Tropical Cyclone Motion

Introduction

To first order, tropical cyclone motion is modulated by the large-scale flow. In the following, we aim to conceptually and mathematically describe this influence. Additional contributions to tropical cyclone motion arise from meso- to synoptic-scale asymmetries driven by a multitude of factors. This lecture closes by describing such asymmetries as well as discussing how and why they impact tropical cyclone motion.

Key Concepts

- What are the factors controlling tropical cyclone motion?
- How do these factors vary as a function of tropical cyclone intensity?

Climatological Perspective on Tropical Cyclone Motion

To first order, tropical cyclones track around the periphery of subtropical anticyclones. In this sense, tropical cyclones originate in the tropics and either 1) track westward to landfall or colder waters and ultimate dissipation or 2) turn poleward and eastward (recurve) into the midlatitudes, ultimately dissipating or undergoing extratropical transition. The global mean latitude of recurvature, defined as the latitude at which the tropical cyclone no longer has a westward component of motion, is approximately 25° (higher in the Northern Hemisphere) The meridional component of motion of tropical cyclones is typically poleward from genesis and increases in magnitude as tropical cyclones enter the midlatitudes. Average translation speeds are quite low in the tropics (~10 kt) but increase rapidly with increasing distance from the Equator; there is greater variability in translation speed for eastward-moving tropical cyclones versus their westward-moving counterparts. Given that track forecast errors are proportional to translation speed, this implies that larger track forecast errors may be expected for tropical cyclones recurving into the midlatitudes.

Several modes of intraseasonal and interseasonal variability influence tropical cyclone tracks across the globe. On time scales of 1-3 weeks, the evolution of the Rossby wave train across the midlatitudes exerts a significant influence on tropical cyclone motion. When synoptic-scale ridging is promoted across the subtropics, tropical cyclones tend to move more westerly; when a synoptic-scale trough is promoted across some portion of the subtropical oceans, tropical cyclones tend to recurve into the midlatitudes. On seasonal time scales, recurving tropical cyclones occur more frequently early and late in the season due to seasonal variations in the equatorward extent of the midlatitude storm track. In the Indian Ocean, the monsoon can impact both tropical cyclone activity and motion. On annual to longer time scales, modes of variability such as ENSO favor preferred longwave patterns across the tropics, subtropics, and midlatitudes that can exert a significant influence on mean tropical cyclone activity and motion *within a given season*.

Large-Scale Influences on Tropical Cyclone Motion

Let us conceptualize a tropical cyclone as a solid rotating body such as a cylinder. The approximate horizontal and vertical scales of the cylinder are approximately 500-1000 km and 10-15 km, respectively. A tropical cyclone is embedded within an atmospheric flow of much larger horizontal scale (*O*(1000-10000 km)). As a result, a tropical cyclone may be viewed as an object that moves largely with the surrounding flow. Such a surrounding flow, often represented by the air flow found 5-7° latitude/longitude away from the center of the tropical cyclone, is referred to as the steering flow. As our focus is on the large-scale flow, the appropriate steering flow here is that with the flow associated with the tropical cyclone itself partitioned out of the total flow.

Dynamically, this can be viewed in terms of the absolute vorticity tendency equation. The local absolute vorticity tendency is related to horizontal advection, vertical advection, stretching (e.g., divergence multiplied by the absolute vorticity), tilting, and friction. In a cylindrical framework, this is expressed as:

$$\frac{\partial \eta}{\partial t} = -\vec{v} \bullet \nabla(\zeta + f) - (\zeta + f)(\nabla \bullet \vec{v}) - \omega \frac{\partial \zeta}{\partial p} + \left(\frac{\partial u}{\partial p} \frac{\partial \omega}{r \partial \lambda} - \frac{\partial v}{\partial p} \frac{\partial \omega}{\partial r} \right) + \vec{F} \tag{1}$$

where derivatives taken with respect to λ (r) are those in the azimuthal (radial) direction. Velocity vectors are two-dimensional, as defined by u = radial wind (positive outward) and v = tangential wind (positive cyclonic). Other variables have their standard meteorological meaning. Terms on the right-hand side of (1) are horizontal advection, stretching, vertical advection, tilting, and friction, respectively. The local absolute vorticity tendency is dominated by the horizontal advection of absolute vorticity (Chan 1984). This can be conceptualized as a tropical cyclone moving toward the area of maximum absolute vorticity tendency, akin to how extratropical cyclones move toward the area of maximum development (e.g., potential temperature advection, as in the Pettersen-Sutcliffe development framework).

As the large-scale flow is not vertically uniform, defining the large-scale flow that steers the tropical cyclone can be challenging. To overcome this problem, a vertically integrated or vertically averaged flow is often used. In this framework, the horizontally averaged large-scale flow between two vertical levels is integrated to obtain an estimate of the large-scale steering flow. The precise bounds, particularly the upper bound, on this integral are a function of tropical cyclone intensity. To first order, the vertical depth of the steering flow increases with increasing tropical cyclone intensity, as more intense tropical cyclones tend to be vertically deeper than their shallower counterparts.

Approximately 50-80% of the variance in tropical cyclone motion can be explained by the steering flow. Several mechanisms result in deviant motion from this steering flow. These are discussed next, with each considered in isolation to clearly identify their individual contributions to tropical cyclone motion.

Other Influences on Tropical Cyclone Motion

Beta Effect

The concept of the beta, or β, effect stems from the meridional variation in the Coriolis parameter f. This effect, sometimes referred to as β drift, superposes a weak northwestward (southwestward) steering current upon the tropical cyclone in the Northern (Southern) Hemisphere.

The β effect can be viewed in the context of the barotropic relative vorticity equation in a Cartesian framework, i.e.,

$$\frac{\partial \zeta}{\partial t} = -\vec{v} \bullet \nabla(\zeta + f) \tag{2}$$

Or, equivalently, assuming that f only varies in the meridional direction,

$$\frac{\partial \zeta}{\partial t} = -\vec{v} \bullet \nabla \zeta - \beta v \tag{3}$$

In (3), the first right-hand side term is a relative vorticity advection term. The second right-hand side term represents planetary vorticity advection, manifest as $\beta = \frac{\partial f}{\partial y}$, by the meridional wind.

Consider a symmetric vortex on a β plane (i.e., β = constant) with no mean environmental flow. A streamfunction can be defined to represent this vortex, such that:

$$u = -\frac{\partial \psi}{\partial y} \qquad \text{and} \qquad v = \frac{\partial \psi}{\partial x}, \qquad \text{where} \quad \zeta = \nabla^2 \psi$$

At the initial time, the streamfunction and relative vorticity are concentric. As a result, there is no horizontal relative vorticity advection and, thus, the second right-hand side term of (3) is the only term impacting the local time rate of change of relative vorticity.

Note that β is positive-definite for the Northern Hemisphere while v is positive (negative) for northward (southward) motion. Thus, for a Northern Hemisphere tropical cyclone, this term will result in a positive (negative) relative vorticity tendency to the west (east). Physically, this can be viewed from the perspective of the conservation of absolute vorticity (which is defined by (2) if you combine the two terms): southward (northward) motion is associated with decreasing (increasing) planetary vorticity. For absolute vorticity to be conserved, relative vorticity must increase (decrease) to the west (east). If a tropical cyclone moves toward regions of increasing relative vorticity and away from regions of decreasing relative vorticity, this gives a westward translation of a tropical cyclone.

The β effect is strongest at large radii from the tropical cyclone's center. Here, the tropical cyclone's circulation extends over a comparatively large meridional distance. As such, although the meridional wind is weaker at large radii, meridional planetary vorticity advection is comparatively large at larger radii. This results in the streamfunction becoming slightly distorted. Near the center, the meridional planetary vorticity advection is approximately zero. As a result, streamfunction and relative vorticity are largely unchanged. At larger radii, the streamfunction expands to the west and contracts to the east. This leads to the westward displacement of the location of maximum streamfunction from the location of maximum relative vorticity.

With the streamfunction and relative vorticity maxima no longer overlapping, the advection term in (3) can no longer be neglected. The flow associated with the streamfunction is now able to advect relative vorticity. With southerly flow atop the maximum in relative vorticity, cyclonic relative vorticity is advected poleward. This results in cyclonic relative vorticity tendency north of the tropical cyclone and anticyclonic relative vorticity tendency south of the tropical cyclone. If a tropical cyclone moves toward the location of the maximum cyclonic relative vorticity tendency, the advective effect leads to the northward displacement of the vortex. In all, the northwestward motion of the vortex is driven by the superposition of the westward displacement resulting from the meridional variation in f and the northward displacement that arises when horizontal advection associated with the westward-distorted streamfunction can act on the eastward-lagging cyclonic relative vorticity maximum. Similar arguments can be made for the Southern Hemisphere, noting that the only fundamental difference from the above is the direction of the planetary vorticity gradient.

Viewed in the context of the streamfunction, the β effect leads to positive relative vorticity tendency northeast and negative relative vorticity tendency southwest of the center. These are often referred to as "β gyres." The northwestward β drift can be conceptualized as resulting from the superposition of the inferred horizontal circulations associated with each tendency anomaly. Although β drift magnitude is sensitive to the outer-core vortex structure, it generally imparts a 1-2 m s^{-1} northwestward steering current on a tropical cyclone and accounts for approximately 10% of the variance in tropical cyclone motion. It should be noted that while observational evidence for the β effect exists, the signal is difficult to extract from the data given the dominance in the flow by that of the tropical cyclone itself.

Non-Uniform Horizontal Flow

In the above discussion on the β effect, we considered the case of no environmental flow. Now, we wish to consider the impact of non-uniform horizontal flow upon tropical cyclone motion. In general, non-uniform horizontal flow distorts the vortex by horizontal advection and deformation. This results in cyclonic (anticyclonic) relative vorticity tendency downstream (upstream), with horizontal variability resulting from the horizontal variation in the horizontal flow. The inferred horizontal circulations associated with each of the resulting relative vorticity tendency anomalies impact a weak localized steering current across a tropical cyclone, the specific details of which depend significantly on the horizontal flow structure.

To illustrate this, let us consider two conceptual examples. In the first, there is easterly flow to the north and westerly flow to the south of the vortex, each maximized at large radii from the tropical cyclone. The magnitude of the flow changes in a linear fashion from north to south, such that there is no horizontal flow through the cyclone's center. In this case, the northern half of the vortex will be distorted westward whereas the southern semicircle of the vortex will be distorted eastward; in both cases, this is downstream. This results in a positive relative vorticity tendency northeast and southeast and a negative relative vorticity tendency northwest and southeast of a tropical cyclone that, in turn, deforms the vortex at large radii but does not induce an anomalous steering current across the center. By itself, linear horizontal shear cannot impact tropical cyclone motion.

In the second example, there is westerly flow north of the vortex that decays to zero to the south. In this case, there is modest westerly flow through the center of the tropical cyclone. The northern semicircle of the vortex is again distorted downstream (here, eastward); however, there is little to no distortion of the southern semicircle. This results in a positive relative vorticity tendency northeast and a negative relative vorticity tendency northwest of the cyclone, leading to a southward steering current imposed on the vortex. Even more complex horizontally sheared flows exist and can influence tropical cyclone motion, but their precise impact is left as a thought experiment for the interested reader.

For simplicity, we have considered the effects of non-uniform horizontal flow in isolation from the β effect. In reality, however, the horizontal shear-induced pattern(s) of relative vorticity tendency superpose on those associated with the β effect. Depending on the structure of the horizontal shear, this can strengthen or weaken the flow associated with the β effect (described above), thereby modifying its impacts upon tropical cyclone motion.

Non-Uniform Vertical Tropical Cyclone Structure

To this point, we have assumed that the tropical cyclone has uniform vertical structure. Now, we wish to consider how the β effect is different when the tropical cyclone structure is baroclinic in nature, i.e., a warm-core vortex that weakens with increasing height. Recall that a key component to the β effect is the horizontal advection of the relative vorticity associated with the tropical cyclone. For a barotropic structure, the magnitude of this effect is equivalent at all altitudes. For a baroclinic structure akin to that of a tropical cyclone, however, the magnitude of this effect is much greater in the lower troposphere (where the vortex is strongest) than it is in the upper troposphere (where the vortex is weakest).

This results in vertical variability in the intensity of the β effect and resultant β gyres. Subsequently, the lower tropospheric portion of the vortex is advected to the northwest more rapidly than it is in the upper troposphere, resulting in a faster-than-before northwestward motion of the tropical cyclone. It also results in a slightly upwind-tilted tropical cyclone. Recall that the quasigeostrophic omega equation enables us to diagnose vertical motion as a function of the differential advection of cyclonic relative vorticity. In this scenario, forcing for descent is found to the northwest and forcing for ascent is found to the southeast. The resulting secondary circulation induces flow from northwest to the southeast in the lower troposphere and from the southeast to the northwest in the upper troposphere, thus counteracting the vertical tilt and tropical cyclone motion impacts that arise from allowing the intensity of the tropical cyclone to vary in the vertical.

Non-Uniform Vertical Flow

Consider a tropical cyclone with a warm-core structure, with maximum cyclonic relative vorticity atop the boundary layer that decreases with increasing height until the flow becomes anticyclonic near the tropopause. For simplicity, consider the case of superposing a purely westerly vertical wind shear atop this tropical cyclone. This displaces the upper-tropospheric anticyclone east of the lower-tropospheric cyclone. Invoking the potential vorticity-based “action at a distance” or “vertical penetration” ideas, we know that the anticyclone will induce anticyclonic flow at lower altitudes and that the cyclone will induce cyclonic flow at higher altitudes from the level at which the intensity of each is maximized. The combined effect of these induced flows leads to the northward displacement of the vertically sheared tropical cyclone. Similar arguments can be made for more complex shears.

Fujiwara Interaction

Fujiwara interaction describes mutual rotation of two vortices about a common locus. This locus is typically the mass-weighted centroid of the two vortices; if they are of equal strength, this center is precisely the middle point between their centers. Consider the case of two vortices of equal intensity. The flow around each steers the other. In the absence of the β effect, the two vortices rotate around each other relative to the fixed locus. In the presence of the β effect, the two vortices still rotate around each other relative to the locus; however, the locus is no longer fixed and, instead, moves northwestward in response to the β effect. If the two vortices are not of equal strength, the locus is closer to the stronger vortex such that the weaker vortex gradually becomes enveloped by its stronger counterpart.

Convective Asymmetries

Asymmetries in deep, moist convection about a tropical cyclone's center can also result in deviant motion from that associated with the synoptic-scale steering flow and β effect. To illustrate this, consider the isentropic potential vorticity tendency equation, neglecting friction but not diabatic heating:

$$\frac{DP}{Dt} = P\frac{\partial\dot{\theta}}{\partial\theta} - g\frac{\partial\theta}{\partial p}\left(\frac{\partial u}{\partial\theta}\frac{\partial\dot{\theta}}{\partial y} - \frac{\partial v}{\partial\theta}\frac{\partial\dot{\theta}}{\partial x}\right)$$

where P is isentropic potential vorticity, θ is the isentropic vertical coordinate (positive upward), $\dot{\theta}$ is the diabatic heating rate (positive for warming), and u, v, and g have their typical meaning. The first right-hand side term is known as the vertical diabatic term given its relationship to vertical diabatic heating gradients, whereas the second right-hand side term is known as the shear diabatic term given its relationship to vertical wind shear (changes in u and v with height in the θ coordinate) and horizontal diabatic heating gradients. In the context of the vertical diabatic term, convective heating will vertically redistribute potential vorticity, such that cyclonic potential vorticity is redistributed toward the surface and anticyclonic potential vorticity is redistributed aloft. Where this is asymmetric about a tropical cyclone, this can lead to a tropical cyclone moving toward areas of enhanced convection. In the context of the shear diabatic term, the specific effect depends on the nature of the convective heating asymmetry and vertical wind shear.

References

- Chan, J. C.-L., 1984: An observational study of the physical processes responsible for tropical cyclone motion. *J. Atmos. Sci.*, **41**, 1036-1048.

- Chan, J. C.-L., 2005: The physics of tropical cyclone motion. *Ann. Rev. Fluid Mech.*, **37**, 99-128.

- Elsberry, R. L., 1995: "Tropical cyclone motion." *Global Perspectives on Tropical Cyclones*, R. L. Elsberry (ed.). World Meteorological Organization, Geneva, Switzerland, Report No. TCP-38. [Available online at http://derecho.math.uwm.edu/classes/TropMet/GPTC/tcmotion.pdf].