Tropical Cyclone Formation

Introduction

Whereas the last section presented the large-scale conditions that are believed to be necessary for a tropical cyclone to develop, here we describe the physical processes by which a tropical cyclone and its warm-core thermal structure develops. We start with an observational overview of tropical cyclone development, from which we segue into contrasting the “top-down” and “bottom-up” perspectives for tropical cyclone vortex formation. We close with an introduction to wind-induced surface heat exchange as the energy mechanism by which a tropical cyclone can intensify.

Key Questions

- 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 some pre-existing synoptic-scale disturbance exists in an environment conducive to tropical cyclone development, tropical cyclone formation (tropical cyclogenesis) can broadly be described as a two-stage process (Zehr 1992). In the first stage, thunderstorms are initiated by persistent lower-tropospheric convergence associated with the pre-existing disturbance in a modestly unstable environment. Over several hours, these thunderstorms grow upscale and result in mesoscale convective system (MCS) formation.

Stratiform precipitation associated with the MCS is characterized by midtropospheric diabatic warming due to condensation and lower-tropospheric diabatic cooling due to evaporation. This results in midtropospheric height falls and the associated development of cyclonic rotation, characterized by a mesoscale convective vortex (MCV).

Evaporatively driven downdrafts associated with the MCS cool and stabilize the boundary layer and weaken the synoptic-scale lower-tropospheric convergence responsible for thunderstorm initiation. These processes result in MCS dissipation over time, albeit not until significant midtropospheric moistening (as water vapor is lofted upward, condenses, and later evaporates) occurs, leaving behind a midtropospheric MCV largely devoid of thunderstorms. In all, this first stage lasts approximately 12-24 h.

Between the first and second stages, the atmospheric boundary layer must be sufficiently destabilized so as to permit renewed thunderstorm development. This is accomplished by enthalpy fluxes (representing both sensible and latent heating) from the underlying surface. Thunderstorm development proceeds in the newly destabilized environment, with these thunderstorms’ development representing the onset of the second stage of tropical cyclogenesis.

This second stage differs from 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 inversely related to inertial stability). Reducing the Rossby radius of deformation laterally constrains the radial extent of the diabatic warming associated with thunderstorms, the implications of which will be discussed shortly. (Note that we are not yet ascribing this diabatic warming to a particular physical process; rather, we are merely stating that it occurs in thunderstorm environments.) Second, the midtroposphere is significantly moister, mitigating the formation and intensity of evaporatively driven downdrafts that would otherwise stabilize the atmospheric boundary layer and disrupt the large-scale convergent circulation associated with the developing disturbance. A tropical cyclone may result from this second stage if the environment is sufficiently favorable for it to form; otherwise, this stage may repeat one or more times on a quasi-daily basis until a tropical cyclone forms – if it can at all.

Tropical-Cyclone Vortex Development

Tropical cyclone development is believed to follow from that of the midtropospheric MCV, but how? There are two theories that attempt to physically describe how the midtropospheric MCV can facilitate the lower-tropospheric tropical cyclone vortex's development. The first emphasizes the downward development or penetration of a midtropospheric MCV toward the surface (Bister and Emanuel 1997; Ritchie and Holland 1997) whereas the second emphasizes the development and eventual organization of deep, moist convective towers (or thunderstorms) within the embryonic environment of the midtropospheric MCV (Montgomery et al. 2006). The former is sometimes colloquially referred to as "top-down", whereas the latter is sometimes colloquially referred to as "bottom-up." Support for both paradigms can be obtained from observations and numerical simulations. We now turn to describing the salient physical and dynamical processes associated with each of these theories.

Downward MCV Penetration

Bister and Emanuel (1997) suggest that tropical-cyclone development occurs in response to the downward development of a midtropospheric MCV toward the surface. Evaporation and sublimation in the MCV's environment increase lower-tropospheric relative humidity and leads to a downdraft that advects the vortex downward to the boundary layer. Increased lower-tropospheric relative humidity mitigates against further evaporatively driven downdrafts, allowing for near-surface cyclonic rotation to persist. Advecting the MCV toward the surface increases the near-surface wind speed, in turn increasing the surface sensible and latent heat fluxes, thus rewarming and remoistening the near-surface air to allow for subsequent thunderstorms to develop.

The MCV's downward development or advection takes as long as it takes air to descend through the layer of evaporational cooling. Thus, precipitation must last as long as is needed to advect the circulation toward the surface and to sufficiently moisten the lower 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 infiltration of subsaturated environmental air into the MCV environment, as often accompanies large vertical wind shear, because it counteracts the evaporation-driven moistening necessary for tropical cyclogenesis.

Convective Tower-Driven Development

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 upscale growth of cyclonic vertical vorticity from thunderstorm or cloud scales to the vortex-scale. Such development occurs in the unstable, cyclonic-vorticity–rich environment of the midtropospheric 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 radially confined region, its downward development is de-emphasized here in favor of the upscale (horizontal) growth of tropospheric-deep towers of cyclonic vertical vorticity.

Specifically, cloud-scale cumulonimbus towers possessing intense cyclonic vertical vorticity in their cores emerge as the preferred coherent features in the MCV’s environment. The cyclonic vertical vorticity in the lower troposphere is initially generated through tilting of the horizontal vorticity associated with the vertical wind shear, itself associated with the midtropospheric MCV’s cold-core structure (with increasing intensity with height). Cyclonic vertical vorticity is subsequently amplified through thunderstorm-updraft–induced stretching. Individual thunderstorms persist for up to several hours, with repeated cycles of thunderstorms in the MCV’s environment gradually reducing buoyancy, humidifying the mid- and upper troposphere, and merging with neighboring convective towers (thereby increasing individual vortices’ horizontal scales).

While each individual thunderstorm is short-lived, the aggregate of thunderstorms in the MCV environment mimic a quasi-steady diabatic warming on the MCV’s scale. 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 warming. Over a period of ~6 h, the convergence associated with this circulation coagulates lower- to midtropospheric 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 an midtropospheric MCV, thunderstorm activity, and lower-tropospheric moistening by stratiform rainfall. However, they differ in the scales that each emphasize: the “top-down” paradigm emphasizes vortex-scale processes whereas the “bottom-up” paradigm emphasizes thunderstorm-scale processes. Studies suggest both are important, with vortex-scale processes setting the environment in which convective-scale processes occur (e.g., Raymond et al. 2011; Davis and Ahijevych 2012). 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. That all aside, neither the “top-down” or “bottom-up” paradigms address the energy source fueling a tropical cyclone’s winds. For that, we need another theory, which we seek to develop in the next section.

Tropical Cyclone Warm-Core Development and Intensification

Tropical cyclones are driven primarily by diabatic warming, particularly that which is maximized in the mid- to upper-troposphere and laterally constrained by strong rotational forces to near the cyclone’s center. Crudely approximating a tropical cyclone with hydrostatic balance, this mid- to upper-tropospheric diabatic warming facilitates lower heights below the level of peak warming. When warming is constrained radially or laterally to near the cyclone's center, it can result in locally large surface pressure falls. However, what is the source of this warming?

The currently accepted theory for the source of this diabatic warming is 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 tropical cyclones gain energy in the form of enthalpy (comprised of heating and moistening) from the underlying ocean surface. The rate at which this energy is gained is directly proportional to both the air-sea disequilibrium (or the difference in temperature and water-vapor mixing ratio between the atmosphere and the ocean, with the water-vapor mixing ratio disequilibrium also being inversely proportional to the surface pressure) and near-surface wind speed.

Several assumptions are implicit to WISHE:

- WISHE does not significantly contribute to the formation of the disturbance that later becomes a tropical cyclone (Zhang and Emanuel 2016 and references therein). That is not to say that WISHE is not ongoing at that stage – just that it is not dominant.
- The troposphere is saturated (or nearly so) near the center of the antecedent disturbance. This helps to mitigate against evaporatively generated downdrafts from transporting low equivalent potential temperature air to the boundary layer (where it would need to be warmed and moistened by surface fluxes that could otherwise support tropical-cyclone development). The requisite tropospheric-deep saturation occurs via moistening that accompanies thunderstorms in the early stages of the tropical cyclogenesis process.
- The troposphere is assumed to be neutrally stratified along angular-momentum surfaces. Near the center of a developing or mature tropical cyclone, angular-momentum surfaces are quasi-vertical with a slight outward tilt. An alternative way of stating this assumption is that there is no slantwise instability present. There may be a small amount of upright, or traditional, instability present within the environment (i.e., if the troposphere is not entirely saturated, such that the lapse rate is not moist adiabatic over its depth); however, WISHE explicitly states that this instability does not contribute to tropical cyclone development.
- Thunderstorms are ongoing and acts to mix or transport air vertically. *Thunderstorms do not impact temperature in the absence of surface fluxes!* They serve only to vertically mix the higher enthalpy or equivalent potential temperature air upward from near the surface, where enthalpy is gained from the ocean, through the troposphere. *Surface fluxes are critical to tropical-cyclone development!*

As noted above, enthalpy exchange is dependent upon air-sea disequilibrium and near-surface wind speed. This can be demonstrated mathematically with bulk formulations for the surface sensible and latent heat fluxes:

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

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

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 $q_{vs,sfc}$ indicates the saturation specific humidity corresponding to the sea surface temperature. These formulations show that surface heat fluxes increase both with increasing wind speed $\|\boldsymbol{v}_{10m}\|$ and air-sea disequilibrium (given by the quantities in parentheses).

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 induce an upward surface enthalpy flux that slowly warms and moistens the atmospheric boundary layer, enabling it to recover from earlier cooling associated with evaporatively driven downdrafts. This intensifies the tropical cyclone's warm thermal anomaly and thus the cyclone itself. This leads to increased upward surface enthalpy fluxes, which further intensify the tropical cyclone's warm thermal anomaly, and so on. The radial extent of the warm thermal anomaly is controlled by the Rossby radius of deformation, a metric related in large part to the radial extent of the pre-existing disturbance, and the inertial stability.

The feedback loop described above can be contextualized by the Carnot-cycle approximation to a tropical cyclone's secondary circulation, comprised of four branches (Emanuel 2019, their section 3c; see also the schematic in the associated lecture materials). We begin by assuming that the temperature and dew-point temperature of inflowing near-surface air parcels are each ~1°C less than the local SST.

- *Isothermal (constant temperature) lower-tropospheric inflow toward the tropical cyclone's center*. Due to friction, near-surface air spirals inward as it rotates around a tropical cyclone. As it does so, it approaches lower pressures (~1010 hPa well away from the center, 900-1000 hPa near the center). Assuming that SST is constant along inflowing air parcels, this reduction of surface pressure inward toward the disturbance's center increases the air-sea disequilibrium for both temperature and water-vapor mixing ratio. For temperature, reduced surface pressure causes air parcels to rise adiabatically and thus to cool by adiabatic expansion. For water-vapor mixing ratio, reduced surface pressure is associated with a larger saturation specific humidity. The increased air-sea disequilibrium, coupled with faster wind speeds near the cyclone's center, results in strong upward-directed enthalpy fluxes that offset the adiabatic cooling, further moisten the air, and maintain isothermal conditions along inflowing air parcels. Altogether, these processes determine the moist adiabat (i.e., direct it further to the right on a skew-T diagram) along which near-center air parcels will ultimately ascend.
- *Moist-adiabatic ascent near and outflow from the tropical cyclone's center*. Inflowing air parcels converge near the cyclone's center and are forced to ascend. Assuming that the troposphere is near saturation, this ascent is approximately moist adiabatic, such that equivalent potential temperature is conserved on ascent and outflow. (Note: this approximation neglects entrainment from either the environment or eye, detrainment, and hydrometeor effects.) The tropical cyclone's warm thermal anomaly aloft results by comparing the moist adiabat of this ascending air to that of the undisturbed environment.
- *Isothermal compression upon initial descent from the upper troposphere/lower stratosphere*. This branch is associated with warming due to adiabatic compression offset by evaporative cooling.
- *Moist-adiabatic descent from the upper- to the lower-troposphere at large radii*. Although the air warms due to adiabatic compression, as in the third branch, it also radiatively cools at an equivalent rate (which necessitates that the associated descent be weak; i.e., it takes a long time for air parcels to return to the atmospheric boundary layer from the upper troposphere). This results in descending air parcels maintaining a moist adiabatic lapse rate (warming, but not as rapidly under dry-adiabatic descent) that matches that of the surrounding environment.

This Carnot-cycle feedback loop forms the fundamental basis for the derivation of a theoretical maximum potential intensity, which is the opening subject of the subsequent lecture. Disruptions to this loop, such as may result from the import of cooler, drier environmental air by vertical wind shear, are described later in that lecture.

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