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 the intensity of the tropical cyclone after it has formed. In so doing, we will consider a wide array of factors, both positive and negative, known to influence tropical cyclone intensity. Some of these, 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?
- Is WISHE the dominant control on tropical cyclone intensity after tropical cyclogenesis?
- How do oceanic eddies and upwelling impact tropical cyclone intensity?
- What is meant by "ventilation" and what are its dynamic and thermodynamic impacts upon the intensity of a tropical cyclone? What are the factors to which these impacts are sensitive?
- How can the interaction of a tropical cyclone with a mid-latitude trough lead to its intensification or its dissipation?

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 the environments of most tropical cyclones. Though higher efficiencies are possible for greater differences between T_s and T_o, such conditions are often not realized in the nearly saturated environments of tropical cyclones unless the tropopause is substantially elevated 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. 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 is being transferred from the ocean's surface to the boundary layer near the ocean's surface. The rate of input of available potential energy increases linearly as a function of the maximum sustained 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.

Typically, the input of available potential energy from the underlying surface exceeds dissipation. The maximum potential intensity of a tropical cyclone is reached when the rate of dissipation becomes equal to the input of available potential energy, a condition that becomes reached at high wind speeds. 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, primarily as viewed in the context of efficiency.
3. The transfer (e.g., surface flux) of enthalpy from the underlying ocean into the boundary layer.

Factors (2) and (3) vary with environmental conditions; factor (1) varies as a function of the intensity of the tropical cyclone itself. 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 at hurricane-force wind speeds, we can obtain a relationship for the lowest-possible sea level pressure of a tropical cyclone:

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

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. A similar expression may be derived using gradient or even geostrophic balance; however, as we are typically interested in the maximum possible intensity for a given tropical cyclone, cyclostrophic balance is typically the most appropriate choice.

As may be inferred from the above, 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. We will discuss these phenomena in more detail in this and subsequent lectures. 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 energy from the vertically-sheared midlatitude flow (e.g., the conversion of available potential energy to kinetic energy) 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.

The Importance of WISHE at Higher Wind Speeds

Traditionally, WISHE has been viewed as an "exponential instability," meaning that as a tropical cyclone becomes more intense, the magnitude of the wind-induced heat and moisture fluxes increases, subsequently leading to the non-linear intensification of the tropical cyclone. However, recent work by Montgomery et al. (2009) and subsequent works, though still the subject of much heated debate within the tropical cyclone community, challenges this viewpoint. In Montgomery et al. (2009), they demonstrate that an intense tropical cyclone develops even when the surface latent heat flux is capped at less than 150 W m^{-2}, approximately equal to the "trade wind" value associated with the easterly 5-10 m s^{-1} trades flowing across a sufficiently warm oceanic surface. They argue that while surface fluxes remain important prior to tropical cyclogenesis (to enable boundary layer equivalent potential temperature to rise sufficiently to permit convective redevelopment), the exponential instability associated with WISHE does not drive the subsequent intensification of the tropical cyclone.

Instead, Montgomery et al. (2009) pose that local buoyancy within individual convective towers serves as the energy source for deep, moist convection and, by extension, the tropical cyclone itself. In this framework, cyclonic vertical vorticity is generated within these towers and subsequently grows upscale through vortex merger and filamentation processes. The resultant upward energy cascade, rather than the vertical lofting of increasingly large heat and moisture fluxes, leads to the intensification of the tropical cyclone. It should be noted that such convective processes occur within all tropical cyclones and thus, in and of itself, such a condition does not "disprove" WISHE theory. Rather, Montgomery et al. (2009) argue that the development of an intense tropical cyclone despite capped surface latent heat fluxes is what acts to disprove WISHE theory. However, more research is necessary to precisely identify and isolate the intensification mechanism(s) for tropical cyclones.

Oceanic Upwelling and Localized Oceanic Eddies

The cyclonic circulation associated with a tropical cyclone acts to upwell water from beneath the ocean's surface. The magnitude of this upwelling is a function of the structure of the cyclone's wind field. Typically, stronger, larger tropical cyclones result in greater upwelling compared to their weaker, smaller counterparts. As a tropical cyclone typically moves at a rate of speed of 2-3 m s^{-1} or greater, the overall magnitude of such upwelling is generally weak (≤ 1°C cooling of the sea surface temperature). Thus, to first order, upwelling is generally only significant for slow-moving or stationary tropical cyclones. 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 an eighteen-year composite of Northern Hemisphere tropical cyclones, Hart et al. (2007) demonstrate that the sea surface temperature in the wake of a tropical cyclone remains below normal for approximately one month after its passage. The greatest impacts are felt within ten days after tropical cyclone passage. Thus, in situations where a tropical cyclone follows a similar path to that a preceding tropical cyclone, its intensity may be limited to some extent by the upwelling of cool water induced by the preceding tropical cyclone.

Conversely, localized oceanic warm eddies can exert a positive influence on tropical cyclone intensity. Such eddies are common within the Gulf of Mexico and are associated with the Loop Current that extends from the Caribbean Sea northward into the Gulf of Mexico. These eddies are characterized by locally 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 so as to continue to provide ample fuel for a tropical cyclone's winds. There may also be a contribution to tropical cyclone intensity from the mesoscale baroclinic zone (i.e., temperature gradient) that develops on the warm eddy's periphery. Regardless, some of the most intense tropical cyclones in the Gulf of Mexico in recent years (e.g., Hurricanes Katrina and Rita in 2005) passed over or near a warm eddy during rapid intensification.

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 the import of low entropy air (i.e., 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:

1. The lower tropospheric ingestion of low entropy air resulting from downdrafts formed in response to the interaction of the tropical cyclone with vertical wind shear (Riemer et al. 2010).

2. The direct middle tropospheric import of low entropy air into the tropical cyclone's inner core.

Dynamical Effects of Vertical Wind Shear

Dynamically, vertical wind shear acts to vertically tilt the circulation of the tropical cyclone. At the onset of the vertical wind shear, this results in the tilt of the vortex in the downshear direction. For westerly vertical wind shear, the upper tropospheric vortex will be tilted to the east of the lower tropospheric vortex. The opposite is true for easterly vertical wind shear. However, after a very short period of time (on the order of one hour), the situation becomes much more complex. In this discussion, we will highlight the resultant mutual rotation of the upper and lower tropospheric vortices; the development of asymmetries in the environment of the tropical cyclone and their role in modulating its vertical tilt; and the sensitivity in these processes to the structure of the vertical wind shear and, in particular, the vortex itself.

Vertical wind shear initially acts to tilt the vortex downshear in the vertical. This results in the misalignment of the lower tropospheric and upper tropospheric vortices. From potential vorticity arguments, the lower tropospheric vortex can induce a cyclonic circulation at higher levels whereas the upper tropospheric vortex can induce a cyclonic circulation at lower levels. The strength of these circulations at any given level decays with increasing 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 in Jones (2004) and references therein:

(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 vertical component of the vorticity vector, L is the horizontal length scale of the vortex, and N is the square root of the static stability. Therefore, a stronger and/or larger vortex, larger Coriolis parameter, or reduced static stability results in a larger Rossby penetration depth. As might be expected, a larger Rossby penetration depth implies a larger depth over which a given 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 circulations at a distance act against the vertical wind shear in two ways. First, the flow associated with these circulations is also vertically-sheared, with strongest winds found near the level of the cyclonic PV anomaly. This can either add to or oppose the environmental vertical wind shear. Similarly, the circulations induced by these vortices act to influence the motion of the other vortex. This mutual rotation of the vortices can also bring them into configurations in which the environmental vertical wind shear can act to either reduce or further enhance the tilt of the vortex.

Vertical wind shear acts to destroy thermal wind balance in the environment of the vortex. In order for balance to be restored, a transverse (i.e., vertical) circulation must develop, the strength of which is directly proportional to the strength of the vortex. From quasi-geostrophic omega arguments, differential cyclonic vorticity advection leads to ascent while differential anticyclonic vorticity advection leads to descent. Thus, ascent occurs downshear while descent occurs upshear. Vertical motions subsequently force the development of potential temperature anomalies via vertical advection, with a cold anomaly forming at the location of ascent and a warm anomaly forming at the location of descent. If potential temperature is (approximately) conserved, vortex-relative flow through these anomalies forces ascent 90° downstream of the warm anomaly and descent 90° upstream of the cold anomaly.

From continuity considerations, the forced vertical motions are associated with divergence and convergence. In regions of ascent, there is lower level convergence and upper level divergence. The inverse is true in regions of descent. The cross-vortex flow from the location of divergence to the location of convergence is noted to counteract the vertical shear within the inner core of the vortex. Thus, the tilt of the inner core of the vortex is typically observed to be less than that of the outer core of the vortex. The subsequent interaction of the outer core of the vortex with the vertically-sheared flow results in the development of large-scale asymmetries at large radii from the center of the vortex (Jones 2000a) that may also act to counterbalance the environmental vertical wind shear.

In the analytical framework of Jones (1995), the 90° out-of-phase relationship between vertical motion and potential temperature anomalies develops within the first six hours after imposing vertical wind shear. This relationship remains robust after the first six hours even as the anomalies and upper and lower tropospheric vortices continue to rotate cyclonically about the middle tropospheric vortex (which, to first order, propagates with the middle tropospheric flow). However, in observed tropical cyclones, the region of ascent is typically found in the downshear-left quadrant (i.e., to the left of the shear vector) of the cyclone. This implies that an approximate steady-state solution for vortex tilt and mutual rotation exists that is not addressed in the framework of Jones (1995).

The resiliency of a tropical cyclone to vertical wind shear also depends in large part upon the radial structure of the tropical cyclone's tangential wind profile, as highlighted by Reasor et al. (2004) and Mallen et al. (2005). Tropical cyclones in which the tangential wind profile only decays gradually with increasing radius away from the inner core of the storm are better able to resist the deleterious effects of vertical wind shear. The reasoning behind this relates to vortex Rossby wave theory, whereby Rossby waves about the vortex at some critical radius from the center act to counteract the vertical wind shear. While the dynamics of vortex Rossby waves are beyond the scope of this course, Reasor et al. (2004) argue that the vortex Rossby wave paradigm provides a more complete framework by which vortex resiliency to vertical wind shear can be understood than does the Jones (1995) framework. This is particularly true in terms of addressing the reduction of vortex tilt and the aforementioned approximate steady-state structure of the tropical cyclone under persistent vertical wind shear.

To summarize, we present a hypothetical thought experiment. By itself, the imposition across the tropical cyclone of a 10 m s^{-1} tropospheric-deep vertical wind shear leads to the horizontal displacement of the lower and upper tropospheric circulation centers by 864 km over one day. As this is generally not observed, other processes must counteract this tilt. As described above, the frameworks of Jones (1995, 2000a,b) and Reasor et al. (2004) describe physical mechanisms by which this shear-induced tilt is (generally) reduced. The precise details of this reduction in the tilt of the vortex depend most significantly upon the radial structure of the vortex and the Rossby penetration depth.

Note that the studies considered to this point are all studies of tropical cyclone-like vortices in dry dynamical frameworks. It is true that a tropical cyclone-like vortex can be maintained for some time in a dry, vertically-sheared environment (e.g., Jones 2004). However, as observed tropical cyclones are, to first order, thermodynamically-driven heat engines, it is also important to consider the effects of vertical wind shear upon the thermodynamics of a tropical cyclone. This is the focus of the next section.

Thermodynamic Effects of Ventilation and Vertical Wind Shear

Thermodynamically, the general impact of low entropy air upon the intensity of a tropical cyclone can be viewed in terms of the Carnot heat engine analog to the tropical cyclone (Tang and Emanuel 2010). The import of low entropy air into the inner core of the tropical cyclone 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 act to spread this low entropy air through a deep, slantwise layer within 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 loss of buoyancy corresponds to a decrease in the height of the outflow layer and an increase of the temperature at the outflow layer. As the efficiency of the tropical cyclone heat engine is a function of the difference between the surface and outflow layer temperatures, this reduces the efficiency of the tropical cyclone.

Mechanically forcing low entropy air into the tropical cyclone's inner core in the middle troposphere is more deleterious upon the intensity of a tropical cyclone than is importing low entropy air into the tropical cyclone's inner core in the lower troposphere. In contrast to the lower troposphere, where entropy may partially be restored via surface fluxes from the underlying ocean, 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 import of low entropy air along the radially-inflowing branch of its secondary circulation in the lower troposphere.

The precise impact of ventilation upon the intensity of the tropical cyclone depends in part upon a substantial number of factors. Three such factors are highlighted by Tang and Emanuel (2010):

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, stronger import, and marginal or unfavorable ambient thermodynamic conditions have a more substantial deleterious impact upon the intensity of the tropical cyclone. As shown by Tang and Emanuel (2010), this can lead to a nearly 60% reduction in the maximum attainable intensity of a given tropical cyclone. If the ventilation magnitude 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 minimal deleterious impact upon tropical cyclone intensity.

Riemer and Montgomery (2011) suggest that the impact of ventilation upon the intensity of a tropical cyclone also depends in large part upon the strength of the tropical cyclone, as manifest by the magnitude of its tangential winds. 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, they suggest that the ability of dry environmental air to enter into the circulation and impact the intensity of a tropical cyclone is dependent upon its location with respect to the vortex and the direction of the vertical wind shear. Consider the case of easterly vertical wind shear, most generally comprised of westerly *storm-relative* flow in the lower troposphere and easterly *storm-relative* flow in the upper troposphere. Here, 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 the tropical cyclone strengthening or weakening. The strong vertical wind shear associated with such a trough can result in the weakening of the tropical cyclone via the mechanisms described 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 (Hanley et al. 2001). The first pathway is associated with quasi-geostrophic forcing for large-scale ascent. As such forcing spreads over a tropical cyclone, stronger ascent is promoted within the inner core of the tropical cyclone. From mass conservation arguments, this is associated with enhanced divergent outflow and convergent inflow into the tropical cyclone. If the vertical wind shear associated with the upper tropospheric trough is relatively weak (≤ 10 m s^{-1}) in the environment of the tropical cyclone, the resultant enhancement to the tropical cyclone's secondary circulation promotes its intensification. Conversely, if the vertical wind shear is relatively strong (≥ 15 m s^{-1}), its deleterious effects upon cyclone intensity dominate and the cyclone weakens.

The second pathway is associated with the superposition of the cyclonic potential vorticity anomalies associated with the upper tropospheric trough and the tropical cyclone. If the upper tropospheric potential vorticity anomaly is of similar horizontal scale to that of the tropical cyclone, the magnitude of the vertical wind shear can be relatively weak. In such a scenario, as the upper tropospheric anomaly approaches the tropical cyclone, the superposition of the two distinct anomalies is favorable for the deepening of the tropical cyclone. This may be manifest by, for example, the superposition of the lower tropospheric tropical cyclone circulation with that of the lower tropospheric cyclonic circulation induced by the upper tropospheric cyclonic potential vorticity anomaly. With time, however, the anomaly associated with the upper tropospheric trough weakens as a result of strong diabatic heating associated with the tropical cyclone.

Hanley et al. (2001) suggest that trough interaction is more often favorable than unfavorable to tropical cyclone intensification. Specifically, over a ten year sample of storms across the North Atlantic basin, they suggest that 78% of cases fitting into the second pathway and 61% of cases fitting into the first pathway led to the intensification of the tropical cyclone. These percentages are likely at the high end of expectations, however, given that they excluded tropical cyclones near land and/or over sub-26°C waters (e.g., likely undergoing extratropical transition). The precise dynamics of how the aforementioned forcing can result in the strengthening of the tropical cyclone's secondary and, subsequently, primary circulation(s) will be explored in a forthcoming lecture on tropical cyclone structure.

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