Tropical Cyclone Structure

Introduction

This section describes a tropical cyclone's structure, emphasizing both its primary (or tangential, describing cyclonic lower-tropospheric rotation) and secondary (or radial, describing lower-tropospheric inflow, deep-layer ascent, and upper-tropospheric outflow) circulations. We also discuss tropical cyclones' axisymmetric (the component that is symmetric in all directions at a given radius) and asymmetric structures.

Key Questions

- What are the salient characteristics of a tropical cyclone's primary circulation?
- What are the salient characteristics of a tropical cyclone's secondary circulation?
- What are the salient characteristics of a tropical cyclone's primary, secondary, and distant rainbands?
- How does a tropical cyclone's secondary circulation respond (to restore thermal-wind balance, and manifest in the cyclone's primary and secondary circulations) to prescribed/external heat and momentum forcing?

Overview

Tropical cyclones are characterized by cyclonic tangential and inflowing radial winds. The cyclonic winds can extend to over 1000 km away from the center in the lower troposphere; this radial extent decays with increasing height. Tropical cyclones are warm-core features, meaning that their intensity (measured by the cyclonic tangential wind speed) decreases with increasing height. A tropical cyclone is most intense just above the atmospheric boundary layer, where the effects of surface friction approach zero, and weakest in the upper troposphere, where the winds become anticyclonic and evacuate mass outward. Radial inflow is typically maximized in the atmospheric boundary layer, with weaker inflow into the midtroposphere. Radial inflow rapidly decelerates as it reaches the eyewall, where convergence leads to tropospheric-deep ascent. Compensatory descent for such strong ascent occurs in a concentrated manner in the eye and in a diffuse manner at radii much larger than the radius of maximum winds.

A tropical cyclone's warm-core structure can be approximated by the hydrostatic response to a localized mid- to upper-tropospheric warm potential-temperature anomaly near the tropical cyclone's center. Surface enthalpy fluxes are the primary contributor to the warm anomaly's development, with a small contribution from adiabatic compression associated with subsidence in the eye. The hydrostatic approximation implies that higher warm anomaly altitudes result in greater surface pressure falls for a given anomaly magnitude; however, the warm anomaly is not always maximized in the upper troposphere and the relationship between the warm anomaly and cyclone intensity is not always linear (Stern and Nolan 2012).

A mature tropical cyclone is characterized by a nearly cloud-free region near its center known as the eye. The minimum sea-level pressure is found at the center of the eye or, for weaker cyclones that lack eyes, collocated with the warm anomaly. The eye is encircled by the primary eyewall, characterized by intense thunderstorms with strong updrafts (~5-10 m s^{-1}) and the location of the cyclone's radius of maximum wind (or RMW). The primary eyewall and RMW are typically located ~10-100 km away from a cyclone's center, with smaller distances often (but not always) being associated with stronger cyclones. The primary eyewall is also characterized by nearly vertical isosurfaces of equivalent potential temperature, such that equivalent potential temperature is nearly constant with increasing height.

The eyewall and RMW within a mature tropical cyclone slope outward with increasing height at an angle approaching 45°, implying that the eyewall's outward displacement in the upper troposphere (relative to its location at the surface) is approximately equivalent to altitude. The physical reasoning behind this sloping structure lies with the conservation of angular momentum and the cyclone's warm-core structure. As air parcels ascend within the eyewall, angular momentum is approximately conserved. For a rotating body such as a tropical cyclone, recall that angular momentum is a function of the radius from the center of rotation (*r*) and the tangential wind speed (*v*), i.e.,

(1)
$$m = rv + \frac{fr^2}{2}$$

Since *v* decreases with increasing height, *r* must increase for *m* to remain constant.

A moat region, or region of predominantly stratiform precipitation, is found radially outward of the eyewall.

Continuing radially outward, mature tropical cyclones may possess a secondary eyewall. These secondary eyewalls form in response to the accumulation of heat energy, angular momentum, and vertical vorticity at some critical radius. The precise dynamics behind secondary eyewall formation remain unclear, however. Kossin and Sitkowski (2009) suggest that secondary eyewall formation is favored when the environment is conductive to high intensities; i.e., with high potential intensities, weak vertical wind shear, a deep layer of underlying warm water, and high mid- to upper-tropospheric relative humidity.

Secondary eyewall formation weakens lower-tropospheric inflow to the inner eyewall, temporarily halting tropical cyclone intensification. This inner eyewall gradually erodes, with the moat region between the two eyewalls clearing out, as the secondary eyewall matures (with strengthening ascent) and contracts to smaller radii. The inner eyewall may dissipate over a several-hour to several-day period, leaving a modestly weaker tropical cyclone with larger eye, broader near-surface wind field, and greater expanse of relatively intense near-surface winds. Reintensification may occur if the environment permits and another secondary eyewall does not immediately begin to form.

The tropical cyclone's rainbands lie beyond the secondary eyewall. These rainbands can be characterized as primary, secondary, or distant. The *primary rainband* is predominantly located in the tropical cyclone's inner-core region ($r < 200$ km). These rainbands typically remain quasi-stationary relative to the cyclone – i.e., they do not continually rotate around the cyclone. How these rainbands form remains unclear, however. Primary rainbands are characterized by new thunderstorms on their upwind flanks, mature thunderstorms in their centers, and decaying thunderstorms with predominantly stratiform precipitation on their downwind flanks. These thunderstorms slope radially outward with height but are generally constrained to within the lowest 8-10 above sea level. Finally, primary rainbands are typically located radially inward of a localized midtropospheric jet sometimes known as a secondary horizontal wind maximum.

Individual thunderstorms in primary rainbands include both updrafts and downdrafts, with the latter usually located radially inward of the former. These result in a sharp horizontal edge to the rainfall associated with the rainband. Low equivalent potential temperature air with these downdrafts can become entrained by the eyewall, wherein it can reduce the eyewall thunderstorms' intensity. Conversely, the updrafts transport high equivalent potential temperature air upward in the rainband. Dynamically, updrafts tilt horizontal vorticity, found beneath the secondary horizontal wind maximum, into the vertical and subsequently amplify it via stretching. Vertical advection accumulates vertical vorticity in the midtroposphere, where it subsequently helps to intensify the secondary horizontal wind maximum.

Secondary rainbands are typically found radially inward of a primary rainband. Secondary rainbands may be manifestations of vortex Rossby waves (e.g., Montgomery and Kallenbach 1997), which are waves that propagate on the radial gradient of cyclonic vertical vorticity outside of the eyewall (like midlatitude Rossby waves, which propagate along the large-scale meridional potential-vorticity gradient). Secondary rainbands propagate cyclonically around and radially outward from the tropical cyclone, both at speeds much smaller than the cyclone's tangential velocity. Energy associated with these features accumulates at which is known as a stagnation radius, whereby the tropical cyclone's strong tangential flow shears and mixes the energy around the tropical cyclone (known as *axisymmetrization*, or the process of symmetrizing the cyclone in all directions at a given radius or radii).

Finally, *distant rainbands* are composed of thunderstorms along confluence lines at large radii ($r > 200$ km) that are driven primarily by the environmental buoyancy. Significant vertical motions and lightning activity are often found with these rainbands. Tornadic activity is also possible along distant rainbands, particularly with those found in the right-front quadrant of landfalling tropical cyclones where the lower-tropospheric helicity is often maximized with environmental westerly vertical wind shear.

Inner-core thunderstorms are typically most intense during the local nighttime hours while outer rainbands are typically most intense during the local daytime hours. The oscillation between these two states is known as the "tropical cyclone diurnal cycle," which is most readily identified in intense tropical cyclones and can be observed as an outward-moving pulse in infrared satellite imagery (Dunion et al. 2014). The most-likely causes of these pulses are cloud-radiation feedbacks that vary between day and night (e.g., daytime heating of the boundary-layer by insolation; nighttime cooling by outgoing longwave radiation) and inertia-gravity waves that are convectively generated by latent heating in and near the cyclone's inner core (Dunion et al. 2014, 2019).

Secondary Circulation: Sawyer-Eliassen Framework

A tropical cyclone's secondary circulation is characterized by lower-tropospheric radial inflow, ascent near its center, and radial outflow near the tropopause. Colloquially, this is sometimes referred to as the cyclone's "in-up-out," or axisymmetric, circulation. This circulation is thermally direct because ascent is found where it is warmest. Compensating descent occurs at larger radii where it is relatively cooler. An exception is in the eye, where locally warm air descends.

The Sawyer-Eliassen secondary-circulation equation enables us to analytically describe how the secondary circulation responds to prescribed heat and/or momentum forcings. A derivation of this equation is provided in Appendix D; here, we simply state the final derived equation,

(2) $$\frac{\partial}{\partial r}\left(N^2\frac{1}{r}\frac{\partial(r\psi)}{\partial r}+B\frac{\partial\psi}{\partial z}\right)+\frac{\partial}{\partial z}\left(B\frac{1}{r}\frac{\partial(r\psi)}{\partial r}+I\frac{\partial\psi}{\partial z}\right)=\frac{g}{\theta_0}\frac{\partial Q}{\partial r}-\frac{1}{r^3}\frac{\partial F}{\partial z}$$

where *r* is radius, ψ is the streamfunction, *Q* is a prescribed heat forcing (positive for warming), and *F* is a prescribed momentum forcing (positive for an inward cyclonic momentum forcing).

Coefficients in (2) are given by:

(3) $$N^2=\frac{g}{\theta_0}\frac{\partial\theta}{\partial z}$$

(4) $$B=-\frac{g}{\theta_0}\frac{\partial\theta}{\partial r}=-\frac{1}{r^3}\frac{\partial m^2}{\partial z}$$

(5) $$I=\frac{1}{r^3}\frac{\partial m^2}{\partial r}=\left(f+\frac{1}{r}\frac{\partial(rv)}{\partial r}\right)\left(f+\frac{2v}{r}\right)$$

where (3) defines the static stability, (4) defines the baroclinicity, and (5) defines the inertial stability.

Finally, the streamfunction ψ is related to the zonal velocity *u* and vertical velocity *w* by:

(6) $$u=-\frac{\partial\psi}{\partial z},w=\frac{1}{r}\left(\frac{\partial(r\psi)}{\partial r}\right)$$

The streamfunction responds to prescribed heat and/or momentum forcing to restore thermal-wind balance that the specified heating and/or momentum forcing disrupts!

Let’s consider the secondary-circulation responses to heat and momentum forcings:

- **Positive heating (warming)**: Warming forces convergence below, ascent through, and divergence above the heat source. Adiabatic expansion (and thus adiabatic cooling) associated with the ascent attempts to mitigate the warming that gave rise to the ascent in the first place.
- **Cyclonic momentum**: This source forces enhanced radial outflow through the region of cyclonic momentum, with ascent below and radially inward from this region. The radial outflow, implying upper-tropospheric divergence and anticyclonic flow, attempts to mitigate the cyclonic momentum forcing that gave rise to the outflow in the first place.

Inertial stability, baroclinicity, and static stability modulate these responses. For localized heat and cyclonic momentum sources, the induced motions are vertically constrained for large static stability (large N_2) and weak inertial stability (small *I*). The induced motions are horizontally constrained for strong inertial stability (large *I*). The induced motions are primarily upright for a barotropic vortex (zero *B*) and vertically tilted for a baroclinic vortex (non-zero *B*), with the degree of this tilting directly proportional to the baroclinicity.

Localized upper-tropospheric heat and cyclonic momentum sources enhance a tropical cyclone’s secondary circulation. Strengthening the secondary circulation enhances the upward mixing of enthalpy gained from the underlying ocean, in turn intensifying the tropical cyclone’s primary circulation. This intensification is maximized near and slightly radially inward of the radius at which the heat or momentum forcing is applied, and these forcings are most efficient at intensifying a tropical cyclone when they are located near the RMW, when the tropical cyclone is relatively small, and when the tropical cyclone is already intense.

A tropical cyclone's response to heat forcing provides a nice context for explaining the RMW contraction that is typically observed with mature tropical cyclones. Warming near or radially inward of the RMW will increase the cyclone's tangential winds radially inward from the RMW, thus resulting in the RMW moving radially inward.

With to cyclonic momentum forcing, it is important to remember that such forcing is often associated with an upper-tropospheric trough, which itself is accompanied by vertical wind shear and cooler, drier mid- to upper-tropospheric air. Tropical cyclones only intensify if the positive impact of the cyclonic-momentum source exceeds the negative impact of dry air import and vertical wind shear, which generally only occurs when the environment is relatively moist and characterized by weak vertical wind shear.

Observations and output from high-resolution numerical model simulations support the analytical insights derived from the Sawyer-Eliassen equation, particularly regarding localized heat forcing. While convective bursts (i.e., vortical hot towers) frequently occur near the center of both weakening and intensifying tropical cyclones, they are preferentially located inward of the RMW for intensifying tropical cyclones, particularly those that are rapidly intensifying (e.g., Rogers et al. 2013, Hazelton et al. 2017). The region radially inward from the RMW is characterized by strong inertial stability, which confines the radial extent of the warming to the region inside the RMW where the cyclone's warm thermal anomaly is strongest. Further, convective bursts in intensifying tropical cyclones are often found upshear, where they may act to reduce the vortex's shear-induced tilt (e.g., Stevenson et al. 2014). This is also connected to the warming's radial extent: a more-upright heating source horizontally concentrates the heating, resulting in a more localized but stronger response in the vortex's mass and wind fields.

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