Tropical Cyclone Intensity Change

Introduction

To this point, we have discussed the dynamics and thermodynamics of the tropical cyclogenesis process. Now, we wish to consider what impacts tropical cyclone intensity after genesis. In so doing, we consider a wide array of factors, both positive and negative, known to influence tropical cyclone intensity. Some, such as vertical wind shear and dry air intrusion, are well-known. Others, like trough interaction, are less well-known but nevertheless important to the study of tropical cyclone intensity change.

Key Concepts

- What is the Maximum Potential Intensity (MPI) of a tropical cyclone, what are its underlying physical principles, and what keeps a tropical cyclone from reaching its MPI?
- How do oceanic eddies and upwelling impact tropical cyclone intensity?
- What is "ventilation," and what are its dynamic and thermodynamic impacts upon tropical cyclone intensity? What are the factors to which these impacts are sensitive?
- How can the interaction of a tropical cyclone with a mid-latitude trough result in intensity change?

Maximum Potential Intensity

The idea that a tropical cyclone can be approximated as a Carnot cycle enables us to determine a relationship for the maximum potential intensity (MPI) of a tropical cyclone. For a system in which dissipative heating is included, we first define an efficiency ε:

(1) $$\varepsilon = \frac{T_s - T_o}{T_o}$$

T_s is the surface temperature whereas T_o is the temperature at the outflow layer just below the tropopause. For a difference of 75 K between T_s (~300 K/27°C) and T_o (~225 K/-48°C), the efficiency of the system is on the order of 0.33. Efficiencies of 0.3-0.5 are common in most tropical cyclone environments. Although higher efficiencies are possible for greater differences between T_s and T_o, such conditions are often not realized in nearly saturated tropical cyclone environments unless the tropopause is substantially colder than or found at elevated altitudes as compared to normal.

The rate of input of available potential energy from the underlying surface can be expressed as:

(2) $$G = \varepsilon C_k \rho V_s \left(k_o^* - k_a\right) T_s$$

where C_k is the enthalpy transfer coefficient, V_s is the surface wind speed, k_o* is the enthalpy of the ocean surface, and k_a is the enthalpy of boundary layer air near the ocean surface. We define enthalpy as equivalent (or nearly so) to moist static energy, which itself is closely related to equivalent potential temperature. For G to be positive, the enthalpy of the ocean surface must be higher than the enthalpy of the boundary layer near the ocean surface. This describes the scenario in which enthalpy, including sensible and latent heating, is transferred (or fluxed) from the ocean's surface to the atmospheric boundary layer. The rate of input of available potential energy increases linearly as a function of the surface wind speed.

The rate of dissipation can be expressed as:

(3) $$D = C_d \rho V_s^3$$

where C_d is the drag, or momentum transfer coefficient. The rate of dissipation increases exponentially with the maximum sustained surface wind speed.

At low wind speeds, the input of available potential energy from the underlying surface exceeds dissipation. The maximum potential intensity of a tropical cyclone is reached at the wind speed where the rate of dissipation becomes equal to the input of available potential energy. The precise wind speed at which this is realized depends on the enthalpy flux, and thus the warmth of the underlying surface. Thus, setting G equal to D and solving for V_s, we obtain an expression for the maximum possible sustained surface wind speed of a tropical cyclone:

(4) $$V_s^2 = \frac{C_k}{C_d} T_s \varepsilon \left(k_o^* - k_a \right)$$

The maximum sustained wind speed is a function of:

1. The enthalpy and momentum transfer coefficients, themselves a function of wind speed.
2. The sea surface and outflow layer temperatures, viewed in the context of efficiency.
3. The transfer (flux) of enthalpy from the underlying ocean into the boundary layer.

Factors (2) and (3) vary with environmental conditions; (1) varies as a function of tropical cyclone intensity. With respect to (1), note that the ratio between C_k and C_d increases linearly with wind speed at low wind speeds but levels off (i.e., begins to remain constant with increasing wind speed) near 30 m s^{-1}.

Of particular interest is factor (2), involving both the surface and outflow layer temperatures. This has important implications for variability of tropical cyclone activity in a warming environment (e.g., even if sea surface temperatures increase, increased outflow layer temperature will counteract the otherwise-expected increase in MPI) and for tropical cyclone activity at higher latitudes (as will be discussed in greater detail with tropical transition).

If we assume cyclostrophic balance, a fair assumption for hurricane-force wind speeds, we can obtain a relationship for the lowest-possible sea-level pressure of a tropical cyclone:

(5) $$p_c = p_m \exp\left(\frac{-V_s^2}{2R_d T_s} \right)$$

where p_c is the pressure at the center of the tropical cyclone, p_m is the pressure at the radius of maximum winds, and R_d is the dry air gas constant.

As may be inferred, MPI theory does not provide a pathway by which tropical cyclones intensify. Rather, in the context of WISHE theory, it provides insight as to the maximum intensity a given tropical cyclone can reach in a perfectly ideal, quiescent environment. Note the vast majority of storms do not reach their MPI. Departures from the aforementioned ideal environment, such as manifest via physical processes such as dry air intrusion, vertical wind shear, interaction with land, and/or cooler sea surface temperatures, are the primary causes behind this failure to reach the MPI. A small percentage of storms – on the order of 1% or less – reach and/or exceed their MPI. Given that the MPI is supposed to provide a maximum possible intensity, how can this be?

For some tropical cyclones, the impacts of asymmetric forcing unaccounted for by MPI theory are the culprit. Examples of asymmetric forcing are manifest via radial momentum flux sources such as mid-latitude troughs and vortex Rossby waves. For other tropical cyclones, the evolving energetics of tropical cyclones undergoing extratropical transition are the culprit. As a tropical cyclone undergoes extratropical transition, it begins to draw available potential energy from the vertically sheared midlatitude flow rather than from the underlying oceanic surface. This typically allows a tropical cyclone to maintain its intensity, particularly in terms of the minimum sea-level pressure of the cyclone, even as surface fluxes wane as the tropical cyclone moves over sub-critical sea surface temperatures (with no accompanying change in outflow layer temperature).

Three charts related to MPI are included within the lecture materials. The first of these charts depicts the minimum-possible sea-level pressure as a function of surface and outflow temperatures. Holding surface temperature constant, values of sea-level pressure decrease for decreasing outflow temperature (i.e., increasing the efficiency of the system). The second of these charts depicts the maximum-possible surface wind speed as a function of surface and outflow temperatures. Holding surface temperature constant, values of maximum surface wind speed increase for decreasing outflow temperature. The third chart utilizes the climatological atmospheric and oceanic conditions for September to construct spatial maps of the climatological minimum-possible sea-level pressure. MPI is at a maximum across the Western North Pacific warm pool north of Indonesia, near the southern Mexican coastline in the Eastern North Pacific, and across the central Gulf of Mexico in the North Atlantic. It decays rapidly away from the tropics. Global maps of daily MPI values are available online at http://wxmaps.org/pix/hurpot.html.

Oceanic Upwelling and Localized Oceanic Eddies

Due to Ekman transport, a tropical cyclone's cyclonic circulation cools the underlying waters due to upwelling from beneath the ocean's surface. Typically, stronger, larger tropical cyclones result in greater upwelling compared to their weaker, smaller counterparts. For all but the slowest-moving ($< 2\text{-}3 \text{ m s}^{-1}$) of cyclones, the overall upwelling magnitude is generally weak ($\leq 1°C$ cooling of the sea surface temperature). However, for slower-moving tropical cyclones, upwelling can significantly cool sea surface temperature, in turn reducing the available heat energy and thus efficiency and maximum potential intensity. Examples of upwelling exerting a substantial influence on the intensity of a nearly stationary tropical cyclone include Hurricanes Roxanne (1995) and Ophelia (2005), both in the North Atlantic basin.

A notable exception to this statement is found with tropical cyclones that follow closely in the path of an earlier tropical cyclone. For example, sea surface temperature after a tropical cyclone passage remains below-normal for as much as one to two months after passage, with the greatest impacts felt in the first two weeks after passage (Hart et al. 2007; Schenkel and Hart 2015). Due to both reduced sea surface temperature and the cyclone-induced meridional flow of warm, moist air out of and cool, dry air into the tropics, lower-tropospheric equivalent potential temperature is also reduced after a tropical cyclone passage (Schenkel and Hart 2015). Thus, in situations where a tropical cyclone follows a similar path to that a preceding tropical cyclone, the intensity of the subsequent tropical cyclone may be limited by the preceding tropical cyclone.

Conversely, localized oceanic warm eddies can positively influence tropical cyclone intensity. One example of such an eddy is that associated with the deflection of the Loop Current in the Gulf of Mexico. Localized warm eddies are associated with warm water over a large depth and an above-normal sea-surface height anomaly. Even in the presence of strong upwelling, water temperature at depth is sufficiently warm with these eddies so as to continue to provide ample fuel for a tropical cyclone's winds. There may also be a positive tropical cyclone intensity contribution from the mesoscale baroclinic zone (temperature gradient) that develops on the warm eddy's periphery. Some of the most intense Gulf of Mexico tropical cyclones in recent years (e.g., Hurricanes Katrina and Rita in 2005) passed over or near a warm eddy as they intensified. Global maps of real-time oceanic heat content and sea-surface height anomalies are available online at http://www.aoml.noaa.gov/phod/cyclone/data/.

Ventilation of the Tropical Cyclone Inner Core

The ventilation of a tropical cyclone's inner core refers to the dynamic and thermodynamic impacts of vertical wind shear upon the tropical cyclone (Tang and Emanuel 2010). The dynamic component deals with how the tropical cyclone vortex interacts with the environmental vertical wind shear. Implicit to this is the tilt of the vortex due to differential vorticity advection by the vertically sheared background flow and the resulting asymmetric structure (e.g., Jones 1995). The thermodynamic component concentrates on the impact upon a tropical cyclone's intensity resulting from importing low entropy air (low equivalent potential temperature air, comprised of relatively cool and/or dry air) from the ambient environment into the tropical cyclone's inner core. Two ways exist by which low entropy air can infiltrate the tropical cyclone inner core: the direct middle-tropospheric import of low entropy air by the vertically sheared flow and indirect lower-tropospheric import of low entropy air facilitated by evaporatively driven downdrafts in thunderstorms at larger radii (e.g., Riemer et al. 2010). We now seek to discuss both dynamic and thermodynamic effects of vertical wind shear on tropical cyclone intensity.

Dynamical Effects of Vertical Wind Shear

Dynamically, vertical wind shear vertically tilts the tropical cyclone's circulation. As expected, this results in a downshear-tilted vortex at the onset of vertical wind shear; e.g., for westerly vertical wind shear, the upper-tropospheric vortex will be tilted to the east of the lower-tropospheric vortex. In isolation, even a moderate vertical wind shear of 10 m s^{-1} will result in a vortex tilt of 864 km over one day. However, after a very short period of time (on the order of 1 h), the situation becomes much more complex, which can add to or counteract the tilt induced by the environmental vertical wind shear.

Consider an initially downshear-tilted vortex. The result is the vertical misalignment of the lower- and upper-tropospheric vortices. Following the "action at a distance" potential vorticity principle, the lower-tropospheric vortex can induce a weak cyclonic circulation at upper levels whereas the upper-tropospheric vortex can induce a weak cyclonic circulation near the surface. The strength of these circulations decays with increasing vertical distance from the inducing vortex and is a function of the Rossby penetration depth. For a rapidly rotating vortex, the Rossby penetration depth can be expressed as (Jones 2004):

(6) $$H_R = \frac{\left[\left(f + \frac{2v_T}{r}\right)(f + \zeta)\right]^{1/2} L}{N}$$

where f is the Coriolis parameter, v_T is the tangential wind, r is the radius, ζ is the relative vorticity, L is the horizontal length scale of the vortex, and N is the square root of the static stability. A stronger and/or larger vortex, larger Coriolis parameter, or reduced static stability results in a larger Rossby penetration depth. In turn, a larger Rossby penetration depth implies a larger depth over which a vortex may induce a meaningful circulation. Note that diabatic processes, such as associated with deep, moist convection, generally act to locally reduce the static stability and increase rotation, thereby increasing the Rossby penetration depth.

The resulting induced circulations can modify vertical wind shear in two ways. The flow associated with these circulations is also vertically sheared, and this can add to, oppose, or not influence environmental vertical wind shear. Similarly, the flow associated with these circulations imparts a steering current that can add to, oppose, or not influence environmental vertical wind shear.

To illustrate these concepts, let us return to our initially downshear-tilted vortex, here assumed to result from westerly vertical wind shear. For this scenario, cyclonic rotation increases with increasing height with the upper-tropospheric vortex whereas it decreases with increasing height with the lower-tropospheric vortex. For the upper-tropospheric vortex, the resulting vortex-induced vertical wind shear is easterly to the north (reducing shear there) and westerly to the south (increasing shear there). For the lower-tropospheric vortex, the resulting vortex-induced vertical wind shear is westerly to the north (increasing shear there) and easterly to the south (reducing shear there). Further, the lower-tropospheric vortex and its upward reflection impart a southerly steering current on the upper-tropospheric vortex, whereas the upper-tropospheric vortex and its downward reflection impart a northerly steering current on the lower-tropospheric vortex. With time, the resulting mutual cyclonic rotation can bring the vortices into a configuration that opposes the westerly environmental vertical wind shear. This process is known as *precession*, and is favored when the differential vortex advection rate (e.g., vertical wind shear magnitude) is smaller than the precession rate (Rappin and Nolan 2012), deep, moist convection is intense and located near the cyclone's center (e.g., Tao and Zhang 2015), and/or vortex Rossby waves are present and able to counteract the vertical wind shear (e.g., Reasor et al. 2004).

The structural asymmetry brought about by vertical wind shear may also modify vertical wind shear and vortex tilt. The vertically sheared flow results in differential cyclonic vorticity advection, which in turn results in ascent downshear and descent upshear. As potential temperature increases with increasing height, this results in positive middle-tropospheric potential temperature advection upshear and negative middle-tropospheric potential temperature advection downshear, leading to a cold anomaly collocated with ascent and a warm anomaly collocated with descent. Assuming potential temperature is approximately conserved, cyclonic rotation through these anomalies results in ascent 90° downstream of the warm anomaly (as parcels ascend along an isentrope) and descent 90° downstream of the cold anomaly (as parcels descend along an isentropes). Note that this structure often develops within the first 6 h after imposing vertical wind shear and remains robust with time, even as the other processes described above may act (Jones 1995). From the continuity equation, ascent is driven by convergence below and divergence above whereas descent is driven by divergence below and convergence above. The resulting cross-vortex flow from divergent to convergent locations counteracts the environmental vertical wind shear within the inner core, reducing tilt in the inner core relative to the outer core (Jones 1995, 2000ab).

Thermodynamic Effects of Ventilation and Vertical Wind Shear

Thermodynamically, the general impact of low entropy air upon tropical cyclone intensity can be viewed in terms of the Carnot heat engine approximation (Tang and Emanuel 2010). Importing low entropy air into the inner core in the lower or middle troposphere locally decreases the entropy along the ascending branch of the Carnot heat engine. Over a sufficient length of time, convective motions spread low entropy air over a deep, slantwise layer in the inner core. This weakens the differential in entropy (or, alternatively, equivalent potential temperature) between the eyewall and ambient environment, reducing the amount of energy available to the tropical cyclone. Concurrently, the implied buoyancy reduction results in a decrease in outflow layer height and an increase of the outflow layer temperature. As the efficiency of the Carnot heat engine is a function of the difference between the surface and outflow layer temperatures, this reduces the tropical cyclone's efficiency at converting available potential energy (heating) into kinetic energy.

There are two ventilation pathways: one direct, and one indirect. In the direct pathway, low entropy air is mechanically forced by vertical wind shear into the inner core in the middle troposphere. In the indirect pathway, low entropy air infiltrates outer rain bands, resulting in evaporatively driven downdraft formation; the resulting low equivalent potential temperature air is then drawn inward along inflowing trajectories. In this indirect pathway, downdraft formation preferentially occurs downshear, as this is where deep, moist convection preferentially forms when vertically sheared due to differential cyclonic vorticity advection. Of these ventilation pathways, direct middle-tropospheric import has a greater negative influence on tropical cyclone intensity than indirect lower-tropospheric import. Whereas entropy may partially be restored via surface fluxes from the underlying ocean in the indirect paradigm, no such recovery is possible in the middle troposphere. As a result, tropical cyclones are substantially less resilient to the direct import of low entropy air, most commonly associated with colder, drier environmental air, via vertical wind shear in the middle troposphere than they are to the indirect import of low entropy air along the radially inflowing branch of its secondary circulation in the lower troposphere. However, inner-core ventilation far more commonly occurs via indirect lower-tropospheric import (Riemer et al. 2010, 2013; Riemer and Laliberté 2015), whether in numerical simulations or in observations.

The precise impact of ventilation upon tropical cyclone intensity depends on several factors:

1. How low is the entropy of the air that is imported into the inner core?
2. How strong is the import of the low entropy air into the inner core?
3. How favorable is the ambient environment (e.g., warmth of the sea surface)?

Lower entropy air (fostering more frequent and more intense evaporatively driven downdrafts), stronger import, and marginal or unfavorable ambient thermodynamic conditions have a more substantial deleterious impact upon the intensity of the tropical cyclone (Tang and Emanuel 2010; Riemer et al. 2013). Together, ventilation can lead to a nearly 60% reduction in the maximum attainable intensity of a given tropical cyclone. If ventilation is sufficiently large, tropical cyclone dissipation is also possible. Conversely, only slightly reduced entropy air, weaker import, and favorable ambient thermodynamic conditions may lead to only a small negative impact upon tropical cyclone intensity.

The impact of ventilation upon tropical cyclone intensity also depends to some extent on intensity itself (Riemer and Montgomery 2011). The strong inertial (or rotational) constraint exerted on a fluid in a region of strong rotation acts to dampen the rate at which asymmetric flows impinge upon the vortex. As a result, as the intensity and/or size of the tropical cyclone increases, the vertical wind shear magnitude necessary for the interaction of the eyewall and outer rain bands with environmental air also increases.

Furthermore, the ability of dry environmental air to enter the circulation and impact tropical cyclone intensity is dependent upon the environmental air's location with respect to the vortex and the vertical wind shear direction. Consider a case of easterly vertical wind shear. For a westward-moving storm (as steered by this flow), easterly vertical wind shear is comprised of westerly lower-tropospheric *storm-relative* flow and easterly upper-tropospheric *storm-relative* flow. If relatively dry lower tropospheric air is found to the west (east) of the tropical cyclone in the lower (upper) troposphere, it will be advected into the tropical cyclone's circulation. In the case of westerly vertical wind shear, these directions are flipped.

Trough Interaction

The interaction of a tropical cyclone with an upper-tropospheric trough can lead to intensification or weakening. At a basic level, strong vertical wind shear associated with an upper-tropospheric trough can result in weakening through the dynamical and thermodynamic means referenced above. This is particularly true for upper-tropospheric troughs of large horizontal scale (i.e., bigger than a tropical cyclone).

Conversely, there are two pathways by which an upper-tropospheric trough may lead to tropical cyclone intensification. The first pathway is associated with quasigeostrophic forcing for large-scale ascent. Forcing for ascent upshear of the center results in deep, moist convection initiation, and the diabatic heating profile associated with such convection in turn redistributes positive potential vorticity toward the surface and reduces the deep-layer vertical wind shear magnitude (Hanley et al. 2001; Fischer et al. 2017). Upshear ascent preferentially occurs when the upstream trough is relatively far removed from the cyclone's center, is zonally narrow, and is intense (Fischer et al. 2017). For upshear ascent, the balance between vortex tilt, ventilation, and convection favors intensification for weak vertical wind shear (≤ 10 m s^{-1}) and weakening for stronger vertical wind shear.

The second pathway, albeit the subject of some debate, is associated with the superposition of the positive potential vorticity anomalies associated with the upstream trough and tropical cyclone. If the upper-tropospheric positive potential vorticity anomaly is of similar horizontal scale to that of the tropical cyclone, the upper-tropospheric anomaly may induce a lower-tropospheric positive potential vorticity anomaly that superimposes with that of the tropical cyclone (Hanley et al. 2001). Again, if deep-layer vertical wind shear magnitude is sufficiently weak, cyclone intensification may occur. However, other investigators suggest no contribution from potential vorticity anomaly superposition (e.g., Galarneau et al. 2015), although there is some evidence to suggest that upper-tropospheric trough structure, as discussed above, is important.

There is disagreement regarding the extent to which trough interaction favors intensification, which may be a function of how trough interaction is defined between studies. For example, Hanley et al. (2001) suggest that trough interaction is more often favorable for intensification, with 60-80% of tropical cyclones that interact with a trough through either of the pathways referenced above intensifying. However, Hanley et al. (2001) excluded tropical cyclones near land and those over sub-26°C waters. Further, rather than using a quantitative metric to define trough interaction, they defined it as any intensity change in the presence of an upper-tropospheric trough. Conversely, Peirano et al. (2016) suggest that trough interaction is more often favorable for weakening, with trough interaction cases 4-14% less likely to intensify and 5-13% more likely to weaken than either all tropical cyclones or those with versus without trough interactions. In their study, trough interaction was defined using a quantitative interaction metric, and no case restriction due to location or sea-surface temperature was considered. Vertical wind shear magnitude is the primary control on tropical cyclone intensity change in trough interaction cases, with the trough interaction itself having a secondary influence (Peirano et al. 2016).

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