Tropical Cyclone Formation

Introduction

Previously, we discussed large-scale conditions believed to be necessary for tropical cyclone development to occur. Now, we focus on describing two physical processes: tropical cyclone vortex and warm-core thermal structure development. We define tropical cyclone formation as the *initial* development of a tropical cyclone, and in subsequent lectures we will discuss processes that influence the intensity of an existing tropical cyclone, including both external and internal influences upon intensity.

Key Concepts

- How do tropical cyclones form?
- How does the tropical cyclone vortex develop?
- How is a warm-core thermal structure acquired by an initially cold-core disturbance during tropical cyclone development?

Observational Perspective on Tropical Cyclone Formation

Presuming that a pre-existing synoptic-scale disturbance of some form exists in an environment conducive to tropical cyclone development, the tropical cyclone formation, or tropical cyclogenesis, process can broadly be described as a two stage process (Zehr 1992). In the first stage, deep, moist convection is triggered by persistent lower-tropospheric convergence associated with the pre-existing disturbance in a modestly unstable environment. Over a period of several hours, such convection grows upscale, resulting in mesoscale convective system (MCS) formation.

Stratiform precipitation (and thus heating) with the MCS results in middle-tropospheric diabatic warming and lower-tropospheric diabatic cooling. This fosters greater vertical packing of isentropes in the middle troposphere. Relating this to potential vorticity, this is associated with high potential vorticity in the middle troposphere, giving rise to the development of a middle-tropospheric mesoscale convective vortex (MCV) in the vicinity of the pre-existing disturbance.

The generation of evaporatively driven downdrafts stabilizes the boundary layer and weakens the large-scale boundary layer convergence driving the deep, moist convection. Over time, these processes result in MCS dissipation, though not before significant middle-tropospheric moistening has occurred. The end result of the first stage of tropical cyclogenesis is thus a middle-tropospheric MCV largely devoid of thunderstorms. The middle troposphere is moister but the boundary layer is cooler and drier than prior to the onset of the first stage of development. In all, this first stage encompasses approximately 12-24 h.

Prior to the second stage, the boundary layer must be sufficiently warmed and/or moistened so as to permit renewed thunderstorm development. Over warm oceans, this is typically accomplished by surface latent heat flux from the underlying surface into the boundary layer. As this occurs, the environment in the vicinity of the MCV can again support deep, moist convection. Such convective activity is often triggered by forcing for ascent triggered by balanced lifting associated with the MCV and/or large-scale convergence associated with the MCV or remnant tropical disturbance. Redeveloping convection signals the onset of the second stage of tropical cyclogenesis, a series of processes that, over ~12-24 h, can lead to tropical cyclone development.

This round of deep, moist convective activity differs from that in the first stage of tropical cyclone development in two important ways. First, the presence of the MCV increases the inertial stability (related to the absolute vorticity), thereby reducing the Rossby radius of deformation, itself approximated by the ratio of the static stability to the inertial stability. Reducing the Rossby radius of deformation acts to laterally constrain the radial extent of the heating associated with the convection, the implications of which will be discussed shortly. Note that we are not yet ascribing such heating to a particular physical process; rather, we are merely stating that it occurs in environments of deep, moist convection.

Second, the middle troposphere is significantly moister as compared to the first stage of tropical cyclone development. This mitigates evaporatively driven downdraft development and intensity. If the middle troposphere is saturated, assuming minimal water loading in regions of active precipitation, then downdraft activity is largely suppressed. In such an environment, the atmospheric lapse rate is typically said to be "moist neutral," describing a saturated profile in which parcels are intrinsically neutrally buoyant to vertical parcel displacements. Such conditions are favorable for the efficient development of a cyclonic circulation within the boundary layer, as is associated with a tropical cyclone. Note, however, that we are not yet describing precisely how this cyclonic circulation forms.

In an environment favorable for tropical cyclogenesis, this observational perspective elucidates the basic physical characteristics and applicable time scale(s) of tropical cyclogenesis quite nicely. However, as noted above, it does not address the two fundamental questions associated with precisely how tropical cyclogenesis occurs. Namely, it does not provide insight into how the tropical cyclone vortex develops in the boundary layer, nor does it describe how the tropical cyclone warm-core develops. To address these questions, we must turn to other resources.

Tropical Cyclone Vortex Development

The development of the tropical cyclone vortex is fundamentally dependent upon the role of the middle tropospheric MCV in the lower tropospheric vortex development process. There are two theories that attempt to address this issue. The first emphasizes the downward development or penetration of a middle tropospheric MCV into the boundary layer (Bister and Emanuel 1997; Ritchie and Holland 1997) whereas the second emphasizes the development and eventual organization of deep, moist convective towers within the embryonic environment provided by the middle tropospheric MCV (Montgomery et al. 2006). The former is sometimes referred to as the "top-down" paradigm, whereas the latter is sometimes referred to as the "bottom-up" or marsupial paradigm. Support for both paradigms can be obtained from both observations and numerical simulations. We now turn to describing the salient physical and dynamical processes associated with each of these two theories.

Downward MCV Penetration

Bister and Emanuel (1997) suggest that tropical cyclone development occurs in response to the downward development of a middle-tropospheric MCV into the boundary layer. A middle-tropospheric MCV first forms in the stratiform rain region of an MCS associated with a pre-existing tropical disturbance. Evaporation of rain in the MCV's environment increases lower-tropospheric relative humidity and leads to a downdraft that advects the vortex downward to the boundary layer. Subsequently, deep, moist convection redevelops, leading to increased lower-tropospheric cyclonic vertical vorticity via tilting and stretching.

Bister and Emanuel (1997) argue that the initial cold-core structure of the MCV is crucial to subsequent tropical cyclone development as such a structure reduces the value of boundary-layer equivalent potential temperature necessary for thunderstorm initiation. Furthermore, increased lower- to middle-tropospheric relative humidity associated with evaporation-driven moistening by an initial thunderstorm cycle mitigates evaporatively driven downdraft formation with subsequent thunderstorm cycles. As the MCV's cold-core structure implies a relatively cool boundary layer, such a structure in and of itself further mitigates against the deleterious impacts of downdrafts upon the near-surface profiles of convergence and vertical vorticity. The downward penetration of the MCV's circulation – and thus increased rotation rate and wind speed at and near the surface – enhances the surface sensible and latent heat fluxes from the underlying ocean, allowing boundary-layer equivalent potential temperature to increase via warming and moistening to the level necessary for deep, moist convection to redevelop.

The downward development or advection of the MCV's cyclonic circulation takes as long as it takes air to descend through the layer of evaporational cooling. This downward extension mechanism thus requires precipitation to last sufficiently long, whether in one or more episodes, so as to moisten/humidify the entire lower to middle troposphere (i.e., the entire layer of evaporational cooling). The relevant time scale for such activity can be as short as a few hours or as long as a couple of days. The required time is longer when there is continual environmental, sub-saturated air infiltration into the MCV environment, as is often found in vertically sheared environments, because it counteracts the evaporation-driven moistening necessary for tropical cyclogenesis.

Convective Tower-Driven Development

Utilizing output from high-resolution idealized numerical simulations of tropical cyclogenesis, Montgomery et al. (2006) suggest that tropical cyclone development occurs in response to cycles of intense convection, termed vortical hot towers due to their association with locally large cyclonic vertical vorticity, and the gradual upscale growth of cyclonic vertical vorticity from the cloud-scale to the vortex-scale. Such development occurs in the unstable, cyclonic-vorticity–rich environment of the middle-tropospheric MCV embryo, colloquially referred to as the "pouch" (Dunkerton et al. 2009). While the MCV plays an important role in focusing deep, moist convection within a laterally confined region, its downward development is de-emphasized in this paradigm in favor of the upscale (horizontal) growth of tropospheric-deep towers of cyclonic vertical vorticity.

Specifically, within the pouch environment, cloud-scale cumulonimbus towers possessing intense cyclonic vertical vorticity in their cores emerge as the preferred coherent convective structures. Initial cyclonic vertical vorticity in the lower troposphere is generated through tilting of the horizontal vorticity associated with the vertical wind shear of the horizontal wind, the latter of which is associated with the middle-tropospheric MCV's cold-core structure (e.g., increasing intensity with height). Thereafter, cyclonic vertical vorticity is amplified through the stretching of both MCV-scale and convective tower-scale vertical vorticity by convective-scale updrafts. Convective towers exhibit lifetimes on the order of 1 h. Repeated cycles of convective tower activity act to overcome the adverse effects of evaporatively driven downdraft activity by reducing the amount of available buoyancy, humidifying the middle and upper troposphere, and merging with neighboring convective towers. Such merger processes act to increase the horizontal scale of the vortices associated with the convective towers.

While each individual convective tower is short-lived, the aggregate of towers about the pouch mimic a quasi-steady heating rate on the scale of the middle-tropospheric MCV. For thermal wind balance to be maintained, a thermally direct circulation with lower-tropospheric convergence, tropospheric-deep ascent, and upper-tropospheric divergence must develop in response to this heating. Over a period of 6 h or greater, the convergence associated with this circulation converges the convectively generated near-surface and MCV-associated middle- to upper-troposphere cyclonic vertical vorticity, building the tropospheric-deep tropical cyclone vortex. Such development typically occurs near what is colloquially known as the "sweet spot," or intersection of the latitude at which the wave's horizontal velocity vanishes and trough axis as viewed in a reference frame moving with the developing disturbance (Dunkerton et al. 2009).

Summary

The top-down and bottom-up paradigms are not necessarily at odds with one another. Both emphasize the presence of a middle-tropospheric MCV and the role of deep, moist convection in building the lower-tropospheric tropical cyclone vortex. They differ in the spatiotemporal scales emphasized in each paradigm. The top-down paradigm emphasizes vortex-scale processes, with convective-scale processes as the final genesis stage; the bottom-up paradigm emphasizes convective-scale processes, with vortex-scale processes serving as facilitators for genesis. Multiple studies (Raymond et al. 2011; Davis and Ahijevych 2012) provide evidence to suggest both are important, with vortex-scale processes setting the environment in which convective-scale processes occur. Specifically, middle-tropospheric moistening accompanying both MCV and thunderstorm development suppresses evaporatively driven downdraft formation, aiding convective vigor and longevity. Stratiform precipitation stabilizes the profile, cooling the lower troposphere and warming the middle troposphere, enabling upward motion to be concentrated more in the lower troposphere and thus amplifying cyclonic vorticity generation though localized vortex stretching. MCV intensification increases inertial stability, restraining convection to a mesoscale area near the disturbance's center and mitigating environmental dry air infiltration. The vertical alignment of the lower- and middle-tropospheric vortices is an indicator of weak vertical wind shear and is generally thought to be necessary for the above processes to occur. Depending on the spatial and temporal scales that one uses in a vorticity budget analysis of tropical cyclogenesis, support for both paradigms can be obtained. Therefore, it is likely that both paradigms are at least partially correct in their depictions of tropical cyclogenesis. This emphasizes the need for continued tropical cyclogenesis research.

Independent of how the near-surface tropical cyclone vortex is built, neither theory is meant to address the underlying energetics of tropical cyclone development. In other words, these theories are not meant to be used to elucidate the source of the diabatic heating that drives tropical cyclones. This problem is addressed in the following section.

Tropical Cyclone Warm-Core Development

Tropical cyclones are driven primarily by diabatic heating, particularly heating found in the middle to upper troposphere and constrained to near the center of the cyclone by sufficiently strong rotational forces. In as much as a tropical cyclone satisfies hydrostatic balance, heating maximized in the middle-to-upper troposphere will result in lower surface pressure below the level of peak heating. When this heating is constrained to near the cyclone's center, it can result in large surface pressure falls. However, what is the source for such heating?

The currently accepted theory for the source of this heating – and, by extension, tropical cyclone warm-core development – is given by the non-linear wind-induced surface heat exchange (WISHE) theory of Emanuel (1986) and subsequent works (e.g., Zhang and Emanuel 2016). In the presence of a pre-existing tropical disturbance over a sufficiently warm ocean surface, WISHE states that latent heat release in the free troposphere is governed by the evaporation of moisture (i.e., the surface latent heat, or more generally enthalpy, flux) from the underlying ocean surface as determined primarily by the magnitude of the surface winds. In other words, tropical cyclones derive the energy that fuels their winds from the underlying ocean, with updrafts acting to loft this latent heating upward, whereupon it is released when the associated water vapor condenses and/or freezes.

Several assumptions are implicit to WISHE. First, we assume that WISHE does not significantly contribute to initial disturbance development prior to tropical cyclogenesis (Zhang and Emanuel 2016 and references therein). This is not to say that it does not occur at those stages – just that it is not the dominant process leading to initial disturbance development. Concurrently, we assume that the disturbance's inner-core is nearly saturated, mitigating against evaporatively generated downdrafts transporting low equivalent potential temperature into the boundary layer. As previously discussed, this occurs via moistening processes associated with cyclic convective activity in the early stages of the tropical cyclogenesis process. Similarly, the atmosphere is assumed to be neutrally stratified along angular momentum surfaces. Within the inner core of a developing or mature tropical cyclone, such surfaces are generally vertically oriented with a slight outward tilt. Therefore, this assumption is equivalent to saying that there is no slantwise instability present. There may be a small amount of upright, or traditional, instability present within the environment (i.e., slightly departing from moist neutrality); however, WISHE explicitly states that this instability does not contribute to tropical cyclone development. Third, WISHE assumes that deep, moist convection is ongoing and acts to mix air vertically. Fourth, WISHE assumes that thunderstorms do not impact temperature in the absence of surface fluxes. Therefore, surface fluxes are critical to the development of the tropical cyclone warm core. Without them, not only would the tropical cyclone warm core not be able to develop, but the boundary layer would never be able to recover sufficiently so as to permit thunderstorm development in the pre-genesis period!

It should be noted that surface fluxes are dependent upon the wind speed. This can be demonstrated mathematically, where bulk formulations for the surface sensible and latent heat flux are given by:

$$Q_h = -\rho c_p c_h \left\|\mathbf{v}_{10m}\right\| \left(T_{2m} - T_{sfc}\right)$$

$$Q_e = -\rho l_v c_e \left\|\mathbf{v}_{10m}\right\| \left(q_{v,2m} - q_{vs,sfc}\right)$$

Here, ρ is density, c_p is the specific heat at constant pressure, l_v is the latent heat of vaporization, c_h and c_e are exchange coefficients for heat and moisture, respectively, and subscripts of 10m, 2m, and sfc represent 10-m, 2-m, and land-surface values of their respective quantities. The subscript of s on $q_{v,sfc}$ indicates the saturation water vapor mixing ratio corresponding to the sea surface temperature. The exchange coefficients themselves are functions of 10-m wind speed, at least to an extent (there is disagreement on exchange coefficient values at hurricane-force wind speeds due to a lack of observations), and near-surface stability. To first-order, surface flux magnitudes increase as wind speed increases.

As a result of the dependence of surface flux magnitudes on wind speed, WISHE can be viewed as a non-linear feedback loop. The weak winds associated with a pre-existing tropical disturbance act to induce weak surface latent (and sensible) heat fluxes. Such fluxes slowly moisten (and warm) the boundary layer, enabling it to recover from earlier cooling associated with evaporatively driven downdrafts. Deep, moist convection transports the fluxed latent heat upward to the upper troposphere, wherein it is released as water vapor condenses and eventually freezes. The radial extent of this heating is controlled by the Rossby radius of deformation, a metric related in large part to the radial extent of the pre-existing disturbance.

In a hydrostatic atmosphere, heating released aloft acts to lower the pressure at levels beneath and raise the pressure at levels above that of the peak heating. We'll use this assumption here, although Stern and Nolan (2012) suggest that it only captures a part of a tropical cyclone's response to its vertical heating distribution. The localized nature of this heating implies that the pressure falls are localized as well, such that surface wind speed becomes larger due to the larger horizontal (radial) pressure gradient magnitude. In turn, the larger surface wind speed acts to enhance the magnitude of the surface heat fluxes. Enhanced heat energy is carried aloft by deep, moist convection, resulting in a further increase of the temperature aloft. This results in an enhanced response in the mass and, subsequently, wind fields above and below the level of peak heating. The feedback loop continues from here through the entirety of the tropical cyclone lifecycle, although this is the subject of some disagreement in the field. The net result of these processes is two-fold. First, a substantial warm anomaly constrained to the inner core of the tropical cyclone develops in the middle-to-upper troposphere. Second, a strong near-surface cyclone and modest anticyclone at the level of the tropopause develop. These are both characteristics of warm-core cyclones.

The WISHE feedback loop describes the Carnot cycle approximation to a tropical cyclone's secondary circulation, comprised by the four branches of the secondary circulation itself. First, there is isothermal inflow toward the center of the cyclone in the lower troposphere. Typically, the temperature (and dew point temperature) of inflowing parcels is approximately 1°C less than the local sea surface temperature, a necessary requirement for the surface sensible and latent heat fluxes to be upward. Second, moist-adiabatic ascent occurs near the center of the cyclone. Moist-adiabatic ascent implies equivalent potential temperature conservation following the motion. Indeed, equivalent potential temperature is nearly constant in the vertical within the tropical cyclone inner core, although departures from this state are thought to be important to vortex intensification. Moist-adiabatic ascent transports latent heat energy to the middle and upper troposphere. Compared to larger radii, where ascent is not entirely moist adiabatic, the net result is a warm anomaly within the cyclone's inner core as the latent heat is released upon condensation and freezing. Third, there is moist-adiabatic outflow away from the center of the cyclone near the tropopause. Finally, there is descent at large radii from the center of the cyclone. As parcels become subsaturated upon descent, whether this descent is forced or unforced, such descent becomes predominantly dry adiabatic. This implies that parcels approximately conserve potential temperature as they descend at large radii.

The Importance of WISHE at Higher Wind Speeds

Traditionally, WISHE has been viewed as an exponential instability: as intensity increases, surface fluxes increase, subsequently leading to non-linear intensification. Recently, however, some studies (e.g., Montgomery et al. 2009 and subsequent works) have suggested that exponential instability is not necessary for tropical cyclone intensification. Supporting evidence from these studies comes in the form of numerical model simulations in which realistic peak tropical cyclone intensities are achieved despite capping surface latent heat fluxes at values associated with trade-wind wind speeds (5-10 m s^{-1}). These studies argue that while surface fluxes remain important prior to tropical cyclogenesis, to enable boundary-layer equivalent potential temperature to increase sufficiently to allow for deep, moist convection to redevelop, an increased flux magnitude as intensity increases is not necessary for subsequent intensification. Rather, Montgomery et al. (2009) and subsequent studies argue that local buoyancy within individual convective towers serves as the tropical cyclone's energy source, with the upscale growth of cyclonic vertical vorticity associated with vortex merger and filamentation leading to tropical cyclone intensification.

This is a minority view, however. Perhaps the best summary of WISHE's role in tropical cyclone warm-core development and intensification is given by Zhang and Emanuel (2016). While WISHE is not explicitly necessary for realistic intensities to be achieved, intensification occurs at a faster pace in idealized and real-data numerical simulations in favorable environments when surface heat fluxes are not artificially limited by wind speed. In environments that are not optimal for development, such as may be associated with large vertical wind shear, WISHE appears essential to tropical cyclone maintenance and development, as the associated surface fluxes act against environmental dry air infiltration. Yet, further research is needed to provide clarity into the physical mechanisms behind tropical cyclone development and intensification.

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