strongly sheared winds quickly smear out any variations in the azimuthal direction, leading to an intense but relatively uniform echo encircling (or partially encircling) the eye of the storm. The CFAD further shows that the central maximum of reflectivity in the eyewall tends to be accompanied by a lot of outliers; a wide region of relatively infrequent echoes, some extending to great heights and some exhibiting very high values of reflectivity appear in the distribution. These outliers of the distribution are apparently the signal of intermittently occurring intense buoyant convective cells superimposed on the slantwise secondary circulation of the tropical cyclone vortex (see sections 7a–c).

According to Fig. 30, the portion of the inner-core region lying outside the eyewall but inside the distant environment region contains the inner portion of the principal rainband, as well as secondary rainbands, and secondary eyewalls (if an eyewall replacement occurs).

This intermediate region is undoubtedly controlled to a great extent by the dynamics of the primary vortex. In several respects, the CFAD for this region (Fig. 32b) resembles that of the eyewall in that it exhibits a frequency distribution that is highly concentrated, or peaked. However, it also differs noticeably from the eyewall region in two respects. First, the CFAD shows that in the inner-core region outside the eyewall echoes do not extend as high as in the eyewall region, probably because they lie directly under and are constrained vertically by the layer of strong radial outflow exiting the top portion of the eyewall cloud (see further discussion in section 8c). The second difference is the lack of outliers of intense reflectivity in the inner-core rainband CFAD, as compared with the eyewall CFAD, which indicates that the inner-core rainbands have fewer extremely intense buoyant convective cells and/or a higher proportion of stratiform precipitation than the eyewall. The minimum of lightning activity at 100–200 km from the storm center (Fig. 29) further indicates relatively subdued convective activity in the inner-core region lying between the eyewall and the distant environment. The dynamics of the inner-core rainbands and secondary eyewalls will be discussed further in sections 8c–e.

The CFAD for the distant-environment region, consisting of echoes in distant rainbands and the extreme upwind portions of principal rainbands, differs substantially from those of the eyewall or inner-core region lying outside the eyewall. Most noticeably, it exhibits a broader distribution of reflectivity (Fig. 32c). This broad, as opposed to peaked, distribution resembles CFADs of non-tropical cyclone buoyant convection in

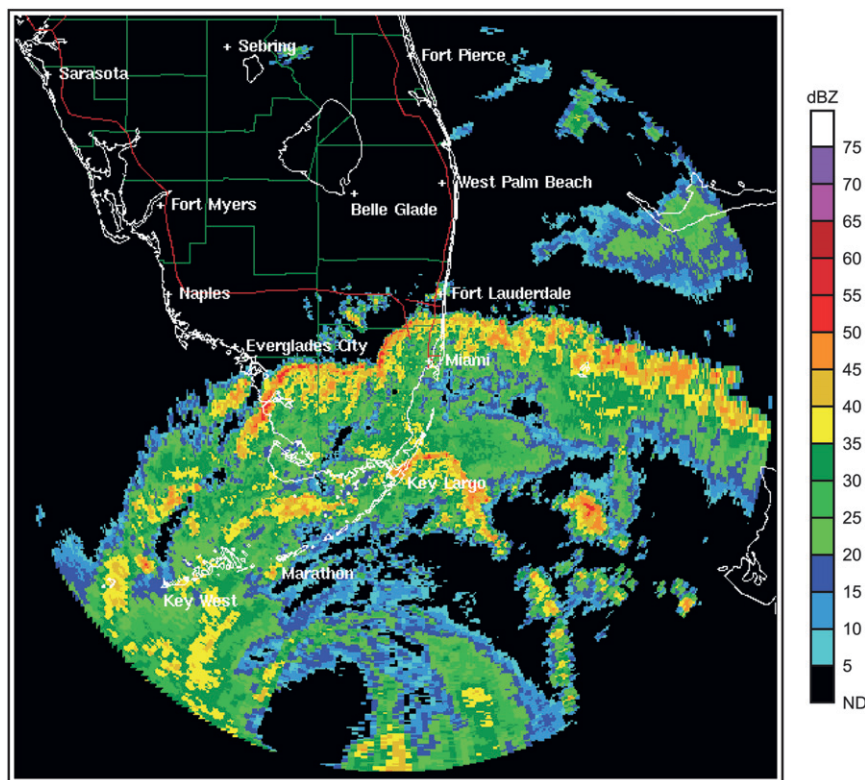


FIG. 33. Pattern of radar reflectivity (dBZ) detected in Hurricane Rita (2005) by the Miami, FL, WSR-88D at 1413 UTC 20 Sep 2005. Arc lines of echo in the distant rainbands were located near Key Largo, Miami, and Everglades City, FL.

an active stage of development (Yuter and Houze 1995). In addition, relative to the intermediate region (Fig. 32b), the distant rainband CFAD extends to greater heights. From these CFAD characteristics, we conclude that the distant rainbands and sometimes the upwind ends of principal rainbands are relatively free of constraints of inner-core dynamics of the tropical cyclone and behave as ordinary buoyant deep convection.

#### *b. Distant rainbands*

Anecdotally, the convective nature of the distant rainbands is well known by pilots and scientists who fly into tropical cyclones. The strong vertical air motions encountered in the distant rainbands are often the roughest part of the flight, with the inner-core rainbands and eyewalls being typically less problematic (except for the occasional extremely intense updraft and/or strongly rotational eyewall feature such as described by Marks et al. 2008; see section 7b). As seen in Fig. 31, the CAPE has its highest values on the outer periphery of the cyclone. These facts lead us to expect that the distant rainbands are different and more traditionally convective than clouds in the inner regions of the tropical cyclone.

Downdrafts as well as updrafts are pronounced in the distant rainbands. As indicated in Fig. 30, the convective elements in the distant rainbands can sometimes develop arc-shaped radar-echo lines, similar to a bow echo (Fujita 1978; Lee et al. 1992; Jorgensen and Smull 1993; Weisman 2001; Davis et al. 2004; Wakimoto et al. 2006a,b), which is a signature of strong downdraft spreading out below a convective cell. Examples of arc lines in the distant rainbands of Hurricane Rita (2005) can be seen northwest of Miami, southeast of Everglades City, and near Key Largo, Florida, in Fig. 33. These arc lines have been noticed primarily via coastal radars, when tropical cyclones are reacting to land surfaces and where the environment may have low enough humidity to support strong downdrafts. The extent to which similar arc lines occur in distant rainbands over the open ocean has not been determined, although arc lines seen emanating from storms in satellite data are sometimes noted by forecasters as an indication of the presence of dry air in the storm.

The maximum of lightning activity seen at distances of  $> \sim 200$  km from the storm center in Fig. 29 further indicates that the distant rainbands are of a more convective nature. As noted in section 7g, lightning in the distant rainbands is probably different from that formed

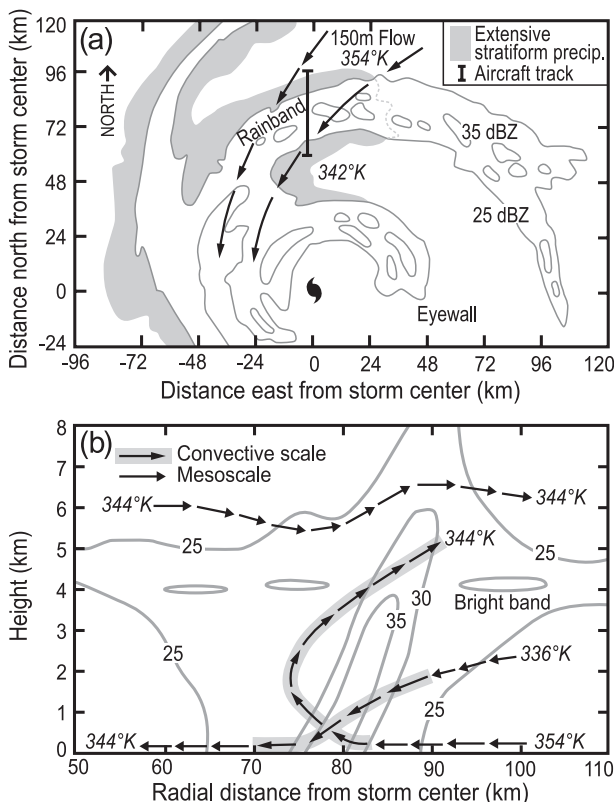


FIG. 34. Rainband observed by aircraft penetration of Hurricane Floyd (1981). (a) A horizontal projection of the data, showing the track. (b) A vertical cross section taken along the aircraft track, which was across the axis of the rainband. In (a) and (b), contours show radar reflectivity (dBZ), temperatures refer to values of equivalent potential temperature, and arrows indicate airflow (on either the convective scale or mesoscale, as indicated by the style of arrow). (From Barnes et al. 1983.)

by the unique electrification process of the eyewall cloud (Fig. 27); it more likely forms by the processes associated with ordinary cumulonimbus (Williams 1988; Houze 1993; Wallace and Hobbs 2006).

### c. The principal rainband

The principal rainband lies mostly within the inner-core region (Fig. 30), and the convection within this portion of the principal rainband is unlike that seen outside the context of a tropical cyclone. Examining data collected on a flight crossing a rainband in Hurricane Floyd (1981), Barnes et al. (1983) found that convective-scale cells in a rainband have the distinctive structure illustrated in Fig. 34. The airflow indicated was the result of three-dimensional transient features, the net result of which was the up- and downdrafts overturning the air crossing the band. Powell (1990a,b) found similar convective structure analyzed from aircraft data obtained in Hurricanes Josephine (1984) and Earl (1986). The

rainband analyzed by Barnes et al. (1983) was spiraling in toward and becoming tangent to the eyewall (Fig. 34a). In a vertical cross section along the flight track (Fig. 34b), the radar reflectivity data showed a sloping cell of high reflectivity surrounded by less intense stratiform precipitation exhibiting a bright band in the melting layer. The radially outward slope of the cell is typical of cells within rainbands. The slope of the rainband cells, moreover, is in the same outward direction as the slope of the eyewall cloud (Figs. 5, 8, 9, 12, 19, and 25). Inward-flowing boundary layer air of high  $\theta_e$  rises out of the boundary and then reverses its radial direction and flows outward along a slantwise path as it continues to rise. This slantwise upward motion in relation to the sloping radar echo maximum is similar to the slantwise motion in the eyewall (Fig. 25). However, the overturning circulation in Fig. 34b is confined to relatively low levels—below 8 km. Although the cells' intensities in terms of vertical velocity and radar reflectivity are considerable, they are vertically confined. The radial outflow from the eyewall dominates the flow above  $\sim 8$ –10 km, and the rainband overturning tends to be confined below. This fact is consistent with the CFADs in Fig. 32, which show that the rainbands in the inner-core region immediately outside the eyewall region do not extend as high as the eyewall or the distant convection.

Downdraft motions also occur in association with the sloping convective-scale precipitation cores (Fig. 34b). Whereas in the case of the eyewall the downdraft appears to play a minor role in the dynamics (section 7d), the downdraft in the rainband can be significant. It transports low- $\theta_e$  air from the midlevel environment into the boundary layer. This type of downdraft is evidently the primary mechanism by which negative flux of  $\theta_e$  occurs at the top of the boundary layer in region III of the idealized tropical cyclone in Fig. 12a. After entering the boundary layer as the downdraft spreads out below cloud, the low- $\theta_e$  air flows on toward the center of the storm, thus reducing the overall potency of the tropical cyclone.

The up- and downdraft structure illustrated in Fig. 34 was inferred indirectly from radar reflectivity and dropsonde data. More recently, radar measurements were obtained in Hurricanes Katrina (2005) and Rita (2005) during the Hurricane Rainband and Intensity Change Experiment (RAINEX; Houze et al. 2006). In comparison with the earlier aircraft flights studied by Barnes et al. (1983) and Powell (1990a,b), the aircraft sampling in RAINEX provided more detailed and more comprehensive coverage of the principal rainbands in Katrina and Rita by employing the dual-antenna Electra Doppler radar (ELDORA; see Hildebrand et al. 1996). The ELDORA system obtained airborne radar data



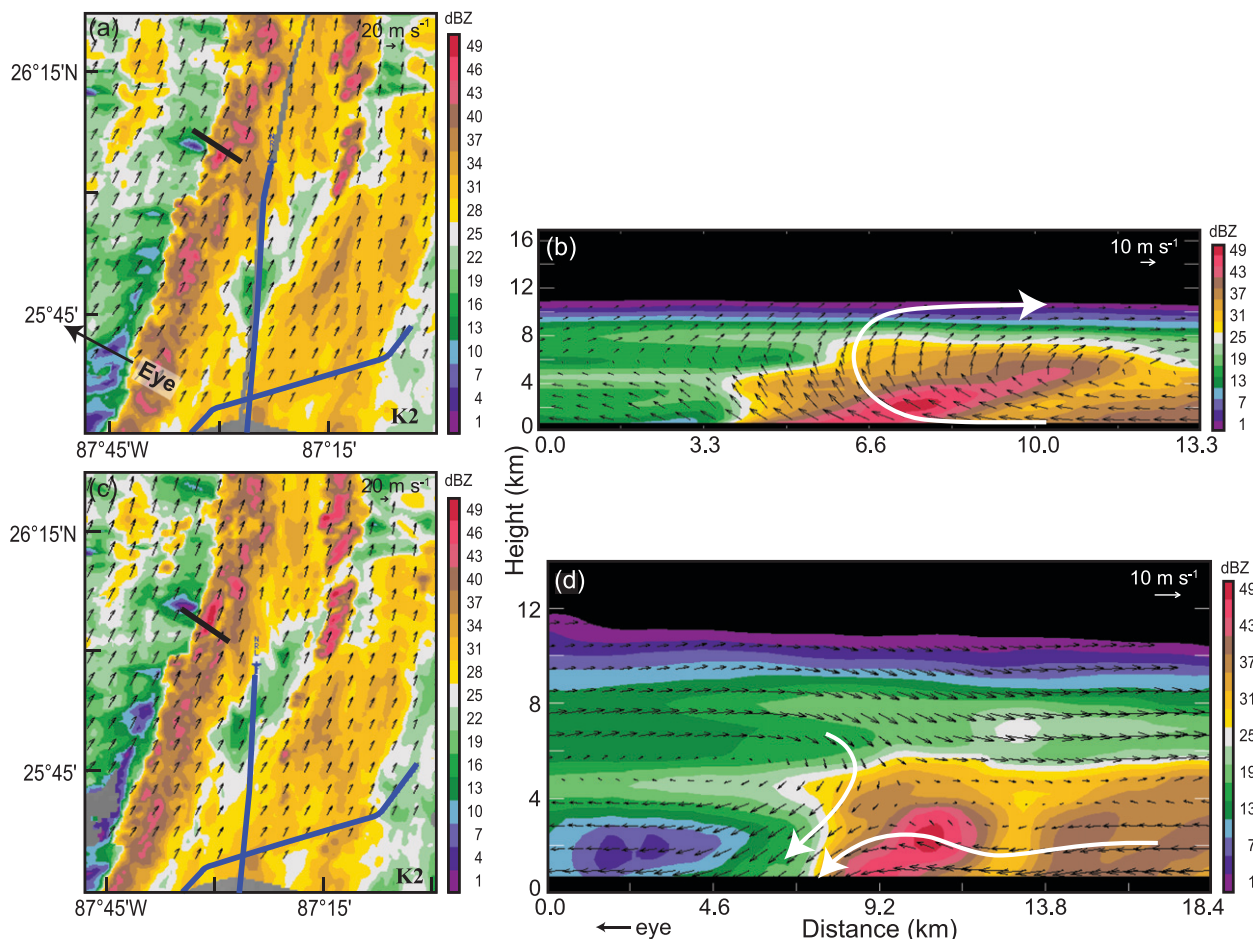


FIG. 35. Composite of Doppler radar data collected in Hurricane Katrina by an aircraft during 2026–2036 UTC 28 Aug 2005 showing the structure of the principal rainband when the storm was at category-5 intensity. (a) Plan view showing the reflectivity pattern (color shading) and Doppler-derived airflow (vectors) at the 3.5-km level observed while the aircraft was flying along the track shown by the blue line. The straight black line passes through a convective-scale updraft associated with one of the intense reflectivity cells. (b) The reflectivity and Doppler-radar-derived airflow in the plane of the cross section along the black line in (a). (c) Plan view showing the reflectivity pattern and Doppler-derived airflow at the 2.0-km level observed while the aircraft was flying along the track shown by the blue line. (d) The reflectivity and Doppler-radar-derived airflow in the plane of the cross section along the black line in (c). (Adapted from Hencé and Houze 2008.)

with higher resolution than previously available, and the RAINEX strategy of flying the Doppler aircraft along rather than across the rainbands allowed for more comprehensive spatial coverage. The RAINEX flights have confirmed that the circulation patterns hypothesized by Barnes et al. (1983) and Powell (1990a,b) really occur, that they are highly three-dimensional, as suggested by the earlier studies, and that these convective structures are repeatable from cell to cell in a particular rainband and from storm to storm (Hence and Houze 2008). Figure 35 shows the air motions and radar reflectivity that they found in a segment of a principal rainband of Hurricane Katrina. The black lines in Figs. 35a,c show the locations of two cross sections across a convective cell in this band (Figs. 35b,d). Although

these two sections, taken at the same time, are only a short distance apart, Fig. 35b shows an updraft structure similar to the idealization in Fig. 34, while Fig. 35d shows the downdraft, also similar to the idealization. The up and down circulations exist in close proximity but are intertwined in a three-dimensional pattern as the updraft transports air of high moist static energy upward and the downdraft simultaneously transports air of low moist static energy downward.

Samsury and Zipser (1995) analyzed aircraft observations showing that rainbands tend to have a “secondary horizontal wind maximum,” or jet located along the axis of the rainband. Hence and Houze (2008) found that the principal rainbands in the category-5 stages of Hurricanes Katrina (2005) and Rita (2005) had well defined jets along

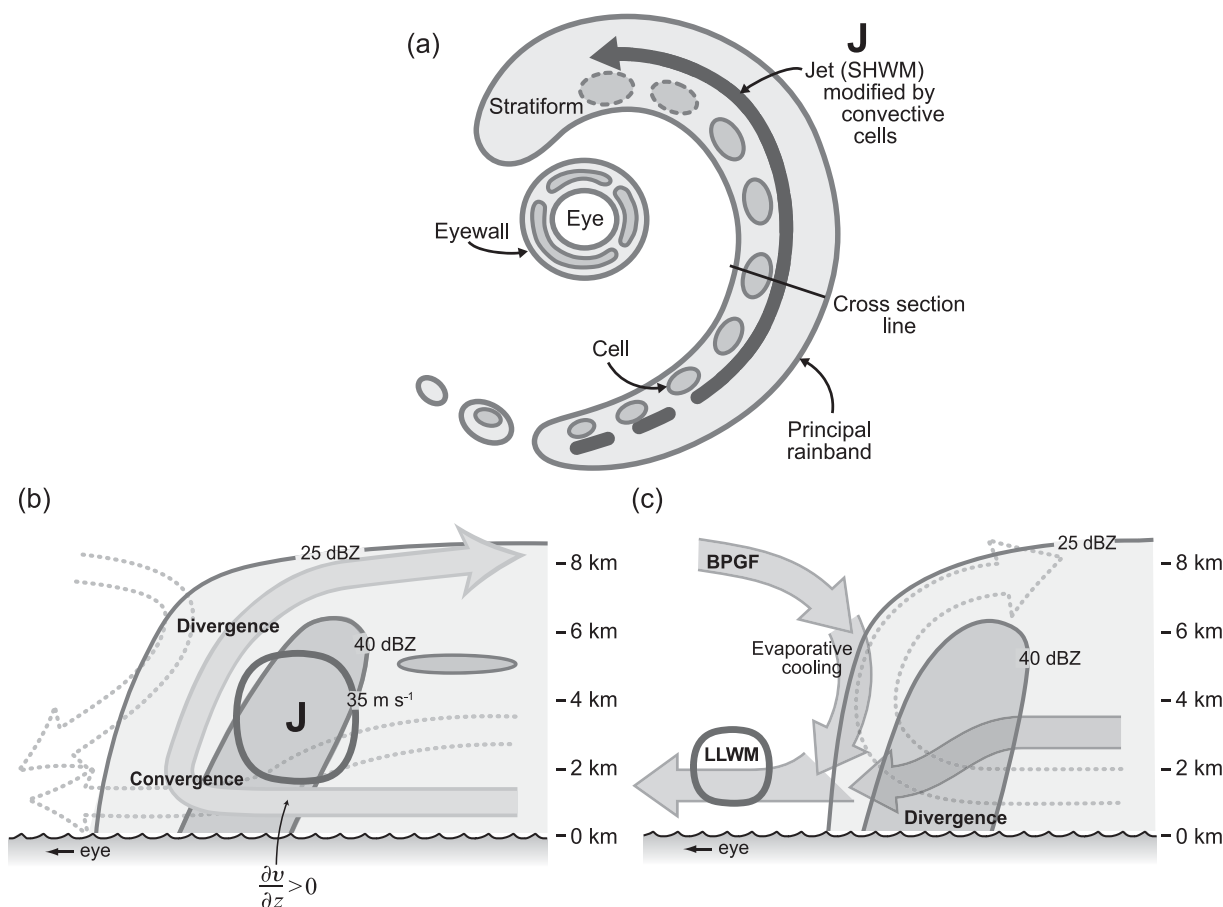


FIG. 36. (a) Plan view of the principal band within an idealized mature tropical cyclone. A jet is located along the axis of the rainband (heavy arrow). Embedded convective cells are located on the radially inward side of the jet. Shading indicates precipitation (dBZ; light shading is 25 dBZ and dark shading is 40 dBZ). The size of the convective cells indicates the level of maturity, with the dashed border indicating collapsing cells. (b) Cross section across the middle portion of the rainband, emphasizing the updraft structure. The wide arrow indicates the in, up, and out circulation of the convective-scale updraft. Low-level convergence, vertical convergence of vorticity flux, and midlevel divergence regions are indicated. The symbol J indicates the location of the jet along the rainband axis, with the associated low-level shear. Dotted arrows in the background indicate nearby downdrafts slightly outside the plane of the cross section. Here,  $\frac{\partial v}{\partial z}$  is the vertical shear of the tangential wind  $v$ . (c) Schematic of the possible convective motions along the cross section of a mature principal rainband, emphasizing the downdraft structures. The series of solid arrows originating in the upper levels indicates an inner-edge downdraft (IED) that is labeled by the two forcing mechanisms that drive it, the buoyancy pressure gradient acceleration (BPGF) field and evaporative cooling, passing through a region of vertical convergence of negative vorticity flux as indicated. The solid arrow originating in the lower levels indicates the low-level downdraft that creates a region of divergence within the rainband as indicated. The dotted arrow in the background indicates the overturning updraft that passes through a region of vertical convergence of positive vorticity flux. The circled region indicates the low-level wind maximum (LLWM) that is impacted by the IED. (Adapted from Hencé and Houze 2008 and Didlake and Houze 2009.)

their axes. The occurrence of the jet suggests that the rainbands have dynamics somewhat in common with eyewalls, which are tied closely to the RMW, and the outward-sloping updrafts seen in Figs. 34 and 35b have a geometry somewhat like the slantwise overturning in the eyewall cloud region. It could be that the rainbands with secondary wind maxima are incipient but somewhat frail attempts to form eyewall-like entities. Hencé and Houze (2008) further inferred from the RAINEX high-

resolution dual-Doppler radar data that the convective cells in the rainbands act to accelerate the jet. Figure 36 illustrates the process that they envisioned. In the vertical cross section sketched in Fig. 36b, an updraft structure similar to those seen in Figs. 34 and 35b is indicated. Under the jet, the horizontal vorticity associated with the shear ( $\frac{\partial v}{\partial z} > 0$ ) points inward. Where the radial inflow (wide arrow) flows under the jet, the magnitude of the horizontal vorticity is large. Where the radial inflow turns

upward this vorticity is converted to vertical vorticity, which is intensified by stretching by the convergence at the base of the updraft. The updraft transports the vertical vorticity upward, and since divergence at the  $\sim 7$ -km level stops the updraft, vertical advection of vertical vorticity leads to accumulation of positive vorticity in midlevels that then strengthens the jet (which is directed into the page). The jet along the axis of the principal rainband may be a key to its coherence as a mesoscale flow entity. If so, the positive feedback to the jet by the convective cells indicates that the principal rainband is a manifestation of cooperative mesoscale-convective dynamics.

The air motions in Figs. 34b and 35b are streamlines in the cross-rainband plane. The cross-band horizontal component of the wind in these cross sections is small relative to the along-band component, which is roughly in the same direction as the tangential component of the tropical cyclone's primary vortex wind. Thus, the trajectories of the air in the overturning updraft circulation deliver high moist-static energy air as well as cloud and precipitation particles to upper levels far downwind of the features seen in the cross-sectional plane (recall Fig. 26). Similarly, the downdraft circulation injects air of low moist static energy into the boundary layer azimuthally far downwind of the cross section. Figure 37, from a classic study of Atlas et al. (1963), illustrates how the ice particles generated in the convective cell of a hurricane rainband can be carried down the length of the rainband while slowly falling. When these ice particles pass downward through the melting layer, they can produce a radar bright band, partly accounting for the stratiform precipitation distributed along the rainband and surrounding the embedded active cells. Note in Fig. 34 the greater prevalence of stratiform precipitation found by Barnes et al. (1983) in the downwind portion of the rainband that they investigated. Hence and Houze (2008) also found that the principal rainbands of Katrina (2005) and Rita (2005) were much more stratiform on their downwind ends.

The downdrafts associated with the convective cells of the principal rainbands were also documented in detail in RAINEX. Using Doppler radar data collected in the principal rainband of Hurricane Katrina (2005) when the storm was at category-5 intensity, Didlake and Houze (2009) showed that the convective cells in the principal rainband induce two distinct convective-scale downdrafts (Fig. 36c). One is the downdraft noted by Barnes et al. (1983) and illustrated in Fig. 34, which enters the rain core at about the 3-km level and sinks as a result of evaporation and precipitation drag. The second is an *inner-edge downdraft*, which they suggest is induced as a result of three different downward forcing

effects, all occurring in response to the updraft cell. At upper levels, the air is forced downward by the pressure gradient force that occurs in the vicinity of the updraft cell as a gravity wave response to the upward accelerating buoyant updraft of the cell. At midlower levels, the air on the flank of the updraft cell is accelerated downward by a dynamically induced pressure gradient force, which occurs where the shear of the wind maximum along the rainband axis interacts with the updraft core (Powell 1990a).<sup>10</sup> Finally, as the inner-edge downdraft penetrates down into the leading edge of the heavy precipitation echo, evaporation of rain into the downdraft air renders the air negatively buoyant, and air accelerates downward on the leading edge of the principal rainband. This downward motion due to pressure forces and evaporative cooling produces a sharp reflectivity gradient on the inner edge of the rainband (bottom-left parts of Figs. 35a,c). The intense inner-edge downdraft produces a low-level divergence of vorticity that builds a low-level wind maximum just inward of the principal rainband. This wind maximum is advected inward toward the center of the storm and thus contributes to intensification of the tropical cyclone.

#### *d. Vortex Rossby waves and secondary rainbands*

Theory of idealized tropical cyclone-like vortices (e.g., Montgomery and Kallenbach 1997) suggests that the flow field near the center of a storm should contain normal-mode disturbances of the general size and shape of the secondary rainbands that occur in the inner-core region outside of the eyewall zone (Fig. 30). Figure 38 from Montgomery and Kallenbach (1997) shows two stages in the evolution of the vorticity field in a dry, barotropic, inviscid, axisymmetric, nondivergent vortex. The idealized basic-state vortex is assumed to have a maximum of vorticity at its center, with the vorticity decreasing monotonically with radius. The idealized vortex is initially disturbed by a wavenumber-2 asymmetry, and in time, the asymmetry takes on the shape of two major spiral bands, as seen in Fig. 38. These bands are packets of *vortex Rossby waves*, which are a generalized version of the planetary-scale Rossby waves that distort the midlatitude westerly jet stream (Holton 2004, 213–217). The comparison to planetary-scale Rossby waves can be seen intuitively by imagining replacing the global jet stream with the tropical cyclone's RMW and letting the center of rotation be located in the eye of the tropical cyclone rather than at the pole. In the planetary

<sup>10</sup> The dynamic pressure perturbation in this situation is proportional (approximately) to the dot product of the shear vector with the horizontal gradient of vertical velocity. See Powell (1990a, p. 913) or Houze (1993, p. 290).

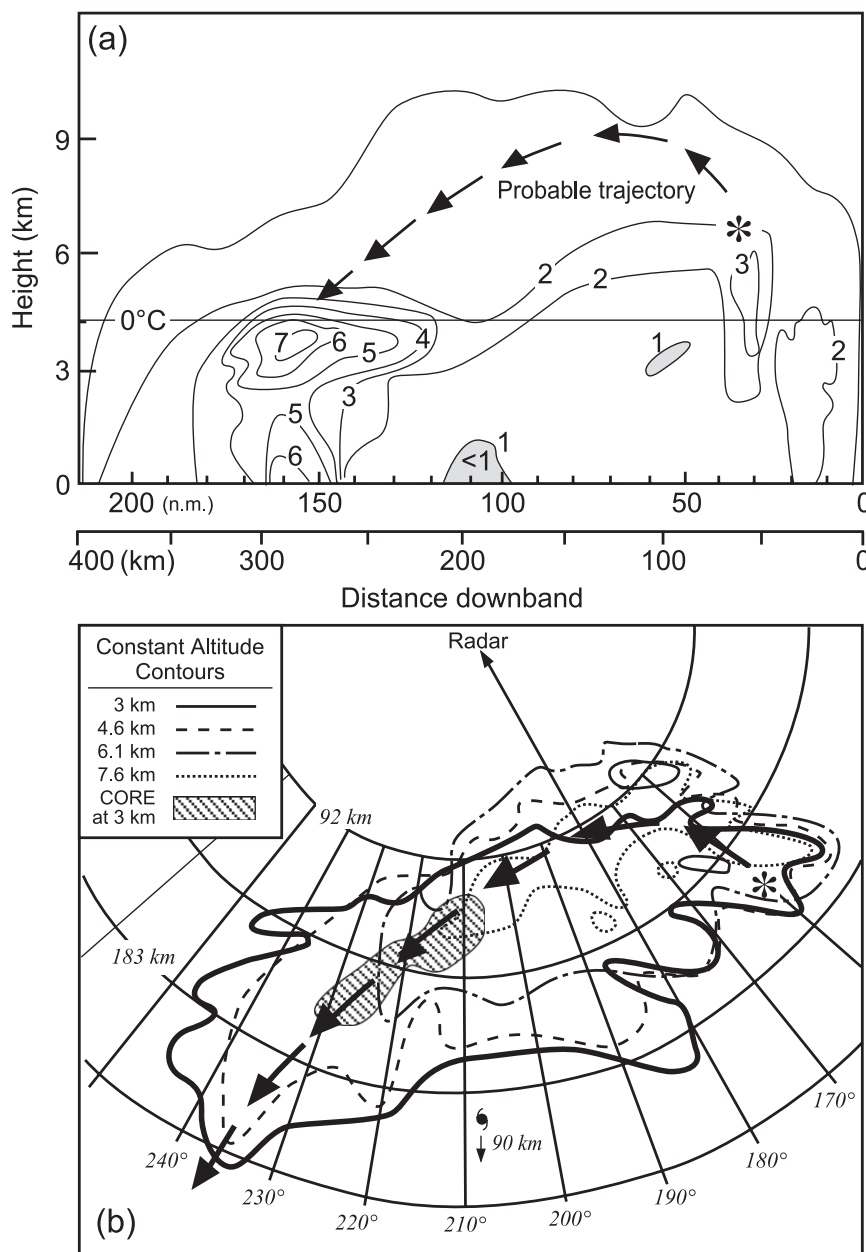


FIG. 37. Hurricane rainband observed by a ground-based radar at North Truro, MA, as Hurricane Esther (1961) moved northward over the Atlantic Ocean just off the coast of the United States. (a) Vertical cross section showing returned power at a sequence of gain settings (numbered 1–7). The horizontal axis is along the axis of the rainband. Only the most intense echoes are seen at the highest gain setting, but the intensity has not been corrected for range from the radar. A hypothesized ice particle trajectory is indicated. (b) Horizontal projection of the hypothesized ice particle trajectory is superimposed on the plan view of the radar echo at several altitudes. For the 3-km level, the location of the intense echo core is shown in addition to the overall echo outline. [From Atlas et al. (1963). Reprinted with permission from *Geofisica Internacional*.]

case, a key parameter affecting planetary-scale Rossby waves is the meridional gradient of the Coriolis parameter. In the idealized tropical cyclone case represented in Fig. 38, the corresponding key parameter is the

radial gradient of the relative vorticity of the basic-state vortex ( $\zeta_0$ ). The dispersion relation for the vortex Rossby waves has been shown by Montgomery and Kallenbach (1997) to be



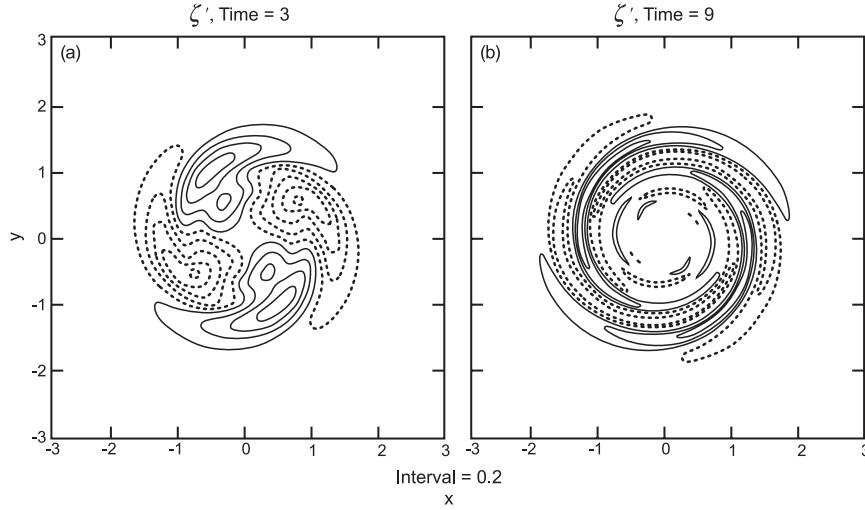


FIG. 38. Vorticity perturbation field at an (a) earlier and a (b) later time in the evolution of a wavenumber-2 vortex Rossby wave perturbation superimposed on a tropical cyclone-like vortex centered in the middle of the domain. [From Montgomery and Kallenbach (1997). Reprinted with permission from the Royal Meteorological Society.]

$$\nu = \frac{\bar{v}_0}{r}n + \frac{n}{R} \frac{\partial \bar{\zeta}_0 / \partial r}{(k_r^2 + n^2/R^2)}, \quad (9)$$

where  $\nu$  is the frequency;  $k_r$  and  $n$  are the radial and azimuthal wavenumbers, respectively;  $\bar{v}$  and  $\bar{\zeta}$  are the tangential velocity and vertical vorticity of the basic-state vortex;  $r$  is the radius measured from the storm center; and  $R$  is a reference radius that satisfies the condition  $k_r R \gg 1$ , which implies that (9) holds for disturbances that have a radial scale that is small relative to  $R$ . The zero subscript indicates quantities evaluated at radius  $R$ . The frequency in (9) implies that the group velocities in the azimuthal and radial directions are

$$C_{g\lambda} = R\bar{\Omega}_0 + \frac{\partial \bar{\zeta}_0 / \partial r}{(k_r^2 + n^2/R^2)^2} \times \left\{ k_{ri}^2 - \frac{n^2}{R^2} \left[ 1 + t^2 R^2 \left( \frac{\partial \bar{\Omega}_0}{\partial r} \right)^2 \right] \right\} \quad \text{and} \quad (10)$$

$$C_{gr} = \frac{-2k_r n (\partial \bar{\zeta}_0 / \partial r)}{R(k_r^2 + n^2/R^2)^2}, \quad (11)$$

where  $\Omega \equiv v/r$  is the angular momentum and  $k_{ri}$  is the initial wavenumber. The radial group velocity  $C_{gr}$  is positive, since the basic-state vorticity  $\bar{\zeta}_0$  in the assumed cyclonelike vortex decreases with radius. An important property of the waves is that the radial wavenumber increases with time according to

$$k_r = k_{ri} - nt \left( \frac{\partial \bar{\Omega}_0}{\partial r} \right), \quad (12)$$

where  $t$  is time. The right-hand side of this equation is positive since the angular momentum of the assumed basic-state vortex decreases with radius. Since the vortex Rossby waves propagate radially outward, the shearing of the basic-state wind field continually reduces their radial wavelengths. The radial squeezing of the wave packets in time is evident from the two snapshots in Fig. 38. Since in time the radial group velocity approaches zero, wave packets collect at a *stagnation radius*. The waves in Fig. 38b have nearly reached their stagnation radius. The waves accumulating at the stagnation radius eventually are homogenized in a ring around the storm as they are sheared by the tangential wind and turbulently mixed. This process is an example of axisymmetrization (section 3d).

Equation (10) describes the azimuthal component of the vortex Rossby group velocity. The rightmost term is negative and grows rapidly in magnitude over time. As a result, the vortex Rossby wave packets slow down and propagate azimuthally at a speed that is considerably less than the mean vortex flow. This behavior is consistent with the fact that the mesoscale features illustrated in Fig. 30, which might be associated with vortex Rossby wave packets, are not simply advected by the rapidly rotating winds in the cyclone.

The theory of vortex Rossby waves exhibiting rain-bandlike behavior in the idealized vortex illustrated by Fig. 38 and (9)–(12) is readily extended to a more realistic divergent flow (Montgomery and Kallenbach 1997). The potential vorticity of the basic-state vortex replaces the basic-state vorticity as the controlling parameter, and

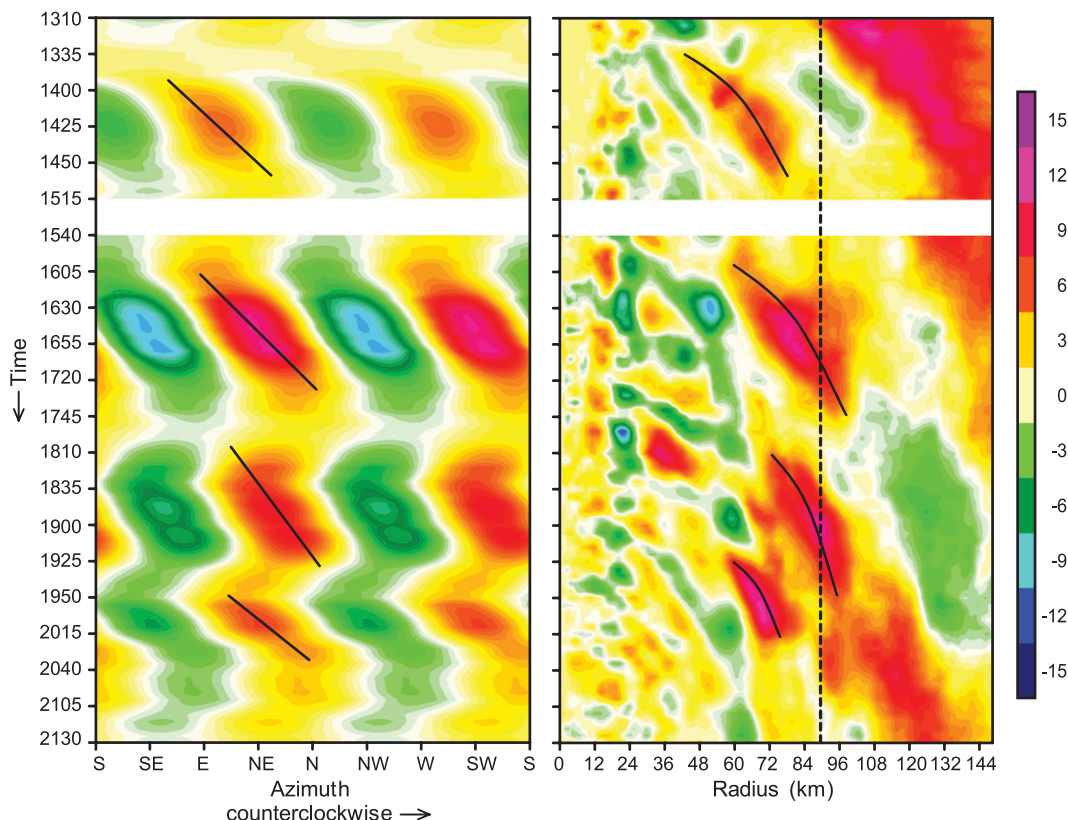


FIG. 39. (left) Azimuth–time plot of the wavenumber-2 asymmetry 80 km from the center of Hurricane Elena (1985). The solid lines track the rotation of the wavenumber-2 asymmetries. (right) Radius–time plot of the wavenumber-2 asymmetry along a radius drawn from the center of Elena toward the southwest. The solid curves track the propagation of the wavenumber-2 asymmetries. The vertical dashed line is the stagnation radius. (From Corbosiero et al. 2006.)

the characteristics of the phase and group velocities change only slightly.

The analysis of Montgomery and Kallenbach (1997) is formulated without moisture effects, which raises questions about its applicability to real tropical cyclones. However, numerical model simulations of hurricanes including moisture, realistic cloud microphysics, and no idealizations of the vortex produce features with the properties of vortex Rossby waves similar to those in the idealized vortex (Chen and Yau 2001; Chen et al. 2003; Braun et al. 2006). There is also support from radar observations. Reasor et al. (2000) analyzed aircraft radar data collected in Hurricane Olivia (1994) at 0.5-h intervals for 3.5 h and found rainbands exhibiting structural and kinematic characteristics of wavenumber-1 and -2 vortex Rossby waves. Corbosiero et al. (2006) examined coastal radar data collected as Hurricane Elena (1985) made landfall. They used two-dimensional Fourier analysis in a polar coordinate system centered on the eye of the storm, and the left panel of Fig. 39 is their time–azimuth plot of the wavenumber-2 component

of the radar echo 80 km from the eye. At this distance from the eye, the echoes had the form of spiral rainbands angling across the 80-km radius circle. At each time of the analysis, two maxima and two minima of reflectivity are apparent, corresponding to the intersections of the rainbands with the 80-km radius circle. In time, the maxima and minima move cyclonically around the storm at a rate of  $\sim 23 \text{ m s}^{-1}$ , in close agreement with the azimuthal group velocity of  $\sim 26 \text{ m s}^{-1}$  estimated by Corbosiero et al. (2006) from the theory of Montgomery and Kallenbach (1997). The right panel of Figure 39 is a time–radius plot of the wavenumber-2 echo pattern along a radius drawn toward the southwest of the storm center. The dashed line is located at the stagnation radius estimated from the theory of Montgomery and Kallenbach (1997; recall section 8d). The solid lines trace the radial propagation of the maxima of the wavenumber-2 pattern of the echoes moving outward across the 80-km circle. The progression of the maxima is at a radial speed of  $\sim 5 \text{ m s}^{-1}$ , which is close to the  $\sim 7 \text{ m s}^{-1}$  estimated from theory. The rapid weakening of the maxima and

minima near the stagnation radius is also consistent with the vortex Rossby wave theory. These group wave packets slow down, thin, and become weaker as they approach the stagnation radius.

The propagation rates in the azimuthal and radial directions seen by Corbosiero et al. (2006) and the seeming agreement of these rates with the vortex Rossby wave theory suggest that at least some of the rainbands in a tropical cyclone are related to these normal modes. Since the principal rainband (see section 8c) is relatively stationary with respect to the tropical cyclone vortex (Willoughby et al. 1984b), it seems that the vortex Rossby wave theory most likely applies to what we have called the smaller, more transient, secondary rainbands depicted in Fig. 30.

#### e. Eyewall contraction and replacement

Often, in intense tropical cyclones, the rainbands in the inner-core region coalesce into a secondary eyewall located radially outward of the original eyewall (Willoughby et al. 1982, 1984b; Willoughby 1988). The schematic in Fig. 30 shows how such a double-eyewall structure might manifest itself. Figure 40 shows a realistic simulation of eyewall replacement in Hurricane Rita (2005). The original eyewall corresponds to the closed annulus of heavy rain in the center of Fig. 40a. To the west and east were spiral rainbands, angling in toward and becoming tangent to the eyewall. A little over a day later (Fig. 40b), eyewall replacement had begun. The rainbands had organized into a coherent circular feature. A more weakly precipitating annular zone, called the *moat*, separated the new outer eyewall from the inner eyewall, which had begun to weaken as the new eyewall cut it off from the low-level frictional inflow of warm moist air.

The process by which the second eyewall forms, although it happens frequently in nature, is not fully understood. One idea is that the intense circulation concentrated at the center of the storm spawns vortex Rossby waves (see section 8c), which propagate radially outward from the original eyewall region (Montgomery and Kallenbach 1997). Depending on the exact structure of the hurricane vortex, the waves may act to concentrate angular momentum at the stagnation radius, where their phase velocity matches the mean swirling flow, and in this way coalesce to form a new outer eyewall (see section 8e). Another suggestion for how the secondary eyewall might form, proposed by Terwey and Montgomery (2008), is that at a radius where radial gradient of mean tangential velocity exhibits certain properties, convectively generated vorticity and kinetic energy anomalies cascade upscale and are axisymmetrized

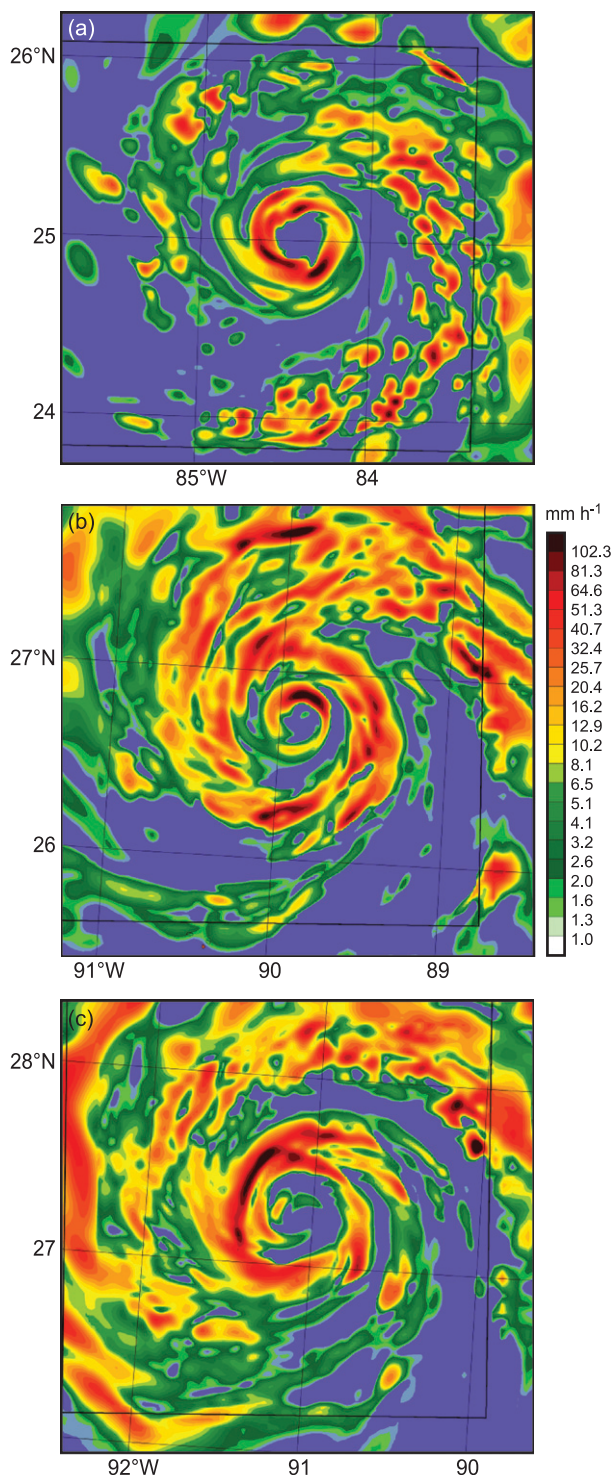


FIG. 40. Forecast of surface rainfall intensity in Hurricane Rita (2005): (a) 0715 UTC 21 Sep, (b) 1115 UTC 22 Sep, and (c) 1715 UTC 22 Sep. Colors show the rainfall rate ( $\text{mm h}^{-1}$ ) at the sea surface generated by a high-resolution numerical model. [From Houze et al. (2007). Reproduced with permission from the American Association for the Advancement of Science.]



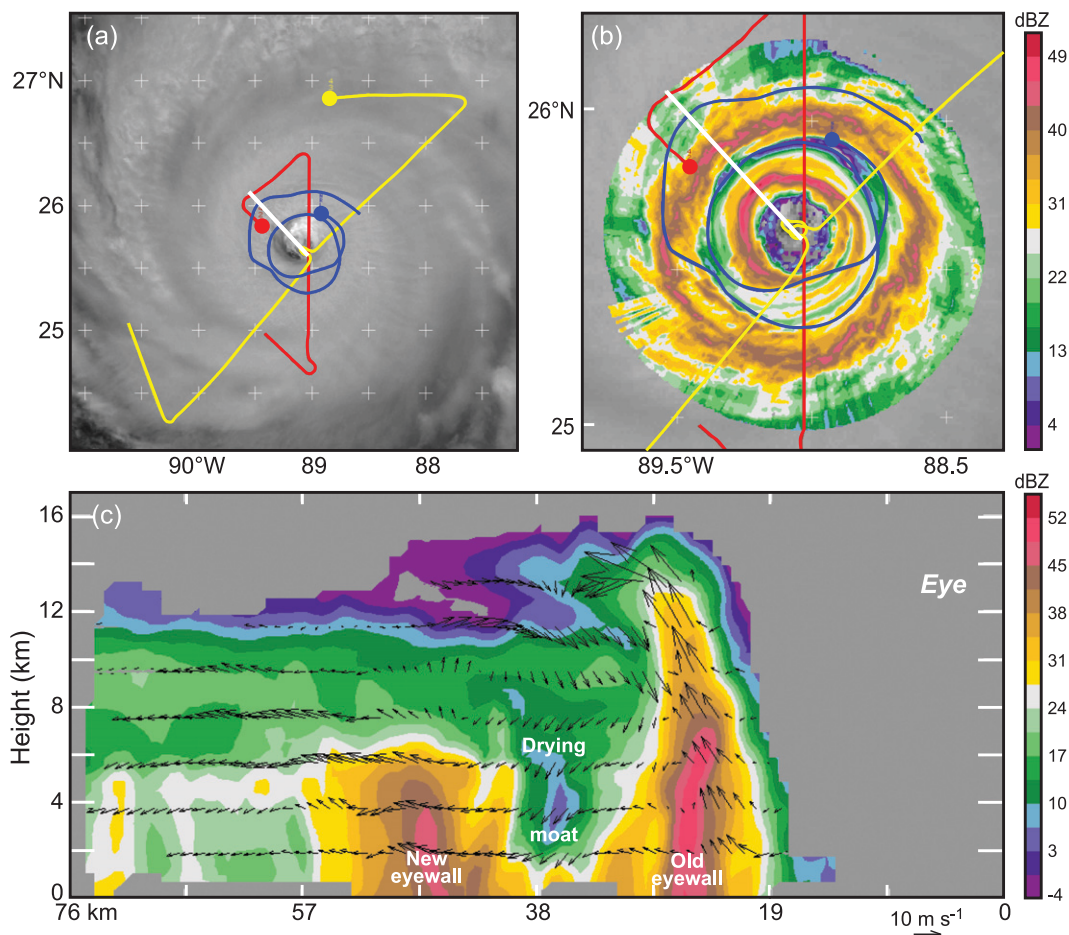


FIG. 41. Aircraft data collected in Hurricane Rita (2005) between 1800 and 1820 UTC 22 Sep. (a), (b) Plan views; (c) vertical cross section across the northwest side of the storm (along the white line in the plan views). Colored lines in (a) denote the flight tracks of three aircraft. The dots show aircraft locations as of 1830 UTC. The yellow track segment is for the 80 min preceding that time; the red and blue track segments are for the preceding 45 min. The yellow track was part of a wide pattern to determine the broadscale structure of the cyclone vortex. The red track was part of an intermediate pattern, with shorter legs across the center of the storm to monitor the two eyewalls. The blue track was a circumnavigation that obtained the key radar and sounding data shown in (b) and (c). The color shades in (b) and (c) correspond to values of radar reflectivity (a measure of precipitation intensity). Vectors in (c) show the component of air motions within the cross sections. [From Houze et al. (2007). Reproduced with permission from the American Association for the Advancement of Science.]

to form a jet that induces locally increased surface fluxes and precipitation generation (i.e., secondary eyewall) at a radius located outside the primary eyewall and RMW. A third idea is that the large-scale humidity field surrounding the storm can favor the formation of rainbands at a certain radius in such a way as to lead to a second eyewall formation (Ortt and Chen 2006, 2008; Ortt 2007). Finally, the convergence caused by storm motion (Chen et al. 2006) might be involved in triggering a new eyewall.

Although the cause of the secondary eyewall is uncertain at this time, recent work has shown some aspects of how the secondary eyewall, once formed, takes over as the new primary eyewall. Figure 40c shows that the inner

eyewall in Hurricane Rita (2005) had nearly vanished, while the outer eyewall, which was replacing it, was taking shape at a greater radius. The maximum winds in the simulated storm had decreased from about  $70 \text{ m s}^{-1}$  to a maximum of  $52 \text{ m s}^{-1}$ , and the new eyewall had a smaller radius in Fig. 40c than in Fig. 40b, indicating that it had begun to contract, as expected from theory (see section 5). RAINEX aircraft radar data collected in Rita (Fig. 41) confirmed the double-eyewall structure suggested by the model forecast (Houze et al. 2007). Two concentric eyewalls were on either side of circular moat of weak echo, and the aircraft data showed that a distinct wind maximum accompanied each eyewall (Houze et al. 2006).



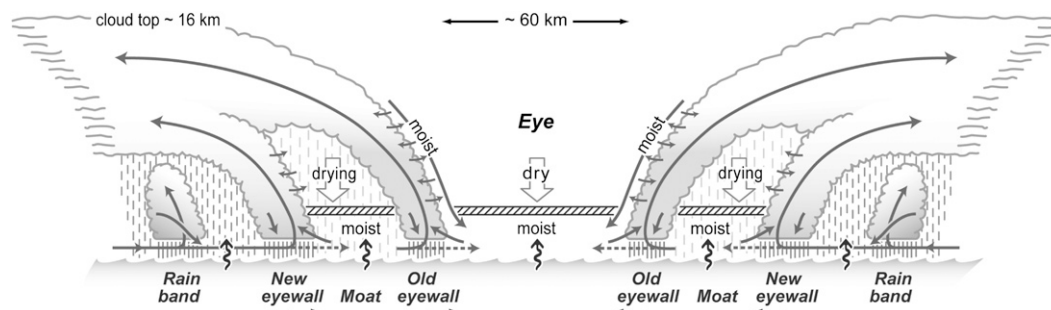


FIG. 42. Conceptual model of the inner region of a tropical cyclone undergoing eyewall replacement. Arrows at the bottom indicate that both eyewalls are contracting inward. The density of vertical lines below the clouds indicates relative precipitation intensity. Thin arrows show the direction of air motion relative to the storm. Dashed segments indicate partially interrupted flow. Small, wavy arrows at the sea surface indicate water vapor mixing upward from the ocean surface. The broad arrows indicate the downward motion induced by the heating in the eyewall and (possibly) momentum mixing across the inner edges of the eyewalls. The hatched zone shows the top of the near-surface moist layer, which is capped by the stabilizing and drying effect of subsiding air above. The moist layer is not to scale; its depth is exaggerated to show details of the airflow at low levels. [From Houze et al. (2007). Reproduced with permission from the American Association for the Advancement of Science.]

Using RAINEX Doppler radar data and dropsondes, Houze et al. (2007) showed that during the eyewall replacement in Rita, the moat took on the characteristics of a hurricane eye (see section 5). Figure 41c shows the detailed wind field inferred from the ELDORA data in a vertical section across the moat and the two eyewalls. The moat region was occupied throughout its volume by downward air motion resembling that in the eye of the storm. Figure 42 conceptualizes the picture. The center region, consisting of the eye and the old eyewall, is the same as the eye structure illustrated in Fig. 9. The new eyewall and moat mimic the old eyewall and eye, respectively. During the replacement, the moat takes on the dynamic behavior of a hurricane eye. Environmental air trapped in the zone between the old and new eyewalls is then forced to sink, warming and drying in the process—just as the eye itself formed dynamically when air from the environment was surrounded by and responded to the original eyewall.

Once the dynamics and thermodynamics of the moat region come to resemble those of an eye, the inner eyewall weakens over time, but only gradually. Even though Rita's second eyewall had enclosed the original eyewall by the time of Fig. 41, the inner eyewall remained strong for 12 h (indicated by tall and intense inner-eyewall radar echoes, as in Fig. 41c). This slow demise of the inner eyewall is typical of tropical cyclones undergoing eyewall replacement (Willoughby et al. 1982, 1984b; Willoughby 1988). The inner eyewall continues to receive radial inflow that underruns the new outer eyewall, and it draws some additional small amount of energy from the eye (Bryan and Rotunno 2009); recall from Figs. 9, 12, 15, and 17 how low-level air can enter the eyewall from the eye, and that this air has very high  $\theta_e$ .

The inner eyewall can apparently survive through this mechanism for several hours before finally expiring. It is perhaps the compensating downward motion from the new eyewall (Shapiro and Willoughby 1982; Pendergrass and Willoughby 2009) that eventually weakens the inner eyewall.

## 9. Conclusions

Because of the underlying dynamics of a strong vortex, the clouds in a tropical cyclone take on forms not seen anywhere else in the earth's atmosphere. They exhibit a wide variety of processes and forms, which have been described in a rich and comprehensive but somewhat scattered literature. This review integrates this body of knowledge into a single reference. It is hoped that this synthesis will guide efforts to represent clouds accurately in models aimed at highly detailed forecasting, warning, and climatic assessment of tropical cyclones. The major characteristics of clouds in tropical cyclones synthesized here are outlined in the following five summaries.

### a. Clouds involved in tropical cyclogenesis

The formation of a tropical cyclone is often, if not always, preceded by a "convective burst" of very deep and persistent cumulonimbus (Steranka et al. 1986). Studies both recent (Houze et al. 2009) and past (Zipser and Gautier 1978) indicate that this convection has unusual properties, which are especially conducive to tropical cyclogenesis. The updrafts of the convection can be extremely deep, wide, and intense. In addition, the updrafts are rotational, as a result of stretching low-level vorticity and advecting the resulting concentrated positive vorticity

upward to produce a convective-scale positive vorticity perturbation that extends from low to midlevels. These “vortex hot towers” operate without strong downdrafts and spreading density currents. The subcloud boundary layer in the developing depression is therefore not diluted by downward transport of lower- $\theta_e$  air. All of these properties are positive contributors to the intensification of a depression, especially in the lower troposphere.

The convective clouds in an intensifying depression are an ensemble of individual vortex hot towers and MCSs containing vortex hot towers and a stratiform precipitation region containing an MCV. As the convective-scale hot tower vortices and the stratiform region MCVs are axisymmetrized, they help to convert the intensifying depression into a tropical storm with an eye, an RMW, an eyewall, and rainbands.

#### *b. Clouds in the eye of the storm*

A fully formed tropical cyclone is marked by an “eye” where subsidence generally suppresses clouds through most of the troposphere. However, over the ocean the combination of high SST and strong mixing below a subsidence-produced stable layer capping the boundary layer creates conditions ideal for the formation of a cloud-topped mixed layer (Lilly 1968; chapter 5 of Houze 1993). As a result, stratus and/or stratocumulus clouds form in the eye of the storm (Fig. 5c). These clouds are thus essentially like other cloud-topped mixed layer clouds.

#### *c. The eyewall cloud*

The eyewall cloud slopes outward (Figs. 5c, 9, 11, 12, and 25), as a result of the powerful tendency of the storm’s annular zone of maximum wind to adjust toward gradient-wind balance via a slantwise moist conditionally neutral circulation (Emanuel 1986). Even when the buoyant instability temporarily appears at a location within the eyewall, inertial forces accelerate the buoyant element radially outward to quickly bring it in line with the slantwise secondary circulation of the balanced vortex (Rotunno and Emanuel 1987). A surprisingly large fraction of the vertical mass flux in the eyewall takes place in regions containing the superimposed buoyant convective elements (Braun 2002). Pockets of enhanced buoyancy in the eyewall region occur in the eyewall for several reasons. As the storm strengthens, the increased low-level radial inflow’s advection of air with high  $\theta_e$  produced by vigorous surface fluxes and mixing in the boundary layer toward the eyewall produces an unstable stratification in the lower troposphere (Rotunno and Emanuel 1987). The maximum wind in the vortex, located in the eyewall region, becomes supergradient, which requires the

eyewall to ingest low-level high- $\theta_e$  air into the base of the eyewall cloud (Smith et al. 2008). High-resolution two-dimensional models indicate that overturning eddies in the radial-vertical plane in the boundary layer occur on the inner edge of the RMW and draw low-level boundary layer air from the eye into the eyewall. This result may be partly an artifact of the two-dimensional model. Three-dimensional modeling shows that the ring of positive vorticity on the inside edge of the RMW is barotropically unstable and breaks down to form small-scale vortices (EVM) in the horizontal flow that entrain high- $\theta_e$  air into the base of the eyewall cloud and from the region of the eye of the storm (Braun et al. 2006). These vortices are collocated with an updraft that results from the entrainment of warm, moist air from the eye’s boundary layer, and this updraft tends to be relatively vertical, because it is rotating and thus inertially resistant to the shear of the wind in the eyewall region. Both the vorticity and updraft velocity in the EVM can be very strong (Marks et al. 2008).

Statistics of the radar reflectivity in the eyewall are different from those of convective systems not associated with tropical cyclones. CFADs of TRMM precipitation radar data show that the echoes are both intense and highly uniform, exhibiting a sharply peaked frequency distribution at all altitudes. The generally uniform echo intensity distribution is undoubtedly controlled by the prevailing slantwise secondary circulation, which is a property of the vortex-scale motions, and by the continual process of axisymmetrization of the air motions in the eyewall around the tropical cyclone vortex. The sharply peaked distribution also displays a wide range of outliers as some echoes become intense and great altitude. These outliers represent the occasional strongly buoyant elements superimposed on the slantwise circulation of the balanced vortex.

The eyewall cloud exhibits asymmetries that result from the inertially stiff tropical cyclone vortex moving through a quiescent environment. Further asymmetry results if the environment exhibits vertical shear of the horizontal wind. These factors lead to stronger upward air motions on the forward side of the eyewall cloud. The precipitation particles resulting from this enhanced formation and growth on the forward side are advected horizontally and distributed around the storm by the tangential winds. Particles may circulate  $\sim 1.5$  times around the storm (Fig. 26) as a result of this effect.

#### *d. Rainbands and secondary eyewalls*

Beyond the eyewall region, the clouds of the tropical cyclone take on four different and distinct mesoscale configurations:

- 1) The principal rainband is a quasi-stationary feature that spirals inward from the outer reaches of the cyclone until it becomes roughly tangent to the eyewall (Willoughby et al. 1984b; Willoughby 1988). Studies by Barnes et al. (1983), Powell (1990a,b), Hense and Houze (2008), and Didlake and Houze (2009) have shown the basic internal workings of the principal rainband. New convection forms on the upwind end of the principal rainband. At its downwind end, where it becomes tangent to the eyewall, the band takes on a more stratiform structure composed of dissipated convective cells. In their active stages, the convective-scale cells embedded in the rainbands display properties that are markedly unlike the convection seen in ordinary mesoscale convective systems (Houze 2004). The cells in the principal rainband are likely limited in height by the layer of strong outflow from the eyewall. The updrafts in the convective-scale cells embedded in the principal rainband have overturning updrafts characterized by radial inflow at low levels feeding an outward-sloping updraft, which resembles the slantwise circulation in the eyewall cloud, except that in the rainband much of the vertical motion is localized to convective cells. The convective-scale updrafts strengthen a wind maximum that characteristically lies along the principal axis of the rainband. A major difference between the rainband convective cells and eyewall convection is that the rainband cell displays deep downdrafts that penetrate into the boundary layer. A convective-scale downdraft fed by radial inflow from low- to midlevels is located in the heavy rainshower of the cell and transports low- $\theta_e$  air into the boundary layer air flowing toward the eyewall, thus acting as a brake on the boundary layer radial flow's transport of otherwise high- $\theta_e$  air toward the center of the storm. In addition, the convective updraft tends to induce an accompanying deep strong convective-scale downdraft along the inner edge of the principal rainband as a combination of vertical pressure gradient and negative buoyancy forces. This downdraft acts to strengthen the tropical cyclone vortex circulation.
  - 2) Secondary rainbands are smaller propagating spiral rainbands in the inner core of the storm and have mesoscale kinematic characteristics consistent with vortex Rossby waves. Although vortex Rossby wave theory has largely been studied in the context of dry vortices (Montgomery and Kallenbach 1997), full-physics models (Chen and Yau 2001; Chen et al. 2003; Braun et al. 2006) and coastal radar data (Corbosiero et al. 2006) show rainbands displaying propagation characteristics similar to idealized vortex Rossby waves.
  - 3) Secondary eyewalls form within the inner-core region outside the main eyewall zone from a coalescence of rainbands within an annular zone (Willoughby et al. 1982). The process controlling this coalescence of convection and rainbands into a secondary eyewall has not been satisfactorily explained, although some combination of axisymmetrization (Montgomery and Kallenbach 1997), mean vortex structure (Terwey and Montgomery 2008), or large-scale environment factors such as the humidity field (Ortt and Chen 2006) is probably involved. The secondary eyewall is qualitatively similar to the primary (inner) eyewall but is shallower since it lies underneath the outflow of the primary eyewall (Fig. 42). The moat region between the primary and secondary eyewall develops strong downward motion and takes on the character of an eye such that the inner eyewall is surrounded by dry warm air (Houze et al. 2007). The inner eyewall persists probably because of being fed by boundary layer inflow from both the eye and the environment.
  - 4) Distant rainbands are bands of convective clouds occurring entirely in the outer reaches of the storm, away from the constraining dynamics of the primary vortex, or portions of the principal rainband extending into this far region of the storm. The distant rainbands contain cumulonimbus clouds that are essentially similar to convective clouds not associated with tropical cyclones. This convection tends to be deeper than the rainband and secondary eyewall convection in the inner-core region where the outflow from the primary eyewall restricts the height of convective elements.
- e. Cloud microphysics and electrification*
- Cloud microphysical data obtained by aircraft in tropical cyclones show these storms to be great producers of ice in the mid- to upper troposphere. The schematic cross section in Fig. 42 indicates how the storm is a massive fountain of upper-level cloud, which is almost entirely glaciated. The eyewalls and rainbands do produce large amounts of rain by coalescence of drops. The eyewall clouds also generate copious graupel. The riming process produces tiny ice splinters by the Hallett–Mossop mechanism (Hallett and Mossop 1974) (Fig. 27). Particles advected upward and radially outward from the eyewall or rainband cells form aggregates that fall out slowly and produce stratiform precipitation where they melt and fall out. The circulating wind carries these aggregates up to 1.5 times around the storm before they fall out (Fig. 26); thus, the stratiform precipitation is spread far from the convective cells generating the ice particles.

The unique dynamics of the eyewall of the tropical cyclone lead to the cloud electrification processes being different from those of ordinary thunderstorms. Aircraft measurements by Black and Hallett (1986, 1999) show that the negatively charged particles evidently produced by small ice particles colliding with graupel in the eyewall cloud are advected in a slantwise manner upward and radially outward to produce an upward electric field gradient that is discharged as lighting (Fig. 27). This electrification pattern in the cloud is much different than the typical tripole pattern produced in ordinary thunderstorms (Williams 1988). Though it has not been studied, the lightning seen on the outer fringes of tropical cyclones, in the distant rainbands, is likely similar to that of ordinary thunderstorms.

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## APPENDIX A

### Equations of a Balanced Vortex

#### a. Gradient-wind balance

When horizontal flow becomes highly circular, as in a hurricane, it is convenient to consider the air motions in a cylindrical coordinate system, with the center of the circulation as the origin. If the circulation is assumed to be axially symmetric and inviscid, then the horizontal components of the equation of motion are

$$\frac{Du}{Dt} = -\frac{1}{\rho} \frac{\partial p}{\partial r} + fv + \frac{v^2}{r} \quad \text{and} \quad (A1)$$

$$\frac{Dv}{Dt} = -fu - \frac{uv}{r}, \quad (A2)$$

where  $u$  and  $v$  represent the radial and tangential horizontal velocity components  $Dr/Dt$  and  $rD\Theta/Dt$ , respectively, with  $r$  being the radial coordinate and  $\Theta$  being the azimuth angle of the coordinate system. Gradient-wind balance is said to occur when the three terms on the right of (A1) are in balance.

It is convenient to write the equations of gradient-wind balance in terms of the angular momentum  $m$  about the axis of the cylindrical coordinate system. For gradient-wind balance, (A1) and (A2), with substitution from (1), become

$$0 = -\frac{1}{\rho} \frac{\partial p}{\partial r} + \frac{m^2}{r^3} - \frac{f^2 r}{4} \quad \text{and} \quad (A3)$$

$$\frac{Dm}{Dt} = 0. \quad (A4)$$

The last relation expresses the conservation of angular momentum around the center of the circulation. Note that it applies regardless of whether or not the circulation in the absence of frictional or other torques is balanced, as long as it is cylindrically symmetric.

#### b. Hydrostatic balance

Vertical force balance is expressed by the hydrostatic equation, which can be written as

$$\frac{\partial \Phi}{\partial p} = -\alpha, \quad (A5)$$

where  $\alpha$  is specific volume and  $\Phi$  is the geopotential, defined as  $gz + \text{constant}$ , with  $g$  being the gravitational acceleration and  $z$  being height.

#### c. Thermal wind balance

When a cylindrically symmetric circulation is in both hydrostatic and gradient-wind balance, the winds (expressed by  $m$ ) at different altitudes are related by the following thermal-wind equation:

$$\frac{2m}{r^3} \frac{\partial m}{\partial p} = -\frac{\partial \alpha}{\partial r}, \quad (A6)$$

where  $\alpha$  is the specific volume. This relation is obtained by taking  $\partial(A5)/\partial r$ , substituting from the gradient-wind equation in (A3), and noting that in hydrostatic balance  $\Delta\Phi = -\alpha\Delta p$ .

## APPENDIX B

### Equivalent Potential Temperature and Saturation Equivalent Potential Temperature

For the purposes of this paper, the *equivalent potential temperature* is defined as

$$\theta_e \equiv \theta(T, p) \exp[Lq_v/c_p T_s(T, q_v, p)], \quad (B1)$$

where  $\theta$  is potential temperature,  $T$  is temperature,  $p$  is pressure,  $L$  is the latent heat of vaporization,  $q_v$  is the water vapor mixing ratio,  $c_p$  is the specific heat at constant pressure, and  $T_s(T, q_v, p)$  is the temperature at which the air would become saturated by lowering its pressure dry adiabatically. According to this definition,  $\theta_e$  is conserved in dry-adiabatic motion and nearly conserved under saturated conditions (see Houze 1993, 27–29, for further



discussion). The *saturation equivalent potential temperature* is defined as

$$\theta_{es} \equiv \theta(T, p) \exp[Lq_{vs}(T, p)/c_p T], \quad (\text{B2})$$

where  $q_{vs}$  is the saturation vapor mixing ratio. Note that the environment is potentially unstable (i.e., unstable if brought to saturation, e.g., by lifting a layer of air) if  $\partial\theta_e/\partial z < 0$  and conditionally unstable if  $\partial\theta_{es}/\partial z < 0$ . The environment is potentially or conditionally symmetrically unstable if these conditions apply along a surface of constant angular momentum (Houze 1993, 53–56).

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