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: the development of the tropical cyclone vortex and the acquisition of a warm-core thermal structure by an initially cold-core incipient disturbance. Thus, we define tropical cyclone formation as the *initial* development of a tropical cyclone. In subsequent lectures, we will discuss processes that influence the intensity of an existing tropical cyclone, focusing upon both external and internal influences upon tropical cyclone intensity.

Key Concepts

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

Observational Perspective on Tropical Cyclone Formation

Presuming that a pre-existing synoptic-scale disturbance of some form (e.g., African easterly wave, monsoon trough, the ITCZ, etc.) 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 the formation of a mesoscale convective system (MCS).

Middle tropospheric warming via latent heat release and lower tropospheric cooling via the generation of evaporatively-driven downdrafts accompany the MCS. This stratiform heating profile 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 acts to stabilize the boundary layer and weaken the large-scale boundary layer convergence driving the deep, moist convection. Over time, these processes result in the erosion of the MCS, though not before significant moistening of the middle troposphere has occurred. The end result of the first stage of tropical cyclogenesis is thus a middle tropospheric MCV devoid of most if not all deep, moist convective activity. 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 a time period of approximately 12-24 h.

Prior to the onset of the second stage of tropical cyclone development, the boundary layer must be sufficiently warmed and/or moistened so as to permit the renewed development of deep, moist convection. Over warm oceans, this is typically accomplished by the flux of latent heat energy 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. The redevelopment of convection signals the onset of the second stage of tropical cyclone development, a series of processes that, over the span of approximately 12-24 h, can lead to the development of a tropical cyclone.

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 the intensity and development of evaporatively-driven downdrafts. If the middle troposphere is entirely saturated, assuming minimal water loading in regions of active precipitation, then downdraft activity is entirely suppressed. In such an environment, the atmospheric lapse rate is typically said to be "moist neutral," describing a situation in which parcels are neutrally buoyant to vertical parcel displacements. Such conditions have been found in observational and numerical modeling studies to be favorable for the efficient development of a cyclonic circulation within the boundary layer, such as is associated with a tropical cyclone. Note, however, that we are not yet describing precisely how this cyclonic circulation forms.

In a quiescent environment, or one that is altogether favorable for tropical cyclogenesis, this observational perspective elucidates the basic physical characteristics and applicable time scale(s) of the tropical cyclogenesis process 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 is constructed within the boundary layer, nor does it provide insight into whether the tropical cyclone warm core is a response of convective heating or some other thermodynamic process. To address these questions, we must turn to other resources.

The Development of the Tropical Cyclone Vortex

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 has been studied extensively from observational and numerical modeling perspectives whereas the latter has largely been studied extensively using numerical model simulations. We now turn to describing the salient physical and dynamical processes associated with each of these two theories.

Downward MCV Penetration

Using a combination of observational analysis, numerical model simulations, and theory, 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. This process can be briefly summarized as follows. A middle tropospheric MCV forms in the stratiform rain region of an MCS associated with a pre-existing tropical disturbance. Evaporation of rain in the environment of the MCV increases lower tropospheric relative humidity and leads to a downdraft that advects the vortex downward into the boundary layer. Subsequently, deep, moist convection redevelops, leading to the increase of cyclonic vertical vorticity in the lower troposphere via vorticity tilting and stretching processes.

Bister and Emanuel (1997) argue that the initial cold-core structure of the MCV is crucial to the subsequent development of a tropical cyclone as such a structure reduces the value of boundary layer equivalent potential temperature necessary for deep, moist convection to occur. Furthermore, increased lower to middle tropospheric relative humidity associated with evaporative moistening associated with initial convective activity mitigates evaporatively-driven downdrafts associated with subsequent convective activity. As the cold-core structure of the MCV implies relatively cool boundary layer conditions, such a structure in and of itself further mitigates against the deleterious impacts of downdraft activity upon the near-surface profiles of convergence and vertical vorticity. The downward penetration of the MCV's circulation enhances fluxes of latent heat energy from the underlying ocean surface, allowing the boundary layer value of equivalent potential temperature to rise (via warming and moistening) to the reduced level necessary for deep, moist convection to redevelop.

The downward development or advection of the cyclonic circulation of the MCV 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. It should also be noted that the downward development process, and thus tropical cyclogenesis as a whole, is significantly hindered by relative flow through the MCV. This is most often attributable to vertical wind shear, which is often associated with the import of dry air into the vortex's circulation. As a result, the relative flow through the MCV and associated precipitation system must be small so as to promote a humid, upright vortex core favorable for subsequent tropical cyclone development.

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 deep, moist convective activity, termed vortical hot towers, and the gradual upscale growth of cyclonic vertical vorticity from the cloud-scale to the vortex-scale. Such development is said to occur in the unstable, cyclonic vorticity-rich environment of the middle tropospheric MCV embryo, colloquially referred to as the "pouch" (e.g., 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 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. Such cyclonic vertical vorticity is generated through two processes. Initial cyclonic vertical vorticity, particularly in the lower troposphere, is generated through the tilting of horizontal vorticity associated with the vertical wind shear of the horizontal wind. This is most generally associated with the cold-core structure of the middle tropospheric MCV. Thereafter, cyclonic vertical vorticity is amplified through the stretching of both MCV-scale and convective tower-scale vertical vorticity. These 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 and middle 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 convective tower-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 wave's critical latitude (i.e., the latitude at which the horizontal velocity vanishes) and trough axis as viewed within a reference frame moving with the developing disturbance (Dunkerton et al. 2009).

Summary

As noted by Dunkerton et al. (2009), the downward MCV penetration and convective-tower driven development paradigms are not necessarily at odds with one another. Both paradigms emphasize the presence of a middle tropospheric MCV and the role of deep, moist convective activity toward building the lower tropospheric tropical cyclone vortex. They differ in the spatiotemporal scales emphasized within each paradigm. The downward MCV penetration paradigm emphasizes processes occurring on the meso-α and larger scales of the middle tropospheric MCV. Conversely, the convective tower paradigm emphasizes processes on the cloud scale and, subsequently, the upscale growth of the vortices resulting from such processes to larger scales. Depending on the spatial and temporal scales that one uses in a budgetary analysis of tropical cyclogenesis, support for both paradigms can be obtained. Therefore, it is possible that both paradigms may be at least partially correct – or, conversely, may be entirely incorrect! – in their depictions of tropical cyclogenesis. This emphasizes the need for further research into the tropical cyclogenesis process.

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 heating that drives tropical cyclones. This problem is addressed in the following section.

The Development of the Tropical Cyclone Warm Core

Tropical cyclones are driven by heating, particularly heating found in the middle to upper troposphere and constrained to near the center of the cyclone by sufficiently strong rotational forces. From thickness arguments, heating maximized in the middle to upper troposphere will result in lower pressure beneath the level of peak heating. When constrained to near the center of the cyclone, the strong, localized heating can result in a strong area of low pressure at the surface. Precisely how this is manifest will be discussed in the context of the Sawyer-Eliassen non-linear secondary circulation model in future lectures on tropical cyclone structure and intensity change. Before delving into such material, however, we wish to describe the source for such heating, or in other words, is such heating associated with latent heat release within deep, moist convection or is it merely transported to the middle and upper troposphere by convective updrafts?

The currently-accepted theory for the source of this heating – and, by extension, the development of the tropical cyclone warm core – is given by the non-linear wind-induced surface heat exchange (WISHE) theory of Emanuel (1986) and subsequent works. 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 from the underlying ocean surface as determined primarily by the magnitude of the surface winds. In other words, latent heat energy used to fuel the tropical cyclone and build the tropical cyclone warm core is obtained from the underlying surface and not through convective heating processes. The tropical cyclone warm core is constructed aloft as updrafts within deep, moist convection carry this latent heat energy from the boundary layer to the middle and upper troposphere.

Several assumptions are implicit to WISHE theory. First, the inner core environment of the developing tropical cyclone vortex must be nearly saturated (i.e., nearly moist neutral) so that evaporatively-generated downdrafts do not import 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. Third, WISHE assumes that deep, moist convection is ongoing and acts to mix air vertically. Fourth, and perhaps key, WISHE assumes that deep, moist convection does 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 deep, moist convective development in the pre-genesis period!

It should be noted that surface fluxes, whether of sensible or latent heat energy, are dependent upon the wind speed. To first order, at sub-hurricane intensities, surface flux magnitudes increase as wind speed increases. (We will consider the impact of surface fluxes on continued tropical cyclone development in a subsequent lecture.) As a result of this dependence, 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 latent heat energy upward to the upper troposphere, where it is accumulated beneath the tropopause. The radial extent of this heat accumulation is controlled by the radial extent of the pre-existing disturbance and accompanying middle tropospheric MCV (i.e., by the Rossby radius of deformation).

Heating released aloft acts to lower the pressure at levels beneath and raise the pressure at levels above that of the peak heating. In time, the wind fields respond to this forcing upon the mass (pressure) field, resulting in an increase in the magnitude of the cyclonic surface winds of the vortex. This acceleration in the 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 at least the tropical cyclogenesis process and, perhaps, through the entirety of the tropical cyclone lifecycle. 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 feedback loop associated with the WISHE paradigm describes the Carnot cycle approximation to the secondary circulation of a tropical cyclone. This representation of the tropical cyclone secondary circulation is 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 dewpoint) of inflowing parcels is approximately 1°C less than the local sea surface temperature, a necessary requirement for the flux of sensible (and latent) heat energy from the underlying surface. Second, moist adiabatic ascent occurs near the center of the cyclone. Moist adiabatic ascent implies the conservation of equivalent potential temperature. Indeed, equivalent potential temperature is nearly constant in the vertical within the inner core of tropical cyclones. Such 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 inner core of the cyclone. 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 forced or unforced in nature, such descent is predominantly dry adiabatic in nature. This implies that parcels approximately conserve potential temperature as they descend to the boundary layer.

Also tied to WISHE and the Carnot cycle approximation is the concept of a maximum potential intensity, or MPI, of tropical cyclones. We will consider MPI more closely in a subsequent lecture on tropical cyclone intensity change.

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