Extratropical and Tropical Transition

Introduction

Extratropical and tropical transition each represents an end of the tropical cyclone lifecycle. Extratropical transition, along with dissipation, represents the end of the tropical cyclone lifecycle. Conversely, tropical transition represents a unique beginning to the tropical cyclone life cycle. In this lecture, we aim to describe both, focusing primarily on the extratropical transition process and its impacts upon both cyclone structure and the downstream weather pattern.

Key Concepts

- What is extratropical transition?
- What factors influence the post-transition evolution of a tropical cyclone?
- What influences do transitioning tropical cyclones exert on the downstream pattern?
- What is tropical transition?
- Under what conditions is tropical transition possible and/or favored?

Extratropical Transition: Overview

Extratropical transition (ET) is the gradual process by which an initially warm-core tropical cyclone transforms into an (initially) cold-core extratropical cyclone (Jones et al. 2003). ET occurs within nearly every basin that experiences tropical cyclones, with the largest number of ET events found in the western North Pacific and the largest percentage of tropical cyclones undergoing ET (~45%) found in the North Atlantic. ET is relatively rare within the eastern North Pacific and northern Indian Ocean basins due to the background synoptic environment and low latitude landlocked nature of the basin, respectively. ET events typically occur at relatively low latitudes early and late in a basin's tropical season and at relatively high latitudes during the peak of a basin's tropical season. The intraseasonal distribution of ET events favors the latter half of a basin's tropical season, when conditions ripe for tropical cyclone development linger into the time period during (and latitudes at) which the midlatitude synoptic-scale environment becomes increasingly conducive to extratropical cyclone development.

As a tropical cyclone moves poleward, it experiences changes in its environment, including increased baroclinicity and vertical wind shear; the presence of meridional humidity gradients; decreased sea surface temperatures associated with strong sea surface temperature gradients; and an increased Coriolis parameter. Oftentimes, a transitioning tropical cyclone interacts with an upper tropospheric trough and/or a mature extratropical cyclone; in so doing, the tropical cyclone typically accelerates poleward and eastward. During ET, the tropical cyclone responds to the aforementioned changes in its environment. The inner core of the tropical cyclone loses its symmetric appearance and gradually takes on the appearance of an extratropical cyclone. Frontal structures, particularly a zonally-oriented warm front, begin to develop and the wind, wave, and precipitation fields associated with the tropical cyclone become increasingly asymmetric. The influence of the wind-induced surface heat exchange process wanes as the cyclone begins to become driven by baroclinic energetics (e.g., those associated with the vertically-sheared flow).

Structural Evolutions Associated with ET

Evolution of the Surface Wind Field

As a tropical cyclone undergoes ET, substantial changes in the structure of its near-surface wind field occur. In response to cooling maximized inside of the transitioning cyclone's radius of maximum winds, the peak intensity of the cyclone decays and its radius of maximum winds moves radially outward (Evans and Hart 2008). Outside of the radius of maximum winds, the wind profile becomes increasingly asymmetric, favoring stronger winds equatorward of the transitioning cyclone. Here, such winds are often stronger than those found at similar radii prior to ET. Evans and Hart (2008) tie this to absolute angular momentum conservation along inflowing trajectories as the cyclone becomes increasingly asymmetric. Further, the intensity of the transitioning cyclone decays more rapidly near the surface than higher up within the boundary layer. The stabilization of the boundary layer over cooler waters enables the near-surface and boundary layer circulations to decouple, mitigating the deleterious impacts of friction upon the boundary layer circulation.

Evolution of the Surface Wave Field

Similar to the near-surface wind field, the near-surface wave field becomes increasingly asymmetric and grows in size during ET. During the tropical phase, waves tend to move faster than the tropical cyclone. However, during ET, the acceleration of the transitioning cyclone locks (or nearly locks) the propagation of the waves to that of the cyclone itself, promoting wave growth. Thus, though cyclone intensity and strength may both decrease during ET, the wave damage potential may become substantially greater. Indeed, as noted by Jones et al. (2003), wave heights associated with a transitioning cyclone can be 50-100% higher than those associated with a similar stationary cyclone.

Evolution of Precipitation and Cloud Fields

At the start of an ET event, heavy precipitation becomes embedded within the large cloud shield associated with the tropical cyclone outflow that extends poleward from the tropical cyclone center. This cloud shield is sometimes referred to as the “delta rain” region of a transitioning cyclone, so named after the “delta” shape that it often takes on satellite imagery (Klein et al. 2000), and is often associated with large-scale forcing for ascent and isentropic upglide atop the transitioning cyclone's developing warm front. Due to the expansion of the area covered by clouds and precipitation during ET, heavy precipitation can occur over land without the transitioning cyclone making landfall. As ET proceeds, the distribution of heavy precipitation changes from that found with a tropical cyclone, where heavy precipitation typically occurs on both sides of the track of the cyclone, to that of an extratropical cyclone, where the heaviest precipitation is typically found to the left (right) of track in the Northern (Southern) Hemisphere. Such a shift is thought to be driven by the forcing for large-scale ascent accompanying the interaction of the tropical cyclone with an upstream trough. Note also that local orographic effects can considerably enhance precipitation during an ET event.

Observational Characterization of ET

Klein et al. (2000) describe a three-stage satellite-derived conceptual model of ET. In stage one, ET commences as the tropical cyclone translates poleward over lowered sea surface temperatures and its outer circulation begins to impinge upon a preexisting midlatitude baroclinic zone. As interaction with the baroclinic zone begins, colder, diver environmental equatorward flow exists to the west of the tropical cyclone. This acts to erode deep convection to the west of the cyclone, promoting the development of a dry slot in its southwest quadrant. Warm, moist environmental poleward flow to the east of the tropical cyclone helps to maintain deep, moist convection there. Such flow turns cyclonically and interacts with the baroclinic zone to produce ascent over vertically-tilted isentropic surfaces. A cirrus shield is also often observed in satellite imagery at this stage, produced by the interaction of the transitioning cyclone's outflow with vertical wind shear associated with the polar jet.

In stage two, the tropical cyclone is located just equatorward of the baroclinic zone. Cyclonic rotation of the baroclinic zone to a southwest-northeast orientation occurs in response to the superposition of the cyclone's flow upon the baroclinic zone itself. Continued cold, dry (warm, moist) flow west (east) of the transitioning cyclone results in the formation of a dipole of lower tropospheric temperature advection, with cold (warm) advection found to the cyclone's west (east). Ascending, cyclonically-turning flow atop the baroclinic zone continues. Some of these ascending parcels subsequently descend into the western quadrant of the cyclone, indicative of the developing descending dry intrusion characteristic of extratropical cyclones. Other such ascending parcels subsequently turn anticyclonically toward the northeast. The upper tropospheric outflow of the transitioning cyclone becomes confluent with the polar jet while vertical wind shear acts to advect the cyclone's upper tropospheric warm core downstream. Deep, moist convection persists within the cyclone's inner core through this stage despite the increasing environmental ventilation.

Stage three is the logical conclusion of the continuation of the physical processes described in stage two as the cyclone becomes embedded within the baroclinic zone. Dry adiabatic descent of parcels west of the cyclone progressively weakens the inner core convection associated with the tropical cyclone, eventually producing eyewall erosion to the south and west. Such descent also acts to suppress cold front genesis. The transitioning cyclone continues to decay from the top-down as the vertically-sheared flow continues to advect its warm core downstream. By the completion of stage three, the cyclone resembles an extratropical cyclone with well-defined cold conveyor belt, warm conveyor belt, and descending dry intrusion structures.

Objective Characterization of ET

Evans and Hart (2003) utilize the salient structural changes observed during ET to describe objective diagnostics for its onset and completion. The onset of ET is marked by the development of a sustained thickness asymmetry of greater than 10 m between 900 and 600 hPa. In the Northern Hemisphere, this thickness asymmetry is measured as the 900-600 hPa thickness in the semicircle to the right of the cyclone's motion vector minus the 900-600 hPa thickness in the semicircle to the left of the cyclone's motion vector. The relationship between thickness and layer-mean virtual temperature enables us to view a horizontal thickness asymmetry as a measure of the horizontal temperature gradient across the tropical cyclone. The completion of ET is marked by the transition of the cyclone's lower-to-middle tropospheric thermal structure from a warm-core to a cold-core structure. This is measured by the 900-600 hPa thermal wind, where positive values of the thermal wind parameter reflect winds that increase with increasing height. For full details on the cyclone phase space defined by these parameters, interested readers are referred to Hart (2003) and references therein.

Downstream Development

The occurrence of an ET event often temporarily compromises the forecast skill across an entire ocean basin. This occurs as the complex interactions between the transitioning cyclone and the midlatitude synoptic-scale pattern that occur during ET can lead to a modification of the synoptic-scale flow pattern, particularly downstream of the transitioning cyclone. Such interactions have the potential to initiate high-impact weather such as explosive cyclogenesis or severe precipitation events far from the ET event itself. To first order, the impact of an ET event is believed to spread downstream through its excitation and the subsequent propagation of a Rossby wave train along the potential vorticity gradient associated with the midlatitude jet stream (Riemer et al. 2008). Herein, we focus on describing the results of two works, Riemer et al. (2008) and Riemer and Jones (2010), that address this problem, hereafter referred to as downstream development, from an idealized numerical modeling perspective.

In Riemer et al. (2008), ET occurs as an idealized tropical cyclone interacts with a zonally-oriented upper tropospheric jet. This interaction leads to the formation of a ridge-trough couplet in the upper troposphere and a jet streak downstream of the transitioning cyclone. The ridge is amplified initially by the divergent flow associated with the outflow of the tropical cyclone. Rapid surface cyclogenesis takes place beneath the left exit region of the jet streak. The ridge-trough pattern extends downstream as a wave pattern and initiates a family of cyclones. The temporal evolution of this wave pattern can be interpreted, as before, as the excitation of a Rossby wave train by the ET event and its subsequent downstream propagation. The transitioning tropical cyclone also induces a significant poleward moisture transport into the midlatitudes, primarily to its east.

The evolution downstream of the transitioning tropical cyclone can be viewed as downstream baroclinic development with the transitioning tropical cyclone acting as the initial perturbation that triggers such development. The diabatically-driven divergent outflow of the tropical cyclone, associated with negative potential vorticity advection on the dynamic tropopause (e.g., Archambault et al. 2013), hinders the downstream propagation of the baroclinic development and promotes phase locking between the upper tropospheric wave and the transitioning cyclone itself. Although the tropical cyclone becomes embedded within the midlatitude flow during ET, it still maintains a coherent structure well after ET rather than rapidly deforms in response to the strong vertical and horizontal wind shears. Thus, the tropical cyclone can be viewed as a long-lived local perturbation on the downstream midlatitude pattern. Archambault et al. (2013, 2015) find that the degree to which the tropical cyclone interacts with the midlatitude flow, rather than the tropical cyclone's intensity, post-ET evolution, or size, exerts the greatest influence upon the downstream response to ET.

The midlatitude atmospheric state exerts a substantial influence upon the propagation of the tropical cyclone-induced Rossby wave train. With strong jets, Rossby wave energy propagates rapidly downstream, resulting in a weakly amplified pattern downstream. With weak jets, the Rossby wave train is of shorter wavelength and, correspondingly, is more amplified. Rossby wave energy still propagates downstream, just not as rapidly as with stronger jets. Diabatic processes, as associated with deep, moist convection (including with the tropical cyclone), are found to be important to the amplitude of the Rossby wave train but not to its wavelength or propagation velocity. In this regard, the Rossby wave train constitutes the upper tropospheric precursor that determines where lower tropospheric cyclogenesis may occur while moist processes act to locally intensify the cyclone (e.g., so as to change its amplitude but not its zonal structure or propagation). A more intense tropical cyclone leads to a stronger deflection of the upper tropospheric flow in the immediate vicinity of the transitioning cyclone, but such a deflection does not necessarily lead to a deeper development of the primary downstream cyclone nor to a higher-amplitude Rossby wave train.

In Riemer and Jones (2010), ET occurs as an idealized tropical cyclone interacts with a pre-existing developing baroclinic wave pattern. To first order, the interaction of the transitioning tropical cyclone with a developing baroclinic wave results in the evolution of the baroclinic wave in a qualitatively similar manner to that described above in the straight jet evolution. However, the development of the wave pattern is sensitive to the initial phasing of the tropical cyclone and the upstream upper tropospheric trough. The case where the transitioning tropical cyclone directly interacts with the leading edge of a trough within the baroclinic wave pattern is believed to represent an optimal configuration of the midlatitude circulation with respect to the tropical cyclone. Displacing the location along the baroclinic wave where the transitioning tropical cyclone exerts its greatest forcing from this optimal configuration results in a reduced downstream impact more closely resembling that which would be expected without the forcing from the tropical cyclone altogether.

Post-Transition Evolution

Hart et al. (2006) quantify the synoptic-scale factors promoting the rapid or slow completion of ET; the intensification or decay of the cyclone after ET; and the evolution of the cyclone's thermal structure after ET. Rapidly (slowly) transitioning tropical cyclones are classified as those that take ≤ 12 h (≥ 48 h) to complete ET. Rapidly transitioning tropical cyclones are associated with a higher amplitude trough than those that slowly transition, thereby promoting the meridional (zonal) translation of rapidly (slowly) transitioning tropical cyclones. Not coincidentally, the sea surface temperatures associated with slowly transitioning tropical cyclones are 3-4°C warmer than those associated with rapidly transitioning tropical cyclones. Rapidly transitioning tropical cyclones tend to be both smaller and weaker (by roughly one category on the Saffir-Simpson Hurricane Wind Scale) than their slowly transitioning counterparts.

Strengthening (weakening) cyclones after ET are classified as those that deepen (weaken) by >4 hPa in terms of minimum sea level pressure over the 24 h period following ET. For post-ET intensifiers (weakeners), a negatively (positively) tilted 500 hPa synoptic-scale trough is advancing on the tropical cyclone. The negative tilt of the trough permits the trough to more closely approach the tropical cyclone, similar to the "favorable trough interaction" composite of Hanley et al. (2001). For post-ET intensifying cyclones, there remains a maximum of 850 hPa equivalent potential temperature in its environment that extends equatorward; no such connection to the tropics exists for post-ET weakening cyclones. The presence of high lower tropospheric equivalent potential temperature and relatively cool temperatures aloft associated with the 500 hPa trough promotes conditional instability that favors robust deep, moist convective development (e.g., Bosart and Lackmann 1995). Sea surface temperatures and horizontal gradients therein are larger for cyclones that strengthen post-ET than for those that weaken post-ET.

Cyclones that reacquire a warm-core structure after ET interact with an upper tropospheric trough that is considerably narrower in horizontal scale, yet more extensive in vertical scale, than a cyclone that remains cold-core after ET. Similarly, tropical cyclones that reacquire a warm-core structure are (on average) 50% larger than those that remain cold-core after ET. Thus, both above-average tropical cyclone size and a below-normal trough width promote the scale matching of the transitioning cyclone and upper tropospheric trough, similar again to that observed with the favorable trough interaction composite of Hanley et al. (2001). It should be noted that cyclones that reacquire a warm-core structure after ET tend to be among the most damaging extratropical cyclones, combining relatively intense maximum sustained winds (common to tropical cyclones) with an expansive wind field (common to extratropical cyclones).

Tropical Transition

Tropical transition (TT), or the transformation of an initially extratropical, cold-core cyclone into a warm-core tropical cyclone, represents a non-traditional pathway to tropical cyclone genesis. TT is most prevalent in the North Atlantic basin, where it led to the development of approximately 28% of all tropical cyclones between 1948-2004 (McTaggart-Cowan et al. 2008). Tropical cyclone genesis associated with TT takes place in environments with relatively warm sea surface temperatures (~26°C), initially high magnitudes of vertical wind shear, moderate to large synoptic-scale forcing for ascent, and moderate to large amounts of lower tropospheric baroclinicity. TT events most frequently occur in the subtropics, particularly off of the eastern coast of the United States, and the tropical cyclones that result from the TT process rarely exceed Category 2 intensity on the Saffir-Simpson Hurricane Wind Scale. The work of Davis and Bosart (2004) forms much of the basis for the following discussion of TT.

To first order, TT events can be classified as “weak TT” or “strong TT” events, following McTaggart-Cowan et al. (2008). These roughly correspond to the “strong extratropical cyclone (SEC)” and “weak extratropical cyclone (WEC)” precursor disturbance paradigms of Davis and Bosart (2004). Fundamentally, TT requires the development of a sufficiently intense lower tropospheric cyclone so as to trigger the non-linear wind-induced surface heat exchange amplification process crucial to tropical cyclone development. For strong TT events, such a cyclone develops in response to large-scale forcing for ascent along a lower tropospheric baroclinic zone. For weak TT events, such a cyclone develops in response to the organization of deep, moist convection, potentially aided by vertical wind shear, about an initially weak precursor disturbance such as a mesoscale convective vortex or a weak baroclinic cyclone.

Furthermore, for both types of TT events, the magnitude of the vertical wind shear must be reduced to sufficiently low values (< 10 m s^{-1}) in order for TT to complete. Two concurrent physical processes, both tied to the diabatic heating associated with deep, moist convection, act to bring about such a reduction. Both upper tropospheric outflow and the vertical redistribution of potential vorticity driven by diabatic heating in deep, moist convection act to reduce the horizontal gradient of potential vorticity across the developing disturbance. As noted by Davis and Bosart (2003), the magnitude of the horizontal gradient of potential vorticity is proportional to the magnitude of the vertical wind shear. Thus, reducing the magnitude of the former acts to reduce the magnitude of the latter. Such a process is promoted when deep, moist convection develops upshear rather than downshear of the incipient disturbance. Subsequently, the developing disturbance occludes in relatively warm air equatorward of the large-scale jet stream, thereby losing its frontal character and enabling tropical cyclone genesis to occur.

A successful TT event requires that the incipient disturbance both occlude and remain over relatively warm water for at least one day following occlusion. With respect to the former, the occlusion process can be prevented if the disturbance continually interacts with transient shortwave upper tropospheric troughs, whereby each transient trough again increases the vertical wind shear and promotes extratropical cyclone development. Thus, it can be stated that TT is promoted under conditions of large-scale blocking or otherwise weakened steering currents. The latter follows directly from our previous discussions related to tropical cyclone energetics. However, it should be noted that TT events can occur with sea surface temperatures below 26°C if the outflow temperature is sufficiently cool so as to permit a relatively efficient tropical cyclone heat engine.

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