Tropical Cyclone Intensity Change

Introduction

Having previously described the necessary conditions for tropical cyclone formation, we now consider how *not* meeting these conditions can negatively influence a tropical cyclone's intensity, particularly as it relates to reduced sea-surface temperatures, increased vertical wind shear, and dry (or low-entropy) air infiltration. Before doing so, however, we introduce potential intensity, relating energy input to dissipation, that governs how intense a tropical cyclone may become within a quiescent environment.

Key Questions

- What defines the efficiency of a tropical cyclone's heat engine?
- What does a tropical cyclone's potential intensity (PI) represent?
- What physical factors influence the potential intensity?
- What are the two paradigms of tropical cyclone inner-core ventilation?
- What are the dynamical impacts of vertical wind shear on tropical-cyclone intensity?
- What are the thermodynamic impacts of vertical wind shear on tropical-cyclone intensity?
- How can a tropical cyclone's interaction with a midlatitude trough result in intensity change?

Potential Intensity

The idea that a tropical cyclone can be approximated as a Carnot cycle enables us to determine a relationship for a tropical cyclone's PI. To do so, we must first define an efficiency ε, which is a measure of the fraction of energy from the underlying surface that is able to be used to fuel the storm's winds. In its most basic of forms, the efficiency takes the form:

$$\varepsilon = \frac{T_s - T_o}{T_s}$$

In the above, T_s is the sea-surface temperature whereas T_o is the temperature at the outflow layer just below the tropopause. Thus, the efficiency is temperature change between the sea-surface and tropopause divided by the sea-surface temperature. However, this formulation does not account for dissipative heating resulting from turbulent kinetic-energy dissipation in the atmospheric boundary layer. Including dissipative heating, the efficiency can be defined as (Bister and Emanuel 1998):

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

Here, the outflow rather than sea-surface temperature appears in the denominator. Given that the sea-surface temperature is typically much warmer than is the outflow temperature (~300 K vs. ~200-225 K), the effect of this change is to *increase* the efficiency relative to the case where dissipative heating is not addressed.

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. For typical sea-surface temperatures, outflow-layer altitudes (and thus temperatures), and tropospheric lapse rates between the surface and outflow layer, most tropical cyclones are characterized by efficiencies between 0.3-0.5.

The available potential energy (analogous to enthalpy) transfer from the underlying ocean is given by:

$$G = \varepsilon C_k \rho V_s (k_o^* - k_a) \tag{2}$$

where C_k is the enthalpy transfer coefficient, V_s is the surface wind speed, $k_0^* = c_p T_{SST} + L_v q_s(p_{sfc}, T_{SST})$ is the saturation enthalpy of the ocean surface, $k_a = c_p T + L_v q$ is the enthalpy of boundary-layer air near the ocean surface, q is the specific humidity, and q_s is the saturation specific humidity (itself a function of the surface pressure p_{sfc} and sea-surface temperature T_{SST}). For a given T_{SST}, reduced p_{sfc} along the inflowing leg of the tropical cyclone's secondary circulation increases the saturation specific humidity, which then increases the difference between the saturation and boundary-layer enthalpies – the wind-induced surface heat exchange feedback loop! G is positive when k_0^* is larger than k_a, scales with the wind speed and with the difference between k_0^* and k_a, and is directly proportional to ε.

The rate of dissipation is given by:

$$D = C_d \rho V_s^3 \tag{3}$$

where C_d is the momentum transfer coefficient, representing the surface roughness. The rate of dissipation increases cubically with the surface wind speed.

The different relationships of V_s with G (linear) and D (cubic) hint at the relationship between G and D. For small V_s, $G > D$; however, for large V_s, $G <$ D. The PI is defined as the V_s at which $G = D$; i.e., energy input equaling energy dissipation. Setting (2) = (3) and solving for V_s, we obtain:

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

The PI (here representing the maximum surface wind speed and not the maximum wind speed at any vertical level in the tropical cyclone; Emanuel 2018) given by V_s in (4) 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, as 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 with the wind speed (increasingly linearly with wind speed at small V_s but beginning to remain constant with increasing wind speed at ~30 m s^{-1}).

Of particular interest is #2, involving the sea-surface and outflow-layer temperatures. This has important implications for variability of tropical-cyclone activity in a warming environment, in which outflow-layer temperatures are projected to warm more than sea-surface temperatures and reduce the expected increase in PI from increased sea-surface temperatures alone. It also has important implications for tropical cyclone activity at higher latitudes, where sea-surface temperatures are colder than their ~26.5°C "necessary" value but outflow-layer temperatures can also be lower than normal (such as with an upper-tropospheric trough cut-off from the synoptic-scale midlatitude flow). In such a case, the efficiency can be sufficiently high as to support a tropical cyclone over colder sea-surface temperatures.

If we assume cyclostrophic balance (relating the horizontal pressure-gradient and centrifugal forces), a fair assumption for hurricane-force wind speeds, we can obtain a relationship for the lowest-possible sea-level pressure of a tropical cyclone. In a natural-coordinate system, where the *s* direction is along the motion and the *n* direction is perpendicular and to the right of the motion, cyclostrophic balance is given by:

$$\frac{V^2}{R} = -\frac{\partial \Phi}{\partial n} \tag{5}$$

where R is radius, $\Phi = gz$ is the geopotential, and n is the normal direction (directed outward). Substituting with the ideal gas law ($p = \rho R_d T_v$) and the hydrostatic equation (in the form $\partial p = -\rho \partial \Phi$), we obtain:

$$\frac{V^2}{R} = R_d T_v \frac{\partial(\ln p)}{\partial n} \tag{6}$$

For $\partial n \approx R_{RMW} - R_{center} = R_{RMW}$, where *RMW* is the radius of maximum winds, we can write:

$$\frac{V^2}{R_{RMW}} = R_d T_v \frac{\ln(p_{RMW}) - \ln(p_c)}{R_{RMW}}, \text{ such that } V^2 = R_d T_v \left(ln(p_{RMW}) - ln(p_c)\right) \tag{7}$$

Finally, solving for p_c, we obtain:

$$p_c = p_{RMW}\, exp\left(\frac{-V_s^2}{R_d T_v}\right) \approx p_{RMW}\, exp\left(\frac{-V_s^2}{R_d T_{SST}}\right) \tag{8}$$

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

Note that PI theory does not provide a pathway by which tropical cyclones intensify. Rather, in the context of WISHE theory, it provides insight as to the potential intensity that a given tropical cyclone can reach in a perfectly ideal, quiescent environment. Most tropical cyclones do not reach their PI. Departures from the ideal environment such as dry-air intrusion, vertical wind shear, landfall, or cooler sea-surface temperatures often keep a tropical cyclone from reaching its PI. Only ~1% of all tropical cyclones reach or, rarely, exceed their PI. The latter is most common when a tropical cyclone is transforming into an extratropical/midlatitude cyclone through *extratropical transition* at higher latitudes, where sea-surface temperatures are low and the cyclone is beginning to be fueled by midlatitude processes.

Oceanic Upwelling and Localized Oceanic Eddies

Ekman transport causes a tropical cyclone's cyclonic circulation to locally reduce sea-surface temperatures by upwelling colder subsurface water. Stronger, larger, slower-moving tropical cyclones are associated with greater upwelling compared to their weaker, smaller, faster-moving counterparts. The effects of upwelling on sea-surface temperatures are generally small (≤1°C cooling) except for very-slow-moving cyclones. For these cases, upwelling can significantly reduce the sea-surface temperature, in turn reducing the enthalpy that can be gained from the underlying surface and thus also the efficiency and PI. Examples of upwelling exerting a substantial influence on intensity include North Atlantic hurricanes Roxanne (1995) and Ophelia (2005).

Tropical cyclones that follow closely in the path of an earlier tropical cyclone can have their PI limited by the oceanic upwelling that accompanied the earlier tropical cyclone. Sea-surface temperature after a tropical cyclone's passage remains below normal for as much as 1-2 months, with the largest impacts felt in the first 2 weeks after passage (Hart et al. 2007; Schenkel and Hart 2015). The accompanying reduction in enthalpy transfer from the ocean to the atmosphere decreases the lower-tropospheric equivalent potential temperature (as does cyclone-induced meridional transport of warm, moist air out of the tropics and cool, dry air to the tropics) and thus stabilizes the environment (Schenkel and Hart 2015).

Conversely, localized oceanic warm eddies can positively influence tropical cyclone intensity. One example of such an eddy is that associated with the Loop Current in the Gulf of Mexico. Localized warm eddies are associated with warm water to abnormally large depths. Even if there is strong upwelling, water temperature is sufficiently warm with these eddies as to continue to provide ample fuel for a tropical cyclone's winds. 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.

Inner-Core Ventilation

Inner-core ventilation refers to the combined dynamic and thermodynamic impacts of vertical wind shear upon a tropical cyclone (Tang and Emanuel 2010). The dynamic component refers to the tropical cyclone's interaction with the environmental vertical wind shear. The thermodynamic component refers to the impact upon a tropical cyclone's intensity from importing low entropy air (air that is relatively cool and/or dry with low equivalent potential temperature) from the ambient environment into the tropical cyclone's inner core. There are two ways that low entropy air can infiltrate the inner core: direct midtropospheric import by the vertically sheared flow and indirect lower-tropospheric import through evaporatively driven downdrafts in thunderstorms at larger radii (e.g., Riemer et al. 2010).

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 shear onset; e.g., westerly vertical wind shear tilts the tropical cyclone to the east. 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 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 a vortex that is initially tilted in the direction of the shear (known as downshear). This misaligns the lower- and upper-tropospheric portions of the vortex. Under the "action at a distance" potential vorticity principle, the lower-tropospheric vortex can induce a weak upper-tropospheric cyclonic circulation whereas the upper-tropospheric vortex can induce a weak near-surface cyclonic circulation. The resulting induced circulations can modify vertical wind shear in two ways. The flow associated with these circulations is also vertically sheared, which can add to or subtract from the environmental vertical wind shear. Similarly, the flow associated with these circulations imparts a steering current that can change the vertical alignment of a tropical cyclone's vortex.

To illustrate, 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 and 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). Conversely, 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 (directly related to 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 (Tao and Zhang 2015), and/or vortex Rossby waves are present and able to counteract the vertical wind shear (Reasor et al. 2004).

Structural asymmetries induced by vertical wind shear may themselves modify the vertical wind shear and the vortex's vertical tilt. Vertically sheared flow results in differential cyclonic vorticity advection, which in turn (under the quasi-geostrophic assumption) results in ascent downshear and descent upshear. Because potential temperature increases with increasing height, this results in positive vertical potential-temperature advection upshear and negative vertical potential-temperature advection downshear in the midtroposphere, leading to a cold anomaly collocated with ascent and a warm anomaly collocated with descent. Assuming that potential temperature is approximately conserved following the motion (neglecting diabatic processes), 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). From mass continuity, the ascent is associated with convergence below and divergence above whereas the descent is associated with 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 the vortex's inner-core tilt (Jones 1995, 2000ab).

Thermodynamic Effects of Ventilation and Vertical Wind Shear

Thermodynamically, the impact of low-entropy air on 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- to midtroposphere locally decreases the entropy along the inflowing branch of the Carnot heat engine. Although surface fluxes may warm and moisten this air as it approaches the cyclone's center, these fluxes are insufficient to increase the entropy to the levels it would have achieved in the absence of low-entropy air infiltration. Over time, thunderstorm updrafts spread low-entropy air over a deep slantwise layer in the inner core. This weakens the entropy differential between the eyewall and ambient environment, reducing the cyclone's intensity. Concurrently, the implied buoyancy reduction decreases the outflow-layer height and increases the outflow-layer temperature, consequently reducing the tropical cyclone's efficiency at converting available potential energy (enthalpy) into kinetic energy.

There are two ventilation pathways: direct and indirect. In the first, low-entropy air is mechanically forced into the inner core in the middle troposphere by vertical wind shear. In the second, low-entropy air infiltrates outer rain bands and results in evaporatively driven downdraft formation, with the resulting low equivalent potential temperature air then being drawn inward along inflowing trajectories preferentially downshear (as this is where thunderstorms preferentially form when the cyclone is vertically sheared). Of these ventilation pathways, the direct pathway has a greater negative influence on tropical-cyclone intensity: though surface enthalpy fluxes can partially restore entropy along inflowing trajectories in the indirect pathway, no such recovery is possible in the midtroposphere. However, inner-core ventilation far more commonly occurs via the indirect pathway (Riemer et al. 2010, 2013; Riemer and Laliberté 2015).

The precise impact of ventilation on 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 intense evaporatively driven downdrafts), stronger import, and marginal or unfavorable ambient thermodynamic conditions have a more substantial deleterious impact on tropical-cyclone intensity (Tang and Emanuel 2010; Riemer et al. 2013). Together, ventilation can lead to a nearly 60% reduction in a tropical cyclone's maximum-attainable intensity. If ventilation is sufficiently large, tropical-cyclone dissipation is also possible. Conversely, slightly reduced entropy air, weaker import, and favorable ambient thermodynamic conditions may lead to only a small negative impact of ventilation on tropical-cyclone intensity. Ventilation's impact also partially depends on the tropical cyclone's intensity (stronger cyclones have greater inertial stability, which dampens the rate at which asymmetric flows impact the vortex; Riemer and Montgomery 2011) and the low-entropy air's location relative to the vortex and the vertical wind shear's direction (low-entropy air upstream will ventilate the cyclone, whereas low-entropy air downstream will not).

Trough Interaction

A tropical cyclone's interaction with an upper-tropospheric trough can lead to intensification or weakening, the latter of which is more common and is physically manifest through the dynamical and thermodynamic ventilation pathways described above. Conversely, when an upstream upper-tropospheric trough is distant from a tropical cyclone's center, is zonally narrow, and is intense, ascent upshear of a tropical cyclone may facilitate tropical-cyclone intensification (Fischer et al. 2017). The upshear ascent facilitates thunderstorm initiation. The associated vertical diabatic-heating profile redistributes positive potential vorticity towards the surface, increasing the lower-tropospheric cyclonic rotation rate and decreasing the deep-layer vertical wind shear (Hanley et al. 2001; Fischer et al. 2017). However, these processes can result in intensification only if the vertical wind shear is sufficiently weak (≤ 10 m s^{-1}).

Studies disagree on the extent to which trough interaction can facilitate tropical-cyclone intensification. For example, Hanley et al. (2001) suggest that trough interaction is often favorable for intensification, with 60-80% of tropical cyclones that interact with a trough intensifying. However, they excluded tropical cyclones near land and over sub-26°C waters. Further, they defined trough interaction as any intensity change in the presence of an upper-tropospheric trough rather than using a quantitative metric to formally define trough interaction. Conversely, Peirano et al. (2016) suggest that trough interaction often favors weakening, with trough interaction cases ~10% less likely to intensify and ~10% more likely to weaken than other tropical cyclones. This study defined trough interaction more rigorously using a quantitative interaction metric and did not exclude cases due to their location or sea-surface temperature. 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).

References

- Bister, M., and K. A. Emanuel, 1998: Dissipative heating and hurricane intensity. *Meteor. Atmos. Phys.*, **65**, 233-240.

- Emanuel, K., 1986: An air-sea interaction theory for tropical cyclones. Part I: steady-state maintenance. *J. Atmos. Sci.*, **43**, 585-605.

- Emanuel, K., 2018: 100 years of progress in tropical cyclone research. *Meteor. Monographs*, **59**, 15.1-15.68.

- Fischer, M. S., B. H. Tang, and K. L. Corbosiero, 2017: Assessing the influence of upper-tropospheric troughs on tropical cyclone intensification rates after genesis. *Mon. Wea. Rev.*, **145**, 1295-1313.

- Galarneau, T. J., R. McTaggart-Cowan, L. F. Bosart, and C. A. Davis, 2015: Development of North Atlantic tropical disturbances near upper-level potential vorticity streamers. *J. Atmos. Sci.*, **72**, 572-597.

- Hanley, D., J. Molinari, and D. Keyser, 2001: A composite study of the interactions between tropical cyclones and upper-tropospheric troughs. *Mon. Wea. Rev.*, **129**, 2570–2584.

- Hart, R., R. Maue, and M. Watson, 2007: Estimating the atmospheric and SST memory of tropical cyclones through PI anomaly evolution. *Mon. Wea. Rev.*, **135**, 3990-4005.

- Jones, S. C., 1995: The evolution of vortices in vertical shear. Part I: Initially barotropic vortices. *Quart. J. Roy. Meteor. Soc.*, **121**, 821-851.

- Jones, S. C., 2000a: The evolution of vortices in vertical shear. Part II: Large-scale asymmetries. *Quart. J. Roy. Meteor. Soc.*, **126**, 3137-3159.

- Jones, S. C., 2000b: The evolution of vortices in vertical shear. Part III: Baroclinic vortices. *Quart. J. Roy. Meteor. Soc.*, **126**, 3161-3185.

- Jones, S. C., 2004: On the ability of dry tropical-cyclone-like vortices to withstand vertical shear. *J. Atmos. Sci.*, **61**, 114-119.

- Peirano, C. M., K. L. Corbosiero, and B. H. Tang, 2016: Revisiting trough interactions and tropical cyclone intensity change. *Geophys. Res. Lett.*, **43**, doi:10.1002/2016GL069040.

- Rappin, E. D., and D. S. Nolan, 2012: The effect of vertical shear orientation on tropical cyclogenesis. *Quart. J. Roy. Meteor. Soc.*, **138**, 1035-1054.

- Reasor, P. D., M. T. Montgomery, and L. D. Grasso, 2004: A new look at the problem of tropical cyclones in vertical shear flow: vortex resiliency. *J. Atmos. Sci.*, **61**, 3-22.
- Riemer, M., and F. Laliberté, 2015: Secondary circulation of tropical cyclones in vertical wind shear: Lagrangian diagnostic and pathways of environmental interaction. *J. Atmos. Sci.*, **72**, 3517-3536.
- Riemer, M., and M. T. Montgomery, 2011: Simple kinematic models for the environmental interaction of tropical cyclones in vertical wind shear. *Atmos. Chem. Phys.*, **11**, 9395-9414.
- Riemer, M., M. T. Montgomery, and M. E. Nicholls, 2010: A new paradigm for intensity modification of tropical cyclones: thermodynamic impact of vertical wind shear on the inflow layer. *Atmos. Chem. Phys.*, **10**, 3163-3188.
- Riemer, M., M. T. Montgomery, and M. E. Nicholls, 2013: Further examination of the thermodynamic modification of the inflow layer of tropical cyclones by vertical wind shear. *Atmos. Chem. Phys.*, **13**, 327-346.
- Tang, B., and K. Emanuel, 2010: Midlevel ventilation's constraint on tropical cyclone intensity. *J. Atmos. Sci.*, **67**, 1817-1830.
- Tao, D., and F. Zhang, 2015: Effects of vertical wind shear on the predictability of tropical cyclones: practical vertical intrinsic limit. *J. Adv. Model. Earth Sys.*, **7**, 1534-1553.