Tropical Cyclones (Hurricanes)

Brian Hong-An Tang, University at Albany, State University of New York, Albany, NY, United States

© 2024 Elsevier Ltd. All rights are reserved, including those for text and data mining, AI training, and similar technologies.

Key Points

- Tropical cyclone formation generally requires an initial cluster of convection, sufficiently warm sea waters, background vorticity, a moist and potentially unstable atmospheric region, and low values of vertical wind shear.
- The precursor disturbances and large-scale patterns by which tropical cyclones form vary between ocean basins and may be influenced by modes of synoptic variability.
- Mesoscale convective systems and deep, rotating convection within tropical disturbances and cyclones are building blocks that are important for development.
- A mature tropical cyclone consists of various types of rainbands, an eyewall, and an eye, whose characteristics are governed by different physical processes.
- Important environmental controls on tropical cyclones include vertical wind shear, ocean mixing, and trough interactions.
- Surface heat fluxes, convective-burst locations, and boundary-layer structure modulate tropical cyclone intensity change.
- Tropical cyclone size is influenced by both internal processes and external properties in which a tropical cyclone is embedded.

 https://doi.org/10.1016/B978-0-323-96026-7.00078-3

Synopsis

A tropical cyclone lifecycle encompasses formation from a cluster of storms, development into well-defined precipitation and wind features, and finally decay or transition as the storm ventures poleward or makes landfall. As a tropical cyclone evolves, it undergoes profound changes to its cloud, thermodynamic, and kinematic structure. These changes feedback on one another to increase the intensity and organization of the system. Tropical cyclones also interact and are affected by the surrounding atmospheric environment and upper ocean, which have strong influences on tropical cyclone formation, intensification, and size. This article overviews fundamental knowledge and some recent advances in these areas.

Introduction

Tropical cyclones are one of the most well-known and destructive natural hazards on this planet. Their presence has defined cultures, shaped history, and altered landscapes. Much has been gained in understanding and forecasting of tropical cyclones over the past half century, aided by increased observational technologies, computational power, and research advances. This chapter overviews the fundamentals and some frontiers in the science of tropical cyclones.

A tropical cyclone is a low-pressure, storm system that exists mainly over the tropical and subtropical oceans. Defining characteristics of a tropical cyclone include deep, organized convection; a well-defined, cyclonic circulation; and a warm-core structure, where temperatures at the center of the storm are warmer than the surrounding environment at a given height. Mature tropical cyclones often have an eye, a relatively calm, circular region in the center of the storm that is devoid of deep convection. Surrounding the eye is the eyewall, where the tallest clouds and most severe weather is located. Beyond the eyewall are spiral rainbands. The structure of tropical cyclones is detailed in the **Tropical Cyclone Structure** section.

Tropical cyclones are classified by intensity, measured by the maximum, sustained, near-surface wind speed. **Table 1** shows tropical cyclone classifications and associated intensities. Tropical cyclogenesis occurs when a tropical disturbance achieves enough of the defining characteristics of a tropical cyclone for a long enough time to be classified as a tropical depression. When a tropical cyclone reaches tropical-storm intensity ($17\ \mathrm{m\ s^{-1}}$), it is assigned a name from an ocean basin-specific list prepared by the World Meteorological Organization. Strong tropical cyclones with an intensity $>33\ \mathrm{m\ s^{-1}}$ are called hurricanes in the North Atlantic Ocean and eastern and central North Pacific Ocean, typhoons in the western North Pacific Ocean, and cyclones in the Indian Ocean and Southern Hemisphere. Despite the different names, the physics of these storms are fundamentally the same. Moreover, these strong tropical cyclones are classified based on a categorical scale. For hurricanes, the Saffir-Simpson scale is used (**Table 1**). Category three or greater hurricanes are known as major hurricanes, which have the potential to cause devastating damages and danger to life when they make landfall in populated areas. Similar scales are used in different ocean basins.

Tropical cyclones bring multiple hazards. One of the most recognizable hazards are the winds. The force of wind, and the damage it causes, increases nonlinearly with wind speed, so the strongest tropical cyclones cause a disproportionate amount of wind damage. Tropical cyclones can also produce tornadoes, particularly in rainbands to the right and in front of the storm's motion. The deadliest hazard is not the wind, but water-related hazards: the storm surge and inundating rains. Storm surge is the push of seawater onshore by a tropical cyclone's winds. Storm surge is influenced not only by the strength of the wind, but also by the size of the storm's wind field, the bathymetry of the seafloor, and the geometry of the coastline. The addition of large waves and storm surge coinciding with high tide exacerbates damage. Tropical cyclones are efficient producers of heavy rainfall, and this hazard can extend far inland. Tropical cyclone rainfall amounts can be especially large if a tropical cyclone has a large size, is moving slowly, is traversing over mountainous regions, and/or when interacting with midlatitude weather systems, resulting in enhanced ascent. Additionally, a tropical cyclone need not be strong to produce excessive amounts of rainfall, which makes it an insidious hazard. There have been recent efforts to better forecast and warn for water-related tropical cyclone hazards, rather than focusing too heavily on the wind and category.

Table 1 Tropical cyclone classifications by intensity and the Saffir-Simpson scale for hurricanes.

Classification	*Intensity ($\mathrm{m\ s^{-1}}$)*
Tropical depression	≤ 16
Tropical storm	17–32
Hurricane/typhoon/cyclone	≥ 33
Category 1	33–42
Category 2	43–49
Category 3	50–58
Category 4	59–69
Category 5	≥ 70

Climatology

Globally, there are about 80–100 tropical cyclones per year, and about 40–50 of them reach hurricane strength. Each ocean basin, however, has different tropical cyclone activity characteristics due to different background states (Fig. 1). For example, the western North Pacific has a plethora of activity, and very intense tropical cyclones frequent that basin, so the average background state is quite favorable for tropical cyclones there. Curiously, almost no tropical cyclones form near the Equator. Additionally, except for some tropical cyclones that rarely form off the east coast of South America, there is a void of tropical cyclone activity in the eastern South Pacific and South Atlantic.

There are several necessary ingredients required for tropical cyclones (Gray, 1968). First, tropical cyclones arise from tropical disturbances, which are coherent clusters of convection and often have weak, broad cyclonic vorticity. While convective clusters are omnipresent in the tropics, only a small percentage of them become tropical cyclones. Second, the energy source for tropical cyclones is surface heat fluxes, primarily evaporation of moisture from sufficiently warm sea waters. A general rule of thumb is sea surface temperatures $>26°C$, although there are some exceptions for tropical cyclones that form from initially cold-core systems and are baroclinically influenced. Third, tropical cyclones can be viewed as concentrators of Earth's background planetary vorticity or angular momentum, akin to a figure skater pulling their arms inward and gaining spin. At the Equator, there is no background planetary vorticity, which makes it difficult for tropical disturbances to acquire enough spin near the Equator. Fourth, the region in and around a tropical disturbance must be sufficiently moist and potentially unstable. Such an environment allows convection to be sustained and feedbacks to initiate that allow the development of a tropical cyclone. Fifth, the vertical wind shear cannot be too large. Vertical wind shear is the vertical change in the environmental winds around a tropical cyclone. If the wind shear is too large, the developing vortex cannot align, and dry air may be advected into the system. Even though these necessary ingredients might all be present at a given location and time, it does not guarantee a tropical cyclone will form. Rather, formation requires an alignment of processes at many scales, which will be explored more in the Tropical Cyclogenesis section.

These ingredients have a seasonality, being more likely to be present in the summer and fall months, so that is why tropical cyclone activity typically peaks then (Fig. 2). The North Atlantic has a more sharply peaked season, with August and September being the most active months (Fig. 2D). The eastern North Pacific has a slightly earlier and broader peak (Fig. 2C). In the western North Pacific, the seasonal distribution of storms is the broadest due to the ingredients for tropical cyclones being favorable over a longer part of the year (Fig. 2B). The combination of activity in the North Pacific and North Atlantic dominates the global seasonal cycle (Fig. 2H). In the North Indian, tropical cyclone activity has a bimodal peak, with the first peak in May and the second peak in November (Fig. 2A). This bimodality is tied to the migration of the South Asian monsoon. Tropical cyclones form from tropical disturbances within the monsoon trough while it is over water and the ingredients are favorable. Once the monsoon trough moves over land in July and August, conditions become unfavorable for tropical cyclones to form. Tropical cyclone activity in the Southern Hemisphere oceans has a peak in January and February (Fig. 2E–G).

The canonical track of a tropical cyclone starts in the deep tropics, moves westward and poleward, reaches an apex, and then turns eastward and poleward (Fig. 1). There are many variations about this canonical track, depending on the concomitant synoptic state around a given tropical cyclone. Regardless, tropical cyclones are steered by two primary mechanisms. First, the environmental wind averaged through much of the troposphere right around a tropical cyclone is largely responsible for steering it. The most pertinent steering layer, though, depends on aspects of the tropical cyclone, like its size and depth (Galarneau and Davis, 2013). Second, due to meridional variations in the background planetary vorticity (i.e., the beta effect), the circulation of a tropical cyclone will create relative vorticity anomalies in its vicinity that impart a poleward and westward motion on the storm itself. This phenomenon

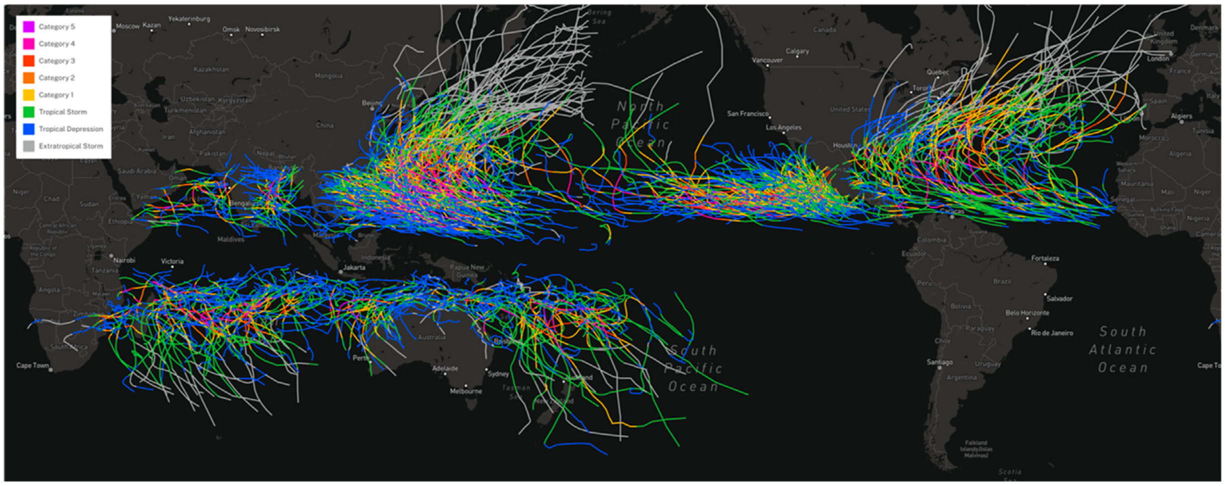

Fig. 1 Tropical cyclone tracks from 2013–2022. Color along track gives an indication of the intensity by tropical cyclone classification. Image generated from the NOAA Historical Hurricane Tracks webpage.

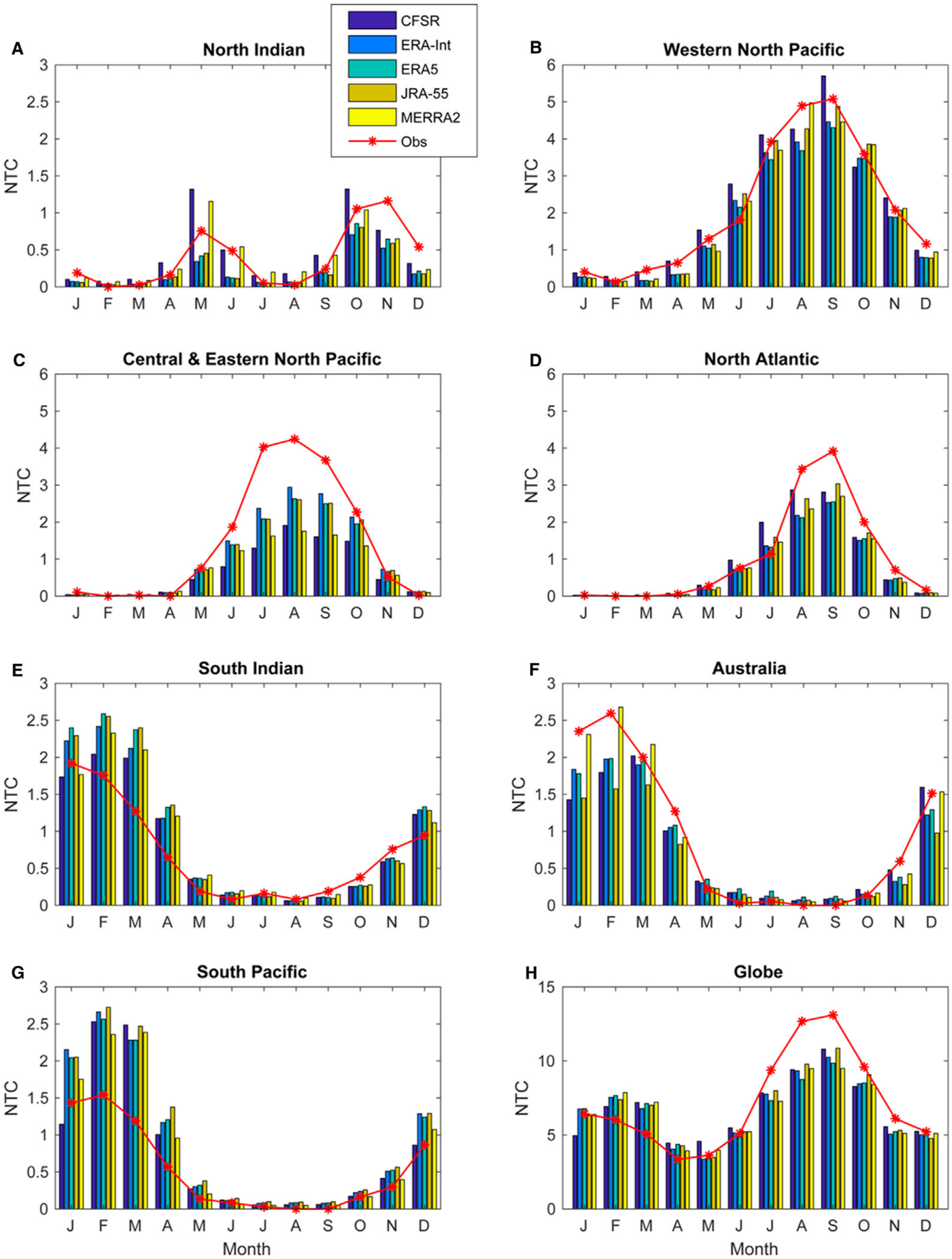

Fig. 2 Average number of tropical cyclones by month in each tropical cyclone ocean basin and globally. The red lines show the observed number, and the various colored bars indicate the number predicted by a tropical cyclogenesis index consisting of a combination of environmental factors, using different reanalysis model output. Figure reproduced from Sobel et al. (2021).

is called the beta drift. The combination of the steering environmental wind and the beta drift is responsible for a large proportion of a tropical cyclone's motion.

As tropical cyclones venture poleward, the environment generally becomes more inhospitable due to cooler sea surface temperatures and greater vertical wind shear. As a result, a number of tropical cyclones simply decay. Some tropical cyclones, however, undergo extratropical transition as they interact with baroclinic zones and can become powerful extratropical cyclones (gray lines in **Fig. 1**) with significant hazards (Evans et al., 2017). These events can also cause sensible weather impacts far downstream due to Rossby wave amplification.

Tropical Cyclogenesis

A fundamental question is what causes a tropical disturbance to develop into a tropical cyclone. The answer is not a straightforward one, involving complex interactions between the large-scale environment, synoptic-scale disturbances, and mesoscale processes that cooperatively lead to tropical cyclogenesis (Nam and Bell, 2021). The necessary ingredients described in the previous section are only part of the equation.

Precursor Disturbances and Patterns

Tropical cyclogenesis occurs via different types of tropical disturbances and synoptic patterns. These disturbance types and patterns differ between ocean basins, making up a tapestry of tropical cyclone formation pathways.

Easterly Waves

Easterly waves are troughs of low pressure that travel westward in tropical latitudes. Given the cyclonic vorticity associated with the trough axis, easterly waves typically have a wavelike undulation in the low-to-mid-level flow. Easterly waves containing sufficient moisture and deep convection are common precursors to tropical cyclogenesis in the North Atlantic and eastern North Pacific, and play a role in other basins. In the North Atlantic, approximately 60%–70% of tropical cyclones have their origins from African easterly waves (Russell et al., 2017).

The origin of African easterly waves is itself of interest. One hypothesized mechanism involves the unstable growth of Rossby waves along the midlevel African easterly jet, by both barotropic and baroclinic processes. A second hypothesized mechanism involves latent heating in mesoscale convective systems (Thorncroft et al., 2008). Local topographic influences in generating convection also are important.

African easterly waves that develop into tropical cyclones have a few contrasting characteristics versus those waves that do not develop. First, developing waves start out stronger and more amplified. Second, waves with more low-to-midlevel moisture are more likely to develop, as they provide a more favorable thermodynamic environment for sustaining convection. Third, in-phase coupling of mesoscale convective systems with the wave trough aid in development (Nuñez Ocasio et al., 2020).

One may view the development of easterly waves into tropical cyclones through the analogy of a baby marsupial developing in a pouch (Wang et al., 2010). The pouch represents a Lagrangian, recirculating area within a wave critical layer that can nurture a developing wave. Within this pouch, there is an aggregation of cyclonic vorticity and moisture, which is protected from disruption by environmental dry-air intrusion. As this aggregation proceeds, the convection, vortex, and warm core of a tropical cyclone can develop.

Monsoon Troughs

The monsoon trough is a seasonally migrating low-pressure zone within which tropical cyclones form (**Fig. 3**). The monsoon trough is rich in background cyclonic vorticity and moisture, so is a favorable area for genesis to occur in the North Indian and western North Pacific. In the western North Pacific, approximately 70%–80% of tropical cyclones form within the monsoon trough (Molinari and Vollaro, 2013).

Perturbations within the western North Pacific monsoon trough may give rise to tropical cyclones. Such perturbations may include wind surges from midlatitude high-pressure systems in both hemispheres, low-pressure systems along the Mei-yu front, trade-wind surges, and synoptic-scale wave trains. These perturbations concentrate cyclonic vorticity over a smaller area, enhance surface heat fluxes, and may lead to tropical cyclogenesis within the larger monsoon trough.

In addition to the seasonal monsoon trough, the monsoon circulation episodically features a large monsoon gyre with a diameter of about 2500 km and monsoon depressions that have a diameter of about 1000 km. These monsoon circulations differ from tropical cyclones in that they have a broader wind field and convective clusters organized into a band, rather than concentrated in a circular area. In a monsoon gyre, the convective clusters do occasionally spawn into one or more tropical cyclones, most commonly in the eastern half of the monsoon gyre (Wu et al., 2013). Monsoon depressions also occasionally develop sufficient convection near the circulation center, along with a more compact vortex, to become tropical cyclones.

Intertropical Convergence Zone Breakdown

Another large-scale pattern that can lead to tropical cyclone formation, sometimes multiple formation events, is breakdown of the intertropical convergence zone. The intertropical convergence zone is a quasi-linear, low-pressure zone where northeasterly and southeasterly trade winds converge and can be typically identified through a lengthy belt of convection in the tropics. While there is a strip of cyclonic vorticity associated with the intertropical convergence zone, it is typically dominated by shear vorticity and deformation, which are not conducive for tropical cyclogenesis.

Occasionally, convective heating within the intertropical convergence zone or synoptic-scale waves propagating into the tropics can initiate a breakdown of the strip of cyclonic vorticity via barotropic instability. The result is the roll up of the vorticity strip into coherent vortices, with greater curvature versus shear vorticity, and can subsequently lead to tropical cyclogenesis.

Tropical Modes of Variability

The tropical circulation also features various wave modes and intraseasonal oscillations that directly affect tropical cyclogenesis.

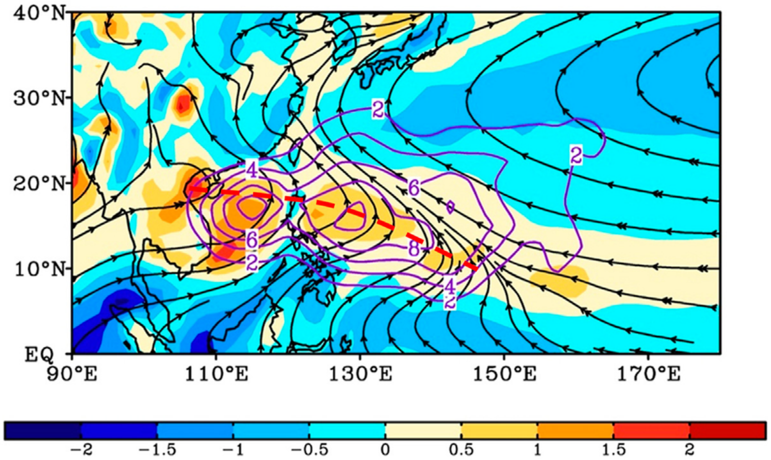

Fig. 3 Structure of the monsoon trough in the western North Pacific. July—September mean 850-hPa flow (streamlines), 850-hPa relative vorticity (shaded; $\times 10^{-5}$ s^{-1}), and frequency of tropical cyclogenesis (contours) in the western North Pacific. The axis of the monsoon trough is given by the dashed, red line. Figure adapted from Feng and Wu (2022).

Equatorial Rossby waves feature an off-equatorial, low-pressure circulation. Similar to tropical cyclogenesis occurring in monsoon gyres, a convective cluster within an equatorial Rossby wave can spawn into a tropical cyclone, and is favored on the more convectively active eastern side of the low-pressure circulation. Since equatorial Rossby waves have symmetric areas of low pressure about the Equator, twin tropical cyclones can sometimes form in each hemisphere (**Fig. 4**). In the Southern Hemisphere, equatorial Rossby waves are the most important precursor disturbance for tropical cyclones (Schreck et al., 2012).

Kelvin waves also influence tropical cyclogenesis, increasing the probability of genesis within a few days after the convectively active part of a Kelvin wave intersects with a tropical disturbance, like an easterly wave. A Kelvin wave may enhance low-level convergence and lift, thereby invigorating convection within a tropical disturbance. Additionally, in the wake of the convectively active part of a Kelvin wave, westerly wind anomalies may increase cyclonic vorticity, and midlevel moisture may be enhanced, giving a window of favorable conditions for genesis (Lawton et al., 2022).

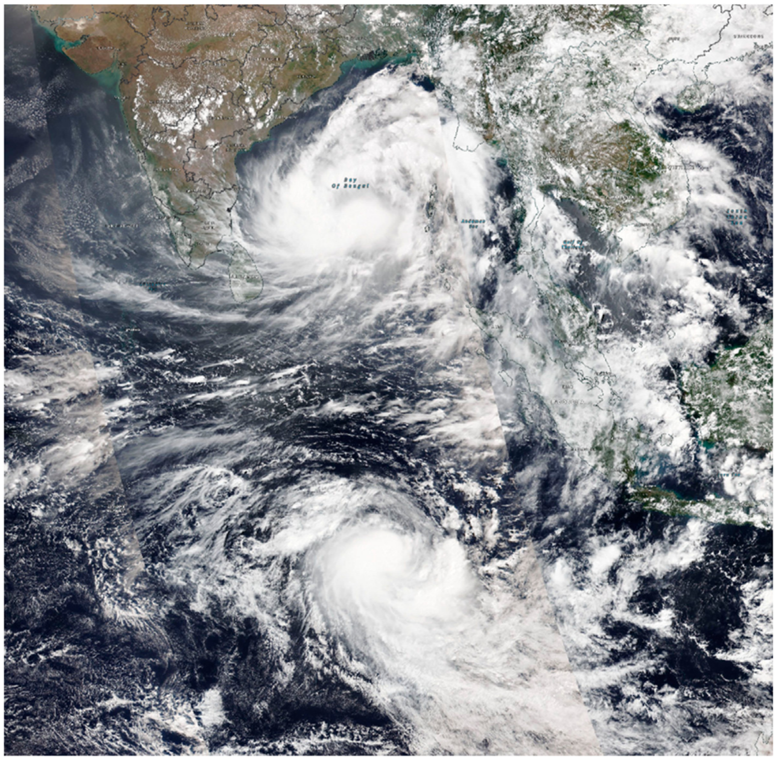

Fig. 4 Visible satellite image from the Visible Infrared Imaging Radiometer Suite of twin tropical cyclones in the Indian Ocean on May 8, 2022. Cyclone Asani is in the Bay of Bengal, and Cyclone Karim is in the southern Indian Ocean. Image generated from the NASA Worldview webpage.

The Madden-Julian Oscillation is a major source of intraseasonal variability in tropical cyclone activity. This oscillation propagates eastward with a period of approximately 30–60 days, affecting rainfall and circulation variability in the tropics. Different phases of the Madden-Julian Oscillation are associated with enhanced or suppressed tropical cyclone activity in each basin. During enhanced activity phases, there is increased large-scale low-level vorticity, moisture, ascent, and upper-level divergence. Additionally, there is less vertical wind shear.

Baroclinic Influences

The canonical picture of tropical cyclogenesis is one that occurs in the deep tropics in a relatively homogeneous temperature environment, i.e., in a nonbaroclinic environment. About 30% of tropical cyclogenesis events worldwide, however, are baroclinically influenced (McTaggart-Cowan et al., 2013). Examples of baroclinically influenced tropical cyclogenesis events are those that occur in association with an initial upper-tropospheric trough or cutoff low, which have a cold-core structure.

Tropical transition occurs when a cold-core cyclone acquires a warm core and develops the characteristics of a tropical cyclone. A number of such events occur over sea surface temperatures $<26°C$, seemingly bucking one of the necessary ingredients for tropical cyclones. The relatively cold temperatures aloft, however, compensate by steepening lapse rates and increasing instability, which allows for the development of deep convection. The convective latent heating profile results in a redistribution of potential vorticity, increasing cyclonic low-level vorticity and decreasing cyclonic upper-level vorticity, reducing the vertical wind shear, and can result in the development of a more compact low-level vortex (Bentley et al., 2016). This process is gradual, first leading to subtropical cyclone formation, where the system contains characteristics of both tropical and extratropical cyclones. If enough tropical cyclone characteristics are gained and extratropical characteristics are lost, then a tropical cyclone results.

Mesoscale Considerations

Regardless of exact precursor disturbance or large-scale patterns leading to tropical cyclogenesis, are there mesoscale pieces that are common to the tropical cyclogenesis puzzle? Many of these pieces are still the subject of active research and ongoing refinements in our understanding.

Moistening of Inner Region

Both numerical modelling studies and composite analyses of dropsonde observations in developing tropical disturbances have shown that deep-column moistening is an important step for genesis to occur (Zawislak and Zipser, 2014). Specifically, relative humidity exceeding 80% within 100–150 km radius of the center precedes genesis. This nearly saturated mesoscale region is important for sustaining convection and mitigating evaporative downdrafts that could disrupt the spinup of the low-level vortex, and allows for more efficient production of kinetic energy. Under such circumstances, a compact (meso-beta-scale) vortex forms and persists with a more sharply peaked tangential wind profile and smaller radius of maximum wind than the precursor disturbance circulation (Nolan, 2007).

The role of moistening in spawning a tropical cyclone vortex may be closely related to a moisture-vortex instability, which has been used to explore the growth of monsoon depressions and easterly waves (Adames, 2021). In a situation with moisture-vortex instability, positive moisture anomalies in phase with the vortex enhances deep convection, which then enhances cyclonic vorticity via stretching. Viewed from a thermodynamic perspective, advective fluxes of equivalent potential temperature (θ_e) import and concentrate θ_e into an inner region of the vortex. Meanwhile, the advective fluxes export θ_e out of the surrounding region, steepening the θ_e gradient. Subsequently, a wind-induced surface heat exchange feedback may be activated to continue amplifying the vortex (Tang, 2017; see **Wind-Induced Surface Heat Exchange**).

Mesoscale Convective Vortices

A topic of continued research is the role of mesoscale convective vortices that are typically observed in tropical disturbances. Mesoscale convective vortices form in the later stages of the lifecycle of mesoscale convective systems, one or more of which may exist in a tropical disturbance in various stages of growth and decay (Houze, 2010). Mesoscale convective vortices have a maximum amplitude at midlevels of the troposphere due to midlevel convergence within stratiform precipitation. The "top-down" hypothesis for tropical cyclogenesis suggests that the midlevel vortex builds downward and directly strengthens the low-level vortex. Although, there have been counterarguments against this hypothesis based on the impermeability principal of potential vorticity. Mergers of multiple mesoscale convective vortices into a larger, stronger midlevel vortex may also result in a stronger projection of the circulation to the surface in a moist-neutral stability setting.

Alternatively, a mesoscale convective vortex may indirectly aid in the development of the low-level vortex by altering the local thermodynamic state. By thermal wind balance, a cold anomaly must exist below the mesoscale convective vortex, and a warm anomaly exists above it. The increased stability results in convective vertical mass flux profiles that are bottom heavy, having a maximum vertical mass flux lower in the troposphere. Then, by conservation of mass, there is greater horizontal convergence at low levels, which more quickly spins up the low-level vortex (Raymond et al., 2014).

The mesoscale convective vortex may assist in tropical cyclogenesis in other ways. First, it may increase the inertial stability and decrease the local Rossby radius, resulting in more efficient conversion of convective heating to the kinetic energy of the circulation. Second, it may better contain midlevel moisture and deflect environmental dry-air intrusions. Some studies argue that the mesoscale convective vortex, while potentially helpful, might not be imperative for tropical cyclogenesis to occur. Instead, deep convection is the critical element.

Vortical Hot Towers

In contrast to the "top-down" hypothesis, the "bottom-up" hypothesis suggests that the vortex builds upward due to deep, rotating convection called vortical hot towers. Deep convection generates and amplifies cyclonic vorticity anomalies via tilting and stretching. Individual cyclonic vorticity anomalies then merge and axisymmetrize into the vorticity core of a tropical cyclone (Hendricks et al., 2004). Axisymmetrization occurs when vorticity perturbations are horizontally sheared and filamented by the background vortex flow. The energy of these perturbations is transferred upscale to the mean vortex flow, strengthening it. Inward transport of air, in association with the collective response to these vortical hot towers, also spins up the circulation above the boundary layer due to conservation of angular momentum.

In addition to vortical hot towers, cumulus congestus also plays a role. While congestus does not have as large of a vertical mass flux nor extends as deep as vortical hot towers, the population of congestus collectively acts to spinup the low-level vortex due to their bottom-heavy vertical mass flux profile. Additionally, detrainment of congestus moistens the low-to-middle troposphere, conditioning the system for deeper convection (Wang, 2014).

Role of Radiation

A research frontier is the role of cloud-radiative feedbacks on tropical cyclogenesis. Cloud-radiative feedbacks involve the interaction of shortwave and longwave radiation with cloud hydrometeors. In tropical disturbances and cyclones, focus has centered on the longwave interaction with ice clouds as a relevant process for tropical cyclone development. At the top of the cirrus canopy, there is strong radiative cooling, which can affect the vertical stability. Below the cirrus canopy, longwave radiative heating occurs within optically thick clouds within a tropical disturbance, while such longwave heating is absent in the clear region around a tropical disturbance. This differential longwave heating results in greater ascent in the cloudy regions of a disturbance, promoting moistening and continued convection. With this greater ascent, there is greater inflow at low-to-mid levels that can help spinup the circulation. As a result, cloud-radiative forcing, while not crucial for tropical cyclogenesis, may possibly accelerate it and be as important as surface fluxes in the pre-genesis phase (Wing, 2022).

Additionally, the diurnal cycle of radiation has been hypothesized to be a factor affecting the timing of genesis. During the day, shortwave heating of the troposphere increases the stability and decreases the relative humidity. At night, longwave cooling decreases the stability and increases the relative humidity, invigorating convection. In some idealized simulations, the longwave cooling at night is important for accelerating genesis, although it is unknown how important this effect is in reality.

Tropical Cyclone Structure

After a tropical cyclone undergoes genesis, presuming it can further mature, it acquires well-defined features in its wind field and cloud structure. Some of the main features of tropical cyclones, focusing on the mature stage, are now described.

Primary Circulation

The primary circulation of a tropical cyclone is the swirling, tangential winds around the center. Above the boundary layer, the primary circulation is largely in gradient wind balance, which is a balance between the Coriolis force, pressure gradient force, and centrifugal force. Within the boundary layer, frictional effects cause deviations of the primary circulation from gradient wind balance. In fact, the strongest tangential winds within the boundary layer are typically supergradient.

The radial profile of the tangential winds may be approximated as a modified Rankine vortex. In a modified Rankine vortex, the tangential wind increases linearly with radius from the center to the radius of maximum wind. This profile is equivalent to solid body rotation, which is characterized by a patch of uniform relative vorticity. Outside the radius of maximum wind, the tangential wind of a modified Rankine vortex decays as $r^{-\alpha}$, where r is the radius and α is the decay scale. For a pure Rankine vortex, $\alpha = 1$, but tropical cyclones have decay scales around 0.2−0.6 (Mallen et al., 2005), with more intense tropical cyclones usually having a larger decay scale. The relative vorticity falls off quickly right around the radius of maximum wind and then more slowly as the radius increases further. This vorticity structure is called a vorticity skirt.

The tangential winds are usually strongest near the top of the boundary layer and decrease with height above the boundary layer. By thermal wind balance, the anticyclonic vertical shear of the primary circulation must coincide with decreasing temperature with radius, accordant with the characteristic warm core of a tropical cyclone. At upper levels of a tropical cyclone, outside the core, the primary circulation is anticyclonic and often covers a large area.

Secondary Circulation

The secondary circulation consists of the radial and vertical winds of a tropical cyclone, or the in-up-and-out component of the circulation. The archetypal secondary circulation consists of radial inflow at low levels, rising motion in rainbands and the eyewall, and then radial outflow at upper levels around the tropopause. Giving the rising motion occurs where it is relatively warm and moist, the secondary circulation is a thermally direct circulation and one that generates mechanical energy.

The radial inflow at low levels is a response to both friction and latent heating in convection. This inflow transports moisture and angular momentum inward. The height of the inflow layer decreases closer to the center. The radial wind is largest at and just outside the radius of maximum wind and at about 100 m height (Zhang et al., 2011).

The rising branch of the secondary circulation occurs in rainbands and the eyewall. As buoyant parcels rise in rainbands and the eyewall, tremendous amounts of condensation, freezing, and latent heating occur. Most of the rising mass in a tropical cyclone occurs in the eyewall.

Air exiting the eyewall then spreads out as a large area of cirrus outflow. As the air spreads out, it conserves its angular momentum, and the air becomes more anticyclonic. The outflow, as it interacts with other upper-level synoptic systems, often develops asymmetries and one or more jets. Outflow interactions with midlatitude Rossby waveguides can produce downstream effects far from a tropical cyclone.

Rainbands

The cloud structure of a mature tropical cyclone consists of multiple types of rainbands. Their characteristics are governed by the amount of buoyancy and vortex flow in which the rainbands are embedded.

The outer region of a tropical cyclone consists of arcs of convective rainbands, resembling curved squall lines. These distant rainbands are largely convective in nature, feeding off buoyancy in the outer core and producing strong downdrafts and cold pools that spread radially outward. Lift of buoyant air along the leading edge of the propagating cold pools initiates new convection in these rainbands. Additionally, these rainbands tilt radially inward with height.

The principal rainband lies inward of the distant rainbands and has a distinct structure. The principal rainband consists of convective cells upwind that transition to a predominately stratiform precipitation region downwind. The stratiform region may connect with the eyewall. Radial inflow of relatively high-θ_e air rises in convective updrafts in the principal rainband. As the air rises, it also turns radially outward, giving the principal rainband an outward tilt with height. The principal rainband only has a moderate depth (about 8 km), in contrast to deeper eyewall convection. Additionally, there is descending motion on the radially inward side of the principal rainband and descending radial inflow through the stratiform region of the rainband. The combination of airflows in the rainband help strengthen and maintain a midlevel, tangential wind maximum along the axis of the rainband through tilting and stretching of vorticity (Didlake and Houze, 2013; Hence and Houze, 2008).

The rainband is asymmetric, encompassing one side of the circulation. The location of the principal rainband lies at the confluent streamlines that form a boundary between the strongly rotational vortex low-level flow and the background environmental low-level flow. Low-level convergence and lift occur along this boundary. This boundary also separates relatively high-θ_e air in the inner core from relatively low-θ_e air in the environment.

Environmental vertical wind shear is an organizing agent for the principal rainband. The moist envelope of the tropical cyclone is distorted due to the interaction of the vortex flow with the environmental, storm-relative flow. Additionally, vertical wind shear tilts the vortex, resulting in asymmetric forcing for ascent. These two processes work together to form the principal rainband, and the radial shear of the primary circulation shapes the rainband (Riemer, 2016).

Inward of the principal rainband are secondary rainbands, which are smaller, transient rainbands that share similarities with the principal rainband. The secondary rainbands also tilt outward with height, transition from more convective to more stratiform going from the upwind to downwind portion of the rainband, and have airflow structures like the principal rainband.

A third class of rainbands are inner rainbands, which originate near the outer edge of the eyewall and propagate radially outward. These rainbands may be associated with vortex Rossby waves. Vortex Rossby waves exist on the large potential vorticity gradient of the vortex, analogous to midlatitude Rossby waves that exist on the potential vorticity gradient of the polar vortex. The phase speed of these waves is slower than the background tangential flow and is radially outward, matching the observed radar characteristics of inner rainbands (Corbosiero et al., 2006). Alternatively, these inner rainbands may simply be advected and shaped by the strong deformation of the vortex flow just outside of the radius of maximum wind (Moon and Nolan, 2015).

Eyewall

The eyewall is a special region in a tropical cyclone due to both the hazards and physics that define the eyewall. The eyewall can be viewed as a circular front. The radial inflow in the boundary layer and surface heat fluxes increase the radial gradient of θ_e relative to the radial gradient of angular momentum. This type of frontogenesis is a unique feature of tropical cyclone eyewalls. As this air erupts out of the boundary layer, the air follows a slantwise path along constant angular momentum and θ_e surfaces that flare outward to the upper troposphere and lower stratosphere. This process manifests visually as a breath-taking eyewall stadium effect (**Fig. 5**).

While the mean, idealized structure of the eyewall is one of symmetry and slantwise-rising motion, the instantaneous structure of the eyewall is highly turbulent with a rich spectrum of eddies. The most prominent eddies are mesovortices that form on the eyewall-eye interface and may sometimes be seen in visible satellite imagery within the eye of intense hurricanes (**Fig. 5**). Mesovortices form due to barotropic instability of an annulus of large potential vorticity, where interacting vortex Rossby waves constructively amplify. This annulus, having a reversal of the gradient of potential vorticity with respect to radius, satisfies the necessary

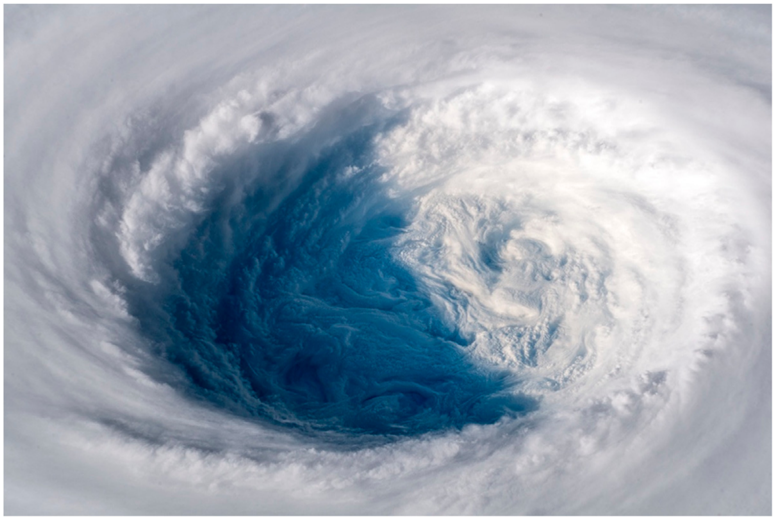

Fig. 5 Eyewall stadium effect and mesovortices seen within the eye of Typhoon Trami (2018). Image credit: ESA/A. Gerst.

condition for barotropic instability. Once barotropic breakdown of the annulus of potential vorticity occurs, higher potential vorticity and momentum are mixed into the eye.

Smaller scale eddies along the eyewall-eye interface include misovortices and tornado-scale vortices. These more transient vortices contain very strong low-level updrafts, forced by intense local rotation aloft and by enhanced buoyancy entrained from the eye. These vortices also contain the strongest wind gusts within tropical cyclones, so can be responsible for localized, extreme wind damage in intense, landfalling tropical cyclones.

Eyewall Replacement Cycles

An eyewall in a tropical cyclone is replaced by an outer, secondary eyewall during an eyewall replacement cycle, which is a common occurrence in mature tropical cyclones. Secondary eyewall formation is preceded by an expansion of the tangential wind field. A secondary eyewall forms as rainbands outside the primary eyewall coalesce into a ring. A secondary wind maximum develops too. Subsequently, the primary eyewall begins to weaken and the secondary eyewall contracts inward, eventually replacing the primary eyewall. The whole eyewall replacement cycle takes on average about 36 h to complete (Sitkowski et al., 2011), but the timescale is influenced by the vortex structure and other external factors. The end result of an eyewall replacement cycle is typically a larger radius of maximum wind and overall wind field. A tropical cyclone may have multiple eyewall replacement cycles over its lifetime.

The formation mechanisms of secondary eyewalls are a subject of active research, and several hypotheses have been put forth. The first hypothesis is that secondary eyewalls form due to the outward stagnation of vortex Rossby waves at some critical radius where the radial group velocity vanishes. At this critical radius, vortex Rossby waves increase the tangential flow through wave-mean flow interactions, and this increased tangential flow can enhance surface heat fluxes and convection at this critical radius. The second hypothesis involves the axisymmetrization of convectively generated vorticity anomalies in the vortex's vorticity skirt. As this axisymmetrization occurs, a low-level, tangential jet materializes and, like the first hypothesis, can enhance surface heat fluxes and convection. The third hypothesis involves unbalanced dynamics within the boundary layer. As the tangential wind field expands, radial inflow increases within the boundary layer and results in the formation of supergradient winds. As air parcels in the supergradient wind area experience an outward acceleration, radial convergence and ascent result, which can form a ring of convection. The fourth hypothesis involves descending inflow in the stratiform region of the principal rainband and the associated surface cold pool. New convective updrafts are generated on the inner and downwind edge of the surface cold pool as it collides with high-θ_e air, leading to an enhancement of vorticity and the development of the secondary eyewall (Yu et al., 2022). These hypotheses, however, are not mutually exclusive.

Eye

The eye is one of the most recognizable features of a tropical cyclone, containing relatively light winds near its center, the lowest pressure in the storm, a lack of deep clouds, and the warmest temperature anomalies aloft. The lowest levels of the eye are warm and moist, and capped with a layer of low clouds below a strong temperature inversion. The warm temperature anomalies above the inversion are the result of subsidence and adiabatic warming.

There are several mechanisms contributing to subsidence in the eye. First, convective heating in the eyewall forces compensating subsidence adjacent to it in the eye. Second, the transfer of momentum into the eye, due to mesovortices and other eddies, results in an ageostrophic response with subsidence within the eye to restore gradient wind balance. Third, supergradient winds erupting out of the boundary layer are accelerated outwards, creating a layer of divergence, which must be compensated with subsidence in the eye to conserve mass.

Diurnal Pulses

A recent topic of investigation has been a cyclical pulsing in the tropical cyclone cloud structure that is tied to the diurnal cycle. Around sunset, the pulse begins as a region of colder cloud tops in the tropical cyclone inner core. The colder cloud tops move radially outward in an annulus overnight, propagating at a speed of about 8–14 m s^{-1}. By the following afternoon, the cooling pulse may extend 500 km from the center (Dunion et al., 2014). These cooling pulses are common in tropical cyclones, occurring about 72% of the time in the North Atlantic, and are more common for more intense tropical cyclones (Ditchek et al., 2019).

The mechanism generating these cooling pulses is an area of active research. While these cooling pulses are observed prominently in the cirrus canopy in infrared satellite imagery, they are often associated with deep convection and lightning activity, implying these pulses are not a feature propagating exclusively within the outflow layer of a tropical cyclone. Although cooling pulses are more common, warming pulses have also been observed with the same timing and propagation characteristics, just with opposite sign cloud-top temperature anomalies. The characteristics of these pulses suggest that they are potentially an inertia-gravity wave feature excited by diurnal variations in radiative and convective heating (Navarro and Hakim, 2016).

Intensity Change

While the mechanisms steering tropical cyclones are well established and rely primarily on large-scale features around tropical cyclones, the mechanisms governing tropical cyclone intensity change occur on a myriad of scales and are inherently linked to the evolution of the tropical cyclone structure. As a result, tropical cyclone intensity change is a much more complex problem. It is common to divide factors affecting intensity change into external and internal influences, but it is important to recognize that these factors do not affect intensity independently of one another.

External Influences

Vertical Wind Shear and Ventilation

An important environmental control on tropical cyclone intensity is environmental vertical wind shear. The vertical wind shear, for tropical cyclone applications, is commonly measured as the bulk environmental wind difference between 850 and 200 hPa. Increasing magnitudes of environmental vertical wind shear are negatively correlated with intensity change. Moderate shear (4.5–11 m s^{-1}; Rios-Berrios and Torn, 2017) regimes feature a greater spread of intensity change outcomes, so is a recent research focus.

Vertical wind shear has two predominate effects on tropical cyclone structure. First, it produces asymmetries in the precipitation structure, with the strongest convection displaced to the downshear side of the storm. In strong wind shear environments, sometimes the low-level circulation of a tropical cyclone can become completely exposed and devoid of deep convection. Second, vertical wind shear tilts the vertical axis of the vortex in the direction of the shear. The tilt direction then precesses cyclonically and is most commonly pointed in the downshear-left direction.

As the tropical cyclone vortex tilts, it opens pathways for environmental, low-θ_e air to intrude into the circulation, which is called ventilation. The low-θ_e air can act as anti-fuel, counteracting surface heat fluxes that power a tropical cyclone (Tang and Emanuel, 2010). There are two classes of ventilation pathways: radial and downdraft ventilation. A radial ventilation pathway may develop upshear at mid-to-upper levels as the displaced mid-to-upper-level vortex advects dry, low-θ_e air over the low-level vortex. If entrained into the eyewall or other deep convection near the center, the low-θ_e air dilutes the buoyancy of rising air and can produce strong, evaporatively driven downdrafts. A second combined radial and downdraft ventilation pathway may develop at low-to-mid levels in the descending inflow region of the principal rainband, which is organized by the vertical wind shear (see **Rainbands**). As precipitation evaporates into the low-θ_e air, it cools and becomes negatively buoyant, depositing cold pools into the boundary layer. A combination of these pathways may operate simultaneously, reducing the upward mass flux in the tropical cyclone inner core and weakening the tropical cyclone (Alland et al., 2021).

The properties of environmental air at the source of these ventilation pathways matters, particularly the moisture content. Drier, lower-θ_e air in the near-environment has more potential to be disruptive to a tropical cyclone, particularly if that drier air is located at midlevels on the upshear side of a tropical cyclone, where it can be more readily entrained into the tropical cyclone circulation via a radial ventilation pathway.

Other factors may also modulate the effects of vertical wind shear and ventilation. Large surface heat fluxes, especially in the left-of-shear region, may effectively restore the low-θ_e air in downdraft-generated cold pools, mitigating their downstream effect on convection. Additionally, stronger, deeper, and larger tropical cyclones are generally more resilient to the effects of vertical wind shear and ventilation. Under such circumstances, a tropical cyclone that is initially tilted by vertical wind shear may realign as the tilt direction precesses towards the upshear direction and the tilt magnitude subsequently decreases. As realignment occurs, a tropical cyclone can more readily intensify. Whether or not realignment occurs, especially in moderate shear regimes, appears to be a key determination in whether a tropical cyclone stays weak or reintensifies.

Ocean Mixing

Interactions between a tropical cyclone's winds with the upper ocean is of fundamental importance to the energy supplied to tropical cyclones. The energy rate supplied is a function of the wind speed and the thermodynamic disequilibrium between air saturated at the sea surface temperature and air a short distance above the sea surface. The warmer the sea surface temperature, the larger the thermodynamic disequilibrium and potential energy rate supply, all other variables held constant.

Compared to undisturbed conditions, the winds of a tropical cyclone will induce turbulent mixing in the oceanic mixed layer, deepening it, which results in cooling of sea surface temperatures by up to several degrees Celsius. The cooling is strongest to the right of the storm track. This process naturally puts a brake on the thermodynamic disequilibrium and surface heat fluxes, particularly for storms that are moving more slowly.

Ocean variability can modulate the thermodynamic disequilibrium. Tropical cyclones traversing over regions with large upper-ocean heat content, having warm waters extending to a greater depth, have more potential to intensify if other conditions are favorable. Examples of ocean regions with large upper-ocean heat content include oceanic warm pools, western boundary currents (e.g., The Gulf Stream), and warm ocean eddies. Additionally, salinity stratification may be important locally; for example, where a layer of fresher water overlays saltier water, representing a barrier to vertical mixing and sea surface temperature cooling (Rudzin et al., 2020).

Trough Interactions

Interactions of tropical cyclones with upper-level troughs, originating from the midlatitudes, can weaken or strengthen a tropical cyclone. Upper-level divergence ahead of a trough can increase lift, especially in the equatorward-entrance region of jet streaks. Additionally, upper-level troughs may induce an eddy flux convergence of angular momentum, providing a source of angular momentum in the tropical cyclone outflow layer. The response to this source of angular momentum is an enhanced secondary circulation. Both effects can invigorate convection in the inner core and increase the intensity. On the flip side, upper-level troughs are a source of vertical wind shear and drier air, so can ventilate tropical cyclones and weaken them. In the balance, troughs are generally more harmful to tropical cyclone intensity because of the effects of ventilation (Peirano et al., 2016).

The configuration of a trough around a tropical cyclone may dictate whether a tropical cyclone can intensify in a trough's presence. Longer wavelength troughs and troughs closer to a tropical cyclone are less favorable for intensification, because such troughs tend to cause greater ventilation. A situation in which a trough fractures, producing a cutoff low equatorward and westward of a tropical cyclone, is a configuration that is more favorable for intensification, compared to troughs located poleward of a tropical cyclone (Fischer et al., 2019).

Internal Influences

Wind-Induced Surface Heat Exchange

Wind-induced surface heat exchange is a positive feedback mechanism between surface heat fluxes and the winds of a tropical cyclone. In this positive feedback cycle, surface heat fluxes increase the radial gradient of θ_e in the boundary layer, which is communicated upwards through convection in the slantwise, saturated eyewall. This increase in the radial gradient of (saturation) θ_e results in an increase in the tangential winds at the top of the boundary layer, through thermal wind balance, and to the surface, through turbulent mixing. As a result, surface heat fluxes increase further, increasing the radial gradient of θ_e and the intensity.

Tropical cyclone energetics may be likened to a Carnot heat engine consisting of the secondary circulation divided into four legs (**Fig. 6**). The energy input occurs along the inflow, isothermal leg, where the moist entropy or θ_e of air parcels increases due to surface heat fluxes. In the ascending, moist-adiabatic leg, large amounts of latent energy are released in the eyewall. The energy output occurs along the outflow, isothermal leg, where radiative cooling balances subsidence warning. A descending, adiabatic leg closes the cycle. This simple model for tropical cyclone energetics is a useful heuristic for intensity.

The work done by the Carnot heat engine is the area encompassed by the cycle in temperature-entropy space (**Fig. 6**). This work powers the winds of a tropical cyclone. The greater the difference in moist entropy and/or temperature across this cycle, the greater the work that is performed. From this perspective, wind-induced surface heat exchange increases the difference in moist entropy. Additionally, the difference in temperature increases as convection reaches greater heights, which increases the thermodynamic efficiency.

The intensification of a tropical cyclone cannot continue indefinitely, however, as a steady state is reached called the potential intensity, a theoretical maximum tangential wind speed that a tropical cyclone can obtain in a given thermodynamic environment under the assumptions of axisymmetry, slantwise neutrality, and gradient wind balance. At this steady state, the mechanical energy produced balances that lost due to frictional dissipation. The potential intensity is a useful metric for estimating the intensity tropical cyclones can obtain under ideal environmental conditions. In some instances, tropical cyclones can exceed their potential intensity due to supergradient winds in the boundary layer.

It has been argued that the wind-induced surface heat exchange hypothesis for tropical cyclone intensification is not the crucial mechanism (Montgomery et al., 2015). In this argument, surface heat fluxes are necessary, but only to provide enough buoyancy for vortical hot towers to intensify the circulation through vorticity stretching. This intensification hypothesis has commonality with the "bottom-up" mechanism for tropical cyclogenesis (see **Vortical Hot Towers**). At the system scale, the radial inflow associated with these vortical hot towers results in the advection of higher angular momentum air inward, intensifying the circulation above the

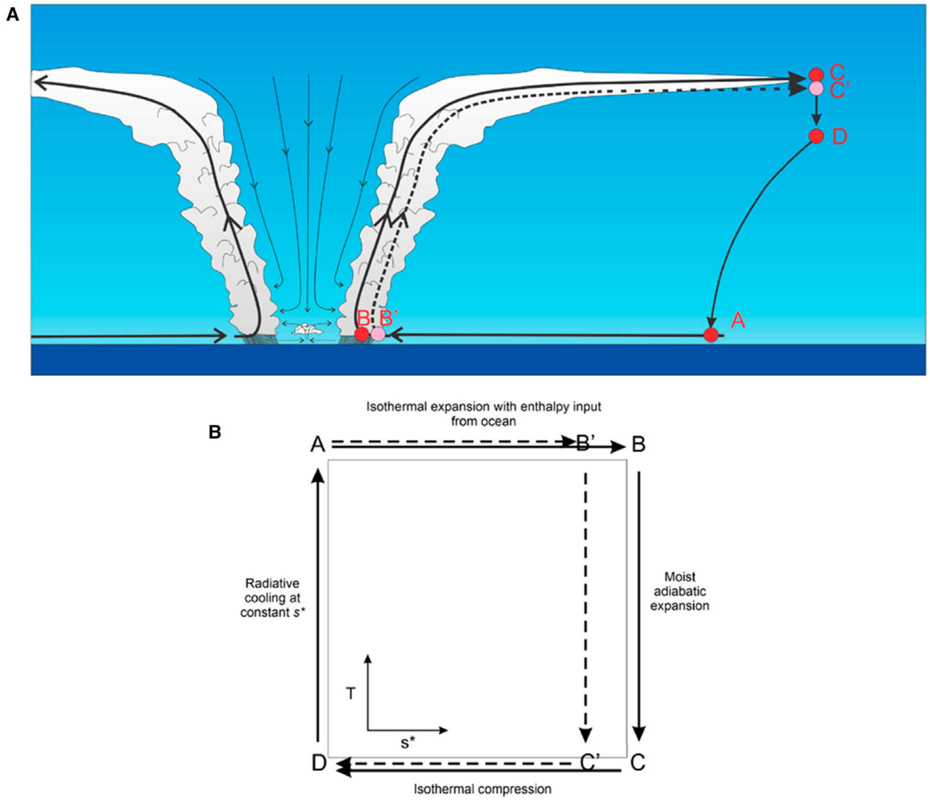

Fig. 6 (A) Axisymmetric secondary circulation of a tropical cyclone divided into four legs (A→B, B→C, C→D, and D→A), representing the legs of the Carnot cycle. (B) The legs represented in entropy-temperature space. Figure reproduced from Emanuel et al. (2023).

boundary layer. The boundary-layer inflow response simultaneously intensifies the circulation there, as the local positive tendency gained due to inward advection of higher angular momentum air exceeds the local negative tendency due to frictional torques.

Location of Convection

From an axisymmetric, balanced-dynamics point of view, the Sawyer-Eliassen equations may be used to diagnose the circulation response to a convective heat source. The response is governed by the background inertial and static stabilities of the vortex, along with the shape and location of the convective heat source. A convective heat source placed at, or slightly inside, the radius of maximum wind results in both a contraction of the radius of maximum wind and an increase in the intensity. Additionally, since the inertial stability is largest in the inner core, convective heating located there is more efficient at increasing the kinetic energy. In contrast, a convective heat source placed far outside the radius of maximum wind increases the tangential winds in the outer regions of a tropical cyclone, but not the intensity of the storm. This diagnostic framework gives a theoretical basis to assess how the location of convection is related to intensifying versus non-intensifying storms.

The distribution of the location of convective bursts, relative to the radius of maximum wind, is different between intensifying and steady-state tropical cyclones of hurricane strength. The majority of convective bursts for intensifying tropical cyclones are located within the radius of maximum wind. This result was shown by both Doppler radar data aboard reconnaissance aircraft (Rogers et al., 2013) and lightning data (Stevenson et al., 2018). In contrast, steady-state tropical cyclones have convective bursts that are shifted radially outward, occurring more outside the radius of maximum wind. These results are consistent with theoretical expectations from the Sawyer-Eliassen framework.

Boundary-Layer Effects

The location at which convection occurs is strongly tied to the boundary-layer structure. Compared to non-intensifying storms, observations show that intensifying storms have a deeper layer of inflow outside the radius of maximum wind. The maximum inflow magnitude is also located just outside of the radius of maximum wind. Both of these boundary-layer inflow characteristics promote greater radial convergence near and within the radius of maximum wind, consistent with there being more convective burst activity there, and supergradient winds. Narrower vortices, having a faster decay of the tangential wind outward of the radius of maximum wind, tend to also have these boundary-layer characteristics, so are more likely to intensify compared to broader vortices. In contrast, non-intensifying and broader tropical cyclones have boundary-layer inflow characteristics that promote greater radial convergence, ascent, and convective activity outside of the radius of maximum wind (Zhang et al., 2023).

At finer scales, the tropical cyclone boundary layer is filled with turbulent eddies and sea spray. Roll vortices are one such type of coherent turbulent structure that is ubiquitous in the boundary layer. They play a role in the vertical transport and mixing of momentum and heat through the boundary layer. Additionally, the effects of sea spray on exchanges of momentum and heat within the boundary layer are uncertain and an area of active research.

Rapid Intensification

Rapid intensification is defined as the top 5% of the observed distribution of overwater intensity changes over a 24-h period. This corresponds to about a 15−20 m s^{-1} increase in intensity over 24 hours, depending on the ocean basin. Better understanding and predictions of rapid intensification have been a recent, major goal due to the drastically escalating danger rapidly intensifying tropical cyclones can pose to coastal communities.

Rapidly intensifying tropical cyclones more frequently occur under the following set of conditions. First, the vertical wind shear and ventilation are small. Second, the upper-ocean heat content is large. Third, there is enhanced divergence at upper levels of the storm. Fourth, the difference between the potential intensity and actual intensity is large. The majority of rapid intensification events begin at tropical storm strength. There are always exceptions to these rules, however. For example, some tropical cyclones in moderate-to-strong shear can undergo rapid intensification due to an asymmetric intensification mechanism involving the restructuring of the vortex (Judt et al., 2023). Even when conditions are favorable, the convection and internal structure of a tropical cyclone must cooperatively interact for rapid intensification to occur.

Recent research advances have given clues as to what aspects of the tropical cyclone's convection and structure may favor rapid intensification. Vertical alignment and precipitation symmetry appear to be important. A symmetric latent heating structure, particularly if concentrated at and within the radius of maximum wind, is conducive for rapid rates of intensification. Additionally, larger surface heat fluxes in the left-of-shear and upshear regions of a tropical cyclone counteract downdraft ventilation and build boundary-layer θ_e (Nguyen et al., 2019). As a result, convection upshear is better supported, resulting in greater precipitation symmetry in sheared environments. Another relevant aspect for rapid intensification is the strength and depth of the vortex aloft. Tropical cyclones with deeper, stronger vortices aloft, and that are more vertically aligned, are more likely to subsequently rapidly intensify (Richardson et al., 2022).

Rapid Weakening

At the opposite end of the intensity change distribution, overwater rapid weakening corresponds to about a 15−20 m s^{-1} decrease in intensity over 24 hours (Wood and Ritchie, 2015). Rapid weakening coincides with tropical cyclones entering environments of larger shear and drier air, or in other words, tropical cyclones experiencing marked increases in ventilation. A tropical cyclone may be especially vulnerable to this effect toward the end of rapid intensification (Finocchio and Rios-Berrios, 2021). Simultaneously, many rapidly weakening storms, especially in the eastern North Pacific, cross over strong sea surface temperature gradients into cooler waters.

Rapid weakening typically also occurs for strong tropical cyclones that make landfall. There are two principal causes. First, increasing surface roughness of land versus ocean results in an increase in frictional torques that spin down the circulation. Second, the loss of sea surface evaporation halts the main energy source of tropical cyclones, weakening convection and the circulation. In some rare cases, tropical cyclones can become temporarily reinvigorated over land if local soil moisture conditions, combined with diurnal heating, result in sufficient surface latent heat fluxes to jump start convection around the vortex center.

Tropical Cyclone Size

Perhaps one of the more overlooked aspects of tropical cyclones is their size. The size of tropical cyclones has significant bearing on the duration, coverage, and severity of the hazards they produce. Fundamental understanding of what controls tropical cyclone size is still being established, perhaps limited by our ability to regularly and reliably map out the outer wind field of tropical cyclones.

Based on satellite scatterometer data, there are variations in the mean tropical size by ocean basin (**Table 2**). Tropical cyclone size is often measured by the average radius of some threshold wind speed, such as minimum tropical storm strength (17 m s^{-1}). The western North Pacific has the largest tropical cyclones on average, while the eastern North Pacific has the smallest tropical cyclones on average (Chan and Chan, 2015). One factor influencing these differences is the incipient circulation size of tropical disturbances (Martinez et al., 2020), along with the background synoptic circulations in which they are embedded, which influences the angular momentum reservoir available to spin up the outer circulation of tropical cyclones.

A multitude of internal and environmental factors may influence tropical cyclone size evolution. The overall tendency of a tropical cyclone is to grow in size after reaching peak intensity. This growth is encouraged by convective rainband activity, including secondary eyewalls, as they repeatedly generate cyclonic vorticity and the associated inflow encourages a gradual inward advection of low-level angular momentum surfaces. Changes in the boundary-layer structure in promoting convergence and ascent out of the boundary layer at greater radii also play a role (Kilroy et al., 2016). The degree of environmental moisture near a tropical cyclone may modulate rainband activity, and the presence of vertical wind shear encourages greater rainband activity farther from the center on the downshear side of a storm. Additionally, trough interactions are known to expand the tropical cyclone wind field. Hurricane

Table 2 Mean and standard deviation of tropical cyclone size by ocean basin. The size is defined as the average radial extent of tropical-storm force winds, as inferred from QuikSCAT satellite data from 1999–2009.

Ocean basin	*Mean size (km)*	*Standard deviation (km)*
North Atlantic	201	83
Eastern North Pacific	124	50
Western North Pacific	234	108
South Indian	179	77
South Pacific	224	78

Data from Chan and Chan (2015).

Sandy (2012) was an impactful example of a massive expansion of the wind field that was brought about by a series of trough interactions. Other factors affecting tropical cyclone size are actively being explored in the research community.

There have also been efforts to understand what parameters set an equilibrium tropical cyclone size in the climate system using idealized simulations. In idealized simulations that do not have any variation in the Coriolis parameter (an f-plane) and that are simulated for long periods of time to radiative-convective equilibrium, the outer equilibrium size scales as the ratio of the potential intensity over the Coriolis parameter. In idealized simulations on an aquaplanet sphere, however, the outer size scales as the Rhines scale (Chavas and Reed, 2019). The Rhines scale varies as $\beta^{-1/2}$, where β is the meridional gradient in the Coriolis parameter. In essence, the Rhines scale serves to limit the expansion of tropical cyclones due to the radiation of Rossby wave energy at scales larger than the Rhines scale. The Rhines scale slowly increases with latitude, so allows for a slow increase in tropical cyclone size with latitude.

Conclusion

A tropical cyclone is an example of one of nature's marvels. The juxtaposition of outward beauty and inner violence that a tropical cyclone can possess seems downright mysterious. Yet, tropical cyclones are no accident. The laws of physics dictate tropical cyclones to be a feature of our Earth system. The tropical cyclone research community has gradually chipped away at the mysteries of tropical cyclones, achieving progress in understanding the formation, structure, intensity change, and size of tropical cyclones. At each stage of a tropical cyclone's lifecycle, there is rich complexity in interactions between the environment, vortex, and clouds. Many questions remain, but the gains in knowledge, coupled with technological advances, have allowed society to better prepare and mitigate against these impactful storms.

References

Adames, Á.F., 2021. Interactions between water vapor, potential vorticity, and vertical wind shear in quasi-geostrophic motions: implications for rotational tropical motion systems. J. Atmos. Sci. 78, 903–923. https://doi.org/10.1175/JAS-D-20-0205.1.

Alland, J.J., Tang, B.H., Corbosiero, K.L., Bryan, G.H., 2021. Combined effects of midlevel dry air and vertical wind shear on tropical cyclone development. Part II: Radial ventilation. J. Atmos. Sci. 78, 783–796. https://doi.org/10.1175/JAS-D-20-0055.1.

Bentley, A.M., Keyser, D., Bosart, L.F., 2016. A dynamically based climatology of subtropical cyclones that undergo tropical transition in the North Atlantic basin. Mon. Weather Rev. 144, 2049–2068. https://doi.org/10.1175/MWR-D-15-0251.1.

Chan, K.T.F., Chan, J.C.L., 2015. Global climatology of tropical cyclone size as inferred from QuikSCAT data. Int. J. Climatol. 35, 4843–4848. https://doi.org/10.1002/joc.4307.

Chavas, D.R., Reed, K.A., 2019. Dynamical aquaplanet experiments with uniform thermal forcing: system dynamics and implications for tropical cyclone genesis and size. J. Atmos. Sci. 76, 2257–2274. https://doi.org/10.1175/JAS-D-19-0001.1.

Corbosiero, K.L., Molinari, J., Aiyyer, A.R., Black, M.L., 2006. The structure and evolution of Hurricane Elena (1985). Part II: convective asymmetries and evidence for vortex Rossby waves. Mon. Weather Rev. 134, 3073–3091. https://doi.org/10.1175/MWR3250.1.

Didlake, A.C., Houze, R.A., 2013. Dynamics of the stratiform sector of a tropical cyclone rainband. J. Atmos. Sci. 70, 1891–1911. https://doi.org/10.1175/JAS-D-12-0245.1.

Ditchek, S.D., Molinari, J., Corbosiero, K.L., Fovell, R.G., 2019. An objective climatology of tropical cyclone diurnal pulses in the Atlantic basin. Mon. Weather Rev. 147, 591–605. https://doi.org/10.1175/MWR-D-18-0368.1.

Dunion, J.P., Thorncroft, C.D., Velden, C.S., 2014. The tropical cyclone diurnal cycle of mature hurricanes. Mon. Weather Rev. 142, 3900–3919. https://doi.org/10.1175/MWR-D-13-00191.1.

Emanuel, K., Velez-Pardo, M., Cronin, T.W., 2023. The surprising roles of turbulence in tropical cyclone physics. Atmos 14, 1254. https://doi.org/10.3390/atmos14081254.

Evans, C., Wood, K.M., Aberson, S.D., Archambault, H.M., Milrad, S.M., Bosart, L.F., Corbosiero, K.L., Davis, C.A., Pinto, J.R.D., Doyle, J., Fogarty, C., Galarneau, T.J., Grams, C.M., Griffin, K.S., Gyakum, J., Hart, R.E., Kitabatake, N., Lentink, H.S., McTaggart-Cowan, R., Perrie, W., Quinting, J.F.D., Reynolds, C.A., Riemer, M., Ritchie, E.A., Sun, Y., Zhang, F., 2017. The extratropical transition of tropical cyclones. Part I: Cyclone evolution and direct impacts. Mon. Weather Rev. 145, 4317–4344. https://doi.org/10.1175/MWR-D-17-0027.1.

Feng, X., Wu, L., 2022. Roles of interdecadal variability of the western North Pacific monsoon trough in shifting tropical cyclone formation. Clim. Dynam. 58, 87–95. https://doi.org/10.1007/s00382-021-05891-w.

Finocchio, P.M., Rios-Berrios, R., 2021. The intensity- and size-dependent response of tropical cyclones to increasing vertical wind shear. J. Atmos. Sci. 78, 3673–3690. https://doi.org/10.1175/JAS-D-21-0126.1.

Fischer, M.S., Tang, B.H., Corbosiero, K.L., 2019. A climatological analysis of tropical cyclone rapid intensification in environments of upper-tropospheric troughs. Mon. Weather Rev. 147, 3693—3719. https://doi.org/10.1175/MWR-D-19-0013.1.

Galarneau, T.J., Davis, C.A., 2013. Diagnosing forecast errors in tropical cyclone motion. Mon. Weather Rev. 141, 405—430. https://doi.org/10.1175/MWR-D-12-00071.1.

Gray, W.M., 1968. Global view of the origin of tropical disturbances and storms. Mon. Weather Rev. 96, 669—700. https://doi.org/10.1175/1520-0493(1968)096<0669:GVOTOO>2.0.CO;2.

Hence, D.A., Houze, R.A., 2008. Kinematic structure of convective-scale elements in the rainbands of Hurricanes Katrina and Rita (2005). J. Geophys. Res. 113, D15108. https://doi.org/10.1029/2007JD009429.

Hendricks, E.A., Montgomery, M.T., Davis, C.A., 2004. The role of "vortical" hot towers in the formation of Tropical Cyclone Diana (1984). J. Atmos. Sci. 61, 1209—1232. https://doi.org/10.1175/1520-0469(2004)061<1209:TROVHT>2.0.CO;2.

Houze, R.A., 2010. Clouds in tropical cyclones. Mon. Weather Rev. 138, 293—344. https://doi.org/10.1175/2009MWR2989.1.

Judt, F., Rios-Berrios, R., Bryan, G.H., 2023. Marathon versus sprint: two modes of tropical cyclone rapid intensification in a global convection-permitting simulation. Mon. Weather Rev. 151, 2683—2699. https://doi.org/10.1175/MWR-D-23-0038.1.

Kilroy, G., Smith, R.K., Montgomery, M.T., 2016. Why do model tropical cyclones grow progressively in size and decay in intensity after reaching maturity? J. Atmos. Sci. 73, 487—503. https://doi.org/10.1175/JAS-D-15-0157.1.

Lawton, Q.A., Majumdar, S.J., Dotterer, K., Thorncroft, C., Schreck, C.J., 2022. The influence of convectively coupled kelvin waves on African easterly waves in a wave-following framework. Mon. Weather Rev. 150, 2055—2072. https://doi.org/10.1175/MWR-D-21-0321.1.

Mallen, K.J., Montgomery, M.T., Wang, B., 2005. Reexamining the near-core radial structure of the tropical cyclone primary circulation: implications for vortex resiliency. J. Atmos. Sci. 62, 408—425. https://doi.org/10.1175/JAS-3377.1.

Martinez, J., Nam, C.C., Bell, M.M., 2020. On the contributions of incipient vortex circulation and environmental moisture to tropical cyclone expansion. J. Geophys. Res. Atmos. 125. https://doi.org/10.1029/2020JD033324 e2020JD033324.

McTaggart-Cowan, R., Galarneau, T.J., Bosart, L.F., Moore, R.W., Martius, O., 2013. A global climatology of baroclinically influenced tropical cyclogenesis. Mon. Weather Rev. 141, 1963—1989. https://doi.org/10.1175/MWR-D-12-00186.1.

Molinari, J., Vollaro, D., 2013. What percentage of western North Pacific tropical cyclones form within the monsoon trough? Mon. Weather Rev. 141, 499—505. https://doi.org/10.1175/MWR-D-12-00165.1.

Montgomery, M.T., Persing, J., Smith, R.K., 2015. Putting to rest WISHE-ful misconceptions for tropical cyclone intensification. J. Adv. Model. Earth Syst. 7, 92—109. https://doi.org/10.1002/2014MS000362.

Moon, Y., Nolan, D.S., 2015. Spiral rainbands in a numerical simulation of Hurricane Bill (2009). Part II: Propagation of inner rainbands. J. Atmos. Sci. 72, 191—215. https://doi.org/10.1175/JAS-D-14-0056.1.

Nam, C.C., Bell, M.M., 2021. Multiscale shear impacts during the genesis of Hagupit (2008). Mon. Weather Rev. 149, 551—569. https://doi.org/10.1175/MWR-D-20-0133.1.

Navarro, E.L., Hakim, G.J., 2016. Idealized numerical modeling of the diurnal cycle of tropical cyclones. J. Atmos. Sci. 73, 4189—4201. https://doi.org/10.1175/JAS-D-15-0349.1.

Nguyen, L.T., Rogers, R., Zawislak, J., Zhang, J.A., 2019. Assessing the influence of convective downdrafts and surface enthalpy fluxes on tropical cyclone intensity change in moderate vertical wind shear. Mon. Weather Rev. 147, 3519—3534. https://doi.org/10.1175/MWR-D-18-0461.1.

Nolan, D.S., 2007. What is the trigger for tropical cyclogenesis? Aust. Meteorol. Mag. 56, 241—266.

Nuñez Ocasio, K.M., Evans, J.L., Young, G.S., 2020. A wave-relative framework analysis of AEW—MCS interactions leading to tropical cyclogenesis. Mon. Weather Rev. 148, 4657—4671. https://doi.org/10.1175/MWR-D-20-0152.1.

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

Raymond, D.J., Gjorgjievska, S., Sessions, S.L., Fuchs, Z., 2014. Tropical cyclogenesis and mid-level vorticity. Aust. Meteorol. Oceanogr. J. 64, 11—25.

Richardson, J.C., Torn, R.D., Tang, B.H., 2022. An analog comparison between rapidly and slowly intensifying tropical cyclones. Mon. Weather Rev. 150, 2139—2156. https://doi.org/10.1175/MWR-D-21-0260.1.

Riemer, M., 2016. Meso-beta-scale environment for the stationary band complex of vertically sheared tropical cyclones. Quart. J. Roy. Meteor. Soc. 142, 2442—2451. https://doi.org/10.1002/qj.2837.

Rios-Berrios, R., Torn, R.D., 2017. Climatological analysis of tropical cyclone intensity changes under moderate vertical wind shear. Mon. Weather Rev. 145, 1717—1738. https://doi.org/10.1175/MWR-D-16-0350.1.

Rogers, R., Reasor, P., Lorsolo, S., 2013. Airborne Doppler observations of the inner-core structural differences between intensifying and steady-state tropical cyclones. Mon. Weather Rev. 141, 2970—2991. https://doi.org/10.1175/MWR-D-12-00357.1.

Rudzin, J.E., Chen, S., Sanabia, E.R., Jayne, S.R., 2020. The air-sea response during Hurricane Irma's (2017) rapid intensification over the Amazon-Orinoco river plume as measured by atmospheric and oceanic observations. J. Geophys. Res. Atmos. 125. https://doi.org/10.1029/2019JD032368 e2019JD032368.

Russell, J.O., Aiyyer, A., White, J.D., Hannah, W., 2017. Revisiting the connection between African easterly waves and Atlantic tropical cyclogenesis. Geophys. Res. Lett. 44, 587—595. https://doi.org/10.1002/2016GL071236.

Schreck, C.J., Molinari, J., Aiyyer, A., 2012. A global view of equatorial waves and tropical cyclogenesis. Mon. Weather Rev. 140, 774—788. https://doi.org/10.1175/MWR-D-11-00110.1.

Sitkowski, M., Kossin, J.P., Rozoff, C.M., 2011. Intensity and structure changes during hurricane eyewall replacement cycles. Mon. Weather Rev. 139, 3829—3847. https://doi.org/10.1175/MWR-D-11-00034.1.

Sobel, A.H., Wing, A.A., Camargo, S.J., Patricola, C.M., Vecchi, G.A., Lee, C.-Y., Tippett, M.K., 2021. Tropical cyclone frequency. Earth's Future 9, e2021EF002275. https://doi.org/10.1029/2021EF002275.

Stevenson, S.N., Corbosiero, K.L., DeMaria, M., Vigh, J.L., 2018. A 10-year survey of tropical cyclone inner-core lightning bursts and their relationship to intensity change. Weather Forecast. 33, 23—36. https://doi.org/10.1175/WAF-D-17-0096.1.

Tang, B.H., 2017. Coupled dynamic—thermodynamic forcings during tropical cyclogenesis. Part II: Axisymmetric experiments. J. Atmos. Sci. 74, 2279—2291. https://doi.org/10.1175/JAS-D-17-0049.1.

Tang, B.H., Emanuel, K.A., 2010. Midlevel ventilation's constraint on tropical cyclone intensity. J. Atmos. Sci. 67, 1817—1830. https://doi.org/10.1175/2010JAS3318.1.

Thorncroft, C.D., Hall, N.M.J., Kiladis, G.N., 2008. Three-dimensional structure and dynamics of African easterly waves. Part III: Genesis. J. Atmos. Sci. 65, 3596—3607. https://doi.org/10.1175/2008JAS2575.1.

Wang, Z., 2014. Role of cumulus congestus in tropical cyclone formation in a high-resolution numerical model simulation. J. Atmos. Sci. 71, 1681—1700. https://doi.org/10.1175/JAS-D-13-0257.1.

Wang, Z., Montgomery, M.T., Dunkerton, T.J., 2010. Genesis of pre—Hurricane Felix (2007). Part I: The role of the easterly wave critical layer. J. Atmos. Sci. 67, 1711—1729. https://doi.org/10.1175/2009JAS3420.1.

Wing, A.A., 2022. Acceleration of tropical cyclone development by cloud-radiative feedbacks. J. Atmos. Sci. 79, 2285—2305. https://doi.org/10.1175/JAS-D-21-0227.1.

Wood, K.M., Ritchie, E.A., 2015. A definition for rapid weakening of North Atlantic and eastern North Pacific tropical cyclones. Geophys. Res. Lett. 42, 10,091—10,097. https://doi.org/10.1002/2015GL066697.

Wu, L., Zong, H., Liang, J., 2013. Observational analysis of tropical cyclone formation associated with monsoon gyres. J. Atmos. Sci. 70, 1023—1034. https://doi.org/10.1175/JAS-D-12-0117.1.

Yu, C.-L., Didlake, A.C., Zhang, F., 2022. Updraft maintenance and axisymmetrization during secondary eyewall formation in a model simulation of Hurricane Matthew (2016). J. Atmos. Sci. 79, 1105—1125. https://doi.org/10.1175/JAS-D-21-0103.1.

Zawislak, J., Zipser, E.J., 2014. Analysis of the thermodynamic properties of developing and nondeveloping tropical disturbances using a comprehensive dropsonde dataset. Mon. Weather Rev. 142, 1250—1264. https://doi.org/10.1175/MWR-D-13-00253.1.

Zhang, J.A., Rogers, R.F., Nolan, D.S., Marks, F.D., 2011. On the characteristic height scales of the hurricane boundary layer. Mon. Weather Rev. 139, 2523—2535. https://doi.org/10.1175/MWR-D-10-05017.1.

Zhang, J.A., Rogers, R.F., Reasor, P.D., Gamache, J., 2023. The mean kinematic structure of the tropical cyclone boundary layer and its relationship to intensity change. Mon. Weather Rev. 151, 63—84. https://doi.org/10.1175/MWR-D-21-0335.1.