Tropical Cyclone Motion

Introduction

To first order, tropical cyclone motion is largely modulated by the large-scale flow. In the following, we aim to describe this influence both conceptually and mathematically. 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 the movement of tropical cyclones?
- How do these 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 and ultimate dissipation or 2) turn poleward and eastward (in other words, recurve) into the middle latitudes, ultimately dissipating or undergoing extratropical transition over land or water. 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°. 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; note, however, that there is greater variability in translation speed for eastward-moving tropical cyclones as compared to westward-moving tropical cyclones. 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.

There exists significant intraseasonal and interseasonal variability influencing 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 a given season owing to the seasonality of the midlatitude storm track (i.e., as manifest by equatorward intrusions of midlatitude troughs). In the Indian Ocean, the monsoon circulation has been shown to impact both tropical cyclone activity as well as tropical cyclone motion. On annual and 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 body, here given by a rotating 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 time rate of change of absolute vorticity is a function of horizontal advection, vertical advection, stretching (e.g., divergence multiplied by the absolute vorticity), tilting, and frictional processes. In a cylindrical framework, this can be expressed by:

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

where derivatives taken with respect to λ (r) are taken in the azimuthal (radial) direction. Velocity vectors are two-dimensional, as defined by (u, v) = (radial flow – positive outward, tangential flow – 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. Chan (1984) demonstrated that the local absolute vorticity tendency is dominated by the horizontal advection of absolute vorticity. If the tropical cyclone moves toward the area of maximum increase in absolute vorticity, this conceptualizes the motion (or steering) of the tropical cyclone.

As the large-scale flow in the real atmosphere is not vertically uniform, defining the large-scale flow that actually steers the tropical cyclone can be challenging. To overcome this problem, a vertically-integrated or vertically-averaged flow is often considered. 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 the intensity of the tropical cyclone. To first order, the vertical depth of the steering flow increases with increasing cyclone intensity, as more intense tropical cyclones tend to be vertically-deeper than their shallower counterparts. Thus, the vertical structure of the horizontal winds and its variability across the tropics, subtropics, and midlatitudes exert a significant influence upon tropical cyclone motion.

As noted above, approximately 50-80% of the variance in tropical cyclone motion can be explained by this relatively simple concept of steering flow. A number of other mechanisms act to result in deviant motion from that given by the large-scale steering flow. These are discussed next, with each considered in isolation so as to clearly identify their individual contribution 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 vorticity equation, i.e.

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

Or, equivalently, in a linearized form,

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

In (3), the first term on the right-hand side is an advection term. The second term on the right-hand side of (3) depicts the linearized advection of Coriolis, manifest as $\beta = \frac{\partial f}{\partial y}$, by the meridional wind.

We first desire to consider a symmetric vortex on a β plane (i.e., β = constant) with no mean environmental flow. At the initial time, the streamfunction (defined by the two horizontal wind components u and v, here referring to their Cartesian forms) and the relative vorticity will initially be concentric. As a result, there is no horizontal advection of the relative vorticity and, thus, the second term on the right-hand side 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. For a tropical cyclone in the northern hemisphere, this term will result in a positive (negative) time rate of change of relative vorticity to the west (east) of a tropical cyclone. Physically, this can be viewed from the perspective of the conservation of absolute vorticity: southward (northward) motion is associated with decreasing (increasing) Coriolis parameter and, thus, planetary vorticity. For absolute vorticity to be conserved, the relative vorticity must increase (decrease) to the west (east). Assuming that 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.

This effect is strongest not in the inner core of the tropical cyclone, however, but outside of the radius of maximum winds, where environmental (large-scale) influences may exert a greater relative influence upon the structure of an intense feature such as a tropical cyclone. This results in the westward expansion and eastward contraction of the circulation envelope of the tropical cyclone. Subsequently, this leads to the westward displacement of the location of maximum streamfunction from the location of maximum relative vorticity. To this point, however, this only provides another perspective on the westward movement of the tropical cyclone in response to a meridionally-varying Coriolis parameter. How do we get from here to the observed northwestward β drift?

With non-overlapping streamfunction and relative vorticity maxima, the advection term in (3) can no longer be neglected. The flow associated with the streamfunction (recall its definition above) 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 (anticyclonic) relative vorticity tendencies to the north (south) of the tropical cyclone. Assuming again that 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 the Coriolis parameter and the northward displacement that arises when horizontal advection associated with the westward-distorted streamfunction is able to act upon the eastward-lagging cyclonic relative vorticity maximum.

Alternatively, the β effect can be viewed as a combination of asymmetric and symmetric processes. At the initial time, the streamfunction is entirely symmetric. The symmetric streamfunction acts upon the asymmetric vorticity, given entirely at the outset of the analysis by the meridional variability in the Coriolis parameter, to distort the vortex. This distortion subsequently manifests itself via asymmetries in the streamfunction, which then act upon the symmetric vorticity, or that associated with the tropical cyclone. The combination of the two processes is identical to that described above and presents an alternate perspective on the β drift-induced northwestward motion of a tropical cyclone in the Northern Hemisphere. Similar arguments can be made for the Southern Hemisphere, noting that the only fundamental difference from the Northern Hemisphere is the direction of the planetary vorticity gradient.

Under the β effect, enhanced cyclonic relative vorticity and tangential winds are found to the northeast of the center. Conversely, weakened cyclonic relative vorticity and tangential winds are found to the southwest of the center. These are often referred to as "β gyres." Further, as one might expect from the above, there exists some sensitivity in the precise magnitude of the β effect to the structure of the outer core of the vortex. In general, however, it imparts a 1-2 m s^{-1} steering current upon the tropical cyclone and accounts for approximately 10% of the variability 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 have considered the case where there is no environmental flow. Now, we wish to consider the impact of non-uniform, non-zero horizontal flow upon tropical cyclone motion. In general, non-uniform horizontal flow acts to horizontally distort the vortex via horizontal advection and deformation processes. This results in cyclonic (anticyclonic) relative vorticity tendencies downstream (upstream) of the vortex for a given horizontal flow. The anomalous cyclonic (anticyclonic) flow induced downstream (upstream) acts to impart a weak localized steering current upon a tropical cyclone. The precise impact of this current depends significantly upon the structure of the horizontal flow, however.

To illustrate this, let us consider two conceptual examples. In the first example, there is easterly flow to the north of the vortex and westerly flow to the south of the vortex, with each maximized at large radii from the center of the tropical cyclone. The magnitude of the flow changes in a linear fashion between the northern and southern extent of the vortex such that there is no horizontal flow through the center of the tropical cyclone. In this case, the northern semicircle of the vortex will be distorted westward whereas the southern semicircle of the vortex will be distorted eastward; in both cases, this is downstream. Anomalous cyclonic (anticyclonic) relative vorticity will be found northwest and southeast (northeast and southwest) of the center. This results in a weak large-scale meridional deformation flow at large radii but no induced motion across the center of the tropical cyclone. By itself, linear horizontal shear cannot impact the motion of the tropical cyclone.

In the second example, there is westerly flow to the north of the vortex that decays to zero south of the vortex. 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. Cyclonic (anticyclonic) relative vorticity tendencies – and, thus, anomalies – are promoted northeast (northwest) of the cyclone. The result is a southward-directed steering current that encompasses the center of the tropical cyclone. Note that more complex horizontally-sheared flows do exist and can influence tropical cyclone motion; their precise impact is left as a thought experiment for the interested reader.

To this point, for simplicity, we have considered the effects of non-uniform horizontal flow in isolation from the β effect. In the real atmosphere, however, the horizontal shear-induced pattern(s) of relative vorticity tendency superpose upon 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 considered cases where 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., where it decays 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 tilted structure to the 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 (ascent) is promoted to the northwest (southeast). The resulting secondary circulation, including flow from the northwest to the southeast in the lower troposphere and flow from the southeast to the northwest in the upper troposphere, works to counteract 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 again a tropical cyclone with a warm-core structure, with maximum cyclonic relative vorticity in the boundary layer that decreases with increasing height until anticyclonic flow is found near the tropopause. For simplicity, let us consider the case of superposing a purely westerly vertical wind shear atop this tropical cyclone. This has the effect of displacing the upper tropospheric anticyclone to the east of the lower tropospheric cyclone. Utilizing again our 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 entire vertically-sheared tropical cyclone. Similar arguments can be made for more complex shears; again, these are left as a thought experiment for the interested reader.

Fujiwara Interaction

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

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].