Influence of upper-tropospheric troughs on tropical cyclone intensity change and structure: observational, reanalysis, and idealized numerical modeling perspectives

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THE INFLUENCE OF UPPER-TROPOSPHERIC TROUGHS ON TROPICAL CYCLONE INTENSITY CHANGE AND STRUCTURE: OBSERVATIONAL, REANALYSIS, AND IDEALIZED NUMERICAL MODELING PERSPECTIVES

by

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ABSTRACT

The interaction between tropical cyclones (TCs) and upper-tropospheric troughs is a common occurrence in the North Atlantic basin; yet, the favorability of such interactions on TC intensity change, as well as the mechanisms by which TC intensity is affected by the trough, remain unclear. These open questions are investigated in this dissertation through the use of reanalysis, satellite observations, and idealized numerical modeling.

A climatology of TC–trough interactions in the North Atlantic basin was compiled from the European Centre for Medium Range Weather Forecasts ERA-Interim (ERA-I) reanalysis. The results of the climatology indicate that TC–trough interactions are more unfavorable for TC intensification compared to the climatology of the basin as a whole and TCs that do not interact with troughs. Although troughs are most often unfavorable for TC intensification, there are still TC–trough interactions that result in intensification.

Analog composites of favorable and unfavorable TC–trough interactions from satellite observations and the ERA–I were made to assess the roles of both the trough morphology and TC convection on intensity change. Relative to the unfavorable TC–trough interaction composite, favorable trough interactions are associated with weaker, shallower, and longitudinally-narrower troughs, which are associated with reduced wind shear. Compared to the unfavorable composite, favorable troughs are also associated with reduced midlevel entropy deficits, especially upshear of the TC, which, combined with the smaller shear values, results in less ventilation of the TC and strengthening convection. During favorable TC–trough interactions, the strengthening convection is able to wrap upshear of the TC center, which is a favorable configuration for TC intensification.
Idealized numerical modeling simulations were used to better assess the physical mechanisms that affect TC intensity change during TC–trough interaction. Two model simulations were first investigated: a simulation of a TC interacting with a westerly jet (TC–jet simulation) and a simulation interacting with a trough embedded in a westerly jet (TC–trough simulation). Despite experiencing larger vertical wind shear, the TC–trough TC was more intense for most of the simulation due to the increased dynamic and thermodynamic favorability imparted by the upstream trough. The presence of the trough results in moistening of the TC environment through three mechanisms associated with large-scale lift and divergence: reduced inertial stability resulting in enhanced outflow, transition of the synoptic flow from subgeostrophic to supergeostrophic, and $Q$-vector forcing for ascent.

Twenty-eight sensitivity experiments were performed to determine the effects of changing the initial parameters of the TC–trough interaction (i.e., trough scale, initial shear affecting the TC / initial distance between the TC and trough, and initial TC vortex intensity) on TC intensification and size. The results indicate that TC size is more sensitive to changes in these parameters, compared to TC intensification. TCs experiencing larger initial shear grew to be bigger because of larger vortex tilts. The simulations with larger tilts are associated with greater moistening of the downtilt TC environment, resulting in greater coverage of precipitation (and concomitant diabatic heating) outside of the RMW, leading to an increase in TC size. The presence of a larger-scale trough may also augment the size of TCs later in the simulation period through increased $Q$-vector forcing for ascent, which moistens the environment and results in greater diabatic heating over a larger area than occurred in simulations with smaller-scale troughs that have weaker forcing for ascent near the TC.
Chapter 2 contains original research previously published in *Geophysical Research Letters* in the article titled “Revisiting trough interactions and tropical cyclone intensity change” ©2016. The author of this dissertation was the lead author and researcher on the article. The original text of the article was slightly modified for this dissertation. Permission to republish was granted to the author by the American Geophysical Union, and the license details are included in Appendix A. The full citation of the article is:

ACKNOWLEDGMENTS

I put off writing the acknowledgments section of my dissertation until the last day, mostly because I had no idea how to start what I consider the most important section. So, with that admission as the first sentence, I suppose that’s as good a start as any!

First, I would like to thank my advisors, Brian Tang and Kristen Corbosiero (B and K, as I liked to address them in my many emails). Separately, anyone would be lucky to have advisors as caring and brilliant, but I had the pleasure of being co-advised. Together they corrected every incorrect comma, poor word choice, and numeral that should have been fully spelled out. I appreciate that they made me a better scientist, even when I was frustrated with the process. Most of all, though, I appreciate that they cared about me and my well-being. It is a rare thing in academia to have as open of a relationship as I did with Brian and Kristen, and it truly made all the difference.

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1. Introduction

1.1 Motivation

Tropical cyclone (TC) intensification during interaction with an upper-tropospheric trough, hereafter trough, remains a significant challenge to both forecasters and researchers. During their lifetime, most TCs interact with at least one trough and often can encounter several troughs (McTaggart–Cowan et al. 2008). Previous case studies, discussed in more detail in section 1.2, have shown that TC intensification following a trough interaction can range from rapid weakening to rapid intensification. In climatological studies, it is also unclear whether trough interactions are favorable or unfavorable for TC intensification. This dichotomy led to the coining of the “good trough / bad trough” problem posed in a climatological study of Atlantic TC–trough interactions by Hanley et al. (2001). The purpose of this dissertation is to use reanalysis, satellite observations, and idealized numerical modeling to determine the trough characteristics and physical processes associated with “good” and “bad” trough interactions.

1.2 Background

This section will begin with an overview of the climatological research on the favorability of TC–trough interactions. The subsequent subsections will provide a summary of the four major processes that have been hypothesized to affect TC intensity during trough interactions. Finally, an overview of other previous idealized modeling studies of TC–trough interactions is presented.

1.2.1 Favorability of TC–trough interactions

Whether TC–trough interactions are favorable for TC intensification remains unclear, with different case studies finding that TC intensity change during a trough interaction can range from rapid weakening to rapid intensification (e.g., McBride and Zehr 1981; Moli-
nari and Vollaro 1989; Bosart et al. 2000; Yu and Kwon 2005; Leroux et al. 2013, 2016; Rios-Berrios et al. 2016; Fischer et al. 2017). Additionally, Atlantic TC–trough interaction climatologies have found conflicting results as to whether troughs have a predominately negative or positive influence on TC intensification (DeMaria et al. 1993; Hanley et al. 2001). DeMaria et al. (1993) found that in a three-year period (39 trough interaction events), TCs intensified 33% of the time following a trough interaction, and weakened 38% of the time. Despite finding troughs slightly unfavorable overall, DeMaria et al. (1993) found intensity changes were highly variable during trough interactions, including several TCs rapidly intensifying under the influence of a trough. Additional results from DeMaria et al. (1993) are included in later subsections.

Hanley et al. (2001) examined trough-interaction events over a 12-year period. Trough interactions were shown to be highly favorable for intensification, with favorable interactions occurring 61–78% of the time, depending on the distance of the trough from the TC. Additional results from this study are discussed in later subsections and chapter 2. Of note, however, is that the apparently highly favorable results of trough interactions in Hanley et al. (2001) are partially an artifact of two methodological choices that bias the results toward intensification, rather than weakening or steady state.

1.2.2 Important physical processes during TC–trough interactions

Most trough interaction studies have focused on four physical processes that can affect TC intensity, three of which are generally considered favorable for intensification: (1) an increase in angular momentum eddy flux convergence (EFC) in the TC outflow layer, (2) an increase in upper-level outflow through the development of an upper-level jet, and (3) trough–TC potential vorticity (PV) merger, a process called superposition. Unfavorable for intensification, enhanced vertical wind shear from the trough acts to ventilate the TC and disrupt its circulation.
1.2.2.1 EFC

The occurrence of a trough interaction event may be defined by an increase in EFC in the upper levels of the TC (DeMaria et al. 1993; Hanley et al. 2001). EFC is defined as

\[
EFC = -\frac{1}{r^2} \frac{\partial}{\partial r} r^2 u_L' v_L'
\]  

(1.1)

where \( r \) is the radius from the TC center, \( u_L \) and \( v_L \) are the storm-relative radial and tangential wind, respectively, primes indicate perturbations from the azimuthal mean, and the overbar indicates the azimuthal average. Increased values of EFC occur in the TC outflow layer as the trough approaches due to the interaction of the TC outflow and the trough flow, which produces a net convergence of eddy angular momentum (DeMaria et al. 1993; Hanley et al. 2001; Leroux et al. 2016). Threshold values of EFC in an annulus (e.g., 300–600-km or 500–900-km) around the TC in the outflow layer, typically 10 m s\(^{-1}\) d\(^{-1}\), have often been used to diagnose the presence of a trough interacting with a TC (DeMaria et al. 1993; Hanley et al. 2001). Enhanced EFC can act to spin up the TC outflow layer through a Sawyer–Eliassen balanced vortex response (Eliassen 1952; Molinari and Vollaro 1989; DeMaria et al. 1993). The Sawyer–Eliassen balanced vortex model applies to a vortex in gradient wind balance, in which the azimuthally averaged tangential wind is in balance with the azimuthally-averaged pressure gradient (Eliassen 1952). The balanced vortex model describes how a secondary (radial–vertical) circulation evolves in the presence of a heat or momentum source in order to maintain balance. It has been hypothesized that in the case of a trough interaction, the trough serves as a momentum source, and the balanced vortex response leads to enhanced upper-level outflow and concomitant evacuation of mass from the TC core, subsequently lowering the TC’s central pressure and increasing its surface winds (Eliassen 1952; Molinari and Vollaro 1989).

Tests of this hypothesis, i.e. whether EFC has a significant effect on TC intensity change during a trough interaction, are mixed. On a case-by-case basis, a strong positive correlation between TC intensity change and EFC have been found (e.g., Molinari and Vollaro
Molinari and Vollaro (1989) found that eddy fluxes of angular momentum at large radii were highly correlated with TC central pressure tendency 27-33 h later (Table 1.1). In a numerical modeling study, Leroux et al. (2016) found that EFC was experimented with changing the position of a trough interacting with TC Dora (2007) and found that TC–trough distance that maximized EFC was significantly correlated with positive intensity change (correlation coefficient = .59).

Given that EFC is a noisy field in time and varies greatly between TCs Leroux et al. (2016); Komaromi and Doyle (2018); Fischer (2018), composite studies and climatologies are likely to yield more information in general than individual case studies. In their climatology, DeMaria et al. (1993) found that EFC was significantly correlated with positive TC intensity change with a 48 h lag, but only in the 100–300-km and 400–600-km TC annuli, and only after using normalized multiple linear regression to control for SST and vertical wind shear. The normalized regression coefficient was less than about 0.1 for EFC (compared to about -0.35 for wind shear and 0.35–0.6 for SST). Finally, having initially included EFC as a predictor in the Statistical Hurricane Intensity Prediction Scheme (DeMaria and Kaplan 1994), EFC was removed in 2005, as it no longer showed a significant correlation to intensity change (DeMaria et al. 2005).

1.2.2.2 Superposition

The superposition of the trough and TC vortex is another favorable process that has been investigated in trough-interaction studies. During the trough’s approach to the TC, the trough’s high PV field (Hoskins et al. 1985), can be deformed by the TC’s low-PV outflow (Molinari et al. 1995, 1998; Hanley et al. 2001; Leroux et al. 2013, 2016; Archambault et al. 2013, 2015). An example of the deformation of a trough’s high PV by the low-PV outflow of a TC is presented in Figure 1.1. If the TC outflow can modify the approaching trough and reduce its size and depth to a scale similar to that of the TC, the PV of the trough and the
PV of the TC can more effectively superpose (Montgomery and Farrell 1993; Molinari et al. 1995, 1998; Hanley et al. 2001; Kimball and Evans 2002; Leroux et al. 2013, 2016). During superposition, the circulation of the TC core increases as the wind and pressure anomalies associated with the PV anomalies begin to interact, which may lead to TC intensification (Molinari et al. 1995, 1998). Emanuel (1997) showed that a tropopause-based PV anomaly on the same scale as a low-level entropy source, here a TC, would cause an increase in the strength of the TC by PV inversion arguments. Nong and Emanuel (2006) also found that the cyclonic vorticity associated with a moist upper-level PV anomaly can project downward along angular momentum surfaces and cause the TC vortex to spin-up. It is necessary, though, to consider the scale reduction of the trough by the outflow of the TC, because a trough that is large and/or strong, is more likely to weaken the TC upon approach due to the large vertical wind shear that would be expected to accompany such a trough (Molinari et al. 1995, 1998; Hanley et al. 2001; Fischer et al. 2017).

Superposition of an approaching trough’s high PV with the TC core has been found to be associated with intensification in individual cases and climatologies. In their climatology of Atlantic TC–trough interactions, Hanley et al. (2001) found that trough superposition led to TC intensification 78% of the time, and, in fact, there were not enough weakening superposition cases to make a meaningful composite. In the strengthening cases composite, the scale of the trough was found to be similar in size to, or smaller than, the TC, which allowed for the favorable effects of superposition, while reducing the negative effects of shear. In individual cases, Molinari et al. (1995, 1998) found that the intensification of Hurricane Elena (1985) and Tropical Storm Danny (1985) occurred concurrently with the near-superposition of an upper-level trough. In these cases, once the trough was within a Rossby radius, the pressure and wind anomalies of the trough and TC could constructively interact, deepening the TC. Molinari et al. (1995, 1998) added, however, that the intensification was able to occur because of the additional effects of diabatic heating, which eroded the approaching trough, reduced shear, and prevented the trough from crossing the TC and reversing the
In a modeling study by Leroux et al. (2013), and a follow-on sensitivity analysis by Leroux et al. (2016), the superposition of an upper-level trough with Southern Hemisphere TC Dora (2007) was investigated. Leroux et al. (2013) presented a complicated picture of how superposition during TC–trough interaction can lead to intensification. In this case, the trough was not able to closely approach the system due to advection of low-PV air by the TC divergent outflow, which heavily deformed the trough. Over a deep layer (450–200 hPa) beneath the outflow, however, radial inflow into the TC allowed for advection of high-PV air from the trough directly into the TC inner core, leading to an increase in circulation. The formation of this PV advection "tongue" is presented in Figure 1.2. This advection allowed for an expansion of cyclonic winds in the TC throughout the entire troposphere, which led to intensification of the TC. As in previous studies, the authors noted that it was important that the bulk of the trough was kept at a distance, allowing for enhanced outflow and reduced shear during the interaction.

Leroux et al. (2016) tested the sensitivity of the initial TC–trough configuration in the TC Dora interaction event. In this event, TC intensity change was found to be sensitive to the initial relative locations of the trough and TC. They found that simulations that allowed for the development and maintenance of the aforementioned PV channel, along which there is inward advection of PV from the trough to the TC, were most favorable for intensification.

1.2.2.3 TC–jet coupling / Divergence

The coupling of a TC to a strong upper-level jet is another possible positive mechanism for TC intensification during a trough interaction. The development of the jet has two major components. First, the upper-level, cold core of the trough approaches the warm core of the TC. This approach results in an increase of the horizontal temperature gradient and, through thermal wind balance, an upper-level jet develops above this gradient (Molinari and Vollaro
Second, the anticyclonic outflow of the TC and the cyclonic trough result in an area of confluent flow between the two that can result in the formation of a jet (Molinari and Vollaro 1989).

There are multiple hypothesized mechanisms that can influence TC spin-up during coupling with an upper-level jet. First, the TC outflow can directly “tap” into this upper-level jet, which enhances outflow, especially if the jet connects to the synoptic-scale westerly flow (Sadler 1976; Holland and Merrill 1984; Hanley et al. 2001; Rappin et al. 2011; Komaromi and Doyle 2018). Second, the jet can act to increase the strength of the TC secondary circulation. If the TC can remain in the right-entrance region of the upper-level jet, where vertical motion is enhanced by the ageostrophic circulation and convection is favored, there may be an increased evacuation of mass from the TC, resulting in lower surface pressures and an increase in TC intensity (Merrill 1988a,b; Molinari and Vollaro 1989; Shi et al. 1997; Hanley et al. 2001; Leroux et al. 2016; Cowan and Hart 2016; Komaromi and Doyle 2018).

The coupling between an upper-level jet and a TC has been shown in multiple climatologies, and has been found to be positive for intensification. Cowan and Hart (2016) showed that during trough interactions, intensifying TCs were able to couple their outflow directly to the trough as the outflow developed, whereas weakening TCs had a distinct outflow jet that was not as effective at merging with the jet of the trough as it approached. In other words, intensifying TCs developed their outflow within the environment of the trough and jet, unlike weakening TCs that developed well-defined outflow and then had the trough impinge upon it. A complicating factor, though, is that weakening TCs were stronger to begin with and, thus, would be more likely to have a well-defined outflow jet before interaction with the trough began.

Hanley et al. (2001) found that in cases of distant trough interaction, where the PV anomaly of the TC was 400 km or more away from the TC, both intensifying and weakening TCs were located in the favorable right-entrance region of the jet. They found that intensifying and weakening TCs had nearly the same amount of divergence and their positions
relative to the jet were similar. Likewise, Leroux et al. (2016) did not find a systematic relationship between divergence and TC intensity in their modeling sensitivity experiments. One possibility to explain the similar position and strength of the divergence in the weakening and intensifying cases is that the formation of the outflow channel could simply be a byproduct of increased downshear convection due to the approach of the trough. Alternatively, the similarity suggests that the upper-level jet, while favorable, is not the dominant mechanism controlling TC intensity during trough interactions.

1.2.2.4 Vertical wind shear

The last major factor that is often considered in TC–trough interactions is the negative influence of vertical wind shear. During a TC–trough interaction, vertical wind shear is enhanced due to the increase in upper-level winds (Molinari and Vollaro 1989; DeMaria et al. 1993; Hanley et al. 2001). Two main mechanisms have been hypothesized to weaken the TC during periods of enhanced shear: (1) the tilting and misalignment of the TC vortex, and (2) the increased ventilation of the TC inner core (Jones 1995; Frank and Ritchie 2001; Riemer et al. 2010a; Tang and Emanuel 2012).

Enhanced vertical wind shear can cause the TC vortex to tilt downshear, away from a favorable upright configuration, in which the upper part of the vortex is advected downshear of the lower vortex (Jones 1995; DeMaria 1996; Riemer et al. 2010a). When this occurs, the tilted warm core vortex will have a lower column-average temperature than an upright vortex through hydrostatic arguments, resulting in rising surface pressure and decreased TC wind speed (Frank and Ritchie 2001).

Second, shear can act to ventilate the TC with cooler and/or drier environmental air. Three main ventilation mechanisms have been proposed (Tang and Emanuel 2012). The first, as described in Frank and Ritchie (2001), posited that enhanced vertical shear can cause the upper-level warm core to be diluted, as eddies flux warm air away from the TC core. This
eddy flux results in cooling of the upper-level warm core, which causes the TC to weaken through hydrostatic arguments. The second, by Tang and Emanuel (2012), describes that there can be direct entrainment of mid-level, low-entropy air into the TC vortex through the development of shear-induced eddies that bring dry environmental air into the TC eyewall. This results in a disruption of the TC heat engine, proposed by Emanuel (1986), and a weakening of the TC. Finally, Riemer et al. (2010a) proposed a low-level pathway, in which convective downdrafts flood the boundary layer with low-entropy air that is then radially advected inward, resulting in reduced convection and latent heating. While it is unclear which shear mechanism is most detrimental to a TC, it is widely accepted that shear is unfavorable for intensification (Gray 1968; Frank and Ritchie 2001; Riemer et al. 2010a; Tang and Emanuel 2012).

Because trough interactions often lead to the weakening of a TC, rather than strengthening (DeMaria et al. 1993), shear is hypothesized to be the dominant mechanism controlling TC intensity during trough interactions. DeMaria et al. (1993) showed that during trough interactions, shear was strongly negatively correlated with intensity change (normalized regression coefficient of about -0.35) after controlling for EFC and SST (normalized regression coefficients of about 0.1 and 0.35–0.6, respectively). Though no multiple linear regression analysis was performed, Hanley et al. (2001) suggested that shear was the dominant process controlling TC intensity in distant trough interaction cases.

It should be noted, however, that while shear generally inhibits TC intensification, some TCs do still intensify during trough interactions under moderate and high shear (e.g., DeMaria et al. 1993; Hanley et al. 2001; Leroux et al. 2013, 2016; Rios-Berrios et al. 2016; Fischer et al. 2017; Fischer 2018). Leroux et al. (2013, 2016) and Rios-Berrios et al. (2016) showed that in the cases of TC Dora (2007) and Hurricane Ophelia (2011), respectively, even under increased shear, the TCs were able to rapidly intensify. For TC Dora, it was argued that the superposition of a trough and the advection of high PV from the trough to the TC allowed the TC to intensify even under otherwise hostile conditions (Leroux et al. 2013,
In the case of Ophelia, enhanced left-of-shear convection allowed the TC to intensify within seemingly unfavorable conditions by preventing the close approach of the trough and a further increase in shear (Rios-Berrios et al. 2016). In a composite of intensifying TCs, many of which were under high shear due to an upper-level trough, Hanley et al. (2001) found that TCs were able to intensify either by the trough superposing with the TC core, or because the convection of the TC eroded the approaching trough enough to lower the shear and permit intensification. Finally, Fischer (2018) showed that TCs that intensified under moderate to high shear in the presence of a trough had large instability and high moisture, and were associated with anomalously high convective activity compared to TCs of similar intensity.

1.2.3 Idealized numerical modeling

In one of the first, comprehensive, idealized modeling studies of TC–trough interactions, Kimball and Evans (2002) sought to identify which processes were important and what allowed TCs to intensify during an interaction with an upper-level cutoff low. Kimball and Evans (2002) used the Pennsylvania State University–National Center for Atmospheric Research fifth-generation Mesoscale Model (MM5), with 45-km, 15-km, and 5-km nested domains, and explicit convection in the inner domain to model the interaction of a TC-like vortex with an idealized upper-level cutoff low. In additional sensitivity tests, the cutoff low was modified to change its strength in three model runs and its vertical depth in one run. The results were then compared to three runs without the cutoff low: (1) the TC was placed in a quiescent environment, (2) the TC was placed in a low-shear environment, and (3) the TC was placed in a high-shear environment.

The results of Kimball and Evans (2002) present a complicated process of TC intensification in the presence of a cutoff low. They found that troughs that were strong, but vertically shallow, were most favorable for TC intensification, because the PV of the trough was able to persist until it could merge with the TC’s vortex. In this case, even though
vertical wind shear was relatively high, about 10 m s$^{-1}$ as the trough approached, the TC was able to recover once the trough and TC merged. Conversely, troughs that were too weak were eroded quickly by the TC’s convection, and the trough could not merge with the TC. In these weaker trough cases, shear also approached 10 m s$^{-1}$, but the TC experienced all of the negative effects of high shear without the positive effect of superposition. In the case of a vertically deep but weak trough, the trough did not become as deformed by the TC as it approached, which precluded superposition and any weakening of the shear (which exceeded 14 m$^{-1}$, and resulted in weakening of the TC.

The results of Kimball and Evans (2002) do not suggest that weak troughs are more favorable for intensification, unlike Hanley et al. (2001). The modeling results of Kimball and Evans (2002) showing that superposition, when it is able to occur, is favorable for TC intensification do, however, fit with Hanley et al. (2001).

In a more recent idealized modeling study, Komaromi and Doyle (2018) found that there is high sensitivity to the initial TC–trough geometry, consistent with the findings of Leroux et al. (2016). In their study, Komaromi and Doyle (2018) used the Coupled Ocean Atmosphere Mesoscale Prediction System for Tropical Cyclones (COAMPS-TC) model at uniform 5-km grid spacing to investigate the interaction between a TC and a trough in a westerly jet. Komaromi and Doyle (2018) showed that there exists a "sweet spot" (a distance between the TC and trough) for trough interaction in which upper-level divergence is enhanced, but shear remains low. In a series of sensitivity tests, Komaromi and Doyle (2018) found that TC intensification occurred when the TC was located between .2–.3 times the wavelength of the trough in longitude and .8–1.2 times the amplitude of the trough in latitude. This "sweet spot" has been hypothesized as a key component in determining the favorability of trough interactions (Hanley et al. 2001; Leroux et al. 2016).

The main finding of Komaromi and Doyle (2018) was that the reduction in inertial stability due to the approaching trough allowed the TC to develop an enhanced region of poleward-directed outflow, which resulted in TC intensification relative to a "no trough"
simulation. As mentioned, however, the initial placement of the trough relative to the TC proved crucial in determining the favorability of the interaction, as troughs placed too close to the TC prevented TC intensification due to high shear. Additionally, even in the favorable interaction simulation, the TC weakened once the trough too closely approached the TC and increased shear dominated any positive effects of the enhanced outflow. Furthermore, Komaromi and Doyle (2018) found no meaningful connection between EFC and intensity change, and they did not explore superposition. In their conclusions, Komaromi and Doyle (2018) hypothesized that the strength of the TC's convection might be a controlling influence in TC–trough interactions, as strong convection would act to weaken the trough through diabatic erosion, thereby keeping the trough at a farther distance and reducing vertical wind shear.

1.3 Research objectives

Based on prior climatologies, it is unclear whether trough interaction is favorable or unfavorable for intensification due to sample size limitations and/or sample biases. Additionally, climatologies and case studies differ greatly on the mechanisms by which TCs intensify under trough forcing. Shear is widely agreed upon to be the most detrimental factor during trough interactions, but EFC, superposition, and enhanced outflow are all suggested to be important mechanisms depending on the study considered. It is possible that given the wide range of possible TC–trough interaction configurations, any of these mechanisms can control intensity change in nature, depending on the case. This possibility motivates the need for idealized simulations in which different TC–trough interaction configurations can be tested. Thus far, only a few idealized studies have attempted this problem, and the results are mixed, with some results not matching well with observational results, and others focusing only on one intensification mechanism or sensitivity.

Given the lack of clarity on both the likelihood of intensification under trough forcing and how TCs intensify when interacting with troughs, this dissertation seeks to use satellite
observations, atmospheric reanalysis, and idealized modeling to answer fundamental questions about TC–trough interactions. Namely,

1. Accounting for sample bias, are troughs climatologically favorable or unfavorable for intensification compared to instances of no trough interaction and compared to the statistics of intensity change in the Atlantic basin as a whole?

2. Which processes are responsible for TC intensity change under upper-level trough forcing?

3. Are certain trough morphologies more favorable for TC intensification?

This dissertation is organized as follows. Chapter 2 will discuss a climatology of TC–trough interactions in the Atlantic, as well as present composites of favorable and unfavorable trough interactions using satellite observations and reanalysis. Chapter 3 presents the idealized modeling methodology and a comparison between a TC–jet interaction and a TC–trough interaction to determine which of the typical TC–trough intensification mechanisms are most closely tied to intensity change. Chapter 4 describes the results of sensitivity testing to the initial strength and size of the trough in the idealized simulations. Chapter 5 concludes this dissertation with a recapitulation of major findings and opportunities for additional study.
1.4 Tables and figures

Table 1.1: Linear correlation coefficient at various radii between eddy momentum flux and central pressure tendency for various lag times in the life cycle of Hurricane Elena (1985), table and caption adapted from Table 2 in Molinari and Vollaro (1989).

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Table 1.2: Regression coefficients for the multiple linear regression of the normalized values of $P$, vertical wind shear magnitude, and EFC at different lag times prior to the time trough interaction was defined, adapted from Fig. 9 in DeMaria et al. (1993). Here, $P = VSST - V$, in which $V$ is the intensity of the TC in knots and VSST is the maximum potential intensity of the TC from an empirically-defined relationship based on SST (Merrill 1988b). EFC was calculated within multiple annuli. Bold values indicate statistical significance at the 95% confidence level using a standard F statistic.

<table>
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<td>.06</td>
<td>.03</td>
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Figure 1.1: Wind vectors and Ertel potential vorticity on the 345 K isentropic surface at (b) 1200 UTC 30 August; (c) 0000 UTC 31 August; (d) 1200 UTC 31 August; and (e) 0000 UTC 1 September 1985. The potential vorticity increment is 1 PVU, and values greater than 1 PVU are shaded. Wind vectors are plotted each 2.25° latitude-longitude, one-half their resolution in the gridded analyses. The 345 K surface is at approximately 200 mb in the hurricane environment and 240–280 mb at the hurricane center. The tropical storm symbol represents the observed position of Hurricane Elena (1985). Figure and caption adapted from Molinari et al. (1995).
Figure 1.2: Radius–pressure cross sections of negative values of PV radial advection (-uPV; PVU m s⁻¹, shaded) after b (28 h) and d (36 h) of simulation and along 160° azimuth. Cross sections are taken along an azimuth spanning between TC Dora (2007), located at 0 km radius, and the trough with which TC Dora is interacting, located between 800–1200 km and labeled B in the figure. Superimposed are PV contours of -.07, 1.5, and .2 PVU. Arrows represent the radial and vertical (-10 x omega) wind vectors. Dashed gray contours indicate regions of vertical velocity lower than -2 Pa s⁻¹. Figure and caption adapted from Leroux et al. (2013).
2. TC intensity change during TC–trough interactions in ERA-I reanalysis and satellite observations

2.1 Introduction

Previous case studies have shown that there can be large variation in intensity change during trough interaction events, and climatologies are split on whether trough interactions are generally favorable or unfavorable for TC intensification. Given that the longest Atlantic climatology of TC–trough interactions is only 12 years (Hanley et al. 2001), and the only other climatology is three years (DeMaria et al. 1993), there is a need for a more comprehensive study stretching back to the beginning of the satellite era (1979). In addition, only one climatological study (Hanley et al. 2001) has explored, in a composite sense, what environmental factors contribute to a “good” versus “bad” trough interaction event for mature TCs.

The remainder of this chapter is organized into three major parts: the first is a discussion of the methodology and datasets used, the second is a climatology of the favorability of Atlantic TC–trough interactions, and the third presents composite characteristics of favorable versus unfavorable trough interaction events.

2.2 Data and methods

2.2.1 Datasets

Six-hourly data for the years 1979–2014 were obtained from the European Centre for Medium Range Weather Forecasts ERA-Interim (ERA-I) reanalysis data set, which has a horizontal resolution of $0.7^\circ \times 0.7^\circ$ and a vertical resolution of either 25 or 50 hPa for the levels used in this study (Dee et al. 2011). Atlantic TC wind speed and minimum central pressure were obtained from the National Hurricane Center Best Track data set. Gridded cloud top infrared (IR) brightness temperature satellite data was obtained from the GridSat
dataset (Knapp et al. 2011).

2.2.2 Methodology

TC centers were calculated from the ERA-I using the 850-hPa vorticity centroid within a $2.8^\circ \times 2.8^\circ$ box surrounding the Best Track center. This center finding method accounts for the fact that the Best Track center may not coincide with the ERA-I reanalysis center. To be included in this study, TCs can never have been subtropical, and they had to be more than 24 h from landfall or extratropical transition at the time period of interest. The deep-layer, environmental vertical wind shear was calculated between 850 and 200 hPa, and within a 200–800-km annulus centered on the TC by interpolating ERA-I winds onto a $5^\circ \times 50$-km cylindrical grid and averaging over the annulus to remove the shear associated with the TC vortex itself.

To determine whether a trough interaction occurred, distributions of 200-hPa EFC were calculated for 300–600-km and 500–900-km annuli for all eligible time periods, consistent with Hanley et al. (2001) (Fig. 2.1). DeMaria et al. (1993) and Hanley et al. (2001) defined a trough interaction as having 200-hPa EFC $\geq 10$ m s$^{-1}$ d$^{-1}$, and Hanley et al. (2001) added that this EFC threshold be met for two consecutive six-hour periods in a given annulus. This study follows a similar procedure, but a trough interaction is defined as when the EFC was in the upper quartile of the full distribution of EFC calculated for all eligible periods (red areas in Figure 2.1) in either of the annuli for two consecutive six-hour periods. The 75th percentile EFC was 7.4 m s$^{-1}$ d$^{-1}$ for the 300–600-km annulus and 10.0 m s$^{-1}$ d$^{-1}$ for the 500–900-km annulus. “No trough interaction” cases were required to maintain an EFC within $\pm 20$ percentiles of where EFC = 0 m s$^{-1}$ d$^{-1}$, which corresponds to a range of -3.5 to 2.7 m s$^{-1}$ d$^{-1}$ for the inner annulus and a range of -6.4 to 3.2 m s$^{-1}$ d$^{-1}$ for the outer annulus (blue areas in Figure 2.1), for 18 consecutive hours. Trough interactions are thus classified into three types: superposition (EFC $\geq 7.4$ m s$^{-1}$ d$^{-1}$ in the inner annulus), distant interaction (EFC $\geq 10.0$ m s$^{-1}$ d$^{-1}$ in the outer annulus), and no trough interaction.
(defined above). To clarify further, the superposition (distant) interactions must have had two consecutive periods of EFC within the red range of the top (bottom) graph in Figure 2.1, and the no trough interactions must have had three consecutive periods within the blue range of both graphs simultaneously.

A major methodological difference between Hanley et al. (2001) and this study is the removal of a minimum SST constraint. DeMaria et al. (1993) found that the effect of EFC during a trough interaction was largely masked by the effect of the changing underlying SST, and Hanley et al. (2001) restricted TCs to SSTs ≥ 26°C to account for this issue. In lieu of this restriction and to control for changes in the thermodynamic environment of the TC, the current study uses a nondimensional intensity metric. By normalizing TC intensity by the potential intensity (PI) (Emanuel 1986), changes in the underlying SST and outflow layer temperature are taken into account in intensity change calculations. For example, a TC with a wind speed of 40 m s⁻¹ and a PI of 60 m s⁻¹ has nondimensional intensity (I) = 0.67. If this TC moves into a less favorable thermodynamic environment (PI decreases by 20 m s⁻¹) and weakens by 10 m s⁻¹ over the next 24 h, the resulting I of 0.75 suggests that the TC is actually strengthening, in this nondimensional framework. This methodology is a significant point of differentiation from DeMaria et al. (1993) and Hanley et al. (2001), in which the previous example would be considered a weakening case due to the reduction of wind speed.

Figure 2.2 shows the distribution of 24-h nondimensional intensity change (ΔI) for all eligible six-hour periods. The value of ΔI for steady state is hereafter defined as ±0.07, which corresponds to a dimensional intensity change of ~5 m s⁻¹ for a mean PI of ~70 m s⁻¹ in our sample. The TCs were then binned into strengthening (ΔI > 0.07), steady (−0.07 ≤ ΔI ≤ 0.07), or weakening (ΔI < −0.07) cases. The sensitivity of the results to different thresholds of steady state was tested. Using steady state values of ±0.04 and ±0.0, the percentage of TCs that remained steady decreased, as expected, but the proportion of intensifying to weakening TCs remained similar.

For the TC-centered composites presented at the end of this chapter, a different method
for choosing the weakening and strengthening TC sample was used. This new methodology is similar to the analog approach used in Rios-Berrios and Torn (2017), and is used to account for differences in the TCs’ initial PI and intensity, which were not controlled for in the nondimensional intensity methodology, and allows for a comparison of spatial variables that is not highly biased by the fact that TCs that strengthen in the subsequent 24-h period tend to be much weaker and farther from their PI than TCs that weaken. To select the sample, six-hour periods were first grouped into trough interaction categories in the same manner as before. Then six-hour periods were categorized as strengthening (weakening) based on whether their subsequent 24-h wind speed change was greater (less) than 10 (–10) kt. For superposition cases, which will be the focus of the composites, this process produced 339 strengthening periods and 258 weakening periods.

Once the weakening and strengthening trough interactions were grouped, a sample size of 150 was selected for each, such that each strengthening case selected had a weakening analog, and vice versa. To choose the samples and their analogs, first 75 strengthening cases were selected at random with replacement. Each case chosen was matched at random with replacement to a weakening case that had both maximum wind speed and PI within 10 kt of the strengthening case at the start of the 24-h period of interest. This analog method controls for initial intensity and PI. In a few cases, no analog could be found and the case was excluded. The process was then repeated, but the weakening cases were chosen first and then matched to strengthening cases. The reason half of the sample was chosen first from strengthening and the other half was chosen from weakening is because the strengthening cases tended to be much weaker than the weakening cases, and choosing all 150 from strengthening (weakening) first would bias the results toward weak (strong) TCs. This analog method produced a strengthening superposition sample with 96 unique cases (64%) and a weakening sample with 110 unique cases (73%).
2.3 Climatology of TC intensity change during Atlantic TC–trough interactions

Figure 2.3 shows the frequency of intensity change for six-hour time periods in each category: superposition trough, distant trough, no trough, and all eligible time periods. The 95th confidence interval was calculated using a Monte Carlo random resample test with replacement, in which 6146 time periods were drawn at random from the “all eligible” category (6146 total periods) and then sorted into the proper interaction category. This test was repeated 10,000 times.

Generally, trough interactions, either superposition or distant, were unfavorable for intensification compared to all eligible time periods. Not only were TCs less likely to intensify following superposition or distant interactions compared to all eligible times (33%, 32%, and 37%, respectively), but TCs were also more likely to weaken following superposition and distant interactions than all eligible times (18%, 19%, and 13%, respectively). These results cannot be compared directly to those of Hanley et al. (2001) in terms of how trough interactions affect intensity change relative to the full sample of all eligible TCs, as such a category was not included. The raw numbers of intensification and weakening cases can, however, be compared between the trough categories in the current study and Hanley et al. (2001). In this study, superposition (distant) interactions were 45% (29%) less likely to intensify, but also 2% (18%) less likely to weaken than in Hanley et al. (2001). Though both superposition and distant trough interactions were slightly less likely to weaken than in Hanley et al. (2001), trough interactions are still much more unfavorable in the present study considering the larger reduction in strengthening frequency. Likewise, DeMaria et al. (1993) did not provide an all eligible category for comparison, but that study found that 33% of TCs intensified following a trough interaction and 38% weakened. The results of this study agree more with the finding of DeMaria et al. (1993) that troughs are unfavorable for intensification, although the values in the present study are not quite as unfavorable.

Because trough interactions comprise 24% of all eligible periods, the intensification
of TCs following trough interactions was also compared against periods with no trough interaction. Superposition (distant) trough interactions were less likely to intensify than TCs without trough interactions, 33% (32%) compared to 47%, respectively, and superposition (distant) trough interactions were also more likely to weaken than TCs without trough interactions, 18% (19%) compared to 5%, respectively. Thus, trough interactions of any sort make weakening more likely, and strengthening less likely, compared to cases of no trough interaction.

It should be noted that more TCs still intensified than weakened following a trough interaction in this study. Given that this result was also true for all eligible periods, it is likely that the eligibility criteria biased the results toward intensification. As stated, the TCs considered eligible had to have been at least 24 h from land or extratropical transition, both of which could contribute to weakening. Additionally, SSTs were controlled for by the use of nondimensional intensity. Therefore, most of the major environmental weakening mechanisms in TCs were removed, with the exception of vertical wind shear, thus biasing TCs in the sample toward intensification. To account for this bias, results in this study have been considered in relation to all eligible periods. Herein lies another differentiation from both DeMaria et al. (1993) and Hanley et al. (2001), which did not define an all eligible category for comparison, making it difficult to account for sampling biases.

It is likely that two major methodological differences are contributing to the discrepancies between the current study and Hanley et al. (2001). First, in Hanley et al. (2001), the times at which 24-h intensity changes were calculated were not chosen based on when periods of enhanced EFC began. Rather, any time TC pressure started rising or falling, the value of EFC was considered under the assumption that the approaching trough must be solely responsible for the pressure change. Such a methodology presupposes that a TC–trough interaction must cause intensity change, rather than hypothesizing that it could. Given the favorable conditions for eligibility in Hanley et al. (2001) (e.g., high SSTs and not over land), most of the TCs in the Hanley et al. (2001) sample were intensifying. Without an all eli-
ble category, these biases lead to the impression that troughs are overwhelmingly favorable, which the results of this study do not support. The second major methodological difference possibly contributing to the differences in results is the definition of steady state. In Hanley et al. (2001), steady state TCs were defined as having a 0-hPa pressure change over 24 h, much smaller than the uncertainty of operational pressure estimates, which are between 3 and 10 hPa, depending on the strength of the TC and the type of instrument(s) used in the estimation (Landsea and Franklin 2013). Thus, even a 1-hPa pressure change, well within observational uncertainty, would cause a TC to be listed as either intensifying or weakening in Hanley et al. (2001). In contrast, this study defines steady state as $|\Delta I| \leq 0.07$ (or, in a dimensional sense, $\sim 5 \text{ m s}^{-1}$) in 24 h.

To determine whether the differences in intensification frequency are the result of the choice of the steady state criterion or the choice of when to define the initial time period of interest, the trough interaction category threshold values, the eligibility criterion, and the steady state value of Hanley et al. (2001) were applied to our data set to define the categories. Additionally, as in Hanley et al. (2001), pressure was used as the intensity change metric. Unlike Hanley et al. (2001), however, the 24-h intensity change was still calculated from when the period of enhanced EFC began, rather than when pressure started changing. Table 2.1 shows the results of this modification and indicates that troughs are still not as favorable as suggested in Hanley et al. (2001). Therefore, it is not the difference in steady state definition that contributes significantly to the differences between the studies, but rather the difference in how the initial times were defined. Thus, a number of the intensifying trough interaction cases in Hanley et al. (2001) may be intensifying due to factors other than enhanced EFC.

To elucidate the relative importance of EFC compared to vertical wind shear on TC intensity change, the influence of 300–600-km EFC and deep-layer shear is investigated through the joint distribution of nondimensional intensity change (Figure 2.4). The overall pattern, which is similar for 500–900-km EFC (not shown), shows that the gradient in nondimensional intensity change is mostly aligned along the shear axis (horizontal), rather than the EFC
axis (vertical). This pattern indicates that nondimensional intensity change is much more sensitive to shear than EFC, in agreement with DeMaria et al. (1993).

To test whether shear is a better intensity change predictor than EFC, the normalized multiple linear regression technique described in DeMaria et al. (1993) is employed. Multiple linear regression allows for the linear dependence of shear (EFC) to be taken into account while controlling for EFC (shear). Using multiple linear regression analysis on the normalized variables indicates that shear (regression coefficient = −0.28) is a strong predictor of 24-h intensity change. Underscoring the importance of shear, the mean 850–200-hPa shear values for weakening superposition interaction cases are about 1 m s\(^{-1}\) higher than strengthening cases starting from six hours prior to through 24 hours following trough interaction; this result is significant at the 95% confidence level using a Mann–Whitney U test (Table 2.2).

Conversely, 300–600-km EFC (regression coefficient = 0.05) and 500–900-km EFC (regression coefficient = 0.04) are weak predictors of 24-h intensity change, although they do support the theory that EFC is favorable for intensification upon controlling for shear. These results are statistically different from zero at the 95% level using a standard F statistic. These findings are similar to DeMaria et al. (1993), with the exception that the EFC was only statistically significant at the 95% level in DeMaria et al. (1993) for 48-h intensity change, and only for EFC calculated in the 100–300-km and 400–600-km annuli. Consistent with the result that EFC is a weak predictor of intensity change, EFC was removed as a predictor in the Statistical Hurricane Intensity Prediction Scheme (SHIPS) in 2002, because it was no longer a statistically significant predictor (DeMaria et al. 2005).

As mentioned in Molinari and Vollaro (1989), trough interaction is not the only mechanism that acts to increase EFC in TCs; an increase in anticyclonically curved outflow can also lead to an increase in EFC. Therefore, it is possible that an increase in EFC might be caused by TC intensification, rather than a harbinger. To test whether this is true, new weakening and strengthening samples were chosen from all of the eligible six-hour periods. The strengthening (weakening) sample was chosen based on whether an eligible period had
a \( \Delta I > .07 (\leq - .07) \). This methodology is the same as for the strengthening and weakening trough interaction samples discussed previously, but now with no criteria related to the value of EFC. Fig. 2.5 shows the 48-h time series of 300–600-km and 500–900-km EFC relative to the initial time from which \( \Delta I \) was calculated. In both, considering both the mean and interquartile range, the time series reveal that strengthening TCs have an increase in EFC as they intensify. In weakening TCs, EFC remains steady or drops slightly. This signal is stronger in the 500–900-km annulus, in which the outflow curvature is more likely to be shaped by an approaching trough, and, thus, exhibit larger changes in EFC as the outflow increases or decreases. Additionally, EFC at or above the 75th percentile in the weakening cases meets the threshold used to identify trough interaction in the 300–600-km annulus (500–900-km annulus) from -18 h (-24 h) through 24 h, while the trough interaction criterion is met in the strengthening cases in the 500–900-km annulus at 24 h. These results suggest the possibility that EFC is at least partially driven by intensity changes, rather than being the driver.

The results presented in this section indicate that trough interactions, using either dimensional or nondimensional intensity change, are not nearly as favorable as suggested in Hanley et al. (2001), but not as unfavorable as found by DeMaria et al. (1993). EFC, as in DeMaria et al. (1993), is found to be a poor predictor of intensity change, especially compared to vertical wind shear. The importance of wind shear as a predictor of intensification agrees well with previous studies (e.g., DeMaria and Kaplan 1999; Zeng et al. 2010); however, the morphology of shear-inducing troughs during TC–trough interactions has not been well studied, and is the focus of the following section.

### 2.4 Composite analysis of TC–trough interactions

Composite analysis of various ERA-I fields and GridSat IR brightness temperatures was performed on strengthening and weakening trough interaction cases using the analog methodology described earlier. TC-centered composites were used to determine if there
were significant differences in the upper-level troughs or environments with which TCs are interacting, or with the convective evolution of the TC, that would cause some to weaken while others intensify. This section will focus only on the superposition trough-interaction cases, as the distant-interaction cases are largely similar for the parameters shown herein, with the exception of the composite trough being located at a greater distance from the TC. Except where noted, only mean values are presented due to median values being similar in magnitude and statistical significance. Statistical significance was calculated using a bootstrap technique with replacement 10,000 times, and results were deemed significant if they exceeded the 95th confidence interval.

2.4.1 Initial TC state

Composites reveal that the mean latitude of TCs during trough interaction is similar between the strengthening and weakening cases. The mean latitude of the strengthening cases is 24.9°N, while the mean latitude of the weakening cases is 25.0°N. The difference in mean initial latitude between the strengthening and weakening cases is not significant. There is a larger difference in initial longitude, however. The mean longitude of strengthening cases is 62.1°W, compared to 54.7°W in the weakening cases. The difference in initial longitude is significant. Even though the initial longitude is significantly farther west in the strengthening cases, and average SSTs are climatologically higher in this region of the Atlantic (Shapiro and Goldenberg 1998), there is not a significant difference in the mean SST of strengthening (27.7°C) vs. weakening (27.5°C) cases. The longitude could matter, though, in terms of moisture, as this region of the Atlantic also has a zonal gradient of relative humidity Braun (2010); moisture will be analyzed in a subsequent subsection.

The initial nondimensional mean potential intensity of superposition trough interaction cases was also analyzed. The initial mean potential intensity of strengthening superposition cases was 68.1 m s⁻¹ vs. 67.7 m s⁻¹ for weakening cases, and the difference is insignificant. For initial intensity, strengthening superposition trough interaction cases had an initial mean
intensity of 33.9 m s\(^{-1}\), and weakening cases had an initial mean intensity of 34.4 m s\(^{-1}\). These differences are not statistically significant. Twenty-four hours following the time of trough interaction, the mean intensity of strengthening cases was 42.3 m s\(^{-1}\) and 26.8 m s\(^{-1}\) for weakening TCs, which is significantly different. It is noteworthy that 24 h prior to the time of trough interaction, the strengthening cases had a significantly lower intensity than weakening cases (27.2 m s\(^{-1}\) vs. 34.9 m s\(^{-1}\)). This result indicates that TCs that strengthened following trough interactions were already intensifying in the 24 h prior to the trough interaction, while weakening cases remained approximately steady or weakened slightly. This result fits with the results of DeMaria and Kaplan (1994), which showed that persistence was among the top three predictors of future intensity change in the SHIPS model.

As vertical wind shear was found to be the dominant mechanism affecting intensity change in the climatology, it was recalculated for the new analog sample (Table 2.3). Twenty-four hours prior to the time of trough interaction, shear for both strengthening and weakening trough interactions was comparable at 9–10 m s\(^{-1}\). By 12 h before the time of interaction, shear began to differ significantly between weakening and strengthening cases, with the mean shear being 1.4 m s\(^{-1}\) higher in weakening cases. The shear continues to differ more following the start of the trough interaction, and the difference between weakening and strengthening cases peaks at 3.8 m s\(^{-1}\) at 24 h following the time of trough interaction. It is not entirely clear why there are much larger differences in the shear values and differences using the analog method (Table 2.3) compared to the nondimensional intensity change method (Table 2.2). Using either method, the shear values are comparable throughout most of the 48-h period for the strengthening cases, with the analog sample experiencing less shear by the end. The weakening cases have larger differences between the methods throughout, with the analog method associated with greater shear. Regardless, both methods show that shear is significantly higher in the weakening cases through most of the 48-h period.
2.4.2 200-hPa streamlines

Analyses of 200-hPa streamlines show that there are differences in both the trough with which the TCs are interacting and the larger-scale environment in which the TCs are situated throughout the 48 h surrounding the time of trough interaction (Fig. 2.6). In strengthening and weakening cases, the composite trough is analyzed northwest of the TC center. This TC–trough configuration is similar to most TC–trough interaction case studies and climatologies (e.g., Molinari and Vollaro 1989; DeMaria et al. 1993; Molinari et al. 1995; Bosart et al. 2000; Hanley et al. 2001). Recent work by Fischer (2018) suggests that at least three major trough configurations exist, and the northwest trough configuration is, in fact, the least favorable for intensification. Nevertheless, composite analysis may help elucidate why some TCs are able to intensify, while others weaken in a similar composite environment.

The composites show that 24 h prior to trough interaction (Fig. 2.6a,d) the upstream trough is comparable in zonal scale in the weakening and strengthening composites based on the distance between the ridges on either side of the trough. The strengthening composite trough is vertically deeper, and the base of the trough in the strengthening composite is situated farther south relative to the TC than in the weakening composite, which results in flow over the TC that is more meridional than in the weakening composite. The weakening-TC composite shows the base of the trough located almost due west of the TC throughout most of the period, which imparts larger zonal flow over the storm center (3–4 m s$^{-1}$ higher than in the strengthening composite). The weakening cases are also associated with higher wind speeds between the TC and the trough, consistent with the higher vertical wind shear in those cases.

Both strengthening and weakening cases initially have a large anticyclone situated southeast of the TC at the time trough interaction is defined, which is enhancing flow across the TC. During the following 24 h, the anticyclone opens up in the strengthening composite, while it remains closed in the weakening composite. The stronger anticyclone throughout the period in the weakening cases may be causing larger shear over the TC core due to a stronger
environmental geopotential gradient across the TC. To test this hypothesis, the gradient of 200-hPa geopotential was taken across the TC center. Various distances and angles were used to calculate geopotential gradients across the TC center. For the same distance across the TC, the geopotential gradient was 10–20% larger at -24 h and 20–30% larger at 0 h and +24 h in weakening TCs compared to strengthening TCs. These differences were driven both by differences in the trough strength and anticyclone strength. The trough is stronger in the weakening composites throughout the 48 h period, but the differences in trough strength diminish throughout the period until the two troughs have comparable geopotential at 24 h (not shown). The composite anticyclone southeast of the TC is also stronger in the weakening cases throughout the period, but, unlike the trough, the differences in anticyclone strength increase with time (not shown), which causes the geopotential gradient to increase even as the differences in geopotential associated with the trough decrease.

The trough in the weakening cases has a slightly positive tilt 24 h prior to trough interaction, which becomes more neutral with time. The strengthening composite trough, however, starts with an approximately neutral tilt and becomes more negatively tilted by the end of the period. The change in tilt in both cases is indicative of cyclonic wave breaking (Thorncroft et al. 1993). This wave breaking is possibly induced by the interaction of the trough with the convection of the TC, which acts to halt the trough’s progression. Consistent with wave breaking, the strengthening composite shows a more amplified flow pattern at 0 h and +24 h, with the trough digging southwest of the TC and the ridge building to the north and northeast. This pattern may be indicative of stronger convection over and near the TC center in the strengthening cases. The increased shear over the weakening TC may be inhibiting upshear convection between the TC and the trough, which could explain why the trough shows less of a wave-breaking signature than the strengthening composite. This hypothesis will be explored in a later section.
2.4.3 PV

Figure 2.7 shows the composite PV structure of intensifying TCs during the 48-h period surrounding the time trough interaction was defined. Focusing on the 200-hPa, plan-view composites (Fig. 2.7a–c), the trough 24 h prior to the time of trough interaction is neutrally tilted. The trough also appears to be meridionally incoherent, as evidenced by the difference in curvature of the 1.5-PVU and 2-PVU contours, in which trough curvature is present in the 1.5-PVU contour, but is missing in the much flatter 2-PVU contour. In fact, there appears to be weak ridging in the 2-PVU contour north of the TC (located northeast of “A” in Fig. 2.7a). There is a gradient of PV across the TC, with low-PV values southeast of the TC and high-PV values associated with the trough to the northwest. This strong PV gradient, and the flow it induces, is consistent with the shear experienced by the TC.

At the time of trough interaction (Fig. 2.7b), the trough has dug and moved farther eastward relative to the TC location, increasing the gradient of PV across the TC and changing the gradient direction to be more zonal. Additionally, there is evidence that the trough is beginning to cyclonically break as high PV associated with the trough is beginning to wrap around the southern portion of the TC, while the progression of the trough has been halted to the north, likely due to the convection of the TC (explored in a later section). To the east-northeast of the TC, low-PV values are being advected northward and/or low-PV values are being diabatically generated due to latent heat release. The slight ridging north of the TC that was present at -24 h, located just east of “C” in (Fig. 2.7b), is now more evident as the TC has moved northward toward this ridge, and the trough at 0 h is situated directly south of this ridge. This setup may be beneficial to the TC because the ridge may be blocking the reinforcement of the trough as the TC convection erodes it, which would result in decreasing shear.

Twenty-four hours after the time of trough interaction, the wave breaking pattern continues and is even more enhanced, as the downstream ridge due east of the TC location has expanded farther north and has lower PV values than at 0 h, while the upstream trough
has dipped farther south and has higher PV values than at 0 h (Fig. 2.7c). The high-PV values of the trough are now well to the south of the latitude of the TC, while low-PV values are wrapping cyclonically around to the north and east of the TC. This evolution results in a PV gradient that induces southeasterly flow over the TC, which could be responsible for the significantly reduced shear in the 24 hours following the time of trough interaction due to the induced flow becoming better aligned with the easterly mean flow in this region. The high-latitude ridge north of the TC, now located just east of “E” in Fig. 2.7c, has moved eastward of the TC, and is perhaps being enhanced by the ridge building occurring directly to its south due to the TC itself.

The strengthening TC–trough interaction vertical cross section composites are presented in Figure 2.7d–f. Note that the evolution of the PV tower of the TC in the cross section composites was not examined, as Schenkel and Hart (2011) showed that TC structure is not well represented in reanalysis datasets. Between 24 hours prior to trough interaction (Fig. 2.7d) and the time of trough interaction (Fig. 2.7e), the cross sections show the progression of the trough toward the TC. The cross sections show relatively little change in the structure of the trough, rather showing the TC and trough moving closer together and the signature of the trough extending over top of the TC PV tower. The closer approach of the trough corresponds with a slight increase in vertical wind shear (Table 2.3). Twenty-four hours after the time of trough interaction (Fig. 2.7f), the trough now extends farther down in the troposphere (the 2-PVU contour has moved from 200 hPa to 230 hPa) and extends slightly more over the TC core. The strength of the trough in the cross section has also increased. Above the PV tower of the TC, there is evidence of the ridge building seen in the plan view composite. There is a kink in the 1.5 PVU contour associated with the approaching trough that indicates the raising of this PVU contour as the ridge atop the TC builds, even while the trough tries to approach the TC.

The 200-hPa weakening composite PV plan-view (Fig. 2.8a,b,c) shows a different evolution of the approaching trough compared to the strengthening case. The most striking
difference between the strengthening and weakening composite is 24 h prior to the time of trough interaction (Fig. 2.8d). The weakening composite shows an anticyclonic wave breaking event associated with a high-amplitude ridge just upstream of the trough with which the TC will interact. This large ridging event could result in enhancement and maintenance of the trough as high-PV air is driven southward; this hypothesis will be addressed in the next section. Unlike in the strengthening composite, the trough in the weakening composite is meridionally coherent into high latitudes. As in the strengthening interaction composite, there is a large gradient in PV across the TC resulting in relatively large shear values.

The trough in the weakening cases is zonally wider and extends farther southward relative to the TC location than in the strengthening cases. Based on the 1.5-PVU contour, the trough 24 h prior to the time of trough interaction has a half wavelength of about 2400 km, compared to about 950 km in the strengthening composite. The trough remains wider from 24 h before trough interaction to the time of trough interaction (Fig. 2.8a,b), before the zonal scale becomes more similar to the trough in the strengthening composite 24 h following interaction (Fig. 2.8c). It has been suggested that larger troughs are unfavorable for intensification because of the high shear they induce and the long time it can take for TCs to diabatically erode the high PV over a larger area (Molinari et al. 1995, 1998). This argument is supported by shear values in the weakening composite that are significantly higher than in the strengthening composite throughout most of the trough interaction event (Table 2.3).

24 h before the time of trough interaction there is a large ridge upstream of the TC and trough in the northwestern corner of the weakening composite (Fig. 2.8a). This ridge was initially hypothesized to be important in building and reinforcing the trough as the TC diabatically eroded it, however, a later section will show that this ridge does not contribute to the strength or maintenance of the trough through PV advection during the time period considered.

At the time of trough interaction (Fig. 2.8b), the trough has dropped south, but
the southern portion of the trough does not appear to be wrapping around the TC as in
the strengthening case. By the end of the following 24 h (Fig. 2.8c), the trough has lost
meridional coherency, based on the 1.5-PVU and 2-PVU contours. Cyclonic wave breaking
also is occurring, as in the strengthening composite, though the ridge building downstream
of the TC is not nearly as large as in the strengthening composite. As at earlier times,
the trough does not wrap around to the south of the TC as much as in the strengthening
composite, but instead the trough has thinned and begun to form a cutoff feature due west
of the TC.

The PV cross sections of the weakening composites (Fig. 2.8d–f) show that 24 h prior
to the time of trough interaction (Fig. 2.8d), the trough is stronger and vertically deeper
than the strengthening composite. The weakening composite has PV values of 1.5 PVU
reaching down to nearly 200 hPa, compared to 175 hPa in the strengthening composite. At
the time of trough interaction (Fig. 2.8e), the weakening composite still has a stronger and
vertically deeper trough, and the TC and trough have approached one another. Twenty-four
hours following the time of trough interaction (Fig. 2.8f), the vertical depths and magnitudes
of the troughs are similar to the time of trough interaction, but the trough has higher PV
values over top of the PV tower of the TC, indicating that the ridging of the TC is not strong
enough to resist the close approach of the trough.

To further illustrate important differences between the composites, Figure 2.9 shows
the difference fields between the strengthening and weakening composites of the 200-hPa
PV and PV cross sections. Focusing first on the plan-view, 200-hPa PV differences (Fig.
2.9a–c), there are large differences between the weakening and strengthening cases. Twenty-
four hours prior to trough interaction (Fig. 2.9a), the trough in the weakening composite
is much stronger and larger than in the strengthening composite. At the time of trough
interaction (Fig. 2.9b), the differences between the trough strength and scale have lessened,
but the weakening trough still is stronger and larger northwest and west of the TC location.
South of the TC, however, the strengthening composite shows an area of higher PV. This area
corresponds to increased wrapping of the trough around the TC as wave breaking occurs. Fischer (2018) found that in rapidly intensifying TCs associated with trough interaction, high-PV values southwest of the TC are more favorable for intensification compared to high-PV values situated either northeast or northwest of the TC. Twenty-four hours after the time of trough interaction (Fig. 2.9c), the same pattern is evident, and fits with the wave breaking occurring in the strengthening composite. Considering the cross sections (Fig. 2.9d–f), weakening TCs are associated with stronger, vertically deeper troughs. These deeper and stronger troughs are likely responsible for the higher shear the weakening TCs experience relative to the strengthening TCs. 24 h after the time of trough interaction (Fig. 2.9f), the higher PV values over top of and just northwest of the TC location in the weakening cross section composite reveal that the trough is able to more closely approach the TC core, likely due to the reduced ridge building compared to the strengthening composite discussed previously.

### 2.4.4 Ventilation

Because upper-level troughs are typically characterized by high shear and cooler, drier air relative to a TC, the presence of a trough may limit the environmental favorability for TC intensification during a trough interaction. It is possible that with large differences in the scale of the trough initially, as shown previously, there may be differences in the thermodynamic impacts of these troughs as well. A useful tool to measure the combined effects of the thermodynamic favorability and shear is the ventilation index of Tang and Emanuel (2012):

\[
\Lambda = \frac{u_{\text{shear}}\chi_m}{u_{PI}}
\]

where \(u_{\text{shear}}\) is the 850–200-hPa vertical wind shear magnitude, \(\chi_m\) is the azimuthally-averaged, mid-level, nondimensional entropy deficit (in this case, 600 hPa), and \(u_{PI}\) is the potential intensity of the TC. Higher ventilation indices indicate that more ventilation of the TC core with lower-entropy air from the environment may be occurring, which is detrimental
to TC intensification.

Figure 2.10 shows the median and various percentiles of the ventilation index for strengthening and weakening TCs from 24 h before trough interaction to 24 h after. The median values were used in lieu of the mean because the mean falls above the interquartile range during some time periods in both strengthening and weakening cases, and, thus, it must be being skewed by large outlier values. The median ventilation index values are statistically significantly higher in the weakening cases compared to the strengthening cases beginning 18 hours prior to the time of trough interaction until 24 hours after. Notably, there is a strong uptick in the ventilation index in weakening cases between -6 h and 0 h, as the trough approaches the TC. The difference in median entropy deficit of strengthening and weakening cases is statistically significant throughout the entire period (not shown), indicating that the environments of weakening TCs are drier and/or cooler than those of strengthening TCs, a similar finding to Rios-Berrios et al. (2016). The differences in the median shear become significantly different starting at -6 h and remain significantly different for the remainder of the period. Because the analog method controlled for potential intensity at the time of trough interaction, the PI between the strengthening and weakening composites is not significantly different, except at -18 h.

The mid-level entropy deficit is examined without azimuthally averaging to better understand the thermodynamic environment within which the TC is embedded. This “local” entropy deficit at 600 hPa is shown for the strengthening and weakening trough interaction cases in Figure 2.11. The data have been shear-rotated such that the shear vector points due east. The thermodynamic favorability of the environment the TC is embedded within is statistically significantly more unfavorable for the weakening TCs during trough interaction at all points outside of 50 km from the TC center (not shown). The largest differences in the favorability of the environment are in the upshear direction, which has been shown to be important for TC intensification (Rios-Berrios et al. 2016). The much more unfavorable environment upshear likely impedes convective growth in the region between the TC and the
trough, which will be discussed further in the next section. The cause of the more unfavorable environment in the weakening composite could be either the approaching trough, and/or, as mentioned previously, the mean location of the weakening TC composite within the basin, wherein there is a large zonal gradient of relative humidity (Braun 2010).

2.4.5 Infrared brightness temperature

To investigate whether the more favorable upshear environment of the strengthening cases results in stronger and more widespread convection, the shear-relative, IR brightness temperature, a proxy for convective strength, in weakening and strengthening TCs was compared (Fig. 2.12). Twenty-four hours prior to the time of trough interaction, the strengthening and weakening composites are similar. Considering that strengthening TCs are on average 15 kt weaker than weakening TCs at this time, the fact that the two are so similar suggests one of the following is true: (1) strengthening TCs have anomalously low IR brightness temperatures relative to TCs of similar strength, (2) the weakening TCs have anomalously high IR brightness temperatures relative to TCs of similar strength, or (3) some combination of the previous two. This result is consistent with the findings of Fischer (2018) and Fischer et al. (2018), who found that TCs with anomalously low IR brightness temperatures compared to TCs of similar strength are more likely to intensify. The displaced area of low IR brightness temperatures downshear is consistent with moderately/highly sheared TCs (Corbosiero and Molinari 2003).

By the time of trough interaction, strengthening TCs have cooler IR brightness temperatures in the upshear left quadrant compared to weakening TCs. These differences are driven by weakening convection in the weakening TCs and strengthening convection left of shear in the strengthening TCs. Enhanced convection upshear, and increased convective symmetry, has been shown to be favorable for TC intensification (Shapiro and Willoughby 1982; Vigh and Schubert 2009; Rogers 2010; Stevenson et al. 2014; Fischer et al. 2018). The same pattern continues in the following 24 h, as the strengthening TCs have a greater degree
of symmetry and cold IR brightness temperatures near and over the TC center, while weakening TCs experience significant cloud-top warming, consistent with Tao and Jiang (2015) and Tao et al. (2017).

To address how IR brightness temperatures are changing in time within each composite individually, the IR brightness temperature 24 h before trough interaction is subtracted from each time lag up to 24 h after interaction (Fig. 2.13). The weakening composites show that convection in the inner 200 km, near the TC core, deteriorates throughout the period. Any new convection that develops in the weakening composite is generated well downshear and downshear-left from the TC center. This convective pattern is consistent with other observational and modeling results of sheared TCs (Corbosiero and Molinari 2003; Riemer et al. 2010b). The heating of this distant convection is well-removed from the core, and, thus, the effect on TC intensity is likely to be low, as previous studies have shown that convection in the TC inner core, especially symmetric convection, is associated with intensification (Shapiro and Willoughby 1982; Vigh and Schubert 2009; Rogers 2010; Stevenson et al. 2014; Fischer et al. 2018).

Strengthening composites show that prior to the time of trough interaction, convection is weakening in the upshear semicircle of the TC, likely a result of the approach cold/dry trough and its associated increased shear. By the time of trough interaction, the composites show the strengthening convection wrapping around the TC toward the upshear semicircle and convection being generated close to the TC center. The intensifying convection within 200 km of the TC center left of shear and upshear continues throughout the time period, which would be favorable for TC intensification if some of this deepening convection is within the radius of maximum wind. Convection wrapping upshear has been shown to be favorable for TC intensification because it acts to realign the TC vortex and results in more symmetric heating (Moon and Nolan 2010; Rogers 2010; Onderlinde and Nolan 2014, 2016; Fischer 2018). Additionally, enhanced convection occurring between the TC and the trough can result in a scale reduction of the trough, a process that has been shown to reduce the
amount, and duration, of shear over the TC and potentially allow for superposition of the trough and TC (Montgomery and Farrell 1993; Molinari et al. 1995, 1998; Kimball and Evans 2002). By 24 hours following the time of trough interaction, the strengthening composite shows a pronounced decrease in IR brightness temperatures close to the TC center, with much of this decrease occurring in the upshear semicircle.

### 2.4.6 Divergence

Divergence is inherently tied to the convection of the TC and to ageostrophic flows associated with synoptic scale features (e.g., troughs). Thus, it stands to reason that because there are differences in the convective evolution of the TC, there may also be important differences in the TC’s divergence. The composite divergence and total wind speed for the strengthening and weakening composites is shown in Figure 2.14. The weakening composite of total wind speed (Fig. 2.14d–f) suggests that a strong, well-defined outflow jet exists northeast of the TC center 24 h prior to the time of trough interaction, and there exists a clear separation between the TC outflow jet and the larger-scale strong westerly flow at high latitudes, which has been shown to be unfavorable for TC intensification compared to outflow jets that connect to the large-scale flow Sadler (1976). By the time of trough interaction, this strong outflow jet is maintained. These results are consistent with Cowan and Hart (2016), who found that weakening TCs generally have a well-defined outflow jet that is impinged upon by the approaching trough. As in that study, it is not clear if the separated structure of the outflow jet from the environment is a unique feature of TCs more likely to weaken, or if it is simply a byproduct of the weakening TCs having a higher mean initial intensity. By 24 h following the time trough interaction was defined, the outflow jet weakened, and there is no longer a discernible outflow jet connected to the TC in the weakening composite. Divergence also substantially decreases 24 h after the time of trough interaction, consistent with both TC weakening and warming IR brightness temperatures presented in the prior section.
The strengthening composite of total wind speed (Fig. 2.14a–c) suggests a different evolution of the TC outflow compared to the weakening composite. Twenty-four hours prior to the time of trough interaction, the TC has a weak outflow jet. This outflow development is again consistent with Cowan and Hart (2016), who found that in strengthening trough interactions, the outflow of the TC is weak prior to trough interaction, and grows stronger as the trough approaches. Cowan and Hart (2016) also found that in intensifying TCs, the TC-outflow jet connects to the jet associated with the trough as it develops, but it is difficult to discern if that is the case in this composite. By 24 h after the time of trough interaction, the strengthening TC composite still has a distinct outflow jet, and the jet streak has elongated away from the TC.

Concerning divergence, it is important to first consider the limitations of the composite methodology. Disentangling the cause of divergence is difficult, as it could be caused by the trough or the convection of the TC. Indeed, there is strong and constant coupling between these two mechanisms, as divergence downstream of the trough, particularly in the right-entrance region of the jet, can support lift and convection, and strong convection can enhance divergent outflow. Twenty-four hours prior to and after trough interaction, the differences in divergence magnitude are likely tied to the intensity of the TC, as the strengthening TCs are weaker initially than the weakening TCs, but are stronger at the end of the period.

As in Hanley et al. (2001), the maximum divergence in the TC is displaced from the TC center toward the north and east, consistent with the west-southwesterly shear associated with the approaching trough. Twenty-four hours before the time trough interaction was defined, the strengthening TCs, as expected, have generally less divergence than weakening TCs (Fig. 2.14g). At the time of trough interaction (Fig. 2.14b,e,h), the divergence maxima are comparable between the strengthening and weakening composites; however, the strengthening TCs have greater divergence near and just east of the TC center (Fig. 2.14h). The strengthening composite also shows that the divergence maximum has rotated cyclonically, matching the cyclonic rotation in the minimum of the IR brightness temperatures (Fig.
2.12). By 24 h following the time of trough interaction (Fig. 2.14i), the strengthening composite has much greater divergence over and around the TC than the weakening composite, consistent with its greater intensity. Oddly, compared to the time of trough interaction, the strengthening composite has weaker divergence at 24 h, even though it is substantially stronger. This result is not the case throughout the entire 24 h period after trough interaction. It could be the case that the weakening of divergence at 24 h is due to the TC moving into higher latitudes, and more unfavorable conditions, and may be the start of weakening.

The composites of divergence and total wind speed taken together provide some insight into favorable and unfavorable TC–trough interactions. Given the initial, higher mean intensity of the weakening TCs and the suggestion in the composites that the TC outflow is stronger and better defined prior to the approach of the trough, well-defined outflow and a greater initial TC intensity do not make the TC necessarily better suited to fend off an approaching trough. Indeed, the opposite is suggested, wherein a weaker initial TC with weak outflow, and a weaker approaching trough, is more likely to intensify. Thus, it is possible that stronger TCs may be more sensitive to approaching troughs than weaker TCs.

### 2.4.7 PV advection by the irrotational wind

To assess the cause of the cyclonic break of the trough in the strengthening and weakening PV composites, and whether the convective outflow and associated divergence of the TC may be causing a reduction in the strength and/or scale of the trough, PV advection by the irrotational wind (Archambault et al. 2013, 2015) was analyzed (Figs. 2.15; 2.16). Twenty-four hours prior to trough interaction (Fig. 2.15a,d; 2.16a), the initial difference in TC intensity is clear in the divergent wind field. Though both the weakening and strengthening TCs have substantially stronger divergent outflow in the northern and eastern quadrants of the TC compared to
other quadrants, consistent with the predominately west-southwesterly shear, the weakening TC composite has divergent outflow nearly twice as strong as the strengthening composite. The negative-PV advection by the irrotational wind between the TC and trough 24 h before trough interaction is much weaker in the strengthening composite (Fig. 2.16a), which is consistent with both weaker TC outflow and a weaker PV gradient between the TC and the trough in the strengthening composite.

Another notable difference is found upstream of both the TC and the approaching trough at this time. In the northwestern part of the weakening composite, where the aforementioned anticyclonic wave breaking event is occurring, there is enhanced divergent outflow associated with the wave break. In the previous section, it was hypothesized that this event could be acting to enhance the trough in the weakening composite, however, there is no signal of meridional, positive-PV advection into the trough. Thus, it is unlikely at the time considered that this upstream wavebreaking event is enhancing the trough through positive PV advection, though it could have driven and enhanced this trough prior to the time period considered.

At the time of trough interaction (Figs. 2.15b,d; 2.16b) the divergent wind in the composites is comparable in magnitude east-northeast of the TC. North and west of the TC, between the TC and the trough, the strengthening composite has 2–4 m s$^{-1}$ stronger divergent wind. This result coincides with stronger negative PV advection between the TC and the trough to the north and west of the TC in the strengthening composite, and stronger negative PV advection closer to the TC center (Fig. 2.16b). The orientation of the divergent wind due west of the TC in the strengthening composite is more perpendicular to the PV contours compared to the weakening composite, acting to enhance the negative-PV advection occurring there. Stronger negative-PV advection between the TC and the trough is consistent with the enhanced ridge building in the strengthening composite discussed previously and the closer approach of the trough to the TC in the weakening composite. The location of the negative-PV advection by the irrotational wind to the northwest of the
TC in both composites is consistent with a slowing of the progression of the northern portion of the trough and forcing the cyclonic wave break seen in the PV fields. In both composites, there is evidence of the downstream ridge building during the cyclonic wave break to the northeast of the TC.

Twenty-four hours after the time of trough interaction (Figs. 2.15c,e; 2.16c), the divergent wind in the strengthening composite remains relatively unchanged. In the weakening composite, however, the characteristic starburst pattern of the TC outflow has disappeared, due to the substantial weakening of the TC. Consistent with high shear, the divergent outflow is located almost entirely to the east of the TC. The pattern of negative-PV advection has changed in both the strengthening and weakening composites. In the strengthening composite, the maximum magnitude of the negative-PV advection has rotated such that the maximum is now west-northwest of the TC center, and the area of negative PV advection extends slightly further to the south. This change is consistent with the elongation and wrapping of the trough. In the weakening composite, the negative-PV advection has decreased in magnitude and is primarily north of the TC, and no longer as large to the west. In all directions near the TC center, the strengthening composite has greater negative-PV advection (Fig. 2.16c). The negative-PV advection associated with the ridge building north and northeast (downstream) of the TC is also still present in the strengthening composite, whereas it has mostly disappeared in the weakening composite.

Even though negative-PV advection is greater in magnitude in the weakening composite 24 h before trough interaction (Figs. 2.15a,d; 2.16a) and comparable in magnitude at the time of trough interaction (Figs.2.15b,e; 2.16b), the TC still weakens. This result suggests that a greater magnitude of negative-PV advection by the irrotational wind does not determine the favorability of a trough interaction early on. Rather, the greater magnitude of the negative-PV advection seems to be a byproduct of the greater initial intensity of the weakening TCs and the greater strength of the approaching trough in the weakening cases. The outflow of the TC does cause the trough to cyclonically break in both cases, but the initially stronger
trough in the weakening composite remains stronger throughout the 48 h period, regardless of the magnitude of negative-PV advection. Thus, it seems the initial strength of the upstream trough is an important factor in determining the favorability of a TC–trough interaction, as the weaker trough in the strengthening composite is more easily deformed and diabatically eroded by the TC convection, which also results in reduced shear.

2.5 Conclusions

This chapter addressed the discrepancies between previous trough interaction climatologies, provided an updated climatology using the ERA-I, and assessed processes that may affect or be associated with TC intensity change during TC–trough interactions. The climatology presented in this chapter compared trough interaction favorability to times of no trough interaction and, most crucially, relative to the intensity change climatology of all eligible times in the Atlantic basin. Trough interactions were shown to be more unfavorable for TC intensification in both instances. The results indicate that trough interactions, using either dimensional or nondimensional intensity, are not nearly as favorable as suggested in Hanley et al. (2001), nor are they as unfavorable as found by DeMaria et al. (1993).

Of the four processes thought to most affect TC intensity change during trough interactions, shear appears to be the dominant mechanism. Wind shear is significantly lower in cases of favorable trough interaction. The importance of wind shear as a predictor of intensification agrees well with previous trough interaction and non-trough interaction studies (e.g., DeMaria et al. 1993; Hanley et al. 2001; DeMaria and Kaplan 1994; Zeng et al. 2010). As in DeMaria et al. (1993), EFC is found to be a poor predictor of intensity change. Even controlling for the influence of shear and SST, EFC is only marginally beneficial to TC intensity change. In composite plan plots, divergence magnitude was shown to be lower in intensifying cases than weakening cases as the trough approached, but slightly higher following trough interaction, similar to the results of Hanley et al. (2001).

The composite morphology of the trough during TC–trough interactions has not been
well-studied, apart from Hanley et al. (2001), and was investigated further in this chapter. The reanalysis composites show that the morphology and evolution of the upstream trough are important in determining the favorability of TC–trough interaction. Generally, favorable trough interactions are found to be those that have a shallower, weaker, and horizontally smaller trough compared to unfavorable interactions, and these features are associated with less wind shear. These results are in line with those found in Hanley et al. (2001). Favorable interactions are not only associated with less shear, but are also associated with a more favorable thermodynamic environment surrounding the TC. The combination of a more favorable environment and lower shear corresponds with less ventilation in the intensifying cases than the weakening cases. Consistent with less ventilation and a more favorable environment, convection is stronger and occurs closer to the TC core in intensifying cases. Convection is also able to wrap upshear, which likely aids in intensification.

The composite results of the divergence and PV advection by the irrotational wind support the conclusion that the initial strength of the trough is important in determining the favorability of the interaction. During the approach of the trough, the strengthening composite has weaker divergence and smaller values of negative-PV advection compared to the weakening composites. Nevertheless, the larger initial magnitude of negative-PV advection, driven by greater divergence and a stronger trough, in the weakening cases does not prevent the close approach of the more unfavorable trough and its concomitant enhanced shear and hostile environment.

Though the characteristics of the approaching trough have been shown to influence the favorability of TC–trough interactions, processes such as superposition, vortex tilt reduction, and internal convective processes, to name a few, may also be important factors in determining the favorability of trough interactions. Indeed, differences in IR brightness temperatures strongly suggest large differences in convective evolution exist and may matter in determining how unfavorable a trough will be to a TC. Those mechanisms will be explored in the next chapter using high-resolution modeling.
2.6 Tables and figures

Table 2.1: Percentage of six-hour periods that strengthen, remain steady, or weaken for each trough interaction category using eligibility requirements, trough interaction thresholds, and steady state definition of Hanley et al. (2001) with this study’s data set. The number of six-hour time periods is listed beside the category name. Values in parentheses are the corresponding values from Hanley et al. (2001), which did not include an all eligible category.

<table>
<thead>
<tr>
<th>Category</th>
<th>% Strengthen</th>
<th>% Steady</th>
<th>% Weaken</th>
</tr>
</thead>
<tbody>
<tr>
<td>Superposition interaction, 464 periods</td>
<td>54 (78)</td>
<td>12 (2)</td>
<td>33 (20)</td>
</tr>
<tr>
<td>Distant interaction, 907 periods</td>
<td>51 (61)</td>
<td>12 (2)</td>
<td>36 (37)</td>
</tr>
<tr>
<td>No trough present, 1368 periods</td>
<td>65 (82)</td>
<td>17 (9)</td>
<td>17 (9)</td>
</tr>
<tr>
<td>All eligible periods, 5425 periods</td>
<td>57</td>
<td>27</td>
<td>15</td>
</tr>
</tbody>
</table>
Table 2.2: Mean deep-layer shear values for strengthening and weakening TCs relative to the time trough interaction was defined (lag = 0). Values in bold indicate strengthening and weakening values are statistically significantly different at the 95% confidence level based on a Mann–Whitney U Test.

<table>
<thead>
<tr>
<th></th>
<th>-24h</th>
<th>-18h</th>
<th>-12h</th>
<th>-6h</th>
<th>0h</th>
<th>6h</th>
<th>12h</th>
<th>18h</th>
<th>24h</th>
</tr>
</thead>
<tbody>
<tr>
<td>Strengthening</td>
<td>9.5</td>
<td>9.8</td>
<td>9.9</td>
<td>10.0</td>
<td>10.0</td>
<td>9.9</td>
<td>9.6</td>
<td>9.3</td>
<td>9.2</td>
</tr>
<tr>
<td>Weakening</td>
<td>9.3</td>
<td>9.9</td>
<td>10.4</td>
<td>10.8</td>
<td>11.1</td>
<td>10.9</td>
<td>10.5</td>
<td>10.3</td>
<td>10.3</td>
</tr>
</tbody>
</table>
Table 2.3: Mean deep-layer shear values for strengthening and weakening TCs relative to the time trough interaction was defined (lag = 0). Values in bold indicate strengthening and weakening values are statistically significantly different at the 95% confidence level based on a bootstrap test with replacement repeated 10,000 times.

<table>
<thead>
<tr>
<th></th>
<th>-24h</th>
<th>-18h</th>
<th>-12h</th>
<th>-6h</th>
<th>0h</th>
<th>6h</th>
<th>12h</th>
<th>18h</th>
<th>24h</th>
</tr>
</thead>
<tbody>
<tr>
<td>Strengthening</td>
<td>9.6</td>
<td>10.1</td>
<td>9.9</td>
<td>10.1</td>
<td>10.0</td>
<td>9.7</td>
<td>9.1</td>
<td>8.8</td>
<td>8.5</td>
</tr>
<tr>
<td>Weakening</td>
<td>9.8</td>
<td>10.5</td>
<td>11.3</td>
<td>11.8</td>
<td>12.2</td>
<td>12.4</td>
<td>12.4</td>
<td>12.1</td>
<td>12.3</td>
</tr>
</tbody>
</table>
Figure 2.1: Distributions of 200-hPa EFC in the 300–600-km annulus (top) and 500–900-km annulus (bottom) for all eligible six-hour time periods. Red shading indicates EFC ≥ 75th percentile. Blue shading indicates EFC within ±20 percentiles of EFC = 0 m s⁻¹ d⁻¹.
Figure 2.2: Distribution of 24-h nondimensional intensity change, $\Delta I$, for all eligible six-hour periods. Red shading indicates strengthening, blue shading indicates weakening, and tan shading indicates steady state.
Figure 2.3: The frequency of six-hour periods that strengthen, remain steady, or weaken in the subsequent 24 h for each of the trough interaction categories: superposition trough interaction (1047 periods, blue), distant trough interaction (1133 periods, red), no trough interaction (790 periods, green), and all eligible times (6146 periods, black). Circles indicate the observed frequency and bars indicate the 95th confidence interval calculated using a Monte Carlo random resample test. Stars indicate values of Hanley et al. (2001) for the corresponding categories. Note: Hanley et al. (2001) did not provide an all eligible category equivalent.
Figure 2.4: Mean 24-h nondimensional intensity change (shaded) for all eligible TC periods binned by 850–200-hPa vertical wind shear and 300–600-km eddy flux convergence at 200 hPa. Black indicates no data.
Figure 2.5: EFC in the 300–600-km (top) and 500–900-km (bottom) annuli for all eligible six-hour periods that strengthen ($\Delta I > 0.07$) and weaken ($\Delta I < -0.07$) in the following 24 h. Red (blue) lines indicate strengthening (weakening) means, and shaded regions indicate interquartile range.
Figure 2.6: 200-hPa, TC-centered, plan-view, composite streamlines (shaded; m s\(^{-1}\)) of strengthening (top row) and weakening (bottom row) trough interaction cases 24 h prior to the time trough interaction was defined (a,d), the time trough interaction was defined (b,e), and 24 h following the time trough interaction was defined (c,f). Red star indicates TC position.
Figure 2.7: 200-hPa, TC-centered, plan-view, composite potential vorticity (shaded; PVU) (top row) and vertical cross sections (bottom row) of strengthening trough interaction cases 24 h prior to the time trough interaction was defined (a,d), the time trough interaction was defined (b,e), and 24 h following the time trough interaction was defined (c,f). Vertical cross section runs along the white, dashed line shown in the plan views. Red star indicates TC position. Black dotted, dashed, solid lines are the 2-PVU, 1.5-PVU, and 1-PVU contour, respectively.
Figure 2.8: As in Figure 2.7 but for the weakening cases.
Figure 2.9: As in Figure 2.7, but for the difference between strengthening and weakening composites, where positive values indicates strengthening composite has larger PV values.
Figure 2.10: Composite time series of the ventilation index of strengthening (red) and weakening (blue) trough interactions from 24 h prior to the time trough interaction was defined to 24 h after. Solid lines indicate the median ventilation index. Dots signify that the difference between the medians of the strengthening and weakening trough interaction ventilation indices are statistically significant at the 95% confidence level. Shading indicates the interquartile range, and the thin dashed lines represent the 10th and 90th percentiles. Note the logarithmic scale on the ventilation index axis.
Figure 2.11: Strengthening (a) and weakening (b) composites of shear-rotated, local entropy deficit at 600 hPa at the initial time a trough interaction was defined. The difference field is plotted in (c), where negative values indicate that the local entropy deficit is larger in the weakening composite. Range rings are every 200 km and extend to 1000 km. Shear vector points due east.
Figure 2.12: Shear-rotated, mean IR brightness temperature from 24 h prior to time of trough interaction to 24 h after (lag time indicated in box at top left of row). Intensifying (weakening) TCs shown in left (middle) column. The right column shows the difference field, where negative values indicate that intensifying TCs have lower IR brightness temperatures relative to weakening TCs. Range rings are every 100 km and extend to 500 km. Shear vector points due east.
Figure 2.13: Shear-rotated, mean IR brightness temperature at 24 h prior to trough interaction subtracted from subsequent 12-h periods (lag time indicated in box at center). Intensifying (weakening) TCs shown at left (right). Blue colors indicate convection strengthening with time. Range rings are every 100 km and extend to 500 km. Shear vector points due east.
Figure 2.14: 200-hPa, TC-centered, composite total wind speed (shaded; m s\(^{-1}\)) and divergence (contoured; s\(^{-1}\)) of strengthening (top row) and weakening (middle row) trough interaction cases 24 h prior to the time trough interaction was defined (a,d), the time of trough interaction (b,e), and 24 h following the time of trough interaction (c,f). Bottom row (g–f) shows differences between strengthening and weakening divergence; positive values indicate strengthening composite has larger divergence. Orange/black dot indicates TC center.
Figure 2.15: 200-hPa, TC-centered, composite PV advection by the irrotational wind (contoured; PVU d$^{-1}$), PV (shaded; PVU), and irrotational wind (colored arrows; m s$^{-1}$) of strengthening (top row) and weakening (bottom row) trough interaction cases 24 h prior to the time trough interaction was defined (a,d), the time trough interaction was defined (b,e), and 24 h following the time trough interaction was defined (c,f). For clarity, irrotational wind values < 2 m s$^{-1}$ not plotted. Red star indicates TC position.
Figure 2.16: 200-hPa, TC-centered, composite differences of PV advection by the irrotational wind (contoured; PVU d$^{-1}$) of strengthening and weakening trough interaction cases 24 h prior to the time trough interaction was defined (a), the time trough interaction was defined (b), and 24 h following the time trough interaction was defined (c). Positive values indicate that strengthening cases have more positive values of PV advection. Red star indicates TC position.
3. TC–trough interactions in idealized numerical modeling: comparison of a TC interacting with a westerly jet and a TC interacting with a trough

3.1 Introduction

While climatologies and composites can help illuminate what environmental differences may lead to favorable or unfavorable TC–trough interactions, the physical processes that cause intensity changes are difficult to resolve with the coarse temporal and spatial resolution of reanalyses. It is also difficult to control for environmental and intensity differences in real TCs. Few studies exist, however, that use idealized numerical modeling to study TC–trough interactions. Compared to modeling a real TC–trough interaction, idealized modeling allows for better control of trough and environmental characteristics, which eliminates many of the degrees of freedom that make intensity and structure change attribution difficult.

Among those studies using idealized modeling, which were reviewed in depth in chapter 1, differing causes of TC intensity change under the influence of a trough were found. Kimball and Evans (2002) suggested that direct superposition of the cutoff low and TC circulations resulted in intensification, but only if shear did not destroy the TC first. Additionally, in three of the four trough simulations performed, the TC was unable to develop a strong outflow channel on the trough side of the TC, hindering intensification.

Leroux et al. (2016) found that midlevel PV advection from the trough into the TC core led to intensification, but they also found that the distance between the TC and trough proved critical due to the heightened effects of shear if the TC was too close to the trough. Finally, Komaromi and Doyle (2018) determined that it was the reduction in inertial stability due to the approaching trough that allowed the TC to intensify through enhanced poleward outflow. Based on this lack of consensus, there exists ample opportunity to explore the intensification and evolution of TCs in the presence of a trough in an idealized modeling framework.
In this chapter, two idealized numerical simulations are presented that show how a TC evolves in the presence of a westerly jet (TC–jet simulation) and how a TC evolves in the presence of a jet plus a trough (TC–trough simulation). The rest of this chapter is organized into two major parts: the first is the idealized model setup, and the second is a comparison of the TC–trough and TC–jet simulations.

3.2 Methodology

3.2.1 Model specifications

This study utilizes the Advanced Research Weather Research and Forecasting (WRF) Model (WRF–ARW) (Skamarock et al. 2008) to simulate TC–trough and TC–jet interactions. The model is fully compressible and non-hydrostatic. The simulations are run on an $f$-plane at 20°N, and they use RRTMG longwave and shortwave radiation (Iacono et al. 2008), the Yonsei University (YSU) boundary layer scheme (Hong et al. 2006), and WRF single-moment, six-class (WSM6) microphysics (Hong and Lim 2006). The grid size, grid spacing, and convective parameterization of the three domains is given in Table 3.1. Of particular note is that the innermost, 3.33-km domain is convection resolving and vortex following. Figure 3.1 shows the heights, and corresponding pressures, of the specified model levels. Forty-eight vertical levels were used in the simulations, ranging between approximately 1015–50 hPa. In addition to the high-resolution boundary layer, higher vertical resolution was added between approximately 9–16 km to better resolve the outflow layer, within which interaction with the trough occurs. The SST was fixed at 28°C, and the simulations were each run for three days.

To initialize the TC, WRF TC bogussing was used with a maximum wind speed of 12.5 m s$^{-1}$ at the 60-km radius. The TC was initialized south of the jet and southeast of the trough (in the TC–trough simulation). The simulation boundaries were set up as a tropical channel, in which the eastern and western boundaries were periodic, and the southern and northern boundaries were closed. Because the simulations were idealized, no appropriate
boundary conditions or tendencies were available. Without this boundary information, the simulations developed jets at the northern and southern boundaries due to latent heat release within the domain, which significantly warmed the domain relative to the static boundary conditions.

To account for this artifact, northern and southern boundary conditions were obtained through the use of WRF’s “nestdown” function. To find the boundary conditions for domain 1, the outermost domain, an initial, single domain run was performed over a vastly larger area (nearly the size of the entire Western Hemisphere) than the area of domain 1. This run contained a TC and a trough embedded in a westerly jet (the setup of which is discussed in subsequent subsections), in keeping with most of the experimental runs. This setup produced latent heating of a similar scale to what occurs in the experimental runs, and while northern and southern jets still formed along the boundaries of the nestdown simulation, the jets were sufficiently far away from the location of domain 1 to not affect the conditions there. The tendencies and conditions of the nestdown simulation were then saved at the boundary locations of domain 1 to be used as boundary conditions in the experimental runs. The same boundary conditions were used for all of the simulations to mitigate boundary jet formation.

### 3.2.2 Initialization of the westerly jet

The background state of the model was initialized with a westerly jet. To create this jet, two soundings were interpolated from north to south across the domain. The first sounding used in the creation of the westerly jet was the Dunion moist tropical sounding (Dunion and Marron 2008), which is presented in Figure 3.2. The other sounding used was a late-October, subtropical sounding taken from ERA-I (Fig. 3.3). This sounding was averaged over a one week period and is located in the central North Atlantic at 32° N, 60° W. This location was chosen because it was subjectively deemed representative of a subtropical environment, while the time of year was chosen so that there would be a large temperature gradient at upper-levels between the moist tropical and subtropical soundings in order to produce a jet.
To interpolate between the moisture and temperature profiles of the two soundings, weighted means were used according to equation 3.1:

\[
x = w \times x_{MT} + (1 - w) \times x_{ST}
\]

where \(x_{MT}\) (\(x_{ST}\)) is the relative humidity or temperature profile of the moist tropical (sub-tropical) sounding, and \(w\) is the moist tropical sounding weight.

The moist tropical sounding was given a weight of one at the southern boundary. Then, an arctangent function (Eq. 3.2) was used to create the weights across the rest of the domain, such that the northern boundary would have a weight of zero for the moist tropical sounding. An arctangent function was chosen because it allowed for weak gradients in the northern and southern portions of the domain, but produced a strong gradient in the center of the domain, crucial for establishing a westerly jet:

\[
w = -\arctan(0.2 \times (\phi - \phi_{jet}))
\]

where \(\phi\) is latitude and \(\phi_{jet}\) is the jet center latitude, which was set to -1°. Note that latitude here is only relevant for scale because of the \(f\)-plane configuration. The coefficient of 0.2, which controls the strength of the gradient, was chosen subjectively to produce a jet with wind speeds between 25-30 m s\(^{-1}\) at jet level (200 hPa).

Figure 3.4 shows a cross section of the temperature field produced after the weights have been applied and Figure 3.5 shows the westerly jet produced by that temperature field, which is maximized near 200 hPa.

### 3.2.3 Initialization of the trough

The trough was inserted into the westerly jet base state by adding a temperature anomaly taken from a trough found in ERA-I. This trough was located at 23° N, 65° W and occurred in late September. Although the trough used for the experiments was chosen
subjectively, its temperature anomaly profile is consistent with other examples examined, albeit with a slightly larger magnitude negative temperature anomaly below the tropopause and a larger magnitude temperature anomaly above the tropopause. This trough was chosen because the temperature anomaly was small at lower levels and maximized around the tropopause. Figure 3.6 shows the 200-hPa temperature associated with this trough and its environment. To find the zonal temperature anomaly, the vertical temperature profile of the trough was first extracted from the center of the trough, denoted by the white star in Figure 3.6. Then, the vertical profiles of temperature were averaged east and west of the trough within $2^\circ$ boxes (red boxes in Fig. 3.6). The vertical temperature profile of the trough was then subtracted from the environmental average to produce the temperature anomaly of the trough (Fig. 3.7a). Ultimately, the magnitude of the temperature anomaly associated with the trough was reduced by half (Fig. 3.7b) before it was inserted, so that wind speed values within the trough were similar to the 25–30 m $s^{-1}$ wind speed of the jet.

Weights were also used to insert the trough temperature anomaly into the westerly jet background state (Eq. 3.3),

$$g = w_T * T'$$

(3.3)

where $g$ is the temperature perturbation at a given grid point, $w_T$ is the weight, and $T'$ is the trough temperature perturbation. The weights were set using an elliptical, Gaussian function (Eq. 3.4):

$$w_T = \left( \frac{1}{m_z} * \frac{\exp(-0.5p^2)}{\sqrt{2\pi}} \right) * \left( \frac{1}{m_m} * \frac{\exp(-0.5q^2)}{\sqrt{2\pi}} \right)$$

(3.4)

where $m_z$ and $m_m$ are the zonal and meridional scale modifiers, respectively, and $p$ and $q$ are defined in Eqs. 3.5 and 3.6, respectively:

$$p = \frac{\lambda - \lambda_{trough}}{m_z}$$

(3.5)

where $\lambda$ is longitude and $\lambda_{trough}$ is the trough center longitude, and
\[ q = \frac{\phi - \phi_{\text{trough}}}{m_m} \]  

(3.6)

where \( \phi \) is latitude and \( \phi_{\text{trough}} \) is the trough center latitude.

The trough center latitude and longitude were set to -5° and -120°, respectively. The values of \( m_z \) and \( m_m \) were set to 7 and 5, respectively. These scale modifiers were chosen so that the wind speed values associated with the trough were comparable to those of the westerly jet, 25–30 m s\(^{-1}\). The resulting trough is zonally large (approximately 30° longitude based on the trough’s half wavelength) compared to cut off lows, but it is not unreasonably large compared to midlatitude and subtropical troughs examined in ERA-I (not shown).

Note that only the temperature anomaly of the trough was added onto the base state; the moisture anomaly of the trough was not included to avoid adding another degree of freedom, which would complicate the analysis. Moisture has been shown to be a major factor in TC intensity change, and it was shown in Chapter 2 of this dissertation, and Fischer (2018), to be important during TC–trough interactions; therefore, the sensitivity to moisture perturbations should be considered in future work.

Figure 3.8a shows the 200-hPa temperature of the initial state after the soundings were interpolated and the trough anomaly was added. Figure 3.8b shows the cross section of the trough temperature anomaly taken through the center of the trough. The cross section shows a warm (cold) anomaly above (below) the tropopause. This temperature profile produces a trough embedded in a westerly jet, with wind speeds maximized near 200 hPa (Fig. 3.9).

### 3.3 Results of the TC–jet and TC–trough simulations: overview of simulations

This section provides an overview of how the TCs and the synoptic environments of the two simulations evolve throughout the 72-h simulations.
3.3.1 Synoptic evolution of simulations: 200-hPa wind speed

Figure 3.10 shows how the 200-hPa flow evolves throughout the three-day TC–jet simulation. Initially, the TC is embedded south of a westerly jet (Fig. 3.10a). Twenty-four hours later (Fig. 3.10b), the TC has moved little, but it is beginning to form a poleward outflow jet. There is also evidence that the westerly jet is being deformed, as a weak trough is forming upstream of the TC. The upstream trough, which has developed in-situ, is being formed due to weak, but persistent stretching deformation, as the TC outflow impinges on, and reduces the speed of, the westerly jet upstream (not shown). This same pattern continues at 48 h (Fig. 3.10c), with the outflow jet becoming stronger and extending farther northward, and the upstream trough expanding southward. The entire flow pattern, in fact, is becoming more amplified, as ridging is occurring downstream of the TC and pushing the westerly jet northward with a jet streak present north of the TC. At 72 h (Fig. 3.10d), there is continued upstream troughing and downstream ridging. The jet streak north of the TC is much stronger, with wind speeds over 40 m s$^{-1}$, and has become more anticyclonically curved.

The 200-hPa wind speed pattern differs significantly between the TC–jet and the TC–trough simulations, though their evolutions share some important similarities. Initially, the trough in the TC–trough simulation is apparent upstream of the TC, which is situated within the inflection point of the trough (Fig. 3.11a). By 24 h (Fig. 3.11b), the longitudinal scale of the trough has been reduced, while the latitudinal scale has increased. A jet streak has formed north of the TC near the inflection point, where the cyclonically-curved flow becomes anticyclonically curved. By 48 h (Fig. 3.11c), the trough has taken on a negative tilt, while the northern outflow jet is stronger and more anticyclonically curved than that of the TC–jet simulation. The same patterns continue at 72 h (Fig. 3.11d), with a negatively tilted trough and a high-amplitude, semicircular, strong, outflow jet.
3.3.2 Synoptic evolution of simulations: vertical wind shear

In the simulations, the differing synoptic environments of the two simulations and their evolutions result in different shear evolutions for the two TCs throughout the three-day period. Shear was calculated for both TCs between 850–200-hPa by averaging around an annulus to remove the symmetric component of the TC vortex. Shear was computed over several annuli varying inner and outer radii. For both simulations, shear was highly sensitive to the outer radius of the annulus. Larger annuli captured more of the westerly jet in both simulations, as well as the trough in the TC–trough simulation, resulting in increased shear (not shown). It is not clear which shear value the TC actually “feels”; thus, for simplicity, only two annuli were used for each simulation. The 0–300-km annulus was used to represent an “inner shear”, while the 200–800-km annulus was used to represent an “outer shear”.

Figure 3.12 shows the shear values for the TC–trough and TC–jet simulations. Initially, the TC–trough simulation is under greater shear, with a 0–300-km shear value of approximately 10 m s\(^{-1}\) and a 200–800-km shear value near 12 m s\(^{-1}\), compared to the TC–jet TC which has shear values near 7 m s\(^{-1}\) and 9 m s\(^{-1}\) in the two annuli, respectively (Fig. 3.12). As the simulation progresses, the 200–800-km shear values in both simulations do not change substantially, generally remaining within 2 m s\(^{-1}\) of the starting shear values. In the 0–300-km annulus, however, the shear values change much more over the course of the simulation, with shear values in both simulations decreasing. The 0–300-km shear in the TC–trough simulation decreases gradually to near 4 m s\(^{-1}\) by the end of the simulation, whereas the 0–300-km shear of the TC–jet simulation remains approximately constant until around 55 h, at which time there is a large decrease from 7 m s\(^{-1}\) to 1.5 m s\(^{-1}\).

In both simulations, the shear direction rotates cyclonically with time (not shown). In the TC–jet simulation, the shear initially points due east, due to the westerly jet, but gradually turns to point northeastward by the end of the simulation as the upstream trough forms in-situ. In the TC–trough simulation, the shear initially points toward the east-northeast, but turns to the north by the end of the simulation, as the trough approaches and
is eventually deformed, and wraps around, to the south of the TC (to be shown).

The direction of the inner shear and the outer shear is similar throughout most of the two simulations, but begins to differ by the end. The 0–300-km shear vector is displaced cyclonically relative to the 200–800-km shear vector at later times. In the TC–trough simulation, vastly different portions of the highly-deformed trough are captured by the two different shear annuli, resulting in the inner-shear annulus capturing more of the southeast-erly flow imparted by the trough as it wraps south of the TC and less of the larger-scale west-southwesterly flow. The differences in the inner and outer shear magnitude of the TC–jet simulation are smaller in magnitude due to the much weaker trough that has formed in-situ.

3.3.3 TC tracks

To generate TC tracks, center locations were determined using a Gaussian smoother on the sea-level pressure field at each output time; the location of the minimum in smoothed pressure was deemed the TC center. Various methods of center finding were tested, including absolute vorticity and pressure centroids, but due to the sheared nature of the convection, these centers were somewhat discontinuous and biased toward the downshear semicircle of the TC, especially while the TC was weak. Figure 3.13 shows the tracks of the TCs in the two simulations. In the TC–jet simulation, the TC tracked slowly east-southeastward at an average speed of 3.8 m s$^{-1}$. The mostly eastward component of motion is unsurprising given the westerly jet dominating the flow, but the reason for the southern component of motion is unclear. It is possible at early times it may be related to the tilted vortex (not shown) imparting southerly flow over the low-level center, and at later times it could be due to the formation of the trough in-situ as shown in Figure 3.10b,c.

As in the TC–jet model run, the dominant component of motion of the TC in the TC–trough simulation is eastward, due to the westerly jet. Unlike in the TC–jet simulation, which drifted southward throughout the three-day period, the TC–trough simulation has a
significant northward component of motion. Barring the first 12 h, in which the TC drifts southeastward, the TC moves generally toward the northeast. This northeast motion is due to the approach of the trough, which imparts southwesterly flow over the TC, as noted previously in the discussion of vertical wind shear. The TC has an average forward speed of 4.9 m s$^{-1}$, which is slightly faster than the TC in the TC–jet simulation. Unlike the TC in the TC–jet simulation, which slowed down at the end of the simulation, the TC in the TC–trough simulation picks up speed throughout the simulation, especially during the last 12 h, likely due to the TC deepening and more strongly interacting with the stronger upper-level flow associated with the upper-level trough. Even though the TC–trough simulation TC ends higher in latitude than the TC–jet TC, the potential intensity of the two TCs remain within 3–4 m s$^{-1}$ of each other throughout their lifetimes (not shown), therefore the track differences do not seem to result in TCs moving into substantially different background state thermodynamic environments.

### 3.3.4 TC intensity

Although the TCs are initialized with the same maximum wind speed and minimum central pressure, the TC–trough run ultimately intensifies to a minimum central pressure of 953.5 hPa and maximum wind speed of 39.1 m s$^{-1}$, compared to the TC–jet run which intensifies to 975.4 hPa and 31.0 m s$^{-1}$ (Figs. 3.14 and 3.15).

For the first 20 h of the simulations, the TCs have similar minimum central pressures and maximum wind speeds (Figs. 3.14 and 3.15, respectively). After this time, however, the minimum central pressures of the TCs diverge, and the TC–trough TC has lower pressure for the rest of the simulation. The pressure of the TC–jet TC remains approximately 5 hPa higher than the TC–trough simulation until 62 h (Fig. 3.14), at which time there is a large drop in minimum pressure in the TC–trough simulation. The difference in maximum wind speed between the two TCs is more varied, with the two TCs having similar wind speeds for the first 20 h, before the TC–trough TC increases in intensity relative to the TC–jet
TC (Fig. 3.15). This maximum wind speed difference is not maintained, however, as the TC–jet simulation increases in intensity, and the wind speed difference returns to near 0 m s\(^{-1}\) between 40–60 h.

As mentioned, significant intensification in the TC–trough simulation occurs between 62–72 h (Figs. 3.14 and 3.15). After 62 h, the magnitude of the difference in minimum pressure increases substantially in just 10 h, from about -5 hPa to -22 hPa (Fig. 3.14). The difference in maximum wind speed between the two simulations between 62 h and 72 h also increases from -2 m s\(^{-1}\) to 8.1 m s\(^{-1}\) (Fig. 3.15). The growing difference in maximum wind speed is due both to intensification of the TC–trough TC, and a pause in intensification of the TC–jet TC.

The increase in the maximum wind speed of the TC in the TC–trough simulation falls just short of the criterion for rapid intensification (RI) in a 12-h period (approximately 10 m s\(^{-1}\); Kaplan et al. 2015). The TC–trough TC takes 13 h, between 59–72 h, to increase 10 m s\(^{-1}\). Given the very sharp increase in wind speed in the final hours of the simulation, it is likely that the TC is undergoing RI.

That the TC–trough TC is more intense than the TC–jet TC starting around 20 h, and that it nearly meets the criterion for rapid intensification when it intensifies late in the simulation, all while experiencing stronger vertical wind shear than the TC–jet TC, suggests that the environment in which the TC–trough TC is embedded is more thermodynamically favorable for intensification than the TC–jet TC. As discussed, the potential intensities of the two TCs do not vary drastically, which suggests that other factors are influencing the thermodynamic environments of the TCs. Various dynamical mechanisms may be contributing to the favorability of the environment through enhanced divergence and lift, which would feedback on the thermodynamic favorability of the environment. A look at the convective evolution of the TC may more clearly indicate the favorability of the TCs’ thermodynamic environments.
3.3.5 Convective evolution of simulations: Maximum reflectivity

Reflectivity is shown in Figures 3.16 and 3.17 in order to document the evolution of the TCs’ precipitation structure. Considering first the TC–jet simulation (Fig. 3.16), at 24 h (Fig. 3.16a) the convection of the TC is displaced downshear, to the northeast of the TC. By 48 h (Fig. 3.16b), the TC has a distinct eye and more symmetric convection, albeit over a fairly small area close to the TC center. At 72 h (Fig. 3.16c), the TC has a large, clear eye and the TC has a larger area of precipitation, especially to the north, which is toward the downshear and downshear-left directions. The improved convective pattern is likely due to the much reduced shear at later times. The precipitation throughout the simulation covers more area on the north-northeast side (downshear) of the TC compared to the other quadrants, which is consistent with previous studies of sheared TCs (Corbosiero and Molinari 2002, 2003; Rios-Berrios et al. 2016; Fischer et al. 2018).

At 24 h (Fig. 3.17a), the reflectivity field of the TC–trough simulation looks similar to that of the TC–jet simulation, albeit with greater overall coverage of precipitation. The TC does not yet have an eye, and the convection is largely displaced downshear, to the northeast of the TC. Between 24 h and 48 h (Fig. 3.17b), the reflectivity appearance of the two simulations diverges. In the TC–trough simulation, there is an eye present, but it is not as well-defined as in the TC–jet simulation. The biggest difference is that there is a much greater area of precipitation in the TC–trough simulation. North of the TC, the TC–trough simulation has a large, approximately north–south oriented rain shield extending over 5° from the TC center. This shield is absent in the TC–jet simulation, as the convection in that simulation extends less than 2° from the center and is associated with the core TC convection. Furthermore, the ring of convection surrounding the TC eye is wider than in the TC–jet simulation, especially on the TC’s western side. Whether these differences in precipitation structure are due to shear is unclear, since the 200–800-km shear is larger in the TC–trough simulation, but the 0–300-km shear is similar (Fig. 3.12).

At 72 h (Fig. 3.17c), the core convection of the two TCs is similar, as both TCs have
clear eyes and symmetric convection, but the differences in the rain shield area between the two simulations becomes more stark by 72 h. At this time, the precipitation shield has expanded, and rotated cyclonically, yet it remains connected to the TC inner core as a large, spiral rainband. The large areal extent of the precipitation north of the TC, and its rotation toward left of track, fits with the precipitation pattern seen in Hurricane Floyd (1999) as it interacted with an approaching trough from the west (Atallah and Bosart 2003). In that case, it was shown that the precipitation shield was situated along a strong baroclinic zone, which formed as the cold-core trough approached the warm-core TC. The relevant mechanisms involved in the formation and maintenance of the precipitation shield in the TC–trough simulation will be explored later in this chapter.

An important component of the convective evolution of the TC–trough TC is that the TC undergoes downshear reformation near 21 h. Downshear reformation is the process by which the TC center reforms in an area of downshear convection (itself containing a nearly upright mesovortex, within the larger-scale, parent vortex), which is sustained, intensifies, and eventually absorbs the parent vortex (Molinari et al. 2006; Nguyen and Molinari 2015). Downshear reformation is then often followed by TC intensification (Molinari et al. 2006; Nguyen and Molinari 2015).

During the first 20 h of the simulation, the TC–trough TC has a strongly-tilted TC vortex (Fig. 3.18) relative to the TC–jet TC, which corresponds to the larger shear in the TC–trough simulation. This tilt reduces in magnitude rapidly, from nearly 80 km to just 20 km beginning near 21 h. To illustrate that this reduction in tilt is a reflection of downshear reformation, Figure 3.19 shows high-temporal resolution, TC–centered, 1-km reflectively and sea-level pressure on the inner-most domain. At 20:36 (Fig. 3.19a) the sea-level pressure minimum is co-located with the TC center; however, most of the TC convection at this time is located downshear, to the northeast. At 20:54 (Fig. 3.19b), a new minimum in sea-level pressure has formed within the downshear convection. Over the following 54 min (Figs. 3.19c–e), this minimum in sea-level pressure rotates cyclonically inward within the
larger TC circulation, axisymmetrizes, and becomes the new center thereafter. The timing of the downshear reformation coincides with a large increase in wind speed in the TC–trough TC after 21 h (Fig. 3.15). The evolution of the downshear reformation, and subsequent TC intensification, is consistent with Nguyen and Molinari (2015), who documented that downshear reformation in WRF simulations of Tropical Storm Gabrielle (2001) led to its subsequent intensification.

### 3.3.6 Diabatic heating

Another way to consider the convective evolution of the TCs is through diabatic heating. Figure 3.20 shows a radius–time Hovmöller of azimuthally-averaged diabatic heating for both TCs. The TCs share a similar initial evolution in which there is an outward propagation of large diabatic heating from the TC center between 0–20 h, which was suggested by the increasing tilt during this time and is consistent with sheared TCs. Differences begin to emerge near 20 h, at which time diabatic heating (Fig. 3.20; purple arrow) suddenly begins to occur closer to the TC–trough TC center compared to before this time, and this closer diabatic heating is sustained thereafter, consistent with the downshear reformation that occurs at this time. Additionally, near 62 h in the TC–trough simulation, the time at which the TC–trough TC’s minimum pressure begins to rapidly decrease, there is sustained, intense diabatic heating within the radius of maximum wind (RMW; Fig. 3.20; blue arrow), which has been shown to be efficient in increasing TC intensity (Shapiro and Willoughby 1982; Vigh and Schubert 2009; Tao and Zhang 2014).

The TC–trough TC from around 20 h to the end of the simulation has significantly more diabatic heating within 500 km of the TC center compared to the TC–jet TC (Fig. 3.20). This result is consistent with the much greater coverage of precipitation in the TC–trough simulation, shown in Figures 3.16 and 3.17.

It is possible that the downshear reformation and the enhanced convection within the RMW at 62 h, which both led to intensification of the TC–trough simulation, could be the
result of random chance, as they are both related to stochastic processes. The larger area of diabatic heating in the TC–trough simulation, however, is indicative of a more favorable environment for convection to occur than in the TC–jet simulation. The fact that the environment is more conducive to convective development suggests that the likelihood of convective bursts occurring near the TC center and resulting in intensification, such as the events that occurred during downshear reformation and later during the period of rapid intensification, is greater in the TC–trough simulation.

It has been shown that the strength and areal extent of a TC’s convection may be influenced by the amount of moisture in the TCs’ environment (Hill and Lackmann 2009; Wang and Hankes 2016). Two related mechanisms, lift and divergence, occurring with greater magnitudes near the TC–trough TC could result in more moistening near the TC, which would contribute to the more convectively-favorable environment shown in the reflectivity and diabatic heating fields of the TC-trough TC relative to the TC–jet TC.

It is hypothesized that the dynamic, synoptic environment of the two TCs, in which one TC is downstream of a large trough and the other is not, results in differences in large-scale lift and divergence near the TCs. In the remainder of this chapter, this hypothesis will be explored, and various dynamical mechanisms through which divergence and lift are enhanced due to the presence of the trough will be examined.

3.3.7 Moisture

To determine whether the environment is indeed more moist in the TC–trough simulation, midlevel RH is explored. Figure 3.21 shows the 700–500-hPa layer-averaged relative humidity (RH) near the TCs at various times during the simulations. Early in the simulation, at 20 h (Figs. 3.21a,b,c), the TC–trough TC already shows a more moist large-scale environment relative to the TC–jet TC. A notable exception is found close to the TC in the upshear semicircle, which is significantly drier in the TC–trough simulation. Both TCs show a similar dipole pattern near the TC center, with drier air upshear and more moist air
downshear, which is indicative of the vertical motion associated with sheared flow interacting with a tilted TC vortex, which forces ascent downtilt and ascent uptilt (Jones 1995; Reasor et al. 2004, 2013). The stronger RH dipole in the TC–trough simulation is likely due to the greater tilt magnitude in that simulation.

Near midway through the simulations, at 40 h (Figs. 3.21d,e,f), and continuing late into the simulations at 60 h (Figs. 3.21g,h,i), the differences in RH between the two simulations grows. The RH dipole in both TCs near the center has disappeared due to the reduction in tilt of the TC vortices. At 40 h, the TC–trough TC has greater moisture in the precipitation shield region downshear of the TC, as well as in the rainband region east and southeast of the TC. The TC–trough TC also has higher values of RH near, and upshear of, the TC center than the TC–jet TC. The only large region where the TC–trough TC is drier is south of the TC. This dry air may be due to subsidence associated with the TC’s outflow approaching the region of high inertial stability in the core of the trough, which acts to impede the outflow and force it to descend. This process will be discussed further in a later section.

At 60 h, the TC–trough TC still has greater RH in the precipitation shield region, though the location of the precipitation shield has rotated cyclonically. Upshear of the TC–trough TC, the region is much more moist than in the TC–jet TC, with RH differences between the two simulations greater than 20%. The greatest areas of relative moistening away from the TC core in the TC–trough TC relative to the TC–jet TC are north, east, and west of the TC. As it relates to convection, the greater RH north of the TC–trough TC corresponds to the precipitation shield. East and west of the TC, the convective differences are less stark, as convection is less organized in both simulations, but there is greater coverage of isolated high reflectivity in the TC–trough TC, indicating that the likelihood of convection is increased in the TC–trough TC due to the greater environmental moisture.

Figure 3.22 shows how the differences in the RH of each simulation evolve with time, rather than how it evolves relative to the other simulation. Between the beginning of the simulation and 20 h (Fig.3.22a,b), both TCs moisten mostly in the downshear semicircle,
where convection is displaced. The dipole pattern in which the TC has dried uptilt and moistened downtilt near the center of the TC discussed previously is also apparent. Between 20–40 h (Fig. 3.22c,d), both TCs have moistened considerably near the TC center upshear, although RH downshear near the center has changed little. The TC–trough TC, as suggested in Fig. 3.21, has moistened more and over a larger area than the TC–jet TC. Between 40–60 h, the TC–trough TC has moistened over nearly the entire western side of the TC, including a substantial upshear portion of the TC. The TC–jet TC moistens over a much smaller area closer to the TC center.

Greater moisture upshear of the TC has been shown to be favorable for TC intensification (Kaplan et al. 2015; Rios-Berrios et al. 2016) and the TC–trough simulation, especially at later times, is more moist upshear. Additionally, RI has been shown to occur sooner in simulations with a more moist environment (Kieu et al. 2014). Therefore, it stands to reason that the more moist TC environment of the TC–trough simulation increased the likelihood of greater intensification in that simulation, which is what occurs in the last 10 h of the TC–trough simulation.

As mentioned in the earlier hypothesis, the reason for the moistening of the large area surrounding the TC could be the result of greater large-scale divergence, and concomitant ascent, due to the presence of the trough near the TC compared to the initially straight jet of the TC–jet simulation. Whether divergence is, in fact, enhanced in the TC–trough simulation is analyzed in the following section.

3.3.8 Divergence

Whether divergence drives changes in moisture, convective strength, and TC intensity, or whether divergence is driven by these changes, is difficult to disentangle, and it is likely they are occurring synergistically. Regardless, in the context of these simulations, there is strong feedback between increased convection and increased outflow, especially in the TC–trough simulation. The strong PV gradient created between the approaching trough
and the TC’s divergent outflow, is associated with a strong, anticyclonic outflow jet near the TC (Fig. 3.11). The divergence resulting from this outflow structure can act to increase the TC’s convection, which strengthens the temperature and height gradients between the trough and TC farther, resulting in a stronger jet and completing the positive feedback loop. Therefore, even small changes in the divergence of the TC, or the convective strength, can result in large changes in the intensity of both the jet and TC divergence with time.

Figure 3.23 shows 200-hPa divergence throughout the simulations at 20, 40, and 60 h. Relatively early in the simulation at 20 h (Figs. 3.23a,b,c), there is much greater divergence over a large area surrounding the TC. The differences are largest in the region of the TC–trough TC’s precipitation shield, which is associated with the largest divergence values. Outside of this region, over almost the entire domain considered, there is more divergence occurring in the TC–trough simulation, indicating that there is also stronger large-scale lift occurring over the entire region. In fact, there are very few locations over which the TC–jet simulation has greater time-averaged divergence.

At 40 h (Figs. 3.23d,e,f) and 60 h (Figs.3.23g,h,i), larger values of divergence in both TCs have expanded in area, but the differences show a similar story to 20 h. The TC–trough TC continues to have greater divergence over almost the entire region. Of note, however, there is a particularly strong area of divergence at 40 h just upshear of the TC–trough TC, which matches the area of large moistening at this time (Fig. 3.22d).

The prolonged time of greater divergence in the TC–trough simulation, nearly the entire 72 h (not shown), fits with the greater moisture and diabatic heating seen in that simulation compared to the TC–jet simulation. The differences in divergence cannot be explained exclusively through differences in TC size or intensity and, as mentioned, separating the effects of greater divergence from stronger convection is impossible. While the TC–trough simulation is stronger after 20 h (Fig. 3.14), and larger after around 25 h (not shown), the differences in divergence appear much earlier in the simulation, as early as 10 h (not shown). This result does suggest that the synoptic setup of the TC–trough simulation is more favorable
for large-scale forcing for divergence and ascent. There are several potential mechanisms for the enhanced divergence and concomitant ascent in the TC–trough simulation, as a result of the differences in synoptic flow, which will be explored subsequently.

3.3.9 EFC and inertial stability

EFC acts to increase TC divergence through a Sawyer–Eliassen balanced vortex response, in which the presence of outflow-layer cyclonic momentum source (due to the trough and anticyclonically-curved outflow jets) results in a secondary circulation response that enhances divergent outflow.

Typically, azimuthally-averaged, annulus-averaged EFC has been considered in relation to the rate of TC intensification at different lag times (Molinari and Vollaro 1989; DeMaria et al. 1993; Hanley et al. 2001). To that end, Figure 3.24 shows the 300–600-km and 500–900-km EFC for each of the two simulations. As mentioned previously, the TC–jet simulation develops a large, weak trough upstream of the TC, and therefore, trough-induced augmentations in EFC are possible in that simulation. There is no initial trough present in the TC–jet simulation, however, which explains why EFC in the TC–jet simulation is initially near 0 m s$^{-1}$ d$^{-1}$. Prior to the intensification of the TC–trough simulation TC at 20 h, EFC is 3–5 m s$^{-1}$ d$^{-1}$ greater in both annuli of the TC–trough simulation than in the TC–jet simulation. The enhanced EFC in the TC–trough case indicates that the presence of the strong, initial trough is augmenting EFC.

In the TC–trough simulation, the 300–600-km annulus EFC begins increasing at about 12 h, while EFC does not begin increasing in the TC–jet simulation until around 18 h. Because the increase in EFC precedes intensification, it is possible that 300–600-km EFC is at least partly contributing to the intensification seen in both simulations about 10–15 hours later. This finding is similar to previous studies that found a positive correlation between EFC and later intensification (Molinari and Vollaro 1989; DeMaria et al. 1993; Leroux et al. 2016); however, the 10–15-h lag between enhanced EFC and intensification in
these simulations is substantially shorter than the 24–48 h lag found in those studies.

It is also possible that there is a cumulative, compounding effect of the enhanced EFC in the TC–trough simulation during the lead up to the intensification of the TC–trough TC. This hypothesis is difficult to test, especially with only two experiments, and the cumulative effect of enhanced EFC has not been examined in previous literature. Theoretically, however, the extended presence of greater EFC values in the TC–trough simulation would prolong the time period during which there is forcing for divergence, which would be favorable to the TC.

As in the 300–600-km annulus, the TC–trough simulation has relatively large values of EFC in the 500–900-km annulus compared to the TC–jet experiment throughout the simulation. Therefore, as before, there could be a cumulative positive effect of this enhanced EFC on divergence and TC intensity. The 500–900-km EFC in the TC–trough simulation increases at 20 h, 8 h later than in the 300–600-km annulus, but this increase does occur prior to the start of a greater rate of intensification in terms of pressure at 28 h, and therefore could be a contributing factor to increased intensification. In the TC–jet simulation, the increase in 500–900-km EFC is also delayed compared to the increase in 300–600-km EFC, but it begins increasing concurrently with a greater rate of intensification at 28 h. The timing of the increase in 500–900-km EFC in the TC–jet simulation suggests that EFC in this annulus could be a result of TC intensification or that EFC is synergistically increasing with increasing TC intensity.

A more comprehensive way of considering what is causing changes in EFC is presented in Figure 3.25, in which EFC is not azimuthally-averaged. One reason for not azimuthally-averaging EFC is that Molinari and Vollaro (1989) pointed to two regions where EFC is enhanced: first, in the region of the approaching trough (a source region of cyclonic angular momentum) and, second, in the region of anticyclonic outflow (a sink region of anticyclonic angular momentum). Second, non-azimuthally-averaged EFC is also considered in Figure 3.25 in relation to inertial stability, as the divergent outflow response to momentum forcing
is modulated by inertial stability (Eliassen 1952). Inertial stability is defined as

\[ I^2 = \left( \frac{2v_t}{r} + f \right) \left( f + \zeta \right) \]  

(3.7)

where \( I^2 \) is the inertial stability, \( v_t \) is the tangential wind, \( f \) is the Coriolis parameter, \( r \) is the radius from the TC center, and \( \zeta \) is the relative vorticity (Alaka 1961; Komaromi and Doyle 2018). Shapiro and Willoughby (1982) and Holland and Merrill (1984) showed that the outflow response to momentum forcing is controlled by the inertial stability of the environment within which the TC is embedded. They showed that lower inertial stability resulted in a greater radial extent of the balanced Sawyer–Eliassen secondary circulation forced by the momentum source because the lower inertial stability provided less resistance to lateral displacement of air parcels. Rappin et al. (2011) and Rappin and Nolan (2012) showed that idealized TCs in lower inertial stability environments had greater divergent outflow, while Komaromi and Doyle (2018) found that in idealized TC–trough interactions, the reduction in inertial stability along the anticyclonic shear side of the trough resulted in enhanced TC divergence in that direction. Therefore, elevated values of EFC are likely to result in a greater divergent outflow response in regions of reduced inertial stability relative to elevated EFC values occurring in regions of greater inertial stability.

At 0 h (Fig. 3.25a,b,c), the location of the low inertial stability region in the TC–trough simulation is displaced toward the west compared to the TC–jet simulation due to the presence of the upstream trough. Low inertial stability values are occurring preferentially toward the trough due to the trough-induced flow resulting in anticyclonic shear along the trough’s eastern edge. In the core of the trough, however, inertial stability is high due to the trough’s positive relative vorticity. Both TC’s have lower inertial stability to the north of the TC due to their location on the anticyclonic shear side of the westerly jet to the north. EFC values at this time are small due to the lack of well-developed TC outflow.

At 20 h (Fig. 3.25d,e,f), once the TC’s outflow has become more established, the two regions of enhanced EFC, suggested in Molinari and Vollaro (1989), are evident, with greater
EFC occurring near the approaching trough to the west and in the strong, anticyclonic outflow to the northeast. Compared to the TC–jet simulation, EFC is greater in the TC–trough simulation in both locations. The elevated EFC values associated with the trough region are not surprising, given the presence of the stronger trough in the TC–trough simulation, but the EFC values northeast of the TC in the outflow region are also much greater in the TC–trough simulation, which is due to the greater anticyclonic curvature and strength of the jet (Fig. 3.11). EFC is not enhanced on either TCs’ south side due to the lack of a strong southern outflow jet. One reason for the lack of a southern outflow jet is that less work is required to develop outflow on the TCs’ northern side, due to the lower inertial stability there, which results in preferential outflow to the north (Rappin et al. 2011; Rappin and Nolan 2012; Barrett et al. 2016; Komaromi and Doyle 2018). Additionally, the displacement of precipitation to the north side of the TCs, due to the southwesterly shear, may also be contributing to the lack of a strong southern outflow jet.

By 40 h (Fig. 3.25g,h,i), both simulations have greater values of EFC, but the EFC values in the TC–trough simulation remain larger than the TC–jet simulation. Both simulations also show an increased areal extent of large EFC values, as the strength and extent of the anticyclonic outflow jets have increased, and the trough in the TC–trough simulation has been elongated and wrapped inward and south of the TC. The same EFC pattern continues at 60 h (Fig. 3.25j,k,l), but again, with increased EFC values in both simulations as the strength and curvature of the northern outflow jets increase and as the trough has more closely approached the TC.

The positive feedback between the large EFC values north of the TC and divergence in the region of strong, anticyclonic curvature is likely synergistically increasing both throughout the simulations. As the outflow strengthens and becomes more anticyclonically curved, EFC increases, which forces even greater divergence. Evidence for this feedback is shown in Figure 3.26. Between 20 h and 60 h, the curvature of the flow in the TC–trough simulation (Fig. 3.26b,d,f); using the 1-PVU contour as a proxy for the shape of the flow),
becomes increasingly anticyclonic. During the same period, north of the TC, the magnitude of EFC increases with time, as does the spatial coverage of large EFC values. Similarly, divergence increases in both magnitude and area between the TC and the EFC maximum, located ~1000 km north of the TC center. In the TC–jet simulation (Fig. 3.26a,c,e), the same behavior is occurring on a weaker scale; the flow is less anticyclonically-curved, and it is associated with weaker values of EFC and reduced divergence. In both simulations, the pattern of EFC located radially outward from the area of enhanced divergence is consistent with Molinari and Vollaro (1989), who found the same pattern in azimuthally-averaged EFC and divergence.

In both simulations, the enhanced values of EFC north of the TC in the anticyclonic outflow jet region are occurring in regions of relatively low inertial stability, which can also help explain the development of the large areal extent of the northern outflow jets. In the TC–trough simulation, the inertial stability is further reduced, especially northwest of the TC, by the presence of the trough and the formation of the downstream ridge. The trough reduces inertial stability along its leading edge due to anticyclonic shear (Komaromi and Doyle 2018), and the ridge decreases inertial stability due to its low relative vorticity. As EFC forces divergence north of the TC, the low inertial stability there allows the air parcels to accelerate away from the TC over a greater distance and with less resistance than in the TC–jet simulation. The combination of greater values of EFC (forcing for divergence) and lower inertial stability (allowing for a greater outflow response to that forcing due to less resistance to lateral motion), results in a stronger northern outflow jet in the TC–jet simulation and, therefore, greater divergence (Figs. 3.10 and 3.11).

On the southern and western sides of the TC, the relatively high inertial stability of the core of the trough reduces the effectiveness of this area of EFC to forcing large lateral divergence. Thus, there is greater resistance to outflow development relative to the northern side of the TC. In the TC–trough simulation, large values of EFC associated with the approaching trough would still be acting to force divergence to the west-southwest of the
TC, but the larger relative values of inertial stability associated with the core of the trough resists lateral motion in the same region.

Thus, there is something akin to an “inertial stability wall” due to the trough itself, which reduces the lateral extent of the westward and southward-directed divergence. The presence of this “wall” does, however, aid the development of the northern outflow jet. As the low-PV outflow of the TC meets the high PV of the trough, it results in a strengthening of the PV gradient between the trough and TC, which is associated with an increase in outflow jet strength, even though the outflow is radially confined. The jet also increases in strength as the divergent wind hits the “wall’ and is directed northward, such that it follows the synoptic-scale jet and flows toward where inertial stability is low. Additionally, the Coriolis force is also contributing the northern redirection of the outflow. The divergent wind is impeded from southward motion due to the high inertial stability in that direction as a result of the PV of the trough being wrapped inward and south of the TC (Fig. 3.25h,k).

As somewhat of an aside, the presence of the inertial stability “wall” southwest of the TC results in the formation of an upper-level front between 400–200-hPa (Fig. 3.27; circled region), which is similar to the PV advection tongue shown in Leroux et al. (2016) that was discussed previously. As discussed in the previous paragraph, the high inertial stability of the trough’s core is causing the radial wind to slow as it approaches the trough, as shown in Figure 3.27. The formation of this front is due to the presence of a transverse circulation, where two branches are part of the TC’s main secondary circulation: the upward motion of the TC eyewall and the TC’s outflow near 200-hPa. The final branch is a combined layer of return flow and sustained downward motion in the layer in which the front is forming. The descent (Fig. 3.28; circled region) is a result of the divergent outflow being impeded by the high inertial stability of the trough. The outflow is stopped from moving upward due to the high static stability of the tropopause and impeded from continuing outward by the high inertial stability of the trough, which results in the outflow being partially forced to descend. The high PV of the trough, which is where the outflow descends, is then pulled.
inward toward the TC core (Fig. 3.27; circled region) along this inward, descending flow. This high-PV frontal feature was analyzed extensively but, unlike in Leroux et al. (2016), who found that the formation of this PV feature was key to TC intensification in sensitivity experiments of TC Dora, no definitive effect on TC size or intensity was found.

The result of the greater EFC occurring in the TC–trough simulation, and that it is occurring in a region of lower inertial stability than the TC–jet simulation, is that there is greater forcing for divergence in the TC–trough simulation. Additionally, the positive feedback between greater EFC and the strengthening outflow jet indicates that forcing for divergence due to EFC is greater in the TC–trough simulation throughout the simulation and may be responsible for the larger, stronger, outflow jet north of the TC–trough TC.

3.3.10 Subgeostrophic to supergeostrophic transition zone

Another reason for the increased moisture by way of enhanced large-scale divergence and ascent in the TC–trough simulation is the initial location of the TC–trough TC, which is initialized near the inflection point of the trough. The curvature of the upstream trough indicates subgeostrophic flow is occurring southwest of the TC, while the developing downstream anticyclone located north-northeast of the TC is associated with supergeostrophic flow (Martin 2006). Thus, the TC is located in the transition zone between subgeostrophic and supergeostrophic flow, which is an area of enhanced divergence and lift (Holton 2004; Martin 2006).

Figure 3.29 shows where the regions of subgeostrophic and supergeostrophic flow are located, as well as streamlines of the full wind, for the two simulations. The spatial pattern of subgeostrophic and supergeostrophic flow for each simulation changes little after 40 h, although the magnitudes increase, but the two runs differ significantly from each other. In both simulations, the areas of subgeostrophic and supergeostrophic flow northeast of the TCs, within approximately 200 km of the TC center, are associated with the TC’s own divergence, and the pattern differs little between the simulations. Considering the larger scale in the
TC–trough simulation, the area near the trough axis is associated with subgeostrophic flow. The ridge to the north is associated with supergeostrophic flow. The full wind shows that the synoptic-flow pattern is around the base of the trough and into the ridge. Thus, the flow accelerates in the subgeostrophic to supergeostrophic transition zone between the trough and ridge, which is near the TC’s location, due to the differences in curvature of the synoptic flow, thereby causing large-scale divergence.

The TC–jet simulation, perhaps surprisingly, shows a somewhat similar pattern of subgeostrophic and supergeostrophic flow. Recall that a trough forms upstream of the TC in-situ in the TC–jet simulation and low-PV, TC outflow, as in the TC–trough case, creates a downstream ridge. Both the trough and ridge are weaker in the TC–jet case and, thus, the magnitudes of the subgeostrophic and supergeostrophic winds, and the gradient between them, are weaker. Therefore, even though the transition zone between subgeostrophic and supergeostrophic flow is closer to the TC–jet TC, given the weaker gradient in that simulation, the contribution to the strength of the divergence due to the transition would be smaller than in the TC–trough TC.

As mentioned previously, the large-scale pattern of subgeostrophic winds and supergeostrophic winds changes little throughout the later part of the period, but the magnitude of the ageostrophic wind, and therefore the gradient between the two regions (in which the TC–trough TC is consistently stronger), does change with time. This difference in the magnitude of the gradient would favor more divergence in the trough case.

3.3.11 Quasi-nondivergent forcing for ascent

Another mechanism that enhances lift is quasi-nondivergent forcing for ascent (Nielsen-Gammon and Gold 2008; McTaggart-Cowan et al. 2008; Molinari and Vollaro 2011). This forcing is similar to quasigeostrophic forcing for ascent (Bosart and Bartlo 1991; Bracken and Bosart 2000; Holton 2004; Martin 2006; Fischer et al. 2017), however, nondivergent wind replaces geostrophic wind, which increases applicability near TCs and at low latitudes.
The TC–trough TC’s location downstream of a trough axis is a preferred location of quasihongdivergent ascent and, in both cases, the TCs’ locations near the right-entrance region of the jet is also a favorable location for large-scale ascent (Bjerknes and Holmboe 1944; Holton 2004; Martin 2006). Previous studies have shown that both quasigeostrophic and quasi-nondivergent forcing for ascent is favorable for TC intensification (Bosart and Bartlo 1991; Bracken and Bosart 2000; McTaggart-Cowan et al. 2008; Fischer et al. 2017).

The effect of the synoptic flow on lift is subsequently analyzed through $\mathbf{Q}$-vector divergence (Bluestein 1992; McTaggart-Cowan et al. 2008; Molinari and Vollaro 2011), which diagnoses areas of forced ascent. The $\mathbf{Q}$-vector is defined as

$$
\mathbf{Q} = (Q_1, Q_2) = -\frac{R}{\sigma p} \left( \frac{\partial v_\psi}{\partial x} \cdot \nabla_p T, \frac{\partial v_\psi}{\partial y} \cdot \nabla_p T \right)
$$

(3.8)

where $R$ is the dry air gas constant, $p$ is the pressure, $v_\psi$ is the non-divergent wind vector, $T$ is temperature, and $\sigma$ is

$$
\sigma = -\frac{RT_0 \partial \ln(\theta)}{p}
$$

(3.9)

where $\theta$ is potential temperature and $T_0$ is the environmental temperature averaged within 1000 km of the TC center (Bluestein 1992). Ascent is related to $\mathbf{Q}$ vector divergence by

$$
L(\omega) = -2\nabla \cdot \mathbf{Q}
$$

(3.10)

where $\omega$ is vertical motion. The left-hand side of eq. 3.11 is assumed to be proportional to $-\omega$ due to the $L$ operator containing the Laplacian; thus, $\mathbf{Q}$-vector convergence (divergence) corresponds to forcing for ascent (descent) (Molinari and Vollaro 2011).

The wind and temperature fields of the TC were filtered out at each time period within 400 km of the TC center, and then the wind and temperature field was re-interpolated using cubic interpolation, to limit the diagnosed forcing for ascent to the synoptic conditions. Multiple radii for filtering and several interpolation routines were tested and the results did
not vary drastically (not shown). As in Molinari and Vollaro (2011), the non-divergent wind vector and temperature fields were spatially smoothed using a Gaussian smoother prior to computation of the Q vector, and Q-vector divergence was calculated between 400–200 hPa.

Figure 3.30 shows the location of forcing for ascent and descent during both simulations. As in the discussion of subgeostrophic and supergeostrophic flow, the TC–trough and TC–jet simulations differ significantly from each other.

At 20 h, in the TC–trough simulation there is Q-vector convergence and, thus, forcing for ascent, downstream of the trough very near and just upstream of the TC. The forcing for ascent covers a large area of the TC’s western semicircle. At this time, the co-location of the right-entrance region of the jet, which is an area favored for ascent, with the location of the TC downstream of the trough, another area favored for ascent, both contribute to forcing for ascent. Northeast of the TC, there is Q-vector forcing for descent along the axis of the developing downstream ridge. This region of descent in a ridge is occurring in an area in which Q-vector forcing for descent is expected (Martin 2006). Closer to the TC, just downstream, there is weak forcing for descent as well. This area of descent is peculiar because part of the TC–trough TC’s large precipitation shield (Fig. 3.17), which is associated with a large area of divergence (Fig. 3.23), extends into this region. There are, however, many other forcings for convection beside Q-vector convergence.

In the TC–jet simulation at 20 h, forcing for both ascent and descent is considerably weaker overall. The weaker forcing is due to both weaker gradients in the non-divergent wind, but also weaker temperature gradients, as there is no strong, upstream, cold-core trough. The forcing for ascent located north of the TC is due to the presence of the weak, developing upstream trough along the axis of the weak, developing downstream ridge.

At 40 h and 60 h (Fig. 3.30c–f), the magnitude of forcing for ascent and descent have increased. The increase in the magnitude of Q-vector forcing is due to both an increase in the magnitudes of the gradients of the non-divergent wind speed and an increase in the magnitudes of the gradients of temperature. These changes in the gradients are magnified
in the TC–trough simulation because of the development of the very strong jet downstream of the TC, the amplification of the synoptic flow, and the approach of the cold-core trough to the warm-core TC. Additionally, static stability upstream of the TC is reduced in the TC–trough simulation due to the approach of the cold-core trough (not shown), further increasing forcing for ascent. These factors are also occurring in the TC–jet simulation, but with much weaker magnitude.

Unlike at 20 h, there is an indication that the TC’s location in the right-entrance region of the jet is not significantly increasing forcing for ascent in the TC–trough simulation. Due to the TC’s diabatic heating and strong outflow, there is warm air advection occurring through the core of the jet streak northeast of the TC (not shown). Warm air advection through the jet core results in the displacement of the favorable region for ascent from the right-entrance region of the jet into the core of the entrance region (Martin 2006), as seen in Figure 3.30d. This result suggests that the TC, although it is near the right-entrance region of the jet, is not being significantly aided in divergence by the circulation induced by the jet due to the TC’s own warm outflow displacing the region favorable for ascent.

There are areas that do change between forcing for ascent and descent between 20 h and later times, such as directly over the TC in the TC–jet simulation (which changes from forcing for ascent to descent), and, in both TCs, the magnitude of the forcing changes. The reason for the sign changes and increases in magnitude are varied and difficult to discern because, as shown in the $\mathbf{Q}$-vector equation, the $\mathbf{Q}$-vector is dependent on first and second derivatives of temperature and non-divergent wind.

In terms of temperature, in both simulations the diabatic outflow of the TC results in a large thermal ridge surrounding the TC in the 400–200-hPa layer (not shown). This area of increased temperature causes a change in the sign of $\frac{\partial T}{\partial y}$ across the TC. Initially, the sign of $\frac{\partial T}{\partial y}$ is negative; but, as the TC warms its environment, the sign of the gradient changes from positive to the south of the TC to negative to the north and these gradients increase with time as the TC further warms its environment. The magnitude of $\frac{\partial T}{\partial x}$ also changes as
the TC warms its environment and as the cold-core trough approaches.

Concerning the non-divergent wind, there are large changes in the non-divergent flow due to the change in orientation of the trough (in the TC–trough simulation) and the formation of a downstream ridge (in both simulations). As the trough becomes negatively tilted and wraps south of the TC, and the ridge builds downstream, changes occur to both $\frac{\partial v_v}{\partial x}$ and $\frac{\partial v_v}{\partial y}$.

These changes to all of the relevant gradients in the Q-vector equation throughout the simulation act to change the relative importance of $Q_1$ (which depends on $\frac{\partial v_v}{\partial x}$) and $Q_2$ (which depends on $\frac{\partial v_v}{\partial y}$) and, therefore, the final sign and magnitude of Q-vector divergence. As previously mentioned, however, diagnosing the exact changes in each derivative and second derivative that lead to changes in Q–vector divergence is beyond the scope of this project.

As an aside, it is interesting to note that comparing the Sutcliffe–Trenberth form of the omega equation, which does not include deformation, (Fig. 3.31; note that in this figure the vertical motion is calculated explicitly as in Fischer et al. (2017)) to the Q-vector form, which includes deformation, (Fig. 3.30d) at 40 h in the TC–trough simulation reveals that deformation clearly plays an important role in modulating ascent / forcing for ascent southwest of the TC, where the Sutcliffe–Trenberth omega equation does not show ascent. As discussed in Martin (2006), deformation is a non-negligible term in thermal ridges and regions of upper-level frontogenesis, which likely explains the sign difference between the two omega equations in the core of the jet and southwest of the TC. It was suggested in Figure 3.27 that upper-level frontogenesis was occurring along the TC’s western side, and a similar feature occurs on the southwestern side of the TC when cross sections are taken through that region, as well (not shown). Both equations do capture the largest values of ascent / forcing for ascent downstream of the trough well.

Throughout the simulation, the TC–trough case has greater forcing for ascent in the large-scale environment of the TC than the TC–jet simulation. This greater forcing for ascent can act to moisten the column underneath it (McTaggart–Cowan et al. 2008), which
aids in the development of convection and generation of diabatic heating. The differences in forcing for ascent in the two simulations is especially apparent west and southwest of the TCs, which is a location that had large differences in 700–500-hPa RH (Fig. 3.21). The location of this synoptic lift upshear of the TC could make the moistening there in the TC–trough simulation even more important to TC intensification, as it has been shown that upshear moistening is favorable to TC intensification (Kaplan et al. 2015; Rios-Berrios et al. 2016). The moistening of the TC–trough simulation later in the simulation west and southwest (upshear) of the TC (Fig. 3.22d,f), also matches the region that develops strong Q-vector forcing for ascent. This forcing for ascent and moistening could be one reason that the core convection of the TC is more radially extensive west of the TC in the TC–trough simulation. Additionally, this region has greater coverage of isolated precipitation than the TC–jet simulation, which could be a result of the moistening of the environment caused by the large-scale forcing for ascent.

3.4 Conclusions

In this chapter, the use of an idealized numerical model allowed for examination of TC–trough interaction with finer temporal and spatial resolution compared to the reanalysis used in the previous chapter. The TC–trough interaction in the modeling framework used in this chapter was found to be beneficial to TC intensity relative to a TC–jet interaction. Even though the TC–trough TC experienced greater shear throughout most of the simulation, especially early in the simulation when it caused significant vortex tilt, the TC–trough TC was more intense than the TC–jet TC starting after 20 h, and it was likely undergoing rapid intensification during the final 10 h. Throughout the simulation, starting between 10–20 h, the TC–trough TC had greater diabatic heating over a larger area. This result is indicative of a more favorable convective environment, in spite of the enhanced shear, which may have increased the likelihood of the downshear reformation and convective bursts that led to intensification. The TC–trough simulation showed several factors that were beneficial
to convective development and diabatic heating.

The TC–trough simulation had greater midlevel moisture in the near-TC environment and the RH differences increased with time. The large differences in moisture were, at least partly, driven by large-scale differences in divergence and ascent near the TC. The TC–trough simulation exhibited much greater divergence and forcing for lift over a large area around the TC.

The differences in divergence were attributed to the synoptic-scale differences between the two simulations. First, EFC, which forces divergence through a Sawyer–Eliassen, balanced-vortex response was larger in the TC–trough simulation. Large values of EFC in the TC–trough simulation also occurred in regions of lower inertial stability than in the TC–jet simulation (excluding EFC occurring within the trough), which likely aided the development of the stronger, more horizontally extensive, outflow jet north of the TC. Second, the location of the TC near the inflection point of the upper-level trough in the TC–trough simulation also provided a mechanism for enhanced ascent. The curvature of the synoptic flow resulted in the TC being located in the region where divergence is favored due to the transition from subgeostrophic flow around the base of the trough to supergeostrophic flow around the downstream ridge. Finally, the synoptic flow also resulted in much greater forcing for ascent, as diagnosed by $Q$-vector convergence, due to the TC’s location downstream of the trough.

Although it was shown that the synoptic environment of the TC–trough simulation increased the dynamic and thermodynamic favorability of the environment, it is difficult to prove that the more favorable environment is directly responsible for the TC–trough TC’s two periods of intensification due to well-placed convective bursts. A conclusion can be drawn, however, that the aforementioned factors contributed to the more favorable environment of the TC–trough simulation, making well-placed convective bursts more likely. For future work, an ensemble of simulations would shed light on whether these stochastic events are more likely in the more favorable environment created by the upstream trough.
An impact of the increased diabatic heating over a larger area in the TC–trough simulation that was not discussed at length in this chapter is the different sizes of the two TCs. The TC–trough simulation TC is much larger (over 2.5 times as large by the end of the simulation based on the 15 m $s^{-1}$ wind contour) than the TC–jet simulation TC, with size differences first appearing around 25–30 h. Previous studies have shown that TCs in more moist environments, as well as those that have large diabatic heating outside of the RMW, can lead to TC size increases (Hill and Lackmann 2009; Wang 2009). The increased divergence and lift over a large area forced by the synoptic flow, as well as the resulting changes in moisture and diabatic heating, thus provides a mechanism for TC size increase; this will be a major focus of the next chapter.
### 3.5 Table and figures

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<th>Domain</th>
<th>Grid Size</th>
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<th>Convective Parameterization</th>
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</thead>
<tbody>
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<td>457 x 199</td>
<td>30 km</td>
<td>Tiedtke (Tiedtke 1989; Zheng et al. 2011)</td>
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<tr>
<td>2</td>
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<td>10 km</td>
<td>Tiedtke (Tiedtke 1989; Zheng et al. 2011)</td>
</tr>
<tr>
<td>3</td>
<td>583 x 370</td>
<td>3.33 km</td>
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Figure 3.27: Radius–pressure, TC–centered, east–west, 5-h time-average cross section of PV (PVU; shaded) and storm-relative zonal wind (m s\(^{-1}\); contoured) at 55 h. Zonal wind is contoured every 4 m s\(^{-1}\) between -12 m s\(^{-1}\) and 12 m s\(^{-1}\), and positive (negative) values are solid (dashed). 0 m s\(^{-1}\) contour not plotted. Inset is 200-hPa PV (PVU; shaded) and the red line indicates the location of the cross section, where the PV for the cross section was averaged from 10 km south of the red line to 10 km north of the red line. Green circle denotes area of the PV frontal feature.
Figure 3.28: Radius–pressure, TC–centered, east–west, 9-h time-average cross section of vertical motion (m s$^{-1}$) for the same time and location as Figure 3.27). Inset is 200-hPa PV (PVU; shaded) and the red line indicates the location of the cross section, where the PV for the cross section was averaged from 10 km south of the red line to 10 km north of the red line. Green circle denotes area of the PV frontal feature.
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Figure 3.31: TC–centered, 200-hPa, non-divergent wind streamlines (m s\(^{-1}\)) and omega at 300 hPa (Pa s\(^{-1}\); shaded) of the TC–trough simulation at 40 h derived from the Sutcliffe–Trenberth omega equation, where non-divergent wind was substituted for geostrophic wind. Red (blue) shading indicates descent (ascent). Horizontal axis is longitude (deg) and vertical axis is latitude (deg). Green dot denotes TC center.
4. TC–trough interactions in idealized numerical modeling: sensitivity experiments

4.1 Introduction

In the previous chapter, the differing effects on TC intensity change of two different large-scale flow patterns was analyzed. Because these initial flow patterns were so different, it raises the question of how TC intensification might change across a spectrum of different trough scales and TC configurations.

As discussed in both the introduction and chapter 3, few idealized modeling studies have explored TC intensification while changing characteristics of the TC and/or the trough. Kimball and Evans (2002) changed the scale and depth of an upper-level cutoff low with which a TC was interacting in four simulations. They found that a strong, vertically shallow trough was more favorable for intensification because this trough morphologies allowed superposition of the TC and trough. Weaker troughs were eroded quickly by the TC’s diabatic heating and could not approach the TC closely enough for superposition to occur. In these weak trough cases, shear increased as the trough approached, which led to TC weakening until the trough was eroded. They also found that more vertically extensive troughs created a deep layer of vertical wind shear, which prevented intensification altogether. Additionally, they showed that in three out of four trough interactions analyzed, outflow was hampered on the trough side of the TC compared to simulations with no trough, reducing intensification.

Leroux et al. (2016) did not change the initial characteristics of the trough, but they did test changing the initial intensity of the TC vortex in their simulations, as well as the relative position of the TC to the trough. As mentioned in previous chapters, they found the existence of a “sweet spot” in terms of distance between the TC and trough, that allowed for a tongue of midlevel, positive PV advection from the trough into the TC core. Moving the TC one degree closer to the trough resulted in stronger intensification compared to the
control run, while moving the TC farther away led to weaker intensification than in the
control run. In terms of initial TC intensity, they found that stronger vortices intensified
most. They attributed the greater intensification to enhanced confluence between the TC
and the trough due to the stronger TC, which allowed for a longer time period of positive
PV advection from the trough into the TC compared to initially weaker TCs.

Komaromi and Doyle (2018) experimented with changing the initial scale of the trough
and the initial location of the TC relative to the trough. As in Leroux et al. (2016), they
found a sweet spot in terms of initial TC location that led to greater rates of intensification.
In this case, they found that the sweet spot was when the relative distance between the
TC and the trough was 0.2–0.3 times the wavelength of the upstream trough in longitude
and 0.8–1.2 times the amplitude of the trough in latitude. This relative positioning allowed
the TC to be close enough for the positive effect of enhanced poleward outflow due to the
trough, but was far enough away not to increase shear to unfavorable values.

The size, and size changes, of a TC interacting with a trough have not been well-
studied. Although there is less research on the topic compared to TC intensity, there have
been some attempts to understand TC size using theory, observations, and modeling. Merrill
(1984) found that TC size and TC intensity are only weakly correlated. He hypothesized,
however, that TCs grow in size not when they intensify, but rather, when there is a net
convergence of relative angular momentum into the TC. He therefore surmised that changes
in the large-scale environment, such as the approach of a trough, can force relative angular
momentum convergence and, thus, expansion of the TC. Kimball and Evans (2002), in their
idealized study of TC–cutoff interactions, found that the import of high relative angular
momentum air at large radii did increase TC size in the first 24 h of the simulations.

Although no study has directly looked at size expansion of TCs interacting with troughs
in observations, Schenkel et al. (2018) used the Climate System Forecast Reanalysis (CFSR)
and the Geophysical Fluid Dynamics Laboratory (GFDL) High-Resolution Forecast-Oriented
Low Ocean Resolution model (HiFLOR) to evaluate the extent of TCs’ 8 m s\(^{-1}\) wind contour
during extratropical transition (ET). They found that in North Atlantic TCs undergoing ET, which occurs in the presence of a trough, there was no change in the TCs’ size (HiFLOR) or the TC slightly decreased in size (CFSR). They did, however, express the caveat that this result may be attributed to the relatively short period of time it takes for a TC to undergo ET (median length of 1–1.5 d) relative to the longer period of time it takes for size expansion to occur (median rates of expansion of the 12 m s\(^{-1}\) contour and outermost radius of the TC are 18.1 and 10.9 km d\(^{-1}\), respectively; Chavas and Emanuel 2010).

Hill and Lackmann (2009) performed idealized modeling experiments in which the relative humidity outside of 100 km from the TC center varied between 20–80%. They found that TCs embedded in drier environments were smaller than those in more moist environments. The TCs in the more moist environments had similar magnitudes of inner core PV and precipitation, but they had greater outer-rainband activity. The greater low-level PV generation in the more active outer rainbands of the more moist TCs led to broadening of the low-level TC PV and, thus, the TC wind field.

In this chapter, 28 individual model runs will be analyzed across a spectrum of trough and TC configurations to explore how trough and TC characteristics affect TC intensification and size. In addition to the TC–jet simulation shown in chapter 3, the initial states of either the TC or the trough will be tested within a multivariate parameter space. The parameter space consists of: (1) the initial TC vortex intensity, (2) the initial trough scale, and (3) the initial shear experienced by the TC. While some previous studies have looked at different parts of this parameter space, only two (Leroux et al. 2016; Komaromi and Doyle 2018) have looked at multiple parts of it together, and only one other study has considered the problem from a strictly idealized modeling perspective (Komaromi and Doyle 2018).

The rest of this chapter is organized into four major sections: Section 4.1 is the methodology of how the parameters are changed; Section 4.2 is an overview of TC intensity and size evolution in the 28 sensitivity experiments, with a focus on size differences; Section 4.3 will explore the causes of the size differences between a representative small TC simulation and
a representative large TC simulation; and Section 4.4 will analyze the size differences of a representative TC that starts growing relatively quickly but then the growth is stunted and a representative TC that grows continuously. Finally, Section 4.5 will provide conclusions.

4.2 Methodology

4.2.1 Initial TC vortex strength

The first parameter is the initial vortex strength. As discussed in the previous chapter, WRF TC-bogussing was used to insert a TC vortex with a 60-km RMW into the simulations. For the sensitivity experiments, simulations were initialized with either a 10 m s\(^{-1}\), 12.5 m s\(^{-1}\), or 15 m s\(^{-1}\) TC vortex. The 15 m s\(^{-1}\) upper-bound was chosen because it was found that in simulations initialized with a 20 m s\(^{-1}\) initial vortex the TCs grew intense quickly, which prevented the close approach of the trough due to the TCs’ strong outflow. The 10 m s\(^{-1}\) lower-bound was chosen because in some simulations initialized with vortices weaker than 10 m s\(^{-1}\), the TCs were unable to develop.

4.2.2 Initial trough scale

The initial trough scale was adjusted by changing the value of the zonal scale modifier, \(m_z\), in Eqs. 3.4 and 3.5, which set the shape of the weights of the trough temperature perturbation. The chosen values of \(m_z\) are four, seven, and 10. These values represent troughs that range in zonal scale, based on their half wavelengths, between \(\sim 1900–3900\) km.

A complicating factor is that changing the zonal trough scale results in changes to the trough’s zonal temperature gradients. Any changes in the gradients between the scales may affect the interactions with the TCs, because the strength of the temperature gradient changes the associated vertical wind shear along the trough’s leading edge, which is where the TC first interacts with the trough.

To address the issue of the varying zonal temperature gradient between the different trough scales, the magnitude of the trough temperature anomaly was adjusted for the dif-
ferent trough scales. Specifically, the temperature anomaly magnitude of the trough was adjusted so that the zonal temperature gradient one standard deviation from the trough temperature anomaly’s center was equal for all simulations.

Using scale 7 as the control, this temperature adjustment resulted in a trough temperature anomaly magnitude that is 39% weaker in the scale 4 trough simulations and 40% greater in the scale 10 trough simulations. It is notable that this methodology results in both the weakest and smallest trough in the scale 4 simulations, and the strongest and largest trough in the scale 10 simulations. Figure 4.1 shows vertical cross sections of the trough temperature anomalies for the three scales, and Figure 4.2 shows the resulting 200-hPa wind speed associated with these anomalies.

An unavoidable consequence of changing the magnitude of the trough temperature anomaly to control for the zonal temperature gradients is that this adjustment intrinsically changes the meridional temperature gradients between the different scales, as well. Because the TCs are located more east of the trough than south and, thus, they are more closely interacting with the trough’s eastern edge, controlling for the zonal temperature gradient was deemed more crucial.

4.2.3 Initial TC shear

The initial shear in which the TC is embedded is based on a point value of 850–200-hPa shear at the TC’s initial location. The point shear values were calculated for each run prior to inserting the TC vortex, but after inserting the jet and trough.

The initial TC latitude was constant in all runs at 8.47° S, which is the same as the initial TC latitude in the TC-jet simulation for consistency, and different initial shear values were achieved by changing the initial longitude of the TC vortex. The TC vortex was moved farther in longitude upstream or downstream relative to the trough until the desired point shear value was reached (Fig. 4.2). Therefore, the initial shear value of the TC is inherently tied to the initial distance between the TC and the trough. In this way, the initial shear

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1 Because the simulations are on an \( f \)-plane, the initial latitude is only a location reference.
choices in this study are similar to the sensitivity experiments by Leroux et al. (2016) and Komaromi and Doyle (2018) in which the relative distance between the TC and the trough was modified.

The initial shear ranges between 7.5 m s$^{-1}$ and 12.5 m s$^{-1}$. The minimum shear value was chosen because shear values below 7.5 m s$^{-1}$ allowed the TCs to develop quickly, and their enhanced outflow prevented strong interaction with the approaching trough, similar to what occurred in the TCs that were initialized as 20 m s$^{-1}$ vortices. The maximum shear value was chosen because shear greater than 12.5 m s$^{-1}$ could only be obtained by initializing the TC almost directly under the core of the trough.

Figure 4.2 provides a visual summary of the different sensitivity experiments. In addition to the initial 200-hPa wind speeds for each of the three trough scales, Figure 4.2 shows the initial locations of the TCs chosen to achieve the three different shear values. The three initial TC vortex strength values are not represented in the figure, but they complete the parameter space, resulting in 27 simulations.

4.3 Results of the sensitivity experiments

This section gives an overview of the TC intensities, both in terms of minimum central pressure and wind speed, and TC size evolutions of the sensitivity experiments.

4.3.1 TC intensity

Table 4.1 shows the minimum central pressures of the sensitivity experiments at the final simulation time (72 h) and Table 4.2 shows the final, maximum, azimuthally-averaged wind speeds. It is important to note, however, that the TCs have not achieved a steady-state by the end of the simulation period. Figure 4.3 shows the 72-h pressure traces of the 12.5 m s$^{-1}$ initial vortex sensitivity experiments. The TCs are almost all still intensifying at 72 h. Likewise, the 72-h maximum wind speed traces for the same set of sensitivity experiments (Fig. 4.4) show that all but one of the TCs are intensifying in terms of the maximum wind speed.
Because the TCs have not reached a steady-state, the goal is not to ascertain which parts of the parameter space (shear/distance from the trough, initial vortex strength, and trough scale) are most favorable for maximum, steady-state TC intensity. It is possible, however, to determine the rate of intensity change during the 72-h period; those TCs with lower pressure or greater wind speed at the end of the period intensified more quickly than other TCs in the parameter space.

In the following discussion, different parameters will be held constant in order to ascertain the influence of each parameter on intensification. Because there are large differences in the TC sizes (which will be the major focus of the latter part of this chapter), the discussion will focus on intensification based on minimum central pressure, rather than maximum wind speed, as maximum wind speed is determined by the strength of the TC’s pressure gradient, which is directly affected by the TC size.

4.3.1.1 Effect of initial vortex intensity

The most coherent pattern in the parameter space was that greater strength initial vortices corresponded to lower minimum pressures at the end of the 72-h period (Table 4.1); only two simulations did not fit this pattern. This result indicates that greater strength initial vortices correspond to faster intensification rates during the 72-h period. This result is similar to the results of sensitivity experiments by Leroux et al. (2016), who found that greater initial TC intensity led to lower minimum pressure at later times relative to weaker initial TCs. A likely reason for this result is that greater initial vortex strength would result in larger surface heat fluxes, which could result in greater TC intensification.

4.3.1.2 Effect of initial shear / distance from the trough

Controlling for initial vortex strength and scale of the trough, the relationship between initial shear and rate of minimum pressure decrease is dependent on the trough scale. The scale-4 simulations showed little sensitivity to the initial shear value in the 10 m s\(^{-1}\) and 15 m s\(^{-1}\) initial vortex simulations and increasing minimum central pressures with increasing
initial shear in the 12.5 m s$^{-1}$ simulations. The rate of intensification of scale-7 and scale-10 simulations, in all but one case, however, is greater when initial shear is larger (Fig. 4.3; Table 4.1).

The result that higher shear results in a greater intensification rate does not match the results of TC–trough studies by Leroux et al. (2016), Komaromi and Doyle (2018), or chapter 2. It is possible that for even higher shear than examined in these experiments, there would be a point at which greater initial shear results in lower intensification rates. Additionally, shear is not fixed in these sensitivity experiments, and the TC can modulate its environment, and therefore potentially reduce the shear it experiences, during the simulation. Alternatively, differences in TC intensification between this study, in which the TCs are initially weak as they interact with the trough, and the aforementioned studies, in which the TCs are more mature, could be a result of these differences in TC intensity at the onset of trough interaction, which likely changes how the TC and trough interact initially.

Because these simulations are quite weak and must develop convection in the first several hours of the simulation, it may be more helpful to compare these results to those of less mature TCs. Bracken and Bosart (2000) looked at the development of tropical cyclones in the western North Atlantic and showed that, in a composite sense, there exists a weak upper-level trough just upstream of the developing disturbance. They also found that developing disturbances experienced, on average, 10 m s$^{-1}$ of vertical wind shear. Taking these results together, they suggested that some vertical wind shear is favorable to developing tropical cyclones because the shear leads to large-scale forced ascent over the disturbance. Similarly, Fischer et al. (2017) and Fischer (2018) found that QG forcing for ascent due to an upstream trough is beneficial to TCs that rapidly intensify soon after genesis, provided this forcing is concentrated upshear and close to the TC center. These TCs were also under relatively large shear. Therefore, there is some evidence that enhanced shear, and closer proximity to a trough (which are tied together in these sensitivity experiments) may be beneficial in the development of weak TCs.
4.3.1.3 Effect of initial trough scale

Controlling for initial shear and initial vortex strength, Table 4.1 shows that there is little coherent signal in how the scale of the trough affects the final minimum pressure of the TC. If, however, the scale-4 troughs are excluded, scale-10 troughs, in all but one case, correspond to slower intensification rates compared to scale-7 trough simulations. This latter result is consistent with the results of chapter 3, in which it was shown that smaller, weaker troughs are more favorable for intensification than larger, stronger troughs. Given that result, it is surprising that the scale-4 simulations are not always the most favorable for intensification, and highlights the diversity of outcomes within this parameter space.

4.3.2 TC size

As mentioned previously, the final minimum pressures and final maximum wind speeds (Tables 4.1, 4.2) in the sensitivity experiments strongly suggest that there exist large differences in TC size; some experiments have similar minimum pressures but dissimilar wind speeds, indicating a difference in their pressure gradients and, therefore, size. For instance, the 12.5 m s$^{-1}$ initial vortex, scale-4, shear 7.5 m s$^{-1}$ simulation and the 12.5 m s$^{-1}$ initial vortex, scale-10, shear 12.5 m s$^{-1}$ simulation have similar pressures at 72 h (960.0 and 960.6 hPa, respectively). They do not, however, have similar final wind speeds (39.2 m s$^{-1}$ and 33.8 m s$^{-1}$, respectively). This discrepancy in wind speed implies a weaker pressure gradient, and therefore a larger TC pressure field, in the 12.5 m s$^{-1}$ initial vortex, scale-10, shear 12.5 m s$^{-1}$ simulation.

To confirm that there are indeed differences in the sizes of the TCs, the surface kinetic energy (KE) of each sensitivity experiment was calculated within the 150–300-km TC-centered annulus. The inner 150 km was excluded to reduce the influence of the strength of the TC’s inner-core maximum wind speeds in the KE calculation, which could mask size differences.

Figure 4.5 shows the surface KE of the experiments. For clarity, only the 12.5 m s$^{-1}$
initial vortex experiments are shown, as the other initial vortex strength experiments showed similar results. Figure 4.5 indicates that there are indeed size differences in the TCs. There are three simulations that are significantly larger than the rest (hereafter the “always large” (AL) group): (1) the scale-7, shear-10 simulation (solid, blue line), (2) the scale-7, shear-12.5 simulation (dashed, blue line), and (3) the scale-10, shear-12.5 simulation (dashed, red line). These same three simulations also have significantly greater KE than the other simulations in the 10 m s$^{-1}$ and 15 m s$^{-1}$ initial vortex simulations (not shown).

The scale-4, shear-10 (solid, black line) and scale-4, shear-12.5 (dashed, black line) simulations (hereafter the “start large end smaller” (SLES) group), have similar KE to the large group for approximately half of the simulation period, until between 40–50 h, at which point their KE begins to lag the AL group. All of the shear-7.5 simulations (dotted lines), and the scale-10, shear-10 simulation (solid, red line) are relatively small beginning around 30–35 h (hereafter the “always small” (AS) group).

Another way to consider TC size is to consider the radial extent of the 15 m s$^{-1}$, azimuthally-averaged surface wind contour. Figure 4.6 reveals, as the KE suggested, that there are two approximate clusters of TCs by the end of the simulation. The same three large KE TCs (the AL group) comprise the large group of TCs, and the other six TCs (the SLES and AS groups), as well as the TC–jet simulation, comprise the smaller group. As with KE, the same three simulations are significantly larger than the other simulations in the 10 m s$^{-1}$ and 15 m s$^{-1}$ simulations (Table 4.3). These large simulations are about twice as large as the smallest TCs.

In terms of patterns in size across the parameter space (Table 4.3), the strongest relationship appears as a function of the initial vortex intensity. The stronger initial vortices typically end up larger in size than weaker initial vortices, which fits with previous work by Ma et al. (2014), who showed that greater surface fluxes can lead to larger TCs.

The other signal in the parameter space is that, for the same scale, TCs under greater initial shear generally develop to be larger in size. This result indicates that there is a shear-
related mechanism that is causing higher shear cases to be larger than their smaller shear counterparts. Alternatively, this signal could be a result of the TC’s placement relative to the trough, since it is tied to initial shear magnitude. Additionally, the synoptic flow pattern seems to be important to TC size not only through a mechanism related to shear, but through the presence of the upper-level trough, as the TC–jet simulation TC is smaller than any of the other 27 TCs in the parameter space based on the 15 m s$^{-1}$ azimuthally-averaged wind contour (Table 4.3), suggesting that the presence of a trough enhances TC size.

In chapter 3, it was shown that the TC–jet simulation TC had radially-confined convective activity that did not extend far beyond the inner core. In contrast, it was shown that the control simulation (12.5 m s$^{-1}$ initial vortex, scale 7, shear 10), which is amongst the cluster of AL TCs, had a large precipitation shield extending north of the TC. Therefore, the differences in radial extent of the 15 m s$^{-1}$ wind contour of the AL, SLES, and AS TCs could be linked to the extent and coverage of precipitation.

Figure 4.7 shows the reflectivity of the 12.5 m s$^{-1}$ initial vortex sensitivity experiments with the 10 and 15 m s$^{-1}$ contours overlaid at 35 h, which is about the time that the AS cluster has separated from the AL and SLES clusters in terms of KE (Fig. 4.5). While all of the simulations show that precipitation outside of the inner core is enhanced north of the TC, there is a striking difference in the structure of this precipitation between the AL/SLES and AS TCs.

The four AS TCs (Fig. 4.7a,d,g,h) have a large, rain-free region between the inner-core convection and the band of precipitation to their north and east (the scale-4, shear-7.5 (Fig. 4.7a) simulation forms this low-reflectivity region closer to 38 h). The AL TCs (Fig. 4.7e,f,i), however, show no such low-reflectivity region. Rather, the AL TCs have an extensive shield of precipitation north of the TC that remains connected to the inner-core.

Given the stark differences in TC reflectivity at 35 h between the TCs that end up large and those that end up smaller, it is hypothesized that the evolution of the TCs’ precipitation plays a key role in determining the TCs’ size evolution.
The evolution of the SLES reflectivity pattern lends support to this hypothesis. At 35 h, the SLES TCs show a reflectivity pattern (Fig. 4.7b,c) similar to that of the AL TCs (Fig. 4.7e,f,i). Later in time, however, the size of the precipitation shield of the SLES TCs shrinks in spatial scale relative to the AL TCs (Fig. 4.29; discussed in greater detail in section 4.5) and the reflectivity pattern becomes more similar to the reflectivity pattern of the AS TCs (not shown). Thus, the SLES TCs begin large, when they have radially extensive precipitation, but ultimately end up more similar in size to the AS TCs as their precipitation area is reduced.

The remainder of this chapter is split into two major parts. First, the cause of the initial split between the AL and AS TCs around 30–35 h is explored using two representative simulations. Second, the cause of the second split in TC size between the AL and SLES simulations between 40–50 h is explored also using two other representative simulations.

4.4 Comparison of representative AL and AS TCs

In this section, a representative AS and AL TC are compared in order to elucidate the causes of their size difference. The chosen AS run is the scale-10, shear-10, vortex-12.5 simulation and the AL run is the scale-7, shear-10, vortex 12.5 simulation. These runs were chosen because the effects of different upstream trough scales are less studied than the effects of different shear magnitudes and keeping the initial shear the same in both runs removes one initial degree of freedom. Additionally, these TCs were small and large, respectively, in both the 10 m s\(^{-1}\) and 15 m s\(^{-1}\) initial vortex simulation, as well.

As discussed in the last section, it is hypothesized that the size differences of the two TCs are due to the greater coverage of precipitation, and thus diabatic heating, outside of the inner core of the TC in the large TC. This hypothesis, and the potential reasons that there is greater precipitation outside of the inner core of the large TC, will be subsequently explored.
4.4.1 Kinetic energy

Figure 4.8 shows the surface KE in the 150–300-km, TC-centered annulus of the representative large and small simulations, as well as the differences between them. Both simulations have similar KE until around 32 h, at which time the large TC begins to increase in KE relative to the small TC. The KE calculation attempts to account for differences in TC intensity by excluding the inner 150 km of the TC. Regardless, the representative large and small TCs are similar in intensity throughout most of the simulation, having similar minimum pressures (Fig. 4.3) and maximum wind speeds (Fig. 4.4) for the first ~60 h. Thus, the differences in KE during the early part of the simulation are due to differences in TC size.

4.4.2 Convective pattern

The reflectively fields of the two TCs surrounding the bifurcation time indicate large differences in precipitation that emerge during this period (Fig. 4.9). Ten hours prior to the bifurcation time (Fig. 4.9a,b), both the large and small TCs have similar reflectivity patterns, in which convection is displaced northeast of the TC center. At the time of the bifurcation (Fig. 4.9c,d), the two TCs begin to diverge in terms of their precipitation patterns. The small TC (Fig. 4.9c) has a large, rain-free region between the inner core of the TC and a curved, outward-propagating rainband northeast of the TC. The large TC (Fig. 4.9d), meanwhile, lacks the rain-free region between the inner-core precipitation and precipitation north and east of the TC.

The large TC does show a similar outward-propagating rainband as the small TC, which is clear in animations of reflectivity in the large\(^2\) and small\(^3\) TCs. This rainband, however, propagates more slowly (~8 km h\(^{-1}\) in the large TC vs. ~11 km h\(^{-1}\) in the small TC), stagnates around 250-300 km from the TC center (compared to between 450-500 km

\(^2\)http://www.atmos.albany.edu/student/cpeirano/Animations/wrf_trough_simulations/diss_runs/sensitivity_cube/vortex_125/scale_7/shear_10/Animation_500_ref.html

\(^3\)http://www.atmos.albany.edu/student/cpeirano/Animations/wrf_trough_simulations/diss_runs/sensitivity_cube/vortex_125/scale_10/shear_10/Animation_500_ref.html
in the large TC), and does not produce a rain-free region behind it, as in the small TC. Instead, as shown in Figures 4.9d,f, the stagnating rainband and the precipitation behind this rainband act to form a large precipitation shield connected to the TC’s core, unlike in the small TC.

By 10 h after the bifurcation time, the large TC (Fig. 4.9f) shows a robust area of precipitation north-northeast of the TC that remains connected to the inner-core of the TC. The small TC (Fig. 4.9e) does not have this active area of northern convection near the TC and, in fact, the outward-propagating rainband has weakened and shrunken substantially as it moved farther from the TC. It should be noted that both TCs maintain similar strength inner-core convection during the period shown.

The wavenumber one pattern in the reflectivity field throughout the 20-h period shown in Figure 4.9 is indicative of vertical wind shear, which displaces convection downshear within 100 km of the center and downshear-right beyond 100 km (Corbosiero and Molinari 2002, 2003; Riemer et al. 2010b). By design, the TCs are both initially experiencing about 10 m s\(^{-1}\) of vertical wind shear to start, but the vertical wind shear can change with time, which may account for the differences in the reflectivity.

Deep-layer wind shear for the 200–800-km and 0–300-km annuli is shown in Figure 4.10. The wind shear remains in the high to moderate range throughout the 20-h period shown in Figure 4.9, and shear direction throughout the period is toward the northeast in both (not shown), consistent with the location of the highest reflectivity to the northeast. That both TCs are experiencing comparable shear directions, as well as comparable magnitudes of 200–800-km shear during the period is somewhat surprising given the large differences in reflectivity. It is possible that the differences in reflectivity are related to the reduced 0–300-km shear in the small TC between 20–30 h, during which shear values in the small TC are 2–3 m s\(^{-1}\) smaller than in the large TC.

One possible explanation for why the reflectivity fields look different under comparable 200–800-km shear is that there may be structural differences in the TCs’ vortices. In mod-
erate shear, TC evolution has been shown to be highly sensitive to random noise and, thus, it is possible that even under comparable 200-800-km shear magnitudes and directions there can be differences in the TC vortex (Zhang and Tao 2012). Given the high sensitivity of the TC vortex in moderate shear, the 2–3-m s$^{-1}$ difference in 0–300-km shear between 20–30 h may allow for a better-organized TC vortex in the small TC during this time. One shear-related mechanism that can affect the placement and amount of convection through forced ascent and concomitant moistening due to an induced secondary circulation, as discussed in chapter 3, is vortex tilt (Jones 1995; Reasor et al. 2004; Zhang and Tao 2012; Reasor et al. 2013; Riemer 2016). Differences in the magnitude and duration of vortex tilt in the small and large TCs could result in differences in moisture, and therefore precipitation coverage, downshear/downtilt in the TCs, where large differences are observed (Fig. 4.9).

### 4.4.3 Vortex tilt

Figures 4.11 and 4.12 show the 0–5-km tilt magnitudes of the simulations stratified by shear and trough scale, respectively. The vortex center at 5 km was determined using the minimum in smoothed pressure, the same as the surface vortex-finding method. The 0–5-km vortex tilt was chosen because during the initial 10 h of the simulations, most of the TC vortices under high shear do not coherently extend above 5 km at all times.

Considering just the two cases being explored in this section (bold lines in Figures 4.11 and 4.12), the small TC does have a smaller maximum tilt magnitude (50 km) than the large TC (85 km). The small-TC tilt is approximately 40% smaller than the large-TC tilt despite both having the same initial shear value and similar shear values for an extended period (Fig. 4.10). The small TC reduces its tilt to less than 20 km approximately 4 h prior to the large TC, but the timing of the tilt reducing to within 20 km, where it becomes approximately steady (24 h in the small TC vs 28 h in the large TC), does not fully explain the differences in TC reflectivity. Differences in reflectivity persist well beyond the 28 h time period, but the tilts are similar in both the small and large TCs after that time.
Given that changes in the tilt have a stochastic component, as shown in chapter 3 and previous studies (Zhang and Tao 2012; Nguyen and Molinari 2015), the tilt magnitudes of all of the 12.5 m s$^{-1}$-initial vortex TCs are also plotted in Figures 4.11 and 4.12 to see if the small and large TCs are representative of their respective clusters.

Indeed, the AL and SLES TCs almost all have greater magnitudes of vortex tilt (mean maximum tilt magnitude is 109 km) and correspond to the TCs with greater precipitation coverage at the 35 h bifurcation point, after which time they are larger in size (Fig. 4.7). The AS simulations have the smallest tilt magnitudes (mean maximum tilt magnitude is 43 km) and correspond to reflectivity patterns having rain-free regions behind the outward-propagating rainband (Fig. 4.7).

Given that the AL and SLES TCs are those with initial shear magnitudes of 10 and 12.5 m s$^{-1}$, the small tilt of the shear-10, scale-10 simulation, which has been the focus of this section, is an exception. A possible explanation is that in moderate shear ranges, as mentioned, the TC’s behavior has greater variability (Zhang and Tao 2012). Indeed, Figure 4.11 shows that the TCs with 10 m s$^{-1}$ of initial shear have the most variability in tilt magnitude. The increased sensitivity explanation is inconsistent, however, with the fact that the scale-10, shear-10 simulations were also small in the 10 m s$^{-1}$ and 15 m s$^{-1}$ initial vortex simulations. This result suggests that there is a consistent physical reason why this TC–trough configuration produces smaller TCs compared to other TCs with the same initial shear magnitude; however, that reason remains an open question and will be revisited in the future work section of Chapter 5.

Stratifying by scale (Fig. 4.12), rather than shear, the scale of the trough does not seem to have a large effect on the tilt magnitude or duration. The scale 4 simulations have tilts as large as, and that persist as long as, the scale 7 and scale 10 (except the scale-10, shear-12.5 simulation) simulations. The similarities in tilt magnitude are likely related to the initial shear value, rather than the trough scale. The duration of tilt not being tied to the scale of the trough is more surprising given that larger, stronger troughs might be expected
to cause enhanced shear over the TC for an extended period of time, as shown in chapter 2 and hypothesized in Hanley et al. (2001). This result is perhaps due to small sample size.

The tilt of the large and small TCs reduce by different means. Figure 4.13 shows the progression of the 5-km vortex of the small and large TCs relative to the surface center. Both simulations show the initial tilting of the vortex, with the large TC experiencing greater tilt. The reduction of the small TC’s tilt, like almost all of the other 12.5 m s\(^{-1}\) initial vortex simulations (not shown), occurs as the 0–300-km shear falls (Fig. 4.10), which allows the vortex to become vertically stacked by 24 h. As an aside, the 200–800-km shear is poorly related to tilt reduction in most simulations (not shown), indicating that it is a poor measure of the shear that affects the TCs’ inner core in these simulations. As discussed in chapter 3, the tilt of the large simulation (which is called the TC–trough simulation in chapter 3) is reduced due to downshear reformation, which is evidenced by the greater than 50% tilt reduction between 22 h and 24 h, and the vortex tilt magnitude becoming more stable around 28 h.

The results of the tilt magnitude traces (Figs. 4.11 and 4.12) indicate that larger tilt is linked to larger TCs. It is hypothesized that the larger, slightly longer-duration tilts in the AL and SLES TCs lead to greater moistening downtilt through persistent, shear-forced upward motion. Enhanced moistening downtilt could account for the larger expanse of precipitation between the band and the TC in the AL and SLES TCs. Greater moistening at early times downtilt could also account for the differences in reflectivity later in the simulation by controlling the propagation speed of the outward-propagating rainband. It is also hypothesized that the relative moistening downtilt in the AL and SLES TCs at early times could cause the slower propagation speed observed in their propagating bands due to the more moist air, through which the band propagates, causing less evaporative cooling and a weaker cold pool beneath the rainband compared to the AS TCs, which are propagating through a drier environment.
4.4.4 Moisture

Raymond (2000) and Raymond et al. (2009) showed that the likelihood of tropical convection are largely controlled by the saturation fraction of the atmospheric column. Figure 4.14 shows the composite saturation fraction of the AL/SLES and AS TCs at 12 h, 22 h, and 32 h. The composites of the AL/SLES and AS TCs were used because they produced a clearer difference signal than just the small and large TCs being examined in this section. Additionally, the tilts of the AL/SLES and AS TCs, which are hypothesized to affect saturation fraction, show high intra-cluster similarity (Figs. 4.11 and 4.12). The 12–32 h time range encompasses the time period during which the rainband propagates outward and large differences in reflectivity begin to appear (Fig. 4.9).

At all three times considered in Figure 4.14, the saturation fraction of the AL/SLES composite (Fig. 4.14b,e,h) is greater than in the AS composite (Fig. 4.14a,d,g) north and northeast of the TC, through where the rainbands propagate. Although the differences are small in Figure 4.14 due to compositing, time-averaging, and smoothing, they are still likely consequential, as values of saturation fraction northeast of the TCs in Figure 4.14 are in the 80% range. Raymond et al. (2009) showed that the greatest sensitivity between saturation fraction and precipitation rate is near 80%. Thus, even small differences in saturation fraction may result in large differences in the extent and amount of convection, given the high sensitivity.

The moist–dry dipole patterns near the TC centers in the raw saturation fraction fields support the hypothesis that differences in TC tilt are responsible for the greater moistening downtilt (northeast) in the AL/SLES TCs. The difference fields between the two composites (Fig. 4.14c,f,i) suggest that the larger tilt in the AL and SLES TCs (Fig. 4.14b,e,h), is causing greater moistening due to enhanced lift over a larger area downtilt. Figures 4.14g,h,i reveal that the differences in moisture persist even after the TC vortices have realigned, which would account for the differences in reflectivity observed that occur after the tilts in the TCs have reduced.
The greater dryness in the AS TCs, which have more rapidly propagating rainbands, to the northeast also provides a mechanism for that faster propagation. The increased dryness could cause greater evaporational cooling and stronger downdrafts as the rainband propagates toward the northeast, which could strengthen the cold pool below it, causing the rainband to propagate more quickly (Rotunno et al. 1988; Bryan et al. 2006).

Figure 4.15 shows soundings taken behind the propagating bands following their passage for the small and large TCs. The small TC sounding is drier overall, and shows significant low-level drying, indicative of evaporative cooling. The surface temperature is also approximately 1 K cooler than the large TC. Thus, the cold pool appears to be slightly stronger in the small TC, which could cause faster propagation. Additionally, the drying and capping behind the rainband would also inhibit precipitation behind it, which could be another cause of the greater dryness and lack of precipitation between the TC core and the rainband in the AS TCs.

Further confirming that the large TC rainband has a stronger cold pool, lowest model level potential temperature ($\theta$) is plotted in Figure 4.16a,b. The differences between the cold pools (Fig. 4.16c) associated with the rainband are apparent east of the TCs. At this time, $\theta$ is lower beneath the propagating rainband in the small TC, which is farther from the TC center, indicating a cooler cold pool and faster propagation speed than in the large TC. These results fit with the hypothesis that the faster propagation speed of the rainband in the small TC is due to a stronger cold pool.

4.4.5 Diabatic heating and surface fluxes

As a result of the greater amount of moisture and precipitation outside the inner core of the large TC, there are also differences in diabatic heating (DH) at large radii. One mechanism by which a TC can expand in size is when DH occurs outside of the RMW, which reduces the pressure at larger radii, resulting in a weakening of the pressure gradient (Wang 2009; Bu et al. 2014). The weakening pressure gradient can decrease the TC intensity
but increase the TC size (Wang 2009). Additionally, greater DH at large radii causes an increase in the import of relative angular momentum into the TC, which can increase the TC size (Xu and Wang 2010; Ma et al. 2014).

Figure 4.17 shows a radius-time Hovmöller of the column-integrated, mass-weighted, DH of the small and large TCs. In both the large and small TCs, the outward propagation of the rainband feature is evident starting around 10 h. The signature of the rainband is strong in the small TC, as it is the only organized convection outside of the inner core for a substantial period of time. In the large TC, there is more DH across most of the radii out to 500 km at nearly all times. Most of the differences in DH, especially in the 20–40 h range during which the size bifurcation occurs, are due to the differences in the convection north of the TC, where the large TC has a robust precipitation shield and the small TC does not (Fig. 4.9). The greater DH in the large TC outside of the TC core is a possible explanation for the expansion of the TC wind field in the large TC relative to the small TC (Wang 2009; Xu and Wang 2010; Bu et al. 2014; Chen et al. 2018).

Once the differences in size emerge around 32 h, non-linear feedbacks, such as surface fluxes, can act to further increase the size of the large TC relative to the small TC. Figure 4.18 shows a Hovmöller of azimuthally-averaged surface heat fluxes with time in the small and large TCs. In the large TC, surface fluxes outside the inner core begin to outpace the surface fluxes in the small TC around 28 h. Surface heat fluxes at large radii have been shown to increase TC size by creating more active convection outside of the TC core, which acts to increase the import of high relative angular momentum air into the TC, increasing the tangential winds (Xu and Wang 2010; Ma et al. 2014). Additionally, the increased surface flux contribution to increased DH outside of the TC inner core can also act to increase TC size through hydrostatic adjustment (Wang 2009). Thus, once the size of the large TC is larger than that of the small TC, a positive feedback loop of increasing surface fluxes and growing TC size causes the differences between the simulations to grow greater with time (Fig. 4.8).
4.4.6 PV

Because DH generates, or destroys, PV (Hoskins et al. 1985), the differences in DH between the small and large TCs also lead to differences in the PV fields, which ultimately affect the TC size. The larger and/or stronger the PV in the TC core is, the larger the wind field it induces (Hoskins et al. 1985).

Figure 4.19 shows a radius-time Hovmöller of PV for the small and large TCs. The differences in PV values indicate that the large TC has more radially extensive high-PV values. The large TC has generally greater PV at almost all times outside of the inner 30 km of the TC compared to the small TC.

The PV generation by DH was computed directly through the use of a PV budget following Wu and Wang (2000) and Ríos-Berríos (2017). The budget was calculated between 1000–500 hPa to capture the component (the vertical gradient) of the DH that most likely results in PV generation. As in Ríos-Berríos (2017), the budget was calculated at 6 min time intervals.

Figure 4.20 shows the average rate of PV generation due to DH in the large and small simulations. Both TCs, unsurprisingly, show the greatest average rate of PV generation due to DH in the TC inner core, where the strongest convection is located. Outside of the core, both TCs show areas with large average rates of PV generation extending from the core toward the northeast. These areas of large average PV generation resemble the locations of highest reflectivity in Figure 4.9, especially at the 42 h time period (Fig. 4.9e,f), after which time the TC rainband largely stays in the same general location (not shown).

Spatially, the differences between the two TCs’ average rate of PV generation by DH are mostly confined to two major areas. First, the region with the largest average rate of PV generation outside of the TC core is in the quasi-stationary rainbands northeast of the TCs. The position of the rainband is farther east in the small TC compared to the large TC, which creates the dipole structure in the PV generation differences northeast of the center (Fig. 4.20). In the rainband region, the large TC has a greater average rate of PV generation.
due to DH than the small TC, which also is consistent with differences in the strength and areal extent of high reflectivity in the rainband (Fig. 4.9e,f).

Second, north of the TC core, there is larger coverage of greater average rates of PV generation due to DH in the large TC, which has an extensive and long-lived precipitation shield. A cross section of PV in the large TC (Fig. 4.21), which is representative of PV-cross sections during the last 40 h of the simulation, indicates that in the precipitation shield region (∼200–400 km) of the TC, there is substantial PV between 800–400 hPa, likely due to the abundance of stratiform precipitation there (Fig. 4.9), which generates PV at midlevels (May and Holland 1999; Murthy and Boos 2019). Similar cross sections of PV in the small TC reveals that there is less coverage and smaller magnitudes of midlevel PV in the small TC (not shown), which is consistent with the smaller average rate of PV generation by DH (Fig. 4.20).

The presence of higher values of spatially extensive, midlevel PV in the larger TC raises the question of whether, and how, this PV can be advected in toward the TC’s core, where it can both increase the TC size and potentially affect intensity (May and Holland 1999; Leroux et al. 2013, 2016). The PV budget proved difficult to use to explore this question because of the sensitivity of budget terms to the domain choice. Instead, animations of planar views of midlevel PV are analyzed to trace subjectively-identified patches of PV. The disadvantage of this methodology is that, although it can be highly suggestive of how PV is behaving at midlevels, the Eulerian framework makes it difficult to prove that PV is accumulating in the TC core.

One hypothesized mechanism by which midlevel PV can act as a source for the TC core is through radial PV advection. May and Holland (1999) showed that there is substantial PV generation in midlevels in TC rainbands due to stratiform precipitation over a large area. Strong shearing deformation then acts to filament the midlevel PV produced in the rainband into a cyclonic band with an inward-curving component as a result of the radial gradient of the tangential velocity in the TC and the effects of trailing spiral geometry (Montgomery
and Kallenbach 1997; May and Holland 1999). May and Holland (1999) also found that a jet formed along this band, which was maximized on its outer periphery. The radial inflow along the high-PV band was hypothesized to transport midlevel PV into the TC core, thus acting as a source of PV for the TC.

To explore whether this mechanism is occurring in these simulations, midlevel PV and radial wind was analyzed (Fig. 4.22). In the small TC (Fig. 4.22a), the PV does appear to be converging into a band northwest of the TC and there is increased radial wind in the same region, but it is disorganized. This result is perhaps because there is much less PV outside of the inner core of the smaller TC, due to the lack of large-scale precipitation north of the center.

The large TC (Fig. 4.22b) has much more midlevel PV north of the center, consistent with Figure 4.21, and the PV coalesces into a band northwest of the TC that wraps around and into the TC core, as described in May and Holland (1999). Along this high-PV band, there is a well-organized, radial inflow jet with radial inflow speeds approaching 10 m s$^{-1}$ along the outer edge of the PV band.

May and Holland (1999) suggested that the midlevel PV generated in the rainband would filament, form a cyclonic band, and spiral inward due to shearing deformation, but they did not suggest what governed where this band would occur, other than that the initial PV would form in the rainband. The tight radial gradient of the high-PV band in Figure 4.22b suggests that there is a preferred location or boundary along which this PV band occurs.

It is hypothesized that the high-PV band forms at the boundary between the TC vortex and its environment. This boundary, called a separatrix, delineates, in a storm-relative framework, a closed, recirculating TC circulation that is mostly isolated from its environment (Dunkerton et al. 2009; Rutherford et al. 2018). As the midlevel PV north of the TC is advected cyclonically around the TC, it would be prevented from crossing the location of the separatrix, thereby resulting in the formation of a high-PV band as the PV
converges at the boundary, as shown in Figure 4.22b.

To determine if such a boundary exists in these simulations that could cause the midlevel PV convergence seen in Figure 4.22, the 500-hPa, storm-relative streamlines, which can provide an approximation of the separatrix, are shown in Figure 4.23. In the large TC (Fig. 4.23), the streamlines on the western side of the TC do suggest the presence of the separatrix, as there is a confluence axis between the vortex flow and the environmental flow along the location of the high-PV band. In the small TC (Fig. 4.23), the separatrix also exists along the high-PV band, but it is not as well-defined as in the large TC, just like the high-PV band itself.

In addition to the presence of the separatrix, the storm-relative streamlines indicate several other factors that may be causing the differences in size of the TCs and differences in the high-PV band. In the small TC, the recirculating region encompassing the inner core is not as large or coherent as in the large TC due to the weaker, less cyclonic flow north of the TC compared to the large TC, and the presence of another cyclonic circulation north of the small TC. The smaller, more poorly-defined recirculating region encompasses less of the area to the northeast of the TC where there is significant PV generation (Fig. 4.20), whereas the larger recirculating region in the large TC encompasses more of the rainband, allowing for the PV to be "gathered" and transported inward.

Animations of midlevel (500-hPa) PV in the large\(^4\) and small\(^5\) TCs show the midlevel PV evolutions of the two TCs and the formation of the high-PV band. As the stronger streamlines in Figure 4.23, which are representative of the strength of the streamlines throughout the latter half of the simulations, suggest, PV in the large TC rotates around the TC more quickly than in the small TC. The greater amount of PV in the large TC, combined with the stronger convergence along the separatrix, leads to a larger, stronger, longer-lived high-PV band.

\(^4\)http://www.atmos.albany.edu/student/cpeirano/Animations/wrf_trough_simulations/diss_runs/sensitivity_cube/vortex_125/scale_7/shear_10/Animation_6min_PV.html
\(^5\)http://www.atmos.albany.edu/student/cpeirano/Animations/wrf_trough_simulations/diss_runs/sensitivity_cube/vortex_125/scale_10/shear_10/Animation_6min_PV.html
Figure 4.24 follows a subjectively-identified patch of PV around the large TC. The patch starts at a large radii (350 km; Fig. 4.24a) and is distorted in shape by the shear of the tangential winds and due to mergers with other PV patches as it wraps around the TC (Fig. 4.24b). The patch of PV ultimately begins to converge in the high-PV band (Fig. 4.24c), where it then begins to move southward along the high-PV band into the TC core (Fig. 4.24d). The evolution (and the PV animation) strongly suggests that the PV in the high-PV band is axisymmetrized and wrapped into the TC core (Möller and Montgomery 2000; Enagonio and Montgomery 2001), but it is not possible to track individual PV maxima very far into this band due to the strong shearing deformation along the band and near the TC core.

In the small TC (Fig. 4.25), a similar evolution occurs, but, as mentioned, the advection around the TC is slower and there is less midlevel PV to advect. The patch of PV (Fig. 4.25a) begins at a similar radius as the patch in the large TC, though more east of the TC, but takes almost 13 h for the PV to reach the high-PV convergence band (Fig. 4.25b; compared to about 9 h in the large TC), indicative of the weak storm-relative flow seen in Figure 4.23. The separatrix does appear to form sooner in the small TC but it breaks down around 53–55 h (Fig. 4.25c), as suggested in Figure 4.23. The result of this breakdown is that the high-PV band separates and begins to move away from the TC (Fig. 4.25c); therefore, the patch of PV stays at large radius and is not able to wrap in toward the TC core (Fig. 4.25d), unlike patches of high PV in the large TC.

Thus, it is likely that the mechanism for midlevel PV generation and advection into the TC core proposed in May and Holland (1999) is occurring in these simulations, but with greater success in the large TC simulation due to the stronger, larger storm-relative vortex at midlevels, associated with the greater amount of PV at midlevels in that TC. The result of the consolidation of PV into the TC core would therefore be expected to increase the size and/or strength of the TC (May and Holland 1999; Leroux et al. 2013, 2016).
4.5 Comparison of representative SLES and AL TCs

This section will focus on a representative SLES and AL TC. The TC chosen to represent the SLES group is the scale 4, shear 12.5 m s\(^{-1}\) TC, and the TC that represents the AL TCs will be the scale 7, shear 12.5 m s\(^{-1}\) TC. These TCs, as for the AL and AS simulations selected for the first section, were chosen because they had the same initial shear value, which reduced the degrees of freedom.

4.5.1 Kinetic energy

To determine at what time the SLES TC diverges from the AL TC, the KE and KE difference was computed (Fig. 4.26). For the first 40 h of the simulation, the KE of the two simulations is similar. The similar values of KE in the SLES and AL TC at early times are consistent with their similar tilts (Fig. 4.11) and reflectivity (Fig. 4.7) for the first \(\sim\)40 h, which suggests that the differences in KE between them after 40 h are a result of different mechanisms than discussed in the previous section. Following 40 h, the KE difference of the AL TC begins to increase compared to the SLES TC. As discussed previously, the effect of the TC intensity on KE is largely removed by the choice of annulus excluding the inner 150 km of the TC but, regardless, the two TCs examined here have similar minimum pressures until around 50–55 h and similar maximum wind speeds throughout the simulation period.

4.5.2 200-hPa PV

Because of the large initial differences in the synoptic flow of the AL and SLES simulations (Fig. 4.2), the 200-hPa PV fields are examined to determine if the differences in TC size can be formed might be due to the differences in the evolution of the synoptic flow.

At 0 h (Figs. 4.27a, 4.28a), the AL simulation has a much larger and stronger trough compared to the SLES simulation. It is also evident at this time that the TC bogussing scheme used to initialize the TC acted to reduce the scale and magnitude of the initial trough in the SLES simulation between the TC and the trough core.

By 24 h, larger differences in the PV structure emerge as the high PV associated with
the trough in the SLES case (Fig. 4.27b) is highly deformed as it approaches the low-PV outflow of the TC. Part of the trough’s high PV fractures and moves toward the south, while the bulk of the high PV is advected north of the TC. At 48 h and 72 h (Fig. 4.27c,d) the high PV of the trough that was north of the TC is advected north and east of the TC. A weak, broad trough forms upstream at later times, as occurred in the TC–jet simulation. In fact, the synoptic PV field more closely resembles the TC–jet simulation at 48 h and 72 h than the scale 7 or scale 10 trough cases.

In contrast, at 24 h, the high PV associated with the trough in the AL case (Fig. 4.28b) is located west-northwest of the TC and is both stronger and more intact than the SLES case, though it is being deformed by the TC’s low-PV outflow. At 48 h and continuing to 72 h (Fig.4.28c), the trough takes on a negative tilt, as its progression is halted by the TC outflow. The trough also begins to wrap south of the TC as it is deformed, possibly due to the larger TC circulation in the AL case. As in the SLES simulation, the TC outflow forms a large, anticyclonic ridge downstream of the trough and TC. While the ridge is similar in scale to the SLES simulation, the larger upstream trough in the AL simulation results in a more amplified flow pattern around the TC.

As noted earlier, the PV evolution of the SLES simulation evolves somewhat similarly to the TC–jet simulation of chapter 3 during the last 48 h (during which time the TC size bifurcation between these TCs occurs), while the AL simulation evolves similarly to the control trough simulation throughout. In chapter 3, TC size was not explicitly explored, but there were large differences between the TC–jet and TC–trough simulation (Table 4.3). It was shown that there was larger DH outside the TC core in the TC–trough simulation (Fig. 3.20), which was likely responsible for its larger size. Therefore, it is hypothesized that similar physical mechanisms (i.e. midlevel moisture, Q-vector convergence) cause increased areal extent of DH in the AL TC relative to the SLES TC during the latter part of the simulations, leading to the increased size in the AL TC by the end of the simulation.
4.5.3 Reflectivity and wind

Twenty h prior to the bifurcation point (Fig. 4.29a,b), neither the reflectivity nor the extent of the 10 m s\(^{-1}\) contour are considerably different between the two TCs. Both have convection displaced toward the northeast, consistent with the shear and the large tilt they are experiencing (Fig. 4.11). By 40 h (Fig. 4.29c,d), the reflectivity field of the AL TC is broader in radial extent north and east of the TC compared to the SLES TC. The wind fields have begun to respond to these differences in convection, as the extents of the 10 and 15 m s\(^{-1}\) contours are both larger in the AL TC. An exception to the AL TC having generally more widespread convection is in the eastern part of the TC core, where the SLES TC has more convection. The core convection on the western side of the SLES TC, however, is weaker than in the AL TC.

At 60 h (Fig. 4.29e,f), the reflectivity patterns of the two TCs is more similar than at 40 h north of the TC, though differences remain. The precipitation shield north of the AL TC encompasses a larger area. The precipitation shield in the AL TC has wrapped slightly farther cyclonically around the TC between these two times, following a cyclonic shift in the shear direction (Fig. 4.30) as the trough wraps south of the TC (Fig. 4.28c). Additionally, there is substantially more scattered convection around the TC outside of the inner core and precipitation shield, and the differences are largest west and southwest of the TC. Both wind radii have expanded in the TCs, but the extent of the 10 m s\(^{-1}\) contour is much larger in the AL TC, extending beyond 600 km from the center in most directions, especially toward the northeast and southwest.

4.5.4 Midlevel moisture

In chapter 3, it was shown that differences in midlevel moisture resulted in differences in the strength and areal extent of precipitation. Therefore, the moisture fields are examined in these two cases to see if similar differences exist that could explain the precipitation differences shown in Figure 4.29.
Figure 4.30 shows the TC-centered, 700–500-hPa layer-average RH surrounding the bifurcation time in KE. At 20 h (Fig. 4.30a,b,c), the moisture fields are not substantially different between the two TCs. The signatures of the shear displacing convection to the northeast and the tilt forcing ascent downtilt (northeast) are apparent in the moisture dipole over the TC centers at this time.

By 40 h (Fig. 4.30d,e,f), the time of the KE bifurcation, there are large changes in midlevel RH. The AL TC is more moist compared to the SLES TC almost everywhere within 500 km of the TC center. The one exception is southeast of the TC core, where the SLES TC is more moist. This pattern fits well with reflectivity at this time, when the AL TC had greater coverage of precipitation over most of the domain, but had reduced convection compared to the SLES TC in the southeast portion of the TC core.

At 60 h (Fig. 4.30g,h,i), there are even larger differences in RH. East of the TCs, the AL TC has a band of greater midlevel moisture, although this difference in moisture is not well-represented in the reflectivity at this time (Fig. 4.29e,f). Even though the AL TC is significantly more moist east of the TC, both TCs have similar coverage of scattered convection at this time. West of the TC, both close to the core and farther away, there are very large differences in midlevel moisture (greater than 30% in some places southwest of the TC). These differences in moisture west of the TC may allow for the cyclonic rotation of the precipitation shield seen in the AL reflectivity (Fig. 4.29f) in addition to the rotating shear vector. Southwest of the TC, the greater moisture in the AL TC is likely responsible for the greater amount of scattered convection compared to the SLES TC.

Considering how the moisture in each respective TC changes in time (Fig. 4.31), the first 40 h (Fig. 4.31a–d) appear similar for the two TCs. The time differences in midlevel moisture between 20 h and 2 h for both TCs (Fig. 4.31a,b) show the effects of shear. The upshear/uptilt sides of the TCs have dried relative to 20 h prior, while the downshear/downtilt sides of the TCs have moistened.

Between 20 h and 40 h (Fig. 4.31c,d), the two TCs have moistened, particularly
upshear due to vortex realignment (Fig. 4.11) as the shear decreases, though the AL TC moistens more than the SLES TC in the western and southern TC core. The moistening of both TCs north and east of the centers is similar, though again the AL TC moistens more and over a slightly larger area. The greater areal extent of the reflectivity in the northeast portion of the domain in Figure 4.29d fits with the moistening in the same area seen in the AL TC at this time. There is drying southwest of both TCs, which is consistent with why there is only scattered convection in this region; however, the SLES TC dries out more.

A major change occurs between 40 h and 60 h southwest of the TC. The SLES TC continues to dry in the southwestern and western parts of the TC. In the same region, however, the AL TC moistens significantly, reversing the drying trend that occurred between 20 h and 40 h. The difference in moistening tendency in this region is consistent with there being more scattered convection southwest of the AL TC. Both TCs have moistened due west of the TC center, and the extent of core convection in both TCs has expanded on the west side, though it is slightly larger in the AL TC (Fig. 4.29e,f). The spatial extent of moistening on the west side is larger in the AL TC, which matches the RH field shown in Figure 4.30 and the greater precipitation on that side.

As in the prior section, the differences in TC size seem to be a reflection of the spatial extent of precipitation, as a result of midlevel moisture differences, which is consistent with Hill and Lackmann (2009). It was shown in the previous chapter that increases in midlevel moisture were driven by greater synoptic forcing in the TC–trough simulation relative to the TC–jet simulation due to the presence of the large, upstream trough. As the PV fields hinted at (Figs. 4.27 and 4.28), there are likely large differences in the forcing for ascent between these two simulations, which are explored next and could be responsible for the differences in midlevel moisture.
4.5.5 Quasi-nondivergent forcing for ascent

Forcing for ascent due to the synoptic, nondivergent flow is diagnosed here through \( \mathbf{Q} \)-vector divergence (Eq. 3.8), as in chapter 3, where \( \mathbf{Q} \)-vector convergence indicates forcing for ascent. At 20 h, although there is a clear difference in trough wavelength between the scale-7 trough in the AL simulation compared to the scale-4 trough in the SLES simulation, there is little difference in \( \mathbf{Q} \)-vector convergence magnitude between the two simulations near and over the TC center. As in chapter 3 (Fig. 3.30a,b), the magnitude of forcing for ascent early in both simulations is weak, which suggests that the temperature and/or wind speed gradients early in the simulations are relatively weak and strengthen with time. The forcing for ascent in the AL simulation covers a slightly larger area and is focused along and west of the TC, whereas the forcing for ascent is smaller in area in the SLES simulation and is mostly north and slightly downstream of the TC. The small differences in forcing at this time are consistent with the differences in midlevel moisture seen in Figure 4.30c.

By 40 h there are much greater differences in \( \mathbf{Q} \)-vector convergence. In the SLES TC (Fig. 4.32c), \( \mathbf{Q} \)-vector convergence has increased in magnitude, and is focused north of the TC. The flow pattern at this time more closely resembles the TC–jet simulation (Fig. 3.30c) compared to the larger-scale trough simulations and the \( \mathbf{Q} \)-vector convergence pattern is similar. Therefore, as in the TC–jet simulation, the \( \mathbf{Q} \)-vector convergence pattern seems to be more due to the jet to the north of the TC and weak, in-situ-forming, upstream trough, rather than to the initial scale-4 trough, which has been advected north and west of the TC by this time. The increase in forcing for ascent at this time matches the moistening north of the TC seen in the RH (Figs. 4.31c) and the increased areal extent of convection north of the TC (Figs. 4.29c).

In the AL simulation, \( \mathbf{Q} \)-vector convergence has also increased in magnitude compared to 20 h prior (Fig. 4.32d), and the forcing for ascent is also more widespread compared to the SLES simulation and the earlier period. The largest area of forcing for ascent is west and southwest of the TC, although there is also forcing for ascent north and south of the TC. At
this time, there is drying southwest of the AL TC in the same region where there is forcing for ascent (Figs. 4.30e and 4.31d). Drying in this area was seen in both TCs in chapter 3, as well as the SLES TC, which indicates that this region is favorable for descent, likely due to the divergent component of the TC outflow (which is excluded from the \( Q \)-vector forcing calculation) where it meets the synoptic flow, which has a significant westerly component there (not shown). Compared to the SLES case, the increased forcing for ascent may offset some of the drying in this region, as the AL TC dries less than the SLES TC west-southwest of the TC (Figs. 4.30d–f and 4.31c,d).

At 60 h, \( Q \)-vector forcing for ascent is strong west and southwest of the AL TC (Fig. 4.32f), and corresponds in time and space to the large amount of moistening west and southwest of the TC seen in Figures 4.30h and 4.31f, and the increase in scattered convection (Fig. 4.29f). This large increase in \( Q \)-vector forcing west of the AL TC corresponds in time to the reversal of the drying trend southwest of the TC seen in Figure 4.31d,f.

In the SLES TC, the magnitude of the forcing for ascent has increased from 40 h, but the area of largest \( Q \)-vector convergence remains northeast of the TC, in the right entrance region of the jet (Fig. 4.32e). There is weak forcing for ascent southwest of the SLES TC, potentially due to the slight troughing upstream. Considering the amount of drying occurring southwest of the SLES TC between 40–60-h (Fig. 4.31e), it seems the weak forcing for ascent is not strong enough to offset the drying.

The strong forcing for ascent across almost the entire western half of the 1000-km x 1000-km, TC-centered domain of the AL TC simulation due to the upstream trough (Fig. 4.32f), starting from around 40 h, matches the timing and location of the greatest moistening in the AL TC (Figs. 4.30h and 4.31f). This moistening corresponds to greater coverage of precipitation on the west and southwest sides of the TC (Fig. 4.29e,f). This increased convection provides a mechanism for TC size increase in the AL TC relative to the SLES TC, which lags behind in terms of size starting around 40 h.
4.5.6 Diabatic heating and surface fluxes

Considering the azimuthal average of DH at large radii, which would act to increase TC size, DH is larger in the AL TC compared to the SLES TC at most times at radii greater than 100 km (Fig. 4.33). The location of this heating, though, is not symmetrically distributed around the TCs, as the average rate of DH shows (Fig. 4.34). The greatest rates of DH in both TCs occurs in the inner core and downshear (northeast of the TC) in the rainband region. The areal extent of positive average rates of DH, as suggested by the DH Hovmöller, is greater in the AL TC at almost all azimuths.

Because the difference in forcing for ascent are most apparent on the western side of the AL TCs (Fig. 4.32d,f), the cumulative sum of mass-weighted DH was calculated in the western domain denoted in Figure 4.34. For the first 30 h of the simulation, DH is similar in both TCs, but, after that time, DH begins to increase in the AL TC relative to the SLES TC (Fig. 4.35). This increase is near the time that Q-vector forcing for ascent begins to differ more markedly between the two simulations and it slightly precedes the size bifurcation that occurs at 40 h.

A complicating factor is the breakdown of the western TC eyewall that occurs in the SLES TC between 32–40 h, which is when the bifurcation in DH occurs. Between these time periods, the eye and eyewall intermittently appear and disappear in the reflectivity field in the SLES simulation (not shown). This breakdown of the TC core convection in the SLES TC would contribute to the plateau in DH seen in the SLES DH trace (Fig. 4.35). The likely reason for the differences in eyewall convection at this time, in which the SLES TC is less coherent than the AL TC for a longer period, however, may relate back to differences in midlevel moisture and forcing for ascent. It was shown previously that there are large differences in midlevel moisture to the west of the TC near this time in the near-TC environment (Figure 4.30d,e) as a result of the enhanced lift due to the trough in the AL TC (Fig. 4.32d). Therefore, it is possible that the closer proximity of this drier environmental air to the TC core in the SLES simulation is causing dry-air entrainment into the TC, resulting
in the eyewall weakening on its western side.

Figure 4.36 shows that the likely cause of the eyewall collapse is, in fact, due to dry air intrusion in the midlevels on the SLES TC’s western side; additionally, animations of midlevel RH (not shown) suggest that this dry air is originating in the region of large RH differences to the west of the SLES and AL TCs. The exact mechanism by which this dry air intrudes into the eyewall, however, is not clear. One possibility is that the shear-induced tilt of the vortex (Fig. 4.12), induces a midlevel inward, eddy flux of this dry environmental air into the TC. The 0–5-km tilt at this time is not particularly large (approximately 20 km) but the close proximity of the dry environmental air to the tilted vortex could still allow this dry air intrusion (Alland 2019). Evaporatively-driven downdrafts could also cause low-level entrainment of low equivalent potential temperature air into the TC vortex (Riemer et al. 2010b).

After the SLES eyewall convection becomes more symmetric again around 45 h, the difference in DH on the west side of the TC continues to increase throughout the remainder of the simulation in both TCs (Fig. 4.35). The role of the trough in moistening the western side of the AL TC likely results in greater DH due to the greater coverage of scattered convection at large radii and to the moistening of the near-TC environment, which prevents the entrainment of dry environmental air into the TC’s core.

It should be noted that because an increase in TC size can both be caused by increased DH at large radii, as well as cause increased DH at large radii through enhanced surface fluxes, it is difficult to disentangle whether the DH differences at later times are due more to the differences in trough forcing or to the differences in TC size. It is reasonable to assume, however, that both factors are working in tandem.

On the east side of the TCs, where there is less trough forcing and coherent eyewalls in both simulations, the DH in both TCs increases steadily with time throughout the simulation (Fig. 4.37). Most of the convection and DH throughout the simulations is downshear, on the northeast side of the TCs (Figs. 4.29 and 4.34); thus, there is significantly more total
DH in both TCs on the east side (Fig. 4.37). Comparing the absolute differences between the west and east sides, however, reveals that the absolute difference in DH is approximately equal between the two sides (Figs. 4.35 and 4.37). Ergo, the two sides are approximately equally contributing to the DH differences. Therefore, the role of the trough in enhancing DH on the AL TC’s west side relative to the SLES TC is a significant factor in the TC’s size evolution.

As in the previous section, once the AL TC surpassed the SLES TC in size at 40 h, large differences in surface heat flux at large radii emerged (Fig. 4.38). The non-linear surface flux feedbacks acted to further cause the size expansion of the AL TC relative to the SLES TC through positive feedbacks with the DH at large radii.

### 4.6 Conclusions

In this chapter, the results of chapter 3 were expanded upon by using the same idealized modeling framework, but expanding it to include 27 simulations in a 3 x 3 x 3 parameter space, as well as the TC–jet simulation. The parameter space tested a range of initial TC shear values, trough scales, and initial vortex strengths.

While the simulation length was too short to draw any conclusions about how the aforementioned parameter space affected the steady-state intensities of the TCs, conclusions could be drawn about their favorability using intensification rate. The initial vortex strength was the most coherent signal in the sensitivity tests. Greater initial vortex strengths resulted in greater intensification rates. It was also found that greater initial shear, for the scale 7 and scale 10 simulations, resulted in greater intensification rates, though this result did not hold for the scale 4 simulations. Finally, concerning the trough scale, in general, the scale 7 simulations had higher intensification rates than the scale 10 simulations, but no patterns were found for the scale 4 simulations.

The size of the TCs also showed sensitivity to the parameters. As with TC intensification, the strongest sensitivity of TC size was to the initial vortex strength, where stronger
initial vortices were larger than their weaker initial vortex counterparts. Another strong signal was that, for the same trough scale, the TCs under greater initial shear values tended to become larger. There was not a clear pattern for differences in TC size with respect to trough scale; however, all of the 27 simulations with troughs had larger TCs than the TC–jet simulation.

Considering only the 12.5 m s\(^{-1}\) simulations, three groups of TCs were found: one group (AL) was large for the entire simulation period after about 30 h, one group (AS) was small during the same time period, and the final group began large but ultimately ended up smaller (SLES). Looking at the TCs that were initially in the large group between 30–40 h (AL and SLES) versus the TCs that were smaller (AS), the size of the TCs was largely found to be related to the areal extent and evolution of precipitation, particularly outside the inner core in the rainbands, during the first 35 h of the simulations. It was found that the areal extent of precipitation was related to the magnitude of the vortex tilt, due to the secondary circulation it induced. In the AL and SLES TCs, the vortex tilts were larger and persisted longer, which resulted in greater moistening over a larger area downtilt. This greater moistening not only allowed greater precipitation coverage north of the AL and SLES TCs, but it slowed the speed of the propagating rainband present in all of the TCs, such that it remained closer to the TC for a longer period, enhancing the size of the precipitation shield in the AL and SLES TCs. This greater coverage of precipitation increased DH, resulting in an increase in the TC winds and surface fluxes outside of the RMW. These effects resulted in a greater expansion of the AL and SLES TCs relative to the AS TCs.

In addition, a mechanism for size expansion at midlevels consistent with May and Holland (1999) was shown in the AL and SLES TCs. The greater amount of PV at midlevels in the large TC, as well as stronger flow and a larger TC-relative midlevel vortex, caused a band of high PV to form at the separatrix boundary of the midlevel vortex. This band of high PV was associated with a radial inflow jet that transported high PV into the TC core, where it axisymmetrized.
Considering the SLES and AL TCs, it was found that the synoptic flow in the SLES simulations evolved similarly to the TC–jet simulation in chapter 3. The lack of forcing for ascent due to a strong upstream trough in the SLES TCs resulted in a drier midlevel environment relative to the AL TCs, especially west of the TC. The drier air, and closer proximity of this drier air to the core, caused a disruption in the SLES TC eyewall midway through the simulation. The more moist air west of the TC in the AL TC resulted in more scattered convection and a stronger western eyewall, both of which contributed to large differences in DH on the western side of the TC, which led to size differences in the AL and SLES simulations.
4.7 Tables and figures

Table 4.1: Minimum central pressure (hPa) of each sensitivity experiment at the final time of the simulation (72 h). The minimum central pressure of the TC–jet simulation, which had a 12.5 m s\(^{-1}\) initial vortex, was 975.4 hPa.

10 m s\(^{-1}\) initial vortex

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15 m s\(^{-1}\) initial vortex

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Table 4.2: Azimuthally-averaged maximum 10-m wind speed (m s\(^{-1}\)) of each experiment at the final time of the simulation (72 h). The maximum wind speed of the TC–jet simulation, which had a 12.5 m s\(^{-1}\) initial vortex, was 31.0 m s\(^{-1}\).

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Table 4.3: Radius of the 15 m s\(^{-1}\) 10-m wind speed contour of each sensitivity experiment at the final time of the simulation (72 h). The radius of the 15 m s\(^{-1}\) wind speed contour in the TC–jet simulation, which had a 12.5 m s\(^{-1}\) initial vortex, was 145 km.

10 m s\(^{-1}\) initial vortex

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12.5 m s\(^{-1}\) initial vortex

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Figure 4.1: Latitudinal (top) and longitudinal (bottom) cross sections of the scale 4 (a,d), scale 7 (b,e), and scale 10 (c,f) trough temperature anomalies (K) relative to the background state of the model.
Figure 4.2: Initial, 200-hPa wind speed (m s$^{-1}$) of the scale 4 (a), scale 7 (b), and scale 10 (c) trough sensitivity experiments. The dots indicate the initial location of the TC in the shear 7.5 (red), shear 10 (orange), and shear 12.5 (yellow) sensitivity experiments.
Figure 4.3: Minimum central pressures (hPa) of the 12.5 m s$^{-1}$ initial vortex sensitivity experiments during the simulation period. Scale 0 denotes the TC–jet simulation.
Figure 4.4: Maximum, azimuthally-averaged, 10-m wind speed (m s$^{-1}$) of the 12.5 m s$^{-1}$ initial vortex sensitivity experiments during the simulation period.
Figure 4.5: Surface kinetic energy ($10^{-6}$ m$^2$ s$^{-2}$) for the 12.5 m s$^{-1}$ initial vortex sensitivity experiments. Surface kinetic energy was calculated within the 150–300-km annulus centered on the TC to account for differences in TC intensity.
Figure 4.6: Radial extent (km) of the azimuthally-averaged 15 m s$^{-1}$ wind contour for the 12.5 m s$^{-1}$ initial vortex sensitivity experiments.
Figure 4.7: Column maximum reflectivity (dBZ) and spatially-smoothed 10 m s\(^{-1}\) (red) and 15 m s\(^{-1}\) (pink) 10-m wind speed contours of the 12.5 m s\(^{-1}\) initial vortex sensitivity experiments at 35 h. Red dot denotes TC center.
Figure 4.8: Surface kinetic energy (*10^{-6} \text{ m}^2 \text{ s}^{-2}; \text{top}) and the difference in surface kinetic energy (*10^{-6} \text{ m}^2 \text{ s}^{-2}; \text{bottom}) for the representative large (scale 10, shear 10 m s^{-1}, vortex 12.5 m s^{-1}; blue) and representative small (scale 10, shear 10 m s^{-1}, vortex 12.5 m s^{-1}; red) TCs. Positive differences indicate that the large TC has higher surface kinetic energy. Surface kinetic energy was calculated within the 150–300-km annulus centered on the TC to account for differences in the TC intensities.
Figure 4.9: Column maximum reflectivity (dBZ) and spatially-smoothed 10 m s$^{-1}$ (red) and 15 m s$^{-1}$ (pink) 10-m wind speed contours of the small (left column) and large (right column) TCs at 22 h (a,b), 32 h (c,d), and 42 h (e,f). Red dot denotes TC center.
Figure 4.10: 0–300-km (top) and 200–800-km (bottom) vertical wind shear (m s\(^{-1}\)) of the small TC (red line) and large TC (blue line).
Figure 4.11: 0–5-km TC vortex tilt (km) for the 12.5 m s\(^{-1}\) initial vortex sensitivity experiments stratified by initial shear, where shear 7.5 is dotted (top), shear 10 is solid (middle), and shear 12.5 is dashed (bottom). The TC–jet simulation (scale 0, shear 7.5) is included in the top panel in green. The representative large and small TCs are denoted by the blue and red bold, solid lines, respectively.
Figure 4.12: 0–5-km TC vortex tilt (km) for the 12.5 m s\(^{-1}\) initial vortex sensitivity experiments stratified by trough scale, where scale 4 is black (top), scale 7 is blue (middle), and scale 10 is red (bottom). The TC–jet simulation (scale 0, shear 7.5) is included in the top panel in green. The representative large and small TCs are denoted by the blue and red bold, solid lines, respectively.
Figure 4.13: 0–5-km tilt distance (km) and tilt angle of the small (left; scale 10, shear 10 m s$^{-1}$, vortex 12.5 m s$^{-1}$) and large (right; scale 7, shear 10 m s$^{-1}$, vortex 12.5 m s$^{-1}$) TCs. Shading indicates simulation time (h).
Figure 4.14: TC–centered, 5-h time-average, surface–200-hPa, column saturation fraction (%) of the small-tilt TC composite (left) and large-tilt TC composite (middle), and the difference between them (right; where blue values indicate that the large-tilt TC composite is more moist) at 12 h (top), 22 h (middle), and 32 h (bottom). Red and black stars denote TC centers.
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Figure 4.38: Radius–time Hovmöller of surface heat flux (K h\(^{-1}\)) of the SLES TC (left) and AL TC (center) simulations, and the difference between them (right), where red values indicate that the AL TC simulation has larger surface heat flux.
5. Conclusions and future work

5.1 Conclusions

In this dissertation, the effects of troughs on TC intensity and structure were examined. In the North Atlantic, TC–trough interaction is a common occurrence and has been shown to result in intensity changes ranging from rapid decay to rapid intensification. Previous climatologies of TC–trough interactions disagreed on the fundamental question of whether TC–trough interactions are generally favorable or unfavorable for TC intensification. Additionally, case studies and previous modeling experiments differed significantly on which processes were most responsible for intensity change under trough forcing. Thus, there is a need to better understand: (1) whether trough interaction is typically favorable or unfavorable for TC intensification, and (2) the dominant mechanisms by which TCs and troughs interact in order to improve forecasts of TC intensity.

In the introduction, three major research questions were posed:

1. Accounting for sample bias, are troughs climatologically favorable or unfavorable for intensification compared to instances of no trough interaction and compared to the statistics of intensity change in the Atlantic basin as a whole?

2. Which processes are responsible for TC intensity change under upper-level trough forcing?

3. Are certain trough morphologies more favorable for TC intensification?

These questions were explored through reanalysis and observations (Chapter 2), as well as through idealized modeling (Chapters 3 and 4). The following subsections will summarize how the results of this dissertation contributed to answering each research question.
5.1.1 Question 1: Accounting for sample bias, are troughs climatologically favorable or unfavorable for intensification compared to instances of no trough interaction and compared to the statistics of intensity change in the Atlantic basin as a whole?

Controlling for sample biases and the climatology of the North Atlantic, TC–trough interactions are unfavorable for intensification relative to TCs that do not interact with troughs, as well as compared to the climatology of the basin.

5.1.2 Question 2: Which processes are responsible for TC intensity change under upper-level trough forcing?

In Chapter 2, analysis of TC–trough interactions using the ERA–I showed that the vertical wind shear is the dominant process determining the favorability of TC–trough interactions. Wind shear is significantly lower in the favorable trough interaction cases throughout almost the entire 48 h period surrounding the time of trough interaction. Though EFC was found to be favorable for intensification, the correlation between intensification and EFC was weak compared to shear. Divergence was found to be slightly higher immediately following favorable trough interactions. Moreover, enhanced TC convection in the favorable interactions was able to wrap upshear, which may have helped to erode the approaching trough and weaken shear; furthermore, convection wrapping upshear has been found to be favorable for TC intensification.

Because of the large temporal and spatial resolution of the ERA-I, idealized numerical modeling at higher temporal and spatial resolution was employed in Chapter 3 to better understand the physical mechanisms that may lead to TC intensification or weakening during TC–trough interactions. In that Chapter, an idealized TC–trough interaction simulation was compared to a TC–jet simulation.

Relative to the TC–jet simulation, the trough was found to be beneficial for TC intensification, as the TC in the TC–trough simulation was more intense from 20 h through the end of the simulation. It was shown that the near-TC environment of the TC–trough
simulation was significantly more moist than the TC–jet simulation. The more moist TC environment increased the likelihood of convective bursts near the TC center, which led to intensification in the TC-trough simulation.

The reason for the more favorable environment in the TC–trough interaction simulation was attributed to greater divergence and lift near the TC. These processes were the result of several mechanisms that were enhanced in the TC–trough simulation due to the presence of the upstream trough. First, enhanced EFC due to the trough was located in regions of low inertial stability (as a result of the presence of the trough itself), which was likely responsible for the TC’s stronger, more radially extensive, northern outflow channel. Second, the upstream trough and downstream ridge surrounding the TC supported lift near the TC due to the TC being located in the transition zone between subgeostrophic and supergeostrophic flow. Finally, forcing for ascent due to the interaction of the TC and synoptic flow was diagnosed via \( Q \)-vector divergence, which showed that lift was enhanced near the TC due to the presence of the trough upstream.

5.1.3 Question 3: Are certain trough morphologies more favorable for TC intensification?

In terms of trough morphology differences between favorable and unfavorable trough interactions, Chapter 2 showed that there are significant differences; troughs associated with favorable interactions were found to be shallower, weaker, and horizontally smaller than the unfavorable interaction troughs, consistent with weaker shear in favorable interactions. Additionally, PV advection by the irrotational wind showed that even though there is greater divergent wind leading up to the trough interaction in the weakening composite, the trough still more closely approaches the TC due to the trough’s greater strength and larger size. This result is consistent with the observed enhanced, persistent shear. Finally, the composite trough associated with unfavorable TC–trough interactions was significantly drier at midlevels. The drier trough resulted in greater ventilation of the TC vortex, and
this increased ventilation corresponded to weakening convection in the unfavorable satellite composite. The more moist trough in the favorable composite, conversely, permitted strengthening TC convection that was able to wrap upshear.

To better answer question 3, a 28-member idealized numerical modeling sensitivity experiment was conducted, and the results were presented in Chapter 4. The sensitivity experiment utilized a 3x3x3 parameter space (with the parameters being trough scale, initial TC vortex intensity, and initial shear / distance between the trough and TC) to understand the effects of different TC and trough configurations on TC and size (structure).

The clearest result in terms of TC intensification was that the intensification rate (based on 72-h pressure change) was greatest when the initial vortex strength was greater. Concerning the sensitivity to the initial shear, it was found that the greater the initial shear, the greater the intensification rate, though this result did not hold for the scale 4 trough simulations. The trough scale had the least coherent effect on TC intensification, but the scale 7 trough simulations were generally more favorable for intensification than the scale 10 simulations, but, again, the scale 4 simulations were quite varied.

The most apparent result of the sensitivity experiment was the large sensitivity of TC size within the parameter space, which was the focus of Chapter 4. As with intensification, the greater the initial TC vortex strength, the greater the size of the TC. TCs with greater initial shear values also tended to be larger. There was not a clear relationship between TC size and initial trough scale, however.

Considering only the TCs with 12.5 m s\(^{-1}\) initial vortices, there were three groups in terms of the size evolution. The always large (AL) group consisted of three TCs that were larger than the rest throughout the simulation time. The starts large but ends smaller (SLES) group was initially in the AL group for the first 40–50 h, before they began lagging behind the AL TCs in size, as measured by the surface KE. Finally, the remaining TCs comprised the always small (AS) group, which were smaller than the AL and SLES TCs beginning early in the simulation period, around 30 h.
An analysis of the AS TCs compared to the SLES/AL TCs found that during the first 30 h of the simulation, the AL/SLES TCs had significantly more vortex tilting than the AS TCs. The result of this enhanced tilting was that the AL/SLES TCs had more moisture over a large area downtilt, likely due to the induced secondary circulation of the vortex tilt. This moisture difference resulted in the AL/SLES TCs having significantly more precipitation northeast of the TCs throughout the simulations. Additionally, the more moist environment resulted in the slower movement of the outward-propagating band seen in the AL/SLES TCs, which resulted in the rainband staying closer to the AL/SLES TCs and contributing to the large extent of their rain shields. The enhanced diabatic heating in the expansive rain shields of the AL/SLES TCs led to the size expansion of those TCs relative to the AS TCs. A conceptual summary of these downtilt differences in moisture and precipitation is provided in Fig. 5.1.

Evidence for a midlevel pathway to TC size expansion, as documented in May and Holland (1999), was also found. In one of the AL TCs examined, there was more midlevel PV north of the TC than in an AS TC due to the greater coverage of precipitation. This PV was advected into a confluence band along the TC vortex’s separatrix, before a radial inflow jet along the band’s outer edge advected the PV into the TC core, where it axisymmetrized.

The difference in the AL and SLES TC sizes was found to be a result of the different initial morphologies of the upstream trough. The SLES TC had a smaller, weaker upstream trough that was advected downstream of the TC within the first 24 h, resulting in the synoptic flow during the last 48 h more closely resembling the TC–jet simulation. The AL TC had a larger trough that took on a negative tilt, but the trough remained upstream of the TC throughout the simulation. These synoptic differences caused greater Q-vector forcing for ascent on the western side of the AL TC in the latter half of the simulation, resulting in greater midlevel moistening and more DH than in the SLES TC. These differences in DH, and the concomitant nonlinear feedback between DH, surface heat fluxes, and TC size expansion, resulted in the observed TC size differences. A conceptual summary of these
differences in Q-vector forcing and their affects on moisture and convection are shown in Fig. 5.2.

One common theme between the reanalysis / observational results and the idealized modeling results was the importance of moisture in the development of the TC. In the reanalysis, the midlevel entropy deficit was found to be significantly larger in the weakening TCs, leading to greater ventilation and weaker convection. In Chapter 3, the modeling results showed that the presence of a trough upstream resulted in greater midlevel moistening, which created a more favorable environment for convection, and likely led to the intensification of the TC in the TC–trough simulation relative to the TC in the TC–jet simulation. In Chapter 4, greater moistening downtilt at early times in the simulation allowed TCs to grow larger than those with less moistening downtilt. Additionally, at later times, differences in the trough-induced forcing for ascent resulted in midlevel moisture differences that resulted in differences in the coverage of convection, ultimately leading to a larger TC in the simulation with greater trough forcing.

5.2 Future work

Given that the TC intensity and size were both highly sensitive to small differences in moisture throughout the TCs life cycle, more realistic moisture profiles within the simulated troughs should be explored in the idealized modeling framework. No moisture perturbation was added to the troughs in the modeling simulations, but, in nature, troughs are generally drier than the mean tropical environment. As shown in the midlevel entropy deficit differences in Chapter 2, the larger, drier trough in the unfavorable composite significantly weakened TC convection. The moistening of the near-TC environments due to the greater trough forcing in the sensitivity experiments shown in Chapters 3 and 4 may take longer, or be thwarted, if there is greater advection of dry air from the trough region into the TC's circulation.

In the TC–jet versus TC–trough simulation, the TC–trough simulation intensified due
to well-placed convective bursts near the TC center. These bursts were shown to be more likely in the TC–trough simulation because of the more favorable environment caused by the upstream trough. Given that these convective bursts have a stochastic component, no matter how much more favorable the environment is in one simulation, future work should use an ensemble of TC–jet and TC–trough simulations. An ensemble of simulations with slightly different initial conditions would reveal whether the differences in intensification in the two simulations examined are, in fact, the result of random differences, or whether the more moist environment of the TC–trough simulation does systematically favor increased core convection and intensification.

Future work should also include simulations of more mature TCs interacting with troughs. The differences in response to troughs in Chapter 2 versus Chapters 3 and 4, in addition to likely being due to differences in moisture associated with the troughs, could also be due to the differences in TC intensity during the trough interaction. In Chapter 2, TCs were generally mature with established convection and wind profiles. The TCs in the sensitivity experiments, however, were bogussed weak TC vortices that took time to develop convection and more realistic wind structure.

In the higher shear cases, where TCs were initialized close to the trough, the TC began immediately interacting with the trough, before convection was established. As shown in Chapter 4, the convective pattern was greatly influenced by the proximity of the trough / initial shear value. Therefore, utilizing a method to place a mature TC in the vicinity of a trough might result in a different response to trough forcing, possibly more in line with the results seen in Chapter 2.

Though the PV animations in Chapter 4 were highly suggestive of positive PV axisymmetrizing around the TC core and causing TC size expansion, the difficulty tracking specific patches of PV once they enter the high-PV band suggests that trajectories of air parcels would be useful to confirm that positive PV that originates north of the TC is axisymmetrizing and contributing to a growth of the circulation. Trajectories would also be
helpful in determining the origin of dry air ventilating the TCs, particularly in the case of
the representative SLES TC, in which dry-air entrainment appeared to cause a significant
disruption of the convection in the western eyewall, a contributing factor to the SLES TC’s
growth being temporarily stunted (Fig. 4.36). Finally, trajectories could illuminate whether
the upper-level frontal feature associated with inflow and a tongue of high PV (Fig. 3.27) is
contributing high PV to the TC core, which was shown to cause TC intensification in Leroux
et al. (2016).

Finally, the open question of why the scale-10, shear-10 TC is small throughout the
simulation period in the 10, 12.5, and 15 m s\(^{-1}\) initial vortex experiments, despite TCs
with comparable shear being larger, should be explored further. Given the sensitivity of
TC evolution in moderate shear (Zhang and Tao 2012), and that the size of the TCs in
the sensitivity simulations in Chapter 4 were sensitive to differences in their initial tilt, a
larger ensemble of simulations of the scale-10, shear-10 simulations (and other parameter
combinations) should be performed. A large ensemble would allow for a comparison of the
synoptic and TC–scale differences between scale-10, shear-10 TCs that are small and those,
if any, that are larger.
5.3 Figures

Figure 5.1: Height–radius conceptual model of the downtilt environments of the AL/SLES (a; large tilts) and AS (b; small tilts) TCs during the first half of the simulation. Blue shading denotes moisture, green shading denotes precipitation, the blue box denotes the cold pool associated with the propagating band, and the arrow denotes the relative speed of the propagating band.
Figure 5.2: Conceptual model of the location of forcing for ascent relative to the SLES (a) and AL (b) TCs. The black star denotes the TC center, black streamlines denote the 200-hPa synoptic flow, purple shading denotes location of largest \( Q \)-vector convergence in the TCs, red/blue shading highlights the areas of the largest differences between the two TCs in midlevel RH, where red (blue) denotes dry (moist) air, and white shading denotes amount of scattered convection (relative to the other TC) in the same area as the moisture differences.
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